Cave and Karst Science

The Transactions of the British Cave Research Association

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Cave and Karst Science

Authors are encouraged to submit articles for publication in the Transactions of the *British Cave Research Association* under four broad headings:

1. Papers

Scientific papers, normally up to 6,000 words, on any aspect of karst/speleological science, including archaeology, biology, chemistry, conservation, geology, geomorphology, history, hydrology and physics. Manuscript papers should be of a high standard, and will be subject to peer review by two referees.

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Personal statements, normally up to 1,000 words, on topical issues; discussion of published papers, and book reviews. Where appropriate, statements should put forward an argument and make a case, backed-up by examples used as evidence.

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Manuscripts may be sent to either of the Editors: Dr D J Lowe, c/o British Geological Survey, Keyworth, Nottingham, NG12 5GG, UK, and Professor J Gunn, Limestone Research Group, University of Huddersfield, Queensgate, Huddersfield, HDI 3DH. UK.
Intending authors are welcome to contact the Editors, who will be pleased to advise on manuscript preparation. Enquiries by E-mail are welcomed, to: d.lowe@bcra.org.uk or j.gunn@bcra.org.uk.

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Authors will be provided with 20 reprints of their own contribution, free of charge, for their own use.

If any problems are perceived regarding the nature, content or format of the material, please consult either of the Editors before submitting the manuscript.

Cave and Karst Science

TRANSACTIONS OF THE BRITISH CAVE RESEARCH ASSOCIATION

Volume 32, Number 1

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Papers

Cover photo:

Spectacular mogote karst in the Vinales Valley of eastern Cuba [see the paper by Mario Parise *et al.* in this Issue].

Photograph by Tony Waltham

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EDITORIAL

David Lowe and John Gunn

To start on a positive note, we are very pleased to report that our Editorial in Volume 31, No.3 was unduly pessimistic in that the whole of this Volume of *Cave and Karst Science* will be produced in the "traditional" paper format. BCRA Council have confirmed that there are no immediate plans to end paper production, although a new electronic format will also be introduced at some point. After this positive start we regret that once again we have to apologize to our readers and institutional subscribers for the ongoing delays in the publication of *Cave and Karst Science*. Though the lag against target dates had its beginnings several years ago and was related to events and actions that are no longer relevant, the after-effects of that initial problem are still being felt. Since the original setback it has occasionally seemed that we were beginning to make good the situation, but on each occasion new sets of factors have conspired to move us back to where we were. One issue underlying the ongoing delays is the relative scarcity and unpredictability of flow of publishable material both into our offices and then through the review and editorial systems. During the past two years we have suffered two unprecedented "lean" periods, during which those few authors who did submit manuscripts have seen their work held back as we waited for submission of additional material to complete the content of the planned publication. Added to this, the review process, which necessarily operates on a good-will basis as and when appropriate reviewers can make time available, can also stall on occasions. When there is already a shortage of material within the system, even small delays in the turn-around of manuscripts can cause hold-ups and frustration that seem out of proportion to the initial problem. This is not unlike the situation where one car driver in a stream of traffic travelling at 70 miles an hour touches the brake pedal – whereas that car will barely decelerate, there is a cumulative "knock-on" effect along the stream and, in the worst cases, the stream may stop and/or collisions might happen at the rear of the queue.

This situation that faces us is, in a way, self-perpetuating. Authors who have seen their potentially groundbreaking results and ideas stagnate for several months will, quite rightly, think twice before committing future manuscripts to the same system. But that's not all. Following recent widespread increases in the potential speed of communications, publishing efficiency and output quality, together with increased accessibility to a number of journals that are raising their international profiles (see below), it is now routine for authors of cave and karst science papers to shop around for advantageous routes towards quick, well-reviewed and well-produced inclusion in respected publications. Hopefully this expansion is a sign of good health in the world of cave and karst science research but, inevitably, it also means that many changes to past practice will need to be made.

Along with the editors and editorial boards of several other leading cave and karst journals we hope that we can move forward within an atmosphere of mutual awareness, respect and cooperation, inevitably flavoured with the necessity to compete within this bustling market in order to survive. Many of our readers will be aware of a long-standing interchange of abstracts between *Karstologia* and *Cave and Karst Science*. This mutually beneficial exchange of information will continue in future when the feedstock is available and when page space allows. Additionally, however, and also when page space allows, we have agreed to exchange title information between ourselves and several other leading journals (including the *International Journal of Speleology* and the *Journal of Cave and Karst Studies*). Additionally, title and abstract details from *Cave and Karst Science* and other leading journals will in future be accessible via the *Speleogenesis and Evolution of Karst Aquifers* website [http://www.speleogenesis.info]. These are new initiatives, and the procedures that will make them happen routinely are still being developed and emplaced, but we hope to play our part in upcoming issues of *Cave and Karst Science*.

All being well we will be able to feature such title information in the next issue (32/2) of *Cave and Karst Science*, which is planned to be a thematic issue with papers deriving from the 2005 Tiankeng Investigation Project in China. Tiankeng are giant collapse dolines that have only recently been recognised as a specific landform type and the Tiankeng Investigation Project provided an opportunity for an international group of leading karst scientists to see the finest of the Chinese tiankengs and to discuss their geomorphology in a wider context. Their papers will form the bulk of the Issue, while hopefully leaving a few pages available for inclusion of a Forum section and the reciprocal publicity outlined above. The Tiankeng issue is planned for publication in May 2006 and we currently have papers submitted that will fill at least half of Volume 32 No.3, which we hope to publish in August 2006. However, we are now seeking submission of papers to appear in the issues of Volume 33, at least one of which should be published during 2006.

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Evolution of caves in response to base-level lowering.

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Abstract: Base-level lowering plays an important role in cave passage development and morphology. Commonly cave conduits are formed at depth below the water table, and sub-horizontal conduits can form at depths of more than 100m below the water table. The depth of formation is a function of flow path length and stratal dip. Subsequent base-level lowering is responsible for evolution from a deep phreatic to a shallow phreatic to a vadose, water-table setting, and is accompanied by gradational features such as bypass passages, vadose entrenchment, and undercapture passages. This pattern of evolution expands on the two-cycle hypothesis of Davis (1930), is compatible with a gradual lowering of base level, and is common. Most caves do not evolve to a vadose passage stretching from sink to spring because the flow is redirected by undercaptures, which are at a lower elevation and are usually formed below the water table. Undercaptures frequently form distributary springs and provide much of the complexity seen in cave maps. Distributary springs and bypass passages can also be formed during short-term rises in base level that also produce wall notches.

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INTRODUCTION

Many accounts have been published about the formation of conduits in caves, and in recent decades their relationship to the water table has been linked to external base-level changes (Palmer, 1987), to the abundance of open fractures (Ford and Ewers, 1978), and to stratal dip and flow path length (Worthington, 2001, 2004). The response of conduits to base-level lowering provides a separate but associated issue. Davis (1930) proposed a "two-cycle" hypothesis of cave development, with conduit formation at depth below the water table. Gradual base-level lowering would result in the water table eventually dropping below the level of the conduit, abandonment of the conduit and the formation of a new, lower-elevation conduit. However, Davis gave no details of conduit evolution during these stages, and Ford (1965) was the first to describe gradational features associated with a falling water table, including vadose entrenchment, bypass passages, and (under)captures.

Many cave passages show clear evidence of vadose modification of an earlier phreatic passage; the main stream passage in Ogof Ffynnon Ddu (South Wales) is an excellent example. This has evolved from being predominantly phreatic with conduits to at least 70m below the water table to an almost completely vadose stream passage (Smart and Christopher, 1989). However, recent summaries such as Ford (2000, 2004), Veni (2005) and Palmer (2005) place little emphasis on gradational processes, implying for instance that water-table passages are usually formed from scratch at the level of the related water table rather than possibly evolving from a pre-existing, formerly deeper phreatic conduit.

The following account seeks to explain how passages evolve over time, how grading mechanisms may be recognized in caves, and how significant such grading is in overall cave development.

CONDUIT EVOLUTION IN RESPONSE TO A FALLING BASE LEVEL

In many caves it is possible to identify two main passages types:

i) predominantly vadose passages, where flow is largely downdip along (or incised below) bedding planes and vertically down joints, and

ii) predominantly phreatic (or phreatic with vadose modification) base-level passages, where flow is largely horizontal.

Passages between King Pot and Keld Head (Yorkshire, England) provide an excellent example. The vadose passages follow the stratal dip towards the north, whereas the phreatic base-level passages drain southwards to Keld Head (Figure 1).

The low-viscosity enhancement of flow deep below the water table suggests that caves, especially in catchments longer than about 3km, should commonly be initiated as a single loop at some depth below the water table (Worthington, 2001, 2004). The simplest pattern is where there is a sinking stream flowing via a cave to a spring, as shown in Figure 2. The initial flow is shown as a curving path below the water table, with the depth of flow being a function of stratal dip and flow path length (Worthington, 2001).

Figures 2b to 2d show changes in the cave in response to a steadily falling base level. In Figure 2b, the vadose, upstream part of the cave has increased in length as the water table has dropped, and the crest of one loop has developed an isolated vadose trench (Ford, 1965). Further increases in the length of vadose passages occur in Figures 2c and 2d. New passages may also form. These are of two types, and were named bypass passages and capture passages by Ford (1965), though Palmer (1969) referred to both types as diversion passages. Bypass passages are new, higher passages that form by sedimentation, roof collapse, or a rising base level. Capture passages have also been called phreatic captures (Smart and Christopher, 1989), diversion passages (Ford and Williams, 1989), undercaptures (Jeannin *et al.*, 2000), and tapoff passages (Veni, 2005). They are new, lower passages that form as a result of steepened hydraulic gradients due to base-level lowering. Sediments aggrading on the cave floor can result not only in bypass passages but also in upward dissolution of the cave ceiling. Such upward dissolution has been called paragenesis by Ford and Ewers (1978), although the term paragenesis was first used by Renault (1968) to signify the reduced rate of dissolution of the walls of a cave passage as a result of shielding by clastic sediments.

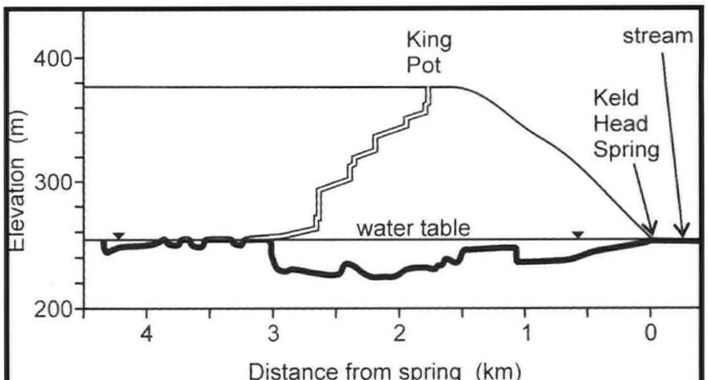

Figure 1. Extended profile of the base-level conduits under East Kingsdale and of the vadose tributary of King Pot (compiled from Brook et al., 1994 and Monico, 1995)

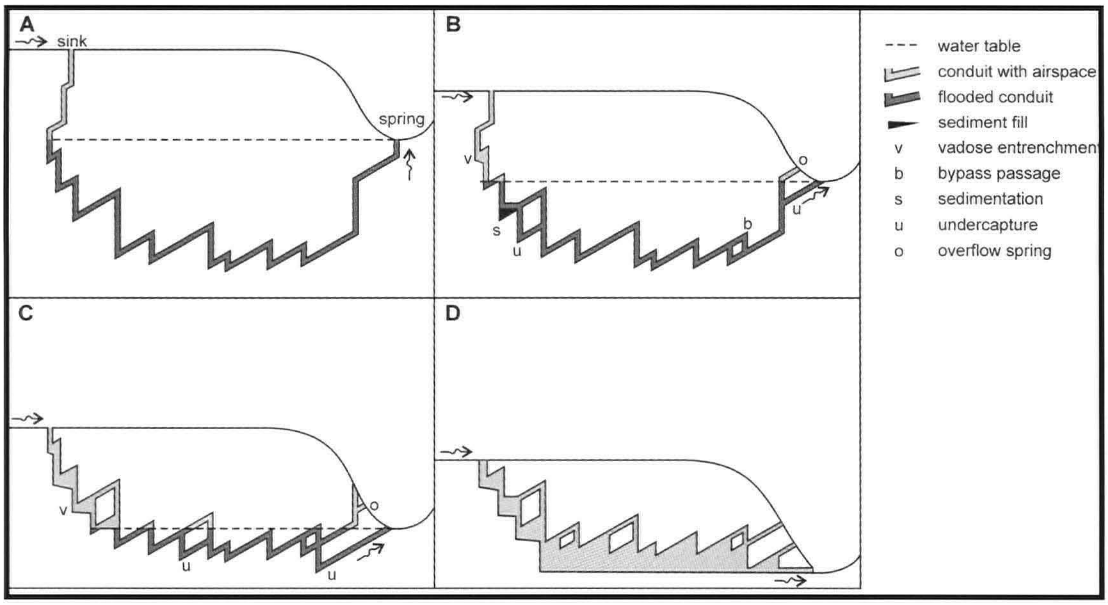

Figure 2. Evolution of a conduit as base level drops: A) initial flow path deep below the water table with a single spring, B) shallow phreatic conduit with gradational features and multiple springs C) mixed phreatic - water-table cave with multiple springs D) water-table cave.

The falling base level in Figures 2a to 2d shows the idealized evolution of the cave from a deep phreatic cave to a shallow phreatic cave to a water-table cave. Figure 2d shows the beginning of the final phase in the evolution of a cave passage – its demise. Unroofing commonly commences close to the sink and less frequently close to the spring. Excellent examples of such unroofing are seen at the Long Churn caves in England (Figure 3), as well as at Porth yr Ogof in South Wales and the Cullaun caves in Ireland (Waltham *et al.*, 1997; Tratman, 1969). Complete unroofed caves have also been described (e.g. Knez and Slabe, 2002).

Figure 2 is idealized, and examples illustrating the features shown are described in the following sections.

FEATURES RESULTING FROM BASE-LEVEL LOWERING

Vadose canyons

The simplest response in a conduit to a falling base level is vadose entrenchment. The eventual product is a stream or river cave that can be followed from sink to spring. If the initial flow path was a single loop deep below the water table, then this may be preserved in the ceiling of a river cave, with the initially deeper centre part of the cave having a lower ceiling than either entrance, as in Figure 4a. Some river caves have passages that reach 100m in height (Figure 4b). It is probable that the deep vadose entrenchment in such caves is favoured by a number of factors, including low uplift rate, low base-level lowering rate, high discharge, high sediment load, and high dissolution rate. However, the great majority of karst springs do not emerge from open caves, and most abandoned vadose passages in caves are less than 20m high. Thus the deep vadose entrenchment and open passages from sink to spring shown in Figures 2d and 4 occur only in a small fraction of conduit pathways.

Porth yr Ogof has one of the best examples in Britain of a vadose river passage. It has a length of 300m from the main upstream entrance to the resurgence entrance and is the underground path of the River Mellte. The cave was formed when the river flowed along the valley (now dry) above the cave (Waltham *et al.*, 1997, p.236). The upstream end of the dry valley is 12m above the cave roof, and so the cave developed at least 12 m below the water table and was large enough to capture surface flow when it was still 12m below the water table. Thus, it clearly records the effect of base-level lowering in transforming a formerly phreatic conduit into a currently vadose one.

In addition to base-level vadose caves, there are numerous vadose canyons that descend steeply to former or present water tables, such as in King Pot (Figure 1) and Swinsto Hole (Waltham *et al.*, 1981). There are many other similar caves in the Yorkshire Dales, as well as numerous examples in areas such as the Burren (Ireland), Waitomo (New Zealand) and West Virginia and Tennessee (USA).

Distributary springs

Springs in carbonate aquifers are commonly located close to base-level streams. Subsequent stream downcutting will result in a spring orifice being raised above base level. The steep hydraulic gradient and short horizontal distance facilitate the formation of new, lower-

Figure 3. Unroofed cave in Yorkshire. The person is standing at the junction of the unroofed passages from Wilson's Cave (on the left) and from Upper Long Churn Cave (on the right). In the background is the entrance to Lower Long Churn Cave.

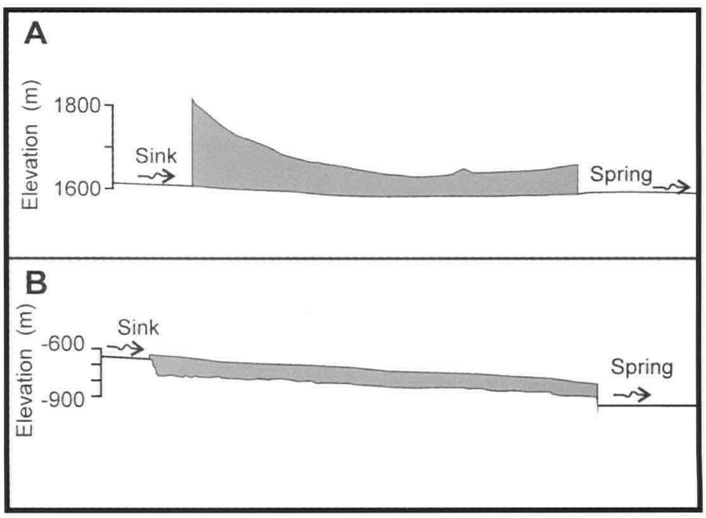

Figure 4. Projected profiles of large river caves in China A) San Cha He Dong (after Waltham, 1986), B) Xio Zhai Tien Ken - downstream section (after Senior, 1995)

level conduits, and such undercaptures are shown in Figures 2b, 2c and 2d. The lower conduit will enlarge steadily until it can discharge all the flow from the conduit system. The diminishing flow in the upper spring will result in it becoming intermittent at first and then abandoned. The resulting patterns of distributary passages associated with groups of springs were named underground deltas by Martel (1894), who also described the association of lower-elevation perennial springs and nearby overflow springs, which are both at higher elevation and are intermittent. Figure 5a shows one of several underground deltas mapped and described by Martel (1894).

Figure 5 also shows three more examples of distributary springs. The Gouffre de Padirac in France is currently the world's longest explored cave stream; this has been followed for 19km to Finou Spring and has four other distributary springs (Figure 5b). The Cheddar caves provide one of the best examples in Britain of distributary outlets, with former springs at elevations of 93m, 61m, and 33m, a current major spring at 25m, and one or more lower-elevation underflow springs (Figure 5c). There is a perennial spring for the Foussoubie System (France) in the bed of the Ardèche River and two more springs just above the river. In addition, there are two overflow springs up to 15m above the river and three former springs up to 118m above the river (Figure 5d). In low-flow conditions there are many pools and sumps, and the passages display an epiphreatic morphology due to the extensive flooding of the cave at times of high flow (Figure 6 and Minvielle, 1977).

Distributary springs are very common in karst aquifers. Some of the best-documented are in the Mammoth Cave area (Kentucky, USA), where individual groundwater basins drain to two to 46 springs (Quinlan and Ewers, 1989). These springs include Gorin Mill Spring and Graham Spring, which are the two largest springs in Kentucky. Both springs have approximately constant discharge, and at low flow are the sole springs for their respective groundwater basins. At high flow there are four overflow springs associated with Graham Spring and 45 overflow springs associated with Gorin Mill Spring (Ray, 1997; Quinlan and Ewers, 1989). Underflow springs have low discharge variance and are typically perennial, whereas overflow springs have high discharge variance and are commonly intermittent (Smart, 1983; Smart and Ford, 1986).

In Britain, three springs draining Penyghent Hill and adjacent areas (Yorkshire) show similar contrasts. The perennial Brants Gill Head has little variation in discharge, but two higher-elevation overflow springs, at Douk Gill Head and Dub Cote Cave, display much larger variations in discharge. The limited capacities of Brants Gill Head and of Gorin Mill Spring and Graham Springs were noted by Waltham *et al.* (1997) and by Ray (1997), respectively, and are probably because these springs are of recent origin and are in the process of capturing flow from their respective overflow springs. There are many other examples of distributary springs in Britain, including those at Leck Beck Head, God's Bridge, White Scar Cave, Turn Dub / Footnaw's Hole, Malham Cove / Aire Head Springs, Sleets Gill Cave and Nidd Heads in the Yorkshire Dales (Waltham *et al.*, 1997), at Ilam and Castleton in the Peak District (Christopher *et al.*, 1977, Gunn, 1991), at Clydach Gorge, Shon Sheffrey, Porth yr Ogof and Nedd Fechan in South Wales (Gascoine, 1989), and at Havant in the Hampshire Chalk (Atkinson and Smith, 1974).

Although distributary springs are common, there are some situations where their development is unlikely, including vadose river caves (e.g. Porth yr Ogof, San Cha He Dong, Xio Zhai Tien Ken - see above and Figure 4) and springs perched on low-permeability strata (e.g. Guiers Mort, France: Lismonde, 1997).

Undercaptures

Undercaptures can form distributaries leading to multiple springs, and these normally form close to the springs, as described above. However, undercaptures can form at any location along a conduit flow path (Figures 7, 8 and 9). Figure 7 shows a small part of Ogof Ffynnon Ddu, and the complexity seen is due to extensive undercapturing (Smart and Christopher, 1989). An initial phreatic tube followed a vertically and horizontally looping course, but successive undercaptures provided a progressively shorter pathway. Most abandoned passages saw little vadose entrenchment, but the modern stream passage has been entrenched by as much as 15m in

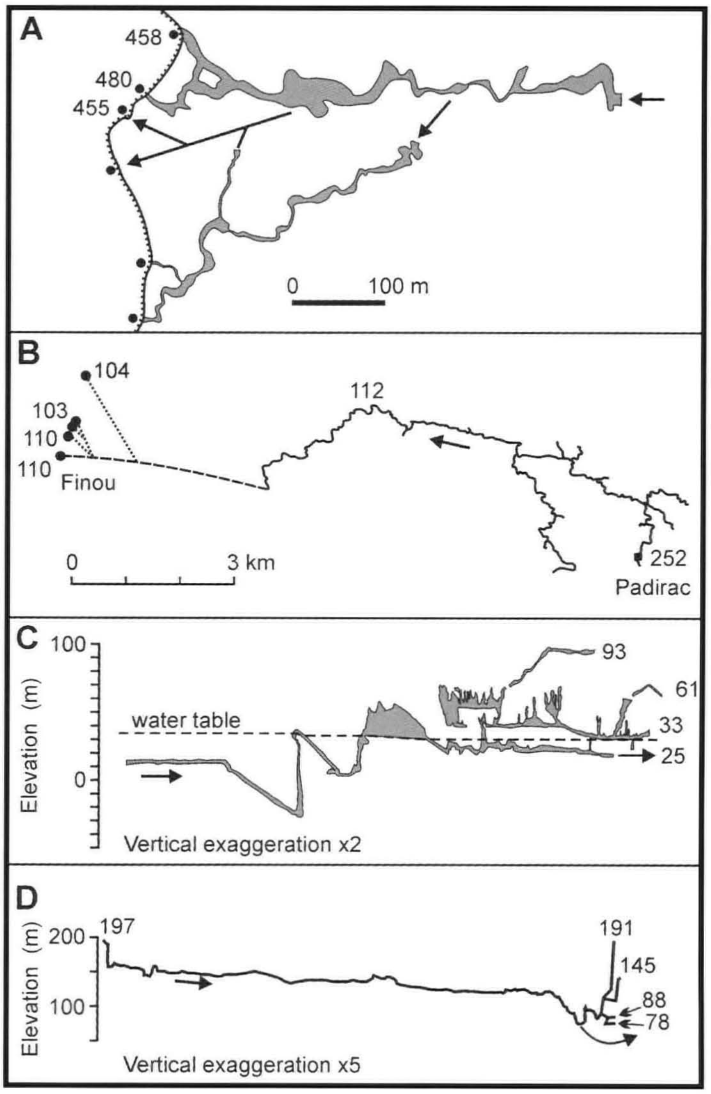

Figure 5. Conduits associated with distributary springs A) Plan of Salles-la-Source, France (after Martel, 1894), B) Plan of Gouffre de Padirac, France (after Anon. 1979, Salomon, 2000) C) Projected profile of Cheddar caves, England (after Ford, 1965 and Farr, 2000), D) Projected profile of Foussoubie system, France (after Le Roux, 1984)

this part of the cave.

The most common morphology observed in a cave passage when the water table drops below conduit level is that of a phreatic tube in the roof and a vadose canyon below. The phreatic tubes might be well developed and account for much of the passage cross-section, as in the upstream part of White Scar Cave (Yorkshire), or might have only rudimentary development, as in the downstream part of this cave (Figure 8a). Undercaptures can also occur before substantial vadose entrenchment can take place, as in the base-level passages in the West Kingsdale System (Yorkshire) and in Hölloch (Switzerland) (Figures 8b and 8c).

Passage development is somewhat more complicated where there is a substantial seasonal variation in water table elevations. In Hölloch the cave passages primarily have phreatic forms, but these are partly due to dissolution during high-flow events when the water table rises more than 100m (Wildberger and Ziegler,1992; Jeannin, 2001). Figure 9a shows high-flow and low-flow water tables in Bärenschacht (Switzerland) at a time when base level was 200m higher than today. There are three main elements to the passage network: phreatic passages about 100m below the water table that were able to transmit low-flow discharge but not all of high-flow discharge; second, epiphreatic passages that discharge the excess high flow (these were formerly phreatic passages when the water table was higher); third, connecting passages called soutirages, which drained the base of the epiphreatic loops (Häuselmann *et al.*, 2003). Figure 9b shows an interpretation of the formation of undercaptures; major ones form some 100–200m below the water

Figure 6. Epiphreatic passage in Foussoubie System, France. Fixed ladders are in place to facilitate a rapid exit when the cave floods.

table and have lengths of several hundred metres.

Some undercaptures in Mammoth Cave have much greater lengths, with diversions to springs that were several kilometres distant from the earlier outlets (Figure 10). For a time there would have been distributary flow to both old and new springs until the new conduit was able to capture all the flow. Modern distributary springs have been widely documented in the Mammoth Cave area, as noted above. The large scale of undercaptures at Mammoth Cave is due the low dip of the limestone and the extensive outcrops of limestone along the Green River, which facilitated the diversions shown in Figure 10. By contrast, caves with steeper stratal dips, such as Ogof Ffynnon Ddu, Hölloch, and Bärenschacht, have much more restricted zones where the caves have discharged into base-level streams or lakes.

The undercaptures shown in Figure 10 occurred towards the upstream end of their groundwater flow paths. Similar examples in Britain include Swildon's Hole (at Tratman's Temple), Giants Hole (below Garland's Pot), Ogof Ffynnon Ddu (at the Crevasse), Easegill Caverns (near Easter Grotto), and Gaping Gill (at Main Chamber). However, such undercaptures are less common than undercaptures closer to springs, where the shorter flow path will result in more rapid creation of new conduits (Figure 5).

The undercaptures discussed so far have been in base-level passages, but they can also occur in the vadose zone substantially above base level. Lost John's Cave (Yorkshire) offers several fine examples, with successive flow paths via Hammer Pot, New Roof Traverse and Old Roof Traverse (Waltham, 1974).

The examples described above display the complexity that can result during the evolution of a single tier of cave passages. Further complexity can occur where a cave has multiple tiers, and this is described below.

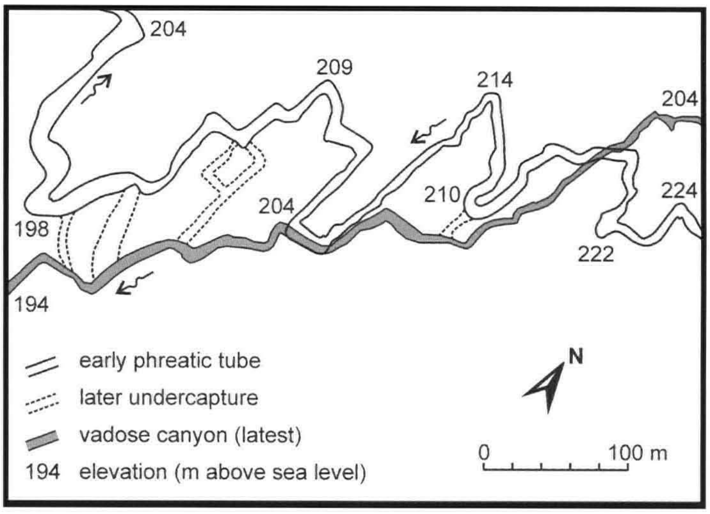

Figure 7. Passages in a small part of Ogof Ffynnon Ddu, Wales, showing early phreatic looping, successive undercaptures, and modern vadose stream flow (after Glennie, 1950 and Railton, 1953)

CAVE TIERS

Relation of tiers to base level

In some caves there are distinct tiers or levels of cave passages, where there appear to be a clustering of major passages at a particular elevation. In discussing Mammoth Cave, Davis (1930) recognized that the formation of four tiers could be explained in two ways:

> "The one-cycle theory, or theory of corrasion and solution by vadose and water-table streams, will .... be here compared, in its modified form as demanded by four pauses in elevation, with the two-cycle theory, or theory of ground-water solution, for which a continuous elevation suffices" (Davis, 1930, p.596).

Davis favoured the two-cycle theory, where there is a steady lowering over time of base level so that passages formed at depth below the water table gradually become vadose as the water table drops. Other workers, both before Davis and more recently, have interpreted tiers as having formed at the same levels as base-level rivers (the one-cycle theory of Davis, 1930) and cave tiers have often been linked to fluvial terraces (see Davis, 1930, pp.595–600; Sweeting, 1950; Palmer, 1987; Anthony, 2005). This latter hypothesis assumes that cave tiers are formed when rivers have long periods when there is negligible base-level lowering, and that such periods alternate with short periods with substantial river incision and thus base-level lowering.

Several tiers are found in some caves and in studies before 1973 tentative correlations were made with Pleistocene glacial episodes. Sedimentation was considered to occur in caves during glacial stages and lowering of base level was thought to occur during interglacial or interstadial periods (Ford, 1964; Atkinson, 1967; Miotke and Palmer, 1972). The concept of a simple correlation between cave tiers and glaciations was, however, found to be inadequate when absolute age dating results became available. First, Shackleton and Opdyke (1973) showed that there were many glaciations during the Pleistocene, with a periodicity of about 100,000 years in the Late Pleistocene. Second, absolute dating methods have shown that base-level lowering rates may be very slow (Atkinson *et al.*, 1978; Gascoyne and Ford, 1984) and that some caves may be several million years old (Ford *et al.*, 1981; Granger *et al.*, 2001; Worthington and Medville, 2005). Consequently, a single cave tier may have been active over a number of successive glacial and interglacial periods. Sea level can drop about 100m during major glaciations and so it seems likely that in many caves there may have been several substantial erosional and aggradational changes in base level during the formation of a single tier. Such rapid changes during the Pleistocene provide a challenge to the assumption of cave tiers being formed at a stable base level.

A second challenge is provided by the discovery that near-horizontal passages can form at substantial depths below the water

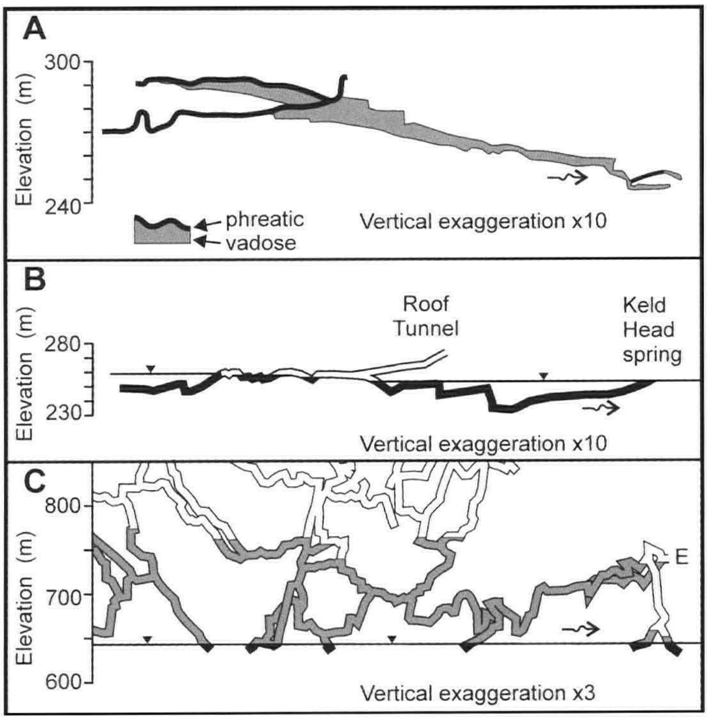

Figure 8. *Profiles of cave passages with contrasting undercaptures. A) White Scar Cave, England, B) West Kingsdale System, England, C) Hölloch, Switzerland (A adapted from Waltham, 1977; B adapted from Brook and Brook, 1976 and Monico, 1995; C after Wildberger and Ziegler, 1992)*

table (e.g. Waltham and Brook, 1980). Figure 11 provides some examples, which are described by Worthington (2004). A hypothesis for the formation of such conduits is described below.

Formation of tiers below base level

Flow deep below the water table is favoured in many settings because geothermal heating reduces viscosity and thus enhances flow. The resultant enhanced flow at depth below the water table results in increased dissolution and preferential conduit enlargement, and is thus a more favourable setting for cave formation than at the water table (Worthington, 2001, 2004). Sub-horizontal flow deep below the water table is likely in particular where flow is along the strike of the strata, as with the examples in Figure 11a-d.

Figure 12 shows a model for how cave tiers may form deep below the water table. Figures 12a, 12b, and 12c are similar to the progression shown in Figure 2, with base-level lowering resulting in a conduit evolving from deep phreatic to shallow phreatic to vadose. Eventually a new, lower-elevation conduit will capture some of the flow (Figure 12c) and over time this will enlarge and pirate all the flow from the upper conduit, leaving it abandoned (Figure 12d).

Worthington (2004) used regression of the data from twenty cave system surveys to show that the depth of conduit development below the water table can be described by:

$$D = 0.18\,(L\ \theta)^{0.81} \qquad (1)$$

where D is the mean conduit depth in metres below the corresponding water table, L is the flow path length in metres and θ is the dimensionless stratal dip (equal to the sine of the dip in degrees). This regression shows that conduit development deep below the water table is associated with long flow paths and with steeply-dipping strata.

Assuming that flow path length and stratal dip remain constant for succeeding tiers, then it follows that each new tier will be formed at a similar depth below the contemporary water table and hence tier spacing will be constant for a given cave. The vertical distance between tiers varies substantially between different caves (Table 1). However, these is a tendency in many of these examples for tier spacing to be near-constant, thus supporting the concept that new tiers form according to Equation 1.

It is likely that the hydraulic gradient is the major driving force behind the formation of a new tier. Hydraulic gradients in mature phreatic conduits are extremely small. Well-documented cases

Cave	Criterion*	Tier spacing (m)	Reference
Carlswark, England	tiers	14, 13, 13	Christopher and Beck, 1977
Mammoth, Kentucky	transition	21, 21, 16	Palmer, 2004
Friars Hole, West Virginia	tiers	18, 17, 15, 24, 27	Worthington, 1984
Demänová, Slovakia	tiers	18, 30, 32, 20	Droppa, 1966
Archway - Terikan, northern Mount Benarat, Mulu, Malaysia	tiers	35, 35, 30	Eavis, 1981
Gua Harimau - Lubang Sakai, southern Mount Benarat, Mulu, Malaysia	tiers	120, 90, 110	Eavis, 1985
Agujas, Spain	tiers	55, 25, 44, 21, 13, 55, 50	Rossi *et al.*, 1997
Daren Cilau - Craig a Ffynnon, Wales	tiers	40, 27, 48, 63	Smart and Gardener, 1989
Nettlebed, New Zealand	tiers	65, 90, 60, 65	Ford and Williams, 1989, p.122
Siebenhengste, Switzerland	transition	150, 80, 135, 80, 65, 550, 85, 45, 60, 40, 102	Jeannin *et al.*, 2005
Hölloch, Switzerland	tiers	160, 120	Bögli, 1980
Nelfastla de Nieva, Mexico	tiers	230, 230	Worthington, 1991

Table 1: *Vertical spacing of cave tiers*
** The criterion for recognition: either the transition in a passage from vadose shape to phreatic shape or the vertical spacing between tiers of*

include gradients of 0.0006 – 0.0015 for Jortulla Cave, Norway (Lauritzen *et al.*, 1985), and 0.0012 – 0.004 for Mangle Hole – Banwell Spring, England (Hobbs, 1988). The geometric mean hydraulic gradient of these phreatic passages is 0.0014.

By contrast, the gradients of vadose cave streams are much steeper. Gradients in some notable vadose stream passages are 0.007 in Sinks of Gandy, West Virginia (Dasher, 2000), 0.019 in San Cha He Dong, China (Figure 4a), 0.022 in Porth yr Ogof, Wales (Lloyd, 1980), 0.03 in White Scar Cave, England (Waltham, 1977), 0.044 in Dan yr Ogof, Wales (Coase and Judson, 1977), 0.045 in Lancaster-Easegill, England (Ashmead, 1974), 0.093 in Ogof Ffynnon Ddu, Wales (Smart and Christopher, 1989), 0.098 in Xio Zhai Tien Ken, China (Figure 4b), and 0.2 in Giant's Hole, England (Ford, 1977). The geometric mean gradient of these vadose streams is 0.064, which is 45 times greater than for the average gradient in the two submerged conduits.

The hydraulic gradient in the partially vadose upper conduit in Figure 11c will be much steeper than in the phreatic conduit in Figure 11a, thus greatly increasing the flow through the lower, immature conduit. It is common for flow to be captured to the lower conduit before extensive vadose development has occurred in the upper conduit, and Figure 11 shows five such examples. Hölloch and Bärenschacht are two additional examples of caves that lack vadose base-level passages because all the passages shown in Figures 8c and 9, respectively, are tubes that were formed and enlarged under phreatic or epiphreatic conditions. In each case, the passages were abandoned by the cave streams that formed them before they could develop beyond stage B in Figure 12.

Conduits in some caves progress to develop vadose canyons before capture to a lower tier. Agujas Cave System (Spain) provides

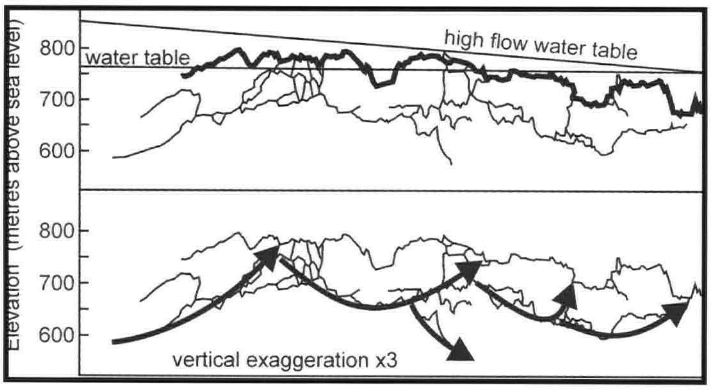

Figure 9. Profiles of Bärenschacht, Switzerland. (a) Profile of the cave when the water table was at 760 m, showing passages up to 200 m below the current water table and in bold the main epiphreatic flood overflow passage (after Häuselmann et al., 2003). (b) Interpretation of the development of five major undercaptures.

a spectacular example of a multi-tier cave where each phreatic passage has evolved to a vadose canyon (Figure 13). The cave is developed in overturned limestones that have a 50° dip, and the passages cut across the bedding. The canyons are up to 20m in height. Other examples of conduits with extensive vadose development are shown in Figures 4, 5a and 5b, 7, and 8a. In Mammoth Cave, Palmer (1987) showed that about half of the base-level passages have negligible vadose development (e.g. Cleaveland Avenue, Marshall Avenue, Swinnerton Avenue, and Turner Avenue in Figure 10) and the remainder have developed substantial vadose canyons (e.g. Echo River and Mather Avenue in Figure 10).

The phreatic conduits shown in figures 9 and 11 were all abandoned before significant vadose erosion took place. These caves are all situated in major mountain chains and it seems likely that the lack of vadose flow may be due largely to rapid base-level lowering rates. Such rates have been measured for several of these caves: 130–1120m/Ma in the Yorkshire Pot area (Ford *et al.*, 1981), <440m/Ma at Nettlebed Cave (Ford and Williams, 1989, p.122), 190m/Ma for the Mulu caves (Farrant *et al.*, 1995), and <600m/Ma at Bärenschacht (Häuselmann, 2002). By contrast, vadose stream passages are more common in lowland karst areas such as Mammoth Cave and the Yorkshire Dales, where much lower base-level lowering rates occur. These average 20m/Ma at Mammoth Cave

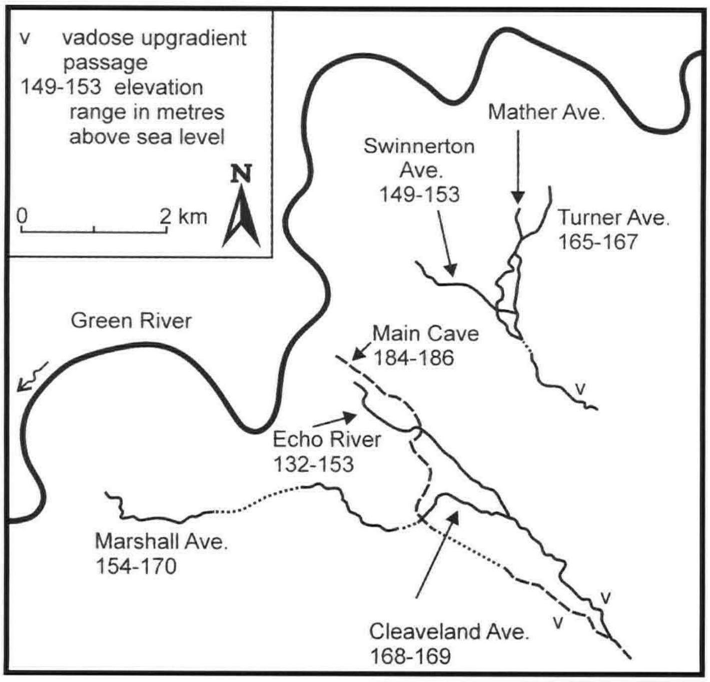

Figure 10. Some major passages in Mammoth Cave, showing undercaptures under Flint Ridge (Mather Avenue then Swinnerton Avenue) and under Mammoth Cave Ridge (Cleaveland Avenue then Echo River), with range of elevations of passages in metres above sea level (adapted from Palmer, 1987 and Palmer, 1989b)

(Granger *et al.*, 2001) and 20–80m/Ma in the Yorkshire Dales (Gascoyne and Ford, 1984). However, this correlation does not hold in all cases; Agujas Cave, for example, is in a mountainous area, yet has substantial vadose development. It is likely that several other factors also influence the transition from one cave tier to a lower one, including stream discharge and variability and stream clastic sediment load.

Testing the two hypotheses of tier formation

It is clear from the above section, and particularly from the examples shown in Figures 5c, 5d, 9 and 11, that there can be extensive sub-horizontal cave development at substantial depths below the water table (the two-cycle theory of Davis, 1930). This contrasts with the evidence that Palmer (1987) found to support tier formation at stable base levels (the one-cycle theory of Davis, 1930). The evidence in support of these two hypotheses in three cave areas (Mammoth Cave, the Mulu caves, and the Yorkshire Dales) is discussed below.

Palmer (1987) carried out extensive levelling in Mammoth Cave to determine the elevations of major passages and of vadose/phreatic transitions, which are an important indicator of former water tables. Palmer identified four major levels of major cave passages at Mammoth Cave at elevations of 210, 180, 168, and 152m. These were later shown to have been active from before 3.3 Ma to 0.7 Ma before present (Granger *et al.*, 2001). Lower passages in the cave, down to river level at 128m, lack well-defined levels.

Davis suggested that evidence to test his one-cycle and two-cycle hypotheses should include determining whether "all the caverns in the district should probably have the same number of gallery-levels separated by essentially the same vertical intervals" (Davis, 1930, p.596). Some major passages associated with levels C (167–169m) and D (152–153m) in Mammoth Cave are shown in Figure 10. Vadose to phreatic transitions occur at these elevations in widely separated passages, which provides strong support for the one-cycle hypothesis. Passages such as Cleaveland Avenue also provide strong support. This passage is an almost horizontal phreatic tube that appears to follow a former water table with exquisite fidelity for 1500m (Palmer, 1981, pp.101–103). Support for the one-cycle hypothesis would be substantially strengthened if major passages in other caves in the area were found to correlate with the Mammoth Cave levels. One possible cave is Ganter Cave, which lies on the opposite side of the base-level river to Mammoth Cave and has been reported to have "five well-defined cave levels" (George, 1989, p.217).

Other evidence at Mammoth Cave supports the two-cycle theory. For instance, Marshall Avenue, the downstream continuation of Cleaveland Avenue, has a phreatic loop that descends 14m below the level of Cleaveland Avenue (Figure 10). Furthermore, when an undercapture from Cleaveland Avenue developed, it descended at Echo River to at least 23m below the water table (Palmer, 1989a). Thus, not all phreatic passages have formed at the main levels described by Palmer (1981). Such variability demonstrates the complexity of cave formation, with different passages at Mammoth Cave supporting either the one-cycle or the two-cycle theories of Davis (1930).

In the Mulu area (Sarawak, Malaysia), cave wall-notches correlate with the benthic oxygen isotope record, and the correlation of wall-notch elevation with time shows that the base-level lowering rate has been a constant 190m/Ma for at least the last 700 ka (Farrant *et al.*, 1995). Many of the passages in the Mulu caves were formed at depths of at least some tens of metres below the water table and such passages may be near-horizontal for substantial distances (Figure 11 and Waltham and Brook, 1980). The wall notches provide evidence for water-table modification of existing passages, but the evidence for the formation of new conduits at the water table is limited and equivocal (Waltham and Brook, 1980). This lack of water-table caves is consistent with the constant base-level lowering rate at Mulu. Water-table caves are thought to occur principally when there is a long period with no base-level lowering (Palmer, 1987). In the absence of such a stable base level it follows that water-table caves will not develop.

Sweeting (1950) correlated cave levels in the Yorkshire Dales with erosion surfaces, but later work has shown that there is a sequence of major inception horizons in the limestone and that these provide the primary guidance for sub-horizontal passages (Waltham, 1970; Lowe, 2000). The concept of base-level control is further challenged by differences in the elevation of major relict cave passages in adjacent caves. For instance, the most prominent level of relict passages under Casterton Fell drained to a spring at 250m and there is also an erosion surface at this elevation (Ashmead, 1974), but Waltham *et al.* (1997, p.37) found the two major relict water tables in Leck Fell caves were at 290m and 225m. Both Casterton and Leck fells drain to the same spring, Leck Beck Head, and so the same relict water tables would be expected in both areas if base-level control were paramount in determining passage elevation. These differences between the cave levels under Casterton and Leck fells suggests that the relict levels were not formed as water-table caves during periods of base-level stability. This view is supported by the lack of association of modern base-level caves with the water table, because the caves under Casterton Fell and Leck Fell have been explored to depths of -32m and -64m, respectively, below the current water table (Monico, 1995).

The evidence from Mammoth Cave, Mulu, and the Yorkshire Dales provides strong evidence that a stable base level is not necessary for the formation of cave tiers. Some passages at Mammoth Cave appear to have formed in proximity to contemporary water tables that may have been stable for long periods, thus supporting the one-cycle theory of Davis (1930). However, other passages at Mammoth Cave and as well as caves at Mulu and in the Yorkshire Dales better support the two-cycle theory of Davis (1930), and this is summarized in Figure 12.

DISCUSSION

Passages possibly associated with a rising base level

The model described above and shown in Figure 12 is associated with a base level that falls steadily over time. However, there are a number of situations where base level can rise. Examples include aggradation in valleys, short-term rises in water level following flooding, and the rise in the Mediterranean Sea following the Messinian regression.

Aggradation in valleys can result in the formation of bypass passages, wall notches, and distributary springs. Bypass passages can result from sedimentation at the base of a phreatic loop, from roof collapse within a cave passage, or from sedimentation at a spring, and the likelihood of occurrence is enhanced where there is a large range in discharge (Ford, 1965; Palmer, 1975, 1991). Wall notches in Mulu caves correlate with aggrading alluvial fans outside the caves (Farrant *et al.*, 1995; Waltham, 2004). The principal explanation given earlier for distributary flow is that it is caused by a lowering of the water table and that this results in undercaptures. This explanation appears best to explain many examples such as those shown in Figure 5. In other cases, such as at Mammoth Cave, it may not be possible to tell whether distributaries were formed by a falling or by a rising base level. However, the long-term trend in any area is usually for base level to fall and so distributaries are more likely to be associated with a falling base level.

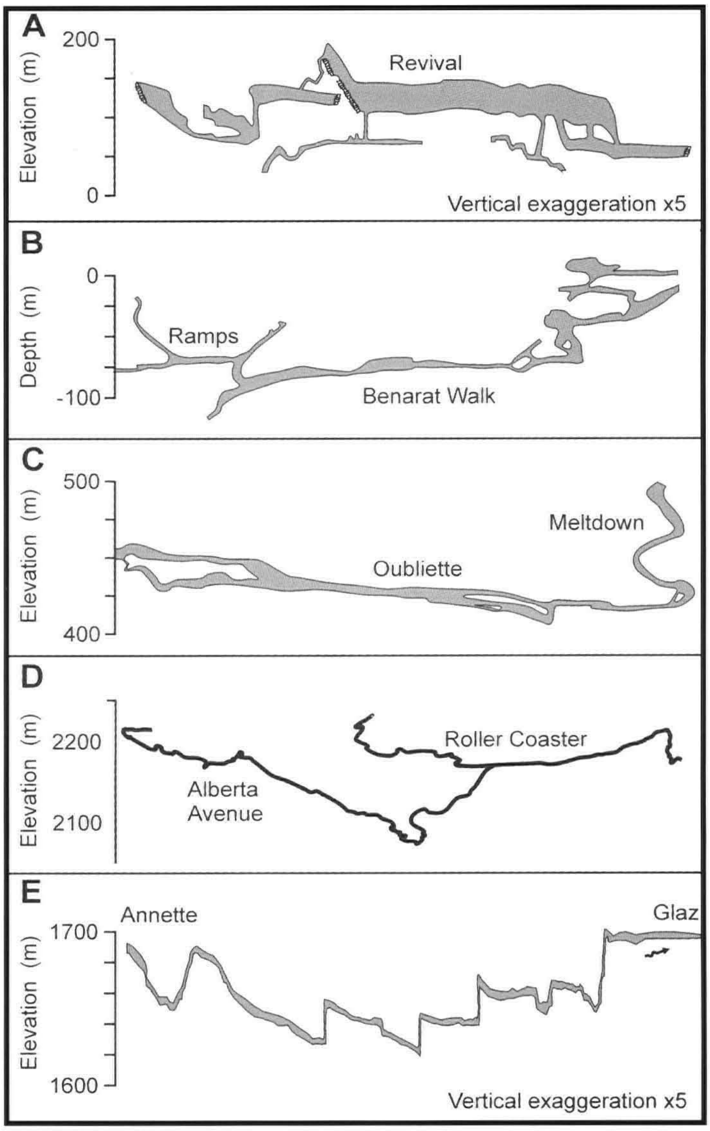

Figure 11. Profiles of conduits formed well below the water table. A) Clearwater Cave, Malaysia, B) Tiger Foot Cave, Malaysia, C) Nettlebed Cave, New Zealand, D) Yorkshire Pot, Canada, E) Dent de Crolles, France (A and B after Waltham and Brook, 1980; C after Pugsley, 1979; D after Worthington, 1991; E adapted from Lismonde, 1997)

Audra (1994, 1997) noted that the water table can rise more than 100m during the snowmelt period in some mountain areas, and he proposed that the epiphreatic or floodwater zone can be a major

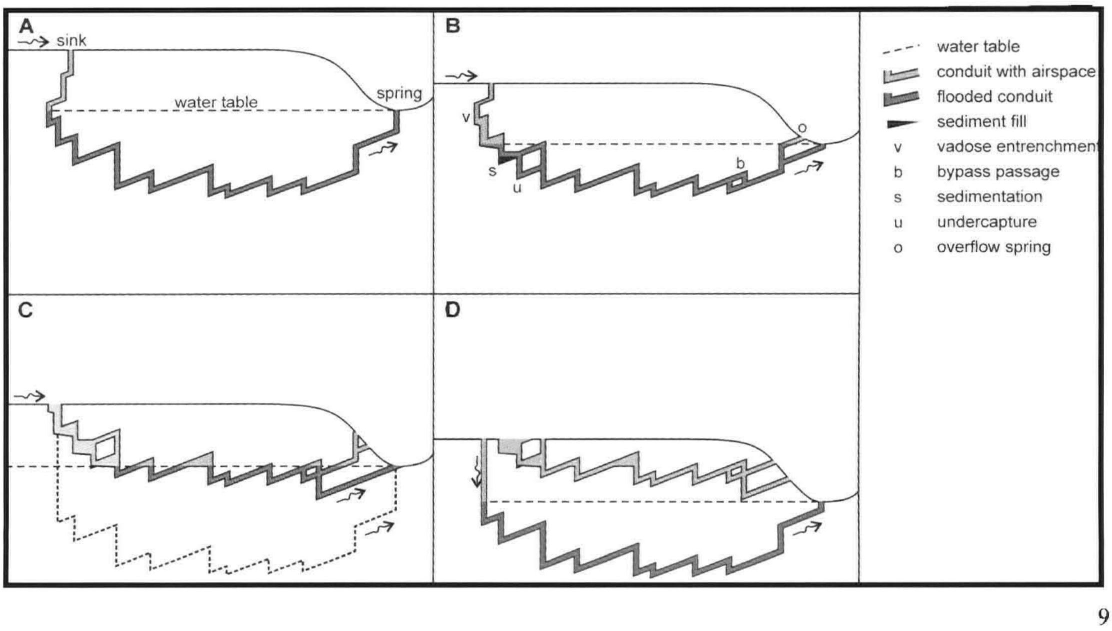

Figure 12. Development of a new conduit as base level drops: A) initial conduit flow path deep below the water table with a single spring, B) shallow phreatic conduit with gradational features and multiple springs C) water-table cave with some flow being pirated to a developing conduit D) all flow captured by the deep conduit.

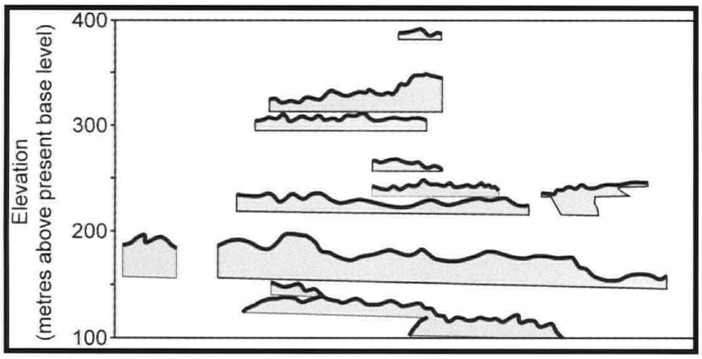

Figure 13. Profile of Agujas Cave System, showing a succession of tiers, with initial phreatic tubes (in bold) and subsequent vadose canyon entrenchment (in grey) (after Rossi et al., 1997)

locus for cave development. This hypothesis is supported by a detailed study at Bärenschacht, where Häuselmann *et al.* (2003) showed that much conduit enlargement takes place in the epiphreatic zone and that some passages (soutirages) are also formed in this zone. Nevertheless, most of the major passage formation takes place some 100m below the water table in the phreatic zone (Figure 9).

There are a number of karst springs in southern France that have been dived to great depths. These include the Fontaine de Vaucluse (-308m), Goul de la Tannerie (-209m), Goul du Pont (-178m), Font Estramar (-167m), Port Miou (-147m) and Source du Lez (-101m). Audra *et al.* (2004) suggested that the karst systems draining to these springs were formed during the Messinian age (5.96 to 5.32 My before present) when the Straits of Gibraltar were closed and sea level in the Mediterranean dropped by at least 1500m. Subsequent transgression would have flooded these caves and extensive alluvial sedimentation would have blocked the original spring outlets. This would have caused water to back up and form springs at what might earlier have been vadose shafts. This explanation implies that the passages feeding these deep springs are more than five million years old; this exceptional longevity is much greater than in most caves, where passages are usually active for much less than one million years.

An alternative explanation for the deep karst springs in southern France is that their deep flow is predicted on hydraulic grounds due to the long flow paths feeding the springs (Worthington, 2004, Equation 10). This also explains the deep flow in French springs such as Source du Bouillant (148m), Fontaine de Lussac (-142m) and Fontaine de Chartreux (-138m), none of which is in the Mediterranean basin and therefore cannot be attributed to the Messinian sea level changes.

In all the cases described above there is a long-term trend of base-level lowering. Consequently, cave formation is associated primarily with falling base levels, and the rising base levels described above add only second-order effects to cave patterns.

Trends in loop amplitude over time

Ford (1965) found evidence for former deep phreatic flow in the caves of the Mendip Hills (England), with flow to depths of -50m at the Cheddar caves, -85m in St Cuthbert's Swallet, -27m or -43m in Swildon's Hole and -43m in Wookey Hole. Processes such as undercapture, vadose entrenchment, and development of bypass passages later reduced the depth of looping over time at Swildon's Hole to -21m, then to -14m, and finally to -5m in the then-known streamway (as far as Sump 6: Ford, 1965). Similar processes occurred at Wookey Hole, so that based on then-available evidence it was inferred that "it now constitutes a water table cave" (Ford, 1965, p.124).

Ford (1968, Figure 2) inferred from the above observations that there had been a general increase in fissure frequency (the number of open fractures that conduits would develop along) over time in the Mendip Hills. This then resulted in a concomitant decrease in the depth of phreatic looping in successive cave tiers because it was thought at that time that conduits were more likely to develop close to the water table. It was later suggested that this model of increasing fissure frequency and decreasing loop amplitude over time was widely applicable (Ford and Ewers, 1978; Ford and Williams, 1989, pp.265–270). However, Hölloch was given as an example of a cave where there has been no tendency towards decreasing loop amplitude over time (Ford and Williams, 1989, p.267).

Over the last 40 years there has been substantial exploration by divers in Mendip caves, and this has now shown that there is no overall trend over time towards shallower phreatic loops. Divers have reached depths of -20m in Sump 12 of Swildon's Hole, -58m at Cheddar (Figure 5c), and -94m in Wookey Hole. Similarly, analysis of loop amplitude at Hölloch and of maximum depth of flow at Mammoth Cave and at Agujas Cave show that there is no trend towards shallower phreatic flow in successive tiers (Figure 14). This lack of a trend in successive tiers contrasts with the trend in a single tier, where decreasing loop amplitude is common. This is caused by gradational processes such as vadose entrenchment, sediment fill, bypass passage development, and undercaptures, which were identified by Ford (1965) in Swildon's Hole.

CONCLUSIONS

The explanation of cave evolution described above expands on the two-cycle hypothesis of Davis (1930) and incorporates the gradational processes described by Ford (1965). This model is supported by extensive evidence from caves.

Cave conduits commonly form as a single loop below the water table, with the depth of flow being a function of flow path length and the dip of the strata (Worthington, 2004). As base-level lowering progresses, the outlet elevation falls and the water table drops, so that conduits evolve from being deep phreatic to shallow phreatic to vadose water-table passages. Only in rare cases does the process evolve to completion, with a vadose passage stretching from sink to spring. In most cases the process is interrupted at an earlier stage by the capture of flow to a new, deeper, phreatic conduit.

Gradational features such as vadose entrenchment, bypass passages, and undercaptures are common. They occur in most caves and account for much of the complexity seen in cave maps. Well-defined cave tiers are found only in a minority of caves and in most cases appear to have formed well below the water table rather than at it. It seems likely that further insight on cave evolution processes will follow from measurement of pertinent variables such as base-level lowering rates, uplift rates, discharge and sediment fluxes through conduits, and conduit wall-retreat rates.

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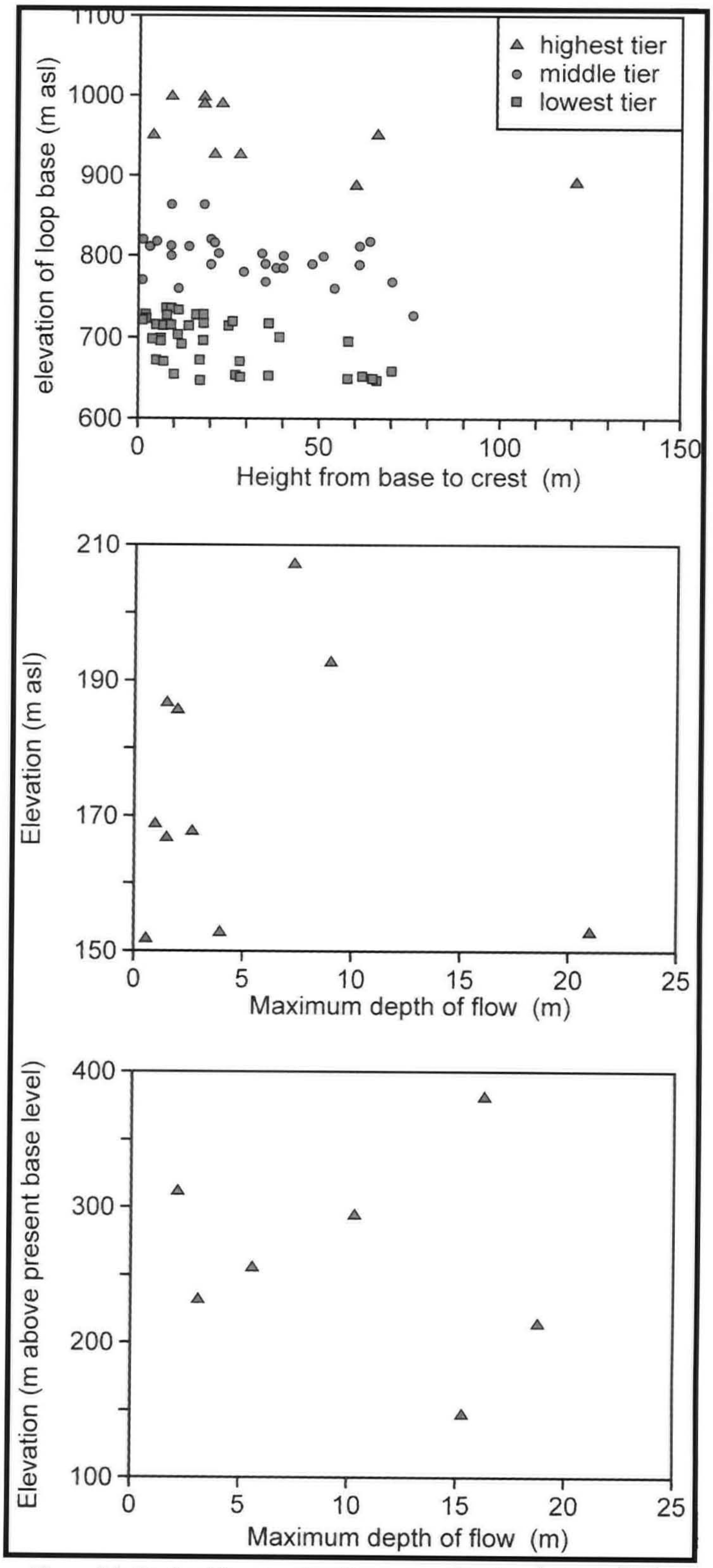

Figure 14. *Depth of flow trends with passage elevation: loop amplitude at Hölloch Cave (top); maximum depth of flow for major passages at Mammoth Cave (middle); maximum depth of flow for the tiers at Agujas Cave (bottom). Data from a) Bögli, 1980, (b) Palmer, 1987, (c) Rossi et al., 1997.*

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Possible fossil cenotes or blue holes in the Carboniferous Limestone of the Derbyshire Peak District, UK.

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Abstract: Karstic pits filled with derived volcanic and carbonate detritus in shallowing-upward cyclic limestones near Wirksworth, Derbyshire, may be cenotes or blue holes analogous to those in the Yucatan peninsula of Mexico and in the Bahamas.

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INTRODUCTION

Brief references to possible cenotes of Carboniferous age in the limestones around Wirksworth have been made in geological literature (e.g. Walkden, 1970, 1972, 1974; Oakman and Walkden, 1982; Oakman, 1984; Ford, 1984, 1989) but no detailed description or analysis of their origin and development has been published. No comparison has been made to the type cenotes of Mexico or to the analogous blue holes of the Bahamas.

In the Yucatan peninsula of Mexico, cenotes (a term derived from the Mayan word "d'zonet" meaning well, and sometimes mis-used for any form of palaeo-cave) are open water-filled shafts in a low-lying karstic terrain, many leading to extensive horizontal submerged cave systems accessible only by cave-divers (Reddell, 1977; Mylroie and Carew, 1990; Beddows, 2004). Around 500km of such submerged caves have now been explored by cave divers at a maximum depth of 100m, though typically 10–30m deep. Allogenic drainage enters some systems from hills farther inland. Stoping of cave roofs causes collapses resulting in the open flooded pits known as cenotes, which generally provide the only access.

The blue holes of the Bahamas include similar water-filled cave systems, seen both on the islands and on the surrounding tide-covered carbonate platforms (Palmer, 1983, 1985a and b, 1986, 1989; Whitaker, 1998; Mylroie, 2004: see also *Cave Science* special issue 1984 and *Cave and Karst Science* special issue 1998). Sub-horizontal caves lie at various depths down to about 50m. Their origin is due to percolating rainfall creating freshwater lenses above salt water, with mixing-corrosion at the halocline, possibly enhanced by bacterially-mediated dissolution aided by hydrogen sulphide. Water movement is caused by outflowing excess rainfall and by tidal pumping effects. Upward stoping by collapse of cave roofs results in both inland and coastal-flat blue holes.

A separate category includes linear, fault-guided, fissure blue holes more or less parallel to the Bahama bank margins (Smart *et al.*, 1988; Whitaker and Smart, 1997), where parts of the banks are beginning to collapse and break away from the main carbonate massif; similar fissures are also known in Yucatan. Some of these fissures are more than a kilometre long and exceed depths of 100m.

Divers have found that both inland cenotes and blue holes are floored by piles of collapsed boulders, indicative of a sub-surface origin caused by roof failure. If the collapses have not blocked them, lateral passages extend from their sides.

Thus, both cenotes and blue holes include phreatic cave systems fed either by influent allogenic water (Yucatan) or autogenic percolating rain-water (Bahamas) moving slowly towards outlets near the coast or rising to springs on the surrounding carbonate platforms. Proximity to the flanks of these platforms could enhance phreatic flow and hence speleogenesis. This environment, near the edge of a carbonate massif, fits the flank-margin model proposed by Mylroie and Carew (1990) and Mylroie *et al.* (1995). Their model requires a substantial rainfall on the karst land areas, be they the low-lying shelf, adjacent hills or offshore islands.

In both the Bahamas and Yucatan, sea-levels were lowered by at least 100m during Pleistocene glacial stages, resulting in the caves being sufficiently drained to act as vadose conduits draining towards platform margins. Little vadose modification of the presently drowned deep cave systems' morphology by the flowing streams has been observed, but percolating rainwater gave rise to numerous speleothem deposits, which could form only in an air-filled environment. However, relict cave systems are known related to high sea-levels in the Turks and Caicos Islands.

The corollary of the above is that anywhere in the stratigraphical record, where shallow-water carbonate successions show evidence

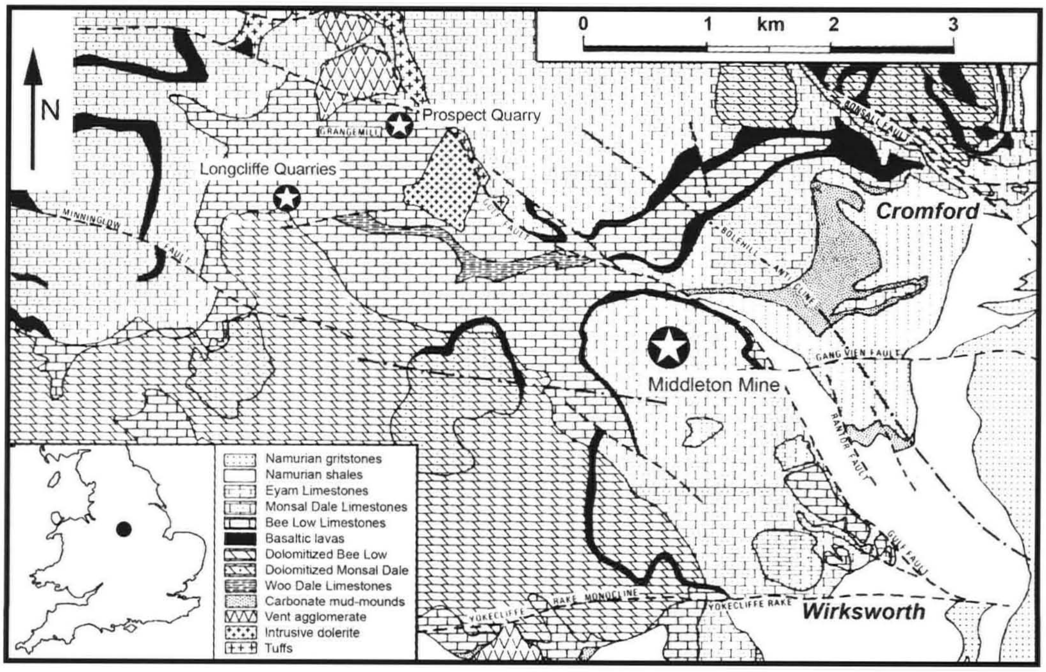


Figure 1. Sketch map of the geology of the Wirksworth-Cromford-Grangemill area (modified after Oakman, 1984).

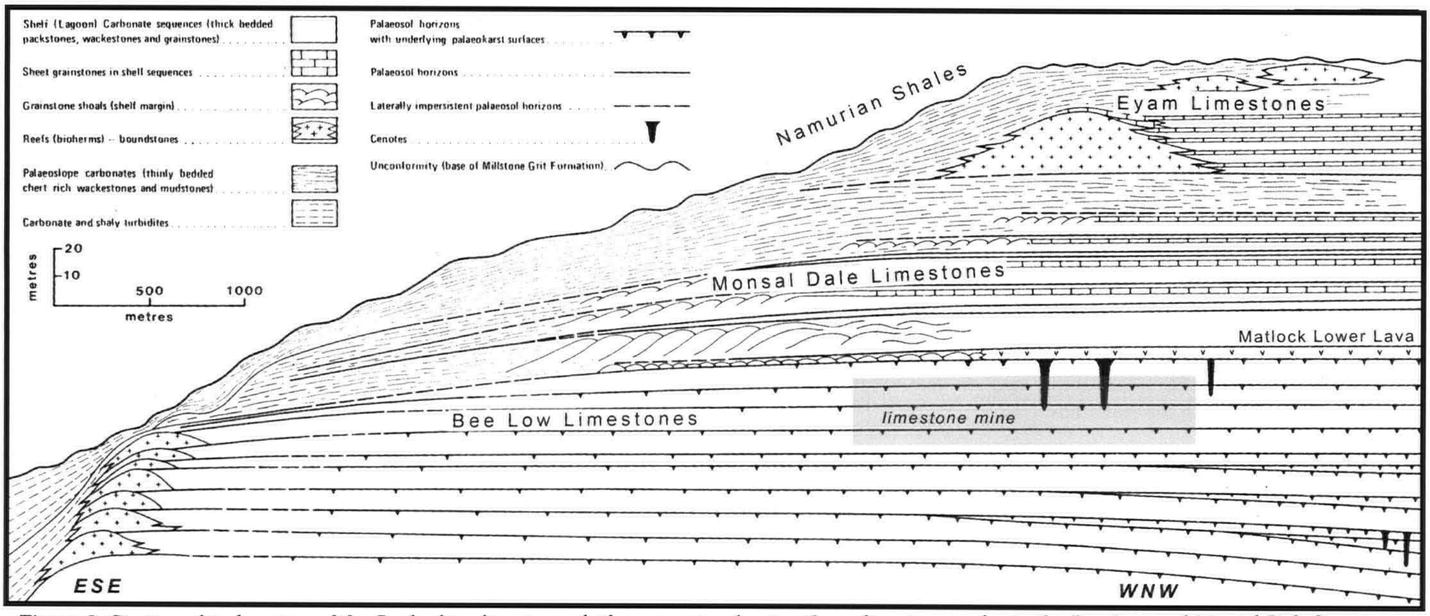


Figure 2. *Stratigraphical section of the Derbyshire limestone platform margin, showing the sedimentary cycles in the Bee Low and Monsal Dale limestones. Cenotes are shown diagrammatically (modified after Oakman, 1984). The position of the Middleton Limestone Mine is shown diagrammatically.*

of intermittent sea-level falls and rises, cenotes and their related cave systems could have developed. A scan of stratigraphical literature and the standard works on palaeokarst and palaeo-caves (James and Choquette, 1988; Bosak, 1989) has failed to reveal comparable examples. However, possible cenotes were reported by Oakman (1984) and Oakman and Walkden (1982) in the Carboniferous Limestone of the Peak District, particularly near Wirksworth, Derbyshire. Others have been revealed by recent quarrying in the Carboniferous Limestone of Cookstown, Co. Tyrone, Northern Ireland (Simms, pers. comm.). It is the aim of this note to draw attention to these Carboniferous examples in the Peak District in the hope that others will be found elsewhere in the stratigraphical record.

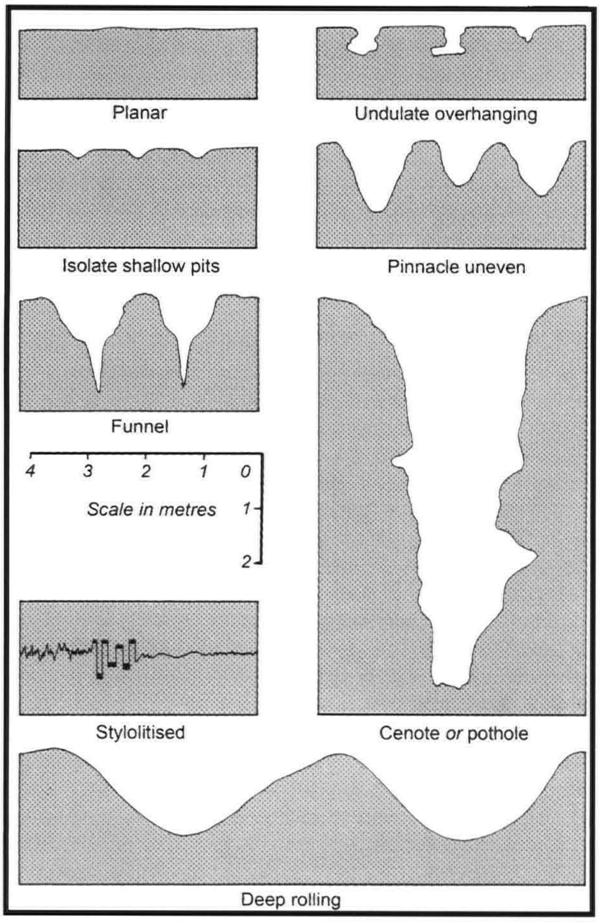


Figure 3. *Diagrammatic sections of bedding planes and palaeokarstic surfaces, including one cenote, around Wirksworth (from Oakman, 1984).*

THE WIRKSWORTH LIMESTONES AND THEIR CENOTES

The general stratigraphy of the Carboniferous Limestone (Dinantian) of the Wirksworth area has been described by Shirley (1959) and in several Geological Survey Memoirs and Reports (Smith *et al.*, 1967; Frost and Smart, 1979; Cox and Harrison,1980; Harrison and Adlam, 1985). Stratigraphical nomenclature was revised by Aitkenhead and Chisholm (1982) and is currently being further refined and formalized as part of a proposed new stratigraphical framework for the Carboniferous rocks of the United Kingdom (Waters *et al.*, in press). Gutteridge (2003) and Cossey *et al.* (2004) have added further sedimentological data. A simplified stratigraphical column is included on the area map (Fig.1).

The stratigraphical units involved comprise the Bee Low Limestone Formation of Asbian age, about 100m thick, overlain by the Monsal Dale Limestone Formation of Brigantian age, with a maximum thickness of about 66m. The latter is capped disconformably by the Eyam Limestone Formation, with a variable thickness between 25 and 55m (Gutteridge, 1983). The Bee Low and Monsal Dale limestones are separated over part of the Wirksworth area by the Matlock Lower Lava, a basalt flow up to 20m thick. Since 1959 the Bee Low Limestone has been mined for high purity carbonates at Middleton-by-Wirksworth. Both limestone formations are also exposed in the Middlepeak quarries, 0.5km down-dip towards Wirksworth, but the lava flow did not reach this area.

The Bee Low Limestone Formation (formerly Hoptonwood Limestones) consists mainly of calcarenites with subordinate calcite mudstones. The Monsal Dale Limestone Formation (formerly Matlock Limestones) includes similar calcarenites but with common chert and with mud-mounds (reefs) towards the top. The Eyam Limestone Formation (formerly Cawdor Limestones) comprises variable calcarenites with increasing mudstone partings, passing up into the Longstone Mudstone (now considered part of the Bowland Shale Formation), which is distinguishable from the overlying Namurian Edale Shale only on the basis of fossil content. Thus, most of the limestone succession shows a sequence of shallowing-upwards carbonate cycles with evidence of intermittent emergence (Walkden, 1970, 1972, 1974, 1977, 1987; Oakman and Walkden, 1982; Oakman, 1984; Walkden and Williams, 1991). Cycles are generally less than 10m thick, and 15 major and several minor cycles have been recognized in the Bee Low Limestone, and ten in the Monsal Dale Limestone. They were deposited on a carbonate ramp (Ahr, 1973) sloping gently towards the south, with intermittent development of marginal reefs southwest of Wirksworth (Oakman, 1984) (Fig.2).

The topmost beds of each cycle are commonly characterized by palaeosol textures, with rhizoliths and vadose cements. Many of these surfaces are capped by thin layers of volcanic dust tuff, known locally as clay-wayboards, generally only a few centimetres thick but locally reaching a metre (Walkden, 1977). Traces of carbonaceous material on some wayboards demonstrate the former existence of terrestrial vegetation and soil-forming processes, which could only develop after emergence above sea level (Vanstone, 1998). Weathering has altered much of the wayboard material, and some associated lavas, to yellow-green or brownish clays. Generally the lavas are considered to have been erupted below sealevel, but some lava tops might have been exposed. Where quarrying or mining has exposed wayboard/limestone contacts the limestone surfaces are commonly potholed, i.e. characterized by rounded pits up to a metre wide and deep, filled with tuff, but without the pebbles that might suggest physical erosion as moulins (Fig.3). Modern examples of potholed surfaces occur in the Bahamas. The term "banana holes" has been applied to such surfaces, but really they are vadose shafts on upland karst. According to Walkden (1977) these pits are due to penecontemporaneous dissolution beneath clay-wayboards and pre-date jointing. Later erosional removal of the clay-wayboards leaves a potholed (mammilated) palaeokarstic surface.

Comparable mammilated palaeokarstic surfaces have been identified in most Dinantian limestone regions of England and Wales, and Vanstone (1998) has suggested that the location of potholes on such surfaces can be related to the effects of tree root growth and conduction of slightly acidic water down the trees. An example occurs in the Hotwells Limestone Formation of White Hole Quarry, East Mendip. However, no other evidence of the presence of trees has been found either at Wirksworth or elsewhere. Vanstone noted that the potholes rarely exceeded 3m in depth and he did not record any cenote-like pits.

Seventeen "sink-holes" or other "possible karstic features" have been found during the working of the Bee Low Limestone in the Middleton Limestone Mine, where mining has been on a pillar-and-stall pattern grid extending about 1.5km from one side of the Middleton Moor plateau to the other (Fig.4). Nine of these appear to be Pleistocene karstic features with fills of glacial outwash sand and gravel, though it is not clear how the latter entered the potholes through a cover of around 80m of limestone and 20m of lava. Eight are regarded as possible cenotes and most were examined by the writer some 20 years ago. Recently they have been re-visited by Dr A C Waltham. The main beds worked are some 8m high with the roof lying 10m beneath the Matlock Lower Lava. Sub-levels have been worked beneath parts of the mine, with intervening floors 4m thick above a further 8m of worked-out galleries. Four faults with a WNW–ESE trend cross the worked area, with downthrows of 10 to 36m towards the southwest. The fault planes show patchy mineralization with intermittent evidence of old lead miners' activity. Pillars are around 10m wide and it is in some of these that the cenotes have been observed. At least one cenote can be traced

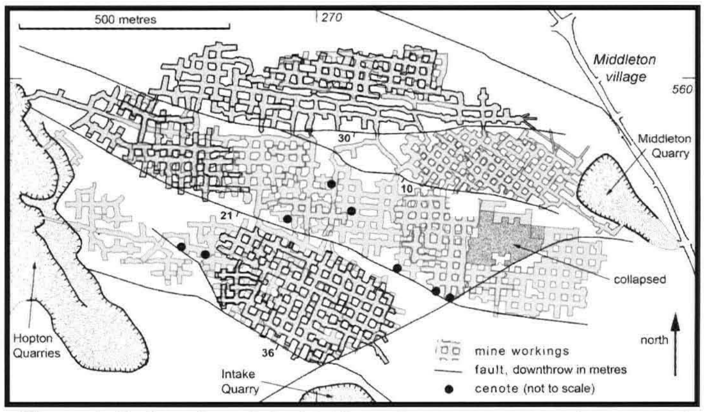


Figure 4. *Outline plan of the Middleton Limestone Mine and quarries of Middleton Moor, with the locations of eight cenotes and fills as observed in early 2005. Heavy outlines to some workings indicate sub-levels. (Adapted from mine plans by courtesy of Omya p.l.c.).*

through two mine levels and the intervening floor. Together with the unseen part up to the lava, the total depth is nearly 30m. The cenotes are roughly-cylindrical pits, 1 to 10m in width (Fig.5). Only one pit can be entered, where the debris has fallen out into the mine. This rises as an uneven cylindrical pipe for about 5m, where there is an arched roof in the remaining fill. Its walls expose a section through clay infill and limestone bedrock, and, if the remaining clay were removed, a complex morphology fretted by dissolution would be revealed. Other cenotes can only be seen as incomplete cross-sections at roof level, as they are concealed by debris cones in the mine galleries. It may be significant that six out of the eight cenotes observed in 2005 lie on or close to faults. However, they show clear dissolutional opening and are not fault fissures. Collapse breccias have not been seen.

The fills are now seen as cones of slumped debris derived from the lava flow above roof level. They consist of greenish or yellowish, lumpy clay, with some blocks that are reddish-brown and more solid, and appear to represent altered or weathered lava (Figs 6 and 7). Some debris cones have more brownish clay in their upper parts. Fallen limestone blocks are locally included, and there are layers of re-deposited lime sediment possibly derived from adjacent limestones. All the pits and fills are thought to extend up to the base of the lava, but mining has not exposed this part of the section. No evidence of concentrated leakage through the lava has been seen. The bases of the pits are not exposed, being below the mine floor. One cenote recorded by Oakman (1984) appeared to have a rounded base (later removed by mining) with no horizontal cave passages in the visible wall sections. On the other hand, one pit appeared to spread out into clay-filled dissolution fissures along the bedding, but excavation would be needed to prove this. Elsewhere in the mine, an unseen cave beneath the floor swallowed the entire product of the

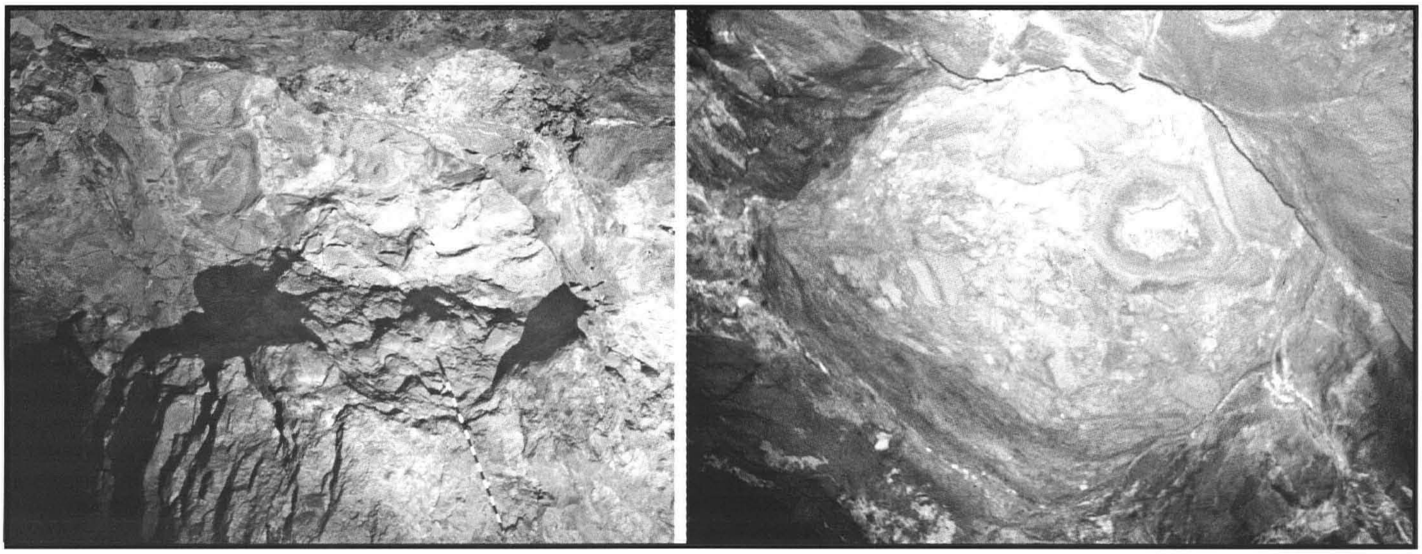


Figure 5. *Two cenotes in the roof of Middleton Limestone Mine, seen looking steeply upwards. The remaining clay fill forms a stable roof in each. The mine gallery is visible at the lower left of the left image, where the scale stick has 10cm graduations (Photos by C Oakman).*

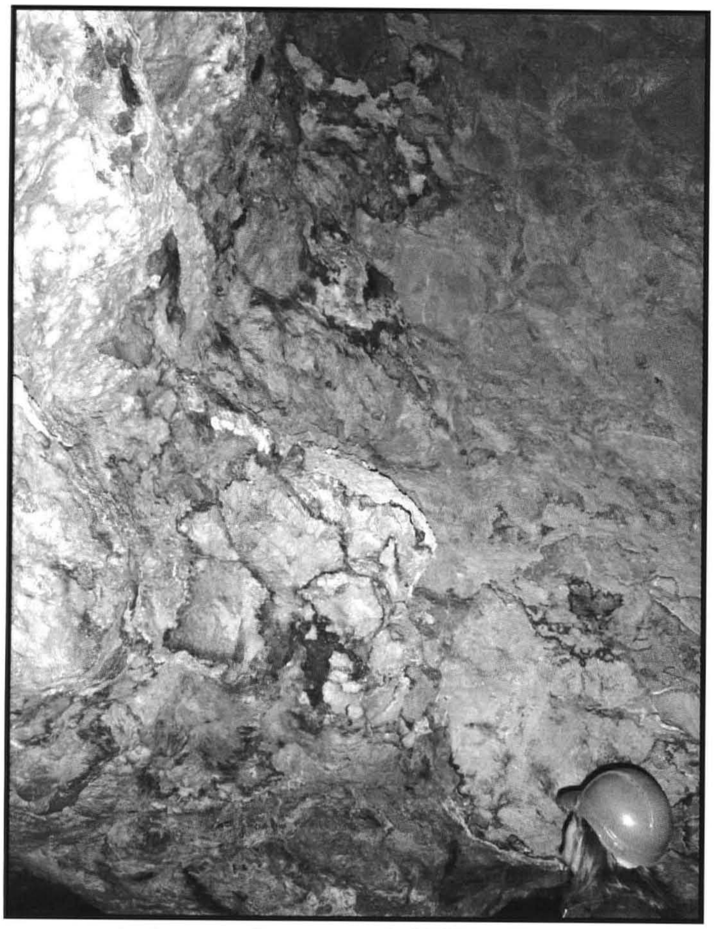


Figure 6. Looking upwards in a cenote in Middleton Limestone Mine. The walls are mainly yellow and green clay. Pale areas are protruding limestone. The mine roof is visible in the bottom left corner. (Photo by T Waltham).

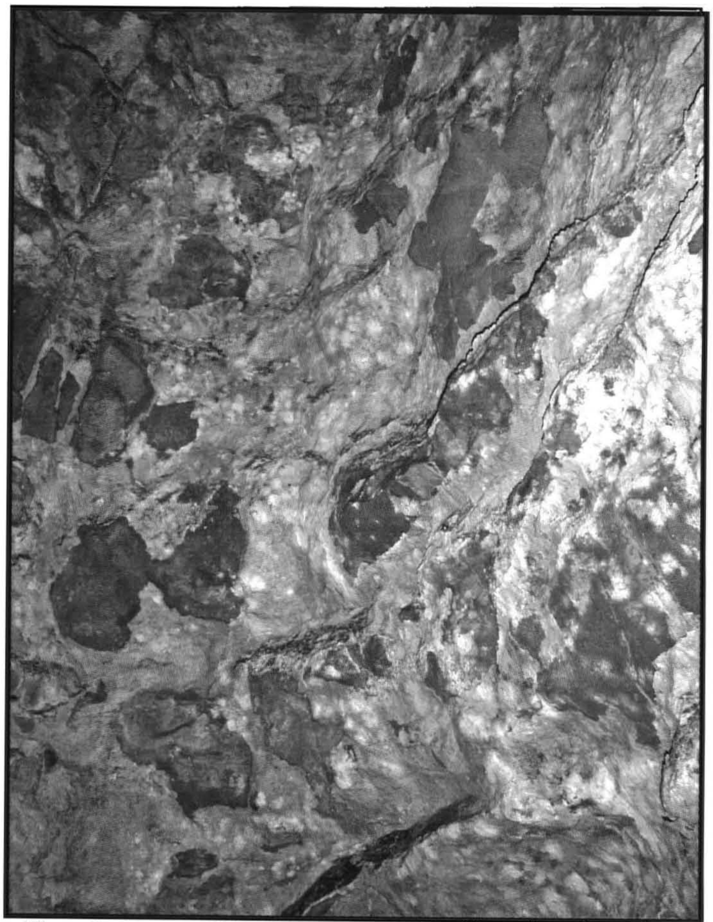


Figure 7. A section of the fill in one cenote. The fill is yellow clay enclosing darker blocks of weathered lava. The white area is part of the limestone wall (Photo by T Waltham).

blasting of the mine level. Walkden (pers. comm.) saw other pits filled with lava boulders.

As far as can be determined the pits are all capped by the lava flow or by a wayboard, but no evidence of a volcanogenic mechanism for their origin has been found. Instead it seems likely that they originated before the lava flow or wayboard was deposited. The fills of blocks and clay derived from these indicates stoping from below in parallel with the roof stoping of typical cenotes.

The site of the Middleton Mine cenotes is about 1.5km from the southern margin of the Derbyshire block, so the flank margin hypothesis of Mylroie and Carew (1990) may be applicable. The Limestone Mine is within a zone of faulting, with downthrows towards the block margin. These faults may represent successors to block margin faults such as are seen in the Bahamas (Smart *et al.*, 1988), though no linear cenotes have been seen at Wirksworth. Neither growth-faulting nor any other syn-sedimentary faulting has been recognized at Wirksworth.

The fills are permeated by minor deposits of galena, baryte and calcite in joints, confirming that the pits were of pre-mineralization age. Mineralization is now generally regarded as being Late Carboniferous in age (Plant and Jones, 1989), and previous suggestions of Permian or Triassic ages of mineralization are no longer in favour.

In the 1970s limestone was extracted from a sub-level beneath the main workings. The sub-level cut into lower carbonate cycles with several thin wayboards. These plastic clays caused instability that resulted in collapse of several hectares in 1975, leaving a substantial crater on the hill top. Amongst the factors alleged to contribute to the collapse were undetected "caves" below the sub-level, but these were not seen and must remain hypothetical.

Several other possible cenotes lie near Wirksworth. One is in a carbonate mud-mound within the upper part of Monsal Dale Limestone, exposed in the "reef" quarry of the National Stone Centre about a kilometre to the east of the limestone mine. This was described by Gutteridge (in Cossey *et al.*, 2004), who reported that the walls were partly encrusted with brownish fibrous calcite probably representing ancient speleothems deposited under vadose conditions, of unknown but pre-mineralization age. The floor and any possible lateral caves are concealed below the quarry floor. This example is close to the top of the Monsal Dale Limestone, where there is evidence of emergence prior to deposition of the Eyam Limestone. The only visible fill is a much later ochreous clay, probably representing inwashed loess of Pleistocene age. The Eyam Limestone is no more than a few metres thick near this mud-mound; erosion precludes arguments that the cenote development might have been downwards through the Eyam Limestone.

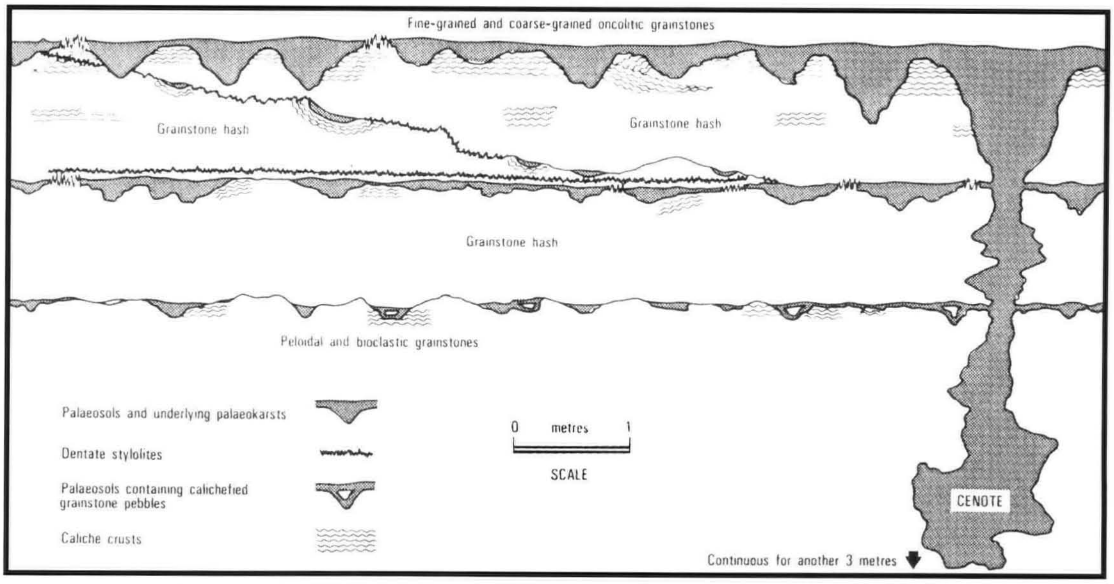
Prospect Quarry near Grangemill cut through several wayboards lying on pot-holed palaeokarstic surfaces. One pit was traced to a depth of at least 6m, and this might also be a small cenote (Oakman, 1984) (Fig.8). Pot-holed surfaces are also widespread beneath wayboards in the nearby Longcliffe quarries.

Other possible cenotes are said to have been encountered in the limestone quarries south of Buxton, but they have been completely removed by quarrying. Two pits filled with derived wayboard material in Millers Dale Station Quarry (Cope, 1933; Walkden, 1977, 1979) may have been cenotes. They are now degraded and heavily overgrown, so that details are no longer determinable, but Cope (1933) described rounded boulders at the bottom. These suggest that they are not cenotes, but it is difficult to see what process could result in rounded limestone boulders in a narrow isolated pit several metres below a wayboard. Another pit lies some 2km to the east, in the Litton cutting on the Monsal Dale trail, at about the same horizon, and was described by Walkden (1970) as having volcanic and limestone pebbles and a few limestone boulders up to 30cm across. It is a shallow basin in shape unlike the usual cenotes and much more suggestive of a scour hollow.

DISCUSSION

The pits with their fills were termed cenotes and illustrated by Oakman and Walkden (1982) and by Oakman (1984) on schematic sections of the cyclic limestones. However, no descriptions were

Figure 8. Diagrammatic sections of wayboards, palaeosols, potholed surfaces and a small cenote in Prospect Quarry, Grangemill (from Oakman, 1984).



included in their reports and no comparisons to the Mexican cenotes or Bahaman blue holes were offered.

Most of the cycles of carbonate sedimentation were concluded by emergence, usually with volcanic material deposited immediately above. Although the fills are dominantly volcanic clays, the evolution of the pits from an emergent limestone surface does not appear to have any direct link to igneous activity. The Cookstown examples in Ireland (noted below) have no associated volcanic material, and nor do other palaeokarstic pit horizons widespread through other regions of England and Wales have any volcanic associations (Vanstone, 1998). If the pits are indeed cenotes, whereas the emergence of the relevant cycles might have been as much as the depth of the observed pits, up to some 30m, it could have been much less, allowing the possibility of speleogenesis in a shallow phreas.

The Wirksworth cenotes are thought to have developed from the bottom upwards by stoping, with collapse of lava fills into them. None of the Wirkworth pits has revealed material collapsed from limestones above the lava or wayboards. Walkden (pers. comm.) suggested a comparison of the Wirksworth pits with solution pipes in the Chalk, but the latter developed from the top downwards, with fills derived from the overlying Tertiary sediments. In Yucatan and the Bahamas, cave development was from the bottoms of the cenotes upwards. Such speleogenesis would be phreatic initially through pene-contemporaneous, porous, partly lithified carbonate sediments. Later development was at a contemporary halocline. The model requires slow movement of percolating water towards the flank margin of the massif. Whether the emergence was enough for vadose modification of any cave passages cannot yet be demonstrated at Wirksworth. Apart from the possible example noted by Gutteridge (2003), no evidence of speleothems has been found in the fills.

During the Carboniferous Period there was a series of worldwide sea-level changes due to the effects of repeated glaciations, and it is tempting to suggest a link between glacial sea-level oscillations and the Wirksworth cycles. However, current knowledge of glacial oscillations is probably not detailed enough for any correlation of the Wirksworth cycles to be established.

If the analogy with Mexican cenotes and Bahamian blue holes is pursued, the Wirksworth cenotes should not be blind pits as illustrated by Oakman and Walkden (1982) and by Oakman (1984). Instead there ought to be cave passages leading off, filled with collapsed volcanic detritus or pene-contemporaneous derived lime sediment. If the Yucatan analogy is followed further, the morphology of such caves should approach that of phreatic tubes, but if the Bahamian blue hole analogy is considered, rather more irregular palaeocaves might be expected. Neither type has yet been found at Wirksworth, which might suggest that the pits there are not cenotes. However, it has been suggested that caves lie below the mine floor, and that these were a contributory factor in the partial mine collapse. Whether the alleged cenotes developed by fresh-water drainage and tidal pumping must remain uncertain at present. No evidence of development of cenotes from potholed surfaces as suggested by Vanstone (1998) has been found. No palaeogeographical evidence for adjacent hills, as in Yucatan, is known, so the Wirksworth examples would be more likely to follow the Bahamian pattern.

Broadly comparable karstic pits exposed by quarrying at Cookstown in Northern Ireland may also be cenotes. They are in limestones of the same age as those at Wirksworth, i.e. Asbian (Mitchell, 2004; Simms, pers. comm.). Pits up to 20m deep have been seen, with fills of fine clay of possible aeolian origin. None has been bottomed at the present state of quarrying, so it is not known whether lateral caves are present. The Cookstown pits differ from those at Wirksworth in that no volcanic horizons have been found, and the pits are capped by horizontal limestones of Brigantian age. The Carboniferous Limestone of Trwyn Dwlban, Red Wharf Bay, Anglesey, exhibits shallowing-upward cycles like those at Wirksworth. Some 160 sandstone-filled pipes within the sequence are exposed on the foreshore and in the cliffs (Walkden and Davies, 1983). The pipes occur in nine different carbonate sediment cycles spread through a thickness of about 150m. No volcanic horizons are present. Pipes are generally 1 to 1.5m in diameter and up to 3m deep. No underlying caves are known. The pits are mostly on palaeokarstic surfaces developed on platforms adjacent to wide shallow channels cut in the contemporary limestone surfaces. They are clearly karstic solution pits, but no evidence has been found that links them to underlying caves, and hence they may not be considered cenotes.

CONCLUSIONS

Whilst the evidence is incomplete, the potential explanation of the pits and their fills is that a process analogous to those at the Mexican cenotes or Bahamian blue holes operated during Dinantian times. On the basis of the evidence presented it is argued that these features should be regarded as palaeo-cenotes. A possible alternative term is "palaeo-collapse sinkholes" comparable with that illustrated in Table 2.1 in Waltham *et al.* (2005). The factors required in the genesis of the Wirksworth cenotes are shallowing-upward cyclic carbonate sedimentation with intermittent emergence and a karstic/phreatic drainage system. Thus, the environment at Wirksworth is directly comparable with those in Yucatan and the Bahamas. If the emergence was sufficient, then karstic processes could have developed both cenotes and attendant phreatic caves. If there was enough emergence, vadose processes might have modified the caves, but little evidence of this has been found. Similarly any limestone formations elsewhere that underwent cyclic emergence could be suitable sites for the development of cenotes at any time in the geological past.

There is an extensive literature on palaeo-caves, much of it related to discoveries made in the course of oil-well drilling (e.g. Loucks, 1999), but reference to fossil cenotes seems to be absent. A perusal of compendium works on palaeokarst (e.g. Esteban and

Klappa, 1983; James and Choquette, 1988; Bosak, 1989; Gunn, 2004) reveals various references to solution pits and other karstic features but no descriptions of fossil cenotes of the Yucatan type or of fossil equivalents to Bahamian blue holes.

Smart *et al.* (1988) have described fissure fills in "neptunean" dykes found during diving in Bahamian blue holes. These fissures appear to be faults up to several kilometres long parallel to the margins of the Bahama banks with no clear comparison to the Wirksworth cenotes. Fossil equivalents of these fissure blue holes have been described from the Permian Capitan reef limestones of New Mexico (Kosa *et al.*, 2003), but no comparable fissures have been found at Wirksworth.

It is possible that fossil cenotes and their related caves are widespread features of palaeokarst surfaces throughout the stratigraphical column but have not been recognized as such.

ACKNOWLEDGMENTS

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Geological and morphological observations in the eastern part of the Gran Caverna de Santo Tomás, Cuba *(results of the "Santo Tomás 2003" speleological expedition).*

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Abstract: This paper deals with activities of the "Santo Tomás 2003" expedition, carried out in December 2003 by Italian and Cuban cavers at *Gran Caverna de Santo Tomás*. The *Gran Caverna de Santo Tomás* is the most famous karst cave on the island of Cuba and it is now more than 46km long. It lies in the Pinar del Rio Province (western Cuba), and comprises seven different levels of caves, the lowest of which is active and occupied by the *Arroyo de Santo Tomás*. Its location is part of the *Sierra de los Organos* mountain ridge, a classical example of cone karst, which is characterized by the presence of a number of *mogotes*, typically isolated carbonate hills with vertical walls and rounded tops. The *Gran Caverna de Santo Tomás* is developed within the *Sierra de Quemado*, and has been explored since 1954 by Antonio Núñez Jiménez, the well-known Cuban explorer and caver. Activity during the 2003 expedition included topographical survey, and geological and morphological observations in the sector of *Gran Caverna de Santo Tomás* closer to the *Valle de Santo Tomás*. Most of the work was devoted to the *Salón del Caos*, one of the largest caverns in the karst system, which enlarged due to a great number of huge rockfalls from the roof and walls of the original chamber.

Key words: Cuba, expedition, tropical karst, Santo Tomás.

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INTRODUCTION

The karst system of the *Gran Caverna de Santo Tomás*, the most famous cave in Cuba, was declared a National Monument on June 5th 1989, reflecting its remarkable natural interest and its karst, palaeontological, historical and archaeological relevance. This complex system comprises seven main levels of sub-horizontal galleries. The lowest of these is active today and lies at an elevation of about 125m above sea level, corresponding to the subterranean course of the *Arroyo Santo Tomás*. Exploration of the system began in 1954, thanks to Antonio Núñez Jiménez, the well-known Cuban explorer and caver, who is considered to be the "Father of the Cuban Speleology". At present the overall length of explored passages in the Santo Tomás cave system is more than 46km, but many galleries still show possibilities for further exploration.

Among the most remarkable discoveries in the cave, it is worth mentioning the discovery of palaeontological remains in several caverns within the karst system (*Cueva de Otero, Cueva de la Incognita, Cueva de Megalocnus*), as well as archaeological finds including the famous wall paintings in the *Galería de Garcia Valdés* of *Cueva de Mesa* (Núñez Jiménez, 1975). The latter were not an isolated find on the island, as several other Cuban caves contain wall paintings (Rivero de la Calle, 1958).

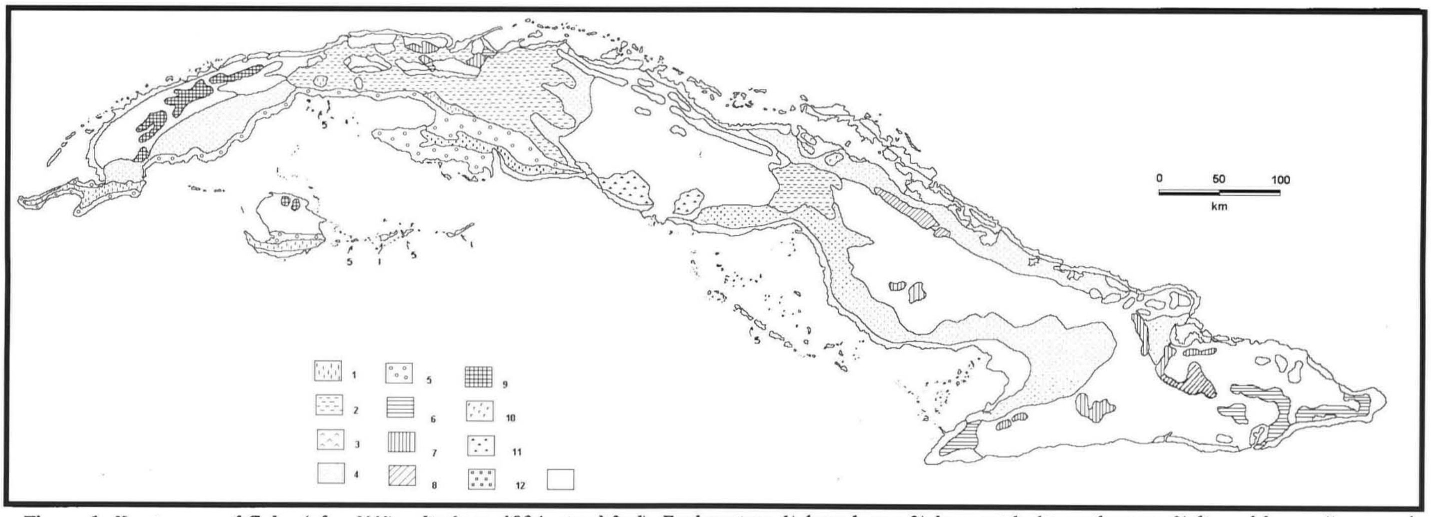

Figure 1. Karst areas of Cuba (after Núñez Jiménez, 1984, simplified). Explanation: 1) bare karst; 2) karst with thin soil cover; 3) littoral karst; 4) covered karst; 5) ponds; 6) highplain; 7) dome karst; 8) tabular karst; 9) cone and tower karst; 10) dome karst; 11) tabular and cone karst in schists and metamorphosized limestones; 12) serpentinite; 13) non-karst areas.

In December 2003 an Italian-Cuban expedition took place to the *Gran Caverna de Santo Tomás*, based around the facilities provided by the *Escuela Nacional de Espeleologia* "Antonio Núñez Jiménez", in El Moncada village. The expedition, which developed from an original idea by members of the Gruppo Puglia Grotte from Castellana-Grotte (Bari, Italy) and involved several others Italian and Cuban caving clubs, was directed primarily toward re-definition of the topography of part of the karst system, and geological and morphological observations (Parise, 2004). The present paper summarizes the main results of the work carried out.

THE CUBAN KARST

The geological structure of Cuba (Fig.1) is among the most complex in the Caribbean. Based upon recent studies (Iturralde-Vinent, 1998; González *et al.*, 2003), this can be rationalized into two main structural zones: the fold belt and the neo-autochthonous terranes. The former encompasses terranes of both continental and oceanic origin, that were formerly parts of the North American, Caribbean, and, probably, Pacific tectonic plates. These rocks range in age from the Neoproterozoic (940–1000 Ma) to Late Eocene (37 Ma). On the other hand, the neo-autochthonous terranes comprise rocks younger than Late Eocene, which were formed in the position now occupied by Cuba. At the time that these younger rocks were deposited Cuba was already part of the passive southern margin of the North American plate.

In southwestern Cuba the most important structure is the Pinar fault, which separates the wide *Pinar del Rìo* Plain from the so-called "south-western terranes" (Guaniguanico, Escambray, Pinos), where the Santo Tomás area is located (Nemec *et al.*, 1966). These terranes are characterized by allochthony and a specific grade of metamorphism, and they include tectono-stratigraphical elements deriving from the continental margin of the Maya block (Yucatán Peninsula). Different lithological and stratigraphical sections have been identified within these terranes. They crop out along NE/SW-striking zones, which resulted from N or NW thrusts, later affected by additional deformation phases (Pszczolkowski *et al.*, 1982).

The Santo Tomás karst system is developed within one of the most ancient of these stratigraphical successions, namely the stratified limestones with chert interlayers of the Guasasa Formation. This overlies the thin beds of the Jagua Formation. The two formations (Guasasa and Jagua) represent the Late Jurassic (Mid Oxfordian – Tithonian) carbonate sequence in the *Sierra de los Organos*, corresponding to the Artemisia Formation in the *Sierra del Rosario* (Fig.2). The Jagua Formation is 160m thick, and represents the basal part of this sequence, being composed of carbonate rocks with intercalations of clay, which were deposited in a shallow marine environment. In contrast the Guasasa Formation represents the upper part of the Late Jurassic sequence, with an overall thickness of 650m. It comprises two distinct members, both essentially made up of carbonates, with the San Vicente Member at its base and the American Member at its top (Pérez Vázquez and Melgarejo, 1998).

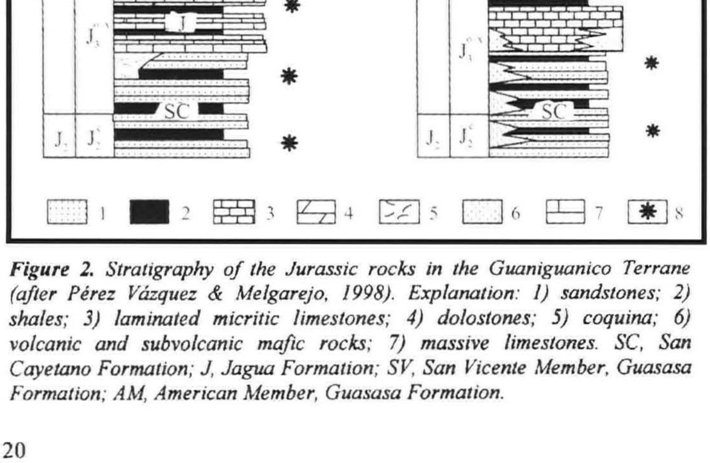

Figure 2. *Stratigraphy of the Jurassic rocks in the Guaniguanico Terrane (after Pérez Vázquez & Melgarejo, 1998). Explanation: 1) sandstones; 2) shales; 3) laminated micritic limestones; 4) dolostones; 5) coquina; 6) volcanic and subvolcanic mafic rocks; 7) massive limestones. SC, San Cayetano Formation; J, Jagua Formation; SV, San Vicente Member, Guasasa Formation; AM, American Member, Guasasa Formation.*

Karst landscapes are widespread on Cuba, due to the presence of soluble rocks, which crop out over 60% of the island (Fig.1). Geology, morphology and tectonics, combined with the local climate, have produced a great variety of landscapes typical of tropical karst. Hence Cuba has become one of the most remarkable examples of tropical karst, and has long attracted the attention of cavers and karst scientists (Lehmann, 1958; Panos and Stelcl, 1968a; Ducloz, 1970; Varona, 1970; Geze and Mangin, 1980; Panos, 1986; Schenck *et al.*, 1999; Corella *et al.*, 2000; Tyc, 2003).

Among the many karst features of the island, the cone karst deserves special mention, and some sectors of Cuba have become type examples, well known all over the world (Gardner, 1987; Tarhule-Lips, 2003). However, Cuba does not only display landforms related to cone karst, but also a great variety of other surface karst landscapes, from coastal karst plains, to dome karst, tower karst and tabular karst (Núñez Jiménez, 1967, 1984; Núñez Jiménez *et al.*, 1968). The presence of so many different karst morphologies within quite a small territory emphasizes the crucial role played by the geological-structural setting in the genesis and evolution of the Cuban karst. The landscape is generally structurally controlled, except where selective erosion prevails due to contact between lithologies showing different response to erosional processes. Karst landscape evolution must therefore be studied by considering the effects of karst processes in combination with effects deriving from other processes.

Sierra de los Organos is part of the karst ridges and mountains of the western sector of Cuba (Fig.1). It is limited by *Sierra del Rosario* to the east, and the *Pinar del Rìo* plain to the south and southwest, the latter being an example of covered karst, with thick alluvial deposits. *Sierra de los Organos* consists of narrow, parallel limestone ridges with rounded tops, and intervening plains or smooth hills. The ridges, extending from *Guane* to *San Diego de los Baños,* with an overall length of about 94km, are made of dark limestones. Several *hoyos* and fertile valleys, locally with narrow communicating gaps, are present, but more commonly connection between the valleys occurs underground. Islands of carbonate rock emerge from the plains, filled by fluvial deposits and by materials derived from the physical and chemical weathering of non-carbonate rocks. Due to its surface karst features, *Sierra de los Organos* has gained the reputation of being one of the most typical examples of mogote karst, deriving from the combined actions of corrosion and erosion. The Viñales Valley is one of the most striking landforms in the area, and has for years been the object of many different interpretations. Originally considered a marginal karst plain, according to Lehmann and his co-workers (Lehmann *et al.*, 1956), who considered the valley to be developed entirely within the limestones of the Viñales Formation, it was later interpreted differently. In fact, other authors doubted the continuity of the limestones, as the valley is developed at the contact between carbonates and non-karstifiable rocks. In particular Panos and Stelcl (1968a, b) were the first to point out the role played by juxtaposition of rocks with contrasting responses to erosion in adjacent outcrops. They actually described the *Sierra de los Organos* as a landscape where karst processes modified features that were originally created by the action of selective erosion across complex structural settings at the contact between shales and limestones. On the other hand, Núñez Jiménez himself (1984) described the Viñales Valley as an intramontane karst valley, with isolated limestone *mogotes* emerging from a mostly clayey plain, thus highlighting the double action of the erosional processes in the clays and dissolution in the carbonate rocks. In this way, the great importance of weathering, which acted in producing huge amount of materials from both the limestones and the shales, was considered together with the development of karst processes.

Mogotes are among the most typical landforms in the *Sierra de los Organos* and the Viñales Valley (Racovitza, 1971; Panos, 1980a). They are carbonate hills and ridges, showing vertical or steeply-dipping walls, with a relief ranging from some tens of meters to more than 100m above the flat plains (Fig.3). *Mogotes* are formed by a combination of several factors, including lithology, tectonics, a high intensity of karst processes, and slope movements. Other surface karst landforms, widespread in the *Sierra de los Organos* and the nearby *Sierra del Rosario*, are isolated dolines known as *hoyos*. These are generally related to collapse, and later evolution, of the roofs of ancient underground cavities. *Uvalas*, which are the result of the coalescence of multiple individual dolines, and which generally present an irregular boundary in plan, are also widespread.

GRAN CAVERNA DE SANTO TOMÁS

The *Sierra de Quemado* is an 8.5km-long, NNW/SSE-striking ridge. In this sector, the *Sierra de los Organos* veers from the E–W trend that is typical of many Cuban mountains to strike approximately southwards, in the shape of a great arch. The *Valle de Quemado* bounds the *Sierra de Quemado* to the west, thus forming, at a smaller scale, a structure similar to that of the Viñales Valley. As regards local hydrology, the *Rio de la Caoba* runs between the *Sierra de Quemado* and the nearby *Sierra de Cabezas* to the west. Once having received the subterranean waters of the *Arroyo Santo Tomás*, the *Rio de la Caoba* joins the *Cuyaguateje* River to enter the *Valle del Sumidero* and, subsequently, the *Valle de Luiz Lazo*, towards the southern coast of Cuba. On the eastern slope of *Sierra de Quemado* the blind valley of Santo Tomás (about 1km wide, and 3.5km long, running from northnorthwest to southsoutheast) has the *Arroyo Santo Tomás* as its main watercourse.

The *Gran Caverna de Santo Tomás* is a complex karst system consisting of seven levels of mostly sub-horizontal caves, with a total length of explored galleries that is over 46km (Figs 4 and 5). The lowest level of the system is active at present, and is occupied by the *Arroyo Santo Tomás*. A small part of the cave (*Cueva de las Avispas*, located within the sixth level of the system, in the Santo Tomás Valley) has been exploited as a show cave since 1994, and is managed by a State society (Parise and Valdes Suarez, 2005). Together with the other Cuban show caves, the *Cueva de las Avispas* plays an important role, both in promoting the development of tourism devoted to the natural features of inland Cuba, and in educating local people to respect and safeguard the local karst environment (Panos, 1980b; Diaz Diaz *et al.*, 1989; Day, 1993; Kueny and Day, 1998). In this latter regard, it could thus represent a suitable site to initiate a cave-monitoring program (Travous and Ream, 2001; Day and Koenig, 2002; Kranjc, 2002), such as those already developed worldwide in other karst caves that are wholly or partially exploited as show caves (Carrasco *et al.*, 2002).

History of exploration

The *Gran Caverna de Santo Tomás* has experienced many different uses during its history. It initially provided shelter for the Cuban aborigines. More recently it provided a refuge for the *Cimarrones*,

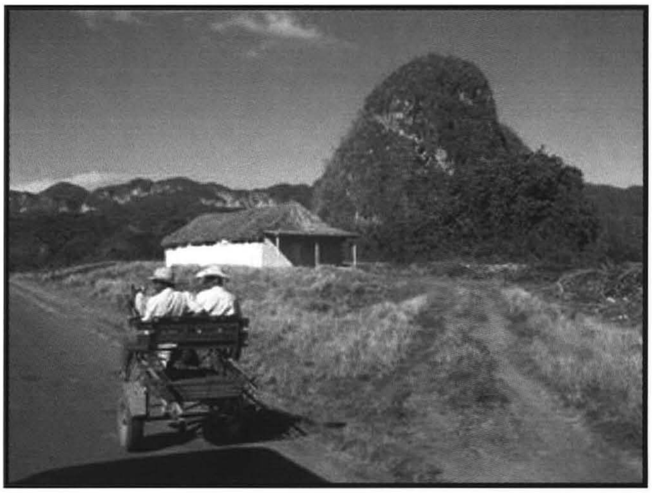

Figure 3. *View of the Viñales Valley, showing typical mogotes (photo: M Parise).*

who were slaves that had escaped from the plantations where they were forced to work. They found sanctuary in many caverns at Santo Tomás during the first half of the 19th century. Farmers from the nearby valleys have frequently used some of the watercourses flowing through the Santo Tomás caverns as a drinking water supply during periods of drought. In addition, they extracted guano as a fertilizer for the tobacco fields, and they used the natural caves as shelters during cyclones. As further evidence of the importance of the *Gran Caverna de Santo Tomás* in the lives of the local inhabitants, it can be remarked that, due to its great size and ease of access from the *Valle de Quemado*, the *Cueva del Salón* was used for holding dances and parties.

Santo Tomás also possesses a remarkable value from the historical point of view. January 1st 1959 is an historic date in Cuba, marking the triumph of the Cuban Revolution. In that same year, on August 31st, Commander Fidel Castro visited the Santo Tomás Valley and its fabulous karst system, guided by his brother-in-law, the expert Antonio Núñez Jiménez. On that occasion the first Rural Militia of the Cuban Revolution was founded at *Cueva de Mesa*, and named *Los Malagones*, reflecting the name of its chief, Leandro Malagòn.

As regards the caving explorations at *Gran Caverna de Santo Tomás* (Valdes Suarez, 2005), Antonio Núñez Jiménez began the first research activities in 1954, from the *Cueva del Salón*, which was at that time the most famous cavern of the system. On January 1st 1955, the full width of the ridge of the *Sierra de Quemado* was traversed underground, with the explorers passing through the *Cueva de Represa*. With continuing activity, the extent of the explored system increased, and Santo Tomás was becoming one of the more important caves in the Americas.

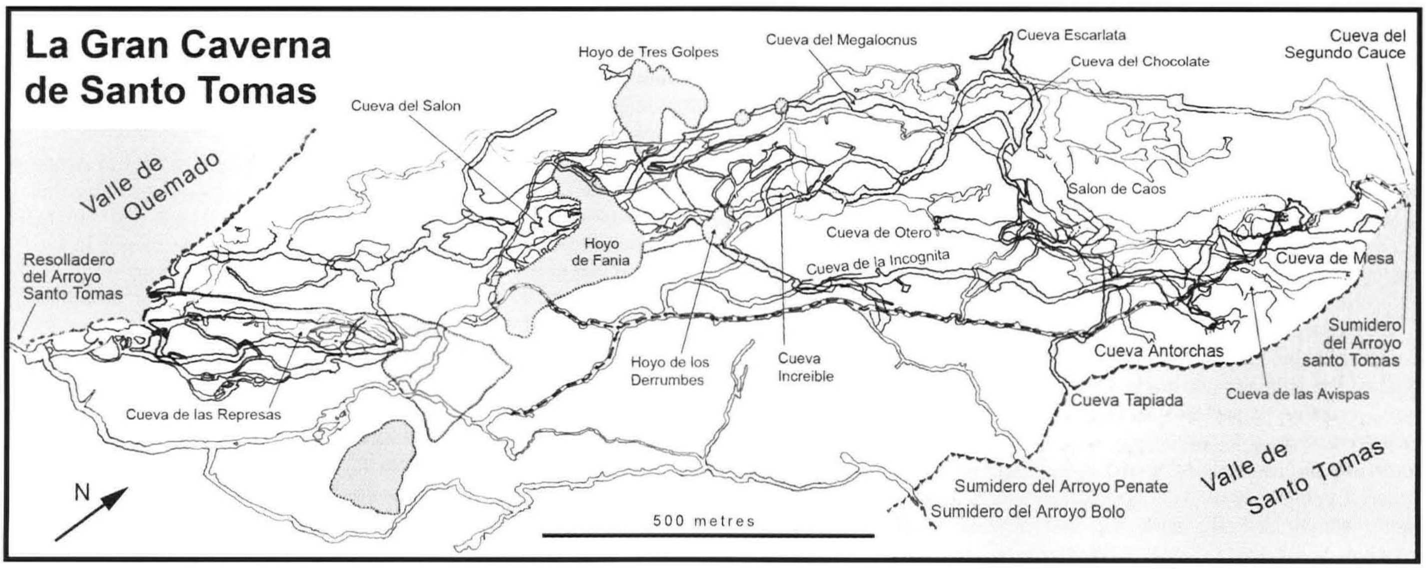

Figure 4. *Overall plan of the Santo Tomás karst system (elaboration and drawing by U Del Vecchio and R Tedesco).*

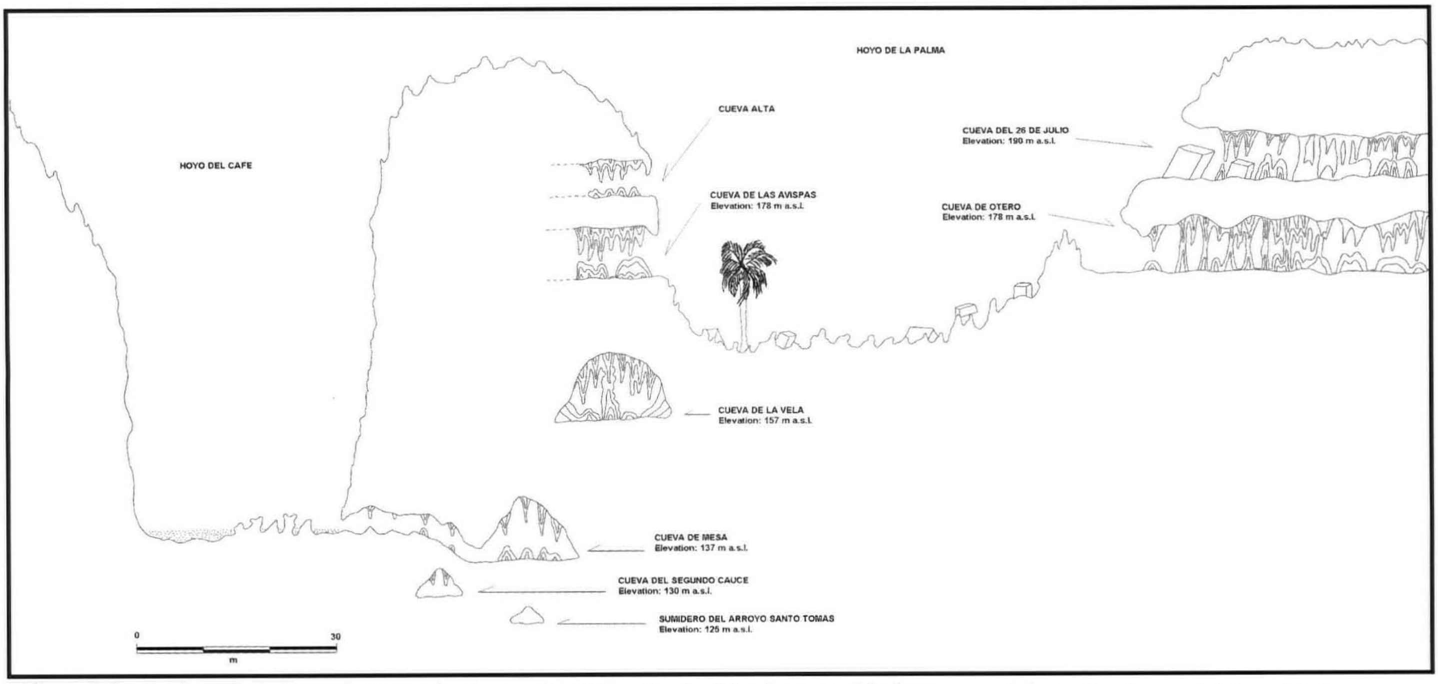

***Figure** 5. Section through the Sierra de Quemado, showing the different levels of development of the karst system (after Nuñez Jimenez, 1990, modified).*

The second exploration phase began in August 1959, and proceeded until 1970. New galleries and caverns were explored, and dye-tracing investigations were undertaken for the first time. Fluorescein was used to confirm the connection between the *Sumidero de los Cerritos* and the *Cueva de Represa* and, in 1969, to prove underground flow to the spring of *Rio Frio* in the *Valle de Quemado*. At the end of the second phase of research the explored caves had reached a total length of 25.550km (Núñez Jiménez, 1990).

The third phase began on September 19th 1983, with an expedition whose aim was to study the effects produced by the flood related to the June 14th 1982, cyclone. On that occasion the water flooded the Santo Tomás Valley to a level 14m above the *Sumidero del Rio Santo Tomás*, and several caverns within the system were inundated, including the *Segundo Cauce* and the *Cueva de Mesa*. During this exploration phase, new caves were found, including the *Cuevas Borras y Valcárcel*, and even the first shaft of the karst system, with a depth of 16m. In April 1985 the *Escuela Nacional de Espeleologia* was inaugurated at the village El Moncada, close to some of the main entrances to the Santo Tomás system. Explorations carried out on the occasion of the school's foundation brought the total length of the cave system to 32.545km.

Eventually, the fourth exploration phase started from the eastern slope of the *Sierra de Quemado* at the beginning of the 1990s, resulting in over 45km of the system being explored.

SANTO TOMÁS 2003

The eastern slopes of the *Sierra de Quemado*, in whose foothills the *Escuela Nacional de Espeleologia* is located, is marked by a number of cave entrances at various heights. The section of the karst system selected for the activities of the "Santo Tomás 2003" expedition was the sector closest to the Santo Tomás valley, focussing on one of the most important caverns of the entire system, the *Salòn del Caos*.

Once within the cave, an enormous richness and variety of concretions and subterranean karst morphology is evident, as well as the prominent influence of tectonics in controlling the main galleries of the system. Joints and faults are extremely widespread and recognizable in environments not covered by secondary deposits. Most of the galleries are marked by tectonic discontinuities in their roofs. Locally, when two or more fracture systems intersect a meandering passage morphology may be observed, as is the case at *Antorchas*, in the final sector of *Tapiada*, and in part of the *Segundo Cauce*. Even the *hoyos* (collapse dolines with a generally circular shape, which alternate with the underground sections in many branches of the Santo Tomás system, Fig.6) are commonly aligned to follow the main tectonic trends of the area (SW–NE, and N–S). For example, *Hoyo de Fanìa, Tres Golpes, Hoyo de los Derrumbes, de los Platanos, del Yagrumìn,* and *del Aire* display such a linear distribution. The collapses that led to the formation of these dolines left relict galleries as hanging levels, present at greater elevations along the slopes of the *hoyos*.

The strong tectonic influence on the development of Santo Tomás is also fundamental to the main theories presented by Nuñez Jiménez (1990) to explain the speleogenesis of the karst system: the theory of "crossing galleries" *(teorìa de las galerìas en crucero)*, and the "T theory". In the former case, Nuñez Jiménez refers to the situation where an intersection (crossing) occurs between two galleries that began their development at different levels. The Cuban explorer claims that, even if not readily visible, continuations of the original galleries must be present, probably concealed by secondary cave deposits. When, on the other hand, two galleries developing at the same level intersect, the resulting plan form is that of a letter "T", rather than a cross as in the previous case. Even though very simple, these two fundamental ideas allowed Nuñez Jiménez to "target" his research and discover many of the presently known ramifications of Santo Tomás.

The expedition effort was concentrated within parts of the lower levels of the system, mainly the second and the third levels, with some local visits to the active cave level. Water was found in some of the normally inactive passages, as for example in the galleries of *Segundo Cauce*, which still functions as an active epiphreatic gallery when the *Sumidero de Santo Tomás* is unable to transmit the full amount of water during flood events. Thus, the most typical phreatic galleries, which range from circular to elliptical in shape (Fig.7), were observed in this part of the system.

Water was also present at the confluence between the *Cueva de Mesa* (which is well developed along a tectonic lineament) and the *Cueva Tapiada*. The Liduvina Lake, with many gours and calcite concretions covering the flooded pavement, is the greatest lake in the system. It can easily be passed by walking close to the rock walls on its western side. The *Cueva Tapiada* is the gallery that probably shows the greatest variety of hypogean morphologies, from small and narrow passages, rich in such concretions as *paletas* (palettes or shields), to wide and high galleries that are partly filled by allogenic deposits (mostly cobble beds) and by rockfalls, to, eventually, many lakes. Beyond the aforementioned Liduvina Lake, the *Cueva Tapiada* leaves its linear course to assume a meandering pattern, while also showing a net increase in the amount of deposits deriving from instability processes within the cave. The section of the *Cueva Tapiada* where the narrow route towards the *Salòn del Caos* starts among huge unstable blocks is one of the areas most intensely

Figure 6. *One of the many hoyos that characterize the karst system (photo: F Maurano).*

Figure 7. *Gallery in the Segundo Cauce (photo: A Marangella).*

affected by instability. Passing beyond the link to the *Salòn del Caos* (which is to the right), and keeping going along the main trend of *Cueva Tapiada*, leads to another of the treasures of Santo Tomás: the Rhinoceros Lake, which is named after a rock at its lower entrance that resembles the face (horn included) of a rhinoceros.

The *Salòn del Caos* is a key feature of the Santo Tomás karst system (Fig.8). Here many of the galleries that developed at different levels join to form this huge cavern, whose irregular floor consists of large blocks deriving from rockfalls and breakdown processes. The cavern is over 100m wide (Fig.9). Estimation of the height is not an easy task, as the huge rockfall deposits make identification of its real floor difficult. However, it is at least 20m high for most of its plan. The Caos is a veritable labyrinth, and its topographical survey required a team completely dedicated to this aim!

The *Antorchas* gallery provides the easternmost access point to the cavern, and lies at about the level where rockfall deposits begin to give way to mud and speleothems, and the bedding of the rock substratum also becomes discernible. The rockfalls have probably intersected several different levels of the system, connecting the multi-level galleries into a unique room, now largely filled by fallen rock detached from the original vaulted ceilings and walls. In fact, from the centre of the cavern, it is possible to see several branch passages at different levels. Some are clearly visible, with wide entrances *(Cueva Escarlata, Cueva Increible)*, others lie in the lower reaches of the cavern, among the rockfall deposits *(Galeria del Chocolate)*, or find a way into the chamber through very narrow passages *(Cueva Tapiada)*.

As previously stated, the *Salòn del Caos* undoubtedly marks a very important site in the Santo Tomás karst system, located as it is in an area where several discontinuity systems and the galleries that they have guided intersect. A result of this high density of fracturation in the rock mass was the occurrence of huge rockfalls that connected the different galleries and created a unique cavern. However, to date the cavern roof has remained stable and thus the development of a big open cavern, that is a *hoyo*, has so far been avoided.

Some of the galleries that are connected to the *Salòn del Caos* are worthy of further mention. Among these, the *Cueva Increible* (Incredible Cave), with a length of some 0.5km, is so named because of the remarkable richness in variety and colour of the concretions that it contains. Helictites are particularly widespread in some parts of the cave (Fig.10). The importance of secondary cave deposits in the Santo Tomás karst system and more generally in the *Sierra de los Organos*, has been highlighted by several researchers in the past few tens of years (e.g. Núñez Jiménez, 1958; Fagundo *et al.*, 1971; Fagundo and Valdes, 1975; Filipov, 1988; Diaconu and Morar, 1994). To the south the *Cueva Increible* ends in the *Hoyo de los Derrumbes*, which is again characterized by wide rockfalls that evoke memories of the fantastic subterranean scenery of the *Salòn del Caos*.

As at the *Cueva Increible*, the nearby access to the *Cueva Escarlata* reveals another site extremely rich in secondary cave deposits. Curtains, *paletas*, different types of stalactites and stalagmites, and an incredible number of helictites characterize the gallery. Here, the morphology is influenced not only by tectonic structures but also by the presence of chert horizons and nodules, whose greater resistance to erosion has enforced narrow points in the development of the gallery.

CONCLUSIONS

The *Gran Caverna de Santo Tomás* is undoubtedly a site of great interest for Cuban speleological and karst research. The 2003 expedition was an attempt once more to focus the attention of cavers and researchers on this karst system, where both further explorations and scientific research programmes are worth carrying out. The hope

Figure 8. *View of the Salòn del Caos (photo: E Loreto).*

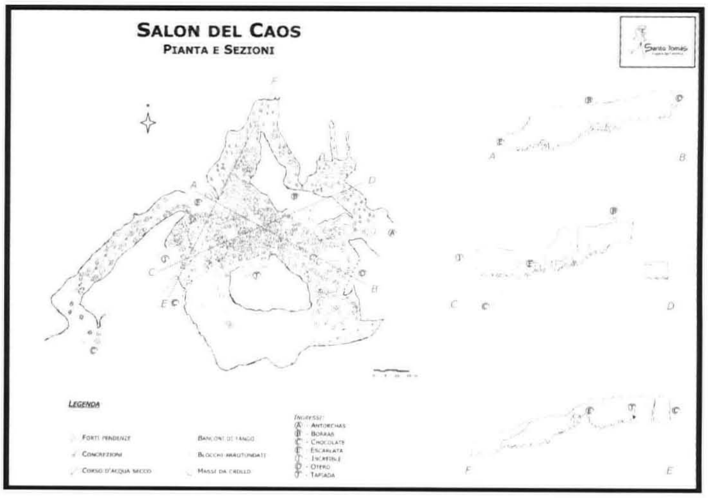

Figure 9. *Map of the Salòn del Caos (elaboration and drawing by U Del Vecchio and R Tedesco).*

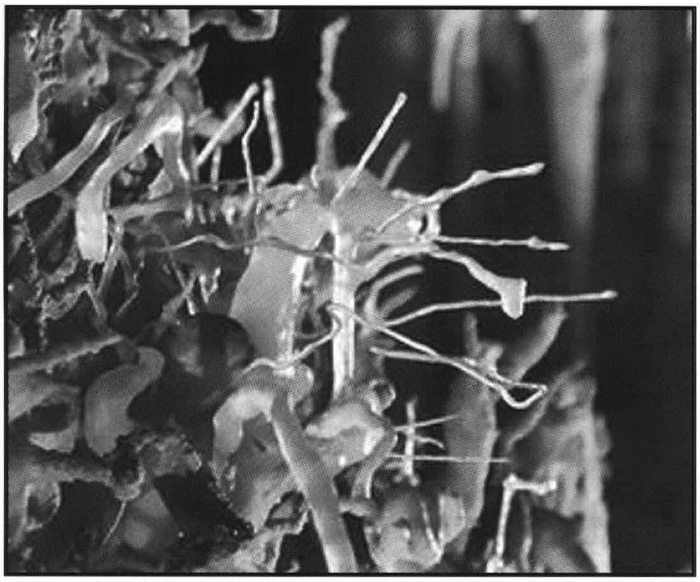

Figure 10. Helictites in the Cueva Increible (photo: G Ruggieri).

is that the work done, both in terms of the topographical re-definition of a part of the cave, and of the co-operation with Cuban cavers, might contribute at this aim.

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Investigating the nature and origins of Gaping Gill Main Chamber, North Yorkshire, UK, using ground penetrating radar and lidar.

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Abstract: The paper reports on a first application of ground penetrating radar (GPR) and lidar (LIght Distance and Ranging) in Gaping Gill Main Chamber (GGMC). The GPR image quality is exceptionally good. Sedimentary structures are clearly recognizable down to 30m below the chamber floor. We compare the sequence stratigraphy with the stepped velocity profile and a century of flood history and suggest a link between the sequence of flow directions and the last five interglacial events. In that scenario, the GGMC roof was breached at the beginning of the last Pleistocene interglacial, reversing floodwater flow directions in the chamber. We compare estimated process rates with that time-scale. Lidar surveys provide accurate spatial measurements. The GPR-estimated minimum sediment volume beneath the chamber floor is about 1.8 times the lidar-measured chamber volume above. As a 3-D "base survey", the 2003 lidar data will allow time-lapse, or '4-D' monitoring of any future changes in GGMC.

Keywords: GPR, lidar, Gaping Gill, karst, sequence stratigraphy, flow.

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INTRODUCTION

On the southeastern flank of Ingleborough, in the Yorkshire Dales National Park, Northern England, the waters of Fell Beck plunge 98m (Cordingley, 2002) down the large, almost vertical Main Shaft of Gaping Gill into the Main Chamber (GGMC). The eminent French speleologist E A Martel first descended the Main Shaft by rope ladder in 1895 (Martel, 1897, translated 1951). Following his first descent a series of trips explored beyond GGMC, often using winches to aid the descent and ascent. Winch meets continue to be popular and are held twice a year, allowing easy access to GGMC and contributing to Gaping Gill's position as the best-known pothole in the United Kingdom.

Beyond GGMC there are over 16km of mapped passages include inlets from at least 16 other entrances and a flooded route through to the resurgences at Clapham Beck Head and Ingleborough Cave (see Brook *et al.*, 1991). The following entrances connect with the Gaping Gill – Ingleborough Cave (GG–IC) System: Bar Pot and a second new entrance to Bar Pot, Flood Entrance, Stream Passage

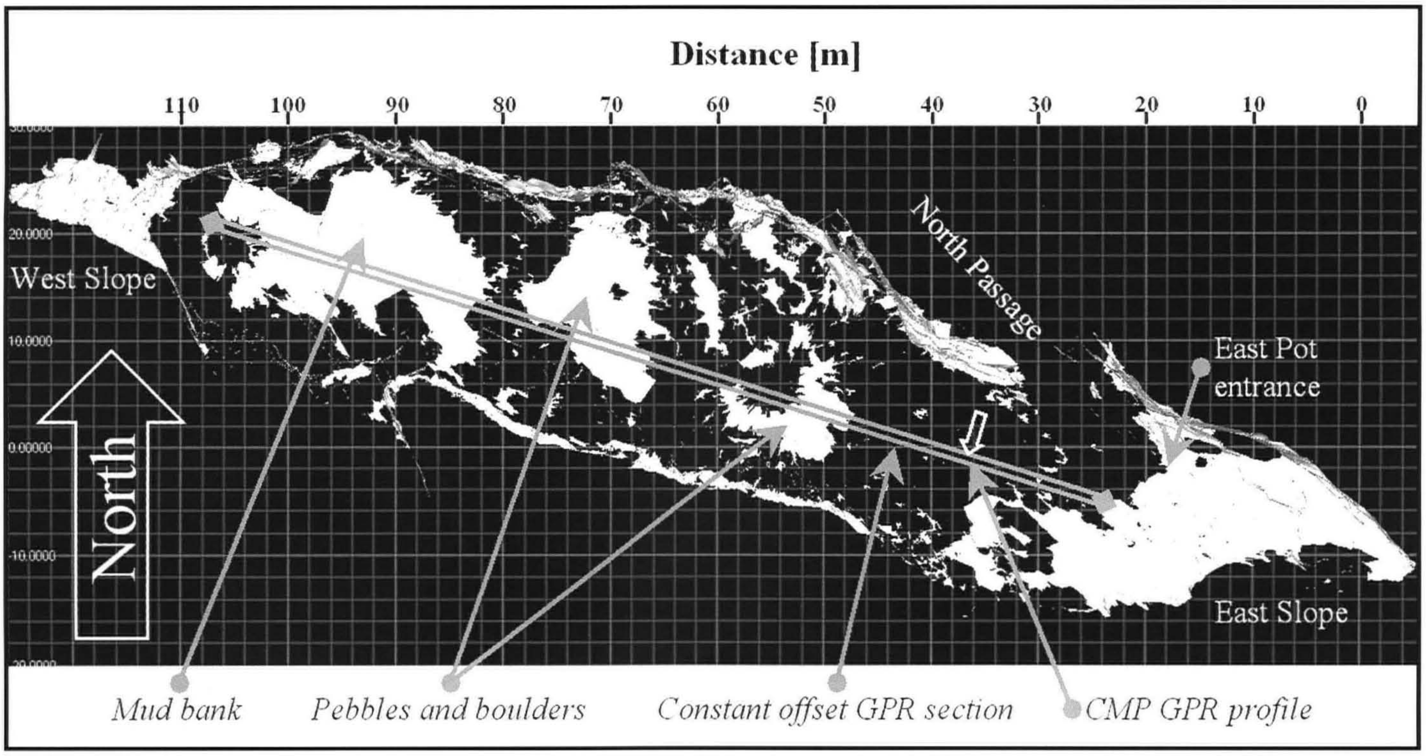

Figure 1. A plan view of Gaping Gill Main Chamber from the 2003 lidar (LIght Detection And Ranging, cf. RADAR) survey by Graham Hunter and Kate Strange of 3D Laser Mapping Ltd., Nottingham.

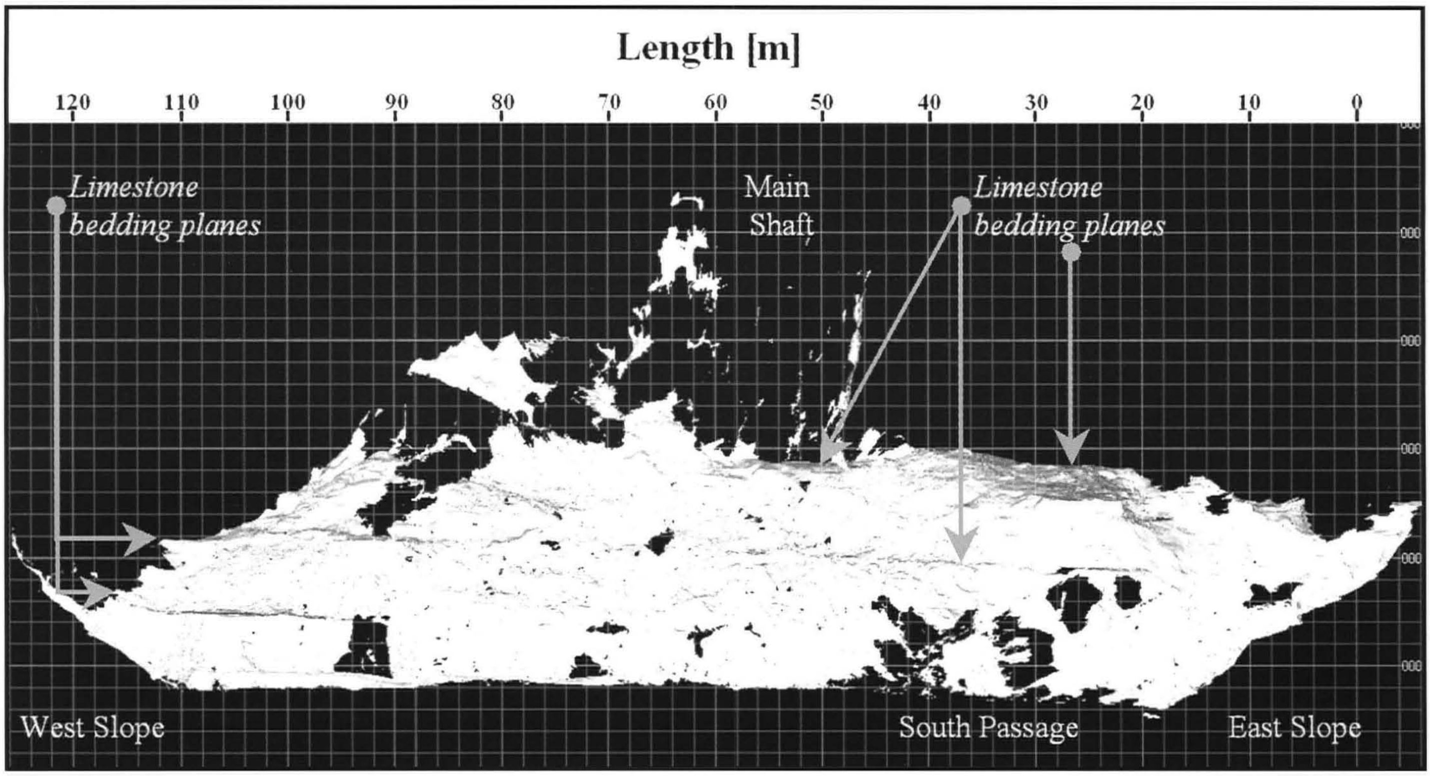

Figure 2. *A lidar longitudinal section of Gaping Gill Main Chamber, as though seen from the south through the country rock. The horizontal and vertical scales are the same.*

Pot, Disappointment Pot, Main Shaft, Lateral/Jib shaft, Rathole, Rathole Sink, Hensler's Pot, Corky's Pot. Additionally, Wades Entrance joins Flood Entrance, as does OBJ, with a voice but impassable link. Clapham Bottoms Pot has a dived connection to Gaping Gill and Car Pot has a voice connection to Far East passage in Gaping Gill.

The Main Chamber is located at OSGB NGR East 375100m North 472700m (SD 751727). It is a large cavern, about 128m long, 25m wide and up to 43m high (cf. figures 1–3; Westerman and Helm, 2003). The long axis aligns approximately east 20° south, parallel to the strike of a guiding fault; see Glover (1973a–c) and Figure 1. The visible part of GGMC has the approximate form of a triangular prism like a "Toblerone" bar, so the cavern may continue to widen with depth, hidden by the boulder fill (figures 3–5 and 13). Apart from East Pot, a 16m-deep shaft in the floor, any passages related to such development at depth remain unknown. Passageways might continue westwards from East Pot through the boulders flooring GGMC (Figure 12).

The floor consists mainly of cobbles and boulders, with banks of fine-grained sediment concentrated at the western and eastern ends. The boulders and cobbles form a low mound in the centre of the chamber adjacent to the main entry shaft, Spencer (1974 to 1994). There is a large mud bank towards the western end and a small patch of silty ground near the eastern end of the chamber (figures 1 and 8). Boulder slopes forming both ends of the chamber can be climbed to gain access to further passages.

Rocks belonging to the Yoredale Group (formerly known as the Yoredale Series or the Wensleydale Group) overlain by the basal sandstone of the Millstone Grit Group form the nearby summit of Ingleborough. The Yoredale Group comprises several (in some cases incomplete) cycles of limestone, mudstone/siltstone and sandstone with a few thin coal seams. This sequence overlies the uppermost beds of the Great Scar Limestone Group, which hosts the GG–IC System. Fell Beck feeds drainage from Clapham Bents on the eastern slopes of Ingleborough into the Main Shaft of GGMC. The water then flows bilaterally away from the centre of the chamber, alongside the northern wall and over the sediment floor before sinking near the eastern and western extremities of the cavern. Most of the sediment floor is therefore relatively impermeable. Ground penetrating radar (GPR) and ground lidar (LIght Detection And Ranging) surveys were undertaken between 18 and 24 August 2003, during the annual Craven Pothole Club (CPC) Winch Meet.

Laboratory measurements could not reliably predict whether GPR might provide a useful sub-surface image of a new prospect, so field-testing was necessary. GPR data acquisition was a logical underground progression of work by Pringle *et al.* (2002), where surveys were undertaken on sediment deposits in the entrance of Peak Cavern, Derbyshire.

PREVIOUS OBSERVATIONS

The nature and position of the sediment in the chamber appear to have changed over the last 110 years (Murphy and Allshorn, 2003), and the points at which the stream has sunk into the floor have also changed over time (Cook, 1957; Spencer, 1974). Although the waters of Fell Beck that sink into the sediment floor at the extremities of the chamber are not seen again in the northern part of the system, a positive dye trace connects GGMC to South East Pot, south of GGMC (Marston and Schoffield,1962).

During times of extreme flood the sinks in the chamber floor can no longer cope with the volume of water entering the system and a temporary lake forms in the chamber. Direct observation of such events is relatively rare. A party that was trapped underground when the winch could not pull against the force of the water coming down the Main Shaft described the power of the flood, yet the water depth did not exceed three feet (<1m) at that time (Rule, 1909).

Strandlines of debris show that the water level has been higher on occasions. Calvert (1899, 1902) recorded sheep bones and wool at least 15 feet (c.4.5m) above the floor level on the East Slope, and baulks of timber including a fence stake 50 feet (c.15m) up the slope. As this was the first exploration of this area, previous visitors cannot have placed them there. Following a severe flood, visitors in 1967 recovered a toilet seat (previously dropped down the shaft) from a position 16 to 18 feet (c.4.8 to 5.5m) above the floor level at the bottom of the Main Shaft (Mitchell, 1967).

The entrance to South Passage is approximately 5m above the floor of GGMC, so it should be the first alternative exit-point for floodwaters once the sinks in the floor of GGMC are overwhelmed. In 1947, visitors found an undisturbed layer of mud in this passage, showing that floodwaters had reached this level and flowed down the passage since the last visit one year before (Beck, 1984, p.71). These various records show that since Martel's first descent, the level of floodwater in GGMC has at least reached the height of South Passage.

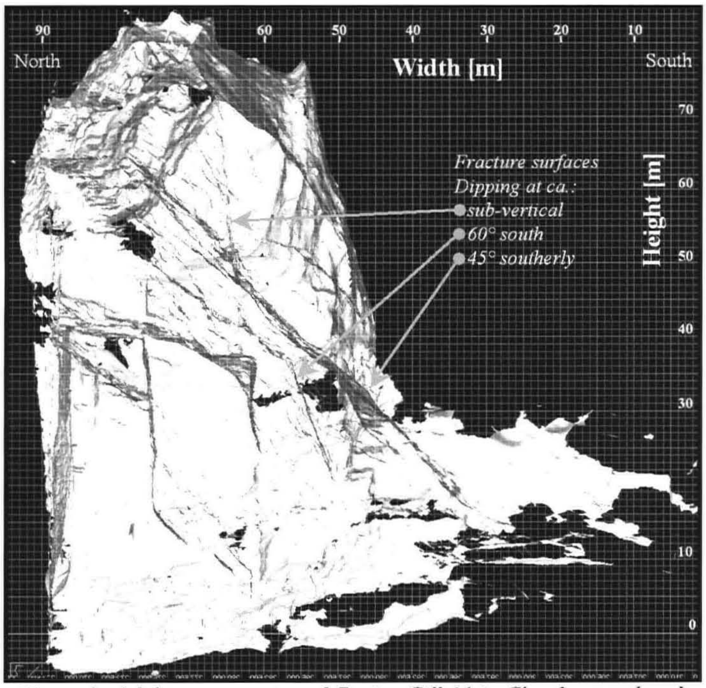

Figure 3. A lidar cross-section of Gaping Gill Main Chamber, as though seen from the west through the country rock. The horizontal and vertical scales are the same.

In 1999, cavers recovered a mid-shaft from a human fibula and a fragment of a probable human right rib from the West Slope of GGMC. Associated with the human remains were bones of red deer and black grouse. Radiocarbon dating of a red deer bone produced a date of 4,016±51 (Wk-12429). Detailed surveying has shown that these bones were 7.45m above the level of South Passage (see Cordingley, 2001 and Figure 2). The most likely explanation for the location of the bones is that they drifted into place on ponded floodwater to the level of an overspill point into West Chamber (Cordingley, 1999). The redistribution of bones in GGMC by ponded floodwater appears to be relatively common. As already mentioned, sheep bones and wool were noted at the eastern end of the chamber by Calvert (1899, 1902) but the most striking example was the discovery of modern human remains in flood debris well above floor level near the entrance to South Passage in 1947 (Crunden, 2002; Beck, 1984; Bottomley, 1979).

The location of the prehistoric bone assemblage suggests that during the Bronze Age GGMC flooded to a much greater depth than has been observed over the past century or so. Perhaps a blockage in the South Passage outlet caused the water to reach such a depth. Such a situation might have been due to increased sedimentation in GGMC after deforestation of the Fell Beck catchment. At the level on West Slope where the bones were found, a stooping passage leads into West Chamber. This large cavity is filled with laminated sands, which appear to have a different origin from the sediments observed in GGMC. West Chamber was discovered in 1896 (Calvert, 1899, 1902). The chamber was revisited in 1907 when a piece of wood was recovered from on top of the sediments (Cutriss, 1908). The description of the piece of wood as an "oak plank lying on silt" suggests it had a worked appearance rather than being a tree trunk from within the sediment fill. As the only previous visit did not descend the climb down into West Chamber, caver activity cannot have emplaced the plank. The only possible points of entry to the system in the area are the Main Shaft, Lateral Shaft (Jib Tunnel) and Rat Hole, which also enter GGMC (see Brook *et al.*, 1991), so the "plank" probably floated into West Chamber on ponded floodwaters. This record of a "plank" is intriguing and perhaps shows ponded floodwaters have reached the depth indicated by the Bronze Age bones in recent times. In the mid 19th century, the geologist Professor T McKenny Hughes threw pieces of wood down the Main Shaft. He wrote the promise of a reward on them for anyone returning one to him, in order to confirm the destination of Fell Beck (Hughes, 1887). If the "plank" observed by Cutriss (1908) was one such piece of wood, then flood waters must have overtopped West Slope into West Chamber in the latter half of the 19th century.

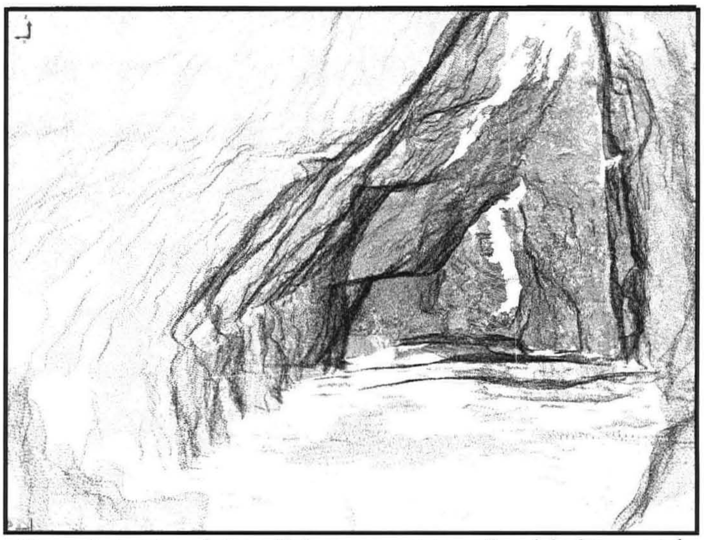

Figure 4. An internal view of lidar survey points, as though looking towards the west from inside Gaping Gill Main Chamber. The points shown are smoothed triangular apices derived from the raw data and they have an average spacing of 6cm.

Direct observation has also shown subsidence and collapse to be ongoing processes within GGMC. Between 1951 and 1956, Spencer (1974 and 1988) records the collapse of part of the GGMC wall below the entrance to South Passage, with the collapse blocks subsiding into the floor sediments. Prior to this development cavers entered South Passage by traversing East Slope rather than by climbing directly up from the chamber floor. The changes led Spencer to postulate the existence of a cavity in the sediments beneath the entrance to South Passage.

Water flow and dampness continue to vary across the chamber floor. The western mud bank arrowed in Figure 1 used to be dry, as seen in a 1952 postcard by Len Cook. In 1989 the CPC held a 60th anniversary dinner on what was then a western sandbank, but which is now glutinous mud. In the late 1960s there was almost no channel between that western mud bank and West Slope, but the Yorkshire Rambler's Club (YRC) survey shows one (Calvert, 1899, 1902). In the early 1970s there was a stream running north to south across the middle of the chamber. That stream is no longer there, but it appears on the YRC survey of a century ago, flowing in the opposite direction. The limited hydraulic conductivity of the sediments forming GGMC floor appears to influence both normal stream-flow and floodwater ponding in the chamber.

Previous work on trying to understand the nature of the fill has relied on direct observations in East Pot, a shaft beneath the eastern end of GGMC (figures 1 and 12). East Pot offers a descent of 16m between fault breccia and a so-called "hanging death" of unconsolidated fill comprising angular limestone blocks and rounded sandstone pebbles (Spencer, 1974 – 1994). There is very little evidence of fine-grained material in the walls of East Pot though Spencer (1986) records the presence of sand at the base of the pot. However, the behaviour of surface water proves, and appearance of the constant-offset GPR section shows, that this is not typical of the chamber fill as a whole. Almost the entire GGMC floor functions as an aquiclude, directing the stream away from the centre and into the far corners of the cavern. Water that sinks intermittently in East Pot during normal conditions, and floodwaters that fill the shaft during periods of flooding could have removed any fine-grained material from the eastern end of the sediment pile around East Pot.

Between 1961 and 1971, government grant-aid paid for drainage ditches (grips) to be dug on the moor above Gaping Gill and elsewhere (Halliwell, 1998; Brook, 1971). Despite the underlying good intentions of this work, such gripping altered the balance of run-off and erosion. Caves started to flood in minutes instead of several hours, bacterial contamination of Clapham Beck Head increased and Clapham Lake silted up. Where the moorland peat was thick or where it covered shale or boulder clay, master ditches cut back rapidly from streams or sink holes to produce canyons up to 20 feet (c.6m) deep, Brook (1971). At that time GGMC flooded severely to a depth of more than 20 feet on at least three occasions.

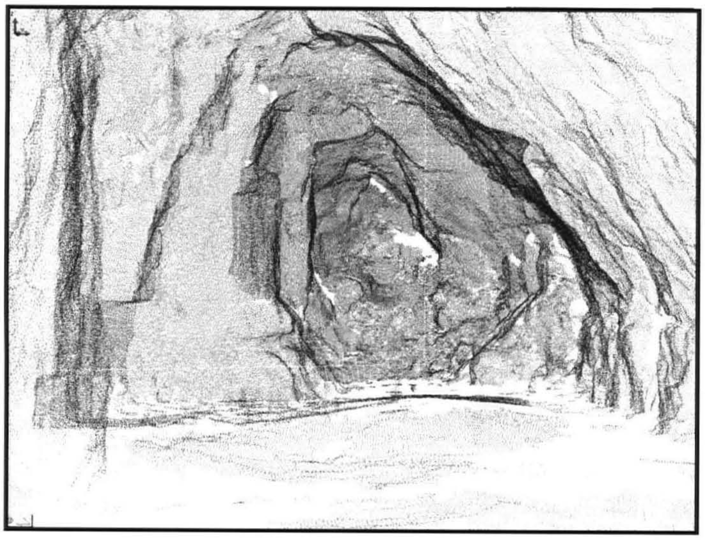

Figure 5. An internal view of lidar survey points similar to that of Figure 4, but looking towards the east.

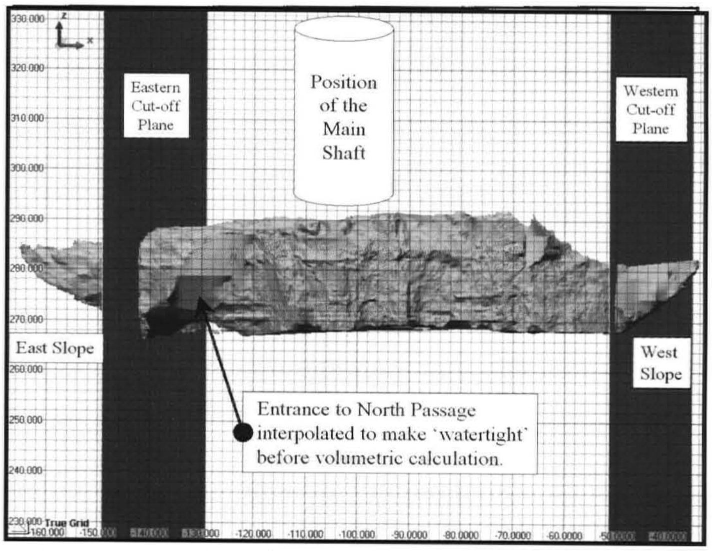

Figure 6. A lidar view of Gaping Gill Main Chamber, as though seen through the rock from the north. Survey gaps are smoothly interpolated prior to volume and area calculations. Vertical cut-off planes separate the East Slope and the West Slope from the Main Chamber floor.

Water ponded in the chamber overflowed along South Passage, filling the passage near T-junction with neck deep water, which was still there in the summer. Concurrently East Slope became unstable and floor sediments between the middle and the eastern end of the GGMC disappeared. Presumably, this series of events was similar to the wartime flood mentioned above, which overflowed along South Passage and drained down D1 at the end of Sand Caverns.

The short-term increase in flood intensity seemed to wane as the moorland grips collapsed in on themselves. However, in one of the wettest summers of recent memory, Cordingley (1985) witnessed a flood at the GG–IC System resurgences of Beck Head Stream Cave and Ingleborough Show Cave. A rainstorm hit waterlogged fells between 10 am and mid-afternoon on 26 July 1985. The river at Beck Head started to rise by 11:30 am. An hour or so later, "...the whole entrance slope had turned into an unbroken stretch of 'white water'." The flood started to subside soon after the rain stopped, when large "whale-backs" of sand were found deposited. In 1998, the European Community and the UK Government paid £5M to regenerate heather and restore upland ponds and woodland, recanting the previous moorland drainage policy.

CONTEXT OF THE PRELIMINARY GPR AND LIDAR TRIALS

The 2003 survey described here acquired both the first lidar and the first GPR datasets for GGMC. It is one of only a few cave surveys to use the exciting new lidar technique (cf. St. Amand, 1999; Day, 2002; Bedford, 2003; Westerman and Pringle, 2003). The integration of the two datasets provides a basis for one of only a few 3-D digital geodata models for cave systems (see Figure 12 and Rossi *et al.*, 2002).

The questions of what lies beneath the floor of GGMC and exactly how big is the cavern have intrigued generations of cavers. Outstanding geological questions include the possible Pleistocene stratification of the GGMC sediments, as there are well-preserved layered sediments elsewhere in the system (Norman, 1995). Several good reasons to carry out the survey gradually emerged. The unstable nature of the GGMC cave floor sediments and sections of the rock wall have been recorded for over a century (e.g. Dwerryhouse, 1907; Horn, 1907; Spencer, 1974 to 1994; Glover and Halliwell, 1983). Investigating the nature of the sedimentary fill should hopefully lead to an understanding of the processes controlling the instability.

LIDAR SURVEY

Ground lidar is essentially automated 3D laser mapping. Lidar data acquisition rates are currently over half a million times faster than manual surveying with a Total Station. Lidar technology is developing rapidly and the image quality of future surveys with more recent equipment will doubtless be even better than that achieved at Gaping Gill in 2003. However, in a dynamic system, there are tremendous advantages in being able to compare future surveys with an existing benchmark.

Acquisition

The Z210 system used was Riegl's first 3D imaging sensor for industrial applications. (Technical data are available at http://www.riegl.co.at/lms-z210/e_lms-z210.htm) Key parameters are a typical measurement accuracy of 25mm with a line scan range of 80° and a frame scan range of 333°.

A scan rate of 18,000 points per second was selected and the first scan was taken from the eastern edge of the Gaping Gill sinkhole. This imaged the surrounding moorland, the Fell Beck valley and the top of the Gaping Gill shaft. In GGMC, three pairs of orthogonal scans were made, distributed along the axis of the floor (Manby, 2003). Reflector patches placed in the overlap zones between survey swaths assisted data integration. Altogether about 15 million xyz points were acquired during the afternoon.

An additional ambitious aim was to link the surface and the underground surveys. With hindsight, if the surface scan had been from the northwestern edge of the sinkhole, a survey point from the floor of GGMC might have seen reflectors or at least common features on the upper southern edge of the shaft. Then the surface and underground surveys could have merged with an overall accuracy of about 25mm (1"). This belated aim was not achieved in 2003, and so the overall measured depth of 98m (Cordingley, 2002) cannot yet be compared with a value based on the lidar results.

Riegl software is able to recognize the artificial reflector patches, then register and join up the separate survey scans. Each least squares fit between overlapping scans involves hundreds of thousands of survey points. This provides unmatched registration accuracy between successive survey stations.

The three pairs of underground lidar scans combine to form the data model available on CD-ROM. The 15 million separate image points re-sampled to 750,000 non-overlapping vertices for rendering. Those points interpolated to 860,000 vertices. The final 30 MB file, 'gaping_model_1.pol' covers GGMC on a spacing of about 6cm (see figures 1 to 6). The 2003 GGMC lidar data remain as originally prepared by 3D Laser Mapping / Riegl UK Ltd. In the 30 MB '.pol' file, the survey is oriented to Grid North.

Volume and area measurements: method

Figure 6 illustrates the calculation of areas and volumes from lidar data. In order to calculate the volume of a model it must be 'watertight'. Due to the rugose nature of the cave walls and limited survey time it was not possible to scan all surfaces to produce a watertight model, so areas of missing data such as the entrance to North Passage were filled in smoothly and, as such, they are easy to spot in the model. Since those surface infill areas are somewhat

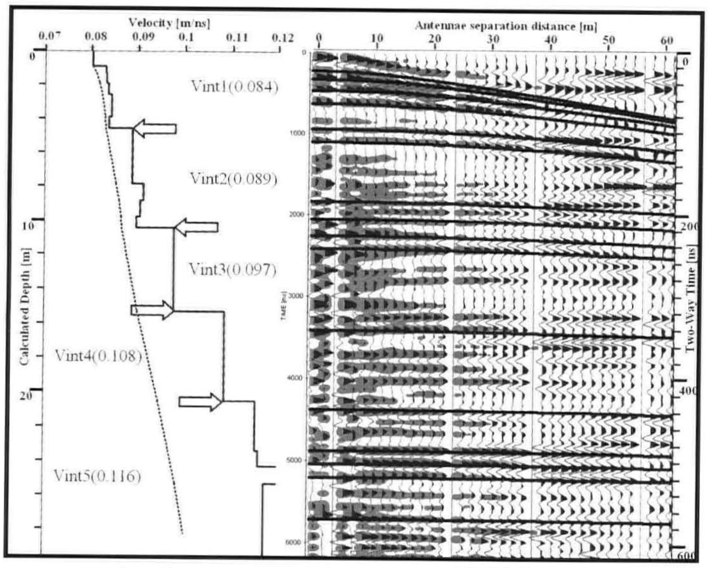

Figure 7. The sole surviving Common Mid Point (CMP) Ground Penetrating Radar (GPR) profile centres at 14m from the eastern end of the line of profile in Gaping Gill Main Chamber.

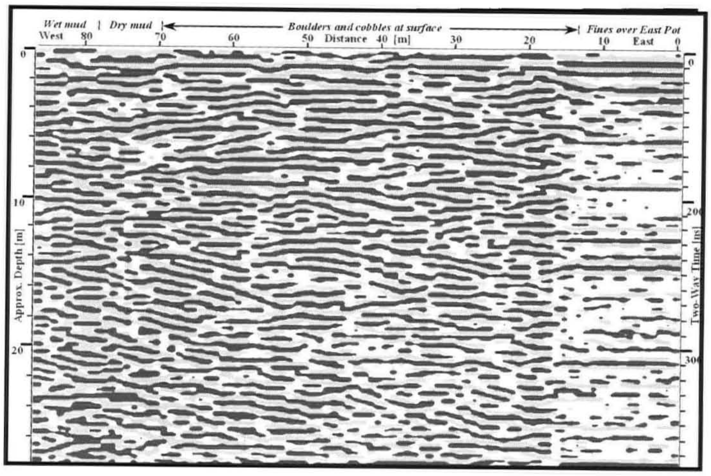

Figure 8. The 87m-long, constant-offset GPR reflection section along the main axis of Gaping Gill Main Chamber. Positive events are shaded dark grey and negative events are pale grey. Near-zero values are white. The horizontal length scale has its origin at the eastern end of the Main Chamber and the surface conditions of the chamber floor are shown along the upper edge.

arbitrary, the accuracy of the volume generated should be treated with caution. Note that the Main Shaft is not included in the volumetric calculations; compare Figure 2 (seen from the south) with Figure 6 (seen from the north). Nevertheless, if we wish to calculate the quantities of sediment removed by dissolution and moved by the migration of voids, volumetric measurements are essential. Future work can always improve on these beginnings.

The volume calculations summarised in Table 1 refer to three segments of GGMC: West Slope, the main floor area and East Slope. Note the cut-off planes in Figure 6. Each segment is then sliced and its volume calculated from the cross-sectional area and the width of section. The larger the number of slices the more accurate is the volume estimate. Each air volume in rows 'a', 'b' and 'c' of Table 1 derives from 1000 slices. If the largest volume, 'c' derives from just 100 slices, the value arrived at is 35,357m^3, which is a difference of only 4m^3, or only 0.01%. A rough check may be made by treating GGMC as a triangular prism and multiplying the floor area by half the height.

In order to calculate the floor surface area, the wall was first trimmed off in the computer model by hand. The floor area is given in row 'e' of Table 1. It refers to the standing area of GGMC only without the slopes and without any vertical faces. A rough check may be made by counting the 10m squares in Figure 1.

Results

The only surprising result in Table 1 is the distance around the main standing area of GGMC, labelled 'f' in Table 1. The approximately 2,100m^2 floor area has a perimeter of 1.89km. However, the lidar survey has a sample interval of about 6cm, the walls of GGMC are rough and the apparent length of fractal surfaces increases in proportion to the detail of their description, Richardson (1961). In any case, we shall only apply the volume and area measurements in what follows.

The 6cm sample interval and 2.5cm accuracy of the 2003 lidar data model is sufficient to define the main geological features. Those features illustrate the role played by geological structures in guiding the development of the exposed upper part of GGMC.

The main faults and joints, the main bedding planes and some overhangs developed below bedding surfaces are clearly visible in figures 1–6. Each geological surface segment may have tens of thousands of lidar survey points. By contrast, a field geologist might aspire to a few dip and strike measurements from representative surfaces.

A remarkable advantage of lidar cave survey data is the ability to view cave morphology from the outside – as though one could see through the solid rock (see for example figures 2, 3, 6 and 12). The geological features controlling cavern development can then be much more obvious than when the cave is viewed from inside. However, the internal, 'immersive' view is better for identifying fine detail and comparing that detail with direct observations (figures 4 and 5).

Figure 2 shows that GGMC has a boat-like longitudinal profile due to the boulder slopes at each end and the near-central Main Shaft. The central floor area of the sedimentary cave fill forms a subtle dome, with an overall fall of 2.5m from west to east, see Figure 12. Three strong bedding plane features dip gently towards the southeast at approximately one-third and two-thirds of the way up the chamber wall and at roof level east of centre, see Figure 2. The Main Shaft is approximately halfway along the axis of the chamber.

Figure 3 shows the strong vertical fractures that control the development of the north wall to GGMC. Southerly-dipping fractures at approximately 45° control the development of the southern wall. A few more steeply dipping fractures are seen at (30m high, 55m north) in the coordinates of the data model.

In contrast to the previously unseen views of figures 2 and 3, figures 4 and 5 are as seen from within the cavern. Of course, the lidar observer is not restricted to the floor or to vantage points afforded by ropes. The overhanging bedding planes near the centre of the internal view in Figure 4 form the near-horizontal lineations in the external view of figures 2 and 6. The triangular West (boulder) Slope is just to right of centre in Figure 4. The near-vertical fault plane forming the north wall of the chamber is to the right and the 45° southerly dipping fractures controlling the development of the southern wall to GGMC are to the left. Cave-fill sediments form the gentle hummocks in the floor.

The bedding plane features in Figure 5, to the east, are weaker than in Figure 4, to the west. The East (boulder) Slope in Figure 5 is larger than the West Slope in Figure 4. The near-vertical fault plane forming the north wall of the chamber is now to the left and the 45° southerly dipping fractures controlling the development of the southern wall to GGMC are now to the right. Cave-fill sediments again form gentle hummocks in the chamber floor.

Note from figures 4 and 5 that, although the southern roof of GGMC is a 'lean-to', there is a significant wall, all the way along the southern side of the chamber. That is, vertical wall rock bounds the cave-fill sediments both to the south and to the north. It therefore seems reasonable to project the observed floor area of GGMC vertically downwards when calculating sedimentary volumes. However, an undercut at depth is possible. Indeed such a feature could underlie the toppling and subsiding wall-rock block near the entrance to South Passage (see sketch cross-section, Figure 13).

Interpretation

Almost all the surfaces that define GGMC are planar and at first glance appear to be little eroded by dissolution, although scallop marks indicating eastward flow have been analysed from the walls and roof (Allshorn, 2003).

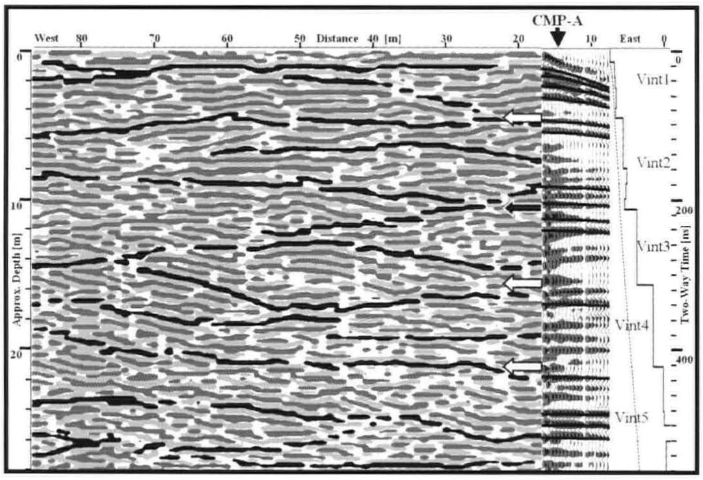

Figure 9. An interpretation of the Gaping Gill Main Chamber constant-offset GPR section in Figure 8. Certain reflection events, interpreted as bedset boundaries, are shaded black.

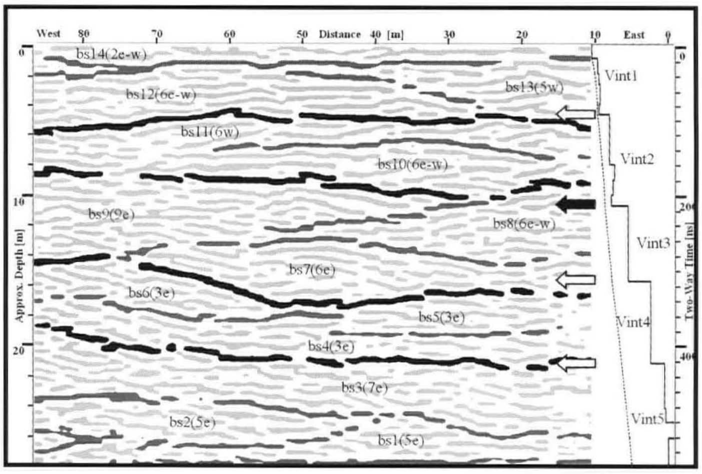

Figure 10. A grouping of the bedsets in Figure 8, using positive events only for clarity and guided by the interval velocity (Vint) steps from Figure 7. The interpretation suggests a three-stage hierarchy of bedding and bedsets (bs) within radar-stratigraphic units (rsu), each of constant interval velocity.

According to Glover (1973a-c, 1974) a fault plane guides the north wall of the chamber. Glover left the issue of the sense of displacement – whether 'scissor' or horizontal wrench – open. The lidar survey reveals a set of sub-parallel, sub-vertical master joints; oriented at an angle to the main fault plane and occupying a zone adjacent to it (cf. Figure 3). Viewed from the outside, as if through the rock from the north, one of the higher joints in this set shows a strong corrugation lineation plunging gently towards the eastsoutheast (cf. Figure 6). This is persuasive evidence for an oblique-slip displacement along the fault guiding the GGMC. The throw of the fault increases towards the eastsoutheast (Glover, op. cit.). The larger horizontal component of the plunge lineation supports Glover's strike-slip option for the master fault displacement. Halliwell (1980) notes Wager's (1931) evidence for strike-slip, or 'tear' displacement along the main Craven Faults to the south of the GG–IC System. Obviously, further similar observations along the length of Glover's fault would be a useful check. Note that this observation of a high-level plunge lineation would not have been possible without a lidar dataset.

In contrast to the northern wall of GGMC, lower-angle fractures and bedding plane overhangs guide the southern wall (cf. figures 3–5 and 13. This provides good evidence for stoping as an influencial mechanism in chamber growth and the upward migration of the void. The stoped and fallen blocks must have made, and be making, a significant contribution to the sediments beneath the cavern floor. The upward migration of voids from old mine workings can gradually choke off, if the detritus 'bulks up' to fill the void and support the roof. However, limestone dissolution by fresh, aggressive peat-bog drainage from Fell Beck must be continually removing material, lowering the floor by dissolution and leaving the chamber roof unsupported (see Figure 13).

The Context section above references the unstable nature of the GGMC floor. Glover and Halliwell (1983), Spencer (1988) and Karley (personal communication, 2003) note the changing position of a large limestone block near to the entrance to South Passage. This could be further evidence that dissolution processes are removing material from deep beneath the floor of GGMC, in this case removing support from the wall and allowing lateral migration of the void (see Figure 13). The 2003 Riegl lidar survey should provide a 3-D benchmark against which future lidar surveys will be able to monitor any further movement of that block, as well as of the West Slope and the floor of GGMC.

GROUND PENETRATING RADAR SURVEY

Most GPR studies in karst have focused on detecting caves within the limestone from the surface, rather than profiling cave earth deposits within the caves (e.g. Al-fares *et al.*, 2002; Beres and Haeni, 2001; Chamberlain *et al.*, 2000; Collins *et al.*, 1994; Doolittle and Collins, 1998; Parr, 2004).

Acquisition

The system used for this survey was the Utsi Electronics 'Groundvue 2' (GV2). The GV2 has a broad frequency bandwidth from 30–100 MHz; (http://www.utsielectronics.co.uk/erica.utsi/gpr.htm). Broad bandwidth signals are most desirable for good quality geophysical imaging. The GV2 can image down to a depth of 11m in peat, arguably one of the most difficult media for GPR investigations (Utsi, 2001, 2003). However, GPR trials conducted on the moor between Gaping Gill and Bar Pot showed strong attenuation in areas of peat and glacial till cover. Those data remain to be analysed more fully.

The GGMC survey plan comprised three CMP (Common Mid Point) surveys linked by a constant-offset profile. All four GPR datasets are co-linear and referenced to the same tape position, with zero at the eastern end of the GGMC cavern floor (see Figure 1, in which some of the main physical features to be discussed are annotated in grey). The 87m-long, constant offset GPR section along the main axis of GGMC is marked by a double grey line with diamond end-points. The eastern end of the tape was the zero point for both the constant offset GPR section and CMP profile locations. A hollow white block arrow at 14m along the constant offset section marks the position of the CMP GPR profile in Figure 7. The CMP surveys were to calibrate the electro-magnetic (e-m) velocity structure of the cave floor sediments.

Every trace in each CMP profile and the 2D section had to be recorded in continuous mode with stationary antennae. Therefore, although the survey took a long time, most traces in each final section derive from up to 300-recorded traces, each stacked 1000 times in the recording system. In principle, that should have increased the signal-to-noise ratio by a factor of about 550. Unfortunately, the GV2 record length was inflexible in 2003, so the large number of traces acquired for each surface position could not provide any advantage in depth of penetration. We have a dataset with very high signal-to-noise ratio and excellent quality control, down to the c.30m depth equivalent of the 640 ns record length (figures 7–10, 12 and 13).

CMP surveys A, B and C centred at 14, 40 and 81m respectively, from the eastern end of the tape at the base of the eastern boulder slope. CMP-A centred on the eastern silt floor but overlapped onto pebbles and cobbles, as marked by the open block arrow in Figure 1. CMPs B and C later proved to have no useable data.

A 2-D constant-offset profile was run from west to east at 1m spacing.

Processing

After data conversion to SEG-Y format and an initial data quality review, Parr (2004) did all the data processing using Reflex-W software from Dr K-J Sandmeier, Karlsruhe. Parr (2004) provides processing workflows for the constant-offset section and the CMP-A profile (Table 4 in Appendix). She applied two separate processing flows, with different gain parameters to enhance reflectors at different depths.

The results displayed in this paper are redisplays for printing in

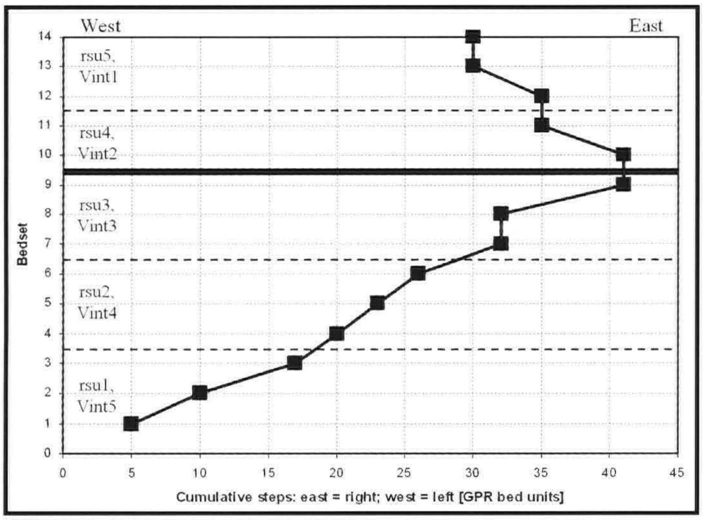

Figure 11. A 'random walk' display to log the cumulative azimuths of apparent bedding fore-set dips grouped by interpreted bedset in the GGMC constant-offset GPR section of figures 8 to 10. A thick horizontal line marks a boundary between predominantly easterly flows below and predominantly westerly flows above.

greyscale only. Mild exponential gain recovery balanced Parr's migrated, F-K filtered version of the constant-offset section. This brought both the shallow and deep reflection events within the reduced dynamic range of a single monochrome plot (figures 7–10 and 12). Positive and negative events were captured separately, grey-scale coded within Adobe PhotoShop and classified with Image-J.

Figure 7 redisplays Parr's (2004) processing results for the CMP profile in a similar manner. PhotoShop layers combined separate Reflex-W displays of grey-scale positive event contours and wiggle traces. The interpreted velocity profile down to 30m depth is in the left track. The average velocity profile is a dotted line and the interval velocity profile is a solid line. The GPR data traces are in the right track, with positive amplitudes highlighted in grey and the picked primary hyperbolic reflection events marked by thick black lines.

Results

The processed CMP-A, located at 14m on tape, enabled a velocity-depth profile by semblance analysis (Figure 7). Note the stepped profile of increasing interval velocity with depth. Hollow block arrows flag the interval velocity steps, to aid comparison with the constant-offset section in figures 8 and 9. Increasing GPR velocity could be due to a decreasing amount of clay (slow e-m velocity) and/ or an increasing amount of air (fast e-m velocity) with depth. Sandstone and limestone matrices have similar GPR velocity ranges, so the void spaces between the boulders must host any changes controlling velocity. The data quality suggests that a much greater depth of penetration would be possible if instrument settings offered longer record lengths.

The right-hand track of Figure 7 shows four dead traces. Acquisition problems affect all traces with transmitter-receiver offsets from 7.8 to 10.2m (not shown). The two further CMP profiles proved to have unusable data, possibly due to the damp and difficult conditions for electronic equipment. However, the surviving CMP gather provides a remarkable, layered velocity profile (figures 7, 9, 10, and Table 2).

There is strong evidence for sedimentary structures down to at least 30m below the floor of GGMC. In the CMP profile, there are deep primary reflection events around 600 ns two-way time, 28–30m depth equivalent (Figure 7). In the constant-offset GPR section, there are persuasive sedimentary bed forms throughout the section (figures 8–10 and 12).

The signal-to-noise ratio is excellent for all the surviving GPR data from GGMC. Figures 7–10 are some of the best GPR records ever published. It would appear that only the restricted record length prevented much deeper penetration. Unfortunately, the present Utsi GV2 GPR system has a pre-set record length suitable for its archaeological design targets within peat. The GV2 system is clearly capable of much greater penetration in freshwater cavern sediments, as at GGMC. Utsi electronics are now increasing the GV2 record length to allow data collection down to about 100m (Erica Utsi, personal communication).

Interpretation

Theoretically, the interpretation of sediment packages to well beyond the maximum constant-offset record depth of c.28m would be possible, given the potential depth penetration in the expected sediments at the frequency used. For the 30–100 MHz antennae frequency range used in this survey, expected penetrations would be 8–12m for clay and 80–100m for limestone; see Reynolds (1997). The vertical resolution of the interpreted beds is also theoretically possible given the antennae frequency and wave velocity. Resolution is taken as one-quarter of the e-m signal wavelength, i.e. c.0.3m, where wavelength is calculated as c.1.3m from a velocity of 0.1m/ns and central frequency of 65 MHz (as velocity = frequency * wavelength).

In Figure 8, the positive subset of GPR reflection events are shaded dark grey and negative events are pale grey. The shaded areas are from ±3 to ±32 (full scale) amplitude units. Amplitude values near to zero are in white. The constant-offset longitudinal GPR section for GGMC thereby reduces to an almost 'sign bit' display, for grey-scale publication purposes. However, seismic and GPR image interpretation is often a matter of opinion. Amplitude variations may appear to support opposing views, but a sign bit display reduces discussion to the existence, size and form of individual and grouped reflection events. Our interpretation below concentrates on those relatively uncontroversial aspects of the image data.

Image-J analysis shows that the constant-offset section has an exponential reflection-event length distribution. Five events are over 40m long. Clinoform reflection events dominate the record from top to bottom. The reflection events have the form of typical sedimentary interfaces, leading to the interpretation proposed by Parr (2004) and developed in this paper. The top edge of the section in Figure 8 notes the present surface sediment types on the floor of GGMC. However, surface conditions do not appear to affect the GPR data quality, except perhaps at the eastern end over East Pot, where the next bullet point below offers an alternative, preferred explanation.

The common-offset profile of figures 8–10, along the long-axis of GGMC shows the following features:

- There are near horizontal, relatively weak reflectors from zero to 16m along the profile, near the eastern edge of the chamber, approaching East Pot. As recorded above, East Pot has no visible fine-grained material, which floodwaters may have washed out. Parr (2004) suggests that the relatively weak arrivals between 0 and 16m along the profile could be due to the surface cover of fine mud/ silt, being a highly attenuating medium for GPR. However, the glutinous mud to the west does not result in weak signals (see the top labels in Figure 6). The issue might be resolved by placing a passive antenna in East Pot at a known depth. Most probably, the well-washed boulders of East Pot and the area around it lack the intercalated sediments that could provide both an aquiclude and good GPR reflection events. Spencer's (1988) hypothesis of a westerly extension to East Pot is a plausible explanation of the change in GPR image quality seen in Figure 8 (see Figure 12). The edge of the fines at surface does not precisely correspond with the change in GPR image quality (Figure 8).
- Coherent events with varying dips and lengths occur throughout the section. Based upon these events, Parr (2004) provided an initial sequence stratigraphy for GGMC. She picked coherent reflectors and truncations to identify 'radar geobodies' or sedimentary packages within the section. Figures 8–10 and Table 2 elaborate upon her insight. We rename the boundaries to her sedimentary packages 'bedset boundaries'.
- The depth to bedrock beneath the present sedimentary cave floor is probably much greater than 30m. Future GPR surveys at GGMC should have much longer record lengths. Indeed, the desire for a longer record length and greater depth of penetration guided the first choice of an S&S GPR system for the 2003 GPR survey.

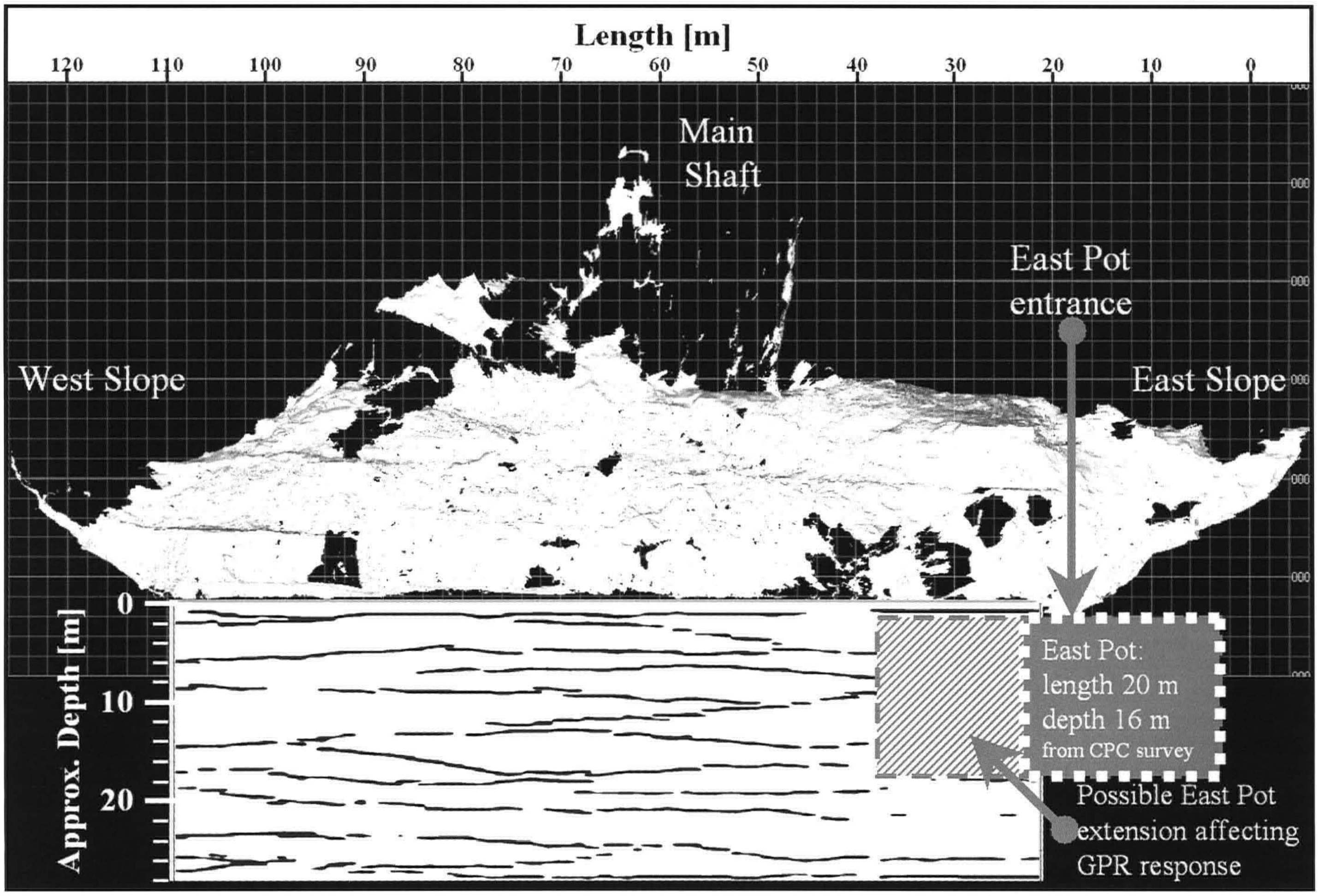

Figure 12. A same-scale composite section of the longitudinal lidar profile in Figure 2, with GPR bedset boundaries from Figure 9 and the approximate profile position of East Pot as surveyed by Craven Pothole Club. A, possible westerly extension (cross-hatched) to East Pot could explain the change in GPR reflection section character near the foot of East Slope.

Figure 9 shows that there is some correspondence between interval velocity and bedset boundaries. The first, fourth and fifth hollow block arrows, marking interval velocity (Vint) steps, correlate well with some constant offset reflection section events. The second and third interval velocity boundaries seem to mis-correlate by about 1m. The correlation between the velocity profile and the constant offset reflection section could have been checked if the other two CMP profiles attempted had recorded usable data.

We therefore suggest a provisional grouping of bedsets into Radar Stratigraphic Units as indicated by interval velocity changes (compare Figure 9 with Table 2). We shall use the nomenclature of sequence stratigraphy, which is normally associated with seismic sections rather than GPR. Sequence stratigraphical concepts were initially associated with sea level change, making them inappropriate for a land-locked system such as GG–IC. However, Khadkikar and Rajshekar (in press) show that climate change can provide an alternative mechanism for cycles of erosion and deposition. Figure 10 develops the interpretation of Figure 9 to suggest a three-level hierarchy of event type, using positive events only for clarity:

a) Bedding planes are now in pale grey. Beds are layers of sedimentary rock separated by breaks called bedding planes (which are not necessarily planar!). A sedimentary bed is a basic unit, or stratigraphical building block (Campbell, 1967). The lithological composition of beds, their geometry, trajectory, stacking patterns and hierarchies are used to determine depositional setting. Because seismic and GPR events are caused by changes in material properties, reflection events mark discontinuities such as stratigraphical bedding planes, rather than the beds themselves.
b) Bedset (bs) bounding surfaces are now in a 50% mid-grey. Bedsets are relatively conformable successions of genetically related beds bounded by bedset surfaces. Bedset surfaces may have characteristics of erosion, non-deposition, or correlative conformity. Most of the bedsets in the GGMC constant offset GPR section comprise bedding planes with parallel top-, fore- and bottom-set dips. Image-J measures several bedset boundary events in Figure 9 as being from 40 to 70m long. These bedset boundaries are long reflection events or lines of reflection events that divide packages of events that look like current, or cross bedding. The clearest examples show sequences of clear onlaps, e.g. labelled 'bs9' at centre left and 'bs13' at upper right. All bedsets have top-, fore- and bottom-set events. Overall, apparent dip azimuths are approximately in a ratio of (4:2:1 :: east: indeterminate: west). This is consistent with an accretionary pile accumulating in the presence of a predominantly easterly current direction. There are only two erosional contacts between bedsets and both are deep in the section overlying the bedset labelled 'bs1' in Figure 10.
c) Radar stratigraphic unit (rsu) boundaries are now in black. These compare with seismic sequence boundaries, which define depositional sequences identified from a seismic, or in this case GPR sections, Mitchum (1977). A Sequence is a relatively conformable succession of seismic reflectors bounded at its top and base by sequence boundaries, Vail *et al.*, (1977). In Figure 10, the lowest and highest radar stratigraphic units show more steeply dipping, higher energy bedsets (labelled bs1, bs2 and bs12, bs13) overlain by gently dipping, lower energy bedsets (bs3 and bs14). (Vint4, rsu2) has gently dipping, lower energy bed forms divided into (bs4, bs5 and bs6). Radar stratigraphic units (Vint2, rsu4) and (Vint3, rsu3) have bedsets of mainly steeply dipping, higher energy events.

Parr (2004) noted that the most obvious break in sequence occurs at a depth of about 10m, at the top of her sedimentary package 8, which corresponds with the base of bedset 10 and between interval velocities Vint2 and Vint3 in this paper (see Figure 10). A sub-horizontal series of reflection events separates the mainly eastward-dipping bedding fore-sets below, from those above, which have a predominance of westward or gently dipping fore-sets. Parr

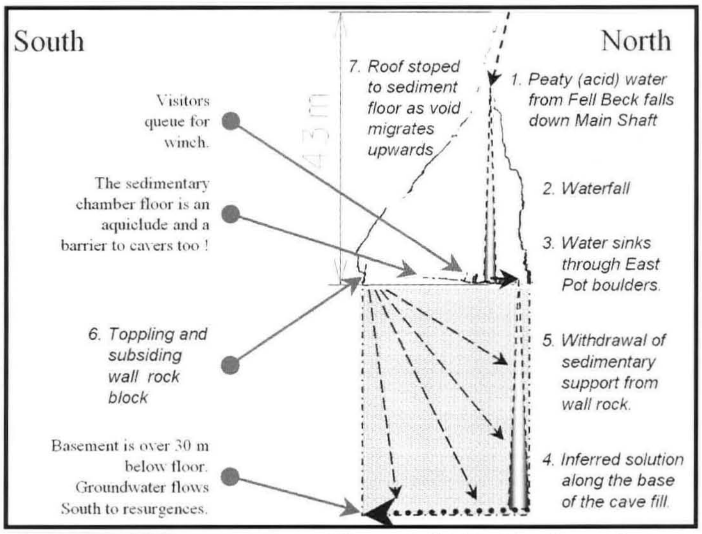

Figure 13. *A lidar cross-section of Gaping Gill Main Chamber to illustrate the seven-step cavern growth process proposed in this paper. Continued void-migration requires sediment removal processes.*

suggested a distinct change of depositional environment, marked by a black solid block arrow in Figure 10 and a thick black line in Table 2 and Figure 11.

Table 2 also logs a rough count of the number of identified fore-set events and their predominant direction of apparent dip within each bedset. The notes refer to the relationship between the bedding surfaces within each bedset and the bedset boundaries. Figure 11 then uses a (cf. random) walk technique to log the cumulative azimuths of fore-set apparent dips. It supports the view of Parr (2004) that the boundary between bedsets 9 and 10 at approximately 10m depth marks a significant change in the depositional processes within GGMC. Below 10m depth, most bedding fore-sets have easterly apparent dips. Above 10m depth, most bedding fore-sets have westerly apparent dips. Further work may confirm or refine the stratigraphical units identified and ideally assign a timescale and environmental setting to each.

Figure 12 is a same-scale composite section comprising:

- The longitudinal lidar profile of GGMC as seen through the rock from the south (compare Figure 2).
- The bedset boundaries identified and described in figures 9–11. Note that previous figures used an exaggerated vertical scale for the GPR constant-offset section to emphasise bedding plane dips, as is commonly done during seismic interpretation. Also, note that the apparent dips seen in the evenly-scaled GPR section are of a similar slope to the topographic changes in the modern cave floor.
- The approximate profile position of East Pot as surveyed by Craven Pothole Club members (Warren, 1986).
- A possible westerly extension to East Pot that might explain the dramatic change in GPR image quality at 38m on the upper scale, or 18m along the constant offset GPR profile of Figure 7 (compare Spencer, 1974 and 1988).

Note that the GPR survey probed at least as deeply below the floor of GGMC as the overall 20–30m height of the cavern (see Figure 12). We next compare the volume of the cave sediment fill with that of the void space above.

SEDIMENT VOLUME ESTIMATION

Table 1 summarises the floor area and void volume of the three main segments of GGMC. Alternative units are provided for petroleum engineers and cavers. The volumes labelled 'a' to 'd' and the floor area labelled 'e' are direct measurements. However, the volume of sediments beneath the central floor area must be calculated, based on a few assumptions. The 2003 GPR survey shows that the sediments are at least 30m deep. As a first instance, we assume that the vertical walls around the floor, shown in figures 4 and 5, project to depth, as in Figure 13. We then multiply the central floor area, 'e' in Table 1, by 30m to estimate a minimum sediment volume, 'g' in Table 1.

A ratio calculation shows that the volume of sediment below the floor of GGMC approaches twice the volume of the air above (see row 'j' in Table 1). The total volume involved in cave formation processes is over two and a half times the volume of the air space alone.

DISCUSSION

It is assumed that the sediments beneath the floor of GGMC should be similar to those displayed on the floor of GGMC. They should therefore consist of clay, sand, sandstone and limestone cobbles from Fell Beck, with limestone cobbles and angular boulders from the cavern roof though the stratigraphy of cave sediments is commonly complex (Osborne,1984). Future surveys should note the evidence for bed erosion at just one horizon, labelled top bedset 'bs1' in Figure 9 at 25–27m depth. Even within the c.110-year recorded history of the GGMC floor, channel shifting and erosion of fine-grained sediment has occurred (Spencer, 1974; Murphy and Allshorn, 2003). However, none of the other GPR bedset boundaries in the sedimentary pile is erosive at the resolution of the 2003 GPR section.

The lidar survey results suggest strongly that the roof architecture of GGMC is due to stoping. Stoping involves roof collapse and upward-migration of the void below. Stoping should leave conical piles of detritus across the cavern floor. Indeed, the bedset boundaries identified in Figure 9 might have arisen in that way. How then, could 'bedding' develop within what would initially have been piles of angular rubble? The GPR reflection events interpreted as bedding are the most convincing of all, with top-, fore- and bottom-sets present, so any hypothesis must first explain their presence in the GPR section. Perhaps the 'bedding' reflection events are due to finer-grained sediment deposited in low flow regimes within each new boulder pile. Cross bedding within fluvial streams and aeolian dunes develops in the lee of flow over existing obstacles. Flow is relatively unrestricted above the fluid / sediment interface. However, perhaps similar structures could develop within the interstices of a boulder pile as fine-grained sediment washes across the top and settles down through the large spaces and slower-moving water within them.

Sub-horizontal reflection events could be due to a shift in the depositional environment to standing, or very low velocity water within the main cavern. Deposition from low velocity water often leads to parallel sediment beds. Alternatively, low apparent dips could be due to flow across the long axis of the cavern. As outlined above, ponded floodwaters have occurred in GGMC over the past century and their flow might be typical of the depositional environment for the upper 10m of sediment. A harder question is what type of environment the deeper sediments represent. However, the overall appearance of all the bedsets in figures 8–10 is quite similar, though their apparent fore-set azimuths set them apart (see Table 2 and Figure 11). The interval velocity steps of figures 7, 9 and 10 are marked, yet show a steady progression.

The Main Shaft shown in figures 2 and 12 breached the roof of the Main Chamber at an unknown date. It is possible that erosion from surface water and roof migration within the cave system led to an instability in which a significant volume of the roof collapsed within a relatively short space of time, Parr (2004). The influx of sediments direct from the surface might then have dominated subsequent deposition and partially blocked major earlier sinks around East Pot and South Passage at the eastern end of the chamber. Flood flows then redirected to West Slope, as indicated by evidence from the past 150 years.

Other possible explanations for the observed GPR reflection events include:

- Glacial and interstadial periods could deposit materials of different composition and proportions. There were at least five events in the last 100,000 years of the Devensian glacial (Sbaffi *et al.*, 2001). The last 100,000 years of glacial events could account for the number of sediment packages, given that the sequence may be much deeper than imaged. The sediment age range present within the cave is not known. Previous work using, for example, speleothem dating suggests that of some the sediments in the GG–IC System could be up to 200,000 years old or even older (Gascoyne *et al.*, 1983). Table 3 offers a tentative correlation, pending future age-dating results.

a	Air volume above West Slope	538	m^3
b	Air volume above East Slope	2,656	m^3
c	Air volume above flattish floor of GGMC	35,361	m^3
d	Total air volume of GGMC	38,555	m^3
		242,503	US oil barrel
		8,480,914	UK gallon
e	Area of flattish floor of GGMC	2,110	m^2
		0.52	acre
f	Distance around floor area 'e'	1,890	m
g	Estimated sediment volume (e * 30m) GPR depth	63,300	m^3
h	Estimated minimum Total Volume (g + d)	101,855	m^3
j	Volumetric ratio (sediment : air above :: g : c)	1.79	
k	Volumetric ratio (estimated min. total : air :: h : d)	2.64	

Table 1. *Gaping Gill Main Chamber cavern volume and floor area measurements prepared from the lidar survey by 3D Laser Mapping Ltd. A minimum estimated sediment volume is then calculated from the floor area and the minimum depth of sediment imaged by the GPR survey.*

- Void migration through layers of different strength could reduce the volumetric rate and particle size of limestone cobble deposited from stronger, more massive beds and increase the rate and particle size of cobbles from weaker, thinner beds. The proportion of fine-grained sediment would then vary, creating relative dielectric permitivity contrasts.
- Anthropogenic effects or interstadial climate change could also control sediment grain size, once the shaft breached the chamber roof – consider the discussion on 'gripping' under *Previous Observations*, above. Significant deforestation has occurred. At present, one percent of the region is tree covered compared to a majority c.5000 BP, (Coulthard, 1999). Vegetation could control the sediment volume and type entering Fell Beck and deposited in the Gaping Gill system.

Within the GGMC sediments, it is possible that each radar stratigraphic unit represents an interglacial event. The cave system should have been inactive under permafrost conditions during glacial periods. However, the clastic Yoredale Group cover rocks could well have eroded faster under ice. The Fell Beck catchment would then be larger after each glacial interval and able to supply an increased proportion of fines through fissures to the developing cave systems below. Table 3 presents a possible correlation between radar stratigraphic units, GPR interval velocities and interglacial periods. Conversely, each glacial period would correspond to a hiatus in sediment deposition and a step in the velocity profile.

If that were the case, (rsu5, Vint1) of figures 7 and 9–11 would correspond to the present Holocene interglacial period, from approximately 12,500 years ago to the present day. The interval velocity step between radar stratigraphic units 'rsu5' and 'rsu4' of Table 2 and Figure 11 would correspond to the last Pleistocene, or Devensian (Wisconsin in North America) Glaciation, a period of approximately 100,000 years. The next radar stratigraphic unit down the section, (rsu4, Vin2), would correspond to the previous interglacial event, and so on. This argument implies that the roof of GGMC breached at the beginning of the last interglacial event in the Late Pleistocene, the Ipswichian (see Table 3). That radar stratigraphic level is flagged by a solid black block arrow in figures 9 and 10 and by a thick black line in Figure 11 and Table 2. Ipswichian sediments occur in other cave sites in the area (Gascoyne et al., 1983).

Fine-grained sediments are presently on the floor of the cavern and they explain both the observed ponding of floodwater and the lateral redirection of surface flow within the chamber. Since fine-grained sediments occur at the surface, it is therefore reasonable to implicate them in the GPR reflection events and the remarkable stepped interval velocity profile.

It is possible that the fine-grained sediments in the upper sequence only arrived with the breaching of the cavern roof. Interval velocities decrease with decreasing depth and so with the 'younging' of the sedimentary pile (figures 7, 9 and 10). This would be compatible with an increasing supply of sedimentary fines as the roof thinned. Should that be the case, the uppermost fines should match components of the glacial detritus blanketing the moorland above GGMC. In contrast, should they ever be available for sampling, the fines lower in the sedimentary section should either have originated from within the system as residues of dissolution or have entered the system via the fissure network as proposed by Bull (1981).

PROCESS REVIEW

We may now compare the themes discussed above using a steady-state process model. We do this with a simple approach, treating GGMC as a "spherical cow" (Harte, 1985). We have volumes, a process and a timescale. Table 1 summarises the volumes both measured and derived. Figure 13 summarises the processes by which GGMC may be growing. Table 3 proposes a tentative time-scale. Further work could refine all three components of the following 'black box' model. However, it may be useful to consider where the main uncertainties lie and so where further effort could be most beneficial.

Fell Beck provides the input to our "spherical cow" model of GGMC. The Clapham Bents catchment has an area of approximately $2km^2$. Meteorological Office figures for Ribblehead between 1941 and 1970 report an annual rainfall of between 1,344 and 2,159mm. Halliwell (1980) decided on an average precipitation input of 2000mm per year with a standard deviation of 360mm per year. His average estimated evapotranspiration losses were 600mm per year. The average effective rainfall is thus about 1,400mm per year. From multiplying 1.4m by $2km^2$, the volume of water entering GGMC from Fell Beck should be about 2.8 million cubic metres per year.

Halliwell (1977, 1979 and 1980) also analysed the calcium carbonate concentrations of waters entering and leaving the Craven cave systems. Water enters the limestone of the area with a $CaCO_3$ concentration of about 50 ppm. It leaves with a wide range of values depending upon residence times controlled by geology, weather systems and vegetation, as discussed in Previous Observations above. The "large risings" – those most similar to the GG–IC situation – have $CaCO_3$ concentrations of around 90 ppm Therefore, the GG–IC System increases the concentration of $CaCO_3$ in its groundwater by about 40 ppm. Multiplying the volume of water entering GGMC by the concentration of limestone removed implies that Fell Beck removes $112m^3$ of limestone from the CC-IC System each year.

The most difficult question is how limestone removal rates vary around the GG–IC System. One could argue that the largest caverns must be near the sites of greatest aggression. Many such large chambers are 'avens', blind to the surface (cf. Dwerryhouse, 1907). The mechanism of Figure 13 invokes longer groundwater residence times along basement lows and fault steps, as shown by Halliwell (1979), along with void migration from depth. There are over 16km known passages in the GG–IC System and about one known inlet for each kilometre of passageway. At present however, Fell Beck is the dominant input.

The proportion of groundwater flow in fissures compared with flow in passageways is hard to estimate. Rats and Chernyashov (1965) found that fracture permeability is an additive function of joint density and noted a lognormal distribution of permeability in oil and water wells. A detailed hydro-geological model of a system like GG–IC would include some representation of all flowing fractures. Future fracture analyses of lidar surveys from representative parts of the GG–IC System might provide such information, but our present "spherical cow" model requires a simple statistic. For groundwater flow, the ratio of path length in fissures to that in passageways could be in the range of one to two orders of magnitude.

Radar Stratigraphic Unit	Vint [m/ns]	Bed-sets	#fore sets	fore-set azimuth	cumul. azim. cf.East	notes on bedding
rsu_5	0.084	bs_14	2	e-w	30	gentle onlaps
		bs_13	5	west	30	strong onlap to west
		bs_12	6	e-w	35	gentle onlap
rsu_4	0.089	bs_11	6	west	35	onlap climbs west
		bs_10	6	e-w	41	onlap climbs east
rsu_3	0.097	bs_9	9	east	41	onlap climbs east
		bs_8	6	e-w	32	onlap climbs west
		bs_7	6	east	32	top, fore & btm. sets
rsu_2	0.108	bs_6	3	east	26	gentle downlap
		bs_5	3	east	23	gentle downlap
		bs_4	3	east	20	gentle downlap
rsu_1	0.116	bs_3	7	east	17	downlap to east
		bs_2	5	east	10	onlap climbs east
		bs_1	5	east	5	eroded top

Table 2. *Notes on the internal features of the hierarchical stratigraphy interpreted in figures 9 to 11.*

From figures 1 and 13, we can take a simple, straight-line flow length of about 50m from the waterfall in GGMC to the peripheral sinks, perhaps 50m fall through the sediment pile and perhaps 50m over the basement unconformity. The ratio of flow length in contact with GGMC sediments to flow length through the GG–IC System could be 150m to 15 or 16km. That would imply that 1%, or 1.12m^3 of the limestone removed by Fell Beck from the GG–IC System each year would come from the base of GGMC. By assuming a greater sinuosity in the maze-like system south of GGMC, the estimated rate of limestone removal from GGMC could reduce by orders of magnitude, to suit the needs of any particular hypothesis! Therefore, groundwater hardness distribution should be the next most useful topic to investigate in the GG–IC System.

The volume of void space created in GGMC is the sum of the pore space within the sediment pile and the directly observed air space above the chamber floor. If we apply a porosity of 30% to the minimum estimated sediment volume of 63,300m^3 from Table 1, we have a void space within the sediments of 18,990m^3. Adding the measured air space of 38,555m^3 due to void migration, gives a volume of country rock removed of 57,545m^3. Of course, this estimate ignores coarse-grained material from outside the GGMC. It also assumes that the fines presumed responsible for the GPR signals were washed in and so are not part of the material removed by dissolution, i.e. they represent original void space.

Finally, we divide the 57,545m^3 void volume by the annual rate of removal to estimate the time taken to excavate GGMC, whilst building the sediment pile within. If the annual rate of removal were 1.12 m^3 per year, it would have taken only about 51,400 years to create GGMC. However, the total time-span of the last five interglacial events is about 220 thousand years and speleothem dates from other parts of the system imply that GGMC was already drained by this time (Gascoyne *et al.*, 1983). If the time-correlation of Table 3 were plausible, the distribution of limestone removal rates around the GG–IC System would need adjusting by a factor of 5. That is, the limestone removal rate beneath GGMC would be about 0.2, rather than 1.12m^3 per year. Residence time and fissure to passageway ratios could easily vary by such a factor. It should be possible to make further measurements that could reduce such a large range of uncertainty.

CONCLUSIONS

A preliminary appraisal of the lidar survey suggests that the cavern roof formed by the migration of voids, as summarised in Figure 13. Most of the cavern's surface segments are relatively planar and follow faults, joints or overhanging bedding planes. One wall block is moving slowly, showing that the processes of roof and wall stoping and the withdrawal of support by the cave-floor sediments are continuing. Aggressive peaty waters may be acting on the deeper sediments, causing the observed instability of the upper sediments. If sediment removal did not occur, void migration would cease when bulked, stoped material supported the roof. Future lidar surveys could measure block displacements by comparison with the 2003 base survey provided by 3D Laser Mapping Ltd. Process rates might then be calculated.

The overall fluvial flow direction within the maze-like Gaping Gill – Ingleborough Cave System is to the south or southsoutheast (cf. Brook *et al.*, 1991, pages 186–187). Marston and Schoffield's (1962) connection between GGMC and South East Pot, the resurgences at Ingleborough Cave and Clapham Beck Head and the southeasterly bedding plane dips in GGMC (Figure 2) are all consistent with that. However, the axis of GGMC is much closer to west–east. At low volume, Fell Beck splits within the chamber and exits to both east and west. Our deep GPR results and Allshorn's scallop analysis suggest that early-stage floodwater flows in GGMC were towards the east. However, recent floodwaters flow westwards within GGMC, as shown by our shallow GPR results, the Bronze Age human, deer and grouse bones and Professor T McKenny Hughes' inscribed Victorian 'plank'.

The GPR CMP shows a 30m-deep, stepped velocity-depth profile. That suggests a sequence of five radar stratigraphic units, each being from 4 to 8m thick. The velocity step changes are most likely due to changes in clay content and porosity between units. The constant-offset GPR profile shows a sequence of about 14 bedsets, each being up to 6m thick. Below 10m depth, fore-set dips agree with the palaeocurrent direction from scallop marks on the chamber walls. Both show that early floodwater flows were mainly towards the east. However, above 10m depth, GPR fore-set dips show a reversal of the previous flows into a westerly direction, consistent with historical evidence on the direction of floodwater flows.

There appears to be a significant change in local depositional controls at 10m depth. As Parr (2004) suggests, the 10m depth GPR event may record the breaching of the GGMC roof and the direct introduction of fine-grained glacial sediment, coarse-grained Yoredale Group sandstone blocks and aggressive, peaty surface waters to the chamber. Earlier deposits might consist purely of limestone blocks derived from the chamber roof and walls, along with a lower proportion of fines derived from within the system. Such a reduction in fines and/or increase in porosity with depth would be consistent with the observed increase in interval velocity with depth. From the argument advanced in the Discussion, we tentatively assign the date at which the GGMC roof breached to the start of the last interglacial event in the Late Pleistocene, i.e. to the Ipswichian. That interglacial was much warmer and wetter than our present climate.

The historical and archaeological evidence also shows that the sediments forming the Main Chamber floor have ponded floodwaters to a far greater depth than has been directly observed. We infer several floodwater sinks of differing capacity and depth at each end of GGMC. Sediment depth and permeability may control the capacity and availability of individual sinks. Such a mechanism could link the breaching of GGMC roof to a change in flood flow direction from east to west. Of course, in order to sweep the complete chamber floor, other inlets to the east of GGMC would need to have contributed to the local westerly flow. Craven Pothole Club members are currently digging such inlets.

The sequence of events seems to have been:

1. GGMC opened by dissolution along a strike-slip, or tear fault at an unknown, possibly very early, date. It then enlarged by dissolution along the basement unconformity and upward void migration, or stoping. The continued dissolution of fallen blocks prevented any propping of the roof. Water flows in a regional south or southsoutheasterly direction, but is locally constrained to flow either almost eastwards or almost westwards along the long axis of the chamber. Fines left by dissolution or washed into the chamber through fissures winnow out between the stoped

Phases in: North America / *(northern) Europe* / **UK** / Alps	Event	Climate	Culture	End [ka_BP]	Start [ka_BP]	Span [ka]	Possible RSU in GGMC	Vint in GGMC [m/ns]
Holocene	interglacial	as at present +/-	Neolithic etc.	0	12	12	5	0.084
Wisconsin / *Weichselian* / **Devensian** / Würm	3-stage glacial	glacial with permafrost	Abevillian	12	70	58		hiatus
Sangamonian / *Eemian* / **Ipswichian** / Riss – Würm	interglacial with short abrupt glacial events	warm and wet, very warm c.125,000 years BP	Palaeolithic, Mousterian (Neanderthal)	70	130	60	4	0.089
Illinois / *Saalian* / **Wolstonian** / Riss	glacial		Acheulian	130	180	50		hiatus
Yarmouthian / *Holsteinian* / **Hoxnian** / Mindel – Riss	interglacial, perhaps multistage	warm	Clactonian and Acheulian	180	230	50	3	0.097
Kansan / *Elsterian* / **Anglian** / Mindel	glacial maximum	very severe glaciation	Oldowan	230	300	70		hiatus
Aftonian / — / **Cromerian** / Günz – Mindel	interglacial	warm	Oldowan	300	330	30	2	0.108
Nabraskan / *Menapian* / **Beestonian** / Günz	glacial		Oldowan	330	470	140		hiatus
[recognized in Europe] Waalian	interglacial		Oldowan	470	540	70	1	0.116
[recognized in Europe] Donau II	glacial		Oldowan	540	550	10		hiatus
[recognized in Europe] Tiglian	interglacial		Oldowan	550	585	35		? below record
[recognized in Europe] ? Donau I	glacial		Oldowan	585	600	15		hiatus

Source: http://www.nationmaster.com/encyclopedia/ice-age

NB: speculative correlation only with no age dating support yet available.

Table 3. *A tentative correlation between the radar stratigraphy of the GGMC sedimentary succession and the last five interglacial events.*

blocks. They and the scallop marks on the chamber walls, kindly recorded the early easterly current direction for the benefit of Holocene speleologists (4).

2. Pleistocene flood flow directions were mainly to the eastsoutheast in GGMC. Each glacial event enlarged Fell Beck catchment. As a result, each successive interglacial stratigraphical unit within GGMC received a greater proportion of fines through fissures from the surface. Increased fines lowered the e-m interval velocity in each successive layer leading to a stepped velocity profile.
3. The last Late Pleistocene, or Ipswichian, interglacial event was particularly warm and wet, but guiltless. The roof of GGMC finally breached, dramatically increasing the sediment supply from Fell Beck. The additional fine-grained sediments partially sealed the previously dominant eastern sinks within GGMC. Floodwaters then had mainly to exit westwards, before rejoining the regional southsoutheasterly flow beyond the chamber.
4. After the last Pleistocene (Devensian or Wisconsin) glaciation, the Fell Beck catchment had again increased in size. The advent of man brought deforestation, further increasing fine-grained sediment input to GGMC. Fines now form a partial aquiclude over the chamber floor. At low flows Holocene water leaves the chamber to both east and west. However, floodwaters still flow westwards, following the pattern set during the warmer and wetter Ipswichian interglacial period.

A more confident understanding of the depositional environments of the cave sediments requires further work. Several relationships appear to be involved: sediment supply and removal; void migration and backfill; the entrapment of fines and the redirection of floodwater flows. We have shown that lidar and GPR can make an effective contribution to such investigations when combined with other evidence.

RECOMMENDATIONS AND FUTURE WORK

The 2003 GPR and lidar surveys have both been a great success. Repeat lidar surveys will now be an option whenever significant changes occur in GGMC. Differences between the cavern state in 2003 and at the time of any future survey will be measurable to within 25mm (1 inch).

For present purposes, a 30 MB data model is appropriate However, future higher-resolution surveys with more recent lidar instruments could have much larger file sizes. At the time of writing, it is difficult to manage a 180 MB lidar data model. The rendering time between poses becomes tedious. Fortunately, the efforts of the 3D computer gaming fraternity are improving the necessary hardware as fast as humanly possible. Future cavers should be able to wear a 3D headset to compare virtual reality with reality.

The eastsoutheasterly increase in throw of Glover's fault might explain why the boulder pile of East Slope is larger than that of West Slope. A structural geological analysis of the exposed fracture surfaces would help understand that relationship, so we urge students of structural geology to enquire. The efficient classification of point subsets within the lidar data requires software that is still in development at 3D Laser Mapping Ltd. and elsewhere.

There is more immediate scope for building upon the excellent but limited 2003 GPR results. A 3-D GPR survey would remove ambiguities associated with apparent dips and make possible a fully 3-D geomodel of the cave sediments and the limestone host beds. Future GPR work should have a much longer record length and include several CMP velocity profiles. If the Utsi GV2 GPR system were to be available, success should be certain, provided it then has a longer record length capability. If alternative GPR systems are used, careful pre-survey parameter evaluation should aim to emulate the superb signal-to-noise characteristics of the Utsi system.

The volume comparisons above and below the floor of GGMC invite further work on flow rates, rates of sediment removal and the provenances of the cave-fill sediments. Our Process Review highlights a pressing need to know the distribution of calcium carbonate concentrations throughout the GG–IC System waters.

A map of sediment provenance for the system should be instructive. The flow rates needed to move each particle size are known. A similar exercise could log vertical sections of East Pot. At 16m deep, it should straddle Parr's environmental boundary at 10m depth. Although fines only occur at the base of East Pot, ratios of indigenous limestone to exotic sandstone blocks would be of interest.

A key outstanding unknown is the local depth to basement. It is now almost certain that GPR is a suitable tool for such a task.

Mapping the basement unconformity throughout the Gaping Gill system would allow a remarkable 3D 'geomodel' to be constructed. Such a 'static' model would provide a basis for 'dynamic' simulations. Flow simulations could evaluate evolutionary hypotheses for the GG–IC System and comparable petroleum reservoirs. Note that a basement horst block underlies Ghawar, the world's largest oil field (Edgell, 1992) and that fracture-controlled secondary porosity and permeability (Westerman, 1981) has dominated its recent production history (Meyer *et al.*, 2000). The remaining reservoir horizons in Ghawar lie closer to the basement, so the GG–IC System could be a useful outcrop analogue for that and similar oilfields.

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APPENDIX

Processing Step	Parameters applied for shallow reflector resolution	Parameters applied for deep reflector resolution	Parameters CMP
Dewow (time window, ns)	100ns	100ns	100ns
Start time zero (shift in time, ns)	-10ns	-10ns	-10ns
Gain – divergence compensation (scaling value assuming a 2D line source)	0.01	0.05	
Gain – AGC (window length, ns, and scaling value)			150ns, 2
Gain – x-distance decay (correction for decay, dB/m)			1
Bandpass frequency filtering (lower cutoff- lower plateau- upper plateau- upper cutoff, MHz)	0-25-150-250MHz	0-25-150-250MHz	0-25-150-250MHz
Background removal (range for determining reference trace)	Whole line	Whole line	
Static corrections (assumed velocity for calculating two way travel time from height differences)	Not yet applied, as no topography data available at time of processing GPR data.		

Table 4. GPR processing parameters used by Parr (2004).

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Hydrology of the Oronte-Sin rivers karst, northwestern Syria.

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Abstract: A hydrological and speleological study of the carbonates of the Alawite Mountains, between the cities of Banias and Jableh, northwest Syria, involved the Sin and Oronte rivers, and the role played by the Ghab Plain. Discharge of the Sin River shows no decrease in the summer, and therefore differs from most other rivers along the Syrian and Lebanese coast. Karst is well developed on fractures, joints, faults, and rifts in the carbonate rocks, with sink holes, shafts and caves formed in many places. Probably the most important water supply to the Sin River is the Ghab Plain and Oronte River, where water seeps through the alluvium, to flow through unknown caves toward the Sin River's head resurgence.

Keywords: hydrology, speleology, karst, Sin River, Syria

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INTRODUCTION

The study area overlooks the northeastern corner of the Mediterranean Sea, between Banias and Jableh, two cities on the Syrian coast, and lies along the slopes of the north-trending Alawite Mountains, which stretch between Turkey and Lebanon. The carbonate hydrological unit includes two perennial rivers, the Sin (emerging from a large spring) and the Oronte (El Assi), which appear to be connected but are separated by the width of the Alawite Mountains (Fig.1).

The Sin's head is at about 28 to 30m above sea level, and its spring lies below the old coastal highway; about 10km north of Banias. From this point, the land rises gently eastwards over the Alawite Mountains to the divide at 1333 m elevation, then lowers to the Ghab Plain and the Oronte River at about 200m elevation. The boundary between the Ghab Plain and the Alawite Mountains is a major north-trending fault system, which constitutes the most northerly part of the African Rift (Mahfoud and Beck, 2003). The plain was originally swampy, but is now drier farmland (Fig.2). The Oronte River rises in the mountains of Lebanon, flows northwards through Hama and then across the Ghab Plain and into Turkey, before draining into the Mediterranean Sea. It loses a good part of its water as it flows northwards across the Ghab Plain.

This study was based on field observations, concepts of karst hydrology in similar carbonate rocks (Lattman and Parizek, 1964; Eakin *et al*, 1976; Ford and Ewers, 1978), and on published geological and topographical maps of Syria. The main objective of the study was to determine the sources of the water that feeds the large spring at the head of the Sin River continuously, and to assess the probable presence of connected caves between the Oronte and the Sin. The Sin River is notable in showing an increase in discharge despite the heavy uses of its water.

GEOLOGY

The eastern Mediterranean coast of Syria is formed of Cretaceous bedded white limestones, overlying Jurassic limestones and dolostones, which have a regional dip of 15° towards the west. The dominantly carbonate sequence includes intercalations of sandstone, clay, shale, marl, and spilitic basalt lava, as results of sea level fluctuations and submarine volcanic activity (Fig.3). The dominant carbonates are therefore anisotropic and non-homogeneous.

Uplift and folding that created the Alawite Mountains during the Late Tertiary and Early Quaternary included processes of rifting, fracturing, shearing, jointing, and faulting, as a consequence of the successive regional stresses broadly related to the opening of the Red Sea rift. With these discontinuities, the gently-dipping, thinly bedded carbonate rocks are unable to retain water, owing to the widening of fractures, joints, faults, and bedding planes by the dissolutional activity of unsaturated meteoric water. The seaward side of the rifted anticlinal structure of the Alawite Mountains has many deep and narrow ravines carved by run-off of the winter rainfall. However, there are scattered springs, probably formed by diffuse flow from dolomite or shaly limestones within the succession (as advocated by White, 1969). The eastern limb of the anticlinal structure was displaced along a normal fault that now forms the western boundary of the Ghab Plain and has a throw of some hundreds of metres. Subsequently, the Ghab valley was filled with alluvium and colluvium (Fig.2).

In the limestones between the Sin River's head and the Ghab Plain, a zone inland from Jableh has a noticeably higher density of recorded faults than the ground inland from Banias. Moreover, the geological map indicates at least three faults that connect the Ghab Plain with the other faults, thus forming a fault-net oriented towards the Sin River's head. Within the mountains, two grabens are

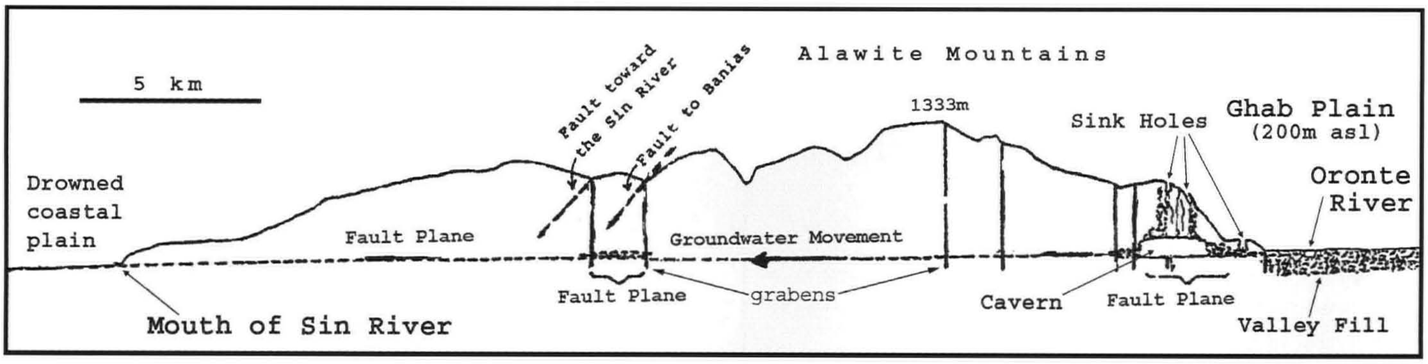

Figure 1. A west–east topographical profile through the Alawite mountains from the head of the Sin River to the Oronte River on the Ghab Plain, showing faults (including two graben structures), sinkholes and a conceptualized cavern.

Figure 2. The Ghab Plain seen from the Alawite Mountains.

recognized (Fig.1); they may be the result of subsidence related to the karstic widening of the fault planes by unsaturated rainwater. Most of the recorded faults are normal, and were probably formed in response to northeast–southwest compression. A major curved fault south of Banias (Fig.3) conforms with the outline of an older dome structure present to the south. Small faults adjacent to the Ghab rift were probably formed during the rifting event, and so post-date those in the mountain interior.

HYDROLOGY

The Mediterranean Sea coastal area, including the western slope of the Alawite and Lebanon mountain ranges, is a region of Mediterranean climate (Mahfoud and Beck, 2003). Summer is generally dry to humid, with occasional light rain. In winter, heavy rains occur, and mountains are snow capped for 2 to 3 months. Snow patches are always preserved on the Lebanese mountains until mid-summer. Annual precipitation may reach over one metre along the western slopes, with temperatures ranging between -5° and more than 30°C (Mahfoud and Beck, 2003). In winter, ten or more rivers flow to capacity seaward across the western slopes of the Alawite and Lebanese mountains; the Sin River is one of these. In addition, many smaller streams also flow during the wet season. Rivers and streams are usually fed by winter precipitation.

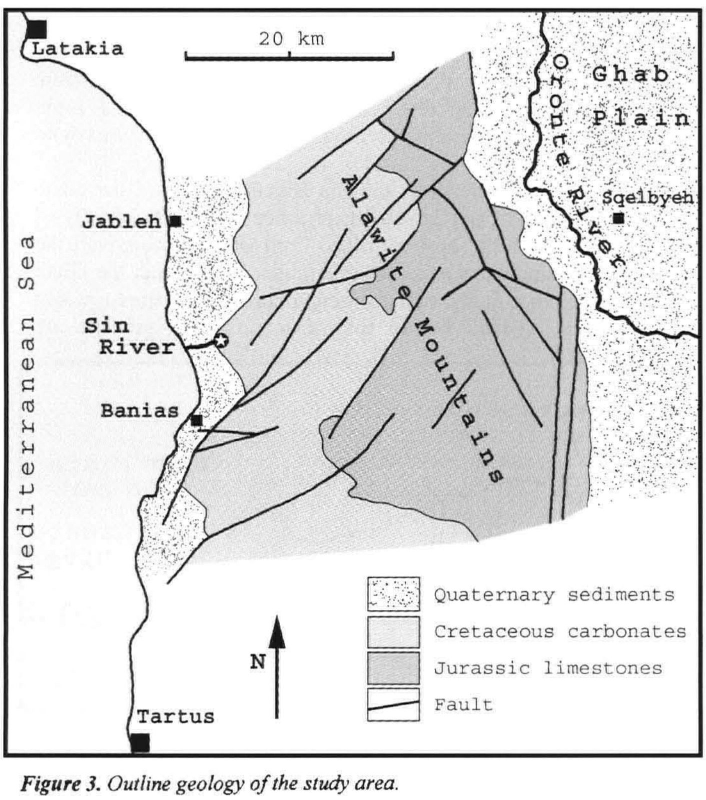

Figure 3. Outline geology of the study area.

Only the Sin and Banias rivers remain flowing to capacity all year long, while the others lose a great deal of their water during the summer. Rivers flow westward in short channels across the narrow coastal plain between the mountains and sea, and their water is used for limited crop irrigation and human consumption. The karstic resurgence at the head of the Sin River maintains a steady flow, which is used to irrigate the plains of Jableh, and also to provide drinking water for the cities of Latakia (whose population has grown from 40,000 to over 500,000 inhabitants during the last forty years) and Tartous (with a population increasing from 20,000 to over 200,000 in the same period). In addition, many coastal towns and villages take water from the Sin River both for irrigation and for home consumption. Fifty years ago, a small creek separated from Sin River's head, and flowed southwestward; today this creek has become a river, despite the heavy uses of the Sin's water. These scenarios would suggest the presence of a year-round continuous feeder to the Sin River.

Near a highway on the southern outskirts of Banias, artesian water suddenly burst out, shooting up a jet of water about one metre high and 100–150mm in diameter. This artesian jet remained active for more than two weeks until it was capped by local engineers. The water reservoir within this confined aquifer could be an off-shoot from the main aquifer that feeds the nearby, and higher, head of the Banias River. In the same area many springs lie offshore close to the coast, and sailors on their boats commonly take their fresh water supplies from those springs.

SPELEOLOGY

The probable presence of cave systems in the karstified carbonate rocks of the Alawite Mountains (Fig.4) is supported by the occurrence of scattered sinkholes, shafts, open cave entrances and the lack of surface streams. At about 1000–1200m altitude, a circular closed depression a few hundred metres across and over 100m deep (Fig.5), lies east of the summer resort of Slenfeh, above the western edge of the Ghab Plain and about 45km east of Latakia. It is not known whether this is a solutional doline or one formed by collapse.

Figure 4. Bedded white limestone forming karst topography.

A vertical shaft, over 10m in diameter and of unknown depth, has been reported near the mountain village of Rum Tarzeh, about 18km northeast of Banias. This shaft was probably formed in the vadose zone. About 4km east of Banias, near a highway passing through Barmaya village, three adjacent small springs issue from heavily fractured and brecciated dolomitized limestone; one of these discharges from a small cave. Larger caves are known in similar karst terrain farther south, including the famous cave of Jeita in Lebanon. Within the study area, a series of sinkholes occurs in a line across the Alawite Mountains, along the northern boundary of the Massiaf topographic quadrangle (Fig.6).

It appears that there is every likelihood of substantial cave passages extending beneath the Alawite Mountains, but none has yet been entered, and much of the trunk drainage route could be flooded beneath the resurgence level at the head of the Sin River. The underground hydraulic gradient between the Ghab Plain (at about 200m elevation) and the Sin River's head (at an altitude of 28m) is 4.8m/km, or 0.0048, over a distance of about 35km.

DISCUSSION

A critical question to ask is, what makes the water flow of the Sin River so abundant and almost inexhaustible when compared with the smaller flows of each of the other coastal rivers and streams, when they all appear to gain their recharge water from almost uniform winter precipitation. All the rivers and streams flow to capacity in winter. However, excluding the Sin and Banias rivers, these watercourses all decline noticeably in flow during the summer, and some streams dry up completely. The Sin River supplies water continuously for irrigation and home consumption, with no significant seasonal decline in flow. In the 1980s, it was reported that the Sin River discharged about 20m^3 of water per second during the summer.

It appears that the extra water derives from a wide area along the western side of the Alawite Mountains. Numerous faults, joints, fractures and bedding planes in the limestone karst have been widened by the unsaturated groundwater flow to form a system of caves that connects the Ghab Plain to the Sin River's head, and thereby sustains the high flow in the Sin River. It appears that there are two trunk drainage systems that feed to the Sin and the Banias rivers. The more densely faulted and fractured zone behind the Sin River feeds the larger flow to that rising, with lesser flows passing through both north and south. In addition, the most important source of the extra water to the Sin and Banias rivers is groundwater that seeps abundantly and continuously from the Ghab Plain and the bed of the Oronte River (Fig.7). It is significant that the Oronte River reduces in size steadily as it flows northwards across the Ghab Plain; in some summers its flow declines to nothing around the town of Jisrash Shughur (northeast of Latakia and just off the Fig.3 map).

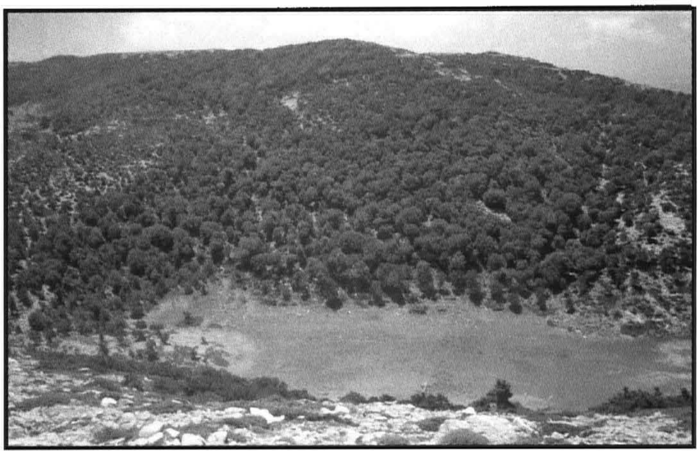

Figure 5. The large karst depression east of Slenfeh, with its wooded slopes and grassy floor, seen from its northern rim.

The catchment basin behind the Sin River is about 250km^2, though this estimate may be greatly modified if karstic groundwater divides do not match the topographical divides. This size of catchment collects a rainfall equivalent to a mean flow of about 8m^3/sec. Allowing for evapo-transpiration and winter surface run-off, it may be estimated that about one third of this sinks to recharge the groundwater reservoir, to yield a mean flow of about 2.7m^3/sec. However, the Sin River flows at around 20m^3/sec. The extra 17.3m^3/sec may be partially supplied from a much wider underground catchment within the karst mountains, but much of it must derive from losses of the Oronte River and infiltration through the Ghab Plain. The unaccounted flow of the Sin River appears to equate well with the losses of the Oronte River.

In recent years there has been a notable increase in the area under irrigation by Sin River water, an increase in the cities and towns that abstract their water supplies from the Sin, and an increase in the flow of water in the new distributary creek near the Sin's head. It would be logical to suggest that there is a steady increase in the Sin's discharge (though records of flow data are not available). This implies that more and more water is leaking from the Oronte

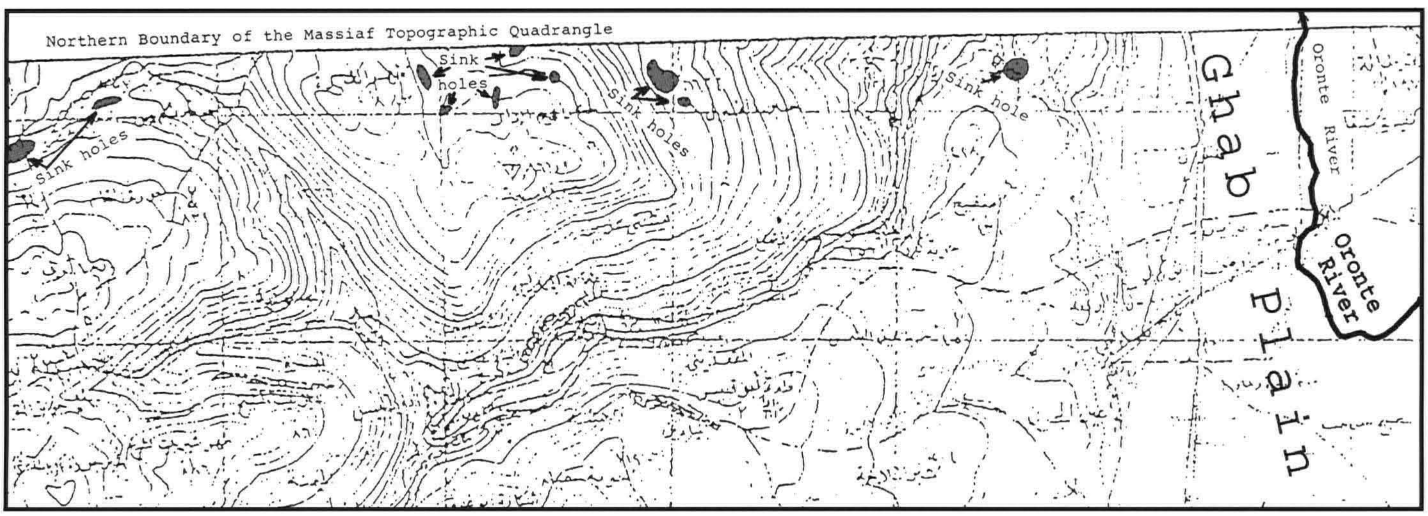

Figure 6 Topography along the northern boundary of the Massiaf quadrangle, showing an alignment of sinkholes across the mountains west of the Ghab Plain.

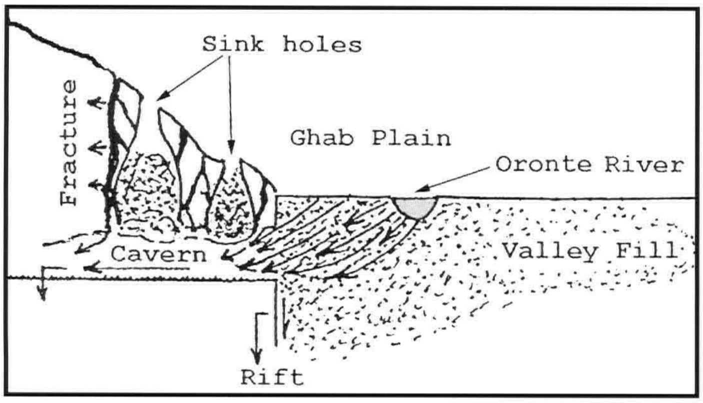

Figure 7. Conceptual model with arrows showing groundwater flow from the Ghab Plain and Oronte River, into the karst towards the Sin River.

River and the Ghab Plain. Karstic maturity therefore appears to be increasing, though such rapid changes may be due to the clearance of clastic sediment from the sinks, rather than to dissolutional enlargement of the conduits.

The main underground channels beneath the Alawite Mountains may lie above an impervious shale that crops out in locations south of Banias, and they may also occupy caverns along fault zones within the karstified carbonate rocks, perhaps beneath observed sinkholes. The distance between the Ghab/Oronte input and the Sin output indicates a karst drainage system extending for at least 35km, but how much of this is potentially accessible cave remains to be determined.

ACKNOWLEDGEMENTS

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ABSTRACTS OF THE 16TH BCRA CAVE SCIENCE SYMPOSIUM SCHOOL OF GEOGRAPHY, EARTH AND ENVIRONMENTAL SCIENCES THE UNIVERSITY OF BIRMINGHAM 5 MARCH 2005

Re-assessment of solutional erosion rates derived from erratic-pedestals and implications for interpretation of karst landscapes in Northern England.

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Interpreting certain surface limestone landforms in Northern England presents problems when using conventionally accepted solutional lowering rates derived from landform relationships between glacial erratics and pedestals in the limestone beneath. These rates are as high as 50 cm of solutional lowering in the past 15k years since the last glaciation, equivalent to 33mm/ka, and suggesting that forms of c. 1m dimensions could be post-glacial. These are mostly in the Yorkshire Dales and Cumbria, at relatively high altitudes, on interfluves, or in sheltered locations with respect to likely ice-flow, or around edges of large surface karst depressions. They include rounded rock outcrops and boulders, wide grikes and grike holes, and tor-like forms, all in circumstances suggesting mature karstification, and problematic to interpret using conventional 'high' solution rates. This is emphasized by their proximity to apparently less 'mature' forms, but other locational characteristics play a part.

The problems prompted re-examination of erratic-pedestal sites. In the literature erratic-pedestal sites include: Farleton Knott, Cunswick Tarn, Norber, Scar Close, Scales Moor, Burren, Leitrim, South Wales, and Marenberg, Switzerland. Of these, Norber is probably the best-known. Its established interpretation is that the pedestals result from protection by erratics of the limestone beneath from rainwater solution.

Norber's field characteristics provide a challenge to this interpretation. It is an interfluve sloping gently to the east. The bedrock is weak, well-bedded limestone inclining at a gentler angle, also to the east. The result is a stepped surface. This supports a new interpretation of Norber. The erratics are on steps, and the so-called pedestals are steps or outliers of steps. This is clear on sideways viewing, where the angular differences are visible. Other landform sites indicating high solution rates have also been reinterpreted: at Scar Close lower parts of pedestals are explained by solution under damp peaty soil. Boulder can only have protected the upper parts, an effect limited here to c. 12 to 18 cm. Other field sites have been re-assessed and data from all sites are summarized.

Conclusions and implications are that there is no evidence for the high solution rates at any re-assessed site. The Norber 'pedestals' do not even give solution rates. 'Real' pedestals on stronger, less-fractured limestones such as at Gaitbarrows, do not exceed c. 20 cm in height, several are much less. Weak limestones are not useful for solution rate assessment due to the effects of mechanical processes. Solution rates on dry interfluve areas are LOW: of the order of 5 to 20 cm in 15 ka, equivalent to 3 to 13 mm/ka.

The implications of lower solution rates for landscape development are that more and smaller karst landforms than previously thought must have, at least partially, survived the last glaciation. This is a more comparable situation to other rock types elsewhere, for example, Cairngorm granites, with similar erosion rates, of a few mm per ka. Larger scale landscape development discussions need modification with lower rates of solutional lowering. The well-established concept of 'karst immunity' needs to be addressed.

There are many lessons here: if old material feels wrong, re-assess it, it might be wrong; never underestimate geological structures; never ignore mechanical processes in limestone areas - on weak limestones they are more important than solution; and consider karstic immunity when thinking about the age and development of any karstified massif.

Historic and Prehistoric hydrological changes in Gaping Gill Main Chamber, North Yorkshire, U.K.

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The 98m-deep shaft of Gaping Gill is Britain's best known pothole. Since the first descent by the eminent French speleologist E. A. Martel in 1895 changes in the sediment distribution on the floor of the Main Chamber can be charted along with the occurrence of active collapse and subsidence of collapse debris into the main Chamber floor. Evidence of ponding of flood water has been recorded on a number of occasions though has rarely been directly observed. Archaeological evidence suggests the Main Chamber may have flooded to greater depths in Prehistoric times. In order to image the sediment fill below the present floor of the main Chamber a ground penetrating radar survey was undertaken. This showed the fill is at least 30m deep and has a complex depositional history.

Can stalagmites capture a record of S pollution in the atmosphere?

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There has been increasing interest in recent years in the wide variety of different chemical species which are preserved within stalagmites. Many of them vary on an annual basis reflecting the seasonal rhythm of changing cave air composition, or water inflow.

Sulphur is a particularly difficult element on which to conduct microanalysis at low concentration, but our recent work at the European Synchrotron Facility at Grenoble has enabled the variation of S with time in two stalagmites to be determined. At one site where there is little vegetation cover, a spiky S record of an interval about 5 thousand years old resembles the volcanic S records found from ice cores. At a forested site, a record covering the last 140 years displays an overall increase reflecting the rise of atmospheric pollution in the 20th century, but in this case there may be modifications because of ecosystem S storage.

Up to now there hasn't been a generally applicable way to examine sulphur pollution records in inhabited regions, but the extension of this approach, coupled with isotopic techniques, should prove fruitful.

Known Nullarbor caves - just scratching the surface? Quantifying unexplored cave volume using microgravity and draught measurements

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The Nullarbor Plain is one of the world's most extensive limestone areas (>200,000km^2) and its subdued topography ranks it also amongst the largest homogeneous land units. Many blowholes exist in the area, which display strong barometric draughts, but generally connect to voids too small for direct exploration. In contrast, only few, but spectacularly large cave systems, accessible through collapse dolines, have been discovered here. The number and size of these caves is at odds with the relative abundance of blowholes and the extent of limestone, and the question of their origin and true extent is highly contentious. Since conventional cave exploration approaches have failed to establish the specific origin of the draughts, a land surface-based microgravity surveying and draught monitoring programme was carried out at three blowholes. This approach allowed remote detection of the presence and shape of inaccessible cave passage, evaluation whether the strong barometric draughts from blowholes are generated by vast unexplored cave systems or arise from air contained within the porosity of the limestone itself as previously suggested.

All blowholes investigated were associated with large gravity anomalies and blowhole draught volumes measured generated by changes in atmospheric pressure were in accordance with the cavity volumes determined by microgravity. The results suggest that blowholes may typically be linked to substantial underground passages, which provide the air reservoirs responsible for draughts. Combined with additional microgravity surveys carried out over potential extensions of several known cave systems, we were able to produce a considerable body of geophysical evidence suggesting that the known deep Nullarbor caves are relatively small parts of much larger systems. Since the number of blowholes on the Nullarbor Plain vastly exceeds known caves (by more than 1000 to 1) it is inferred that the karstification of the Nullarbor has produced much more extensive cave passage than previously thought. Given the nature of local karstification processes and age of the Nullarbor Plain, these findings have considerable implications for our understanding of karst processes in low-lying carbonate platforms, and also climate and sea level history in the region.

A High-Resolution Climate Record for the Last Millennium From a Scottish Stalagmite

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Stalagmites are recognised as important continental archives of palaeoenvironmental information, they can be accurately dated and their subterranean location means that they can accumulate undisturbed for thousands of years. This work takes advantage of the recent developments in the High-Resolution analysis technique of Laser Ablation Gas Chromatography Isotope Ratio Mass Spectrometry (LA-GC-IRMS) to produce High-Resolution stable isotope records and hence a climate record for Northwest Scotland. NW Scotland is a site located on the North Atlantic Seaboard and an area that is thought to be climatically sensitive, particularly in the relationships between precipitation and the NAO and temperature and ocean circulation via the North Atlantic Drift. Initial laser ablation $\delta^{18}O$ results on a 1000 year old annually laminated stalagmite show oxygen isotope variations in the stalagmite calcite which is greater than can be attributed to temperature change alone. This suggests that the stalagmite is recording a combination of environmental factors and highlights the need for detailed calibration of the modern cave system. An understanding of how these surface climate signals are transmitted through an ^{18}O proxy to a stalagmite via the soil and groundwater system is essential. The cave system from which the stalagmite was sampled was monitored over a 12-month period. Surface and cave drip waters were collected and analysed for a variety of isotopic and geochemical parameters including $\delta^{18}O$, δD, $\delta^{13}C$. Cave climatology was monitored for temperature, humidity and carbon dioxide concentration. Drip rates were monitored in order to understand the hydrology of the overlying karst system. Initial monitoring results indicate a seasonally variable cave temperature (4–9 degrees C), high relative humidity (100 %); cave air CO_2 concentration also follows a seasonal pattern. Waters collected from the surface show seasonal variation in $\delta^{18}O$ and δ^2H composition with more positive values in summer. Drip waters show little isotopic variation indicating a substantial degree of mixing in the epikarst. Monitoring suggests that the cave is sensitive to a number of parameters whose dominance may change through time. The role of ventilation and hence kinetic effects upon isotopic fractionation may also play a part.

Symmetry and asymmetry in mean dimensions of cave passages on each side of ridges aligned N–S in central Scandinavia

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The mean lengths, vertical ranges, cross-sections and volumes of 'relict' caves in the metalimestones of central Scandinavia that are on the western sides of major N–S ridges and that are above marine limits (below which cave enlargement by marine erosion at the onset and culmination of glaciation can complicate the picture) are greater than those on the eastern sides, whereas the mean dimensions of the active 'mainly vadose' caves are similar on each side. The mean dimensional differences of the 'combination' caves, which contain relict passages above active streamways, are commonly intermediate between the two extremes. However, there are far fewer caves of all three hydrological classes on the western sides of major ridges, despite similar occurrences of limestone outcrops.

The likely explanation is that cave inception started from the tectonic creation of relatively large fractures by a pulse of local isostatic rebound earthquakes that followed the eastward recession of the ice margin at the end of each major glaciation. The seismic effects were more muffled by the weight of the continuing ice sheet on eastern sides, resulting in the creation of shorter, shallower, narrower but more numerous fractures there than on the western sides. The relict passages, which are independent of present catchment areas, enlarged phreatically under deep ice-dammed lakes, primarily during Weichselian deglaciation, with asymmetrical occurrence frequencies and mean dimensions that follow those fractures created during previous deglaciations. The active passages primarily developed in the Holocene after the disappearance of the ice, in quite different interglacial, mainly vadose, conditions (after a phreatic 'kick start'), with maximum dimensions related to present catchment areas. They utilised the subset of fractures that provided a suitable hydraulic gradient within the local topography, which therefore had more-symmetrical W and E distributions, and also utilised fractures created by the later postglacial neotectonic seismicity of the decaying regional isostatic uplift, which were probably completely symmetrical relative to western and eastern slopes.

A mid Holocene high resolution ^{18}O stalagmite record from Ethiopia

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Ach-1, a 24 cm length stalagmite from the Mechara region of Ethiopia has been dated by lamina-tuned TIMS U-Th dating to 4760±60 BP until 4275±25 BP. The sample is visibly laminated, with 451 annual laminations. Lamina width varies between 1.66-0.12mm, averaging at 0.53 mm. The lamina widths are in agreement with those expected from the modern drip water temperature of the cave (21 °C) and measured drip water calcium ion concentrations (~200 ppm).

Oxygen isotopes from within speleothems can potentially record paleoenvironmental signals. Previous studies from Oman have shown oxygen isotopes reflect the moisture availability when the speleothem was deposited, with negative values signifying increasing moisture, and with a strong correlation between lamina width and ^{18}O. The $\delta^{18}O$ isotopes from Ach-1 fluctuate between -2.4 to -4.5 ($\delta^{18}O$), with no correlation with annual lamina width (R^2 = 0.01) due to the significant within lamina thickness variability due to fluctuating stalagmite shape. Variations along individual growth layers demonstrate greater lateral variation in ^{13}C and ^{18}O than temporal variations, suggesting that the sample was deposited out of isotopic equilibrium. However spectral analysis shows that for both ^{18}O and ^{13}C, there are statistically significant periodicities of 65-81 years and 20-22 years.

Ach-1 ^{18}O is isotopically lighter than modern straw stalactites (-0.4 to -1.2 ‰), and than that predicted (-2.8 to -1.5 ‰) to form from modern day precipitation in Ethiopia (from Addis Ababa, 175 km to the W, IAEA monthly dataset). Modern drip data and straws all fall on the MWL, suggesting there is no evaporative effect, which agrees with humidity measurements made within the modern cave environment (RH > 85%). The isotopically lighter ^{18}O observed in Ach-1 than predicted from waters or observed in modern straws can be explained by lighter drip water ^{18}O, deposition of calcite that was closer to isotope equilibrium, or less evaporation in the cave or overlying soil. Given the lack of observed evaporative effects in the cave today, we interpret the $\delta^{18}O$ record as one of wetter conditions in the mid Holocene, possibly due to a stronger ITCZ forced by solar variations.

Rapid speleothem growth in New St Michaels Cave, Gibraltar

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New St. Michaels Cave was discovered in 1942 when an access tunnel driven into the lowest part of the well known St Michaels Show Cave exposed a new rift system that led to a series of highly decorated chambers containing a lake. Approximately 150m from the entrance rift is a wet area with active flowstone and straw formation with small stalagmites overgrowing broken debris. A specimen of an actively growing stalagmite approximately 10cm tall representing growth post-1942 was collected for dating, high resolution oxygen isotope profiling by laser ablation and trace element analysis.

The specimen consists of 45mm of pale amber laminated calcite overgrowing what is thought to be a broken stump that was damaged during early exploration of the system. U-Th dating of the base of the new growth gives an age of 90±20 years and this, coupled with an estimated 70 annual growth bands counted from the top of the specimen, is reasonably consistent with growth commencing around 1942. The lower portion of the specimen has a complex growth history with U-Th dates obtained so far giving Pleistocene ages around 180ka.

As historic weather records and monthly rainfall oxygen isotope data have been monitored within 3km of the cave entrance this site may provide an opportunity for calibration with the instrumental climate record at a sub-annual resolution. The annual variation of Gibraltar monthly rainfall oxygen isotope data from 1960-2000 shows decadal variations of around 1‰ but systematic variations of up to 4‰ are seen from the wet winter season to the dry summer season. In order to better understand the seasonal timing and speleothem growth forcing mechanisms, a monthly monitoring and sampling program was established in June 2004 in collaboration with members of the Caves section of the Gibraltar Ornithological and Natural History Society (GOHNS). Samples of external air, soil gas and a transect of air samples into the cave have been collected for CO_2 and CH_4 isotope analysis; drip and lake water samples are collected for oxygen isotope, anion and trace element analysis. A novel aspect of this monitoring is the analysis of methane whose behaviour strongly contrasts with that of CO_2. CH_4 concentrations in cave air are expected to be controlled by open ventilation since soil is a methane sink and methane is not transported into the cave via solution in water. There is concern that calcite deposition may be enhanced by elevated Ca concentrations derived from man-made sources related to leaching of lime cement in the surrounding area. If this is the case then isotope records must be treated with caution. However trace element and stable isotope compositions of post 1942 calcite are not unusual and fall within the range of variation observed in Pleistocene calcite from the same site. Oxygen isotope compositions of young calcite vary from –3‰ to –5‰ and individual growth laminae possess less 0.3‰ variation in oxygen isotopes suggesting that isotopic equilibrium has not been strongly affected by degassing or evaporation. Although is no evidence yet for anthropogenic effects related to building and change of land use contributing to the rapid deposition of calcite at the site, this issue continues to be carefully investigated. Oxygen isotope composition of calcite has been obtained by laser ablation at 500 micron intervals providing an isotope record of calcite growth at a sub-annual resolution. Work is in progress to investigate how these data relate to the local rainfall record.

A microclimatology study of two caves: Shatter Cave, SW England and Uamh an Tartair, NW Scotland

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Microclimate data were collected in two contrasting caves with the objectives of (1) establishing temperature and humidity change with distance into the caves (2) establishing if a time lag between changes in surface temperature and cave air temperature is in operation and (3) establishing cave air temperature response to changes in external weather parameters.

Using a Kestrel 3000 multi-parameter, observations were collected at progressive distance into Shatter Cave (OS Grid ST657475) and Uamh an Tartair (OS Grid NC276206) at selected points such as change in passage shape or major decorative features. Gemini Tinytag Plus data loggers were set to record every five minutes for two weeks at Shatter Cave and four days at Uamh an Tartair during July 2004. The loggers were placed outside, in the first chamber, last reachable chamber and approximately midway within the caves. Surface temperature was collected on site and wind direction, wind speed, wind gust and barometric pressure data were provided by weather stations at Curry Rivel in Somerset and Stornoway, NW Scotland.

The results show that air temperature in both caves generally behaves similarly to the profile created by Wigley and Brown (Wigley, T M L and Brown, M C, 1971, Geophysical applications of heat and mass transfer in turbulent pipe flow, *Boundary-Layer Meteorology*, 1, 300-320). However, the profiles for Shatter Cave and Uamh an Tartair do not include a slight increase in temperature near the entrance of the cave as does Wigley and Brown's (1971) model. The profile for Uamh an Tartair shows two increases in air temperature. This does not fit the exponential decline in air temperature that is occurring with distance into the cave. These selected points are in close proximity to the stream that runs throughout the cave. This suggests the stream has a regulating influence upon air temperature within the cave. Shatter Cave does not have a stream and thus is not subject to this air temperature regulation, but temperature logger data suggest a mid cave (currently unknown) surface climate connection exists.

In conclusion for a cave with simple morphology such as Shatter Cave then Wigley and Brown's (1971) air temperature and relative humidity profile can be applied. For Uamh an Tartair, which is a much more complex cave, then the models do not apply quite so well. The stream has a regulating influence on air temperature, and relative humidity reaches 100% at a short distance into the cave. Lag times between change in surface and cave air temperature do not operate within either of the caves studied. Another entrance to the free atmosphere is present within the middle section of Shatter Cave where air temperature frequently correlates strongly with external weather parameters.

Statistical Processing of Speleothem Time Series

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Since the initial discovery that stalagmites are capable of providing an archive of past climatic conditions, there has been a significant advance in both the quantity and quality of the data collected. However, the utilisation of rigorous statistics to assist in the evaluation of results and to communicate with the climate modelling community has lagged behind. Frequently, the data are not conducive to the application of common statistical techniques due to the non-uniform nature of the series over time. However, this does not qualify as a reason to neglect the statistical processing of data from stalagmites and the rigorous evaluation of speleothem time series remains an important goal for researchers.

Presented here are the results of modelling and statistical analysis of two stalagmite climate proxies, annual growth rate and subannual trace element series. The Northern Hemisphere temperature series for the past 500 years has been successfully modelled using annual stalagmite growth rate as a proxy, with respect to the instrumental record and a number of previously published paleoclimate reconstructions.

Annuality in speleothem records can be displayed by both laminae and trace element variations. Spectral analyses for unevenly spaced time series reveal a significant annual signal in Mg, Sr and P series. Further processing of trace element series using wavelet analysis show the annual signal to be temporally unstable and also display evidence of some low frequency behaviour. The presence of annual cycles in the trace element series offers a potential chronology building tool, enabling automated cycle counting.

Evidence for drought in southern Europe 1100-1200 yr. BP from annually layered cave precipitates

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Annually resolved trace element records from speleothems are quickly becoming established as useful and reliable palaeoenvironmental proxy indicators, in particular with reference to palaeoprecipitation records. This study uses high resolution chemical analyses to examine the behaviour of selected trace elements over a known climatic event during the Holocene in speleothems from hydrologically sensitive regions in southern Europe - Grotte de Clamouse, southern France and El Refugio, Malaga, southern Spain.

The stalagmite samples from Clamouse contain discrete layers of aragonite dated at 1100-1200 years BP. These layers are unique as they occur within calcitic speleothems and have been suggested as representing an period of aridity in the region lasting ~100 years. U-Th dating of samples from Refugio in southern Spain revealed an apparent growth hiatus at a similar time to the growth of aragonite at Clamouse. Lamina counts combined with U-Th dates from Refugio samples show the date of the apparent hiatus layer is 1104 yr. BP.

Ion microprobe work was carried out across this time interval in both sets of samples to investigate trace element behaviour, which can allow an insight into the behaviour of the karst system overlying the cave. In addition trace element analyses are used to infer possible similarities between the two sites in terms of palaeoprecipitation patterns and the timing of maximum aridity. This period in time 1100-1200 years BP also correlates with a severe and prolonged drought in central America which has been linked to the demise of the Mayan civilisation.

Suspended sediment dynamics in the Castleton karst, Derbyshire, U.K.

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It is well known that carbonate dissolution is the dominant process in the earliest stages of speleogenesis in limestone aquifers and that flow is laminar and therefore unable to transport suspended clastic sediment. However, after breakthrough is achieved and flow becomes turbulent sediment can be transported through the dissolutionally enlarged conduits. Newson (1970) was one of the first to undertake detailed studies of sediment transport and mechanical erosion by streams in limestone areas but the focus of most subsequent process studies was firmly on solute dynamics and the flux of clastic sediments in carbonate aquifers was identified by White (2002) as an area that had been largely overlooked. More recently a small number of studies have been published which add to this knowledge but none in the UK. Water quality is profoundly influenced by mobile particulates, and particulate matter, especially colloidal, has a high capacity of transporting bacteria. Hence the study of suspended sediment in karst aquifers has both a theoretical and an applied dimension. In this paper we report the initial stage of a study of suspended sediment dynamics in the Castleton karst.

Earlier studies in the Castleton karst, including a paper with a similar title (Bottrell et al., 1999), have established that modern stream sediments in the Castleton caves have a similar mineralogy to the solifluction deposits in the Rushup Vale allogenic catchment but the mineralogy of the loessic soils in the autogenic catchment is substantially different. Erosion in the allogenic catchment is therefore considered to be the primary source of modern sediments. Hardwick (1995) provided initial estimates of erosion rates in two of the allogenic catchments : P6 (improved pasture, 16.1 - 29.5 t km^{-2} a^{-1}) and P10 (unimproved pasture, 1.5 t km^{-2} a^{-1}) and our study aims to quantify sediment outputs from the three Castleton Springs : Peak Cavern Rising (PCR), Slop Moll (SM) and Russet Well (RW). On the basis of a preliminary rating curve the total outputs for the 2004 calendar year are estimated at ~232 t. Potential sources of error are the rating curve, which does not cover the highest flows, and the anomalous discharge pattern that is thought to result from flow switching (Bottrell & Gunn, 1991). However, even if the errors are in the order of +/- 25% the erosion rate in the 5 km^2 allogenic catchment must be at least 35 t km^{-2} a^{-1}, substantially higher than estimated by Hardwick. Alternatively, sediment stored within the conduit system is being eroded and removed from the system, i.e. output is presently greater than input.

A more detailed analysis was undertaken of an individual storm event from 21 to 22 January 2005. The total yield over a 15.5 hour period was ~5.9 t of which ~52% was discharged from Slop Moll, ~31% from Russet Well and ~17% from Peak Cavern Rising. The average minimum, median and maximum particle sizes for samples from Slop Moll were greater than for samples from Russet Well and from Peak Cavern Rising.

Bottrell, S and Gunn, J. 1991. Flow switching in the Castleton karst aquifer. *Cave Science*, 18, 47-49.

Bottrell, S, Hardwick, P and Gunn, J. 1999. Sediment dynamics in the Castleton karst, Derbyshire, UK. *Earth Surface Processes and Landforms*, 24, 745-759.

Hardwick, P. 1995. The impact of agriculture on limestone caves. Unpubl PhD thesis, University of Huddersfield.

Newson, M D. 1970. Studies in chemical and mechanical erosion by streams in limestone terrains. Unpubl PhD thesis, University of Bristol.

White, W B. 2002. Karst hydrology: recent developments and open questions. *Engineering Geology*, 65, 85-105.

Preliminary evidence for the 8.2 ka cold event in a stalagmite from Pippikin Pot, Yorkshire Dales

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A 59cm-long stalagmite was collected in October 2001 from Pippikin Pot in the Yorkshire Dales to study Holocene palaeoclimate variability. Stable isotope samples were taken at 3-4mm intervals over the full length of the speleothem. Carbon isotope values were high and highly variable indicating that the calcite was not deposited in equilibrium with soil CO_2. The oxygen isotope record was largely invariant with a mean $\delta^{18}O$ value of -4.75 ‰, similar to modern-day rainfall in this area. The only significant oxygen isotope excursion occurs at 37.7cm from the top of the stalagmite and reaches a maximum of -5.55 ‰ (a negative excursion of up to 1‰). This event was further investigated using ICP-MS U-series dating.

Three preliminary U-series ages for the initiation, peak and close of the isotopic excursion (a 5 cm long section) enabled this event to be confidently assigned to the 8.2 ka event (dating errors are less than 0.1 ka). This event lasted for approximately 200 years and is comparable in magnitude and duration to other circum-Atlantic climate records of the 8.2 ka event, including the Greenland ice cores and the Hawes Water lake record (approx. 30km west of Pippikin Pot). The Pippikin Pot record is significantly different in magnitude and duration from the 37 year and 8 ‰ $\delta^{18}O$ excursion recorded at 8.2 ka in the Crag Cave record from South West Ireland.

Further work is planned to obtain a high-resolution stable isotope and trace element record over the entire 8.2 ka excursion in order to resolve the structure and magnitude of the event. We also plan to take radiocarbon measurements across the excursion with the possibility of tying the isotopic record to the absolute ages of the tree-ring chronology. Ongoing monitoring of oxygen isotope variation of rainwater and cave drip water in the vicinity of Pippikin Pot is aiding our interpretation of the Holocene oxygen isotope record.

Cave science in Great Britain: past successes and future prospects

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The year 2005 will be a very important one for the BCRA as the organisation seeks to establish a new identity following the establishment of the British Caving Association (BCA) by merger of the NCA with the 'old' BCRA. Inevitably there has been some concern over the support that cave science will receive under the new arrangements and the purpose of this paper is to encourage debate amongst those attending the annual cave science meeting. There is no intention to provide a history of cave science in Britain but rather to remind attendees of some past highlights that show the importance of maintaining a link between 'recreational/sporting' and 'scientific' cavers. To that effect we reproduce below the first three sentences from the Introduction to one of the most seminal works "Caving is the most absolute of sports: It matches the thrill of exploring the unknown and defying physical obstacles with the intellectual challenge to explain how the unfamiliar shapes and beauties of underground scenery have evolved. Most speleologists have started caving for sport, and have found fuller and richer experiences because so many scientific questions called out for an answer" (Cullingford, 1953, p.1).

Cullingford, C.H.D. (editor). 1953. British Caving : An introduction to speleology. Routledge and Keegan Paul, London.

An investigation into the controlling factors of cave formation in the Cnoc Nan Uamh cave system, Inchnadamph, northwest Scotland

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The caves of northwest Scotland are unique in that they occur within a localised outcrop of the Durness Group close to Assynt. Cnoc Nan Uamh is the largest cave in the Traligill Basin and is currently one of a few caves in the world that is used for palaeoclimate research using speleothem laminae. Theories on cave formation have been superseded through time and it is a credit to the continued interest in cave speleogenesis that these theories are constantly modified. These theories have never been applied in any great detail to this cave and with an increasing interest from a palaeoclimate point of view in this cave, it would prove useful to consider formation of the Cnoc Nan Uamh cave.

Nine chambers were studied including 'the grotto' which is the main site of palaeoclimate research. Predominantly the chambers appeared to be strongly controlled by bedding and jointing however there was a localised example of tectonic activity overriding these features. Tectonic activity has largely been demoted to purely localised impacts on cave formation according to Jennings (1971, p.147) but identified as a superior factor in controlling cave development by Sweeting (1972) against that posed by bedding planes. When considering all the chambers together to form an opinion on cave formation on the large scale in this area, the influence of bedding and jointing remained but shallow phreatic flow appeared to become a dominant feature as illustrated by Jennings (1971, p.202). The influence of collapse has greatly altered the cave, which in itself is viewed as an important control on cave formation (Ford, 1989, p.309). Another possibility is the development of Inception Horizons (Gunn, 2004, p.438) which seems like it can be applied to this part of the cave in conjunction with the theories put forward by Jennings (1971, p. 202) on shallow dipping limestones and the role of shallow phreatic flow.

ERRATUM

Apologies from Tony Waltham to members and readers, for an error that crept into the table comparing Encyclopaedias in the book review pages of the last Issue (Vol.31(3), p.141). In fact, the *Speleogenesis* volume by Klimchouk *et al.* sells at just $60 direct from the National Speleological Society (plus $8 shipping, which could be more outside the USA). With the dollar at recent levels with respect to sterling this puts the price at around £40, which makes the book excellent value compared to the two encyclopaedias from mainstream publishers.

BOOK REVIEW

Carbonate Sediments and Rocks. By Colin Braithwaite. Published by Whittles (Caithness) and Orsa (Texas). 2005. ISBN 1-870325-39-7 (0-9710427-5-6 in USA). 184pp. £40.

Everything you ever wanted to know about limestone as a rock (but you never knew quite where to look it up). This is it (or nearly so). Written by a geologist, it covers all aspects of limestone origins. The core chapters on marine carbonate environments and on the diagenesis from sediment to rock are comprehensive and informative, covering just about all the types and distinctive features of carbonates that may somewhere develop karst terrains. Another chapter explains the classifications of carbonates, and defines the various terms that appear in the geological literature. Then there are competent overview chapters on carbonate mineralogy, dolomites (and the processes of dolomitisation) and calcretes. Another on land-based carbonate environments makes the clear distinction between tufas (deposited from normal karst waters) and travertines (deposited from geothermal waters); let us hope that this terminology continues to permeate through the karst literature.

These early chapters combine to offer a very accessible overview of limestone geology. They originated as a section of a mighty project on engineering conditions on carbonate ground, which never reached publication. The remaining third of the book was therefore added by the author to cover all the related aspects (that were planned for elsewhere in the original volume). Sadly, this shows, in that much of it makes a good read, but offers only less exhaustive glimpses that lack the authority of the earlier chapters.

The two chapters on karst are rather minimal, and are noticeably dated. A chapter on conservation is little more than a set of personal notes that just prompt a few thoughts. The chapter on engineering properties tends to list features for the various British limestones, but is a little lost without tighter correlation to the earlier chapters and lacks any worldwide context. The chapter on hydrocarbons includes mention of the Ekofisk oilfield, but without reference to the remarkable compaction of its chalk (which had such expensive consequences to its North Sea platform). A chapter of case histories is subtitled "the hazards of karst"; the examples are taken from the well-documented classics, and they do provide an interesting read for a cave scientist. Sadly though, the Vaiont and St Francis disasters are included, whereas they were due to landslides unrelated to karst; references to both are very out of date. Brief mention of geophysical detection of caves is so dated that it is misleading.

Illustrations include just 36 line drawings distributed through the 141 pages of text. They are mostly clear and concise, but the reader could have wished for more. In addition, 101 colour photographs, all quite small, are bunched into a central colour section of 20 pages. This is due to economies in printing costs, but the colour is welcome, and it does enhance many of the photos of limestone textures and structures. There is a set of very good microphotographs of limestones, but the karst photographs are notably weak and include three that have incorrect locations.

Most readers of *Cave and Karst Science* would not refer to this book for its chapter on karst, as they will have ready access to far more extensive and informative material. However, they will find it a valuable reference for data on limestone geology, where only professional geologists might wish to delve deeper. Shortcomings in the later chapters should not obscure the values of the authoritative earlier chapters. Covering a rather specialist subject, the book is not in the bargain basement, but its price is reasonable by scientific book standards, and no obvious recent competitor comes to mind. It merits a place on the bookshelf.

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January 2006

THESIS ABSTRACT

Cave inception and development in Caledonide metacarbonate rocks

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PhD, 2005
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Abstract

This is the first comprehensive study of cave inception and development in metacarbonate rocks. The main study area is a 40000km^2 region in central Scandinavia that contains over 1000 individual metacarbonate outcrops, and has nearly 1000 recorded karst caves (with passage lengths up to 5.6km). The area, which was repeatedly glaciated in the late Cenozoic, comprises a suite of nappes in the Cambro–Silurian Caledonides, a paleic range of mountains with terranes presently occurring on both sides of the northern Atlantic. Information about the stripe karst and non-stripe karst outcrops and their contained caves was assembled into computer-based databases, enabling relationships between the internal attributes of the caves and their external geological and geomorphological environments to be analysed. A rather consistent pattern emerged. For example, karst hydrological system distances are invariably shorter than 3.5km, and cave passages are positioned randomly in a vertical dimension, whilst commonly remaining within 50m of the overlying surface. This consistency is suggestive that the relevant cave inception, development and removal processes operated at a regional scale, and over long timescales. A consequence of the *epigean* association of caves with the landscape is that cave development can only be understood in the context of the geomorphological evolution of the host region. A review of the latest knowledge of the inception and development of caves in sedimentary limestones concluded that the speleogenesis of the central Scandinavian caves cannot be explained by these ideas. Five new inter-related conceptual models are constructed to explain cave development in metacarbonate rocks in the various Caledonide terranes. These are:

1. The *tectonic inception model* - this shows that it is only open fracture routes, primarily created by the seismic shocks that accompany deglaciation, which can provide the opportunity for dissolution of metalimestone rocks that have negligible primary porosity.
2. The *external model of cave development* - this black-box approach reveals how the formation, development and destruction of the karst caves are related to the evolution of their local landscape. During the Pleistocene, these processes were dominated by the cycle of glaciation, leading to *cyclic speleogenesis*, and the development of ever-longer and deeper systems, where the maximum distance to the surface commonly remains within one-eighth of the extent of change in local relief.
3. The *hydrogeological model* - this demonstrates that the caves developed to their mapped dimensions in timescales compatible with the first two models, within the constraints imposed by the physics and chemistry of calcite dissolution and erosion, primarily in almost pure water. *Relict* caves were predominantly formed in phreatic conditions beneath active deglacial ice-dammed lakes, with asymmetric distributions on east- and west-facing slopes. *Mainly vadose* caves developed during the present interglacial, primarily vadose, conditions, with maximum dimensions determined by catchment area. *Combination* caves developed during both deglacial and interglacial stages. The cross-sections of phreatic passages obey a non-fractal distribution, because they enlarged at maximum rates in similar timescales. Phreatic cave entrances could be enlarged at high altitudes by freeze / thaw processes at the surface of ice-dammed lakes, and at low altitudes by marine activity during isostatic uplift.
4. The *internal static and dynamic model of cave development* - this white-box approach demonstrates that many caves have 'upside-down' morphology, with relict phreatic passages overlying a single, primarily vadose, streamway. Both types of passage are guided along inception surfaces that follow the structural geology and fractures of the carbonate outcrops. Dynamically, the caves developed in a 'Top-Down, Middle-Outwards' (TDMO) sequence that may have extended over several glacial cycles, and passages in the older *multi-cycle* caves were removed downwards and inwards by glacial erosion.
5. The *Caledonide model* - this shows that the same processes (with some refinements) applied to cave development in most of the other (non-central Scandinavian) Caledonide areas. The prime influences on cave dimensions were the thicknesses of the successive northern Atlantic glacial icesheets and the positions of the caves relative to deglacial ice-dammed lakes and to local topography. Other influences included contact metamorphism, proximity to major thrusts, and marine incursions. With knowledge of these influences for each area, mean cave dimensions can be predicted.

The thesis provides the opportunity for the five models to be extended, so that cave development in other glaciated metamorphic and *sedimentary* limestones can be better understood, and to be inverted, so that landscape evolution can be derived from cave data.

RESEARCH FUNDS AND GRANTS

The BCRA Research Fund

The British Cave Research Association has established the BCRA Research Fund to promote research into all aspects of speleology in Britain and abroad. A total of £2000 per year is currently available. The aims of the scheme are primarily:

a) To assist in the purchase of consumable items such as water-tracing dyes, sample holders or chemical reagents without which it would be impossible to carry out or complete a research project;

b) To provide funds for travel in association with fieldwork or to visit laboratories that could provide essential facilities;

c) To provide financial support for the preparation of scientific reports. This could cover, for example, the costs of photographic processing, cartographic materials or computing time;

d) To stimulate new research that the BCRA Research Committee considers could contribute significantly to emerging areas of speleology.

The award scheme will not support the salaries of the research worker(s) or assistants, attendance at conferences in Britain or abroad, nor the purchase of personal caving clothing, equipment or vehicles. The applicant must be the principal investigator, and must be a member of the BCRA in order to qualify. Grants may be made to individuals or groups (including BCRA Special Interest Groups), who need not be employed in universities or research establishments. Information about the Fund and application forms for Research Awards are available from the Research Fund Administrator (address at foot of page or e-mail research-fund@bcra.org.uk).

Ghar Parau Foundation Expedition Awards

An award, or awards, with a minimum of around £1000 available annually, to overseas caving expeditions originating from within the United Kingdom. Grants are normally given to those expeditions with an emphasis on a scientific approach and/or pure exploration in remote or little known areas. Application forms are available from the GPF Secretary, David Judson, Hurst Barn, Castlemorton, Malvern, Worcestershire, WR13 6LS, e-mail: d.judson@bcra.org.uk. Closing dates for applications are: 31August and 31 January.

The E K Tratman Award

An annual award is made for the most stimulating contribution towards speleological literature published within the United Kingdom during the past 12 months. Suggestions are always welcome to members of the GPF Awards Committee, or its Secretary, David Judson (see above for contact details), not later than 31 January each year.

BRITISH CAVE RESEARCH ASSOCIATION PUBLICATIONS

Cave and Karst Science – published three times annually, a scientific journal comprising original research papers, reports, reviews and discussion forum, on all aspects of speleological investigation, geology and geomorphology related to karst and caves, archaeology, biospeleology, exploration and expedition reports.

Editors: Dr D J Lowe, c/o British Geological Survey, Keyworth, Nottingham, NG12 5GG, UK, (e-mail d.lowe@bcra.org.uk) and Professor J Gunn, Limestone Research Group, University of Huddersfield, Queensgate, Huddersfield, HDI 3DH, UK (e-mail j.gunn@bcra.org.uk).

Speleology - published three times annually and replacing BCRA's bulletin *'Caves & Caving'*. A magazine promoting the scientific study of caves, caving technology, and the activity of cave exploration. The magazine also acts as a forum for BCRA's special interest groups and includes book reviews and reports of caving events.

Editor: David Gibson, 12 Well house Drive, Leeds, LS8 4BX, (e-mail: speleology@bcra.org.uk).

Cave Studies Series - occasional series of booklets on various speleological or karst subjects.

No. 1 *Caves and Karst of the Yorkshire Dales*; by Tony Waltham and Martin Davies, 1987. Reprinted 1991.
No. 3 *Caves and Karst of the Peak District*; by Trevor Ford and John Gunn, 1990. Reprinted with corrections 1992.
No. 4 *An Introduction to Cave Photography*; by Sheena Stoddard, 1994.
No. 5 *An Introduction to British Limestone Karst Environments*; edited by John Gunn, 1994.
No. 7 *Caves and Karst of the Brecon Beacons National Park*; by Mike Simms, 1998.
No. 8 *Walks around the Caves and Karst of the Mendip Hills*; by Andy Farrant, 1999.
No. 9 *Sediments in Caves*; by Trevor Ford, 2001
No. 10 *Dictionary of Karst and Caves*; by D J Lowe and A C Waltham, 2002.
No. 11 *Cave Surveying*; by A J Day, 2002.
No.12 *Underground Britain-Legal + Insurance Issues*; (2nd extended/revised edition) by David Judson, 2005.
No 13 *Exploring the Limestone Landscapes of Upper Wharfedale*; by Phillip Murphy, 2003.
No.14 *Swildon's Two and Three*; by Dave Irwin, 2004.
No.15 *Exploring the Limestone Landscapes of The Three Peaks + Malham*; by P Murphy, 2005.

Numbers 2 and 6 are out of print, but have been updated by numbers 11 and 10 respectively.

Speleohistory Series – an occasional series.

No.1 The Ease Gill System – Forty Years of Exploration; by Jim Eyre, 1989.

BCRA SPECIAL INTEREST GROUPS

Special Interest Groups are organised groups within the BCRA that issue their own publications and hold symposia, field meetings, etc.

Cave Radio and Electronics Group promotes the theoretical and practical study of cave radio and the uses of electronics in cave-related projects. The Group publishes a quarterly technical journal (c.32pp A4) and organises twice-yearly field meetings. Occasional publications include the Bibliography of Underground Communications (2nd edition, 36pp A4).

Explosives Users' Group provides information to cavers using explosives for cave exploration and rescue, and liaises with relevant authorities. The Group produces a regular newsletter and organizes field meetings. Occasional publications include a Bibliography and Guide to Regulations, etc.

Hydrology Group organizes meetings around the country for the demonstration and discussion of water-tracing techniques, and organizes programmes of tracer insertion, sampling, monitoring and so on. The Group publishes an occasional newsletter.

Speleohistory Group publishes an occasional newsletter on matters related to historical records of caves; documentary, photographic, biographical and so on.

Cave Surveying Group is a forum for discussion of matters relating to cave surveying, including methods of data recording, data processing, survey standards, instruments, archiving policy, etc. The Group publishes a quarterly newsletter, Compass Points (c.16pp A4), and organizes seminars and field meetings.

Copies of BCRA Publications are obtainable from: Ernie Shield, Publication Sales, Village Farm, Great Thirkleby, Thirsk, North Yorkshire, Y07 2AT, UK.

BCRA Research Fund application forms and information about BCRA Special Interest Groups can be obtained from the BCRA Honorary Secretary: John Wilcock, 22 Kingsley Close, Stafford, ST17 9BT, UK.