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Entrance of El Zacatón, Tamaulipas, Mexico, a 319-m deep water-filled hypogenic shaft; see Gary and Sharp, pages 27-39 for details (photo courtesy of Stone Aerospace and Jose Antonio Soriano).

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PREFACE

In 1998, the National Cave and Karst Research Institute (NCKRI) was established by a U.S. Congressional mandate to facilitate and support collaborative cave and karst research, stewardship, education, and data consolidation. This publication marks the beginning of a new publication series, NCKRI Symposium, collections of peer-reviewed manuscripts focused on a common theme in cave and karst science.

In 2007, NCKRI published *Hypogene Speleogenesis: Hydrogeological and Morphometric Perspective* by Alexander B. Klimchouk. In 2008, NCKRI and the Edwards Aquifer Authority sponsored a topical session entitled “Hypogenic Karst: Shedding light on once poorly understood hydrologic and morphologic features” at the Geological Society of America annual meeting in Houston, Texas. This publication evolved out of that topical session. It highlights recent advances in hypogene cave and karst science, including some widely discussed and debated issues.

While speleogenesis has been viewed traditionally from an epigenic perspective, where subsurface processes are directly coupled with surface processes, research on hypogene karst is shedding light upon the importance of fluid interactions in confined and semi-confined hydrogeologic regimes. For several decades, hypogene karst has been associated with special fluid properties, such as thermal and sulfuric acid-rich waters, or associated with highly soluble evaporite rocks. Current studies show hypogene speleogenesis is not limited to special conditions of fluid and rock chemistry, but includes a broad range of intrastratal processes which are not directly coupled to surface processes. We are now beginning to recognize the importance of hypogene processes in a much broader sense, which this publication demonstrates.

Throughout this publication, examples of hypogene processes demonstrate the variability and breadth of hypogene cave and karst systems with respect to speleogenetic evolution. The first section focuses on general theories of hypogene speleogenesis. The second section provides detailed studies of hypogene processes at specific sites, either individual caves or limited areas. The third section investigates hypogene processes from a broader perspective and regional-scale. The final section discusses the relationship between natural resources and hypogene processes.

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PRINCIPAL CHARACTERISTICS OF HYPOGENE SPELEOGENESIS

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Abstract

Hypogenic and epigenic karst systems are regularly associated with different types, patterns and segments of flow systems, which are characterized by distinct hydrokinetic, chemical and thermal conditions. Epigenic karst systems, which had long been the focus of most karst/speleogenetic research, are predominantly local systems receiving recharge from the overlying or immediately adjacent surface. Hypogenic karst is associated with discharge regimes of regional or intermediate flow systems dominated by upward flow, although mixing with local systems is commonly involved. Hypogenic speleogenesis tends to operate over long time spans, continuously or intermittently. Its main characteristic is the lack of genetic relationship with groundwater recharge from the overlying or immediately adjacent surface. Hypogenic karst may not be expressed at the surface and is largely climate-independent.

Hypogenic speleogenesis involves the formation of solution-enlarged permeability structures by waters ascending through a cave-forming zone from below under leaky confined conditions. Vertical hydraulic communication across lithological boundaries and different porosity systems allows deeper groundwaters in regional or intermediate flow systems to interact with shallower and more local systems supporting various dissolution mechanisms. There is a specific hydrogeologic mechanism inherent in hypogenic transverse speleogenesis (restricted input/output), which slackens the positive flow-dissolution feedback and speleogenetic competition in initial flowpath networks characteristic of epigenic systems. This mechanism accounts for the more uniform and pervasive conduit development typical of hypogene karst.

Hypogenic caves are identified in various geological and tectonic settings, being formed by different dissolutional mechanisms operating in various lithologies. Despite these variations, resultant caves demonstrate remarkable similarity in patterns and mesomorphology, strongly suggesting that the type of flow system is the primary control. Hypogenic caves commonly demonstrate a characteristic suite of cave morphologies resulting from rising flow across the cave-forming zone, with distinct buoyancy dissolution

components. Cave patterns in hypogenic speleogenesis are guided by the initial permeability structure, its vertical heterogeneities (discordance in the permeability structure between adjacent beds) and the mode of water input to, and output from, the cave-forming zone. The latter again depends on relationships between permeability structures in the cave-forming zone and formations that lie below and above. Because of its "transversal" nature, hypogenic speleogenesis has a clustered distribution in plan view although clusters may merge and extend through large areas.

Recognition of the wide occurrence, importance and specific characteristics of hypogenic speleogenesis represents a major paradigm shift in karst science that answers many questions not satisfactorily addressed previously.

Principal environments for speleogenesis and karst development

Three basic genetic settings are broadly recognized now for dissolution caves (Klimchouk et al., 2000; Gunn, 2004; Ford, 2006; Ford and Williams, 2007; Palmer 2007): 1) syngenetic/eogenetic (coastal and oceanic), which occurs in young rocks with high matrix porosity and permeability where caves are formed in the zone of mixing of marine and meteoric water; 2) hypogenic, predominantly confined (semi-confined), which occurs when water enters a soluble formation from below, and 3) hypergenic (epigenic), which occurs under unconfined conditions, where water is recharged from the overlying surface. The order of how these categories are listed above broadly correlates to their normal evolutionary sequence from the perspective of the geological, diagenetic and hydrogeologic evolution of a bedrock formation (Klimchouk and Ford, 2000). This creates potential for some inheritance of solution permeability structures from the earlier stages to the later ones. Syngenetic/eogenetic speleogenesis in coastal and oceanic settings, although also created by unconfined circulation (thus falling into the realm of epigenic karst in the hydrogeological sense), results in distinct porosity patterns because of the specific conditions determined by the dissolution of porous, poorly indurated carbonates by mixing of waters of contrasting chemistry at

the halocline. This is why it is commonly distinguished as a separate category.

Epigenic karst systems have historically been the focus of karst research. Most concepts and models explaining karst porosity invoke epigenic settings with descending and lateral (phreatic zone) groundwater circulation; they dominate the current karst paradigm. Epigenic karst systems are predominantly local systems or parts of recharge segments of intermediate and regional systems, receiving recharge from the overlying or immediately adjacent surface. Aggressiveness of water with regard to carbonate rocks is mainly acquired in the soil zone. In terms of hydrogeology, epigenic systems are largely unconfined. Flow and cave development in epigenic settings are predominantly lateral, from a recharge boundary to an outlet (Figure 1A), although considerable vertical components may exist where a thick vadose zone is present due to great local relief such as in high mountains. Epigenic speleogenesis is directly related to the contemporary surface topography and commonly results in hierarchical dendritic conduit systems, a pattern which never develops in hypogene settings.

The idea that some caves could form at depth by rising thermal waters had been introduced in the mid-19th century (e.g. Desnoyers, 1845; see also Bosak, 2000). However, the concept was established only during the second half of the 20th century with regard to hydrothermal karst. The term and concept of hypogenic speleogenesis has seen increasing use during the past two decades, although limited to dissolution mechanisms peculiar to cooling thermal waters and waters containing hydrogen sulfide (Ford and Williams, 1989; Palmer, 1991; Hill, 2000). The current burst of interest in hypogene speleogenesis is driven by 1) the establishment of a hydrogeological rather than a geochemical approach to its definition, which highlights the common hydrogeological genetic background and similarity of caves previously seen as unrelated, thus broadening the family of hypogenic caves; 2) the appearance of many detailed regional studies, which demonstrated that hypogene speleo-

genesis is much more common in nature than previously thought; and 3) rapidly growing recognition of the great scientific and practical significance of hypogene speleogenesis (Klimchouk, 2007).

Hypogenic karst is associated with discharge regimes of regional or intermediate flow systems dominated by upward flow, although mixing with local systems is commonly involved. Hypogene speleogenesis is the formation of solution-enlarged permeability structures by waters ascending through a cave-forming zone from below, driven by hydrostatic pressure or other sources of energy, independent of recharge from the overlying or immediately adjacent surface (Ford, 2006; Klimchouk, 2007). This definition, in contrast to previous views of hypogenic speleogenesis as a specifically geochemical phenomenon, is largely hydrogeological: it refers to the source of fluid recharge to the cave-forming zone, and to the type of flow system. The term “hypogenic” here does not necessarily mean the occurrence at great depth but, rather, refers to the origin of the cave-forming fluids from depth. The hydrogeological approach to the definition of hypogene speleogenesis places hypogenic karst in the systematized context and hierarchical structure of regional groundwater flow (sensu Tóth, 1999) and highlights the powerful role of speleogenesis in the organization of regional flow systems, a consequence of its unique capacity to dramatically alter the primary porosity and permeability of soluble formations.

Hypogene speleogenesis occurs in various geological and tectonic settings, at different depths (ranging from a few tens of meters to several kilometers), due to different dissolutional mechanisms operating in various lithologies. Thus, hypogene speleogenesis is not a single physical or chemical process but rather is a multitude of processes, or a complex geological process. The unifying factor is the type of flow system where water rises across a sedimentary sequence that includes soluble rocks. Because of the inherent vertical heterogeneity of sedimentary sequences, an upwelling flow system implies a certain degree of

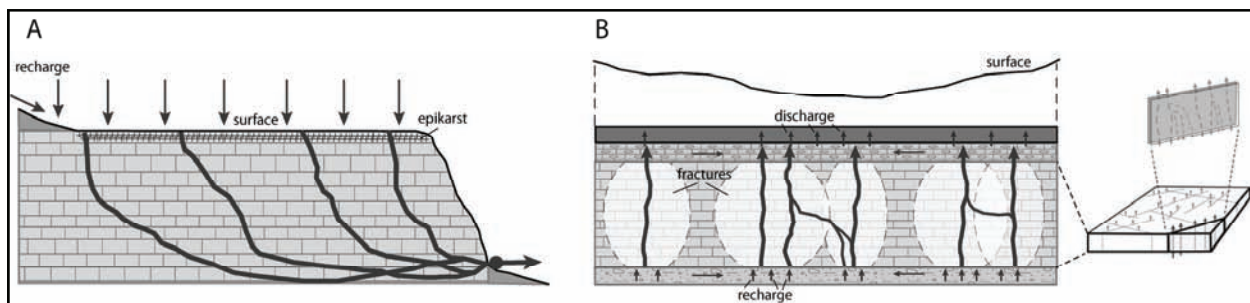


Figure 1. Conceptual representation of epigene (A) versus hypogene (B) speleogenesis.

hydrogeological confinement. Vertical hydraulic communication across lithological boundaries and different porosity systems allows deeper groundwaters in regional or intermediate flow systems to interact with shallower and more local systems, supporting multiple dissolution mechanisms that differ in their physical and chemical parameters.

Where hypogenic caves are shifted to the shallower, unconfined situation due to uplift and denudation but continue receiving upwelling recharge from deep systems, this would still be classified as hypogenic development, although unconfined. Unconfined hypogene development can be considered as the extinction stage of hypogene speleogenesis.

Transverse nature of hypogene speleogenesis and the role of vertical heterogeneities

In a stratified sedimentary succession (multi-aquifer system), soluble units are commonly vertically continuous with non-soluble or less soluble units of initially higher permeability. Due to their low matrix permeability, indurated soluble rocks often serve as separating beds (aquitards) prior to speleogenesis and during its early stage. Flow and speleogenesis is predominantly vertical (transverse; Figure 1B), although considerable lateral components can develop within more transmissible beds. Hypogenic caves evolve to facilitate cross-formational hydraulic communication across initially less pervious soluble beds. The overall result is the switching of the hydrostratigraphic function of the soluble unit from a separating aquitard to an aquifer that can be more prominent than the adjoining non-soluble aquifer units.

In hypogenic transverse speleogenesis, input and output of water to/from the cave-forming zone occurs through underlying and overlying insoluble or less soluble formations and is determined by the conductivity of the least permeable member. Thus there is an external conservative hydraulic control on the amount of flow through the evolving conduits. When conduits have evolved (i.e. after kinetic breakthrough), the flow across the cave-forming unit is limited not by the hydraulic resistance of the conduit system but by the permeability of the least permeable member and by the boundary conditions of the flow system. This suppresses the positive flow-dissolution feedback and speleogenetic competition, the main mechanism acting in unconfined (epigenic) speleogenesis, and promotes more pervasive enlargement of initial permeability structures. Conditions and manifestations of this effect depend on different variables that have been studied by numerical modeling of speleogenesis

in a gypsum bed under artesian hypogenic settings (Birk, 2002; Birk et al., 2003, 2005; Rehl et al., 2008, 2009). In hydrothermal speleogenesis, in carbonates there is another specific flow-temperature controlled mechanism that suppresses speleogenetic competition, caused by the thermal coupling between the fluid and rock (Andre and Rajaram, 2005).

Forced-flow regimes in confined settings commonly have very low velocities, and water with lesser density enters the cave-forming zone from below. This creates buoyancy driven flow patterns powered by either solute or thermal density gradients. Various morphological effects of free convection are recognized in hypogenic caves, among which upward-pointed dissolution morphs are most common (Klimchouk, 1997; 2000a, 2007). Speleogenesis at the base of the soluble unit due to buoyancy dissolution may operate even without forced hydraulic communication across the unit. More common, however, are mixed convection systems, where buoyancy dissolution effects are particularly pronounced during the mature stage of speleogenesis when vertical hydraulic gradients across the cave-forming zone diminish.

Vertical heterogeneity in a sedimentary sequence is normally greater than lateral heterogeneity within individual units. This holds true not only with regard to distinctly different lithologies, but within overall relatively uniform lithological divisions, such as bedded carbonate or gypsum formations. Geological variations in individual lithological units such as diagenetic processes, grain size, mineralogy, texture, structure and porosity determine hydrogeological properties of the matrix. Through their control over rock mechanical properties, these variations influence the structure and distribution of fracture networks in 3D space, confining them to certain horizons (mechanical units; Cooke et al., 2006). Therefore the distribution of the permeability structures of pore, fracture and mixed types relates to lithostratigraphy and so does the flow architecture. Studies that exemplify this are abundant in the modern hydrogeological and sedimentological literature; see Budd and Vacher (2004) for stratigraphic heterogeneity in pore type aquifers and Cooke et al. (2006) for stratigraphic heterogeneity in fracture type aquifers.

One or more lithostratigraphic units may comprise a distinct permeability structure (a hydrostratigraphic unit, or hydrofacies sensu Eaton, 2006) which is relatively independent of other units. Contrasts and discordance between permeability structures of different hydrostratigraphic units are common so that vertical hydraulic connection between them is varied

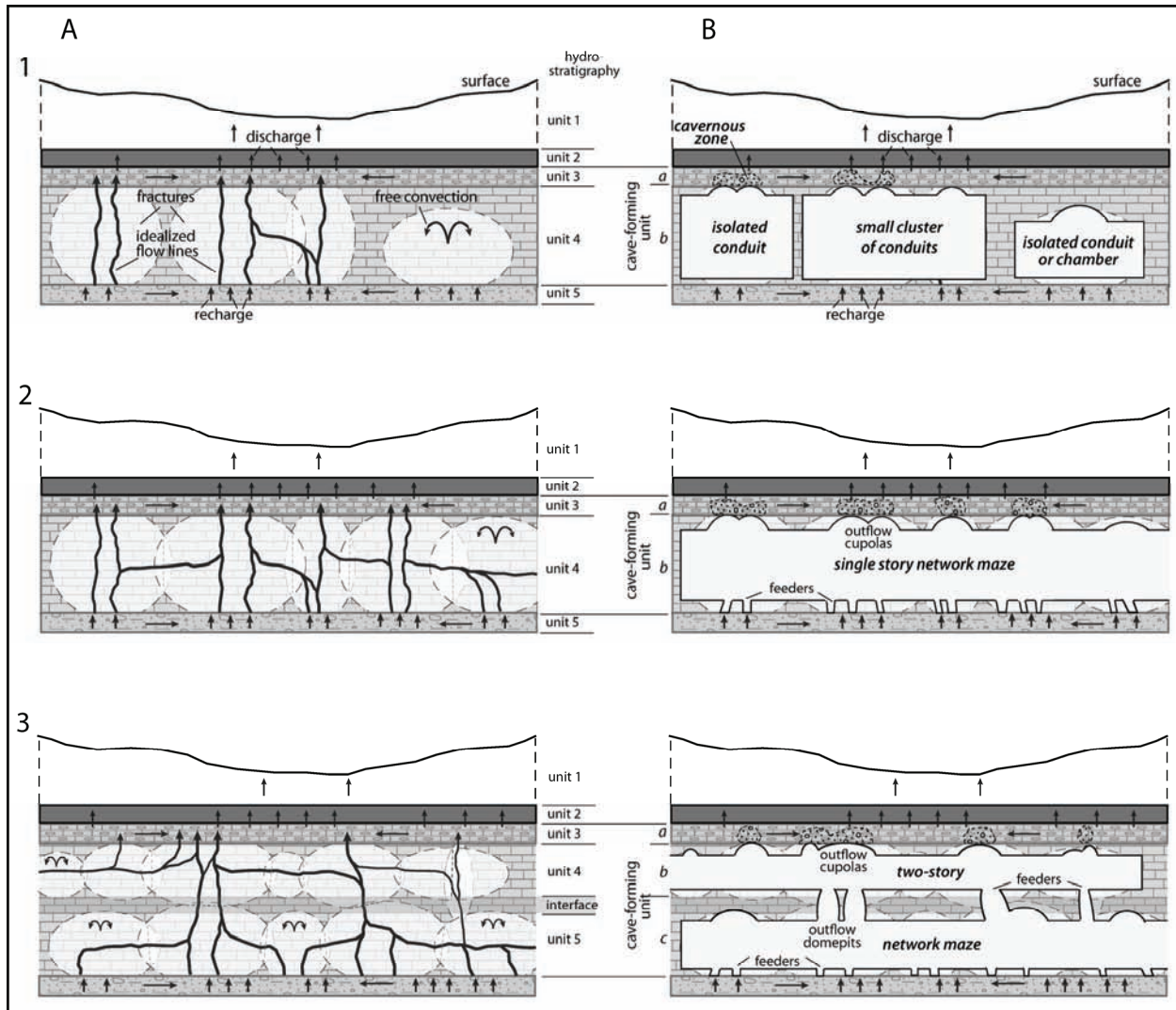


Figure 2. Some variants of hydrogeological settings of hypogene transverse speleogenesis (A) and their translation into cave patterns (B). See text for explanations.

and often limited even if units of relatively high lateral bulk permeability are immediately adjoining (hydrostratigraphic interface). Leakage through the hydrostratigraphic interface occurs via a limited number of points where permeability features of adjacent units vertically intersect (Figure 2, A3). Between units of higher bulk permeability, horizons of contrastingly lower bulk permeability often intervene, comprising vertical permeability barriers. Such barriers can be variously breached by cross-cut permeability features (e.g. prominent joints and faults). Thus, systematic heterogeneities in vertical conductivity are the inherent condition of transverse flow and hypogene speleogenesis.

Ascending transverse flow across such vertically heterogeneous sequences utilizes various kinds of

original (pre-speleogenetic) porosity in different hydrostratigraphic units: pores, touching-vug porosity, bedding planes and fractures. Limited vertical connectivity across the sequence caused by hydrostratigraphic interfaces and barriers redirects overall upwelling flow along relatively high permeability units, adding considerable lateral components to flow architecture. The resultant structure of the flow domain in leaky confined systems can be remarkably complex, especially where the original stratigraphic structure is deformed and complicated by cross-cutting permeability components such as faults. Adding to the complexity are zones of preferential recharge at depth that are often laterally shifted with regard to zones where the entire flow system discharges to the surface. Speleogenetic development of the initial porosity structures occurs in certain seg-

ments of the flow domain (cave-forming units), depending on the interplay of hydraulic, thermobaric and geochemical factors, all varying in the course of hydrogeological and speleogenetic evolution.

The complexity described above is reflected in the multiplicity of elementary cave patterns that occur in hypogenic settings. Figure 2 illustrates some relatively simple variants of hydrostratigraphic and structural settings of hypogene speleogenesis and how these translate into conduit patterns. Case 1 depicts a soluble bed in which fractures span its entire thickness but the fracture network is highly discontinuous (unit 4). Isolated passages or chambers and small clusters of passages develop in this bed due to transverse speleogenesis. Case 2 shows the same lithostratigraphic framework but fractures in the soluble bed are laterally well connected. This favors a lateral flow component to develop through the fracture network and may result in the formation of a single-story network maze. In case 3, continuous largely independent fracture networks are encased in two distinct horizons in the overall uniform lithostratigraphic unit, dividing it into two hydrostratigraphic units with a leaky interface in-between. The resultant cave pattern is a two-story network maze. In the overlying unit with predominantly diffuse porosity and permeability, cavernous zones may develop above outflow features of the cave systems (around occasional fractures or other flow-conducting structures that disrupt this unit).

The diagrams in Figure 2 are generic and elastic; they can be stretched vertically and further hydrostratigraphic and structural heterogeneity can be added. Complex 3D cave patterns, such as multi-story mazes in the Black Hills, South Dakota, USA and in the Western Ukraine, or still more vertically extended structures in variably deformed settings, comprised of large chambers, stratigraphically-conformable or disconformable clusters of network or spongework mazes at different levels, and sub-vertical conduits and other morphs connecting them, such as caves in the Guadalupe Mountains, New Mexico, USA and caves in Central Italy illustrate well the role of vertical heterogeneities in determining cave patterns.

Modeling of hypogene speleogenesis

During the recent two decades, the modeling approach in speleogenetic studies has greatly advanced our understanding of the development and evolution of solution porosity (see Dreybrodt et al. 2005 for a comprehensive overview). However, the great majority of numerical models imply that conduit development is based on lateral flow along fractures from a recharge boundary at the surface to an outlet, i.e.

essentially unconfined epigene settings. These models demonstrated that early cave genesis is governed by a positive-feedback mechanism involving the mutual enhancement of flow rate and solutational conduit enlargement. Modeling of speleogenesis in hypogene settings is still at its beginning, although several works attempted so far have been useful to gain a more detailed insight into the mechanisms of hypogene speleogenesis and demonstrate important differences between epigene and hypogene speleogenesis. In particular, they confirmed an assumption of the early conceptual models that the selectivity in conduit development in hypogenic settings is suppressed under certain conditions.

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

Several modeling studies have been performed to investigate hypogene speleogenesis under artesian conditions, based on settings in which conduit development is driven by cross-formational flow between aquifers across a heterogeneous soluble formation beneath topographical lows, where the vertical hydraulic gradient is maximized. As these settings are typified by the gypsum karst of the Western Ukraine, dissolution of gypsum was the key dissolution mechanism considered. Birk et al. (2003, 2005) translated the conceptual model of transverse speleogenesis in confined settings into generic numerical model scenarios. The model combines a coupled continuum-pipe flow model, representing both diffuse-flow and

conduit-flow components of karst aquifers, with a dissolution-transport model calculating dissolution rates and corresponding widening of karst conduits. These studies revealed, in particular, that maze cave development is favored by the presence of systematic heterogeneities in vertical conductivity of a fracture network. In addition to structural preferences, the variation of boundary conditions in time, e.g. increasing hydraulic gradient across the soluble unit due to river incision into the upper confining bed, further influences the development of maze patterns.

More site-related models that are closer to the actual settings found in the Western Ukraine were considered by Rehr et al. (2008, 2009). These models included various characteristic features concerning hydraulic boundary conditions as well as hydraulic properties of both aquifers and soluble units to investigate how they control the geometry of cave patterns. The models demonstrated effects of the spatial extension of discharge areas, the hydraulic conductivity of the rock units, the chemical saturation of water and the variability of initial apertures of proto-conduits on evolving cave patterns. The model scenarios with reduced hydraulic conductivities of all units in the system suggest that permeability of the entire rock formation is a crucial factor that controls the frequency distribution of conduit diameters in hypogene speleogenesis. If permeabilities are sufficiently low, the aperture distribution will be characterized by a smooth transition from nearly undeveloped, narrow protoconduits to well-developed, large-diameter conduits. The aperture variability is determining the temporal development of cave patterns and generally decelerates the karstification process, but appears to be of minor relevance regarding the general structure and geometric properties of evolving cave patterns. With an increasing degree of heterogeneity the number of conduits developed after a given time period decreases. In the long-term, however, differences in initial heterogeneity appear to be overridden and conduit systems approach similar patterns (Rehr et al. (2008, 2009).

The numerical models realized so far simulate only a few of a number of settings characteristic for hypogene speleogenesis. Even for these settings, the models are still highly simplified. For instance, the type of vertical heterogeneity caused by the discordance of fracture networks between different units of the soluble bed have been mimicked using different frequencies of vertical protoconduits in different horizons of the two dimensional model. Thus future work will have to develop more complex models, e.g., three-dimensional models explicitly representing the

observed 3D fracture networks. In addition, more advanced models should be developed to address the effects of density driven flow (free convection), which is of tremendous importance in hypogenic speleogenesis. It is generally assumed so far that these effects are particularly important at the late stage of speleogenesis when conduit apertures are substantial and hydraulic gradients are low. However, some recent studies demonstrate the pronounced role of buoyancy in rather small aperture fractures (Dijk and Berkowitz, 2000, 2002).

Solution porosity patterns produced by hypogene speleogenesis

Cave patterns in hypogenic speleogenesis are the result of a complex interplay of structural, hydraulic and geochemical conditions, all varying in the course of geological evolution. Patterns are strongly guided by vertical heterogeneity in the initial permeability structures of a sedimentary sequence, spatial distribution of these structures within cave-forming and adjacent units, mode of water input to, and output from, cave-forming units and the overall recharge-discharge configuration in the system. Modes of recharge and discharge, again, depend on relationships between permeability structures within cave-forming units and units that lie below and above. The presence of cross-cutting permeability features, such as faults, can exert a strong effect over cave patterns. Geochemical interaction of flow components guided by transverse and lateral permeability pathways determines zones of pronounced speleogenetic development and influences resultant patterns. General evolution factors such as regional tectonic and geomorphic development that change flow architecture and conditions, as well as timing of speleogenesis, also affect cave patterns forming in hypogenic settings. In contrast to epigene settings where initial effective permeability structures are exploited by speleogenesis in a very selective manner, hypogenic speleogenesis tends to exploit most of these permeability structures within cave-forming zones.

In this paper, only a brief outline of the variety of hypogenic cave patterns is given. For more extended discussion the reader is referred to Palmer (1991, 1995, 2000, 2007), Ford and Williams (2007) and Klimchouk (2000, 2007, 2008).

The following elementary cave patterns are typical for hypogenic speleogenesis:

- Zones of cavernous porosity
- Network maze
- Spongework maze

- Isolated passages or small clusters of passages
- Irregular isolated chambers
- Rising steeply-inclined passages or shafts
- Collapse shafts over large hypogenic voids

Laterally extended (multiple-story) or vertically extended composite 3D patterns are representative. It is difficult to say whether these composite patterns are the most common for hypogene speleogenesis or are just most prominently represented amongst the hypogene caves that have been explored, because they are exemplified by many of the world's largest cave systems. Among maze patterns, stratigraphically-conformable rectilinear network mazes predominate, often arranged in multiple (up to 5-6) storeys connected via rising shafts or cross-cutting prominent rift-like passages. Spongework mazes are less frequent. Vertically extended 3D patterns comprise various elementary patterns at different levels such as chambers, clusters of network or spongework mazes, and sub-vertical conduits and other morphs connecting them. A variant of composite 3D patterns is distinguished as a ramiform (ramifying) pattern composed of irregular rooms with various branches that extend outward from a central area (Palmer, 1991, 2000).

Probably the most common cave structures for hypogene speleogenesis are single isolated passages and chambers, or small clusters of a few intersecting passages. Also common, although less appreciated in cave science traditionally focused on larger voids, are zones of cavernous porosity. Such zones are often clustered around transverse permeability features such as prominent fractures, but are constrained to certain lithostratigraphic units that favor diffuse lateral flow. Such zones are apparently related to dissolution due to mixing of deep and shallower flow components.

Large isolated chambers commonly occur at the bottoms of cave-forming units. Their development is typically associated with deep hydrothermal systems or with dissolution of evaporites from below and greatly promoted by free convection dissolution effects. Rising, steeply-inclined passages or shafts are outlets of deep hypogenic systems in which the "root" structure remains unknown in most cases. Collapse shafts (including breccia pipes or 'geologic organs', and perhaps surface expression as sinkholes) often occur over large hypogenic voids.

Because of its "transverse" nature, hypogene solution porosity tends to have a clustered distribution in plan view although clusters may merge and extend through large areas. For instance, laterally extensive multi-story maze caves such as in the Western Ukraine or

the Black Hills, or vertically extensive 3D structures such as in the Guadalupe Mountains, are in fact combinations of many clusters of passages representing relatively independent sub-systems of transverse flow.

Meso-morphology of hypogenic caves

Hypogenic caves can be formed by a number of dissolution mechanisms, occur in various geological and structural conditions and may have different patterns. Despite this variability, meso-morphological features of hypogenic caves exhibit remarkable similarity, generally indicating sluggish flow conditions. Individual occurrences of some of these features are also found in epigenic caves where they form in phreatic or subaerial conditions. Specific for hypogene speleogenesis, however, is that in hypogenic caves they commonly occur in related suites where fluid flow paths, including distinct density flow components, can be visually traced from rising inlet conduits and vents ("feeders", or "risers"), through transitional wall and ceiling features (rising wall channels and ceiling half-tubes), to outlet features (cupolas and domepits). This particular combination has been termed the morphologic suite of rising flow (MSRF) and provides diagnostic evidence for hypogene speleogenesis (Figure 3). An extensive discussion of the morphology of hypogenic caves is provided by Klimchouk (2007, 2008).

On Figure 3, the geometry of a cave segment, the relative scale of features and hydrostratigraphy is directly representative for Ozerna Cave in Western Ukraine, the second greatest gypsum maze cave in the world, but this regular combination of forms is found in hundreds of hypogenic caves across the globe. Similar to Figure 2, this diagram is also generic and elastic; it can be stretched vertically, and a complexity can be added to account for multiple stories. The arrangement of the forms will repeat itself on each story, and functional relationships between the forms will remain the same. Hypogene caves may consist of a few elementary segments like this, or combine hundreds and thousands of them within a single system.

Distribution of hypogene speleogenesis

Recent compilation and overview by Klimchouk (2007) made it apparent that hypogenic speleogenesis is much more widespread in nature than previously thought. It has stimulated an intense re-evaluation of the origin of many caves and previously accepted regional paleohydrogeological and karstological concepts according to the new understanding of hypogene speleogenesis and methodology of its

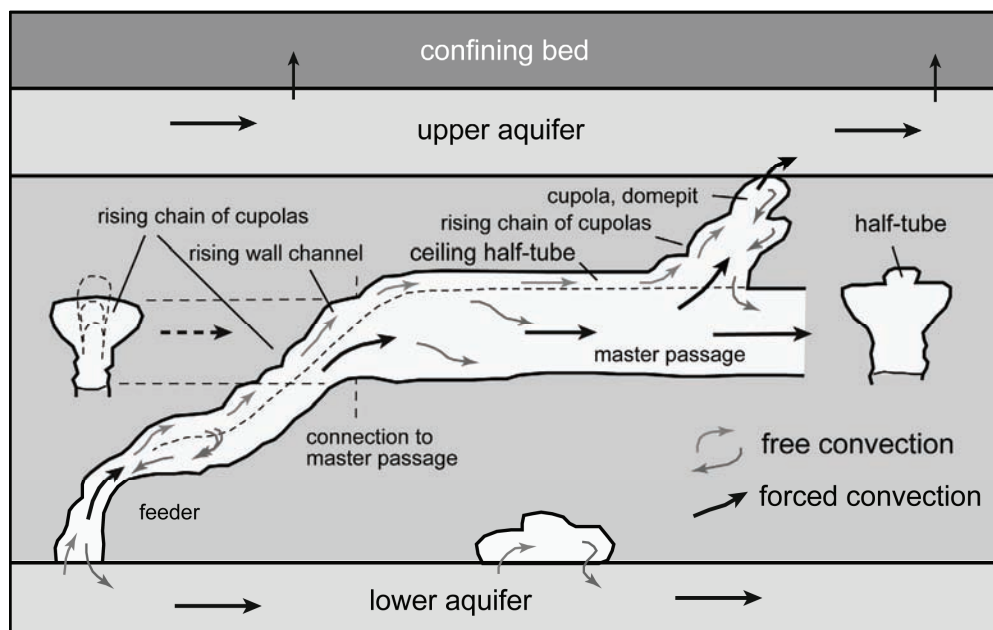


Figure 3. The morphologic suite of rising flow, diagnostic of the hypogenic transverse origin of caves (adapted from Klimchouk, 2007).

identification, as illustrated by many recent publications, including this volume. These new data and efforts will allow refining and further developing criteria for identification of hypogene speleogenesis.

Criteria to identify hypogene speleogenesis

In identifying hypogenic caves, the following approaches and criteria are used in combination (see Klimchouk, 2007 for details):

1. Regional paleo-hydrogeological analysis, regional and local hydrostratigraphy,
2. Morphogenetic analysis of caves: patterns and meso-morphology in relation to hydrostratigraphy,
3. Cave sediments and minerals,
4. Geochemistry of the host rock from the perspective of geochemical alterations in the course of hypogene speleogenesis.

Hypogenic karst and paleokarst

Within the conventional, predominantly epigenetic, karst paradigm, instances of deep-seated karst have commonly been interpreted as paleo (epigenetic) karst because the possibility of karstification in deep environments without recharge from the immediately overlying or adjacent surface has been neglected for a long time.

Paleokarst is not a particular type of karst but is, rather, a condition - fossilized. Features become paleokarst as they get hydrologically decoupled from

contemporary systems, in contrast to relict features that exist within contemporary systems but are removed from the environment in which they developed (Ford & Williams 1989). True paleokarst is buried karst, which is a complete infilling and burial of epigenetic (including coastal/oceanic) karst by later rocks such as transgressive marine sediments. Paleokarst horizons are reliably recognized where they underlie unambiguous stratigraphic unconformities related to subaerial exposure.

With growing recognition of hypogenic speleogenesis, it becomes increasingly obvious that in many cases features previously interpreted as paleo (epigenetic) karst (including coastal/oceanic karst) can be better explained as active or relict hypogenic features. Review of the international literature, especially related to carbonate-hosted hydrocarbon reservoirs, reveals that in many cases paleokarst had been doubtfully assumed based on its occurrence beneath distinct formational contacts, and interpreted as evidence of subaerial exposure through reciprocal reasoning of the presence of paleokarst. Other common cases of problematic paleokarst are stratiform breccia horizons, which are often the ultimate result of hypogenic speleogenesis, namely – collapsing of laterally extensive stratigraphically-comformable maze caves.

Hypogene speleogenesis tends to operate over long time spans, intermittently or being repeatedly suspended and activated. Hypogenic features may become relict but still remain within the contemporary systems, e.g. in a system where original confinement was breached and flow pattern reversed from upwell-

ing to descending. Hypogenic features are not paleokarst unless their evolution is completely halted by the removal of the cave-forming unit from the geological section (stratiform breccia horizons), complete sealing by cementation (mineralization) or lithification of the fill material, or by closure of the contemporary hydrogeological cycle with a new marine transgression. However, establishing the paleo-status of hypogenic features and their distinction from eogenic and epigenic paleokarst requires additional discussion.

Many important hydrocarbon and mineral deposits are karst-related. Better understanding of hypogene speleogenesis is of paramount importance for their proper genetic interpretation, which in turn is crucial for the development of more adequate approaches to prediction, prospecting and exploitation of these resources.

Mineral deposits

The significance of fluid migration and groundwater flow systems in the genesis of mineral deposits is now well recognized (Sharp and Kyle, 1988). Proper flow models are crucial for understanding ore genesis since it is the appropriate conditions of the flow system that allow particular geochemical processes to operate and produce massive mineral accumulations. The emerging theory of hypogenic speleogenesis places hypogenic karst in the systematized context and hierarchical structure of basinal groundwater flow, and highlights the powerful role of speleogenesis in the organization of regional flow systems, a consequence of its unique capacity to dramatically alter primary porosity and permeability of soluble formations. In sequences containing soluble formations, it is hypogenic speleogenesis that creates zones of enhanced ascending cross-formational communication and converges flow to them, a condition commonly seen as most important for the flow-induced accumulation of transported mineral matter and hydrocarbon migration.

Because of its transverse nature and ability to enhance cross-formational communication, hypogene speleogenesis facilitates the interaction of waters of contrasting chemistries and different geochemical environments. Thus, it often creates transitional environments and geochemical thresholds and the necessary pattern of migration of reactants and reaction products between them to produce mineral accumulations.

Hydrocarbon deposits

As with ore deposits, the role of hypogenic transverse speleogenesis in converging flow and enhancing cross-formational hydraulic communication between

stories in layered reservoirs can be demonstrated for migration and concentration of hydrocarbons. The difference with respect to ore formation is that entrapment of hydrocarbons, and the formation of deposits, is caused not by geochemical barriers but stratigraphic and hydrodynamic barriers at overlying insoluble low permeability units.

Many important deposits of hydrocarbons throughout the world are associated with karstified formations. An important issue in prospecting and exploration of carbonate-hosted hydrocarbon deposits is the characterization of karst porosity in production horizons. This problem is presently approached almost exclusively on the basis of general epigenic karst concepts, taken in the context of paleokarst. The most popular model is an island hydrology model implying speleogenesis at the fresh water/salt water mixing zone beneath a limestone island. The concept of hypogenic transverse speleogenesis opens new perspectives for interpreting karst features in oil and gas fields and suggests that many deposits previously thought to be associated with paleokarst may be of hypogenic karst origin. An instructive example of re-interpretation of the karst origin in an oil field from the perspective of hypogene speleogenesis is provided for the Yates oil field in Texas by Stafford et al. (2008; see also Chapter 5.3 in Klimchouk, 2007).

Synopsis

The list below is an attempt to summarize essential features and roles of hypogene speleogenesis. In brief, hypogene speleogenesis:

1. Operates mainly under confined (semi-confined) conditions,
2. Involves various dissolution mechanisms at transitional geochemical environments and thresholds that commonly occur along cross-formational flowpaths,
3. Creates solution porosity structures which are quite distinct from those formed by epigenic speleogenesis. They range from concordant structures controlled by individual strata to discordant cross-cutting structures within tens to hundreds of meters of sedimentary sequences. Combination of discordant and concordant elements is common,
4. Has much wider distribution than previously presumed,
5. Has no genetic relationship with groundwater recharge from the overlying or immediately adjacent surface, and may have no surface manifestation at all,
6. May operate during long time spans (millions to hundreds of millions of years),

7. Creates aerially extensive (although commonly clustered) or localized zones of high vertical permeability,
8. Serves to enhance cross-formational communication and to concentrate flow to such zones by opening migration paths across soluble formations and overlying non-karstic aquitards,
9. Plays an important role in (re)organization of regional flow systems.

The current burst of recognition of hypogene speleogenesis and its theoretical implications signifies an ongoing shift from the predominantly epigenetic, largely geomorphological paradigm in karst science to a more comprehensive geological paradigm, where karst and speleogenesis are viewed from the perspective of the entire evolutionary history of a sedimentary sequence containing soluble formations.

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References

- Andre, B.J., and H. Rajaram. 2005. Dissolution of limestone fractures by cooling waters: Early development of hypogene karst systems. *Water Resources Research* 41: doi:10.1029/2004WR003331.
- Birk, S. 2002. *Characterisation of karst systems by simulating aquifer genesis and spring responses: Model development and application to gypsum karst*. Tübinger Geowissenschaftliche Arbeiten C60.
- Birk, S., R. Liedl, M. Sauter, and G. Teutsch. 2003. Hydraulic boundary conditions as a controlling factor in karst genesis: A numerical modeling study on artesian conduit development in gypsum. *Water Resources Research* 39 (1): 1004: doi: 10.1029/2002WR001308.
- Birk, S., R. Liedl, M. Sauter, and G. Teutsch. 2005. Simulation of the development of gypsum maze caves. *Environmental Geology* 48 (3): 296-306.
- Bosak, P. 2000. Notes on the history of some karstological terms - hydrothermal karst, geysermite, vadose zone. *Acta Carsologica* 29: 233-240.
- Budd, D.A., and H.L. Vacher. 2004. Matrix permeability of the confined Floridan Aquifer, Florida, USA. *Hydrogeology Journal* 12: 531-549.
- Cooke Michele L., J.A. Simo, Chad A. Underwood, and Peggy Rijken. 2006. Mechanical stratigraphic controls on fracture patterns within carbonates and implications for groundwater flow. *Sedimentary Geology* 184: 225-239.
- Desnoyers, J.P.F.S. 1845. *Recherches géologiques et historiques sur les cavernes et particulièrement sur les cavernes à ossements*. Paris: 1-83.
- Dijk, P.E., and B. Berkowitz. 2000. Buoyancy-driven dissolution enhancement in rock fractures. *Geology* 28 (11): 1051-1054.
- Dijk, P.E., and B. Berkowitz. 2002. Measurement and analysis of dissolution patterns in rock fractures. *Water Resources Research* 38 (2): 1013, doi: 10.1029/2001WR000246.
- Dreybrodt, W., F. Gabrovsek, and D. Romanov. 2005. *Processes of Speleogenesis: A Modeling Approach*. Ljubljana, ZRC Publishing.
- Eaton, T.T. 2006. On the importance of geological heterogeneity for flow simulation. *Sedimentary Geology* 184: 187-201.
- Ford, D.C. 2006. Karst geomorphology, caves and cave deposits: A review of North American contributions during the past half century. In *Perspectives on Karst Geomorphology, Hydrology and Geochemistry*, ed. R.S. Harmon and C.W. Wicks, 1-14. Geological Society of America Special Paper 404.
- Ford, Derek C., and Paul W. Williams. 1989. *Karst Geomorphology and Hydrology*. London: Unwin Hyman.
- Ford, Derek C., and Paul W. Williams. 2007. *Karst Hydrogeology and Geomorphology*. Chichester: Wiley and Sons.
- Gunn, J., ed. 2004. *Encyclopedia of Caves and Karst Science*. New York-London: Fitzroy Dearborn/Taylor and Francis.
- Hill, C.A. 2000. Sulfuric acid, hypogene karst in the Guadalupe Mountains of New Mexico and West Texas, USA. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer, and W. Dreybrodt, 309-316. Huntsville: National Speleological Society.
- Klimchouk, A.B. 1997. Speleogenetic effects of water density differences. In Proceedings, 12th International Congress of Speleology, La Chaux-de-Fonds. La Chaux-de-Fonds, Switzerland, 161-164.
- Klimchouk, A.B. 2000. Speleogenesis under deep-seated and confined settings. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer, and W. Dreybrodt, 244-260. Huntsville: National Speleological Society.
- Klimchouk, A.B. 2007. *Hypogene Speleogenesis: Hydrogeological and Morphogenetic Perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Klimchouk, A.B. 2008. Morphogenesis of hypogenic caves. *Geomorphology*: doi:10.1016/j.geomorph.2008.09.013.

- Klimchouk, A.B., and D.C. Ford. 2000. Types of karst and evolution of hydrogeologic settings. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer, and W. Dreybrodt, 45-53. Huntsville: National Speleological Society.
- Klimchouk, A.B., D.C. Ford, A. Palmer, and W. Dreybrodt. 2000. *Speleogenesis: evolution of karst aquifers*. Huntsville: National Speleological Society.
- Palmer, A.N. 1991. Origin and morphology of limestone caves. *Geological Society of America Bulletin* 103: 1-21.
- Palmer, A.N. 1995. Geochemical models for the origin of macroscopic solution porosity in carbonate rocks. In *Geochemical models for the origin of macroscopic solution porosity in carbonate rocks*, American Association of Petroleum Geologists, Memoir 63, ed. D. A. Budd, P. M. Harris and A. Sailer, 77-101. Tulsa: American Association of Petroleum Geologists.
- Palmer, A.N. 2000. Hydrogeologic control of cave patterns. In *Speleogenesis: Evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer, and W. Dreybrodt, 77-90. Huntsville: National Speleological Society.
- Palmer, A.N. 2007. *Cave Geology*. Dayton: Cave Books.
- Rehrl, C., S. Birk, and A.B. Klimchouk. 2008. Conduit evolution in deep-seated settings: Conceptual and numerical models based on field observations. *Water Resources Research* 44: W11425, doi:10.1029/2008WR006905.
- Rehrl, C., S. Birk, and A.B. Klimchouk. 2009. Influence of initial aperture variability on conduit development in hypogene settings. *Zeitschrift für Geomorphologie*, submitted.
- Sharp, J.M., Jr., and J.R. Kyle. 1988. The role of ground-water processes in the formation of ore deposits. In *The Geology of North America, Hydrogeology*, ed. W. Back, J. S. Rosenshein and P. R. Seaber. 0-2: 461-483. Boulder: Geological Society of America.
- Stafford, K.W., F.H. Behnken, and J.G. White. 2008. Hypogene speleogenesis within the central basin platform: Karst porosity in the Yates Field, Pecos County, Texas, U.S.A. In *Karst from Recent to Reservoirs: Karst Water Institute Special Publication 14*, ed. I.D. Sasowsky, C.T. Feazel, J.E. Mylroie, A.N. Palmer, and M.V. Palmer, 174-178. Leesburg: Karst Waters Institute, Inc.
- Tóth, J. 1999. Groundwater as a geologic agent: An overview of the cases, processes and manifestations. *Hydrogeology Journal* 7: 1-14.

DIAGNOSTIC FEATURES OF HYPOGENIC KARST: IS CONFINED FLOW NECESSARY?

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Abstract

The term hypogenic cave, popularized in the cave science community by Arthur Palmer in 1991 to explain cave development de-coupled from surface hydrology, was recently refined by Alexander Klimchouk to primarily mean the outcome of slow flow resulting from confined karst aquifer conditions. A key aspect of this argument is the existence of diagnostic bedrock dissolutional forms and morphologies in caves: rising wall channels, ear-like or dome orifices, point and rift floor infeeders, ceiling channels, cupolas, and thin bedrock partitions, among others. The existence of such features in continental epigenic caves is argued to represent inheritance from an earlier confined stage, with an epigenic overprint. However, is confined flow the essential condition to produce these diagnostic dissolutional morphologies?

Because of the inheritance and overprinting arguments, the dissolutional morphology question can only be addressed by examining caves where confined or mesogenetic conditions never existed, as on eogenetic carbonate coasts and islands. These caves contain suites of bedrock dissolutional morphologies remarkably similar or identical to those reported to be solely the result of confined or semi-confined flow, and therefore these morphologies cannot be true diagnostic indicators of confined conditions. The key factor in developing these bedrock morphologies is slow flow, as occurs in mixing dissolution in coastal carbonates. In epigenic cave systems, stagnant ponding from certain flood conditions also allows slow flow to repeatedly participate in the bedrock dissolutional processes, creating an overprint on epigenic dissolution, instead of the other way around as Klimchouk's hypogenic cave model postulates.

Introduction

Hypogenic caves have been part of a new frontier in speleology for over two decades, particularly since the publication of Arthur Palmer's landmark paper, "Origin and morphology of limestone caves", placed such caves in a continuum of dissolutional void development in carbonate rocks (Palmer, 1991). Klimchouk (2007 and references therein) expanded the concept to include evaporite rocks, focusing on

gypsum. Because hypogenic caves are decoupled from direct surface hydrology, they create a class of caves strongly influenced by explorational bias. Mylroie (2003) reviewed the cave distribution data of Palmer (1991) to demonstrate that the under representation of hypogenic caves was a result of their isolation from the surface environment. Klimchouk (2007) has also emphasized this point. Of the five longest caves in the world (<http://www.caverbob.com/wlong.htm> as of January 19, 2009), four are hypogenic (Jewel, 225 km; Optymistychna, 215 km; Wind, 209 km; and Lechuguilla, 201 km), yet two, Jewel Cave and Lechuguilla Cave, have one known natural entrance, and Wind Cave has only two. These observations indicate that discovery and entry into extremely long relict hypogenic caves is very fortuitous. The existence of the three American examples in the dry western United States also indicates that survival of these caves, and the ability to enter them, may be contingent on climate. For many years following the 1980's development of the H₂S hypogenic model surrounding the Guadalupe Mountains, New Mexico caves (e.g. Hose and Pisarowicz, 2000), and the various hypogenic models for the Black Hills, South Dakota caves (e.g. Bakalowicz et al., 1987), eastern U.S. cavers wondered why none of these spectacular and long caves were in their region of the country. The consensus in the American cave science community seemed to be that in the humid environments of the eastern United States epigenic caves and the large karst water fluxes they generated had resulted in overprinting, segmentation, and infilling of relict hypogenic caves. This last aspect is a key tenet of Klimchouk's model of hypogenic cave development: that many cave features seen in epigenic caves today are inherited from an earlier hypogenic phase (Klimchouk, 2007).

Klimchouk's model of hypogenic cave development (Klimchouk, 2007 and references therein) centers on dissolution of carbonates and evaporites in confined or semi-confined settings. The full model is dealt with elsewhere in this volume, and will not be extensively reviewed in this paper. Both confined and semi-confined conditions will be considered confined in this paper, as upward leakage through a confining layer is the primary control of the model. The central

point of Klimchouk's model is that karst aquifers, when confined in sedimentary basins, undergo dissolutional modification. That modification is necessarily in a slow flow condition, as one component of water flow through these karst aquifers is a slow leakage upward across confining layers. Rapid turbulent flow does not exist, and the slow leakage creates a non-competitive environment for water flow; there is no organized lateral flow to create integrated conduit systems. Through time, dissolution may cause independently-forming voids to intersect and link up, but such connections have little impact on overall flow patterns. Void systems from a few meters across to those with hundreds of kilometers of passage may form. The bedrock dissolutional morphologies produced by this process are different from those expected in epigenic stream caves, where turbulent flow hydrodynamics govern dissolution.

Some aquifers, such as the Edwards aquifer, are now considered to be hypogenic (Schindel et al., 2008), however the Edwards links directly via dissolutional macroporosity to an adjacent epigenic recharge system and is quite different from the classic isolated, deep basin and confined hypogenic cave-producing karst aquifer presented by Klimchouk (2007). In addition, discharge points for artesian aquifers, such as artesian wells and springs, commonly exhibit turbulent flow, but such turbulent flow cannot be assumed for the entire aquifer feeding those discharge points. Deep confined sandstone aquifers, when initially drilled, commonly show voluminous and turbulent flow from the producing well as a result of high pressure heads.

Carbonate rocks can be classified based on their position in the diagenetic cycle. Choquette and Pray (1970) divided the post-depositional evolution of carbonate porosity into three time-porosity stages as elements of the rock cycle. They defined "the time of early burial as eogenetic, the time of deeper burial as mesogenetic, and the late stage associated with erosion of long-buried carbonates as telogenetic" (Choquette and Pray, 1970, p. 215). There is a tendency when thinking of the confined hypogenic cave model to assume that the speleogenesis takes place solely in the mesogenetic realm, the site of deep-burial diagenesis. As the gypsum maze caves of the Ukraine demonstrate, neither deep burial nor extensive diagenesis is necessary to create a confined speleogenetic environment. Confined conditions can occur in young carbonate rocks near the surface, as seen in the eogenetic Oligocene limestones of Florida, which are locally confined by the Hawthorn Formation clays (Martin et al., 2006). One can argue that caves formed

in the mesogenetic environment are obligatory hypogenic caves, but the converse is not necessarily true (Mylroie and Mylroie, 2009). None the less, the soluble rock units found in large and deep sedimentary basins are commonly confined, and the great areal extent of the basins makes lateral water transfer slow and inefficient (Klimchouk, 2007). As a result, very slow ascending flow across soluble units confined by insoluble, mostly impermeable rocks has both the opportunity, and perhaps more importantly, the time to operate.

The term "hypogene cave" was introduced to the larger speleological community by Ford and Williams (1989), primarily in the context of rising deep-seated H₂S-rich waters mixing with shallow phreatic O₂-rich water, as they reported was observed in the Guadalupe Mountains of New Mexico, or in Italy and Georgia in Europe. Arthur Palmer went on to use the term "hypogenic cave" to explain limestone cave formation as a result of: "...acids of deep-seated origin, or epigenic acids rejuvenated by deep-seated processes" (Palmer, 1991, p. 4). Ford (1995) felt rising waters were a required aspect for hypogenic cave development (his *per ascensum* water). The concept described by the term "hypogene" has been around a long time. Krauskopf (1967) indicates that Lindgren's 1933 text, "Mineral Deposits" 4th edition (McGraw Hill, NY) used the term "hypothermal" for minerals formed at the greatest depths and closest to the igneous source. The Glossary of Geology defines hypogene as: "(a) Said of a geologic process, and its resultant features, occurring within and below the crust of the earth. Cf. epigenic; endogenetic. Syn: hypogenic; hypogeal; hypogeic. (b) Said of a mineral deposit formed by ascending solutions; also said of those solutions and of that environment. Cf. supergene; mesogene. (c) rarely used syn. of plutonic." (Neuendorf et al., 2005, p. 315). Interestingly, and in contrast to the quote from Palmer (1991) above, the Glossary of Geology defines a hypogenic cave as: "A dissolution cave formed by waters whose solutional capacity is derived from sources beneath the land surface, most commonly by hydrogen sulfide or cooling of thermal waters (Palmer 1991). Cf. epigenic cave." (Neuendorf et al., 2005, p. 315). The change was written for the Glossary by Palmer (personal communication). The term hypogenic cave was quickly expanded to include caves where mixing of waters in coastal carbonate aquifers produced flank margin caves, with passage configurations closely similar to those found in the Guadalupe Mountains (Mylroie, 1991; Mylroie and Carew 1995; Palmer, 2007). The argument was made in this case that the idea of "depth" was relative, and that the over riding

factor was renewed dissolutional aggressivity by mixing of different waters in an environment decoupled from surface, or epigenetic processes. Klimchouk (2007 and references therein) considered the Palmer (1991) definition too restrictive, as its emphasis on acids failed to consider dissolution of evaporites at depth, but Palmer (1991) was a paper on limestone cave (not overall dissolutional cave) origin so Palmer's acidity focus was appropriate. It is interesting to note that the Glossary of Geology definition of hypogenic caves is worded by Palmer so as to discuss "solutional capacity", and not "acids", which anticipated Klimchouk's concern. A Google® search (January 19, 2009) shows 58,900 responses for "hypogene". Narrowing to "hypogene caves" yields 3,340 responses; "hypogenic caves" yields 1,830 responses. The term is well established in the geoscience literature. Klimchouk (2007) reviews recent usage of the term by a variety of authors. Because the authors of this paper feel that cave development in a confined environment (whether mesogenetic or not) is one of several ways caves can develop under hypogenic conditions, we have used the term confined hypogenic cave in this paper for those caves described by Klimchouk (2007).

To what degree do hypogenic caves survive introduction into the epigenetic environment, and secondarily, to what extent do those surviving hypogenic caves influence epigenetic cave development and hydrology? As hypogenic caves enter the epigenetic environment, they should be considered paleokarst features (Ford, 1995). Ford (1995) considered the involvement of epigenetic waters (his per descensum waters) with paleokarst to be random and to not influence modern

ground-water flow trends. Rising water (his per ascensum water), on the other hand, could be expected to utilize paleokarst features as one of the few high permeability routes available. The physical translation of confined hypogenic caves from the deep-burial diagenesis of the mesogenetic environment to the epigenetic processes of the telogenetic environment (e.g. Klimchouk, 2007, his Figure 1) is a long-term activity, working essentially at the cratonic scale. It seems unlikely that these caves would be continually influencing hydrologic flow; at some point they represent paleokarst.

Palmer (2007), and Ford and Williams (2007) both define paleokarst as epigenetic karst that has been buried. Interestingly, Ford (1995, p. 138) in a paper focused on hypogenic caves, said: "Paleokarst phenomena are here defined as those that at the present time or for some significant duration in the geologic past have been decoupled from all active hydrogeochemical systems." Burial is not a requirement to become paleokarst as hypogenic caves are already buried. After a traditional explanation of the burial of epigenetic karst to produce paleokarst, Ford (1995, p. 138) further elaborates: "If they [caves] were formed at depth (subadjacent, interstratal or intrastratal karst resulting from e.g. salt dissolution) the definition requires a long period when there was no significant chemical interaction with any fluids moving through them; probably fluid circulation ceased to all intents and purposes." Ford (1995) implies that hypogenic caves leave their environment of formation and transit to the telogenetic realm as hydrologically-inactive paleokarst features.

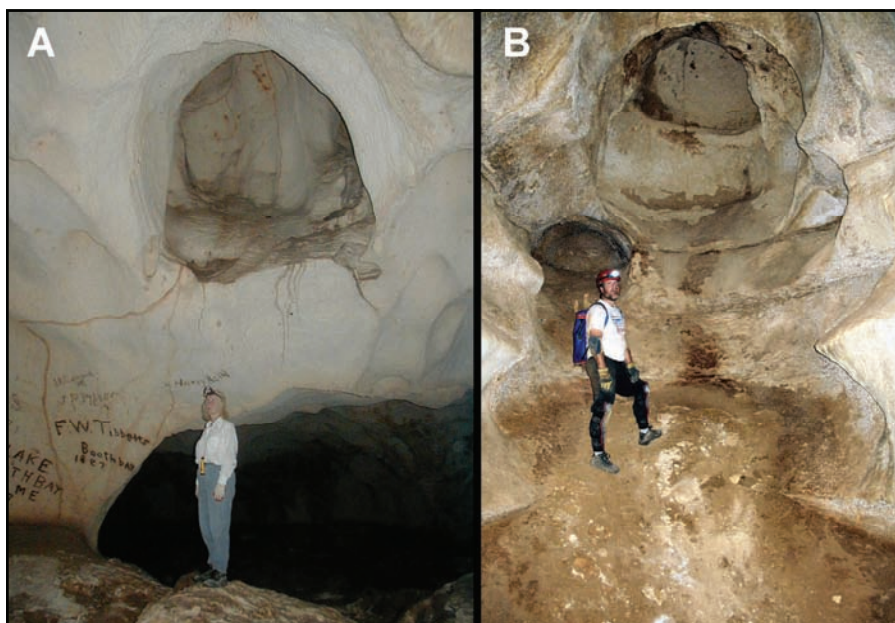


Figure 1. Comparison of large cupolas from flank margin caves and caves classified as confined hypogenic caves. A) Hatchet Bay Cave, Eleuthera Island, Bahamas, a flank margin cave. B) Deep Cave, Texas, a confined hypogenic cave (from Klimchouk, 2007, Plate 6D). For all figures, photos are by the authors unless otherwise noted; all rocks are carbonates unless otherwise noted; images from Klimchouk 2007 used with permission.

The formation of these hypogenic caves between confining units by ascending flow means that the caves have little or no influence on flow rates or directions. If they did, then competitive forces would be established between functioning flow routes, and consolidation of flow would result. Such consolidation of flow is specifically denied by the confined hypogenic cave model of Klimchouk (2007).

For confined hypogenic caves to successfully influence epigenic karst flow, they would need to act as preferential flow paths during their ascension to the epigenic realm, and impose those flow paths on epigenic cave development. Given that models developed to show how cave conduits develop in the epigenic environment have been highly successful in explaining these subsurface flow paths (e.g. Palmer, 1991), organizational control by relict hypogenic caves seems unlikely. On the other hand, the intersection of relict hypogenic caves by epigenic conduits seems almost a guaranteed occurrence at some scale. Assuming some finite number of relict hypogenic caves emerging from the mesogenetic environment into the telogenetic (or surficial) environment, the interaction with epigenic flow routes would have to happen from time to time. Since these epigenic flow routes are controlled by the interaction of geologic structure with topography (e.g. dip and strike; faults and folds), only rarely would the relict hypogenic caves be positioned and oriented to provide a competitive advantage to ground-water flow. Confined hypogenic caves develop in a non-competitive hydrologic environment, as no flow route can gain a hydrologic transmission advantage over any other flow route due to slow leakage upward through the confining layer (Klimchouk, 2007). As a result, internal connections are of an essentially random nature, and it is therefore unlikely that they carry any pre-adaptation to assist in epigenic flow. Therefore their influence on epigenic flow should be minimal.

Klimchouk (2007) re-evaluates the epigenic models of maze cave formation, especially those of Palmer (1975; 1991). Klimchouk's argument, based on contrasts between alpine gradients and lowland gradients, where the former can produce floodwater mazes but the latter cannot, are unconvincing. If recharge exceeds discharge, vadose and epiphreatic caves will flood, and if base level rises above the cave, caves will flood, regardless of the amount of local relief. Similar arguments are made regarding flood mazes adjacent to rivers, where Klimchouk (2007) claimed that such maze caves can enlarge but not originate in these flood settings. This approach by Klimchouk (2007) assumes floodwaters cannot

penetrate joints laterally in a low confining pressure epigenic environment to initiate cave genesis, but that slow upward leakage across aquicludes in a high pressure confining environment can.

Klimchouk (2007) takes an expanded view of confined hypogenic caves, and considers a large number of caves previously thought to be epigenic to be actually mostly hypogenic in origin, with later epigenic overprinting. Part of the reason for these re-evaluations is the identification of features thought to be solely the outcome of confined hypogenic processes. To assess the degree to which such overprinting has occurred, and the degree to which many epigenic caves, especially maze caves, are hypogenic in origin, requires being able to unambiguously identify hypogenic cave features.

Cave deposits might be a way to differentiate a hypogenic versus epigenic origin for a cave, especially if those deposits could be demonstrated to be quite old. Cave deposit data from the Guadalupe Mountain caves of New Mexico, involving datable clays of clear hypogenic origin (Polyak, et al., 1998), are an excellent example. Some data, such as that from the Jenolan Caves in Australia (Osborne, 2007), come from cave deposits that indicate an early phreatic origin at ~340 million years ago. Such data are only secondary evidence. While it is highly unlikely that a cave formed 340 million years ago would survive in an epigenic environment for that long a period of time, the cave could as easily be an epigenic cave subsequently buried and re-exhumed instead of a hypogenic cave making its first appearance at the earth's surface. In any event, it is clear that the passages involved in the Osborne (2007) discussion are not part of the current epigenic flow regime, and provide little evidence of past control of that flow regime.

The confined hypogenic cave model considers the dissolutional bedrock forms produced in the confined, hypogenic environment to be diagnostic (Klimchouk, 2007). The appearance of such bedrock dissolutional forms in both active and inactive epigenic caves today is considered in this model to be the result of inheritance of hypogenic cave voids by some, or most, of the cave as currently observed. Klimchouk (2007) cites numerous examples of caves currently functioning in the epigenic environment as examples of this hypogenic inheritance, such as Mystery Cave, Minnesota and Skull Cave, New York. These examples are based on cave pattern and on the dissolutional bedrock morphology of individual passages. The morphological signatures of confined hypogenic caves are

displayed in Plates 1 – 18 of Klimchouk (2007) as key forms and shapes that can be used as diagnostic indicators. Extensive or continuous appearance of these features is interpreted to indicate actual flow control of the epigenic waters in the cave by inherited hypogenic voids. The critical factor is the bedrock dissolutional form. Are these forms truly unique to the hypogenic environment, or are they the result of laminar or slow flow hydrodynamics, an event that can occur in a variety of phreatic karst settings, not just in confined ones? If the forms are not diagnostic of confined hypogenic cave origin, then the widespread re-evaluation of many cave systems as being under the control of relict hypogenic caves (e.g. Klimchouk, 2007; Alexander et al., 2008) is questionable.

Mylroie (2008) has made some strong criticisms, not of the basic confined hypogenic cave model, but of its widespread application to the global cave environment by Klimchouk (2007). Mylroie and Mylroie (2008) have questioned the use of dissolutional rock sculpture to assume a confined, hypogenic origin for caves. As the Mylroie and Mylroie (2008) material is a GSA abstract and subsequent oral presentation, the issues brought up in that work are the focus of this paper, to create an accessible and permanent record and initiate scientific discussion needed to clarify impact of hypogene processes on observed karst features at the earth's surface.

Diagnostic bedrock forms

The dissolutional bedrock forms described as diagnostic for confined hypogenic cave origin by Klimchouk (2007), as listed and shown in plates 1-18 of that publication, are: side and floor feeders, rising wall channels, ceiling cupolas and rising chains of ceiling cupolas, ceiling channels, outlets, dome pits, and bedrock partitions, among others. The list is a mix of descriptive terms (ceiling channels) and interpretive terms (feeders).

A critical point in assessing the diagnostic bedrock dissolutional forms used to establish a confined hypogenic cave origin is to demonstrate that such forms can or cannot occur in a karst aquifer that has never been confined. Such conditions are difficult to demonstrate for any cave developed in telogenetic rock, because by definition that rock has been down to the mesogenetic environment. Its reappearance in the telogenetic environment is accompanied by inherited characteristics from the mesogenetic stage: loss of primary porosity, re-crystallization of cements and allochems, and compositional changes (e.g. dolomitization). It is relatively straightforward to then argue

that macroscopic features such as confined hypogenic caves have also been inherited.

There are some environments in which the karst aquifer has never been confined, let alone has descended to the mesogenetic environment. Eogenetic karst aquifers are formed in rocks that are still in or near their environment of deposition, and have not undergone burial, or been moved out of the influence of meteoric diagenesis (Vacher and Mylroie, 2002). Some of the best examples of such an aquifer are the eolian calcarenites of the Bahamian Archipelago. Not only have these rocks never been buried or confined, they were actually deposited subaerially by wind in an environment free of phreatic water. Only as a result of glacioeustatic sea-level rise have these dune limestones subsequently experienced shallow phreatic conditions. In addition, the rocks are very young in geologic terms, generally less than 500 ka (Carew and Mylroie, 1995a; 1997).

These carbonate rocks contain dissolutional voids of macroscopic size, the largest of which are termed flank margin caves. Dissolution occurs in the phreatic environment of a floating Ghyben-Herzberg-Dupuit fresh-water lens. The dissolution is maximized at the distal margin of the lens, under the flank of the enclosing landmass, hence the name flank margin cave. This dissolution is a result of: 1) mixing with vadose freshwater descending to the lens, and mixing with phreatic marine water below and adjacent to the lens; 2) oxidation of organics trapped at the density interfaces at the top and bottom of the lens; and 3) the increased flow velocity associated with the distal lens margin (Mylroie and Mylroie, 2007).

The Bahamas are tectonically stable (Carew and Mylroie, 1995b), so the flank margin caves that are above sea level today are the result of a past, higher glacioeustatic sea-level position. Because of the age of the rocks, only the Quaternary glaciations, as identified by the Marine Isotope Stage (MIS) system, can have driven the necessary sea-level change. Based on isostatic subsidence rates of the Bahamas, only the MIS 5e highstand (last interglacial) could have provided the sea-level highstand needed to lift a fresh-water lens into the position to make the flank margin caves as now observed in a subaerial setting (Carew and Mylroie, 1995b). That highstand lasted from 131 ka to 119 ka (Chen et al., 1991), a 12,000 year time window to make the caves. The caves formed extremely rapidly in an unconfined condition.

As a contrast, Isla de Mona contains many flank margin caves (Frank et al., 1998). This island, located

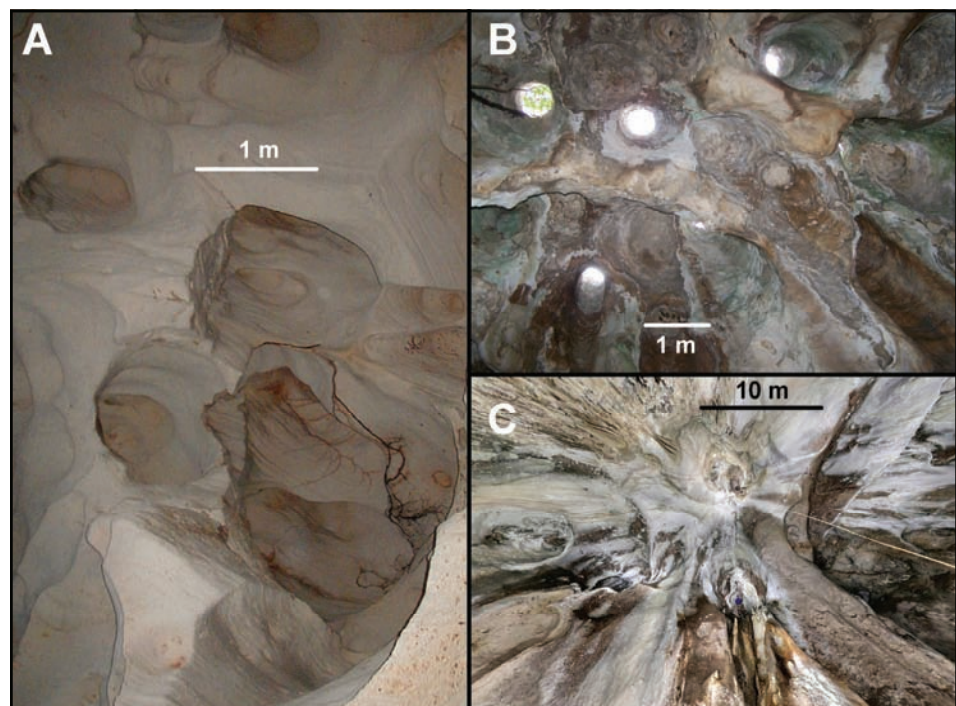
in Mona Passage between Hispaniola and Puerto Rico, is made up of Mio-Pliocene marine limestones. It contains the largest known flank margin caves in the world, with survey totals of 20 km for the Lirio Cave system. It has been tectonically uplifted, a consequence of its proximity to the North American – Caribbean plate boundary (Gonzalez et al., 1997). Paleomagnetic analysis of cave deposits has demonstrated that the caves are up to 1.8 million years old or older (Panuska, et al., 1998). The large cave size is apparently a result of cave genesis prior to the onset of high amplitude, short wavelength glacioeustasy of the Quaternary. As a result, unlike the Bahamas, the fresh-water lens was able to stay in a single position for an extended period of time, creating exceptionally large flank margin caves (Myroie and Myroie, 2007). Tectonic uplift of the island after cave genesis preserved the caves, and removed any opportunity for mesogenetic diagenesis, and confined conditions. The Bahamas show that flank margin caves can form rapidly in thousands of years; Isla de Mona shows that the caves can persist for millions of years.

Myroie and Carew (1995) and Palmer (2007) consider flank margin caves to be hypogenic, as they form as the result of mixing of waters at depth within the bedrock mass, isolated from direct hydrologic connection with surface hydrology. It can be argued that they are not hypogenic by some definitions of hypogene, as ascending waters are not required for their development (Ford, 1995); however, in confined conditions, a karst aquifer could leak downward as

well as upward to create flow as it is in a pressurized system, so purely ascending water may not be necessary. In any event, the flow within flank margin caves is not turbulent, and features such as bedrock ablation scallops, or stream-laid sediments, are absent.

Comparing the bedrock dissolutional forms of flank margin caves in the Bahamas and Isla de Mona with those forms listed as diagnostic of confined hypogenic speleogenesis by Klimchouk (2007) is revealing. Cupolas are a common feature in flank margin caves. Figures 1 and 2 compare cupolas, and Figure 3 cupolas with outlets and domepits, to those displayed by Klimchouk (2007). Figure 4 compares rising cupolas, and Figure 5 compares ceiling channels. Figure 6 compares bedrock wall partitions. Figure 7 compares fissure and rift-like feeders. In all cases, bedrock dissolutional forms exist in flank margin caves that have a remarkable similarity to those classified as diagnostic by Klimchouk (2007) for confined hypogenic caves. The features shown from the Bahamas and Isla de Mona are common in many caves in both locations. It is not necessary that the features from eogenetic carbonate islands be exactly identical to the diagnostic forms displayed by Klimchouk (2007). It is sufficient to say that these dissolutional forms are so close in configuration and appearance that if these forms were found later in an epigenic cave in the telogenetic environment, one could not tell the forms apart. As a result, such forms cannot be considered diagnostic of confined hypogenic cave origin. Can flank margin caves survive deep burial downward

Figure 2. Comparison of cupolas on overhead ceilings. A) Hatchet Bay Cave, Eleuthera Island, Bahamas, a flank margin cave. B) Generator Cave, Crooked Island, Bahamas, a flank margin cave. Some cupolas are breached by surface denudation. C) Lechuguilla Cave, Guadalupe Mountains, New Mexico, a confined hypogenic cave. Note the scale difference with A and B (National Park Service from Klimchouk, 2007, Plate 11A).



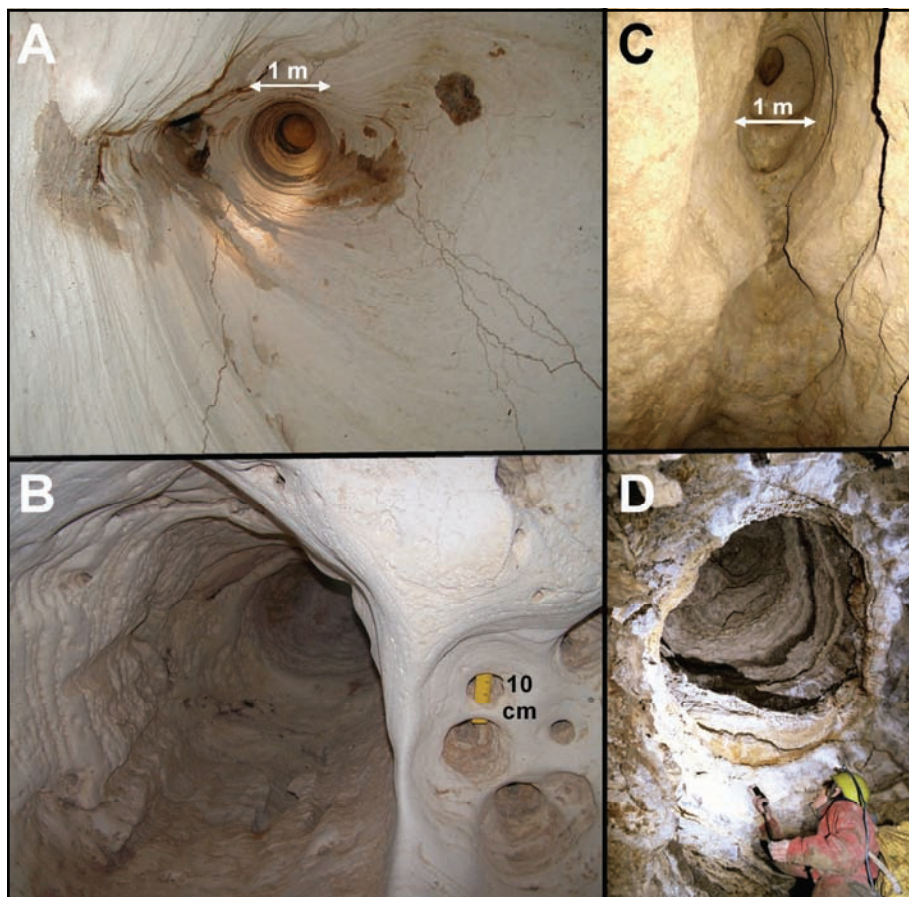


Figure 3. Comparison of "cupolas with outlets" and domepits. A) Cupola with outlet, Hamilton's Cave, Long Island, Bahamas. B) Domepit (possible bell hole) and subsidiary cupolas, Cumulous Cave, Crooked Island, Bahamas. C) Cupola with outlet, Mystery Cave, Minnesota (from Klimchouk, 2007, Plate 9D). D) Domepit, Slavka Cave, western Ukraine, in gypsum (from Klimchouk, 2007, Plate 9J).

Figure 4. Comparison of rising cupolas. A) Cupolas rising towards the viewer, Harry Oakes Cave, New Providence Island, Bahamas. B) Cupolas rising towards the viewer, Cueva del Agua, Punte Los Ingleses, Isla de Mona, Puerto Rico. C) Cupolas rising away from the viewer, Caverns of Sonora, Texas (from Klimchouk, 2007, Plate 6B). D) Cupolas rising away from the viewer, Spider Cave, Guadalupe Mountains, New Mexico (from Klimchouk, 2007, Plate 6C).

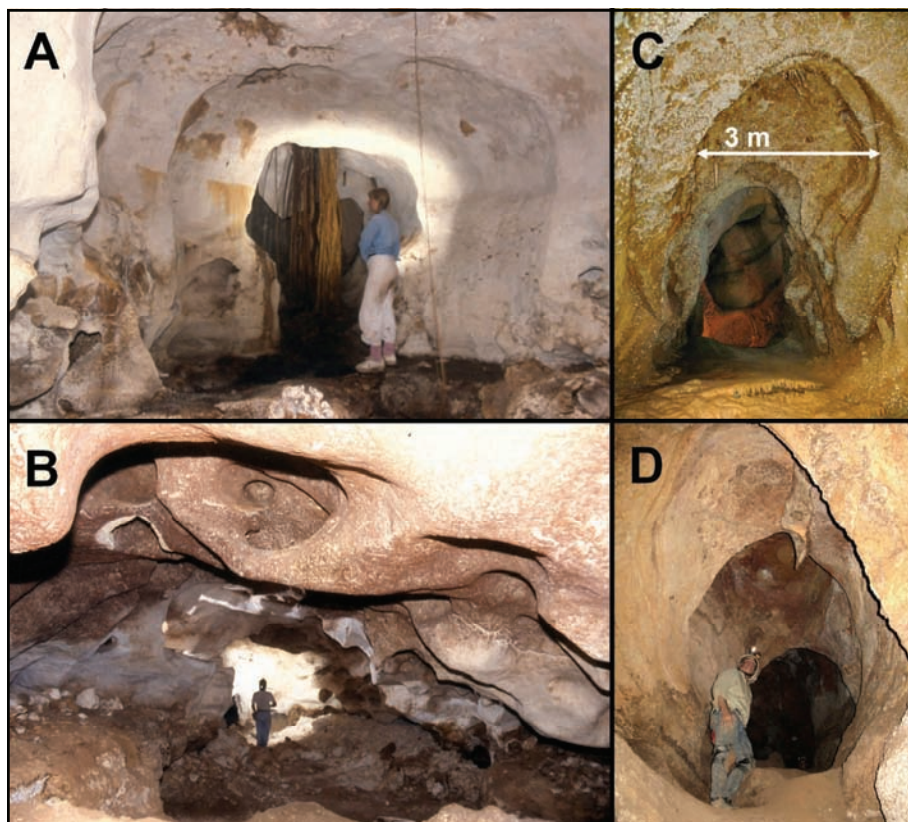


Figure 5. Comparison of ceiling channels. A) Ceiling channel, Cueva del Agua Sardinera, Isla de Mona, Puerto Rico. B) Ceiling channel, Cueva Nuevo, Isla de Mona, Puerto Rico. C) Ceiling channel in Atlantida Cave, western Ukraine, in gypsum (from Klimchouk, 2007, Plate 7A). D) Ceiling channel in Dzhurinskaja Cave, western Ukraine, in gypsum (from Klimchouk, 2007, Plate 6G).

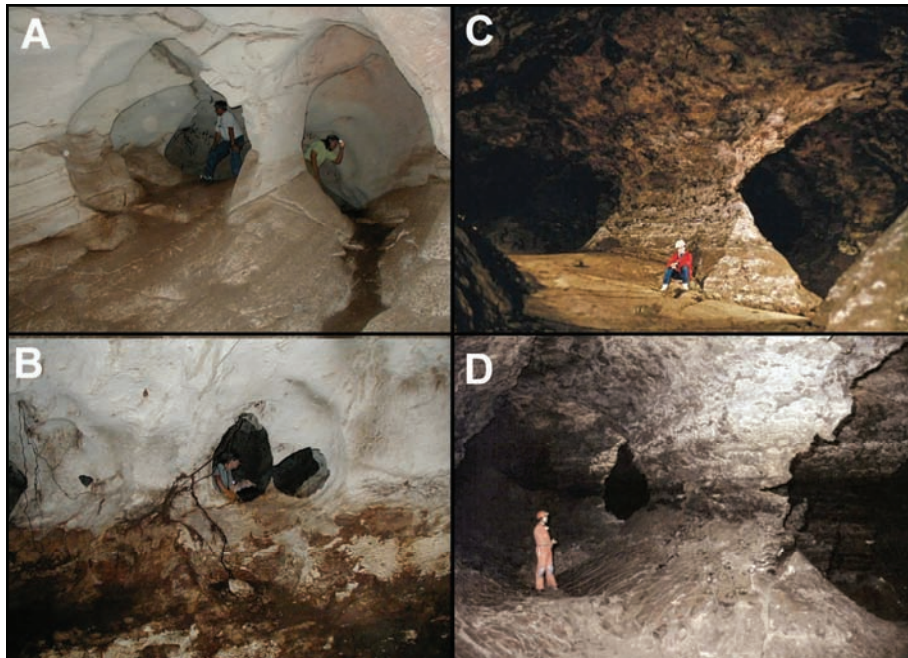
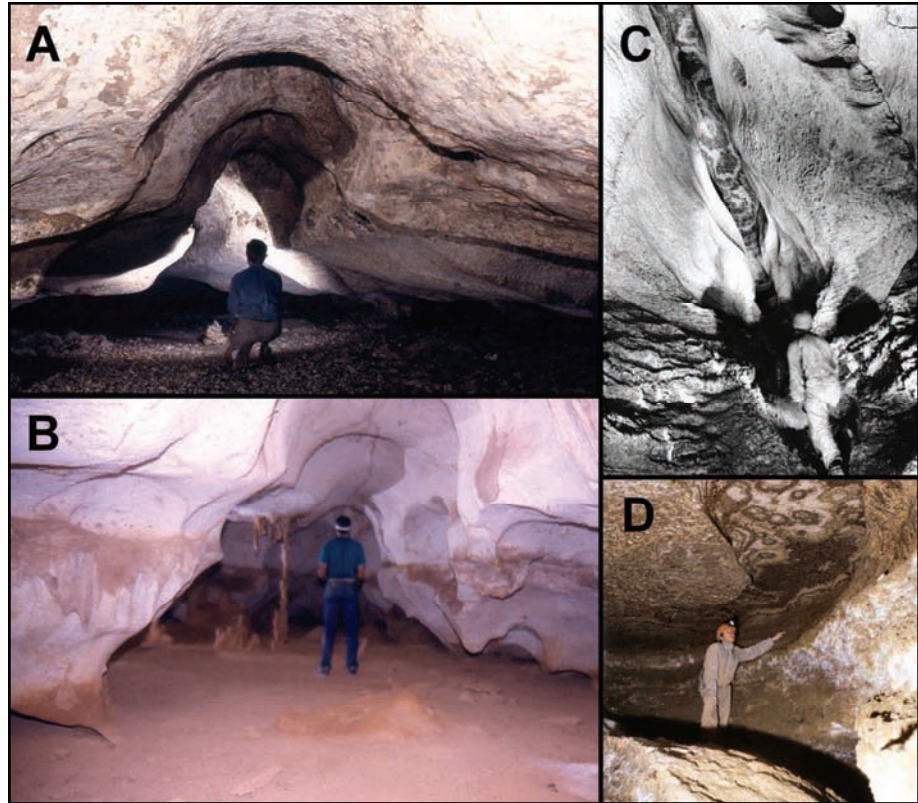


Figure 6. Comparison of bedrock wall partitions. A) Goat Cave, Long Island, Bahamas. B) Eight Mile Cave, Abaco Island, Bahamas. C) and D) Zoloushka Cave, western Ukraine, in gypsum (photos by B. Ridush and V. Kisselev, from Klimchouk, 2007, Plates 12E and 12G).

from the eogenetic environment? Isla de Mona indicates that large flank margin caves can survive for millions of years under near-surface conditions. In the Bahamas, Meyerhoff and Hatten (1974) report striking cavernous porosity at many depths ranging from 21 to 4082 m, the deepest void accepted 2430 m of broken drill pipe. The carbonate rocks involved were all

shallow water deposits that had subsided to depth. That large voids can survive at great depth in isostatically-subsiding carbonate rocks is without question. The question that cannot be answered is whether these voids were inherited from the shallow eogenetic environment, or were created at depth in this unconfining condition.

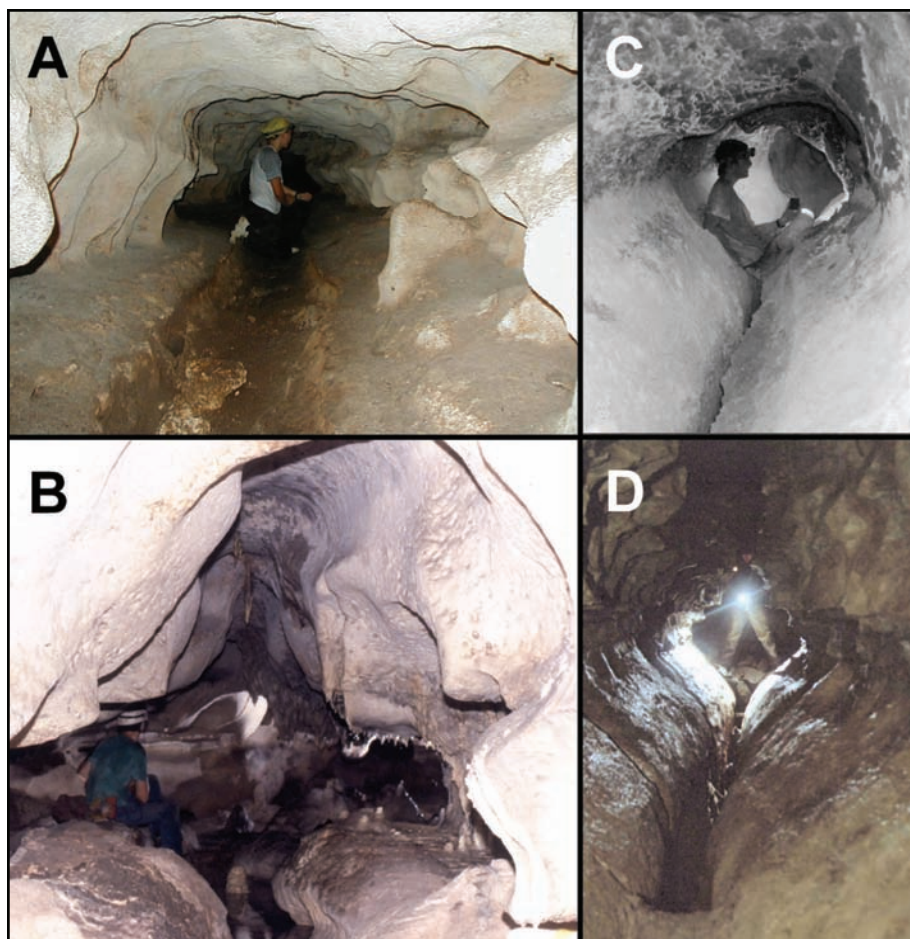


Figure 7. Comparison of fissure and rift-like passages. A) Eight Mile Cave, Abaco Island, Bahamas. B) Cueva del Agua Sardinera, Isla de Mona, Puerto Rico. C) Aneva Cave, Israel (photo by Amos Frumkin from Klimchouk, 2007, Plate 5H). D) Knock Fell Caverns, Northern Pennines, United Kingdom (from Klimchouk, 2007, Plate 5I).

Flank margin caves contain other interesting lessons. Figure 8 shows breccia facies from flank margin caves on Isla de Mona and the Bahamas. The facies are actually paleosol material either transported into the caves by vadose processes, or true paleosols within the carbonate sequence that have been cut through by dissolution. Such facies can give the impression of great age, or great depth of burial for the rock material. The discovery of voids with surfaces formed under phreatic conditions cutting such breccias could be misinterpreted as implying the voids formed in the mesogenetic environment.

Creation of similar bedrock dissolutional forms in dissimilar environments

What has caused the convergence of the dissolutional forms seen in Figures 1 to 7? One can postulate an inheritance argument, in which all such forms originated in the eogenetic environment as flank margin caves and later were carried down to the mesogenetic environment and subsequently upward into the telogenetic environment. Their appearance in epigenetic caves is therefore a consequence of inheritance from an initial eogenetic condition, and not a later hypogenic

origin in a mesogenetic setting. Confined hypogenic cave maps from the Guadalupe Mountains show a great similarity to flank margin cave maps from the Bahamas or Isla de Mona (Figure 9). Maps of flank margin caves from diagenetically-mature rocks, such as in New Zealand (Mylroie, et al., 2008), show patterns similar to joint-controlled confined hypogenic caves (Figure 10). The authors do not actually suggest that dissolutional forms found in epigenetic caves that mimic those from hypogenic caves actually originated in flank margin caves. The point to make is that the inheritance argument for either eogenetic or confined hypogenic caves carries its own problems if extended too far.

What is much more likely is that the similarity of bedrock dissolutional forms found in caves originating in the eogenetic, mesogenetic and telogenetic environments is the result of similar hydrodynamic flow conditions. That argument is easy to make for flank margin caves and confined hypogenic caves. Both environments involve slow movement of waters in a non-competitive flow setting. The dissolution envisioned for confined hypogenic cave development is

necessarily slow as described by Klimchouk (2007). If it were fast, then entire soluble rock units would disappear completely in the time span of the creation and exhumation of cratonic basins, even if the aquifer was not incised by erosion and stayed confined the entire time. Evidence exists for substantial removal of both gypsum and halite units in basins, followed by regional collapse and brecciation (Johnson, 1997). Completely opposite is the formation of flank margin caves, where entire caves in the Bahamas, with tens of thousands of cubic meters of void space, developed in the time span of twelve thousand years. Time, it would appear, is less important than the geochemical and hydrodynamic properties of the fluid involved. If slow flow occurs in both cases, then the dissolutional potential would need to act inversely to the time available: low dissolutional potential needs a lot of time to create the observed bedrock forms (confined hypogenic caves), high dissolutional potential needs little time to make the same forms (flank margin caves).

What does this say about the epigenic cave environment? Under what conditions do these caves achieve a slow flow regime, and for how long? If such a regime can be established, what are the likely geochemical conditions of that flow regime? The classic example has been floodwaters in carbonate caves. Under flooding conditions, epigenic caves, which do not have floodplains, experience exceptional increases in head, and high flow velocities (Palmer, 1972). The flooding described by Palmer (1972) is of input floods, in which the cave is confronted with more

incoming water than it can handle, and hydraulic head relative to the discharge point remains high. Epigenic caves also experience output floods, where hydraulic heads decrease, or even reverse, as a result of flood-induced rising base level. During an output flood, the epigenic conduit can become essentially stagnant. Water flow can become extremely slow for hours or even days at a time. The water can be remarkably aggressive as its initial residence time is short, and in carbonate systems, organic decay can help maintain that dissolutional potential. Output flooding of caves is a potentially repetitive occurrence for epigenic caves, and may therefore leave a recognizable dissolutional overprint.

Some of the epigenic caves reinterpreted by Klimchouk (2007) as rejuvenated confined hypogenic caves, such as Mystery Cave, Minnesota, or Skull Cave, New York, are each caves that experience significant output flooding. Mystery Cave was described by Klimchouk (2007, p. 70): “The cave has perfectly expressed the morphological suite of rising flow as described in Section 4.4 (see Plates 1-E; 6-I; 7-C; 9-D and 14 for photographs of the cave’s hypogenic morphology).” Mystery Cave is a meander cutoff cave located in a meander neck of the South Branch of the Root River (Myroie, 1991). When the river floods, the hydraulic head between the two sides of the meander diminishes, and the cave floods. The initial flooding, and the drain phase as the flood recedes, may occur under high head conditions and create fast flow conditions inside the cave. But in between those stages, once the base level rise is

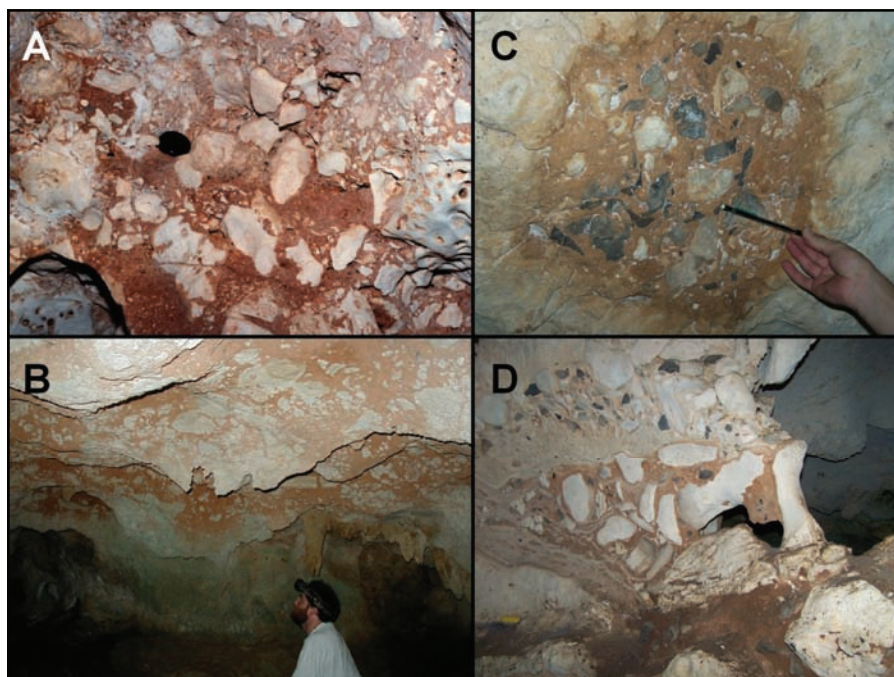


Figure 8. Paleosol breccias in flank margin caves. A) Cueva del Parajos, Isla de Mona, Puerto Rico (6 cm lens cap in left-center for scale). B) Red Roof Cave, Long Island, Bahamas. C) Beach Cave, San Salvador Island, Bahamas. D) 1702 Cave, Crooked Island, Bahamas. Images A and C show a paleosol infiltrate from the surface. Image B shows a paleosol layer between two eolianites. Image D is a complex feature combining elements of both paleosol infiltrates and paleosol layers (10 cm ruler in lower left for scale). In all four images the paleosol deposits are cut by smooth, phreatic dissolution surfaces that ignore the heterogeneity of the paleosol material.

complete, then the cave water is relatively stagnant. Skull Cave is a pre-glacial epigenic cave system which has had its outlets blocked by glacial drift (Kastning, 1975; Hesler, 1990). Based on map pattern, and a single photograph from Palmer (2001, his Figures 10 and 11, respectively), Klimchouk (2007, p. 34) interpreted the map and photograph: “An alternative possibility is that clusters of hypogenic transverse mazes, inherited from the confined stage, are encountered by invasion stream passages during the subsequent unconfined stage. A photograph of a typical floodwater (supposedly) passage on the cited figure shows a hole with a smooth edge in the bedrock floor, which is typical example of a feeder (riser) in hypogenic transverse caves (see next section, photos F, H and I on Plate 3).” (The quote incorrectly conflates Palmer (2001) Figures 10 and 11 as a single figure). As with Mystery Cave, the fill and drain phases may have high flow velocities, but the middle stagnant stage does not. Skull Cave shows modern flood debris over 25 m above the active stream passages in the cave. This flood debris contains human-manufactured items such as tires, tin cans and plastics, indicating that flooding is a modern activity. If both caves were truly hypogenic in origin, one would expect that they would contain extensive passages not explicable by the current hydrology, and that adjacent, unconnected

hypogenic cave fragments would exist in the surrounding area.

One of the consequences of slow flow or stagnant conditions in any cave, compared to the usual rapid turbulent flow found in epigenic stream caves, is the opportunity for convective flow to occur. In stagnant cave waters, vertical convection can be driven by density contrasts within the fluid established by thermal or solute differences (Ford and Williams, 2007). Flow that mimics vertical convection can occur from vertical hydraulic pressure injection of water from overlying or underlying units, or by evolution and entrainment of gas bubbles (Palmer, 2007). Vertical convective flow could be responsible for the many cupolas, ceiling channels, and related dissolution features seen in both confined hypogenic caves and flank margin caves. The methods of convection are probably different. Shallow flank margin caves are unlikely to show thermal convection, whereas mesogenetic confined hypogenic caves, given their pressure regime at depth, are unlikely to evolve gases. Provided there is dissolutional aggressivity to the water, vertical convective flow, if occurring at discrete sites, would be expected to produce cupolas and ceiling channels. Cupolas would occur where the convective cell rise and return limbs were close and

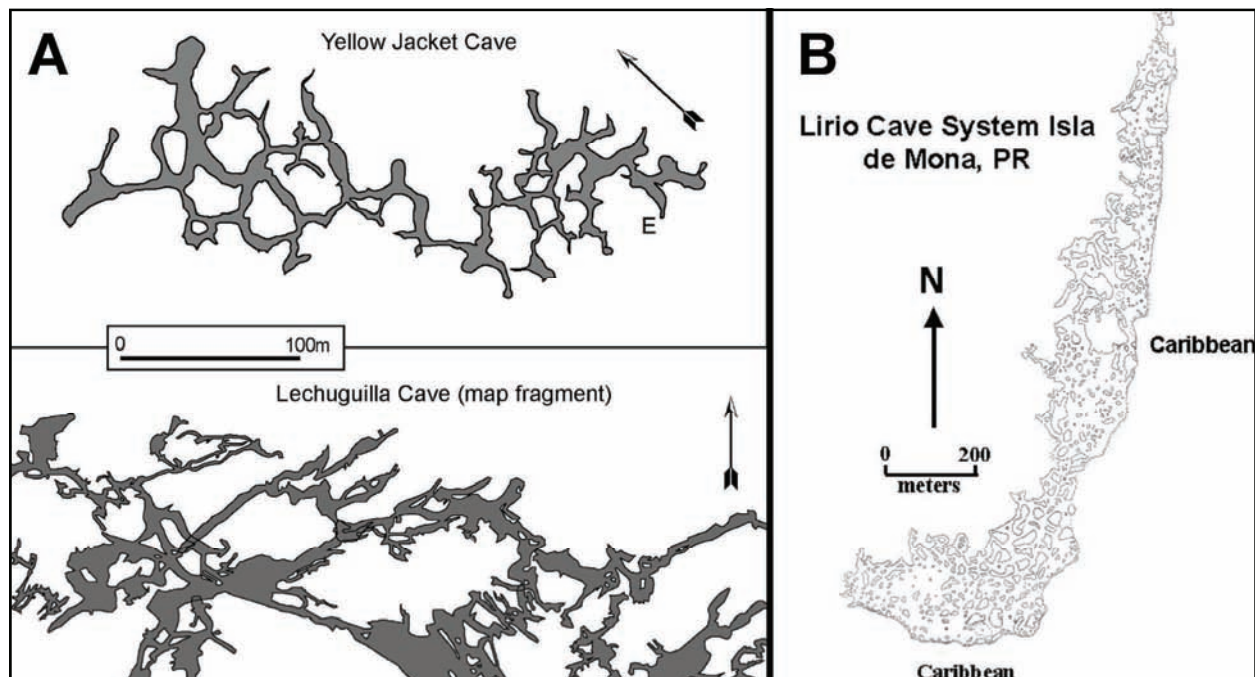


Figure 9. A) Maps of two Guadalupe Mountain, New Mexico caves from Klimchouk (2007, Figure 16). B) Map of Lirio Cave from Isla de Mona, Puerto Rico. Polygonal, anastomotic and related cave patterns (spongework and ramiform patterns of Palmer, 1991) have been suggested as diagnostic for confined hypogenic caves as in A by Klimchouk (2007), but flank margin caves, as in B, commonly show similar patterns. Flank margin caves, being tied to the fresh-water lens margin, only show the 3-D complexity found in confined hypogenic caves if sea level change has been slow and gradual.

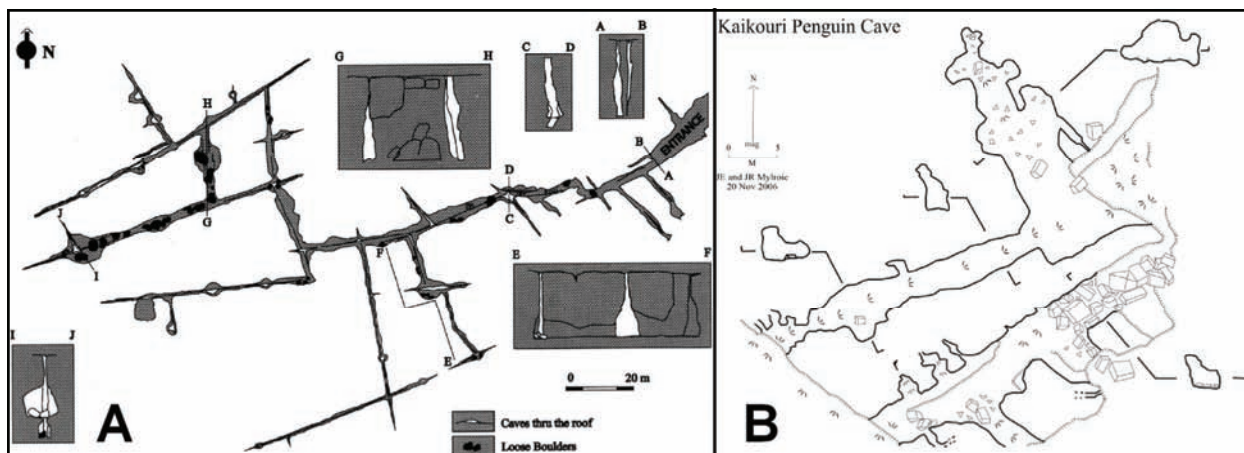


Figure 10. A) Map of Jabal Al Qarah Cave, northeastern Saudi Arabia (from Klimchouk, 2007, Figure 37). Map of Kaikouri Cave, South Island, New Zealand (from Mylroie et al., 2008). Kaikouri Penguin Cave is a flank margin cave developed in telogenetic rocks, as a result primary porosity is low and cave patterns are joint and fracture controlled, similar to confined hypogenic caves, as in A.

parallel. Ceiling channels would form where the rise and return limbs were separated by a lateral difference such that the upper horizontal leg was against the cave roof.

In epigenic caves, stagnant flood conditions could allow vertical convection to occur, to generate dissolutional features that mimic those associated with flank margin caves or confined hypogenic caves. Stagnant waters in a cave create a condition of non-competitive water flow, so bedrock dissolution would be expected on all surfaces to produce maze-like passage development and thin wall partitions. As a result of allogenic sediment transport into the cave, epigenic caves also support paragenesis, which forces water flow against cave roofs and walls to create a variety of dissolutional forms that could appear hypogenic. Paragenesis is not a major factor in confined hypogenic caves, or in flank margin caves, as allogenic sediment transport does not occur during cave genesis. Any sediment load in these two cases would be autogenic material left as an insoluble residue. In flank margin caves in Bahamian eolianites, the limestones are 99.9% CaCO_3 , so insoluble sediment load is not a factor in armoring floor surfaces.

Conclusions

The argument that epigenic caves, especially maze caves, are primarily relict confined hypogenic caves translated to the telogenetic environment is questionable for two reasons. First, hypogenic caves formed in response to a set of hydrologic and geochemical conditions different from the epigenic karst environment. To assume that they would be positioned and oriented to make a significant contribution to epigenic flow routes is not viable. As previously discussed,

models that explain epigenic maze cave development appear viable, and the criticisms of those models appear invalid. Second, the diagnostic features stated to be solely the result of confined hypogenic conditions are not unique to the hypogenic environment. Flank margin caves developed in eogenetic rocks, which have never been confined, display similar dissolutional features. Arguments can be presented to generate these same diagnostic features within the epigenic realm by stagnant floodwaters. Being phreatic in an epigenic system is not sufficient, it must be phreatic and stagnant to prevent lateral shear of vertical convection. The vast differences in geologic time between the long duration necessary for confined hypogenic cave development, and the comparatively short time necessary for flank margin cave development indicate that hydrodynamic and geochemical conditions are a greater control of cave morphology than simple duration of hydrological activity.

That confined hypogenic caves exist, and that they are introduced in relict form to the telogenetic environment is most certainly true. That they are intersected by epigenic caves is also true. What is not as apparent is that they form a major class of cave within the epigenic cave environment, and that they play a major role in epigenic cave development. It is not true that they can be uniquely identified by a suite of dissolutional features.

The major hypogenic caves of the United States, such as in the Guadalupe Mountains of New Mexico or the Black Hills of South Dakota, do not participate in the current epigenic karst water flow of those regions.

Their large explorable extent is directly related to their lack of participation in local hydrology, such that they

are not overprinted, segmented, and infilled by epigenic stream caves. The climatic regime of the American West has created a preservational bias for hypogenic caves that has helped overcome the explorational bias that limited hypogenic understanding until the 1980's. The confined hypogenic cave model is an important new view of cave development at depth. The degree to which relict hypogenic caves exist in the epigenic environment, and influence epigenic cave genesis and function, is still open to debate.

Acknowledgments

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Appendix - Discussion of terminology

Klimchouk (2007) argues that hypergenic or hypergene is more appropriate than epigenic for traditional stream caves coupled to the surface hydrology, as those terms are antonyms for hypogenic and hypogene, respectively. He notes that those terms have been used in Eastern Europe. However, correct use of the term "hyper" means over, above or beyond, and in this case the frame of reference is the earth's surface, hypergenic is therefore above or over the earth's surface. The term hypergene is given in the Glossary of Geology as a synonym for supergene; hypergenesis as: "A term introduced by Fersman (1922) and persisting to the present day in Russian geology, for surficial alteration (weathering) of sedimentary rocks. Little used in the English literature...." (Neuendorf et al., 2005, p. 314). Ford and Williams (2007, p. 3-4) report that the European literature has used the terms "hyperkarst" and "hypokarst" as subcategories of "endokarst", where endokarst is karst developed underground. Hyperkarst is then underground dissolution by circulating meteoric waters, and hypokarst underground dissolution by juvenile or connate waters. These terms are completely relative as regards vertical position, which could vary widely from place to place, and the terms conflict in North America with the meaning of hypergene as geologic activities on the earth's surface. So while the term hypergenic may be useful for describing karren on the land surface, it is not useful for stream caves in the subsurface that are coupled to surface hydrology. Epigene is defined in the Glossary

of Geology as: "(a) said of a geologic process, or its resultant features, occurring at or near the earth's surface. Cf: hypogene. Syn: epigenic." (Neuendorf et al., 2005, p. 213). The Palmer (1991) choice of the term epigenic seems the correct one. It is also well-established in the literature. A Google® search (January 19, 2009) yields 38,600 responses for the term "epigene". Narrowing the search to "epigene cave" yields 9,560 responses; "epigenic cave" yields 29,400 responses. "Hypergene cave" yields 430 responses; "hypergenic cave" 458 responses.

This review of terminology may be tedious, but hypogenic caves, as discussed in this volume, represent the introduction of important new ideas into the speleological literature. It is useful that such new ideas, when defined, can be seen to fit into the continuum of speleological thought. Changing epigenic cave to hypergenic cave would not only be incorrect usage and abandon a well-established term, it would also be in conflict with current North American usage.

References

- Alexander, E.C., K. Barr, and S. Alexander. 2008. Goliath's and Mystery Caves Minnesota: Epigenic modifications and extension of preexisting hypogenic conduits. *Geological Society of America Abstracts with Programs* 40 (6): 343.
- Bakalowicz, M.J., D.C. Ford, T.E. Miller, A.N. Palmer, and M.V. Palmer. 1987. Thermal genesis of dissolution caves in the Black Hills, South Dakota. *Geological Society of America Bulletin* 99: 729-738.
- Carew, J.L., and J.E. Mylroie. 1995a. A stratigraphic and depositional model for the Bahama Islands. In *Geological Society of America Special Paper 300, Terrestrial and Shallow Marine Geology of the Bahamas and Bermuda*, ed. H.A. Curran and B. White 5-31. Denver: Geological Society of America.
- Carew, J.L., and J.E. Mylroie. 1995b. Quaternary tectonic Stability of the Bahamian Archipelago: Evidence from fossil coral reefs and flank margin caves. *Quaternary Science Reviews* 14: 144-153.
- Carew, J.L., and J.E. Mylroie. 1997. Geology of the Bahamas. In *Geology and hydrogeology of carbonate islands, Developments in Sedimentology 54*, ed. H.L. Vacher and T.M. Quinn, 91-139. New York: Elsevier Science Publishers.
- Chen, J.H., H.A. Curran, B. White, and G.J. Wasserburg. 1991. Precise chronology of the last interglacial period: ^{234}U - ^{230}Th data from fossil coral reefs in the Bahamas. *Geological Society of America Bulletin* 103: 82-97.

- Choquette, P.W., and L.C. Pray. 1970. Geologic nomenclature and classification of porosity in sedimentary carbonates. *American Association of Petroleum Geologists Bulletin* 54: 207-250.
- Fersman, A.E. 1922. *Geokhimiia Rossii*. St. Petersburg: Nauchnoe Khimicheskoe-Tekhnicheskoe izdatel'stvo.
- Ford, D.C. 1995. Paleokarst as a target for modern karstification. *Carbonates and Evaporites* 10 (2): 138-147.
- Ford, D.C., and P.W. Williams. 1989. *Karst geomorphology and hydrology*. New York: Chapman and Hall.
- Ford, D.C., and P.W. Williams. 2007. *Karst hydrogeology and geomorphology*. Chichester: John Wiley & Sons.
- Frank, E.F., J. Mylroie, J. Troester, E.C. Alexander, and J. L. Carew. 1998. Karst development and speleogenesis, Isla de Mona, Puerto Rico. *Journal of Cave and Karst Studies* 60 (2): 73-83.
- Gonzalez, L.A., H.A. Ruiz, B.E. Taggart, A.F. Budd, and V. Monell. 1997. Geology of Isla de Mona, Puerto Rico. In *Geology and hydrology of carbonates islands, developments in sedimentology* 54, ed. H. L. Vacher and T. M. Quinn, 327-358. New York: Elsevier Science Publishers.
- Hesler, D.J. 1990. *A hydrologic study of the Knox-Skull Cave System, Albany County, New York*. Bulletin 4. New York: New York Cave Survey.
- Hose, L.D., and J.A. Pisarowicz, ed. 2000. The caves of the Guadalupe Mountains research symposium. *Journal of Cave and Karst Studies* 62: 53-157.
- Johnson, K.S. 1997. Evaporite karst in the United States. *Carbonates and Evaporites* 12 (1): 2-14.
- Kastning, E.H. 1975. *Cavern Development in the Helderberg Plateau, East-central New York*. Bulletin 1. New York: New York Cave Survey.
- Klimchouk, A.B. 2007. *Hypogene Speleogenesis: Hydrogeological and Morphogenetic Perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Krauskopf, K.B. 1967. *Introduction to Geochemistry*. New York: McGraw-Hill.
- Lindgren, W. 1933. *Mineral Deposits, 4th ed.* New York: McGraw-Hill.
- Martin, J.M., E.J. Screaton, and J.B. Martin. 2006. Monitoring well responses to karst conduit head fluctuations: Implications for fluid exchange and matrix transmissivity in the Floridan aquifer. In *Geological Society of America Special Paper* 404, 209-217. Denver: Geological Society of America.
- Meyerhoff, A.A., and C.W. Hatten. 1974. Bahamas salient of North America: Tectonic framework, stratigraphy, and petroleum potential. *American Association of Petroleum Geologists Bulletin* 58: 1201-1239.
- Mylroie, J.E. 1991. Cave Development in the Glaciated Appalachian Karst of New York: Surface-Coupled or Saline-Freshwater Mixing Hydrology? In *Proceedings of the Appalachian Karst Symposium*, ed. E.H. Kastning and M.K. Kastning, 85-90. Huntsville: National Speleological Society.
- Mylroie, J.E., and J.L. Carew. 1995. Chapter 3, Karst development on carbonate islands. In *Unconformities and Porosity in Carbonate Strata: American Association of Petroleum Geologists Memoir* 63, ed. D. A. Budd, P. M. Harris and A. Saller, 55-76. Tulsa: American Association of Petroleum Geologists.
- Mylroie, J.E. 2003. The interaction of hypogenic caves and explorational bias: under representation of cave data: *GEO²* 30 (2-3): 8.
- Mylroie, J.E. 2008. Review Of: Hypogene Speleogenesis: Hydrological and Morphological Perspective. *Journal of Cave and Karst Studies* 70 (2): 129-131.
- Mylroie, J.E., and J.R. Mylroie. 2007. Development of the Carbonate Island Karst Model. *Journal of Cave and Karst Studies* 69: 59-75.
- Mylroie, J.E., J.R. Mylroie, and C.N. Nelson. 2008. Flank Margin Cave Development in Telogenetic Limestones of New Zealand. *Acta Carsologica* 37 (1): 15-40.
- Mylroie, J.E., and J.R. Mylroie. 2008. Diagnostic features of hypogenic karst: Is confined ascending flow necessary? *Geological Society of America Abstracts with Programs* 40 (6): 343-344.
- Mylroie, J.E., and J.R. Mylroie. 2009. Flank margin cave development as syndeositional caves: Examples from the Bahamas. *Proceedings of the 15th International Congress of Speleology* (in press).
- Neuendorf, K.K.E., J.P. Mehl, and J.A. Jackson. 2005. *Glossary of Geology, Fifth Edition*. Alexandria: American Geological Institute.
- Osborne R.L. 2007. The world's oldest caves: How did they survive and what can they tell us? *Acta Carsologica* 36 (1): 133-142.
- Palmer, A.N. 1972. Dynamics of a sinking stream system: Onesquethaw Cave, New York. *National Speleological Society Bulletin* 34 (3): 89-110.
- Palmer, A.N. 1975. The origin of maze caves. *National Speleological Society Bulletin* 37 (3): 56-76.
- Palmer, A.N. 1991. Origin and morphology of limestone caves. *Geological Society of America Bulletin* 103: 1-25.

- Palmer, A.N. 2001. Dynamics of cave development by allogenic water. *Acta Carsologica* 30: 14-32.
- Palmer, A.N. 2007. *Cave Geology*. Dayton: Cave Books.
- Panuska, B.C., J.M. Mylroie, D. Armentrout, and D. McFarlane. 1998. Magnetostratigraphy of Cueva del Aleman, Isla de Mona, Puerto Rico and the species duration of Audubon's Shearwater. *Journal of Cave and Karst Studies* 60 (2): 96-100.
- Polyak, V.J., W.C. McIntosh, N. Güven, and P. Provencio. 1998. Age and origin of Carlsbad Caverns and related caves from $^{40}\text{Ar}/^{39}\text{Ar}$ of alunite. *Science* 279: 1919-1922.
- Schindel, G.M., S.B. Johnson, and E.C. Alexander. 2008. Hypogene processes in the Balcones fault zone Edwards aquifer in south-central Texas, a new conceptual model to explain aquifer dynamics. Geological Society of America Abstracts with Programs 40 (6): 344.
- Vacher, H.L., and J.E. Mylroie. 2002. Eogenetic karst from the perspective of an equivalent porous medium. *Carbonates and Evaporites* 17 (2): 182-196.

VOLCANOGENIC KARSTIFICATION: IMPLICATIONS OF THIS HYPOGENE PROCESS

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Abstract

Numerous geologic conditions facilitate a setting for hypogenic karst processes to evolve, including the interaction of igneous rocks and groundwater in carbonate rocks. Hydrothermal, deep-seated karst is documented, but the mechanisms are not always applied in geologic evaluations of the karst. Volcanic activity provides conditions that can effectively dissolve large voids deep below the Earth's surface. Volcanogenic karstification relies on four components to initiate and develop deep, subsurface voids: 1) thick carbonate strata, 2) preferential groundwater flow-paths (fractures), 3) volcanic activity that releases acids, and 4) flux of groundwater through the system. The order of occurrence (from 1 to 4) is critical to develop the karst. Components 1, 2, and 4 are common to almost all karst, but component 3 can accelerate dissolution processes in volcanogenic karst systems (VKS). High fluxes of carbon dioxide and/or hydrogen sulfide from volcanic rocks create hyper-aggressive subsurface conditions that rapidly dissolve carbonate rocks. Volcanogenic karstification has produced the Earth's two deepest underwater cave systems, Pozzo del Merro (Italy) and Sistema Zacatón (Mexico). Studies of these processes require evaluation of systems currently active on or near the surface (directly accessible by humans or robots). Volcanogenic karstification can produce deep solutional porosity and high permeability where older carbonate rocks are juxtaposed to younger volcanic rocks. VKS examples are discussed and some potential VKS identified.

Introduction

Hypogene speleogenesis has been defined several ways in karst literature. Palmer (1991, 2007) states that hypogenic karst is based on the source of the acid dissolving the soluble rock. Carbonic acid derived from soil and atmospheric carbon dioxide (surface sourced) is involved in epigenic karstification process whereas acids from deep seated sources are defined as hypogenic. These include volcanic gases of CO₂ and H₂S, reduced sulfur and carbonic acids from petroleum laden rocks, and other sources. Klimchouk (2007) favors a more broad approach to hypogene

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speleogenesis, incorporating not only deep-seated geochemical factors to the process, but including most settings with rising waters independent of the source of acid.

Most of the karst systems used to study and model hypogene speleogenesis are currently vadose caves, while the processes used to form most hypogene caves occur in phreatic conditions. A future paradigm that focuses on applying phreatic karst systems as the standard to model the geologic mechanisms forming karst is needed. The two deepest explored phreatic caves in the world, Pozzo del Merro, Italy, and El Zacatón, Mexico, (Table 1) (Gary et al., 2003; Knab, 2008) are both hypothesized as developing in hypogene settings, specifically as a subset defined as a volcanogenic karst system (VKS) (Gary and Sharp, 2006, 2008).

Volcanogenic karstification occurs when a groundwater system in soluble rocks interacts with volcanic activity in the subsurface. This interaction can produce conditions favorable for intensive karstification in focused geographical areas and at considerable depth. Although surface expression of karst features can be present, it is possible that no recognizable karst development exists on the surface above areas with volcanogenic karstification. This can make identification and investigation of VKS difficult and limited scientific literature exists on the subject.

The geologic processes involved in development of VKS are well documented, and include fundamental components common to all karst systems. These are:

1. Sufficient thickness of carbonate strata.
2. Preferential groundwater flowpaths (fractures).
3. Flux of groundwater (pump for mass transfer of dissolved rock).
4. Volcanic activity – influx of proton (H⁺). CO₂ and/or H₂S accelerated kinetics and thermodynamics.

Components 1-3 are common to all active karst systems, and some form of acid must be present.

Table 1. *Deepest Underwater Caves in the World (compiled by Knab (2008); Zacatón data from Gary et al., 2008).*

Cave Name (country)	Depth (m/ft.)	Method of exploration
Pozzo del Merro (Italy)	392 / 1286	Telenaute ROV Prometheus
El Zacatón (Mexico)	319 / 1046	DEPTHX AUV
Fontaine de Vaucluse (France)	315 / 1033	Telenaute ROV
Bushmansgat (South Africa)	282 / 927	Nuno Gomes (SCUBA)
Crveno Jezero (Croatia)	281 / 921	ROV

Component 4 is unique to VKS because the major source of acid (H^+) is derived from the interaction of groundwater with volcanic rocks at depth. Klimchouk (2007) modified a conceptual model of groundwater flow from Tóth (1999) to show regional settings of hypogenic and epigenic karst systems (Figure 1) and includes interaction of a magmatic zone adding CO_2 and heat to the groundwater system. The addition of heat is inherent to VKS that create hydrothermal conditions that can provide upward convection of fluids through pre-existing fractures and increase the flux of groundwater through a developing karst system. Dublyansky (2000b) discusses karst development by geothermal waters with the effects of rising thermal water, H_2S and mixing with H_2S and CO_2 to accelerate dissolution. A general review of the interaction of volcanic activity and karst is given by Salomon (2003) who defines several cases where different landforms result in this geologic setting. While Salomon mentions CO_2 emissions from volcanic settings accelerating dissolution, the focus of this discussion relates to the variety of morphologies and not specifically to karst processes, but also structural deformation related to volcanic terrain.

Integration of volcanogenic karstification as a specific, defined type of hypogene speleogenesis is useful when characterizing karst systems on a global scale. Several karst systems have been identified directly with volcanic activity in modern settings, and others may have been influenced by volcanism during earlier time periods. Such systems may have features and characteristics that are difficult to relate directly to volcanogenic processes, but applying the VKS model improves interpretation of the geologic framework when modeling the modern hydrogeologic system.

Sistema Zacatón – VKS type locale

We analyze a well-developed karst system, Sistema Zacatón, with observable major features and settings common to this karst type to define VKS common characteristics. Sistema Zacatón is a deep, phreatic karst system in northeast Mexico which formed in the Pleistocene. The system remains active in present time, although the most significant phase of karstifica-

tion occurred during periods when nearby volcanic activity produced active lava flows and cinder cone volcanoes. The primary features of Sistema Zacatón are the large water-filled sinkholes (cenotes) that are oriented in two primary linear patterns: north-south and east west at the southern end of the chain of cenotes. This linear orientation reflects preferential formation of the cenotes along fractures originating in the early Cenozoic (Gary and Sharp, 2006). These features are large enough to be seen from satellite imagery (Figure 2). The geologic setting of Sistema Zacatón includes mid-late Cretaceous argillaceous limestone, Laramide related uplift and fracturing, mid-Cenozoic intrusive igneous bodies, and Pleistocene extrusive volcanic activity (Figure 3) (Suter, 1987; Camacho, 1993; Goldhammer, 1999; Vasconcelos and Ramírez Fernandez, 2004; Ramírez Fernandez et al., 2007). These components together provide the unique combination which initiated and accelerated karstification of this deep hydrothermal cave system.

The generalized model of groundwater flow including interaction with volcanic rocks is shown in Figure 4. This resembles the idealized setting shown in Figure 1 by Klimchouk. Dublyansky (2000a) defines settings and features characteristic of hydrothermal karst systems, and Sistema Zacatón displays this model extremely closely. A major trait of hydrothermal systems is that of exaggerated zones of carbonate dissolution and zones of carbonate precipitation. Sistema Zacatón has some of the largest explored karst phreatic voids in the world with over 2.7 million cubic meters of water in the four largest cenotes, representing massive zones of dissolution. These underwater caves have been explored by humans and recently mapped in great detail by the DEPTHX (DEep Phreatic THERmal eXplorer) project sponsored by NASA (Figure 5) (Gary, 2007; Gary et al., 2008). At least 4 phases of travertine precipitation occurred at Sistema Zacatón, including large-scale hot spring deposition pre-dating cenote opening to the surface and subsequent closing of some cenotes in the form of travertine lids. The first stage of travertine contains fossils of late-Pleistocene mammoth bones indicating the cenotes opened to the surface in recent geologic

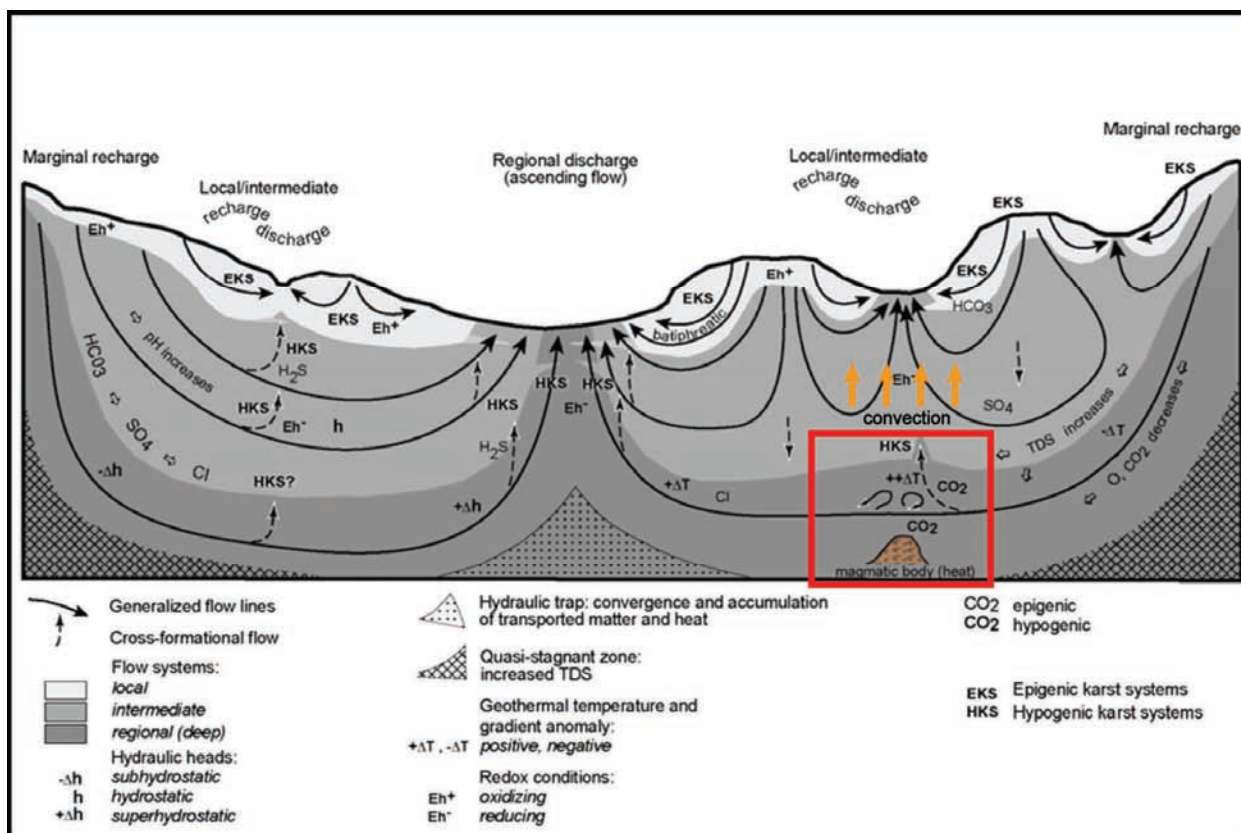


Figure 1. Idealized groundwater flowpaths and basic geochemical interactions adapted by Klimchouk (2007) from Tóth (1999). Klimchouk shows the interaction of a deeply placed magmatic body (red box) with the enrichment of CO₂ and heat to groundwater. Convection has been added to this model and may contribute to focused dissolution and mass transport in a developing volcanogenic karst system.

time (Gary et al. 2007; Gary et al. 2006). The second stage involves sealing over open cenotes as water supersaturated with calcite off-gasses CO₂ at the surface. Geophysical investigations of this phenomenon are discussed in Gary et al. (2006), and Gary et al. (2009, in press). The third and fourth phases involve broad overland travertine deposition from springs and speleothems precipitated in dry caves formed within a matrix of travertine from phase one. An additional attribute found in the phreatic zones at Sistema Zacatón are the diverse microbial habitats of bacteria and Archaea that are similar to those in other hydrothermal settings that cycle sulfur and carbon in the underwater environment (Sahl and Spear, 2007).

Geochemical and isotopic evidence

The features and characteristics of Sistema Zacatón, including the unique ability to access deep segments of the karst aquifer for direct study, make it an ideal system to model VKS. A variety of initial investigations have been conducted at Sistema Zacatón to characterize the karst, including geochemical and isotopic studies. Complete results of these studies are

not presented in this general overview of Sistema Zacatón, and initial data were previously published (Gary and Sharp, 2006), or are in the publication process. However, some convincing evidence exists with physical and chemical profiles, limited carbon-13 and strontium isotopes, dissolved CO₂ values, some trace elements, and extensive microbial activity.

Geochemical profiles were collected using two methods at Sistema Zacatón: 1.) multi-parameter sonde profiles (typically with a Hydrolab Series 4a instrument); and 2.) a titanium multi-parameter instrument developed by HydroTech as part of the DEPTHX project (Gary et al., 2008). Both methods revealed similar geochemical conditions in the main cenotes of Zacatón, Caracol, Verde, and La Pilita and results of over 1.5 million data points are presented in Figure 6. With the exception of Verde, the deeper cenotes are all extremely homogeneous with respect to temperature, pH, dissolved oxygen, and specific conductance. A sensor on the DEPTHX probe was designed to measure sulfide at a frequency of 1 Hz, but was only able to measure relative variations. True

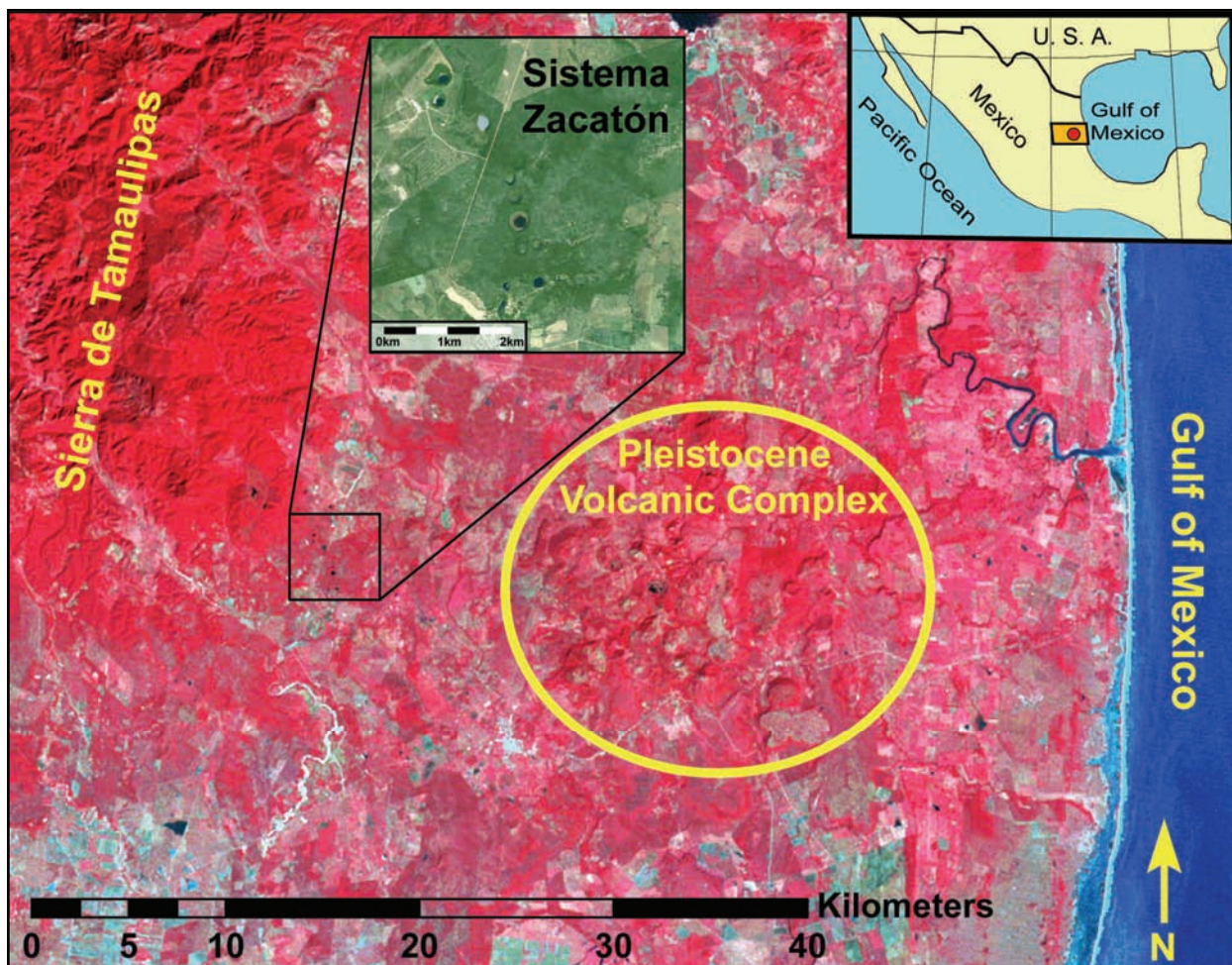


Figure 2. Landsat image of the area including Sistema Zacatón (upper right inset). The larger scale image shows the karst area formed between the Sierra de Tamaulipas to the west and Volcanic Complex of Villa Aldama to the east. Shapes of cinder cone volcanoes and water-filled craters are evident in the volcanic field. Major karst features are isolated in the area bounded by the yellow box. Spatial resolution of the data is 30 meters/pixel indicating the karst features are large (Data from North American Land Coverage, 1980).

sulfide values are not reported in this dataset. The temperature data show Zacatón (30.1° C), Caracol (29.6° C), and La Pilita (31.6° C) to remain constant from top to bottom. These values are significantly above the mean annual surface temperature of the region (24° C) indicating a constant influx of geothermal energy to the groundwater. The uniform temperature profiles are attributed to convective mixing in the water column. The dissolved oxygen concentrations in the three deeper cenotes are just above 0 mg/L throughout the water columns except for some slightly higher values at the surface due to entrainment of air bubbles into the water from the DEPTHX probe thrusters. This lack of oxygen results from accelerated microbial consumption of oxygen in the water, likely by sulfur oxidizing bacteria (Sahl et al., 2007). The pH of the system is low for a carbonate buffered system with profiles of the deeper cenotes as meas-

ured by DEPTHX ranging from 6.6 to 6.9, and values as low as 6.2 in some water samples collected from the cenote Caracol. The low pH is reflective of the very high levels of CO₂ dissolved in the water discussed below. Specific conductance data reflect the convectively well-mixed water column relative to bulk ion geochemistry. Relative sulfide values show some variation with depth and concentrations have been measured in the water of Zacatón and Caracol in ranges from 0.1 – 3.1 ppm using a Chemetrics spectral photometer.

Verde reflects quite different geochemical profiles from these five parameters due to the unique morphology of the cenote. Verde to be the shallowest cenote (Figure 5) and has an extremely flat floor atypical to one of a collapse sinkhole structure. This flat floor is hypothesized to be a travertine “barrier” formed when

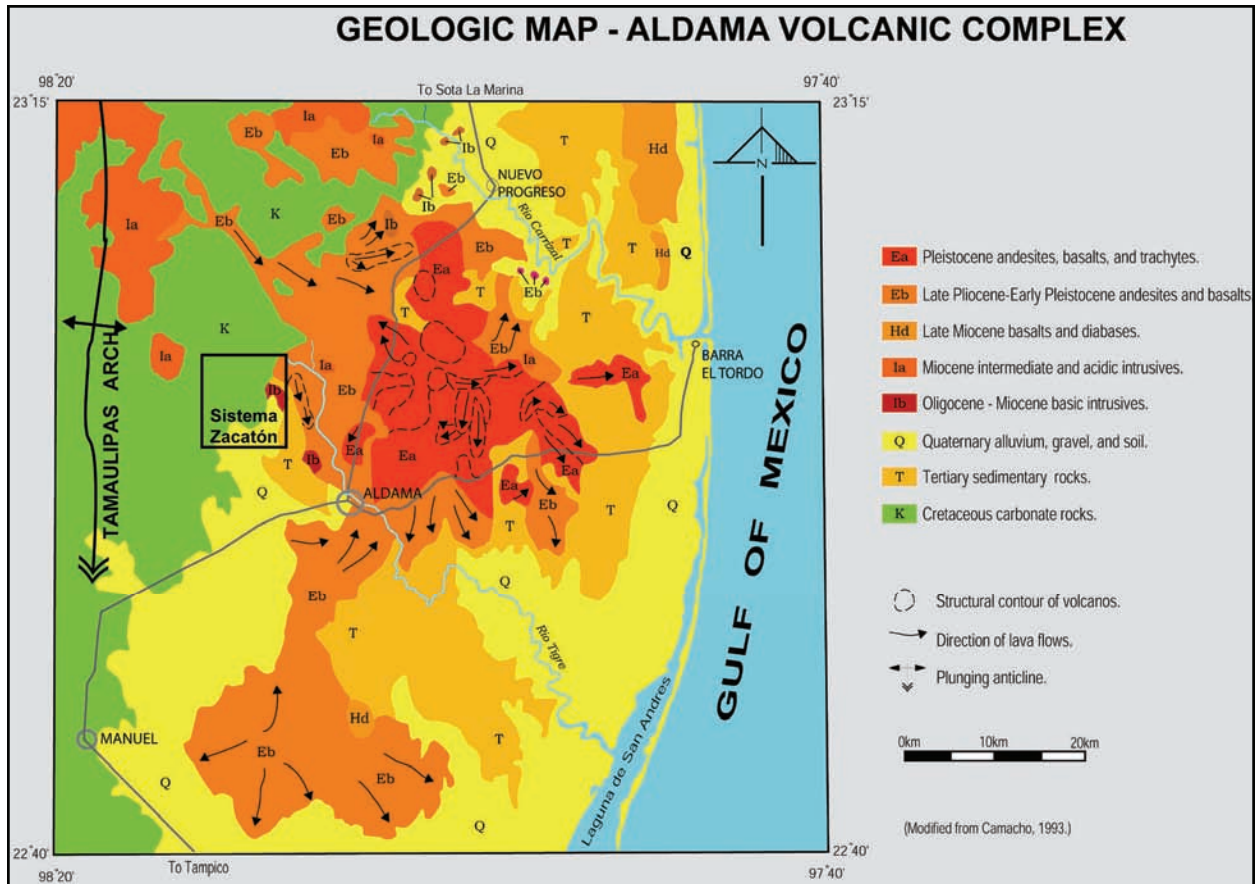


Figure 3. Geologic map of the Villa Aldama Volcanic Complex shows extent and relative ages of igneous rocks east of the study area of Sistema Zacatón (study area). The Sierra de Tamaulipas to the west is part of the large domal anticline of the Tamaulipas Arch (modified from Camacho, 1993).

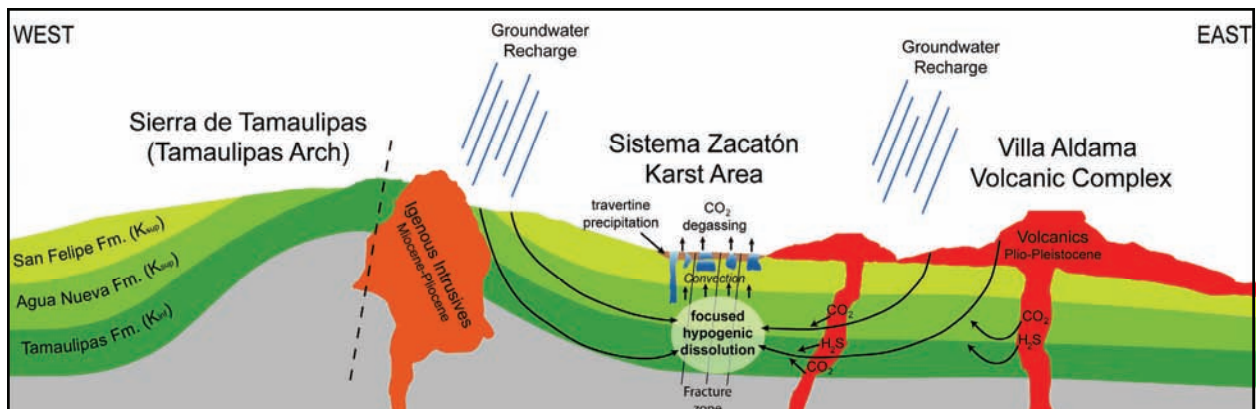


Figure 4. Generalized cross-section of groundwater flowpaths to Sistema Zacatón shows enrichment of CO₂ and H₂S from the Pleistocene volcanic complex on the right. The intrusive rocks at the left have had little geochemical effect on groundwater compared to extrusive volcanic rocks, but create a geographic high which contributes a primary source of recharge to Sistema Zacatón. The karst features have formed in a fracture zone which created preferential flowpaths during inception of dissolution process. Heat sourced from volcanic activity drives convection in the groundwater system resulting in upward stopping and eventual collapse of cenotes to the surface.

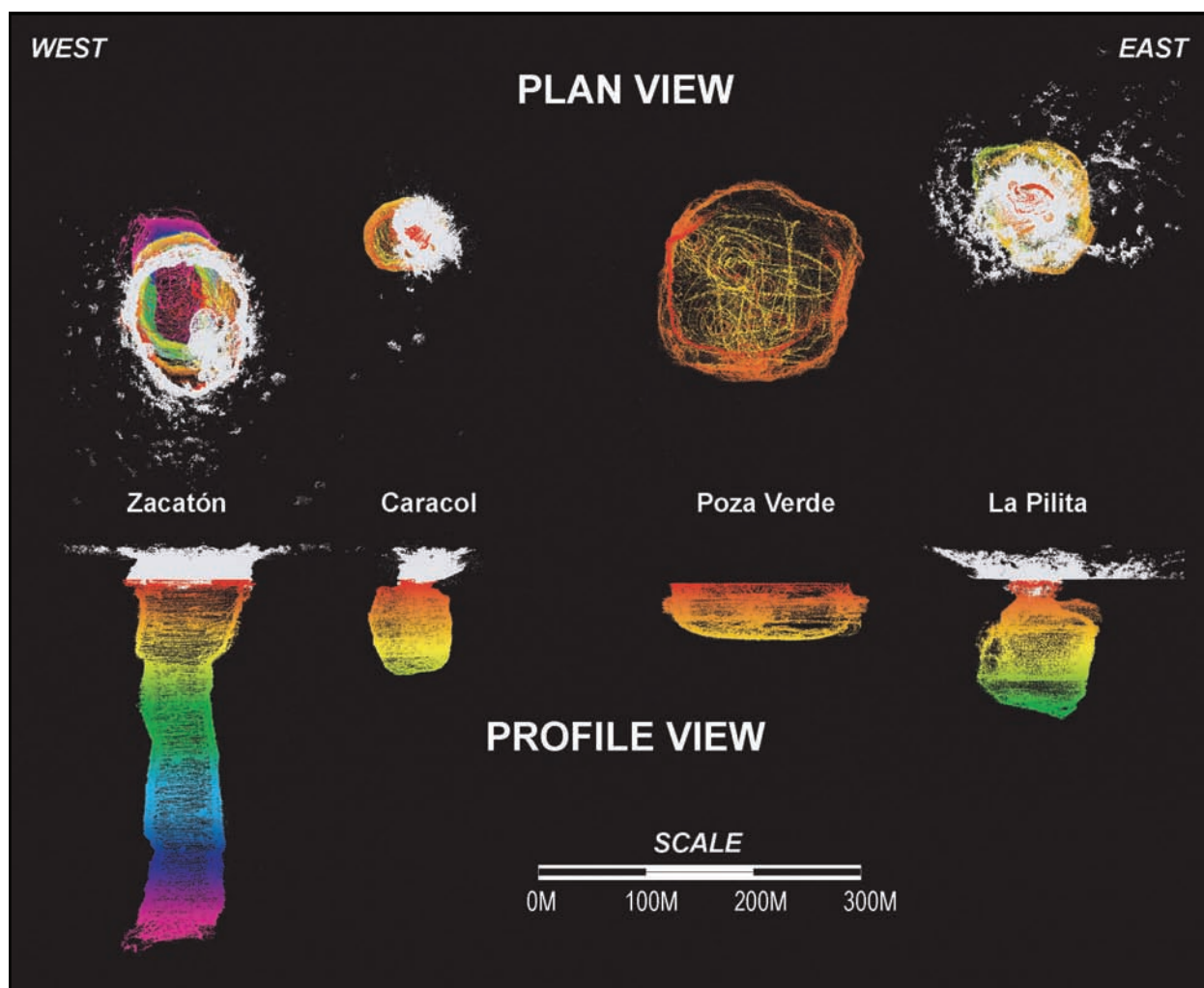


Figure 5. Plan and profile view of the southern water-filled sinkholes of Sistema Zacatón show the depth of Zacatón at over 300 meters. This map was generated from underwater sonar data collected from the DEPTHX probe and laser scanning data above the water surface. The white point cloud data points represent walls above the water and the color scaled points represent different water depths within the phreatic zone. The sinkhole Zacatón is the second-deepest explored phreatic cave in the world. All data are to scale and geographic position (Gary et al., 2008).

paleo-water levels persisted 45 meters lower than present day levels isolating a perched body of water disconnected from the deeper, hydrothermal groundwater system (Gary and Sharp, 2006). The resulting geochemistry in Verde has thermoclines, chemoclines, higher levels of dissolved oxygen, higher pH, and lower water temperatures (Figure 6) that vary seasonally with climate.

Analysis of the water in the cenotes of Sistema Zacatón reveals a calcium carbonate signature typical of most karst waters. However, very high levels of dissolved carbon dioxide have been measured in the deeper cenotes, ranging from $\log \text{PCO}_2 = -1.66$ to $\log \text{PCO}_2 = -0.86$. Typical values in the water column of the cenote Caracol have concentrations of $\log \text{PCO}_2 =$

-1.35 . These values are many times higher than water at equilibrium with the atmosphere ($\log \text{PCO}_2 = -3.5$) and explain the very low pH values. The source of the CO_2 is hypothesized to be a combination of microbial and volcanogenic origin, and ^{13}C isotopes of total dissolved inorganic carbon (TDIC) reveal values clustered around -11‰ PDB . This value fits within a mixing model with end members of TDIC of 0‰ PDB (volcanogenic) and -30‰ PDB (biogenic). Strontium isotopes also indicate groundwater interacts with mantle derived rocks with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios lower than would be expected strictly from contact with Cretaceous limestone (Gary and Sharp, 2006).

The geologic, morphologic, and geochemical evidence strongly support the hypothesis that Sistema Zacatón

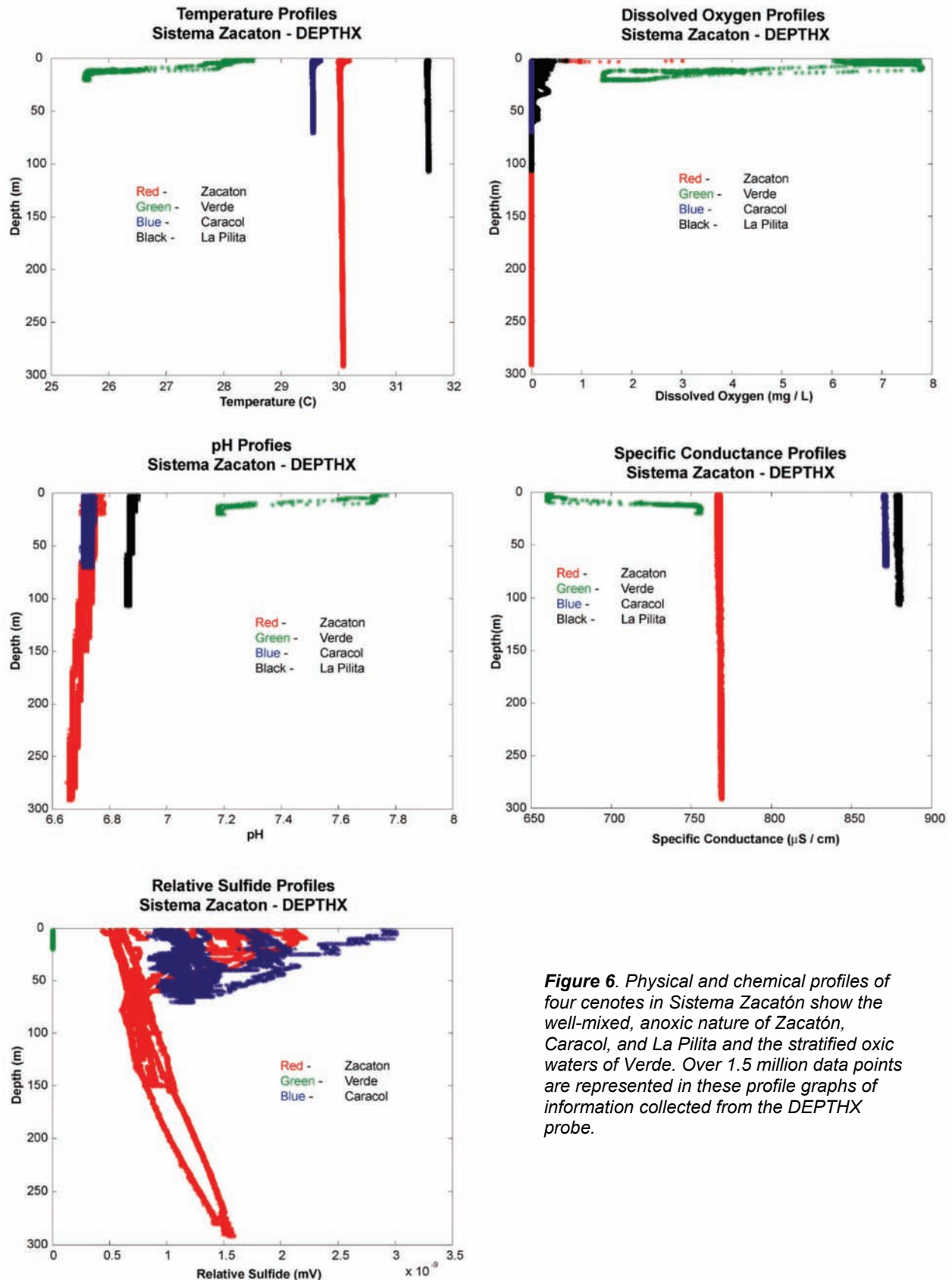


Figure 6. Physical and chemical profiles of four cenotes in Sistema Zacatón show the well-mixed, anoxic nature of Zacatón, Caracol, and La Pilita and the stratified oxic waters of Verde. Over 1.5 million data points are represented in these profile graphs of information collected from the DEPTHX probe.

is a VKS. The unique nature of this karst system, including the extreme expression of the features and localized karst development makes it ideal as a model system for volcanogenic karstification as defined here.

Other modern (active) VKSs

There are a number of other modern or active VKS. These include Pozzo del Merro, Turkish Obruks, Mammoth Hot Springs, and Cueva de Villa Luz.

Pozzo del Merro, Italy

Pozzo del Merro, the deepest underwater cave in the world, developed in geologic conditions similar to those at Sistema Zacatón (Gary et al., 2003). Pozzo del Merro extends 392 meters below the water table, which is 50 meters below the land surface. It was explored in 2002 by a remotely operated vehicle (ROV), where a floor was encountered at the bottom, but with ongoing lateral passage (Caramana, 2002). This region of Italy includes Triassic and Cretaceous limestones that have been structurally altered in the area of the Cornicolani Mountains as a part of the central Apennine fold-thrust belt. This resulted in faulting and fracturing during the Pliocene and Pleistocene, which created preferential groundwater

flowpaths to circulate deep karst water (Billi et al., 2007). Pleistocene volcanic activity accelerated the karst processes here so that dissolution occurred at great depths (Figure 7). Salvati and Sasowsky (2002) noted enrichment of CO₂ to deep groundwater from interaction with the nearby Alan Hill Volcanic Complex as one possible source. Similar to Sistema Zacatón, central Italy has extensive hot-spring travertine deposits, which represent the end-member of a deep-seated carbonate mass-transfer system. Caves and springs in the area have also been identified to have microbial communities involved in sulfur and carbon cycling, such as in Frasassi Cave (Jones et al., 2008) and other sulfur springs near Pozzo del Merro (Caramana and Gary, 2006). Pozzo del Merro and Sistema Zacatón have been compared as similar systems qualitatively (Caramana and Gary, 2006), but a detailed, quantitative study is needed to document more precisely the karst processes common to these two deepest underwater caves.

Turkish Obruks

Bayari et al. (2009) have recently documented karst development in central Turkey that strongly resembles that observed at Sistema Zacatón, although on a

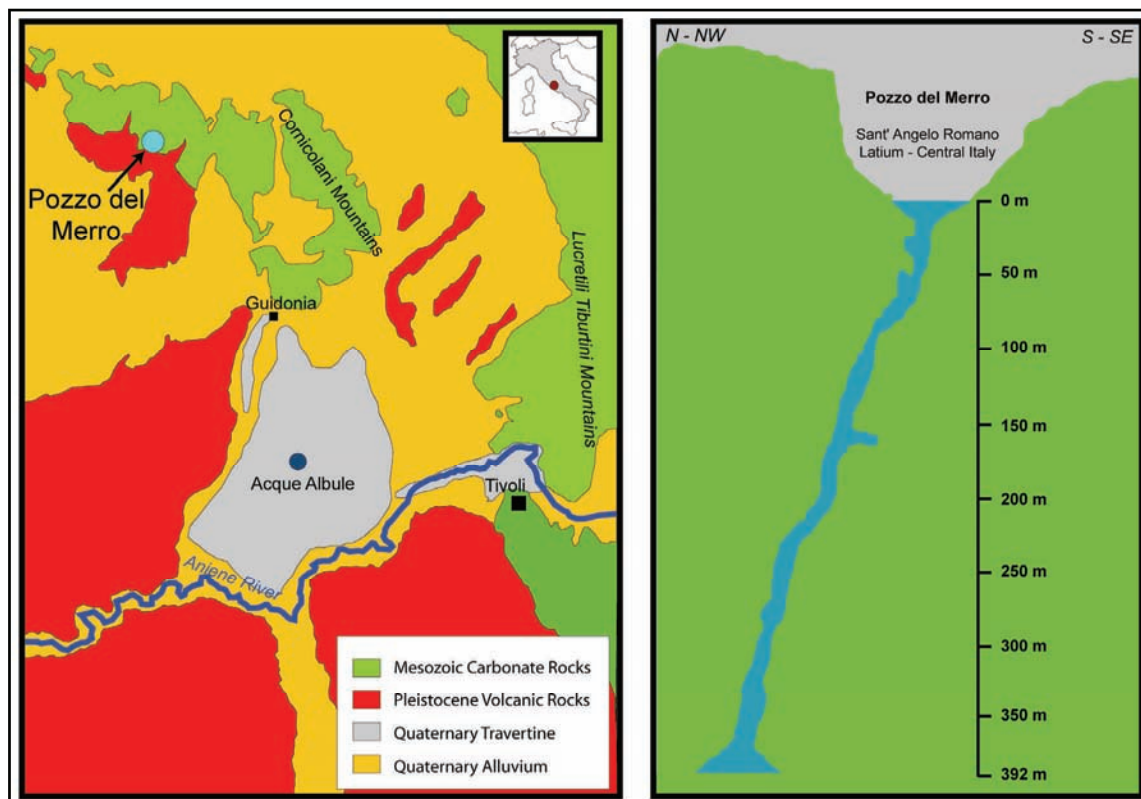


Figure 7. Geologic map of the area (left) around Pozzo del Merro show the juxtaposition of Pleistocene volcanic rocks against Mesozoic carbonates within which the underwater sinkhole formed (right). Pozzo del Merro is the deepest explored underwater cave in the world at a depth of 392 meters (from Caramana, 2002).

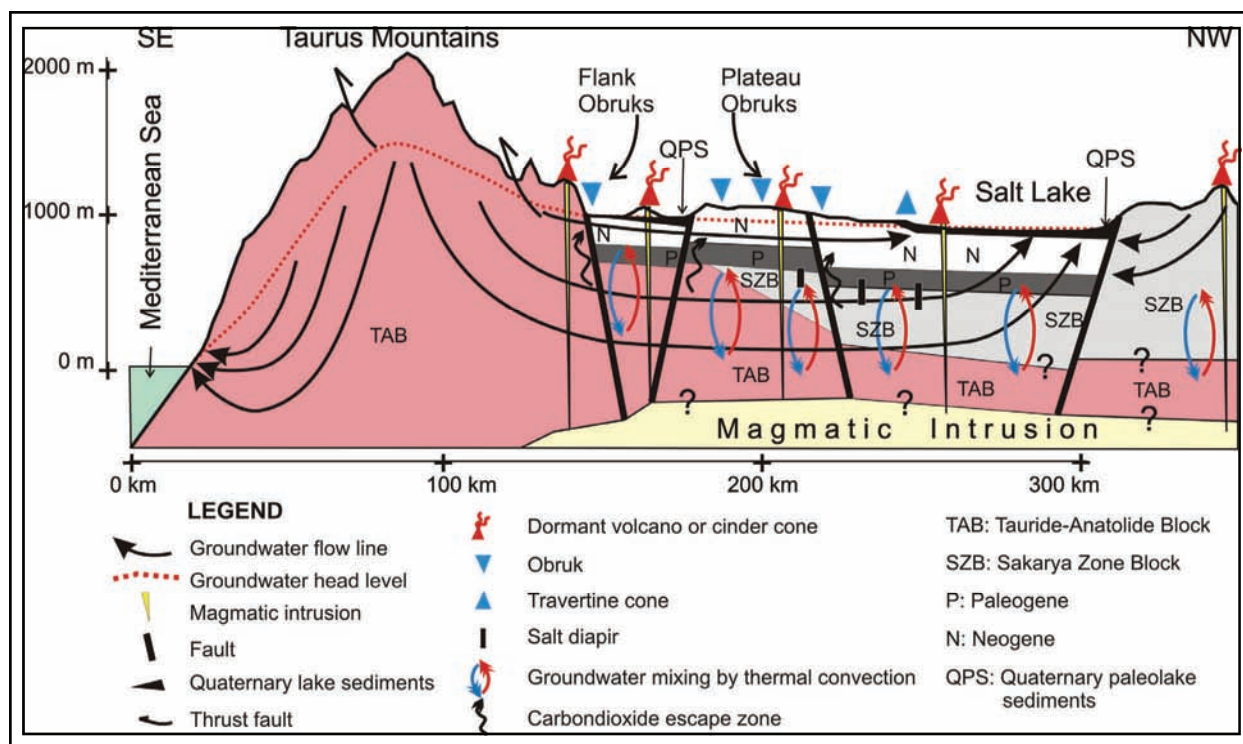


Figure 8. Schematic cross-section of the regional groundwater system and zones of Obruk formation in central Turkey shows strong resemblance to Sistema Zacatón (from Bayari et al., 2009).

broader spatial scale. Large sinkholes, regionally known as Obruks, developed in the Konya Closed Basin, and have morphologies similar to the cenotes of Sistema Zacatón. The features reach diameters in excess of 600 meters and depths up to 125 meters below the water table. This karst system has formed through hypogene processes hypothesized to be driven by volcanic activity. Bayari et al. conducted geochemical and geomorphic investigations with results similar to those observed at Sistema Zacatón. Carbon dioxide concentrations are high ($\log \text{PCO}_2 = -0.9$ to -2.2) and stable carbon isotopic signatures of dissolved inorganic carbon ($\sim 2\text{‰}$ PDB) are reflective of volcanically influenced groundwater systems. Obruk formation appears to remain extremely active as recent development of new features is witnessed every 5-10 years. The conceptual model for Obruk formation is shown (Figure 8), and shows many of the same geologic conditions as Sistema Zacatón (Figure 4). Future studies directly linking these two systems should improve the ability to identify and characterize other VKS in a global context.

Mammoth Hot Springs, Wyoming, USA

Mammoth Hot Springs in Yellowstone National Park, WY is a well-documented hydrothermal groundwater system that developed in direct response to volcanic activity (Sorey, 1991; Sorey and Colvard, 1997;

Pisarowicz, 2003). Many features and characteristics observed here are similar to those of Sistema Zacatón. There is a thick sequence of travertines deposited by hydrothermal waters supersaturated with calcium carbonate. The water is heated by magma at depth and rises through a zone of fissures and joints. It becomes supersaturated with calcium carbonate at depth due to highly elevated levels of CO_2 and flows to the surface (Barger, 1978). Hot springs are found north of Mammoth Hot Springs and groundwater flows through Mississippian Upper Madison Limestone. Other caves and karst features, including travertine deposits, are found in this region. The groundwater had been identified as meteoric from isotopic signatures (Bargar, 1978). Studies have been made in this hydrothermal system on the distribution and diversity of geomicrobial communities, and the connection between volcanism and CaCO_3 dissolution and precipitation processes is likely interconnected with biologic activity (Fouke, 2000). Surface karst features are not widespread in the Mammoth Hot Springs area, but the mass of calcium carbonate precipitation as travertine on the surface indicates that large scale dissolution is occurring in the subsurface in this mass transfer system, possibly creating large voids that remain unexplored.

Cueva de Villa Luz, Mexico

Cueva de Villa Luz in Tabasco, Mexico, is another biologically rich sulfur cave which may have been influenced by volcanic activity. Lagarde Rosales et al. (2008) investigated direct groundwater influence from the Chichón Volcano located 50 kilometers west of the cave (Figure 7). This volcanic system is active with the last known eruption occurring in 1982. The host limestone of the cave is Cretaceous in age, and significant strike-slip faulting has occurred in the region (Lagarde Rosales et al., 2008). Water now flowing in this primarily vadose cave is rich in H₂S and feeds a dynamic community of microorganisms (Hose, 1999; Hose et al., 2000). Although it is not determined what role volcanic activity has had on the formation of Cueva de Villa Luz, karst development in the region may contain VKS.

Other possible VKS

Other VKS may include Rhodope Mountain hydrothermal cavity of Bulgaria (which may still be active), and the Edwards Aquifer of Texas.

Rhodope Mountain hydrothermal cavity (Mandan Chamber), Bulgaria

Dublyansky (2000c) reports of a voluminous cavity 1000 + meters below the surface in the Rhodope Mountains of Bulgaria. It was discovered from boreholes drilled during exploration of ore deposits; the boreholes encountered a mass of Archaean marble containing the void in an area of Paleozoic igneous rocks in the form of stocks and dikes. The deepest borehole encountered the void at a depth of 667.9 meters below land surface, and extended to a depth of 2009 meters without encountering the lower boundary of the void. Dublyansky inferred a minimum height of the chamber at over 1341 meters. This immense cave, known as the Mandan Chamber, is filled with hydrothermal water reaching temperatures of over 129° C. This high temperature water is thought to result from recent magmatic activity nearby or from deep water flowing up through an ancient fault. If the former of these two hypotheses is correct, the deep cavity in the Rhodopes would be a VKS, and by far the deepest phreatic karst system on Earth. Very little is known about this karst system, but the advancement of robotic exploration in extreme karst environments (Gary, 2007) could lead to detailed documentation of the specific geochemical characteristics of deep karst development.

Western Edwards aquifer, Texas, USA

The Edwards Aquifer in central Texas is one of the most prolific karst aquifers in the world, and supplies water to millions of people in and around San Anto-

nio, Austin, and surrounding rural communities (Sharp and Banner, 1997). The Edwards Aquifer has been intensely studied, particularly the eastern and northern regions of the Balcones Fault Zone (BFZ) Segment, and recent hypotheses include hypogene development as an important process in karstification (Schindel et al., 2008). The western segments of the Edwards BFZ, particularly in Medina, Uvalde, and Kinney counties, have a unique geologic framework not present in the eastern and northern segments that includes extensive volcanic activity. This area is referenced as the Uvalde igneous field (Figure 8), and contains late Cretaceous rocks dated from 82-72 Ma (Smith et al., 2007; Blome et al., 2007). This volcanic activity occurred well after deposition of the Edwards Limestone (105-95 Ma). There are anomalous groundwater conditions that exist in this segment of the Edwards Aquifer, including “channeling of groundwater” in a feature known as the Knippa Gap, and valleys with extremely prolific artesian wells in Kinney County from which the water is primarily sourced from cavernous voids 200-300 meters deep (Green et al., 2006). One explanation for these unique hydrogeologic occurrences could be alteration of the geologic framework during periods of active volcanism, creating a VKS deep in the subsurface that strongly influenced subsequent karstification throughout the Cenozoic. Some preliminary evidence of a VKS in Uvalde County may lie in paleosprings with microbially mediated phreatic precipitants. Further investigation of these types of features may provide insightful information when evaluating if the western segment of the Edwards Aquifer was a VKS at one time.

Summary

Volcanogenic karst systems are an important subcategory of hypogene karst that can develop cavernous porosity deep in the subsurface, often without surface expression of karst features. They result as hyper-acidic groundwater conditions occur from enrichment of volcanic CO₂ and H₂S dissolve voids, creating deep-seated karst features. As in the case of Sistema Zacatón, these features have opened to the surface as phreatic sinkholes, providing direct access deep into the aquifer. As the relevance of VKS becomes realized as an important karst process occurring globally, the challenges of investigating such systems can be difficult. Accessing deep, phreatic environments, often through a man-made borehole, limits the type of observations and measurements made. Recent advances in underwater robotics tested at Sistema Zacatón (Gary, 2007) may provide future technological advancements that will expand our knowledge of VKS. It is important to realize the karst forming processes of modern volcanogenic karst systems such

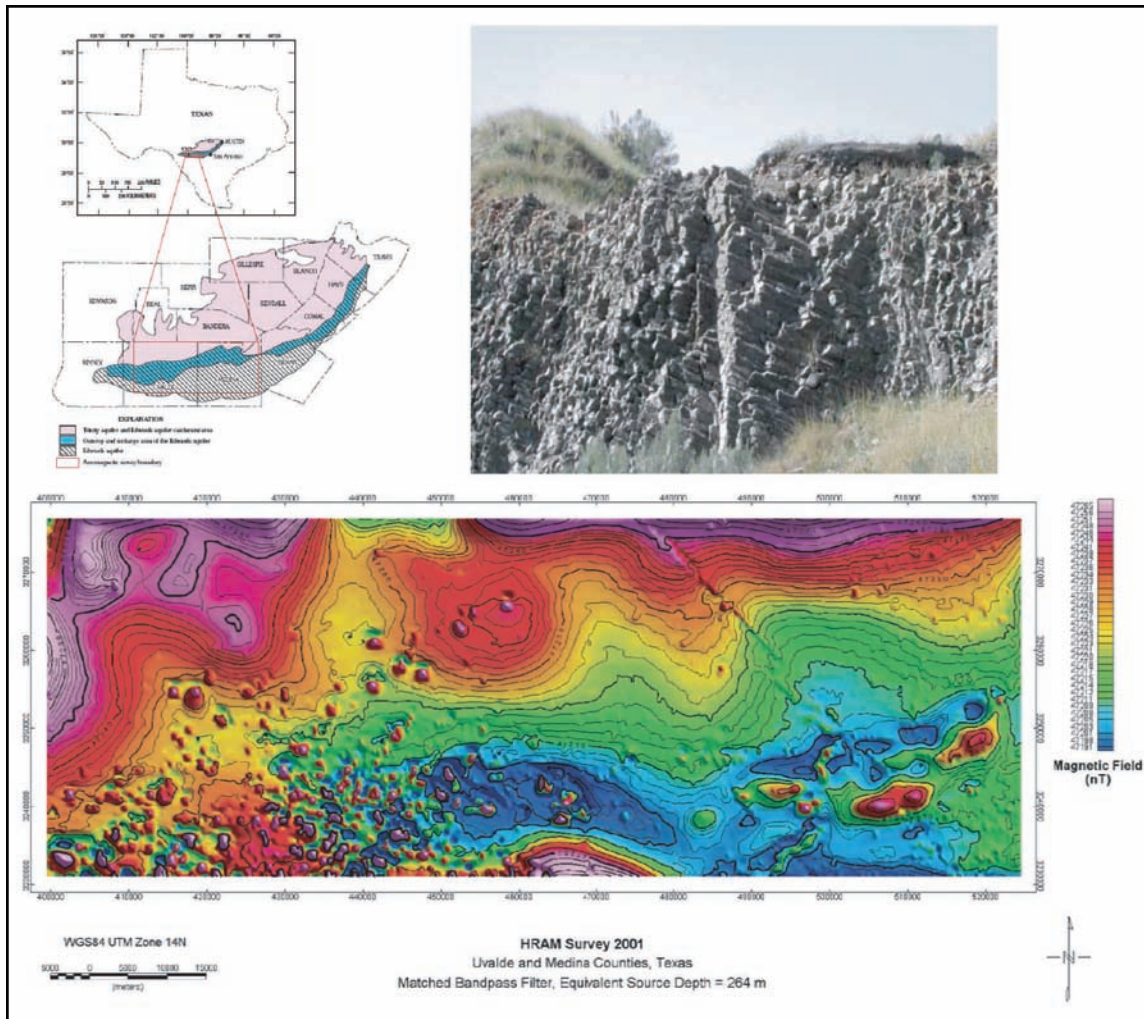


Figure 9. Aerial electromagnetic survey (lower inset) of Uvalde and Medina Counties, Texas conducted by the U.S. Geological Survey show numerous late Cretaceous buried igneous bodies, particularly in Uvalde County (western half). Kinney County (upper left inset), immediately west of Uvalde County also contains late Cretaceous volcanic rocks exposed at the surface. The upper right photograph is of columnar joints in eastern Uvalde County near the Knippa Gap (modified from Smith et al., 2007; Blome et al., 2007).

as Sistema Zacatón and the Obruks, investigate the speleogenesis and dynamics of these active settings, and use their models to aid in interpretation of older, more cryptic karst phenomena.

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References

- Barger, K.E. 1978. Geology and thermal history of Mammoth Hot Springs, Yellowstone National Park, Wyoming. *U.S. Geological Survey Bulletin* 1444: 49-50.
- Bayari, C.S., E. Pekkan, and N.N. Ozyurt. 2009. Obruks, as giant collapse dolines caused by hypogenic karstification in central Anatolia, Turkey: analysis of likely formation processes. *Hydrogeology Journal* 17: 327-345.
- Billi, A., A. Valle, M. Brilli, C. Faccenna, and R. Funiciello. 2007. Fracture-controlled fluid circulation and dissolutional weathering in sinkhole-prone carbonate rocks from central Italy. *Journal of Structural Geology* 29: 385-395.
- Blome, C.D., J.R. Faith, and G.B. Ozuna. 2007. Geohydrologic framework of the Edwards and Trinity aquifers, South-Central Texas, U.S. *Geological Survey Fact Sheet* 2006-3145.
- Camacho A.F. 1993. *Compilación Geologica de la Vertiente del Golfo de Mexico, Area 1*. Comision Federal de Electricidad, Subdireccion Tecnica, Gerencia de Estudios de Ingenieria Civil, Subgerencia de Estudios Geologicos, Departamento de Geologia. G-43: 123-130.
- Caramana, G. 2002. Exploring of the world's deepest sinkholes: The Pozzo del Merro (Italy). *Underwater Speleology* February: 4-8.
- Caramana, G., and M.O. Gary. 2006. *Applicazioni di metodologie di immersione scientifica e ROV (Remote, Operated, Vehicle) nello studio geologic comparato die due sinkholes allagatie più profondi del pianeta: Pozzo del Merro (Lazio, Italia Centrale), El Zacatón (Tamaulipas, Mexico)*. Italian Institute for Environmental Protection and Research, http://www.apat.gov.it/site/_files/sinkhole/211_228.pdf
- Dublyansky, Y.V. 2000a. Hydrothermal speleogenesis – Its settings and peculiar features. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D. Ford, A. Palmer and W. Dreybrodt, 292-297. Huntsville: National Speleological Society.
- Dublyansky, Y.V. 2000b. Dissolution of carbonates by geothermal waters. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D. Ford, A. Palmer and W. Dreybrodt, 158-159. Huntsville: National Speleological Society.
- Dublyansky, Y.V. 2000c. A giant hydrothermal cavity in the Rhodope Mountains, Bulgaria. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D. Ford, A. Palmer and W. Dreybrodt, 317-318. Huntsville: National Speleological Society.
- Fouke, B.W., J.D. Farmer, D.J. Des Marias, L. Pratt, N.C. Sturchio, P.C. Burns, and M.K. Discipulo. 2000. Depositional facies and aqueous-solid phase geochemistry of travertine-depositing hot springs (Angel Terrace, Mammoth Hot Springs, Yellowstone National Park, U.S.A.). *Journal of Sedimentary Research* 70 (3): 565-585.
- Gary, M.O., J.M. Sharp, Jr., G. Caramana, and R.H. Havens. 2003. Volcanically influenced speleogenesis: Forming El Sistema Zacatón, Mexico, and Pozzo Merro, Italy, the deepest phreatic sinkholes in the world. *Geological Society of America Abstracts with Programs* 34 (7): 52.
- Gary, M.O., and J.M. Sharp, Jr. 2006. Volcanogenic Karstification of Sistema Zacatón. In *Perspectives on Karst Geomorphology, Hydrology, and Geochemistry: A Tribute volume to Derek C. Ford and William B. White: Geological Society of America Special Paper 404*, ed. R.S. Harmon and C.M. Wicks, 79-89. Boulder: Geological Society of America.
- Gary, M.O., T. Halihan, J.M. Sharp, Jr., S. Mouri, and J. Thorstad. 2006. Electrical resistivity imaging of travertine capped sinkholes: Deep lakes with lids. *Geological Society of America, 2006 Philadelphia Annual Meeting Program with Abstracts*, Topical Session T65, Paper No. 218-4.
- Gary, M.O., J.M. Sharp, Jr., T. Halihan, and J.A. Ramirez Fernandez. 2007. Mammoth discovery: Paleontological and geophysical evidence for timing and sequence of karstification at Sistema Zacatón, Mexico. *Geological Society of America, Abstracts with Programs (South-Central Section Meeting)* 39 (3): 28.
- Gary, M.O. 2007. DEPTHX – The DEep PHreatic THERmal eXplorer: Robotic exploration and characterization of Sistema Zacatón on the mission path to Europa. *Geological Society of America special session*, <http://www.geosociety.org/meetings/2007/DEPTHX-GSAspecial.pdf>
- Gary, M.O., N. Fairfield, W.C. Stone, D. Wettergreen, G. Kantor, and J.M. Sharp Jr. 2008. 3-D mapping and characterization of Sistema Zacatón from DEPTHX (Deep Phreatic Thermal Explorer). *Proceedings of the 11th Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst*. American Society of Civil Engineers Geotechnical Special Publication no. 183: 202-212.
- Gary, M.O., and J.M. Sharp, Jr. 2008. Volcanogenic karstification: Implications of this hypogene process. *Geological Society of America Abstracts with Programs*, 2008 Joint Annual Meeting, Houston, Texas: 343.
- Gary, M.O., T. Halihan, and J.M. Sharp, Jr. 2009. Detection of sub-travertine lakes using electrical resistivity imaging, Sistema Zacatón, Mexico. In

- Proceedings of the 15th International Congress of Speleology, Kerrville, Texas (in press).
- Goldhammer, R.K. 1999. Mesozoic sequence stratigraphy and paleogeographic evolution of Northeast Mexico. In *Mesozoic sedimentary and tectonic history of north-central Mexico: Geological Society of America Special Paper 340*, ed. C. Bartolini, J.L. Wilson, and T.F. Lawton, 3-14. Denver: Geological Society of America.
- Green, R.T., F.P. Bertetti, N.M. Franklin, A.P. Morris, D.A. Ferrill, and R.V. Klar. 2006. *Evaluation of the Edwards Aquifer in Kinney and Uvalde counties, Texas*. Report for the Edwards Aquifer Authority: 53.
- Hose, L. D. 1999. Cueva de Villa Luz, Tabasco, Mexico: Reconnaissance study of an active sulfur spring cave and ecosystem. *Journal of Cave and Karst Studies* 61 (1): 13-21.
- Hose, L.D., A.N. Palmer, M.V. Palmer, D.E. Northup, P.J. Boston, and H.R. DuChene. 2000. Microbiology and geochemistry in a hydrogen-sulphide-rich karst environment. *Chemical Geology* 169: 399-423.
- Jones, D.S., E.H. Lyon, and J.L. Macalady. 2008. Geomicrobiology of biovermiculations from the Frasassi Cave System, Italy. *Journal of Cave and Karst Studies* 70 (2): 78-93.
- Klimchouk, A.B. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Knab, O. 2008. *Die tiefsten Unterwasserhöhlen der Welt*. Zürich: Im Tiergarten.
- Palmer, A.N. 2007. *Cave Geology*. Dayton: Cave Books.
- Palmer, A.N. 1991. Origin and morphology of limestone caves. *Geological Society of America Bulletin* 103: 1-21.
- Pisarowicz, P. 2003. The Mammoth Hot Springs area beneath Yellowstone. *Rocky Mountain Caving* 20 (4): 12-19.
- Ramirez Fernandez, J.A., F. Velasco Tapia, F. Viera Decida, J.M. Vasconcelos Fernandez, and M.O. Gary. 2007. Tertiary interplate mafic magmatism in the Eastern Mexican alkaline province: Villa Aldama volcanic complex. Geological Society of America abstracts with programs: 390.
- Rosales-Lagarde, L., P.J. Boston, A. Campbell, and K.W. Stafford. 2008. Possible structural connection between Chichón Volcano and the sulfur-rich springs of Villa Luz Cave (a.k.a. Cueva de las Sardinas), southern Mexico. *AMCS Bulletin 19/ SMES Boletín* 7: 177-184.
- Sahl, J.W., and J.R. Spear. 2007. A depth profile of microbial diversity and community structure in cenote El Zacatón. Geological Society of America special session, DEPTHX – The DEep Phreatic THERmal eXplorer: Robotic exploration and characterization of Sistema Zacatón on the mission path to Europa. <http://www.geosociety.org/meetings/2007/-DEPTHX.htm>.
- Salomon, J.N. 2003. Karst system response in volcanically and tectonically active regions. *Annals of Geomorphology: Karst in a Changing World*, supplement volume 131: 89-112.
- Salvati, R., and I.D. Sasowsky. 2002. Development of collapse sinkholes in areas of groundwater discharge. *Journal of Hydrology* 265: 1-11.
- Schindel, G.M., S.B. Johnson, and E.C. Alexander, Jr. 2008. Hypogene processes in the Balcones Fault Zone Edwards Aquifer in south-central Texas: A new conceptual model to explain aquifer dynamics. Geological Society of America Abstracts with Programs, 2008 Joint Annual Meeting, Houston, Texas: 344.
- Sharp, J. M., Jr., and J.L. Banner. 1997. The Edwards aquifer—a resource in conflict. *GSA Today* 7 (8): 1-9.
- Smith, D.V., R.R. McDougal, B.D. Smith, and C.D. Blome. 2007. Distribution of igneous rocks in Medina and Uvalde Counties, Texas, as inferred from aeromagnetic data. *U.S. Geological Survey, Scientific Investigations Report 2007-5191*. Washington: U.S. Geological Survey.
- Sorey, M.L. 1991. Effects of potential geothermal development in the Corwin Springs known geothermal resources area, Montana, on the thermal features of Yellowstone National Park. U.S. Geological Survey, *Water-Resources Investigations Report* 91-4052: 159-182.
- Sorey, M.L., and E.M. Colvard. 1997. Hydrologic investigations in the Mammoth Corridor, Yellowstone National Park and vicinity, U.S.A. *Geothermics* 26: 221-249.
- Suter, M. 1987. Structural traverse across the Sierra Madre Oriental fold - thrust belt in east-central Mexico. *Geological Society of America Bulletin* 98: 249-264.
- Tóth, J. 1999. Groundwater as a geologic agent: An overview of the causes, processes, and manifestations. *Hydrogeology Journal* 7: 1-14.
- Vasconcelos Fernandez, J.M., and J.A. Ramirez Fernandez. 2004. Geología y petrología del complejo volcánico de Villa Aldama, Tamaulipas, *Ciencia UANL VII* (I): 40-44.

FOLIA SPELEOTHEMS, A HYPOGENIC DEGASSING ORIGIN

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Abstract

Bubble trails are subaqueous features in carbonate caves, which are made by the corrosion by ascending carbon dioxide bubbles. Folia are calcite deposits resembling inverted rimstone dams in saturated pools. Based on morphological studies in Adaouste Cave (Provence, France) and on studies elsewhere in the world, we propose a new genetic model for folia, close to the model of Green (1991). The association of bubble trails and folia, occurring on overhanging walls, is interpreted to be an indicator of hypogenic degassing occurring just below the water table. The association is the result of juxtaposed processes composed of corrosion along bubble trails and calcite deposition in calcite-saturated pools.

Introduction

Bubble trails are small channels developed subaqueously in the walls of carbonate caves by corrosion by carbon dioxide bubbles. Folia are among the most curious and rarest calcite speleothems: these subaqueous calcite coatings cover overhanging walls and resemble inverted rimstone dams. Their origin is still debated and several hypotheses have been suggested, among which the two main hypotheses are: deposits at the surface of water bodies (pools with oscillating water level) or subaqueous speleothems in thermal caves. In Adaouste Cave, bubble trails and folia are

intimately associated, which is a paradox involving both corrosion and deposition, respectively. This paper presents a detailed analysis of their morphology and spatial distribution that provides a better understanding of their origin and significance.

Our observations of bubble trails and folia from Adaouste Cave (Provence, France), and comparison with other occurrences worldwide, allow us to ascribe to them a hypogenic origin by degassing. Moreover, the association between bubble trails and folia provides us with an indicator for hypogenic processes by carbon dioxide degassing that is highly useful for the interpretation of inactive hypogenic caves.

Previous genetic theories

Bubble trails have been only recently identified (Chiesi and Forti, 1987). The discovery of folia occurred earlier (Emerson, 1952), but the understanding of their genesis is still debated.

Bubble trails (bubble-flow canals)

Bubble trails are smooth and regular solution channels cut in overhanging walls. The cross-section is roughly half-circular, with diameters of 0.5 to 10 cm, ranging from a shallow print (Figure 1C) to a deep channel (Figure 1D). The course is essentially straight, with some gentle curves following the steepest slope

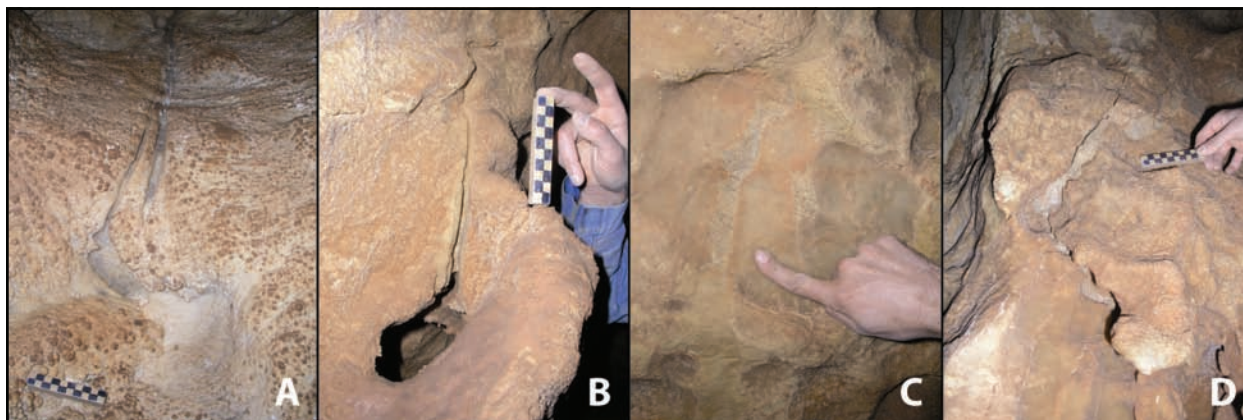


Figure 1. Bubble trails in Adaouste Cave. Corrosion is limited to the channel, whereas the rest of the wall is covered with a subaqueous calcite coating. A and D: views from below; B and C: front views (photos. J.-Y. Bigot, <http://catherine.arnoux.club.fr/photo/13/adao/adao.htm>).

(Figure 1A). Bubble trails were first identified in the caves of the Iglesias metallic district in Sardinia (Chiesi and Forti, 1987). They are also mentioned in some sulfuric acid caves, such as the Frasassi Caves in Italy (Galdenzi and Sarbu, 2000). In Hungary, they were first identified in the cave Ferenc-hegy barlang, and later in the Buda Hills caves, other karst massifs (Bükk, Pilis), and close to the Balaton Lake in Tapolca barlang (Szabó, 2005). Their development is due to carbon dioxide degassing (Chiesi and Forti, 1987). In the Santa Barbara 2 Cave, the degassing is due to the oxidation of sulfur ore deposits (De Waele and Forti, 2006). At depth, carbon dioxide remains dissolved because of the high pressure. During the rise of water in the phreatic zone, water degasses CO_2 as depth decreases, typically 15-30 m depth at the maximum (Luiszer, 1997). Bubbles converge into the steepest upward courses along overhanging walls. The carbon dioxide bubbles locally enhance the solutional aggressivity of the adjacent water. Bubble trails are produced by the continued corrosion along these unchanging routes. Such corrosion features are probably more frequent than suggested by the sparse sites mentioned above. Theoretically, such phenomena should be present in most hypogenic caves where carbon dioxide degassing occurs. On the contrary, they do not seem to exist in any other type of cave. Bubble trails should not be confused with other types of wall and ceiling channels such as paragenetic wall channels, convective channels in deep-seated caves (Klimchouk, 2007), hydrothermal condensation-corrosion channels (Audra, 2007), etc.

Folia (Hill and Forti, 1997)

Folia are speleothems resembling inverted rimstone dams or mushroom caps. They occur as undulating ribbons, stacked like leaves, and hence their name (Figure 2). They grow downward as a continuous coating, exclusively on overhanging walls. The lower rim is horizontal or gently tilted about several degrees. Individual ribbons are on average of 1 cm thick, less than 10 cm wide, and separated vertically by empty spaces up to 5 cm. In places they are influenced by currents and are arranged parallel to the flow direction (e.g. in Indian Burial Cave, Nevada; Green, 1991). At the micro-scale, the calcite is deposited in a dendritic microcrystalline fabric. Eventually, this porous dendritic texture is overgrown by large columnar crystals, up to 10 mm in length (Kolesar and Riggs, 2004). Such dendrites and skeleton crystals are the result of a rapid growth and a limited supply of material. For carbonate solutions, this is often caused by the pressure fluctuations associated with mechanical degassing, e.g. the growth tips of stalactites (Maltsev, 1999), or by high supersaturation under

periodic very low-flow-regime periods that result in prolonged outgassing (Frisia et al., 2000). Folia are formed below the water table, down to several meters or dozens of meters deep. More rarely they mark an

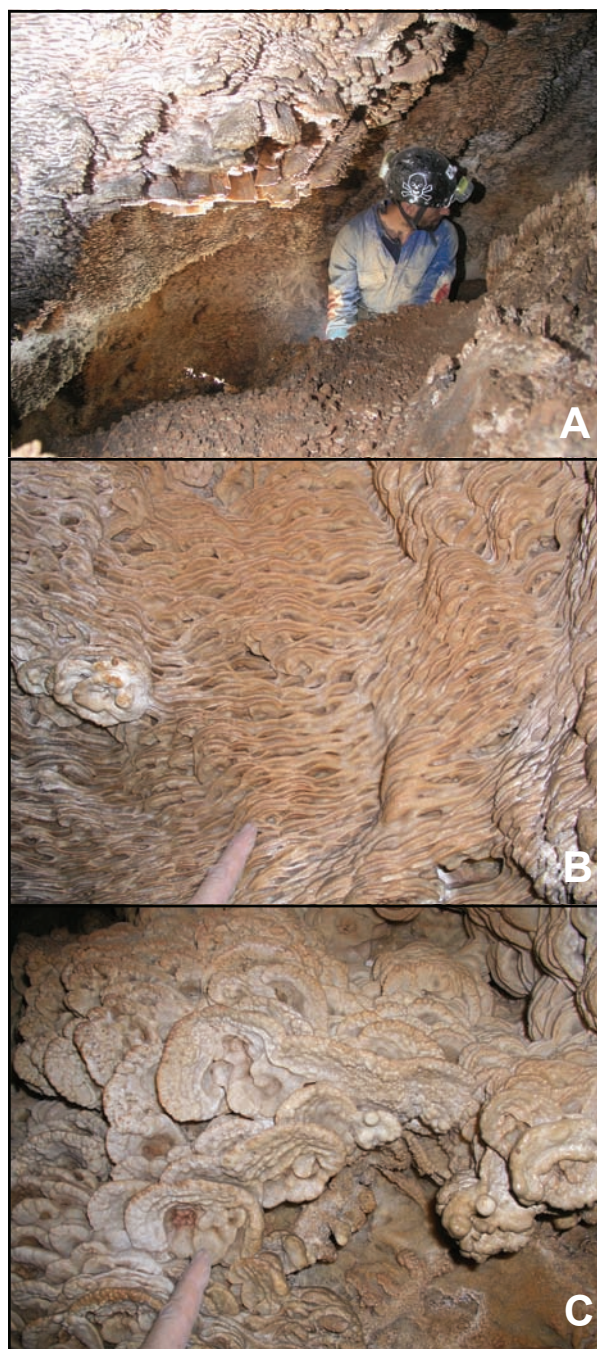


Figure 2. Folia in Adaouste Cave. Top: folia occur exclusively below overhanging walls; middle: front view toward the top showing the “inverted rimstone” morphology; bottom: view from below showing the outward and downward development as inverted cups (photos by J.-Y. Bigot, <http://catherine.arnoux.club.fr/photo/13/adao/adao.htm>).

Table 1. Folia occurrences, about 25 sites worldwide.

Cave	Location	Active / fossil	Hydrothermalism	CO ₂ degassing	Tectonic / hydrogeology	Reference
Indian Burial Cave	USA, Nevada	F	86-119° C (T _H of the fluid inclusions)			Emerson, 1952; Green, 1991; Halliday, 1957
Hurricane Crawl Cave	USA, California	A	No (river cave)	From upwelling phreatic water		Davis, 1997
Crystal Sequoia Cave	USA, California					Davis, 1997
Goshute Cave	USA, Nevada	F				Halliday, 1954b
Devil's Hole	USA, Nevada	A	Water table 34 °C	x	Non karstic active extensional fault	Kolesar and Riggs, 2004
Gneiss Cave	USA, Utah	F			Calcite folia coating onto a gneiss wall	Green, 1997
Bida Cave	USA, Arizona					Davis, 1965; Hill, 1982
Groaning Cave	USA, Colorado					Davis, 1973
Agua Caliente Cave	USA, Arizona	A	Cave at 38 °C			McLean, 1965
Carlsbad Cave	USA, New-Mexico	F	Carlsbad 20-25 °C		Basin margin	Davis, 1970
Lechuguilla Cave	USA, New-Mexico	F	Carlsbad 20-25 °C		Basin margin	Davis, 2000; Hose, 1992
Cinco Cuevas (caverna de las)	Cuba					Núñez-Jimenez, 1975
Pulpo (sima del)	Spain, Murcia	A	Water table 21 °C			Ferrer Rico, 2004
Benís (sima de)	Spain, Murcia	A	Water table 21 °C			Ferrer Rico, 2004
Ermite (grotte de l')	France, Pyrénées	A	Water table 19 °C Thermal sulfidic spring 38 °C	x	Artesian flow path in deep syncline	Bigot and Nobecourt, unpub.
Adaouste (grotte de l')	France, Provence	A	Abandoned at 11 Ma Thermal springs along Durance fault	x	Durance active transcurrent fault	Audra et al., 2002
Pál-Völgy - Matyas-Hegy barlang	Hungary, Buda Hills	F	Springs 20-27 °C		Rim of the Danube rift	Takacsne Bolner, 2005
József-hegy barlang	Hungary, Buda Hills	F	Springs 20-27 °C		Rim of the Danube rift	Takacsne Bolner, 2005
Molnar-Janos	Hungary, Buda Hills	A	Water table 20-27 °C	x	Rim of the Danube rift	Takacsne Bolner, 1993
Matyas-Farras	Hungary					Takacsne Bolner, 1993
Giusti (grotta)	Italy, Tuscany	A	Thermal spring 34 °C	x		Forti and Urti, 1984; Piccini, 2000
Ryan Imperial Cave	Australia, Queensland					Jennings, 1982
Cupp-Coutunn Cave	Turkmenistan	F	80-170° C (T _H of the fluid inclusions, from fluorite and calcite)		Basin margin	Maltsev, 1997; Maltsev and Self, 1993

oscillating water table, as a horizontal ring that covers the walls, within a vertical range of several centimeters to decimeters as in Hurricane Crawl Cave, California (Davis, 1997), and Devil's Hole, Nevada (Kolesar and Riggs, 2004).

Folia are frequently associated with subaqueous speleothems originating from calcite deposition in supersaturated pools, such as cave clouds, calcite rafts, tower cones, and coral towers. All of these deposits are generally, but not systematically, induced

by hypogenic degassing producing oversaturation in the pools (Audra et al., 2002). Folia were first identified in Indian Burial Cave, Nevada (Emerson, 1952). Currently, they are recorded in fewer than 25 caves worldwide, some active, some inactive (Table 1). Many of these sites are in thermal caves. However, hydrothermal conditions do not seem to control the presence of folia: the temperature of water and the temperature of calcite crystallization deduced from fluid inclusions range across a large spectrum (from 20 to 120 °C; Table 1), and Hurricane Crawl is a cold epigenic cave.

Two main hypotheses have been proposed for the origin of folia:

1. The oscillation of a saturated pool surface (Davis 1997). At the surface of supersaturated pools, precipitation is due to evaporation (Halliday, 1954a; Davis, 1973; Jennings, 1982). Precipitation can be due to strong degassing, trapped gas is involved in shaping the folia growth around bubbles, and water-level oscillations are essential to explain upper and lower limits of the folia (Davis 1997). Calcite precipitation occurs either directly by particle accretion, or by accretion of calcite rafts. Hill (1987) specifies that after the lowering of the water, the sloping folia edges developed like draperies, where calcite botryoids ("popcorn") may grow. Kolesar and Riggs (2004) correlate folia with the water oscillation due to earth-tidal waves, which would favor calcite precipitation by degassing and subsequent saturation of the capillary film during the lowering of the water.
2. Phreatic degassing in thermal water (Green, 1991, 1997): the CO₂ bubbles originating from degassing are trapped below wall irregularities. These small gas pockets focus the precipitation of calcite around the bubbles with downward-orientated growth. Successive upward "sloughing" produces the structure of inverted rimstones. Thermal rising flow concentrates the deposition below overhanging walls.

Regarding these hypotheses and observations as a whole, Hill and Forti (1997) do not distinguish the variety of physical settings (e.g. in perched pools, water table, or shallow depth in the phreatic zone) and come to no single explanation for folia genesis. However, they admit a close relationship with the water table and with subaqueous speleothems such as cave clouds, as well as with pool speleothems such as rafts.

New evidence in The Adaouste Cave

The Adaouste Cave opens below the top of the Mira-beau anticline. This fold is cut by the Durance water gap, along an active transcurrent fault, lined by thermo-mineral springs (Audra et al., 2002). Hypogenic flow tends to converge toward the highest places of buried aquifers, where discharge can occur. Consequently, the outflow is located at the intersection between the anticline hinge and the water gap, which acted as the regional base level. The Adaouste Cave was probably active in the Tortonian (11 Ma), at the beginning of the Durance water gap entrenchment. The main entrenchment phase of the water gap occurred in the Messinian (Clauzon, 1979). The cave drained, became perched, and consequently has been preserved from any further reworking. The temperature of homogenization (TH) of fluid inclusion, even if not reliable, tends to show a trend of temperatures of crystallization higher than cold meteoric environment (Audra and Häuselmann, 2004). Some ore deposits in the neighborhood (Fe-Mn) are associated with hypogenic karst and have preserved microbial evidence in metallic pool fingers (Audra and Hofmann, 2004). The cave is made of two steep passages following the anticline dip, which rise from about 200 m depth (Figure 3). These steep passages connect to the horizontal upper levels (-18 and -27) which record past water-table elevations (Audra et al., 2002). These horizontal levels display intense condensation-corrosion features, suggesting the presence of a nearby thermal body (Audra et al., 2007). Below the water table, corrosion concentrated in bubble trails. At depth, carbon dioxide gas was trapped in blind bells with a high pressure, giving rise to hyper-corrosive atmospheres, where condensation-corrosion produced boxwork at the ceiling, as well as drip holes (Figure 4). Simultaneously, degassing led to supersaturation and massive calcite deposition, as various morphologies: botryoids (popcorn) above the water table; and rafts, tower cones, folia, and coral towers, beneath the water table (Audra et al., 2002). The water/gas interfaces are very clear, showing transition between atmospheric corrosion and subaqueous precipitation (Figure 5)

Folia and bubble trails in Penitents Chamber

Observations have been made in Penitents Chamber (-124 m), together within the main passage above the chamber (Figure 4). The diversity of subaqueous corrosion and deposition features in this area allow us to establish some relationships, particularly between bubble trails and folia. The bubble trails converge upward and can reach several meters in length before connecting to blind domes (Figure 1A) or disappearing where the overhanging walls become vertical. No

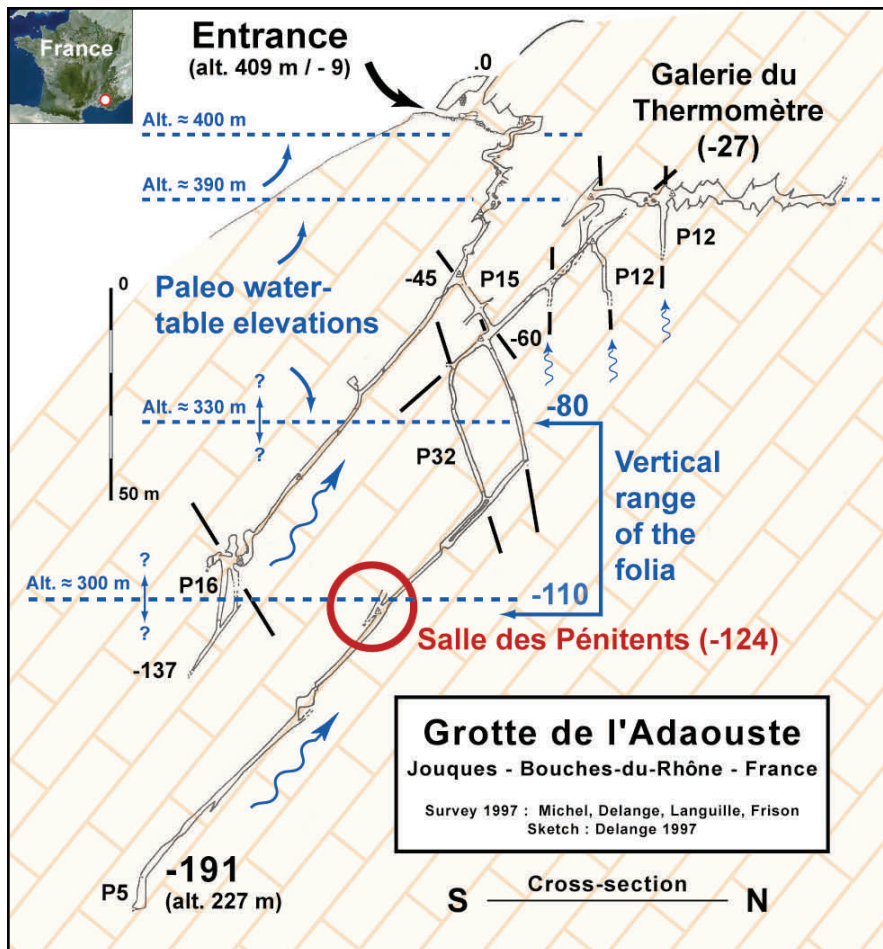


Figure 3. Adaouste Cave survey. Hypogenic flow (undulated arrows) welled up through conduits along dip and through fissures (black lines); horizontal levels record past water table positions.

Figure 4. The Penitents Chamber, an ancient blind phreatic conduit where the cupolas were filled with CO₂ from degassing. Distribution of features originating from atmospheric corrosion and subaqueous calcite deposition.

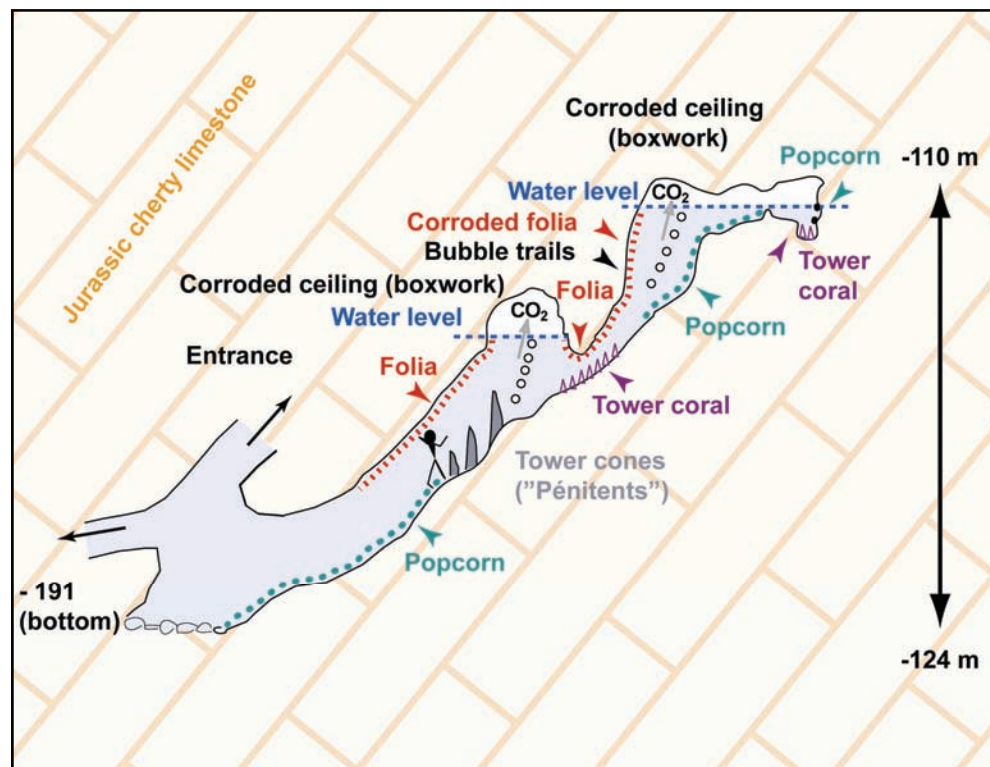




Figure 5. Interface between a carbon dioxide-rich gas pocket with walls corroded as boxwork (top) and subaqueous calcite deposits below: folia below overhanging walls, tower cones below the caver (photo by J.-Y. Bigot).

deposits are present in the bubble trails. When a calcite coating occurs on neighboring walls, it is cut by the bubble trails (Figure 1C).

Folia are developed continuously between -80 and -120 m (Figure 3). In the upper part, they are uniformly corroded and partly covered with subaerial popcorn. In the lower part, folia are not corroded. A standing water level is visible in the Penitents Chamber, around -110 m. The size of folia increases upward in sequences several meters high. From these observations, we can deduce that:

- After their deposition, the highest folia (-80 to -95 m) have been exposed to air by a water-table lowering, then they have been corroded by condensation-corrosion;
- The influence of condensation-corrosion is intense in the upper horizontal levels, it decreases downwards, and seems to be absent below -95 m.
- The lowest folia (-95 to -120 m) have been deposited either simultaneously with upper ones, with a vertical range of about 30m, or subsequently as the result of a water-table lowering;
- There is no evidence of any subsequent water-table rising, since popcorn covering folia is not covered with any new subaqueous calcite;

- The upward increase of folia size reflects the upward increase of gas volume, both by the degassing and by the addition of rising bubbles.

Folia bubbles

We observed several calcite bubbles inside the folia hollows. We call these speleothems folia bubbles. They are composed of calcite that forms at the water-air contacts of bubbles, by centripetal growth. The development of such features needs the presence of a solution shifting to oversaturation at the water-bubble interface. We suggest the following origin (Figure 6):

- A film of condensation water appears at the vaulted solid top of the bubble, due to the thermal gradient between the thermal water and the rocky ceiling. The gradient is maintained by thermal flux through the rock;
- In the high-CO₂ atmosphere of the bubble, condensation water becomes hyper-aggressive;
- The corrosive water dissolves the calcite and flows along the wall. This migration makes the solution progressively saturated;
- At the base of the bubble, evaporation leads to supersaturation;
- The calcite precipitates on the lower edge of the folia. The precipitation zone propagates along the bubble, at the water-bubble interface. Since this process involving calcite redistribution inside the cavity seems to be limited, calcite particle accretion from the degassing water-body may also participate to the building of the calcite bubble.

Discussion: folia genesis and association with bubble trails, a record of thermal water-table dynamics

Rebuttal to the previous hypotheses

Oscillation of a saturated pool surface (Davis 1997). The main argument supporting this hypothesis is that the constant upper boundary of the deposit is a record of the slow rise and fall of the water body. During these oscillations, accretion of solid material grows downward to enlarge the speleothem. Trapping of bubbles when the water rises, or from degassing, would accentuate this process. However, we identify numerous arguments negating this hypothesis:

- The oscillation of a saturated pool is not sufficient by itself to form folia, because such a condition is present in all karst beneath a vegetal cover, and yet folia are on the contrary extremely rare.

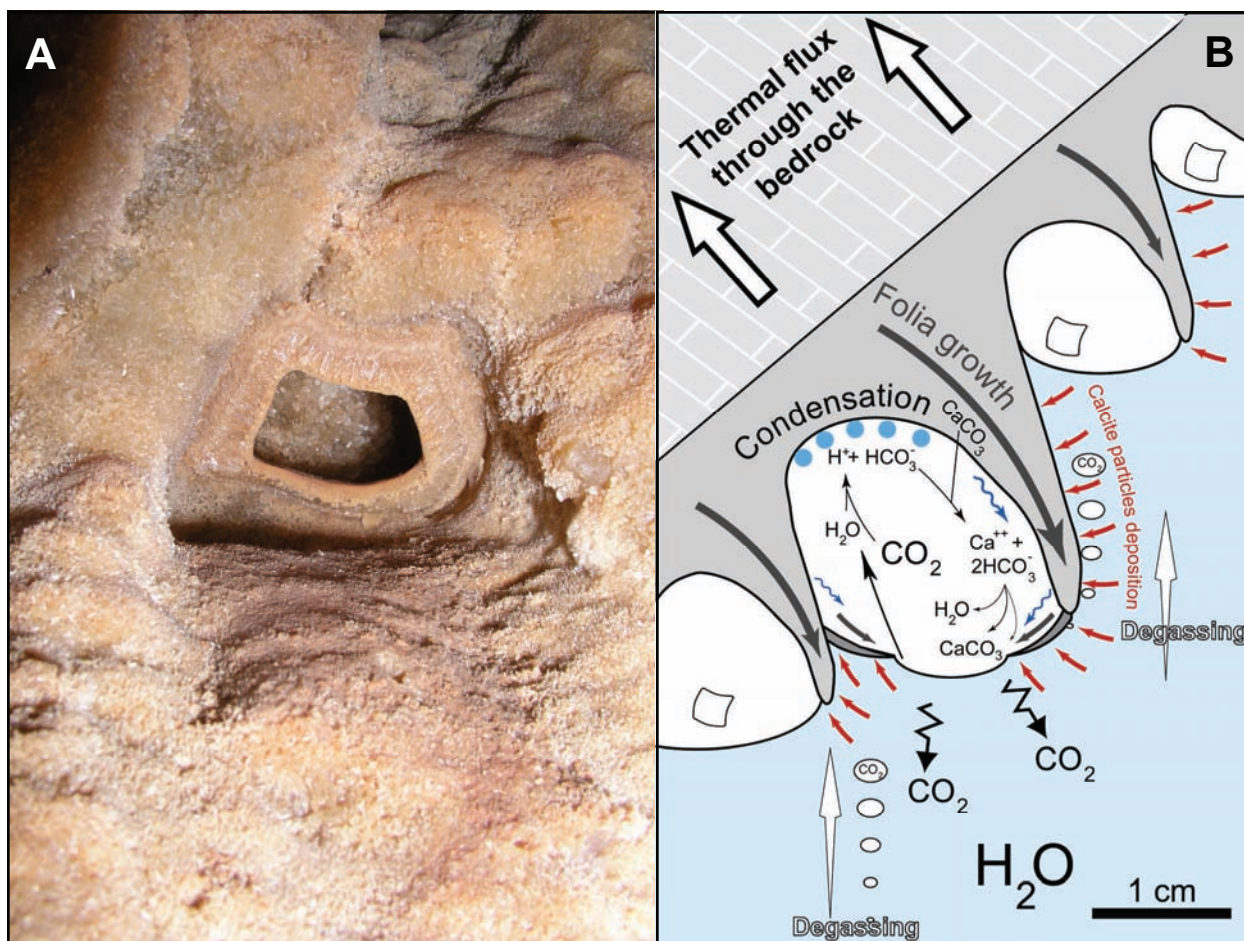


Figure 6. Left: folia bubble (photo by J.-Y. Bigot). The aperture is about 4 cm wide. Right: folia and folia bubble development by condensation-corrosion at the top of the carbon dioxide bubble and then by calcite deposition through evaporation at the base of the bubble, at the interface between gas and saturated water. The calcite closing the bubble is shown in dark grey. White arrows show CO_2 degassing; red arrows show subaqueous calcite particle accretion; grey arrows show the growing direction of the folia.

- The systematic distribution of folia below overhanging walls excludes a simple mechanism of an oscillating pool, which would have similar consequences on non-overhanging surfaces.
- Except for their upper limit, folia never display horizontal patterns which would record some lower water surface.
- Regarding the calcite texture, a random texture would be expected from the accretion of particles or rafts; on the contrary, a dendritic texture is observed deriving from the mechanical effect of degassing (Maltsev, 1999).
- The particular morphology of inverted rimstone is never clearly explained.

Phreatic degassing in thermal water (Green, 1991, 1997). Following this author, calcite particles appear after degassing due to sudden pressure reduction when

the fluid expands from a small orifice; the thermal buoyancy raises the particles and accretion occurs around gas bubbles that are trapped below irregularities of the overhanging walls. This hypothesis is consistent with our observations on the whole; however some statements do not seem to be relevant:

- Folia are frequently associated with thermal fluids, but not systematically (Table 1).
- Decompression at constrictions sufficient to cause bubbling by cavitation is not feasible because cavitation requires flow velocities far above those reasonably possible at all observed sites.

New perspectives for the genesis of folia

Since previous hypotheses are not or only partly relevant, respectively, we propose a new process to

explain the genesis of folia, partly based on Green's (1991, 1997) hypothesis:

- Below the water table at shallow depth, CO₂ degassing produces bubbles rising toward the surface;
- The bubbles are trapped beneath irregularities in the overhanging walls, and when the spaces are filled the bubbles slough upward into the next hollow; We effectively observed such underwater process in grotta Giusti (Tuscany, Italy) (Piccini, 2000).
- Degassing makes the solution supersaturated, which produces ubiquitous calcite precipitation, except in hollows filled with bubbles. Calcite protrudes outward and downward from wall projections to form the lower edges of the folia. By positive feedback, the size of each trapped bubble increases, and so on.

The folia bubbles clearly show crystallization at the gas-water interface. Consequently, the genesis of folia in subaqueous conditions, due to hypogenic degassing, is demonstrated here. Trapping of bubbles is admitted by both Green (1991) and Davis (1997), as the main consequence for degassing, or as a secondary factor occurring when water rises, respectively. We demonstrate that the trapping of bubbles from a strong degassing is essential to obtain the typical hollow morphology of inverted rimstones. Their size increases upward by addition of rising bubbles. Consequently, we think that folia are almost exclusively associated with degassing from a hypogenic origin. Only two extremely rare cases mimicking the hypogenic environments were identified in non-hypogenic settings: (1) Hurricane Crawl Cave, where degassing into saturated pools is due to the inflow of epigenic water originating from a mountain karst with dense vegetation; (2) cuevas de Bellamar - El Jarrito, Cuba (Hall, 2008), where the air is mechanically trapped by tidal fluctuations.

The thermalism does not seem to be a necessity, even if frequent. Rising hypogenic flow looks more adequate to explain both the folia morphology and the degassing.

We also demonstrate that folia develop by degassing at shallow depth below the water table, no deeper than about 30 m (Luiszer, 1997). Sudden

degassing deeper in the phreatic zone by decompression at the outlet of constriction is unlikely. The water table acts as the upper boundary, the lower boundary corresponds to the lower limit of degassing. Both boundaries determine the vertical range of the folia deposits. In these conditions, a water-table oscillation (such as in Devil's Hole) is not necessary; it is only occasionally present. The small vertical range of folia in Hurricane Crawl Cave could correspond to some sediment filling which may since have been removed. Regarding the maximum vertical range of the folia, few data are available. They reach about 40 m in Lechuguilla Cave, and 30 m at Adaouste. Such a depth is compatible with the depth of degassing, and also with the amount of water-table lowering that is documented in Adaouste.

Association between bubble trails and folia

Many bubble trails emerge from folia or cupolas where carbon dioxide has been concentrated (Figure 7, 1A). The juxtaposition of both features, with a sharp transition lacking any overlap (in the stratigraphic sense), shows that they are coeval. Moreover, their association shows they originate from a common process, i.e. carbon dioxide degassing. Their association is controlled by wall geometry: if an overhanging wall is absent, the bubbles rise vertically in the water: neither folia nor bubble trails develop; below a convex-downward overhanging wall, the bubbles pour from folia into folia while diverging; below a concave-

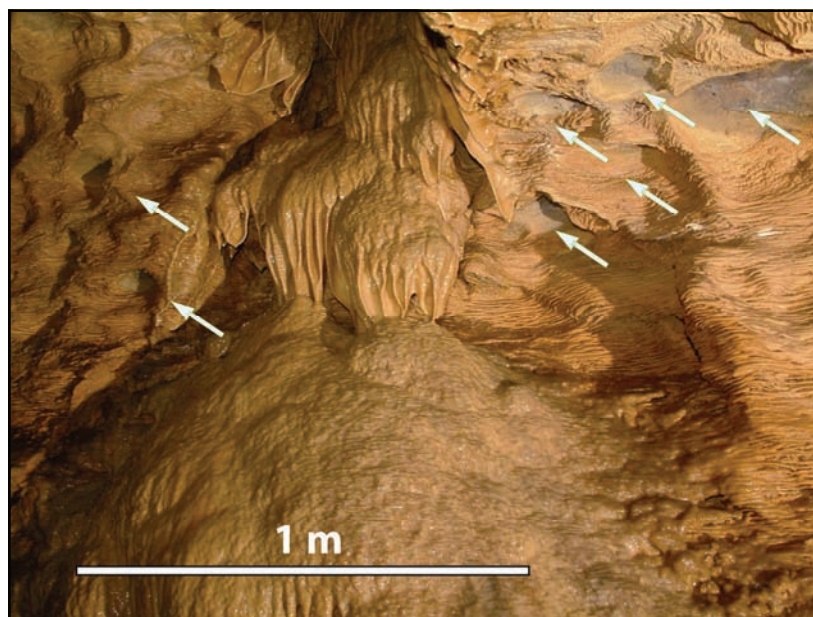


Figure 7. Folia in Pál-Völgy barlang, Hungary. Arrows indicate cupolas where gas bubbles are trapped and where intense corrosion showing bare rock is clearly visible, whereas walls are covered with a thick folia coating. The flowstone in the center developed after draining (photo by A. Kiss, with permission).

downward overhanging wall, the bubbles converge toward an invariable trajectory, and a bubble trail gradually appears in the wall; the bubble trail could reach a cupola, from which another bubble trail emerges.

Record of paleo-water tables

In the Adaouste Cave, the vertical range of the folia, their corrosion in the upper part, and their absence below -120 m, allows us to reconstruct the lowering of the thermal water table (Figure 3). It demonstrates:

- The presence of a thermal water table at about -80 m, at the boundary between popcorn corrosion above and of folia deposition below;
- The lowering of the water table to the top of the Penitents Chamber ($\approx$ -110 m). This lowering caused the smoothing by condensation-corrosion of the overlying folia and the development of the underlying folia.
- The distribution of folia record the recession of thermal activity in the Adaouste Cave.

Non-carbonate features and the term folia

Some features resembling folia are made of minerals other than calcite:

- In clay: Cave of the Winds and Orient Mine Cave, Colorado (Davis, 1982, 1984, 1997), Vass Imbre barlang (Maucha, 1993), and Matyas-Farras, Hungary (Takacsné Bolner, 1993). However, their morphology differs significantly: rims do not overlap and do not form individual inverted cusps. Since they are much too soft to be generated by oscillating water, these clay rims are apparently produced by the regular lowering of a turbid water body (Green, 1997);
- In halite: Liquid Crystal Cave, Israel. Initially published as folia (Frumkin, 1997). Actually they correspond to rimstone shelves (Frumkin, personal communication).
- In sulfur: Cueva de Villa Luz (Hose et al., 2000; Hose, pers. comm.). Soft sulfur forms inverted and overlaid hollowed cusps. Some are similar to coalescent mushroom caps with a flat bottom. They develop in confined areas in very acid atmosphere close to the sulfur springs. Their genetic process is still enigmatic; however they appear to develop in the air, maybe under the influence of acid gas convection.

Despite the visual similarities of these features in non-carbonate material, they are not related to the process of bubble trapping. Sulfur features would develop

from sulfuric gas convection by crystallization of sulfur. For the mud features, the genetic process is not physico-chemical, it rather involves mechanical processes, by the transport of solid particles (in phreatic conditions by lowering of levels of turbid water, or by thin sheets of runoff over walls). Their morphology, roughly as horizontal ribs, with flat bottoms in the case of sulfur folia, do not display holes capable of trapping bubbles. Their genetic process, although still largely unknown, is hardly similar to that of folia. Consequently, we think that true folia are exclusively made of calcite; we suggest restricting “folia” to a precise descriptive term for downward-concave calcitic features, and by extension to a genetic term, since these features are associated with a given process. Similar features that owe their existence to different genetic processes should have different names, to avoid confusion.

Conclusions

Folia clearly seem to be connected to hypogenic flow. Hydrothermal conditions are typical; however they do not seem necessary, as long as a strong carbon dioxide degassing is present. The original hydrothermal hypothesis was initially proposed at a time when hypogenic caves were mainly considered hydrothermal, which is only one aspect of hypogenic speleogenesis, neither systematic nor a principal feature (Klimchouk, 2007).

Folia cover large areas and have unambiguous morphology. Consequently, their occurrence seems to be strongly correlated to a carbonic acid, hypogenic context, involving degassing at shallow depth below the water table. Similar features that occur in non-hypogenic context (e.g., in halite or clay) are clearly different, from both a morphological and a genetic point of view. Their inclusion in the term “folia” should be abandoned.

The appearance of folia results from the following combination: (1) strong (hypogenic) degassing below the water table making bubbles and leading to supersaturation; (2) overhanging walls; (3) trapping of carbon dioxide bubbles, making calcite precipitate at the periphery of bubbles, in a downward growth. If the overhanging wall geometry is dihedral, then there is an association between folia and bubble trails. Folia, and moreover the association between folia and bubble trails, can be considered a very reliable record of hypogenic conditions.

The hypothesis of an oscillating supersaturated water pool must be abandoned. First, because folia formed in this way should be widespread, when in fact they

are extremely rare. Moreover, this hypothesis does not give a global explanation for the specific morphology of folia, such as inverted rims.

Since degassing occurs at shallow depth below the water table, folia and bubble trails can also be considered a precise record of the water-table position, located at the top of the folia zone.

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References

- Audra, P. 2007. Karst et spéléogénèse épigènes, hypogènes, recherches appliquées et valorisation. Operative research and valorisation. Thesis, University of Nice Sophia-Antipolis.
- Audra, P., and P. Häuselmann. 2004. Hydrothermal origin of two hypogenic karst caves in French Provence: Preliminary results from fluid inclusions. *Actes des Journées européennes de l'AFK, Le karst de la craie en Normandie, Rouen* 2003: 32-34.
- Audra, P., and B.A. Hofmann. 2004. Les cavités hypogènes associées aux dépôts de sulfures métalliques (MVT). *Le Grotte d'Italia* 5: 35-56.
- Audra, P., J.-Y. Bigot, and L. Mocochain. 2002. Hypogenic caves in Provence (France): Specific features and sediments. *Acta Carsologica* 31 (3): 33-50.
- Audra, P., F. Hobléa, J.-Y. Bigot, and J.C. Nobécourt. 2007. The role of condensation-corrosion in thermal speleogenesis. Study of a hypogenic sulfidic cave in Aix-les-Bains, France. *Acta Carsologica* 37 (2): 185-194.
- Chiesi, M., and P. Forti. 1987. Studio morfologico di due nuove cavità carsiche dell'Iglesiente (Sardegna Sud occidentale). *Ipoantropo* 4: 40-45.
- Clauzon, G. 1979. Le canyon messinien de la Durance (Provence, Fr.): une preuve paléogéographique du bassin profond de dessiccation. *Palaeogeography, Paleoclimatology, Palaeoecology* 29: 15-40.
- Davis, D.G. 1965. *Observations in Bida Cave, Grand Canyon National Park*. Unpublished Report: Grand Canyon National Park.
- Davis, D.G. 1970. *Folia in Carlsbad Cavern*. Unpublished Report: Carlsbad Caverns National Park.
- Davis, D.G. 1973. Miniature folia in Groaning Cave, Colorado. *Caving in the Rockies* 15 (1): 1.
- Davis, D.G. 1982. Virgin passage found in Cave of the Winds. *Caving in the Rockies* 24 (5): 54-55.
- Davis, D.G. 1984. Mysteries in mud: ancient frost-crystal impressions and other curiosities in Cave of the Winds, Colorado. *Rocky Mountain Caving* 1 (3): 26-29.
- Davis, D., G. 1997. Folia in Hurricane Crawl Cave and Crystal Sequoia Cave. National Speleological Society, San Francisco Bay Chapter, 40 (5): <http://www.caves.org/grotto/sfbc/news/vol40/issue-40-5.html> (accessed December 2007).
- Davis, D. 2000. Extraordinary features of Lechuguilla Cave, Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62 (2): 147-157.
- De Waele, J., and P. Forti. 2006. A new hypogean karst form: the oxidation vent. *Zeitschrift für Geomorphologie, supplement* 147: 107-127.
- Emerson, D. 1952. Labor day Nevada Cave trip, 1951. *California Caver* 4(8): 3-5.
- Ferrer Rico, V. 2004. Grandes cuevas y simas del Mediterráneo. De Gibraltar a Catalunya Nord. Guía fotográfica. <http://www.cuevasmediterraneo.com> (accessed December 2007).
- Forti, P., and F. Utili. 1984. Le concrezioni della Grotta Giusti. *Speleo* 7 (7): 17-25.
- Frisia, S., A. Borsato, I. J. Fairchild, and F. McDermott. 2000. Calcite Fabrics, Growth Mechanisms, and Environments of Formation in Speleothems from the Italian Alps and Southwestern Ireland. *Journal of Sedimentary Research* 70 (5): 1183-1196.
- Frumkin, A. 1997. Liquid Crystal Cave, Israel. In *Cave minerals of the world*, ed. C. Hill and P. Forti, 319-322. Huntsville: National Speleological Society.
- Galdenzi, S., and S.M. Sarbu. 2000. Chemiosintesi e speleogenesi in un ecosistema ipogeo: i rami sulfurei delle grotte di Frasassi (Italia centrale). *Le Grotte d'Italia* 1: 3-18.
- Green, D.J. 1991. On the origin of the folia and rims. National Speleological Society, *Salt Lake Grotto Technical Note* 88: 182-196.
- Green, D.J. 1997. The origin of folia. National Speleological Society, *Salt Lake Grotto Technical Note* 96: 51-60.
- Hall, L. 2008. Cuban Caves Photo Page, El Jarrito - Bellamar Cave system, Matanzas province, Cuba. The Pittsburg Grotto, <http://www.karst.org/pgrotto/cubapics.htm> (accessed May 2008).
- Halliday, W.R. 1954a. Basic geology of Crystal Cave, Utah. National Speleological Society, *Salt Lake Grotto Technical Note* 16: 13.

- Halliday, W.R. 1954b. Basic geology of Goshute Cave, Nevada. National Speleological Society, *Salt Lake Grotto Technical Note* 12: 1-4.
- Halliday, W.R. 1957. The Snake Creek Caves, White Pine County, Nevada. National Speleological Society, *Salt Lake Grotto Technical Note* 39: 1-4.
- Hill, C.A. 1982. Mineralogy of Bida Cave, Grand Canyon National Park, Arizona. *Cave Research Foundation Annual Report* 15: 29-30.
- Hill, C.A. 1987. *Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains*. New Mexico Bureau of Mines and Mineral Resources Memoir 117. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hill, C.A., and P. Forti. 1997. Folia. In *Cave minerals of the world*, ed. C.A. Hill and P. Forti, 73-74. Huntsville: National Speleological Society.
- Hose, L.D. 1992. To Lechuguilla Depths. *Rocky Mountain Caving* 9 (2): 15-18.
- Hose, L.D., A.N. Palmer, M.V. Palmer, D.E. Northup, P.J. Boston, and H.R. DuChene. 2000. Microbiology and geochemistry in a hydrogen-sulphide-rich karst environment. *Chemical Geology* 169: 399-423.
- Jennings, J.N. 1982. Karst of northeastern Queensland reconsidered. Tower Karst. *Chillagoe Caving Club Occasional Paper* 4: 13-52.
- Klimchouk, A.B. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Kolesar, P.T., and A.C. Riggs. Influence of depositional environment on Devil's Hole calcite morphology and petrology. In *Studies of cave sediments: physical and chemical records of paleoclimate*, ed. I. D. Sasowsky and J. Mylroie. New York: Kluwer/Plenum Academic Press.
- Luiszer, F.G. 1997. *Genesis of Cave of the Winds, Manitou Springs, Colorado*. PhD Dissertation, University of Colorado.
- Maltsev, V. A. 1997. Cupp-Coutunn Cave, Turkmenistan. In *Cave minerals of the world*, ed. C.A. Hill and P. Forti. 323-328. Huntsville: National Speleological Society.
- Maltsev, V.A. 1999. Stalactites with "internal" and "external" feeding. *Proceedings of the University of Bristol Speleological Society* 21 (2): 149-158.
- Maltsev, V.A., and C.A. Self. 1993. Cupp-Coutunn cave system, Turkmenia, Central Asia. *Proceedings of the University of Bristol Speleological Society* 19 (2): 117-149.
- Maucha, L. 1993. *A Vass Imre-barlang természetvédelmi célú állapotfelmérés*. Unpublished Report: National Authority for Nature Conservation (Hungary).
- McLean, J.S. 1965. Folia found in Agua Caliente. *Arizona Caver* 2 (6): 125.
- Núñez Jiménez, A. 1975. Nivelitas nuevas formaciones espeleológicas. *Simp. 35e Aniv. Soc. Espeleol. Cuba, Resúmenes, Isla de Pinos*.
- Piccini, L. 2000. Il carsismo di origine idrotermale del Colle di Monsummano (Pistoia – Toscana). *Le Grotte d'Italia* 1: 33-43.
- Szabó, Z., 2005. Morphological and hydrological relationship of the Tapolca lake cave. Genesis and formation of hydrothermal Caves. International Conference, Budapest 2004, Papers, 100-105. Budapest: Hungarian Speleological Society.
- Takácsné Bolner, K. 1993. Rare types of carbonate speleothems. *Karszt és Barlang* 1-2: 29-36.
- Takácsné Bolner, K. 2005. Rare speleothems found in Pál-völgy cave. Genesis and formation of hydrothermal Caves. International Conference, Budapest 2004, Papers, 118-124. Budapest: Hungarian Speleological Society.

EXAMPLES OF HYPOGENIC KARST COLLAPSE STRUCTURES: TWIN CITIES METROPOLITAN AREA, MINNESOTA, USA

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Abstract

Three collapse structures in the Twin Cities Metropolitan Area, USA, may be products of hypogenic karst processes. The features are associated with deeply entrenched bedrock valleys, one the current Mississippi River and the other two buried by glacial drift. These valleys have been discharge points for regional groundwater flow systems.

The most recent occurred in October, 2005 in Woodbury, Minnesota in a newly constructed infiltration pond excavated into the subcropping St. Peter Sandstone. The collapse event created several sinkholes on the sides and bottom of the pond. Excavation of three of twelve sinkholes revealed a major collapse structure comprised of several breccia pipes. These pipes are at least 10 to 15 m in diameter and extend up

through almost 20 m of friable St. Peter from the underlying Prairie du Chien Group.

A second occurred in 1989 in Mahtomedi, MN. A hole 12.8 m wide and 11 m deep opened suddenly. Both of these collapses developed adjacent to river valleys buried under glacial sediments.

The oldest feature is in eastern Minneapolis near the Mississippi River gorge. This closed depression in the land surface was incorporated into a park when Minneapolis was platted. This depression extends through the subcropping Platteville Formation to a large void in the underlying St. Peter. A 250 meter long cave in the St. Peter extends between the collapse and the Mississippi River with typical passages 15 meters wide and 6 meters high.

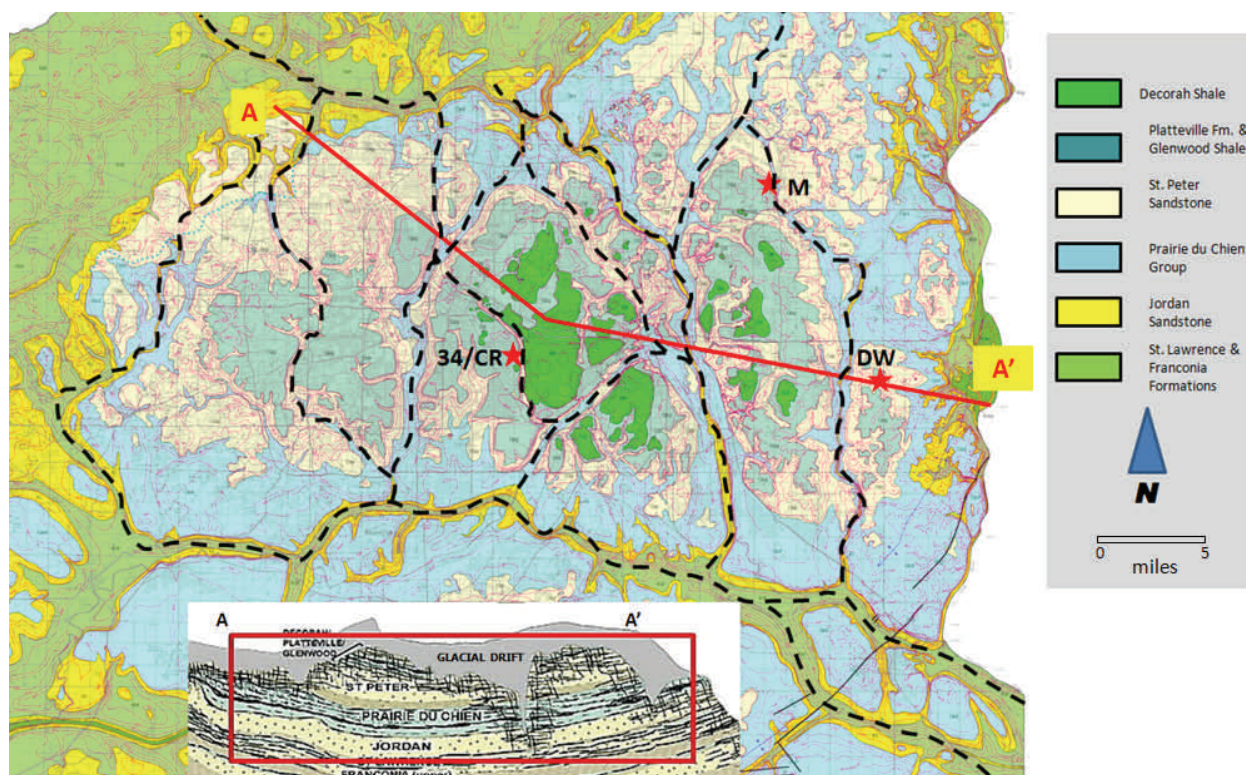


Figure 1. Locations and bedrock geology of collapse structures in Twin Cities metropolitan area. DW=Dancing Waters, M=Mahtomedi, 34/CR=34th Street/Channel Rock. Present or buried bedrock valleys are denoted by dashed lines. Base map from Mossler and Tipping (2000); cross section from Runkel et al. (2006).

Recent investigations of the Prairie du Chien have found extensive solution enlargement features on a regional scale; these features have several characteristics diagnostic of hypogenic speleogenesis. The possibility of other breccia pipes in the St. Peter within the metropolitan area is both probable and problematic.

Introduction

The recent work by Klimchouk (2007) has advanced the understanding of the formational processes operating in hypogenic karst. Application of these insights has assisted in the re-interpretation of karst features within several caves (Alexander *et al.*, 2008) and buried karst features (Barr and Klimchouk, 2007; Barr *et al.*, 2008) in the Paleozoic formations of Minnesota. In addition, recent work by Runkle *et al.* (2003, 2006), Tipping *et al.* (2001, 2006) and others have begun reinterpreting the hydrostratigraphy of southeastern Minnesota using both outcrop and downhole information.

In particular, this examination of the Prairie du Chien Group, a predominantly dolomitic unit of Ordovician age, has found extensive development of secondary porosity throughout much of its areal extent in southeastern Minnesota. This unit overlies the Jordan Sandstone and underlies the St. Peter Sandstone, both orthoquartzite sandstone aquifers with silicic or minimal calcic cementation (*cf.* Section A-A', Figure 1). The Prairie du Chien is situated generally in the middle of the Paleozoic sequence of sedimentary formations with additional aquifers near the base and the top of the sequence. Thus, the Prairie du Chien can be seen as an ideal case for hypogenic speleogenesis in the discharge portions of the regional flow systems. Indeed, where Prairie du Chien karst has been denuded and is accessible, the passageways are anastomosing and maze-like with the cave porosities and areal coverage in conformance with hypogenic settings.

Within the Twin Cities metropolitan area the Prairie du Chien is subjacent to denuded environments (*cf.* Klimchouk, 2007), but generally is located beneath other geologic units and the water table. A series of valleys have been eroded into the bedrock formations (see Figure 1); most of these valleys are presently filled with glaciofluvial deposits. Like the bedrock valleys in the metropolitan area that are presently occupied by rivers, these buried bedrock valleys are thought to have been regional discharge points for groundwater flow systems when the valleys were occupied by functional rivers. Consequently, the portions of the Prairie du Chien in the discharge

portion of the regional flow system have changed through geologic time, likely causing the development of hypogenic karst within the Prairie du Chien to occur throughout most of the Prairie du Chien in the metropolitan area.

As can be seen in the Twin Cities bedrock geology map in Figure 1, the Prairie du Chien is overlain by the St. Peter over much of the Twin Cities. The cross section inset in Figure 1 shows that almost all of the bedrock is overlain by glacial deposits in the area, obscuring karst features that may be present on the bedrock surface. However, one implication of the formation of hypogenic karst is the possible development of breccia pipes in the formations overlying the hypogenic karst. These vertical collapse structures can form from outlet cupolas and dome pits within the karstified unit (Klimchouk and Andrejchuk, 2005). Several collapse structures associated with possible breccia pipes have been identified within the Twin Cities metropolitan area. The locations of these structures are shown on Figure 1 and their attributes are discussed below.

Dancing Waters collapse structure

The most recent collapse structure is in Woodbury, Minnesota, in the southeastern portion of the metropolitan area. On October 4th and 5th, 2005 a newly constructed infiltration basin in the Dancing Waters subdivision, named Dancing Waters Pond, filled for the first time; by October 8th more than 6,000 m³ of water had disappeared into twelve sinkholes that developed on the floor and walls of the basin (Figure 2). The subsequent investigation of the newly created sinkholes found that the basin had been excavated through the surficial glaciofluvial sediments and



Figure 2. Sinkholes in Dancing Water Pond, Woodbury, Minnesota (looking southwest). Picture taken October 12, 2005 by E. C. Alexander, Jr. Long axis of left sinkhole (longest arrow) is approximately 9 m.



Figure 3. Western breccia pipe, Dancing Waters Pond. Note open fractures at base of excavation. Tape measure (circled in red) is extended into the fracture for its maximum length (7.6 m). Photo by K. Barr.

approximately 10 meters into the subcropping St. Peter Sandstone and that this basin floor was approximately 20 meters above the contact with the underlying Prairie du Chien Group.

Further investigation found that the sinkholes clustered on the floor of the basin were associated with a series of open fractures, many filled with glaciofluvial sediments as well as the earthen materials used for the basin's liner. These fractures' orientations were largely different from the major joint orientations in the intact sandstone and were predominantly dipping at $45^{\circ} \pm 10^{\circ}$. Furthermore, these fractures were generally found around the periphery of the excavation of the main collapse site shown in Figures 3 and 4. These open fractures were dipping away from at least two common centers, spaced around a circular perimeter with a diameter of 10 to 15 meters. Inside the perimeter the sandstone was found to be comprised of numerous large blocks of sandstone with their interstices filled with glaciofluvial sediments and reworked St. Peter sands (Figures 3 and 4).

It was concluded that the sinkholes were created by the stormwaters' flushing and reactivation of the fractures and voids associated with a breccia pipe or several breccia pipes extending upward from the Prairie du Chien, the top of which had been intersected by the excavation of the basin. Examination of topographic maps and aerial photographs of the vicinity did not reveal any surface expression of the breccia pipe(s) at the basin location prior to its excavation. The Dancing Waters collapse structure is approximately 2 ½ kilometers east of a major north-south trending buried bedrock valley (Lake Elmo



Figure 4. Eastern breccia pipe, Dancing Waters Pond. Note that the outer perimeter of the truncated top of the breccia pipe is filled with dark topsoil/pond liner material. (Western breccia pipe is in hole in front of far orange fence.) Photo by K. Barr.

Valley) eroded through the Prairie du Chien and approximately 7 kilometers west of the north-south trending St. Croix River valley (see Figure 5). Bedrock topography of the area mapped by Patterson *et al.* (1990) and Mossler and Tipping (2000) indicate that the basin is underlain by a minor valley extending from the St. Croix River valley to the west-northwest and intersecting with the Lake Elmo Valley; a complementary tributary valley is found on the opposite side of the Lake Elmo Valley, also trending west-northwest. Although this area is overlain by up to 100 meters of glacial deposits, obscuring most features, the consistent tributary valleys' trend which is orthogonal to the strike of the Afton Anticline to the southeast suggests structural control of the valley which may contribute zones of weakness for the formation of breccia pipes within the tributary valley and the Dancing Waters Pond vicinity.

Mahtomedi collapse structure

A second collapse structure developed in 1989 in Mahtomedi, Minnesota (see Figure 6). A sinkhole appeared overnight, creating a hole 12.8 m wide and 11 m deep. While photographs exist showing the walls of the collapse in glacial drift (Figures 7A and 7B), the report of the sinkhole's characteristics and the subsequent subsurface investigation is no longer available. However, the location is mapped by Mossler and Tipping (2000) and Mossler and Bloomgren (1990) as being underlain by subcropping St. Peter Sandstone. The collapse structure was one-half kilometer to the west of a buried bedrock valley, the west fork of the Lake Elmo bedrock valley that extends generally to the south-southwest.

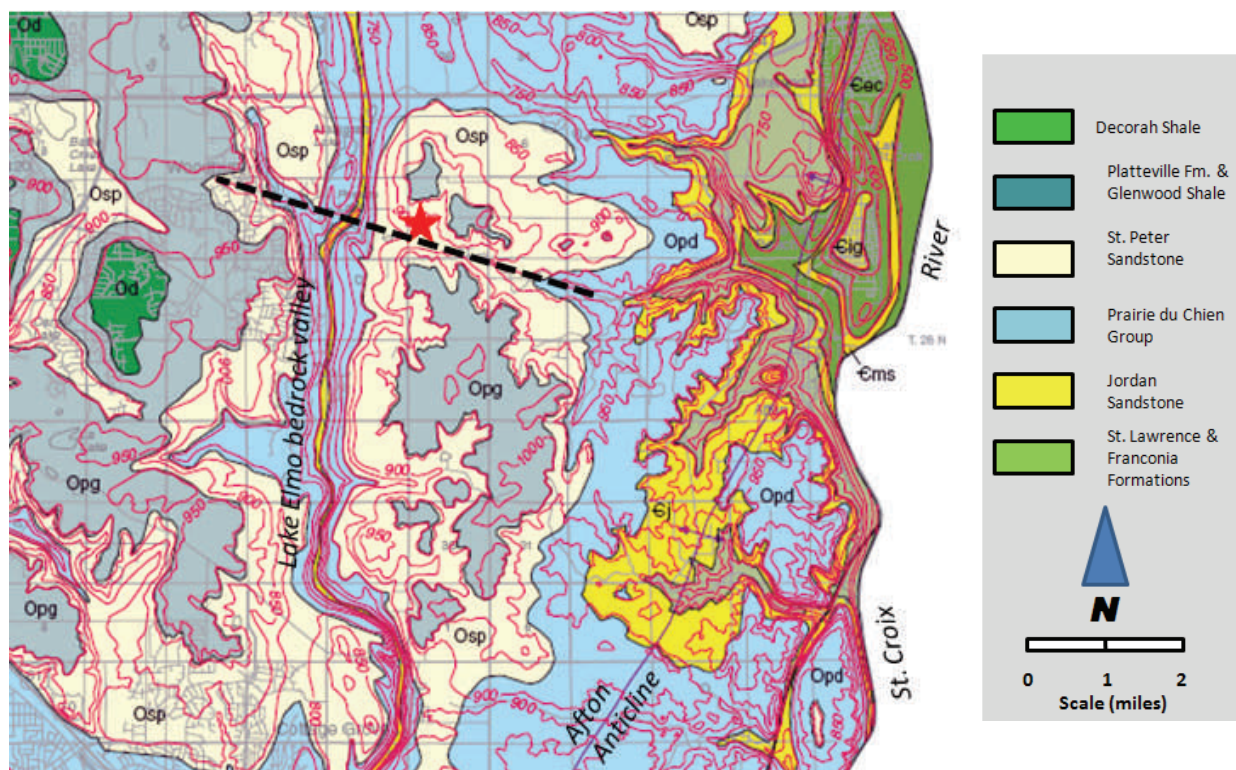


Figure 5. Geology of Dancing Water vicinity. Dancing Waters collapse structure is indicated by star. Location of nearby buried bedrock tributary valley is indicated by dashed line. Base map from Mossler and Tipping (2000).

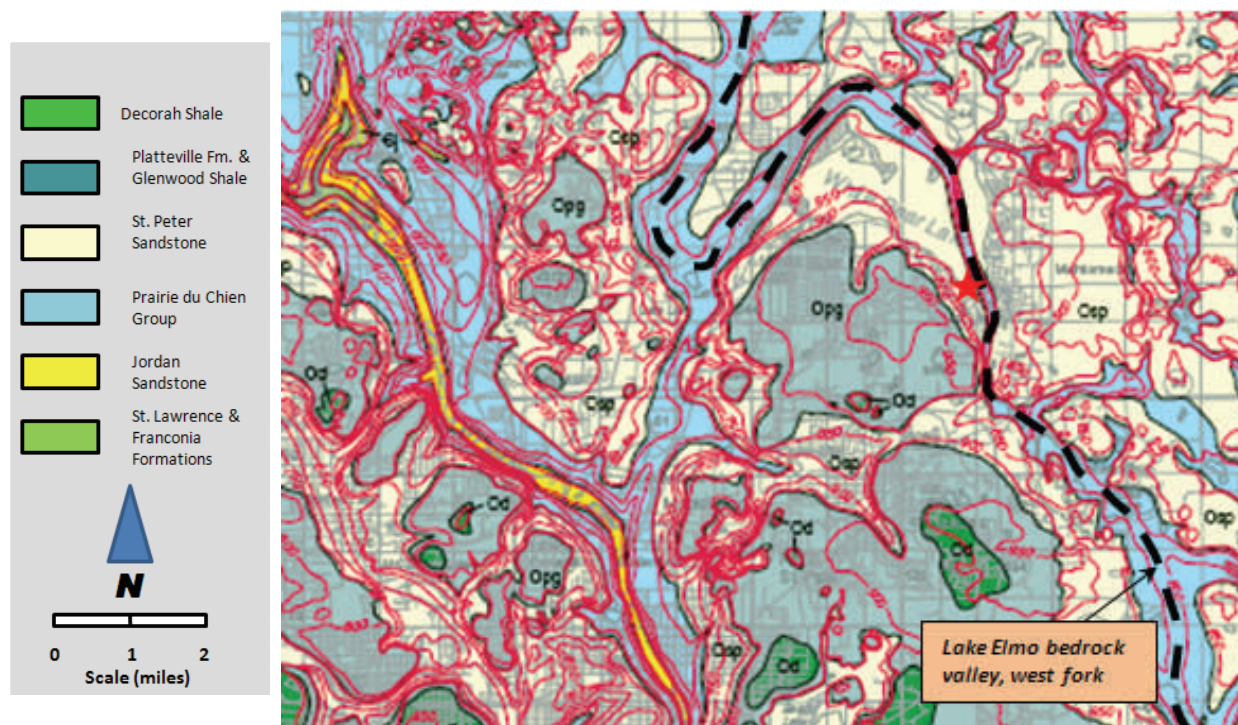


Figure 6. Geology of Mahtomedi vicinity. Mahtomedi collapse structure is indicated by star. Location of nearby buried bedrock valley is indicated by dashed line. Base map from Mossler and Tipping (2000).



Figure 7. Collapse structure, Mahtomedi, Minnesota. Both 7A (left) and 7B (right) show the glacial till walls of the collapse. Note the small tree in 7B (right) that dropped into the collapse. Photos from D. Setterholm.

34th Street Sinkhole, Minneapolis

A much older collapse structure is found in eastern Minneapolis near the center of the metropolitan area, just south of 34th Street East and a few blocks west of the Mississippi River. A sinkhole in the subcropping Platteville Formation predates the settling of Minneapolis, and the surficial depression comprises the circular Seven Oaks Park. The Platteville Formation is an Ordovician dolomitic limestone unit approximately 10 meters in total thickness in the metropolitan area. In 1935 as a sewer tunnel was excavated along the river bluff approximately 130 meters east of the park another associated feature was encountered, a large void within the St. Peter. The “L”-shaped void, called the Channel Rock Cavern, extended approximately 120 meters west, then approximately 170 meters to the south (Figure 8A). The passageway’s ceiling is the base of the Platteville Formation, or remaining vestiges of the 1-meter-thick Glenwood Shale underlying the Platteville. The floor of the passageway was originally quite irregular with large blocks of sand-

stone, so that the height of the passageway varied from approximately 5.5 to 10 meters. Shortly after its discovery, the tunneling operation used the cavern for disposal of mined sand from the St. Peter, half-filling the cavern for most of its length. The width of the passageway ranges from approximately 11 to 18 meters.

The location of Channel Rock Cavern with respect to the 34th St. Sinkhole is shown in Figure 9. While both features are very close to the Mississippi River valley, the cavern was not found to have a naturally occurring opening to the river gorge. (An opening was excavated shortly after the cavern’s discovery; subsequently it has been barricaded.) Along the passageway walls still exposed there is one side passageway (Figure 8B). This passageway trends east-west and extends approximately 10 meters before narrowing to several open, dipping fractures.



Figure 8. Channel Rock Cavern, Minneapolis, Minnesota. 8A (left) shows a transverse perspective of the sand deposited in the cavern and the base of the Platteville Fm. Comprising the ceiling. 8B (right) shows the side passageway which continues via the hole above the seated person. Photos by J. Lovaas.

There has been no further investigation of these features. However, it can be seen from Figures 10A and 10B that the sinkhole and cavern align with Shadow Falls, a major side valley of the Mississippi

River gorge in the Twin Cities, on the opposite side of the valley. Also, the alignment of the back portion of the passageway correlates well with other features further south along the gorge wall and the general

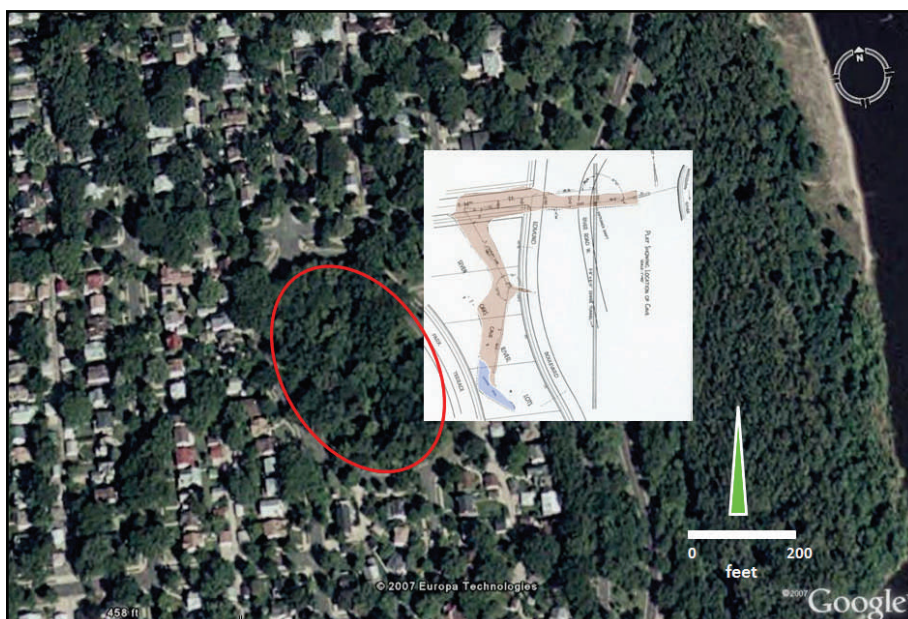


Figure 9. Location of the 34th Street sinkhole (circled in red) with respect to Channel Rock Cavern (pink and blue areas on inset map. Mississippi River on far right side of figure.

Figure 10A.
Locations of 34th
Street sinkhole and
Shadow Falls side
valley. Note the
pair of side valleys
aligned with each
other south of
these features.



Figure 10B.
Alignment of
land surface and
cavern features.



alignment of the river reach immediately downstream of the sinkhole. Both of these alignment trends correlate closely with joint patterns observed in the Platteville at two other locations in eastern Minneapolis shown in Figure 11. These correlations suggest that the occurrence and location of the sinkhole and cavern may be due to local weaknesses defined by the structural geology.

The adjacent Mississippi River gorge is the youngest bedrock valley in the Twin Cities metropolitan area, and the upstream progression of St. Anthony Falls – now situated 11 km upstream – passed the sinkhole’s vicinity approximately 7,500 years ago (*cf.* Wright, 1972). Since that time, the Mississippi River gorge

has been a regional discharge point for the metropolitan area flow systems. An older buried bedrock valley, generally trending north-south, that underlies Lakes Nokomis, Hiawatha, and Powderhorn, is approximately 2¼ km west of the sinkhole.

Summary

Reexamination of the hydrostratigraphy of the Paleozoic units in Minnesota, along with the application of hypogenic mechanisms for speleogenesis indicate that hypogenic karst has likely developed in many portions of the Prairie du Chien Group in the Twin Cities metropolitan area. A preliminary examination of the surficial geology of the metropolitan area for collapse features has yielded three such features that,

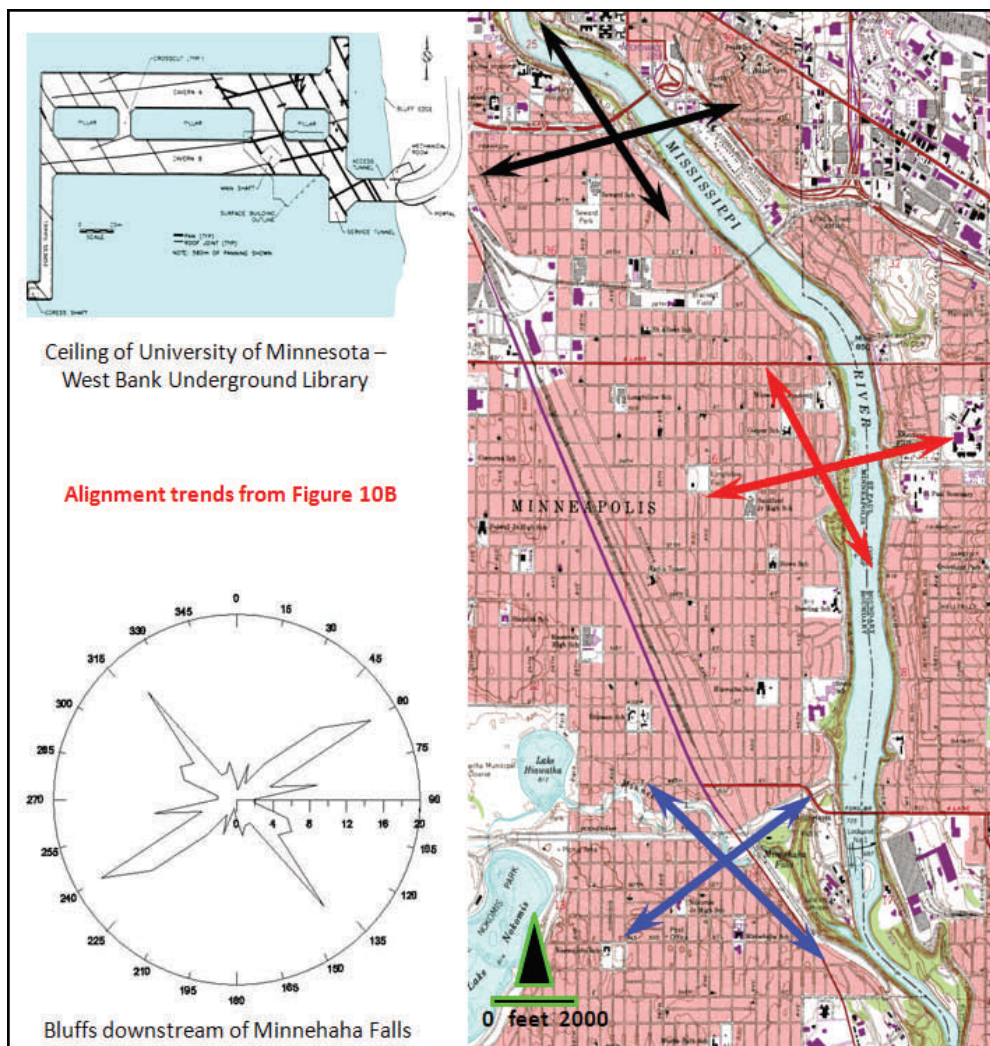


Figure 11. Comparison of alignment trends from vicinity of 34th Street sinkhole and Channel Rock Cavern with major joint orientations from two locations in eastern Minneapolis. Measurement location is at the intersection of trend arrows. (Joint pattern insets from E. C. Alexander, Jr., Bruce M. Olsen unpublished data, and Kelton Barr Consulting, 2000.)

with varying degrees of rigor based upon the current status of investigation of the features, indicate the presence of breccia pipes extending through the St. Peter, any overlying bedrock units, and the glacial overburden. The locations of these breccia pipes may be related to their proximity to a bedrock valley that is currently or was formerly a regional groundwater discharge point. There is also suggestive evidence of structural geologic features providing points of weakness for breakthrough of upwardly migrating groundwater. Considering that there are a number of buried bedrock valleys in the metropolitan area and the potential complexity of structural features within the Twin Cities Structural Basin, these combined hydrogeologic and structural conditions may have occurred in a number of other locations around the metropolitan area. Any breccia pipes already formed in those locations may have been buried by the glacial deposits now covering the area's surface. The possibility for other breccia pipes in the St. Peter within the metropolitan area is both probable and problematic.

Acknowledgements

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References

- Alexander, Jr., E.C., K.D. Barr, and S. Alexander. 2008. Goliath's and Mystery Caves, Minnesota: Epigenic modifications and extensions of preexisting hypogenic conduits. *Geological Society of America Abstracts with Programs* 40 (6): 341.
- Barr, K.D., and A.B. Klimchouk. 2007. Hypogenic karst and its implications for Minnesota hydrogeology: Minnesota Ground Water Association Fall Conference, November 2007. St. Paul: Minnesota Ground Water Association.

- Barr, K.D., A.B. Klimchouk, and E.C. Alexander, Jr. 2008. Hypogenic karst and its implications for Minnesota hydrogeology. In Proceedings of the Eleventh Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, Tallahassee, Florida: American Society of Civil Engineers Geotechnical Special Publication no. 183, ed. L.Y. Yuhr, E.C. Alexander, Jr. and B.F. Beck, 43-53. Reston: American Society of Civil Engineers.
- Kelton Barr Consulting, Inc. 2000. *Bluff area summary report*. Report to Minnehaha Creek Watershed District.
- Klimchouk, A.B. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Klimchouk, A., and V. Andrejchuk. 2005. Karst breakdown mechanisms from observations in the gypsum caves of the western Ukraine: implications for subsidence hazard assessment. *Environmental Geology* 48 (3): 336-359.
- Mossler, J.H., and B.A. Bloomgren. 1990. Bedrock geology, plate 2. In *Geologic atlas of Washington County: Minnesota Geological Survey County Atlas C-5*, ed. L. Swanson and G.N. Meyer. Minneapolis: Minnesota Geological Survey.
- Mossler, J.H., and R.G. Tipping. 2000. Bedrock geology and structure of the seven-county Twin Cities metropolitan area, Minnesota. *Minnesota Geological Survey Miscellaneous Map M-104*. Minneapolis: Minnesota Geological Survey.
- Patterson, C. J., J. H. Mossler, and B.A. Bloomgren. 1990. Bedrock topography and depth to bedrock, plate 4. In *Geologic atlas of Washington County: Minnesota Geological Survey County Atlas C-5*, ed. L. Swanson and G.N. Meyer. Minneapolis: Minnesota Geological Survey.
- Runkel, A.C., R.G. Tipping, E.C. Alexander, Jr., and S.C. Alexander. 2006. Hydrostratigraphic characterization of intergranular and secondary porosity in part of the Cambrian sandstone aquifer system of the cratonic interior of North America: Improving predictability of hydrogeologic properties. *Sedimentary Geology* 184 (3-4): 281-304.
- Runkel, A.C., R.G. Tipping, E.C. Alexander, Jr., J.A. Green, J.H. Mossler, and S.C. Alexander. 2003. *Hydrogeology of the Paleozoic bedrock in south-eastern Minnesota*. Minnesota Geological Survey Report of Investigations 61. St. Paul: Minnesota Geological Survey.
- Tipping, R.G., J.A. Green, and E.C. Alexander Jr. 2001. Karst features, Plate 5. In *Geologic atlas of Wabasha County, Minnesota: Minnesota Geological Survey County Atlas C-14, Part A*, ed. A. C. Runkel. St. Paul: Minnesota Geological Survey.
- Tipping, R.G., A.C. Runkel, E.C. Alexander Jr., S. . Alexander, and J. A. Green. 2006. Evidence for hydraulic heterogeneity and anisotropy in the mostly carbonate Prairie du Chien Group, south-eastern Minnesota, USA. *Sedimentary Geology* 184 (3-4): 305-330.
- Wright, Jr., H.E. 1972. Quaternary history of Minnesota. In *Geology of Minnesota: A centennial volume*, ed. P.K. Sims and G.B. Morey, 515-547. St. Paul: Minnesota Geological Survey.

STRUCTURAL AND FACIES CONTROL OF HYPOGENIC KARST DEVELOPMENT IN THE GUADALUPE MOUNTAINS, NEW MEXICO, USA

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Abstract

The Guadalupe Mountains are the exhumed remnants of a Permian reef complex uplifted beginning around 21 million years ago. The caves of the Guadalupe Mountains were formed by the mixing of deep-circulating meteoric water and hydrogen-sulfide-rich brine derived from surrounding oil and gas deposits. Cave development was controlled by fracture zones, faults, and structures associated with Permian and Tertiary tectonics.

Cave development shows strong linear trends that are correlative to broad structural trends in the Guadalupe Mountains and reflect fracturing, faulting, and folding during uplift. Some anticlinal features reflect deposition of Permian sediments across syndepositional faults. Many of these syndepositional faults can be observed in the caves and exhibit a strong influence on both overall passage trends and on passage character.

There are large breccia zones associated with syndepositional faults, forereef deposits, and possibly

development of paleokarst. In these areas, cave passages typically change from large, linear trunk passages to complex spongework mazes of smaller passages. Forereef deposits, paleokarst, and paleochannels through the reef can also be preserved as breccia zones and have a similar effect on passage character.

While overall speleogenesis crossed formational boundaries, lithology had some influence on passage character. The backreef units contain more rectilinear maze-type passages than the underlying reef and forereef units. Large trunk passage development is prevalent in the Capitan formation, especially along the reef/forereef transition. Other facies changes such as alternating layers of siltstone and dolomite, and development of tepee structures exhibit significant, but more localized controls on passage character.

Introduction

The Guadalupe Mountains (Figure 1) are the exhumed remnants of a Permian reef complex formed along the edge of an inland sea beginning around 254 Ma and

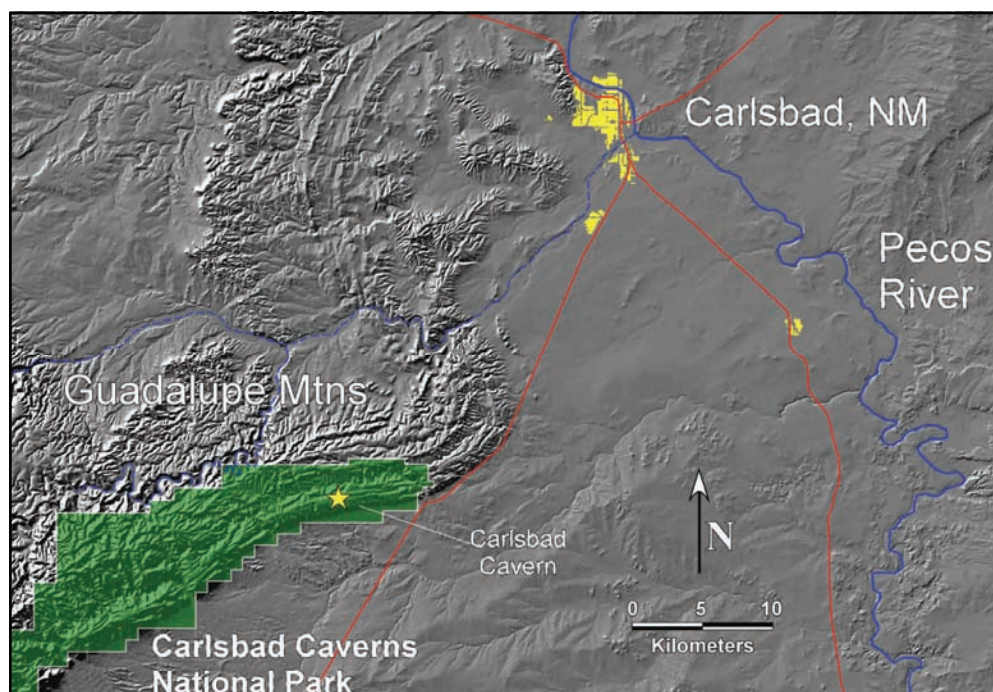


Figure 1. Area map of Guadalupe Mountains and Carlsbad Caverns National Park.

exposed during middle to late Tertiary uplift and erosion (King, 1948; Newell et. al, 1953; Hill, 1996). The reef (the massive member of the Capitan Formation) was composed primarily of a calcareous sponge and algae framework with some bryozoans, bivalves, marine snails, and various microorganisms (Figure 2; King, 1948; Newell et. al, 1953). The majority of the reef volume is made of very early marine cement which has led to the excellent preservation of the fossils and lack of wholesale diagenetic alteration of the limestone.

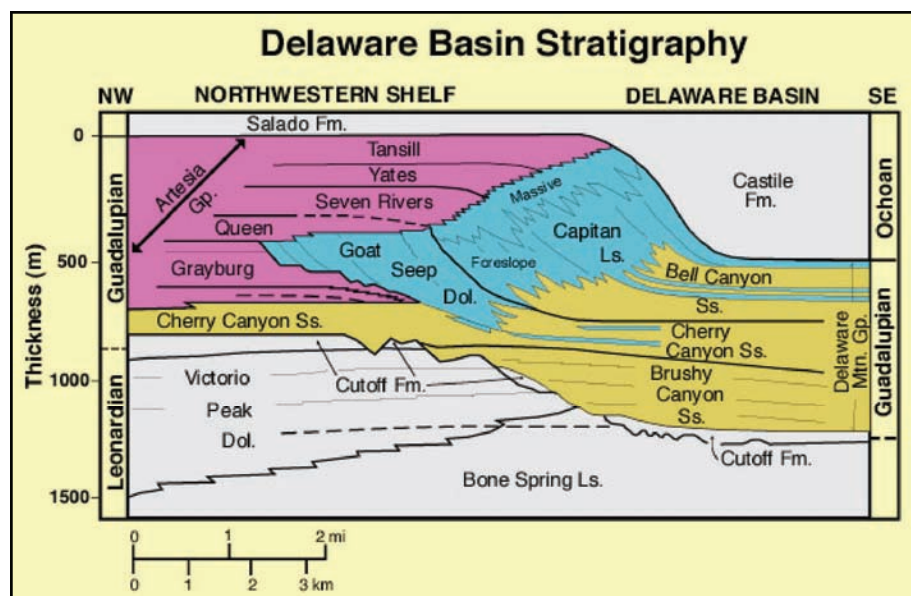


Figure 2. General stratigraphy of the Guadalupe Mountains and

During growth of the reef, large blocks of cemented material and other gravity-driven deposits were shed into the deeper water of the basin. These deposits make up the breccia or foreslope member of the Capitan Formation and are also characterized by early marine cementation (Newell et. al, 1953).

On the shelf landward of the reef, alternating layers of carbonate siltstones (lower sea level) and dolomite (elevated sea level) were deposited in lagoonal environments that varied from unrestricted to restricted marine conditions. These deposits are represented by the backreef Seven Rivers, Yates, and Tansill formations (Newell et. al, 1953).

Around 251 Ma, the connection between the inland sea and the open ocean began to close off, leaving a thick sequence of evaporite deposits subsequently overlain by an accumulation of younger sediments through the early Tertiary (Hill, 1996). These younger sediments began to be eroded away during the development of the Laramide-aged Alvarado Ridge 35-38 Ma. This ridge was a broad topographic feature that extended from southern Wyoming, through west Texas and into northern Mexico (Eaton, 1987). During Basin and Range extension beginning around 21 Ma, the Alvarado Ridge was broken into smaller structural blocks, including the Guadalupe Mountains. Neither of these tectonic periods resulted in major deformation, but caused minor tilting (less than 10 degrees) and fracturing of the Permian rocks. It was these fractures that became the major pathway for fluid flow and cave development.

The Guadalupe Mountains possess few characteristics of karst areas; there are few sinkholes and karst springs and the lack of surface stream flow is more a reflection of the area being part of a desert than the presence of subterranean pathways for the water. Early explanations of cave development still focused on underground streams and traditional models of carbonic acid speleogenesis. The presence of large deposits of gypsum and other unusual mineral deposits coupled with the lack of fluvial features led researchers to develop a model of hypogenic cave development where sulfuric acid was the primary agent of dissolution. Early ideas examined the source of sulfuric acid being oxidation of pyrite from the backreef units (Hill, 1987). More recent studies have shown that sulfuric acid is related to the oxidation of hydrogen sulfide generated from the oil and gas deposits surrounding the Guadalupe Mountains (Jagnow et. al., 2000).

Initial fracture enlargement and solution of the Permian rocks took place as the result of strong groundwater flow from the upland areas of the Alvarado Ridge developed between 35 and 38 million years ago (DuChene and Cunningham, 2006). Larger passages were then formed by the mixing of deep-circulating meteoric water and hydrogen-sulfide-rich brine derived from surrounding oil and gas deposits beginning 12-15 million years ago (Palmer and Palmer, 2000, Polyak et al. 1998). The sulfuric acid speleogenesis formed large, irregular rooms and three-dimensional spongework passages (Palmer and Palmer, 2000).

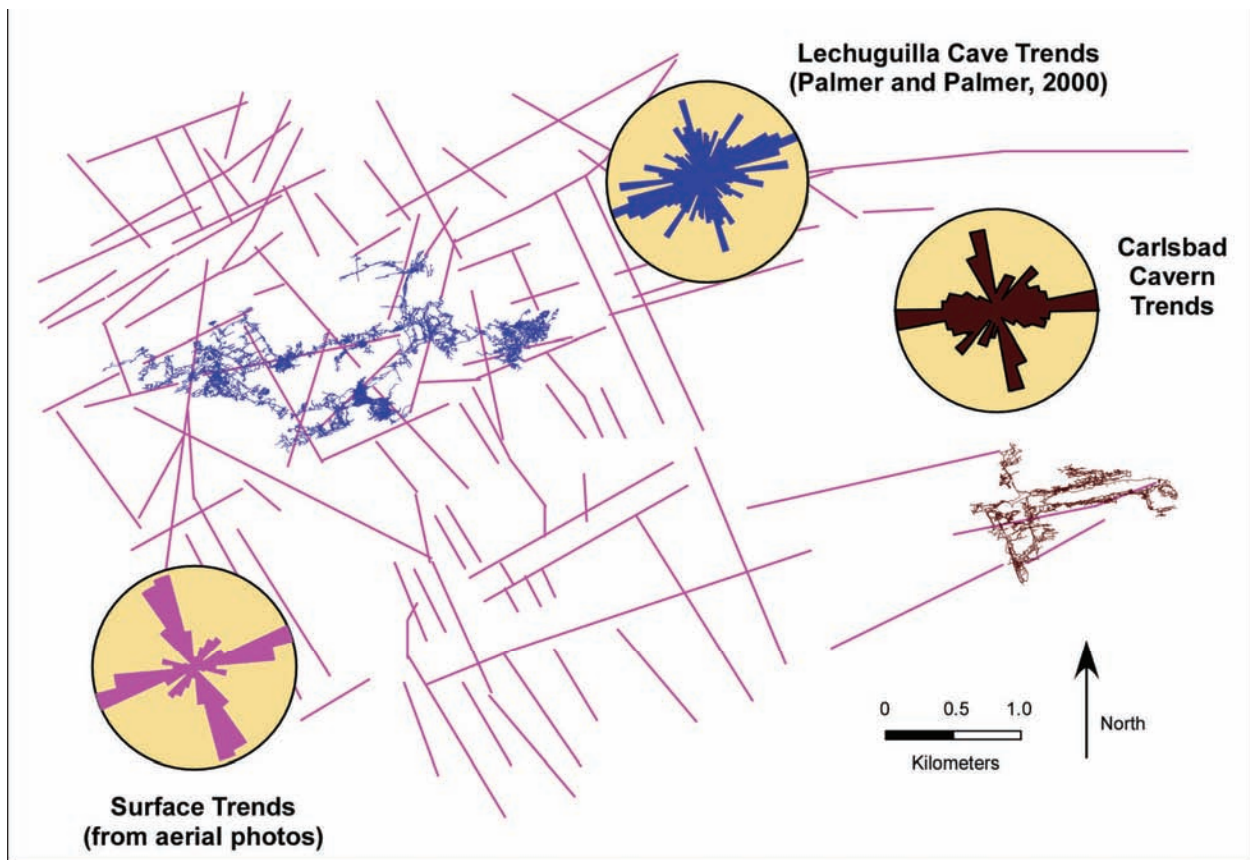


Figure 3. Surface and cave structural trends.

Figure 4. Dry Cave showing bedrock orientations (Allison and Stockton, 2009).

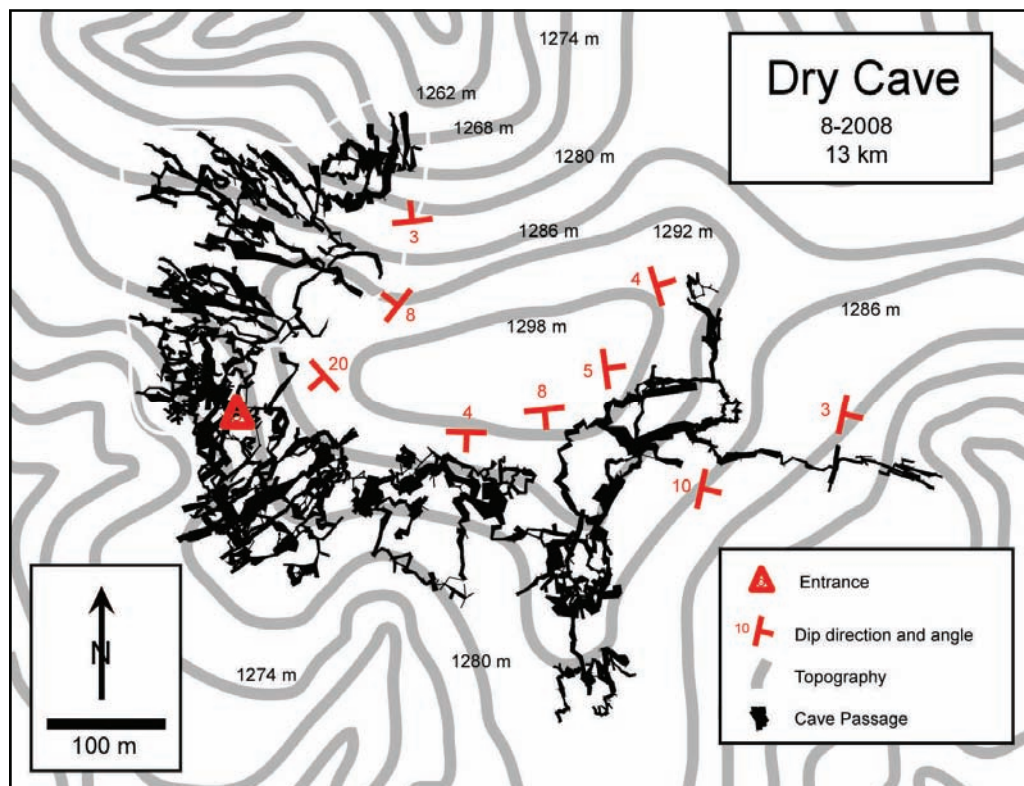


Table 1. Paleokarst fill and breccias (after Kosa, 2003).

Type	Characteristics	Matrix	Depositional Environment
Limestone / dolomite and related breccia (LB)	Massive, laminated fill or chaotic breccia. Breccia contains angular clasts of platform carbonates and siliciclastics, and earlier paleocavern fills	Lime mudstone to lime/dolomite grainstone with Permian bioclasts	Solution collapse breccias with platform-derived carbonate matrix; highstand deposits
Microspar-lithified breccia (MSB)	Chaotic or layered breccia with angular to rarely rounded or etched clasts of platform carbonates and siliciclastics, and earlier paleocavern fills	Fine calcite spar with Permian bioclasts	Chaotic and traction-reworked solution/collapse breccia; spar replacing primary carbonate or evaporite matrix
Spar-lithified breccia	Breccia with angular clasts rarely etched on edges made up of platform carbonates and siliciclastics	Blocky calcite spar	Fault-related cataclastic breccia, solution collapse breccia, sedimentary breccia; calcite spar replacing primary carbonate or evaporite cement/matrix
Beige dolomitic silt/sandstone and related breccia (DS, DSB)	Massive, crudely laminated; breccia is chaotic, layered, and graded with irregular, etched clasts of platform carbonates, siliciclastics, and earlier paleocavern fills	Dolomitic silt/sandstone with Permian bioclasts	Solution collapse breccias with platform-derived siliciclastic matrix; lowstand deposits
Carbonate-rich breccia (CB)	Massive and laminated; breccia is chaotic, layered, and graded with angular to rounded clasts of host-rock carbonates, siliciclastics, and earlier carbonate-rich fills	Siliciclastic-bearing dolomudstone-packstone with Permian bioclasts	Chaotic and traction-reworked solution/collapse breccia with matrix of mixed platform-derived lime mud and lowstand siliciclastics; transgression/regression deposits
Pink and red dolomitic silt/sandstone to silt/sand-bearing dolostone and related breccia (P, PB, Rd, RdB)	Massive and laminated; breccia is chaotic, layered, and graded with angular to rounded (rarely etched) clasts of host-rock carbonates, siliciclastics, and earlier carbonate-rich fills	Siliciclastic-bearing dolomite to dolomitic silt/sandstone with Permian bioclasts	Origin as of DS, DSB, and mainly CB lithologies modified by mineralization by oxidized Fe and Mn and by early dolomitization

Hydrologic control

The vertical distribution of cave passages was controlled by changing groundwater flow conditions caused by uplift and faulting of the meteoric catchment area (Polyak, et. al 1998; DuChene and Cunningham, 2006) and downcutting of the Pecos River Valley. Changes in hydrologic conditions lowered the mixing zone with time, creating distinct cave development levels identified by many authors (Jagnow, 1977; Hill, 1987; Polyak et al., 1998). The primary focus of this paper, however, is on the spatial distribution of cave patterns and the controls on lateral cave development.

Structural control

Major faults and folds in the Guadalupe Mountains were controlled by sediment compaction in the Permian, and Permian to Tertiary tectonics. The east-northeast trend reflects the development of fractures parallel to the shelf margin and developed as a result of sediment loading during deposition, as the massive reef overstepped reef talus. The north-northwest trend

is younger, and is related to later uplift associated with the growth of the Alvarado Ridge (Harvey DuChene, written communication, 2009).

Jagnow (1977) documented the coincidence of major fault and fold trends in the Guadalupe Mountains with fractures and cave passage orientation mapped in Carlsbad Cavern. Similar relationships between regional structural trends and Lechuguilla Cave passage orientation have also been well documented (Jagnow, 1989 and Palmer and Palmer, 2000). While the timing of some of the structures in the Guadalupe Mountains is still problematic (Hill, 1996), there is clear influence of the structure and fracture patterns on overall cave development (Figure 3).

Local structures can also strongly influence cave development. Dry Cave on McKittrick Hill west of Carlsbad appears to be developed around the center of a dome in the backreef (Figure 4). Dips measured at points throughout the cave show a consistent trend away from a central high point (Allison, 2009). This

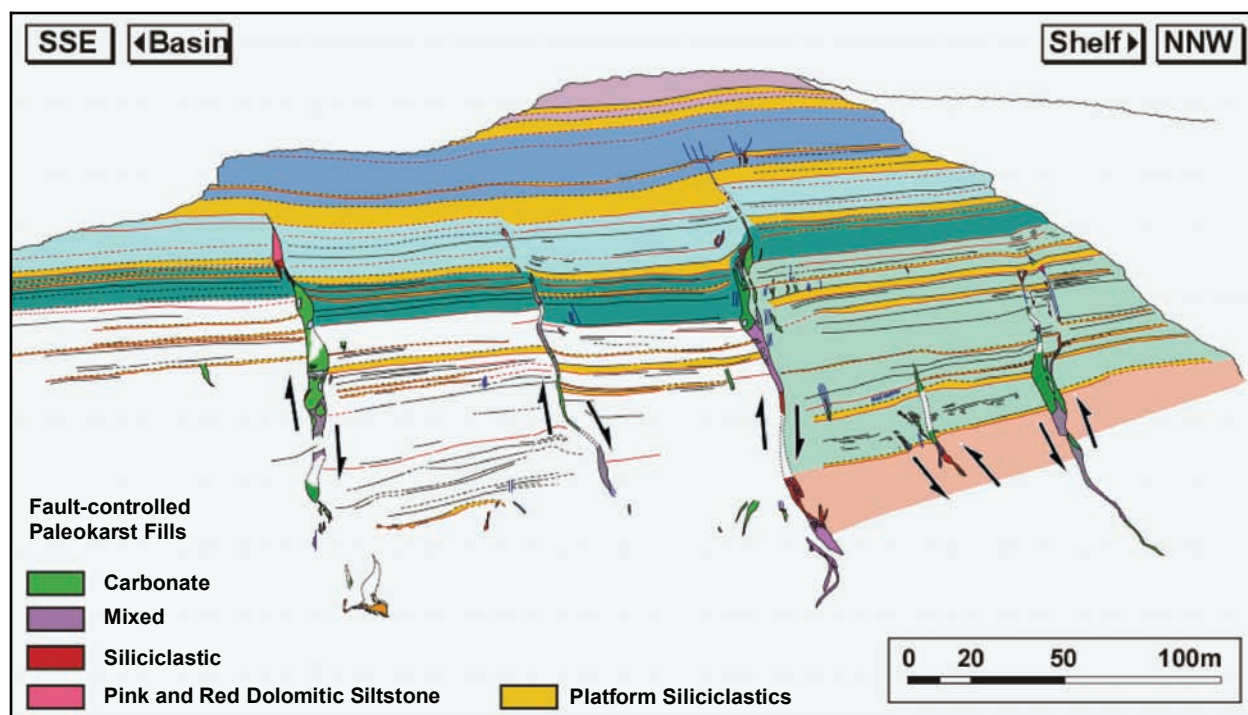


Figure 5. Profile view of Slaughter Canyon outcrop showing variable fill types in solution-modified syndepositional faults (from Kosa and Hunt, 2006). Also note the way bedding appears to drape over fault tips resembling shallow antiforms.

feature may be a bioherm, patch reef, or other positive feature underlying the backreef sediments (Noe and Mazzullo, 1994).

Syndepositional faults

Hunt and others (2002) did extensive mapping of stratigraphy in Slaughter Canyon and Rattlesnake Canyon in Carlsbad Caverns National Park. They found significant syndepositional faulting and associated breccia zones, filled fractures, mineralization, and localized paleokarst development (Kosa, 2003). Many of these faults are represented by chaotic breccias at depth but grade upward into sediment filled fractures and in some cases are represented only as small folds in the backreef units (Figure 5).

While many of these structural features can be observed in outcrop and on aerial photographs, their influence on cave development is best mapped in the subsurface. The most extensive caves in the Guadalupe Mountains are Carlsbad Cavern (48 km long) and Lechuguilla Cave

(202 km long), both within the boundary of Carlsbad Caverns National Park. Carlsbad Cavern penetrates the Tansill, Yates, and Capitan formations. Lechuguilla Cave (Figure 8) penetrates the Yates, Seven Rivers, and Capitan formations and possibly the Goat Seep and Queen formations (Palmer and Palmer, 2000). Both caves provide access to extensive three-



Figure 6. Fault exposed in Glacier Bay (oblique view) with person for scale – NPS (photo by Stan Allison).

dimensional views of the geology and allow for direct observation of geologic structure and features that have influenced cave development. Many of the features seen in these two caves can be observed in caves throughout the Guadalupe Mountains, but Carlsbad and Lechuguilla will be used as representative examples.

The entrance of Lechuguilla Cave has long been thought to have formed along the crest of a small anticline (Jagnow, 1989). Breccia seen in the entrance pit and throughout the entrance series of the cave suggests that passage development is controlled by a syndepositional fault that is represented by an antiform near the cave entrance and by breccia deposits within the cave. Boulder Falls was mapped to coincide with a vertical breccia “dike” (Palmer and Palmer, 2000). An obvious fault can be seen in the wall in Glacier Bay and appears to be aligned with the trend of the entrance series (Figure 6). The coincidence in orientation of these features suggests they are related to one significant fault that can be traced from the Entrance to Glacier Bay over 470 m.



Figure 7. Fault Breccia in Quasimodos Lair, Far East, Lechuguilla Cave (photo by Paul Burger).

The Firefall Hall – Quasimodos Lair in the Far East section of Lechuguilla (Figure 8) is a major passage that cuts 122 meters vertically through formational boundaries from the Capitan all the way into the backreef. At the upper terminus of the room, it is clear that the passage has developed along a significant

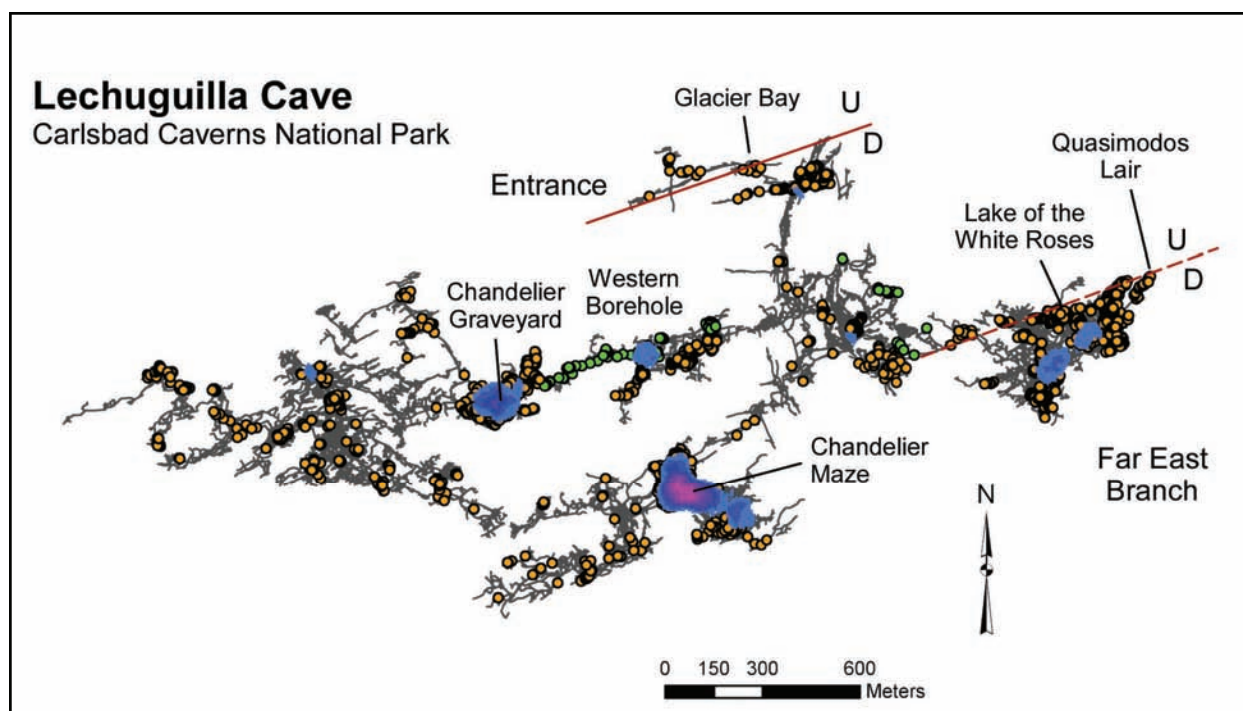


Figure 8. Lechuguilla Cave (orange: mapped breccia; green: forereef breccia; red: observed fault trends; blue to purple shows increasing concentration of maze passages).

fault (Figure 7). This fault has no expression on the surface above the cave but has clearly influenced passage development. This shows both the difficulty in using surface mapping to predict cave development and illustrates the usefulness of the underground outcrops accessible in Guadalupe Mountain caves.

The route down to Lake of the White Roses in the Far East part of Lechuguilla Cave is a steeply descending fissure. The hanging wall of this passage is a breccia with subangular clasts up to 20 cm wide and the footwall contains no breccia. This fissure is not oriented parallel to the main reef trend and is unlikely to represent the contact between the massive and forereef members of the Capitan Formation. More likely this fissure represents dissolution along a linear fracture or fault developed prior to the onset of speleogenesis.

Breccia bodies

While the controls on overall orientation and patterns are generally understood, there has been little study of the controls of passage character in Guadalupe Mountain caves. Some cave passages show dramatic changes in character over a very short distance within the same stratigraphic unit. Specifically, some large trunk passages more than 15 meters in diameter terminate in spongework mazes with passages typically less than 2 meters in diameter.

Many of the maze areas in Lechuguilla Cave contain a significant amount of breccia. Mapping and inventory data were used to determine if there was a direct correlation between the transition into maze passages and these breccia zones. To date, more than 13,000 individual stations throughout the cave have been

inventoried for geologic and speleogenetic character, including the presence of breccia, bedrock geology, and mineralization.

Survey data were used to characterize parts of the cave into maze and non-maze passages. Places in the cave that were the junction of four or more other passages were considered to be maze (Figure 8). These data were then correlated to locations where breccia was indicated on the inventory. A density plot was then made showing areas where both maze passage and breccia were concentrated. Analysis of the data shows a definite correlation between the maze areas of the Chandelier Maze, the Chandelier Graveyard, and portions of the Far East with large breccia zones.

The character of all the different breccia zones has not yet been determined. Much of the breccia in the Chandelier Graveyard shows evidence of transport and sorting, with a general dip to the southwest. In some areas, the clasts are uniformly cobble-size and have imbricate stacking. The lithology of the clasts include reef rock, sandstone, pisolitic limestone, and carbonate grainstone typical of the Seven Rivers Formation and are interpreted to represent deposition in a channel through the reef (Harvey DuChene, written communication, 2009).

DuChene (2000) classified the breccia bodies found in Lechuguilla Cave based on his detailed mineral inventory (now integrated into the broader, more general inventory database). The breccia bodies were divided into depositional breccias, related to the growth of the Permian Reef and part of the bedrock, and speleogenetic breccias, related to the development

Table 2. *Depositional breccias in Lechuguilla Cave (after DuChene, 2000).*

Type	Characteristics	Matrix	Occurrence	Depositional Environment
<u>Mosaic Breccias</u>				
	Clasts sharply angular with no rounding of corners; very slight or no rotation of clasts; unsorted; well-cemented	Calcite (predominant); silt (minor)	Fracture fill (fractures open and sub-vertical)	Tension fractures in massive reef and backreef limestone
			Collapsed beds (deposits sub-horizontal)	Bedding collapse and fragmentation due to dissolution of underlying beds
<u>Chaotic Breccias</u>				
	Clasts angular to subrounded, poorly sorted, mixed lithologies		Large masses parallel to dip of forereef slope	Forereef slope
	Clasts angular to subrounded, poorly to moderately sorted, mixed lithologies including bedded pisolitic limestone and siltstone; moderately imbricate; well cemented	Calcite (predominant); silt (minor)	Channel fill	Passes and channels through reef

of the cave. The depositional breccias were subdivided into mosaic and chaotic types and characterized to aid in the identification of their possible depositional environments (Table 2).

Syn depositional faults also produced depositional breccias during the Permian (Hunt and Kosa, 2002; Kosa, 2003). Some of these were unmodified fault breccias and others were heavily modified by karstification. Some faults contain various fills representing deposition during high and low sea level stands and reworking of earlier fill material (Table 1).

In Virgin Cave, located in the Lincoln National Forest, the main trend of the front section of the cave is oriented along a linear, near-vertical, tabular breccia body approximately 217 meters long and up to 46 meters wide oriented northeast-southwest (Figure 9). Passage developed within the breccia consists of spongework maze passage redissolved out of breccia fill and clasts more than 2 meters in diameter. Individual stratigraphic units cannot be mapped across this feature so it is uncertain whether this feature is a fault or an enlarged fracture.

Davis (1980) and Queen (1981) identified spar and cemented-sediment-filled vugs in Carlsbad Cavern that may represent pre-Tertiary karstification of the reef rocks. There is some evidence that early karstification of the Permian rocks occurred during brief times of subaerial exposure and may have been concentrated along syn depositional faults (Kosa and Hunt, 2006).

Early karstification could be a source of breccia bodies exposed in cave passages. Paleokarst breccias have been identified in other parts of the basin and have been shown to influence later fluid flow and emplacement of petroleum (Loucks and Anderson, 1980; Kerans 1990). Kosa and Hunt identified two types of paleocaverns in the Capitan platform: 1) narrow, planar, elongate paleocaverns that mimic precursor fault and fracture orientations and are dominated by siliciclastic and mixed, carbonate-siliciclastic fills; and 2) bulbous paleocaverns that extend far from their host fault and fracture zones, are up to 90 m wide, and are dominated by carbonate fills. From the lateral and vertical relationships between the two types of paleokarst, they concluded that the

bulbous paleocaverns pre-date the planar features and may have acted to localize subsequent faulting and later karst development. How these karst types may relate to those identified by Davis (1980) and Queen (1981) is unclear.

Since many of these breccias are similar and can vary vertically and laterally, the only way to differentiate them is through detailed mapping of breccia zone geometry. Reef channels and passes would tend to run perpendicular to the reef trend and would have specific diagnostic features such as clast rounding and imbrication. Fault breccias would tend to be more tabular and oriented with major structural trends in the area. Paleokarst breccias would be more limited in horizontal and vertical extent and could contain mineralization consistent with deposition in an open cavity rather than a sediment-filled fault or channel. Forereef breccia would parallel the reef trend, would be more spatially extensive, and would be limited to parts of the cave developed stratigraphically lower in the limestone.

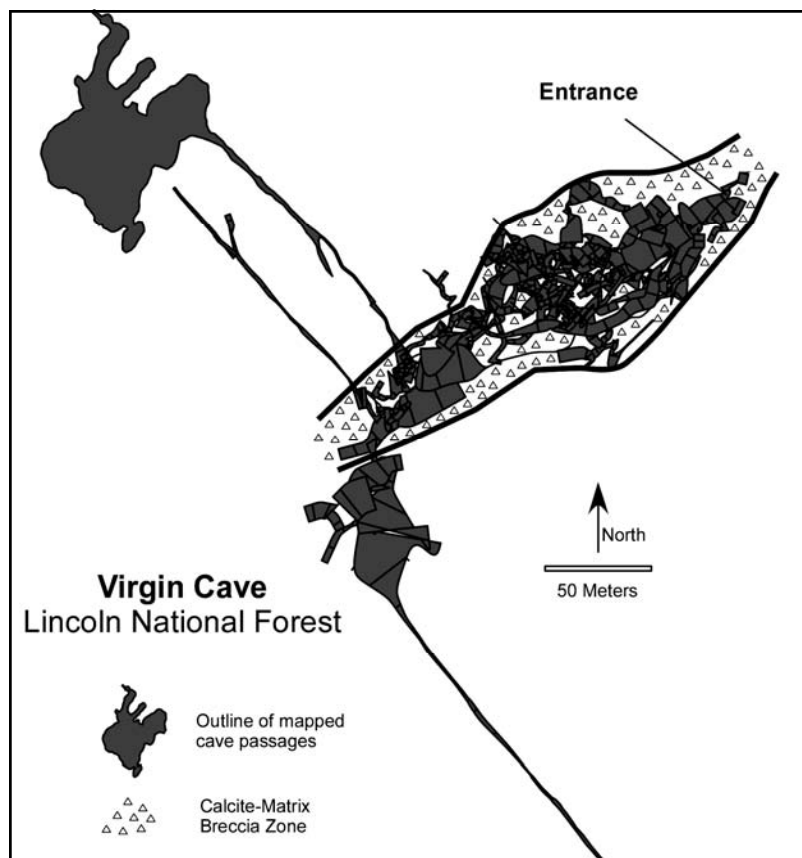


Figure 9. Plan map of Virgin Cave showing large breccia zone where cave is developed as a spongework maze.

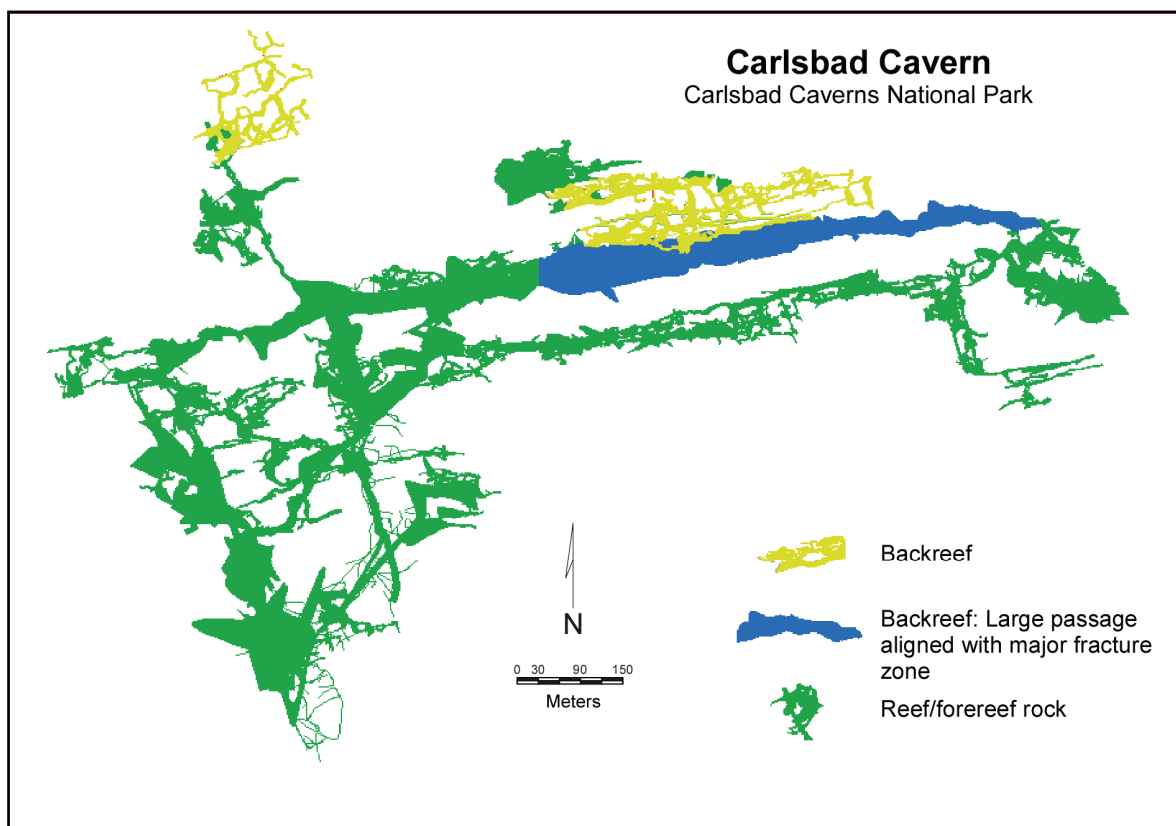


Figure 10. Map of Carlsbad Cavern showing rectilinear cave development in backreef units (yellow) and ramiform passages in reef rocks and along primary conduits (green and blue respectively).

Lithology

Most of the major passage trends in Lechuguilla Cave and Carlsbad Cavern cut through different rock types, so there is little lithologic control on overall cave

pattern. However, local passage character can be affected by changes in lithology. Jagnow (1989) recognized that most of the Western Borehole is developed along a linear breccia deposit that parallels the reef trend for 2.1 km (Figure 8). He attributed

these deposits to dissolution of Permian evaporites and subsequent collapse of the overlying limestone. DuChene (2000) identified the bedrock breccia deposits in the Western Branch below 250 m as belonging to the forereef talus or breccia member of the Capitan. It seems likely the latter is the correct interpretation and that the Western Borehole was developed at the contact between the massive and breccia members of the Capitan.



Figure 11. Tepee structure located near the natural entrance to Carlsbad Cavern.

Bedding thickness has had some influence on local

passage character. In the bedded backreef units, the fracture and cave patterns follow the regional trends but are more closely spaced than in the more massive Capitan. This has resulted in the development of smaller (typically less than 3 m in diameter) rectilinear maze passages in the backreef units. A plan map of Carlsbad Cavern (Figure 10) illustrates the differences in passage character between the passages developed in the backreef and those developed in the Capitan.

Passage development can also be influenced by sedimentary structures and diagenesis. Jagnow (1977) documented that passages in Tepee Cave and Jurnigans Cave #2 were aligned along the axes of tepee structures, polygonal features in the backreef units thought to be expansion features related to fluid flow and mineral precipitation in lagoonal carbonate sediments (Figure 11; Scholle, 2008; also see discussion of tepee origins in Hill, 1996, p. 75-78). The passages in Yellowjacket Cave in Dark Canyon are also oriented to the axes of tepee structures (J.M. Queen pers. comm., 2006). While the overall pattern of the cave is oriented to major fractures, many passages in Spider Cave are also aligned with the axes of tepee structures. In most cases, the influence of these sedimentary structures is minor, but it does have some influence on cave passage development.

Conclusions

Bedrock heterogeneities, including faults and fractures and changes in lithology, have influenced hypogenic speleogenesis in the Guadalupe Mountains. The patterns of cave development display a hierarchy of influence.

1. Regional paleohydrology and barriers to flow have controlled the location of caves and the vertical distribution of cave passages.
2. Overall orientation and distribution of cave passages are controlled by regional structures that were active at least since the Permian and possibly earlier. Structures such as small domes have had localized, but important influence on fracturing and cave development in some areas.
3. Permian-aged syndepositional faults have had strong influence on passage development (e.g. Lechuguilla and Virgin Caves), both in major passage orientation and by changing dissolution patterns. Breccia bodies associated with these faults and other changes in facies have resulted in the development of extensive spongework mazes.
4. Smaller scale changes in lithology, bedding thickness, and facies (e.g. tepee structures) can

affect passage character and can be locally important to dissolution (e.g. Spider and Yellowjacket caves). However, these small-scale variations are not major factors in overall cave development in the Guadalupe Mountains.

Further work

To determine the significance of syndepositional faults and brecciation on cave development, more detailed mapping is needed. Brecciation within caves must be mapped with respect to their overall geometry, lithology variation, and influence on speleogenesis. Paleokarst and karst-modified faults must be identified and characterized to determine what, if any, influence they have had on later cave development.

Detailed mapping of these discrete features, coupled with further general bedrock mapping should provide additional information on the paleohydrology and speleogenesis of Guadalupe Mountain caves. Since many of these features also have an influence on modern water infiltration and nutrient pathways, mapping may provide some insight into the microbial variability in these caves.

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References

- Allison, S.D., and A.J. Stockton. 2009. Exploration in Dry Cave 2005-2009, Guadalupe Mountains, New Mexico. Proceedings of the 2009 International Congress of Speleology (in press).
- Davis, D.G. 1980. Cave development in the Guadalupe Mountains: a critical review of recent hypotheses. *NSS Bulletin* 42 (3): 42-48.
- DuChene, H.R. 2000. Bedrock features of Lechuguilla Cave, Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62 (2): 109-119.
- DuChene, H.R., and K. Cunningham. 2006. Tectonic influences on speleogenesis in the Guadalupe Mountains, New Mexico and Texas. In *New Mexico Geological Society, Guidebook 57*, ed. L. Land, V.W. Lueth, W. Raatz, P. Boston and D. Love, 211-217. Socorro: New Mexico Geological Society.
- Eaton, G. 1987. Topography and origin of the Southern Rocky Mountains and Alvarado Ridge. In *Continental Extension Tectonics: Geological Society Special Publication 28*, ed. M. Coward,

- J.F. Dewey and L. Hancock, 355-369. London: The Geological Society.
- Hill, C.A. 1996. *Geology of the Delaware Basin – Guadalupe, Apache, and Glass Mountains, New Mexico and west Texas*. Midland: Society of Economic Paleontologists and Mineralogists, Permian Basin Section.
- Hill, C.A. 1987. *Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains, New Mexico and Texas*. New Mexico Bureau of Mines and Mineral Resources, Bulletin 117. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hunt, D.W., W.M. Fitchen, and E. Kosa. 2002. Syndepositional deformation of the Permian Capitan reef carbonate platform, Guadalupe Mountains, New Mexico, USA. *Sedimentary Geology* 154: 89–126.
- Jagnow, D.H., C.A. Hill, D.G. Davis, H.R. DuChene, K.I. Cunningham, D.E. Northup, and J.M. Queen. 2000. History of the sulfuric acid theory of speleogenesis in the Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62 (2): 54-59.
- Jagnow, D.H. 1989. The geology of Lechuguilla Cave, New Mexico. In *Subsurface and outcrop examination of the Capitan shelf margin, northern Delaware Basin: SEPM Core Workshop No. 13*, ed. P.M. Harris and G.A. Grover, 459-466. Tulsa: Society for Sedimentary Geology.
- Jagnow, D.H. 1977. *Cavern development in the Guadalupe Mountains*. Albuquerque: Adobe Press.
- Kerans, C. 1990. *Depositional systems and karst geology of the Ellenburger Group, (Lower Ordovician), subsurface West Texas*. University of Texas at Austin, Bureau of Economic Geology Report of Investigations No. 193. Austin: Bureau of Economic Geology.
- King, P.B. 1948. *Geology of the southern Guadalupe Mountains, Texas*. U.S. Geological Survey Professional Paper 215. Boulder: U.S. Geological Survey.
- Kosa, E. 2003. Heterogeneity in the structure, diagenesis and fill of syndepositional faults in carbonate strata: Upper Permian Capitan Platform, Guadalupe Mountains, New Mexico, USA. PhD diss., University of Manchester.
- Kosa, E., and D.W. Hunt. 2006. Heterogeneity in fill and properties of karst-modified syndepositional faults and fractures: Upper Permian Capitan platform, New Mexico, USA. *Journal of Sedimentary Research* 76: 131 - 151.
- Loucks, R.G., and J.H. Anderson. 1980. Depositional facies and porosity development in Lower Ordovician Ellenburger dolomite, Puckett field, West Texas. In *Carbonate reservoir rocks: SEPM Core Workshop No. 1*, ed. R.B. Halley and R.G. Loucks, 1-31. Tulsa: Society for Sedimentary Geology.
- Newell, K.J., A.G. Fischer, A.J. Whiteman, J.E. Hickox, and J.S. Bradley. 1953. *The Permian reef complex of the Guadalupe Mountains region, Texas and New Mexico*. San Francisco: W.H. Freeman and Co.
- Noe, S.U., and S.J. Mazzullo. 1994. Patch reef-dominated outer shelf facies along a non-rimmed platform, middle to upper Tansill Formation, northern Guadalupe Mountains, New Mexico. *West Texas Geological Society Bulletin* 33 (5): 5-11.
- Palmer, A.N., and M.V. Palmer. 2000. Hydrochemical interpretation of cave patterns in the Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62 (2): 91-108.
- Polyak, V.J., W.C. McIntosh, N. Güven, and P. Provencio. 1998. Age and origin of Carlsbad Cavern and related caves from $^{40}\text{Ar}/^{39}\text{Ar}$ of alunite. *Science* 279: 1919-1922.
- Queen, J.M. 1981. *A discussion and field guide to the geology of Carlsbad Caverns*. Preliminary report to the National Park Service for the 8th International Congress of Speleology. Carlsbad: National Park Service.
- Scholle, P.M. An Introduction and virtual geologic field trip to the Permian Reef Complex, Guadalupe and Delaware Mountains, New Mexico-West Texas: <http://geoinfo.nmt.edu/staff/scholle/guadalupe.html> (accessed December 10, 2008).

THE ROLE OF HYPOGENE PROCESSES IN SULFATE REDUCTION AND SPELEOGENESIS IN THE CASTILE FORMATION: EDDY COUNTY, NEW MEXICO AND CULBERSON COUNTY, TEXAS, USA

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Abstract

The Castile Formation crops out over ~1800 km² in Eddy County, New Mexico and Culberson County, Texas. GIS-analysis has indicated that over 9,000 karst features are likely to exist in this area. Many of these features are epigenic in origin. However, evidence of hypogene processes is widespread throughout the area. Dense clusters of hypogenic caves are typically associated with calcitized evaporites and selenite masses, suggesting a genetic relationship. Within these caves, specific morphologic forms (i.e. risers, cupolas, and half-tubes) provide evidence of their hypogenic origin.

Brecciation is common throughout the Castile Formation, indicating subsurface dissolution and collapse. Blanket breccias and subsidence troughs are suggestive of confined horizontal flow, while breccia pipes have been formed from upward stoping of subsurface voids. Calcitized evaporites are extensive throughout the region and are frequently associated with zones of brecciation. Deposits of native sulfur have also been observed in association with zones of brecciation, calcitized evaporites, and selenite masses.

The close proximity of hypogene caves, breccias, native sulfur, selenite masses, and calcitization suggests that hypogene processes have dominated sulfate diagenesis in the Castile Formation. Hypogene processes have provided pathways through which hydrocarbons from the underlying Bell Canyon formation have migrated upward and contributed to the calcitization of the Castile Evaporites. Hydrogen sulfide or native sulfur produced in this process has continued to migrate upward or laterally into oxic regions where selenite has been deposited. Significant overprinting due to surface denudation and epigenic processes has resulted in complex speleogenetic evolution within the Castile Formation.

Introduction

The Castile Formation crops out in the western edge of the Delaware Basin (Figure 1). The Delaware Basin

is part of the larger Permian Basin and lies at the boundary of the Basin and Range and the Great Plains provinces. The outcrop area lies at the northern end of the Chihuahuan Desert. The area averages 26.7 cm. of rainfall annually and has an average annual temperature of 17.3°C, ranging from a low of 9.2°C to a high of 25.2°C (Sares, 1984). Most of the annual rainfall occurs as afternoon thunderstorm showers or as

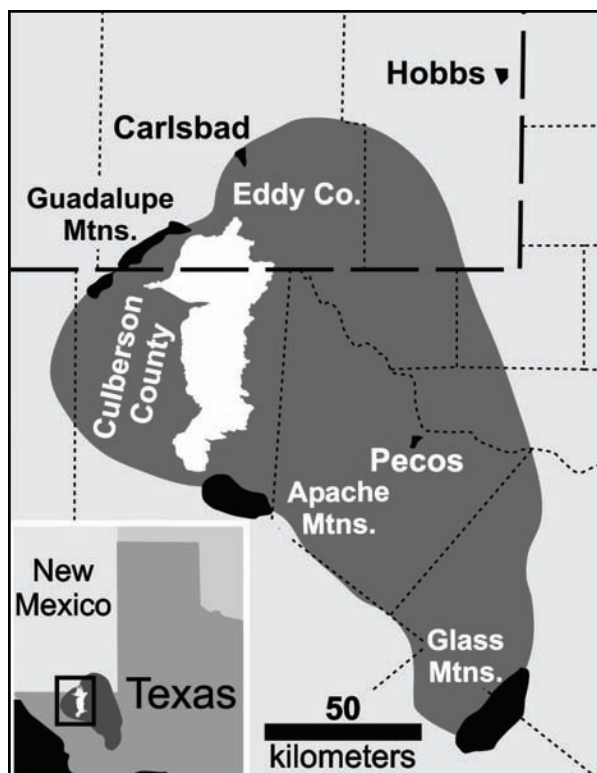


Figure 1. Location map showing location of Gypsum Plain including outcrop areas of the Castile Formation (solid white) and the Capitan Reef (solid black) within the Delaware Basin (dark gray), Eddy County, NM and Culberson County, Texas. Location of the Delaware Basin in relation to Texas and New Mexico is illustrated in bottom left corner, with the enlarged region outlined by the small black rectangle (adapted from Kelley, 1971, and Hill, 1996).

monsoonal storms. The outcrop covers an area of ~1800 km² in a band extending from Eddy County, New Mexico into Culberson County, Texas. The Castile Formation is a basin-filling evaporate sequence encircled by the Capitan Reef, a nearly continuous carbonate reef complex (Hill, 1996). The Castile is underlain by the clastic Bell Canyon and Cherry Canyon formations. After deposition, it was covered by the Permian evaporites of the Salado and Rustler Formations (Kelley, 1971) (Figure 2).

The Castile formation was deposited late in the Permian period and was cyclic, corresponding to periods of ocean input from the southwest (Hill, 1996). Lithologically, it contains anhydrite/gypsum, halite, and limestone beds. Additionally, vertical and horizontal breccias are found throughout the area. Most of the Castile Formation is varved anhydrite and halite or varved anhydrite and calcite (Anderson et al., 1972). The varves resulted from seasonal salinity cycles in the shallow inland Permian Sea. The horizontal breccias observed in the western end of the outcrop correlate to halite beds found in boreholes further east (Anderson et al., 1972, 1978). Carbonates diagenetically altered from anhydrite are found with the breccias (Anderson and Kirkland, 1980; Kirkland and Evans, 1976). Oil and gas exploration is abundant in the area. Sulfur has been economically produced from mines in the area as well.

Regional uplift began with the Laramide Orogeny, tipping the basin to the east/northeast at approximately 3-5° (Horak, 1985; Kirkland and Evans, 1976). During the mid-Miocene, roughly 30-40 Ma, basin-and-range expansion resulted in the development of

east-northeast trending faults. Hentz et al. (1989) describe these faults extending down into the Bell Canyon Formation and possibly deeper. Horak (1985) described the development of zones of west-east permeability throughout the basin resulting from tectonic activity. On the surface, grabens formed along these faults are evident along the western edge of the study area. Solution/subsidence valleys have formed along the surface between these fractures as a result of dissolution of halite by waters rising from depth along the fractures (Olive, 1957).

Horak (1985) described igneous activity that began during this period of basin-and-range expansion. According to Barker and Pawlewicz (1993), the Delaware Basin experienced high heat flow from 40-30 Ma related to igneous intrusions and from 23-0 Ma as a result of basin-and-range type block faulting. As a result, the basin has exhibited a higher than average thermal gradient (Hentz and Henry, 1989).

Diagenesis within the Castile Formation has been significantly influenced by hypogene karst processes (Stafford et al., 2008a,b). Hypogene karst develops in regions where rocks are confined or partially confined and the vertical or lateral migration of solutionally aggressive fluids produces porosity without any direct connection to meteoric processes (Klimchouk, 2007). Sulfate diagenesis associated with these zones of porosity has resulted in the development of carbonate masses frequently associated with deposits of sulfur and selenite (Anderson et al., 1980; Kirkland et al., 1976). Evidence of hypogene and hypergene (epigenic or meteoric) speleogenesis has been observed in numerous caves throughout the study area.

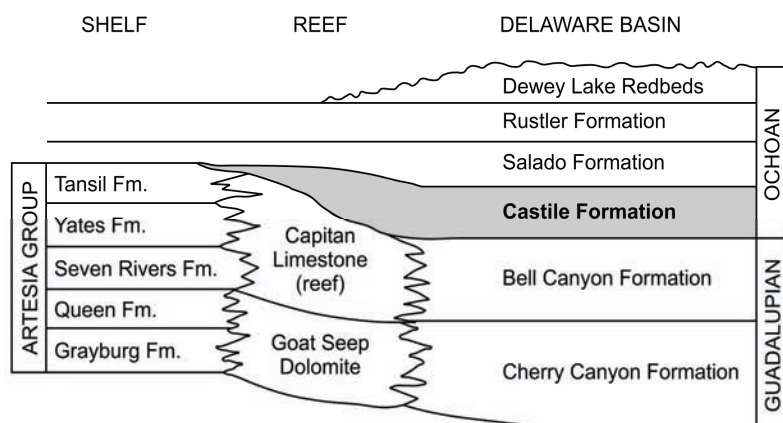


Figure 2. Diagrammatic representation of late Permian (Guadalupian and Ochoan) deposits associated with the Guadalupe Mountains (left) and Delaware Basin (right). Note that the Castile Formation fills in the basin and marks the beginning of the Ochoan (adapted from Hill, 1996).

Stafford et al. (2008a) described multiple solutional features formed by meteoric waters descending along zones of permeability. To evaluate the karst potential of the study area, they studied karst features found in fifty randomly selected 1-km² areas. In these areas, they described blind valleys and pirated arroyos, closed depressions, sinkholes, karst windows, karst springs and cenotes. The epigenic caves described typically had a large opening and closed rapidly, typically less than 100 meters in length and less than 5 meters in depth.

The most impressive karst in the area developed as hypogene karst. Hypogene karst develops in confined or

partially confined beds where lateral and vertical migration of fluids produces solutional porosity without direct connection to surface meteoric processes (Klimchouk, 2007; Stafford et al. 2008a). In hypogene processes, solutionally aggressive water is continually supplied to the dissolution front. Kirkland and Evans (1976) proposed that calcitized masses and sulfur deposits are the product of rising water. Large deposits of selenite produced by artesian water have been described (Crawford, 1993; Hill, 1996; Stafford et al., 2008a). All of these have been formed in and near breccias. Some of these breccias are in the form of breccia pipes and others are bedded, or blanket breccias. Both have been described as containing dissolution breccias and collapse breccias (Anderson et al., 1978). In some locations, these breccias are entirely in varved anhydrite. In others, the breccias are composed of calcitized clasts. In some instances, voids between the clasts are lined with anhydrite.

Hypogene karst development

Klimchouk (2007) defined karst as “an integrated mass-transfer system in soluble rocks with a permeability structure dominated by conduits dissolved from the rock and organized to facilitate the circulation of fluids.” He defined speleogenesis as “the creation and evolution of organized permeability structures in a rock that have evolved as the result of dissolutional enlargement of an earlier porosity.” These definitions expand the concept of karst and speleogenesis to include features with no surface connection. Using digital ortho photo quads and field observation, Stafford et al. (2008a) developed a karst potential map of the Castile outcrop in which they predicted over 9000 karst features in the outcrop area (Figure 3). Many epigenic features were noted, but a large number of the features were determined to be hypogenic in origin. Using nearest neighbor analysis (Ford and Williams, 2007), Stafford et al. (2008a) determined a nearest neighbor index of 0.439, indicating significant clustering of karst features. Klimchouk (2007) and Frumkin and Fischhendler (2005) indicated that hypogene karst tends to form in dense clusters, separated by regions of minimal karst. Among the hypogene features observed are calcitized buttes and masses, large deposits of selenite, sulfur, and breccias (Figure 4). Hypogene processes have been critical in the speleogenesis of several caves in the study area.

Tectonic effects

Tilting during the Laramide Orogeny resulted in a regional dip to the northeast. Recharge to the Bell Canyon aquifer was from the Guadalupe and Delaware Mountains to the west (Anderson and Kirkland, 1980). Faulting during the mid-Miocene provided

pathways for water from basinal aquifers to rise into the overlying Castile and Salado formations. Anderson and Kirkland (1980) proposed two prerequisites for brine density flow: 1) Pressurized or artesian source of relatively fresh water and 2) permeable fracture systems and salt mass normally isolated from groundwater. They cited the Delaware Basin as a good example of dissolution by brine density flow. Dijk and Berkowitz (2000) modeled buoyancy-driven dissolution of evaporate rocks using nuclear magnetic resonance imaging and also cited the Delaware Basin as a good example of such systems.

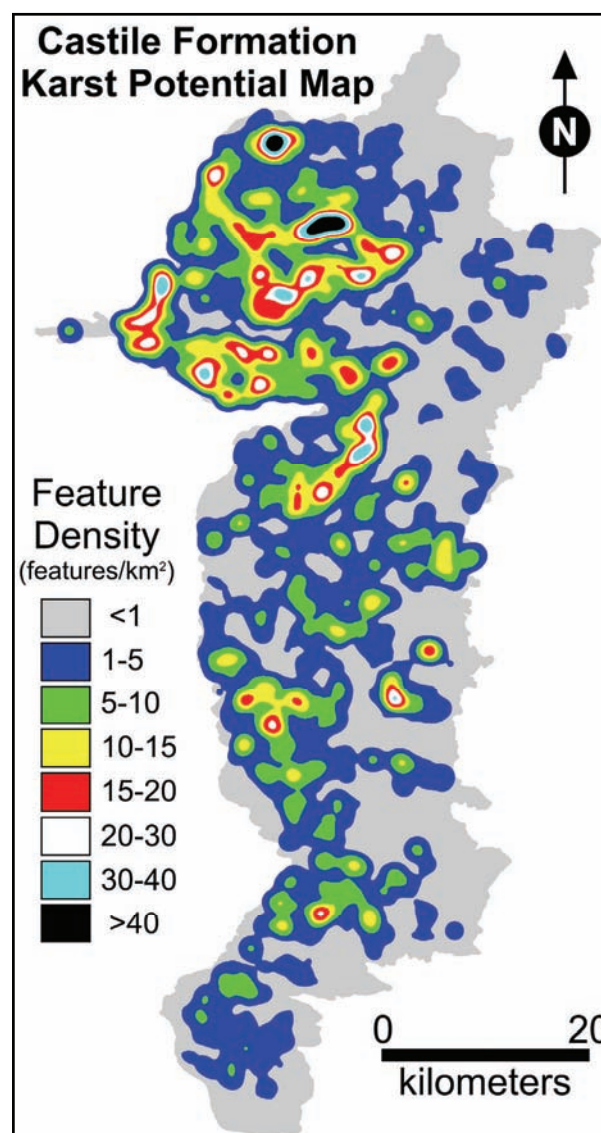


Figure 3. Karst potential map of the Castile Formation outcrop region defined in this study. Note the two dense areas of karst development within the northern portion of the study area with densities greater than 40 features/km² (from Stafford et al., 2008c).

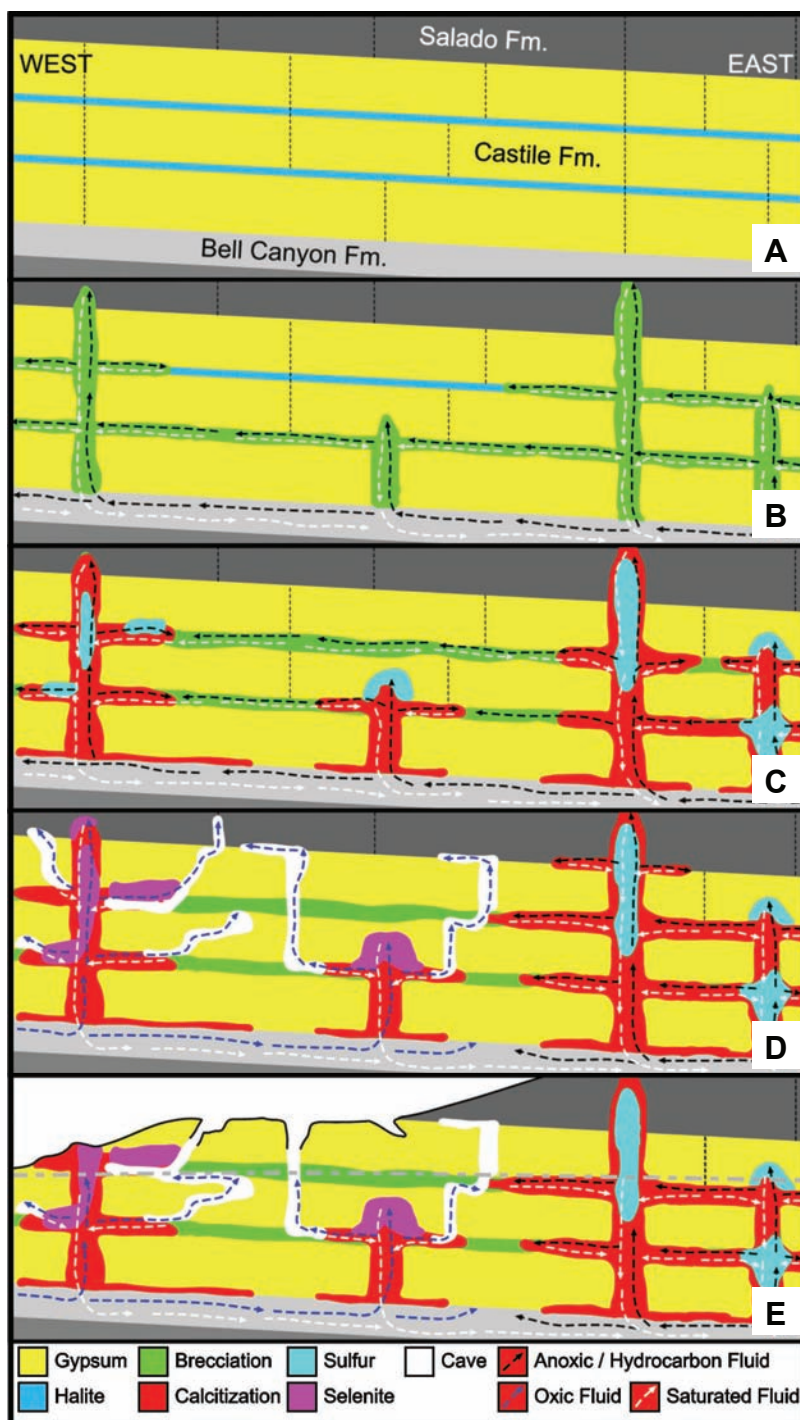


Figure 4. Hypogene karst development in the Delaware Basin.

A: Initial setting

B: Intrastratal, hypogene dissolution forming vertical breccia pipes and lateral blanket breccias (halite dissolution)

C: Evaporite calcitization resulting from bacterial sulfate reduction and deposition of native sulfur deposits.

D: Introduction of oxygenated waters resulting in additional hypogene porosity and oxidation of native sulfur to selenite.

E: Surficial breaching of hypogene features with subsequent epigene overprinting.

Undersaturated water would rise up the middle of the fracture to the top where dissolution would occur (Anderson and Kirkland, 1980). The resulting denser brine would then sink back to the aquifer along the edges (Figure 5). In the aquifer, the sulfate brine would sink to the bottom and continue to the east, where it was eventually discharged through the San Andres formation (Hiss, 1975). Maley and Huffington (1953) described accumulations of Cenozoic fill that overlie areas of mid-basin subsidence where the underlying halite beds are thinner. In these areas, the Salado halite beds extend further west than the underlying Castile halite beds. Maley et al., (1953) state that this indicates removal of the Castile beds from beneath by undersaturated waters.

Potential source aquifers

Several aquifers in the basin have been described as potential sources of the hypogene waters responsible for speleogenesis in the study area. The low permeability sandstones of the Bell Canyon Formation have been cited as the source of ascending waters (Smith, 1978; Klemmick, 1993; Guilinger, 1993). Hentz et al. (1989) described the Delaware Mountain Group as an artesian system. Crawford and Wallace (1993) described artesian flow from permeable sandstones within the Manzanita member of the Cherry Canyon formation. Lee and Williams (2000) described the third sand of the Bone Springs sandstone as the most likely source of water. Davis (1993) described an aquifer observed in the basal regions of the Castile. The aquifer was studied in wells drilled in the vicinity of Virginia Draw, an area significant in the deposition of sulfur in the Rustler Hills area. The aquifer formed along dissolution-replacement channels. Preliminary studies indicated high transmissivity. Potential discharge from the aquifer was described as being in a marshy area in Virginia Draw, near the Culberson Mine.

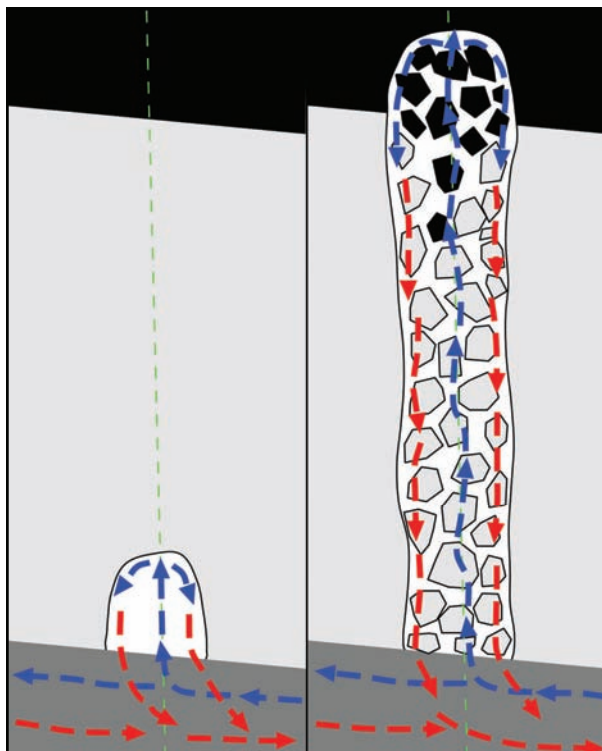


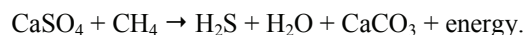
Figure 5. Model of breccia pipe formation by brine density flow. Flow of returning, saturated water down the outer edges of the void results in maximum dissolution occurring at the top: (left) initiation of the brine density dissolution process, (right) development of breccia pipe as a result of bedrock collapse or slump into the resulting void.

Bachman (1984; 1987) cited low permeability as a limiting factor in the ability of the Bell Canyon to provide adequate water to form the karst features observed. Using permeability data from Hiss (1975) and the method described by Lohman (1972), Anderson and Kirkland (1980) demonstrated that the upper 30 meters of the basin aquifer in the Bell Canyon formation has sufficient flow to remove the salt from 100 dissolution chambers of $1 \times 10^6 \text{ m}^3$ in a period of about 30,000 years. This suggests that the low permeability of the basin aquifer is not an overall limiting factor to the development of breccia pipes or dissolution breccias by brine density flow in the mid-basin region, as suggested by Bachman (1984; 1987).

Origin of calcitized masses

Calcitized evaporite masses are a dominant diagenetic feature in the western

Delaware Basin. Prior to their formation, the basin was structurally modified by Laramide uplift and basin-and-range expansion. At some point in the past 30 Ma, hydrocarbons began to migrate up dip to the west. Several authors have described possible sources for these hydrocarbons (Davis and Kirkland, 1970; Hill, 1996; Kirkland and Evans, 1976). Most cite the nearby hydrocarbon reserves of the Bell Canyon Formation. However, Lee and Williams (2000) have suggested the third sand aquifer of the Bone Springs Formation as a likely source. Hydrocarbons were trapped at fault boundaries where the lighter hydrocarbons, primarily methane, rose into karstic voids being created along fractures. Lee and Williams (2000) suggested that light carbon isotopes in calcitized masses indicate a hydrocarbon origin for the carbons in the masses. Isotope data from this study supports the hydrocarbon origin of carbon in the carbonate masses. Hentz et al. (1989) have described the reaction between the hydrocarbons and anhydrite in which sulfate is altered to carbonate and H_2S is released:



The resulting carbonate masses are observed in the subsurface and form the distinctive “castles” that give the formation its name (Figure 6).

On the surface, carbonate masses and solutional depressions have been observed to form linear features that are parallel to or lie along the faults (Hentz et al., 1989). Stafford et al. (2008b) demonstrated the linear nature of the carbonate masses. Most sources

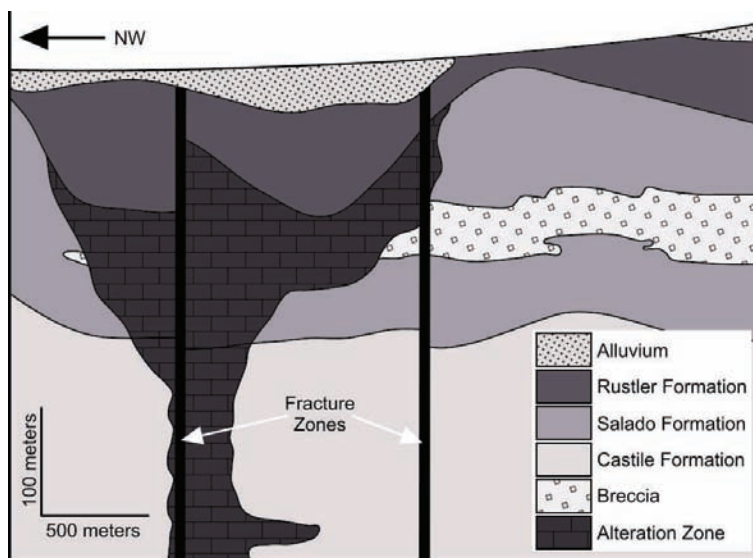


Figure 6. Cross section of sulfur deposit and calcitized mass at the Culberson mine site. Due to faulting at the site, basinal waters were able to rise through the formations, resulting in the formation of carbonate masses and sulfur deposits (from Stafford et al, 2008c).

cite isotope data indicating that these carbonates are biogenic (Hentz et al., 1989; Hill, 1996; Kirkland and Evans, 1976.). Worden and Smiley (1986) indicated that bio-sulfate reduction (BSR) and thermal-sulfate reduction (TSR) fractionate carbon and sulfur isotopes to similar degrees. Isotopic data does not provide definitive support for BSR over TSR based on. The presence of local igneous intrusives and higher than normal thermal flux are sufficient to support thermogenesis of the carbonate deposits (Calzia and Hiss, 1978).

Anhydrite diagenesis

The diagenesis of anhydrite to carbonate resulted in a volume decrease of 20% (Kirkland and Evans, 1976). Hydrocarbons were able to migrate through the resulting permeability to a dissolution face where dissolved anhydrite was being immediately replaced by calcite (Kirkland and Evans, 1976). In the resulting diagenetic boundary layer, sulfates dissolved and were replaced by carbonates as the result of either biogenic or thermogenic processes. The authors have observed areas where this boundary zone has been preserved in Dead Bunny Cave. Varves were observed continuing from gypsum, through the diagenetic zone, and into the carbonate section (Figure 7). The hydrogen sulfide produced oxidized in place, forming elemental sulfur (Smith, 1978; Hill, 1996; Hentz et al., 1989) or migrated back into the fracture where it continued to rise until it was trapped or vented to the surface (Hentz et al., 1989). In some areas, hydration of anhydrite to gypsum sealed fractures, creating a trap for rising gases such as methane or hydrogen sulfide (Smith, 1978; Guilinger, 1993)



Figure 7. Diagenetic Boundary between gypsum and calcite. Darker, lower half is calcitized material. Note the boundary layer and continuation of varves from gypsum into calcite.

Breccia development

Extensive solution breccias, collapse breccias, and breccia pipes in the basin were also the result of hypogene processes. Several halite beds were described in the Castile Formation (Anderson et al., 1972, 1978). The primary lithology of the Salado Formation is halite, as well. Rising undersaturated waters, on reaching a halite bed, would dissolve laterally into the halite beds (Figure 8). Anderson et al. (1978) described the Big Sinks depression, east of the study site, in which the halite beds of the Castile show significantly more thinning than the overlying Salado formation, indicating dissolution by rising undersaturated waters. Anderson et al. (1978) have correlated breccia beds observed in the west with halite beds found in the east. As the breccia beds were forming outward from the fractures, water was moving into the Castile from the Capitan formation or from exposures of the Bell Canyon to the west and dissolving halite beds. Erosion exposed beds to the west and allowed meteoric waters to descend downdip, along breccias, to halite beds. As halite beds dissolved, anhydrite beds within the halite or adjacent to it formed dissolution breccia beds (Anderson et al., 1978). In these beds, most of the breccia still lies in beds parallel to the plane of dissolution. Overlying these beds are collapse breccias. These breccias are typically clasts of 2+ cm. Most exhibit evidence of rotation and minor movement from the original location. In areas not exposed to diagenetic processes, these breccias are almost entirely anhydrite. However, breccias composed entirely of calcitized evaporites and breccias of calcitized material with anhydrite breccias nearby were observed by the authors. Anderson et al. (1978) described a limit to eastward dissolution of halite that occurred due to fresh water input from the Capitan Formation or exposed beds.

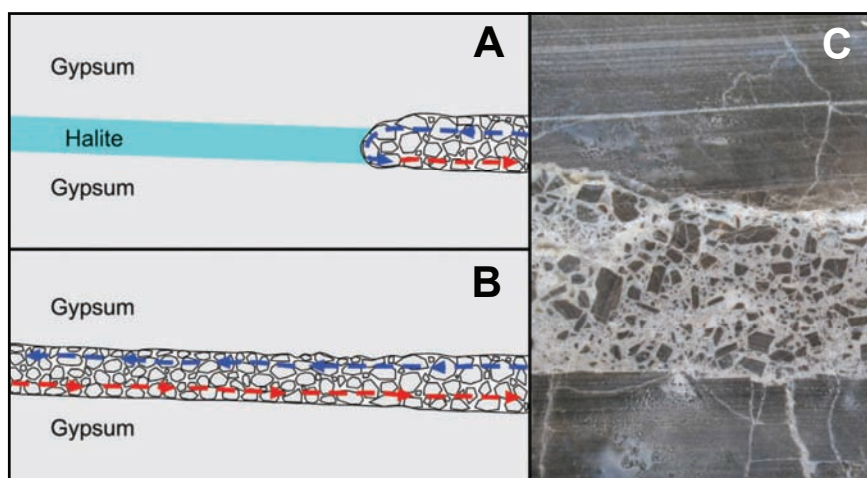
Observations by the authors indicate that blanket breccias have originated at breccia pipes as well as along the western dissolution front. Blanket breccias are the result of dissolution of nearby, more soluble beds (intrastratal halite). However, they give no indication of the origin of the dissolving fluids. These fluids could be the result of downdip flow from the western dissolution front or basinal fluids rising as a result of brine density flow (Anderson et al., 1980). Dissolution and collapse breccias formed from calcitized evaporites are formed by dissolution by rising basinal fluids, rather than meteoric waters along the western halite dissolution front.

Sulfur deposits

Sulfur was originally described precipitating in the water of Delaware Creek by the Pope expedition

Figure 8. Model of blanket breccia dissolution by brine density flow. Descending water input from the western dissolution front will result in anhydrite breccia. Ascending water input from a breccia pipe or fracture will result in calcitized or mixed anhydrite and calcite breccias.

- A. Initiation of breccia processes by convective flow.
- B. Well developed breccia with continued convective flow.



while surveying for a railroad route to the Pacific Ocean (Smith, 1896). Porch (1917) documented early exploration of sulfur deposits in Culberson County, Texas. He described test pits that vented hydrogen sulfide gas and others that vented methane gas. He also described one pit in which methane flow would reverse with changes in atmospheric pressure. He documented oil standing at a 10.7 m depth in a well in the Rustler Hills near the New Mexico/Texas border. Smith (1978) described important factors in the origin of elemental sulfur deposits: 1) presence of sulfate rocks, 2) presence of hydrocarbons, 3) fracturing or brecciation of rocks to facilitate fluid flow, 4) oxygenated groundwater or other agents capable of oxidizing hydrogen sulfide, 5) sealing to prevent the loss of hydrogen sulfide gas, 6) absence of halite. In describing the origin of Ukrainian sulfur deposits, Klimchouk (1997) also noted that factors 1-4 are prerequisites for karst development and that karstification must occur for sulfur deposits to form. According to Hentz and Henry (1989), all known ore bodies in the Rustler Springs sulfur district are associated with zones of fractured evaporate strata. Alignment of the Pokorny deposit with faults in the Rustler Springs deposit suggests a genetic relationship. All native sulfur observed by the authors was associated with calcitized evaporites or large masses of selenite in fractured evaporite zones.

Most carbonate masses and sulfur ore bodies are surrounded by selenite deposits. At Yeso Hills quarry, one crystal measures approximately 2 m x 7 m. The authors have observed numerous selenite crystals in excess of two meters in length. In Crystal Cave, the lowest extent of the surveyed cave is in a relatively homogenous mass of selenite in excess of 20m in height. Mohammed (1988) suggested that sulfur was oxidized with the help of sulfur oxidizing bacteria to form sulfuric acid. This then reacted with the limestone to produce carbon dioxide and gypsum. In a

confined, limited flow system, calcium sulfate would recrystallize from this saturated solution in the form of selenite. Hill (1996) has suggested that the selenite formed concurrently with the deposition of sulfur. The authors have observed native sulfur in association with selenite at several locations (Figure 9).

Delaware Basin hypogene model

In the subsurface, dissolution of halite beds occurred simultaneously from the west and from locations in the basin where halite beds had been breached by breccia pipes as a result of brine density flow. Hydrocarbons rising through the breccia pipes were converted, either biogenically or thermogenically, to calcitized evaporite masses resulting in the production of hydrogen sulfide gas. The hydrogen sulfide was trapped vertically as a result of volume expansion as anhydrite rehydrated to gypsum (Guilinger, 1993). Laterally, hydrogen sulfide migration was limited to the extent of blanket brecciation originating at the breccia pipe. Breaching of the seals, either through erosional removal of gypsum seals in fractures, or breaching of blanket breccias originating at breccia



Figure 9. Elemental sulfur crystal in a selenite mass found beside a calcitized 'dike' in southern Eddy County, New Mexico (note: small unit on scale = 1cm.)

pipes by the western dissolution front, resulted in the movement of oxygenated meteoric water into the hydrogen sulfide rich zones. Mixing with oxygenated water resulted in the deposition of elemental sulfur. Crawford and Wallace (1993) suggested that deposition of sulfur occurred in at least two cycles. During one cycle, sulfur and carbonate were formed concurrently with microscopic sulfur crystals forming within the carbonate mass. Hill (1996) stated that native sulfur deposits in the Delaware basin were found with bioepigenetic limestone because the two are genetically related. Crawford and Wallace (1993) noted the presence of orthorhombic voids in calcitized masses where elemental sulfur had been removed, supporting the idea of syngenetic origin of calcite and some sulfur. In a second cycle, sulfur was deposited in vugs and openings in the bedrock, the result of inflooding by oxygenated meteoric waters. At the Grant Prospect, Smith (1978) described a 30 meter deep shaft lined with a four inch layer of crystalline sulfur. The authors have observed minor deposits of sulfur in calcite lined vugs at this site. Kirkland and Evans (1976) have suggested that the shaft formed as a result of oxidation of native sulfur to sulfuric acid and subsequent dissolution of the surrounding carbonate to form the void.

The reaction converting anhydrite to calcite is exothermic. The hydrogen sulfide produced is converted

to elemental sulfur by oxidation. Further oxidation, possibly supported by sulfur oxidizing bacteria, would produce sulfate. Release into oxygenated water or water containing oxidizing agents would convert hydrogen sulfide directly to sulfate. This water would already be rich in sulfate, resulting in saturated sulfate brine. Sulfuric acid produced by the oxidation of sulfur would react with calcite and release calcium ions into the brine. In a low flow environment, this mix of calcium sulfate saturated water would begin to precipitate selenite crystals.

Calcitized masses in the study area are currently venting sulfide gases. "Acid earth" has been described on these hills (Evans, 1946). At the Grant deposit, also known as Sulfur Mine Hill, sulfur was mined from a series of large vugs and the resulting shaft extended down to more than 30 meters in depth. Gas venting from the site has been described as hydrogen sulfide. One of the authors entered the shaft with a hydrogen sulfide meter set at a sensitivity of 1 ppm but did not detect hydrogen sulfide. On his first visit to the mine, biochemist Hilliard (2008, pers. comm.) stated that the odor resembled mercaptan, a compound similar in structure to alcohol, but in which sulfur has replaced oxygen. Jaco, a former chemist at the Pennzoil Sulfur Mine (2009, personal communication) described a spring approximately 3 km north of the Culberson mine at which he had measured hydrogen sulfide in

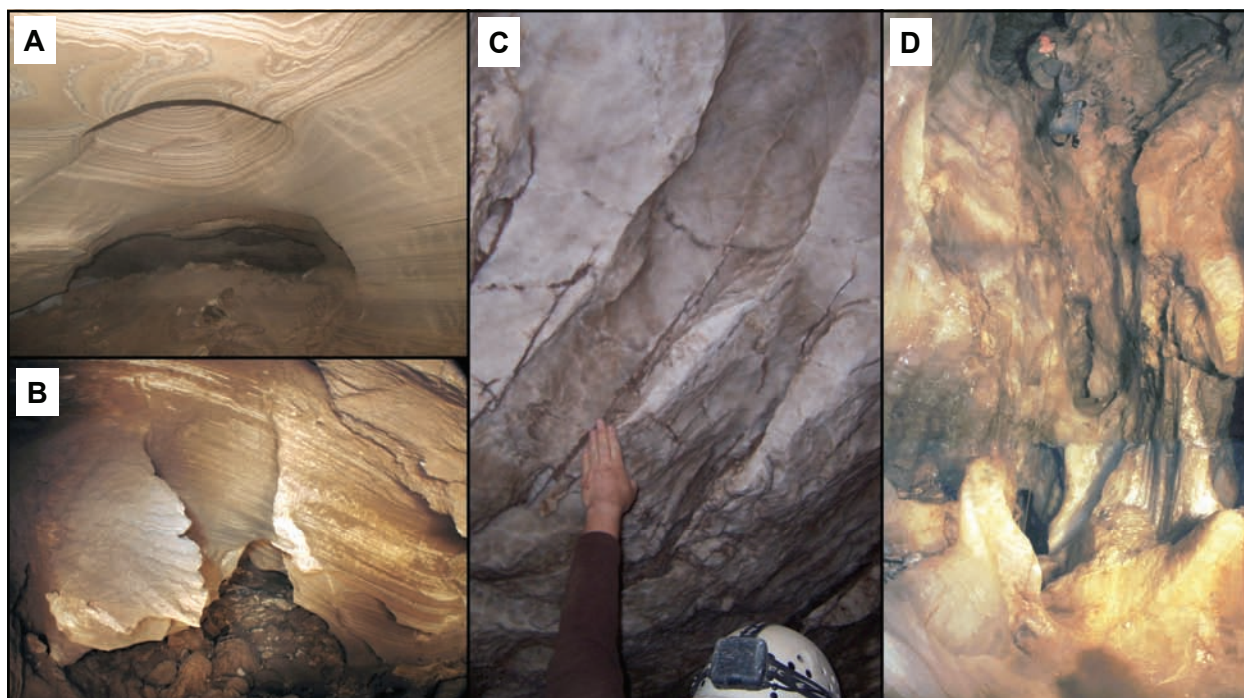


Figure 10. Hypogene features from study area: A) Cupola in Dead Bunny Hole; B) Riser in Dead Bunny Hole; C) Half-tube wall channel in Black Dog Cave; and D) Risers and wall channel in Glacier Bay II, Crystal Cave.

the spring water. This would be near the area described by Davis (1993) as a possible discharge zone for the basal Castile aquifer.

Most springs in the study area are what would be expected in a gypsum plain. The water has high levels of sulfate and mud in pools is hydrogen sulfide rich as a result of anaerobic sulfate reducing bacteria. However, a number of springs exhibit characteristics that would not normally be expected. Bell (2008) described a spring in the northern end of the study area in which unusual bacterial activity has been observed. Bell and Hilliard (2009, personal communication) describe this spring, and another in Culberson County, Texas, as unique due to their microbial diversity. Current studies are focusing on sulfide gases rising with the water. One of the authors measured a pH of 6.5 for groundwater at the spring in Culberson County. These unusual springs indicate the possibility of artesian flow from the lower Castile or deeper. Hentz et al. (1989) has suggested that sulfur mineralization might be an active process at depth in the Delaware Basin. As such, hypogene speleogenesis is likely an ongoing process as well.

Recent hypogene processes

Several caves in the study area exhibit a diagnostic suite of hypogene features. These features, described by Klimchouk (2007), include wall channels, ceiling channels, risers, and cupolas (Figure 10). In most of these caves, epigenic overprinting of these features has resulted in complex, mixed morphologies.

Dead Bunny Cave (Figure 11) lies in the basal Castile Formation and has been formed in varved gypsum and calcitized evaporites. The boundary zone between the two lithologies is found throughout the cave. Dead Bunny Cave has developed along three horizontal horizons. The upper portion of the cave is formed along the fold axis of several anticlines. Near the midsection of the cave, a boneyard area, typical of deep phreatic flow, was also documented. A clay 'dike' extends through the lower portion and indicates an earlier period of speleogenesis in which the passages were later filled with clays. This is consistent with other clay filled paleokarst observed in the study area. The system was breached in the up-dip area of the anticlines and epigenic overprinting continues to modify the pre-existing features, resulting in a complex speleogenetic history.

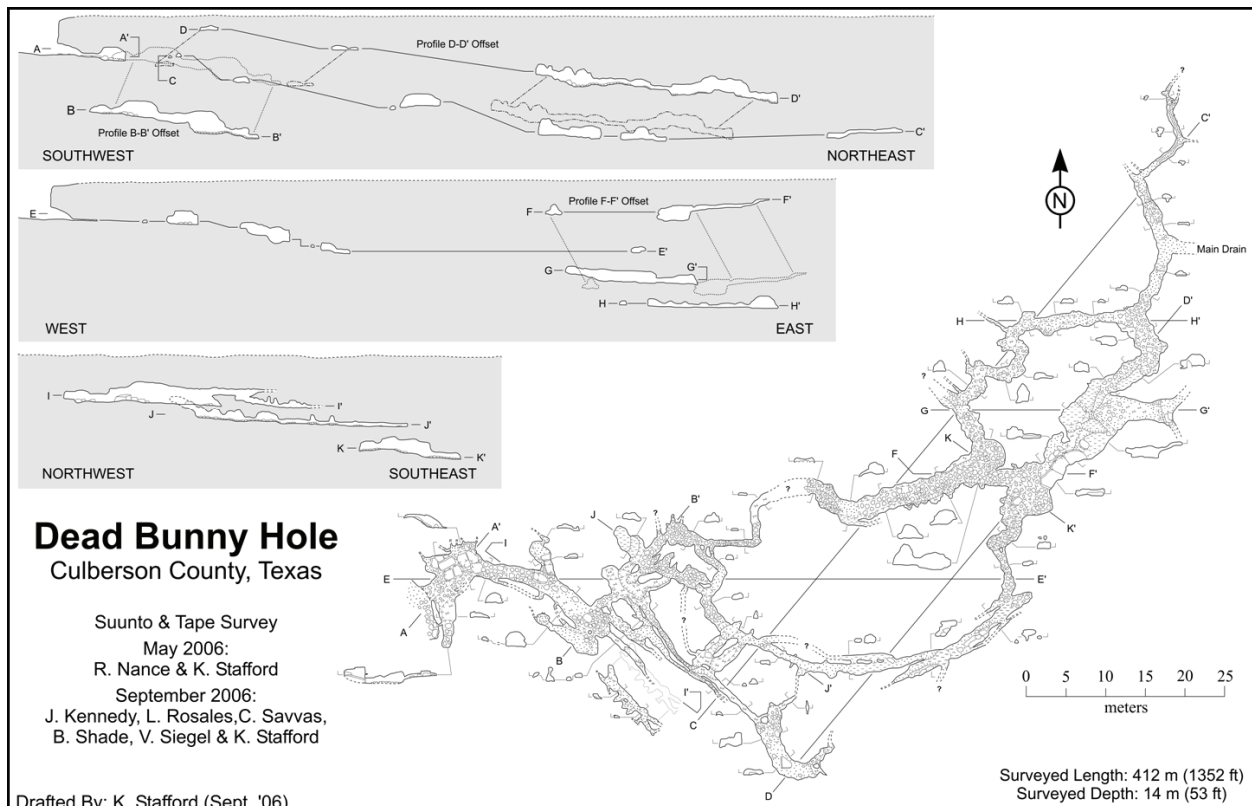
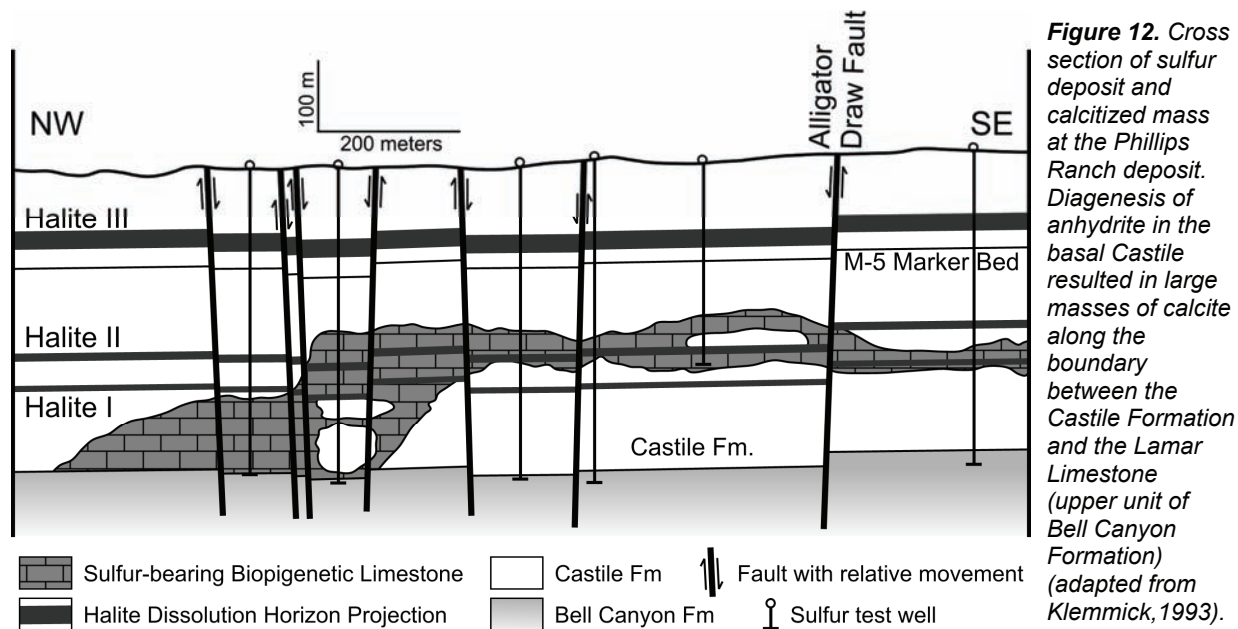


Figure 11. Dead Bunny Hole. Cave was formed by deep phreatic flow in the basal Castile Formation. Epigenic overprinting is currently an active process due to surface breaching in the vicinity of NW-SE trending anticlines.



Davis (1993) described a basal Castile aquifer located in the area of the Culberson sulfur mine and suggested a marshy area in Virginia Draw as a discharge. He described the aquifer as having good transmissivity and suggested that it could extend further under the Castile. The contact between the Lamar Limestone and the Castile Formation was observed approximately .5 km west of Dead Bunny cave. The cave lies very close to the base of the Castile Formation and the calcification observed is very similar to that described at the base of the Castile in the Phillips Ranch sulfur deposit (Figure 12). Dead Bunny cave exhibits features that would be expected in an exhumed aquifer of the type described at the Culberson Sulfur mine, and is assumed to have initially formed under deep phreatic conditions as part of a hypogene system.

Crystal cave is the deepest gypsum cave in Texas. It has been surveyed to a depth of 90 meters and is formed along extensive selenite deposits. Crystal Cave is a single riser, hypogene cave that can be described in three sections (Figure 13). Ford (2006) commented on the existence of simple, single conduit caves formed by rising water, noting that they were rare. The upper section is a vadose entrenched passage that eventually reaches a depth of approximately 10 meters below entrance level. In the area of Quarryman Pit, vadose entrenchment cut into the midsection passage. Quarryman pit is formed along a zone of minor faulting that predates deposition of selenite beds observed in the mid section of the cave (Figure 14). The mid section of the cave is an epigenically modified passage that follows regional dip to the east/northeast at approximately 6°. It follows a vein of

selenite that also lies along the regional dip. The selenite is underlain by nodular gypsum in several locations and lies within breccia through much of its length, indicating formation along a dissolved halite bed.

The lowest section of the cave is formed in homogeneous selenite. The passage drops approximately 10 meters through a vertical pit (Glacier Bay II) and then through a series of smaller drops to a terminal sump at a total depth of 92.5 meters. Risers and cupolas were observed above the terminal sump, leading upward into the pit. The terminus of the mid-section of the cave lies directly above these riser features. At the lip of Glacier Bay II, slotting, or downcutting, is minimal. In most caves, this rapid transition from horizontal passage to vertical pit is attributed to heterogeneous lithology, with a less soluble layer, such as chert, forming the lip of a pit. Without such a resistant layer, epigene water would entrench the edge of the pit, forming a slope in the passage, rather than a vertical pit. Glacier Bay II is formed in homogeneous selenite and contains no resistant layer. The vertical nature of the pit at Glacier Bay II and the minimal pit-edge slotting indicates a rising water origin with slight epigenic overprinting.

Stafford (2008) suggested episodic development of carbonates and sulfur deposits. Crystal Cave presents features supportive of the idea of episodic development. Early hypogene processes produced the riser void which later filled with selenite and became Glacier Bay II. A halite bed was dissolved and subsequent formation of breccias occurred. Hydration of anhydrite on either side of the breccias resulted in a

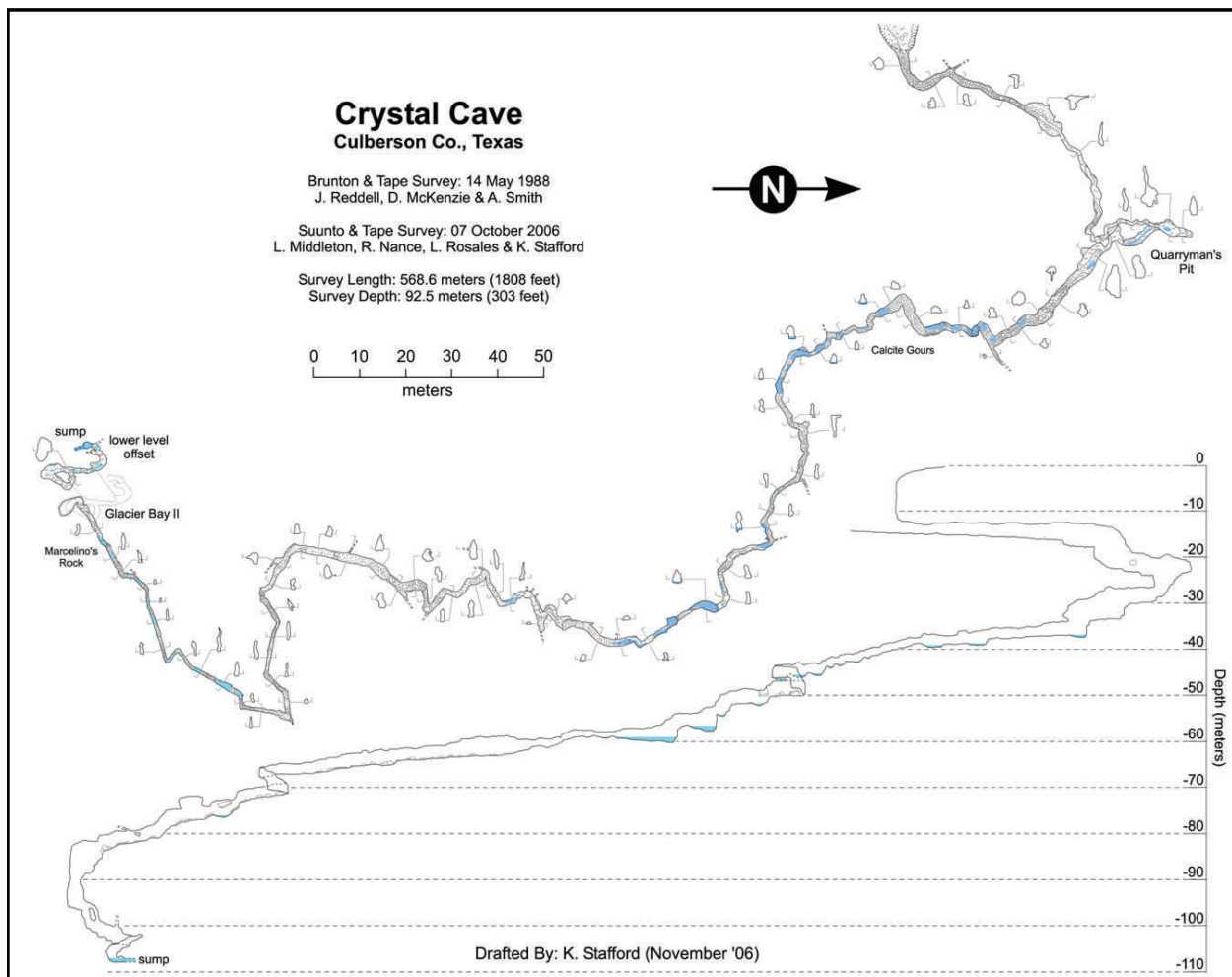


Figure 13. Crystal Cave. Upper section of the cave extends from the surface level to a depth of 25 meters. The mid-section of the cave extends from 25 meters to a depth of 80 meters below the surface. The lower section of the cave lies below 80 meters depth.



Figure 14. Fractures in Quarryman Pit, Crystal Cave, at a depth of 25 meters below the surface. Note the faulting and the selenite beneath it (image ~3m wide).



Figure 15. Breccia and selenite replacement of boundary anhydrite during rehydration. Note nodular anhydrite in lower left corner of photo (image ~2 m wide).

selenitic rather than alabastrine texture in the sulfate bedrock (Figure 15). Selenite began to precipitate from void-filling sulfate brines. A period of minor, possibly localized, tectonism produced minor faulting in the area of Quarryman's Pit. These fractures provided a discharge point for the artesian waters still filling minor porosity in the area of Crystal Cave. This breach allowed rising water to begin forming the passages observed in the lower and mid-sections of Crystal Cave. Lowering of the potentiometric surface eventually lowered the water level in the cave. Epigenic processes enlarged fractures in the bedrock until they breached the existing voids found in the area of Quarryman's Pit. At that point, epigenic processes dominated, resulting in a mixture of speleogenic features.

Summary

In the Delaware Basin of southeastern New Mexico and west Texas, the study of sulfur deposits has yielded information on a suite of features indicative of hypogene processes. The conversion of anhydrite to carbonate and the resulting masses, associated breccias, large selenite deposits and associated deposits of elemental sulfur are all found together in the study area. These features resulted from aquifer fluids deeper in the Delaware Basin moving upward through fractures in the overlying evaporate beds. Hydrocarbons in the rising water reacted with anhydrite resulting in conversion to carbonate. This process was likely biogenic, but may have been thermogenic. The hydrogen sulfide produced as a result was later oxidized to produce deposits of elemental sulfur and selenite.

Caves in the study area exhibit characteristics of initial formation by hypogene or deep phreatic conditions. These hypogene caves appear to have formed along zones of permeability resulting from previous hypogene episodes. Overprinting by epigene processes has resulted in complex morphologies, making speleogenetic studies difficult. Numerous sulfide vents and sulfur springs in the area indicate ongoing processes at depth.

Speleogenesis in the study area needs to be evaluated from its position within the overall evolution of the Delaware Basin. Processes at depths in excess of 300 meters, some of which may have been active as far as long as 30 Ma, have been critical in the development of karst features seen in the basin today. Karst processes in the basin have played a crucial role in the economy of the region by forming aquifers, hydrocarbon reservoirs and economic sulfur deposits.

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References

- Anderson, R.Y., W.E. Dean, D.W. Kirkland, and H.I. Snider. 1972. Permian Castile varved evaporite sequence, West Texas and New Mexico. *Geological Society of America Bulletin* 83: 59-85.
- Anderson, R.Y., K.K. Kietzke, and D.J. Rhodes. 1978. Development of dissolution breccias, northern Delaware Basin, New Mexico and Texas. In *Geology and Mineral Deposits of Ochoan Rocks in Delaware Basin and Adjacent Areas*, ed. G.S. Austin, 47-52. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Anderson, R.Y., and D.W. Kirkland. 1980. Dissolution of salt deposits by brine density flow. *Geology* 8: 66-69.
- Bachman, G.O. 1984. *Regional geology of Ochoan Evaporites, northern part of Delaware Basin*. New Mexico Bureau of Mines and Mineral Resources Circular 184. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Bachman, G.O. 1987. *Karst in evaporites in southeastern New Mexico*. Sandia National Laboratories Contractor Report SAND86-7078. Albuquerque: Sandia National Laboratories.
- Barker, C.E., and M.J. Pawlewicz. 1993. Post-tectonic reheating of portions of the Permian Basin as expressed by iso-reflectance lines on regional structural sections. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin and J.M. Barker, 29-30. Socorro: New Mexico Geological Society.
- Bell, T. 2008. An analysis of microbial life and the effects on water chemistry and microenvironment in a karst spring, southern Eddy County, New Mexico. *Geological Society of America Abstracts with Programs* 40: abstract 158-15.
- Calzia, J.P., and W.L. Hiss. 1978. Igneous rocks in northern Delaware Basin, New Mexico and Texas. In *Geology and mineral deposits of Ochoan Rocks in Delaware Basin and adjacent*

- areas, ed. G.S. Austin. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Crawford, J.E., and C.S. Wallace. 1993. Geology and mineralization of the Culberson Sulfur Deposit. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin, and J. M. Barker, 301-316. Socorro: New Mexico Geological Society.
- Crawford, J.E. 1993. K Hill and Yeso Hills Selenite occurrence. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin and J. M. Barker, 8-10. Socorro: New Mexico Geological Society.
- Davis, J.B. and D.W. Kirkland. 1970. Native sulfur deposition in the Castile Formation, Culberson County, Texas. *Economic Geology* 65:107-121.
- Davis, T.D. 1993. Progress report on ground-water movement within the sulfur-bearing zones of the lower Castile Formation at Phillips Ranch deposit, Culberson County, Texas. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin and J.M. Barker, 23-26. Socorro: New Mexico Geological Society.
- Dijk, P.E., and B. Berkowitz. 2000. Buoyancy-driven dissolution enhancement in rock fractures. *Geology* 28: 1051-1054.
- Evans, G.L. 1946. *The Rustler Springs sulphur deposits as a source of fertilizer*. Bureau of Economic Geology Report of Investigations No. 1. Austin: Bureau of Economic Geology.
- Ford, D.C. 2006. Karst geomorphology, caves, and cave deposits: A review of North American contributions during the past half century. In *Perspectives on karst geomorphology, hydrology and geochemistry: A tribute volume to Derek C. Ford and William B. White: Geological Society of America Special Paper 404*, ed. R.S. Harmon and C.M. Wicks, 1-14. Boulder: Geological Society of America.
- Ford, Derek C., and Paul Williams. 2007. *Karst hydrogeology and geomorphology*. Hoboken: John Wiley & Sons, Ltd.
- Frumkin, A., and I. Fischhendler. 2005. Morphometry and distribution of isolated caves as a guide for phreatic and confined paleohydrological conditions. *Geomorphology* 67: 457-471.
- Guilinger, J.R. 1993. Geology of the Pokorný sulfur deposit, Culberson County, Texas. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin and J.M. Barker, 21-23. Socorro: New Mexico Geological Society.
- Hentz, T.F., and C.D. Henry. 1989. Evaporite-hosted native sulfur in Trans-Pecos Texas: relation to late-phase Basin and Range deformation. *Geology* 17: 400-403.
- Hentz, T.F., J.G. Price, and G.N. Gutierrez. 1989. *Geologic occurrence and regional assessment of evaporite-Hosted native sulfur, Trans-Pecos Texas*. Bureau of Economic Geology Report of Investigations No. 184. Austin: Bureau of Economic Geology.
- Hill, C.A. 1996. *Geology of the Delaware Basin – Guadalupe, Apache, and Glass Mountains, New Mexico and west Texas*. Midland: Society of Economic Paleontologists and Mineralogists, Permian Basin Section.
- Hiss, W.L. 1975. Stratigraphy and groundwater hydrology of the Capitan Aquifer, southeastern New Mexico and Western Texas. PhD diss., University of Colorado.
- Horak, R.L. 1985. Trans-Pecos tectonism and its effect on the Permian Basin. In *Structure and tectonics of trans-Pecos Texas*, ed. P.W. Dickerson and W. R. Muelberger, 81-87. Midland: West Texas Geological Society.
- Kelley, V.C. 1971. *Geology of the Pecos Country, Southeastern New Mexico*. New Mexico Bureau of Mines and Mineral Resources Memoir 24. Socorro: New Mexico Bureau of Mines and mineral Resources.
- Kirkland, D.W., and R. Evans. 1976. Origin of limestone buttes, gypsum plain, Culberson County, Texas. *American Association of Petroleum Geologists Bulletin* 60: 2005-2018.
- Klemmick, G.F. 1993. Geology of the Pokorný sulfur deposit, Culberson County, Texas. In *Carlsbad Region, New Mexico and West Texas: New Mexico Geological Society 44th Annual Field Conference, Guidebook*, ed. D.W. Love, J.W. Hawley, B.S. Kues, J.W. Adams, G.S. Austin, and J.M. Barker, 18-19. Socorro: New Mexico Geological Society.
- Klimchouk, A. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Lee, M.K., and D.D. Williams. 2000. Paleohydrology of the Delaware Basin, Western Texas: Overpressure development, hydrocarbon migration, and

- ore genesis. *American Association of Petroleum Geologists Bulletin* 84: 961-974.
- Lohman, S.W. 1972. *Ground-water hydraulics*. U.S. Geological Survey Professional Paper 708. Boulder: U.S. Geological Survey.
- Maley, V.C., and R.M. Huffington. 1953. Cenozoic fill and evaporite solution in the Delaware Basin, Texas and New Mexico. *Geological Society of America Bulletin* 64: 539-546.
- Mohammed, K. 1988. The original phase of Castile calcium sulphates: A study of the classic-outcrops, southeastern New Mexico and west Texas. MS Thesis, Sul Ross State University.
- Olive, W. W. 1957. Solution-subsidence troughs, Castile Formation of gypsum plain, Texas and New Mexico. *Geological Society of America Bulletin* 68: 351-358.
- Porch, E.L. 1917. *The Rustler Springs sulphur deposits*. University of Texas Bulletin No. 1722. Austin: University of Texas.
- Sares, S.W. 1984. Hydrologic and geomorphic development of a low relief evaporite karst drainage basin, Southeastern New Mexico. MS Thesis, University of New Mexico.
- Smith, A.R. 1978. Sulfur deposits in Ochoan rocks of southeast New Mexico and Texas. In *Geology and mineral deposits of Ochoan Rocks in Delaware Basin and adjacent areas*, ed. G.S. Austin, 71-77. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Smith, E.A. 1896. Notes on native sulphur in Texas. *Science* 3: 657-659.
- Stafford, K.W. 2008. Cavernous porosity and associated sulfate diagenesis in the Castile Formation: Eddy County, New Mexico and Culberson County, Texas. *West Texas Geological Society Bulletin* 47: 14-26.
- Stafford, K.W., R.G. Nance, L. Rosales-Lagarde, and P.J. Boston. 2008a. Epigene and hypogene gypsum karst manifestations of the Castile Formation: Eddy County, New Mexico and Culberson County, Texas, USA. *International Journal of Speleology* 37: 83-98.
- Stafford, K.W., L. Rosales-Lagarde, and P.J. Boston. 2008b. Castile evaporite karst potential map of the gypsum plain, Eddy County, New Mexico and Culberson County, Texas. *Journal of Cave and Karst Studies* 70: 35-46.
- Stafford, K.W., D. Ulmer-Scholle, and L. Rosales-Lagarde. 2008c. Hypogene calcitization: Evaporite diagenesis in the western Delaware Basin. *Carbonates and Evaporites* 23: 89-103.
- Worden, R.H., and P.C. Smalley, 1996. H₂S-producing reactions in deep carbonate gas reservoirs: Khuff Formation, Abu Dhabi. *Chemical Geology* 133: 157-171.
- Zimmerman, J.B., and E. Thomas. 1969. *Sulfur in west Texas*. Its geology and economics: Bureau of Economic Geology Geological Circular 69-2. Austin: Bureau of Economic Geology.

HYPOGENIC ORIGIN OF ROBBER BARON CAVE: IMPLICATIONS ON THE EVOLUTION AND MANAGEMENT OF THE EDWARDS AQUIFER, CENTRAL TEXAS, USA

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Abstract

Robber Baron Cave is formed within the Upper Cretaceous Austin Chalk, in Bexar County, Texas, USA. The cave exhibits features that demonstrate a hypogenic origin, including a 1.51-km-long network maze pattern, fissure-floored passages, passage ceilings laterally enlarged adjacent to a contact with an upper confining unit, and authigenic sediments. The Edwards Aquifer provides the only source of water that could create the conditions necessary to form the cave, and provides modern analogs through artesian flows from the nearby San Antonio and San Pedro Park Springs. Anecdotal reports from the early 20th century describe flowing streams and pools in sections of the cave no longer accessible.

The Edwards Aquifer enlarged westward by stream incision along the Balcones Fault Zone, exposing down-faulted permeable units to allow groundwater discharge from lower elevation locations. Stream incision rates indicate that the hypogenic conditions necessary to form Robber Baron Cave occurred 2.0 to 2.5 Ma, and thus set a minimum age for accretion of the Bexar County portion of the aquifer. The presence of this and other hypogenic caves and artesian springs in the Austin Chalk, above the upper confining unit of the Edwards Aquifer, demonstrates areas of significant localized upward flow into the Austin and the paleo land surface. Identification of these areas is important in establishing areas of more stringent land use regulations to prevent aquifer degradation through these highly permeable features situated outside of the recognized aquifer recharge zone.

Introduction

Robber Baron Cave is located within the city of San Antonio, in central Texas, USA. It is by far the most extensive and best known cave in Bexar County. The earliest reports of the cave date to the 1910s. From 1926 -1933, about 160 m of its passages were developed for tourists. This included the placement of electric lights and the partial filling and leveling of the floors to create easy walkways. The developed areas were passages in the northern third of the cave.

Currently the cave has 1.51 km of known passages within a square area approximately 100 m on each side (Figure 1). It is a predominantly horizontal network maze of linear passages intersecting at 30-90° angles. The entrance is a sinkhole measuring 11 m long by 9 m wide by 3-9 m deep and formed by collapse at the intersection of least three passages. The sinkhole had



Figure 1. Silhouette map of Robber Baron Cave and aerial photo lineaments, updated from Veni (1988, 1997a) with surveys conducted in 2008.

been half filled with dumped dirt and debris since at least the 1950s until cleaned out from 2003-2004, when a trench was cut into the north side of the sinkhole for easier access (Mitchell and Palit, 2009). Two of the three passages in the sinkhole are open to the surface; one is blocked by bars and the other, at the deepest point of the sinkhole, is gated to protect the cave from vandalism and unauthorized visitation.

The majority of passages in Robber Baron Cave are about 1 m wide by 1-3 m high. Many walls are covered with a layer of flowstone usually <2 cm thick. Several walls taper to “V”-shaped cross sections or narrow rifts 1-3 m deep, so the floor areas in these passages are relatively small. Clay is common on the floors. Gray clay is either washed or carried in material and brown clay is composed predominantly of the insoluble residual from the Austin Chalk that dissolved to form the passages (demonstrated in the following sections). The clay usually ranges from about 3 cm to a meter thick, although a systematic sounding of clay depths has not been conducted. Excavation of a pit into the floor of the Lighted Passage in 1977 demonstrated the sediment was at least 3 m deep. However, most of this was artificial fill. Sediment depth decreases with distance from the sinkhole, and the cave’s mean depth generally increases with that distance, probably a direct correlation to the thinner sediments. The northern third of the cave averages roughly 10 m below the surface, the west-central portion averages a depth of about 12 m, and passages in the southern third of the cave rise to a depth of 8-10 m. A sediment-filled pit along the west-central margin of the cave was excavated to reveal the “Lower Level,” which has at least 75 m of passages averaging 17 m below the surface.

Most of the cave’s reported extent is not accessible for exploration or study. During its commercial development, the owner at the time deliberately collapsed many passages. Road building, underground utility lines, and other construction associated with the area’s urbanization invariably added to the collapse. Prior to the collapses, Robber Baron was known to extend at least 100 m farther east to a water well, 600 m southwest to a now-sealed, extensive, but poorly explored maze cave known as Holmgreen’s Hole, and about 1.2 km to the southwest to underground streams, pools, and another well that pumped water from the cave. Two caves sealed by urbanization have also been reported within 500 m of Robber Baron, and could potentially have connected to

the cave. Since the late 1970s, cave explorers have excavated the collapsed passages in hopes of rediscovering the “lost” parts of the cave. Much of that effort and the cave’s history is recorded by Veni (1988) and updated by Veni (1989, 1997a).

Geologic setting

Robber Baron Cave is located at the southeastern corner of the karstic Edwards Plateau. Its entrance is at an elevation of 250 m above mean sea level at the northern edge of an effectively level upland area between Olmos and Salado creeks, situated 3 km to the southwest and east, respectively. Local surface drainage at the cave flows northeast to Salado Creek, while regional surface water flow is to the southeast.

The surface above Robber Baron Cave is about 4.5 m below the top of the Upper Cretaceous (Coniacian) Austin Chalk Group. Locally comprising the Atco, Vinson, and Dessau formations (Table 1), the Austin ranges in thickness in Bexar County from 40 to 73 m. Barnes (1983) described the Austin as “chalk and marl, chalk mostly microgranular calcite with minor foraminifera tests and *Inoceramus* prisms, averages about 85% calcium carbonate, ledge forming, grayish white; alternates with marl, bentonitic seams locally, recessive, medium gray, sparsely glauconitic, pyrite nodules in part weathered to limonite common, occasional beds with large-scale cross-stratification; locally highly fossiliferous.” Its stratigraphy in the Bexar County area has not been well described but has been generally discussed by Holt (1956), Arnow (1959), Pessagno (1969), Cloud (1975), Waddell (1977), Dravis (1979), Corbett (1982), Barnes (1983), Young (1985), and Corbett et al. (1991). Detailed descriptions of the Austin have not been published, and its formation boundaries are not locally mapped, but the cave’s gross

Table 1. Stratigraphic column of the Bexar County, Texas, area (wide lines represent unconformities).

Age	Group	Formation	Average Thickness (m)
Upper Cretaceous/ Coniacian	Taylor	Pecan Gap Chalk	30-122
Upper Cretaceous/ Coniacian	Austin		40-73
		Dessau	inadequately mapped
		Vinson	inadequately mapped
		Atco	inadequately mapped
Lower Cretaceous/ Cenomanian-Albian	Washita		32.5-37.5
		Buda Limestone	16.5
		Del Rio Clay	16.0
		Georgetown	0-5
Lower Cretaceous/ Albian	Edwards		147.1

stratigraphic position at the top of the group indicates it lies within the Dessau Formation.

Textural and mineralogical observations and analyses of the Austin Chalk

This study provides the first published microprobe imaging, chemical mapping, and quantitative analyses of the Austin Chalk from the Bexar County outcrop. Eight samples described below were collected from the cave. Two sediment samples from within the cave were also collected and are described later in this paper.

All samples were examined using a Cameca SX-100 microprobe located at the New Mexico Institute of Mining and Technology. Polished samples of all but the two sediment (clay) samples were prepared. Samples were examined using backscattered electron (BSE) imaging. Chemical maps showing the distribution of Ca, Si, and Fe were collected in representative areas of three rock samples (RBC 8 through RBC 10). Quantitative analyses were performed on all polished samples. An accelerating voltage of 15 kV and a 10 nA beam current were used. Analyses of carbonate or mixed matrix material were done using a broad (10 or 20 micron) beam. A point beam (~1 micron) was used for all other analyses. Standard reference materials, including

amphiboles and carbonates were run in each analytical session to assess and monitor calibration accuracy and reproducibility.

Because their tight matrix and consequent low permeability inhibited infiltration of the epoxy, relatively flat unpolished surfaces of the clay samples were mounted on carbon tape and qualitative scans performed to determine their composition; a few quantitative analyses were performed on RBC 4 to determine any distinction between observed color variations in the clay.

Two distinct lithologic horizons occur in the Austin Chalk at Robber Baron Cave. Approximately the upper 13 m of the Austin, extending about 8 m below the surface at the cave, are a pale yellow (Munsell 2.5Y 8/2), soft, massive, highly fractured chalk. All solutionally-formed passages occur below the chalk within a hard, fossiliferous limestone horizon. The chalk is often exposed in the cave at the top of collapse-formed domes that stoped up from the underlying limestone. The chalk horizon has been referred to as a marl in some previous reports, which is consistent with chemical maps and BSE images of sample RBC 10, which show microscopic fossils and carbonate fragments with occasional glauconite grains in a finer matrix of dominantly broken shell fragments and glauconitic clay (Figure 2).

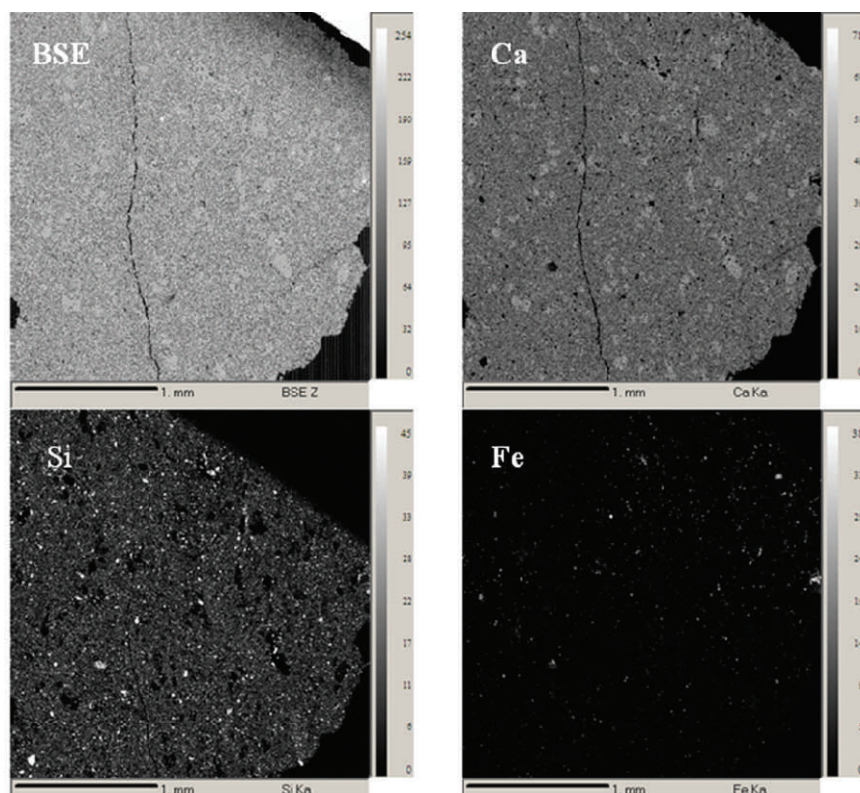


Figure 2. BSE image and Ca, Si, & Fe element maps of chalk horizon of Austin Chalk (sample RBC 10).

Goethite nodules (referred to in the literature by the generic term “limonite”) occur in both the chalk and limestone horizons. RBC 2 is a nodule from the chalk horizon with a yellow (Munsell 10YR 7/6) rind. BSE images and quantitative analyses show a dense core of dominantly goethite with minor siliceous inclusions, rimmed by a porous 500 micron thick outer rind (Figure 3-A). The rind is dominantly composed of fine calcium carbonate fragments, grains, and cement intermixed with goethite crystals and cement. Glauconite pellets and quartz grains are incorporated in the rind in some areas and some carbonate fossils are replaced by goethite (Figure 3-B). RBC 3 appears typical of weathered goethite nodules in the cave walls. They have a weathered reddish brown color (Munsell 7.5YR 6/6), and RBC 3 was found to be made up

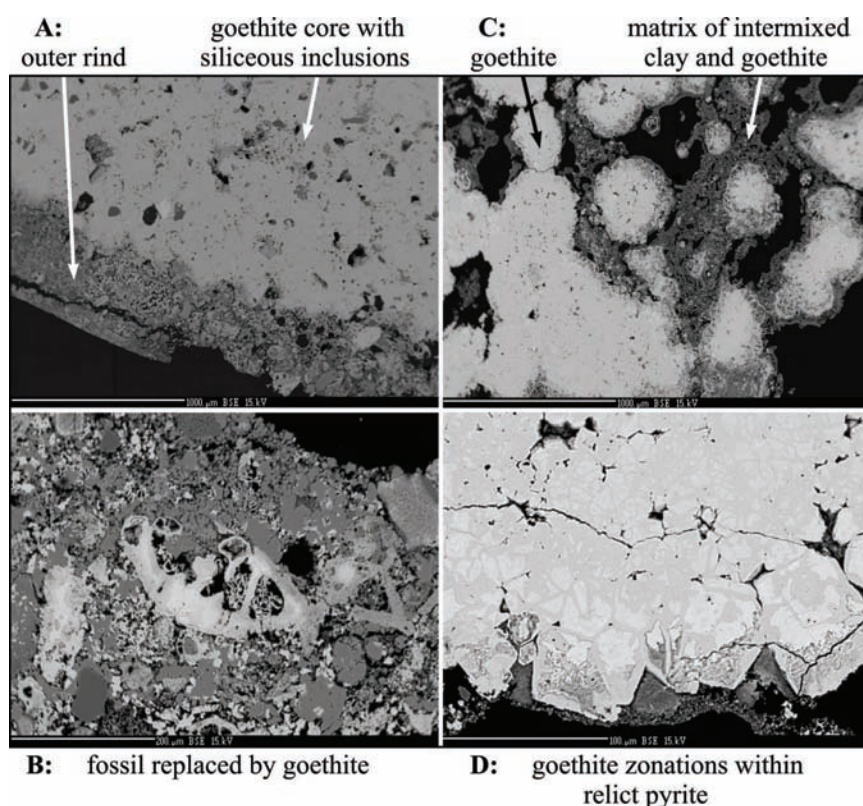


Figure 3. Sample RBC 2; A: Goethite nodule with distinct particle-rich rind. B: Close up of rind showing mix of carbonate (medium gray) and goethite (bright areas) grains and cement. Sample RBC 3; C: Aggregate clusters of zoned goethite crystals rimmed by a porous matrix of glauconitic clay and goethite. D: Close up of goethite crystals showing higher Z (bright) and lower Z (dark) zonations and highly altered rims of outer crystals.

of clusters of aggregate crystals (altered pyrite) with distinct zonations loosely rimmed by a porous matrix dominantly composed of glauconitic clay intermixed with fine iron-rich material (goethite) (Figure 3-C). Well-developed crystal faces along outer edges of clusters are significantly altered. Quantitative analyses within goethite zonations with a lower average atomic number or “Z” (appear darker in BSE images) have higher concentrations of Si and Al, and lower Fe content than higher Z zonations (Figure 3-D, Table 2).

The Austin Chalk’s limestone horizon at Robber Baron Cave is at least 9 m thick. Its upper 2 m are a highly fossiliferous zone, increasing in fossil content downward to become a dense accumulation of *Exogyra* (?) *laeviuscula* gastropods. It rests unconformably on a burrowed surface extending about 50 cm into the underlying limestone (Figure 4), which generally weathers in the cave to a predominantly yellow (Munsell

Table 2. Quantitative analyses of light (higher average atomic number or Z) and dark (lower Z) zonations in goethite crystals (% weight oxide).

Goethite Analyses	P ₂ O ₅	SiO ₂	SO ₂	TiO ₂	Al ₂ O ₃	MgO	CaO	MnO	FeO	Na ₂ O	K ₂ O	Total
RBC 3-01	0.26	4.80	0.05	0.00	1.04	0.20	0.31	0.01	68.04	0.01	0.01	74.72
RBC 3-03	0.26	4.81	0.05	0.06	1.10	0.25	0.36	0.00	67.34	0.02	0.00	74.24
RBC 3-05	0.27	4.95	0.01	0.02	0.98	0.21	0.46	0.01	67.71	0.02	0.02	74.64
RBC 3-18	0.24	4.91	0.04	0.01	1.14	0.24	0.31	0.03	66.59	0.01	0.00	73.50
RBC 3-19	0.45	6.11	0.04	0.13	2.57	0.21	0.60	0.01	65.08	0.01	0.02	75.22
RBC 3-26	0.33	5.98	0.03	0.00	1.53	0.26	0.41	0.00	66.28	0.03	0.01	74.86
RBC 3-27	0.33	6.64	0.04	0.03	1.45	0.22	0.48	0.01	67.53	0.00	0.00	76.73
RBC 3-34	0.18	5.12	0.02	0.00	1.09	0.23	0.31	0.02	67.46	0.00	0.00	74.42
Avg. low Z	0.29	5.41	0.03	0.03	1.36	0.23	0.41	0.01	67.00	0.01	0.01	74.79
RBC 3-02	0.26	3.11	0.06	0.04	0.53	0.18	0.34	0.00	72.15	0.00	0.01	76.69
RBC 3-04	0.34	3.20	0.06	0.03	0.71	0.18	0.45	0.04	73.59	0.00	0.03	78.61
RBC 3-17	0.39	3.43	0.04	0.02	0.69	0.13	0.51	0.00	72.85	0.00	0.00	78.06
RBC 3-25	0.36	3.52	0.02	0.03	0.73	0.08	0.45	0.00	74.06	0.02	0.01	79.28
RBC 3-35	0.27	3.25	0.07	0.02	0.55	0.14	0.39	0.00	71.98	0.00	0.01	76.66
Avg. hi Z	0.32	3.30	0.05	0.03	0.64	0.14	0.43	0.01	72.93	0.00	0.01	77.86

10YR 7/6) surface. While the burrowed interval has no apparent effect on cave development, most passages in the cave occur in its host and underlying limestone. This limestone is grainy and contains some fossils; abundant glauconite and quartz grains and occasional goethite grains appear in the BSE images of sample RBC 8 (Figure 5). The limestone in the cave's lower levels has not been sampled or carefully examined; frequent high percentages of atmospheric carbon dioxide preclude regular access to that part of the cave.

The limestone's fossiliferous zone has abundant glauconitic grains and pellets, as illustrated in BSE images of sample RBC 9 (Figure 5). The burrow fill material (sample RBC 5) is dominantly composed of clasts of poorly cemented, predominately finely crystalline calcium carbonate with some larger crystals and fossil remnants. Some clasts in the sample contain abundant glauconite pellets intermixed with carbonate; porous matrix material of intermixed glauconitic clay and carbonate surrounds some of the clasts (Figure 6). RBC-6, a sample of pale brown (Munsell 10YR 6/3) crust that lines some of the burrows, is composed of glauconite grains in a porous matrix of clay intermixed with carbonate cement. Goethite appears to preferentially coat and fill fractures within many of the glauconite grains (Figure 5). RBC 7, a gray (Munsell 10YR 5/1), <1 cm long siliceous crystalline boxwork structure (Figure 5) was



Figure 4. Blue arrow spans the width of filled burrows in the Austin Chalk, below the highly fossiliferous zone at the top of the limestone horizon.

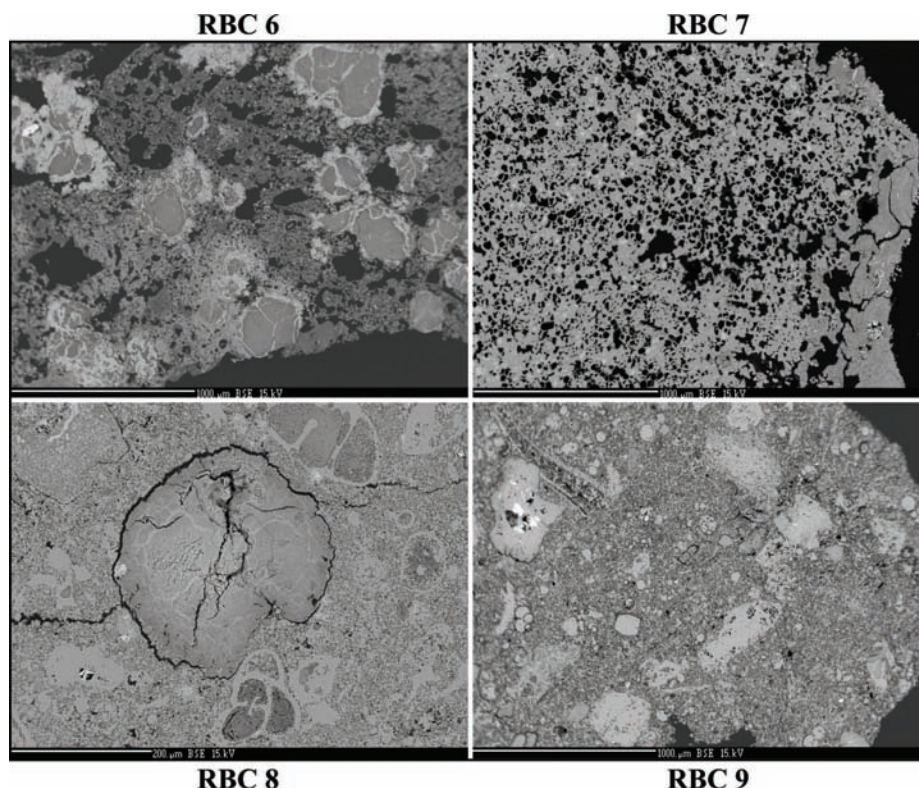


Figure 5. BSE images; RBC 6) goethite coated glauconite grains in mixed carbonate and clay matrix; RBC 7) siliceous crystalline boxwork structure; RBC 8) close up of glauconite grain and carbonate tests in mixed carbonate-dominant and clay matrix; RBC 9) carbonate grains, tests and glauconite pellets in mixed carbonate-dominant and clay matrix.

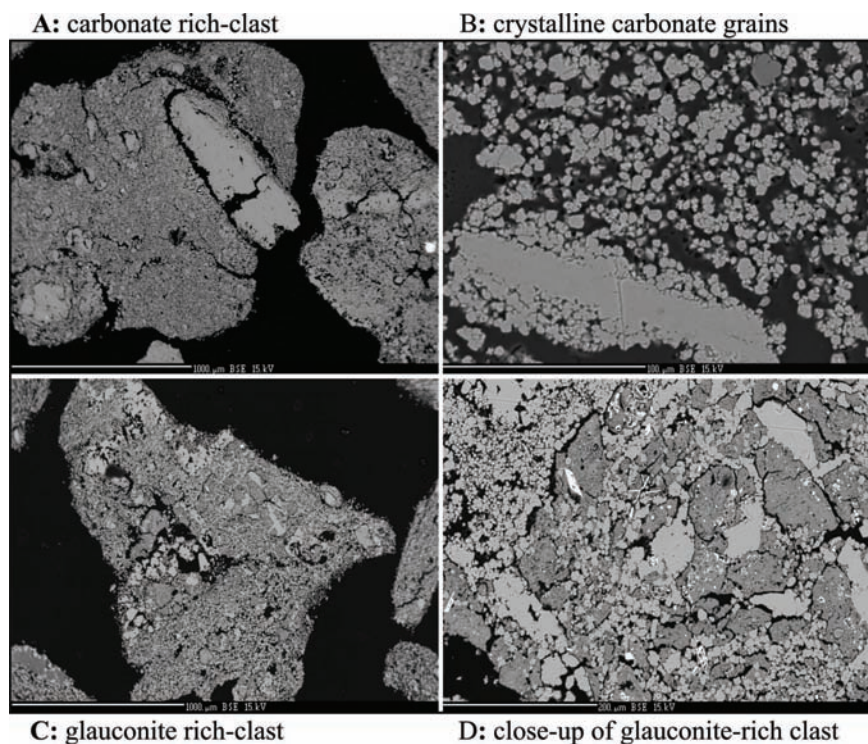


Figure 6. Sample RBC 5; A: Finely crystalline, dominantly carbonate clast with some larger carbonate grains or pellets. B: Close up of carbonate matrix showing loosely cemented, possibly recrystallized grains. C: Clast composed of abundant glauconitic pellets in carbonate cement. D: Close up of central portion of glauconite-rich clast. Bright specks are Fe-rich (goethite) grains.

found associated with RBC 6, but its origin and significance are not known.

Geologic structure

Robber Baron Cave is in the Alamo Heights horst, an upthrown block of Austin Chalk that is the dominant structural feature of the study area (Figure 7) and which topographically stands 20-30 m above the surrounding terrain. The horst formed by tectonic stresses associated with its location within the Balcones Fault Zone, which developed along the homoclinal hinge between the relatively flat-lying strata of the Edwards Plateau to the northwest and the more steeply dipping strata in the Gulf of Mexico Basin to the southeast. The Balcones Fault Zone is characterized by a series of en echelon normal faults, mostly downthrown toward the gulf.

Five major faults define the Alamo Heights horst. About 700 m north of the cave, one fault strikes N60°E. Three faults extend southwest from the northern fault. Respectively 1.8 and 2.7 km west of the cave, two near-parallel faults strike N25°E and N30°E while the third fault is 2.4 km east of the cave strikes a mean N47°E. These three faults are truncated by a fault 4.2 km south of the cave which strikes N55°E. Cumulative average fault

displacement is 75-110 m (Small, 1986). In the central section of the horst near the cave, Corbett et al. (1991) found that joints are generally vertical and trend N40° E. The beds within the horst are almost horizontal, but their exact attitude has not been precisely measured. No faults have been reported or mapped near the cave.

The geologic outcrop near Robber Baron Cave is almost entirely covered by urban development. No definitive fractures, sinkholes, outcrops, or other geologic features were found or are known within a 100 m radius. Cliffs and quarries in Olmos Creek and Salado Creek, which flank the Alamo Heights horst, provide the only opportunities (except within the cave) to closely examine the bedrock geology. No major geologic features were clearly apparent in aerial photographs, but weak lineaments that align with passages at the southwest corner of the cave almost certainly reflect joints in the bedrock. A 13-m-diameter circular feature located

about 17 m south of the cave's southeast corner is also faintly visible on the air photos (Figure 1). While there is insufficient information to determine if it is natural or artificial, the feature's similar size to the entrance sinkhole, and three radiating lineaments that parallel major fracture trends in the cave, suggest it may be a collapse of the cave beyond the current limit of exploration. No evidence of the possible collapse structure or fractures could be seen at ground level, but the surface is highly disturbed by decades of human activities (Veni, 1997a).

Most passages in Robber Baron are strongly joint-controlled. The orientation of linear passage segments can be measured from the map to approximate the joints' effect on the cave's development. The following discussion summarizes the analyses by Veni (1997a). The most dominant fractures guiding passage development range in orientation from N60-74°E, reflecting the nearby fault to the north, and comprise 27.9% of the cave's passages. The next most dominant group ranges twice as broadly from N25-54°E to include 24.5% of the passages, and reflects the N40°E jointing observed by Corbett et al. (1991) and the trends of the three faults

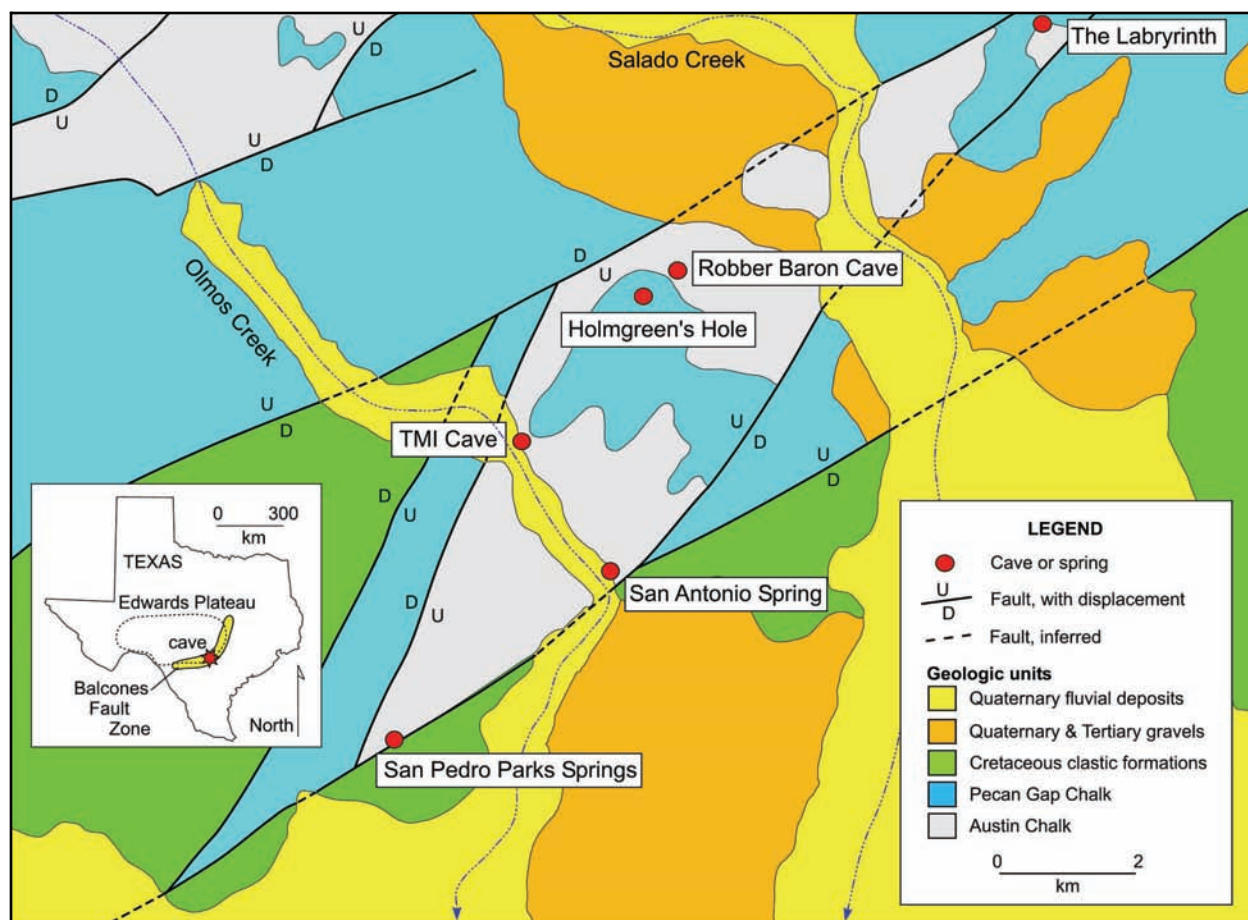


Figure 7. Geologic map of the Robber Baron Cave area (simplified from Barnes, 1983).

east and west of the cave. While these segments are secondary in terms of total passage length, they include the longest segments in the cave, with the N60-74°E segments being slightly shorter. A third important group of passage segments include two ranges: N41-45°W and N16-25°W. These trends do not correlate to the known structural features of the horst and may be secondary products of the tectonic strain. These two groups differ significantly in their horizontal and vertical distribution. Segments oriented between N41-45°W occur in the northern part of the cave, and generally include the cave's highest elevation passages. In contrast, the N16-25°W segments are predominantly grouped at the west-central and southwestern side of the cave, comprise most of the deepest passages, and are smaller, generally having about half the cross sectional area. The depth differentiation in passage orientation may suggest elevations where certain fractures are more permeable; additional study is needed.

Austin Chalk hydrogeology

Little research has been done on groundwater in the Austin Chalk, and no potentiometric mapping has been

conducted. Arnow (1959), George (1952), and Holt (1956) respectively described typical Austin groundwater in Bexar and neighboring Comal and Medina counties as yielding only small volumes to water wells, and as commonly high in hydrogen sulfide from the oxidation of the pyrite nodules within the chalk. However, Livingstone, Sayre, and White (1936) found that in some places in Bexar County, artesian water from the Edwards (Balcones Fault Zone) Aquifer (hereafter, "Edwards Aquifer") flowed upward along fractures to become Austin groundwater.

Data on groundwater flow between the Austin Chalk and the Edwards Aquifer are sparse (Veni, 1995). Contamination of the Edwards Aquifer from a landfill in the Austin Chalk reveals at least some downward interformational flow (Buszka, 1987). In contrast, significant volumes of rising artesian flow from the Edwards into the Austin is demonstrated at the San Antonio Spring in Olmos Creek, 4.5 km southwest of Robber Baron Cave, and San Pedro Park Springs 8 km southwest. Discharge records for San Antonio Spring

from 1892-1978 show Edwards water discharging from this Austin Chalk spring at rates up to 5.8 m³/s with a mean discharge of 1.2 m³/s; San Pedro Park Springs discharged up to 0.8 m³/s with a mean discharge of 0.22 m³/s from 1895-1977 (Brune, 1981). While Austin Chalk groundwater is certainly supplemented by artesian Edwards Aquifer water in some areas, there is insufficient data to determine its volume or rate of inflow, exact areas of contribution, storage, and seasonal and geographic variations.

Speleogenesis

The initial model on the origin of Robber Baron Cave (Veni, 1988) was that of vadose groundwater seeping down joints through the chalk horizon of the Austin Chalk to uniformly dissolve passages in the underlying more soluble limestone strata, per the model proposed by Palmer (1975) for some maze caves covered by permeable but insoluble sandstone caprocks. The passages in Robber Baron enlarged linearly along the joints until they interconnected to form the maze.

This model explained the gross morphology of the cave and several specific features, but not all. The most problematic feature involves solutionally enlarged and occasionally anastomosed bedding planes approximately 0.3 to 1.0 m below the top of the limestone horizon (Figure 8). If the cave had been formed by vertically descending groundwater, it would have flowed from the bedding planes into the main passages, incising the passage walls. Instead, little or no incision occurs and often the passage walls undercut the bedding planes. However, when Klimchouk (2007) proposed a hypogenic origin for the cave, by reversing the initially hypothesized downward direction of vertical flow through the cave, all of its features were easily explained. Klimchouk briefly identified side feeders and point features in the cave's floor as evidence of hypogenic flow. Three additional features discussed below in greater detail further support hypogenic development: sediments, solutionally enlarged bedding planes, rifts in the floors, and cusped passage connections.

Sediments

Two sediment samples were collected from the Pavilion Room at the north-central end of Robber Baron Cave; sediments observed in other parts of the cave had no discernible differences in appearance or setting to warrant

sampling. RBC 1 is a strong brown (Munsell 7.5YR 5/6) nodular clay from a pocket high in a passage wall. RBC 4 is a finely laminated reddish brown (Munsell 2.5YR 5/4) clay, with some distinctly yellow-brown sections, from the undisturbed floor of a humanly inaccessible passage. It contains some white specks, probably calcite crystals, and black specks.

While the hand specimen of each clay sample appears different in color and somewhat in texture, BSE images and qualitative scans revealed that they are essentially identical and residual dissolution products of the Austin Chalk (Figure 9). Qualitative scans for both clays showed dominant Si and Al peaks with some Fe and minor concentrations of Ca, Mg, and K. Quantitative analyses done on smooth areas of RBC 4 were consistent with the qualitative scans and additionally showed trace Ti. Based on the qualitative and quantitative data, the clays appear to be smectitic in composition and the observed color variation likely reflects differences in Fe oxidation not discernible with the microprobe. Black specks observed in RBC 4 appear bright in BSE images and qualitative scans show a Mn phase (likely a Mn oxide coating) superimposed on the clay signal. The Al peak appears slightly suppressed, possibly due to leaching of Al from the clay when the Mn-rich phase was deposited. A black (Munsell 5Y 2.5/1) veneer on RBC 1 may also be a manganese coating. While some hypogene caves are famously associated with sulfuric acid dissolution and the deposition of exotic minerals, carbonic acid dissolution is expected from artesian



Figure 8. Solutionally enlarged bedding plane and anastomoses in the Graffiti Room of Robber Baron Cave. Yellow line marks the contact between the chalk horizon and the underlying limestone horizon (the dip of the line is due to the photograph's wide angle, not a dip in the strata).

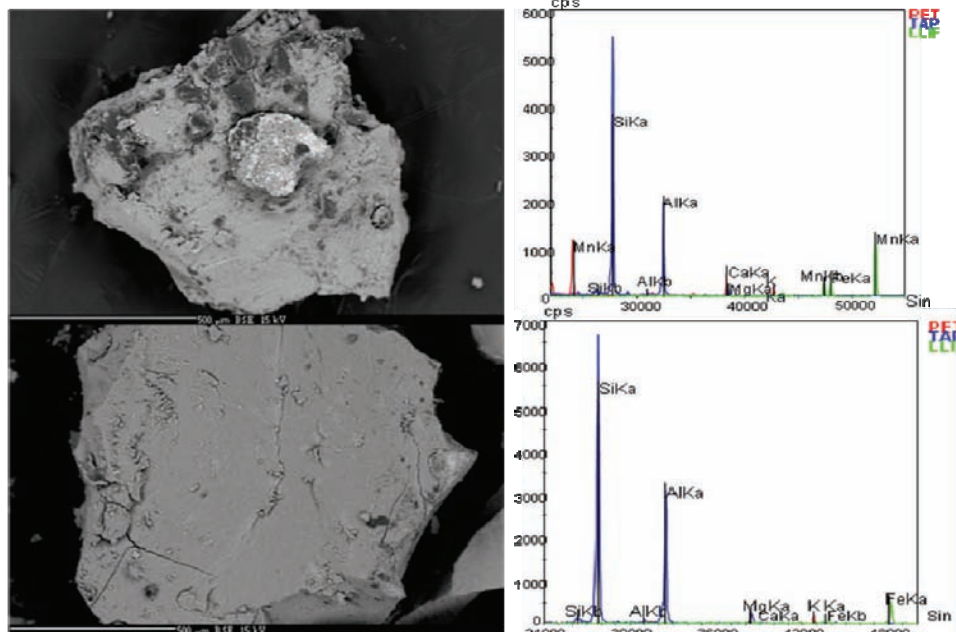


Figure 9. (top) BSE image and qualitative scan of dark particle in RBC 4 (appears bright in BSE image). Qualitative analysis shows presence of Mn-bearing phase overprinting clay signal; (bottom) BSE image and qualitative analysis of typical clay in RBC 1 and RBC 4.

Edwards Aquifer waters and the sediment analyses yielded no unusual minerals to suggest otherwise.

Solutionally Enlarged Bedding Planes

The solutionally enlarged and anastomosed bedding planes developed as upwelling water could not efficiently flow through the overlying chalk horizon and pressure forced it laterally into the bedding planes at the top of the limestone. Enlarged bedding planes that decrease in size with distance from the main passage, until they eventually pinch, represent areas of groundwater storage, while those that connect to other passages represent a pressure gradient along which water flowed to discharge upward through fractures in the chalk. A precise leveling survey of these enlarged bedding planes (some are humanly passable) may identify the past hydrologic gradients, but their general prevalence near the northeast and southeast ends of the cave suggests the sinkhole and the circular aerial lineament may have been foci of groundwater flow through the chalk horizon into overlying strata now removed by erosion. Direct evidence of this flow through the chalk has not been found and is not expected. The chalk is slightly ductile, and where collapsed it often seals open fractures and gaps, so any hypogenically enlarged fractures or conduits in the chalk probably closed long ago.

Rifts and Cusps

Rifts in the cave floor (Figure 10) were described by Klimchouk (2007) as “point” sources of rising water, but are actually elongated along joints. Most range from 3-5 m in length, and the longest is at least 30 m. Some rifts are separated by narrow fins of limestone but hydrologically and morphologically function as undi-

vided features. Natural and artificial sediment that covers many passage floors doubtlessly hides many rifts. Their importance, as compared to point sources of rising water, is that they are less prone to forming distinctive wall and ceiling channels but are more likely to create cusped passage connections. Cusped connections are usually expressed as abrupt, blade-like rises of the limestone floor with concurrent lowering of the ceiling along the axis of a passage to form relatively small windows within the passage (Figure 11). They can also occur between near-parallel passages. These windows are the most diagnostic feature of vertical groundwater movement; they would not exist in a regime with notable horizontal flow. Initially, individual passage segments developed along a joint above each rift. As upward flow was slowed by the overlying chalk, the segments enlarged laterally along the joints until they connected to form a cusped window (Figure 12). In many locations, continued passage growth completely removed the windows.

Not all of Robber Baron’s passages formed by rifts. Passages along the west-central and southwest side of the cave exhibit more classical elliptical phreatic morphologies and probably transmitted groundwater along less steep gradients, although more detailed surveys and studies are required to be certain of their origin. Ultimately, the Edwards Aquifer is the only viable source for the water, and the modern presence of Edwards water flowing from Austin Chalk wells and springs suggests that the cave developed when Edwards groundwater levels were significantly higher in the past.



Figure 10. Example of a 5-m-long floor rift.



Figure 11. Cusped window along the axis of passage.

Edwards Aquifer: evolutionary and management considerations

Groundwater flow in the Edwards Aquifer has developed in response to stream incision of the Balcones Escarpment, creating locations where springs can discharge Edwards water (Woodruff and Abbott, 1979). Preliminary assessment of cave distribution, morphology, origin, and regional hydrogeology suggests the aquifer grew west-to-east-to-northeast, as the groundwater drainage basins for ever-lower springs expanded headward until they captured water from older, higher elevation areas to progressively integrate the aquifer (Veni, 2009). The timing of these accretionary events has not been established, but the study of Robber Baron Cave provides some boundaries on the initial origin of San Antonio Spring and associated hydrologic changes within the aquifer.

The down-cutting of Olmos Creek can be used to constrain the time of the cave's origin. We considered the creek's elevation as an approximation for the elevation of the modern and past water table of the Edwards Aquifer in that area. Using the elevation of San Antonio Spring to reflect the mean modern

water table, we found that Olmos Creek was 47.7 m higher in elevation to create hypogene conditions in Robber Baron Cave. This elevation is the current land surface at the cave and was used to approximate minimum artesian conditions. We slightly modified the mean

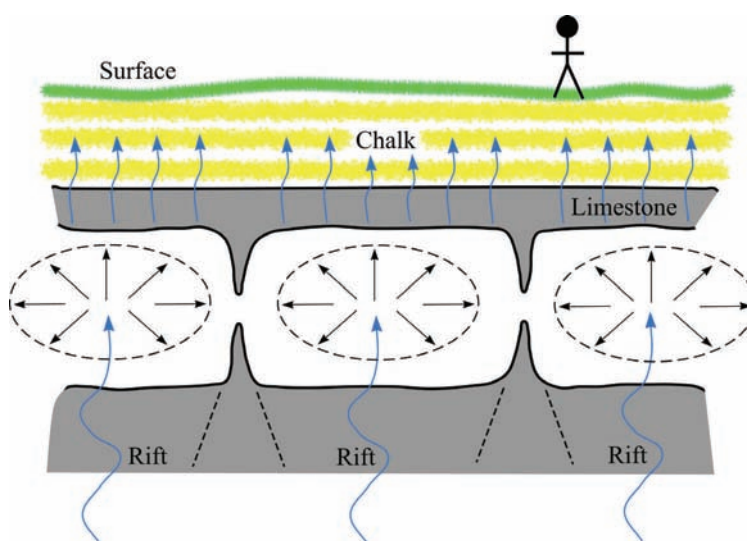


Figure 12. Schematic diagram of water rising from rifts in passage floors to create individual passage segments (ovals of dashed lines with arrows showing direction of growth) which linearly connect along joints to form cusped windows; groundwater continues to ascend by leakage through the chalk.

incision rates for a nearby section of the Balcones Escarpment region (Veni, 1997b) to 19-24 mm/ka to better fit the conditions in Olmos Creek, and determined that Robber Baron Cave would have developed about 2.0-2.5 million years ago. This is significantly older than previously calculated by assuming an epigenic origin for the cave (Veni, 1994).

Robber Baron Cave aligns with other caves and springs in the area in a pattern suggestive of changing base levels for the aquifer (Figure 13). The cave reportedly deepens and connects with other caves and a well roughly 1.2 km to the southwest. Another kilometer southwest is a former quarry where more caves have been reported, some rumored to connect to Robber Baron. Another 1.2 km further southwest is TMI Cave. This 58 m long multi-entrance maze is in a cliff overlooking Olmos Creek and was initially considered a product of backflooding by the creek (Veni, 1988). While TMI Cave has been vadosely modified by floodwaters, further study of the area now strongly suggests it is a fragment of a once more extensive hypogenic cave. It is located at the junction of Robber Baron's southwest trend with the southeast trend of a series of springs that extend 2.3 km southeast to San Antonio Spring.

About 2 to 2.5 million years ago, Robber Baron Cave transmitted flow to the first or one of the first ancestral San Antonio Springs. There is no direct evidence to demonstrate that Olmos or Salado Creek extended over the cave in the past, except that as a hypogenic cave, Robber Baron most likely discharged into a significant topographic low which would probably be an ancestral Olmos or Salado Creek; Salado Creek is suggested by gravel deposits in a paleo-channel 1.1

km northeast of the cave. As both streams deepened and shifted to their present positions, the potentiometric surface for the Edwards Aquifer lowered, and cave development, and presumed overlying springs, extended southwest toward TMI Cave and Olmos Creek. Spring development then predominantly followed the northeast side of the creek as progressively younger springs arose to the southeast, with flow through the older springs becoming more intermittent with continued potentiometric declines. This evolutionary scenario poses the management considerations discussed below.

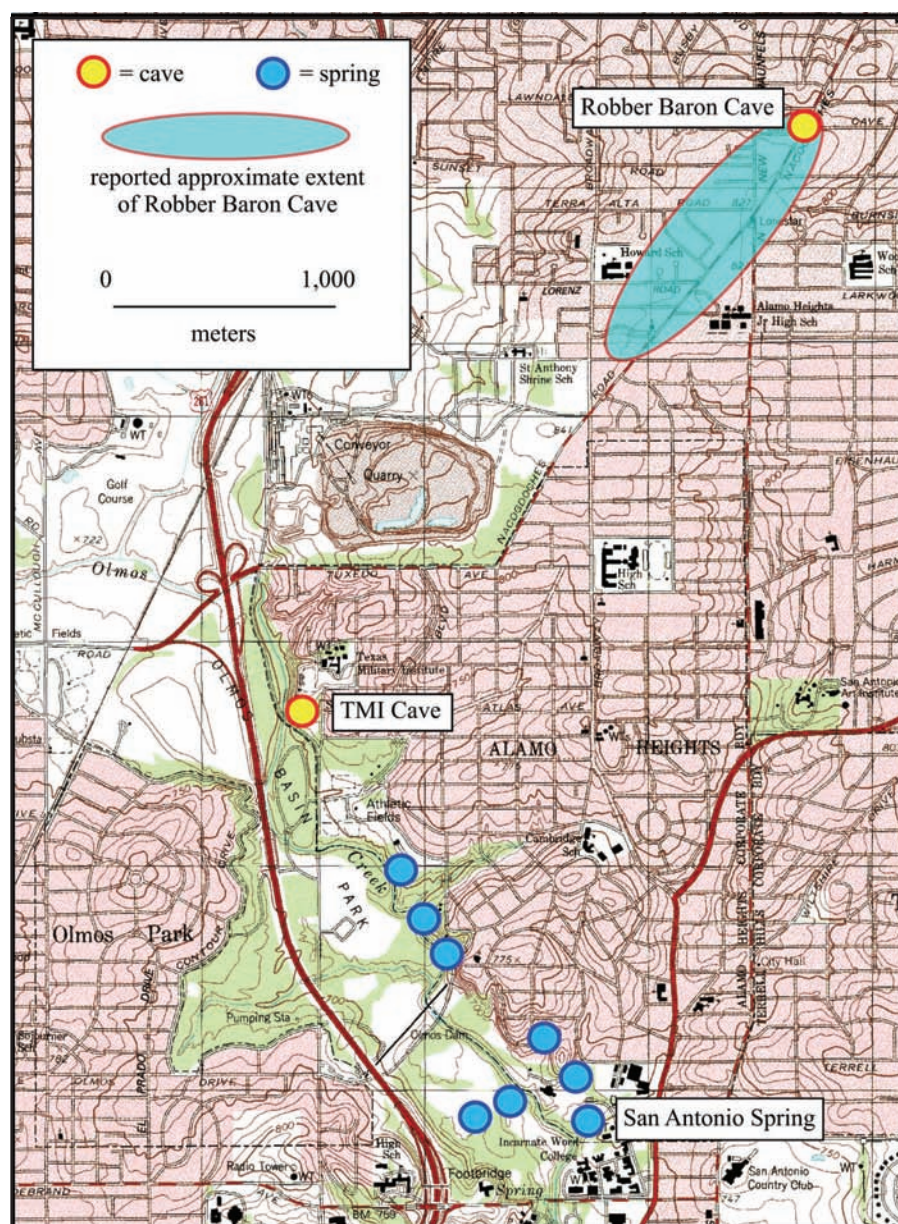


Figure 13. Topographic map showing the location of Robber Baron Cave and its estimated extent relative to TMI Cave and San Antonio and associated springs in Olmos Creek.

Groundwater contamination

Cavernous permeability from past hypogenic conditions creates a potential for contaminants to easily reach the modern Edwards Aquifer under current vadose conditions. The absence of significant contamination of the aquifer in this area is probably the function of three related factors. First, because the caves and related features were created to discharge, not recharge water, they generally do not possess surface catchment areas to channel water and contaminants into the Austin Chalk in sufficient volumes to pose imminent threats to Edwards water quality. Second, the chalk horizon covering much of the Austin results in a poorly developed epikarst. Additionally, the poorly permeable clayey soils of the area promote moderate to rapid runoff, and some form a caliche-like layer at their base which further inhibits recharge (Taylor et al. 1966). Third, due to the first two factors, most contaminants that enter the Austin Chalk are diffusely distributed through the area and with little hydraulic head, slowly entering the Edwards Aquifer at a rate where they are diluted below detection limits.

Despite the factors that protect against groundwater contamination in the Austin Chalk, significant pollution risks exist. Robber Baron Cave is home to six endemic invertebrate species, two of which are federally listed as endangered (U.S. Fish and Wildlife Service, 2000). Nothing is directly known about the tolerance of these species to pollutants; potentially even small and diffuse amounts of contaminants may adversely impact their populations. Significant threats to the aquifer occur at sites that store large amounts of hazardous materials, such as gasoline stations. If they occur over hypogenic karst, the potential danger is due to the volume and toxicity of pollutants that could be released in point locations. Detailed studies of the karst and hypogenic conditions are crucial to establish the probable areas where contaminants on the surface might enter a cave like Robber Baron, and to place hazardous materials where less karstification has developed for better containment in case of an accidental release.

The most significant risk of Edwards Aquifer contamination through the Austin Chalk occurs along Olmos Creek between TMI Cave and the San Antonio Spring. This is an area of enhanced permeability due to past and modern spring flows. Due to pumping of the aquifer, all of the springs periodically cease flowing and become estavelles, episodic sites of both recharge and discharge depending on potentiometric levels. When the springs aren't flowing, flooding of Olmos Creek recharges the Edwards Aquifer through these springs and paleosprings in the Austin Chalk. Turbid water and organic debris

have been discharged from nearby artesian Edwards wells after such floods (Veni, 1985), which drain 88.2 km² of highly urbanized northern San Antonio. The volume of recharge has not been quantified, but whirlpools have been observed for days over the San Antonio and associated springs. No study has been conducted to identify which other Edwards wells intercept this recharge, or to determine the time of travel, dilution, and dispersal of contaminants before they reach those wells. A major contaminant release (e.g., tanker spill, ruptured sewer line) into Olmos Creek during such a recharge event could severely impair the drinking water quality in some parts of the Edwards Aquifer.

Risk delineation

Hypogenic karst poses a certain higher risk of environmental and land use problems because it is often poorly expressed on the land surface. Since environmental vulnerability in karst is frequently measured by the presence of solutional sinkholes and related features, such as with the state's regulations governing the Edwards Aquifer region, the vulnerability of hypogene karst may be significantly underestimated with regard to groundwater contamination, ground stability, endangered subsurface fauna, and related issues.

The presence of hypogenic caves is a clear indication of hypogenic conditions, but evolutionary studies of regional karst development is necessary for construction of conceptual models to define their probable range. This study has focused on Robber Baron Cave and illustrates that at least the southwest portion of the Alamo Heights horst, including adjacent sections of the Olmos Creek valley, have or still experience hypogenic groundwater flow. It is beyond the scope of this report to conduct digital modeling of groundwater movement to try and better define the area, but the conceptual description in this paper provides the framework for such modeling and especially for the development of a focused research program to more accurately delineate the area and its conditions.

At least two and possibly four other Austin Chalk areas in Bexar County are hypogenic and require further study: San Pedro Park Springs, The Labyrinth, probably Salado Creek, and possibly Culebra Anticline. The San Pedro Park Springs were previously mentioned in this report as discharging artesian water from the Edwards Aquifer. Several caves are reported nearby, but none are accessible, mapped, or studied; the area has been urbanized for about 100 years, and all reported caves are filled. The Labyrinth is an incompletely explored maze cave with at least 248 m of surveyed passage and morphologic features report-

edly similar to Robber Baron Cave that strongly suggest it also has a hypogenic origin (Texas Speleological Survey, unpublished reports and data). It is located 6.4 km northwest of Robber Baron and is probably related to the Edwards Aquifer. Edwards Aquifer springs have been reported anecdotally along Salado Creek within and downstream of the Alamo Heights horst. A study verifying these reports and determining the distribution of probable paleosprings is warranted to better evaluate the potential loss of contaminated urban stream flow into permeable hypogene features that may recharge the Edwards Aquifer. The outcrop of Austin Chalk on the Culebra Anticline in western Bexar County holds several extensive and hydrologically active caves (Veni, 1997a). Phreatic morphologies are known in these caves, but no clearly hypogenic features have been noted. No springs have been identified for the area, and there is speculation that it may recharge the Edwards Aquifer. If this area proves hypogenic or a non-hypogenic recharge area for the Edwards Aquifer, it is potentially more vulnerable to groundwater contamination than the Alamo Heights horst area due to significant differences in lithology, soil, and abundant, focused, surface drainage into karst features.

Conclusions

The longest cave in Bexar County, Robber Baron Cave is also without a doubt the most complicated. This maze cave has a long history of exploration, yet most of its passages are currently inaccessible. It formed hypogenically from artesian water in the Edwards Aquifer that rose to the surface 2 to 2.5 million years ago, to flow as possibly the first location of the ancestral San Antonio Spring. The alignment of the cave with other caves, springs, major fracture patterns, and the likely direction of the lowering potentiometric surface with the incising Olmos Creek, strongly suggest a physical and hydrological connection as the cave's conduit network and its overlying spring system migrated 3.4 km southwest to TMI Cave and then 2.3 km southeast to the current site of the San Antonio Spring.

Understanding the cave's origin and how the evolution of the Edwards Aquifer and San Antonio Spring are related allow better delineation and study of previously unrecognized areas that pose potential risks to aquifer quality. Understanding of hypogenic processes and identification of key features and conditions also make it possible to study other areas in the region to determine if currently unrecognized hypogenic connections exist with the Edwards Aquifer that warrant investigation and protection. Existing hazardous material storage facilities over the known, reported, and projected extent of Robber Baron Cave should be at be more carefully

monitored for leaks or spills; proposed facilities should not be sited without detailed hydrogeologic investigations, including geophysical surveys for possible underlying hypogenic conduits.

Additional research in Robber Baron Cave and the surrounding area would better define and delineate its hydrogeologic history and physical limits. Recommended studies include detailed stratigraphic mapping, a precise leveling survey of key passage levels, geophysical surveys to determine the density of passages beyond the limits of modern exploration, speleothem dating to establish the upper boundary on the timing of vadose conditions, hydrologic investigation of groundwater within the Austin Chalk, including tracer studies when San Antonio Spring and other estavelles in Olmos Creek are recharging, and most especially, further exploration of the cave and other caves in the area.

Acknowledgments

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References

- Arnow, Ted. 1959. *Ground-water geology of Bexar County, Texas*. Texas Board of Water Engineers Bulletin 5911. Austin: Texas Board of Water Engineers.
- Barnes, Virgil E. 1983. *Geologic atlas of Texas, San Antonio sheet*. Austin: Bureau of Economic Geology.
- Brune, Gunnar. 1981. *Springs of Texas, volume 1*. Fort Worth: Branch-Smith, Inc.
- Buszka, Paul M. 1987. *Relation of water chemistry of the Edward Aquifer to hydrogeology and land use, San Antonio region, Texas*. U.S. Geological

- Survey Water-resources Investigations Report 97-4116. Austin: U.S. Geological Survey.
- Cloud, K.W. 1975. The diagenesis of the Austin Chalk. M.S. thesis: University of Texas-Dallas.
- Corbett, K. 1982. Structural stratigraphy of the Austin Chalk. M.S. thesis, Texas A&M University.
- Corbett, K., M. Friedman, D.V. Wiltschko, and J.H. Hung. 1991. Controls on fracture development, spacing, and geometry in the Austin Chalk Formation, central Texas – considerations for exploration and production. In *Dallas Geological Society Field Trip #4, American Association of Petroleum Geologists Convention Guidebook*. Tulsa: American Association of Petroleum Geologists.
- Dravis, J.J. 1979. Sedimentology and diagenesis of the upper Cretaceous Austin Chalk Formation, south Texas and northern Mexico. PhD diss., Rice University.
- George, W.O. 1952. *Geology and ground-water resources of Comal County, Texas*. U.S. Geological Survey Water-Supply Paper 1138. Austin: U.S. Geological Survey.
- Holt, C.L.R., Jr. 1956. *Geology and ground-water resources of Medina County, Texas*. Texas Board of Water Engineers Bulletin 5601. Austin: Texas Board of Water Engineers.
- Klimchouk, Alexander. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Livingstone, P.P., A.N. Sayre, and W.N. White. 1936. *Water resources of the Edwards Limestone in the San Antonio area, Texas*. U.S. Geological Survey Water-Supply Paper 773-B. Austin: U.S. Geological Survey.
- Mitchell, Joe N., and Linda K. Palit. 2009. Robber Baron: Restoring an urban cave preserve. In *Proceedings, 15th International Congress of Speleology*: Kerrville, Texas (in press).
- Palmer, Arthur N. 1975. The origin of maze caves. *National Speleological Society Bulletin* 37 (3): 57-76.
- Pessagno, E.A., Jr. 1969. *Upper Cretaceous stratigraphy of the western Gulf Coast area of Mexico, Texas, and Arkansas*. Geological Society of America Memoir 111. Denver: Geological Society of America.
- Small, Ted. 1986. *Hydrogeologic sections of the Edwards Aquifer and its confining units in the San Antonio area, Texas*. U.S. Geological Survey Water-Resources Investigations Report 85-4259. Austin: U.S. Geological Survey.
- Taylor, F.B., R.B. Hailey, and D.L. Richmond. 1966. *Soil survey of Bexar County, Texas*. Soil Conservation Service, Series 1962 (12). Washington: U.S. Department of Agriculture.
- U.S. Fish and Wildlife Service. 2000. Endangered and threatened wildlife and plants; final rule to list nine Bexar County, Texas invertebrate species as endangered. *Federal Register* 63 (248): 81,419-81,433.
- Veni, George. 1985. Effects of urbanization on the quantity and quality of storm-water runoff recharging through caves into the Edwards Aquifer, Bexar County, Texas. M.S. thesis, Western Kentucky University.
- Veni, George. 1988. *The caves of Bexar County, second edition*. Texas Memorial Museum Speleological Monographs 2. Austin: University of Texas.
- Veni, George. 1989. The history, legend, and continued exploration of Robber Baron Cave. *NSS News* 47 (6): 133-138.
- Veni, George. 1994. *Geologic controls on cave development and the distribution of endemic cave fauna in the San Antonio, Texas, region*. Report for Texas Parks and Wildlife Department and U.S. Fish and Wildlife Service. Austin: Texas Parks and Wildlife Department.
- Veni, George. 1995. Revising the boundaries of the Edwards (Balcones Fault Zone) Aquifer recharge zone. In *Proceedings, Water for Texas Conference*. College Station: Texas Water Resources Institute.
- Veni, George. 1997a. *Evaluation of areas of potential influence on karst ecosystems for certain caves in Bexar County, Texas (part 2 of 2)*. Report for U.S. Fish and Wildlife Service. San Antonio: George Veni and Associates.
- Veni, George. 1997b. Geomorphology, hydrogeology, geochemistry, and evolution of the karstic lower Glen Rose Aquifer, south-central Texas. PhD diss., Pennsylvania State University: Texas Speleological Survey Monographs 1. Austin: Texas Speleological Survey.
- Veni, George. 2009. Caves and karst of the Great Plains. In *Caves and karst of the USA*, ed. A.N. Palmer and M.V. Palmer. Huntsville: National Speleological Society (in press).
- Waddell, R.K. 1977. Environmental geology of the Helotes Quadrangle, Bexar County, Texas. M.S. thesis, University of Texas.
- Woodruff, C.M., and P.L. Abbott. 1979. Drainage basin evolution and aquifer development in a karstic limestone terrain, south-central Texas, U.S.A. *Earth Surface Processes* 4: 319-334.
- Young, Keith. 1985. The Austin Division of central Texas. In *Austin Chalk and its type area – stratigraphy and structure: Austin Geological Society, Guidebook 7*, ed. K. Young and C. M. Woodruff. Austin: Austin Geological Society.

REACTIVATED BASEMENT FAULTING AS A HYDROGEOLOGIC CONTROL OF HYPOGENE SPELEOGENESIS IN THE SOUTHERN OZARKS OF ARKANSAS, USA

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Abstract

Integration of data from recent, diverse research thrusts has enabled development of a conceptual model of hypogene speleogenesis for Chilly Bowl Cave—a system representative of Ozark Plateaus karst developed over deep basement faults in northern Arkansas, USA. The model is based on and is constrained by geologic mapping, cave mapping, structural reconstruction, gravity mapping, hydrogeologic framework characterization (non parallel construction), ground-water tracing, endangered-species distribution mapping, cave mineralogy, stable-isotope and fluid inclusion geochemistry. Preexisting basement faults—reactivated during tectonic pulses of the Ouachita orogeny, the core belt of which was centered about 150 kilometers to the south—are recognized as a critical framework element in control of regional karst development. Confined, geothermally heated water was expelled from the foreland basin, flowed under pressure toward the Ozark Dome, and recharged overlying regional Paleozoic aquifers from below via integrated faults and systematic joints; the geochemistry of these waters was established in the deeply buried, clastic-dominated sediments of Ouachita forebasin, which upon entering the carbonate-rich Paleozoic sequences of the Ozarks, created a striking diagenetic regime, creating caves and depositing minerals—including those of the MVT ores. Cave erosional features and deposits in Arkansas have been found that indicate past episodes of thermal ground-water circulation. These include upper-level paleo-cave passages and chambers, now reactivated by present-day stream drainage, scallops in conduits and flow tubes evincing upward flow, dolomite breccia linings, and remarkable dog-tooth spar (calcite) crystals, some of which are as long as 1.9 meters. The stable isotopic signature of the early-formed core of these crystals is anomalous in context with local host

rocks, and inconsistent with a meteoric-fluid origin of precipitation and stream recharge percolating downward from the surface. The conceptual model incorporates all known information regarding this hydrogeologic system, and integrates seemingly disparate observations into a coherent, consistent model that requires hypogene speleogenesis as an initial step in a remarkably complex history of overprinting by numerous stages of karst development and evolution.

Introduction

The Ozark Plateaus (Figure 1) are an ancient, variably karstified region of the mid-continent of the U.S. that have more than 8500 reported caves, tens of thousands of springs, and a diverse suite of accompanying karst landforms and hydrogeologic features (Imes and Emmett, 1994; Brahana et al., 2005). Basement rocks in the Ozarks have a northeast-northwest orthogonal fracture pattern expressed in geophysical signatures, as well as major faults at land surface (Hendrix et al., 1981; Haley et al., 1993; Hudson, 2001; Johnson, 2008).

Whereas much of the karst of the Ozark Plateaus was formed by epigenetic processes, hypogenic karst also is present, although it seldom has been recognized nor seldom has been the object of major research (Tennyson et al., 2008). The term hypogenic karst (or hypogene speleogenesis) used herein follows the definition of Ford (2006), in which "the formation of caves is by water that recharges the soluble formation from below, driven by hydrostatic pressure or other sources of energy, independent of recharge from the overlying or immediately adjacent surface." The lack of discernible genetic relation with recharge from the overlying surface is the main characteristic of this process (Klimchouk, 2007). Epigenetic karst is formed by aggressive recharge descending from the

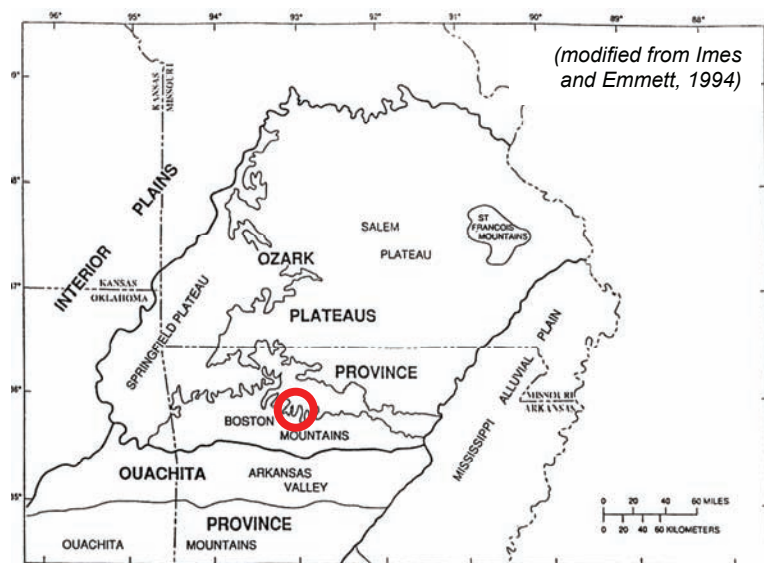


Figure 1. The Ozark Plateaus, showing the approximate location of northeast-trending basement faults in northern Arkansas. The Upper Flatrock Creek fault study area in Figure 2 is shown by the circle (modified from Imes and Emmett, 1994).

land surface (Ford and Williams, 2007). Understanding the distinction between these types of karst and the dominant factors that control them has evolved rapidly over the past several decades, and only recently has been clearly elucidated (Hill, 2000; Klimchouk et al., 2000; Klimchouk, 2007).

This report documents a subset of hypogenetic features that overlie deep northeast-trending faults that extend from the land surface to the basement in northern Arkansas, and specifically, the area of Chilly Bowl Cave overlying the Upper Flatrock Creek fault in Newton County, highlighted by the open circle in Figure 1. We herein describe the integration of previous data from diverse research efforts incorporating geologic mapping (Hudson, 1998; Hudson et al., 2001; Braden and Ausbrooks, 2003; and Hudson and Murray, 2004), cave mapping (Jim Terry, Rodney Tennyson, and David Taylor, 2008, written communication, Clarksville, Arkansas), structural geology (Granath, 1989; Hudson, 2001), gravity mapping (Hendricks et al., 1981), hydrogeology (Brahana, 1997), ground-water tracing (Dawn Cannon, 2001, written communication, Fayetteville, Arkansas; Chuck Bitting, National Park Service, 2008, written communication, Harrison, Arkansas), endangered-species distribution mapping (Brown and Graening, 2003), cave mineralogy, stable-isotope geochemistry, economic mineral occurrence (McKnight, 1935; Smith, 1977), and fluid inclusion studies (Leach, 1979; Leach and Rowan, 1986) to constrain our understanding of

the factors that control hydrogeology and karst development in the region, both hypogenic and epigenetic. We also introduce previously unpublished stable-isotope data ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) and interpretations to elucidate the hydrogeochemical environment under which the spar crystals were precipitated.

Purpose and scope

The purpose of this study is to introduce and document a conceptual model of hypogene speleogenesis for areas of the Ozark Plateaus that overlie reactivated Precambrian faults. This report is limited in scope to the area of northwest Arkansas, where major faults are well-defined by mapping and remote sensing (Haley et al., 1993; Hudson, 1998; Hudson et al., 2001; Braden and Ausbrooks, 2003; Hudson and Murray, 2004; and Johnson, 2007). Emphasis and most examples from this paper are drawn from Newton County, and specifically the Flatrock Creek fault (Figure 1-area of the open circle).

Conceptual model

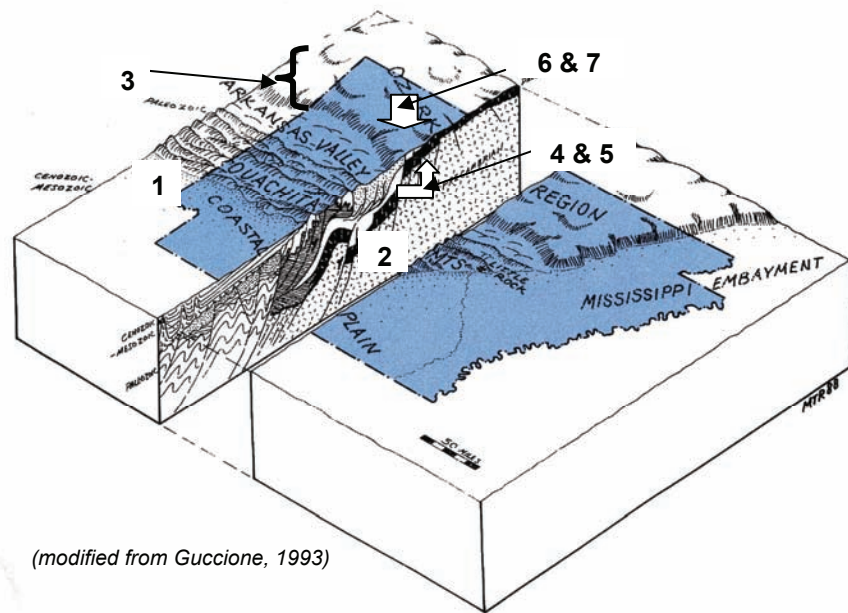
The conceptual model of hypogene speleogenesis in the southern Ozarks of northern Arkansas (Figure 2) is consistent with all available data from a wide range of disciplines. Simply, the Ouachita orogeny was the driving force for differential uplift of basement rocks along preexisting, major northeast-trending faults. The greater strength of the crystalline basement rocks allowed much of the vertical energy transmission, acting as broad, thick fault-bound levers that facilitated uplift. Differential uplift was responsible for fracturing the relatively thin overlying sedimentary cover in a systematic stress field of orthogonally jointed and faulted rocks. The northeast-trending basement faults also served as preferred pathways for fluid migration from the south. Thick basinal sediments in the Arkoma Basin and the Ouachita orogen served as the source of organics, metals, and aggressive solutions. The oblique closure of the Ouachita orogeny from east to west expelled the heated solutions toward the north. Deep burial and the geothermal gradient supplied heat to the fluids as they moved in the deep subsurface from the Ouachitas into the carbonate rocks of the Ozark Plateaus Province. The aggressive nature of the fluids dissolved and enlarged preferred flow paths from below, creating hypogenic karst features. Younger minerals, including sphalerite and calcite (dog-tooth spar), were precipitated in voids in carbonates below the shale and sandstone cover by

later-stage thermal fluids. The terrigenous sedimentary rock cover provided insulation and pressure confinement such that as the heated water was advectively transported, it cooled slowly, and retained an anomalously warm signature far into the Ozarks. The overlying sedimentary-rock cover was karstified by multiple pulses of hot water, especially in areas overlying the deep faults.

Epigenetic karstification was later overprinted on the hypogenic karst, and in many cases, meteoric recharge extensively modified original hypogenic karst as local flow conditions responded to capture of surface streams. After the pulses of uplift from the Ouachita orogeny ceased, base level was successively lowered over a long interval of geologic time, estimated to be hundreds of millions of years. Several ancient segments of the original hypogenic caves have been preserved, owing to the protection by the sandstone and chert cover layers, and the diversion of epigenic recharge to lower components in the flow system that fortuitously have kept flow away from fragile and reactive formations. Geologically recent activity, mostly Pleistocene in age, includes precipitated and evaporative speleothems that were formed in the vadose zone.

Discussion

A recent program to refine geologic mapping at a 1:24,000 scale within and near the study area has been undertaken by several agencies and universities, most notably the U.S. Geological Survey, the National Park Service, the Arkansas Geological Commission, and the University of Arkansas (Hudson, 1998; Hudson et al., 2001; Braden and Ausbrooks, 2003; Hudson and Murray, 2004, and Johnson, 2008). Surface geology in the study area ranges in age from Ordovician to Pennsylvanian. Lithologies include a thick sequence of carbonate rocks in the deeper part of the section, with dolomite dominant up to the Ordovician, and limestone dominant from the Ordovician through the middle Mississippian (Figures 3 and 4).



(modified from Guccione, 1993)

Figure 2. Conceptual model. 1) compression caused by oblique closure (toward northwest) of Ouachita orogen; 2) reactivation of deep basement faults, resulting in differential uplift that accommodates shortening from compression; 3) faulting and tilting of relatively thin sedimentary-rock cover into orthogonally jointed and faulted blocks; 4) expulsion of deep, hot, aggressive water from the south upward along fractures, creating hypogene speleogenesis; 5) later pulses of hot, mineralized water deposited MVT minerals and accessory minerals in fractures and voids in pores in carbonate rocks and previously created hypogene caves; 6) very long history of weathering (on the order of hundreds of millions of years) lowered base level of main streams and drains; 7) multiple stages of overprinting of hypogenic karst by epigenic karst processes from captured surface water and meteoric recharge; partial plugging of preexisting voids with speleothems and insoluble debris washed into caves from the surface.

The Boone Formation, a major karst-forming rock unit in the region, has significant (as much as 70%) chert throughout much of the area. Chert and other insoluble materials remaining after weathering mantle surface-karst features, and provide a distinct nonkarst appearance to the untrained eye, especially in that physiographic province of the northwest Arkansas Ozarks classified as the Springfield Plateau (Figure 1). Clastic sediments characterize Mississippian though Pennsylvanian rocks. Shales are the dominant lithology, and where these are interlayered with siltstones, they form confining units; where sandstones and limestones occur, they form locally important aquifers. Outcropping shale allows passage of little recharge, diverting most rainfall laterally to surface streams; karst development below shale is limited, although locally it may be well-developed where fracturing prevails.

Structurally, the rocks in this area are nearly flat-lying (Hudson, 2001). Dips typically are less than 3°, except

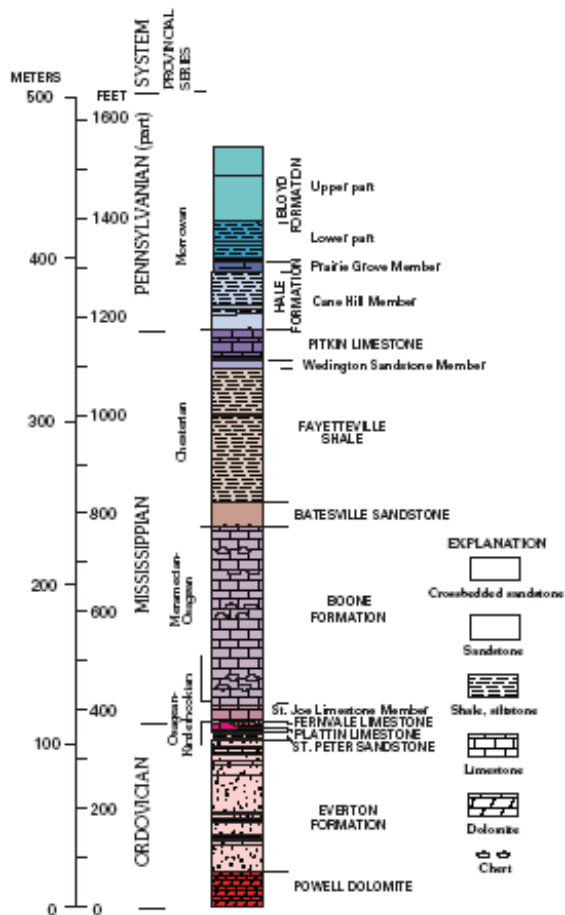
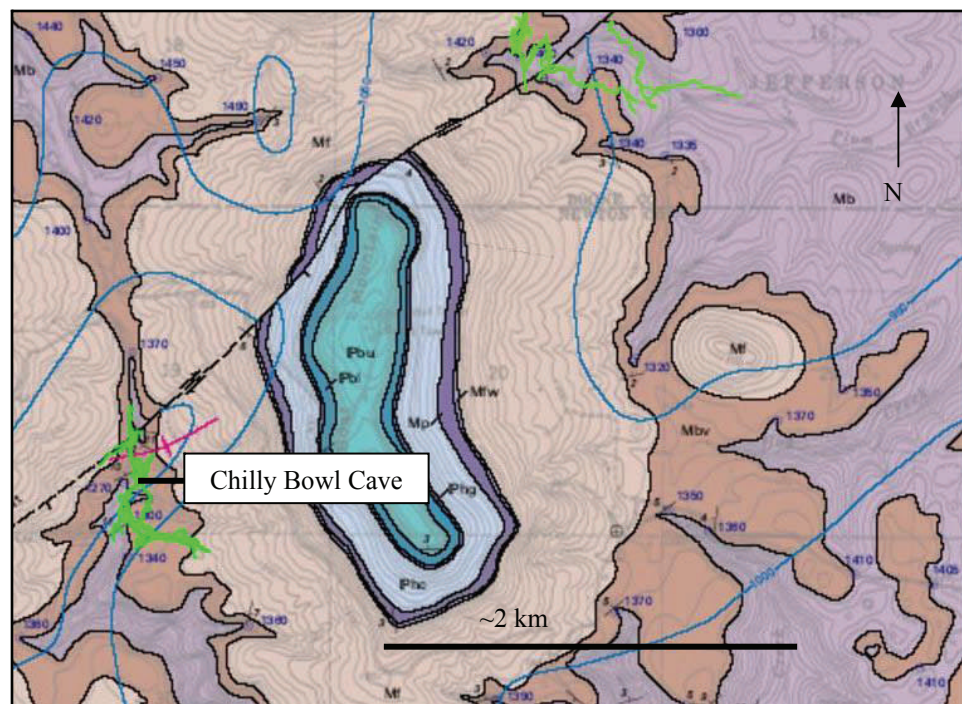


Figure 3. Partial stratigraphic column in the area of interest (modified from Hudson and Murray, 2004).

for locations where faulting has occurred (Figure 4). Uplift is observed to increase from west to east across the study area, and near Flatrock Creek fault, in Newton County, and the nearby bluffs of the Buffalo National River. Vertical cave entrances are consistent with uplift of tens to hundreds of meters (Figure 5) compared with the same formations in nearby counties (Fanning, 1994). Brittle fracturing (systematic jointing) dominates the structural geology (Hudson, 2001) suggesting thin overburden during deformation of these formations.

Timing of the faulting is constrained by surface mapping, and by olistoliths. Soft-sediment deformation in the form of olistoliths is documented from one recently excavated road cut within 50 meters of a major northeast-trending basement fault (Bella Vista fault). This outcrop lies near the highway 412 bridge that crosses the Illinois River near Robinson, in Benton County (~120 km east of the Flatrock Creek fault), indicative that movement on this basement fault was active as early as Osagean time (about 360 million years ago), when St. Joe sediments had yet to be indurated (Chandler, 2001). Energy released from the surface waves associated with this faulting had to be significant enough to initiate sliding on basal bedding plane with a dip of no more than 3°. Within Chilly Bowl Cave, faults are common (Figure 6) and clearly visible. Only a few of these faults are of a magnitude to be shown on published geological quadrangle maps at a scale of 1:24,000 (Hudson, 1998; Hudson et al., 2001; Braden and Ausbrooks,

Figure 4. Relation of the northeast-trending Flatrock Creek fault to the surface geology (from Hudson and Murray, 2004) and cave locations (after Tennyson et al., 2008). Structural contours on top of the Boone Formation are shown in blue, and cave locations are shown in green. Chilly Bowl Cave is furthest west and is labeled, Endless Cave is furthest east, and Big Hole Cave is in between.



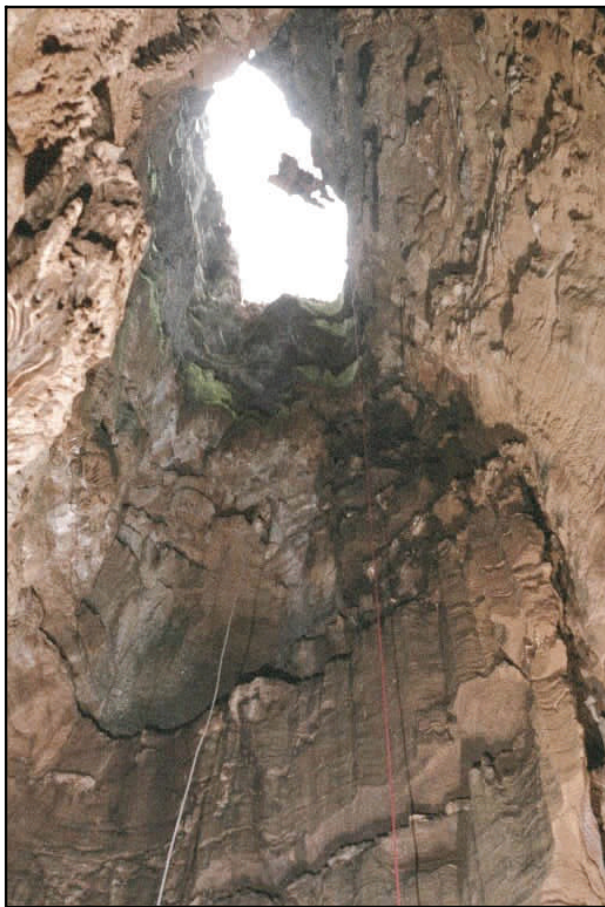


Figure 5. View from inside Big Hole Cave looking up at the entrance about 38 meters above.

2003; and Hudson and Murray, 2004). Most of these individual faults appear to have been important at controlling ground-water flow that formed cavern passages.

Bouguer gravity anomalies suggest that reactivation of basement faults has been a major control on fracturing in this part of the Ozarks (Hendricks et al., 1981), owing to the steep gradients in gravity near known northwest-trending surface faults. The occurrence, symmetry, and magnitude of the gravity anomalies is such that basement rocks are required to be involved; rocks of the sedimentary cover alone cannot explain the observed data. Deep drilling records are yet additional data sources that reflect the displacement of basement rocks across major faults (Hendricks et al., 1981).

Hydrogeologically, flow pathways and cave passages tend to be aligned and to lie on near-horizontal bedding planes of slightly-tilted blocks (Brahana, 1997), or along near-vertical shafts that follow joints and fractures (Jim Terry and Rodney Tennyson written

communication, 2008). Brahana (1997) has used orthogonal fracturing to successfully delineate spring-basin boundaries in shallow karst aquifers elsewhere in the region, and the mechanism proposed appears to be applicable to Newton County as well. In a cave such as Chilly Bowl, dozens of faults and fracture features are common (Figure 6).

Hydraulic gradients of the present ground-water flow system, which generally appear to follow the tilt (structural dip) of the rock formations, act independently (are decoupled) from surface-water bodies, particularly where chert horizons in the Boone Formation act as effective confining layers. As such, this decoupling is lithologically controlled, but in areas of faults and major joints, is dependent on structure. Stream piracy is one manifestation of combined lithologic and structural control that is obvious.

Dye tracing studies within (Dawn Cannon, University of Arkansas, 2000, written communication, Fayetteville, Arkansas) and near the study area (Tom Aley, 2004, written communication, Protom, Missouri) indicate that ground-water flow across surface-water drainage divides is common. Water-level maps and cave diving explorations indicate that portions of some faults currently act as ground-water dams.

The distribution of most known occurrences of endangered species of cave crayfish and cavefish [*Cambarus setosis*, *Cambarus aculabrum*, and *Amblyopsis rosea*] (Brown and Graening, 2003; David Kampwerth, U.S. Fish and Wildlife Service, 2008, written communication, Conway, Arkansas) along the major northeast trending faults in northwest Arkansas is consistent with deep faulting. It is suggested that the oblique closure of the Ouachitas was transferred to the distal part of the stress field by the stronger basement rocks, resulting in reactivation of Precambrian faults. Differential movement of these basement faults resulted in uplift (Figure 4), during which brittle sedimentary rocks were faulted, systematically jointed, and tilted slightly; overlying the major basement faults, fracturing was continuous from basement to land surface. These dominant fault zones underwent dissolution and later served as conduits for ground-water flow as well as pathways for dispersal of subterranean species. Endangered species are found only in caves that lie at the distal end of current flow systems that are large enough for human entry. It is suspected that these organisms also occur within deep aquifers that are inaccessible to humans. Hydrocarbon spills in the southern Missouri part of the Ozarks support this: endangered species live within carbonate aquifers, and are not restricted only to caves. In several reported

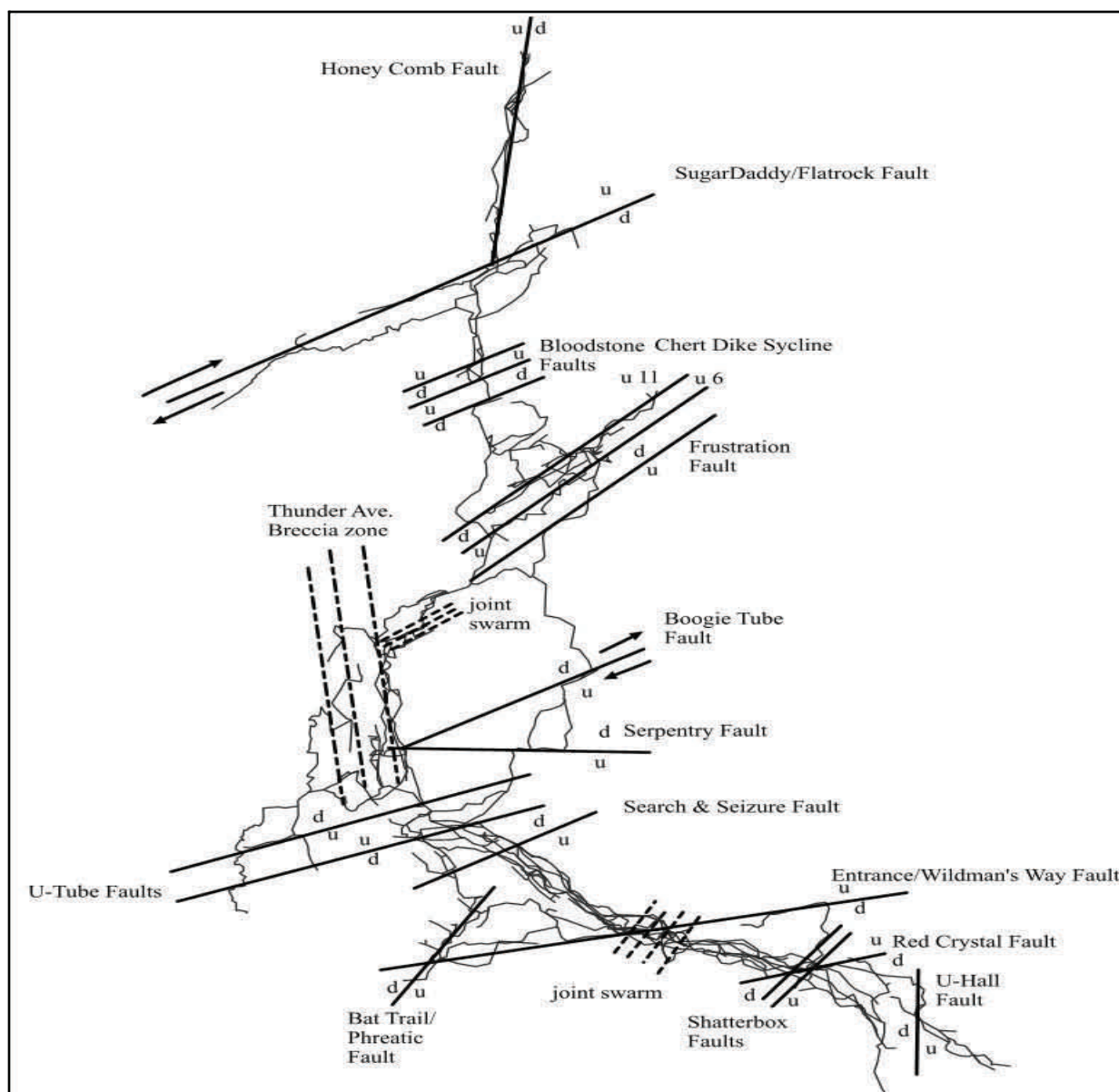


Figure 6. Plan view of Chilly Bowl Cave illustrating the tectonic control and faulting that are present and mapped within the cave. Relative motion across faults is u, upthrown; d, downthrown (after Tennyson et al., 2008).

instances, dead cavefish were observed where spills occurred yet no caves were present, nor were any cavefish ever previously observed (Carol Wicks, 2001, written communication, Columbia, Missouri).

Cave mapping (Figure 6) indicates that passages in Chilly Bowl Cave appear to have been formed by deep flow from below along integrated systematic joints and faults (Figures 6 and 10), driven by hydrostatic pressure. Erosional features and deposits observed indicate past episodes of thermal ground-water circulation, with directional scallops in conduits and flow tubes showing upward flow. Cave mapping (Jim

Terry, 2008, written communication, Rogers, Arkansas; Rodney Tennyson, 2008, written communication, Carrolton, Arkansas) also indicates that passages formed by hypogene processes likely have been utilized by later epigenetic processes, with a reversal of flow so that presently flow in the upper portions of the cave is downward.

Additional documentation of thermal hypogene flow includes upper-level paleo-cave passages and chambers, now reactivated by present-day ground-water drainage, dolomite breccia linings on cave walls, and remarkable, beautifully preserved, large crystals of

dog-tooth spar (Figures 7 and 8). These crystals occur in the highest level of the cave, protected from dissolution and erosion from above by a sandstone caprock. Present-day recharge from the surface has been diverted to lower levels in the cave by older hypogene features, bypassing those few remaining passages where the crystals have been spared. Uranium-series radionuclide dating of these crystals is planned but has not been completed.

Stable-isotopic analyses of samples taken from a transect of one of the dog-tooth spar crystals from Chilly Bowl Cave shows a crystal section (samples 6-8 in Figure 9 and Table 1) that exhibit anomalously light $\delta^{13}\text{C}$, consistent with hypogenic calcite derived from the oxidation of methane to CO_2 . Most other samples fall on a simple mixing line near the margins of thermal



Figure 7. An upper-level passage in Chilly Bowl Cave, uniquely preserved with large dogtooth spar crystals that are thought to have been precipitated by thermal brines from hypogene sources. The caver's head serves as scale.



Figure 8. Backlit dogtooth spar crystal from Chilly Bowl Cave showing incipient cleavage planes. A broken sample from an in situ crystal was used for stable isotope analysis (see Figure 9 for sampling locations, and Table 1 for isotopic values for each location).

carbonate minerals (Palmer, 2007). Samples 6-8 reveal an anomalous $\delta^{13}\text{C}$ signature (Figure 10 and Table 1) in that $\delta^{13}\text{C}$ is about -20 ‰, unlike other calcite samples from the region (P.D. Hays, 2008, written communication, Fayetteville, Arkansas). Although speculative at this time, one hypothesized source is the organic-rich, Arkoma hydrocarbon basin that lies on the flow path from the Ouachita Mountains in the south to the study area in north Arkansas (Figure 1).

Economic minerals (sphalerite, galena, and chalcopryrite) are present in the Rush mining district (McKnight, 1935; Smith, 1973), north and east of the study area about 30 km; the same mineralogy is disseminated throughout much of the area in noncommercial quantities. Although these are not directly related to the karst focus of this report, the source of the metals and the details of the hydrogeology associated with mineral emplacement are hypothesized to be related to the hypogenic fluids responsible for early karst initiation along Flatrock Creek Fault.

Fluid-inclusion research (Leach, 1979; Leach and Rowan, 1986) provides additional insight into the hot temperatures of mineralizing fluids that were hypothesized to have been expelled northward from the Arkoma Basin through permeable Paleozoic carbonate rocks toward the Ozark uplift. Using homogenization temperatures of fluid inclusions in sphalerite, these authors showed a northward cooling of deep basinal brines that deposited sphalerite. Temperatures ranged

Table 1. Stable isotopes of carbon-13 ($\delta^{13}\text{C}$) and oxygen-18 ($\delta^{18}\text{O}$) for each of the data points measured on the cleavage face (labeled A) and outside of the spar crystal (labeled B) from Chilly Bowl Cave. Sampling locations on the cleavage face are keyed to approximate locations in Figure 9. Anomalous samples are shaded. DUP reflects separate, duplicate samples from the same location in the crystal, with RERUN reflecting lab analyses rerun on a single sample. NS indicates no sample.

Sample ID (QA/QC)	$\delta^{13}\text{C}$ (in VPDB)	$\delta^{18}\text{O}$ (in VPDB)
A-1	0.83	-8.09
A-2	1.03	-8.76
A-3	-0.53	-8.71
A-4	-1.59	-8.50
A-4 (RERUN)	-1.75	-8.51
A-4 (DUP)	-1.22	-8.39
A-5	0.03	-9.83
A-6	-20.28	-9.97
A-7	-19.02	-9.84
A-8	-18.83	-9.86
A-8 (DUP)	-18.74	-9.59
A-9	-4.48	-9.71
A-10	-0.09	-9.38
A-11	-1.79	-8.84
A-12 (DUP)	-6.71	-6.79
A-12 (RERUN)	-6.88	-6.70
A-13	-4.07	-8.27
A-14	-4.28	-8.65
A-15	-4.19	-8.40
A-15 (DUP)	-4.03	-8.16
A-16	NS	NS
A-15 (RERUN)	-4.16	-8.32
A-17	-5.29	-9.25
A-18	-5.37	-9.76
B-1	-0.29	-7.98
B-2	-6.35	-7.59
B-2 (RERUN)	-6.27	-7.40
B-2 (DUP)	-6.18	-7.29
B-3	-5.29	-7.31
B-4	-4.39	-7.72
B-5	-5.24	-6.99
B-5 (DUP)	-5.25	-6.97

from $\sim 120^\circ\text{C}$ at ~ 100 kilometers north of the Ouachita Mountains to $\sim 80^\circ\text{C}$ in deposits ~ 400 kilometers north of the Ouachitas. At about 170 km, the approximate location of this study area, the temperature is interpolated to be greater than 100°C . The hydrogeology associated with these deposits in the mining areas appears to be quite similar to the study area.

Summary

The compilation briefly discussed herein documents a general conceptual model of hypogene speleogenesis (Figure 2), and relates it to a specific north-south section of Chilly Bowl Cave along the Flatrock Creek

fault (Figure 11 and Table 2) that is consistent with all available data from a wide range of disciplines. Simply, the Ouachita orogeny was the driving force for differential uplift of basement rocks along pre-existing northeast-trending faults, for faulting and tilting of the overlying sedimentary cover, and for fluid migration from the south; basinal sediments of the Arkoma Basin were the source of organics, metals, and aggressive solutions. Faulting created preferred paths for fluid flow; the geothermal gradient supplied the heat, and the shale and sandstone cover of terrigenous sedimentary rocks supplied the insulation and pressure confinement such that the hot water was advectively transported, slowly cooled, and retained an anomalously warm signature far north into the Ozarks. The strength of the crystalline basement rocks allowed much of the energy transmission, acting as a broad, thick lever to facilitate uplift. Uplift was responsible for fracturing the relatively thin overlying sedimentary cover in a systematic stress field of orthogonal jointed and faulted rocks. This overlying cover was karstified by multiple pulses of hot water, especially in areas overlying the deep faults.

Epigenetic karstification was later overprinted on the hypogenic karst, and in many cases, meteoric recharge

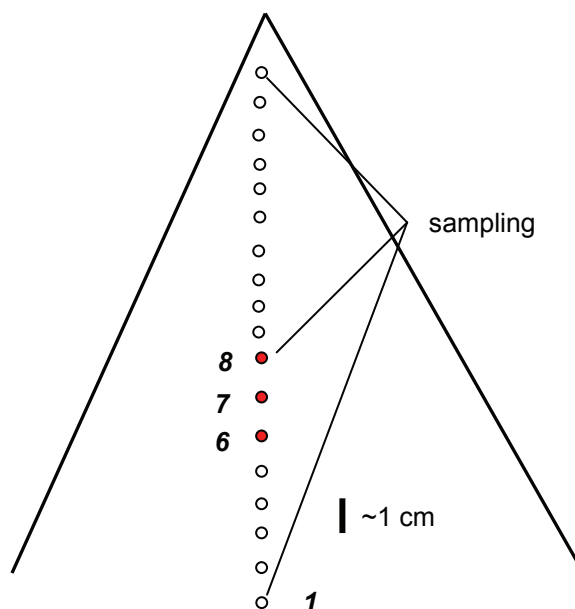


Figure 9. Sketch showing approximate locations of sampling sites for stable isotope analyses on a cleavage face of a spar crystal similar to that shown in Figure 7. Samples are numbered from 1 at the bottom and increase sequentially upward. Samples 5 through 8, indicated by solid red points, are anomalously light in $\delta^{13}\text{C}$, consistent with hypogenic calcite derived from the oxidation of methane to CO_2 .

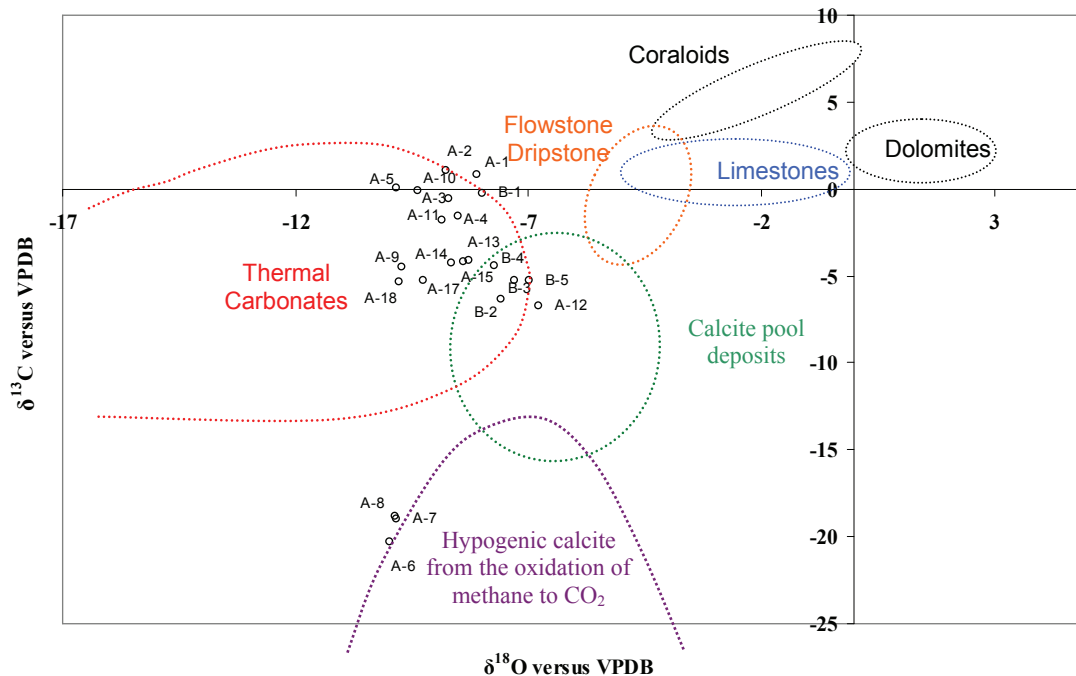


Figure 10. Plot showing the relation of $\delta^{13}\text{C}$ versus $\delta^{18}\text{O}$ (per mil) superimposed on the range of ratios expected from different environments of formation (modified from Palmer, 2007). Numbers are from sampling points on the spar crystal, shown in Figure 9, and reported in Table 1. Samples 5 through 8 are anomalously light in $\delta^{13}\text{C}$, consistent with hypogenic calcite derived from the oxidation of methane to CO_2 . Most other samples fall on a simple mixing line near the margins of thermal carbonate minerals.

extensively modified original hypogenic karst as local flow conditions responded to capture of surface streams. After the pulses of uplift from the Ouachita orogeny ceased, base level was successively lowered over a long interval of geologic time, estimated to be hundreds of millions of years. Several ancient segments of the original hypogenic caves have been preserved, owing to the protection by the sandstone

and chert cover layers, and the diversion of epigenic recharge to lower components in the flow system that fortuitously have kept flow away from fragile and reactive formations. Geologically recent karst activity, mostly Pleistocene in age, represents the final stage of cave evolution, and includes precipitated and evaporative speleothems that were formed in the vadose zone.

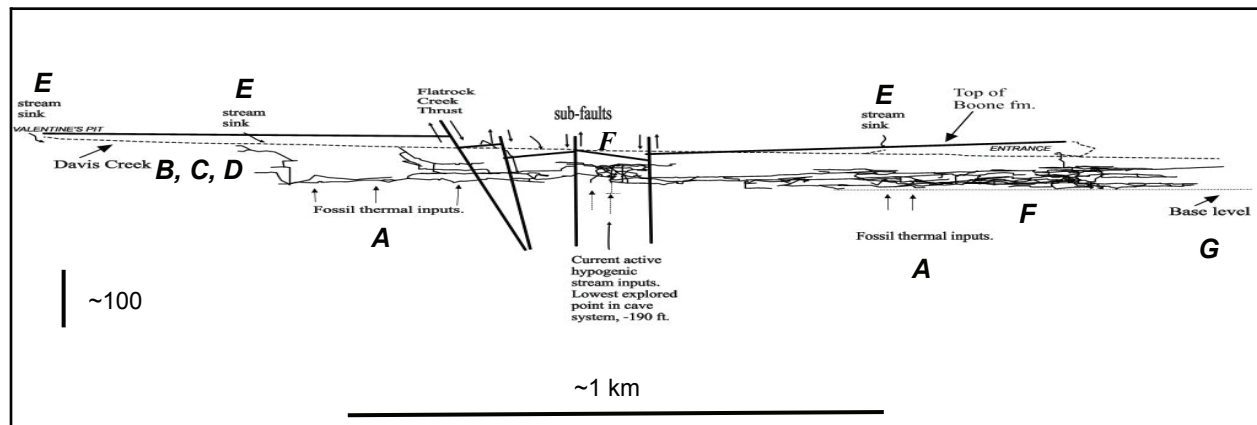


Figure 11. North-south section of Chilly Bowl Cave illustrating the vertical speleogenetic development, major faulting associated with the Flatrock Creek fault, current level of Davis Creek, current base level, the top of the Boone Formation, and sources of thermal input that are preserved within the cave. Letters on the figure show the relative sequence of events keyed to documentation provided in text and summarized in Table 2.

Table 2. Features of the conceptual model, including interpretation relevance, and timing of key events that affected speleogenesis of Chilly Bowl Cave. Attribute locations within cave are approximated on Figure 11.

Attribute ID	Feature/Evidence	Interpretation	Relative Timing	Comments
A	near-vertical shafts cutting across bedding and chert layers; scallops indicate upward flow	upward, cross formational flow, along sections of fractures	likely old, formed by vertically upward flow in confined conduits	strong evidence of hypogenic flow; scallops indicate flow direction from below.
B	amber dogtooth spar crystals	precipitated from hot water expelled from Ouachita Basin along major basement faults.	very old, although younger than attribute A.	likely part of paragenetic sequence associated with lead and zinc deposits
C	stable isotopes—particularly carbon-13	carbon-13 data in core of measured crystal consistent with methane source	samples 6-8 (Table 1) are likely on the order of more than 300 M years, consistent with olistoliths related to basement faulting (Chandler, 2000)	carbon-13 data and methane source are consistent with flow from an organic-rich environment, such as Arkoma Basin
D	radionuclide dating of spar	uranium-lead dating feasible if background lead concentration is low	these analyses would provide strong absolute age date for spar, if samples have not been contaminated with lead	radionuclide dating is planned, but not yet completed
E	surface erosion, land-surface lowering, and capture of meteoric surface water	karst drainage of surface water was opportunistic, and followed flow paths of least resistance	stable isotopes are consistent with mixing of hypogenic and meteoric waters; this mixing was present in the outer margins of the spar crystal, so this represents a significant interval of time	multiple periods of stream capture appear to have overprinted earlier zones of focused flow from hypogenic sources
F	cavern morphology (Figures 6 and 11) and erosional features	typical morphology in plan and section for hypogenic cave (Klimchouk, 2007)	developed and evolved over long period of time; overprinting of old hypogenic features by younger epigenic features dominates cave, but is far from complete	strongly influenced by initial hypogenic development
G	fluid inclusions	homogenization temperatures of nearby sphalerite deposits indicate $T > 100^{\circ}\text{C}$ (Leach, 1979); areal distribution shows cooling northward	fluid expulsion from the south during Ouachita orogeny	thermal water from a non-meteoric source is supported by these data and are consistent with flow and geochemical signatures

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References

Braden, A.K., and S.M. Ausbrooks. 2003. *Geologic map of the Mt. Judea quadrangle, Newton County, Arkansas*. Arkansas Geological Commis-

- sion Digital Geologic Quadrangle Map DGM-AR-00590, scale 1:24,000. Little Rock: Arkansas Geological Commission.
- Brahana, J.V. 1997. Rationale and methodology for approximating spring-basin boundaries in the mantled karst terrain of the Springfield Plateau, northwestern Arkansas. In Sixth multidisciplinary conference on engineering geology and hydrogeology of karst terranes, ed. B. F. Beck, and J. B. Stephenson, 77-82. Rotterdam: A.A. Balkema.
- Brahana, J.V., J. Thrailkill, T. Freeman, and W.C. Ward. 1988. Carbonate rocks. In *Hydrogeology: The Geology of North America*, v. O-2, ed. W. Back, J. S. Rosenshein and P. R. Seaber, 333-352. Boulder: Geological Society of America.
- Brahana, J.V., P.D. Hays, M. Al-Qinna, J.F. Murdoch, R.K. Davis, J.J. Killingbeck, E. Szilvagy, M. Doheny-Skubic, I. Chaubey, T.E. Ting, and G. Thoma. 2005. Quantification of hydrologic budget parameters for the vadose zone and epikarst in mantled karst. In U.S. Geological Survey Karst Interest Group proceedings, Rapid City, South Dakota, September 12-15, 2005: U.S. Geological Survey Scientific Investigations Report 2005-5160, ed. E. L. Kuniansky, 144-152. Boulder: U.S. Geological Survey.
- Chandler, S.L. 2001. Carbonate olistoliths, St. Joe and Boone Limestones (Lower Mississippian), northwestern Arkansas. M.S. thesis, University of Arkansas.
- Fanning, B.J. 1994. Geospeleologic analysis of cave and karst development within the Boone and St. Joe Formations of Benton and Madison counties, Northwest Arkansas. M.S. thesis, University of Arkansas.
- Fischer, M.P., and R.D. Christensen. 2004. Insights into the growth of basement uplifts deduced from a study of fracture systems in the San Rafael monocline, east central Utah. *Tectonics* 23: TC1018.
- Fischer, M.P., and P.B. Jackson. 1999. Stratigraphic controls on deformation patterns in fault-related folds: A detachment fold example from the Sierra Madre Oriental, northeast Mexico. *Journal of Structural Geology* 21: 613-633, doi: 10.1016/S0191-8141(99)00044-9.
- Ford, D.C. 2006. Karst geomorphology, caves and cave deposits: A review of North American contributions during the past half century. In *Perspectives on karst geomorphology, hydrology, and geochemistry: Geological Society of America Special Paper 404*, ed. R.S. Harmon and C.W. Wicks, 1-14. Boulder: Geological Society of America.
- Granath, J.W. 1989. Structural evolution of the Ardmore Basin, Oklahoma: progressive deformation in the foreland of the Ouachita collision. *Tectonics* 8: 1015-1036.
- Guccione, M.J. 1993. *Geologic history of Arkansas through time and space*. National Science Foundation unnumbered publication, Grant No. ESI-8855588. Fayetteville: University of Arkansas.
- Hale-Erlich, W.S., and J.L. Coleman, Jr. 1993. Ouachita-Appalachian juncture: A Paleozoic transpressional zone in the southeastern U.S.A. *American Association of Petroleum Geologists Bulletin* 77: 553-568.
- Haley, B.R., E.E. Glick, W.V. Bush, B.F. Clardy, C.G. Stone, M.B. Woodward, and D.L. Zachry. 1993. *Geologic map of Arkansas: scale 1:500,000*. Arkansas Geological Commission and U.S. Geological Survey.
- Hendricks, J.D., G.R. Keller, and T.G. Hildenbrand. 1981. *Bouguer gravity map of Arkansas*. U.S. Geological Survey Geophysical Investigations Map GP-944, scale 1:500,000. Boulder: U.S. Geological Survey.
- Hill, C.A. 2000. Sulfuric acid hypogene karst in the Guadalupe Mountains of New Mexico and West Texas. In *Speleogenesis: evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer and W. Dreybrodt, 309-316. Huntsville: National Speleological Society.
- Houseknecht, D.W., and S.M. Matthews. 1985. Thermal maturity of Carboniferous strata, Ouachita Mountains. *American Association of Petroleum Geologists Bulletin* 69: 335-345.
- Hudson, M.R. 1998. *Geologic map of parts of the Jasper, Hasty, Ponca, Gaither, and Harrison quadrangles in and adjacent to the Buffalo National River, northwestern Arkansas*. U.S. Geological Survey Open-File Report 98-116, scale 1:24,000. Boulder: U.S. Geological Survey.
- Hudson, M.R. 2001. Coordinated strike-slip and normal faulting in the southern Ozark dome of northern Arkansas—Deformation in a late Paleozoic foreland. *Geology* 28: 511-514.
- Hudson, M.R., and K.E. Murray. 2004. *Geologic map of the Hasty Quadrangle, Boone and Newton Counties, Arkansas*. U.S. Geological Survey Scientific Investigations Map 2847, Version 1.0, scale 1:24,000. Boulder: U.S. Geological Survey.
- Hudson, M.R., K.E. Murray, and D. Pezzutti. 2001. *Geologic map of the Jasper quadrangle, Newton and Boone Counties, Arkansas*. U.S. Geological Survey Miscellaneous Field Studies Map MF-2356, scale 1:24,000. Boulder: U.S. Geological Survey.

- Imes, J.L., and L.F. Emmett. 1994. *Geohydrology of the Ozark Plateau aquifer system in parts of Missouri, Arkansas, Oklahoma, and Kansas*. U.S. Geological Survey Professional Paper 1414-D. Boulder: U.S. Geological Survey.
- Johnson, T.C. 2008. Geologic map of Forum Quadrangle, with a karst inventory, Madison County, Arkansas. M.S. thesis, University of Arkansas.
- Klimchouk, A. 2007. *Hypogene speleogenesis: hydrogeological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Klimchouk, A., D.C. Ford, A.N. Palmer, and W. Dreybrodt. 2000. *Speleogenesis: evolution of karst aquifers*. Huntsville: National Speleological Society.
- Kranjc, A., R. Gabrovsek, D.C. Culver, and I.D. Sasowsky. 2007. *Time in karst*. Karst Waters Institute Special Publication 12. Charles Town: Karst Waters Institute.
- Leach, D.L. 1979. Temperature and salinity of the fluids responsible for minor occurrences of sphalerite in the Ozark region of Missouri. *Economic Geology* 74: 931-937.
- Leach, D.L., and E.L. Rowan. 1986. Genetic link between Ouachita foldbelt tectonism and the Mississippi Valley-type lead-zinc deposits of the Ozarks. *Geology* 14: 931-935.
- McKnight, E.T. 1935. *Zinc and lead deposits of northern Arkansas*. U.S. Geological Survey Bulletin 853. Boulder: U.S. Geological Survey.
- Meckel, L.D., D.G. Smith, and L.A. Wells. 1992. Ouachita foredeep basins: Regional paleogeography and habitat of hydrocarbons. In *Foreland basins and fold belts: American Association of Petroleum Geologists Memoir 5*, ed. R. W. Macqueen and D. A. Leckie, 427-444. Tulsa: American Association of Petroleum Geologists.
- Palmer, A.N. 2007. *Cave Geology*. Dayton: Cave Books.
- Smith, D.A. 1977. Lead-zinc mineralization in the Ponca-Boxley area, Arkansas. M.S. thesis, University of Arkansas.
- Tennyson, R., J. Terry, V. Brahana, P. Hays, and E. Pollock. 2008. Tectonic control of hypogene speleogenesis in the southern Ozarks—Implications for NAWQA and beyond. In *U.S. Geological Survey Karst Interest Group proceedings, Bowling Green, Kentucky: U.S. Geological Survey Scientific Investigations Report 2008-5023*, ed. E.L. Kuniansky. Boulder: U.S. Geological Survey.
- Viele, G.W., and W.A. Thomas. 1989. Tectonic synthesis of the Ouachita orogenic belt. In *The Appalachian-Ouachita orogen in the United States: The Geology of North America*, v. F-2, ed. R.D. Hatcher, Jr., W.A. Thomas and G.W. Viele, 695-728. Boulder: Geological Society of America.
- Withjack, M.O., and W.R. Jamison. 1986. Deformation produced by oblique rifting. *Tectonophysics* 126: 99-124.

THE RELATIONSHIP OF OIL FIELD-DERIVED HYDROGEN SULFIDE IN THE PERMIAN (GUADALUPIAN) ARTESIA GROUP TO SULFURIC ACID SPELEOGENESIS IN THE GUADALUPE MOUNTAINS, NEW MEXICO AND TEXAS, USA

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Abstract

Hydrogen sulfide (H_2S) is common in oil, gas and connate water associated with the Capitan aquifer and equivalent strata of the Artesia Group (Permian, Guadalupian) in southeast New Mexico, USA, east of the Pecos River. This H_2S is a byproduct of the microbial oxidation of hydrocarbons and reduction of sulfate in Artesia Group oil reservoirs. Most of the reservoir facies of the Artesia Group crop out on the north side of the Guadalupe Mountains, where oil fields in Guadalupian strata are absent, except near the Pecos River. In this area, H_2S is found in connate water and native sulfur occurs in Artesia Group strata. Fluid inclusions containing hydrocarbons in late stage calcite cements on the north side of the Guadalupe Mountains, and in speleogenetic sulfur in Lechuguilla Cave indicate the former presence of oil and gas accumulations in Artesia Group reservoirs west of the Pecos River. Hydrogen sulfide derived from the microbial alteration of crude oil in these reservoirs is the most likely source of the sulfuric acid that drove Late Tertiary speleogenesis and formed large passages and galleries in caves of the Guadalupe Mountains.

Introduction

The Guadalupe Mountains of New Mexico and Texas, USA (Figure 1), are renowned for the caves they contain, including Carlsbad Cavern and Lechuguilla Cave. These caves are a result of the dissolution of limestone and dolomite by sulfuric acid (Hill, 1987). This process occurred within the Capitan aquifer where meteoric water mixed with connate water rich in hydrogen sulfide to form sulfuric acid (Hill, 1987; DuChene and McLean, 1989; DuChene and Cunningham, 2006). The volume and geometry of the caves is a consequence of sulfuric acid dissolution at or near the water table during the last 14-12 Ma (Polyak et al., 1998), where dissolution of carbonate modified and enlarged early stage vertical caves that formed along faults and fractures (Palmer, 2000; DuChene

and Cunningham, 2006). For the sulfuric acid reaction to occur, two critical ingredients are required; hydrogen sulfide (H_2S) and oxygen (O_2). Hydrogen sulfide was formed in the subsurface by the biogenic alteration of hydrocarbon and sulfate, which is a hypogenic process. Since the reaction occurred at or near the water table, O_2 was most likely supplied from the atmosphere in air-filled porosity in the vadose zone and by oxygenated meteoric water seeping downward from the surface of the earth, which is an epigenic process. Sulfuric acid dissolution of limestone in the Guadalupe Mountains is therefore a combination of hypogenic and epigenic processes. The presence of caves in the Guadalupe Mountains that contain abundant mineralogical and speleogenetic evidence of the sulfuric acid reaction proves that hydrogen sulfide must have been present. Polyak et al., (1998) show that sulfuric acid speleogenesis in the Guadalupe Mountains ended 3.8 Ma. Therefore, conditions that supported dissolution of limestone and dolomite by sulfuric acid in the Guadalupe Mountains were differ-

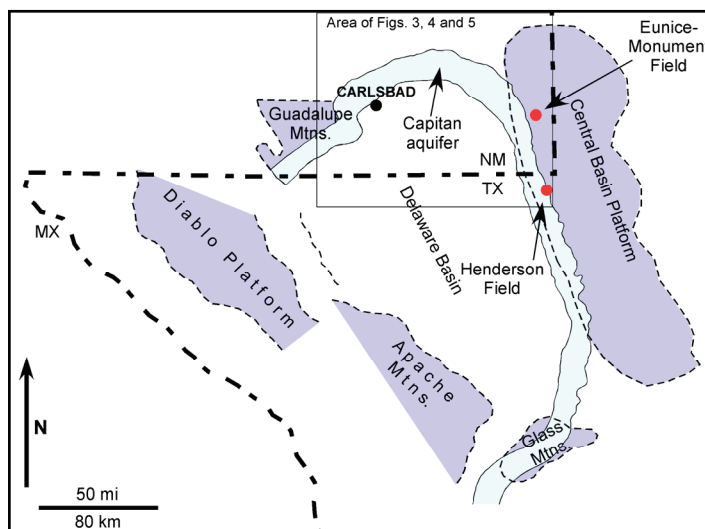


Figure 1. Index map showing the location of the Delaware basin and surrounding features, including the Guadalupe Mountains and Capitan aquifer.

ent during Late Tertiary speleogenesis than they are today.

In her 1987 treatise on speleogenesis in the Guadalupe Mountains, Carol Hill presented a convincing array of data in support of a sulfuric acid origin for the caves. She proposed that the source of H_2S was in the Delaware basin, south of the Guadalupe Mountains. In her model, H_2S was generated at the contact between the Bell Canyon Formation and the Castile Formation, where hydrocarbons were altered by anaerobic bacteria. In this process, bacteria oxidized hydrocarbons and reduced sulfate, converting anhydrite to calcite and generating H_2S . Hydrogen sulfide generated by this process migrated up-dip into the Capitan Limestone and associated backreef carbonates to mixing zones where it was converted to sulfuric acid (Hill, 1987, her Figure 87).

The purposes of this paper are: 1) to show that H_2S is a common constituent of connate water in Permian (Guadalupian) strata on the northwest shelf of the Delaware basin, and 2) to demonstrate that H_2S on the shelf north of the Capitan aquifer was a major contributor of the sulfuric acid that enlarged Guadalupe Mountain caves.

Definition of the Capitan aquifer and the Artesia Group

Hiss (1980) defined the Capitan aquifer as a lithosome that includes the Capitan and Goat Seep formations and the Carlsbad facies of the Artesia Group, as defined by Meissner (1972) (Figure 2). The Artesia Group is comprised of the Grayburg, Queen, Seven Rivers, Yates and Tansill formations in ascending order. The Carlsbad facies of the Artesia Group

consists of near shelf edge dolomites and thinner, intercalated sandstones that grade into or pinch out into reef strata of the Capitan and Goat Seep formations (Meissner, 1972, p. 206). Artesia Group formations grade from a predominantly carbonate facies with some clastics near the shelf margin to red shale, sand and anhydrite toward the shelf.

The Capitan aquifer nearly encircles the Delaware basin, and is exposed in the Guadalupe, Apache, and Glass Mountains (Figure 1). The portion of the Capitan aquifer pertinent to this discussion lies in southeastern New Mexico where it is buried beneath Ochoan and younger sediments east of the Pecos River, and is exposed to the west in the Guadalupe Mountains.

The Capitan aquifer is now and has been a primary conduit for transport of meteoric water into the subsurface of the Permian basin (Hiss, 1980). East-directed hydrodynamic flow was strongest in early- to mid- Tertiary time, prior to faulting associated with the opening of the Rio Grande Rift (Lindsay, 1998). Hydrodynamic flow today is diminished due to the reduction in hydrostatic head caused by the opening of the Rio Grande rift (Lindsay, 1998; DuChene and Cunningham, 2006) and breaching of the Capitan aquifer by the Pecos River (Hiss, 1980, p. 294).

Relationship of hydrocarbons to the Capitan aquifer

Figures 3, 4 and 5 are maps showing the distribution and characteristics of fluids within the Capitan aquifer and laterally equivalent strata. Figure 3 shows the distribution of H_2S in oil trapped in reservoirs of the Artesia Group. Figure 4 shows the concentration of chloride ions, reflecting the salinity of connate water. Figure 5 is a map of the API gravity of oils recovered from traps in strata associated with the Capitan aquifer. Data for Figures 3 and 5 were extracted from symposia published by the Roswell Geological Society (1956, 1960). These data are incomplete because not all of the oil fields in southeastern New Mexico are included in the symposia, and many of the individual field reports do not provide information on H_2S or sulfur. The chloride concentration map (Figure 4) was adapted from Hiss (1980).

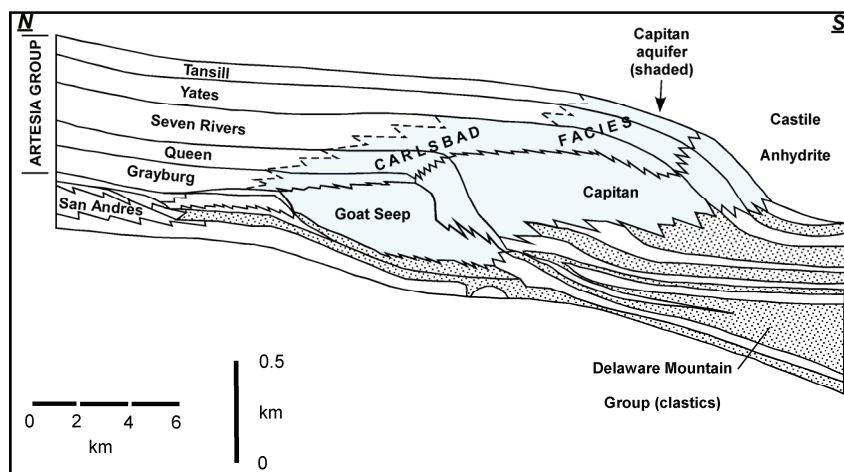


Figure 2. Diagrammatic cross section through the Capitan shelf margin showing the Capitan aquifer (shaded), which is comprised of the Goat Seep and Capitan formations and the Carlsbad facies of the Artesia Group.

Distribution of sulfur and H₂S

Figure 3 shows the Capitan aquifer and the location of major oil fields in the Artesia Group in southeastern New Mexico. Much of the oil production from the Delaware basin east of the Pecos River is from Permian (Guadalupian) reservoirs in the Artesia Group along an arcuate trend north of the Capitan shelf margin. Ward et al. (1986) reported that 7.4 billion barrels of oil have been produced from Artesia Group reservoirs. Much of the oil produced from these reservoirs is stratigraphically trapped and contains sulfur and/or H₂S. Stratigraphic traps in the Artesia Group are formed where porous and permeable carbonates grade up-dip into non-porous siltstone, shale and anhydrite.

Several oil fields where sulfur and/or H₂S have been reported are highlighted along the Artesia Group oil-producing trend (Figure 3). The percentage of H₂S in each field is plotted on the map and summarized in Table 1. It is likely that most, if not all, of the oil and gas produced from Artesia Group reservoirs is sour but information regarding the sulfur content has not been published for all wells. The western limit of production from Artesia Group reservoirs is located near the Pecos River between the towns of Carlsbad and Artesia, New Mexico. West of the Pecos, many of the oil-productive Artesia Group formations crop out

on the north flank of the Guadalupe Mountains (Figure 1). Although there are no Artesia Group oil fields on the north flank of the Guadalupe Mountains today, there is evidence that oil was once trapped in Artesia Group reservoirs. Scholle, et al. (1992) found fluorescent oil inclusions in Artesia Group rocks in backreef facies near the Capitan shelf margin in the Guadalupe Mountains. Spirakis and Cunningham (1992) reported light chain aliphatic hydrocarbons from fluid inclusions in sulfur from Lechuguilla Cave. Porous limestone samples collected by the author from Grayburg outcrops on the western side of the Guadalupe Mountains have a strong petroleum odor when freshly broken.

East of the Pecos River and north of the Capitan shelf margin there are several small oil fields that lie within the boundaries of the Capitan aquifer (Figure 3). Most of these fields produce from the Yates Formation and have oil with low API gravity (Figure 5). Formation water containing H₂S is produced with oil from Capitan aquifer and Artesia Group reservoirs extending from the Pecos River to Henderson field in Texas and beyond (Table 1).

The origin of sulfur in Castile Buttes of the Delaware basin has been attributed to the biogenic oxidation of oil and associated reduction of sulfate (Kirkland and

Figure 3. Distribution of sulfur and H₂S in the Artesia Group of southeastern New Mexico. The Capitan aquifer is shown in blue. Major oil fields that produce from Artesia Group reservoirs are shaded yellow. Oil fields where information on H₂S content of oil, gas or water has been reported are shaded green. Squares symbols are drill holes where native sulfur has been reported. Triangles note occurrences of speleogenetic sulfur in caves.

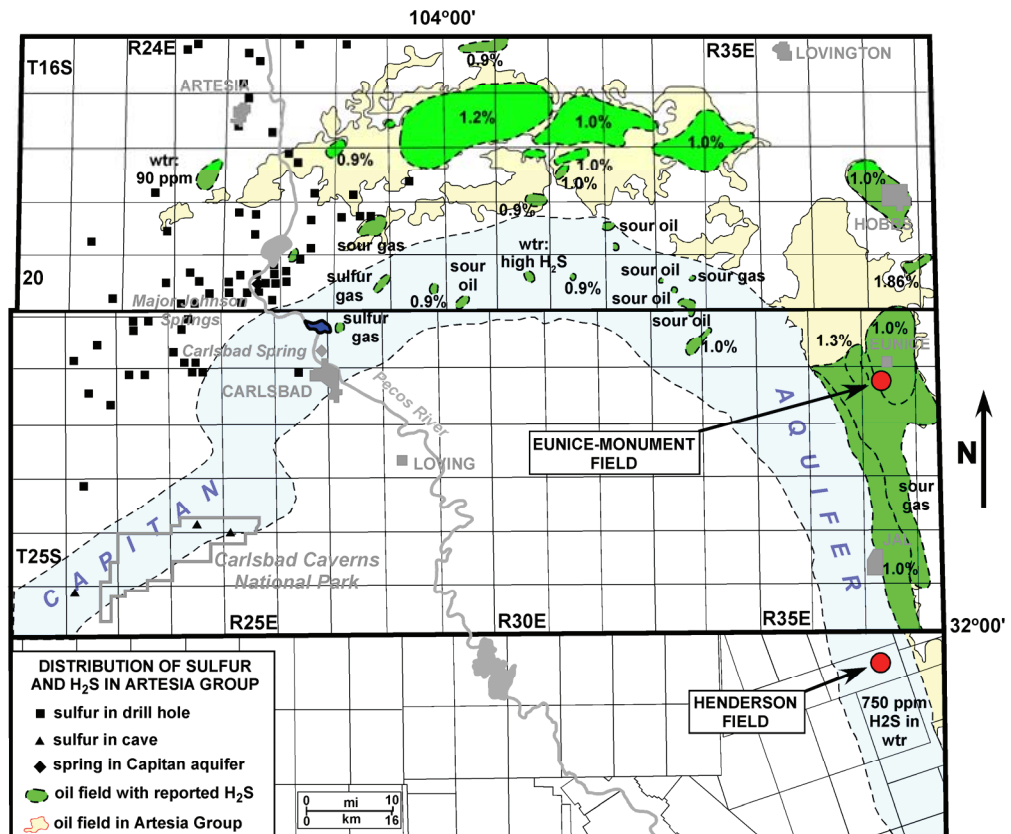


Table 1. Oil and gas fields in Artesia Group reservoirs with reported H₂S.

Field	Formation	Location	Depth to producing zone		API gravity	H ₂ S content	Comments: fluids	Reference
			ft	M				
Aid	Yates	T19S, R28-29E	792	241	39.7°		sour oil	E.E. Kinney 1956a
Arrowhead	Queen-Grayburg	T21-22S, R36-37E	3560	1085	34°	1.3%	8 ppm H ₂ S in water	Roswell Geol. Soc. Sym. Comm. 19656a
Atoka	San Andres	T18S, R26E	1520	463	37°		90 ppm H ₂ S in water	Roswell Geol. Soc. Sym. Comm. 19656b
Cedar Hills	Yates	T21S, R27E	550	168	21°		low btu sulfur gas	H.F. Schram 1956a
Dos Hermanos	Yates	T20S, R30E	1650	503	27°		sour crude	H.W. Puckett 1960
Gem	Yates	T19S, R33E	3240	988	31°		590 ppm H ₂ S in water	D. Wilson 1956
Getty	Yates	T20S R29E	1320	402	19°-24°	0.9%	sour, 550btu	Roswell Geol. Soc. Sym. Comm. 19656c
Grayburg-Jackson	Grayburg-San Andres	T17S, R29-31E	2600	792	36°	1.2%		A.A. Fryman 1956
Halfway	Yates	T20S, R32E	2496	761	26°	9.0%		H.N. Sweeney 1956a
Henderson	Yates-Seven Rivers	Block 26, PSL Survey	3017	920			low btu gas, high nitrogen	Wiggins et al. 1993
Hobbs	Grayburg-San Andres	T18-19S, R37-38E	3700	1128	34°	1.0%	750 ppm of H ₂ S in formation water	K.D. McPeters and J. M. Kelly 1956
House	San Andres	T20S, R38-39E	4365	1330	34°	1.9%		M.L. Zoller 1956
Jalmat	Yates-Seven Rivers	T21-26S, R35-37E	2800	853	29°		sweet and sour gas	Roswell Geol. Soc. Sym. Comm. 19656d
Lanlie-Mattix	Yates-Seven Rivers	T22-26S, R36-38E	2500	762	37°		sour gas in south part of field	B. Stringer 1956
Littman	San Andres	T21S, R38E	4350	1326	31°		trace of H ₂ S in formation water	H.N. Sweeney 1956b
McMillan	Seven Rivers-Queen	T19S, R27E	550	168	29°		slightly sour	D. Fryman 1956
Millman East	Queen-Grayburg	T19S, R28E	1700	518	38°		sour gas	P.M. Mershon Jr. 1960
P.C.A.	Yates	T20S, R30E	1450	442	20°		water with high H ₂ S content	Roswell Geol. Soc. Sym. Comm. 19656e
Penrose-Skelly	Queen-Grayburg	T21-22S, R36-37E	3435	1047	36°	1.0%		W.R. Green 1956
Red Lake	Grayburg-San Andres	T17-18S, R27-28E	1780	543	29.2°	0.9%		Roswell Geol. Soc. Sym. Comm. 19656f
Rhodes	Yates-Seven Rivers	T26S, R37E	3270	997	36°	1.0%		Roswell Geol. Soc. Sym. Comm. 19656g
Robinson	Grayburg-San Andres	T16-17S, R31-32E	3550	1082	34.5°	0.9%		E.S. Dietrich 1956
Russell	Yates	T20S, R28W	800	244	38.2°	0.9%	123 ppm H ₂ S in formation water	H.F. Schram 1956b
Square Lake	Grayburg-San Andres	T16-17S, R30-31E	3325	1013	38.7°	0.9%		J.M. Goodger 1956
Tonto	Yates	T19S, R33E	3465	1056	33°		sour oil	E.E. Kinney 1956b
Vacuum	Grayburg-San Andres	T17-18S, R33-35E	4680	1426	36°	1.0%		R. Milks 1956
Young	Queen	T18S, R32E	3725	1135	37°	1.0%		H.N. Sweeney 1956c

Evans, 1976; Hill, 1987). A similar origin is documented for late-stage calcite cement and H_2S in the Henderson (Yates and Seven Rivers) field, located in Winkler County, Texas near the southeastern corner of New Mexico (Figure 3) (Wiggins et al., 1993). Although there are few Artesia Group oil fields west of the Pecos River, there is evidence of sulfur and H_2S in the area. Hinds and Cunningham (1970, p. 4) state that "many of the early drillers in Eddy County reported "sulphur water" and "black sulphur rock" in numerous wells. Water saturated with hydrogen sulfide is common in wells throughout much of the county." On Figure 3, core and drill holes where sulfur was encountered are plotted as black squares. Hinds and Cunningham (1970, p. 9) attributed these sulfur occurrences to "...the alteration of anhydrite to carbonate and sulfur by the metabolic processes of bacteria in the presence of hydrocarbons." Sulfur has also been found in Carlsbad Cavern (Hill, 1987), Cottonwood Cave (Davis, 1973) and Lechuguilla Cave (Cunningham et al., 1993), where it is interpreted to be a byproduct of the dissolution of limestone by sulfuric acid (Hill, 1987). It is likely that oil and gas containing sulfur and H_2S were present in traps on the north side of the Guadalupe Mountains prior to uplift of the Alvarado Ridge and opening of the Rio Grande rift (Scholle et al., 1992; DuChene and Cunningham, 2006).

Distribution of chlorides in the Capitan aquifer and age-equivalent strata

Figure 4 shows the Capitan aquifer and major Artesia Group oil fields. Contour lines show the chloride ion concentration in mg/liter (modified from Hiss, 1975; 1980). The API gravity of the Artesia Group oil is also shown. Within the Capitan aquifer, chloride ion concentration is uniformly low, but the concentration north and south of the Capitan aquifer gradationally increases to more than 100,000 mg/l. Hiss (1980) regarded the aquifers north and south of the Capitan shelf margin as separate from and poorly connected to the Capitan aquifer. One anomaly is the low chloride concentration near Hobbs, New Mexico in the northeast part of the map. The contours suggest the area near Hobbs is now, or was, an outlet for water flowing through the Capitan aquifer, allowing for east directed hydrodynamic flow Hiss, (1980).

API Gravity of oils in strata associated with the Capitan aquifer

The quality of oil is commonly measured by its API gravity and its sulfur content. API gravity is a measurement of the weight of crude oil compared to water. There is an inverse relationship between API gravity and density.

Figure 4. Chloride-ion concentrations in Guadalupian strata with API gravity of oils in the Capitan aquifer and Artesia Group reservoirs. East of the Pecos River, chloride-ion concentrations are low within the Capitan aquifer. To the north and south, chloride-ion concentrations are high in Guadalupian aquifers that are poorly connected to the Capitan aquifer. Symbols for the Capitan aquifer and for major oil fields in the Artesia Group are the same as Figure 3.

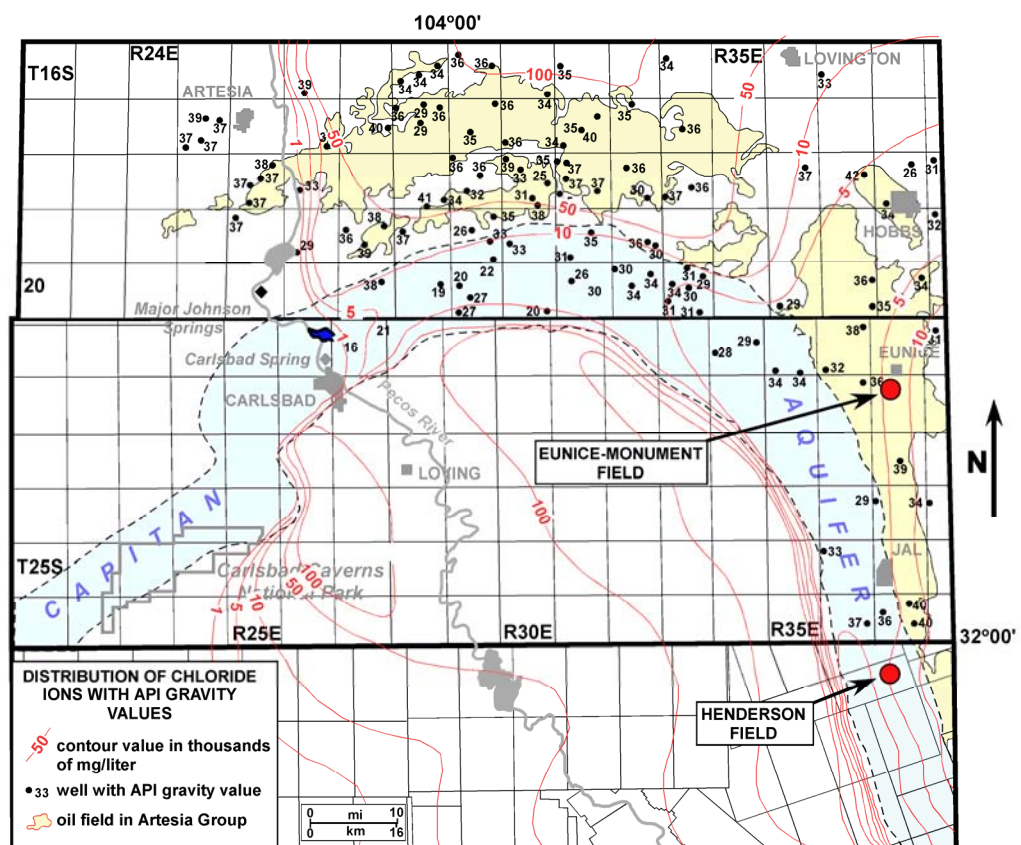


Figure 5 shows the API gravity of crude oil in and near the Capitan aquifer in southeastern New Mexico. Lowering of API gravity can be caused by oxidation, water-washing and microbial degradation (Bailey et al., 1973). API gravity values are also plotted on Figure 4, which shows that most of the lower values are associated with relatively fresh water in the Capitan aquifer. The lowest values are found near the Pecos River at Carlsbad, and API Gravities gradually increase to the east, within the Capitan aquifer, and to the north and northeast in laterally equivalent strata.

Degraded oil has been found in fluid inclusions in late-stage calcite at Henderson field (Wiggins, et al., 1993, p 783-784). The oil in the inclusions is more degraded than the oil that is currently produced from the reservoir, leading Wiggins et al., (1993, p. 788) to conclude that the Henderson reservoir was either replenished by a second pulse of oil after calcite cementation or the original oil column was displaced and oil migrated back into the reservoir after calcite cementation. The oil column at Eunice-Monument field was displaced by strong, eastward directed flow of meteoric water prior to the opening of the Rio Grande rift (Lindsay, 1998). When hydrodynamic flow diminished, oil migrated back into the trap at Eunice-Monument, partly refilling the reservoir (Robert F. Lindsay, personal communication, 2008).

Discussion

Sulfur and H_2S are common constituents of oil, gas and water in the Capitan aquifer and Artesia Group strata in southeastern New Mexico. They are most likely derived from the alteration of sulfates and hydrocarbons by microbes in Artesia Group reservoirs. For example, Henderson field, near the southeastern corner of New Mexico (Figure 3, 4 and 5), produces oil from Yates and Seven Rivers reservoirs. Oil in this reservoir is degraded due to oxidation by bacteria, which resulted in precipitation of late stage calcite and production of H_2S (Wiggins et al., 1993). Henderson field is one of many oil fields in Artesia Group reservoirs on the shelfward side of the Capitan aquifer. Most of these fields are stratigraphically trapped by the gradual up-dip increase of anhydrite pore filling. Calcium sulfate ($CaSO_4$) is an integral part of the trapping facies in these fields and provides a source for the sulfur released by microbial metabolism. Hydrogen sulfide is a documented constituent of the oil in many Artesia Group oil fields, and is likely found in all of them. The biogenic degradation of hydrocarbons generated huge amounts of H_2S which is found today in oil, gas and formation water in Artesia Group strata and the Capitan Aquifer. Diffusion is the probable mechanism for the widespread distribution of H_2S in these reservoirs.

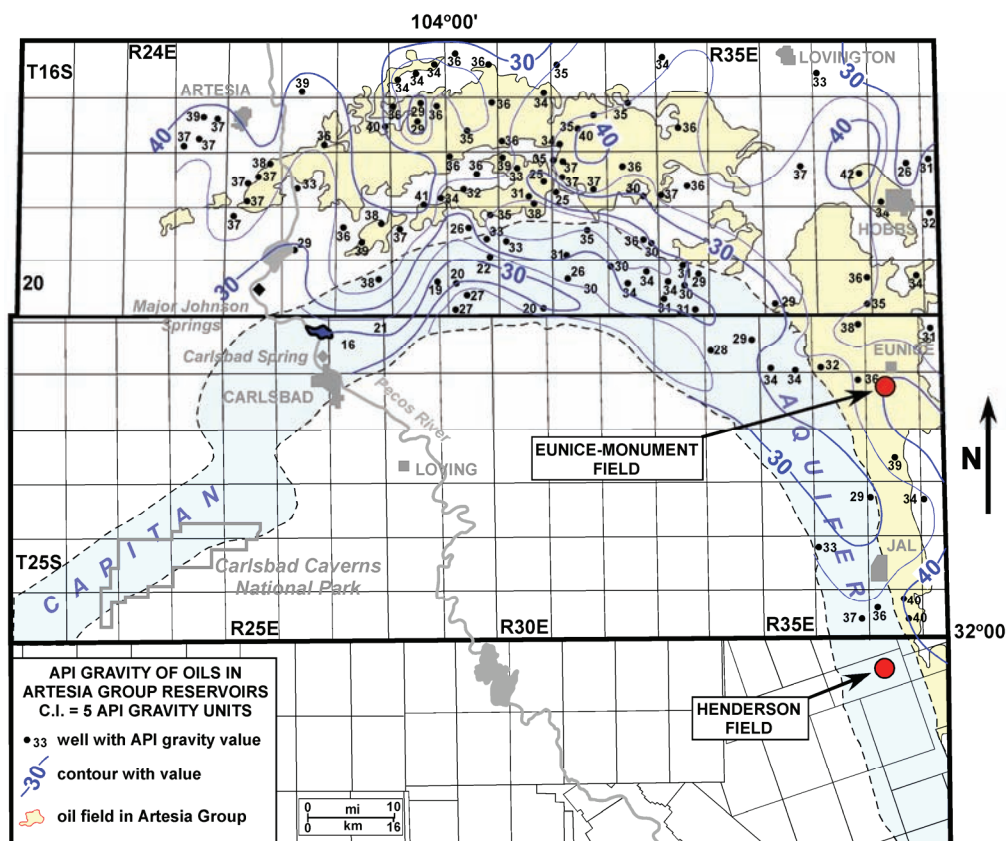


Figure 5. API gravity contour map for oils in the Capitan aquifer and Artesia Group. API gravity values are lowest in the Capitan aquifer near the Pecos River. API gravity contours reflect eastward movement of groundwater through the Capitan aquifer and show a systematic degradation in oil quality from east to west. Symbols for the Capitan aquifer and for major oil fields in the Artesia Group are the same as Figure 3.

West of the Pecos River, on the northwest flank of the Guadalupe Mountains, there are numerous drill holes where sulfur has been documented (Figure 3), and water containing H_2S is commonly encountered by drillers in this area (Hinds and Cunningham, 1970). The presence of sulfur and H_2S suggests that microbial alteration of hydrocarbons and sulfate may have occurred in the past. Today, the hydrocarbons have been consumed or leaked from the reservoirs, but there is evidence that they once existed (Scholle et al., 1992; Spirakis and Cunningham, 1992).

Iso-API gravity contour lines show a pronounced plume of lower gravity oils coincident with the Capitan aquifer (Figure 5). The chloride ion concentration in strata of Guadalupian age is lowest (10,000 mg/l or less) within the Capitan aquifer. This suggests a relationship between water within the aquifer and low API gravity of the oils. The API gravity of oil can be decreased by oxidation, which normally occurs in near surface environments, by water-washing, and by biogenic alteration and degradation of hydrocarbons (Bailey, et al., 1973). The depth of hydrocarbon accumulations in Permian (Guadalupian) reservoirs generally decreases from east to west (Table 1), with most of the reservoir formations cropping out on the northern and western sides of the Guadalupe Mountains. Oil in the shallowest Artesia Group reservoirs may have been degraded in part by oxidation, and in part by microbial metabolism. Oil in traps that directly overlie the Capitan aquifer is exposed to the strongest hydrodynamic flow. Oil in these traps may have been degraded by microbial metabolism, oxidation and water-washing.

DuChene and Cunningham (2006) presented a two-stage model for speleogenesis in the Guadalupe Mountains. Prior to the opening of the Rio Grande rift, there was strong, east-directed hydrodynamic flow through the Capitan aquifer. The displaced oil column at Eunice-Monument field (Lindsay, 1998) and remobilized oil at Henderson field (Wiggins et al., 1993) are evidence of this flow regime. As the evaporite cover that confined Capitan aquifer flow in the present-day Guadalupe Mountains was stripped away, artesian springs formed along vertical and sub-vertical fractures in the Capitan Formation and Carlsbad facies of the Artesia Group. Water flowing upward dissolved limestone and enlarged the fractures, forming early stage caves that may not have involved sulfuric acid. At the time these early caves were forming, the Rio Grande rift had not yet opened and the present geomorphology of the Guadalupe Mountains had not formed. Beginning 35-38 Ma, the regional water table began to fall in response to erosion and the

opening of the Rio Grande rift (DuChene and Cunningham, 2006). Approximately 14-12 Ma, the lowering regional water table intersected the Capitan aquifer at the west end of the modern Guadalupe Mountains. Hydrogen sulfide derived from oil fields in Artesia Group reservoirs west of the Pecos River was present in connate water. Atmospheric O_2 in the vadose zone and O_2 dissolved in meteoric water mixed with sulfidic water at the water table to form sulfuric acid, which dissolved carbonate rocks in the Capitan aquifer. Focal points for dissolution were early stage vertical caves where the supplies of air and oxygenated meteoric water were greatest. Here, large horizontal passages and galleries were formed at the water table. As tectonism progressed and the Rio Grande rift continued to open, the graben on the west side of the Guadalupe Mountains periodically shifted downward, causing the water table to fall. During the period from 14 Ma to 3.8 Ma, the locus of sulfuric acid speleogenesis shifted downward and eastward through the Guadalupe Mountains (Polyak et al., 1998; DuChene and Cunningham, 2006). At the time that sulfuric acid speleogenesis was active, H_2S concentrations within the Capitan aquifer were similar to conditions east of the Pecos River today. Some hydrogen sulfide may have been supplied to the Capitan aquifer from the Delaware basin, as proposed by Hill (1987), but the largest supply was derived from the biogenic degradation of oil in Artesia Group reservoirs.

It is possible that sulfuric acid speleogenesis is occurring today in the Capitan aquifer east of the Pecos River. Water reported from P.C.A. (Yates), Russell (Yates), Gem (Yates) and Henderson (Yates-Seven Rivers) fields contains H_2S . Water that enters the aquifer along the Pecos River is oxygenated, but by the time it has moved a short distance into the subsurface, O_2 has reacted with H_2S to form sulfuric acid. The acid immediately reacts with limestone to form sulfate and carbon dioxide. In a system where H_2S is dissolved in groundwater, the critical ingredient for the formation of acid is O_2 , which enters the system from the surface. Carbonate rocks of the Capitan Formation and Artesia Group east of the Pecos River are covered with evaporite beds that inhibit the downward percolation of oxygenated meteoric water. Consequently, if speleogenesis is occurring east of the Pecos River today, it is happening at a very slow rate.

In the Guadalupe Mountains, caves formed because of the conjunction of tectonic, petrochemical and hydrological factors in an area with abundant, soluble carbonate rock. Sulfuric acid speleogenesis in the Guadalupe Mountains was a dynamic process that took place over at least 10 my. There was abundant

H₂S dissolved in connate water in the Capitan Formation and the carbonate rich Carlsbad facies of the Artesia Group. Tectonic activity, accompanied by erosion, caused evaporite beds that covered the Capitan aquifer to be gradually stripped away, resulting in the lowering of the water table. As the water table fell, the volume of rock above it became part of the vadose zone and was filled with air, providing a source of O₂. Oxygenated meteoric water percolated downward to the water table where it mixed with H₂S rich connate water to form the sulfuric acid responsible for Late Tertiary speleogenesis.

Conclusions

Hydrogen sulfide and sulfur are common constituents of connate water in southeastern New Mexico, where they are associated with oil fields in Artesia Group reservoirs north of the Capitan aquifer. Hydrogen sulfide was formed by the oxidation of hydrocarbons and associated reduction of sulfate by microbes in Artesia Group reservoirs. This reaction created a large volume of H₂S which is dissolved in oil, gas and connate water contained in Artesia Group strata and in the Capitan aquifer east of the Pecos River. The presence of sulfur, H₂S, and hydrocarbons in fluid inclusions west of the Pecos River and north of the Capitan shelf margin is evidence that hydrocarbons were present prior to the opening of the Rio Grande rift. The H₂S that contributed to sulfuric acid speleogenesis in the Guadalupe Mountains is most likely derived from the biogenic alteration of hydrocarbons in these oil fields in Late Tertiary time. Artesia Group reservoirs occur in rocks that are laterally equivalent to the strata that comprise the Capitan aquifer. In Late Tertiary time, H₂S generated by the microbial alteration of oil in Artesia Group reservoirs was widely distributed in the Capitan aquifer and associated Carlsbad facies carbonates of the Artesia Group in the Guadalupe Mountains, as it is today in the subsurface east of the Pecos River. Oil accumulations in the Artesia Group are the most likely source of the H₂S that was converted to sulfuric acid in the Guadalupe Mountains and caused Late Tertiary speleogenesis.

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References

- Bailey, N.J. L., H.R. Krouse, C.R. Evans, and M. . Rogers. 1973. Alteration of crude oil by waters and bacteria – evidence from geochemical and isotope studies. *American Association of Petroleum Geologists Bulletin* 57: 1276-1290.
- Cunningham, K.I., H.R. DuChene, and C.S. Spirakis, C.S. 1993. Elemental sulfur in caves of the Guadalupe Mountains, New Mexico. In *Carlsbad region, New Mexico and West Texas: New Mexico Geological Society Guidebook 44*, ed. D. Love, J. W. Hawley, B. S. Kues, J. W. Adams, G. S. Austin and J. M. Barker, 129-136. Socorro: New Mexico Geological Society.
- Dietrich, E.S. 1956. Robinson (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 304-305. Roswell: Roswell Geological Society.
- Davis, D.G. 1973. Sulfur in Cottonwood Cave, Eddy County, New Mexico. *National Speleological Society Bulletin* 35 (3): 89-95.
- DuChene, H.R., and K.I. Cunningham. 2006. Tectonic influences on speleogenesis in the Guadalupe Mountains, New Mexico and Texas. In *Caves and karst of southeastern New Mexico: New Mexico Geological Society Guidebook 57*, ed. L. Land, V.W. Lueth, W. Raatz, P. Boston and D.L. Love, 211-218. Socorro: New Mexico Geological Society.
- DuChene, H.R., and J.S. McLean. 1989. The role of hydrogen sulfide in the evolution of caves in the Guadalupe Mountains of southeastern New Mexico. In *Subsurface and outcrop examination of the Capitan Shelf Margin, northern Delaware Basin: SEPM Core Workshop No. 13*, ed. P. D. Harris and G. A. Grover, 475-481. Tulsa: Society for Sedimentary Geology.
- Fryman, A.A. 1956. Grayburg-Jackson (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 40-41. Roswell: Roswell Geological Society.
- Fryman, D. 1956. McMillan (Queen) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 275-276. Roswell: Roswell Geological Society.
- Goodger, J.M. 1956. Square Lake (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 328-329. Roswell: Roswell Geological Society.
- Green W.R. 1956. Penrose-Skelly (Queen-Grayburg) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 294-295. Roswell: Roswell Geological Society.

- Hill, C.A. 1987. *Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains, New Mexico and Texas*. New Mexico Bureau of Mines and Mineral Resources Bulletin 117. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hinds, J.S., and R.R. Cunningham. 1970. *Elemental sulfur in Eddy County, New Mexico*. U.S. Geological Survey Circular 628. Boulder: U.S. Geological Survey.
- Hiss, W.L. 1975. *Chloride-ion concentration in ground water in Permian Guadalupian rocks, southeast New Mexico and west Texas*. New Mexico Bureau of Mines and Mineral Resources, Resource Map 4. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hiss, W.L. 1980. Movement of ground water in Permian Guadalupian aquifer systems, southeastern New Mexico and western Texas. In *Trans-Pecos region, southeastern New Mexico and west Texas: New Mexico Geological Society Guidebook 31*, ed. P.W. Dickerson, J.M. Hoffer and J.F. Callender, 289-294. Socorro: New Mexico Geological Society.
- Kinney, E.E., 1956a. Aid (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 40-41. Roswell: Roswell Geological Society.
- Kinney, E.E., 1956b. Tonto (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 342-343. Roswell: Roswell Geological Society.
- Kirkland, D.W., and R. Evans. 1976. Origin of limestone buttes, Gypsum Plain, Culberson County, Texas. *American Association of Petroleum Geologists Bulletin* 60: 2005-2018.
- Lindsay, R.F. 1998. Meteoric recharge, displacement of oil columns and the development of residual oil intervals in the Permian basin. In *West Texas Geological Society Publication 98-105, The Search Continues into the 21st Century*, ed. W.D. DeMis and M.K. Nelis, 271-274. Midland: West Texas Geological Society.
- McPeters, K.D., and J.M. Kelly. 1956. Hobbs (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 206-207. Roswell: Roswell Geological Society.
- Mershon, P.M., Jr. 1960. East Millman (Queen-Grayburg) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 152-153. Roswell: Roswell Geological Society.
- Meissner, F.F. 1972. *Cyclic sedimentation in middle Permian strata of the Permian Basin, west Texas and New Mexico*. West Texas Geological Society Publication 72-18, second edition, 203-232. Midland: West Texas Geological Society.
- Milks, R. 1956. Vacuum (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 352-353. Roswell: Roswell Geological Society.
- Palmer, A.N., and M.V. Palmer. 2000. Hydrochemical interpretation of cave patterns in the Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62: 91-108.
- Polyak V.J., W.C. McIntosh, N. Güven, and P. Provencio. 1998. Age and origin of Carlsbad Cavern and related caves from ⁴⁰Ar/³⁹Ar of alunite. *Science* 279: 1919-1922.
- Puckett, H.W. 1960. Dos Hermanos Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 68-69. Roswell: Roswell Geological Society.
- Roswell Geological Society. 1956. *A symposium of oil and gas fields of southeastern New Mexico*. Roswell. Roswell Geological Society.
- Roswell Geological Society. 1960. *A symposium of oil and gas fields of southeastern New Mexico*. Roswell. Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956a. Arrowhead (Queen-Grayburg) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 52-53. Roswell: Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956b. Atoka (San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 56-57. Roswell: Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956c. Getty (Yates) Field. In *Roswell Geological Society Symposium. Oil and gas fields of southeastern New Mexico*, 180-181. Roswell: Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956d. Jalmat (Yates-Seven Rivers) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 218-219. Roswell: Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956e. P.C.A. (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 290-291. Roswell: Roswell Geological Society.
- Roswell Geological Society Symposium Committee. 1956f. Red Lake (Grayburg-San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 298-299. Roswell: Roswell Geological Society.

- Roswell Geological Society Symposium Committee. 1956g. Rhodes (Yates-Seven Rivers) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 300-301. Roswell: Roswell Geological Society.
- Scholle, P.A., D.S. Ulmer, and L.A. Melim. 1992. Late stage calcites in the Permian Capitan Formation and its equivalents, Delaware Basin margin, west Texas and New Mexico. *Sedimentology* 39: 207-234.
- Schram, H.F. 1956a. Cedar Hills (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 102-103. Roswell: Roswell Geological Society.
- Schram, H.F. 1956b. Russell (Yates) Field. In *Roswell Geological Society Symposium. Oil and gas fields of southeastern New Mexico*, 306-307. Roswell: Roswell Geological Society.
- Spirakis, C., and K.I. Cunningham. 1992. Genesis of sulfur deposits in Lechuguilla Cave, Carlsbad Caverns National Park. In *American Institute of Mining, Metallurgical and Petroleum Engineers (AIME) Special Volume: Native sulfur – developments in geology and exploration*, ed. G. Wessel and B. Wimberley, 139-145. Littleton: American Institute of Mining, Metallurgical and Petroleum Engineers.
- Stringer, B. 1956. Langlie-Mattix (Yates-Seven Rivers-Queen) Field. In *Roswell Geological Society Symposium: Oil and gas Fields of southeastern New Mexico*, 226-227. Roswell: Roswell Geological Society.
- Sweeney, H.N. 1956a. Halfway (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 193. Roswell: Roswell Geological Society.
- Sweeney, H.N. 1956b. Littman (San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 238-239. Roswell: Roswell Geological Society.
- Sweeney, H.N. 1956c. Young (Queen) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 374-375. Roswell: Roswell Geological Society.
- Ward, R.F., St. C. Kendall, and M. Harris. 1986. Upper Permian (Guadalupian) facies and their association with hydrocarbons – Permian basin, west Texas and New Mexico. *American Association of Petroleum Geologists Bulletin* 70: 239-262.
- Wiggins, W.D., P.M. Harris, and R.C. Burruss, 1993. Geochemistry of post-uplift calcite in the Permian Basin of Texas and New Mexico. *Geological Society of America Bulletin* 105: 779-790.
- Wilson, D. 1956. Gem (Yates) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 179. Roswell: Roswell Geological Society.
- Zoller, M.L. 1956. House (San Andres) Field. In *Roswell Geological Society Symposium: Oil and gas fields of southeastern New Mexico*, 214-215. Roswell: Roswell Geological Society.

THE PECOS RIVER HYPOGENE SPELEOGENETIC PROVINCE: A BASIN-SCALE KARST PARADIGM FOR EASTERN NEW MEXICO AND WEST TEXAS, USA

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Abstract

Since the mid-Tertiary, lateral migration and entrenchment of the Pecos River Valley in eastern New Mexico and west Texas, USA, has significantly influenced regional groundwater flow paths, providing a focus for ascending flow in multi-storey artesian systems and a powerful potentiometric driving force for hypogene speleogenesis. Individual occurrences of hypogene karst phenomena associated with the central Pecos River Valley are widespread throughout the greater Delaware Basin region, including development in a wide range of Permian carbonate and evaporite facies. Hypogene occurrences are well-documented as far north as Santa Rosa, New Mexico and as far south as Lake Amistad, Texas. Throughout the northern shelf, intrastratal dissolution and brecciation of the San Andres formation is widespread as a result of eastward migration of the Pecos River. Proximal to the current river, hypogene dissolution in interbedded carbonate/evaporite facies of the Seven Rivers Formation has produced three-dimensional network caves and vertical collapse structures. In the carbonate reef facies of the Guadalupe Mountains, complex three-dimensional caves are common, as well as stepped terraces associated with eastward migration of the Pecos River. Although these caves have been attributed to sulfuric acid dissolution, they are the result of hypogene speleogenesis in which solutional aggressivity was increased by the addition of both thermal and sulfuric-acid components. Within the interior of the Delaware Basin, hypogene karst in basin-filling evaporite facies of the Castile and Salado Formations is widespread, including development of large solution subsidence troughs associated with the lateral migration of the Pecos River. On the far eastern margin of the Delaware Basin, at the southeastern tip of the Central Basin Platform, persistent downcutting of the Pecos River Valley contributed to the development of hypogene karst within the Yates Petroleum

Field, providing cavernous reservoir porosity for the largest individual oil field known within the Permian Basin region. Immediately below the confluence of the Pecos River and the Rio Grande, the large first order magnitude spring, Goodenough Spring, flows from a deep phreatic cave under extreme artesian conditions, even as 45 meters of pressure head has been added over the spring from Amistad Reservoir.

Introduction

Hypogene processes have been recognized throughout far west Texas and southeastern New Mexico, USA, for many decades, but these processes have generally been associated with unusual fluid chemistries in Permian age carbonate units of the Guadalupe Mountains (Figure 1), specifically sulfuric acid speleogenesis. While sulfuric-acid karst is often hypogene, the two are not interchangeable terms. Instead, what has been defined as sulfuric acid karst in the Guadalupe Mountains is porosity produced by hypogene processes that has simply been enhanced by solutionally aggressive fluids enriched with a sulfuric acid component. This phenomenon is not limited to the reef and forereef facies of the Capitan Formation in the Guadalupe Mountains, but also extends into the carbonate backreef facies as seen in the upper portions of some of the Guadalupe Mountains caves (Hose and Pisarowicz, 2000) and in strata deposited farther shelfward, such as found associated with the caves of McKittrick Hill (Kunath, 1978).

The voluminous carbonate caves of the Guadalupe Mountains are usually invoked as typical examples of hypogenic speleogenesis in the Delaware Basin region; however, there is extensive and even more widespread karst development within the associated evaporite facies of southeastern New Mexico and west Texas (Stafford and Nance, 2009). Breccia pipes in evaporite strata, several hundred meters in vertical

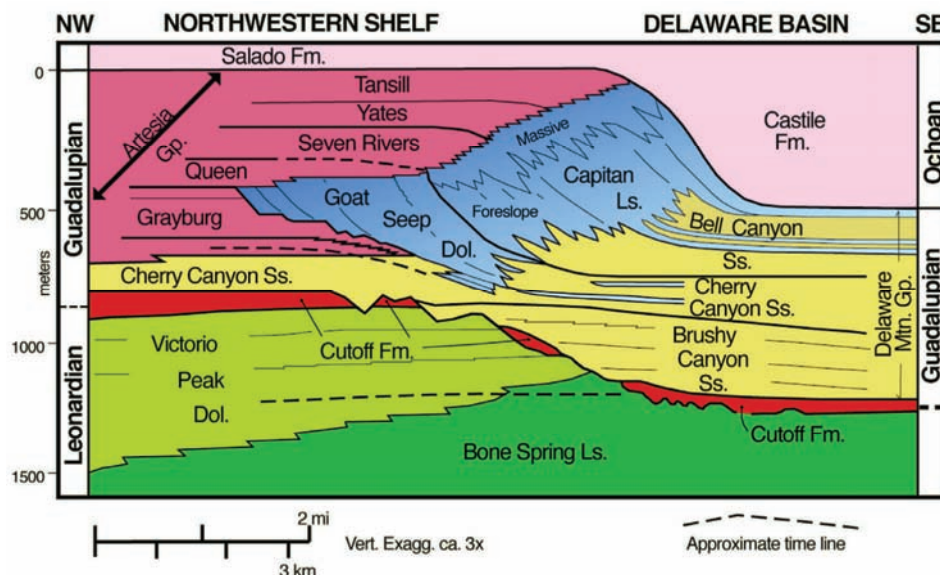


Figure 1. Late Permian stratigraphic nomenclature showing relationship to deposits on the Northwestern Shelf, Capitan Reef and Delaware Basin (from Scholle, 2004).

extent, have been associated with brine density convection, where hypogene processes are driven by variations in the solute concentrations of intrastratal fluids (Anderson and Kirkland, 1980). Recent research has shown that evaporite calcitization, native sulfur deposits, and speleogenesis within the Castile Formation are largely the result of hypogene processes (Stafford et al., 2008d, e), an association similar to that recognized in the western Ukraine (Klimchouk, 1997). However, all of these features show varying degrees of epigenetic overprinting, as do the carbonate caves of the region. The very nature of these caves, breached, drained and thus available for human exploration, has removed them from the hypogenic environment in which they formed.

Most of the course of the Pecos River is across Permian age carbonate and evaporite facies with associated karst development in these strata. However, throughout the southeastern portion of the Pecos River Valley near-surface strata are primarily Cretaceous carbonate rocks of the Edwards Plateau. Klimchouk (2007) argues that maze caves (e.g., Amazing Maze Cave) in this region are the result of hypogene processes involving sulfuric acid-rich

these speleogenetic phenomena in the context of related, basin-scale processes dominated by a unifying potentiometric driving force, the Pecos River.

fluids with a thermal component. Similarly, various caves beyond the Pecos River Valley in Cretaceous strata west of San Antonio (Kunath, 1995) exhibit morphologies that are suggestive of formation by hypogene processes.

Previously known evidence of hypogene karst occurs throughout the lower Pecos region of eastern New Mexico and west Texas, but most of these features have been regarded as unique, isolated occurrences. This paper is a first attempt to view

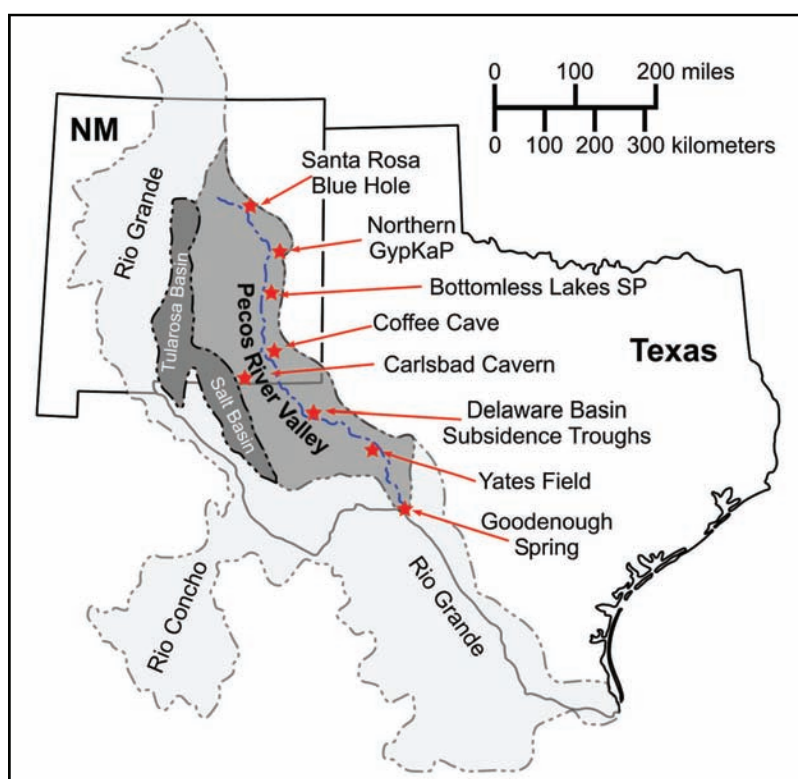


Figure 2. Location of karst features throughout the current Pecos River Valley, with comparison to the modern Rio Grande and Rio Concho basins (adapted from Thomas, 1972).

Pecos River

The headwaters of the Pecos River are located in the Sangre de Cristo Mountains of northern New Mexico (Figure 2). After exiting the mountains in western San Miguel County, the Pecos flows southward across eastern New Mexico into west Texas. The course of the river then turns southeast and ultimately flows into the Rio Grande west of Lake Amistad on the international border. The river's length is 1,320 km. Over this distance the Pecos crosses portions of the Southern Rockies, High Plains and Edwards Plateau physiographic provinces. Most of the river's route is across Permian age strata, including: 1) evaporites, mudstone and carbonates representing far backreef facies on the Northwest Shelf of the Delaware Basin ; 2) carbonates of near backreef, reef, and forereef facies deposited on the Basin margin; and 3) evaporites that filled the Permian inland basins and extended beyond the basin margin onto the Northwest Shelf by the end of the Permian. These sediments were deposited in equatorial conditions along the western edge of Pangaea with a limited connection to the Panthalassa Ocean (Scholle et al., 2004). Today, these strata not only host groundwater resources and numerous caves, but also a plethora of oil fields, both large and small.

As the Pecos River passes out of the Delaware Basin and onto the Central Basin Platform, the Permian strata are progressively buried in subsurface and surficial units dominated by Triassic conglomerates and sandstones, and by overlying carbonate strata. The carbonate strata were deposited during a Cretaceous-age continental transgression, which covered the western interior of North America with a shallow epicontinental sea (Richey et al., 1985). Through much of the flowpath across these carbonate strata, the Pecos River has entrenched into the Cretaceous units of the Edwards Plateau (Thomas, 1972). Today, the Pecos River is a major tributary of the Rio Grande, and is deeply-entrenched, near the rivers' junction west of Lake Amistad.

Thomas (1972) states that by the end of the Cretaceous or earliest Paleogene, the ancestral Pecos River had already developed in eastern New Mexico and west Texas as a result of Laramide Orogeny and uplift of the San Juan Mountains in southwestern Colorado (Figure 3). This early Pecos River was likely the dominant fluvial system at this time as it flowed across northern New Mexico, down through west Texas and discharged into the Gulf of Mexico possibly

by joining with an ancestral Conchos River. This early phase began its incision into the Edwards Plateau (Thomas, 1972). During the late Paleogene and throughout much of the Neogene, most of the northern reaches of the ancestral Pecos River were diverted toward eastern New Mexico and northwest Texas where vast accumulations of clastic sediments were deposited to produce the Ogallala Formation (Bretz and Horberg, 1949). This northern piracy, resulting from increased sediment production from the Rocky Mountains, reduced the area of

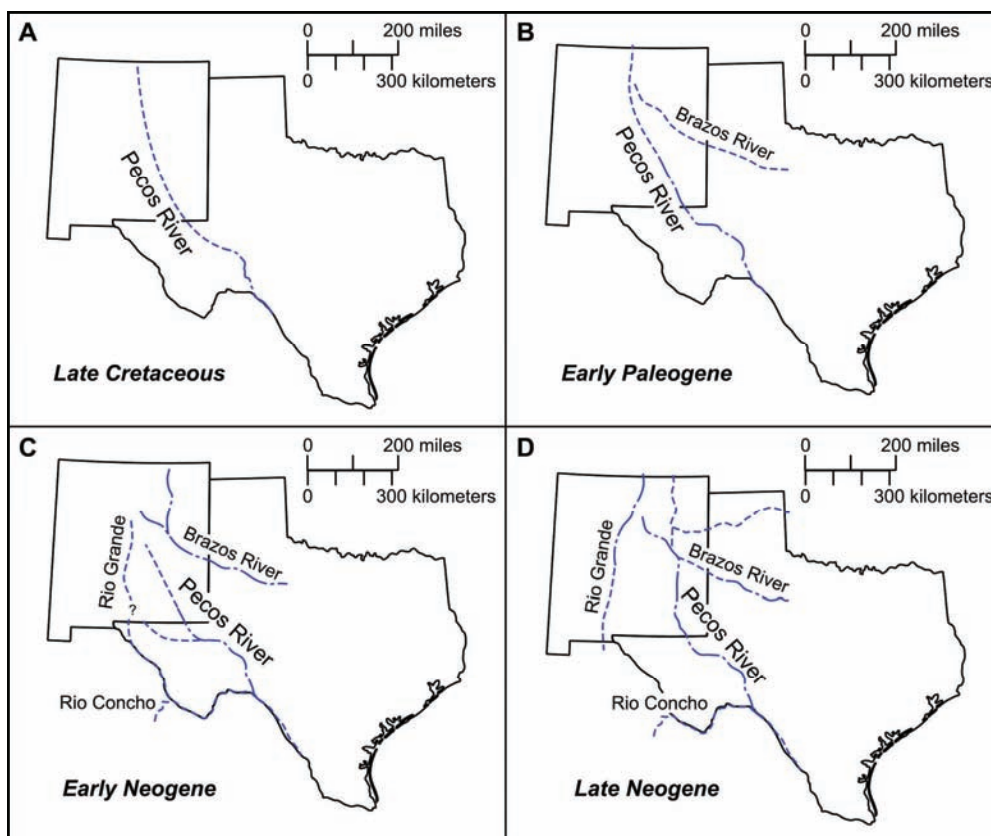


Figure 3. Reconstructions of the ancestral Pecos River in the A) Late Cretaceous, B) Early Paleogene, C) Early Neogene and D) Late Neogene (adapted from Thomas, 1972).

the Pecos River Valley; however its general course persisted throughout southeastern New Mexico and west Texas, with continued entrenchment into the Edwards Plateau (Thomas, 1972).

By the end of the Neogene (Figure 3), uplift of the Sacramento and Sangre de Cristo Mountains had begun, shaping the current route of the Pecos River. At the same time, the ancestral Rio Grande was forming and at this time probably flowed across the Delaware Basin of west Texas as a tributary of the Pecos, or it may have continued farther south as a tributary of the ancestral Conchos River (Thomas, 1972). Contemporaneously, the ancestral Brazos River was draining most of northeastern New Mexico, continuing to supply alluvium to north Texas. During the Pleistocene the lower Pecos Valley extended farther to the north by karstification and headward erosion, culminating in capture of the ancestral Brazos near Fort Sumner, and thereby established the modern configuration of the Pecos River valley (Galloway, 1956; Reeves, 1972).

Throughout its history of the Pecos River, it has had a profound influence on the geomorphology and hydrology of west Texas and eastern New Mexico. Since the late Cretaceous or early Paleogene, the Pecos River has incised into the Edwards Plateau of southwest Texas, providing a consistent, major potentiometric low across this region that focused groundwater migration and discharge. Throughout west Texas and southeastern New Mexico the Pecos River has persisted since its inception, but in this region there have been greater lateral shifts as the river has migrated across a large alluvial floodplain (Thomas, 1972), again acting as a regional potentiometric low. The northern reaches of the Pecos River have seen the most change during its history. While the northernmost reaches of the ancestral Pecos are now part of the Rio Grande, its current headwaters are positioned in a region that has been occupied by either the ancestral Pecos River or the ancestral Brazos River for at least several million years (Thomas, 1972). This long presence of a river in a relatively persistent location must have had a significant influence on the evolution of regional groundwater flow paths.

Selected Karst Phenomena

In the following sections, eight specific areas are discussed that

demonstrate the dominance of hypogene speleogenesis throughout the greater Pecos River basin. While many more individual examples occur throughout the region, these eight show that hypogene karst development associated with the Pecos River exists from the northern reaches of the Pecos River Valley of north-central New Mexico, through the entire length of the river as it passes through eastern New Mexico and west Texas, and even at its confluence with the Rio Grande in south Texas. These eight occurrences likely represent only a small fraction of the total hypogene karst that is associated with the evolution of the Pecos River (Figure 2).

Santa Rosa Blue Hole

Santa Rosa, known as the “City of Natural Lakes” (McLemore, 1989), is located on the Pecos River in Guadalupe County, New Mexico (Figure 2). The city is known for its many spring-fed lakes and marshes, including the well known Blue Hole. Blue Hole is a collapse feature more than 25 meters deep, with an entrance approximately 24 meters in diameter that bells out to approximately 40 meters in diameter at the bottom (Kelley, 1972). Blue Hole is developed in the Triassic Santa Rosa Sandstone, which crops out throughout the Santa Rosa area. While the surficial sandstones are largely insoluble, the underlying Permian San Andres carbonates contain abundant solutional voids that stope upward to create cenote-like features that pierce the overlying strata (McLemore, 1989). Blue Hole, the largest of 16 small natural lakes in the Santa Rosa area, discharges more than 16.3 million liters of potable spring water per day (Kelley, 1972).

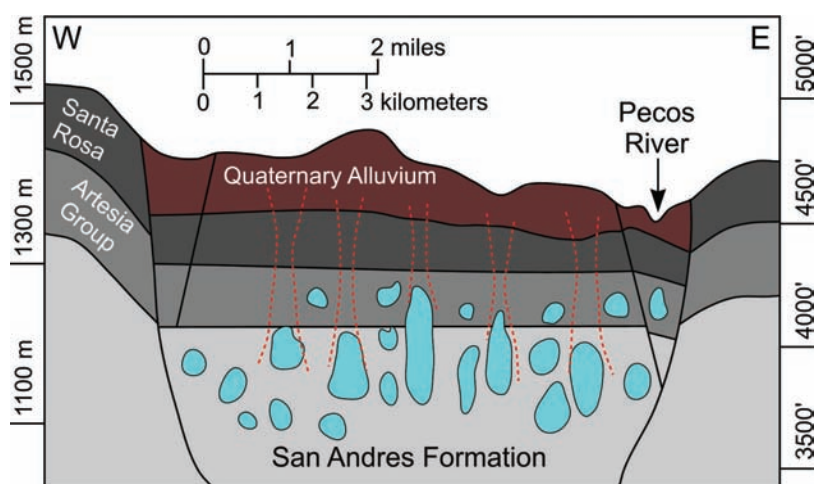


Figure 4. Simplified cross-section through the Santa Rosa Sink, showing intense dissolution in the San Andres and massive stoping to produce this immense collapse structure. Dashed red lines indicate zones of stoping (adapted from Kelley, 1972).

Blue Hole and the other flooded sinkholes of Santa Rosa are only a small part of speleogenesis in the area. These features and the town of Santa Rosa are all positioned inside a broad karst subsidence depression 10 kilometers in diameter and more than 120 meters deep (Figure 4) (Kelley, 1972). The depression consists of a series of convex-outward fractures, suggesting that the feature has expanded outward from a central collapse area. According to Kelley (1972), collapse of the massive sink began in the early Pleistocene, based on sediment fill and the deep gorges that the Pecos River cut through the northern and southern edges of the sink as a result of incision related to Pleistocene regional uplift. Most of the interior of the sink is mantled with Pleistocene gravels as the Pecos River deposited more than 75 meters of sediment into this expanding collapse structure (McLemore, 1989). Today, more than 190 individual collapse features in and around the periphery of the Santa Rosa depression attest to extensive vertical stoping of voids throughout the region. The large size of the subsidence basin and the abundance of springs and sinkhole lakes suggest that significant cavernous porosity exists at depth, probably related to deep circulation flow paths, which are focusing artesian discharge locally along the potentiometric low created by the Pecos River.

Northern GypKaP

The San Andres Formation crops out extensively from Roswell to Vaughn, New Mexico on both the east and west sides of the Pecos River valley. Here interbedded gypsum and dolomite host a plethora of simple and

complex caves in what is commonly termed “Northern GypKaP,” the northern extent of the Gypsum Karst Project of the National Speleological Society (Figure 2) (Eaton, 1987). While karst regions to the north and south currently exhibit abundant features that actively discharge artesian waters, attesting to hypogene influences, the caves in this region have a more complex history, with sections that clearly exhibit morphologies similar to those observed in hypogenic caves, and other sections that are dominated by epigenic, vadose features.

Many caves in Northern GypKaP exhibit sections of rectilinear maze and/or large isolated chambers that have been intersected by entrenched vadose canyons (Figure 5). Scrooge Cave contains a large section of rectilinear maze developed along northeast and northwest trending fracture sets, while Montecito Cave contains long sections of narrow vadose canyons that occasionally intercept large isolated chambers, generally at significant lithology changes (Stafford and Nance, 2009). Most of the longer surveyed caves exhibit passage lengths that are disproportionately long compared to the size of the watersheds that feed the entrances, especially for caves developed in gypsum facies. Individual morphometric features suggestive of hypogene dissolution, or at the very least sluggish flow regimes, are common throughout both the small and large caves of the region. However, scallops are also commonly observed, indicating that in recent times high velocity, unconfined flow conditions have existed in many of these caves.

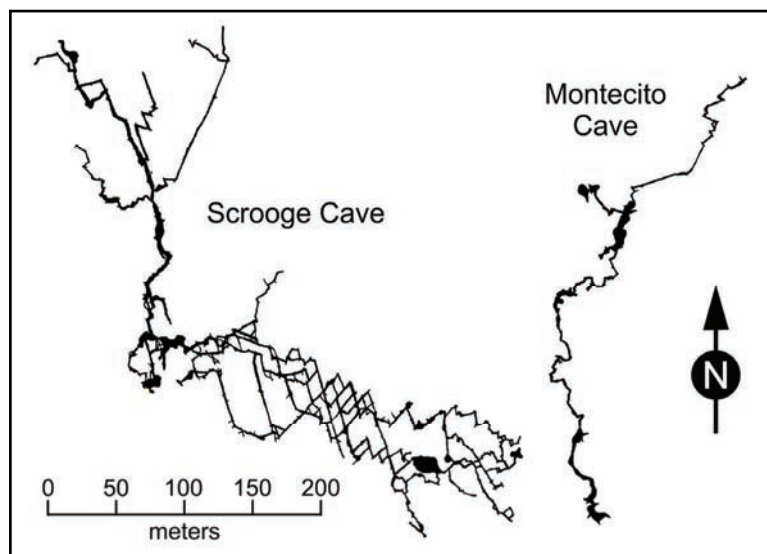


Figure 5. Plan view map outlines of Scrooge and Montecito Caves, which illustrate the complex cave patterns documented in Northern GypKaP. Note that caves are depicted at the same scale but not spatially projected in relation to each other (adapted from Lee, 1996).

Many of the caves within the Northern GypKaP region have complex speleogenetic histories that reflect epigenic overprinting onto originally hypogenic features. Complex maze patterns and extensive cave networks in gypsum facies likely formed when these units were buried more deeply and the Pecos River was migrating across them, possibly tens of millions of years ago at the same time intense dissolution was actively proceeding near Santa Rosa to the north (Kelley, 1972). As surface denudation and down cutting of the Pecos River continued, these caves systems became unconfined and exposed to epigenic processes. Surface breaching introduced focused recharge into the cave networks, which was followed by significant epigenic overprinting, including the development of incised canyons connecting zones of more intense hypogene porosity.

Bottomless Lakes State Park

Bottomless Lakes State Park is located on the eastern margin of the Pecos River Valley southeast of Roswell, New Mexico (Figures 2, 6). The park includes a series of eight cenote-like features, formed in gypsum and mudstone of the Permian Seven Rivers Formation, that extend for several kilometers along the Seven Rivers Escarpment (Kelley, 1971; Martinez et al., 1998; Land, 2003). These steep-walled, vertical collapse structures occur at the downstream end of the Roswell Artesian Basin, a regional artesian aquifer system formed in the San Andres limestone (Welder, 1983; Land and Newton, 2007; 2008). Meteoric

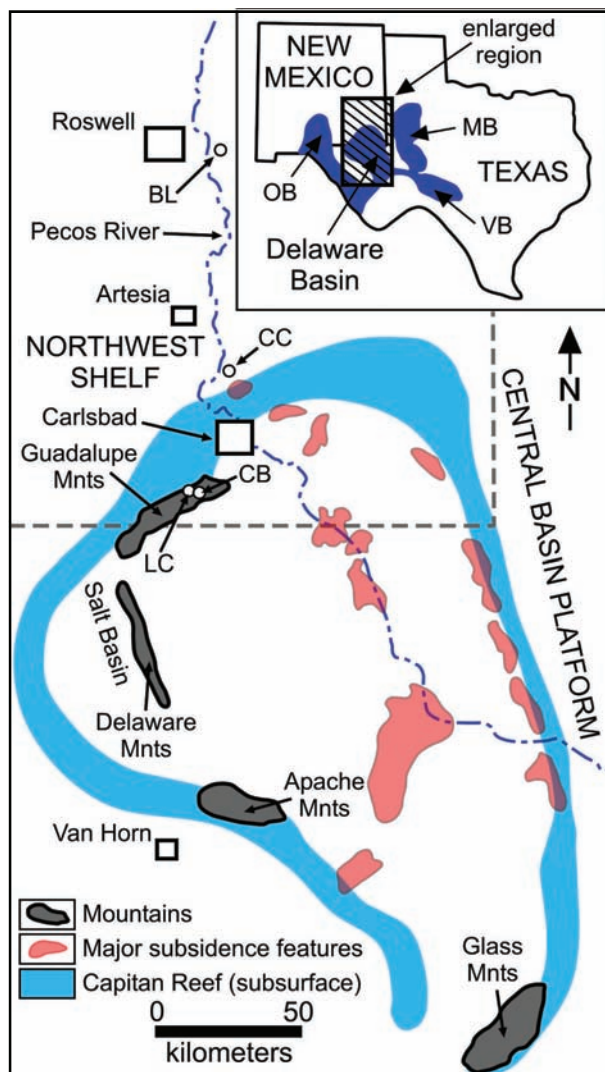


Figure 6. Map of Delaware Basin showing the geographic relationships between the Capitan Reef, major subsidence features and the Pecos River. BL = Bottomless Lakes State Park, CC = Coffee Cave, CB = Carlsbad Cavern, LC = Lechuguilla Cave, OB = Orogrande Basin, MB = Midland Basin, VB = Val Verde Basin (adapted from Klimchouk, 2007).

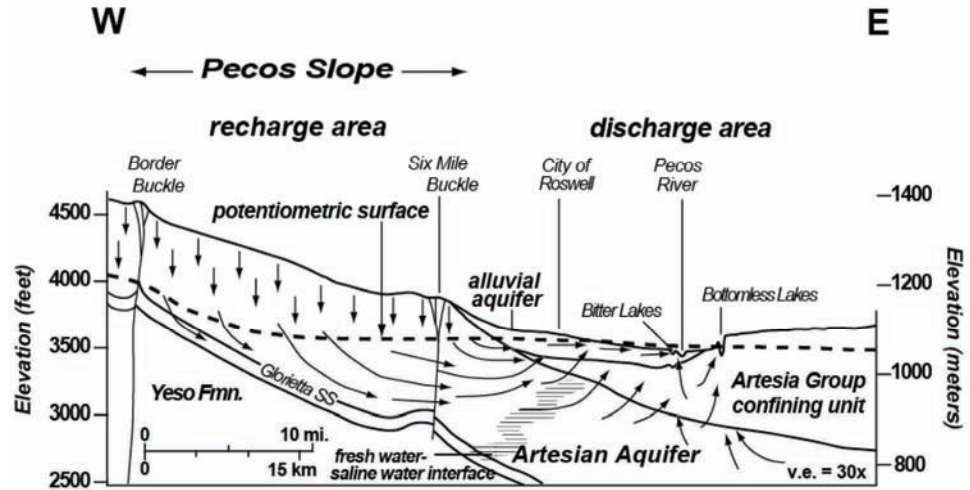
waters recharge the aquifer where the San Andres Formation crops out on the Pecos Slope west of Roswell, and migrate down gradient to the south and east. The San Andres aquifer is under water table conditions on the Pecos Slope, but becomes confined ~10 km west of Roswell where it dips beneath gypsum and mudstones of the Seven Rivers Formation (Figure 7.). In the vicinity of Bottomless Lakes, the potentiometric surface within the San Andres aquifer is above ground level. Forced convection drives lateral and upward migration of groundwater toward the Pecos River, and focused solution occurs along fractures in the overlying gypsum and mudstone. As artesian water migrates upward through the leaky confining beds of the Seven Rivers Formation, it is saturated with calcite but undersaturated with respect to gypsum. The result is the development of large hypogene voids that collapse and slope upward to discharge high salinity, artesian waters into the Bottomless Lakes cenotes. Some of these cenotes overflow into wetlands west of the park, which are hydraulically connected to the Pecos River (Land, 2003; 2006).

Historically, high volume springs that fed extensive wetlands were common in the Roswell area. Decades of intensive pumping for irrigated farming caused substantial declines in hydraulic head in the artesian aquifer, and most of these springs are now dry (Welder, 1983; Land and Newton, 2007; 2008). However, substantial discharge still occurs into springs and cenotes along the Pecos River, indicating that hypogene processes remain active along this reach of the Pecos River valley. This natural artesian discharge amounts to ~37 million m³/yr, but was much greater prior to pumping (Barroll and Shomaker, 2003).

Coffee Cave, Lake McMillan

Coffee Cave is located along the eastern edge of old Lake McMillan in northern Eddy County, New Mexico (Figure 2, 6). Coffee Cave, along with several smaller caves in the area, is developed in interbedded dolomites and gypsum of the Seven Rivers Formation. Coffee Cave is a rectilinear maze cave developed along northeast and northwest trending fracture sets (Figure 8) (Stafford et al., 2008b). The cave consists of at least 4 levels of passage development in gypsum facies, each separated by a significant dolomite bed. The uppermost two levels tend to be small and often not humanly navigable, while the lowest level, and potentially additional lower levels, is currently flooded. Throughout the cave, abundant hypogene morphometric features are found, including individual occurrences of well-developed cupolas, risers, half-

Figure 7. Stratigraphic cross-section showing the groundwater circulation flow paths associated with Bottomless Lakes State Park, where meteoric water is recharged on the Pecos Slope and discharge as artesian springs along the Pecos River (from Land, 2006).



tubes and ceiling channels (Stafford et al., 2008b). Most importantly, these morphometric features are commonly found in related suites where fluid flow paths can be visually traced from lower riser inlets, up wall half-tubes, along ceiling channels, and out ceiling cupolas. The presence of these features provides diagnostic evidence of dissolution in confined, hypogene conditions (Klimchouk, 2007). The morphol-

ogy of Coffee Cave shows considerable similarities to the classic hypogene maze caves of the Western Ukraine.

Stafford et al. (2008b) conclude that speleogenesis within Coffee Cave is largely the result of hypogene processes. The cave is located at the southern end of the Roswell Artesian Basin. Groundwater discharge

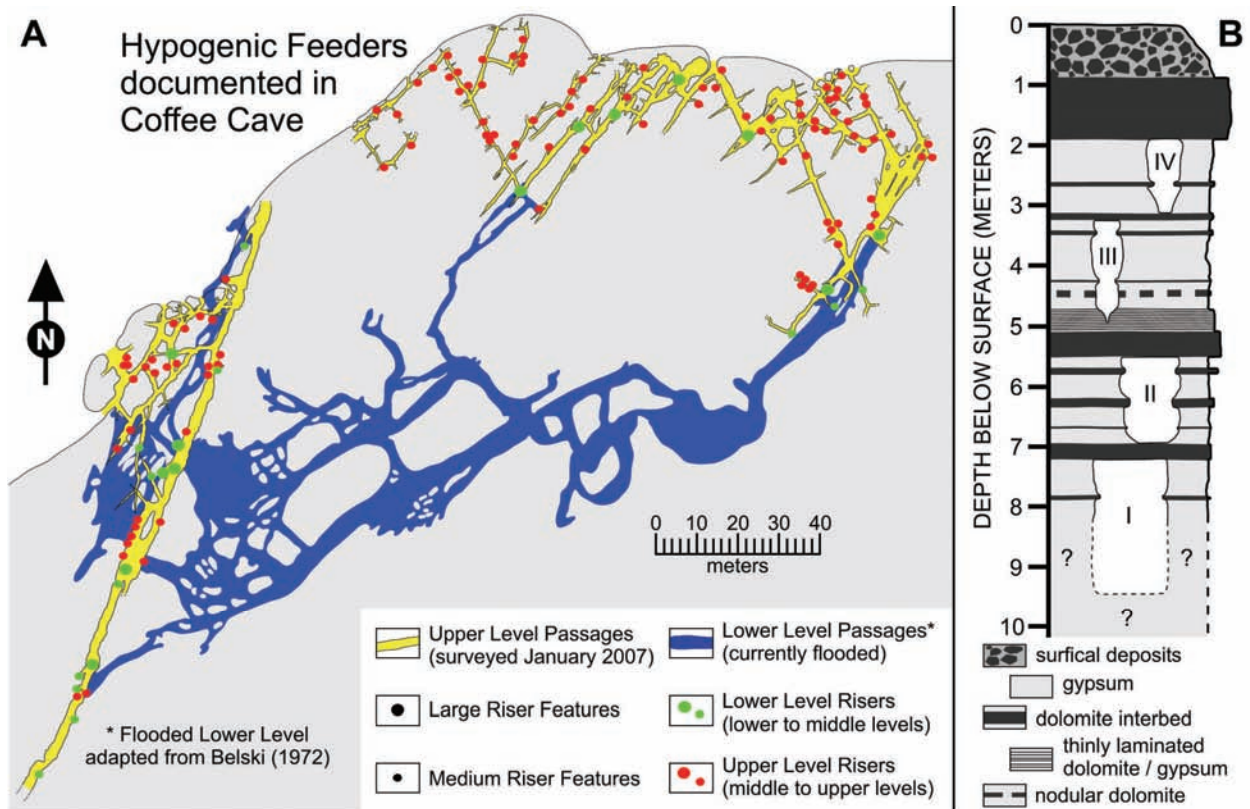


Figure 8. Coffee Cave, A) plan view map showing distribution of cave passages, levels and hypogene features and B) stratigraphic section through Coffee Cave showing relationship between the four documented cave levels and dolomite interbeds (from Stafford et al., 2008b).

from the underlying San Andres artesian aquifer has been focused upward along the Pecos River, similar to the hydrologic system described above at Bottomless Lakes State Park (Land, 2003; 2006) (Figure 7). Meteoric recharge on the Pecos Slope to the west infiltrates down through the Permian San Andres and Grayburg formations and moves laterally, down-gradient to the east and south (Land and Newton, 2007; 2008). Proximal to the Pecos River, these fluids migrate upward through the Seven Rivers Formation as forced convection is induced by the potentiometric low of the river (Stafford et al., 2008b). As surface denudation and eastward migration of the modern Pecos River occurred, Coffee Cave was breached and the semi-confined conditions that formed the bulk of the cavernous porosity were replaced by unconfined, epigene conditions. While early models for the formation of Coffee Cave suggest the cave resulted from dissolution associated with back-flooding along the Pecos River, gypsum solution kinetics do not support this theory. Average cave passage cross sectional area increases with depth and distance from the numerous collapse entrances into Coffee Cave (Stafford et al., 2008b), suggesting that solutionally aggressive fluids were delivered from below instead of laterally or above as a result of back flooding.

Lake McMillan is no longer an active reservoir, but has been drained and replaced by Brantley Lake farther downstream. Since its construction in 1893, Lake McMillan was constantly plagued with leakage problems, including the development of large sinkholes in the lake bed and massive leakage through karst conduits (Cox, 1967). It is likely that the entire region along the Pecos River near Coffee Cave contains abundant cavernous porosity created by hypogene processes, many of which provided direct bypass networks for the movement of fluids beneath the Lake McMillan dam.

In the early 20th century, many irrigation wells in the vicinity of Coffee Cave displayed strong artesian flow (Fiedler and Nye, 1933; Welder, 1983). Today, flowing artesian wells are less common in this area because of

reductions in hydraulic head due to decades of agricultural groundwater pumping. However, in 2007 water levels in Coffee Cave rose more than 2 meters between the summer irrigation season when groundwater pumping is most intense and mid-winter when most irrigation pumping has ceased (Stafford et al., 2008b). This rise in water levels suggests that hypogene processes are still active in interbedded evaporites and carbonates of the Seven Rivers Formation near the Pecos River.

Carlsbad Caverns National Park

The caves of Carlsbad Caverns National Park in the Guadalupe Mountains of southernmost New Mexico (Figure 2, 6) are the most studied karst features within the western United States (Hose and Pisarowicz, 2000). The caves contain large isolated chambers, maze sections and multiple levels, all characteristics of dissolution involving sluggish waters in a confined or semi-confined setting. Cavernous porosity is developed largely in the Permian age Capitan Formation (Figure 1), including the reef and fore reef sections, and extends into the equivalent near back reef facies (Hill, 2000). Massive accumulations of secondary gypsum in caves attest to the role of sulfuric acid (Palmer and Palmer, 2000), which increased solutional aggressivity. Regional tectonic studies indicate

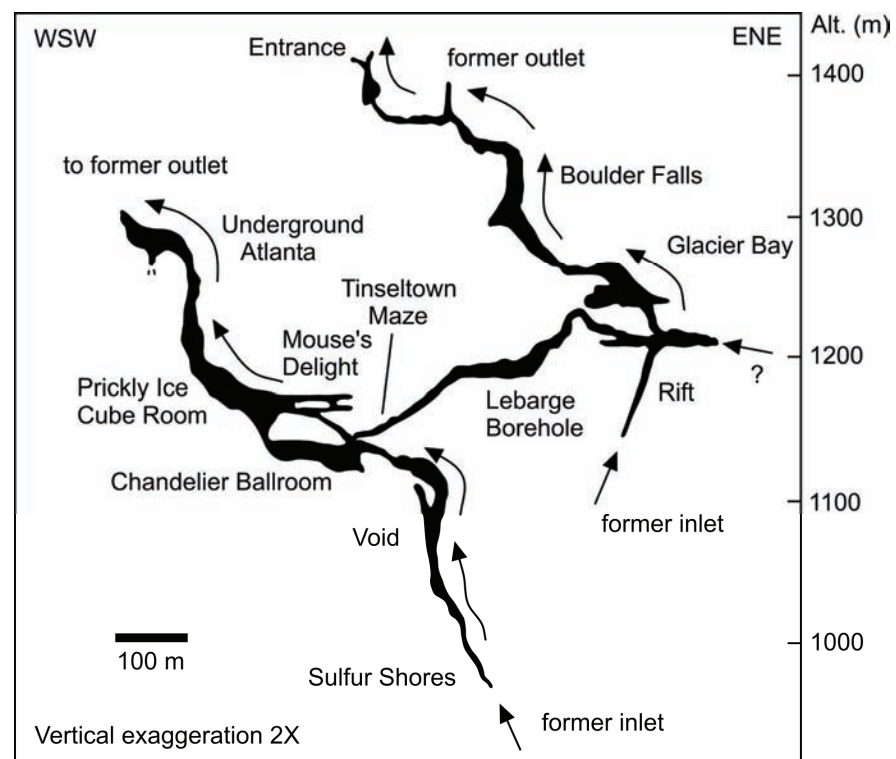


Figure 9. Simplified profile view through part of Lechuguilla Cave shows the flow paths of ascending fluids (from Palmer and Palmer, 2000).

that geothermal gradients were as high as 40-50°C/km (Barker and Pawlewicz, 1987), which must have added a thermal component to the hypogene speleogenesis of the Guadalupe Mountains.

Klimchouk (2007) and Palmer and Palmer (2000) suggest that most of the volume of major caves in the Guadalupe Mountains (e.g., Carlsbad Cavern, Lechuguilla Cave) is the result of speleogenesis in deep-seated conditions wherein rising solutional fluids are delivered from depth. Palmer and Palmer (2000) trace distinct flow paths through the caves that show fluids migrating from the deepest portions (e.g. Sulfur Shores in Lechuguilla Cave; Lake of the Clouds in Carlsbad Cavern), through intermediate passages, to the uppermost levels of the caves and even to the current cave entrances, which they interpret as outlets for ascending fluids (Figure 9). Klimchouk (2007) further shows abundant morphometric evidence throughout the caves of Carlsbad Caverns National Park that attests to dissolution within semi-confined, hypogene conditions. The intense hypogene dissolution was driven by steep density gradients established by thermal convection (Hill, 1996) and high solute loads resulting from increased dissolution by sulfuric acid-rich fluids. Hill (2000) concluded that sulfate reduction within the Castile Formation of the Delaware Basin interior was the source of hydrogen sulfide that produced the sulfuric acid-rich fluids. However, DuChene and Cunningham (2006) argue for hydrogen sulfide originating in the Artesia Group of the Northwest Shelf, which is supported by recent studies of evaporite calcitization within the Castile Formation (Stafford et al., 2008e).

While there is compelling evidence for hypogene origins for the Guadalupe Mountains caves, and specifically caves in Carlsbad Caverns National Park, most of the speleogenetic models overlook regional paleohydrology and the potentiometric driving forces that induced focused dissolution of hypogene caves within this region. Instead they have focused on the unique mineralogy and geochemistry associated with the sulfuric acid model of speleogenesis (Hose and Piarowicz, 2000). During the Neogene, the Pecos River was positioned farther west than it is today (Thomas, 1972) and since has migrated eastward, acting as a major, hydrologic driving force across the Capitan Reef massif. As a result, entrenchment and the erosion of several terraces occurred in the Guadalupe Mountains (Thomas, 1972), as is well marked by three distinct topographic levels that decrease in elevation from west to east over the length of the mountain ridge (DuChene and Martinez, 2000). While thermal and sulfuric acid components drove increased

dissolution rates, locally focused forced flow was likely induced by the ancestral Pecos River as it migrated eastward. This model is supported by age dates from alunite deposits in caves that show increasing age to the west with increasing elevation (Polyak and Provencio, 2000).

Delaware Basin Subsidence Troughs

Within the interior of the Delaware Basin, caves and solution subsidence features are common in Ochoan evaporites (Hill, 1996) (Figures 2, 6). Along the western edge of the Delaware Basin large subsidence troughs have been identified, associated with hypogenic fluids ascending from the subsurface Capitan Reef (Anderson and Kirkland, 1980). In the eastern Delaware Basin, extensive caves, evaporite calcitization and native sulfur accumulations are associated with hypogene processes primarily in the Castile Formation, but also extending into the overlying Rustler and Salado Formations (Stafford et al., 2008 c,d,e). In the middle of the Delaware Basin, several large subsidence features filled with Quaternary sediments occur along the direct flowpath of the Pecos River (Figure 6) (Malley and Huffington, 1953).

Anderson and Kirkland (1980) show that vertical breccia pipes occur throughout the Delaware Basin as a result of brine density convection combined with upward stoping. Their model was developed for features observed over the buried Capitan Reef on the eastern edge of the basin, and was extrapolated throughout the region to other breccia pipes. Earlier, Anderson et al. (1978) had also recognized widespread blanket brecciation throughout the Castile and Salado formations as a result of intrastratal dissolution of halite interbeds. Both breccia pipes and breccia blankets are the result of hypogene dissolution driven by steep density gradients established by differences in fluid saturation. These early observations indicate that the potential for extensive hypogene speleogenesis within evaporite facies of the basin interior is extremely high.

While the Pecos River has migrated laterally throughout the Delaware Basin since the early Paleogene, it has been located close to its current location since at least the late Neogene (Thomas, 1972). These large subsidence features appear to be the result of locally intense, intrastratal dissolution of Castile and Salado evaporites, resulting in intrastratal collapse and upward stoping to create large closed depressions that were subsequently filled by Quaternary sediments delivered by the Pecos River (Malley and Huffington, 1953). This intrastratal dissolution was driven by brine density free convection and forced convection,

with flow directed toward the potentiometric low of the Pecos River. Because of the high solubility of evaporites and the lack of upper Permian carbonate or clastic interbeds within the interior of the Delaware Basin, the development of large hypogene voids is possible.

It is possible that some subsidence features may have also developed along the ancestral Rio Grande, which is postulated to have flowed across the Delaware Basin during the Neogene as a tributary of the Pecos River (Thomas, 1972). It is likely that intrastratal dissolution is still occurring today throughout the Delaware Basin in association with the Pecos River and other natural hydrologic phenomena. However, modern catastrophic collapse features within this region are often attributed to anthropogenic causes such as leaky casing and improper well installation (e.g. Johnson et al., 2003; Powers, 2003), even though natural subsidence features and collapse structures permeate these evaporite units.

Yates Field

The Yates Unit Oil Field is located in eastern Pecos County, Texas (Figure 2) on the southeastern tip of the Central Basin Platform. It is the largest oil field in the Permian Basin and has been characterized as a karstic reservoir since the first well was drilled in 1926 (Hennen and Metcalf, 1929). Production is from middle Permian strata, primarily the upper San Andres

Formation but extending into the overlying Grayburg and Queen formations, at depths of approximately 300 meters (Stafford et al., 2008a). The field is subdivided into low permeability Westside Yates and high permeability Eastside Yates units (Figure 10). Structurally, the Yates Field is dominated by a horseshoe-shaped anticline, with an axial ridge centered in Eastside Yates and extending into the northern and southern edges of Westside Yates (Craig, 1988). Stratigraphically overlying the Yates Field along the northeastern edge, the Toborg Field is developed in uppermost Triassic and Cretaceous strata (Franklin, 1966). Both fields are located on the southwest side of the Pecos River and were initially discovered because of oil seeps along the river.

Cavernous porosity within the Yates Field includes open caves, solution enhanced fractures and extensive brecciation, as well as secondary mineralization including clastic sediment fill, calcite and dolomite spar, and more rarely native sulfur, albite, galena and sphalerite (Stafford et al., 2008a). Cave distribution within the field, based on petrophysical analyses and documentation of 1,566 individual cave occurrences, is highly clustered and tends to be focused along the structural axis of the field at the upper San Andres contact (Figure 10) (Stafford et al., 2008a). Isotope analyses of secondary calcite spar indicate deposition in thermal fluids associated with methane oxidation (Stafford et al., 2008a). While no Permian strata are

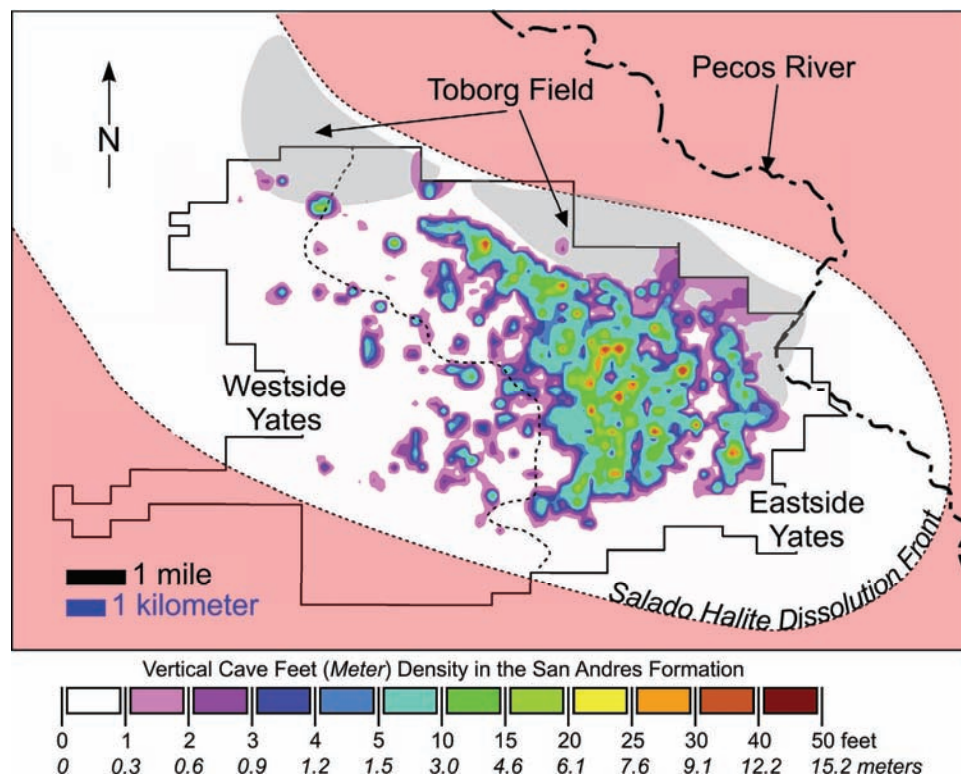


Figure 10. Distribution of caves within the San Andres Formation of the Yates Field in comparison to the overlying Salado Halite Dissolution Front, Toborg Field and Pecos River. Note that the most intense karst development forms an arch through Eastside Yates, which is the structural axis of the anticlinorium that dominates the region (adapted from Stafford et al., 2008a).

exposed at the surface in this region, in the overlying Cretaceous carbonates morphometric features observed in Ess Cave within the Yates Unit Oil Field indicate that hypogene processes have been locally active (Stafford et al., 2008a).

Early models for the karst porosity of the Yates Field invoked eogenetic, island karst processes for cave origins (Craig, 1988; Tinker and Mruk, 1995; Tinker et al., 1995). These models suggest that subaerial exposure at the end of San Andres deposition created a series of islands on the southern end of the Central Basin Platform in which freshwater horizons developed where caves formed as a result of mixing of fresh and saline waters. However, current studies (Stafford et al., 2008a) suggest that most of the cavernous porosity within the Yates Field is probably attributable to hypogene processes. The distribution of cavernous zones from petrophysical analyses shows karst development centered along the crest of the anticlinal structure that dominates Yates Field (Figure 10), not as an elliptical band of intense karst as would be expected from island karst dissolution along a coastal margin. It is significant that cavernous porosity is not limited to the San Andres Formation but extends upward into the Grayburg and Queen Formations (Tinker and Mruk, 1995; Tinker et al., 1995). Clastic sediments filling San Andres vugs are the same composition as the insoluble residue of the overlying strata (Tinker and Mruk, 1995), suggesting an autochthonous origin, in contrast to allogenic sediments derived from surface environments in an epigenic system. Secondary minerals filling vugs and lining fractures indicate thermal and sulfuric-acid components to the fluid history of the field (Stafford et al., 2008a). Associated with the Yates Field, overlying halite in the Salado Formation has been removed through intrastratal dissolution (Wessel, 2000), which is likely associated with upward fluid migration from Yates Field units. Finally, overlying Cretaceous rocks contain caves resulting from hypogene speleogenesis.

We contend that Yates Field developed through hypogene speleogenesis with dissolution focused along structural and lithologic boundaries. As fluids migrated upward toward the Pecos River, which has been persistently downcutting in this region since the early Paleogene, they passed through San Andres carbonates, overlying Permian strata and through Cretaceous carbonates in which hypogene caves have been locally documented. As solutionally aggressive fluids were replaced with hydrocarbons within the Yates Field during the Neogene, it is probable that solutional pathways created through the overlying interbedded evaporites and carbonates of the Seven

Rivers Formation began to close and provide a leaky seal for hydrocarbon entrapment. The Yates Field has a complex diagenetic history, but it appears that much of the cavernous porosity and the location of the field are related to proximity of the Pecos River.

Goodenough Spring – Amistad Reservoir

The distal end of the Pecos River reaches the Rio Grande in Val Verde County, Texas, at the International Amistad Reservoir, which was constructed in the 1960's impounding the Rio Grande, including the lowest reach of the Pecos River. The geology of Val Verde County is dominated by early-mid Cretaceous limestone of the Edwards Plateau, associated with the Maverick Basin. This large, closed marine basin developed as the Devils River Formation formed a reef bank surrounding basinal facies of the West Nueces, McKnight, and Salmon Peak Formations (Figure 11) during Fredericksburgian through Washitan time. The Salmon Peak and Devils River Formations are significantly karstified through this region of Texas and Mexico (Boghici, 2004; Barker et al., 1994).

The impoundment of Amistad Reservoir flooded numerous caves and springs along the Pecos and Rio Grande rivers. The most significant feature inundated was Goodenough Spring, 21 kilometers south-southeast of the confluence of the Pecos and Rio Grande (Figure 11). This spring was the third largest in Texas prior to inundation in 1968 with an average annual discharge of 3.9 meters³/second and a maximum recorded flow of 18.4 meters³/second (Brune, 1981). The water flowing from Goodenough Spring is 28° C, which is over 7 degrees above average surface temperatures and has low total dissolved solids of 208 mg/l (Kamps et al., 2009).

In the 1990's underwater cave exploration began at Goodenough Spring by cave divers using SCUBA. A tight restriction was encountered over 100 meters into the cave with very high water velocity hindering further exploration (Milhollin and Laird, 1996). This restriction was passed in 2004, and passage continued down at a steep angle. The passage was extended in 2008 to a water depth of 157 meters (GSEP, 2008). The reservoir water level was 45 meters over the spring cave entrance at this time, pushing the total explored depth within the cave to 112 meters (Figure 11), making it the third deepest cave in Texas, either air or water-filled (Elliott and Veni, 1994).

Goodenough Spring continues to flow significant volumes despite additional head above the cave. In 2005, over 2 meters³/second of flow was measured

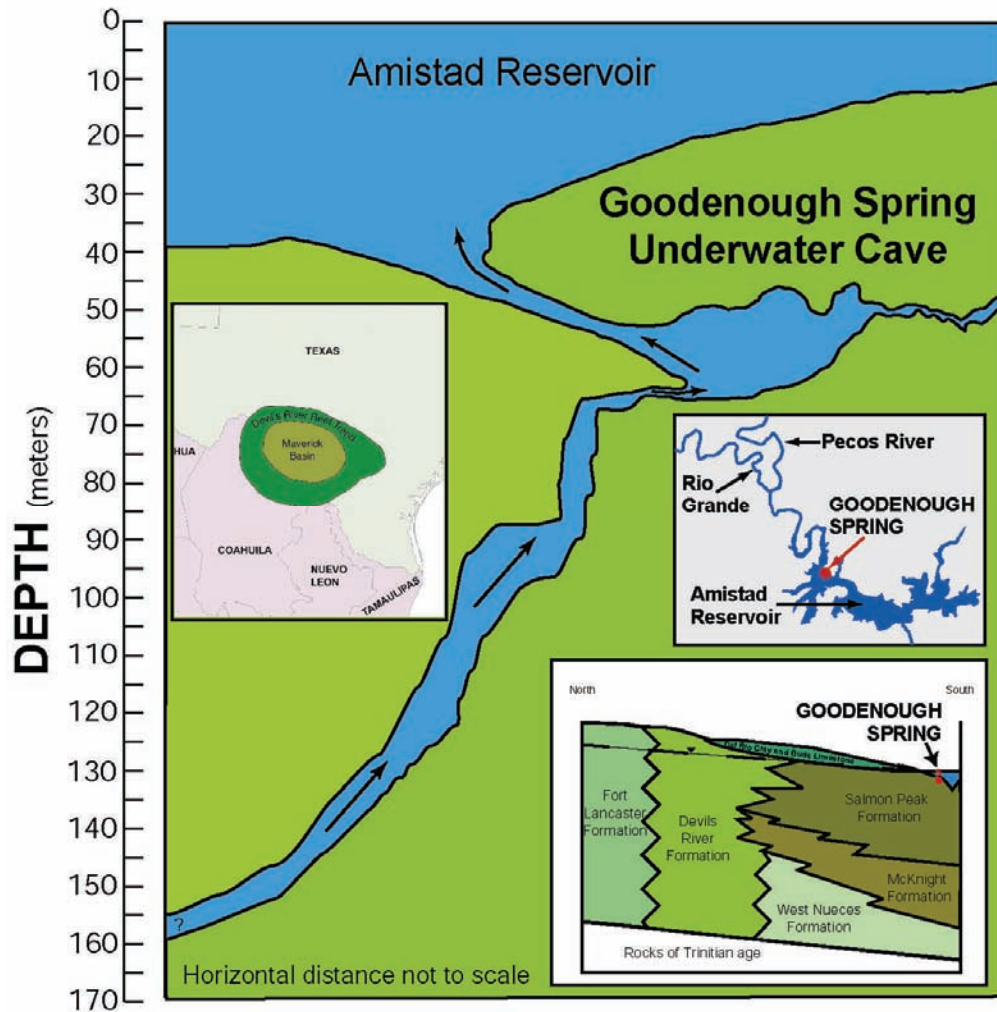


Figure 11. Profile cave map of Goodenough Spring underwater cave (main image) shows vertical extent of the artesian spring extending to water depths beyond 160 m. The water depth above the spring cave entrance is 45 m due to impounded water from Amistad Reservoir (upper right inset). The cave has formed in the Salmon Peak Formation (lower right inset), which is the top unit of the Maverick Basin (upper left inset).

from the spring (Kamps et al., 2009). Geologists active in the area prior to inundation of Goodenough Spring noticed responses in spring flow to precipitation events occurring to the south in Mexico, when no event occurred to the north in Texas (T.A. Small, 2000, personal communication). Uplifted limestone with extrusive igneous rocks 30 kilometers south of Goodenough Spring is hypothesized as a possible recharge zone. The extreme artesian conditions and elevated water temperatures indicate a deep groundwater flow route through conduits likely formed from hypogene karst processes, possibly influenced by subsurface volcanic activity. Entrenchment by the Pecos and Rio Grande rivers exposed the deep flow path of Goodenough Spring, creating the modern karst system.

Conclusions

The Pecos River has played a dominant role in shaping the geomorphology of eastern New Mexico and west Texas, but it has also had a profound impact on

the hydrologic and speleogenetic evolution of the region since the late Cretaceous. Karst features located in the Pecos River Valley suggest that the continuous potentiometric low this fluvial system created has been a major driving force for basin-scale groundwater movement throughout the Cenozoic Era. Modern artesian springs in eastern New Mexico attest to continued hypogene processes acting today in many areas of the Pecos River Basin. Relict caves in the Guadalupe Mountains signify intense periods of hypogene karst development associated with the ancestral Pecos River. Cavernous porosity in the Yates Field is probably only a small fraction of the cavernous porosity associated with Cenozoic entrenchment of the Pecos River in southwest Texas. It is likely that future research will demonstrate an even greater dominance of the Pecos River on speleogenetic processes throughout eastern New Mexico and west Texas.

References

- Anderson, R.Y., K.K. Kietzke, and D.J. Rhodes. 1978. Development of dissolution breccias, northern Delaware Basin and adjacent areas. In *New Mexico Bureau of Mines and Mineral Resources Circular 159*, ed. G. S. Austin, 47-52. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Anderson, R.Y., and D.W. Kirkland. 1980. Dissolution of salt deposits by brine density flow. *Geology* 8: 66-69.
- Barker, C.E., and M.J. Pawlewicz. 1987. The effects of igneous intrusions and higher heat flow on the thermal maturity of Leonardian and younger rocks, western Delaware Basin, Texas. In *The Leonardian Facies in W. Texas and S.E. New Mexico and Guidebook to the Glass Mountains, West Texas*, ed. L. Mazzullo and D. Cromwell, 69-84. Midland: Permian Basin Section – SEPM.
- Barker, R.A., P.W. Bush, and E.T. Baker. 1994. Geologic history and hydrogeologic setting of the Edwards-Trinity aquifer system, west-central Texas. In *U.S. Geological Survey Water Resources Investigations Report 94-4039*, 18-23. Boulder: U.S. Geological Survey.
- Barroll, P., and J. Shomaker. 2003. Regional hydrology of the Roswell Artesian Basin and the Capitan aquifer. In *Water resources of the lower Pecos region, New Mexico: New Mexico Bureau of Geology and Mineral Resources, 2003 New Mexico Decision Makers Guidebook*, ed. P. Johnson, L. Land, G. Price and F. Titus, 23-27. Socorro: New Mexico Bureau of Geology and Mineral Resources.
- Boghici, R. 2004. Hydrogeology of Trinity-Edwards aquifer of Texas and Coahuila in the border region. In *Aquifers of the Edwards Plateau Conference, Report 360 proceedings*, ed. R. E. Mace, E. S. Angle and W. F. Mullican III, 91-114. Austin: Texas Water Development Board.
- Bretz, J.H., and L. Horberg. 1949. The Ogallala Formation west of Llano Estacado. *Journal of Geology* 57: 447-490.
- Brune, Gunnar. 1981. *Springs of Texas, vol. 1*. College Station: Texas A & M University Press.
- Cox, E.R. 1967. *Geology and hydrology between Lake McMillan and Carlsbad Springs, Eddy County, New Mexico*. U.S. Geological Survey, Water Supply Paper 1828. Boulder: U.S. Geological Survey.
- Craig, D.H. 1988. Caves and other features of Permian karst in San Andres dolomite, Yates Field reservoir, west Texas. In *Paleokarst*, ed. N.P. James and P.W. Choquette, 342-363. New York: Springer.
- DuChene, H.R., and K.I. Cunningham. 2006. Tectonic influences on speleogenesis in the Guadalupe Mountains. In *Caves and Karst of Southeastern New Mexico: New Mexico Geological Society Guidebook 57*, ed. L. Land, V.W. Lueth, W. Raatz, P. Boston and D. Love, 211-218. Socorro: New Mexico Geological Society.
- DuChene, H.R., and R. Martinez. 2000. Post-speleogenetic erosion and its effect on cave development in the Guadalupe Mountains, New Mexico and west Texas. *Journal of Cave and Karst Studies* 62 (2): 25-29.
- Eaton, J., ed. 1987. *GYPKAP 1987 Annual Report*. Alamogordo: Southwestern Region of the National Speleological Society.
- Elliott, William R., and George Veni. 1994. *The Caves and Karst of Texas*. Huntsville: National Speleological Society.
- Fiedler, A.G., and W.S. Nye. 1933. *Geology and ground-water resources of the Roswell Basin, New Mexico*. U.S. Geological Survey Water-Supply Paper 639. Boulder: U.S. Geological Survey.
- Franklin, D.W. 1966. *Yates Field, Pecos County, Texas*. Midland: Marathon Oil Company, Unpublished internal report.
- Galloway, S.E. 1956. *Geology and Ground-Water Resources of the Portales Valley Area, Roosevelt and Curry Counties, New Mexico*. M.S. thesis, University of New Mexico.
- GSEP (Goodenough Springs Exploration Project). 2008. News Release. <http://www.goodenoughsprings.org/newsreleases.htm> (accessed January 26, 2009).
- Hennen, R.V., and R.J. Metcalf. 1929. Yates oil pool, Pecos County, Texas. *American Association of Petroleum Geologists Bulletin* 13 (12): 1509-1556.
- Hill, C.A. 1996. *Geology of the Delaware Basin, Guadalupe, Apache and Glass Mountains*. New Mexico and West Texas. Midland: Permian Basin Section – SEPM.
- Hill, C.A. 2000. Overview of geologic history of cave development in the Guadalupe Mountains, New Mexico and west Texas. *Journal of Cave and Karst Studies* 62 (1): 10-21.
- Hose, L., and J.A. Pisarowicz, ed. 2000. The Caves of the Guadalupe Mountains. *Journal of Cave and Karst Studies* 62 (2).
- Johnson, K.S., E.W. Collins, and S.J. Seni. 2003. Sinkholes and land subsidence owing to salt dissolution near Wink Sink, West Texas, and other sites in western Texas and New Mexico. In *Evaporite karst and engineering / environmental problems in the United States*, ed. K. S. Johnson and J. T. Neal, 183-196. Norman: Oklahoma Geological Survey.
- Kamps, R.H., G.S. Tatum, M. Gault, and A. Groeger. 2009. Goodenough Spring, Texas, USA: Discharge

- and water chemistry of a large spring deeply submerged under the binational Amistad Reservoir. *Hydrogeology Journal* 17 (4): 891-899.
- Kelley, V.C. 1971. *Geology of the Pecos Country, southeastern New Mexico*. Socorro. New Mexico Bureau of Mines and Mineral Resources.
- Kelley, V.C. 1972. Geology of the Santa Rosa Area. In *New Mexico Geological Society Fall Field Conference Guidebook 23: East-Central New Mexico*, ed. V. C. Kelly and F. D. Trauger, 218-220. Socorro: New Mexico Geological Society.
- Klimchouk, A.B. 1997. The role of karst in the genesis of sulfur deposits, Pre-Carpathian region, Ukraine. *Environmental Geology*, 31 (1/2): 1-20.
- Klimchouk, Alexander B. 2007. *Hypogene speleogenesis: hydrological and morphogenetic perspective*. National Cave and Karst Research Institute Special Paper No. 1. Carlsbad: National Cave and Karst Research Institute.
- Kunath, Carl E., ed. 1995. *The Caves of Carta Valley*. Austin: Texas Speleological Survey.
- Kunath, Carl E. 1978. *The Caves of McKittrick Hill, Eddy County, New Mexico*. Austin: Texas Speleological Survey.
- Land, L. 2003. Evaporite karst and regional ground water circulation in the lower Pecos Valley. In *Evaporite Karst and Engineering/Environmental Problems in the United States: Oklahoma Geological Survey Circular 109*, ed. K. S. Johnson and J. T. Neal, 227-232. Norman: Oklahoma Geological Survey.
- Land, L. 2006. Hydrogeology of Bottomless Lakes State Park. In *Caves and karst of southeastern New Mexico: New Mexico Geological Society Guidebook 57*, ed. L. Land, V. W. Lueth, W. Raatz, P. Boston and D. Love, 93-94. Socorro: New Mexico Geological Society.
- Land, L., and B.T. Newton. 2007. *Seasonal and long-term variations in hydraulic head in a karstic aquifer: Roswell Artesian Basin, New Mexico: New Mexico Bureau of Geology and Mineral Resources Open-File Report no. 503*. Socorro: New Mexico Bureau of Geology and Mineral Resources.
- Land, L., and B.T. Newton. 2008. Seasonal and long-term variations in hydraulic head in a karstic aquifer: Roswell Artesian Basin, New Mexico. *Journal of the American Water Resources Association* 44: 175-191.
- Lee, J., ed. 1996. *GYPKAP Report Volume 3*. Alamo-gordo: Southwestern Region of the National Speleological Society.
- Maley, V.C., and R.M. Huffington. 1953. Cenozoic fill and evaporite solution in the Delaware Basin, Texas and New Mexico. *Geological Society of America Bulletin* 64: 539-546.
- Martinez, J.D., K.S. Johnson, and J. T. Neal. 1998. Sinkholes in evaporite rocks. *American Scientist* 86: 38-51.
- McLemore, V.T. 1989. Santa Rosa Lake. *New Mexico Geology* 11 (1): 8-10.
- Milhollin, R.D., and R. Laird. 1996. *Goodenough Springs Survey and Exploration Report # 2*, prepared for U.S. National Park Service Amistad National Recreation Area.
- Palmer, A.N., and M.V. Palmer. 2000. Hydrochemical interpretation of cave patterns in the Guadalupe Mountains, New Mexico. *Journal of Cave and Karst Studies* 62 (2): 41-58.
- Polyak, V. J., and P. P. Provencio. 2000. Summary of the timing of sulfuric acid speleogenesis for the Guadalupe Mountains, New Mexico and west Texas. *Journal of Cave and Karst Studies* 62 (2): 22-24.
- Powers, D.W. 2003. Jal Sinkhole in southeastern New Mexico: Evaporite dissolution, drill holes, and the potential for sinkhole development. In *Evaporite karst and engineering/environmental problems in the United States: Oklahoma Geological Survey Circular 109*, ed. K. S. Johnson and J. T. Neal, 211-218. Norman: Oklahoma Geological Survey.
- Reeves, C.C., Jr. 1972. Tertiary-Quaternary stratigraphy and geomorphology of west Texas and southeastern New Mexico. In *East-Central New Mexico: New Mexico Geological Society Guidebook 23*, ed. V. C. Kelley and F. D. Trauger, 108-117. Socorro: New Mexico Geological Society.
- Richey, S.F., J.G. Wells, and K.T. Stephens. 1985. *Geohydrology of the Delaware Basin and vicinity, Texas and New Mexico*: U.S. Geological Survey Open-file Report 88-4502. Boulder: U.S. Geological Survey.
- Scholle, P.A., R.H. Goldstein, and D.S. Ulmer-Scholle. 2004. *Classic upper Paleozoic reefs and bioherms of west Texas and New Mexico*. Socorro: New Mexico Institute of Mining and Technology.
- Stafford, K.W., F. Behnken, and J. White. 2008. Hypogene speleogenesis within the Central Basin Platform: Karst porosity in the Yates field, Pecos County, Texas, U.S.A. In *Karst from recent to reservoirs, Karst Waters Institute Special Publication 14*, ed. I. Sasowsky, C. Feazel, J. Mylroie, A. Palmer and M. Palmer, 174-178. Leesburg: Karst Waters Institute, Inc.
- Stafford, K.W., L. Land, and A. Klimchouk. 2008. Hypogenic speleogenesis within Seven Rivers evaporites: Coffee Cave, Eddy County, New Mexico. *Journal of Cave and Karst Studies* 70 (1): 47-61.
- Stafford, K.W., and R. Nance. 2009. Evaporite speleogenesis of the Gypsum Plain: New Mexico and far

- west Texas. In *The caves and karst of the USA*, ed. A. Palmer and M. Palmer. Huntsville: National Speleological Society (in press).
- Stafford, K.W., R. Nance, L. Rosales-Lagarde, and P. Boston. 2008. Castile evaporite karst potential map of the Gypsum Plain, Eddy County, New Mexico and Culberson County, Texas: A GIS methodological comparison. *Journal of Cave and Karst Studies* 70 (1): 35-46.
- Stafford, K.W., L. Rosales-Lagarde, and P. Boston. 2008. Epigene and hypogene gypsum karst manifestations of the Castile Formation: Eddy County, New Mexico and Culberson County, Texas, USA. *International Journal of Speleology* 37 (2): 83-98.
- Stafford, K.W., D. Ulmer-Scholle, and L. Rosales-Lagarde. 2008. Hypogene calcitization: Evaporite diagenesis in the western Delaware Basin. *Carbonates and Evaporites* 23 (2): 89-103.
- Thomas, R.G. 1972. *The Geomorphic evolution of the Pecos River system: Baylor Geological Studies, Bulletin No. 22*. Waco: Baylor University.
- Tinker, S.W., J.R. Ehret, and M.D. Brondos. 1995. Multiple karst events related to stratigraphic cyclicity: San Andres Formation, Yates Field, west Texas. In *Unconformities and porosity in carbonate strata: AAPG Memoir 63*, ed. D. A. Budd, A. H. Staller and P. M. Harris, 213-237. Tulsa: American Association of Petroleum Geologists.
- Tinker, S.C., and D.H. Mruk. 1995. Reservoir characterization of a Permian Giant: Yates Field, west Texas. In *Hydrocarbon reservoir characterization – Geologic framework flow unit modeling: SEPM short course No. 34*, ed. E.L. Stoudt and P.M. Harris, 51-128. Tulsa: Society for Sedimentary Geology.
- Welder, G.E. 1983. *Geohydrologic framework of the Roswell ground-water basin, Chaves and Eddy Counties, New Mexico*. New Mexico State Engineer Technical Report 42. Santa Fe: New Mexico State Engineer.
- Wessel, G.R. 2000. Stratigraphic relationships in the Yates Field area, Pecos and Crockett Counties, Texas – Part II. *West Texas Geological Society Bulletin* 41 (7): 4-8.

CARBONATE-HOSTED MASSIVE SULFIDE DEPOSITS AND HYPOGENE SPELEOGENESIS: A CASE STUDY FROM NANISVIK ZINC / LEAD MINE, BAFFIN ISLAND, CANADA

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Abstract

Nanisivik (Inuit – “the place where they find things”) zinc/lead mine is located at Lat. 73° N in northwestern Baffin Island. The host rock is a Proterozoic platform carbonate 260-800 m thick, medium to massively bedded and pervasively dolomitized. It rests on mixed shales and shaly dolomites, and is overlain by 150+ m of further shales functioning as an aquitard. These formations were buried by later Proterozoic strata, uplifted, eroded and buried again in a Cambrian sedimentary basin. The ore-grade deposits are contained within a horst block of the dolomites dipping NW at 15° across it. Graben to the north and south are roofed in the overlying shales.

The principal deposit, the Main Ore, is of zinc, lead and iron sulfide precipitates plus gangue minerals, chiefly secondary dolomite. It extends for three km E-W along the horst. It is horizontal, at ~300 m above sea level and terminated at both ends by modern valley entrenchments. The Main Ore body is consistently ~100 m in width and five-seven m in depth. This wide ceiling is a nearly planar, horizontal corrosion bevel. The sulfides scarcely extend above it anywhere. Within the Main Ore two or more generations of tapered fins of dolomite *in situ* extend from both south (updip) and north (downdip) walls into the cavity. Fin surfaces truncate the bedding. Edges of fins are sinuous, some meandering with a wavelength of ~50 m. Very sharp, horizontal corrosion notches 20-30 cm high extend into the dolomite walls for at least 20 m (the limit of deep crosscuts in the mine). They are filled with layered pyrites which continue out into the ore body as regular sheets truncating earlier, dipping mineral layers until they themselves are truncated by later fillings. One exceptional notch, one meter deep, is at least 350 m in breadth. The ore displays four sedimentary modes: (i) regular layers settled or precipitated onto the cavity floor; (ii) chaotic polymict breccias suggestive of channel cut-and-fill episodes; (iii) the horizontal pyrite sheets in corrosion notches; (iv) minor metasomatic replacements of dolomite.

The ore cavity was created by paragenesis in a channel flow mode, with ore and gangue deposition on the floor taking place in tandem with dissolutional cavity creation upwards. Principal deposition took place when a fluid interface could be rigorously maintained. Fluid inclusions indicate derivation of the metals from exchange reactions with metalliferous sediments (the underlying shales), indicating low water/rock ratios and moderate temperatures. The ore fluids were similar to oilfield brines. Sulfur isotope fractionations indicate temperatures of 90-150 +/-40° C, suggesting that the Main Ore formed along a gas/brine interface at a depth of at least 1600 m as a consequence of fluid expulsion in the subsiding Cambrian sedimentary basin.

Introduction

Carbonate rocks, chiefly dolomites, host a large proportion of the world's economic deposits of zinc, lead and associated elements, in the form of massive accumulations of sulfide minerals that are intermingled with pene-contemporaneous (“secondary”) precipitates of dolomite. Historically, there have been two principal views on the origin of these deposits: first, that they are ‘epigene’, implying that the ores are fillings of pre-existing karst voids and were deposited by mechanical inwash or chemical precipitation from surface waters passing underground, much like the accumulation of bauxite ore in sinkholes or similar traps; alternatively, that they are ‘hypogene’, precipitated from solutions flowing upwards or laterally into the carbonate formations, often having derived from (or passed through) other types of rocks such as shales that supplied the metals, and arriving via fractures. Although a few deposits may be genuinely epigene, today it is widely recognized that some form of hypogene origin applies to most of those that have ores in economic concentrations (Dzulynski and Sass-Gutkiewicz, 1989).

There remains controversy on several important questions concerning the hypogene creation and morphology of the ore bodies - do the precipitating fluids (i) invade and infill pre-existing, open karst features, perhaps modifying their form a little by

minor solution and collapse of bedrock? or (ii) is the morphology largely or entirely created by the dissolution and precipitation activities of the ore-bearing fluids themselves ('syngenetic development')? If the latter is true, is the process (at one extreme) one of progressive molecule by molecule replacement of dolomite by sulfide as the fluid permeates the rock widely, a process known as "metasomatism"? – or, at the other extreme, does the fluid first open a cave by some form of channel or chamber dissolution and then infill it with the ore and secondary (gangue) minerals?

Most of the hypogene ore bodies in carbonate rocks display one or two of three basic morphologies. In the West, the more common of these were first described in detail in the Tri-State mining region west of the Mississippi River and so are known as "Mississippi Valley type" deposits (MVT). The global distribution of well studied MVT mines is shown in Figure 1 (from Bradley and Leach, 2003). The first morphologic type is a domelike ('beehive-shaped' – Sass-Gutkiewicz, 1974) breakdown chamber, commonly tens to a few hundred meters in diameter. The ore fills interstices in the breakdown and may exhibit a variety of sedimentary and crystalline forms (e.g. Sass-Gutkiewicz, op. cit; Sangster, 1988; Loucks et al. 2004). The second morphology is a reticulate pattern of long, narrow and quite high, corridors of breakdown that are filled with the ore and gangue: this is very similar to the patterns seen in many open maze caves, and in a very few instances there are some truly open passages mingled with the ore galleries (e.g. Devil's Hole Mine, England; Ryder, 1975). The third, less common, morphology is described by Sangster (op. cit.) as 'columnar', being curvilinear or irregular breakdown bodies that can vary considerably in width, and are usually discontinuous. Sometimes they occur in arcuate groupings. The principal example of this morphology is to be found in the Pine Point zinc/lead mines along the south shore of Great Slave Lake, Canada (Rhodes et al. 1984): today I would interpret much of the form there as a product of ore fluid invasion and enlargement of the reefal flank margin type of caves described by Mylroie and Carew (1990) in the modern Bahamas. To conclude this brief review it is important to stress that in all three morphologies the ores

occur chiefly as interstitial fillings in bedrock breakdown. Sangster (op. cit.) emphasizes the additional point that there is usually very little alteration of this bedrock by metasomatic processes, merely narrow rims in a few places.

One of the designated MVT sites in Figure 1 is Nanisivik zinc/lead mine, northwestern Baffin Island, Canada. The author had the opportunity to investigate it in some detail in 1981 and 1982, shortly after it had opened, and later co-supervised a PhD study there (Ghazban, 1988). The morphology of the "Main Ore" (the principal ore body) differed quite markedly from the MVT types summarized above because it displays sinuous, meandering channel flow forms such as are common in meteoric water caves, especially those in the most soluble rocks, gypsum and salt. There was little breakdown. It is presented and interpreted here as an example of hypogene speleogenesis with syngenetic ore deposition, in a setting that is paragenetic and hyperacidic.

Nanisivik geography and geology

Nanisivik (Inuit – "the place where they find things") zinc/lead mine is located close to the shore of Strathcona Sound, a fiord at Lat. 73° N in northwestern Baffin Island. The mine operated between 1976 and 2002, producing ~500,000 – 600,000 tons of ore per year, reduced to 120,000 – 140,000 tons of concentrate for shipping. The pertinent geographical and geological features are shown in Figures 2 and 3.

There is a horst and graben topography that has been much scoured by glaciers of the Island ice caps. The



Figure 1. A world map of known MVT deposits, from Bradley and Leach, 2003. The Nanisivik deposit is Number 4, in northwest Baffin Island.

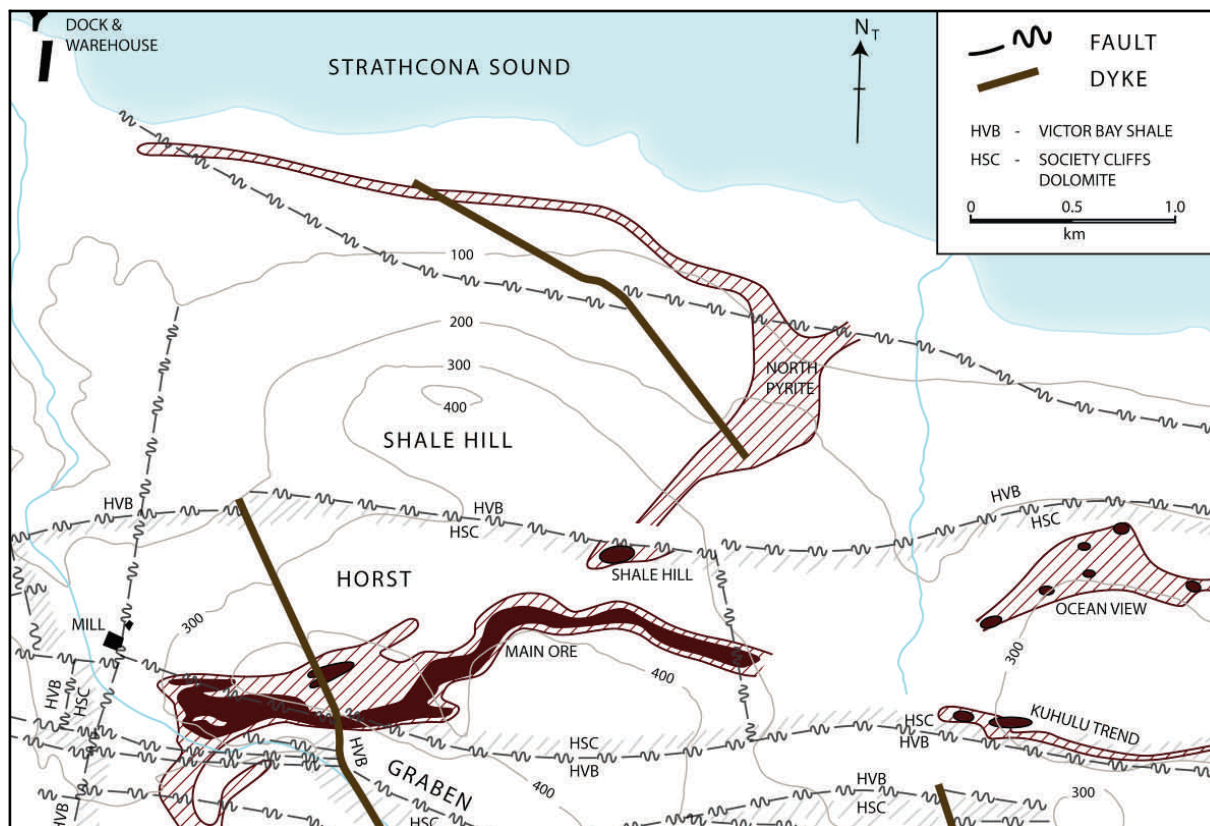


Figure 2. Plan view of the pyrite and zinc-lead deposits at Nanisivik, Baffin Island, Canada. Sulfide bodies are indicated by cross-hatching, with ore-grade deposits shown in deeper red. HVB, HSC delineate outcrop of the principal geological formations. Contour interval is in meters above sea level.

Main Ore and other economic deposits are contained within a horst block that rises to a little above 400 m above sea level today. Two valleys have been eroded across the horst, truncating the Main Ore at its west end and separating it from a pair of smaller deposits in the east, the Ocean View and Kuhulu Trends. There is an arctic climate, with permafrost present to depths greater than 100 meters today and a temperature of -12°C in the mine.

The basement rocks are granites and gneisses of Archaean age, now exposed to the south by the block faulting (Figure 3). They were unconformably overlain by thick sequences of clastic and carbonate sediments of middle Proterozoic age (1200-1000 Ma). The host rock is the Society Cliffs Formation (HSC), a subtidal to tidal algal platform carbonate containing some gypsum and organic interbeds. It displays evidence of occasional marine emergent epikarstification during its accumulation

(Ghazban et al. 1993). The formation is ~460 m thick in this region, medium to massively bedded, and pervasively dolomitized. It rests on >900 m of mixed shales, siltstones, sandstones and impure dolomites (HAB – the Arctic Bay Fm) and is overlain by the Victor Bay Fm (HVB), 150+ m of shales and minor limestones that function as an aquitard or aquiclude. These formations were buried by >1000 m of later Proterozoic

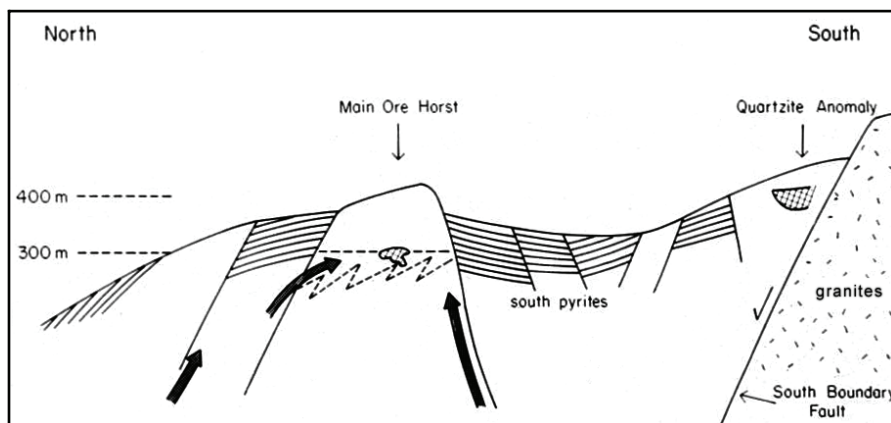


Figure 3. Sketch cross-section of the geological structure along a N-S transect at Nanisivik (Baffin Island, Canada), showing location of the principal sulfide bodies.

strata, then uplifted with some local folding and block faulting as shown in Figure 3. Considerable erosion followed before re-burial and the unconformable deposition of >500 m of Cambrian sedimentary rocks. In an early geological report, Clayton and Thorpe (1982) asserted that all of the formations were then cut by gabbro dykes which they dated at 476 Ma and which also intersect and displace the ore bodies, indicating that the latter must be older: both the age of the dyke and its displacing effects are now disputed (Sherlock et al. 2004 - see below).

In the Main Ore horst block the Society Cliffs dolomites dip consistently to the northwest at an angle of 15°, across the trend of the ore bodies. Graben to the north and south are roofed in the Victor Bay shales (Figure 3).

Morphologic features: Main Ore Mine *The Main Ore body*

The Main Ore extends for three km E-W along the horst (Figure 2). It is horizontal, oriented close to the strike of the steeply dipping dolomites but sinuous so that it meanders discordantly across any strictly strike-oriented path. It is ~300 m above sea level and terminated at both ends by the modern valleys entrenched across it. Ocean View and the Kuhulu Trend, the two

lesser bodies to the east, are preserved at the same elevation on the further side of the valley. To the north a narrow, elongated ore body, the North Pyrite, has been mapped by exploration drilling as ascending at the Society Cliffs – Victor Bay contact through a series of graben until it enters the horst at the same elevation as the Main Ore; only at that elevation do its concentrations of sphalerite and galena attain the ore grade. To the south of the horst further, minor and below-grade, pyrite bodies extend to higher elevations (Figure 3).

The Main Ore body is a quite remarkable piece of underground morphology. Figure 4 attempts to display this with a series of cross-sections. The body is consistently ~100 m in width and five-seven m in height from ceiling to floor. The ceiling is a nearly planar, horizontal bevel that does not differ in its elevation of ~300 meters above sea level by as much as two meters except where it is displaced a little by the gabbro dykes (Clayton and Thorpe 1982). This ceiling has not collapsed despite being a great, flat span cross-cutting dolomite beds that were dipping in cantilever array with respect to it. The sulfides scarcely extend above it anywhere. When inspected by the author in 1981 and 1982 a few small (half meter-sized) solution pockets could be seen in one

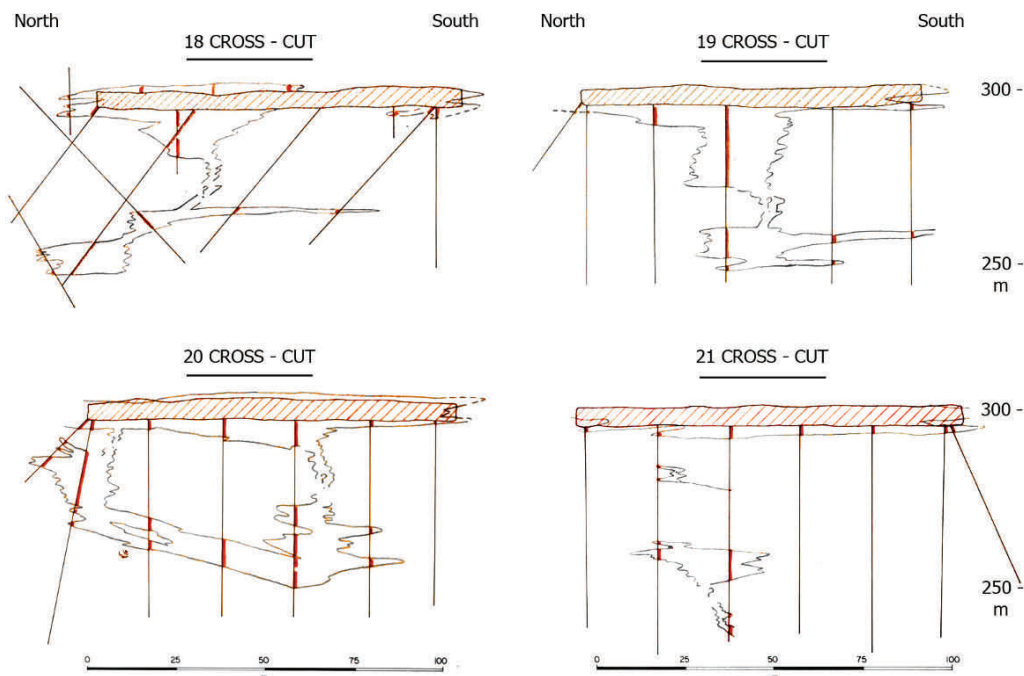


Figure 4. The form of the Nanisivik Main Ore body and underlying manto pyrite deposits at a sample of mined adits (cross-cuts, shown in red hatching) across the East-West trend of the ore body that, in each case, extended to or very close to the limits of the ore. Location of exploratory, six inch drill core holes below the Main Ore are shown in black, with the presence of ore and gangue minerals marked in red; the morphology of the manto deposits that is shown are the mine geologists' interpretations of these drill core interceptions.

small area of the mine ceiling; they followed dip or strike joints and did not appear to contain any large amounts of pyrite or exhalative precipitates. They may be recent features but below one example mining operations had left a sloping fragment of corrosional surface (the root of a fin – see below) with a section of layered clay resting upon it that looked very much like flood clay on any ledge in a meteoric water cave: the clay appeared altered (blanched and hardened) but was not accessible for sampling to confirm this. A particularly striking feature was the apparent absence of any substantial quantities of breakdown anywhere in or under the sulfide ore masses, with one exception - for two hundred meters along the southwest side of the mine, both bedrock and ore had collapsed in large blocks that were now firmly cemented by large masses of ground ice; this collapse is attributed to Quaternary sub-glacial dissolution, long after the ore was deposited (Ford 1987).

Irregularly along its length the Main Ore mass is underlain by a “keel” or “manto” (mine geologists’ expressions) of below-grade ore that is up to ten meters or more in depth (Figure 4). However, this keel

is discontinuous, as are other quasi-horizontal bodies detected by drilling that, where they are encountered, are rather consistently centered at depths of 40 - 50 m below the Main Ore. Olson (1984) interpreted the keel as a later stage vadose entrenchment beneath a Main Ore phreatic cavity and likened the pair to the morphology of some of the major passages in the Mammoth Cave System, Kentucky. However, there is nothing like the Main Ore cavity (if such it was) in Mammoth Cave and the discontinuous nature of the keel appears to rule out continuous vadose entrenchment. An alternative explanation is offered below.

Fins and corrosion notches in the Main Ore

Two striking features in the Main Ore are the fins and the mineral-filled corrosion notches found there. The fins are considered first, and illustrated in Figures 5-8. Fins of bedrock projecting into a passage are common features in meteoric water caves. Figure 5A shows the typical relationship between rock and channel that may be expected there – the fins are concordant to the

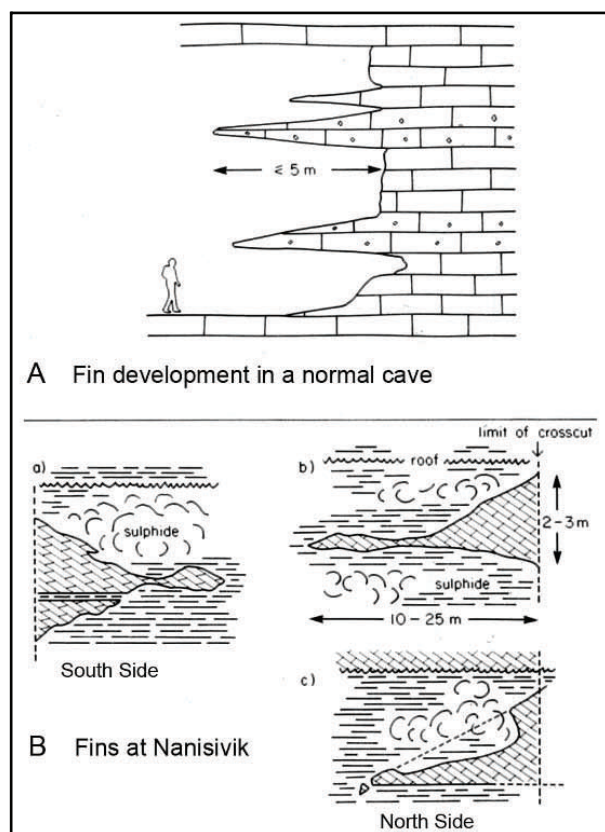


Figure 5. The morphology of bedrock fins in typical meteoric water caves (A) compared to examples of dolomite fins in the Nanisivik Main Ore that were measured by the author in 1981 and 1982 (B).

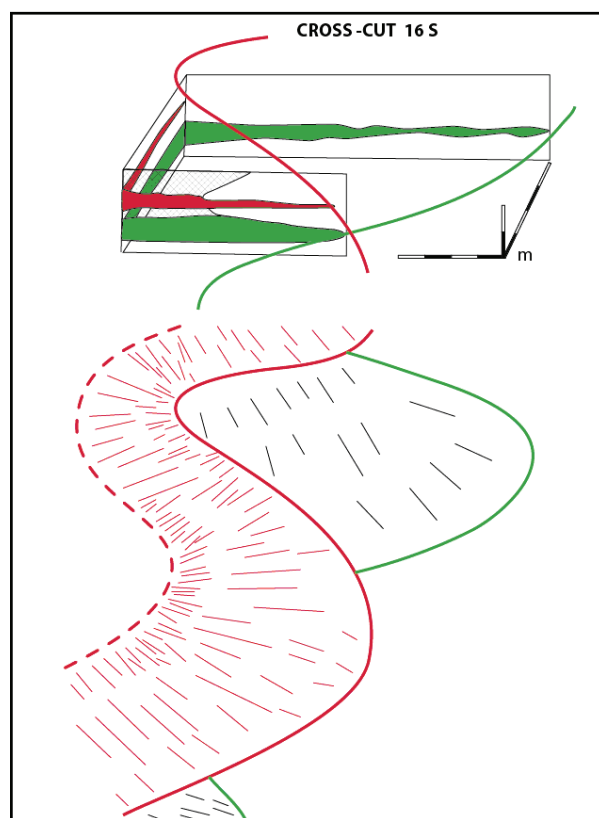


Figure 6. Schematic drawing to illustrate the meandering forms of a pair of bedrock fins (lower in green, upper in red) measured in 1981 at Cross-cut 16 and east of it, Nanisivik Main Ore. Scale bar at the cross-cut is marked at one meter intervals. The true wavelengths of the pair of meanders in the lower sketch were about 50 meters.

bedding, it is the slightly less soluble beds that project, and projection is rarely greater than one or two meters. In contrast, within the Main Ore, fins of dolomite *in situ* extend from both south (updip) and north (downdip) walls into the cavity (Figure 5B; Ford 1985). Their form varies in detail but all are tapered. Some extend more than 20 m in from the boundary walls of dolomite. Note that fin surfaces truncate the bedding. In plan view the edges of the fins are sinuous or meandering, with a meander wavelength of approximately 50 m (Figure 6). At several places in the mine in 1981 and 1982 it could be seen that there were (or had been before the mining) at least two generations of fins one above the other and with the meanders out of phase with each other, as can sometimes be seen in salt mine cavities that have been subjected to occasional flooding (Figure 7).

The meandering, tapered form of the fins leaves little doubt that they were fluid dynamic flow forms. Their large size and discordant relationships with the dolomite bedding imply that they should be very fragile, requiring mechanical support from either the continuous buoyant presence of dense fluid, or solid underpinning with co-precipitated ore and gangue minerals, or both. Figure 8 shows the author's proposed explanation of their formation. They are remains of intersecting 'paragenetic' cavities. Standard models of paragenesis in meteoric water caves are shown in Figure 8A: the depositional environment is one that permits net accumulation of insoluble sediments on the floor, causing the cave roof to be dissolved upwards to maintain the channel cross-sectional area, until a water table is intersected (Ford and Williams, 2007; page 232). In Figure 8A steady dissolution



Figure 7. Upper left. The tip of a small dolomite fin overlain and underlain by Type i ore deposits, Main Ore, Nanisivik. Note the very sharp upper contact between dolomite and ore. There is some Type iv metasomatic alteration dolomite (bright white) on the underside. Patches of ground ice are seen in the ore above the fin. The scale bar is graduated in decimeters. Lower left. A display of Type ii chaotic ore in the central flow channel of the Main Ore, overlain by white sparry dolomite at the top. Right. Waterline corrosion notches, fins and ceiling bevels in the Cardona Salt Mine, Spain. Notching and beveling are completely discordant with tight concertina folding in the diapiric salt (photo courtesy of J. Cardona).

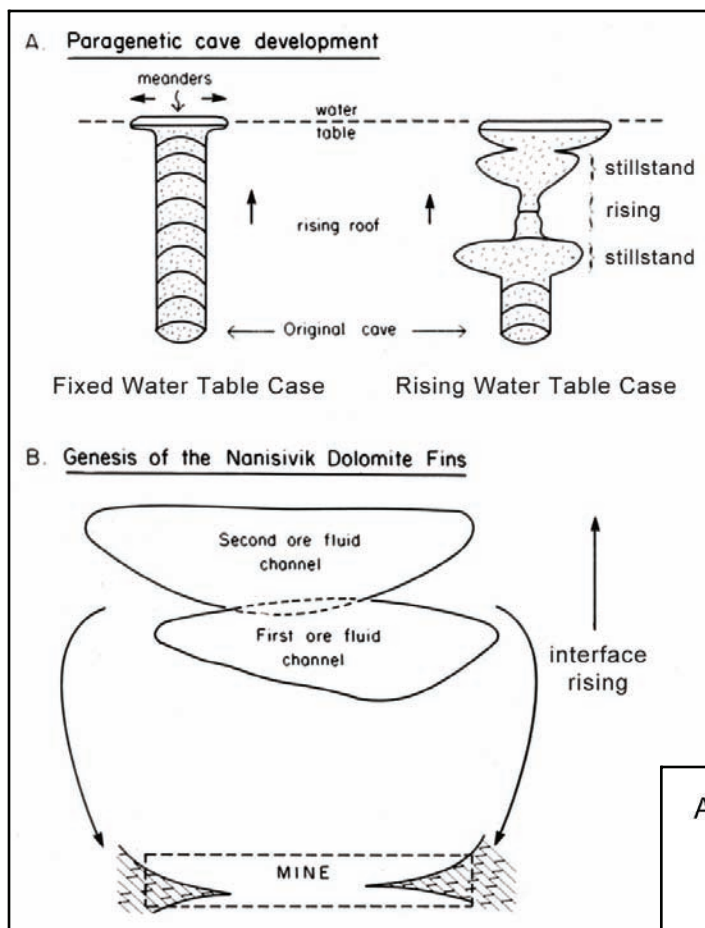


Figure 8. A) The development of paragenetic cave morphology towards water table in meteoric water caves: on left, where the water table is fixed throughout evolution; on right, where it is shifted upwards episodically. B) The author's interpretation of the genesis of the fins seen at Nanisivik Mine.

upwards to one fixed water table is shown on the left and, on the right, the more complex case where the water table itself shifts upwards from time to time, as it might do, for example, in the flanks of a river valley that is being aggraded. The meandering fin morphology in the Main Ore walls was created by the overlap of two successive paragenetic channels in an upward shifting interface scenario (Figure 8B). Contemporary precipitation of ore and gangue minerals may have helped to support portions of the roof spans. The nature of the fluid interface itself at Nanisivik (whether a water table or other) will be considered later.

Corrosion notching at a water line is a common feature in meteoric water caves. Typical examples are illustrated in Figure 9. At its most simple, a notch may form at the water line where there is a slow drip into a

pool that introduces fresh carbonic acid (H_2CO_3) to its surface (Figure 9A). In limestone caves the author has seen corrosion notches of this kind that are as much as two meters deep (Figure 9B). Where the water is ponded behind some natural dam entire sections of submerged cave roof may be dissolved up to the water line, creating a ceiling bevel (Figure 9C). More extreme notching is known to be created where H_2S from hypogene sources is released at or just below a water table, the Big Room at Carlsbad Caverns (one of the world's largest known open cave chambers) being an example (Ford and Williams 2007; Palmer 2007).

At Nanisivik, very sharp corrosion notches that are horizontal and 20-30 cm high extend into the dolomite walls for at least 20 m (the limit of deep crosscuts in the mine). They are filled with layered pyrites which continue out into the ore body as regular sheets truncating earlier, dipping mineral layers until they themselves are truncated by later fillings (Figure 10A, B, C). Note that they often have a discordant relation-

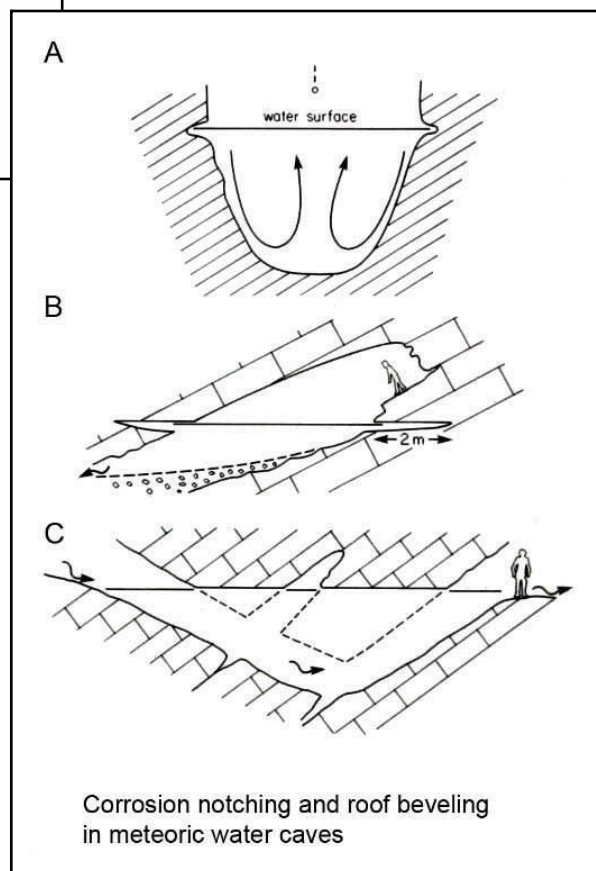


Figure 9. Corrosion notching and roof beveling in meteoric water caves; see the text for details.

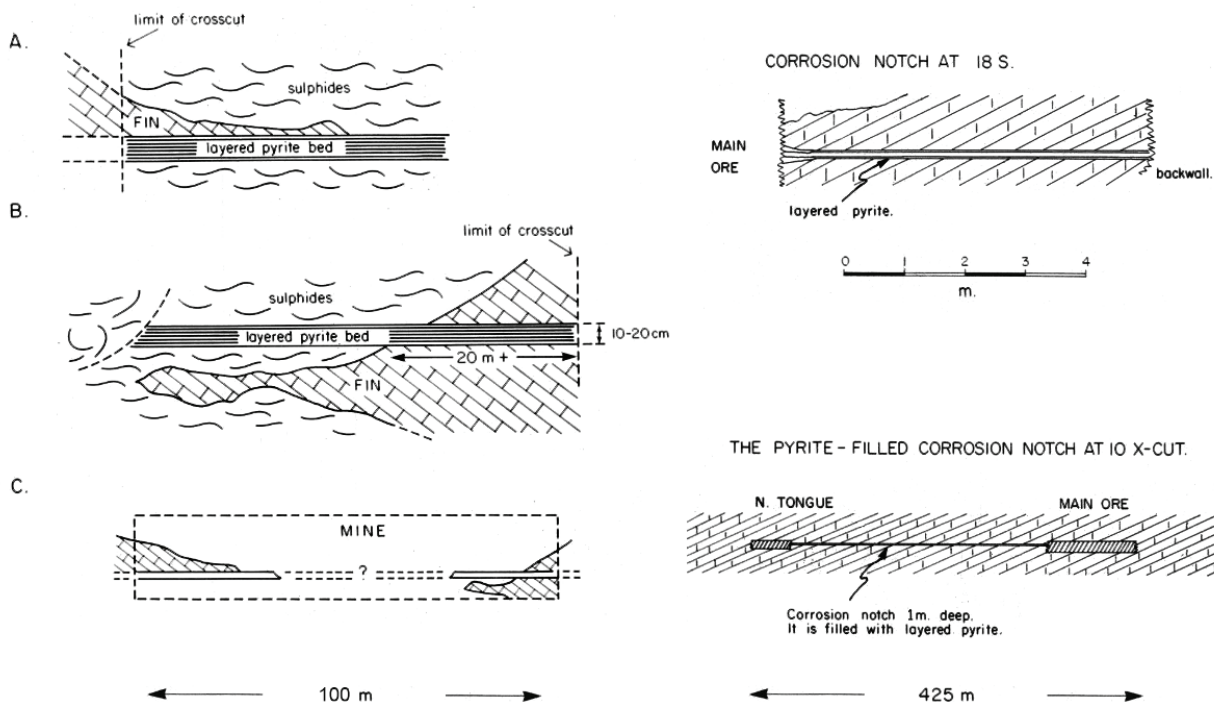


Figure 10. Examples of pyrite-filled corrosion notches in the Nanisivik Main Ore body, based on field sketches by the author in 1981, and mine drilling records at No. 10 Cross-Cut.

ship with the fins, suggesting that they are later features in most examples (Figure 10B). If the mine geologists' interpretation of their frequent drill cores was correct, in the western half of the cavity is one exceptional corrosion notch that is filled with layered pyrite. This feature is horizontal, one meter deep, and extends across the dipping dolomites for a distance of 350 m towards Shale Hill north of the main cavity (Figure 2; Figure 10, lower right); it is the greatest example of corrosion notching that I am aware of anywhere. The roof of the Main Ore cavity itself can be considered to be a ceiling bevel formed at the close of a single stage; its modern extent is 300,000 m² (~3000 x 100 m), having been reduced by surface erosion at both ends.

Sedimentology and geochemistry of the ore fillings

The fillings are of pyrite (FeS₂), galena (PbS), sphalerite ZnS), minor accessory minerals, and the gangue minerals - secondary dolomite (CaMg₂CO₃, which is volumetrically predominant, chiefly as white sparry dolomite), calcite (CaCO₃) and quartz (SiO₂). The bulk porosity is quite high but most pores today are filled with ice because, as noted, the mine is entirely within permafrost and has a temperature of -12° C. Generalizing, the ore displays four sedimentary modes (Figure 11) : (i) as regular layers settled or

precipitated on the floor, or lapping onto the upper faces of fins where they lie sub-parallel to the sloping fin surfaces: (ii) as chaotic polymict breccias; these were chiefly in the centre of the body, strongly suggestive of channel cut-and-fill episodes of erosion and deposition, including avalanching of earlier deposits off of the lateral fins into an abruptly emptied centre: (iii) as the horizontal, thin but layered, pyrite sheets noted above (Figure 10), that crosscut earlier type i and ii sediments and fill corrosional slashes into the dolomite walls; (iv) as metasomatic replacements of dolomite. Ghazban (1988) found metasomatism to be more widespread than the author observed in exposures laid bare in the earlier mining (1981 and 1982) but ore-bedrock contacts remain generally sharp (as seen in the example in Figure 7) and the depth of isotopic alteration of bedrock where we have examined it is trivial (Figure 12).

The isotopic compositions of the dolomite gangue suggest that they were partly a by-product of dissolution-replacement of the host rock, and partly due to sulfate reduction by hydrocarbons and related organic matter derived from the large volumes of shaly rocks above and below the Society Cliffs dolomite in the region (Ghazban et al.1992). Throughout the mine freshly blasted rock stank of bitumen. The stable isotopic compositions of fluid inclusions from more

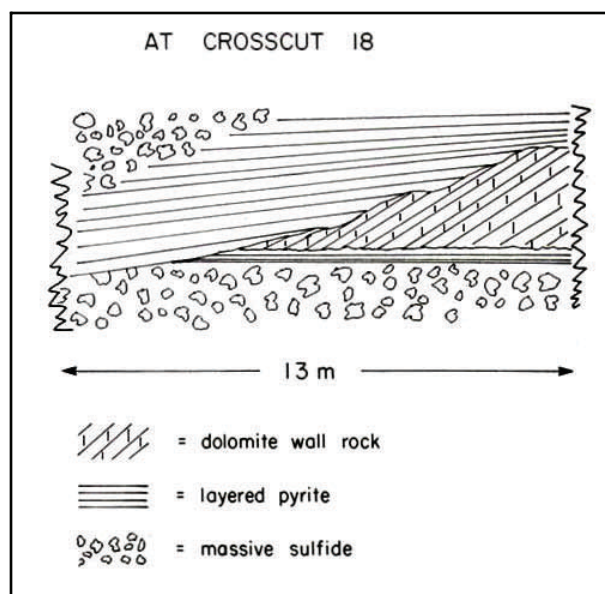


Figure 11. Illustrating three modes of ore and gangue mineral deposition in the Nanisivik Main Ore body. Type i as relatively thick beds onlapping a fin of dolomite; Type ii, chaotic polymict breccias (a) early stage, beneath the fin, and (b) later stage, occupying a channel cut into Type i deposits: Type iii, layered pyrite corrosion notch filling, with thin beds. From the author's field measurements at Cross-Cut 18 in 1981.

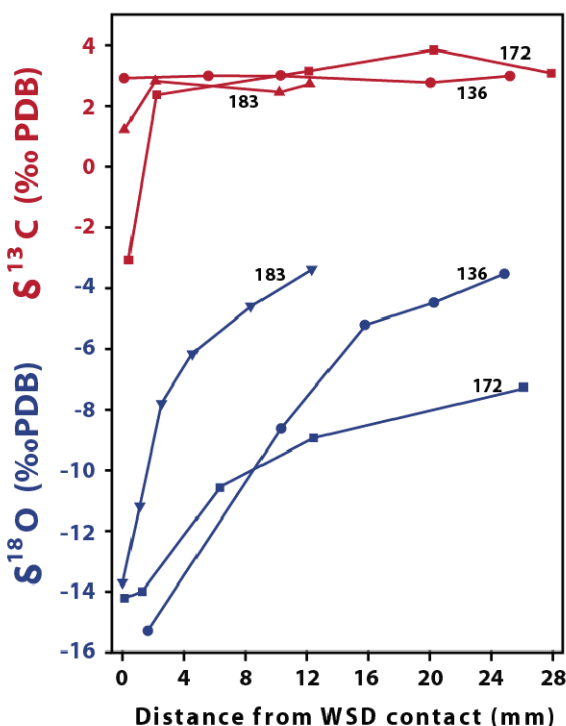


Figure 12. Plot illustrating the depth of isotopic alteration beneath the surface of a fin of dolomite bedrock that was overlain by white sparry dolomite (WSD). The scale is in mm (from Ghazban, 1988).

than 400 published analyses of gangue dolomite (principally), plus sphalerite, galena, calcite and quartz (Olson, 1984; McNaughton and Smith, 1986; Ghazban et al. 1991) are consistent with derivation of the metals primarily from exchange reactions with large volumes of metalliferous sediments, indicating low water/rock ratios and generally moderate temperatures in the system. The ore fluids were similar to many modern oilfield brines (Figure 13). Based on fluid inclusion homogenization temperatures and isotope fractionations between co-precipitating minerals, the various authors estimated temperatures to be in the range 90-170 $\pm$ 40° C in most samples, but rising as high as 250 $\pm$ 50° C in some of the late stage precipitates. These findings point firmly to the conclusion that the Nanisivik ore deposits are of hydrothermal origin, in waters with high concentrations of dissolved solids (brines); CaCl₂ was the most abundant solute in the fluid inclusions, suggesting that the waters originated as sea water trapped in interstices in the regional shales and siltstones.

Modeling the genesis of the Main Ore cavity and its deposits

Syngeneses and paragenesis

In the Introduction it was noted that there are two schools of thought amongst the scientists who consider that economic massive sulfide deposits are of hypogene origin. One school advocates the infilling of pre-existing karst, with relatively little modification of the karst morphology. Sangster (1988) writes that "The other school, advocated chiefly for the Polish deposits, proposes that 'the emplacement of ores and formation of underground karst were parts of the same formative processes and essentially simultaneous (Sass-Gutkiewicz et al. 1982)". At the Main Ore the field evidence is overwhelmingly in favor of the Polish syngenetic model, as indicated by its horizontality, width, ceiling beveling and general lack of bedrock breakdown. Further, the presence of two or more levels of sharp fins exhibiting meandering channel flow form, and the sedimentological evidence of onlapping sulfides covering the fins before avalanching into channels to form the polymict breccias there strongly support the notion of a paragenetic mode of dissolution and deposition. The dolomite was dissolved upwards (and laterally outwards where the corrosion interface was stable in its elevation for substantial periods of time) with more or less simultaneous precipitation of sulfides and gangue minerals underneath.

Lebedev (1974) and others have shown that many modern hydrothermal systems are strongly episodic in their behavior, with long periods of comparatively low

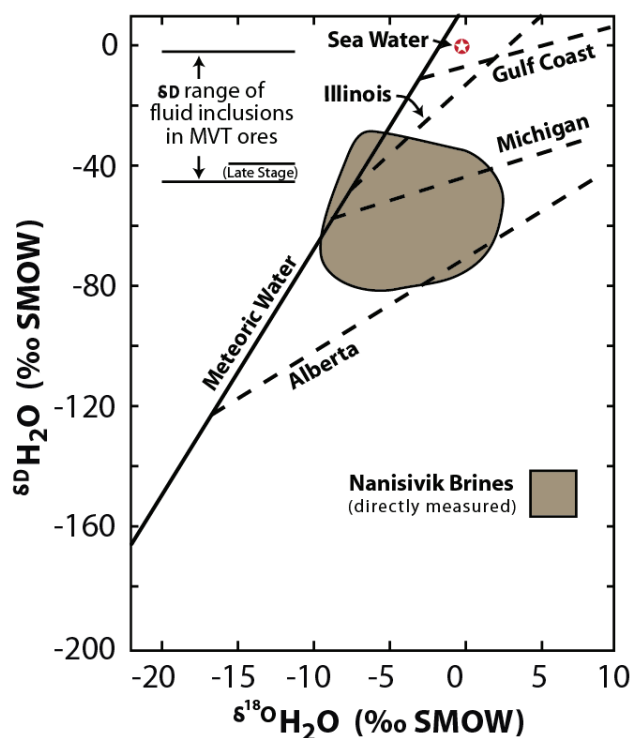


Figure 13. The range of the oxygen and hydrogen isotopic composition of fluid inclusions extracted from ore and gangue minerals samples of the Nanisivik Main Ore compared to modern seawater, worldwide rain and snow (the Global Meteoric Water Line), and brine samples from selected oilfields. (from Ghazban, 1988).

and stable discharge being interrupted by brief spells of abundant flow (surges): in some instances the surges are correlated with regional earthquakes. In the Main Ore, the contrast between the dynamic nature of the channel cut-and-fill sulfide breccia events on the one hand and the long periods of stability suggested by the astonishing pyrite-filled corrosion notches slashed horizontally for tens to hundreds of meters laterally into the bedrock on the other, are clear evidence of a sequence of many, dynamically very contrasted, episodes in the history of development.

The age and depth of formation of the Main Ore?

The question of age remains controversial. Symons et al. (2000) accepted Clayton and Thorpe's (1982) contention that the gabbroic dykes displace the Main Ore, which thus must be older. Using paleomagnetic measurements of the ores and gabbros and correlating the results with polar wander curves, they placed the age of the dykes at ~720 Ma and the ore as early as 1095±10 Ma, which is soon after deposition of the Arctic Bay shales that cover the host dolomites must have ceased. In flat contradiction, Sherlock et al.

(2004) contend that the dykes do not displace the ore, which thus is younger. They discovered small quantities of altered siliciclastic deposits scattered throughout the ores and interpreted them as insoluble residues of dolomite solution plus, perhaps, other clay colloids from nearby shales. The ^{40}Ar : ^{39}Ar dating method was applied to samples of orthoclase derived from the altered material, giving ages around 461 Ma (Middle Ordovician). It is possible that the altered clays the author observed resting on a fragment of fin near the roof in the Main Ore in 1981 (see above) belonged to this category of deposits.

Based on their thermal water temperature findings, McNaughton and Smith (1986) estimated that the Main Ore emplacement took place at a depth of at least 1600 meters underground. By the Middle Ordovician the host dolomites at Nanisivik had been deeply buried, then deformed, uplifted and eroded, followed by a second burial beneath 500+ m of Cambrian sedimentary rocks. This history accords well with a 1600+ m depth for the ore emplacement.

McNaughton and Smith (1986) also derived a vertical thermal gradient as great as 60° C/100 m from the fluid inclusion measurements at the Ore; this is probably an overestimate but it does underscore the very dynamic nature of the flow system that is also indicated by the morphologic and sedimentary evidence outlined above.

A model for the Main Ore emplacement

Figure 14 presents the author's model for the syngenetic dissolution of host dolomite and emplacement of the Main Ore at Nanisivik. Frame A shows the regional picture. Readers must imagine the entire structure to be sinking beneath a load of many hundreds of meters of later Proterozoic and Cambrian sediments that are not shown. Basin waters are being expelled through the more permeable formations and the major fractures.

Figure 14 B is a close-up of the Main Ore horst. It is known that there were minor episodes of karstification during deposition of the dolomite. After it was folded and faulted, further karstification may have occurred during the long erosional time interval before the Cambrian deposition began. This is suggested by two pieces of circumstantial evidence. The first is the sharp northwards jog in the middle of the Main Ore, which shifts the channel from a southside strike-oriented position in the horst to a similar central position (Figure 2); this kind of behavior often indi-

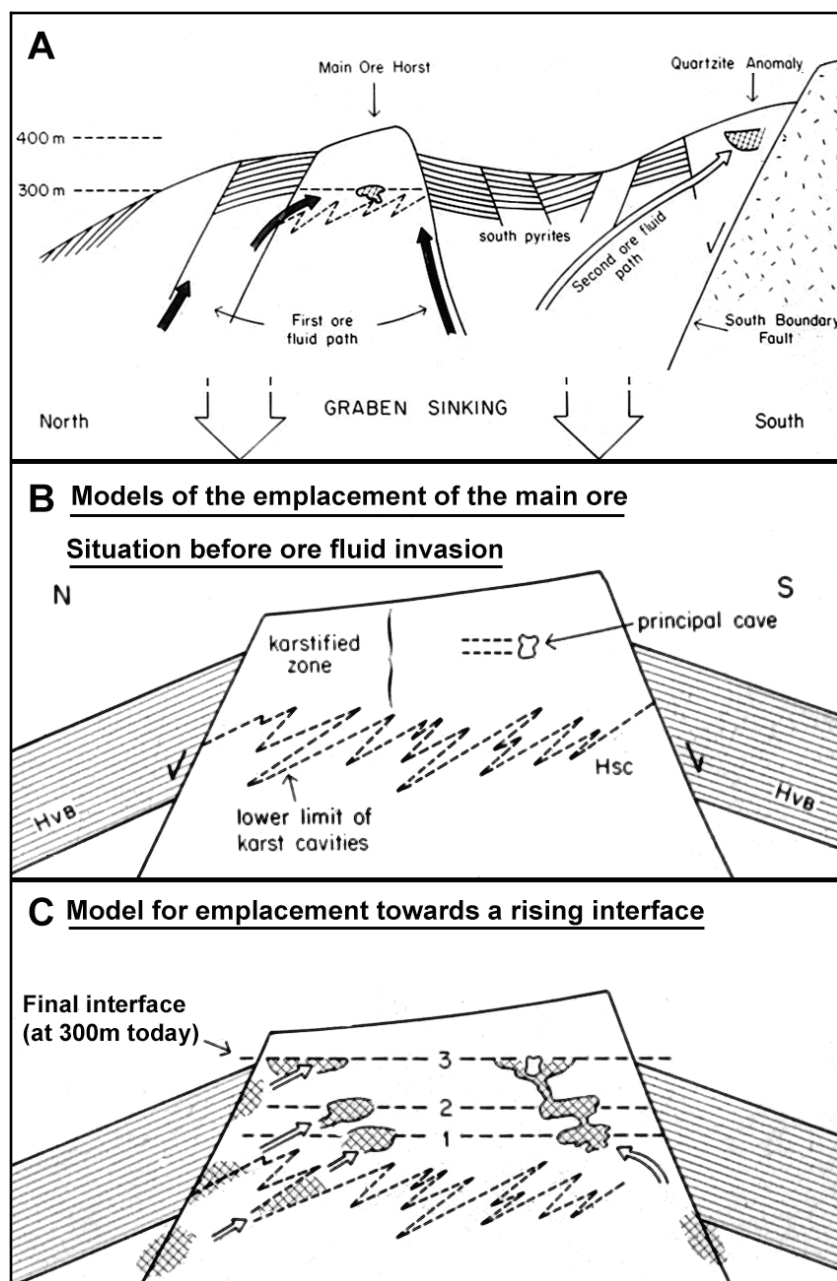


Figure 14. Model for the emplacement of the Main Ore and manto massive sulfide deposits at Nanisivik zinc-lead mine. See the text for details.

cates that the hypogene flow was ascending towards an open target (a cave) in the rock, and became diverted along it (Ford, 1995). Second, and associated with the first, is the very irregular spatial distribution and size of the sulfide deposits detected in the manto below the Main Ore body (Figure 4). It suggests the presence of scattered, small dip tubes (proto-caves; Ford and Williams, 2007; 214-7) in the north-dipping strata underneath the Main Ore level. Ore fluids rising up the northern boundary fractures were able to migrate out into the body of the dolomite horst block

via these tubes, and begin sulfide deposition as the pressure dropped.

Figure 14C completes the story. The graben is sinking, with episodic stillstands of the kind illustrated in Figure 8A. One such (Number 2 in the figure) is a little longer in duration and gives rise to the larger concentrations of sulfides seen from place to place at ~250 m asl in the manto zone. Then the fluid interface is stable but oscillating for a long period in a seven meter high zone that is now centered at 300 m asl. At least two successive paragenetic channels were created within this narrow range, forming the bulk of the Main Ore body. Flow discharge was markedly episodic during its formation, including periods of steady paragenetic excavation/deposition, stronger surges that temporarily emptied portions of the central channel, and long periods of quiescence when very deep corrosion notches of little height could be cut into the host rock at precisely fixed levels of the fluid interface.

Further sinking eventually uncoupled the Main Ore body from its fluid sources, some of which were evidently re-directed upwards to the south, where the final surviving pyrite body in the local area was deposited at 350-380 m asl in the Quartzite Anomaly (Figure 14A).

Dissolution and the fluid interface

Ghazban (1988) showed that degradation of the local shale bodies (Arctic Bay and Victor Bay formations) to produce the kerogen

predecessors of oil would release more than sufficient CO_2 to supply the carbonic acid necessary for the volumes of dissolution of the Society Cliffs dolomite that occurred during emplacement of the ore bodies. However, the exceptionally deep corrosion notches suggest that there was some contribution from stronger acids: H_2S from the reduction of gypsum, interacting with the metal chloride solutions to precipitate the sulfides, probably yielded the extra H^+ in abundance.

What was the nature of the fluid interface? This question is not within the author's area of competence but the combination of high temperatures with stability over time suggests that it must have been either an oil/brine interface or (more probably?) a gas/brine interface. The geographical area covered by the Main Ore horst, including the Ocean View and Kuhulu Trend extensions, is similar to that of many small oil or gas fields, and the horst and graben geometry with shale aquitards will have offered opportunities for structural trapping.

The regional tectonic setting

Bradley and Leach (2003) have carefully reviewed the tectonic settings of the worldwide MVT deposits shown in Figure 1. The Nanisivik massive sulfide deposits are atypical in that they appear to have accumulated in an extending (sinking) tectonic setting rather than one of the variety of collisional settings (orogenic forelands) which predominate. This may help to explain the very unusual and spectacular speleomorphic features (both dissolutional and depositional) described in the Main Ore.

Conclusions

The massive sulfides at the Nanisivik Mine were emplaced in a syngenetic mode by thermal waters (brines) of hypogene origin. Host dolomite was dissolved and the sulfide and gangue minerals were precipitated essentially simultaneously at a given site. The principal mode of cavity excavation and mineral deposition was paragenesis, wherein a corrosional ceiling is dissolved upwards in balance with the rising elevation (increasing thickness) of deposits on the floor below. Most dissolution took place at or close to an interface between brine and oil or brine and gas. Rates of discharge of the brine ranged widely, sometimes surging to carve channels through previously deposited ore, at other times being semi-static and creating deep corrosion notches precisely at the interface. Deposition took place in an extensional (sinking) basin setting in which metal-rich formation fluids from clastic rocks could be expelled through the host dolomites.

Acknowledgements

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References

- Bradley, D.C., and D.L. Leach. 2003. Tectonic controls of Mississippi Valley-type mineralization in orogenic forelands. *Mineralium Deposita* 38: 652-667.
- Clayton, R.H., and L. Thorpe. 1982. Geology of the Nanisivik zinc-lead deposit. In *Precambrian sulphide deposits: Geological Association of Canada special paper 25*, ed. R.W. Hutchinson, C.D. Spence and J. M. Franklin, 739-758. Toronto: Geological Association of Canada
- Dżulynski, S., and M. Sass-Gutkiewicz. 1989. Pb-Zn ores. In *Paleokarst – a systematic and regional review*, ed. P. Bosák, D. C. Ford, J. Głazek and I. Horáček, 377-397. Prague, Amsterdam: Academia Praha / Elsevier.
- Ford, D.C. 1986. Genesis of paleokarst and strata-bound zinc/lead sulfide deposits in a Proterozoic dolostone, northern Baffin Island - a discussion. *Economic Geology* 81: 1562-1563.
- Ford, D.C. 1987. Effects of glaciations and permafrost upon the development of karst in Canada. *Earth Surface Processes and Landforms* 12 (5): 507-522.
- Ford, D.C. 1995. Paleokarst phenomena as "targets" for modern karst groundwaters: The contrasts between thermal water and meteoric water behaviour. *Carbonate and Evaporites*, 10(2): 138-147.
- Ford, Derek C., and Paul Williams. 2007. *Karst Hydrogeology and Geomorphology*. Chichester: John Wiley and Sons, Ltd.
- Ghazban, F. 1988. Stable isotope studies of sulphides, gangue minerals, wall rocks and their fluid inclusions at Nanisivik, Baffin Is., Canada. PhD diss., McMaster University.
- Ghazban F., D.C. Ford, and H.P. Schwarcz. 1993. Multistage dolomitization in the Society Cliffs Formation, northern Baffin Island, NWT, Canada. *Canadian Journal of Earth Sciences* 29 (7): 1459-1473.
- Ghazban, F., H.P. Schwarcz, and D.C. Ford. 1991. Stable isotopic composition of the hydrothermal fluids responsible for the Nanisivik Zn-Pb deposits, Northwest Territories, Canada. *Applied Geochemistry* 6: 257-266.
- Ghazban, F., H.P. Schwarcz, and D.C. Ford. 1992. Correlated strontium, carbon and oxygen isotopes in carbonate gangue at the Nanisivik zinc-lead deposits, northern Baffin Island, Canada. *Chemical Geology (Isotope Geoscience Section)* 87: 137-146.
- Lebedev, L.M. 1974. *Recent ore-forming hydrothermal systems* (in Russian). Moscow: Nedra.

- McNaughton, K., and T.E. Smith. 1986. A fluid inclusion study of sphalerite and dolomite from the Nanisivik lead-zinc deposit, Baffin Island, Northwest Territories, Canada. *Economic Geology* 81: 713-20.
- Myroie, J.E., and J.L. Carew. 1990. The flank margin model for dissolution cave development in carbonate platforms. *Earth Surface Processes and Landforms* 15: 413–24.
- Olson, R.A. 1984. Genesis of paleokarst and strata-bound zinc-lead sulphide deposits in a Proterozoic dolostone, northern Baffin Island, Canada. *Economic Geology* 79: 1056-1103.
- Palmer, A.N. 2007. *Cave Geology*. Dayton: Cave Books.
- Rhodes, D., E.A. Lantos, J.A. Lantos, R.J. Webb, and D.C. Owens. 1984. Pine Point orebodies and their relationship to structure, dolomitization and karstification of the middle Devonian barrier complex. *Economic Geology* 70: 991–1055.
- Ryder, P.F. 1975. Phreatic network caves in the Swaledale area, Yorkshire. *Transactions, British Cave Research Association* 2 (4): 177-192.
- Sangster, D.F. 1988. Breccia-hosted lead-zinc deposits in carbonate rocks. In *Paleokarst*, ed. N.P. James and P.W. Choquette, 102-116. New York: Springer-Verlag.
- Sass-Gutkiewicz, M. 1974. Collapse breccia in the ore-bearing dolomite of the Olkusz mine (Cracow-Silesian region). *Roczniki Polskiego Towarzystwa Geologicznego* 44: 217-226.
- Sass-Gutkiewicz, M., S. Dzulynski, and J.D. Ridge. 1982. The emplacement of zinc-lead sulfide ores in the Upper Silesian district – a contribution to the understanding of Mississippi Valley-type deposits. *Economic Geology* 77: 392-412.
- Sherlock, R.L., J.K.W. Lee, and B.L. Cousens. 2004. Geologic and geochronologic constraints on the timing of mineralization at the Nanisivik Mississippi Valley-type deposit, northern Baffin Island, Nunavut, Canada. *Economic Geology* 99: 279-93.
- Symons, D.T.A., T.B. Symons and D. F. Sangster. 2000. Paleomagnetism of the Society Cliffs dolostone and the age of the Nanisivik zinc deposits, Baffin Island, Canada. *Mineralium Deposita* 35: 672-82.

THE IMPACT OF HYPOGENIC PROCESSES ON WATER RESOURCES IN THE ARID SOUTHWEST: EXAMPLES FROM THE LOWER PECOS REGION OF NEW MEXICO, USA

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Abstract

Hypogenic phenomena are well-developed throughout the lower Pecos region of southeastern New Mexico, USA, occurring over a distance of more than 300 km. Artesian flow of groundwater from these features makes a substantial contribution to stream flow in surface drainages such as the Pecos River and its tributaries. In addition, groundwater stored in karstic artesian aquifer systems in the lower Pecos basin supports a robust agricultural community essential to the economy of the region. Human settlement in this part of the arid southwest would be much more limited without the benefit of hypogenic groundwater flow systems on local and regional scales.

Introduction

Manifestations of hypogenic processes extend across almost three degrees of latitude in the lower Pecos region of southeastern New Mexico, USA (Figure 1). Hypogenic features include springs and cenotes fed by upward artesian flow of groundwater from carbonate aquifers developed in the middle Permian San Andres limestone and Capitan Reef. Artesian flow from these features contributes directly to surface drainages such as the Pecos River as base flow, and indirectly by overflow into wetlands that are hydraulically connected to the Pecos (Land, 2003; 2006). This paper summarizes the impact of hypogenic groundwater discharge on human settlement patterns in the lower Pecos region of New Mexico.

With respect to water use, the lower Pecos region is conventionally regarded as extending from Santa Rosa, NM (Figure 1) to its confluence with the Rio

Grande in Texas (e.g., Hall, 2002, p. 40-41; Johnson et al., 2003). Farther upstream the upper Pecos is a high mountain stream. The lower Pecos basin includes the northern edge of the Chihuahuan Desert. The climate of the region is semi-arid, with mean annual rainfall less than 33 cm/yr, and thus can sustain only a limited population. The U.S. Census Bureau (2009)

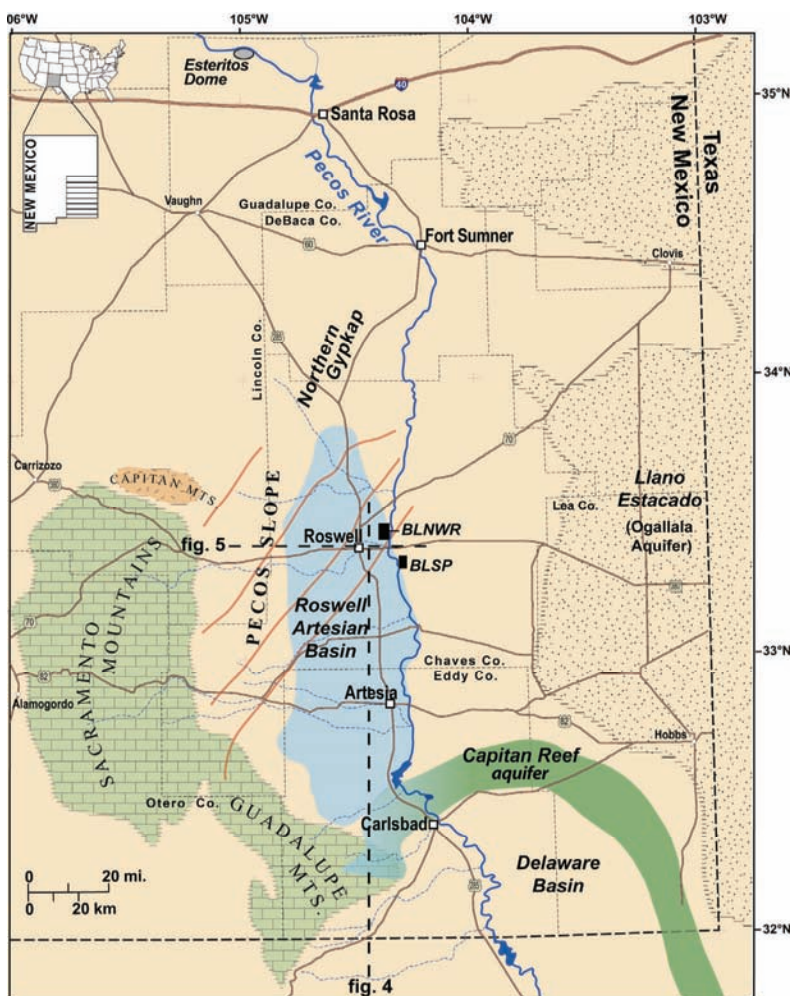


Figure 1. Regional map of the lower Pecos basin, showing major aquifer systems and surface drainages. BLNWR = Bitter Lakes National Wildlife Refuge. BLSP = Bottomless Lakes State Park (modified from Land and Newton, 2007)

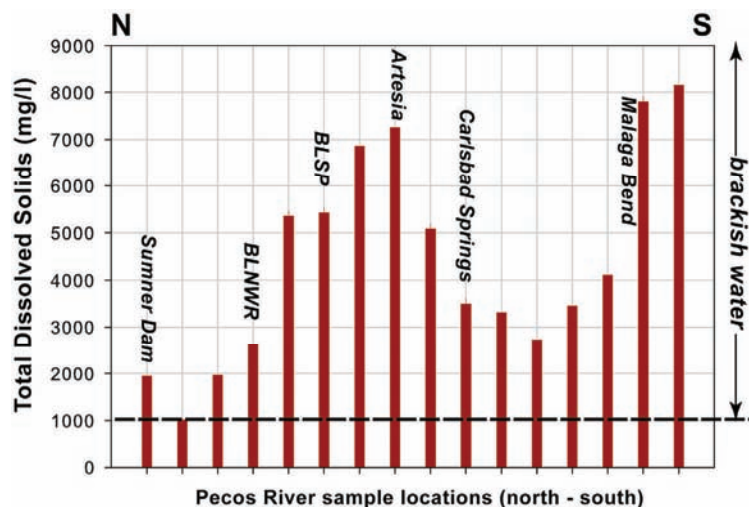


Figure 2. North-south variation in surface water quality in the Pecos River (New Mexico Environment Dept., 2003, unpublished data).

estimates that the total population in 2006 was only about 120,000 in the four New Mexico counties through which the lower Pecos River flows (Figure 1). The economy of the region is primarily agricultural, relying to a certain extent on surface water resources. However, water quality in the Pecos River is generally poor, with high sediment load and a dissolved solids content ranging from 1000 to 8000 mg/l (Figure 2) (New Mexico Environment Dept., unpublished data). While adequate for irrigation and watering livestock, surface water in the lower Pecos basin is for the most part unfit for human consumption. In most of the region, groundwater stored in karstic artesian aquifers is the sole source of drinking water and municipal water supply. Population centers on the lower Pecos are concentrated in areas where artesian water resources are available, separated by vast areas where only isolated ranches occur.

Artesian water resources in the Lower Pecos Basin Carlsbad area

The town of Carlsbad in central Eddy County represents the southernmost community in the lower Pecos basin where artesian water resources are available. Farmers in the Carlsbad area rely on surface water stored in artificial impoundments on the Pecos River for irrigation, but water in that reach of the Pecos is too saline for human consumption, with dissolved solids

>3000 mg/l (Figure 2). Carlsbad is located a few km northeast of the Guadalupe Mountains (Figure 1). The middle Permian Capitan Reef is exposed along the southeast escarpment of the mountains, and serves as host rock for the famous Carlsbad Cavern and Lechuguilla Cave, both formed by hypogenic speleogenesis (Hill, 1987; Stafford et al., 2009). A few km northeast of the entrance to Carlsbad Cavern, the Capitan Reef plunges into the subsurface and extends beneath Carlsbad (Figure 3), where it forms a karstic aquifer that provides the potable water supply for the town (Hiss, 1975; 1980). A substantial component of recharge to the Reef aquifer occurs by direct infiltration into cavernous zones in the Capitan limestone during flood events, in those areas where the Reef is exposed in arroyos in the Guadalupe Mountains.

Groundwater flows northeastward through the reef and discharges from springs along the Pecos River within the town of Carlsbad (Bjorklund and Motts, 1959; Land and Burger, 2008).

The Capitan Reef can be followed in the subsurface east of Carlsbad in southeastern New Mexico, where it defines the northern margin of the Delaware Basin (Figure 3). However, throughout most of its extent the Reef is a brine aquifer, with chloride concentrations >5000 mg/l (Hiss, 1975; 1980). Despite the vast extent of the Capitan Reef, Carlsbad, because of its proximity to recharge areas in the Guadalupe Moun-



Figure 3. Surface and subsurface extent of the New Mexico portion of the Capitan Reef aquifer. Contours show chloride concentration in the Reef aquifer, in mg/l (modified from Hiss, 1975).

tains, is the only community in the lower Pecos region positioned to exploit the artesian water resources in the fresh water portion of the aquifer. For this reason, Carlsbad is also home to ~50% of the population of Eddy County (U.S. Census Bureau, 2009), and has a substantially higher percentage of green lawns than many other parts of the state. A few km north of Carlsbad the Reef changes facies into backreef evaporites and redbeds with limited potential to store or transmit fresh water (Figure 4). Absent artesian base flow from the Reef, the Pecos River is a losing stream for several km upstream (Cox, 1967). The town of Artesia, the only other significant population center in Eddy Co., derives its freshwater resources from a separate, underlying aquifer system formed in karstic limestones of the San Andres and Grayburg Formations (Figure 4).

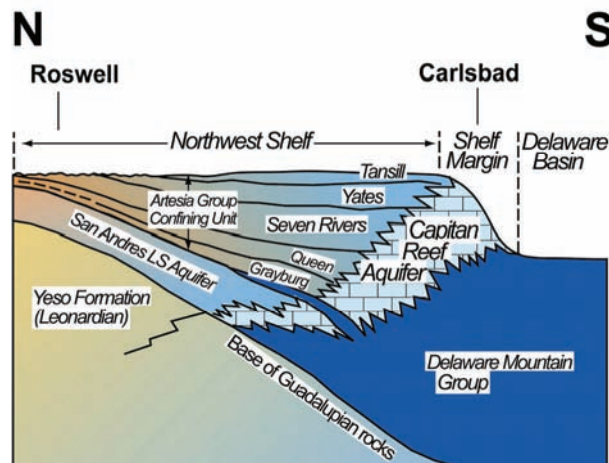


Figure 4. Diagrammatic cross-section showing shelf-to-basin facies relationships within middle Permian Guadalupian strata in southeastern New Mexico. Backreef units of the Artesia Group change facies from carbonates in the near-backreef section to interbedded evaporites and mudstone farther north, where they serve as confining beds for the San Andres artesian aquifer. Line of section shown in Figure 1 (modified from Hiss, 1975).

Roswell Artesian Basin

The Roswell Artesian Basin, located in central Chaves and northern Eddy Counties (Figure 1), is one of the most intensively farmed areas in New Mexico, and has been described by many workers as a world-class example of a rechargeable artesian aquifer system (e.g., Havenor, 1968). The principal crops in the Basin are alfalfa, cotton, sorghum, pecans, and chiles. Dairy farming is also an important component of the agricultural economy. Total dissolved solids along this reach of the Pecos River range from 2000 to 7000 mg/l (Figure 2), and for this reason, communities in the

Roswell Basin derive virtually all of their irrigation, industrial, and municipal drinking water from groundwater in an artesian carbonate aquifer and a shallow alluvial aquifer. Groundwater in the carbonate aquifer is stored in multiple erratically developed, highly porous and transmissive zones within the middle Permian San Andres limestone, and to a lesser extent in the overlying Grayburg Formation of the Artesia Group (Figure 4).

Recharge to the artesian aquifer occurs by direct infiltration from precipitation, and by runoff from intermittent losing streams that flow eastward across the Pecos Slope, a broad area east of the Sacramento Mountains where the San Andres limestone is exposed in outcrop (Figures 1 and 5). Enhanced recharge occurs through sinkholes and solution-enlarged fractures associated with the Pecos Buckles, wrench fault zones of probable Laramide age that extend SW-NE for several tens of km across the Pecos Slope (Motts and Cushman, 1964; Havenor, 1968; Kelley, 1971). The artesian aquifer becomes confined ~10 km west of the city of Roswell, where the eastward-dipping San Andres limestone passes beneath gypsum and mudstones of the Artesia Group (Figure 5). Most of the agricultural activity in the Basin is concentrated east of the confined-unconfined boundary in a 20 km wide strip west of the Pecos River.

In pre-development times, groundwater flowed east and south, down-gradient from the recharge area, then upward through leaky gypsum confining beds into the alluvial aquifer, and ultimately to the Pecos River (Figure 5). Since the inception of irrigated agriculture in the early 20th century, most of the down-gradient flow is now intercepted by irrigation wells in the Artesian Basin. Total discharge from artesian wells for irrigation and municipal water supply is approximately 432 million m³/yr (Barroll and Shomaker, 2003). In the early history of development of the artesian aquifer, many wells flowed to the surface with yields as high as 21,500 l/min (Welder, 1983), and high-volume springs fed tributary streams flowing into the Pecos River from the west.

Substantial natural discharge from the artesian aquifer still contributes to base flow in the Pecos River, through fractures and solution channels formed in the overlying evaporitic confining beds. This natural discharge amounts to roughly 37 million m³/yr, but was much greater before irrigated farming began (Barroll and Shomaker, 2003; Land and Newton, 2008). Natural discharge from the artesian aquifer has caused the formation of a complex of karst springs, sinkhole lakes and extensive wetlands along the west

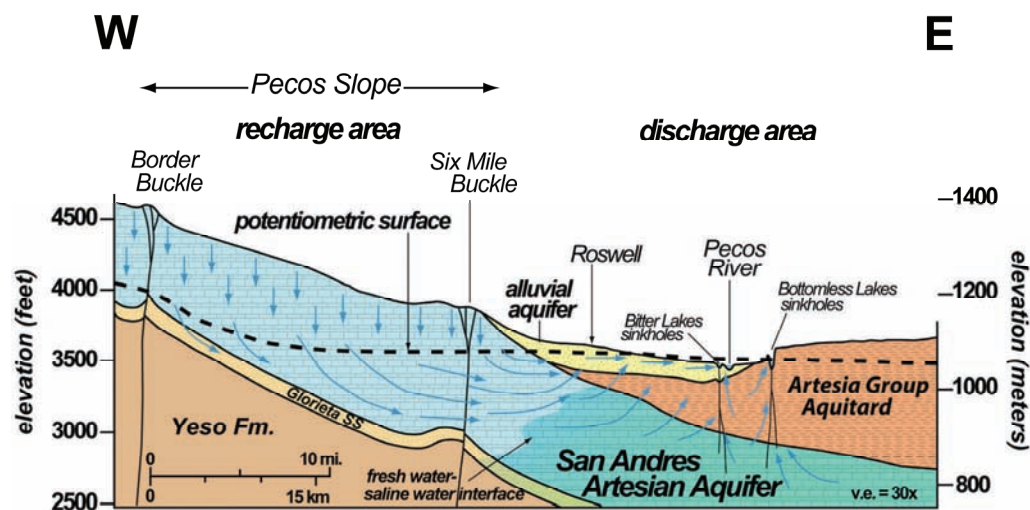


Figure 5. West-east hydrostratigraphic section illustrating regional ground water flow patterns within the artesian and shallow aquifers. Arrows indicate general direction of ground water flow. Line of section shown in Figure 1 (adapted from Land and Newton, 2007).

bank of the Pecos at Bitter Lakes National Wildlife Refuge, east of Roswell (Figure 1) (Land, 2005; Land and Huff, 200-). Along the eastern margin of the Pecos River valley, discharge from the artesian aquifer has formed a chain of large gypsum cenotes in the Seven Rivers Escarpment at Bottomless Lakes State Park (Figure 6) (Land, 2003; 2006). Discharge from the springs and cenotes at Bitter Lakes and Bottomless Lakes contributes indirectly to flow in the Pecos River.

In the early history of settlement in the Artesian Basin, most of the cenotes at Bottomless Lakes overflowed into wetlands along the eastern shore of the Pecos River, but the progressive decline in hydraulic head in the artesian aquifer (up to 70 m in some areas) caused lake levels to fall, so that now only Lea Lake, the largest of the cenotes, overflows. In 1975, a catastrophic rockslide occurred on the steep eastern wall of Lea Lake, which apparently opened new spring sources in the lake bed, as indicated by a significant increase in flow from the lake and the flooding of adjacent grazing lands with several thousand m^3 per day of saline water. A culvert was installed to convey the increased flow into the wetlands west of the park, but the lake continued to



Figure 6. Aerial view of gypsum cenotes at Bottomless Lakes State Park, east of Roswell, showing overflow from the Lea Lake cenote.

flood an adjacent parking lot and camping area during the winter. In 2002, the Park completed construction of a more efficient drainage canal to capture all of the discharge, resulting in a substantial increase in measured flow volume from the lake. As discharge continued to increase, a second drain was installed in 2005. On January 16th, 2008, the New Mexico Interstate Stream Commission measured a combined discharge of 41 million l/day from both drains. The increased flow from Lea Lake, amounting to roughly 15 million m³/yr, has caused an expansion of wetlands to the west, which are now hydraulically connected to the Pecos River, and a net gain in stream flow downstream from the park, an interesting phenomenon in a semi-arid region that experiences periodic drought.

The broader societal implications of this local increase in water supply relate to the interstate use of surface water resources in the lower Pecos valley. New Mexico shares the water in the Pecos River with its downstream neighbor, the state of Texas, and is obligated by interstate compact to deliver a specific volume of water to Texas every year (Hall, 2002). The increase in artesian flow from Lea Lake will ultimately be used by the state of New Mexico to help meet its compact obligation.

94% of the population in Chaves and northern Eddy Counties lives in the cities of Roswell and Artesia, or in one of the small farming communities concentrated along the Pecos River between them (U.S. Census Bureau, 2009). Outside the area of irrigated agriculture in the Roswell Artesian Basin only isolated ranches occur.

Northern Gypkap

North of Roswell lies an extensive area in DeBaca and northern Chaves Counties known to cave explorers as the northern Gypkap (Gypsum Karst Project; Figure 1; Eaton, 1987). Artesian groundwater resources are not available in this region, in part because of the much greater distance from potential recharge areas to the southwest, and because of regional groundwater flow direction away from the Gypkap area. However, there is extensive evidence of past hypogenic processes, as indicated by the widespread distribution of maze caves and sinkholes formed in gypsum bedrock of the upper San Andres Formation and Artesia Group (Stafford et al., 2009). Regional facies changes within the San Andres Formation may also contribute to the limited availability of groundwater in this area. Regional stratigraphic models of the San Andres show deposition of mixed carbonates and evaporites occurring over an extremely broad shelf of very low relief in east central and southeastern New Mexico, west

Texas and the Texas Panhandle. The formation is comprised mostly of carbonates in southern New Mexico, where it is bounded to the southeast by a low-relief shelf margin with ramp-like geometry (Silver and Todd, 1969; Elliott and Warren, 1989). The Glorieta sandstone member of the lower San Andres thickens to the north at the expense of the carbonate lithology. The upper San Andres becomes increasingly evaporitic to the north and east (Kelley, 1971; Ward et al., 1986) and is correlative with the Blaine gypsum in western Oklahoma.

In the subsurface of the Roswell Artesian Basin carbonate lithologies predominate, and the evaporitic component of the San Andres is represented by intraformational solution-collapse breccias created by dissolution of interbedded evaporites by circulating groundwater. These breccia zones provide much of the groundwater storage capacity for the artesian aquifer system (Motts and Cushman, 1964; Welder, 1983). North of Roswell, shelfward facies change results in a San Andres section that consists mostly of gypsum with relatively thin interbeds of dolomite or limestone (Eaton, 1987; Forbes and Nance, 1997; Stafford and Nance, 2009). Common features of the San Andres in the northern Gypkap area are thus sinkholes and gypsum caves with dolomite ceilings or floors, rather than the extensive solution collapse breccias found farther south in the Artesian Basin. These caves are located above the regional water table, which is several hundred m deep in most of the northern Gypkap area, and do not provide significant potential for storage of groundwater other than as highly localized perched aquifers.

Because of the absence of abundant water supply, the northern Gypkap region is very sparsely populated. Most of the population of DeBaca County (>60%) is concentrated in the small farming community of Fort Sumner, which uses Pecos River water for irrigation (Figure 1) (U.S. Census Bureau, 2009).

Santa Rosa area

Solution-collapse processes have profoundly affected the topography and hydrology of western Guadalupe County, primarily due to subsurface dissolution of Permian carbonates and gypsum (Dinwiddie and Clebsch, 1973). The small community of Santa Rosa (2000 population ~2700), located in central Guadalupe County at the northern end of the lower Pecos valley (Figure 1), occupies a broad solution-subsidence depression about 10 km in diameter and 120 m deep. The Pecos River enters the depression through a narrow gorge north of town and flows southward across the eastern floor, exiting through the

southern drape wall of the sink. Kelley (1972) reports that subsidence results from subsurface dissolution of limestone and evaporites in the San Andres Formation, which is ~180 m deep beneath the floor of the basin. The Santa Rosa depression has been filled with 70 – 120 m of Tertiary and Quaternary sand and gravel deposited by the Pecos River, and is rimmed by an inward-sloping drape of sandstone beds of the Triassic Santa Rosa Formation (Kelley, 1972).

Groundwater is the chief source of water supply in Guadalupe County. Surface water is sparse and available perennially only along the Pecos River. Thus most of the population of Guadalupe County (62%) is concentrated in Santa Rosa (U.S. Census Bureau, 2009). The San Andres limestone is the principal source for large quantities of groundwater in the western and central parts of the county. Irrigation wells completed in the San Andres may yield up to 9500 l/min, and spring flow in the area ranges from 35 to 11,000 l/min (Trauger, 1972; Dinwiddie and Clebsch, 1973).

About 25 km upstream from Santa Rosa the Pecos River flows across outcrops of the San Andres Formation along the north flank of Esteritos Dome (Figure 1). Along this reach of the river the Pecos loses water through fractured and cavernous San Andres limestones. Groundwater flows southeast through these highly transmissive carbonates and discharges through springs near the river, and by upward artesian flow into sinkhole lakes in the vicinity of Santa Rosa. The general route of subsurface flow is indicated by a line of lakes and sinkholes between Santa Rosa and the San Andres outcrops at Esteritos Dome (Dinwiddie and Clebsch, 1973).

The San Andres

Formation grades from a limestone facies in western Guadalupe County to anhydrite-gypsum-halite facies as it dips eastward into the Tucumcari Basin (Trauger, 1972). The precise nature and location of this subsurface facies change is not well-constrained, but is clearly reflected in mineral content in the Pecos River. Sulfate begins to increase upstream from Santa Rosa, indicating that circulating groundwater that contributes to base flow is in contact with beds of gypsum in the San Andres Formation. Near Santa Rosa, an increase in chloride content in the river suggests that beds of salt are present in the subsurface (Dinwiddie and Clebsch, 1973). The location of the Santa Rosa subsidence depression is very likely controlled by subsurface dissolution of evaporites in the vicinity of this facies change.

In addition to the large solution-subsidence depression, a great many smaller sinkholes occur in the Santa Rosa area, primarily along the north and west sides of the basin. Where soluble rocks in the underlying San Andres are overlain by relatively thin sandstone beds of the Santa Rosa Formation, upward stoping and collapse has produced vertical-walled sinkholes of typical cenote morphology formed in



Figure 7. Aerial view of the Santa Rosa karst district, showing location of the Blue Hole cenote. Discharge from the Blue Hole sink is reported to be 11,350 l/min (16,350,000 l/day).

siliciclastic rocks (Kelley, 1972). These sinks contain lakes sustained by upward artesian flow of groundwater from the San Andres aquifer, the best known of which is the Blue Hole sink (Figure 7). Groundwater from the underlying artesian aquifer discharges from Blue Hole at a rate of 11,350 l/min, ultimately flowing into the Pecos River (Dinwiddie and Clebsch, 1973). This artesian discharge contributes more than 5.8 million m³/yr to stream flow in the Pecos.

Conclusions

Water quality in the Pecos River, while generally adequate for irrigation and watering stock, is too saline for human consumption (Figure 2). For this reason, groundwater derived from karstic artesian aquifers is the sole source of fresh drinking water for the agricultural communities of the lower Pecos basin. The small population centers of the lower Pecos are concentrated in those areas where artesian water supplies are available. Human settlement in this part of the arid southwest would be much more limited without the benefit of hypogenic groundwater flow systems on local and regional scales.

References

- Barroll, P., and J. Shomaker. 2003. Regional hydrology of the Roswell Artesian Basin and the Capitan aquifer. In *Water resources of the lower Pecos region, New Mexico: 2003 New Mexico Decision Makers Guidebook*, ed. P. Johnson, L. Land, G. Price and F. Titus, 23-27. Socorro: New Mexico Bureau of Geology and Mineral Resources.
- Bjorklund, L.J., and W.S. Motts. 1959. *Geology and water resources of the Carlsbad area, New Mexico*. U.S. Geological Survey Open File Report. Washington: U.S. Geological Survey.
- Cox, E.R. 1967. *Geology and hydrology between Lake McMillan and Carlsbad Springs, Eddy Co., New Mexico*. U.S. Geological Survey Water-Supply Paper 1828. Denver: U.S. Geological Survey.
- Dinwiddie, G.A., and A. Clebsch Jr. 1973. *Water resources of Guadalupe County, New Mexico. Hydrologic Report 3*. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Eaton, J., ed. 1987. *GYPKAP 1987 Annual Report. Alamogordo*. Southwestern Region of the National Speleological Society.
- Elliott, L.A., and J.K. Warren. 1989. Stratigraphy and depositional environment of lower San Andres Formation in subsurface and equivalent outcrops: Chaves, Lincoln, and Roosevelt Counties, New Mexico. *American Association of Petroleum Geologists Bulletin* 73: 1307-1325.
- Forbes, J., and R. Nance. 1997. Stratigraphy, sedimentology, and structural geology of gypsum caves in east central New Mexico. *Carbonates and Evaporites* 12: 64-72.
- Hall, G.E. 2002. *High and dry: The Texas-New Mexico struggle for the Pecos River*. Albuquerque: University of New Mexico Press.
- Havenor, K.C. 1968. *Structure, stratigraphy, and hydrogeology of the northern Roswell Artesian Basin, Chaves County, New Mexico*. New Mexico Bureau of Mines and Mineral Resources, Circular 93. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hill, C.A. 1987. *Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains, New Mexico and Texas*. New Mexico Bureau of Mines and Mineral Resources, Bulletin 117. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hiss, W.L. 1975. *Chloride-ion concentration in ground water in Permian Guadalupian rocks, southeast New Mexico and west Texas*. New Mexico Bureau of Mines and Mineral Resources, Resource Map 4. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Hiss, W.L. 1980. Movement of ground water in Permian Guadalupian aquifer systems, southeastern New Mexico and western Texas. In *Trans-Pecos Region, Southeastern New Mexico and West Texas: New Mexico Geological Society Guidebook 31*, ed. P.W. Dickerson, J.M. Hoffer and J.F. Callender, 289-294. Socorro: New Mexico Geological Society.
- Johnson, P.S., L. Land, L.G. Price, and F. Titus, ed. 2003. *Water resources of the lower Pecos region, New Mexico: Science, policy, and a look to the future*. New Mexico Bureau of Geology and Mineral Resources, 2003 New Mexico Decision Makers Guidebook. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Kelley, V.C. 1971. *Geology of the Pecos country, southeastern New Mexico*. New Mexico Bureau of Mines and Mineral Resources, Memoir 24. Socorro: New Mexico Bureau of Mines and Mineral Resources.
- Kelley, V.C. 1972. Geology of the Santa Rosa area. In *East central New Mexico: New Mexico Geological Society Guidebook 23*, ed. V.C. Kelley and F.D. Trauger, 218-220. Socorro: New Mexico Geological Society.
- Land, L. 2003. Evaporite karst and regional ground water circulation in the lower Pecos Valley. In *Evaporite karst and engineering and environmental problems in the United States: Oklahoma Geological Survey Circular 109*, ed. K.S. Johnson and J.T. Neal, 227-232. Norman: Oklahoma Geological Survey.

- Land, L. 2005. Evaluation of groundwater residence time in a karstic aquifer using environmental tracers: Roswell Artesian Basin, New Mexico. In *Proceedings of the tenth multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, San Antonio, Texas, 2005: American Society of Civil Engineers Geotechnical Special Publication No. 144*, 432-440. Reston: American Society of Civil Engineers.
- Land, L. 2006. Hydrogeology of Bottomless Lakes State Park. In *Caves and karst of southeastern New Mexico: New Mexico Geological Society Guidebook 57*, ed. L. Land, V. Lueth, B. Raatz, P. Boston and D. Love, 95-96. Socorro: New Mexico Geological Society.
- Land, L., and P. Burger. 2008. Rapid recharge events in a karstic aquifer: An example from Lake of the White Roses, Lechuguilla Cave, NM. In *Proceedings of the eleventh multidisciplinary conference on sinkholes and the engineering and environmental impacts of karst, Tallahassee, Florida: American Society of Civil Engineers Geotechnical Special Publication No. 183*, 396-403. Reston: American Society of Civil Engineers.
- Land, L., and B.T. Newton. 2007. *Seasonal and long-term variations in hydraulic head in a karstic aquifer: Roswell Artesian Basin, New Mexico: New Mexico Bureau of Geology and Mineral Resources Open-File Report no. 503*. Socorro: New Mexico Bureau of Geology and Mineral Resources.
- Land, L., and B.T. Newton. 2008. Seasonal and long-term variations in hydraulic head in a karstic aquifer: Roswell Artesian Basin, New Mexico. *Journal of the American Water Resources Association* 44: 175-191.
- Land, L., and G.F. Huff. 20-- . Multi-tracer investigation of groundwater residence time in a karstic aquifer: Bitter Lakes National Wildlife Refuge, New Mexico, USA. *Hydrogeology Journal* (in review).
- Motts, W.S., and R.L. Cushman. 1964. *An appraisal of the possibilities of artificial recharge to ground-water supplies in part of the Roswell Basin, New Mexico. U.S. Geological Survey Water-Supply Paper 1785*. Boulder: U.S. Geological Survey.
- Silver, B.A., and R.G. Todd. 1969. Permian cyclic strata, northern Midland and Delaware Basins, west Texas and southeastern New Mexico. *American Association of Petroleum Geologists Bulletin* 53: 2223-2251.
- Stafford, K.W., and R. Nance. 2009. Evaporite speleogenesis of the Gypsum Plain: New Mexico and far west Texas. In *The caves and karst of the USA*, ed. A. Palmer and M. Palmer. Huntsville: National Speleological Society (in press).
- Stafford, K.W., L. Land, A.B. Klimchouk, and M.O. Gary. 2009. The Pecos River hypogene speleogenetic province: A basin-scale karst paradigm for eastern New Mexico and west Texas, USA. In *Advances in hypogene karst studies, NCKRI Symposium No. 1*, ed. K.W. Stafford, L. Land and G. Veni, (this volume). Carlsbad: National Cave and Karst Research Institute.
- Trauger, F.D. 1972. Ground water in east-central New Mexico. In *East central New Mexico: New Mexico Geological Society Guidebook 23*, ed. V. C. Kelley and F. D. Trauger, 201-207. Socorro: New Mexico Geological Society.
- U.S. Census Bureau. 2009. <http://quickfacts.census.gov/qfd/states/35000.html>, (accessed January, 2009).
- Ward, R.F., C.G. St. C. Kendall, and P.M. Harris. 1986. Upper Permian (Guadalupian) facies and their association with hydrocarbons – Permian Basin, west Texas and New Mexico. *American Association of Petroleum Geologists Bulletin* 70: 239-262.
- Welder, G.E. 1983. *Geohydrologic framework of the Roswell Ground-Water Basin, Chaves and Eddy Counties, New Mexico*. New Mexico State Engineer Technical Report 42. Santa Fe: New Mexico State Engineer.

UPPER ORDOVICIAN TRENTON-BLACK RIVER HYDROTHERMAL DOLOMITE RESERVOIRS OF EASTERN NORTH AMERICA

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Abstract

In the past decade, more than thirty new natural gas fields have been discovered in laterally discontinuous brecciated, vuggy dolomites of the Upper Ordovician Black River Group in south-central New York. These dolomitized reservoirs are virtually identical to those found in the Albion-Scipio Trend in Michigan, the Lima-Indiana Trend in Ohio and many smaller fields across eastern North America. The dolomitized reservoirs form around basement-rooted transtensional faults, some of which are detectable on seismic data. Most fields occur in and around elongate fault-bounded structural lows interpreted to be negative flower structures. Away from these faults, the formation is composed of impermeable limestone and forms the lateral seal for the reservoirs. In most cases the faults die out within the overlying Trenton Limestone and Utica Shale. Most porosity occurs in saddle dolomite coated vugs, breccias and fractured zones.

The patchy distribution around basement-rooted faults and geochemical and fluid inclusion analyses support a hypogene, fault-related hydrothermal origin for the saddle and matrix dolomites. The breccias are interpreted to have formed in space created by transtensional faulting with a component of dissolution by hydrothermal fluids flowing up the

faults. Using the appropriate integrated structural-stratigraphic-diagenetic model, more hydrothermal dolomite natural gas reservoirs are likely to be discovered in the Trenton and Black River Groups of eastern North America and in carbonates around the world.

Background

Oil and gas have been produced from laterally discontinuous brecciated, vuggy, dolomitized zones in the Upper Ordovician Trenton and Black River carbonates in eastern North America for more than a century. The first discoveries were made in the dolomites of the Lima-Indiana Trend of Ohio and Indiana (Figure 1) in 1884. Most of the dolomites are demonstrably fault-related as they align with the NNW-SSE trending Bowling Green Fault and other fault zones (Wickstrom et al., 1992). More than 500 million barrels of oil and over 28 million cubic meters of gas (one trillion cubic feet) of gas were produced from the Lima-Indiana Trend (Wickstrom et al., 1992), making it one of the first giant oilfields in the world. Smaller yet still significant discoveries were subsequently made in fault-controlled dolomites at Dover Field, Ontario, and Deerfield and Northville Fields in Michigan (Hurley and Budros, 1990).

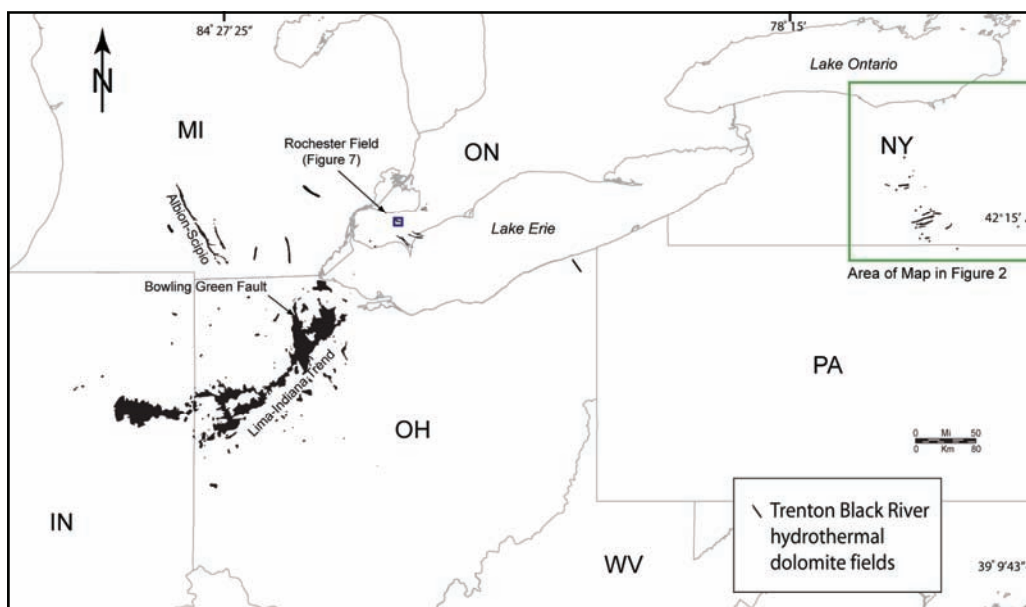


Figure 1. Map of Trenton and Black River hydrothermal dolomite fields in eastern North America.

The next major discovery in Trenton-Black River laterally discontinuous dolomites was the Albion-Scipio Trend in southern Michigan (Figure 1). Porous matrix dolomite, saddle dolomite-cemented breccias, saddle dolomite-lined fractures and vugs, occur in a trend of *en echelon* structural lows that is ~50 km long (30 miles) long and about 1.6 km (1 mile) wide (Hurley and Budros, 1990). Albion-Scipio Field has produced more than 250 MMBOE (million barrels of oil) since its discovery. Several smaller fields have subsequently been discovered in Michigan and Ontario.

Natural gas was discovered in laterally discontinuous dolomite in the Black River Group of south-central New York in 1986. Since then, at least thirty new fields have been discovered in laterally discontinuous dolomites of the Trenton-Black River (Figure 2). Most of the fields are between 2,000 and 3,000 m deep. Several wells have produced at sustainable rates of > 283, 000 m³/day (10 million cubic feet per day (MMCF/D)), and the best well produced in excess of 1 million m³/day (35MMCF/D for several months. As of the end of 2007, the well with the best cumulative production had produced more than 622 million m³ (22 billion cubic feet (BCF)), the best field (Quackenbush Hill) had produced more than 2.46 billion m³ (87 BCF) and the trend had produced more than 6.57 billion m³ (232 BCF) of gas.

These dolomitized fields occur in long, narrow, fault-bounded *en echelon* structural depressions (Harding, 1974; Prouty, 1988; Hurley and Budros, 1990; Colquhoun, 1991) and along other wrench and possible normal faults (Wickstrom et al., 1992). The reservoir facies consists of matrix dolomite with saddle dolomite cemented vugs, breccias and fractures around the faults, and the reservoirs are laterally sealed by tight, undolomitized limestone. Although some

fields have a significant component of cavernous porosity (Albion-Scipio is famous for its bit drops of up to ~18 meters (60 feet)), others, including those in New York, have little or no cavernous porosity.

Loucks (1999, 2003a, 2003b) recently suggested that the brecciation in Trenton-Black River reservoirs occurred when the Trenton and Black River carbonates collapsed into meteoric caves in the underlying Cambro-Ordovician Beekmantown Group. Later, deep-burial fluids then were interpreted to be responsible for the dolomitization and mineralization in porosity that stayed open for more than 100 million years (Loucks 1999, 2003a, 2003b).

Most other workers agree that the breccias, vugs and patchy dolomite formed due to faulting and associated fluid flow up basement-rooted faults (Harding, 1974;

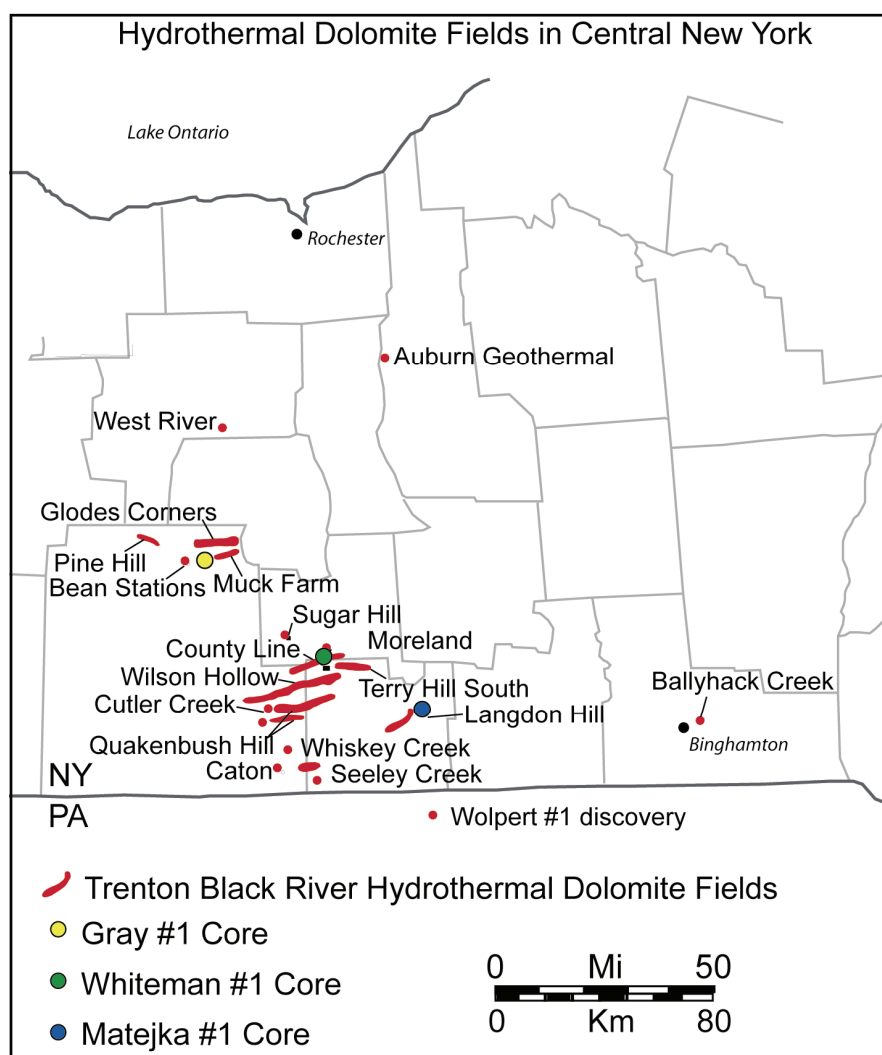


Figure 2. Map of producing dolomite fields in Black River Group of New York with locations of three cores discussed in the paper.

Prouty, 1988; Gregg and Sibley, 1984; Taylor and Sibley, 1986; Hurley and Budros, 1990; Colquhoun, 1991; Wickstrom et al., 1992; Coniglio et al., 1994; Davies, 2001; Smith, 2006; Smith and Davies, 2006) and has nothing to do with meteoric karst. Harding (1974), Prouty (1988) and Hurley and Budros (1990) suggested that the elongate Albion-Scipio Field in southern Michigan formed over a left-lateral strike-slip fault system and that hot fluids flowed up the faults and fractures and dolomitized the adjacent limestone. The structural sags were interpreted to have formed by transtensional faulting and the development of negative flower structures (Harding, 1974; Hurley and Budros, 1990). Brecciation is interpreted to have occurred due to space creation in transtensional fault zones (Davies and Smith, 2006). Wickstrom et al. (1992) suggested a similar origin for dolomite around the NW-SE trending Bowling Green Fault Zone in northwestern Ohio and Colquhoun (1991) made similar interpretations about the Hillman Field in southern Ontario.

The recently discovered natural gas fields in New York are similar to the Trenton-Black River Fields in Ohio, Michigan and Ontario. Saddle dolomite-lined vugs, fractures and breccias, and matrix dolomitization occur around subtle basement-rooted faults that are visible on seismic data. Away from the faults, the Black River is composed of unaltered limestone.

The purpose of this paper is to present structural, stratigraphic, petrographic and geochemical data from the dolomite fields in New York that together demonstrate that they formed entirely in the subsurface due to wrench faulting and bottom-up hydrothermal fluid flow and to show the similarity between the fields in New York with those in Ohio, Ontario and Michigan. The timing and depth of alteration is an ongoing controversy and data will be presented that demonstrates that the dolomitization and other hydrothermal alteration mainly occurred in the Late Ordovician, soon after deposition when the Trenton and Black River strata were buried less than 500 meters.

Geologic setting

The Grenville Basement, which underlies New York State and much of the Eastern United States, has been subjected to numerous tectonic events that produced numerous fault and fracture trends. The Grenville Orogeny occurred approximately 1.1 billion years ago (Moore, 1986) and during this event, thrust faults and associated tear faults formed. After the Grenville Orogeny, North America was part of a supercontinent that underwent a long-lasting episode of rifting in the Late Precambrian (620-550 ma: van Staal, 2005).

There were numerous failed rift zones such as the Rome Trough, Reelfoot Rift and many other features (Burke and Dewey, 1973; Thomas, 1991), some of which extend into New York. Failed rift extensional faults are interpreted to extend from Pennsylvania into south-central New York in an ENE trend (parallel to many of the trends of the Black River hydrothermal dolomite occurrences) and then turn toward a more NNE trend as they extend farther north and east toward the Mohawk Valley and the Adirondack Mountains (Jacobi et al., 2008). Along with the extensional faults there were associated strike-slip transfer faults (Thomas, 1991) which are typically oriented close to perpendicular to the extensional fault systems. The faults formed during these earlier times were likely reactivated during subsequent tectonic events.

After the Late Precambrian rifting of the supercontinent, New York was situated on a passive margin developed over the New York Promontory (Thomas, 1991). The Middle Cambrian Potsdam Sandstone (0-180 m thick) rests unconformably on the rifted basement and is overlain by the Cambro-Ordovician Beekmantown Group carbonates and siliciclastics (Figure 3). In New York, the Beekmantown is composed of the Cambrian Little Falls Dolomite, which has common vugs, breccias and fractures partially filled with saddle dolomite, "Herkimer Diamond" quartz crystals and anthraxolite, a carbonate mineral with solid hydrocarbon in the matrix. The upper part of the Beekmantown consists of the Tribes Hill Formation, which is dolomitized near some faults and mostly limestone away from faults where it outcrops in the Mohawk Valley of New York (Landing et al., 1996; Smith et al., 2004b; Slater et al., 2006). The Beekmantown and its equivalents across North America (Knox, Ellenburger, Arbuckle, St. George and Romaine) commonly host Mississippi Valley Type Lead-Zinc sulfide deposits and dolomitized oil and gas reservoirs. The global Lower Ordovician Knox Unconformity overlies the Beekmantown.

The overlying Black River Group is primarily composed of muddy and fine-grained shallow marine carbonates. The Black River is overlain by the Trenton Group, which is composed of deeper water argillaceous limestones and calcareous shales and high-energy shallow marine grainstones and packstones (Brett and Baird, 2002). The Trenton and Black River thin toward eastern New York, where they are absent in some places, and thicken into the south-central part of the state where most of the recent production occurs (Rickard, 1973). The Trenton is overlain by the deeper-water Utica Shale, which is a black shale that

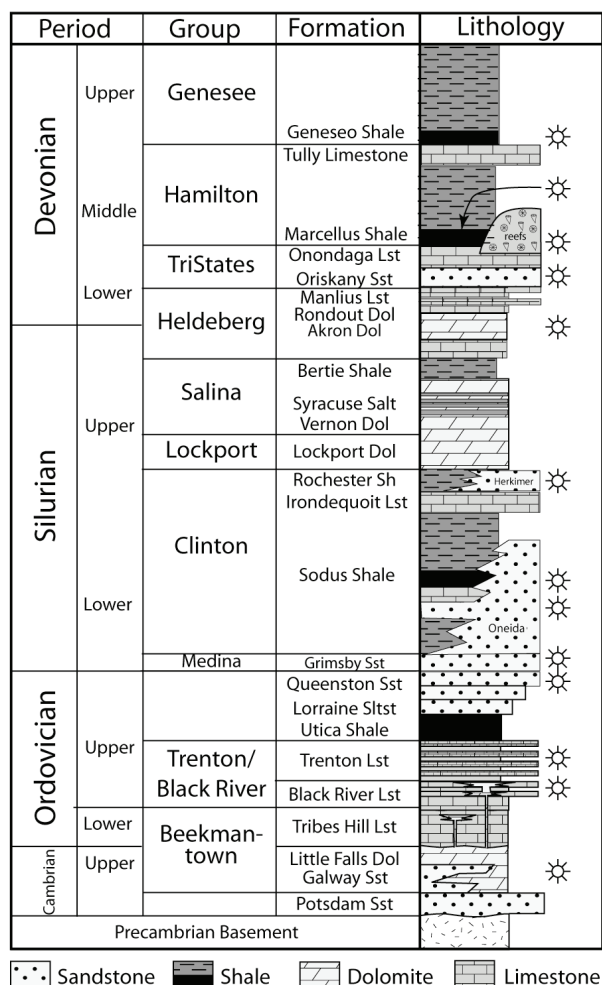


Figure 3. Generalized stratigraphic column for central New York. The Black River Group is earliest Late Ordovician in age.

blankets much of the eastern North America. This contact is diachronous as the Trenton Limestone grades laterally into the lower members of the Utica Shale to the south and east. The Utica Shale grades up into the Ordovician Lorraine Siltstone and Queenston Sandstone, which are siliciclastics that prograded from the Taconic Mountains across most of New York in the Late Ordovician.

The Taconic Orogeny began in the Late Ordovician when an island arc collided with proto-North America east of present-day New York and continued throughout the Late Ordovician and into the Early Silurian (Ettensohn and Brett, 2002). Both the Black River and Trenton Groups have bentonite (volcanic ash) beds in them, some of which can be correlated for great distances (Kolata et al., 1996). Further evidence for tectonic activity at this time includes development of a

foreland basin to the east, Late Ordovician extensional faulting in the foreland area of the Mohawk Valley (Bradley and Kidd, 1991), spatial and temporal variations in differential subsidence in the Trenton Group and the occurrence of seismites in the outcrops of Trenton and Utica-aged rocks in New York, Kentucky and Ohio (Pope et al., 1997; McLaughlin and Brett, 2002; Ettensohn et al., 2002).

Methods

Logs were analyzed and maps were constructed using *Petra* Software and then imported into *Adobe Illustrator* where they were manipulated for presentation. Cores were described using slabs and thin sections. Samples of matrix dolomite, saddle dolomite, calcite and quartz were collected separately for each set of analyses. In some cases, samples were split four ways so we could get fluid inclusion, stable isotope, trace element and strontium isotope data for the same samples.

Fluid inclusions were analyzed by Fluid Inclusion Technologies, Inc. using commonly accepted techniques and equipment. Polished slabs of rock material were prepared and studied optically with a petrographic microscope. Samples were placed into a gas-flow temperature stage (manufactured by Fluid Inc.) and individual inclusions in aqueous inclusion populations were viewed optically during heating and cooling (-196°C to +200°C or higher). Phase equilibria within the trapped fluids reflect their composition and bulk density, which, in turn, is a function of trapping temperature, pressure and fluid composition for each inclusion.

Stable isotopes values were measured by Peter Swart at the University of Miami. Samples were reacted for 10 minutes using the common acid bath method at 90°C and the CO₂ produced analyzed using a Finnigan MAT-251. Standard isobaric corrections were applied. Data for both C and O isotopes are reported relative to the Vienna Pee Dee Belemnite (V-PDB) using the conventional notation.

Trace elements and Strontium isotopes were analyzed by Mihai Ducea at the University of Arizona. About 100 mg were dissolved for each sample in 2.5M nitric acid. About 10% of that was used for trace element analyses; the remainder was taken up in 3.5M nitric acid and passed through Sr Spec chromatographic columns for rapid Sr elution. The separated Sr cut was then redissolved in 1% nitric acid.

Trace elements were analyzed in a mild nitric acid using a Perkin Elmer Elan DRC-II instrument. Stron-

tium isotopes were analyzed on a VG Sector 54 instrument. During the period of our analysis, the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were 0.71025 ± 1 for standard NBS 987.

Data

Maps and cross-sections

Figure 2 shows the distribution of Trenton-Black River producing hydrothermal dolomite fields and discoveries in New York. Many of these fields are in the process of being drilled and extended and this map is likely to look quite different when all drilling is completed.

The distribution of dolomite in the Trenton and Black River Groups helps explain its origin. The Trenton Group carbonates are primarily limestone and shaly limestone throughout most of New York and are only rarely dolomitized. This is in contrast to Ohio, Michigan and Ontario where the Trenton is commonly dolomitized and produces more hydrocarbons than the

underlying Black River. There is a thick shaly limestone at the base of the Trenton, where the fields are located in south-central New York, that thins and pinches out to the west. This may be the primary reason that the Trenton is rarely dolomitized in central New York. There is virtually no porosity in the limestone facies of both the Trenton and Black River facies and these impermeable limestones are thought to form a vertical and lateral seal on the hydrocarbon reservoirs. Furthermore, there is no “cap” dolomite in New York (*sensu* Wickstrom et al., 1992), which is a tight regional dolomite in the uppermost Trenton between 3-15 m thick that blankets much of north-western Ohio and southwestern Michigan.

Most of the dolomitization in New York occurs in the Black River Group and is laterally discontinuous. Where there are enough wells to map out dolomite bodies, the dolomite occurs in sub-linear trends that are up to 20 km long and 2 km wide (Figure 2). Undolomitized wells occur within short distances of

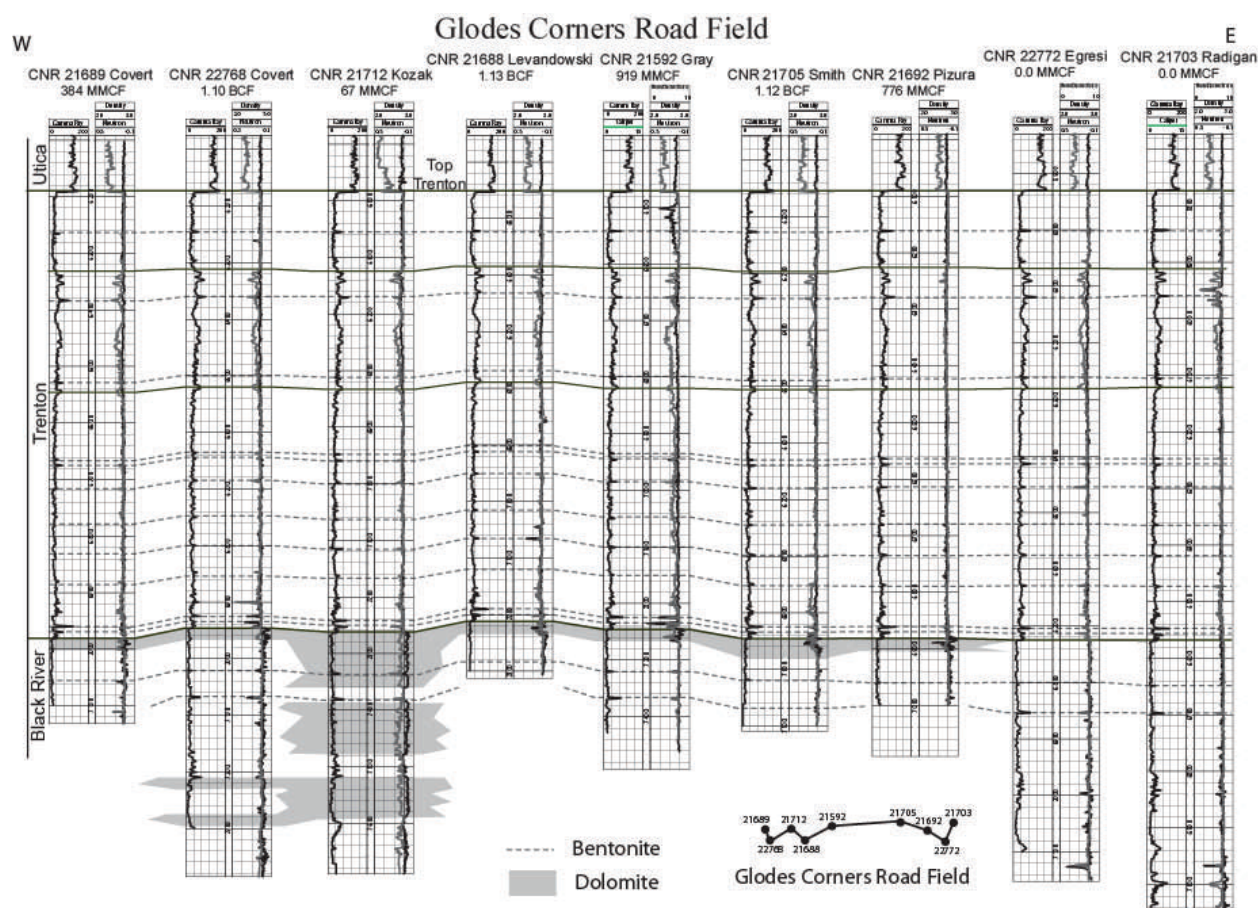


Figure 4. Cross section of Trenton and Black River Groups at Glodes Corners Road Field, New York. Dolomitized intervals are colored gray. The rest of the Trenton and Black River Groups are composed of limestone and shaly limestone. Well labels include the operator (Columbia Natural Resources or CNR), the shortened API number, lease name and cumulative production as of the end of 2007.

wells with (meters) of dolomite in the same stratigraphic interval. All productive wells have at least some porous dolomite in the upper half of the Black River Group and almost all wells have dolomite in the top 6 m. Dolomite thickness within the Black River ranges from a few meters to tens of meters (a few to hundreds of feet).

Figure 4 is a cross section of wells in and near the Glodes Corners Road Field (the discovery field for the New York trend). The field is about 10 km long and 0.7 km wide and has produced about 8 BCF. Within the Black River, the cross-section includes several dolomitized producing wells, two tight dolomite wells and a tight limestone well. Dolomite is picked using the PEF curve or where that is absent, in zones where the density log reads greater than 2.75 g/cc or where the density log (on a limestone scale) reads about 5% less porosity than the neutron log. When the Black River is productive, it is almost always dolomitized in the uppermost 15 meters (50 feet) under the argillaceous limestones at the base of the Trenton (Figure 4).

The thickening and thinning and laterally discontinuous nature of the dolomite suggests that the source of the dolomitizing fluids was highly localized.

Figure 5 is a cross section of some of the better wells in some of the major producing fields. The cross section trends from Glodes Corners Road in the northwest to Seeley Creek Field in the southeast. In many cases, once gas was discovered, operators would stop drilling and hook the wells up, so some logs do not go all the way through the Black River. Note the variable dolomitization from just a few meters in Glodes Corners Road and Muck Farms to near pervasive dolomitization in the Lant #1 (Terry Hill South Field) and Parker #1 (Wilson Hollow Field) to the more interbedded dolomitization in several other wells. This variability in dolomitization demonstrates again that the dolomitization was not regional but highly localized. The porosity distribution also varies from well to well, but the zone that is most consistently porous is in the top 15 m (50 feet) of the Black River.

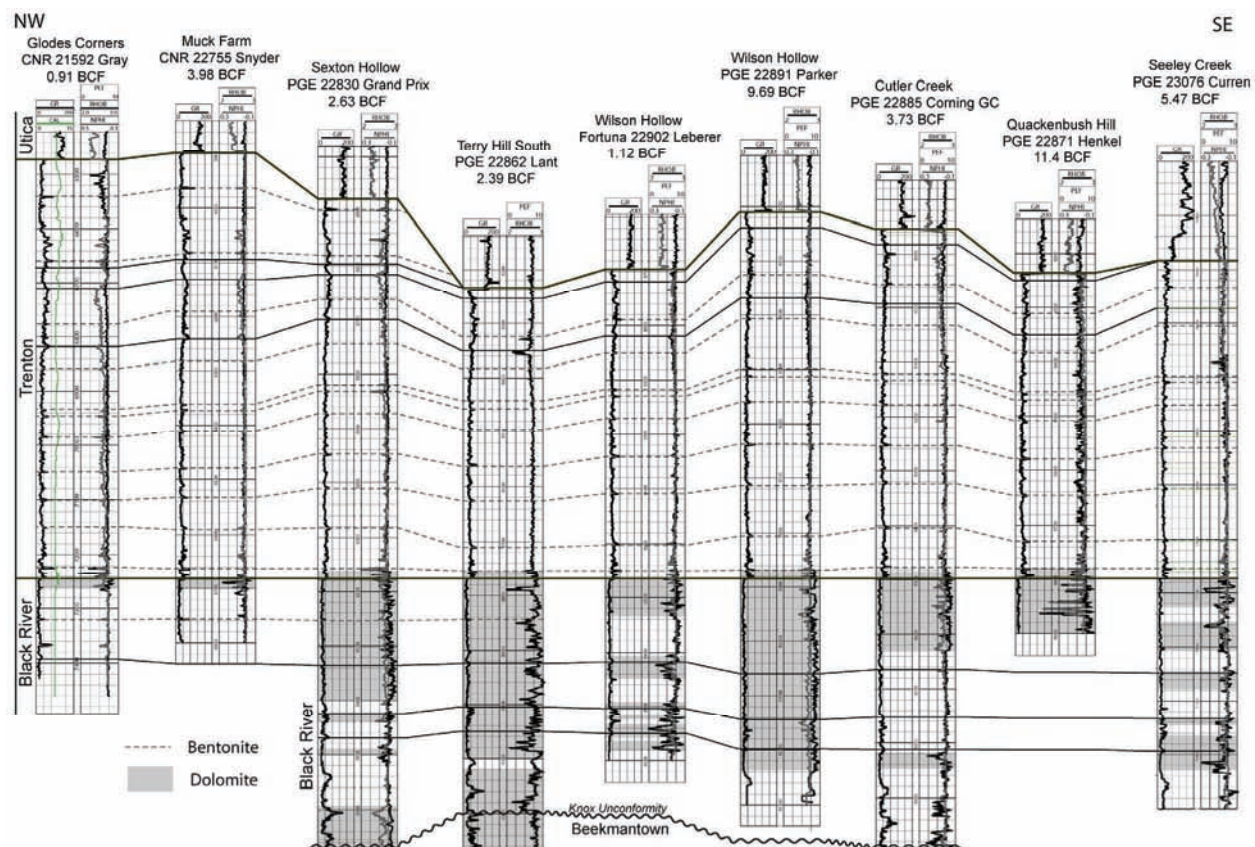


Figure 5. Cross section of Trenton and Black River Groups from NW to SE through eight of the producing fields in the trend. Note variable dolomitization and porosity development. Well labels include Field name, operator, shortened API number, lease name and production as of December 2007. The wells have produced for varying amounts of time and all continue to produce.

Seismic data

Trenton-Black River hydrothermal dolomite reservoirs are primarily discovered using seismic data. The dolomitized zones occur primarily in and around fault-bounded structural lows (Prouty, 1983; Hurley and Budros, 1990; Davies, 2004). These features are commonly called “grabens” or “sags.” Figure 6 is a time structure contour map of the top of the Trenton from a producing hydrothermal dolomite oil field in Ontario that shows *en echelon* sags. Oil was produced from fractured, vuggy dolomite located within the structural lows and most of the Trenton and Black River strata outside the sags are impermeable limestone (Horvath et al, 2004). The *en echelon* sags are interpreted to be negative flower structures formed as a result of transtensional (strike-slip with extensional component) faulting (*c.f.* Hurley and Budros, 1990).

Figures 7 and 8 show the seismic signature of several different productive Black River hydrothermal dolomite fields in New York. Figure 7 shows Muck Farms Field (Figures 7A, B and C) and Wilson Hollow Field (Figures 7D, E and F). The dolomite occurs in the Black River, typically within the structural lows. The fault picks in the basement (Figures 7B and 7E) are equivocal, but reflectors are demonstrably offset and faulted. Most of the vertical offset of the seismic traces occurs within the Black River with little offset in the overlying Trenton or underlying Beekmantown.

Most of the faults with associated dolomitization die out in the Trenton or overlying Utica Shale. Also note on the Wilson Hollow line that much of the structural sag affecting the lower Trenton and Black River is filled in by earliest Utica time (Figure 7E).

Figure 8 shows the seismic line across three different productive Black River hydrothermal dolomite fields in New York (seismic data courtesy of Talisman Energy). The dolomite occurs in the Black River, typically within the structural lows. Typical of some negative flower structures, reflectors immediately overlying the basement are not as obviously offset vertically by the faults while those higher in the section have clear vertical displacement (see cross sections in Dooley and McClay, 1997 for examples of how these might form). Interpretations in 2D are equivocal; the interpretations in Figures 7 and 8 are almost certainly wrong to some degree, but the features are almost certainly linked to some sort of basement-rooted strike-slip faults. Most of the vertical offset of the seismic traces occurs within the Black River with little offset in the overlying Trenton or underlying Beekmantown. Most of the faults with associated dolomitization die out in the Trenton or overlying Utica Shale and the sags are filled in by the Trenton and Utica sediments, which constrain the timing of faulting to Trenton and Utica time. Note that in both Figures 7 and 8 there is no evidence for collapse into the underlying Beekmantown.

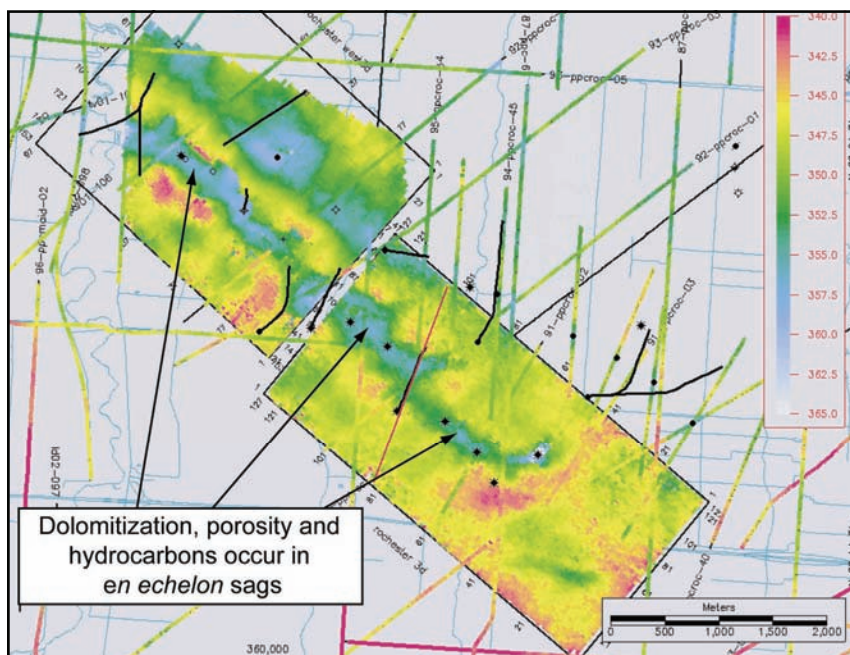


Figure 6. Time-structure map of the top of the Trenton Group, Rochester Field, Ontario based on 3D seismic data (courtesy Talisman Energy). Note *en echelon* sags where dolomitization, porosity and oil occur. Color bar scale is in meters below sea level. See Figure 1 for location of Rochester Field.

Cores

There are three publicly available cores from the Black River hydrothermal dolomite play in New York. Each of them has features that are important to understanding the play’s sedimentology, stratigraphy and diagenesis. The locations of the cores are presented in Figure 2.

Rock types and stratigraphy

The Black River Group carbonates were deposited on a shallow low-relief tropical carbonate ramp. Rock types include mudstone (sometimes with fenestrae and clay drapes), fine-to coarse-grained skeletal wackestone, and very fine-to-fine peloidal packstone and grainstone (see Cornell and Brett, 2000 for more detailed facies analysis). All of these rock

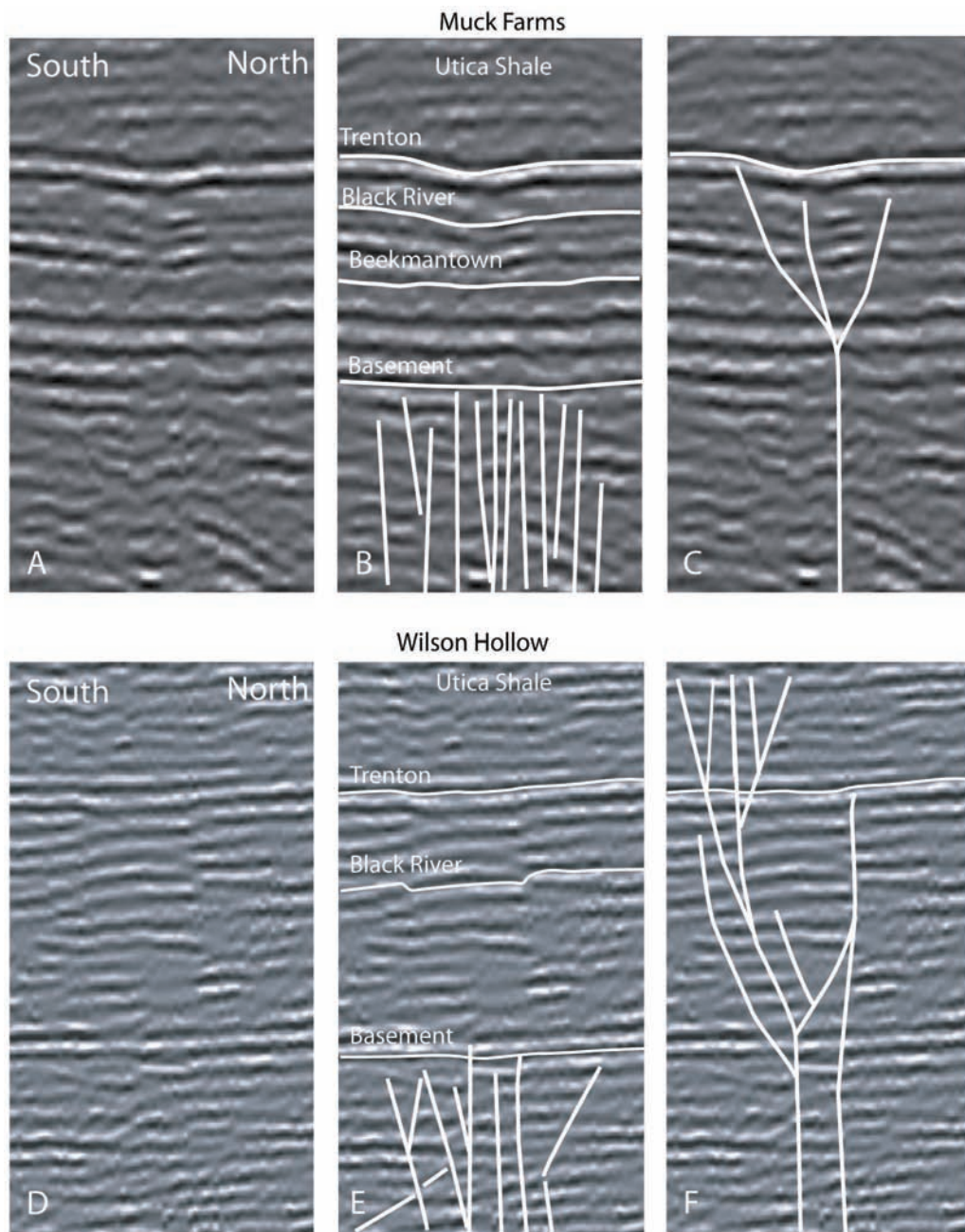


Figure 7. Seismic lines from Muck Farm and Wilson Hollow Fields, New York. A) Uninterpreted line over Muck Farm Field. B) Formation picks and possible faults in basement. Note sag at top of Trenton and lack of sag in Beekmantown. C) Interpreted flower structure with faults bounding structural sag. D) Uninterpreted line over Wilson Hollow Field. E) Formation and possible basement fault picks. Note that sag is accommodated in Trenton and lower Utica. F) Possible basement-rooted flower structure with faults bounding sag in Trenton and Black River.

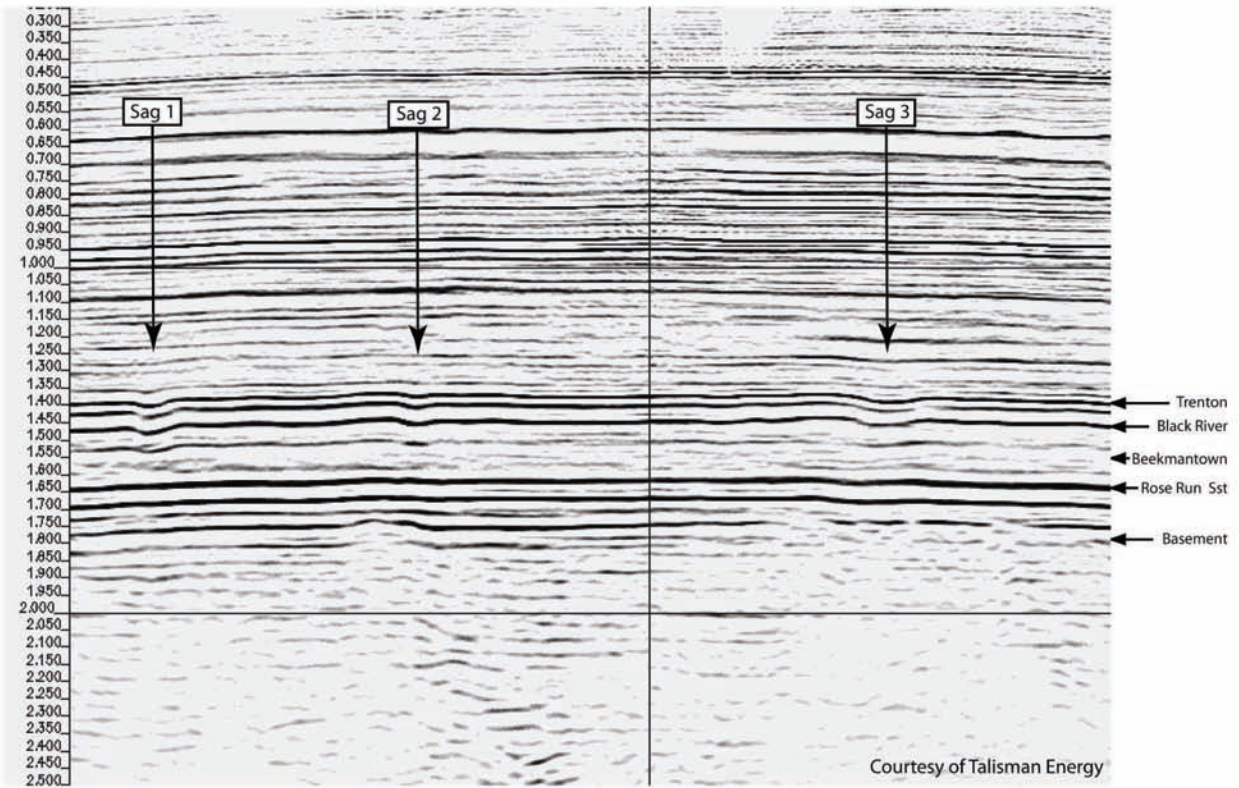
types can be dolomitized, but dolomitization of the peloidal grainstones and packstones may be more common. The clay drapes superficially resemble stylolites but are unrelated to pressure solution. Burrowing is common in most rock types. The Black River Group carbonates are remarkably consistent in rock type across much of eastern North America. None of these rock types is a reservoir facies in the absence of dolomitization.

The Trenton has grainstones and packstones that are coarser and more fossiliferous than those in the underlying Black River and also has common dark

gray to black shale (Brett and Baird, 2002). Deeper water rock types include black or dark gray shale, fossiliferous shale and skeletal wackestone. Shallower water facies occur in some locations and they include coarse-grained skeletal grainstone and packstone.

Matrix dolomite description

The reservoir rock types in the Trenton-Black River are hosted in laterally discontinuous dolomites. There is both matrix dolomitization and void filling saddle dolomite. Analysis of stained, impregnated thin sections from the cores shows that more than 95% of the dolomite is medium to coarse (50-400 micron)



Courtesy of Talisman Energy

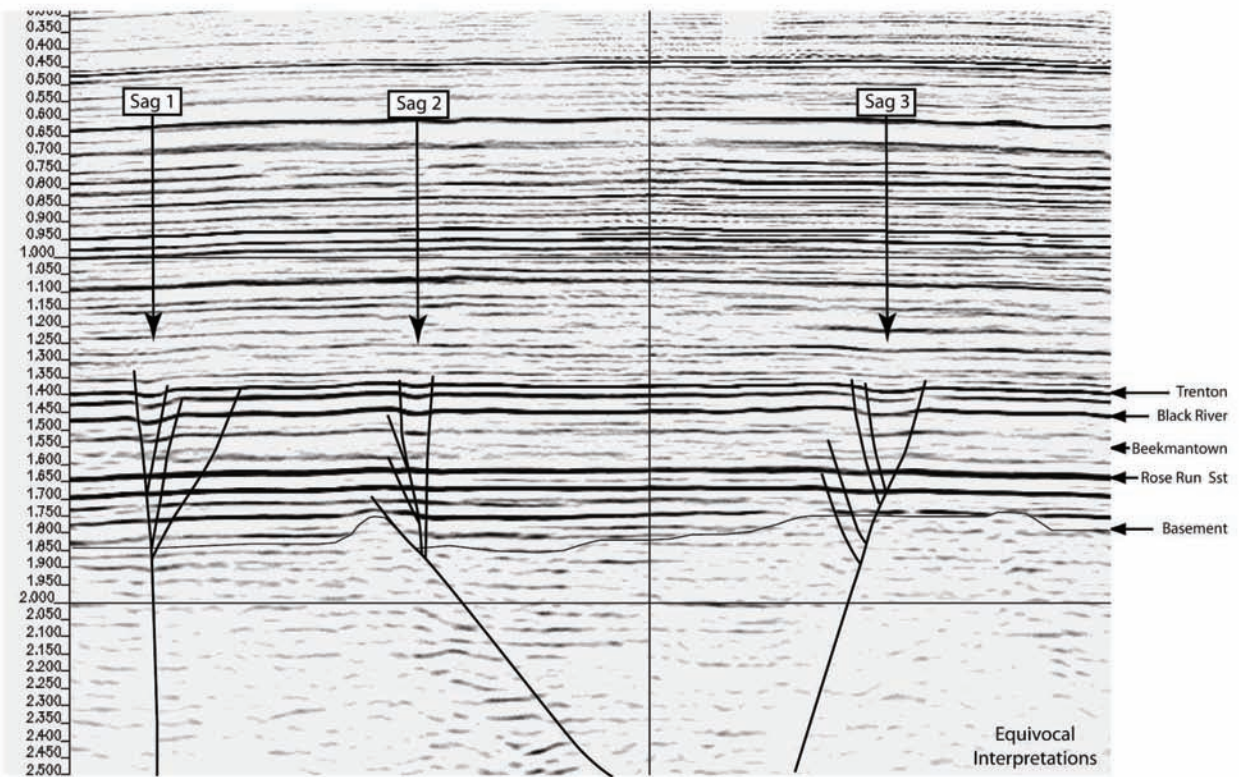


Figure 8. Seismic line over three producing fields in New York (courtesy of Talisman Energy). A) Uninterpreted line. B) Same line with possible interpreted fault scenarios.

matrix-replacive dolomite. The matrix dolomite has little or no porosity in the cores studied from New York. Matrix porosity may occur in other wells, but was not present in the cores studied. This lack of matrix porosity is in contrast to Trenton-Black River dolomitized reservoirs in Ontario and Ohio where matrix porosity is more common and makes up a significant part of the reservoir (Wickstrom et al., 1992; Colquhoun, 1991).

Much of the matrix dolomitization is fabric-destructive, anhedral and nonplanar (*sensu* Sibley and Gregg, 1986). Finer dolomites are here interpreted to have been mudstones and mud-dominated wackestones and the coarser dolomites are interpreted to have been peloidal or pelletal grainstones and packstones (c.f. Lucia, 1995).

Even though this play has significant structural and diagenetic aspects, stratigraphy is still important. Strata with higher porosity and permeability remaining at the time of dolomitization were probably more likely to have been dolomitized and to have had some porosity preserved after dolomitization. As more core data are collected, a major emphasis will be placed on learning what each of the dolomitized rock types were prior to dolomitization and what the impact of original facies is on reservoir quality.

Reservoir facies description

Most of the open pores in the Black River dolomites occur in saddle dolomite-lined fractures, vugs, zebra fabrics, boxwork fabrics and between clasts in breccias (Figures 9 and 10). Some vugs are sub-

spherical and several centimeters in diameter. Other vugs are elongate (a few millimeters wide and several centimeters long) and may be solution-enlarged fractures. Some of these elongate vugs are horizontal while others are inclined. Porous zebra fabrics and boxwork fabrics are both present (Figure 10). Breccias are common but not abundant and consist of millimeter-to centimeter-scale angular to corroded clasts cemented with saddle dolomite.

The Matejka #1 core (Figure 888) is in Chemung County (Figure 2) and was drilled in 1975, eleven years before the first official discovery. It encountered

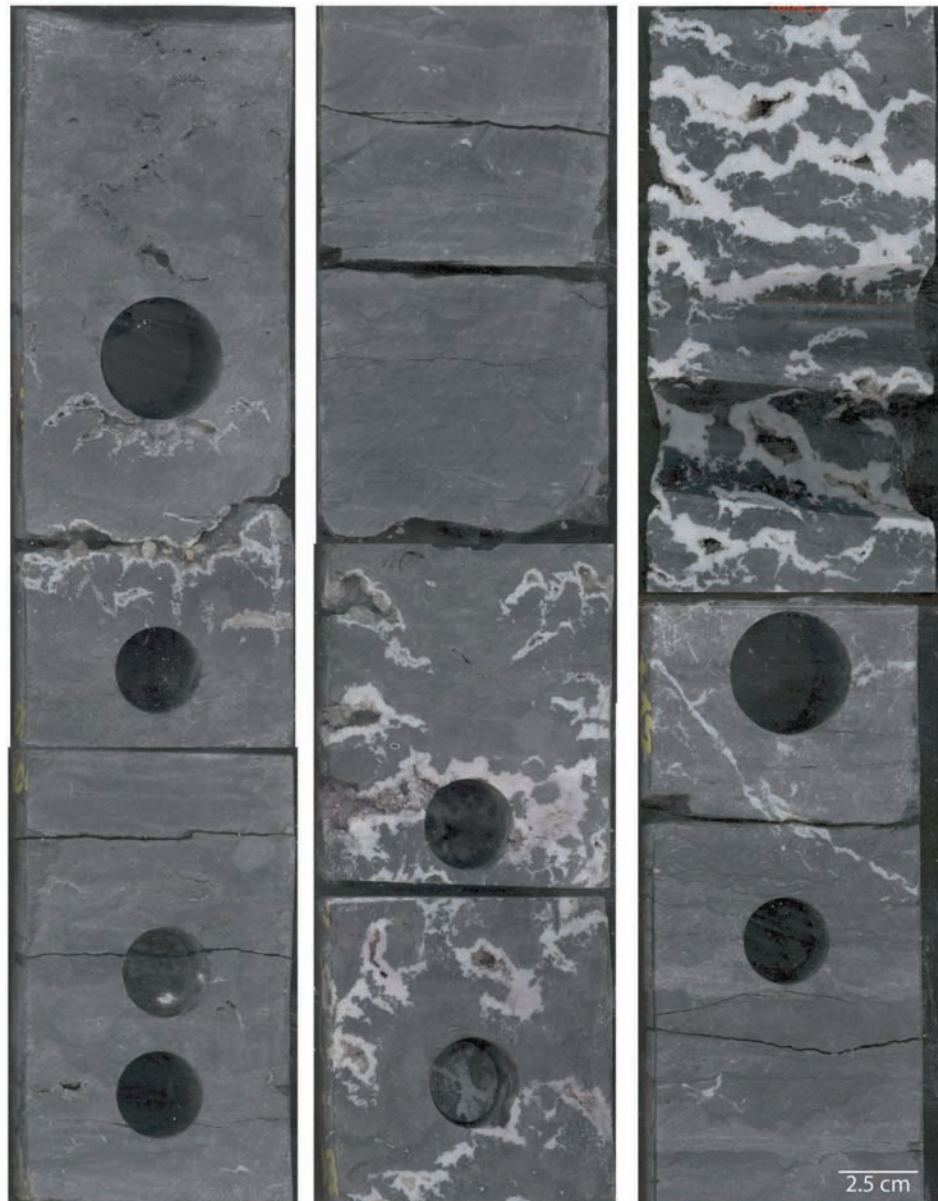


Figure 9. Core photographs from Gray #1 well, Steuben County, NY. Gray matrix dolomite with vugs lined with white saddle dolomite. Core is 9 cm (3.7 inches) wide.

dolomite in the Black River and heavily fractured limestone in the overlying Trenton Group. There is some porosity in fractures in the basal Trenton limestone that are mostly cemented with coarse calcite cement (Figure 11A). The dolomite in the Black River has no porosity (Figures 11B and 11C) and no open vugs, fractures or breccias. Saddle dolomite and calcite plug fractures within the dolomitized interval. The well was plugged as a dry hole, but it is located within a few hundred meters of a producing dolo-

mitized well and may well have produced from the fractures in the basal Trenton limestone. The Gray #1 core (Figure 9) was drilled to the west of Muck Farm Field and did not produce economic quantities of gas. The matrix in the upper half of the core is dolomitized and the lower half is predominantly limestone. Saddle dolomite-lined vugs occur in the dolomitized interval (Figure 9) but there are no open fractures or breccias. This well tested a very small amount of gas but not enough to justify facility

costs and may be sidetracked.

The Whiteman #1 core (Figure 10) is pervasively dolomitized and has common saddle dolomite-lined vugs, zebra fabrics, breccias and fractures. Porosity primarily occurs in vugs and fractures, and there is little or no matrix porosity. The well has produced approximately 0.5 BCF in three years as of the end of 2004 and is still producing today.

Figure 12 shows core samples from the Lima-Indiana Trend in Ohio and the Albion-Scipio Trend in Michigan, which are the two largest oil fields in the Trenton Black River laterally discontinuous dolomite play. They show similar saddle dolomite lined vugs and breccias with saddle dolomite cement.

Paragenetic sequence

The paragenetic sequence of the Black River dolomitized reservoirs is presented in Figure 13 and supported with photographs in Figure 14. Prior to dolomitization,



Figure 10. Core photographs from Whiteman #1 well, Chemung County, NY. This core has common fractures, vugs, zebra fabrics (ZF), boxwork fabric (BF) and an overall brecciated appearance. White saddle dolomite lines vugs and fractures in gray matrix dolomite. Core is 5 cm (2 inches) wide.

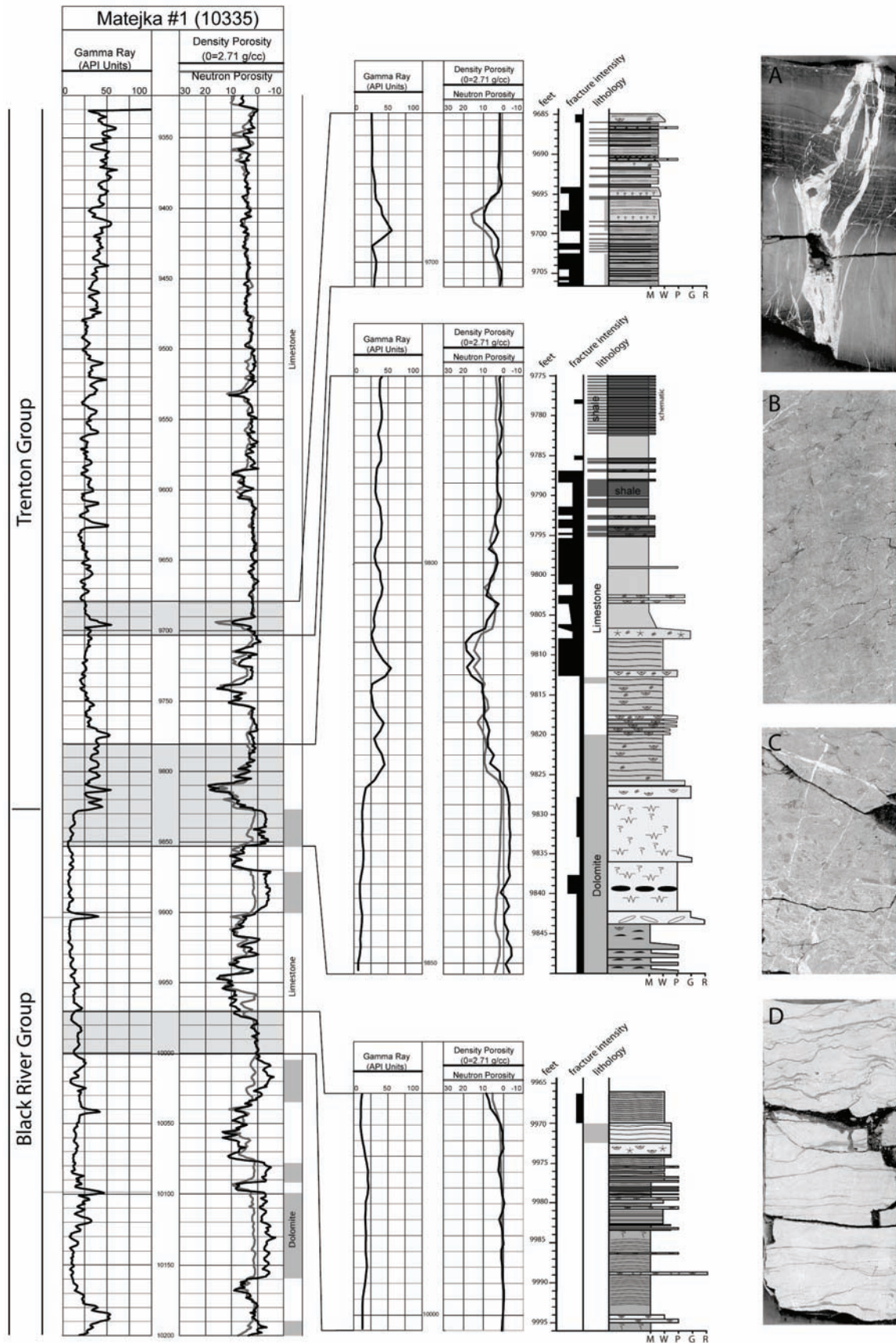


Figure 11 (opposite page). Core description with logs and photos from Matejka # 1 well, Chemung County, New York. Dolomitized intervals in core match with intervals where density log reads 5% less porosity than neutron log. Color code for core description: grays are deeper water mudstone, wackestone and shale, dark blue is mid ramp skeletal wackestone, light blue is shallow marine peloidal and skeletal grainstone (interpreted in dolomitized interval), tan is shallow marine argillaceous mudstone and wackestone. (M = mudstone; W = wackestone; P = packstone; G = grainstone) Photographs are lettered: A) Lower Trenton Group interbedded fossiliferous shale and mudstone is faulted and cemented with calcite. This sample is from 9787' and has some porosity in the fracture. B) Uppermost Black River dolomite with relict clay fills, minor fractures and tight, gray matrix dolomite. C) Dolomitized conglomerate from upper Black River Group with calcite cemented fractures and no porosity. D) Typical Black River Limestone from lower core is shallow marine mudstone with clay laminae.

some early marine and shallow burial calcite cementation and compaction occurred along with some minor grain suturing. This early diagenesis was followed by matrix dolomitization (Figure 14A), faulting, fracturing and brecciation (Figure 10), and then saddle

dolomite (Figure 14A), quartz (Figure 14B), rare authigenic feldspar (Figure 14C), pyrite, bitumen and calcite precipitation (Figure 14D). In some cases, the late calcite and/or saddle dolomite are leached by a later event (Figure 14E). Figure 14E shows bitumen in a pore that apparently was precipitated between dolomite rhombs that were later leached. Major stylolitization follows all mineralization (Figure 14F). There are no dolomite filled fractures that cut across major stylolites and matrix and saddle dolomite rhombs are clearly consumed at stylolites. All these are postdated by hydrocarbon migration, which likely occurred in the Late Paleozoic. The bitumen found lining many pores and fractures is not thought to represent a full-scale migration event.

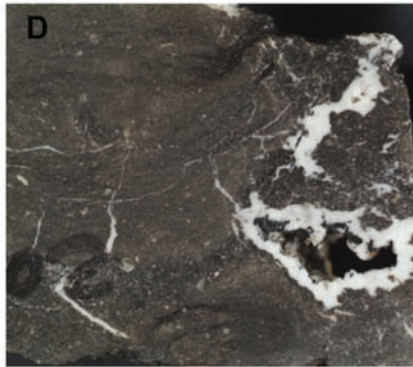
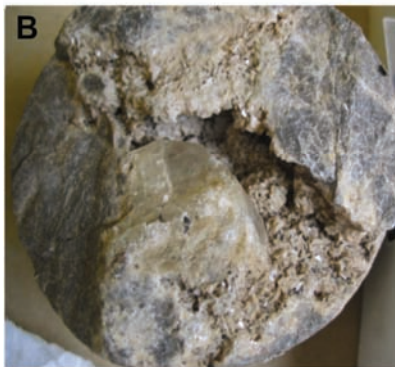


Figure 12. Core Photographs from Trenton/Black River brecciated, vuggy laterally discontinuous dolomite reservoirs in Bowling Green Fault Trend, NW Ohio (A, B) and Albion-Scipio Field, S. Michigan (C,D – photos courtesy of Mike Grammer).

The timing of the development of the vugs is equivocal. It may be that most vugs formed before the matrix dolomitization (the preferred interpretation) during a period of pre-dolomitization leaching. Or it may be that they formed after matrix dolomitization but before saddle dolomite was precipitated through matrix dolomite dissolution. There is no evidence of significant dolomite dissolution such as corroded rhombs lining the vugs. In some cases, vugs appear to be enlarged fractures (Figure 11), which would make their development syn- or post-fracturing.

Geochemistry and fluid inclusions
Geochemical and fluid inclusion analysis of the dolomites helps to understand their origin (Allan and Wiggins, 1993). We have con-

Geochemistry and fluid inclusions
Geochemical and fluid inclusion analysis of the dolomites helps to understand their origin (Allan and Wiggins, 1993). We have con-

Event	Surface and near surface	Fault-Related Hydrothermal	Deep Burial	Impact on Reservoir
Micritization	—			
Marine calcite	—			degrade
Syntaxial+ blocky calcite	—			degrade
Faulting, Fracturing		—		enhance
Leaching of limestone, development of vugs		—		enhance
Brecciation		—		enhance
Matrix Dolomitization		—		degrade
Saddle Dolomite		—		degrade
Quartz Cement			—	degrade
Authigenic Feldspar			—	degrade
Sulfides			—	degrade
Bitumen			—	degrade
Dolomite Leaching			—	enhance
Blocky Calcite Cement			—	degrade
Stylolitization	—	—	—	degrade
Hydrocarbon migration			—	

Figure 13. *Paragenetic sequence of events in dolomitized cores from Black River Group of New York.*

ducted stable isotope, strontium isotope, trace element of the saddle dolomite, blocky calcite, matrix dolomite and matrix limestone and fluid inclusion analysis of the saddle and matrix dolomite and blocky calcite.

Fluid inclusions

Primary fluid inclusions from the saddle dolomite in the New York wells have homogenization temperatures between 110 and 170°C with an average of 130°C (Figure 14). Inclusions of an equivocal primary or secondary origin from the matrix have a similar range of homogenization temperatures with a higher average of 145°C. Secondary fluid inclusions from the saddle dolomites have homogenization temperatures up to 180°C with an average of 173°C. Matrix dolomites have equivocal primary/secondary inclusions with fluid inclusion homogenization temperatures that range from 135-165°C with an average of 145°C. Post-dolomite primary and secondary quartz fluid inclusions have homogenization temperatures ranging from 155° to more than 200°C with an average of 177°C. There were no petroleum inclusions found in any mineral.

The salinities of the primary fluid inclusions in both the saddle dolomite range from 13.2 to 15.5 wt % with an average of approximately 14.4 wt % (approximately four times normal seawater). The salinity values of the equivocal inclusions in the matrix dolomites are similar with an average of 14.9 wt%. The secondary fluid inclusions in the post-

dolomite quartz cement average 16.7 wt. %. These values show that the fluid that made the dolomites and quartz were saline brines. The values for the New York dolomites are not as saline as many brines thought to have precipitated hydrothermal dolomite in the Trenton-Black River in Ontario, Ohio and Michigan and elsewhere in the world, which average closer to 20 wt.% (Allan and Wiggins, 1993; Coniglio et al., 1994).

Stable isotopes

Stable isotopes of carbon and oxygen were analyzed from matrix dolomite, saddle dolomite and limestone where present (Figure 16). The limestones in the Black River of New York have consistent $\delta^{18}\text{O}$ values around -6.5‰ . Due to fractionation, dolomite that precipitated from the same water at the

same temperature should have $\delta^{18}\text{O}$ values of around -3.5‰ (Friedman and O'Neil, 1977). The matrix and saddle dolomites in the Black River have $\delta^{18}\text{O}$ values between -9 and -12.5‰ . Increased temperature and mixing of fresh water both can drive $\delta^{18}\text{O}$ values toward more negative values. The salinity of the primary fluid inclusions in the dolomites averages four times greater than seawater, which eliminates fresh water as a source of lighter (more negative) $\delta^{18}\text{O}$ values.

Increased temperature is then the likely source for the depletion of $\delta^{18}\text{O}$. The actual temperature of formation cannot be determined without first plotting the $\delta^{18}\text{O}$ values vs. fluid inclusion homogenization temperatures from the same samples and backing out the oxygen isotope composition of the fluid. Figure 17 shows the $\delta^{18}\text{O}$ values and fluid inclusion homogenization temperatures for the matrix and saddle dolomites in the Trenton-Black River in New York. The homogenization temperatures from each sample were averaged and plotted with $\delta^{18}\text{O}$. It suggests that the fluid was somewhere between 0 and +4 with an average of +2 ‰. Note that this is considerably heavier (more positive) than the calculated composition of Late Ordovician seawater (around -6‰ based on limestone values). If one assumes that the fluid composition did not vary much from around +2 ‰, the temperature of formation can be estimated with some degree of precision from stable isotope values alone using the graph in Figure 16.

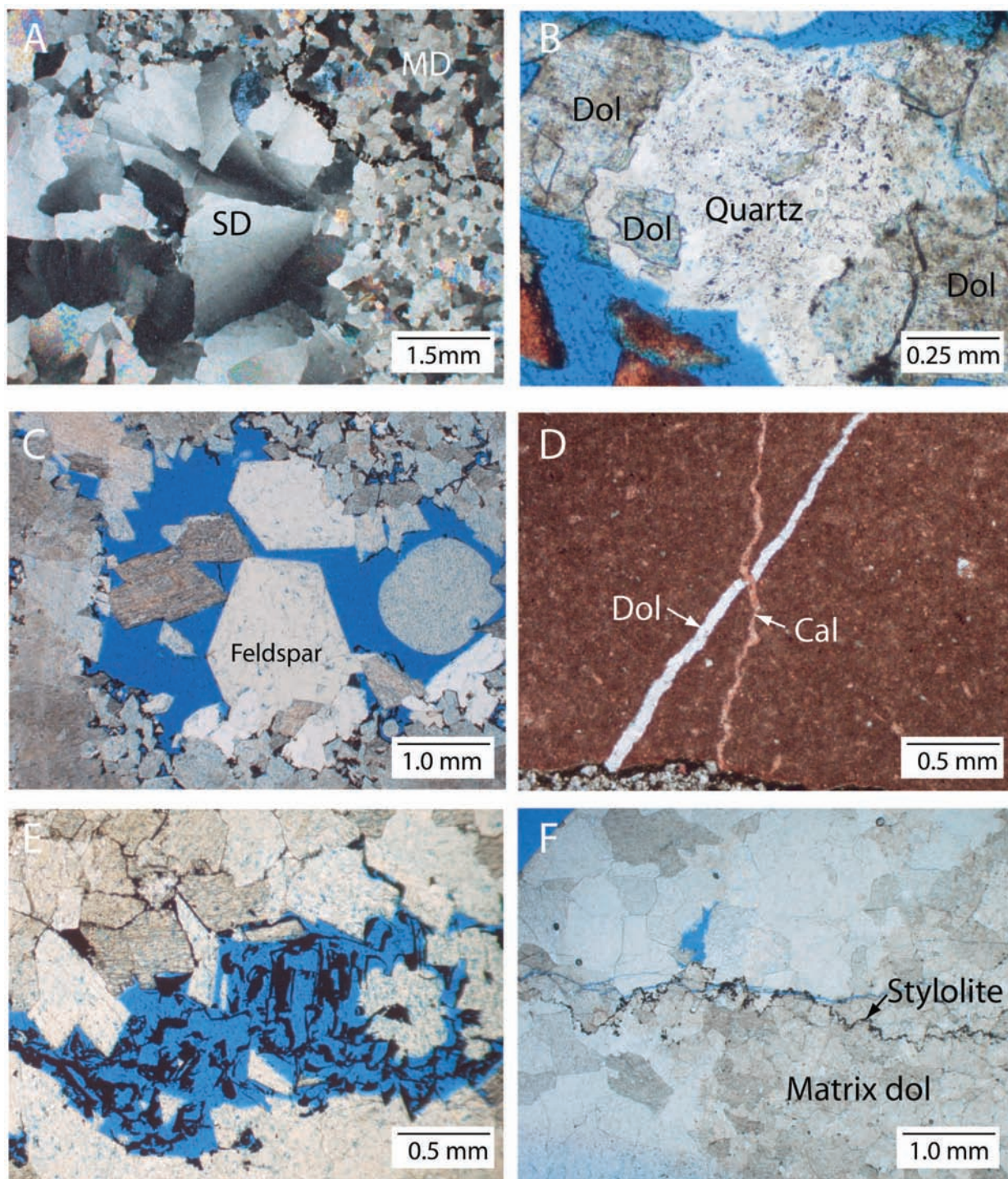


Figure 14. Thin section photographs from Black River Group, New York. A) Saddle (SD) and matrix (MD) dolomite. Gray #1 core, 2,378 m. B) Quartz cement in fractured ferroan dolomite. Auburn geothermal well cuttings, Onondaga County, NY, 1,265 m. C) Authigenic feldspar in vug. Whiteman #1 core, 2,904 m. D) Calcite-filled fracture clearly postdates dolomite-filled fracture. Late calcite commonly plugs porosity in Black River reservoirs. Gray #1 core, 2,382 m. E) Bitumen distribution in vug suggests that dolomite may have been leached after bitumen was precipitated. Whiteman #1 core, 2,906 m. F) Stylolite clearly postdates dolomitization. No dolomite grows across stylolites, all rhombs are consumed at stylolites. Gray #1, core 2,378 m.

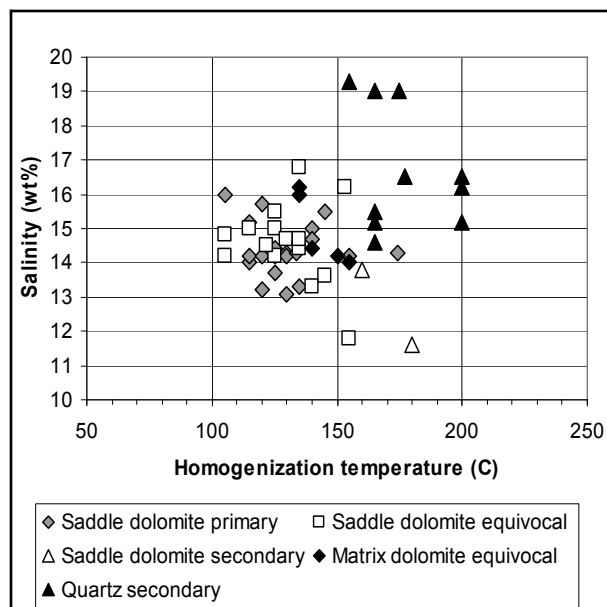


Figure 15. Fluid inclusion homogenization temperatures vs. salinity for samples from the Whiteman #1 and Gray #1 cores.

Trace elements

The trace element data show that the Trenton-Black River dolomites are enriched in iron and manganese relative to the limestones (Figure 18). The dolomites are enriched in iron and manganese relative to those that might have formed in normal seawater. The iron content of the dolomites ranges from 1600 to 13,804 ppm and has a median value of ~4250 ppm. In contrast to the dolomites, the limestones from the same cores have a median value of ~550 ppm Fe. The

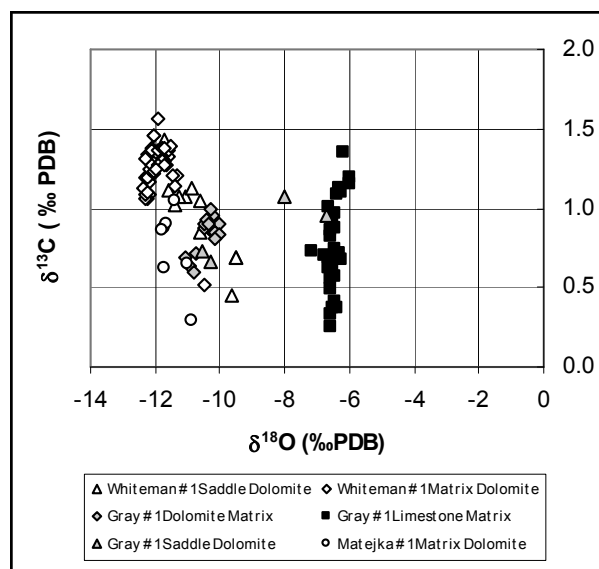


Figure 16. Stable isotope values for samples from the Whiteman #1, Gray #1 and Matejka #1 cores. Values reported versus PeeDee Belemnite standard (PDB).

dolomites are also enriched in manganese and have a median value of 880 ppm Mn while the median value for the limestones is only 52 ppm. The relatively high Mn and Fe values of the dolomites support a subsurface origin for the dolomites (*c.f.* Montanez, 1994).

Strontium isotopes

Dolomites that formed from subsurface brines commonly (but not always) have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that are higher (more radiogenic) than seawater for the time

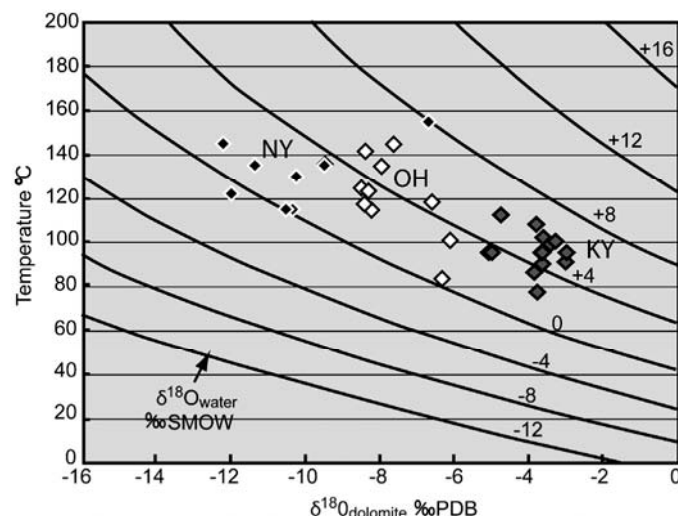


Figure 17. Average fluid inclusion homogenization temperature vs. $\delta^{18}\text{O}$ for selected samples in Gray #1 and Whiteman #1 wells. Water values measured versus standard mean ocean water (SMOW).

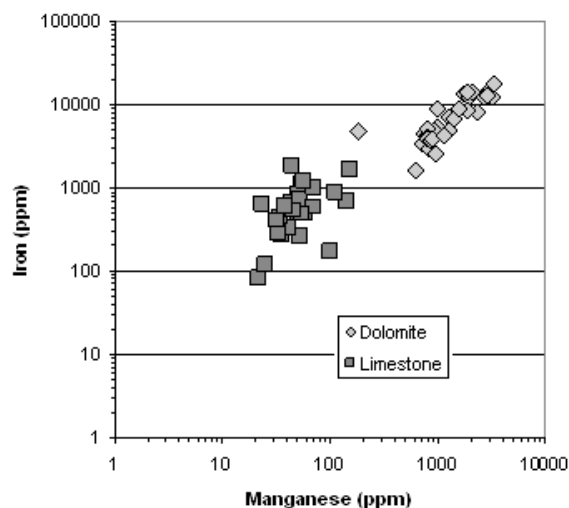


Figure 18. Trace element composition of limestones and dolomites in Gray #1 and Whiteman #1 wells. Dolomites strongly enriched in Mn and Fe relative to limestones.

that they formed (Allan and Wiggins, 1993). The range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of seawater for Trenton and Black River time is between 0.7078 and 0.7085 (Burke et al., 1982). The strontium isotope values for the dolomites in the Black River range from 0.7085 and 0.7092 for both the matrix and saddle dolomites and are just above the range for this time interval. While not strongly radiogenic (enriched in ^{87}Sr), the dolomites do plot above the range for seawater at the time of deposition, which suggests that the fluid that formed the dolomite passed through basement rocks or immature feldspar-rich siliciclastics prior to precipitating the dolomite.

Summary of geochemistry

The geochemical and fluid inclusion data suggest that the dolomites formed from a hot (110–200°C), saline (3–6 times normal seawater), iron- and manganese-rich, fluid that passed through basement rocks or immature siliciclastics prior to making the dolomite.

Discussion

Hydrothermal diagenesis occurs when fluids are introduced to a given formation at a temperature that exceeds the ambient temperature of that formation (*sensu* White, 1957; Davies and Smith, 2006). By this definition, there is not a set temperature range at which hydrothermal alteration occurs, the water simply must be warmer than the ambient temperature of the formation due to the local geothermal gradient. For example, if a formation is buried to a depth where the ambient temperature is 50°C and a fluid is introduced that is 60°C, that fluid would here be called a hydrothermal fluid. In the absence of local igneous intrusions, the most likely way for a hydrothermal fluid to be introduced to a formation is via rapid upward fluid flow from greater depths through high-permeability faults and fractures (Deming, 1992). Lateral and vertical unfocused fluid flow through porous and permeable formations, in the absence of faults and fractures, is in most cases too slow to generate hydrothermal conditions because fluids equilibrate with the ambient temperature as they migrate laterally through the formation (Deming, 1992).

Solubility of carbonates (and other minerals) is directly affected by changes in temperature, pressure, partial pressure of CO_2 (P_{CO_2}), pH, and salinity, and all of these are fluctuating on short time scales in fault-related hydrothermal systems (Rimstidt, 1997). Subsurface fluids flowing rapidly up faults and fractures can maintain most of their physical and chemical attributes until they are introduced to the formation where they are then capable of producing

significant diagenesis in short periods of time.

Hydrothermal alteration is in this case thought to occur when relatively high-pressure, high-temperature fluids flow up active transtensional faults and laterally into permeable formations that underlie sealing shales or other low permeability strata. Mixing with in situ fluids may also produce significant diagenetic alteration. The main diagenetic features produced by the hydrothermal fluids in this case are leaching and dolomitization along with less abundant quartz, bitumen and feldspar mineralization.

Timing and depth of burial during alteration

Hydrothermal fluid flow is thought to be most common while faults are active and much less common during periods of tectonic quiescence (Sibson, 1990, 2000; Davies, 2001; Knipe, 1993; Muir Wood and King, 1993; Davies and Smith, 2006). The timing of hydrothermal alteration is closely linked to the timing of fault movement.

Most of the faults that have associated dolomitization in New York appear to have been active during Trenton and Utica time (during the Late Ordovician Taconic Orogeny) and largely inactive after that time (Smith et al., 2003a, 2003b). On seismic data, most dolomitized wrench faults in New York die out in the Trenton and Utica and sags are commonly filled in during Trenton or Utica time (see Figures 7 and 8). This suggests that most of these faults were active during the Taconic Orogeny but were not reactivated during subsequent mountain building events.

During Trenton and Utica time, when most of the faulting appears to have occurred, the Black River was buried to depths of less than 350 meters (1100 feet) in New York. If hydrothermal alteration occurred during the period of active faulting, the Black River was buried to depths of less than 350 meters when the faulting, brecciation, vuggy porosity development and dolomitization occurred.

Demonstration of a hydrothermal origin

As a first pass, if dolomitization is highly localized and patchy, the patchiness can be linked to faults and the geochemistry and fluid inclusions support a high-temperature subsurface origin, the mineralization is likely to be hydrothermal in origin. This is the case with the Black River dolomites in New York. The dolomite is patchy, occurs near faults visible on seismic and has geochemical attributes that suggest a hot, subsurface origin. This strongly suggests that fluids flowed up the faults and precipitated dolomite. The sags associated with the faults were filled during

Trenton and Utica time when the Black River was only buried to a depth of a few hundred meters. If fluid flow up faults is most likely to occur when the faults are active, as has been suggested by many who study this process (Knipe, 1993; Sibson, 2000; Muir Wood and King, 1993), dolomitization of the Black River is likely to have occurred as a result of fault-related hydrothermal fluid flow during the first few hundred meters of burial.

The field relations between the faults and dolomitization and the hot subsurface origin of the dolomite strongly suggest a fault-controlled hydrothermal origin that can be argued support a strong enough case to approach these as hydrothermal dolomite reservoirs. Some workers require more rigorous proof (Machel and Lonee, 2002), insisting that some dolomites with the same characteristics might have formed due to slow lateral or vertical flow within a deeply buried formation.

A more rigorous method to demonstrate a hydrothermal origin for dolomite is to determine when dolomitization occurred during the burial and thermal history of the formation in question and compare that to the fluid inclusion homogenization temperatures in the dolomites (Davies, 2001; Machel and Lonee, 2002). If the homogenization temperatures exceed the maximum temperatures that the formation has ever been exposed to during burial, or the known depth and ambient burial temperature at the time of dolomitization, the dolomite or other minerals can be called unequivocally hydrothermal (Machel and Lonee, 2002).

Weary et al. (2001) showed that the area where the dolomite fields are located in New York has experienced very high burial temperatures using conodont alteration indices (CAI) from the overlying Utica Shale. All of the cores examined for this study are located in the area where the CAI values from the overlying Utica Shale are 4.5, which suggests that the Utica was heated to between 187-354°C (Hulver, 1997) or 150-300°C (Harris, 1979). Therefore, in this case, the Black River was buried to a depth where the temperature was equal to or greater than the fluid inclusion homogenization temperatures in the dolomites (Figure 19). This does not mean that the dolo-

mites in New York are not hydrothermal in origin, just that the formation was at some point in time buried to a temperature that exceeded the temperature at which the dolomites formed. If the dolomitization was during early burial, when the ambient temperature was still <100°C (which it appears to have been based on the timing of fault movement and interpreted fluid flow) the dolomite would still be considered to be hydrothermal in origin.

Confirmation of hydrothermal origin for similar Trenton-Black River dolomites comes from Ontario, Michigan and northwest Ohio (Figure 19). In most respects, the dolomitized horizons look the same with matrix dolomitization, breccias, saddle dolomite, fractures and vugs and the clear link to faults. At the Hillman Field in Ontario, fluid inclusion homogenization temperatures for the dolomites ranged from 100-220°C (Coniglio et al., 1994), but CAI analysis in that area suggests that the Trenton Group was never buried more than a kilometer in that area (Colquhoun, 1991). Using a geothermal gradient of 25-30°C/km and a surface temperature of 20°C, the maximum burial temperature was 45-50°C and the homogenization temperatures in the dolomites exceed the maximum ambient burial temperature by 50-170°C. A similar scenario occurs in Michigan where fluid inclusion

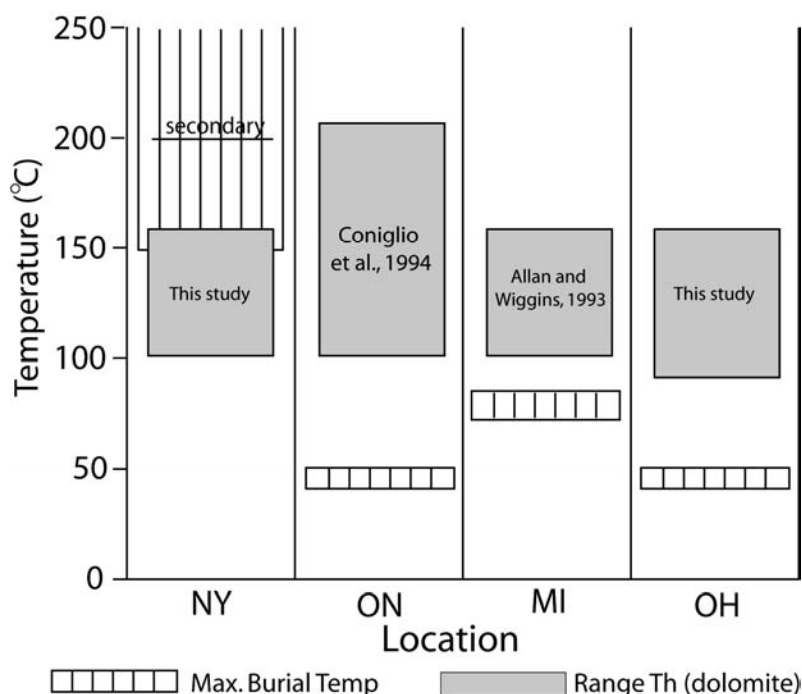


Figure 19. Fluid inclusion homogenization temperature vs. maximum burial depth. Ohio CAI data from Rowan et al., 2004. Ontario fluid inclusion data from Coniglio et al., 1994 and CAI data from Colquhoun, 1991. Michigan fluid inclusion data from Allan and Wiggins, 1993 and CAI data from Repetski et al., 2004.

homogenization temperatures from Trenton-Black River dolomites at Albion-Scipio Field exceed the maximum burial temperature by 40-90°C (using fluid inclusion data from Allan and Wiggins (1993) and CAI data from Repetski et al. (2004)). Fluid inclusion homogenization temperatures from the Lima-Indiana Trend in northwest Ohio for matrix and saddle dolomites range from 100-160°C, but the Trenton was never buried more than 800-1000 meters in this area (Fluid inclusions analyzed for this report from core in Bowling Green Fault Trend and CAI data from Rowan et al., 2004). Therefore, the Trenton-Black River dolomites are hydrothermal beyond a reasonable doubt in Ohio, Michigan and Ontario.

The dolomites from the fields in Ohio, Michigan and Ontario are virtually identical to those found in New York in appearance, association with wrench faults and geochemical attributes. It is unlikely that they would have formed by a different process. Further evidence of relatively early alteration comes from the salinity data. All reservoirs in New York below the halite in the Silurian Salina Formation have salinities near 300,000 ppm or 30 wt% (Matsumoto et al., 1992). The fluid inclusions in the Black River dolomites have salinities of approximately 15 wt%. If they had formed after the Silurian halite, the fluid inclusions in the dolomites would probably have much higher salinity values.

Origin of sags

The structural sags could form as a result of the combination of the development of negative flower structures, dissolution of limestone or dolomite and/or the volume reduction associated with the dolomitization of limestone. Negative flower structures form in

transtensional parts of strike-slip fault zones. Negative flower structures can show an apparent volume loss because strata are displaced laterally as well as vertically (e.g., Dooley and McClay, 1997). It has been convincingly demonstrated that many Trenton-Black River Fields form in structural sags associated with strike-slip faults and negative flower structures (e.g., Prouty, 1988; Hurley and Budros, 1990; Davies, 2001). This model can explain most or all of the sag development in the Trenton-Black River reservoirs (Figure 20).

Some part of the sags could have been produced by

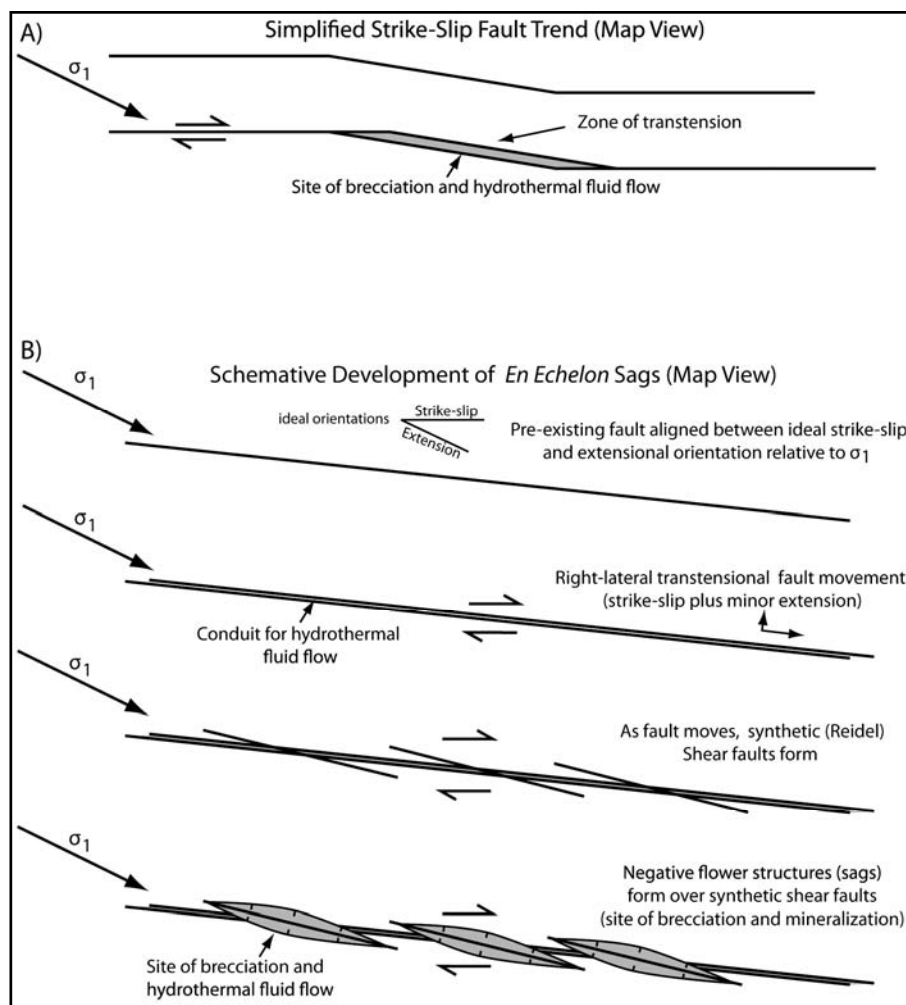


Figure 20. Transtensional faulting scenarios. A) Right lateral strike-slip fault has transtensional segment at dilational jog. Brecciation and hydrothermal fluid flow most likely at dilational jog. B) Possible evolution of en echelon sags such as those found in Rochester Field (Figure 7) and other Trenton Black River laterally discontinuous dolomite fields. Pre-existing fault is oriented between ideal strike-slip and extensional orientations to a compressive stress. This orientation leads to transtensional movement and the development of synthetic Riedel shear faults that trend even closer to the ideal extensional orientation. Because of this orientation, transtensional negative flower structures form over each shear fault making sags visible on seismic and creating space where brecciation and hydrothermal fluid flow occur.

dissolution or by volume reduction during dolomitization. The occurrence of vuggy porosity in most or all Trenton-Black River hydrothermal dolomite fields shows that there has been significant dissolution of limestone, dolomite or both during hydrothermal diagenesis. If the dissolved carbonate was carried away by fluids exiting the altered zone, or was precipitated in tight matrix dolomite and limestone outside of the sag, it is possible that some volume loss may have occurred due to dissolution. The presence of significant vuggy porosity shows that dissolution has occurred but its impact on sag development is difficult to assess.

Mole-for mole replacement of limestone with dolomite results in a volume loss of 12-13% because dolomite molecules are smaller than calcite molecules (Weyl, 1960). This would have a minor impact on sag development, if any. In many wells, only 10 meters (33 feet) of section are dolomitized, but they still occur in sags visible on seismic. If 10 meters of dolomite were left after dolomitization of 11.3 meters of limestone, this minor volume loss (1.3 meters) would not be detectable on a seismic line.

There is no evidence that the Black River has collapsed into underlying caves in the Beekmantown as was suggested by Loucks (2003a, 2003b). This model is not supported by the seismic (Figures 7 and 8) and can be discarded based on field relations in Ontario where the Black River Group carbonates directly overlie Cambrian siliciclastics which sit directly on the basement and the Beekmantown and equivalents are absent. If there is no Beekmantown present, there can be no caves in the Beekmantown in which the overlying Black River might collapse. The sags, breccias and mineralization in the Trenton and Black River in Ontario (such as Rochester Field – Figure 7) are virtually identical to those in New York, Michigan and Ohio (Colquhoun, 1991; Coniglio et al., 1996; Bonnar, 2001). Furthermore, there is only minor offset on the top of the Beekmantown underlying many of the sags in the Black River play in New York (see Figures 7 and 8).

Fault style

The *en echelon* sags in Figure 6 are interpreted to have formed as a result of transtensional faulting or oblique divergent slip (Smith and Nyahay, 2005) as initially proposed by Harding (1974) for similar *en echelon* sags at Albion-Scipio Field. The main component of fault movement is thought to be strike-slip, but with a very important extensional component (Figure 20). The individual sags are thought to be negative flower structures that formed over synthetic shear

faults that tie back to a master transtensional fault at depth (Figure 20). The extensional component would tend to open the synthetic fractures and make them conduits for upward flowing hydrothermal fluids (Harding, 1974). This might occur at a dilational jog on a longer strike-slip fault (Figure 20A) or on a pre-existing fault that is reactivated in a transtensional sense rather than pure strike-slip because it is slightly misaligned toward the extensional orientation with respect to the principle compressive stress (Figure 20B).

It appears that there are many variations on this theme in the Trenton Black River Play. There may be both right-lateral and left-lateral fault movement (Smith et al., 2003) on different faults and in different parts of the basin associated with the same tectonic event. Fault intersections are common locations for sag development and hydrothermal alteration because they commonly set up quadrants of transtension and transpression. Again, zones of transtension are the most likely to be affected by hydrothermal fluid flow.

Hydrothermal dolomitization appears to primarily occur around basement-rooted faults and is rare or absent around faults that die out within the sedimentary column. Strike-slip faults may extend far down into the basement and long strike-slip faults may even extend to the brittle-ductile transition. If fluids flow deep down into the basement before they flow back up to the zone of alteration this could produce some of the high temperatures found in the fluid inclusion data. The basement might also be a source of some elements such as magnesium or metals found in genetically related Mississippi Valley Type ore deposits (Davies and Smith, 2006).

Fault-related hydrothermal alteration model

This model for the hydrothermal alteration of the Black River Group carbonates in New York is speculative, but is supported by all of the known facts at this time (Figure 21). Black River Group carbonates were deposited on a relatively stable craton (Figure 21A). Major collision between North America and Volcanic Island Arc begins during earliest Trenton time and continues through the Late Ordovician and into the Silurian (Ettensohn and Brett, 2002). This collision led to reactivation of appropriately oriented older faults or activation of new faults. Near the thrust front, thrust loading led to normal faulting oriented subparallel to the orogenic belt (Bradley and Kidd, 1991). In the more distal parts of the craton, strike-slip faulting is initiated along faults oriented at an angle of between roughly 10° and 45° to the principal compressive stress. Faults or segments of faults that are

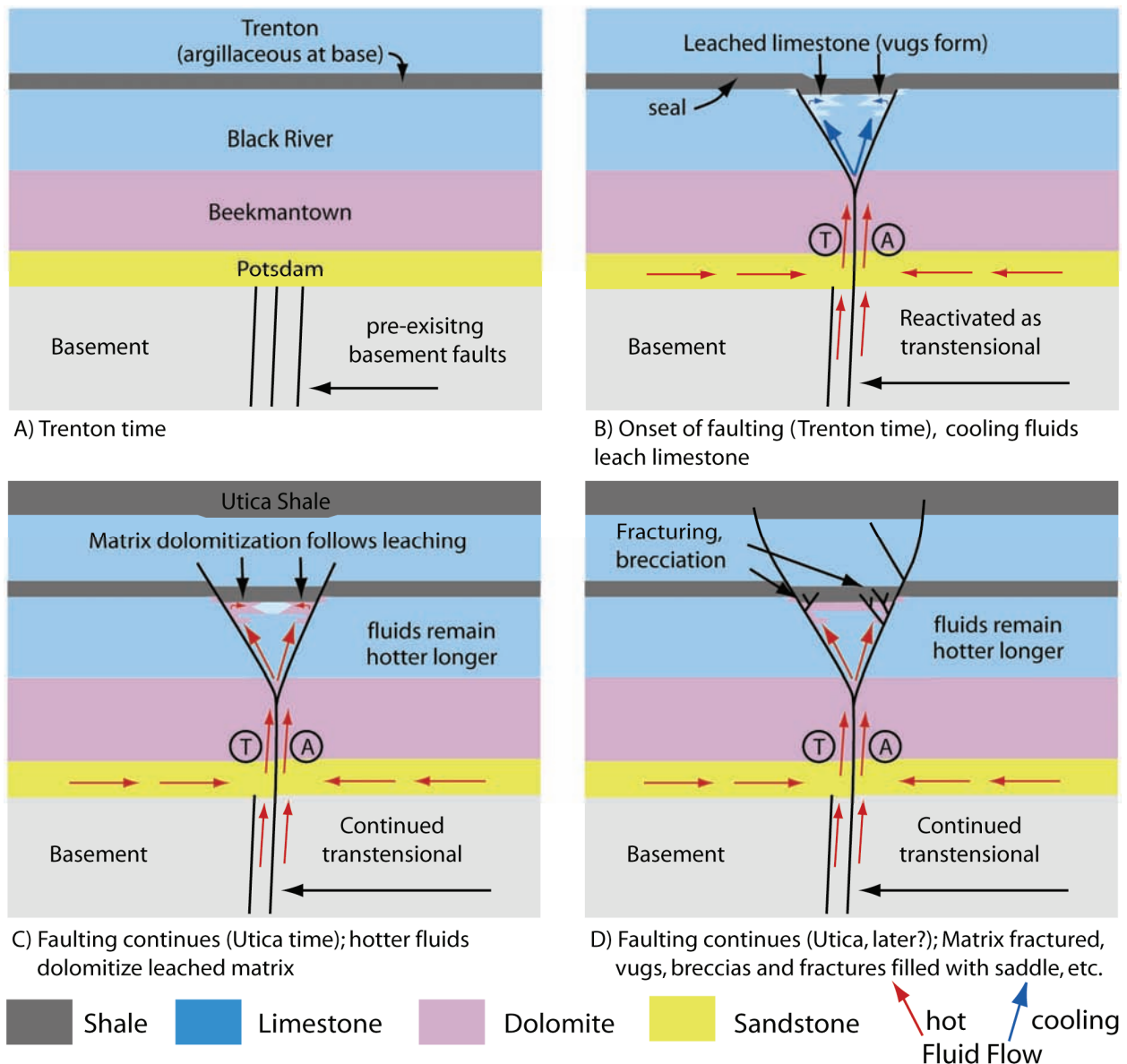


Figure 21. Schematic fault-related hydrothermal alteration model for Black River Group dolomite reservoirs, New York. This view would be one of the negative flower structures shown in map view in Figure 20B.

oriented closer to parallel to the principal compressive stress (closer to 10°) will have a greater extensional component leading to more intense alteration.

High-pressure, high-temperature fluids flowed up active basement-rooted strike-slip and transtensional faults (particularly in dilational parts of fault zones) during the time of Trenton and Utica deposition, hit low permeability beds at the base of the Trenton and flowed out laterally into the more permeable limestones of the Black River. Each time the faults moved there may have been another pulse of high-rate fluid flow. Cooling, possibly acidic hydrothermal fluids

leached the limestone and produced vugs in a migrating front moving away from the fault zone (Figure 21B). As permeability was enhanced by fracturing and leaching, hotter fluids were able to migrate farther from fault zone (Figure 21C). These fluids may have directly precipitated dolomite or may have mixed with *in situ* modified seawater to precipitate dolomite. This dolomitization first produced a halo of matrix dolomite, particularly on the downthrown sides of faults in negative flower structures. Because the fluids flowed up from greater depths where pressures are higher, the elevated pressure of the fluids may have led to hydrofracturing (*sensu* Phillips, 1972), enlargement of

existing fractures and further brecciation. Some dissolution vugs may have formed prior to and during matrix dolomitization. Matrix dolomitization was followed by further fracturing, brecciation and vug development as tectonic activity continued (Figure 21D). Fractures and vugs were lined or filled with saddle dolomite soon after their formation.

As time passed, fluids evolved and precipitated a range of other minerals including quartz, bitumen, sulfides and calcite. Bitumen may have formed when kerogen within the altered formation and near the faults was heated by the hydrothermal fluids and small quantities of oil formed that coated some pores and fractures (“forced maturation” of Davies, 2001). If the faulting was over by Late Ordovician or Early Silurian time (as it appears to be on many seismic lines), that would make most of the diagenesis Late Ordovician to Early Silurian in age. If the faulting continued or faults were reactivated during the Devonian Acadian or Pennsylvanian Alleghenian Orogenies, some of the later stages of mineralization may have occurred during those times. Some calcite cementation may have occurred during later pressure solution of the adjacent limestones under normal burial conditions.

Hypogene karst

Variations on the fault-related hydrothermal alteration model presented here could be the origin of many mineralized breccias and other diagenetic features from around the world. Many carbonate-hosted Mississippi Valley Type ore deposits are likely to have formed due to transtensional faulting and upward hydrothermal fluid flow (Davies and Smith, 2006). In some carbonate reservoirs, the main impact of fault-related fluid flow is dissolution of limestone and the formation of microporosity. Dissolution might occur due to cooling of hydrothermal fluids, the introduction of acidic fluids sourced from deeper in the sedimentary column or the basement or from mixing of fault-sourced hydrothermal fluids with in situ fluids (Dublyansky, 2000)

Calcite has retrograde solubility and cooling hydrothermal fluids should become progressively undersaturated. If they are introduced to the formation at 100°C and cool slowly to 40° or 50°C as they move out away from the fault the fluids should continue to dissolve carbonate until they equilibrate with ambient burial temperature. CO₂ also has retrograde solubility so cooling fluids can hold progressively more CO₂. If CO₂ gas is present in the hydrothermal fluid, it could go into solution as the fluid cools forming carbonic acid, causing the pH to decrease and increasing the capacity for dissolution of carbonate.

The pH of formation waters generally decreases with depth in many sedimentary basins (see Hanor, 1994). Fluids flowing up faults in basins where this is the case should be more acidic than in situ fluids and could dissolve carbonate. The presence of H₂S in upward flowing fluids could lead to significant dissolution if it is oxidized to form sulfuric acid (Palmer, 1991; Hill, 1995; Dublyansky, 2000). Mixing of hydrothermal fluids with in situ modified seawater could also lead to an undersaturated fluid capable of dissolution of limestone (Salas et al., 2007).

The dissolution caused by cooling, lower pH fluids and fluid mixing could create vuggy porosity and breccia development that could be called hypogene karst. Brecciation and the development of vuggy porosity is likely to be more common where the host rock has relatively low permeability and diagenesis and fracturing are intensified around the fault zone. When higher permeability strata are altered, fluids might flow farther from the fault and enhance matrix porosity but cause less intense brecciation and less vuggy porosity near the faults.

The same fluid might precipitate high-temperature minerals such as dolomite, sulfides, sulfates, quartz and more near the fault as the pressure of the fluid drops and then starts dissolving the carbonate host rock as it moves away from the fault and cools. Any breccias where the matrix is primarily composed of high-temperature minerals or that are localized around faults should be considered candidates for fault-related hydrothermal or hypogene diagenesis. Many breccias previously interpreted to be of a meteoric karst origin may in fact have formed entirely in the subsurface or been significantly overprinted by fault-related hydrothermal processes.

Exploration model

Hydrothermal dolomite (and associated leached limestone) reservoirs represent one of the most significant remaining resources in North America and other mature regions of the world. These reservoirs are likely to have been bypassed in many cases because of their common occurrence in structural lows, which are unlikely to have been drilled during earlier exploration phases. The exploration model for Trenton and Black River hydrothermal dolomite reservoirs is to look for subtle basement-rooted wrench faults and negative flower structures that cut the regional limestone with evidence for movement in the first kilometer of burial. Faults with relatively minor offset that do not extend far above the target formation are typically the best candidates because the faults have not breached the seal for the hydrothermal fluids or the hydrocarbons.

Most documented hydrothermal dolomite reservoirs occur in regional limestones, but that does not mean that hydrothermal alteration is restricted to this scenario. The effects of the hydrothermal alteration may be less obvious where entire formations are dolomitized regionally, but saddle dolomite cemented breccias, fractures and vugs and sulfide ore deposits are very common in regional dolomites and in most cases are probably fault-related hydrothermal in origin (Smith, 2004a). Hydrothermal leached limestone reservoirs may be as common or more common than hydrothermal dolomite. Although leached limestone is not a common component of the Trenton-Black River Play, it does occur in many other settings and similar exploration strategies may lead to success in leached limestone plays as well (Davies and Smith, 2006; Wierzbicki et al., 2006).

Conclusions

1. Newly discovered natural gas reservoirs in the Black River Group carbonates of New York occur in laterally discontinuous, linear, fractured, vuggy dolomite bodies that occur in structural lows associated with wrench faults.
2. Both matrix and saddle dolomite formed from hot (110-170°C), saline (14.5 wt.%), Fe- and Mn-rich brines with radiogenic strontium isotope values. All of these support a subsurface origin for the dolomite.
3. In Ohio and Michigan, fluid inclusions from similar fault-related dolomitized bodies in the Trenton and Black River have homogenization temperatures that exceed the maximum ambient burial temperature by 40-120°C. That makes the dolomite there unequivocally hydrothermal in origin. In New York, the Black River was buried to a temperature equal to or greater than the homogenization temperatures in the dolomites so the origin of the dolomite is equivocal in that sense. However, the New York dolomites are similar in their link to wrench faults, appearance and geochemical attributes to those in Ohio and Michigan and are here interpreted to have formed in the same way and at roughly the same time.
4. The dolomite, vugs, breccias and fractures are interpreted to have formed when high-pressure high-temperature fluids flowed up dilational portions of strike-slip faults when the faults were active, hit a seal at the base of the Trenton and flowed laterally, altering the formation. Vugs formed when limestone was leached by a front of cooling hydrothermal fluids, and dolomitization and other mineralization followed soon afterward. Alteration is thought to have occurred during the Late Ordovician Taconic Orogeny

when the formation was buried to a depth of less than 500 meters.

5. Because these reservoirs occur in structural lows, they have been bypassed for many years. More of these reservoirs (and associated leached limestone reservoirs) can and will be found in the Trenton and Black River groups and in carbonates around the world using the appropriate integrated structural-stratigraphic diagenetic model.
6. A fault-related hydrothermal or hypogene origin should be considered for any breccia that is cemented with minerals precipitated from high-temperature fluids.

References

- Allan, J.R., and W.D. Wiggins. 1993. *Dolomite reservoirs: Geochemical techniques for evaluating origin and distribution: AAPG continuing education course notes no. 36*. Tulsa: American Association of Petroleum Geologists.
- Bonnar, B. 2001. Talisman Energy Inc. and the Ontario oil patch: Does the shoe fit? Ontario Petroleum Institute (OPI) 2001 Annual Meeting Program.
- Bradley, D.C., and W.S.F. Kidd. 1991. Flexural Extension of the Upper Continental Crust in Collisional Foredeeps. *Geological Society of America Bulletin* 103: 1416-1438.
- Brett C.E., and G.C. Baird. 2002. Revised stratigraphy of the Trenton Group in the type area, central New York State: Sedimentology and tectonics of a middle Ordovician shelf-to-basin succession. In *Taconic convergence: Orogen, foreland basin and craton*, ed. C.E. Mitchell and R. Jacobi. *Physics and Chemistry of the Earth* 27: 231-263.
- Burke K., and J.F. Dewey. 1973. Plume-generated triple junctions: key indicators of applying plate tectonics to old rocks. *Journal of Geology* 81: 406-433.
- Burke, W.H., R.E. Denison, E.A. Hetherington, R.B. Koepnick, H.F. Nelson, and J.B. Otto. 1982. Variation of seawater $^{87}\text{Sr}/^{86}\text{Sr}$ throughout Phanerozoic time. *Geology* 10: 516-519.
- Colquhoun, I. M. 1991. Paragenetic history of the Ordovician Trenton Group carbonates, South-western Ontario. M.S. thesis, Brock University.
- Coniglio, M., R. Sherlock, A.E. Williams-Jones, K. Middleton, and S.K. Frape. 1994. Burial and hydrothermal diagenesis of Ordovician carbonates from the Michigan Basin, Ontario, Canada. In *Dolomites, a volume in honor of Dolomieu*: International Association of Sedimentologists Special Publication 21, ed. B. Purser, M. Tucker and D. Zenger, 231-254. Oxford: Blackwell.

- Cornell, S.R., and C.E. Brett. 2000. K-Bentonite and sequence correlations of Upper Black River and lower Trenton Limestones from Lake Simcoe, southern Ontario, Canada to Watertown, northern New York State. *American Association of Petroleum Geologists Bulletin* 84 (9): 1381.
- Davies, G.R. 2001. *Hydrothermal (thermobaric) dolomite reservoir facies: Global and western Canadian perspectives*, unpublished report. Calgary: Graham Davies Geological Consultants Ltd.
- Davies, G.R. 2004. Hydrothermal (thermobaric) dolomite and leached limestone reservoirs: General principles, genetic connections and economic significance in Canada. American Association of Petroleum Geologists, Annual Convention Abstracts 14: A32.
- Davies, G.R., and L.B. Smith Jr. 2006. Structurally controlled hydrothermal dolomite reservoir facies: An overview. *American Association of Petroleum Geologists Bulletin* 90 (11): 1641-1690.
- Deming, D. 1992. Catastrophic release of heat and fluid flow in the continental crust. *Geology* 20: 83-86.
- Dooley, T., and K. McClay. 1997. Analog modeling of pull-apart basins. *American Association of Petroleum Geologists Bulletin* 81: 1804-1826.
- Dubljansky, Y.V. 2000. Dissolution of carbonates by geothermal waters. In *Speleogenesis: Evolution of karst aquifers*, ed. A. Klimchouk, D.C. Ford, A.N. Palmer and W. Dreybrodt, 158-159. Huntsville: National Speleological Society.
- Ettensohn, F.R., M.A. Kulp, and N. Rast. 2002. Interpreting ancient marine seismites and apparent epicentral areas for paleo-earthquakes, middle Ordovician Lexington Limestone, central Kentucky. In *Ancient seismites: Geological Society of America Special Paper 359*, ed. F.R. Ettensohn, N. Rast and C.E. Brett, 177-190. Denver: Geological Society of America.
- Ettensohn, F.R., and C.E. Brett. 2002. Stratigraphic evidence for continuation of Taconian Orogeny into Early Silurian time. In *Taconic convergence: Orogen, foreland basin, and craton*, ed. C.E. Mitchell and R. Jacobi. *Physics and Chemistry of the Earth* 27: 279-288.
- Friedman, I., and J.R. O'Neil. 1977. *Compilation of stable isotope fractionation factors of geochemical interest: U.S. Geological Survey Professional Paper 440-KK*. Boulder: U.S. Geological Survey.
- Gregg, J.M., and D.F. Sibley. 1984. Epigenetic dolomitization and the origin of xenotopic dolomite texture. *Journal of Sedimentary Petrology* 54: 908-931.
- Hanor, J. S. 1994. Origin of saline fluids in sedimentary basins. In *Geofluids: Origin and migration of fluids in sedimentary basins: Geological Society of London Special Publication No. 78*, ed. J. Parnell, 151-174. London: Geological Society of London.
- Harding, T.P. 1974. Petroleum traps associated with wrench faults. *American Association of Petroleum Geologists Bulletin* 58: 1290-1304.
- Harris, A.G. 1979. Conodont color alteration, an organo-mineral metamorphic index, and its application to Appalachian Basin geology. In *Aspects of diagenesis: Society of Economic Paleontologists and Mineralogists Special Publication No. 26*, ed. P. A. Scholle and P.R. Schluger, 3-16. Tulsa: Society of Economic Paleontologists and Mineralogists.
- Hill, C.A. 1995. H₂S-related porosity and sulfuric acid oil-field karst. In *Unconformities and porosity in carbonate strata, AAPG Memoir 63*, ed. D.A. Budd, A.H. Saller and P.M. Harris, 301-306. Tulsa: American Association of Petroleum Geologists.
- Horvath, E., L. Davies, and M. Molgat. 2004. Example of a Trenton-Black River reservoir: a case study of the Rochester Pools, Ontario. Ontario Petroleum Institute Annual Meeting Program with abstracts.
- Hulver, M.L. 1997. Post-orogenic evolution of the Appalachian Basin mountain system and its foreland. Ph.D. diss., University of Chicago.
- Hurley, N.F., and R. Budros. 1990. Albion-Scipio and Stoney Point Fields-USA Michigan Basin. In *Stratigraphic traps I, treatise of petroleum geology, atlas of oil and gas fields*, ed. E.A. Beaumont and N.H. Foster, 1-37. Tulsa: American Association of Petroleum Geologists.
- Jacobi, R., P. Agle, M. Evans, S. Coulter, S. Lowenstein, T. Mroz, and J. Martin. 2008. Taconic tectonics and the Trenton/Black River carbonate gas play in the northern Appalachians of New York state: Evidence from seismic data and the first oriented core in the Appalachian Basin T/BR. Canadian Society of Petroleum Geologists Annual Meeting Extended Abstracts: 23-25.
- Knipe, R.J. 1993. The influence of fault zone processes and diagenesis on fluid flow. In *Diagenesis and basin development, AAPG Studies in Geology #26*, ed. A. D. Horbury and A.G. Robinson, 131-156. Tulsa: American Association of Petroleum Geologists.
- Kolata, D.R., W.D. Huff, and S.M. Bergstrom. 1996. Ordovician K-bentonites of eastern North America, Geological Society of America Special Paper 313. Denver: Geological Society of America.

- ica.
- Landing, E., S.R. Westrop, and L.A. Knox. 1996. Conodonts, stratigraphy and relative sea-level changes of the Tribes Hill Formation (Lower Ordovician, east-central New York). *Journal of Paleontology* 70: 656-680.
- Loucks, R.G. 1999. Paleocave carbonate reservoirs: Origins, burial-depth modifications, spatial complexity, and reservoir implications. *American Association of Petroleum Geologists Bulletin* 83: 1795-1834.
- Loucks, R.G. 2003a. Origin of Lower Ordovician Ellenburger Group Brecciated and Fractured Reservoirs in West Texas: paleocave, thermobaric, tectonic or all of the above? AAPG National Meeting Program with Abstracts: 107.
- Loucks, R.G.. 2003b. Understanding the development of breccias and fractures in Ordovician carbonate reservoirs. In *The Permian Basin: Back to basics: West Texas Geological Society Fall Symposium: West Texas Geological Society Publication #03-112*, ed. T.J. Hunt and P.H. Lufholm, 231-252. Midland: West Texas Geological Society.
- Lucia, F.J. 1995. Rock-fabric/physical classification of carbonate pore space for reservoir characterization. *American Association of Petroleum Geologists Bulletin* 79: 1275-1300.
- Machel, H.G., and J. Lonee. 2002. Hydrothermal dolomite- a product of poor definition and imagination. *Sedimentary Geology* 152: 163-171.
- Matsumoto, M.R., J.F. Atkinson, M.D. Bunn, and D.S. Hodge. 1992. *Disposal/recovery options for brine waters from oil and gas production in New York: New York State Energy Research and Development Authority Report No. 1591-ERER-RIER-91*. New York. New York State Energy Research and Development Authority.
- McLaughlin, P.I., and C.E. Brett., 2002. Upper Ordovician seismites of Kentucky: a sedimentological and sequence stratigraphic approach. *Geological Society of America Abstracts with Programs* 36: 102.
- Montañez, I.P. 1994. Late diagenetic dolomitization of lower Ordovician upper Knox carbonates: A record of the hydrodynamic evolution of the southern Appalachian Basin. *American Association of Petroleum Geologists Bulletin* 78 (8): 1210-1239.
- Moore, J.M. 1986. Introduction: the 'Grenville problem' then and now. In *The Grenville Province: Geological Association of Canada Special Paper 31*, ed. J.M. Moore, A. Davidson and A.J. Baer, 1-12. St. John's: Geological Association of Canada.
- Muir-Wood, R., and G.C.P. King. 1993 Hydrological signatures of earthquake strain. *Journal of Geophysical Research* 98: 22035-22068.
- Phillips, W.J. 1972. Hydraulic fracturing and mineralization. *Journal of the Geological Society of London* 12228: 337-359.
- Palmer, A.N. 1991. Origin and morphology of limestone caves. *Geological Society of America Bulletin* 103: 1-21.
- Pope, M.C., J.F. Read, R.K. Bambach, and H.J. Hofmann. 1997. Late middle to late Ordovician seismites of Kentucky, southwest Ohio and Virginia: Sedimentary records of earthquakes in the Appalachian Basin. *Geological Society of America Bulletin* 109: 489-503.
- Prouty, C.E. 1988. Trenton exploration and wrenching tectonics in Michigan Basin and environs. In *The Trenton Group (upper Ordovician Series) of eastern North America: American Association of Petroleum Geologists Studies in Geology* 29, ed. B. D. Keith, 207-236. Tulsa: American Association of Petroleum Geologists.
- Repetski, J.E., L.F. Ruppert, R.T. Ryder, D.J. Weary, E.L. Rowan, and M.H. Trippi. 2004. Thermal maturity patterns for Ordovician, Devonian, and Pennsylvanian rocks of the central Appalachian Basin, based on CAI and vitrinite reflectance (% Ro). American Association of Petroleum Geologists Eastern Section Meeting Program with abstracts (unpaginated).
- Rickard, L.V. 1973. *Stratigraphy and structure of the subsurface Cambrian and Ordovician carbonates of New York*. New York State Museum map & chart No. 18. New York: New York State Museum.
- Rowan, E.L., R.T. Ryder, C.S. Swezey, J.E. Repetski, R.D. Crangle, M.H. Trippi, and L.F. Ruppert. 2004. Burial and thermal history model of a cross-section through the central Appalachian Basin, Ohio and West Virginia. American Association of Petroleum Geologists Annual Meeting Expanded Abstracts: 121.
- Rimstidt, J.D. 1997. Gangue mineral transport and deposition. In *Geochemistry of hydrothermal ore deposits*, ed. H. L. Barnes, 487-516. New York: John Wiley and Sons.
- Sibson, R.H. 1990. Faulting and fluid flow. In *Fluids in tectonically active regimes of the continental crust: Mineralogical Association of Canada, Short Course Handbook, v. 18*, ed. B. E. Nesbitt, 93-132. Quebec: Mineralogical Association of Canada.
- Sibson, R.H. 2000. Fluid movement in normal faulting. *Journal of Geodynamics* 29: 469-499.
- Smith, Jr., L.B. 2006. Origin and reservoir characteristics of upper Ordovician Trenton-Black River

- hydrothermal dolomite reservoirs in New York. *American Association of Petroleum Geologists Bulletin* 90 (11): 1691-1718.
- Smith, L.B., and R.E. Nyahay. 2005. Wrench fault architecture of Trenton Black River hydrothermal dolomite reservoirs. American Association of Petroleum Geologists Eastern Section Meeting Program with Abstracts (unpaginated).
- Smith, L.B. 2004. Hydrothermal alteration of carbonate reservoirs: How common is it? American Association of Petroleum Geologists National Meeting Program with Abstracts: A130.
- Smith, L.B., R.E. Nyahay, and B. Slater. 2004. Outcrop analogs for hydrothermal dolomite reservoirs, Mohawk Valley, New York. American Association of Petroleum Geologists Eastern Section Meeting Program with Abstracts (unpaginated).
- Smith, L.B., C.M. Lugert, and R.E. Nyahay. 2003. Integrated characterization of hydrothermal dolomite reservoirs in Trenton-Black River carbonates of New York: American Association of Petroleum Geologists National Meeting Program with Abstracts: 160.
- Smith L.B., and R.E. Nyahay. 2003. Trenton-Black River hydrothermal dolomite reservoirs: An integrated structural stratigraphic diagenetic play. American Association of Petroleum Geologists Eastern Section Meeting Program with Abstracts (unpaginated).
- Taylor, T.R., and D.F. Sibley. 1986. Petrographic and geochemical characteristics of dolomite types and the origin of ferroan dolomite in the Trenton Formation, Ordovician, Michigan Basin. *Sedimentology* 33: 61-86.
- Thomas, W.A. 1991. The Appalachian-Ouachita rifted margin of southeastern North America. *Geological Society of America Bulletin* 103: 415-431.
- van Staal, C.R. 2005. The Northern Appalachians. In: *Encyclopedia of geology, Vol. 4*, ed. R.C. Selley, L. Robin, M. Cocks and I. R. Plimer. Oxford: Elsevier.
- Weyl, P.K. 1960. Porosity through dolomitization: Conservation of mass requirements. *Journal of Sedimentary Petrology* 30: 85-90.
- Weary, D.J., R.T. Ryder, and R.E. Nyahay. 2001. Thermal maturity patterns in New York State using CAI and %R_O. *Northeastern Geology and Environmental Science* 23: 356-376.
- White, D.E. 1957. Thermal waters of volcanic origin. *Geological Society of America Bulletin* 68: 1637-1658.
- Wickstrom, L.H., J.D. Gray, and R.D. Stieglitz. 1992. *Stratigraphy, structure, and production history of the Trenton Limestone (Ordovician) and adjacent strata in northwestern Ohio*. Ohio Geological Survey Report of Investigations 143. Columbus: Ohio Geological Survey.
- Wierzbicki, R., J.J. Dravis, I. Al Aasm, and N. Harland. 2006. Dolomitization and porosity evolution of upper Jurassic Abenaki platform carbonates, deep Panuke reservoir, Nova Scotia: Hydrothermal or deep burial diagenetic processes or both? *American Association of Petroleum Geologists Bulletin* 90 (11): 1843-1861.

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