Salt Dissolution (2 of 5): Caves in Evaporites

Properties of evaporite karst

At first sight, medium-scale evaporite karst surface landforms, such as dolines, polje-like depressions, subsidence bowls and collapse dolines, appear near-identical to those found in and on carbonate karst (Part 1). Likewise, all the smaller-scale intracavern features, both dissolutional (pipes and cavities) and constructive calcite speleothems), can be preserved as karst zones in and above an evaporite host, even when the formative hydrology becomes inactive and the system is buried (paleokarst). But, the much higher rates of dissolution of halite and gypsum compared to limestone or dolomite means there are significant differences in rates of formation and cave geometries compared to carbonate caves in non-evaporitic hosts. One cannot simply take models of caves developed by studies of carbonate karst and use them to unreservedly interpret evaporite caves.

Differences reflect the inherently higher solubility and flowability of the host salts at earth surface conditions and mean most evaporite-hosted caves show peculiarities related to higher dissolution rates compared to carbonate karst. For example, one most significant in a vadose halite cave is that evaporation of a near-saturated brine is aided by airflow and means many halite stalactites tend to curve into the direction of air flow (anemolites) rather than the subvertical dripstone mechanism that controls the shape of most calcite stalactites and stalagmites. Forti (2017) is an excellent summary of evaporite speleothems and formative mechanisms in both gypsum and halite caves.

The high solubility of subsurface evaporite beds atop deeply-circulating pressurised and jointed aquifers (confined hydrologies) means deep-phreatic chimneying (vertical shafts) and deep dissolution with upward-stopping are a common deep cavity-forming mechanism in and atop buried salt beds. This, in turn, means high volumes of overburden sediment can be swallowed quickly, especially in suprasalt regions with strong roof spans covered by loose sediment (Figure 1). In such settings, large at-surface natural sinkholes and circular structures (breccia chimneys) can daylight in days or even hours after roof span collapse and so constitute significant geohazards (see Part 3). This type of rapid point-sourced vertical stoping atop evaporites doesn't just happen in continental settings. Similar circular structures atop breccia-filled vertical shafts or fault-bound chimneys form beneath deep waters of the Eastern Mediterranean via substratal breaching of the Messinian evaporite, followed by upward stoping through the overlying sediments (Bertoni and Cartwright, 2005).

Worldwide, much of the dissolution action in uplifting salt beds and masses begins well below the watertable. It takes
place hundreds of metres or more below the surface at bathyphreatic depths where; 1) halite is leaching, 2) anhydrite is reconvert ing to gypsum and 3) where caprocks and other dissolution residues are forming atop diapiric salt features. Early dissolution is greatest at contacts between the salt edges and joints or fractures in adjacent aquifers. Thus, patterns of jointing or fracturing control the extent and style of caves in a salt host and typically create high-density maze caves. With halite, the only setting where a NaCl mass makes it to the surface is as an active or recently active diapir crest or namakier. Before its dissolution, the salt matrix hosting a cave-hosting diapiric halite is largely impervious, flowing and re-annealing. Hence, most of the karst action accessible for study in halite caves is vadose and tied to perched water tables; this is most obvious near the margins of namakier salt sheets and located well away from locations of active salt fountains.

If phreatic caves ever do form near to rising salt fountain domes, these early cavities are quickly closed by the pressurised salt flow needed to bring salt to surface. Rather than being telogenetic features, many of the metre-scale blebs of finely-layered gravitationally-aligned laminar halite seen within diapirs intersected in deep salt mines, and floating in a matrix of flow foliated coarse-crystalline halite, are the result of mesogenetic salt precipitating in gas-filled open cavities within the diapir mass (see Salty Matters October 31, 2016). These laminar cavity filling growth-aligned halite textures are not telogenetic, neither are the entrained salt clasts preserving primary (relict) depositional texture of the mother salt. Instead, they indicate the existence of former or present-day gas-filled cavities (N₂, CH₄ or CO₂), which are a hazard occasionally encountered during salt and potash mining, for example in the diapirs of NE Germany (Hedlund, 2012). Their presence and their ability to blow out mine walls show that pressurised gas pockets and associated cavern fill textures are not unusual mesogenetic features in a flowing salt mass. Away from deep mine intersections of mesogenetic gas-filled cavities, most of the shallow cavities and fill textures of a flowing salt body and studied by speleologists are telogenetic and mostly vadose.

Cave-forming processes in a telogenetic salt cave in bedded salt bodies are either vadose or phreatic (both shallow and deep), with many gypsum/anhydrite caves showing textural evidence indicating transit from one realm to the other. At any one time, the transition from phreatic to vadose landforms is tied to depth to watertable. This is seen in the gypsumiferous cap karst to diapirs and allochthons of Triassic salt in the Betic Cordillera of Spain (Calaforra and Pulido–Bosch, 1999). The crests of the salt structures are dominated by collapse dolines (vadose), this passes radially out into the belt of solution dolines occupied by seasonal saline lagoons and further out into rim of saline springs that form wherever the watertable intersects the landsurface (phreatic). Some caves pass from phreatic to vadose a number of times in their histories in response to watertable fluctuations tied to varying climate and tectonics (Columbu et al., 2015).

Vadose processes characterise the uppermost part of a karst aquifer and have air in pores above a water surface or perched watertable. Water drainage is free-flowing under gravity; cave passages drain downslope and show strong

<table>
<thead>
<tr>
<th>Cave location and host lithology</th>
<th>Length (km)</th>
<th>Area of cave (m²x10⁶)</th>
<th>Volume of cave (m³x10⁶)</th>
<th>Area of cave field (km²)</th>
<th>Volume of rock (m³x10⁶)</th>
<th>Specific volume (m³/m)</th>
<th>Passage density (km/km²)</th>
<th>Cave porosity (%)</th>
<th>Areal coverage (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Optymistychna Cave, W. Ukraine, Neogene gypsum (Figure 4a)</td>
<td>214</td>
<td>0.260</td>
<td>0.52</td>
<td>1.48</td>
<td>26.03</td>
<td>2.8</td>
<td>127.03</td>
<td>2.00</td>
<td>17.57</td>
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<tr>
<td>Ozernaja Cave, W. Ukraine, Neogene gypsum (Figure 4b)</td>
<td>111</td>
<td>0.330</td>
<td>0.665</td>
<td>0.74</td>
<td>13.20</td>
<td>6.0</td>
<td>150.00</td>
<td>5.04</td>
<td>44.59</td>
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<tr>
<td>Zoloushka Cave, W. Ukraine, Neogene gypsum</td>
<td>89.5</td>
<td>0.305</td>
<td>0.712</td>
<td>0.63</td>
<td>18.93</td>
<td>8.0</td>
<td>142.06</td>
<td>3.76</td>
<td>48.41</td>
</tr>
<tr>
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<td>24</td>
<td>0.047</td>
<td>0.080</td>
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<td>3.8</td>
<td>141.18</td>
<td>3.36</td>
<td>27.65</td>
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<td>Kristalna Cave, W. Ukraine, Neogene gypsum</td>
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<td>0.038</td>
<td>0.110</td>
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<td>0.008</td>
<td>0.064</td>
<td>0.06</td>
<td>0.71</td>
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<td>59.32</td>
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<td>1.386</td>
<td>8</td>
<td>36.78</td>
<td>3310.2</td>
<td>14.5</td>
<td>14.95</td>
<td>0.24</td>
<td>3.77</td>
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<tr>
<td>Lechuguilla Cave, NM, USA, Permian limestone</td>
<td>193.4</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Carlsbad Cavern, NM, USA, Permian limestone</td>
<td>49.6</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
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</tr>
</tbody>
</table>

Table 1 Representative morphometry of gypsum maze caves that formed as confined aquifer phreatic systems, compared to Mammoth Cave (unconfined aquifer system) and Carlsbad and Lechuguilla caves (now both unconfined aquifer systems but first formed as sulphuric acid caves with gypsum precipitates; after Klimchouk, 2003, 2007; Palmer and Palmer, 2000; and listed references).
gravitational orientations (numerous sub-vertical features). Because they form above the watertable, vadose cave walls are subject to surface seepage, evaporation and drying. Speleothems decorating vadose cave walls indicate varying combinations of gravity and airflow and include; stalactites, stalagmites, cave popcorn, helicites and flowstones.

Caves can form in a salt bed on its way down (eogene-syndepositional to early burial phreatic), on its way up (evaporite telogenesis, generally begins at bathyphreatic depths) and at the bottom of the bed’s burial history (mesogenetic; e.g., gas-driven bathykarst).

**Gypsum caves**

Active caves beneath gypsum karst landforms first formed as deep phreatic maze caves and are characterised by dense passage networks with numerous contemporaneously closed loops, typically in bedded evaporites. Vadose cross sections of the same system can be quite large due to the high solubility and relative homogeneity of the host, especially if formed in a thick capstone. Heimkehle cave in the Zechstein anhydrites in the Harz Mountains of Germany has an overall passage length of more than 2 km, with large rooms up to 22 metres high and 65 metres wide. It was large enough to be used in World War II to house a factory manufacturing parts for JU88 aeroplanes (Knolle et al., 2013).

Other gypsum caves, some more than 150–200 km long, are well documented in Miocene gypsum hosts in west Ukraine (Table 1; Figure 2; Klimchouk, 2000; Klimchouk and Andrejchuk, 2003; Andrejchuk and Klimchouk, 2004; Klimchouk, 2007) and the Gypsum Plain of West Texas and New Mexico (Stafford et al., 2008, 2009, 2017) and much of our current understanding of the gypsum karst process comes from these regions.

Gypsum cave walls in the vadose realm are relatively undecorated compared to carbonate caves. But calcitic speleothems can form in a gypsum cave and are well documented, as are alabastrine flowstones and selenite rinds on cave walls, for example; in northwest Texas (McGregor et al., 1963), in Alabaster Cave and others in the Blaine Gypsum in western Oklahoma (Johnson, 1996; Bozeman and Bozeman 2002), and in quarries intersecting the gypsum karst system in the Kirschberg Evaporite member near Fredericksburg, Texas (Warren et al., 1990).

**Phreatic gypsum caves**

Phreatic caves in bedded evaporites atop artesian aquifers typically grow as upward stoping and branching blind flow loop caverns (phreatic cupolas) and chimneys, which can begin to grow deep below the watertable as bathyphreatic or hypogenic karst. For example, deep artesian systems drive cupola karst, and blind chimney stopes in the Black Hills of Dakota, the Elk Point Basin of Canada and the Optymistychna Cave, western Ukraine (Figures 2, 4a). Deeper in a basin, where bathyphreatic caves are bathed by centripetal mesogenetic crossflows, water flow is slow and phreatic karst is influenced by the escape of H2S- and CO2-rich basal water, not necessarily by the confined meteoric head. Fluid flow at these greater depths is driven by pore water gradients that reflect potentiometric variations in temperature, organic maturation, pressure and salinity of basal water.

Passages in mesogenetic and telogenetic evaporite-hosted caves tend to first develop near fractures and joints in ad-
adjacent aquifers and then expand into maze cave networks. Once growing in the main evaporite mass, some phreatic maze passages can show internal upward-directed switchback gradients independent of jointing in adjacent aquifers. Cave orientations within the evaporite bed and away from the aquifer inflow level are tied to internal inhomogeneities in the host bed such as less soluble or more soluble intrasalt beds and changes in mineral proportions.

Phreatic tubes are the most common passage shape in smooth-walled blind-dissolution pockets and cupolas in bathypreatic gypsum caves. Tube or channelway shapes range in cross section from near-circular (isotopic dissolution of a soluble host) to elliptical tubes to canyon-shaped keyholes (Figures 3, 4a, b; Klimchouk, 1992, 1996). Ornamentation is minimal where undersaturated crossflow drives the rapid dissolutional breakdown of the cave wall and the resulting passages are smooth, with local scalloped dissolution irregularities. Dense intersecting maze cave networks first form in the early stages of exhumation at the contact between a tight but soluble gypsum/anhydrite unit and a less soluble carbonate or siliciclastic aquifer (Figure 2, 3). This less soluble bed is the supply conduit for groundwater crossflows that dissolve the edges of the initially impervious anhydrite/gypsum bed. The greatest rate of water supply to the dissolving gypsum contact is along joints and fractures in the adjacent aquifer bed (especially with limestone aquifers). Accordingly, the meshwork of caves penetrates the gypsum bed and tends to follow joint and fracture patterns of the adjacent aquifer.

More stagnant phreatic portions of telogenetic CaSO₄ cave systems can be saturated and so precipitate isopachous crusts and crystal rinds. Gypsum is the commonplace isopachous precipitate in voids in this setting, while anhydrite tends to dominate in cavities in the deeper mesogenetic realm (Garcia-Guinea et al., 2002). By the time phreatic maze caves are exhumed and enter a vadose (epigene) setting, where they are accessible for speleological study, much of the earlier phreatic ornamentation has already been dissolved by the increasing undersaturated throughflow and cavern interconnection. This is associated with uplift into the more hydrologically-active phreatic realm, which always precedes entry into the vadose realm. (Warren et al., 1990).

Some of the longest and most complex phreatic maze cave systems in the world are found in Miocene gypsum in west Ukraine. Optymistychna Cave, with more than 214 km of surveyed passages, is the world’s longest gypsum cave and the second or third longest cave of any type (Table 1; Figure 2; Klimchouk, 2007). The world’s longest cave, at 550 km, is the carbonate-hosted Mammoth Cave of Kentucky. The West Ukraine region contains the five longest known gypsum caves in the world, accounting for well over half of the total known length of known gypsum caves on Earth. By area and volume, the world’s largest gypsum caves are: Ozernaja (330,000 m² and 665,000 m³; with 122 km of documented passages it is also the world’s 10th longest), Zoloushka (305,000 m² and 712,000 m³), followed by Optimisticheskaja Cave (260,000 m² and 520,000 m³). They are all complex joint-controlled maze caves, formed under confined aquifer conditions that existed from the Pliocene to the Early Pleistocene (that is karstification on the way up – telogenesis or gypsum exhumation, initially focused on the underside of the bed). Their growth

**Figure 3. Maze caves.** A) Formation of upward expanding dissolution cupolas by pressurized buoyant currents, such a system drives the formation of gypsum caves in the western Ukraine. The diagram shows the schematic relationship between lines of natural and forced convection flow on the mature stage of artesian speleogenesis, when conduit connection has already been established through the gypsum, but forced flow is slow due to the major constraint of the upper confining bed. B) Cross section of Kempe’s (1972) model for phreatic tube cavern development via dissolution due to natural density convection. C) Examples of typical cross-sections from the gypsum caves of the western Ukraine showing varieties of notching or faceting effects from differing rates of lateral expansion (dissolution). (after Klimchouk, 1992, 1996; Kempe, 1972).
patterns indicate upward-transverse phreatic groundwater circulation, with ultimate cavern fusion across the gypsum bed. All these west Ukrainian caves were fed by artesian crossflow in sub-gypsum and supra-gypsum aquifers that were sourced in the Carpathian Mountains.

High rates of dissolution in phreatic gypsum caves, relative to rates of water crossflow, are indicated by bevelled, faceted and “keyhole” cross sections in the Ukrainian caves along with a lack of vadose wall ornamentation (Figures 3a 4; Klimchouk, 1996; Pfeiffer and Hahn, 1976). Keyholes indicate density stratification and convectional circulation in cave-forming waters, with shapes sometimes complicated by lithological discontinuities in the gypsum bed (Figure 3b; Kempe, 1972). Convection in caves is most pronounced where sluggish artesian flow and low flow velocities dominate.

This was the case in the Pliocene to early Pleistocene history of the maze caves of the western Ukraine (Klimchouk and Andrejchuk, 2003). At that time the deeply buried gypsum dissolved via upward growing but blind, phreatic cavities, with a reflux of somewhat denser “spent” waters sinking toward the base of the cave. Spent water was replaced by less dense inflow waters supplied from the lower aquifer (Figure 3a). This set up a natural density-stratified convection, which maintained fresher (less dense) waters near the phreatic cave roof. Upward growing blind caves tend to expand more at their tops driving the transition from subcircular to keyhole caverns along the cave conduit (Figure 4b). After dissolving gypsum and increasing in density, a portion of the “spent” cave water sank all the way back into the underlying aquifer where it once again joined the regional throughflow in the lower aquifer. Once a stopping cave breached the top aquifer water flow direction in the cave was controlled by temperature, pressure and density contrasts between aquifers on either side of the gypsum bed. It seems that post-breath most of the cave water continued to rise through the cave system into the overlying aquifer (Figures 3a, 4) Similar keyhole and cupola morphologies are developed in low flow rate bathyphreatic sulphuric acid caves in carbonate hosts (e.g., early stages in the formation of the Carlsbad and Lechuguilla caverns).

Formation of keyhole passages is not an exclusively phre-
atic phenomenon in density-stratified gypsum caves. Keyholes in the vadose portions of many telogenetic carbonate caves indicate the transition of the cave passage from phreatic, with circular cross sections, to vadose with deepening drainage slots at the base of the passages (e.g. Calaforra and Pulido-Bosch, 2003). Phreatic stoping, followed by vadose cavern enlargement, probably explains the close correlation between caprock sinkhole distribution and position of underlying vadose passages in Permian gypsum subcrop in the Kungur Cave region in the Russian Urals, where the caprock thickness is a little as 25 m (Figure 5).

Vadose gypsum caves

A lowering of the watertable, either by uplift or climatic change tied to increasing local aridity, converts a former hypogene to meteoric phreatic cave into a vadose cave. In the latter, the walls become ornamented with gypsum and calcite speleothems. This is the recent history of the accessible portions of the gypsum maze caves worldwide, including documented examples in New Mexico, the western Ukraine and Saudi Arabia where the passage into the middle-upper Pleistocene marks the transition from phreatic to vadose in most of the accessible caves (Stafford et al., 2008). It is characteristic also of the very recent history of some natural phreatic caves where water tables were artificially lowered to allow quarrying of gypsum (Warren et al., 1990; Klimchouk, 2012).

Climate shifts or watertable fluctuations at the early end of the burial cycle (salt on the way down) can create alternating vadose-phreatic conditions in evaporite beds in the early stages of burial and so create an early watertable-associated karst level in the accumulating evaporites. That is, a gypsum bed hosting a vadose hydrology on its way down into burial, may pass through the watertable a number of times before its final burial and passage into the mesogenetic realm. This is the case today in the captured recharge playas in central Australia and about the edges of some halite-filled salars in the Andes and many Canadian salt lakes (Last, 1993; Warren, 2016), where fluctuating hydrologic conditions alternate between vadose and phreatic. Similarly, Quaternary climate shifts have variably karstified the gypsiferous sediments of many Sinic playas. For example, Yaoru and Cooper (1997) document Pleistocene lake basins in north-west China, such as in the Chaidamu Basin, where exposed gypsum beds evidence karst overprints that include: corroded flutes, fissures, small caves and associated collapse breccias and roof falls, followed by phreatic evaporite cement overprints. Watertable fluctuation is a hydrological overprint that is preserved as alternating ornamented surfaces in gypsum caves, disconformities and cave fills in many ancient lacustrine gypsum units.

Halite Caves

Because of its high solubility, halite does not make it into outcrop or shallow subcrop as easily as gypsum/anhydrite. Where halite is at the surface, it tends to be in regions of Pleistocene halite deposition (salt on its way down) or in zones of active diapirism (salt coming up very quickly). Chabert and Courbon (1997) noted caves in ancient rock salt in several regions: Algeria (in diapirs, mostly as vertical to sub-vertical shafts and short caves up to 28 m long), Chile (diapirs, with caves 250-500 m long), Israel (in Mt Sedom diapir as vadose caves and tube caves several hundred metres in length and subvertical vadose shafts that are metres across), Romania (in diapirs, with caves up to several hundred metres long), Spain (in diapirs, with

Figure 5. Kungur ice cave, Siberia. A) Cave map B) Cave interior. C) Sinkhole.
Dead Sea karst

Halite caves occur in the Mt. Sedom diapir, where halokinetic Miocene salt is sporadically exposed beneath a weathering and fractured gypsiferous caprock (Figure 6a,b; Frumkin, 1996; Frumkin and Ford, 1995; Frumkin 2009). Water enters the various caves in Sedom diapir through breaches in the caprock (Figure 7b). Most of the caves in the higher parts of Mt. Sedom salt are vadose inlet caves; these are meandering steeply inclined tubes and canyon slots located within or immediately below the caprock. They form where salt solution quickly carves out near vertical slots and shafts (typically < 2m broad and much deeper) that lead down from the surface, sometimes along pre-existing fractures and shears in the salt. Inlet caves in the central portions of the mountain can only be accessed through their sinks and appear to have no distinct outlet (Figure 6c). All terminate several tens of metres above the regional wattertable (e.g. Karbolot Cave; Figure 6c, d; Frumkin, 1994a, b, 1996).

The lower parts of inlet caves often contain steep silt and clay banks with surge marks that indicate occurrence of low energy water ponding, with variable residence times. Silt and clay sediments settling at the bottoms of inlet caves impede infiltration, extending the residence time of pond water (Frumkin, 1994a, 1996). Three of the studied caves in northern Mount Sedom had perennial ponds throughout the period 1984-1995. The ponds are perched, without any lithologic control, tens of metres above the nearest potential outlet at the foot of the mountain (Figure 6c). The water level in each pond differs from the others by tens of metres. All pond waters are highly concentrated, up to 324 g/l, with solutes consisting mainly of sodium and chlorine. Fresh inflow waters reach halite saturation within a few hours of reaching the pond. Both dissolution and precipitation features form the pond edges, and their equivalents can be seen on cave walls wherever ponds have dried out. Dissolution is indicated by horizontal notches, which connote density stratification in the ponds when aggressive fresh flood waters are temporarily diluting the upper parts of the pond. Subsequent saturation of holomictic pond waters is indicated by the growth of cm-scale halite crystals on the bottom and sides of the ponds.

Table 2. Landscape relief around active salt structures (after Warren, 2016).

<table>
<thead>
<tr>
<th>Location</th>
<th>Relief (m)</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abu Thama, Arabian Gulf</td>
<td>60</td>
<td>Neogene compressional reactivation</td>
</tr>
<tr>
<td>Cabo St Tome, Brazil</td>
<td>97-144</td>
<td>Gravity gliding</td>
</tr>
<tr>
<td>Allochthons, Gulf of Mexico</td>
<td>130-260</td>
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<tr>
<td>Offshore Mississippi Delta</td>
<td>100-240</td>
<td>Gravity gliding</td>
</tr>
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<td>Al Salif, Yemen</td>
<td>45</td>
<td>Extensional with gravity gliding</td>
</tr>
<tr>
<td>Jabal al Milh, Red Sea</td>
<td>45</td>
<td>Extensional</td>
</tr>
<tr>
<td>Mt Sedom, Dead Sea</td>
<td>200</td>
<td>Transtensional</td>
</tr>
<tr>
<td>Saharaan Atlas, Algeria</td>
<td>300</td>
<td>Compression with likely salt source cut-off</td>
</tr>
<tr>
<td>Kuh-e-Namak, Qom</td>
<td>305</td>
<td>Transtensional in compression belt</td>
</tr>
<tr>
<td>Zagros Chain, Iran</td>
<td>300-1500</td>
<td>Compression, some with salt source cut-off</td>
</tr>
</tbody>
</table>

The high solubility of halite means a halite cave system can form in a few hundred years rather than the thousands to tens of thousands of years needed to form carbonate karst (Bruthans et al., 2010). Not only is rate of cavern formation swift (dissolution/downcutting of 20 mm/year), these rapidly forming karst features are hosted in and atop a flowing rock mass. This means some of our notions of karst process and cave stability, related to the rate and density of cavern expansion, need to be modified when dealing with halite karst. For example, unlike gypsum and carbonate karst, jointing is not significant in controlling cavern orientation in namakier karst.
Figure 6. Salt karst features on Mt Sedom (salt diapir), Southern Dead Sea edge, Israel. A) Location map showing Malham cave area and Lot’s Wife. B) Lot’s Wife, view from north; human statue-like appearance of a degrading caprock section of the Sedom salt mound (salt diapir) that the superstitious give divine significance within Judeo-Christian biblical mythologies (aka apophenia and pareidolia mindsets). See also Figure 7a and Frumkin, 2009). C) East-West cross section through the southern part of Mount Sedom, showing ages of upper levels of caves in 14C yr B.P. Dashed lines indicate dip of salt layers within the diapir. Caves are coloured black. The oldest cave date comes from Lashleshet Cave, which lies at the top of the eastern escarpment, implying this was the earliest region on the Sedom uplift to experience vadose flushing. D) Vertical section along part of Malham South Cave, showing several dated levels. The age of the uppermost level (5570 ± 110 yr B.P) indicate the minimum exposure age of this site created by the halokinetic uplift of the landsurface (after Frumkin, 1996, 1998).
Towards the periphery of Mt Sedom, the inlet caves lead down to laterally expanding vadose cave levels that drain onto the Dead Sea plain (Figure 6c,d, 7c). Sedom Cave, is the longest laterally expanding diapir cave, with an aggregate length between two subparallel conduits of 1.8 km. Malham Cave, another large perched and laterally expanding cave, lies a few hundred metres south of Sedom Cave (Figure 6a, c, d; Frumkin, 1996). It has an aggregate passage length of more than 5.5 km and reaches to some 194 m below the landsurface. There is an upper tier of mostly inactive passages and a lower active channel level. 14C dates on fossil wood in the upper cave level shows meteoric waters began to sculpt the upper cave more than 5,500 years ago. Ongoing uplift of Mt Sedom salt means the active channel level in Malham Cave is now downcut some 10-12 metres lower than when it began. Malham Cave passages quickly developed an open outlet through which floodwater escaped directly to the Dead Sea floor, proving that during this period some 4000 years ago rock salt had already risen above region hydrological base level at the Malham outlet point within the eastern escarpment (Figure 6c). Lashelshet Cave, an inlet cave on the highest point on the diapir cross section, has an even older age of more than 7,000 years since cave initiation. Caves in the northern part of Mt Sedom did not begin to form until some 3,000 years later (Frumkin, 1996).

Based on their study of the caves of Mt Sedom, Frumkin and Ford (1995) concluded cave passages develop in two main stages: (1) an early stage characterized by inlet caves with high downcutting rates into the rock salt bed, and steep passage gradients; (2) a mature laterally expanding stage characterized by lower downcutting rates and the establishment of a wider subhorizontal perched stream bed armoured with alluvial detritus. This style of cave tends to develop toward the periphery of the diapir mound. In the mature expanding stage downcutting rates are controlled by the uplift rate of the diapir and changes of the level of the Dead Sea. Passages may aggrade to create wide flat bevelled passages and slots with thick sediment armoured bases (Figure 7c; Frumkin, 1998). A lack of a consistent phreatic level in the blind bottoms of perched water levels and the presence of the horizontal slots in the lower levels of Sedom Cave means dissolution in both types of caves is largely

Figure 7. Karst Mt Sedom, Israel. A) Lot’s Wife, view from west, showing it is a dissected caprock (buff) with underlying diapiric salt (grey) (see also Figure 6b and Frumkin, 2009). B) A collapse doline formed at the swallet of an ephemeral stream draining the Sedom caprock carapace and entering the vadose Malham Cave. C) Vadose canyon in Sedom Cave. All images courtesy of A. Frumkin (see https://serc.carleton.edu/69535)
restricted to times of flooding and perched or backed up freshwater in the vadose zone. This explains the tapering passages of inlet caves and the widespread alternation of armouring and bevelling as well as formation of narrow horizontal meandering slots toward parts of the top of the meandering channel that is now Sedom Cave.

Mass balance calculations in the halite caves of Mt Sedom yield downcutting rates of 0.2 mm s⁻¹ during peak flood conditions, this is about eight orders of magnitude higher than reported rates in any limestone cave stream (Frumkin and Ford, 1995). However, floods have a low recurrence interval in the arid climate of Mount Sedom so that long-term mean downcutting rates are lower: an average rate of 8.8 mm/ year was measured for the period 1986-1991, while Frumkin (2000) estimate the average regional vadose downcutting rate in the Mt Sedom karst region to be 20 mm/year. This is still at least three orders of magnitude higher than rates established for limestone caves and more than able to cope with the rate of supply of the diapiric salt.

The highly impervious nature of halite and its resupply in actively growing diapirs means that, unlike carbonate and gypsum caves, there is no real watertable level to define maximum cave development in a rising salt stem. Rather, the inlet caves are simple dissolution tubes where rainwater has accessed halite and sank until it was saturated and then dissolution stopped until the next flood. Toward the edge of the rising stem, these inlet caves breached the edge of the salt mound and vented their perched groundwaters to the surrounding plain (Figure 6; Sedom and Malham caves). This creates a downcutting and laterally expanding cave system, which is still some metres to tens of metres above the base level of the regional watertable in the surrounding plain. The expanding cave level is dominated by mostly horizontal growth, often with a sediment-armoured floor. It has numerous benches in the walls that probably reflect changes in the hydrological base level.

Dense anastomose cave networks that characterise gypsum caves are not found in the halite caves of Mt Sedom. This reflects the ability of diapiric halite to re-anneal and the fact that all exhumed halite that makes it to the surface is diapiric, not bedded. At-surface halite is not sandwiched between jointed aquifers above and below the dissolving layer. Rather it is a growing mound subject to dissolution at its top and sides.

Away from Mt Sedom, there are active collapse sinkholes and caverns forming in the alluvial fans and clastic aprons that overlie the bedded Quaternary lacustrine halite of the Dead Sea (Figure 8a, b; salt on its way down in the burial cycle). In the sediments around the lakeshore, the pace of karst collapse has accelerated in the last 60 years due to a drastic lowering of the circum-lake watertable and the associated lakeward migration of the saline-fresh water interface (Figure 9). For example, a series of collapse dolines 2-15 m diameter and up to 7 m deep, appeared in 1990 in the New Zohar area. In January 2001 a large sinkhole, some 20 m deep and 30 m wide, cut through the asphalt surface of the main road along the western shore of the
Dead Sea. It was opened by the passage of a busload of tourists on their way from Ein Gedi to the Mineral Beach solarium. Existing tourist facilities, such as the Ein Gedi beachside parking, were shut down after the road was damaged and several buildings have since collapsed into sinkholes. Sinkholes have since developed in other areas about the Dead Sea Margin including Qalia, Ein Samar and Ein Gedi. The process began in the southern part of the Dead Sea coast and slowly spread northward along the Israeli coast. Collapse is more localised in the northern and southern regions on the Jordanian side, and across the region, continues to increase in frequency as the sea level falls (Ezersky and Frumkin, 2013).

Three main types of sinkhole or doline fill have been recognized atop the dissolving Holocene salt beds of Mt Sedom; 1) Gravel holes in alluvial fans, 2) mud holes in the intervening bays of laminated clay deposits between fans, and 3) a combination of both types at the front of young alluvial fans where they overlap mud flats. Fossil, relict sinkholes have been observed in the wadi channels cutting into some old alluvial fans, showing this is a natural and ongoing process. While lake levels continue to fall (Figure 9b,c), the potential for subsidence hazards related to karst collapse is ongoing.

Sinkholes and related subsidence have been the focus of much geological study of the halite caves, but Closson et al. (2010) pointed out that an even more significant and ultimately damaging environmental effect of the ongoing water level lowering is the hectometre and larger scale landslides along the retreating shorezone. In the 1990s, international builders created major tourist resorts and industrial plants along the Jordanian and Israeli shore while, during the same period, geological hazards triggered by the level lowering spread out. From the beginning of the year 2000, sinkholes, subsidence, landslides, and river erosion damaged infrastructures more and more frequently: dykes, bridges, roads, houses, factories, pipes, crops, etc. all suffered as a result.

There is evidence of an older set of widespread ground collapses, sinkholes and caves that are tied to an earlier substantial fall in the Dead Sea water level some 4 ka. It may

Figure 9. Karstification along the newly exposed margin of the Dead Sea. A) Plan view showing zone of karstic collapse zones near a wadi outflow fan on the eastern margin of the Dead sea (31.333265°N, 35.542658°E). B) Collapse process via removal of salts due to lowering of water level and associated flow of freshwater along the interface of the sandy gravels and Lisan Formation salt at the Dead Sea margin. (after Salameh and El-Naser, 2000). C. Growth rate of sinkholes along the west coast of the Dead Sea, both in numbers and area coverage. The sinkhole development significantly accelerated since 1997; currently there are more than 1000 sinkholes, and their numbers are increasing at a rate of 150–200 per year (after Yechieli et al, 2006).
even be that the events described in Genesis 14 in the Christian Bible took place at the time of a substantially lowered sea level. The described battle, which occurred prior to the fall of Sodom and Gomorrah, perhaps took place on the subaerially exposed flats of the Southern Basin of the Dead Sea. The “pits of slime” described in the fall perhaps were solution collapse sinkholes activated by the 4000 ka fall in the Dead Sea water level (Frumkin and Elitzur, 2002).

Halite karst in diapiric Hormuz salt, Middle East

Namakier outcrops in and about the Arabian Gulf range from structures actively extruding salt to those in ruins where salt has not flowed for tens of thousands of years (Figure 9). Likewise the halite caves developed in the namakiers of Iran, or their offshore island counterparts, show a broad range of ages and styles of salt cave development tied to the time since cessation of salt flow (Filippi et al., 2011; Bruthans et al., 2010; Talbot et al., 2009).

Surfaces of actively flowing namakiers on the Iranian mainland are characterised by karren flutes and pinnacles, with numerous small-medium dolines, collapse structures, swallow holes and small caves at their base. As at Mt Sedom, caves tend to be sediment-armoured meandering tube caves or subvertical canyon slots that are centred on joints in the salt beneath a thin suffusion mantle. In contrast, the halite caves in the diapiric cores of the many islands in the Arabian Gulf have a more mature bevelled meandering style with thicker sediment armouring on the cavern floor. Many of these caves breach the retreating edges of former namakiers and salt fountains.

Salt movement in the various diapiric cores of these islands is inactive, or is greatly reduced, compared to the Miocene when these structures were active namakiers. For example, Dragon Breath Cave on Hormuz Island is a linear meander tube cave fed by an ephemeral stream in a shallow valley filling with alluvium (Figure 11; Bosak et al., 1999; Filippi et al., 2011). The surrounding landscape is classic salt karst with numerous depressions, blind valleys, ponors and subrosion sinks. Together they form a highly pockmarked centripetally-ringed topography, which outlines those central parts of the island underlain by shallow subcropping Hormuz salt. The cave itself is one of a number of tube caves exiting about the edge of the zone of diapiric salt. Hosted in steeply dipping diapiric salt, it is around 100 m long with its main passage created by a minor ephemeral stream. Its near flat roof, with an average inclination of 5.4%, is a notable feature and reflects either joint-related dissolutional spalling of the roof or an earlier watertable slot related to backup of a freshened water body.

The current cave passage has cut down a metre or more into earlier cave floor sediments (sediment armour), which contain clasts up to 50 cm in diameter. The cave formed by initial ingress along a linear joint, which was then widened by salt dissolution, so allowing meandering of the stream trace within the salt. It is a cave system very similar to the mature stages of laterally expanding caves in Mt Sedom. The base level of the cave correlates with the surface of a widespread marine terrace, which is now uplifted some 20 metres above sealevel and defines much of the periphery of Hormuz Island. The raising of the marine terrace is related to the ongoing raising of the island via salt flow.

Bruthans et al. (2000, 2010) show that the style of karst landform developed in dissolving diapiric salt in the Arabian Gulf Islands reflects the thickness of the carapace that caps the dissolving salt core (Figure 10). They distinguished four classes of diapir cap, each with a particular association of superficial and underground karst forms, namely: 1) outcropping salt, 2) thin capping (0.5-2 m), 3) capping with moderate thickness (5-30 m), 4) capping with greater thickness (more than 30 m). Cap thickness controls or reflects: 1) the density of recharge points, with high densities of recharge points in the thinner caps; 2) the amount of concentrated recharge which occurs at each recharge point, with suffusion karst characterising thinner caps; 3) the rate of lowering the ground surface atop the salt, with the faster rates of lowering occurring beneath thinner caps, and 4) the amount of load transported by underground flood-streams into cave systems. The volume of sediment load tends to be locally higher and focused beneath the thicker caps, particularly when inflow streams abut the edges of a dissolving salt dome. The thickness of caps atop expanding halite caves does not appear to influence the shape or style of the cave developed within

Figure 10. Landform evolution in an area where salt is flowing at the surface in a salt glacier (namakier). Based on various salt plugs in the Zagros foreland belt near the Straits of Hormuz.
the salt mass; more important seems to be the thickness of cap in the recharge area of the cave and the type of recharge into the salt environment. That is, how much water is passing into the salt and is its flow ongoing or ephemeral?

Halite caves in the relatively mature salt stems of the various islands of the Arabian Gulf, unlike carbonate systems, can swallow and store huge volumes of clastic sediment, volumes that would clog the entrance to a carbonate system. The extreme solubility of halite enables the pace of dissolution/corrosion enlargement in a salt cave to keep pace with large amounts of sediment carried into the cave by external inflows (Figure 11b). Stream sediments arriving at the cave entrance, including boulders, move inside and are trapped within the salt itself. Sediment is not dumped outside the cave entrance, which is the typical situation in blind valley river mouths at carbonate caves (Bruthans et al., 2003). For example, coarse-grained sediment fractions are carried hundreds of metres into the cave by two large intermittent streams entering the upper part of the Ponor Cave (Hormuz Island). The clasts in the resulting intra-cave alluvial fan conglomerates range from cobble of several centimetres up to 1 m diameter boulders; only sand-sized particles make it to the lower part of the same cave.

Caves capable of storing such coarse alluvium within the cavern itself are halite-specific with no equivalent in a carbonate karst terrain. There the boulder-size fraction in the cave itself is the result of roof fall, and almost all stream-borne coarse alluvium is deposited outside the cave.

Evaporite Speleothems

Cave walls in zones of less intense dissolution and stream crossflow are decorated with halite, gypsum and anhydrite speleothems (Figure 12). Halite has a much higher potential to form macro- or mono-crystalline speleothems than calcite and gypsum (Forti, 2017). Therefore, in most of the studied halite caves around the world, relatively large euhedral or hopper halite crystals have been observed as in the Iranian and Atacama caves (Forti, 2017; De Waele et al., 2009). The preferred location for these crystals are the pools in the cave entrances, where evaporation is sufficiently low to allow the development of euhedral crystals up to 10 cm in size (Figure 12, 13; Fillipi et al., 2011).

In the Iranian caves halite macrocrystals normally form also along streams, whereas in the Atacama Desert they
are completely lacking. This is because they need time to develop and therefore the stream must remain active for at least a few days (Figure 12a; Filippi et al., 2011).

Monocrystalline stalactites are widespread in the Iranian caves; the most common of which are the skeletal forms (Figure 12a; Filippi et al., 2011). An idealised skeletal stalactite normally consists of a central rounded “stalactite” from which, at different heights, three smaller and shorter rounded branches develops, being equally spaced at an angle of 120° (Figure 12b). Moreover, each branch and the central stalactite form an angle of ~70°. These values show that the directions of the central columns and the side twigs correspond to that of the four cube diagonals. Finally, at the end of each twig, there is a small halite crystal with one of its diagonals perfectly coincident with the twig (Figure 12b).

It is therefore evident that the entire structure of these peculiar stalactites consists of a single crystal lattice, albeit with a fractal appearance, and this fact is also confirmed by the presence along the rounded column of evident crystal facets oriented in the same direction (Figure 12b). Air currents and other local perturbing factors may cause a deflection from the theoretical direction of both the main column and of the side twigs (anemolites; Figure 12a). Finally, the rounded structure of the central column and of the external twigs is normally covered by glazy halite suggesting that cycles of deposition and dissolution alternate, while no inner feeding tube is present within the central column.

By these observations, the genesis of these skeletal monocrystalline stalactites is induced as driven by solutions mainly coming from brines and sprays, that then flow via gravity and capillarity only on the external surface of the stalactites. The amount and the composition of these solutions must change in time, becoming sometimes slightly undersaturated, probably during rain falls. The location of these speleothems close to waterfalls, where sprays are easily formed support that interpretation (Figure 13; Filippi et al., 2011; Forti 2017).

Compared to the growth rates of calcite speleothems in carbonate caves the growth rates of evaporite speleothems are phenomenal. Halite stalactites, several metres long and curving into the direction of airflow, formed in the mouth of Dragon Breath cave in a few years rather than millennia needed for carbonate counterparts (Bokacs et al., 1999). In August 1997, a network of numerous halite stalagmites and stalactites blocked the entrance to Dragon Breath Cave. In March 1998 there were no remains of the speleothem meshwork, while in February 1999, the stalagmites had reappeared (Bosak et al., 1999). Similar halite structures occur in the caverns of Mt Sedom and on the wet roofs of some salt mines.

Telogenetic halite deposits forming in namakiers encompass a range of mechanisms and speleothem textures (Figure 13; Filippi et al., 2011): i) via crystallization in/on streams and pools, ii) from dripping, splashing and aerosol water, iii) from evaporation of seepage and capillary water, and iv) other types of evaporative deposits. The following examples of halite textures are distinguished in each of the above-mentioned groups: i) euhedral crystals, floating rafts (raft cones), thin brine surface crusts and films; ii) straw stalactites, macromorpholine skeletal and hyaline deposits, aerosol deposits; iii) microcrystalline forms (crusts, stalactites and stalagmites, helictites); iv) macromorpholine he-
licrites, halite bottom fibres and spiders, crystals in fluvial sediments, euhedral halite crystals in rock salt, combined or transient forms and biologically induced deposits. The occurrence of particular forms depends strongly on the environment, in particular on the type of brine occurrence (pool, drip, splashing brine, microscopic capillary brine, etc.), flow rate and its variation, atmospheric humidity, evaporation rate and, in some cases, on the air flow direction. Combined or transitional secondary deposits can be observed if the conditions changed during the deposition. Euhedral halite crystals originate solely below the brine surface of supersaturated streams and lakes.

Macrocrystalline skeletal deposits occur at places with abundant irregular dripping and splashing (i.e., waterfalls, or places with strong dripping from the cave ceilings, etc.). Microcrystalline (fine-grained) deposits are generated by evaporation of capillary brine at places where brine is not present in a macroscopically visible form. Straw stalactites form at places where dripping is concentrated in small spots and is frequently sufficient to assure that the tip of the stalactite will not be overgrown by halite precipitates. If the tip is blocked by halite precipitates, the brine remaining in the straw will seep through the walls and helictites start to grow in some places.

Macrocrystalline skeletal halite deposits and straw stalactites usually grow after a major rain event when dripping is strong, while microcrystalline speleothems are formed continuously during much longer periods and ultimately (usually) overgrow the other types of speleothems during dry periods. The rate of secondary halite deposition is much faster compared to the carbonate karst. Some forms increase more than 0.5 m during the first year after a strong rain event; however, the age of speleothems is difficult to estimate, as they are often combinations of segments of various ages and growth periods alternate with long intervals of inactivity.

Anhydrite forms speleothems in preference to gypsum in those rare parts of a cave with very high salinity waters. But overall gypsum speleothems dominate. Some of these gypsum speleothems can be quite large, up to a few metres
Unlike halite and gypsum caves, which are rich in halite and gypsum formations respectively, anhydrite caves do not host anhydrite speleothems at all. This is a direct consequence of the CaSO$_4$ – CaSO$_4$.2H$_2$O solubility disequilibrium, which makes the hydrated mineral (gypsum) less soluble than the anhydrous one (anhydrite) at normal cave temperatures, thus totally hindering the development of secondary anhydrite formations. Most of the gypsum produced by hydration replaces anhydrite within the rock structure, and therefore anhydrite does not form any speleothems. Nonetheless, a minor part of this secondary gypsum may develop some small deposits. In the caves of the Upper Secchia Valley, small gypsum crusts and flowstones were observed where condensation water, after dissolving anhydrite, flows over the gallery roof or walls where air currents induce evaporation (Chiesi & Forti, 1988). In the same caves, when per ascensum capillary flow and evaporation are possible, euhedral aggregates of small gypsum crystals may develop on top of rock. The interested reader is referred to a comprehensive paper by Forti, 2017 dealing with the great variety of halite and gypsum speleothems.

The Upper Secchia Valley anhydrite caves and their chemical deposits were already known when the first monograph on speleothems in gypsum caves was published (Forti, 1996). However, at that time these caves were incorrectly considered as formed in gypsum and therefore their deposits were described along with those hosted in classical gypsum karst.

The genesis and evolution of the German and Italian anhydrite caves are completely different; in fact, the first are hypogenic caves (Kempe, 2014) and lack any natural entrance, whereas the second ones are epigenic and often develop very close to the surface (Malavolti, 1949). Therefore, chemical deposits are different in the two locations and restricted to the peculiar environment that controlled the evolution of the caves.

Leaving aside the widespread secondary gypsum produced by the hydration of the host rock, anhydrite caves are extremely poor in chemical deposits. The lack of minerals in the hypogenic caves is because they were filled with near-stagnant water for most of the time during their development. In the epigenic caves, instead, the absence of cave minerals is mainly attributed to the strong increase in volume caused by hydration of anhydrite (that turns into gypsum), which makes the wall and the ceiling of these cavities extremely fractured. In this latter setting, the rather continuous breakdown normally inhibits the development of even small chemical deposits, which, in any case, are easily washed away by the frequent floods that characterise the Upper Secchia Valley. Despite all these restrictions, the anhydrite caves proved to be interesting not only from a mineralogical point of view, as they host one cave mineral (clinochlore, Chiesi & Forti, 1985) restricted to this environment, but also for the presence of a unique gypsum/anhydrite speleothem, i.e., the huge “leather-like sheets” of Barbarossa Cave (Figure 14).

**Implications**

Part 1 and Part 2 of this set of articles dealing with evaporite dissolution emphasise the importance of rapid rates of volume loss in creating a unique set of karst landforms and speleothems. This rapidity creates cavities in a hydrological milieu of contrasting brine salinity and temperature interfaces and permeability contrasts. This inherent association of voids in a setting with abrupt chemical interfaces facilitates the enrichment levels of economic commodities (part 4) and drives rapid bed stoping and foundering that forms zones of significant geohazard in the landscape (parts 3 and 4).
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