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PALEOENVIRONMENTAL DATA FOR N. W. GEORGIA, U.S.A., FROM FOSSILS IN CAVE SPELEOTHEMS

George A. Brook*, Eugene P. Keferl** and Rudy J. Nickmann***

SUMMARY

Pollen grains and gastropod shells in two speleothems from Red Spider Cave, Georgia indicate that c. 10,000 yr B.P. the vegetation near the cave was Mixed Mesophytic Forest. Conditions were cooler and moister than today and a shallow pond existed in the doline above the cave. As these findings support palynologic evidence from nearby pond sites it is clear that cave speleothems are a potential source of paleoecological data to c. 350,000 yr. B.P.

INTRODUCTION

Cave speleothems (stalactites, stalagmites and columns) have become increasingly important in paleoenvironmental research because they can provide paleomagnetic, paleotemperature and paleohydrologic data (SCHWARCZ ET AL., 1976; LATHAM ET AL., 1979; BROOK, 1982). Most importantly, any data obtained from speleothems can be placed in an accurate chronologic framework because the calcite of these formations is dateable by the $^{14}$C and $^{234}$U/$^{230}$Th methods to c. 50,000 and c. 350,000 yr. B.P., respectively. However, despite the finding that speleothems may contain large numbers of pollen grains (BASTIN, 1978), there have been few studies to ascertain the usefulness of fossils in cave formations as paleoenvironmental indicators. Studies of Red Spider Cave, Georgia, begun in 1982, have demonstrated that speleothems may contain abundant fossils that can provide important, dateable, paleoenvironmental information.

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Red Spider Cave (elev. 274 m) is located in Johnson's Crook, Dade County, Georgia (Fig. 1). The entrance is on the northwest-facing slope of a resistant ridge at the base of Lookout Mountain. The cave, in limestones, is 135 m long with a vertical relief of 8 m. One passage, Red Spider Crawl, passes beneath a doline 15 m deep in the sandstones and shales which cap the ridge in which the cave has developed. After rain, water enters Red Spider Crawl through the main swallet of the doline. The cave was once almost totally filled with clastic sediments up to 3 m thick. Remnants of this fill are still evident on many passage walls. The mean annual temperature at the cave is 14.9°C and the annual precipitation is 1,427 mm.

When sectioned, a stalagmite (RS-3) and a column (RS-4) recovered from Red Spider Crawl were found to contain numerous gastropod shells (Fig. 2). Samples of calcite were taken from these formations for pollen analysis and for \(^{14}\text{C}\) dating. Assuming that the speleothem calcite was deposited with 85% modern carbon (Franke and Geyh, 1971; Cooke and Verhagen, 1977; Hennig et al., 1980), RS-3 is of very early Holocene age (9,900±260 yr B.P.-UGa 3341) and RS-4 of late Glacial age (10,880±990 yr B.P.-UGa 3986).

THE GASTROPOD SHELLS

Speleothems RS-3 and RS-4 were cut into 1 cm thick slabs, 110 intact gastropod shells or gastropod shell fragments were observed on the slab surfaces (Table 1). Five different gastropod species were identified. Eighty specimens were of the Genus Carychium. Only a small number could be identified to the species level but the available evidence suggests that all are the species Carychium exile (Lea, 1842). Ten specimens belong to the Family Zonitidae, at least seven of these have been identified as Hawaii

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<th>Speleothem</th>
<th>Carychium</th>
<th>Hawaii minuscula</th>
<th>Unknown Zonitidae Species 1</th>
<th>Unknown Zonitidae Species 2</th>
<th>Unknown Fresh-water Species</th>
<th>Misc. Unknown Species &amp; Fragments</th>
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<td>RS-3</td>
<td>39</td>
<td>2</td>
<td>0</td>
<td>1</td>
<td>0</td>
<td>7</td>
<td>49</td>
</tr>
<tr>
<td>RS-4</td>
<td>41</td>
<td>5</td>
<td>2</td>
<td>0</td>
<td>1</td>
<td>12</td>
<td>61</td>
</tr>
<tr>
<td>Total</td>
<td>80</td>
<td>7</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>19</td>
<td>110</td>
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Table 1 — Gastropod Shells in Two Speleothems, Red Spider Cave.
Fig. 1 - Location and profile of Red Spider Cave showing the positions of speleothems RS-3 and RS-4.
*minuscula* (Binney, 1840), the other specimens could not be identified to the species level. Two Zonitidae specimens in speleothem RS-4 are too small and the whorls too narrow to be *H. minuscula*, and one specimen in speleothem RS-3 is too large to be this species. In speleothem RS-4 another unknown species 1.6 mm high, 0.75 mm wide, with 3.5 whorls, appears to be a freshwater gastropod.

Shells and living specimens of *H. minuscula* and *C. exile* have been discovered previously in caves (Morrison, 1939; Hubricht, 1941, 1964), and also in cave clastic sediments of Pleistocene age (LaRoque, 1967; Parmalee, 1967; Guilday et al., 1978). Although Hubricht (1964) reports *C. exile* in seven caves in Kentucky, Tennessee, and Alabama in sufficient numbers to suggest that they lived and bred there, it appears that in most cases specimens in caves are washed in through the entrance or through fissures in the roof. As far as we are aware, there has been no previous report of large numbers of gastropod shells being found in cave stalagmites and columns, although numerous land snail shells have been discovered in layers of flowstone deposited on cave floors (Gillieson and Mountain, 1983).

As some of the shells in speleothems RS-3 and RS-4 were broken, the gastropods were clearly not living on these formations. The shells were transported into the cave (probably through fissures in the roof) after the animals had died. *C. exile* and *H. minuscula* shells are extremely delicate so that the perfect preservation of several specimens is convincing evidence that the shells are contemporary with speleothem deposition and do not represent reworked older deposits. The shells and shell fragments show no evidence of the degradation to be expected if they had resided in the acid soil layer (on shales) above the cave for any significant length of time.

*H. minuscula* and *C. exile* are most commonly found in leaf mould and forest debris in wet and moist habitats. Today, both species inhabit moist litter in the doline above Red Spider Cave. Of 254 gastropod shells in five litter samples taken from the slopes and floor of the doline, 57 (22.4%) were of the species *C. exile*, and 5 (2.0%) were *H. minuscula*. No fresh-water gastropod shells were recovered from any of the samples. *H. minuscula* has been recorded from numerous mesic forest associations in various parts of North America. River and stream floodplains and lake and pond shores appear to be common *H. minuscula* habitats (Baker, 1911; Oughton, 1948; Dexter, 1950; Leonard, 1959). *C. exile* has been recorded in a wide variety of plant communities, particularly deciduous forest, and was collected from the moist areas within these communities (Ba-
KER, 1911; SMITH, 1928; KEEFERL, 1975; CONEY ET AL., 1982). Significantly, both *H. minuscula* and *C. exile* are conspicuously absent from most coniferous forest communities and from xeric deciduous forest communities.

Abundant shells of *H. minuscula* and *C. exile* in speleothems RS-3 and RS-4 suggest that by the late Glacial and early Holocene, the predominantly coniferous forest of full Glacial time (DELCOURT AND DELCOURT, 1981; WATTS, 1983) had been replaced by a mesic deciduous forest. The gastropods were probably living in moist forest litter on the slopes of the doline above the cave and were washed into Red Spider Crawl through fissures in the roof (Fig. 1). The presence of what appears to be a freshwater gastropod shell in speleothem RS-4 raises the possibility that water was ponded in the doline above Red Spider Cave during late Glacial times. Ponding may have occurred because swallets, which normally drain

![Fig. 2 - Section of speleotherm RS-4 showing shells of *H. minuscula* (center), and *C. exile* (lower right).]
the doline, became blocked with sediments perhaps at a time when the passages of the cave were also largely filled with clastic debris. If ponded water was present in the doline c. 11,000-10,000 yr B. P., it is possible that large numbers of *H. minuscula* and *C. exile* were living along the moist shore of the pond. Pond shores are favored habitats of both species today.

**THE POLLEN SPECTRA**

Samples of calcite from speleothems RS-3 and RS-4 yielded 537 and 384 pollen grains, respectively (Fig. 3). Based on studies by BASTIN (1978) it is evident that speleothem pollen spectra reflect the local vegetation near the cave rather than the regional pollen rain. Therefore, pollen spectra for the two Red Spider Cave speleothems are considered to provide evidence about vegetation immediately above and near the cave at the time of speleothem deposition. Long residence in the oxidizing environment of the soil above the cave, where pollen degradation is rapid, is not indicated by the good preservation of the pollen grains recovered. Therefore, these plant microfossils are believed to be contemporary with speleothem deposition and, like the much larger gastropod shells in these formations, are believed to have been washed into the cave through fissures in the roof.

Speleothem RS-4, dated to 10,880 ± 990 yr B. P., contained 70.6% arboreal and 21.2% non-arboreal pollen. *Quercus* (oak) dominated the pollen sum with 32.6% of the pollen. *Pinus* (pine), *Ostrya-Carpinus* (hornbeam), and *Ulmus* (elm) followed with 16.2%, 11.5% and 6.3% of the pollen, respectively. *Gramineae* (grasses), *Cyperaceae* (sedges), and *Ambrosia* (ragweed) made up the bulk of the non-arboreal pollen content contributing 11.2%, 5.0%, and 2.6% of the total, respectively. Speleothem RS-3 contained 87.9% arboreal and only 8% non-arboreal pollen suggesting a more closed forest cover near the cave by 9,900 ± 260 yr B. P. The forest must have been largely oak (42.3% of the pollen sum) with pine contributing only 6.7% of the pollen rain. Oak and pine made up only 63.3% of the arboreal pollen in speleothem RS-3, compared to 88% in speleothem RS-4, reflecting the increased importance of other deciduous taxa such as *Ulmus*, *Fraxinus* (ash), *Fagus* (beech), *Liriodendron* (tulip tree), *Carya* (hickory), *Corylus* (hazel), and *Ostrya-Carpinus* which contributed 14.9%, 6.4%, 3.5%, 4.5%, 2.8%, 1.3% and 1.1% to the pollen sum, respectively. *Gramineae* (3.2%) and *Cyperaceae* (1.7%) were the main non-arboreal pollen types in speleothem RS-3.
Spore percentages are calculated using Pollen Sum and Spores as a sum.

Aquatic percentages are calculated using Pollen Sum and Aquatics as a sum.

* = Abundant

Fig. 3 - Pollen and spore spectra for speleothems RS-3 and RS-4.
The speleothem pollen spectra indicate that in late Glacial and early Holocene times the vegetation near Red Spider Cave was a largely deciduous forest. Lower non-arboreal pollen percentages and lower percentages of pine pollen in speleothem RS-3 suggest a trend in late Glacial times from a relatively open Oak-Hickory-Pine Forest at 10,880±990 yr B.P. to a closed Mixed Deciduous Forest more dominated by deciduous taxa at 9,900±260 yr B.P. *Picea* (spruce) pollen in both speleothems and *Tsuga* (hemlock) pollen in RS-3 imply more available moisture and a cooler climate than today in late Glacial and early Holocene times. Small percentages of aquatic pollen (*Typha latifolia* (cat-tail) and *Sparganium* (bur-reed)) suggest that there was standing water nearby — the most likely location being in the floor of the doline above the cave.

**CONCLUSIONS**

The pollen grain and gastropod shell assemblages in two speleothems from Red Spider Cave, Georgia both indicate that in late Glacial and early Holocene times the cave was deciduous forest. The pollen grains and gastropod shells were probably transported into the cave by water percolating through fissures in the roof. The pollen evidence suggests that between 10,880 and 9,900 yr B.P. the vegetation changed from an Oak-Hickory-Pine Forest to a Mixed Deciduous Forest and that throughout the period water may have been ponded in the doline above the cave. A possible freshwater gastropod shell in speleothem RS-4 also suggests standing water in the doline.

**DELCOURT and DELCOURT** (1981) and **WATTS** (1983) contend that in the late Glacial and early Holocene a cool, moist climate with abundant moisture in the growing season favored the widespread expansion of species-rich, Mixed Deciduous Forest from 34° to 37° N latitude in eastern North America. This area was dominated in the winter by the Pacific Airmass and in the summer by the Maritime Tropical Airmass (**DELCOURT AND DELCOURT**, 1984). The speleothem pollen evidence of Mixed Deciduous Forest at Red Spider Cave (lat. 34° 53’ N) c. 10,000 yr B.P. is in essential agreement with this argument.

In summary, studies at Red Spider Cave have shown that macrofossils and microfossils may be present in cave speleothems in large numbers. The evidence obtained from two Red Spider Cave speleothems, indicating a Mixed Mesophytic Forest vegetation near the cave in early Holocene ti-
mes, is in essential agreement with palynological data obtained from nearby pond sites (DELCOURT AND DELCOURT, 1981; WATTS, 1983). This suggests that fossils in cave speleothems, which have not drawn significant scientific interest to date, could be a valuable source of paleoecological data to c. 350,000 yr B.P. — the range of the $^{234}\text{U}/^{230}\text{Th}$ dating method.

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REFERENCES


Manuscript received: October 1986
STRATIGRAPHIC SECTIONS IN THE STE. GENEVIEVE FORMATION (MIDDLE MISSISSIPPAN) EXPOSED IN GARRISON CHAPEL KARST AREA CAVERNS - WESTERN MONROE COUNTY, INDIANA USA

Garre A. Conner*

SUMMARY

The Ste. Genevieve Formation and related strata in the Blue River Group comprise more than 75 meters of Middle Mississippian carbonate deposition across the Indiana portion of the Eastern Interior Basin in Valmeyeran seaways. Forty kilometers of subterranean caverns occur in this carbonate rock sequence in the Garrison Chapel karst area where blind valleys are a striking topographical feature. The bedrock floor of a karst valley is locally accordant with a continuous horizon of lithographic limestone named Indian Creek Beds and illustrated on five cavern stratigraphic reference profiles.

INTRODUCTION

The Garrison Chapel karst area in western Monroe County, Indiana has been a popular area of subterranean exploration and scientific investigations pertaining to cavern origin and development. More than 40 kilometers of subterranean streams and canyons have been surveyed from water catchment sinkholes and blind valleys downward to the spring resurgences. These caverns lie along the western margin of a distinctively karst erosional valley occupying 35 square kilometers located southwest of Bloomington, Indiana in the Crawford Upland physiographic province (MALOTT, 1922, p. 197-203). The Garrison Chapel karst area of about 10 square kilometers and the karst valley at the headlands of Indian Creek are illustrated in Fig. 1.

The bedrock host or floor of this karst valley consisting of sinkholes and downstream segments of sinking streams is locally accordant with a solutionally incised horizon of lithographic limestone that is laterally continuous throughout the headland area of Indian Creek. The lithographic limestone beds are generally 3 to 5 meters thick, thin and evenly bedded

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Fig. 1 - Karst valley-headlands, Indian Creek and Garrison Chapel karst area, Monroe Co., Indiana.
cryptocrystalline micrite. They are white with tan and green clay stains. Clay intercalations occur along bedding planes and vertical fractures are filled with clay and recrystallized calcite. Styolites are observed at some exposures coinciding with bedding planes. Stratified nodular chert may be present at some exposures. The lithographic texture distinguishes these strata essentially comprising the lower half of the Levias Member of the Ste. Genevieve Formation. The Ste. Genevieve Formation in the Blue River Group is middle Mississippian in age. These lithographic beds are named the Indian Creek Limestone Beds (CONNER, 1986) (Fig. 2).

Recognition of the Indian Creek Beds as a mappable unit is attributed in part to exposures in cavern sections in the Garrison Chapel karst area and comparison to established surface exposures described in the literature. Five measured stratigraphic reference sections in cavern exposures associated with the karst valley are presented with this investigation and illustrate the cycle of subterranean cavern development and related surface valley drainage accordance with the structural form the the Indian Creek Beds (Fig. 3).

THE GARRISON CHAPEL KARST AREA

The Garrison Chapel karst area (Fig. 4) described in the Convention Guidebook of the National Speleological Society (1973, p. 16-34), refers
generally to four major subterranean cavern systems. These westward draining caverns are oriented along the strike axis of the karst valley from Garrison Chapel southward to the Illinois Central Railway grade. The northernmost cavern system considered here is comprised of the hydrologically integrated streams flowing through Salamander, Shaft, Grotto and Coons Caves. Wayne and Buckner Cave lie to the south as separate hydrological drainage systems. Farthest south is the Blair Springs system comprised of Triple J, Brinegar, Trap Door, and King Blair Caves. Many smaller caves are related to these systems. Other major systems lie to the north and east of this area of the karst valley, but were not included in this investigation. POWELL (1960) published a statewide survey of known caverns in Indiana. The periodical Newsletter of the Bloomington Indiana Grotto is an important source for articles and maps pertaining to caverns of the Garrison Chapel area.

RECOGNITION OF THE KARST VALLEY

The fifteen minute topographical map of the Bloomington quadrangle (MARSHALL, 1910), revealed the karst landforms in the headlands of Indian Creek. J. W. BEEDE (1911), investigated the origin of the karst valley there with description of the area including the Garrison Chapel.

[Diagram image]

Fig. 3 - Measured cavern sections and stratigraphic position.
Fig. 4 - Indian Creek Beds structure contours in karst valley.
area while focusing on the eastern side of the valley in the area of Leonard Springs. The springs and related landforms were illustrated with photographs and a pen sketch revealing Beede's interpretation of subterranean stream piracy. Sinkholes above the caverns intercept surface waters from short segments of Indian Creek diverting them downward and southwestward under the Garrison Chapel area through cave streams discharging into Richland Creek. Trending in the opposite direction other cave streams flow southeastward below the Leonard area discharging waters into Clear Creek to the east.

C. A. MALOTT (1922) described the landforms and drainage related to the karst valley of Indian Creek and the related phenomena of subterranean stream piracy. MALOTT's sketch map outlined the entire karst valley showing internal surface drainage divides and sinkhole in the blind valleys similar to the outlines in Fig. 1. The origin and age of the karst valley were discussed by MALOTT in relation to the Kirksville Peneplain (MALOTT, 1919, p. 23) of the region. Stratigraphic names Paoli and Ste. Genevieve Limestone were in use at that time, but MALOTT's reference to the strata employed the earlier terms Mitchell Limestone and the overlying Coal Measures of the Carboniferous strata.

W. J. WAYNE (1949) described a karst plane or sinkhole plane as a region underlain by limestone in which all of the drainage is underground. Wayne's map illustrated principally the same area as MALOTT's map. WAYNE discussed the jointed and oolitic nature of the Paoli and Ste. Genevieve limestones; stating the operational stream piracy of the headwaters of Indian Creek by Richland and Clear Creeks.

STRATIGRAPHIC NOMENCLATURE FOR THE CAVERN SECTIONS

Stratigraphic names employed in the illustrations of the measured rock sections in the caverns represent (Fig. 2) units of current use in rock-unit stratigraphy in Indiana.

Emphasis on the word member in the stratigraphic sense, used here, for measured sections rather than the words limestone or sandstone following a proper name is intended to maintain conformity with established nomenclature for the Blue River Group in Indiana while omitting an exiguous and reiterative discussion of the attendant nomenclatural revisions over past decades.

The Blue River Group established by GRAY & AL. (1960), includes in ascending order the St. Louis and Ste. Genevieve Formations succeeded
by the Paoli Limestone. Divisions of the Ste Genevieve Formation include the Fredonia Member, with the Lost River Chert Bed, the Rosiclare Member, and the Levias Member including the Indian Creek Beds and the Bryantsville Breccia Bed at the top.

The term Spar Mountain Member of Illinois use has been synonymously substituted for the Rosiclare Member of Indiana which has been agreeably accepted by most investigators.

The Paoli Limestone and lower sandy beds were considered by N. M. Smith (SHAVER ET AL., 1970; p. 125-128), to be the out-crop equivalent of the Renault Formation. The lower sandy beds were consistently recorded at Indiana exposures under the name Aux Vases sandstone by MALOTT (1952) and by PERRY & SMITH (1958). Later an exposure of this unit in Lawrence County Indiana was named Popcorn Sandstone Bed by SWANN (1963, p. 31-32). The Popcorn Bed is continuous through the Garrison Chapel area as a calcareous sandstone and shale, but is less than 2.5 cm. in thickness in many exposures there. Illustrated on the figures it is not labeled at its position overlying the Bryantsville Breccia Bed.

DESCRIPTION OF THE CARBONATE AND RELATED LITHOLOGIES

A legend of the lithologic and stratigraphic characteristics of the Blue River Group exposed in the caverns is illustrated in Fig. 5. Legend symbols represent the main lithotypes, bedding, and jointing characteristics which allow the beds to be recognized and correlated among isolated exposures.

Carbonate rocks were classified by relative size of crystallinity, grainstone component, and cementation. Sparite generally refers to a cementing material or cavity filling. Crystal size was not directly measured during microscopic examination of hand samples. Granular is a term applied to dolostones, but also to certain calcite and partially dolomitized beds. Micrite was recognized in crystalline and detrital modes, but this detail was omitted from the descriptions substituting the terms crystalline or granular.

Algal structures and related sub-areal laminated crust were frequently observed in the top of the Levias associated with the Bryantsville Breccia Bed. Low and steeply inclined joints were commonly exhibited in the oolitic and bioclastic beds in the upper half of the Levias Member and the Paoli Limestone. Thin laminar and cross laminar bedding is characteristic of the sandy beds in the Spar Mountain Member. Massive argillaceous detri-
tal limestones with nodular chert and silty laminations are typical of the Fredonia Member including dolostone beds.

DESCRIPTION OF MEASURED CAVERNS SECTIONS

Description and measurement of the carbonate rock sections in the caverns were made at Breakdown Mountain in Salamander Cave (Fig. 6); the entrance and lower offset pit in Shaft Cave (Fig. 7); the entrance and offset chambers in Grotto Cave (Fig. 8); entrance, Signature Room, and waterfall in Buckner Cave (Fig. 9); and the entrance crevice in Triple J Cave (Fig. 10).

Bedrock surfaces exposed in cavern walls and shafts are frequently more amenable to recognition of various lithologies than corresponding surface exposures. Surface frost wedging, vegetation, and direct sunlight
are not a major influence on modification of cavern walls that are sculptured by water or gravity fall of jointed blocks.

Irregularities and surface relief exposed on the cavern walls results from solution features which are generally smooth and curved or from breakdown falls which leave straight and angular fracture surfaces. Both types of features reveal some degree of surficial leaching as a response to differential moisture after the bed is left above the high water level of the cave stream.

Individual beds or groups of beds are observed to stand in relief from the plane of the wall or are reentrant. The reentrant zones are recessed behind the wall plane. Harder and more indurated carbonates and those high in insoluble grains as well as the more coarsely crystalline beds tend to stand in relief from walls in contrast to the oolitic and bioclastic beds. Clayey shales tend to be more reentrant where plastic flow, pressure unloading, and differential moisture influence them.

Sandstones and silty shales usually are exposed in relief from the walls, but are not always conspicuous. Dolostone is usually found in relief and is often sculptured into erosional forms extending across the entire width of canyons as bridges because of its lower solubility rate and a pro-
pencency to become coated with precipitated carbonate and manganese dioxide. Where dolostone is not coated it typically develops a soft chalky surface. Carbonate precipitated on cavern walls results in various forms referred to as speleothems frequently covers silty or sandy beds where porosity permits water seepage. Wall forms are important in recognizing the textural types of carbonate strata when tracing from one reference section to the next.

Breakdown Mountain in Salamander Cave (Fig. 6) is located about 150 m. upstream from the entrance and is generally accessible except for rapid short term crest during thunderstorms. The section profile is drawn facing outward or downstream. A floor elevation of 227 m. mean tide was
established by Paulin altimeter. The lower canyon walls reveal the top sandy beds of the Spar Mountain Member exposed nine feet above the floor. The Indian Creek Beds in the lower Levias Member above measure 2.9 m. in thickness with the base resting on the Spar Mountain. Above in the breakdown chamber the oolitic, bioclastic and jointed beds of the upper Levias Member are well exposed in the ceiling. The Bryantsville Breccia Bed was not observed, but would lie several feet above the ceiling.

The entrance pit to Shaft Cave (Fig. 7) is located about 180 m. southwest of Breakdown Mountain in Salamander Cave. The entrance lies at an elevation of 251 m. mean tide and drops 23 m. to a canyon developed in the middle of the Spar Mountain Member leading to an offset lower pit 9 m. away where 6 meters of strata are exposed down to a lower stream canyon in the top of the Fredonia Member. At the top of the entrance pit a finely crystalline micritic limestone is exposed. Below the lip of the pit 0.75 m. of calcareous sandstone is exposed consisting of brown weathered very fine grained quartz. The stratigraphic names of these two units is discussed in the summary of correlations. The Paoli Limestone measures 7.1 m. thick. The lower sandy bed, the Popcorn Bed of Swann, is obscured by flowstone, but measures 9.1 cm. thick resting on the Bryantsville Breccia Bed. The entire Levias Member measures 9 m. thick with the Indian Creek Beds comprising 2.9 m. The Spar Mountain Member measures 10.4 m. including a thin silty bed near the floor of the lower pit. Below, one foot of finely crystalline limestone is exposed in the floor of the stream canyon marking the upper unit of the Fredonia Member. Shaft may be entered safely in dry weather with modern rope ascending equipment.

Grotto Cave section (Fig. 8) was measured in the entrance chamber about 180 m. southwest from Shaft. A steep slope descends to a small dome and adjacent chamber where the section was continued. The surface reference elevation was 244 m. mean tide measured with Paulin Altimeter. From there 6 m. of Paoli Limestone was measured downward to a 3 cm. grey silty lamination marking the Popcorn Bed. No Bryantsville Breccia was observed. The silty unit rests directly on a micritic bioclastic bed of upper Levias Member. The Levias measures 9.5 m. feet including 2.8 m. of Indian Creek Beds exposed at the boulder which must be climbed over in decent near the bottom of the chamber. The exposed section of Spar Mountain measures 8.6 m. from the base of the Indian Creek Beds to the top of the Fredonia 22 m. away in the adjacent chamber. The top and base of the Spar Mountain Member are marked by thin resistant sandy lentils in relief. These sandy lentils are exposed in the adjacent chamber, but were
not observed in the entrance chamber. Instead in the lower part of the entrance chamber there is 1.9 m. of argillaceous limestone exposed, with no sand, immediately below the Indian Creek Beds. Again as at Salamander these argillaceous beds are included in the Spar Mountain Member in an effort to establish the base of the Indian Creek Beds in contact with the Spar Mountain Member. The sandy lentils in the adjacent chamber include three modes of detrital quartz.

Buckner Cave section (Fig. 9) lying 1.8 km. to the south is an extended vertical profile starting at the entrance, elevation 247 m. mean tide, continuing downward through the crawlway and Signature Room then through the stream canyon to the base of the waterfall. There is 0.9 m. of
Paoli Limestone exposed in the ravine above the entrance. The Popcorn Bed measures 15 cm. of silty shale and the micritic laminar crusts of the Bryantsville Breccia Bed form the ceiling. The Levias Member measures 11.3 m. including 3.7 m. of Indian Creek Beds. Near the Signature Room the base of the Indian Creek Beds is well exposed near the ceiling where 2.4 m. of slightly sandy crystalline limestone underlies them marking the contact with the Spar Mountain Member. The entire Spar Mountain Member measures 4.9 m. with detrital chert in the sands. The Fredonia Member measures a total exposed thickness of 11.4 m. traced through the stream canyon, with a very steep floor gradient, and ending in a waterfall. The Lost River Chert Bed is exposed as a resistent ledge below the lip of the waterfall.

Fig. 9 - Buckner Cave section.
The Triple J Cave section (Fig. 10) lies 900 m. south of the Buckner Cave entrance and is estimated to lie at an elevation of 245 m. mean tide. One meter of Paoli limestone is exposed inside the entrance in the sinkhole. On a ledge 36.6 cm. of sandstone is exposed which is identified as the Popcorn Bed, the best development of the Popcorn in the Garrison Chapel area. Alternatively, this sandstone may be considered to be the upper sandstone exposed in Shaft Cave and consequently the identification of the Paoli Limestone at Triple J would be affected. At Triple J the Bryantsville Breccia Bed is not recognized. Considering the observation that the Popcorn Bed is much more thickly developed several miles southeast of Triple J Cave and the exceptionally thick oolitic section in the upper Levias Member at Triple J where the Bryantsville is absent it is preferable to refer the sandstone of question to the Popcorn horizon. The Levias Member is unusually thick measuring 12.0 meters attributable to a thick upper oolitic section; perhaps near the center of a convex oolite body. The Indian Creek Beds below measure 3.6 meters and rest directly on a sandy breccia bed marking the top of the Spar Mountain Member. Chert sand is also present in the breccia, however the section is different than at the Signature Room in Buckner Cave. In Triple J Cave the Spar Mountain measures 4.7 m. to the floor at the entrance. Below the sandy breccia section lies 3.2 meters of micritic lime with sparse chert nodules. Farther downstream the lower Spar Mountain Member was recognized beyond Aqua Avenue at survey stations C5 and C11 through to the Cherty Channel. There a 0.5 m. thick green sparry gastropod and brachiopod limestone capped by a 15.2 cm. silty clay bed is recognized as the lower boundary of the Spar Mountain Member overlying a micritic limestone with concentrations of chert nodules representing the upper Fredonia. This same silty clay bed and gastropod limestone association is observed four miles to the southeast in the Mountain Room and Blue Pool Canyon in Reeves Cave. Total thickness of the Spar Mountain Member in Triple J was not determined.

SUMMARY OF CORRELATIONS

Correlation of the reference cavern sections in the Garrison Chapel karst area to the units of Indiana out-crops is established by recognition of the Indian Creek Beds in the lower Levias Member which are recognized twenty miles away at the Cataracts on Mill Creek where MALOTT (1946) subdivided the Ste. Genevieve Formation in Indiana. Additionally MALOTT (1952, p. 57) described the lithographic beds now named Indian
Creek Beds where they appear in the "old tunnel section" three miles east of the Garrison Chapel area.

The entire stratigraphic interval of the Spar Mountain Member was measured downward from the base of the Indian Creek Beds in Shaft, Grotto, and Buckner revealing an average thickness of 7.3 m., with a maximum of 10.4 m. in Shaft and a minimum of 4.9 m. in Buckner.

The upper portion of the Levias Member, the oolitic and bioclastic beds, overlying the Indian Creek Beds was measured in all five sections averaging 6.7 m. except for a thickness of 8.3 m. in Triple J entrance which is interpreted as the center of a convex oolite bar.

The upper sandstone near the top of Shaft Cave entrance is 7 m. above a shaley bed interpreted as the Popcorn Bed of Swann. The intervening oolitic limestone is lower Paoli and the sandstone is agreeably equivalent to either the Basin Aux Vases sandstone or the higher Renault sandstone. The micritic limestone above is upper Paoli and confirmed by a thick massive sandstone higher on the slope; the Mooretown of Indiana out-crops or the Bethel of Basin usage. Two kilometers to the west on the Leininger farm 6.8 m. of Bethel sandstone was logged between the Beaver Bend and

![Diagram of Triple J Cave section](image)
Paoli limestone in Indiana Geological Survey Drillhole No. 155. Bethel sandstone in the area is consistently more than 3.6 meters thick in known exposures.

Correlation of the individual beds within the Ste. Genevieve Formation and the Paoli Limestone in the Garrison Chapel area and karst valley is relatively straightforward with exception to the thin clastic beds within the Paoli Limestone which may be locally discontinuous within the area of an individual cave system.

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Manuscript received: October 1986
DIATOM, CONTRIBUTORS OF CORALLOID SPELEOTHEMS, FROM TOGAWA-SAKAIDANI-DO CAVE IN MIYAZAKI PREFECTURE, CENTRAL KYUSHU, JAPAN

Naruhiko Kashima*, Teruo Irie** and Nobuhiro Kinoshita***

SUMMARY

Coralloid speleothems are commonly distributed in Togawa-Sakaidani-do Cave in Miyazaki Prefecture, Central Kyushu, but their speleological study has not heretofore been achieved. Light and scanning microscopes analyses revealed that coralloid speleothems consist of alternating layers of diatom colonies, detrital minerals and clay. Electron microprobe analysis shows coralloid speleothems to be siliciferous.

This paper assert that diatom (genus Melosira) is one of the important contributors to siliceous coralloid speleothems in the threshold zone at noncalcareous caves.

INTRODUCTION

At the start of this study, Mr. T. IRIE, one of the writers, found the curious coralloid speleothems 2-3 cm in height and 0.5-1 cm in diameter from Togawa-Sakaidani-do Cave. The powder X-ray diffractive pattern of coralloid speleothems contains no mineral peaks. The other hand, this specimen is unreactive to hydrochloric acid.

Recently, the writers visited Togawa-Sakaidani-do Cave for a resurvey of coralloid speleothems, and reexamined by light and scanning microscopes and electron microprobe analysis.

The purpose of this paper is to report on a detailed nature of newly discovered diatomaceous coralloid speleothems from the noncalcareous cave.

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Fig. 1 - Locality map shown in the portion of cave studied. Aso pyroclastic flows in black.

STUDIED CAVE AND CORALLOID SPELEOTHEMS

The geological survey of this area was carried by Metallic Minerals Exploration Agency of Japan in 1967-1968. The stratigraphy of the studied area consists of 1) the mainly Permian strata are made up of slate, sandstone, conglomerate, chert, limestone and greenstones, so-called the Kiurato Togawa facies of southern subbelt of the Chichibu Belt unconformably overlain by 2) the Quaternary Aso pyroclastic flows.

Togawa-Sakaidani-do Cave is located near the south boundary of the Aso pyroclastic flows which is distributed along a small tributary of the Hinokage River, at latitude: 32° 42' 07'' N and longitude: 131° 23' 53'' E (Fig. 1).

Erosive Togawa-Sakaidani-do Cave entirely developed in the lowest part of the Aso pyroclastic flows. The cave entrance lies about 75 m above the water of the Hinokage River and opens the foot of the active water fall that hangs on the edge of the Aso pyroclastic flows.

Coralloid speleothems described here, were distributed on the cave ceiling, walls and on the surfaces of breakdown blocks in the threshold zone and characteristically oriented with respect to the cave entrance.
CORALLOID SPELEOTHEMS

ANALYTICAL RESULTS

In light microscopic observation of the thin sections of coralloid speleothems, they apparently had constructed with concentric different colored layers of diatom colonies, detrital minerals and alternate clay in the growth crust.

By scanning microscopic scrutinies, coralloid speleothems are revealed to be encrustes of diatom colonies. These diatom specimens identified with a genus *Melosira* sp. in their general clindrical morphology, their cell size up to 0.02-0.03 mm wide, and their cell frameworks; surface sculptures and thorny cell terminals (Figs. 2, 3).

Results of electron microprobe analysis of diatom cells are given Table 1. Analysis of 8 specimens were resting yielded mean chemical values of 91.7% for SiO₂, 1.08% for Al₂O₃, 0.17% for FeO and 0.06% for CaO.

CONCLUSIVE REMARKS

The nature and distribution of the flora in the cave environment was described by CUBBON (1976). Summarized discussion has been given to the influence the activity of microorganisms such as bacteria, algae and fungi may exert over the construction of speleothems by HILL and FORTI (1986).

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*Total Fe as FeO
Fig. 2 - Scanning photomicrographs of the surface of coralloid speleothems showing increased magnification views of diatom (*Melosira* sp.) colonies.
Fig. 3 - Scanning photomicrographs of the surface of coralloid speleothems showing increased magnification views of diatom (Melosira sp.) colonies.
To our knowledge, diatom algae speleothems has not been reported from the cave environment. The common occurrence of diatom (genus *Melosira*) in Togawa-Sakaidani-do Cave, merging into concentric layers and covering up coralloid speleothems, suggests that coralloid speleothems growth is triggered, directly or indirectly, by the activities of diatom algae.

At the present stage, it is not possible to attest that the living diatom algae actually distributes on the surface of coralloid speleothems, cave walls and drip water, so that this question must be subjected to future study.

It is to be hoped that the diatom algae speleothems in noncalcareous caves will encourage other researchers to the study of biogenic speleogenesis.

ACKNOWLEDGEMENTS

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LES MÉSAVENTURES DES SOURCES DE L’ESTAVELLE ET DE L’INVERSAC EN LANGUEDOC MÉDITERRANÉEN

Bernard Gèze*

RESUMÉ

L’Estavelle et l’Inversac sont deux sources célèbres du Languedoc méditerranéen (Sud de la France). La première a malheureusement servi de type pour les cavités karstiques alternativement absorbant ou dégorgeant des eaux, suivant les saisons, ce qui n’a jamais été son cas. La seconde peut, au contraire, servir de modèle pour ce fonctionnement alternatif de perte ou d’émergence.

SUMMARY

The Estavelle and the Inversac are two celebrated springs in the mediterranean Languedoc (South of France). Unfortunately, the first one has been chosen as a type for the karstic cavities alternatively absorbing or discharging the waters, in accordance with the season, that had never been the case. On the opposit, the second one can be taken as model for this alternation as swallow hole or emergence.

Les sources de l’Estavelle et de l’Inversac sont citées depuis fort longtemps dans les ouvrages d’”Histoire Naturelle” des provinces méridionales de la France, car leur fonctionnement hydrologique a toujours beaucoup intrigué les observateurs. Elles se trouvent toutes deux dans le département de l’Hérault, à l’Ouest de Montpellier, la première dans la commune de Cabrières, près de l’extrémité orientale du massif ancien dit de la Montagne Noire, la seconde dans la commune de Balaruc-les-Bains, au bord de l’étang de Thau (fig. 1).

Leur description sommaire fera comprendre comment leurs étranges comportements ont provoqué des explications plutôt douteuses et comment l’accumulation de textes mal lus a entraîné d’invraisemblables er-

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Les Monts de Cabrières, le Causse d'Aumelas et la Montagne de la Gardiole sont à dominante karstique.

reurs, fâcheusement répétées par trop d’hydrologues et de karstologues. Mais leurs mésaventures sont aujourd’hui bien terminées, puisqu’aucune ne fonctionne plus!

LA SOURCE DE L’ESTAVELLE

Il semble que ce soit J. Fournet, alors professeur de Géologie à la Faculté des Sciences de Lyon, qui ait le premier, en 1859, attiré l’attention sur l’Estavelle dans son excellent travail sur l’Hydrographie souterraine, particulièrement consacré aux régions “caverneuses” du Jura, mais aussi du Bas-Languedoc où il avait étudié avec M. Graff, à partir de 1844, le petit bassin houiller de Neffies, ainsi que les terrains paléozoïques qui le dominent dans ce que nous appelons maintenant les Monts de Cabrières, partie de la Montagne Noire célèbre par ses faunes fossiles allant de l’Ordovicien au Carbonifère et par sa tectonique d’une extrême complication.

C’est 1500 m au Nord-Ouest du village de Cabrières, par conséquent en plein Languedoc méditerranéen et non dans le Jura, que se trouve la
source de l’Estavelle. Suivant les auteurs, son nom (qui s’écrit aussi Esta-
bel) signifierait qu’il y avait une étable auprès d’elle, ou bien viendrait de
l’Occitan “estervel”, qui veut dire tourbillon. Quoi qu’il en soit, la source
débouche sur un énorme dépôt de tuf calcaire (1000 m × 500 m) formant
un plateau en faible pente au-dessus de la vallée de la Boyne, dont le lit,
déjà creusé, a été dévié par lui. D’après l’altitude relative du plateau (40 à
50 m) et le fait que de petites cavités dans le tuf ont été habitées par l’ours
des cavernes, on peut supposer que le début de la construction de cet édifi-
cement remonte à peu près au Quaternaire moyen et que la source a toujours
débité une eau très riche en calcaire.

Or, le point de sortie se localise dans des schistes dinantiens (Viséen),
sur le trajet d’une faille injectée de quartz, et non dans les calcaires et do-
lomies du Dévonien qui encadrent la région, ce qui est déjà curieux. On
est conduit à penser que l’eau a suivi un assez long parcours en profondeur
et n’a pu gagner l’extérieur qu’à la faveur de la faille, qui a servi soit de
barrière, soit de drain.

Mais le plus étrange est que, malgré l’activité déployée forcément par
la source pour construire le plateau de tuf, elle ne débite depuis longtemps
que tout-à-fait rarement. Déjà Fournet notait quelle avait “vomi beau-
coup d’eau” en 1856, mais qu’elle n’avait point fonctionné depuis dix
ans. Personnellement, j’ai entendu parler de périodes de 25 ans sans
qu’elle ait craché. La notice de la carte géologique de la France à 1/50.000

![Diagramme](image.png)

Fig. 2 - Coupe des chevauchements et séries renversées au voisinage de la source de l’Estavelle
(l’échelle des hauteurs est double de celle des longueurs).

- Σ = schistes ordoviciens; d = calcaires et dolomies du Dévonien; h¹ = lydiennes du Tournai-
sien; h² = schistes du Viséen (faciès culm); Q = faille quartzifiée; ϕ = surface de chevauche-
ment.

Variantes orthographiques sur les cartes récentes: le Caragnas = les Cayraignasses; Estavelle
= Estabel; Bissous = Vissou.
B. GÈZE

(Feuille Lodève, datant de 1982) précise: "l’Estabel est une source temporaire ne coulant que très exceptionnellement; son débit peut alors atteindre, en quelques jours, 1 m$^3$/s et il est aussitôt suivi d’une période de tarissement pouvant s’étendre sur plusieurs mois".

Sans être troublé par un comportement aussi étrange, Fournet pense que l’on est en présence de "fontaines caractérisées par deux bouches en quelque sorte jumelles ... Dans tous les cas, la destination des unes, qui sont habituellement à sec, est de servir à l’évacuation du trop-plein des cavernes, du moment où l’orifice, dont l’écoulement est permanent, devient insuffisant part suite de l’exubérance des eaux ... Dans le Languedoc, ces bouches supplémentaires sont désignées sous le nom d’Estavelles, dénomination que j’ai jugé à propos de généraliser, en l’appliquant à tous les pertuis du même ordre, disséminés dans les autres contrées".

Quelques pages plus loin dans son mémoire, Fournet étudie des sources proches de Porrentruy, dans les calcaires du Jura franco-suissé et dit qu’en remontant une vallée "aux sources pérennes succède une première estavelle, puis viennent des estavelles d’estavelles, largement espacées, de plus en plus intermittentes, conformément à leurs hauteurs, et il me semble qu’un pareil enchaînement est suffisamment démonstratif pour ne plus rien laisser à désirer à l’égard de la parfaite solidarité de ces divers débouchés".

Il est donc parfaitement clair que, pour Fournet, créateur du terme, une estavelle est une source de trop-plein fonctionnant temporairement au-dessus d’une source pérenne.

Mais a-t-il eu raison de prendre comme modèle d’un tel fonctionnement l’Estavelle type de Cabrières? Ce n’est pas évident pour plusieurs raisons: Au-dessous de sa bouche, il n’existe que de fort petites sources qui ne paraissent pas avoir le moindre rapport avec elle. Ensuite, la rareté du fonctionnement paraît incompatible avec un rôle normal de trop-plein, car les précipitations ne sont tout de même pas négligeables sur les monts de Cabrières où les fortes pluies de printemps et d’automne relèvent habituellement le débit de toutes les sources. Des décades sans débordement d’un trop-plein paraissent invraisemblables, aussi bien qu’un dépôt de tuf sans un écoulement assez régulier.

Depuis une douzaine d’années, on est certain qu’il n’y a plus eu de débordement, mais la raison en est évidente: un forage, profond de 55 m a été exécuté au voisinage du griffon et, dans les fissures des schistes sous la carapace de tuf, a rencontré de l’eau en quantité suffisante pour assurer l’alimentation du village de Cabrières. Cette eau a tous les caractères
d’une eau karstique, ce qui confirme l’hypothèse d’une provenance relativement profonde, mais sans que l’on puisse garantir l’origine dans l’un ou l’autre des paquets de dolomies dévoniennes situés dans des position tectoniques particulièrement compliquées aux alentours. De toutes façons, l’exemple de l’Estavelle a été vraiment mal choisi par Fournet pour désigner de simples trop-pleins.

La fâcheuse histoire du terme “estavelle” ne s’arrête malheureusement pas là. Dans son célèbre ouvrage “Les Abîmes”, E.A. Martel cite un peu en vrac des exemples de trop-pleins par lesquels “l’eau interne s’extravase” et a la mauvaise idée de mentionner les “Estavelles de Porrentruy” (puisées dans Fournet) en même temps que le “lac de Zirknitz en Carniole”, dont il précise bien, quelques chapitres plus loin, le fonctionnement alternatif, rempli ou vidé par les cavités y aboutissant.

C’est alors que Jovan Cvijic, après avoir lu un peu vite les textes antérieurs, écrit: “dans les profondes dépressions karstiques il y a des fissures et des avens qui fonctionnent alternativement comme sources ou comme gouffres absorbants ou ponors; j’ai proposé de les appeler “estavelles” en utilisant la dénomination que Fournet a donné à une source semblable”.

Cette désastreuse initiative explique pourquoi le terme est largement employé dans la littérature géographique de langues slaves, d’où elle nous est revenue au point que Marjorie Sweeting écrit: “The term estavelle was first used in the Jura but is now used fairly widely for a hole which is at one time of the year a swallow hole and another time a spring” et que le lexique de J. Margat, publié pourtant en France, donne pour estavelle la définition: “orifice, en terrain karstique, alternativement absorbant ou émissif, selon les saisons”.

En définitive, on se trouve en présence d’un terme mal choisi par Fournet pour signifier trop-plein temporaire, reproduit avec une erreur géographique par Martel et mal compris au sens hydrologique par Cvijic. Il faut donc rejeter formellement “estavelle” hors de la littérature géographique et hydrologique.

Cependant, son succès prouve d’abord que l’on a besoin d’un terme simple pour désigner une cavité karstique à fonctionnement hydrique alternatif, dite parfois “perte-émergence”, ensuite qu’aucun pays ne dispose de ce terme dans sa langue nationale. La proposition que nous allons faire nécessite l’étude de la deuxième source à laquelle nous avons fait allusion en introduction.
LA SOURCE DE L'INVERSAC

Cette source a fait l'objet d'innombrables descriptions. Il semble que sa première étude tant soit peu sérieuse soit celle d'Astruc, dans son "Mémoire pour l'Histoire Naturelle de la Province du Languedoc", datant de 1740. On pourrait encore citer les observations de Marcel de Serres et Louis Figuier, publiées à Montpellier en 1848, et qui ont l'avantage de réduire la part de l'imagination dans l'explication d'un fonctionnement hydrique à première vue surprenant.

Son nom d'Inversac provient peut-être de l'Occitan "Enversar", qui signifie renverser, mais la plupart des auteurs le font carrément remonter au Latin "Inversae aquae", les eaux qui s'inversent, ce qui est en somme une bonne définition du phénomène anciennement observé dans la source, comme nous allons voir ci-après.

Elle se localise 1500 m à l'Est de Balaruc-les-Bains, dans le quartier dit aujourd'hui Balaruc-lès-Usines, 3 km au Nord de la ville de Sète, en bordure de l'étang (ou bassin) de Thau, qui est au niveau de la Méditerranée et en relation directe avec elle. Son exutoire est dans les calcaires miocènes, au contact de l'extrémité des calcaires jurassiques qui forment le chaînon de la Gardiole entre Montpellier et Sète. Le bassin émissif débouchait autrefois sur le minuscule ruisseau de Colobres, qui n'avait guère qu'une vingtaine de mètres de longueur entre rocher et rivage de l'étang, le tout à la cote zéro, non troublée ici par des marées.

Actuellement, un chenal maritime et de larges quais édifiés devant une usine qui a capté l'eau de la source, l'ont définitivement séparée de l'étang, mais le fonctionnement hydrique antérieur avait été sérieusement

![Diagram](Fig. 3 - Schéma du fonctionnement hydrologique de l'Inversac.

\[ j = \text{calcaires et dolomies jurassiques; } m = \text{calcaires molassiques miocènes.} \]
étudié. Les anciens auteurs disaient qu’après un écoulement normal vers l’extérieur pendant l’hiver, au commencement de l’été, les eaux de l’étang se précipitaient sous terre avec une telle impétuosité que l’on entendait à une distance assez éloignée le bruit de leur engouffrement! Même en faisant la part de l’exagération, on doit reconnaître la réalité du phénomène puisqu’en 1925, à la suite de trois années de sécheresse persistante, l’étang se déversait dans l’Inversac à raison d’environ 10.000 mètres cubes par jour. L’eau de l’étang titrait alors 36 grammes de chlorure de sodium par litre; l’eau de la source ne renfermait pourtant que 10 grammes en septembre et atteignait en octobre le maximum de 23 grammes seulement. De là vint l’idée que, même lorsque toutes les sources du pays étaient asséchées, ou coulaient si peu que la consommation devait être sévèrement réglementée, il arrivaît encore à l’Inversac une quantité d’eau douce suffisamment importante pour en tirer un utile parti.

Le problème était d’éliminer l’eau salée absolument impropre aux emplois industriels. On se mit à rechercher en amont de la source une cavité dont on supposait l’existence après un vieux plan daté du 9 octobre 1894. On découvrit en effet, à 4 m, 50 au-dessous du sol, une petite salle, haute de 1 m, 50, large au plus de 6 mètres et entièrement occupée par l’eau calme, profonde de 5 mètres par endroits (descendant donc plus bas que le niveau de la mer). En outre, les travaux produisirent un providentiel effondrement de rochers qui, croit-on, obstruèrent partiellement les conduits par lesquels se faisait l’absorption des eaux salées vers la profondeur. De fait, à partir de ce jour, la teneur en sel marin de la source ne cessa de décroître jusqu’à 3 g, 8 au maximum, ce qui ne gênait plus dans la pratique.

Les quelques variations que l’on a observées par la suite continuaient à être curieusement causées par la pluie et le vent. En effet, la moindre pluie provoque une augmentation de débit qui, brassant sans doute les eaux salées préalablement engouffrées, entraîne une remontée du sel, contrairement à ce que l’on aurait pu imaginer. De même, suivant que le “mistral” (vent du NW) ou le “marin” (vent du SE) soufflent sur l’étang, celui-ci subit une sorte de crue ou au contraire de dessèchemen au voisinage de la source, qui reçoit alors des infiltrations salées plus ou moins abondantes.

En juillet 1937, les conditions naturelles n’étaient encore qu’assez peu modifiées lorsque j’ai pu observer l’étrange phénomène de l’inversion du courant qui a donné son nom à l’émergence. Le flot d’eau salée pénétrant dans le ruisseau de Colobres refoulait d’abord l’eau douce qui conti-
nuait à sortir en s'élevant progressivement. Ensuite, l'eau de source gagnait l'extérieur en coulant en surface, tandis que l'eau salée courbait les algues vertes de la profondeur et se précipitait en sens inverse vers les fissures du rocher dans lesquelles elle s'engloutissait.

Bien entendu, comme mes prédécesseurs, j'ai imaginé d'ingénieux mécanismes pour expliquer ce phénomène, notamment une liaison possible avec les sources thermales de Balaruc-les-Bains (température atteignant 47°), ainsi qu'avec la source de l'Abyss (ou de la Vise) qui jaillit violemment au fond de l'étang de Thau à 3 km de l'Inversac. Mais depuis, les études sérieuses faites en Provence à la source sous-marine de Port-Miou, puis à la Roubine de Vic-la-Gardiole non loin de l'Inversac, tout aussi bien que celles consacrées aux célèbres "moulins de la mer" d'Argostoli et à plusieurs autres pertes-emergences du littoral hellénique, ont montré qu'il s'agissait d'un fonctionnement hydrique assez normal, qui s'explique simplement par la différence de densité entre l'eau douce et l'eau de mer, tandis que leur remontée commune peut se produire n'importe où vers le large.

CONCLUSION GÉNÉRALE

Bien que le cas qui vient d'être considéré soit un peu particulier et que le fonctionnement ait cessé d'être observable, le nom de la source paraît tellement favorable qu'il y a probablement intérêt à en généraliser l'emploi. Je reprends donc les termes d'une note publiée en 1971 dans Spelunca; elle servira de conclusion à cette petite révision historique des "mésaventures" de deux sources célèbres:

Un INVERSAC est une cavité karstique alternativement absorbante ou émissive, selon les saison. L'inversac peut être horizontal ou vertical, pénétrable ou impénétrable et fonctionner d'une façon pérenne ou temporaire. L'inversion des eaux, qui explique le nom de son prototype languedocien, peut avoir des causes variées qui n'interviennent pas dans la définition. Le terme d'inversac doit remplacer celui d'estavelle qui était employé a tort.
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RELATIONSHIPS BETWEEN THE INTERNAL AND EXTERNAL EVOLUTION OF THE MONTE CUCCO KARST COMPLEX. UMBRIA, CENTRAL ITALY.

Fausto Guzzetti

SUMMARY

The relationships between the internal and external evolution of the Mt. Cucco karst complex are studied.

A classic set of equations, involving the oxidation of hydrogen sulphide, originated at depth in an evaporitic formation, is used to explain the presence of massive gypsum deposits in the Mt. Cucco and the Faggeto Tondo caves.

The distribution and the morphology of more than 30 caves in the system, the presence of gypsum, always located along faults, and the presence of broken stalactites and columns, suggest that the evolution of the karst system has been controlled by tectonic movements.

Relationships between the development of the caves and the geomorphic evolution of the area are proposed.

INTRODUCTION

The purpose of this report is to relate knowledge about the location, shape, internal deposits and relative age of different levels of caves to the geomorphic and tectonic history of the area in order to understand the origin and evolution of the Mt. Cucco karst complex. Since the karst complex, as it appears today, is the result of a long sequence of tectonic, geomorphic and hydrologic events, understanding its origin and development is important in understanding the geomorphic evolution of a whole section of the Apennines.

The karst complex is located in the northern sector of the Umbria-Marches fold belt, in central Italy. It is developed on the western limb of one of the major anticlines that, in an en échelon pattern, form the anticlinorium known as the “Ruga Interna” (SCARSELLA, 1951).

The complex is developed almost entirely in the Calcare Massiccio Formation (Upper Lias), a platform limestone that is the lowest outcrop-
ping formation of the Umbria-Marches stratigraphic sequence (see Fig. 1 and Fig. 2 I, II). As pointed out by PASSERI (1972) caves are formed in the Calcare Massiccio limestone because of its high original porosity and permeability. Caves rarely develop in the overlying limestones, marls and shales because these rocks have a low, or very low, permeability, and a low secondary porosity of tectonic origin (PASSERI, 1972).

THE KARST COMPLEX

The Mt. Cucco karst complex, is developed almost entirely underground despite the presence of the deep Rio Freddo canyon, few superficial caves, few sinkholes and karrens. More than thirty caves are present, ranging in length from a few meters to several kilometers. Two in particular are important in understanding the geomorphic and hydrologic history of the area and will be discussed in this paper: the Mt. Cucco cave and the Faggeto Tondo cave.

The Mt. Cucco cave, more than 25 kilometers long and 922 meters deep, represents one of the most impressive cave complex in Italy. The cave, as shown in a Southwest-Northeast vertical cross section of the area (see Fig. 2 III), consists of several sets, or levels, of galleries connected by deep pits. The galleries dip gently westward along the western limb of the anticline. They have been formed along zones of higher synsedimentary porosity, developed between oolites, oncites and birdseyes. The galleries in each level are mostly sub-parallel, few are anastomized, with a rounded, cylindrical shape suggesting a phreatic origin. Only a few of the existing galleries are now accessible by cavers; most of them are plugged with mud or flowstones. Their existence can be inferred from the morphology and the distribution of the other known galleries in the cave.

The absence of any sign of gully erosion on the gallery’s floors shows that the phreatic levels have not been affected by any important vadose erosive process. This fact, as will be shown later, is important in understanding the evolution of the entire karst complex.

The different levels of the cave are connected by pits developed mainly along major tectonic discontinuities such as faults and joint systems. These pits are never older than the sub-horizontal levels that they intersect. Good examples of pits intersecting older sub-horizontal levels are: the Saracco pit, which is developed along a fault or a set of joints and intersects the galleries coming from the Staffa region and going to the Barba-
Fig. 1 - The Mt. Cucco karst area. The gray pattern represents the area where the Calcare Massiccio and the Bugarone formations (Condensed Sequence) outcrop.
ri area; the Gitzmo and Px pits that intersect three different levels; and the Bicco, Berro and Torino pits that, having formed along an East-dipping fault, cut the galleries coming from the Orco gallery and going to the Italian area (see Fig. 2 III).

The shape and the spatial distribution of the galleries suggest that two different drainage systems have been active in the Mt. Cucco cave; an older, entirely phreatic system and a younger, mostly vadose one. The old phreatic system is very complex. It has been active for a long period of time forming different levels of galleries and pits. Today the system is entirely obsolete; the galleries are dry and, locally, partially filled or completely plugged by mud, boulders or flowstones.

The presence of several small caves on the surface, at the projected extensions of most of the phreatic levels, suggests that the old galleries have been in direct communication with the surface, the actual obstruction being only a temporal and later event. Some small caves, representing the paleo-entrances * of the system, are present on the surface only a few meters or tens of meters from the Terminale gallery and the Staffa room, on the eastern side of the Mt. Cucco ridge. Also several small caves are present on the western side of the ridge, at the tip of the Orsara valley.

The shape and the relative position suggest that these caves represent the paleo-springs of the cave complex. The trend of the phreatic galleries and the presence on the surface of both paleo-entrances and paleo-springs suggests that some, or even all, of the phreatic levels presently known can be considered as a former hydrological tunnel that cuts across the entire ridge.

The new drainage system is much different and completely independent from the old phreatic one. It is developed completely under vadose conditions along tectonic discontinuities. The Meandrino gallery (see Fig. 2 III), the longest known section of the vadose system, is a sequence of characteristically narrow, meandering galleries and small pits shaped by the retrogressive erosion of waterfalls. Its overall shape is vaguely hyperbolic, suggestive of the equilibrium profile of an external river; it is completely different from the almost linear, structurally controlled trend of the old phreatic levels.

The Faggeto Tondo cave, 1.5 kilometers long and 350 meters deep, is far less complex than the Mt. Cucco one. The cave developed during at

* The term paleo-entrances is used to mean paleo-entrances to the phreatic system.
Fig. 2 - I) Southwest-Northeast geologic cross-section across the Mt. Cucco anticline.

II) Stratigraphic sequences present in the area. 1 = Calcare Massiccio, 2 = Corniola, 3a = Calcare Diasprigni, Calcare and Marne del Sentino, 3b = Bugarone, 4 = Maiolica, 5 = Scisti a Fucoidi, 6 = Scaglia Bianca and Scaglia Rossa, 7 = Scaglia Variegata and Scaglia Cinerea, 8 = Bisciaro and Schilier, 9 = Quaternary alluvial sediments.

III) Vertical cross section of the major caves projected on a Southwest-Northeast topographic section across the karst complex.
least two different periods. The upper level of the cave, in the upper section of the Calcare Massiccio Formation, represents the first stage of development, in a phreatic environment. This part of the cave is no more than 20 or 30 meters below the topographic surface. A later event resulted in the formation of deep sub-vertical pits along tectonic discontinuities, always in a phreatic environment as suggested by their morphology. Unlike the Mt. Cucco cave, the new drainage system in the Faggeto Tondo cave follows the old galleries, eroding gullies and small ravines on their floors.

At the present time almost all of the karst complex is under vadose conditions. All the internal streams show a characteristic seasonal behavior with very low runoff during the summer and high or very high runoff during the rainy season. The only part of the karst complex still under phreatic conditions is probably the deepest part of the complex, approximately at the level of the Scirca springs, the only known springs for the entire karst complex (BERTOLUCCI ET AL., 1975; BOILA ET AL., 1983; DRAGONI ET AL., 1982). Only a few and local perched aquifers are present inside the caves in form of small lakes and siphons. This is typical of the sub-surface karst hydrology.

THE ORIGIN OF GYPSUM DEPOSITS

A peculiar characteristic of the Mt. Cucco karst system is the presence of gypsum inside the caves. Gypsum is present both inside the Mt. Cucco cave and in the upper level of the Faggeto Tondo cave. The deposits are located everywhere near faults or joints, along two levels. The total volume of the gypsum can be estimated as slightly in excess of $1 \times 10^3 \text{ m}^3$.

The presence of abundant gypsum deposits is peculiar and raises a number of questions. How can gypsum, a calcium sulphate with a solubility product higher than that of calcium carbonate ($K_{sp} \text{CaCO}_3 = 4.2 \times 10^{-7} \text{ mol/l}, T = 20^\circ\text{C}; K_{sp} \text{CaSO}_4 \cdot 2\text{H}_2\text{O} = 2.3 \times 10^{-4} \text{mol/l}, T = 25^\circ\text{C}$) precipitate inside a cave? How can it be present at more than 1300 meters of elevation inside a cave when the nearest external gypsum deposit is more than 10 km to the East and at no more than 450 meters of altitude (SERVIZIO GEOLOGICO D’ITALIA, 1979)? Moreover, where does all this gypsum come from? Does any connection exist between gypsum deposits and the cave development?

The Mt. Cucco karst complex is not the only one in the world that contains gypsum. In the United States, several caves contain gypsum de-
posits: the Carlsbad caverns in the Guadaloupe Mountains of southeastern New Mexico, the Crevice cave in Iowa and the Kane cave in Wyoming are some of them.

A number of hypotheses have been proposed to explain the origin of the massive gypsum deposits inside these caves. In particular, two different groups of hypotheses are available in the literature. A traditional group considers gypsum as a deposit from sulfate-laden waters during a secondary event in the cave history (J.H. BRETZ, 1949; L. HOBBERG, 1949; GOOD, 1957). A second group considers gypsum as the result of the replacement of the calcium carbonate by sulfate, or the solution of calcium carbonate by a mechanism involving sulfate precipitation. The deposition of gypsum is thought to be contemporaneous to the cave development. At least three different mechanisms have been proposed:

MOREHOUSE (1968) and JAGONOW (1978) proposed a mechanism involving the oxidation of pyrite (FeS$_2$) and the production of sulfuric acid (HOWARD, 1960). The chemistry can be summarized as follows:

$$4\text{FeS}_2 + 15\text{O}_2 + 14\text{H}_2\text{O} + 16\text{CaCO}_3 \rightarrow 8\text{CaSO}_4 + 2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O} + 8\text{Ca}^{++} + 16\text{HCO}_3^-$$

PALMER ET AL. (1977) suggested a large-scale gypsum replacement of carbonate in the deep phreatic zone, following the reaction:

$$2\text{H}_2\text{O} + \text{CaCO}_3 + \text{SO}_4^{--} \rightarrow \text{CO}_3^{--} + \text{CaSO}_4 \cdot 2\text{H}_2\text{O}$$

Finally EGEMEIER (1973) proposed a 3-step gypsum replacement mechanism acting near the air-water interface and based on a sulphuric acid reaction produced by oxidation of hydrogen sulfide in the cave forming water. The chemical equilibria are as follows:

$$2\text{H}_2\text{S} + \text{O}_2 \rightarrow 2\text{S} + 2\text{H}_2\text{O}$$

$$2\text{S} + 2\text{H}_2\text{O} + 3\text{O}_2 \rightarrow 2\text{HSO}_4^- + 2\text{H}^+$$

$$\text{HSO}_4^- + \text{H}^+ + \text{H}_2\text{O} + \text{CaCO}_3 \rightarrow \text{CaSO}_4 \cdot 2\text{H}_2\text{O} + \text{CO}_2$$
DAVIS (1980) suggested that hydrogen sulfide could originate at depth in oil-bearing evaporitic rocks where conditions were anaerobic. The reaction, with or without the presence of sulfur-reducing bacteria (KIRK-LAND & EVANS, 1976; DAZY & GRILLOT, 1982), is as follows:

\[
\text{CaSO}_4 + \text{CH}_4 (+ \text{Bacteria}) \rightarrow \text{H}_2\text{S} + \text{H}_2\text{O} + \text{CaCO}_3 + \text{Energy} \quad (4)
\]

DAVIS (1980) proposed that the hydrogen sulfide originated via the reaction (4) and then ascended in solution along faults to the cave level. On becoming aerated at the water-air interface, the hydrogen sulfide oxidized to sulphur and/or sulphuric acid which, on encountering limestone, reversed the replacement process and converted calcite to gypsum according to the Egemeier proposal (reactions (3)).

In the Mt. Cucco cave, three deposits of massive gypsum are known to date. They are located in the first part of the Barbary gallery, in the final section of the Burella gallery and in the Orco gallery (Fig. n. 2 III). In the Barbari and Orco galleries, gypsum seems to be laying on the cave floor whereas in the Burella gallery gypsum seems to be part of the cave floor.

The amount of gypsum is much greater in the Faggeto Tondo cave, despite its small size. Gypsum is spread along the upper part of the cave and, as in the Mt. Cucco cave, it is located near to or downslope from faults or major joints.

These sulphate deposits have not originated by precipitation from standing waters, either in a vadose or phreatic environment. They are scattered and too far away from any possible paleo-entrance of the system. If they originated from cooling waters coming from outside the caves, the deposition should have taken place in the part of the caves closest to the surface, not at depth. Moreover, the only known external source of gypsum is the Gessoso Solfitifera Formation (Messinian) with the closest outcrops more than 10 km to the East and at only 450 meters of elevation (SERVIZIO GEOLOGICO D’ITALIA, 1979). It is difficult to think of a mechanism that could transport gypsum for such a long distance and across a major ridge.

Similarly, the gypsum is probably not derived from pyrite oxidation via the mechanism proposed by Morehouse and Jagonow. The Umbria-Marches sedimentary sequence does not contain pyrite or any other sulfide in sufficient quantities for the process to take place.

The presence of gypsum in the area can probably be explained by the Egemeier-Davis proposal. A 2000 meter thick sequence of anhydrite and dolomite at depth under the Calcare Massiccio Formation could be the source for the hydrogen sulphide. This sequence, the Anidridi di Burano
MONTE CUCCO KARST COMPLEX

Formation, never outcrops, but it has been drilled for more than 1000 meters of thickness 20 kilometers Northwest of the studied area and it is commonly believed to underlie the Calcare Massiccio Formation in the Umbria-Marches area. A very small amount of hydrocarbon in the anhydrite could generate hydrogen sulphide (Kirkland & Evans, 1976; Dazy & Grillot, 1982) which could rise in water along normal faults to the level where it could be mixed with fresh water, causing in turn the oxidation of hydrogen sulphide, the production of sulphuric acid and the replacement of limestone by gypsum (Egemeyer, 1973; Davis, 1980).

This replacement process, limited in space and time by the quantity of hydrogen sulfide available, could take place both inside already existing caves or outside the caves. Since the solubility of gypsum in water is high and the total volume of the deposits was probably small, only the deposits inside the caves could survive surface weathering and dissolution.

This model can explain why gypsum is always located near faults, the difference in volume among different deposits and their morphological appearance.

TECTONIC AND GEOMORPHIC IMPLICATIONS

The hydrologic conditions in the area have been changed several times during the formation of the caves. Each level of galleries in the Mt. Cucco cave can be considered as an old and independent hydrological level, suggesting that the karst complex has not been developed all at the same time but instead through several steps, at least one for each main level of galleries.

Most of the galleries have been formed under phreatic conditions; both the rounded shape and the linear trend support this idea. Moreover the gallery’s floors do not show evidence of later gully erosion, suggesting that no important vadose event ever took place inside the caves. The transition from an old phreatic level to a deeper, new one has always been fast enough not to leave ravines or canyons.

Only a few galleries in the Mt. Cucco cave have formed under vadose conditions. These galleries represent the actual underground drainage system. They generally follow tectonic discontinuities and they are everywhere younger than the old phreatic levels that they intersect.

In order to understand the genesis and development of the karst complex, two different models of evolution of the Mt. Cucco cave have been
prepared by the author (Fig. 3). The first model (A) is based on the hypothesis that the discharge area drops while the recharge area remains at a relatively constant, higher level. The fall of the discharge area can be the result of the erosion of an impermeable layer (i.e. Bungarone or Scisti a Fucoidi Formations) as well as the down-dropping action of an active normal fault, or even a combination of both mechanisms. Which mechanism has been active is not important from a speleogenetic point of view; the erosive mechanism being probably only more gradual than the tectonic one.

The model shows that each time the discharge area drops, the active level of phreatic galleries is suddenly abandoned in favor of a deeper, new one. The model, however, cannot explain the presence of some galleries, such those in the Staffa area.

The Staffa area is characterized by phreatic and anastomized galleries, close to the surface at 1300 meters of elevation. The origin of these galleries and the presence of paleo-entrances at their projections to the surface can only be explained by assuming that the recharge area, as well as the discharge area, drops in elevation. This is the hypothesis used to prepare the second model (B) of figure 3. The recharge and discharge areas probably did not always drop simultaneously or with the same amount of offset. The differential drop can explain why the phreatic levels are not perfectly parallel.

In the second model (B) (Fig. n. 3), the drop of both the recharge and discharge areas is thought to be due to the action of normal faults. The evidence that gypsum deposits are always located near tectonic discontinuities supports the idea of an important tectonic activity and suggests that the rise of sulphate-laden water along normal faults can be related to hydrothermal pumping, caused by an unusually intense tectonic activity in the area. In this view, each main level of sub-horizontal galleries represents a period of relative quiet in between periods of intense seismic and tectonic activity that down-dropped the recharge and/or the discharge areas, and eventually resulted in the deposition of gypsum.

The presence in the upper level of the Mt. Cucco cave (Cattedrale room and Galleria Terminale) of several broken stalagmites and columns suggests that the area has been affected by recent tectonic activity (this section of the Apennines is currently active). The dripstones do not seem to be broken for instability reasons (i.e. overweight). Columns up to 1.5 meters in diameter have been broken and shifted several centimeters along horizontal planes. Broken and non linear stalagmites have been observed in other caves of the Umbria-Marches area (FORTI ET AL., 1983; FORTI &
Fig. 3 - Two different models for the evolution of the Mt. Cucco karst system.
Model A is based on the hypothesis that the discharge area drops while the recharge area remains at a relatively constant, higher level. The fall of the discharge area is the result of the down-dropping action of a normal fault. This model cannot explain the presence of the galleries in the Staffa area.
Model B is based on the assumption that both the recharge and discharge areas drop because of the down-dropping effect of normal faults. This model can explain the origin and evolution of all the known galleries in the cave system. The discharge and the recharge areas are only outlined. Their real position during each step may be different from what shown in the figure.
POSTPISCHL, 1984). These authors proposed that the dripstones were tilted, and eventually broken by neo-tectonic movements. The presence of broken or non linear stalagmites and columns cannot be related to any particular tectonic event, but it suggests that the area has been recently active.

The history of the Mt. Cucco karst complex is related to the tectonic and geomorphic evolution of the entire area. The karst complex, as it is today, is the final product of a complex series of events that changed the hydrology of the area several times.

The active hydrologic condition, even if not completely clear, is relatively simple. The watershed strikes North following the highest ridges and only in a few places it is slightly shifted. Where a major river (i.e. the Sentino river) cuts across the entire anticlinorium, the divide is shifted to the west and water is diverted from the Tyrrhenian side to the Adriatic side of the watershed. Where an underground, structurally controlled, drainage system is acting, as in the Mt. Cucco area, water is diverted to the West and the divide is shifted to the East.

The present situation is probably, in many aspect, similar to those in the past during periods of relative tectonic quiescence. During those periods, a well-developed underground drainage system could slowly form along the west-dipping strata of the Calcare Massiccio, diverting an increasing amount of water from East to West. The intermediate periods of intense tectonic activity resulted in a rapid change of the quiescent situations, causing a relative uplift of the anticlinorium. In the karst complex, the current phreatic levels were suddenly abandoned in favor of new, deeper ones; while the Sentino river, because of the relative uplift of the anticlinorium, lost the upper part of its water basin. The result was a more linear watershed, with no, or few, deviations from the highest ridge line. At the end of each tectonically active period, the entire system gradually adjusted to the new conditions.

FINAL REMARKS

More work needs to be done to fully understand the origin and evolution of the Mt. Cucco karst complex; nevertheless a few conclusion can be drawn:

— The karst complex has not formed all at one time, but instead it is the result of the tectonic, hydrologic and geomorphic evolution of the entire area.
— The gypsum deposits present inside the caves are the result of the reaction of sulphuric acid, derived from the oxidation of hydrogen sulfide with limestone. The hydrogen sulfide originated at depth in the Anidridi di Burano Formation and rose to the caves level along normal faults.

— The tectonic behaviour of the Mt. Cucco area, as suggested by the karst complex development, can perhaps be related to a stick and slip movement along major local faults.

— If gypsum is dated at least one of the tectonic events that took place in the Umbria-Marches Apennines can also be dated.

The model proposed for the evolution of the studied karst complex does not take into account any major change in the climatic conditions of the area. If the climate and, in particular, the amount of rain changed very much during the development of the cave, as it is probably the case, the evolution of the entire complex would be much more complex than that predicted by the model.

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CARBONATE SURFACE SOLUTION
IN THE CLASSICAL KARST

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RIASSUNTO

Le ricerche attualmente in corso sulla dissoluzione delle rocce carbonatiche sul Carso di Trieste indicano che l’abbassamento medio delle superfici esposte agli agenti atmosferici è di 0.028 mm/anno con una piovosità media di 1350 mm. I valori massimi (0.031 mm/anno) competono ai calcari microcristallini, quelli minimi (0.014 mm/anno) alle dolomie.

SUMMARY

The current research on the dissolution of carbonate rocks in the Karst of Trieste indicates that the average degradation of surfaces exposed to atmospheric agents is 0.028 mm/year with an average rainfall of 1350 mm. The maximum levels (0.031 mm/year) correspond to micro-crystalline limestones, the minimum values (0.014 mm/year) to dolomites.

FOREWORD

For 9 years, at eight experimental stations located in the Karst of Trieste (see Fig. 1), direct measurements have been taken on the degradation of the karst surface exposed to atmospheric agents.

The collection of data, performed every six months, relies on a special

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micrometric instrument\(^{(2)}\) resting on stainless-steel nails driven into the rock (FORTI, 1981; see Fig. 2 and 3).

The preliminary results of the research have been reported on several occasions (see enclosed bibliography): the aim of this paper is to illustrate the available results graphically, to discuss the meaning of the measurements briefly and to compare data with the results available in the relevant literature, despite the different methods of data collection used.

MEASUREMENT STATIONS

Station nr. 1 is situated near the Grotta Gigante, next to the weather station chosen (see CLIMATIC FEATURES). It lies on an outcrop of Cretaceous fossiliferous limestone (wackestone sensu Dunham, bimicrite sensu Folk) with a 15° slope. From the mineralogical point of view, the rock is made up of Calcite (92.5%), Dolomite (2.4%) with an insoluble residue of 5.1%.

![Map of measurement stations](image)

Fig. 1 - Location of measurement stations; the weather station is located next to measurement station nr. 1.

\(^{(2)}\) The instrument is similar to that one developed in Bristol (HIGH & HANNA, 1970).
Station nr. 2 can be found in the surroundings of Borgo Grotta Gigante, standing on an outcrop of Cretaceous fossiliferous limestone (wackestone sensu Dunham, biomicrite sensu Folk) with a 10° slope. From the mineralogical point of view, the rock is composed of Calcite (91.0%), Dolomite (2.5%) with an insoluble residue of 6.4%.

Station nr. 3 is just a few centimeters away from station nr. 2 (and therefore lies on the same lithotype), but in the middle of a “‘kamenitza’” 15 cm in diameter.

Station nr. 4 is located on the flank of a vast doline, on an outcrop of Cretaceous fossiliferous limestone with a 17° slope. The rock has the same petrographic features as those of stations nrs. 2 and 3, with Calcite (92.6%), Dolomite (2.4%) and 5.0% of insoluble residue.

Station nr. 5 is situated near the conurbation of Opicina on an outcrop (see Fig. 2) of Cretaceous limestone (wackestone sensu Dunham, intrabiomicrite sensu Folk) which, after mineralogical analysis, have resulted to contain Calcite (91.9%), Dolomite (2.8%) and 5.3% of insoluble residue.

Station nr. 6 is located in a small valley on Cretaceous dolomites (anedral dolomite from biomicrite recrystallization) which, after chemical analysis, have shown to be made up of Dolomite (85.6%), Calcite (10.1%) and 4.3% of insoluble residue.

Station nr. 7 can be found on the eastern slopes of a ridge (Mount Lanaro) on an outcrop (see Fig. 3) of Cretaceous limestone (mudstone sensu Dunham, fossiliferous micrite sensu Folk) which, after chemical analysis, turned out to be composed of Calcite (91.9%), Dolomite (2.8%) and 5.3% of insoluble residue.

Station nr. 8 is located on the northern side of Mount Carso on the right handside of the Val Rosandra, on a subhorizontal outcrop of Palaeocene limestone (packestone sensu Dunham, recrystallized biomicrite sensu Folk) which are mainly made up of Calcite (93.0%) and Dolomite (2.1%) with 4.9% of insoluble residue.

CLIMATIC FEATURES

The Karst of Trieste is characterized by a climate which is half way between the “‘Mediterranean type’” and the “‘Continental type’”, with a long and cold winter, an unsettled spring and a hot summer extending over part of the autumn.

An official weather station is located by the entrance to the Grotta Gigante, i.e. very near to the measurement points 1, 2 and 3. The relevant
Fig. 2 - The micrometric instrument and station nr. 5.

Fig. 3 - The micrometric instrument and station nr. 7
observations have been under way for twenty years; the text makes reference to them and from them the daily precipitation data, reported for each of the various measurement stations, are drawn (GASPARO, 1980 ... 1987).

The average rainfall is 1350 mm/year, with daily peaks of 105 mm recorded in autumn, in the month of November. The minimum values are recorded in winter, in the month of February. On average, there are 130 days per year when precipitation occurs, with a remarkable prevalence of rain (rain: 31.5%; snow: 3.0%; hail: 1.1%; none: 64.4%) (TOMMASINI, 1979; GASPARO, 1980 ... 1988).

Humidity generally ranges within rather low values, with maximum values in the late autumn and minimum values in July and August.

The average yearly temperature is 12°C, with minimum and maximum daily temperatures of -15°C and +34°C respectively; during the winter months the average value is 3.5°C, during the summer 19.5°C.

The winds of the 1st Quadrant are the prevailing winds in terms of intensity and prevalence. In particular, the bora (ENE) is the most recurrent wind, with an average frequency of 78 days/year.

RESULTS

The diagrams illustrating the results of the measurements are given below (Figs. 4-11).

The measurements made in the various stations have been processed so as to obtain information on the features of the degradation due to karst phenomena.

It should also be underlined that the extremely limited values of the degradation can easily lead into errors when making measurements or, rather, they can sometimes produce measurements falling short of expectations.

However, all measurements have been transposed into diagrams, as very high degradation values or "accretion" values are all the same meaningful. Furthermore, instrumental or human errors are part of the experience.

During the 8 years of the experiment, 17-18 measurements were made at each station, which made possible to calculate an average 0.028 mm/year degradation of calcareous surfaces. The minimum degradation (0.018 mm/year) was recorded at station nr. 8, located on compact, fossiliferous, sparitic limestone; the maximum degradation (0.031 mm/year) was recorded at station nr. 4, situated on compact, micritic and partially fossiliferous limestone.
In this respect, it should be mentioned that stations nrs. 2, 3 and 4 are located on rocks virtually having the same petrographic and mineralogical features. The absolute minimum degradation measured at station nr. 3 (0.012 mm/year) is to be ascribed to the fact that the station lies on a "kamenitza" and, therefore, on a morphotype with a very limited vertical evolution.

Data referring to station nr. 6, located on dolomitic rocks, confirm the lower liability to karst phenomena of these lithotypes, although the recorded value (0.014 mm/year) is about 2.5 times lower than the values concerning calcareous lithotypes. These results have been corroborated by a recent research (STEFANINI et al., 1985), consisting of a five-year exposure to atmospheric agents of about 40 "bricks" of various carbonate lithotypes.

If the measurements indicate a substantial increase in the degradation with the passing of time, it should also be recognized that when calculating the ratio between degradation (mm) and quantity of precipitation in the period between measurements (mm), the final result was nonetheless unexpected (Fig. 12). In short, the result is that there is no correlation between quantity of fallen precipitation and corresponding degradation, so much so that, for example, univocal degradations do not correspond to fundamentally equal precipitations.

The fact that the experience on the site contrasts with theoretical forecasts can have several explanations: different type and/or chemism of precipitations in the period considered, different average temperature, pressure, etc. Furthermore, and this subject will be dealt with later, there could also be differences in the exposed surfaces, which go unnoticed at a first observation. Preferring one explanation to another, however, seems premature. In the continuation of the experiment, once the data relating to 10 years of homogeneous measurements have been collected for all stations, all possibilities will be considered, also making correlations with stations of regions characterized by climatic features different from those of the Karst of Trieste.

As evidence of the research carried out, the reported diagrams indicate the results of the measurements taken at all stations, including the ratios between overall degradation and overall precipitation.

With regard to the ratios between relative degradation and quantity of precipitation in the period, only a few were chosen to represent non correlability.
Fig. 4 - Station n. 1: relationships between relative degradation (in mm, left Y-axis), rainfall (in mm, right Y-axis) and time of exposure (in days, X-axis).

Fig. 5 - Station n. 2: relationships between relative degradation (in mm, Y-left axis), rainfall (in mm, Y-right axis) and time of exposure (in days, X-axis).
Fig. 6 - Station n. 3: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).

Fig. 7 - Station n. 4: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).
Fig. 8 - Station n. 5: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).

Fig. 9 - Station n. 6: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).
Fig. 10 - Station n. 7: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).

Fig. 11 - Station n. 8: relationships between relative degradation (in mm, Y - left axis), rainfall (in mm, Y - right axis) and time of exposure (in days, X - axis).
FINAL REMARKS

The data assumed indicate that the current ratio of degradation of carbonated surfaces on the Karst of Triest is equal to nearly 2.8 mm every 100 years, implying an average weight loss due to surface erosion and corrosion of $18 \times 10^3$ mg/sq.cm/day.

Such value is about one unit of magnitude higher than the value obtained during the observations on "standard limestone tablets" performed under the aegis of the U.I.S. Commission on Karst Denudation (GAMS, 1985).

The average weight loss of rock samples having the same features and exposed to nearly identical morphological and climatic conditions, namely samples taken in Jugoslavia in the surroundings of Sesana and Divaca, is more or less equal to $2.99 \times 10^3$ mg/sq.cm/day, although samples which had been exposed to atmospheric agents at a height of $+250$ cm from ground level experience a loss ranging from $7.59 \times 10^3$ to $11.32 \times 10^3$ mg/sq.cm/day (GAMS, 1985).

During an observation, similar to the one performed by U.I.S., made on the Karst of Triest by using rectangular tablets having size of $6 \times 3$ cm (STEFANINI et al., 1985), corrosion was reported to cause average weight losses equal to $3.5 \times 10^3$ mg/sq.cm/day, which is very similar to the results arrived at by Gams in the standard limestone tablets experiment.

The values are very similar to the one measured by KUNAVER (1979) on Dachstein limestones of the M. Canin (YU), a high alpine Karst with a total rainfall of $3.500$ mm/year. Kunaver, using a Micro Erosion Meter, reports a mean ratio of degradation $0.035$ mm/year (maximum value $0.102$ mm/year, minimum $0.026$ mm/year).

Other observations, such as the works by CRATBEE and TRUDGILL (1985), have produced values which cannot be correlated owing to the different methods used: in general, the weight loss values are substantially lower than those reported by the Authors.

PERNA and SAURO (1979) reported that, following direct monitoring on Karstic samples in Trentino (Italy), a degradation of calcareous surfaces in Italian alpine areas over the last 1,000 years can be assumed, ranging from $0.2 \times 10^3$ to $35 \times 10^3$ mg/sq.cm/day, although the Authors deem that average values of $10 \times 10^3$ mg/sq.cm/day are more likely.

The "artificial" experiences, such as those performed on tablets, "plaquettes", etc. seem to yield dissolution values in default owing to the non-natural features of the surfaces exposed to atmospheric agents: when
Fig. 13 - Stations n. 1, 5, 6: relationship between absolute degradation (in mm, Y-axis) and rainfall (in mm $\cdot 10^{-3}$, X-axis) relative to the same time period.
samples are prepared in advance surface alterations, biological films, external intercrystalline porosity, etc. are eliminated with the result that the surface obtained substantially slows down the rate and extent of surface degradation.

Therefore "in situ" observations on samples which have not been previously prepared, are likely to yield more reliable results even though it must be acknowledged that the average weight loss values are defined indirectly and by means of methods which would have to be checked and compared.

REFERENCES


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