Cave and Karst Science

The Transactions of the British Cave Research Association

BCRA

Volume 30 Number 1 2003

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:

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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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Shorter contributions, normally 500-3,000 words, on aspects of karst/speleological science, as listed above, or more descriptive material, such as caving expedition reports and technical articles. Manuscripts will be reviewed by the Editorial Board unless the subject matter is outside their fields of expertise, in which case assessment by an appropriate expert will be sought.

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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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Speleological expeditions have a moral obligation to produce reports (contractual in the case of recipients of awards from the Ghar Parau Foundation). These should be concise and cover the results of the expedition as soon as possible after the return from overseas, so that later expeditions are informed for their planning. Personal anecdotes should be kept to a minimum, but useful advice such as location of food supplies, medical services, etc, may be included. preferably as a series of appendices.

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Cave and Karst Science

TRANSACTIONS OF THE BRITISH CAVE RESEARCH ASSOCIATION

Volume 30, Number 1

Contents

Papers

Reports

Cover photo:

Spectacular drowned tower karst off the Andaman Coast of southern Thailand has been used as a backdrop for many movie productions. The scenery shown here, for example, appeared in '*The Man with the Golden Gun*'. These islands have evolved partly as a result of rising and falling sea levels, a topic that is the subject of a Report by Dean Smart in this Issue.

Photographed by Dean Smart.

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ISSN 1356-191X

Printed by The Sherwood Press, Nottingham, UK: November 2003
DTP by Rebecca Talbot

EDITORIAL

John Gunn and David Lowe

Although this is the first issue of the 2003 Volume of *Cave and Karst Science* we send all our readers Christmas greetings and best wishes for 2004! This seems more positive than apologising for the fact that, yet again, and despite our best efforts, this issue has been unavoidably delayed. However, we are pleased to record that Volume 30 Number 2 is close to completion and should be with you very early in 2004. Graham Proudlove has kindly agreed to provide a Guest Editorial for that Issue, which includes the first major review of the subterranean aquatic Crustacea recorded in Britain and Ireland.

Delayed though it may be, we hope that readers will enjoy the present issue and as editors we are particularly pleased at the mix of papers presented. Authors from inside and outside of academia cover various aspects of earth science, hydrology and palaeontology, based on studies in Britain, Iran, the Mariana Islands, Thailand and Turkey.

As the 'merger' between BCRA and NCA to form a new BCA gradually proceeds, some thought is being given to the future of this publication, particularly in terms of content and format. As editors we feel that the present balance between the different aspects of cave and karst science, and the geographical coverage (as reflected in this issue) are about right. We also feel that by publishing both fully refereed scientific papers and less rigorously reviewed reports we provide an opportunity for both 'academic' and 'amateur' cave scientists to contribute. As always we welcome the views of readers as to how well we succeed and whether there are areas that we are not covering and where we should be actively soliciting papers. The format of the journal has also become well established and seems to work well, although again we would be interested to hear if there are any contrary views or suggestions for improvement.

More controversially, there have been some suggestions that the journal should not appear as hard copy but as electronic copy on the World Wide Web. A good example of an electronic journal, which readers with an interest in the earth sciences might wish to examine, is *'Speleogenesis'*, published jointly by the Commission on Karst Hydrogeology and Speleogenesis of the International Speleological Union and the IGU Karst Commission (http://www.speleogenesis.info/). The journal publishes a mixture of 'new' articles, which are refereed by members of the Editorial Board, and 'old' articles, which have already been published elsewhere but are deemed to be worthy of drawing to the attention of a larger audience. The latter include a number of papers that were previously published in *Cave and Karst Science*. As the name implies, *'Speleogenesis'* focuses on the origin and development of dissolutional caves, although this is placed within the somewhat broader context of the hydrogeological setting and the evolution of karst. Although the web site is very professional, and provides useful information in addition to that in the actual papers, personally, both of us prefer our journals as hard copy. However, we recognise that the Web may provide opportunities for swifter, cheaper, publication, and these benefits are certainly worth pursuing.

CAVE AND KARST SCIENCE Vol. 30 No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Towards defining a scallop dominant discharge for vadose conduits: some preliminary results

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Abstract: a well-established inverse relationship exists between mean scallop length and flow velocity for a given population of scallops. Previous authors have suggested that one or more 'scallop dominant discharges' can be identified at which erosion by dissolution proceeds at the greatest rate, since scallop populations usually indicate a single flow velocity whereas discharge and velocity are unsteady through time. For vadose conduits, a scallop dominant discharge is difficult to define because of the unconstrained cross-section; this causes problems in determining the discharge at which scallops are formed, although recent developments in instrumentation allow greater flexibility in monitoring flows on a continuous basis. Here the relationships between monitored flow velocity and depth are compared with the scallop velocity for an active vadose streamway in Poulnagollum, Co. Clare, Ireland. From these initial results, a complex relationship is seen to exist between the velocity and depth of flow as discharge changes. Thresholds occur over discrete depth ranges where there is little or no change in velocity; these are observed during both rising and falling stage. It is suggested that these thresholds may be related to changes in hydraulic radius, and hence flow resistance at different depths of flow. The scallop-derived velocity is related to the recorded flow data, with reference to the various controls on erosion, most notably the degree to which the flow is undersaturated with $CaCO_3$, and ongoing research is outlined.

(Received 26th August 2003; Accepted 16th October 2003)

INTRODUCTION

Scallops are commonly-occurring dissolutional features that wholly or partly cover the boundaries of many active and fossil conduits. These concave forms are longitudinally asymmetrical, indicating the direction of the flow that formed them, and they vary in length from a few millimetres to several metres, depending upon the passage dimensions and prevailing flow conditions. A well-established inverse relationship exists between the Sauter mean scallop length and flow velocity, established by Curl (1966) using dimensional analysis and later confirmed experimentally using plaster blocks (Goodchild and Ford, 1971; Blumbeg and Curl, 1974). This relationship has been widely applied in determining flow conditions for active and fossil cave conduits.

The fact that scallop populations usually have a log-normal, unimodal distribution, while discharge and velocity are unsteady through time, has led to the suggestion that a 'scallop dominant discharge' can be defined at which dissolutional erosion occurs at the greatest rate and is therefore most effective in passage formation (Smart and Brown, 1981; Lauritzen, 1982 and 1983; Ford and Williams, 1989). The concept of an 'effective', or 'dominant', discharge was first proposed by Wolman and Miller (1960) for (surface) alluvial channels in humid temperate regions where the frequency, or return period, of a given flow is considered together with its magnitude in determining geomorphological effectiveness in terms of bedload transport rates. Whereas larger, rarer flows may individually carry out a considerable amount of geomorphological work, the infrequent occurrence of such events in comparison to the cumulative effect of lower magnitude, higher frequency flows may be more effective in determining channel capacity and form over a given period of time. This builds on the work of Leopold and Wolman (1957), who correlated meander wavelength with bankfull width and bankfull discharge, demonstrating the morphological significance of frequently occurring flows.

A rather different set of controls applies to cave conduits since, in the absence of a significant clastic load, erosion is almost entirely due to dissolution. The relationship between passage width and meander wavelength was examined by Smart and Brown (1981) for vadose streamways in Ireland and New Zealand using passage width as a surrogate for bankfull discharge (which cannot be defined for a vadose canyon). The relationship was found to be anomalous compared with the widely observed inverse relationship that exists for alluvial channels, where adjustment of channel form is accomplished through the erosion and deposition of sediment. The authors highlight the importance of local base levels in determining relative rates of vertical and horizontal incision, together with variations in the aggressiveness of the streamflow – erosion is effectively zero when the water is saturated, regardless of discharge.

Dominant discharges have been determined from measurements made by divers in phreatic conduits in marble stripe caves in Norway (Lauritzen, 1982, 1989 and 1995). Discharges calculated from scallop-derived velocities and measurements of conduit cross-section were analysed in conjunction with flow data from a gauging station. The scallop-derived discharges were found to represent flood flows that occur between 2 and 15% of the time, a flow with a magnitude three times that of the mean annual flood. By integrating flow and chemical measurements, it was estimated that half of all the chemical work occurs for only 10% of time, during the highest discharges (Lauritzen, 1989). There are a number of possible applications of the concept that include developments in understanding of erosion controls and rates of erosion and analysis of palaeoflows. Lauritzen (1989) suggested that a direct linear relationship exists between scallop dominant discharge and the corresponding drainage area – similar to the relationship observed for surface catchments – which may have potential as a powerful tool for deducing the area of palaeo-watersheds from relict caves. It is also possible to derive hydraulic parameters such as boundary friction, stream power and boundary shear stress from scallop and sediment properties (Gale, 1984).

Determining the dominant discharge for a vadose streamway has always been problematical – although not impossible – because both the cross-sectional area and velocity vary with changing discharge,

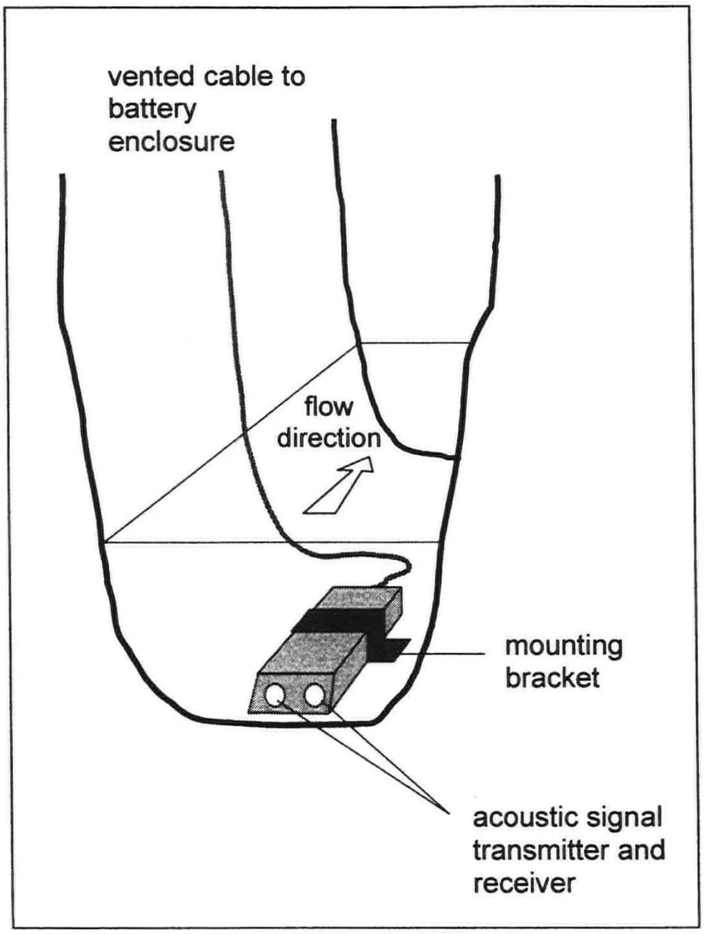

Figure 1. *The Starflow ultrasonic Doppler instrument. The dimensions of the Starflow are 290mm (L) x 70mm (W) x 25mm (H).*

making it difficult to calculate a meaningful discharge(s) from scallop measurements. However, recent developments in flow monitoring mean that it is now possible to monitor depth and velocity in vadose conduits without the need to install gauging structures. The research described here is at an initial stage, the main aims being to attempt to identify the flow conditions under which scalloping develops at the greatest rate and to determine whether a scallop dominant discharge is a valid concept in this instance.

METHODS

Flow monitoring

A Unidata 'Starflow' ultrasonic Doppler instrument (Fig.1) was selected for installation in a section of vadose canyon streamway in Upper Poulnagollum, Slieve Elva, Co. Clare, between Poll Binn Pot and Main Entrance (Fig.2). In selecting this site, several factors were taken into consideration. A straight section of actively incising streamway with a regular cross-section was chosen, care being taken to ensure that there were no obvious structural or hydraulic controls that might, for instance, cause flow to back up during high flows. Further criteria included the presence of well-developed scalloping and an absence of clastic load. The site also had to be suitable for the installation of the Starflow, which is connected via a 15m vented cable to an enclosure containing a 12V battery and computer interface, which had to be placed away from any risk of inundation. Flow monitoring is an ongoing process, with data downloaded from the Starflow at regular intervals.

The Starflow is able to record flow data for a period of up to three months, and was set to scan at a rate of once a minute, logging time-averaged data every ten minutes. Depth is measured by means of a hydrostatic pressure sensor that is vented via a cable to the atmosphere, while velocity is determined by means of an incoherent, or continuous, ultrasonic Doppler. During a scan, a continuous ultrasonic signal is transmitted at a fixed frequency in an upstream direction. The centreline of the beam is aligned at an angle of 30° from horizontal and the beam has a width, or spread, of 10°. The transmitted signal is reflected by particles and air bubbles carried in the flow, and the frequency of the signal is changed as a result of the Doppler shift. A measuring circuit detects changes in frequency in the reflected signal arriving at a receiver and a processing system accumulates and analyses frequency changes to calculate a representative Doppler shift from the range received. Since the Starflow is an incoherent Doppler, velocity is depth-integrated and it is not possible to obtain a velocity profile; instruments with this capability are not widely available. The operating ranges of the instrument are, for depth, 0m to 2.0m at a 2mm resolution; and velocity, 21mm s^{-1} to 4500mm s^{-1}, at a resolution of 1mm s^{-1}, with an accuracy of 2% of the measured velocity.

Determining scallop velocity

Scallops form at a stable scallop Reynolds number (Re*) of ~2200 (Curl, 1974; Blumbeg and Curl, 1974), where Re* is related to the mean boundary shear velocity $\bar{u}*$ (obtained by dividing boundary shear stress by fluid density), Sauter mean scallop wavelength ($\bar{\lambda}$) fluid density (ρ_f) and fluid dynamic viscosity (μ) by:

1. $$\mathrm{Re}* = \frac{\bar{u}*\bar{\lambda}\rho_f}{\mu}$$

Where:
D = passage width
$\bar{u}*$ = friction velocity
$\bar{\lambda}$ = Sauter mean scallop length
B_L = Prandtl's bed roughness constant

A theory of scallop formation was proposed by Curl (1966). The layer of fluid next to the boundary is slow moving and saturated with respect to $CaCO_3$. However, dissolution can start to occur once a critical scallop Reynolds number has been reached – assuming the bulk fluid is not saturated. At this scallop Reynolds number, flow separation starts to occur at the site of small surface irregularities, forming a jet of fluid that undergoes a transition to turbulent flow, becoming unstable after a certain distance, at which point reattachment occurs. This allows aggressive bulk fluid to reach the boundary at the point of reattachment, where erosion proceeds at the greatest rate. The frequency of detachment increases with increasing velocity, thus reducing the erosion length available to each individual scallop. The characteristic scaling of scallops is a hydrodynamic mechanism, and fluid dynamic equations can be used to describe their formation and the flow conditions under which they were formed (Curl, 1966 and 1974).

Velocity conditions near the wall depend on the conduit size for a given mean conduit velocity and may be described by a 'law of the wall' type turbulent analysis, where the velocity distribution within the boundary layer (the thickness of flow affected by boundary drag) is assumed to be semi-logarithmic. Curl (1974) used a modified version of Prandtl's universal velocity distribution law, from which the mean velocity ($\bar{u}$) in a channel may be computed by substituting values in the equation for parallel-walled conduits:

2. $$\bar{u} = \bar{u}*\left[2.5(\ln\frac{D}{2\bar{\lambda}} - 1) + B_L\right]$$

Scallop dimensions were measured on both sides of the passage below the depth of the maximum flow and within an area on each wall that was delimited upstream from, and within range of the beam of the Starflow. This was determined by calculating the distance from the Starflow at which the beam, inclined at an angle of 30° from horizontal, would intersect the water surface at different depths of flow; this varies from 0.12m for a depth of 0.1m to 0.84m for a depth of 0.7m. Although the Starflow was installed during low flow conditions (approximately 0.1m depth), the water depth was greater

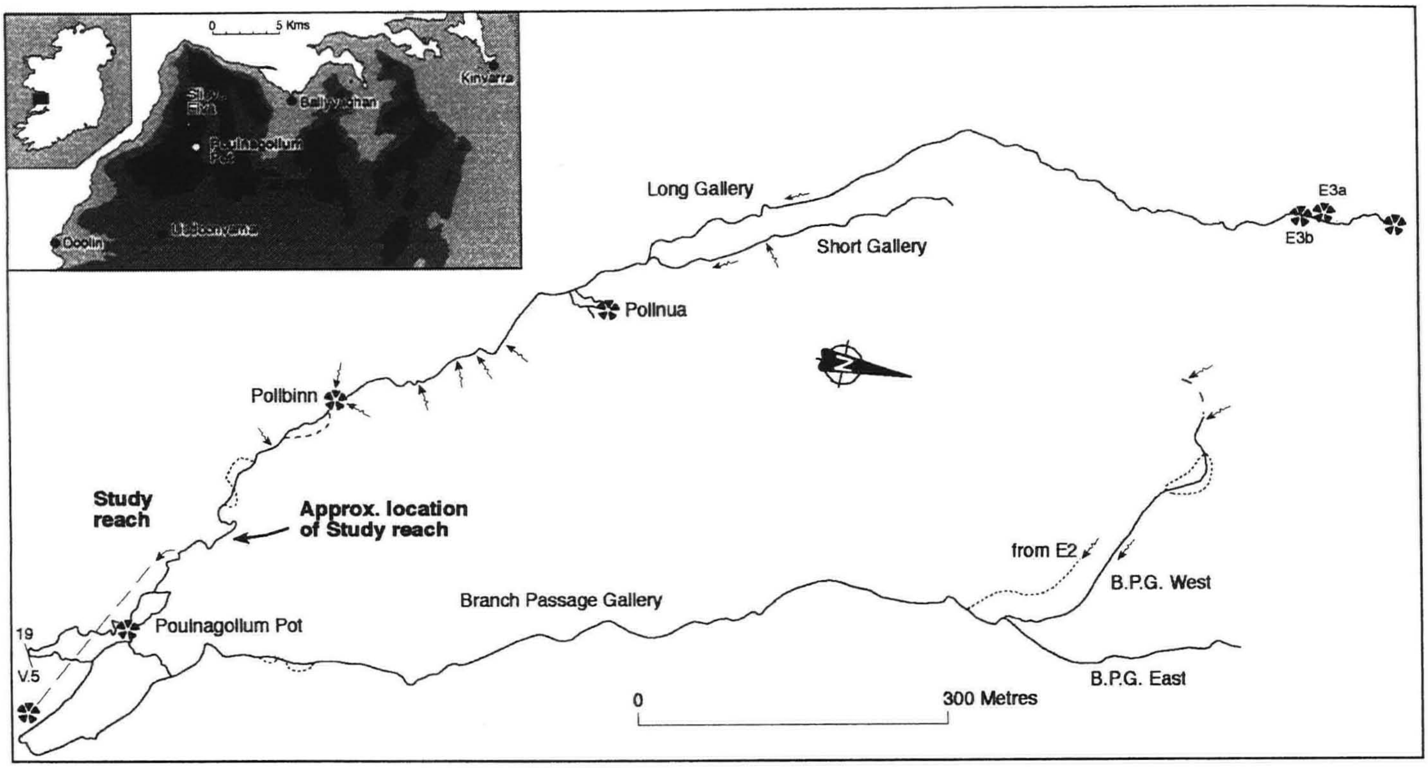

Figure 2. Map and survey showing the location of the study site.

than 0.2m on subsequent visits so it was not possible to measure the dimensions of scallops below this level. The width of the passage was measured at vertical intervals of 0.1m for a cross-section located at the Starflow and within the range of flow depths recorded. From these measurements an average width of 0.67m was calculated. These width measurements also enabled an estimate to be made of the discharge from recorded velocity and depth data, although width measurements alone do not provide sufficient information about the cross-section to calculate accurate discharge values. For this reason depth hydrographs are referred to in the following section. The friction velocity was calculated from Equation 1 using a value of $0.013 cm^2\ s^{-1}$ for μ/ρ_f (kinematic viscosity), which assumes a temperature of 10°C (the range of water temperatures recorded by the Starflow was between 5.8°C and 12.1°C). The mean conduit velocity was then calculated using a value of 9.4 for B_L (Blumberg and Curl, 1974).

RESULTS AND DISCUSSION

Flow data have been analysed for the eight-month period between September 2002 and May 2003. The peak flow recorded, approximately $1.2m^3\ s^{-1}$, occurred on 5/12/02, with a maximum depth of 0.76m and a maximum velocity of $2.27m\ s^{-1}$. The minimum depth was 0.02m, but it is not possible to define a minimum discharge because a value of zero is recorded when the mean velocity falls below $0.022mm\ s^{-1}$; problems also arise when the flow depth falls below the level of the acoustic transmitter and receiver. It should be noted that measurement of velocity at very low flows is difficult using most instruments. Fig.3 shows depth and velocity hydrographs for a one-month period from 18/12/02 to 18/1/02, selected as being representative of the high and low flow conditions recorded over the monitoring period. Each of the data points on

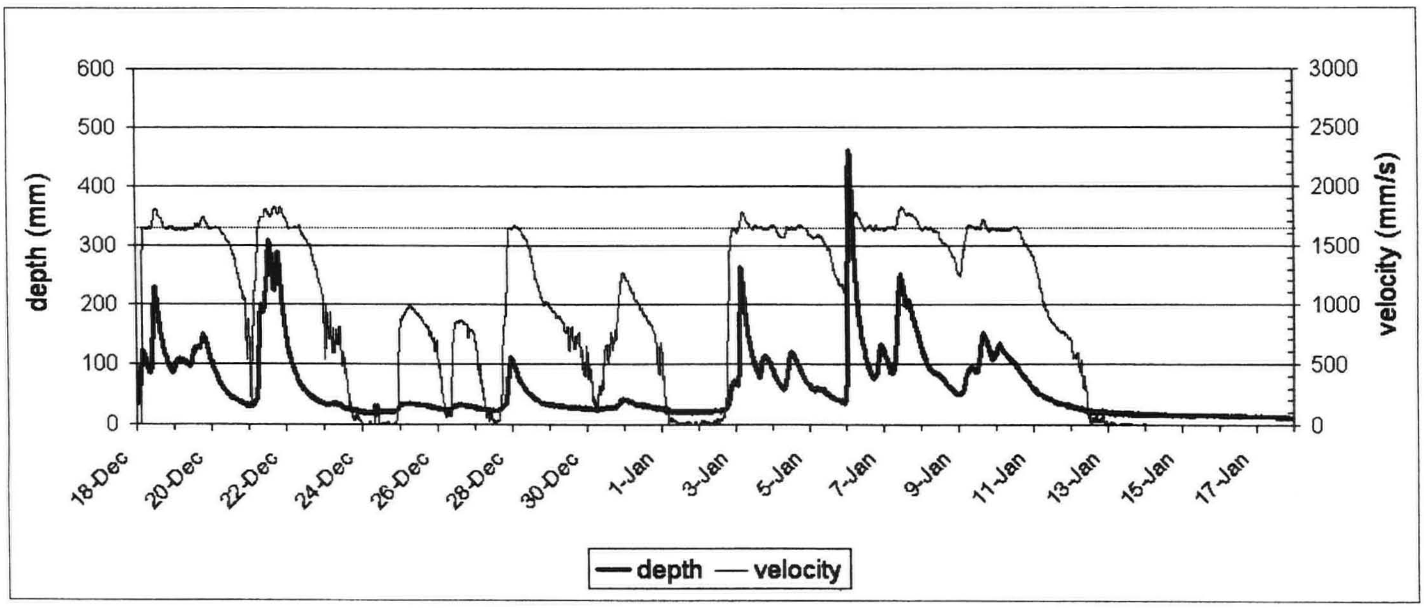

Figure 3. Depth and velocity hydrographs for 18/12/02 to 18/01/02. The dotted line indicates the threshold velocity of approximately $1.65m\ s^{-1}$.

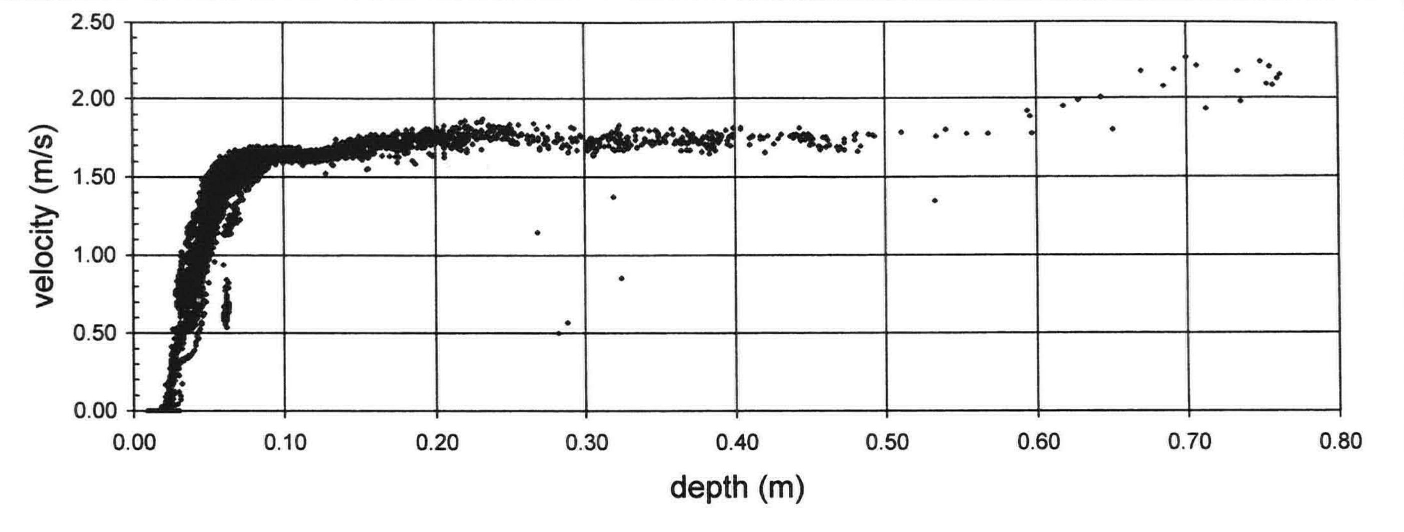

Figure 4. *Graph of velocity against depth. The data include the rising and falling limbs of selected storm hydrographs.*

these graphs is the logged 10-minute time-averaged value derived from scans made at one-minute intervals. Although rainfall data are not currently available, the depth hydrographs indicate a catchment with a rapid or 'flashy' response to rainfall. This might be expected, as the catchment is small (less than 1km^2), underlain by shale, and covered by blanket peat in which drainage channels have been cut for forestry.

An interesting characteristic of the velocity hydrographs is a 'plateau' that occurs once the velocity reaches a value of approximately 1.65m s^{-1}; this is in contrast to the rapid increase in velocity with discharge (depth) up to this point. This 'plateau' is indicated by a dashed line in Fig.3, and the effect can be seen for several of the events shown, where it appears that some threshold must be reached for further increases in velocity to occur. The multi-peaked events from 03/01/03 to 13/01/03 are a good example, with the velocity exceeding the velocity threshold where there are corresponding peaks in the depth hydrograph, although the duration of these peaks is relatively short. On the rising limb of hydrographs, the velocity appears to stop increasing once the depth exceeds a value of approximately 0.1m, and further increases in velocity do not occur until the depth is greater than about 0.15m. Examination of all hydrographs for the monitoring period indicates that further thresholds exist, although these are not so distinct and are difficult to define because of the relatively short duration of the higher velocities. The relationship between depth and velocity can be seen more clearly in Fig.4, which shows the relationship between depth and velocity for a number of selected events. The data include values from the rising and falling limbs of each hydrograph, and over 6000 data points are shown (many are superimposed on the graph). Here the thresholds in velocity can be seen, despite the fact that there is a hysteresis effect (this can be seen from Fig.3) where, for the same depth, the velocity is greater on the falling limb than it is on the rising limb of the hydrograph. In Fig.4 the isolated points at the upper end of the distribution correspond to the event of 5/12/02, for which an anticlockwise hysteresis occurred, with lower velocities on the rising limb. It is suggested that the velocity thresholds observed may be due to variations in total boundary resistance. Although the roughness of the actual boundary itself, defined by B_L in equation 2, can be assumed to be constant, the amount of contact between the body of flowing water and the boundary – the hydraulic radius – changes with discharge. The hydraulic radius is calculated by dividing the cross-sectional area of the flow by the wetted perimeter (the combined length of bed and banks in contact with the flow), with higher values indicating a greater hydraulic efficiency. If the depth of the stream shown in Fig.1 were doubled, the cross-sectional area would more than double. However, the wetted perimeter would increase by a lesser amount, leading to an increased hydraulic radius. The enhanced hydraulic efficiency reduces the proportion of total energy expended in overcoming boundary friction, enabling water to be transmitted more rapidly. At very low flows, flow resistance would be considerable, and it is suggested that for flows of less than 0.1m depth any variation in depth results in a substantial change in flow resistance, accounting for the large variations in velocity with depth in this range. For flow depths greater than 0.15m, it is likely that changes in flow resistance for a given change in discharge are not so great, although variations in channel width with depth would affect the cross-section and may affect the different rate at which velocity changes with depth seen in Fig.4. From the approximate cross-section shown in Fig.5 it can be seen that as flow depth increases from 0.2m to 0.5m there is a gradual widening of the passage, with an increase of more than 50% between 0.6m and 0.8m. Flows with depths in this range have only been recorded for one event so far: that of 05/12/02. At these flows, a given increase in depth represents a proportionately greater increase in discharge than at lower flows. Additionally, the cross-sectional area increases at a much greater rate than the wetted perimeter, increasing the hydraulic radius, and may account for the high velocities recorded for this event.

A scallop mean velocity of 1.43m s^{-1} was derived for the conduit and was compared with the distribution of recorded velocities. Each ten-minute velocity record was assigned to a class with a range of 0.09m s^{-1}. The 1.40 to 1.49m s^{-1} class had a relative frequency of 2.1%, and flows within this range were equalled or exceeded 10.0% of the time. The most frequently recorded flow velocities were those less than 0.1m s^{-1}, which were logged for 43.6% of the time. All other classes had a frequency of less than 4.0%, with the exception of flows with velocities between 1.60 and 1.69m s^{-1}, for which the frequency was 11.6%; the main velocity threshold observed (between depths of 0.1m and 0.15m) falls into this class. In comparing the scallop mean velocity with recorded flows several factors must be taken into consideration, such as the possibility that the scallops were formed under hydraulic and hydrological conditions that are outside the range of the recorded data. A critical control on rates of dissolution is the concentration of $CaCO_3$ and the way in which this varies with discharge. Limited water chemistry data exist for upper Poulnagollum although the $CaCO_3$ content of flow from individual swallets in the Poulnagollum system has been observed to decrease with increasing discharge (Ingle Smith *et al.*, 1969). The same authors found that in the central part of this system, the $CaCO_3$ content of streams with no direct surface feeders can be much higher, with aggressive flows only occurring during flood conditions. The mixing of water from these two different types of source leads to a complex pattern of variation between sites. In order to examine variations in $CaCO_3$ at the research site, a conductivity logger has recently been installed to provide data at ten-minute intervals, and will allow the range of scallop forming conditions to

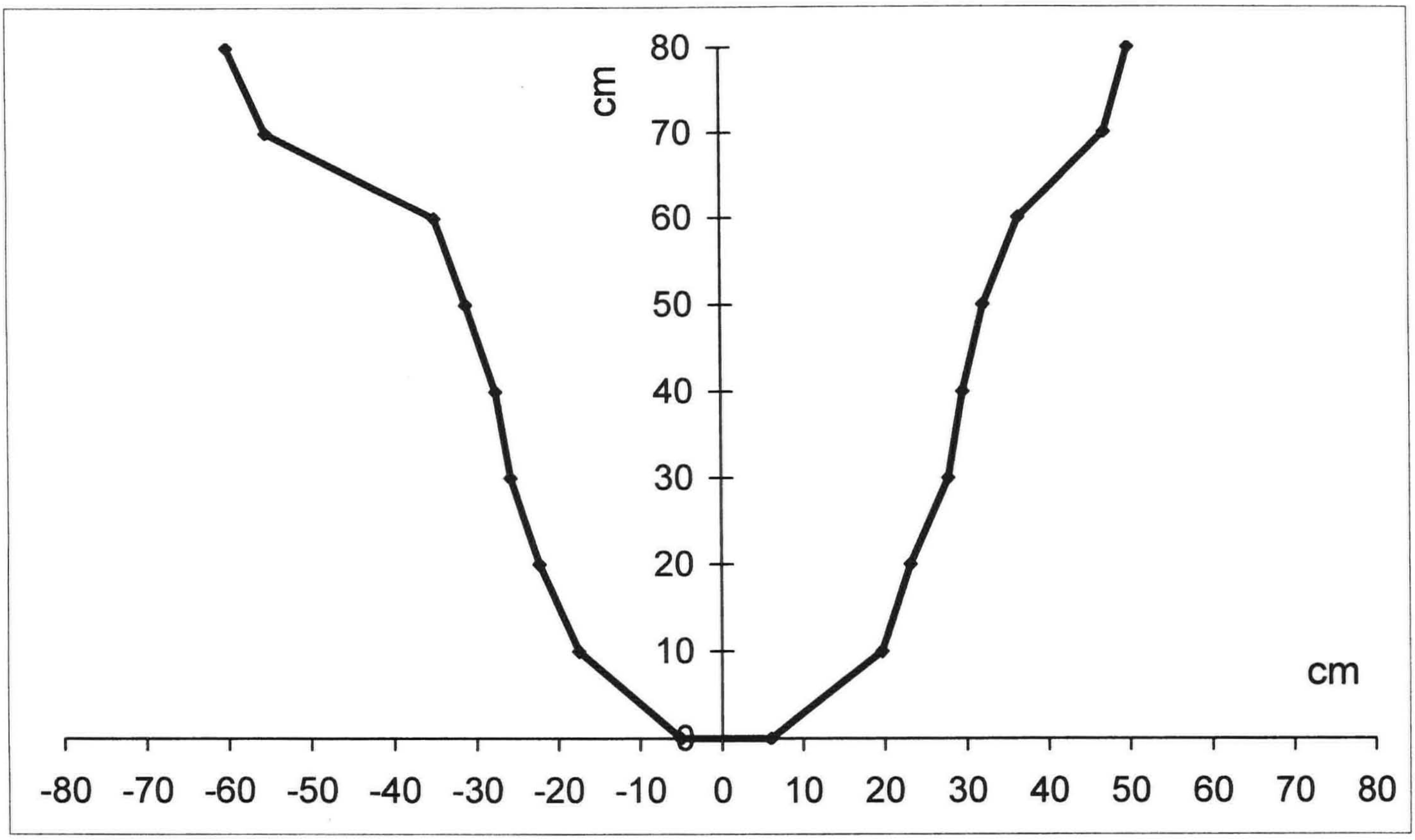

Figure 5. *Approximate cross-section derived from measurements of passage width made at the installation site.*

be refined. Further measurements of scallop dimensions are necessary as the original sample size was relatively small (100 scallops), although care will be taken to ensure that individual scallops are not included twice in the sample. At the same time, the size distribution corresponding to different depths of flow should be examined as it is not known if the scallops found at a given height are fossil features or are formed by contemporary high flows. Visual observations made at the site suggest that the scallop size does not change with increasing height above the floor of the passage, but this has not been confirmed quantitatively.

CONCLUSIONS

Determining a dominant discharge for vadose conduits is problematical because of the unconstrained cross-section, which makes it difficult to identify the depth(s) of flow at which the scallop dominant velocity occurs. The preliminary results of the monitoring programme described here indicate that the relationship between velocity and depth with changing discharge is complex, and appears to be controlled by the shape of the passage cross-section and the resultant variations in flow resistance with depth. The depth-velocity relationship is characterised by thresholds, with a similar velocity occurring over a range of depths. This could mean that it is not possible to define a single dominant discharge on the basis of scallop measurements, rather a range of discharges. The calculated scallop velocity was found to lie towards the upper end of the flows monitored, being equalled or exceeded approximately 10% of the time and corresponding to a flow depth of less than 0.1m. Although this flow velocity was exceeded by many of the hydrograph peaks recorded, its value was less than that of the main threshold observed in the depth-velocity relationship. However, it is not really possible to determine the flow conditions leading to the development of scalloping in the absence of $CaCO_3$ concentration data. A conductivity logger has recently been installed at the site to address this.

ACKNOWLEDGEMENTS

The help provided by Adrian Thomas and Tom Twomey with installing the loggers and accessing the cave is very much appreciated.

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CAVE AND KARST SCIENCE Vol. 30, No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

The Çamlik Caves System (Konya, Turkey)

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Abstract: Çamlık Caves are an active and semi active cave system located in the Western Taurus in Çamlık town, south of Beyşehir Lake. The caves were enlarged when a palaeo river (a palaeo tributary of the Manavgat River), flowing on a Pliocene denudational surface in Late Pliocene to Early Pleistocene times met a northwest-southeast fault line, incised deeply and integrated into a pre-existing underground water system. During the Early Pleistocene the system, which now comprises Körükini and Suluin caves, was a single unit. A collapse doline formed where the Pliocene palaeo valley intersected an existing conduit, resulting in ceiling collapse, and the original cave was divided into the two parts. The system was unable to extend deeper because the limestones were not very thick and the base comprised insoluble formations. Because a sinking river developed the cave system it displays single conduit cave features and a sinuous passage. The cross-section of the cave is of elliptical, canyon type, reflecting the guidance of the fault line and the effects of hydraulic control.

Keywords: Cave, karst, Çamlık, Turkey

(Received 21st July 2003; Accepted 10th September 2003)

INTRODUCTION

Karstic areas developed on carbonate and sulphate rocks make up about one third of the total area of Turkey. A large part of these areas lies in the Taurus Karst Belt, which stretches 200km from north to south. The Taurus Karst Belt is highly faulted, folded and fractured, as a result of Alpine folding movements. There has been extensive underground and surface karst development within this belt, and considerable cave dimensions and depths are reached,. However, due to the fact that the lithostratigraphical features of the structural units and the effects of tectonic movements were not the same everywhere, the degree and intensity of karstification varies within the belt. There are relict semi-active, semi lateral/vertical and swallowhole or rising caves in the slopes of macro-karstic features such as poljes, uvalas and intermontain basins that have been developing since the start of the palaeokarstic period (post-Mid Miocene). There are active and semi-active caves, such as Çamlık Caves, close to base level, which exhibit both spring and swallowhole characteristics (Nazik and Törk, 1998).

The cave system comprising the Körükini and Suluin caves is generally referred to as the Çamlık Caves. The system is located in Çamlık town, in Konya, in the northern part of Western Taurus Mountains, facing towards the Central Anatolia region (Fig.1).

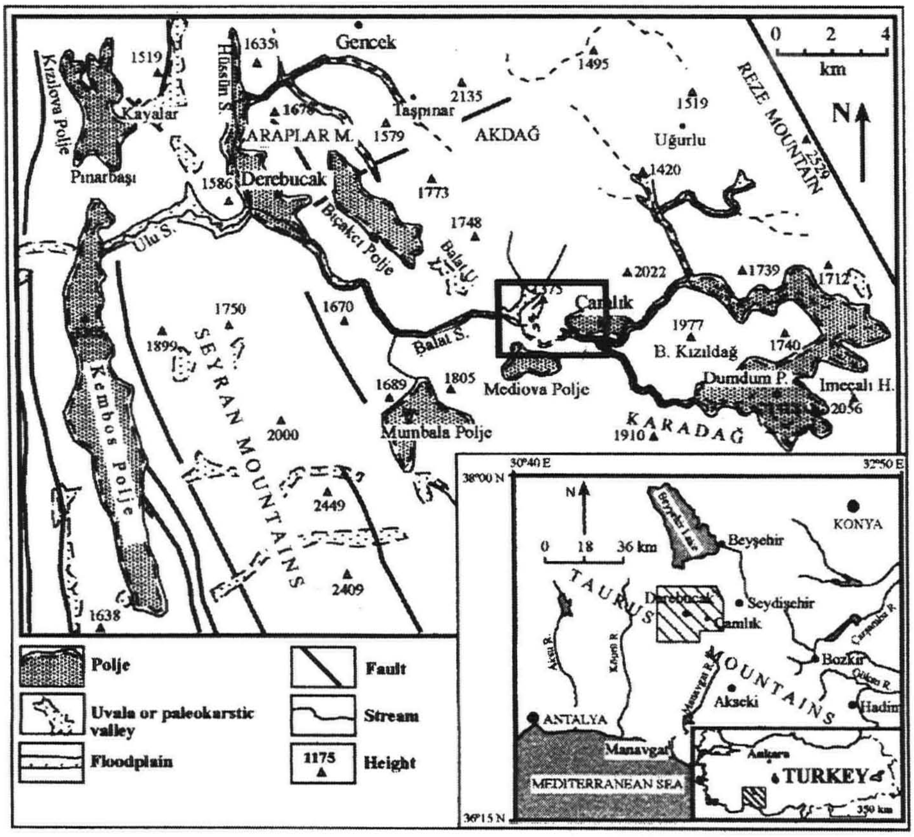

Figure 1. Location map of the study area.

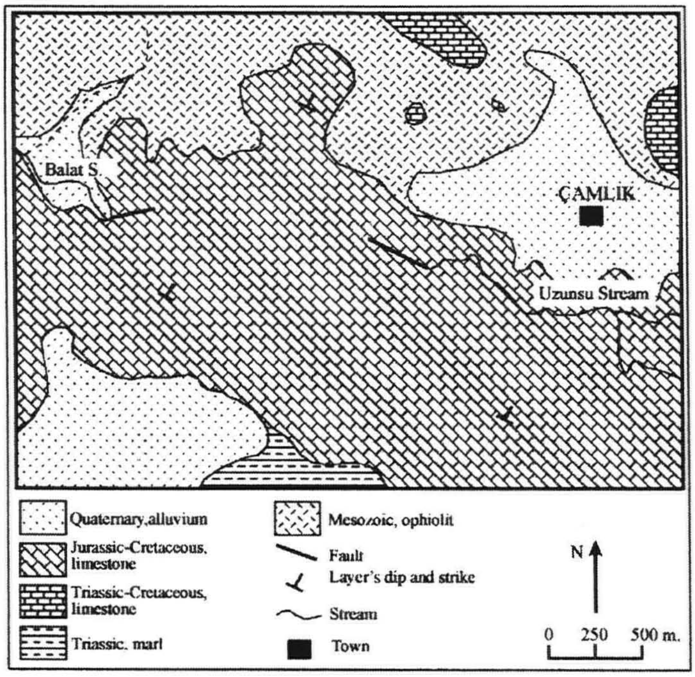

Figure 2.. Geological map of the study area.

Körükini Cave is 750m and Suluin Cave 2.5km away from Çamlık town. Körükini and Suluin, which are now separate caves or underground water galleries with separate entrances and exits were originally a single gallery that was divided into two parts due to the effects of a subsequent collapse in its middle section. The Uzunsu Stream, which is directly connected to Çamlık Caves and connects various basins in the region (Çamlık, Mumbala, Mediova poljes, Balat uvala and Kembos Polje) goes underground at Körükini Cave and resurfaces at Suluin Cave. This stream stays on the surface for 18km before entering a swallowhole near Kembos Polje (Fig.1), a feature characteristic of the region. It then flows underground again before draining into the Manavgat River and then into the Mediterranean. Çamlık Caves have been investigated previously (Aygen, 1970; Nazik, 1992; Nazik et al., 1993). The current study is concerned with the investigation of the Çamlık Caves and the formation and development, karst morphology, hydrology and speleology of the Uzunsu Collapse Doline, which separates the component caves.

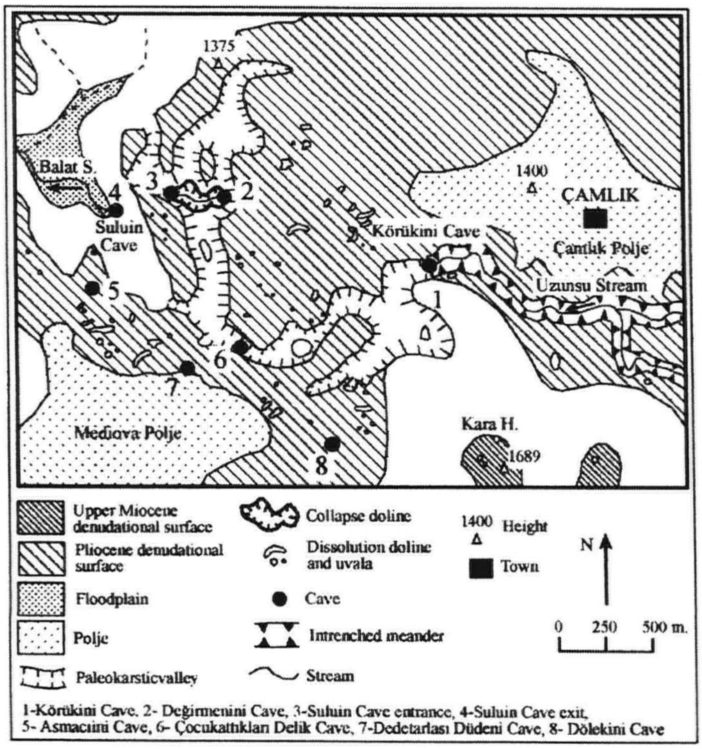

Figure 3. Geomorphological map of the Çamlık Caves area.

GEOLOGY

The Western Taurus Mountains, where the study site is located, comprise autochthonous and allochthonous structural units that reflect different environmental conditions with regard to their stratigraphical, lithological, structural and metamorphic features (Özgül, 1976, 1983; Monod, 1977, 1979; Akay, 1981). The area surrounding Çamlık Caves is very complex geologically, reflecting the structural juxtaposition and superimposition of different rock successions.

Çamlık Caves are surrounded by Mesozoic ophiolites, Triassic marls, Triassic-Cretaceous limestones, Jurassic-Cretaceous limestones and Quaternary alluviums (Fig.2). All these formations except the alluviums are incorporated within the Beyşehir-Hoyran Nappes, which reached their present positions due to movements during the Eocene (Monod, 1977; Akay, 1981; Nazik, 1992; Nazik *et al.*, 1993). Çamlık Caves developed within the Jurassic-Cretaceous limestones, which exhibit extensive fractures capable of guiding karstification. At the base of the Jurassic-Cretaceous limestones are impermeable marls of Triassic age (Fig.2).

Neotectonic movements affected the region from late Tortonian times, causing block faulting (Koçyiğit, 1984; Nazik *et al.*, 1993; Atalay, 2001; Erol, 2001). These neotectonic movements, which continued into the Quaternary, shaped the geomorphological and karstic development of the region (Nazik, 1992; Doğan, 1997).

GEOMORPHOLOGIC DEVELOPMENT SURROUNDING THE ÇAMLIK CAVES

Around Çamlık Caves, in the Lakes District of the Taurus Mountains, orogenic belts extend along northwest–southeast and north–south trends, with geomorphological forms such as poljes, palaeo-valleys and young valleys between them (Fig.1). Different climates and tectonic-epirogenic processes, fluvial erosion and karstification that started during the Oligocene region have influenced shaping of the region. The area between Beyşehir Lake and the Manavgat River was subjected to rapid erosion after it acquired a continental character during the Oligocene. Four denudational surfaces correspond with relief systems that developed during Mid Miocene, Late Miocene, Pliocene and Early Pleistocene times. Continuing neotectonic activity (uplift and lowering) kept fluvial and karstic erosion active in this region. As a result the altitude difference between denudational surfaces and intermontain basins increased continuously, supporting the development of karstic landforms.

During evolution of the region in Pre-Late Miocene times the Mediterranean (Fig.1) provided the palaeo base level. Geomorphological development of the region south of the site during the Late Miocene was related to the Mediterranean, that for areas north of the site was related to the palaeo Beyşehir Lake (Nazik, 1992). Flow lines in the area were influenced by the fault lines extending along northwest–southeast trends. Formations including "wild" flysch, Eocene flysch and conglomerates between the carbonate rocks and located according to orogenic effects created suitable areas for the rivers. Rivers located on the outcrps of these formations incised their beds both deeply and laterally, depending upon the local effects of physical erosion.

The study site passed through an area-wide erosional phase with no over-riding structural change after the present land features were imprinted. This large denudational surface, called the Anatolian Peneplain (Erol, 1979, 1983, 2001) is generally preserved at heights of 1800m or above. Changes that took place due to neotectonic movements, which started with the collision of the Anatolian and Arabian plates (Şengör, 1980; Koçyiğit, 1984), orogenic elevation and block faulting, ended the development of this surface. On the basis of surface tectonic data and regional correslations (Erol, 1990; Nazik, 1992; Doğan, 1997) this surface, the result of long-term erosional processes, is of Early–Mid Miocene age. Regions where the peneplain cuts limestones take the form of karstic plateaus.

A second relief system and denudational surface of this system was observed at 1500–1700m, related to a base level that appeared as a result of Tortonian tectonic movements. This surface, which developed in places on the slopes of the Early–Mid Miocene peneplain and in front of Late Tortonian fault lines is Late Miocene in age (Erol, 1990, 2001). Development of this denudational surface was ended by Late Miocene–Early Pliocene tectonic movements. The Pliocene denudational surfaces began to develop in depressions or along the rivers as a result of tectonic movements in Late Miocene–Early Pliocene times.

In some parts of the study site there are extensive karstic features on the denudational surfaces present as karstic plateaus, or in the palaeo valleys between them, where conditions are suitable. Among the macro karstic features that dominate at these sites are poljes and uvalas, which are of fluvio-karstic origin and formed as a result of the karstification of Pliocene palaeo valleys. Among the poljes in this group are Kembos, Çamlık, Derebucak, Kızılova, Bıçakçı and Mumbala poljes (Fig.1). Kembos Polje, the largest polje in the area, started to form by the separation of a tributary of the Pliocene Manavgat River, as a result of Late Pliocene–Early Pleistocene tectonic movements and karstification of a closed basin (Penck, 1918; Alagöz, 1944; Aygen, 1967; Bakalowicz, 1970; Monod, 1977; Biricik, 1982; Eroskay, 1982; Nazik, 1992). The poljes in this region are of fluvio-karstic origin.

Active processes and other factors during the Pliocene–Pleistocene period played a significant part in the geomorphological development process and the establishment of the current situation. During this period new river networks were established with reaches guided in east-west and northeast-southwest directions. The Uludere and its tributaries, the only significant and certainly the most powerful river in the basin, captured drainage from most of the poljes and Pliocene valleys and connected them to the Kembos basin (Fig.1). Balat Stream, which drains the waters of Çamlık Caves, was also captured by the Uludere in early Pleistocene times. Later, the Balat Stream incised quickly as a result of the lowering of the fault lines at the base of the Kembos Polje, which was the local base level. With this incision the Pliocene Çamlık Polje, Mediova, Balat Uvala and Mumbala Polje were connected to Kembos Polje both above and below the ground, and a new karstic period was initiated (Fig.1). The Uludere Stream, which enters the Kembos Polje after collecting the waters of the study site, goes underground through the swallowholes at the west of the polje and joins the Manavgat River after passing through Altınbeşik Cave, near the Manavgat River (Nazik, 1992). Early Pleistocene denudational surfaces were formed around this river network.

ÇAMLIK CAVES SYSTEM (KÖRÜKINI-SULUIN CAVES SYSTEM)

At the study sites different cave systems were developed according to the influences of hydrogeological zones during palaeo and neokarstic periods. The laterally extensive caves developed in relation to the Pliocene relief system, and became fossilized by remaining in the vadose zone. These caves have been developing since the Pliocene. In contrast well-formed "vertical caves" in the high karstic areas (upon the Mid and Late Miocene surfaces) belong to the neokarstic period (post Late Pliocene) and developed in the vadose zone (young vadose caves). The third group of caves in the region comprises active to semi-active caves. These, formed after the Late Pliocene, are of "swallowhole" or "spring" type (Nazik, 1992). Çamlık Caves are in the latter category.

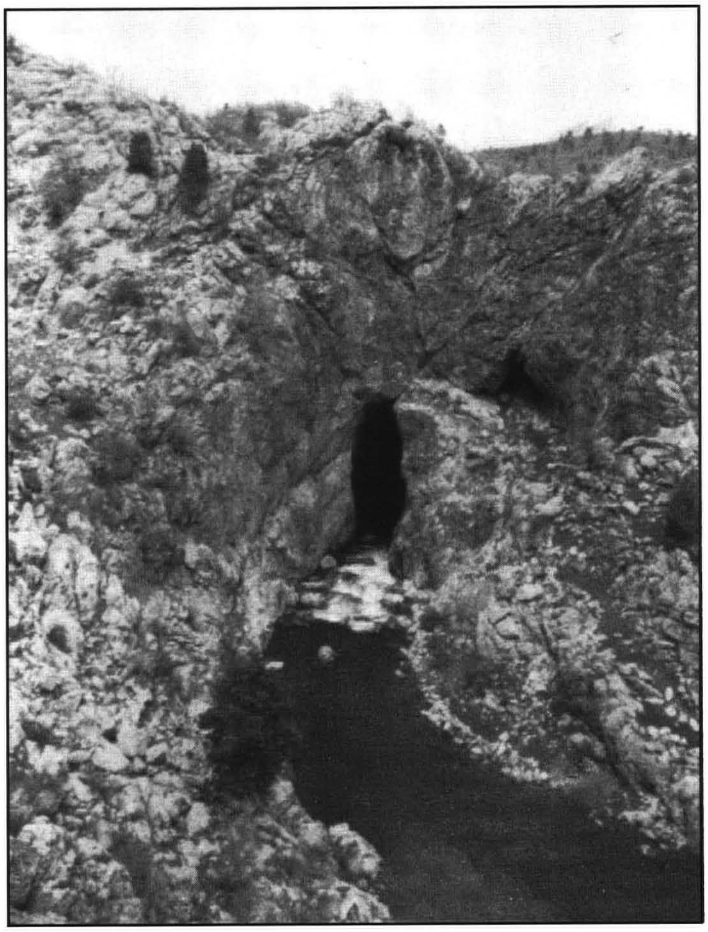

Figure 4. The entrance of the Körükini Cave, the starting point of Çamlık Caves. Uzunsu Stream joints the underground system in Körükini Cave. The cross section of Körükini Cave is of canyon type.

The Uzunsu Stream, which is fed by karstic springs to the west of Reze Mountain, influenced the formation of the Çamlık Caves and also led to them being of "swallowhole" or "spring" type. The stream flows in an entrenched meander with a depth of 5 to10m, and joins the underground system after it enters Körükini Cave southwest of Çamlık town (Figs 3 and 4).

Körükini and Suluin caves, which developed in an east-west direction, are parts of the underground system that were separated from each other by a collapse doline (Figs 3 and 5). The width and height of the 1330m-long Körükini Cave range between 8-15m and 4-20m respectively (Fig.6). There are many large and small lakes in the cave and, in places where underground water doesn't reach, speleothem deposition is observed. Speleothem pools in particular reach large proportions. The depth of the lakes ranges between 0.5 and 5m, and there are sand and pebble islets and blocks between them (Fig.6).

Körükini Cave, which drops 37m between its start and the end points terminates at a collapse doline (Fig.6). The exit of this cave, located in the collapse doline, is called Değirmenini (Fig.7). The Uzunsu Stream flows underground in the Körükini Cave, resurfaces from Değirmenini and courses across the 315m-long floor of the collapse doline before passing into Suluin Cave (Fig.8). The total length of Suluin Cave, which is 52m below the entrance of Körükini Cave is 290m. Its width and height are 5 to 20m and 10 to 15m respectively. A large part of the cave is floored by lakes with a depth of 4 to 5m (Fig.6). Water leaving the cave is called the Balat Stream, and it flows to the Kembos Polje in the west (Fig.1). The cave system, which has an altitude difference of 67m between its starting point (Körükini entrance) and its termination (Suluin exit), has a length of 1936m (Nazik, 1992; Nazik et al., 1993).

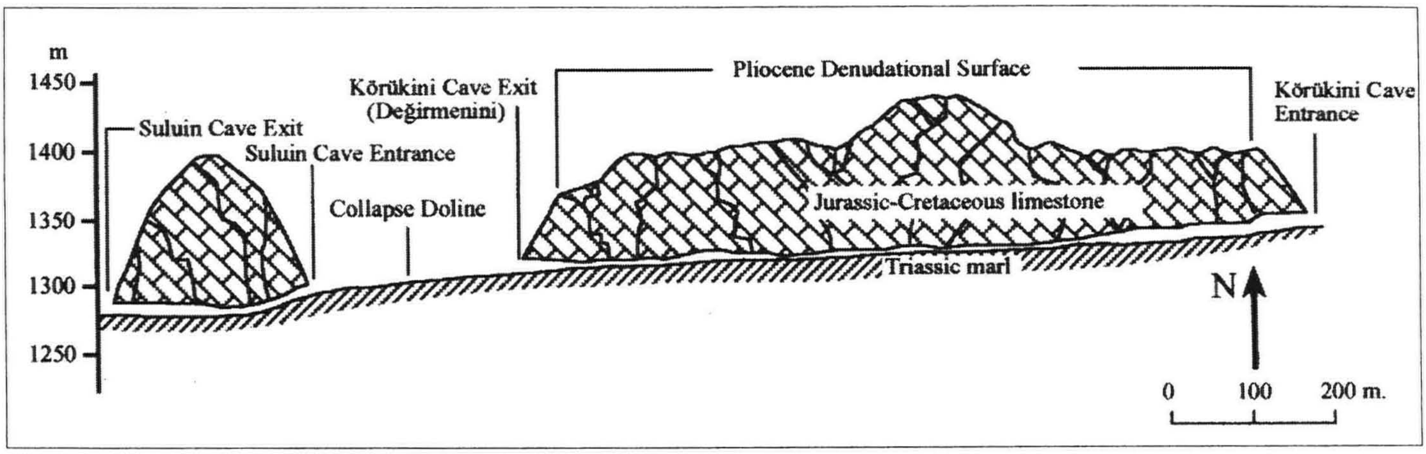

Figure 5. Profile of Çamlık Caves.

The Körükini-Suluin Cave System is the product of a series of geomorphological developments that began in the Pliocene and continue today. Influences of the Mediterranean and the Manavgat River have been very important to the process. The Çamlık Caves area is a Pliocene denudational surface with poljes (Fig.3). Çamlık Polje and Mediova Polje date from this period. A river that flowed on the denudational surface after the Pliocene (the Uzunsu River) was rejuvenated and incised into its bed due to Late Pliocene–Early Pleistocene tectonic movements (epirogenic elevation of the Taurus). The Uzunsu Stream became part of the karstic conduit as a result of this rejuvenation, and modification of the caves was initiaqted. This stream enlarged the caves after being directed and deepened along a fault line extending in a southeastern–northwestern direction at the site of the Körükini Cave. Thus, Çamlık Caves is a cave system that started its enlargement in the Early Pleistocene and its development continues today.

The old stream bed, located on the denudational surface between Suluin and Körükini caves underwent karstification and was converted into a palaeokarstic valley (Fig.3). This palaeokarstic valley displays old flow lines and erosion corridors, as well as Çocukattıkları Delik Cave, containing large dimension dolines and swallowholes, which characterize multi-staged development. The karstification that took place after the river passed underground caused the valley floor to lose its original profile.

Çamlık Caves have developed in the Jurassic–Cretaceous Çamlık Chalk (Monod, 1977, 1979), an important rock unit in the Beyşehir-Hoyran Nappes. The Çamlık Chalk surface covering a large area around the study site is very prone to karstification and the formation of caves. As a result of deep tectonic fissuring of the Eocene flysch the Çamlık Chalks that remained upstanding formed large cornices. Generally these limestones, covered by impervious marl and shale (Figs 2 and 5), create a very suitable environment for the development of lateral caves (Nazik, 1992).

Analyses of samples taken from various regions of the Çamlık Chalks revealed that the unit contains 98% $CaCO_3$ and an average of 1% $MgCO_3$. The Formation has a porosity of 2 to 5%, which is largely due to the effects of tectonics and karstification (Nazik, 1992). Because of the lithological features and the structure, with a large number of fractures, there has been intensive karstification of these limestones. The underground thickness of the limestones isolated by impervious units is very small, which is why the karstic features developed laterally rather than vertically (shallow karst). Triassic marls underlying the limestone guide the position of the base level. The Körükini-Suluin cave system, and Küçük Suluin, Dedetarlası Düdeni, Çocukattıkları Delik, Dölek Düdeni and Asmacıini caves (Fig.3) have very similar geological and geomorphological features (Nazik, 1992). These caves were formed within allochthonous Lower Jurassic–Cretaceous limestones.

From the hydrological point of view Körükini and Suluin caves are active/semi active cave systems. Abundant water flows in the winter and spring months because of the Uzunsu Stream, whereas the amount of water decreases significantly in drought periods. However, the lakes in the caves remain regardless of the conditions. The waters of Dedetarlası Düdeni, Dölek Düdeni and Çocukattıkları Delik caves also flow into this cave.

When the Körükini and Suluin Cave System is evaluated with regard to drainage network or gallery features they are

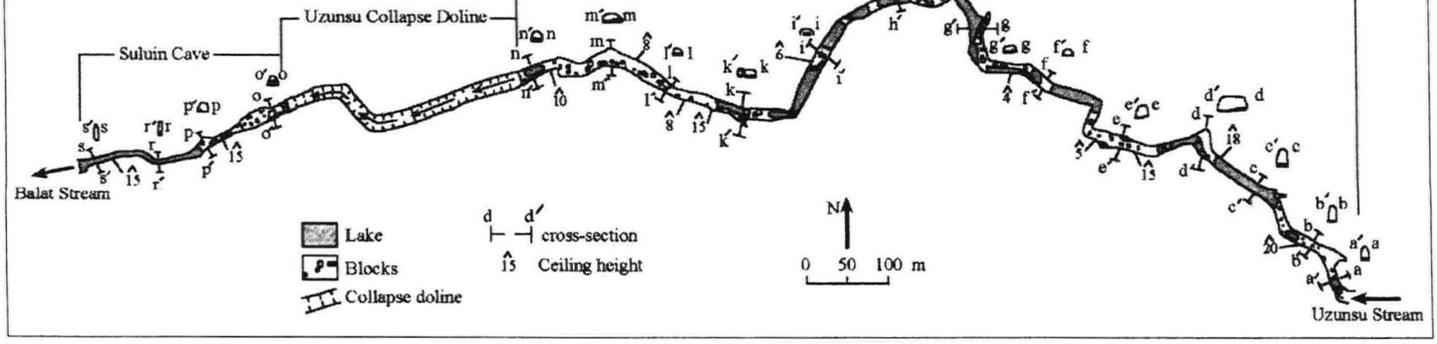

Figure 6. Plan of Çamlık Caves.

observed to be of single conduit cave type, with sinuous galleries. The fact that development of the Çamlık Caves System depended upon a river explains why the cave gallery is sinuous. This agrees well with literature (White, 1988; Ford and Williams, 1989).

Cross-sections of the Çamlık Caves are of elliptical-canyon tube type. Passages formed under hydraulic control are of two end-member types – the canyon and the elliptical tube. Canyons show evidence of fast-moving water. The cross-section of the cave has a canyon tube as the entrance of Körükini and the exit of Suluin caves are located on a fault line. The cave cross–section is an elliptical tube in the middle of the cave, especially at the entrance and exit of the Uzunsu Collapse Doline. Elliptical tubes form as groundwater flows under pipe- full conditions.

UZUNSU COLLAPSE DOLINE

The Uzunsu Collapse Doline was formed by the collapse of the roof of the cave at 1331 to 1646m of Çamlık Caves, which was a single gallery during the Early Pleistocene. The doline starts at the exit of Körükini Cave (Değirmenini) and then returns to the underground at the entrance of Suluin Cave. The doline is 315m long and 40m deep. Because the floor of the doline is also the bed of the Uzunsu Stream, there is a 15m altitude difference across the doline as a result of the long profile of the stream.

Although the Uzunsu Collapse Doline is similar to other dolines described in literature as regards its formation mechanism, from a karstic point of view it has a markedly different shape. In other word it does not look like a circular or elliptical collapse doline (obruk). The shape of the Uzunsu Collapse Doline accords well with the shape of the pre-existing cave gallery. As the cave gallery in which the collapse doline formed makes a small turn it is not possible to see the whole of the doline (Fig.9). As the base of the collapse doline is the floor of an ancient cave gallery it slopes in the same direction as the flow of the Uzunsu Stream. There are potholes and interconnected or independent pools at the base of the doline. Additionally there are many big and small limestone blocks and other collapse debris (Fig.10).

The fact that there is an underground flow system at the base of the Uzunsu Collapse Doline, and hence detailed knowledge of the associated gallery system, makes explanation of the formation of the doline straightforward. The biggest reason for the collapse of the roof of the Çamlık Cave System at this point is the development of the Uzunsu Stream. The palaeo valley of the Uzunsu Stream on the Pliocene denudational surface and the cave gallery that the stream started to form after it was included into the underground system, coincide at this point (Fig.3). Stress relaxation above the cave gallery that played the biggest part in the formation of the collapse doline, was active before doline development. As a result of elevation of the cave ceiling the existing fractures were enlarged and new fractures formed as a result of stress relaxation. As the elevation of the cave roof increased, the ceiling quickly reached a point where it could no longer support the load upon it and ceiling balance was lost resulting in its collapse and the formation of the doline

.CONCLUSIONS

The Çamlık Cave System started to form as a result of the capture of the river that was flowing upon the overlying denudational surface into an existing underground system during the Late Pliocene after being directed along a northwest–southeast fault line. The Pliocene valley of the stream that formed the Çamlık Caves was karstified and converted into a palaeokarstic valley containing dolines. Today Çamlık Caves comprise two parts, namely Körükini Cave and Suluin Cave. The collapse doline developed at the intersection of the cave

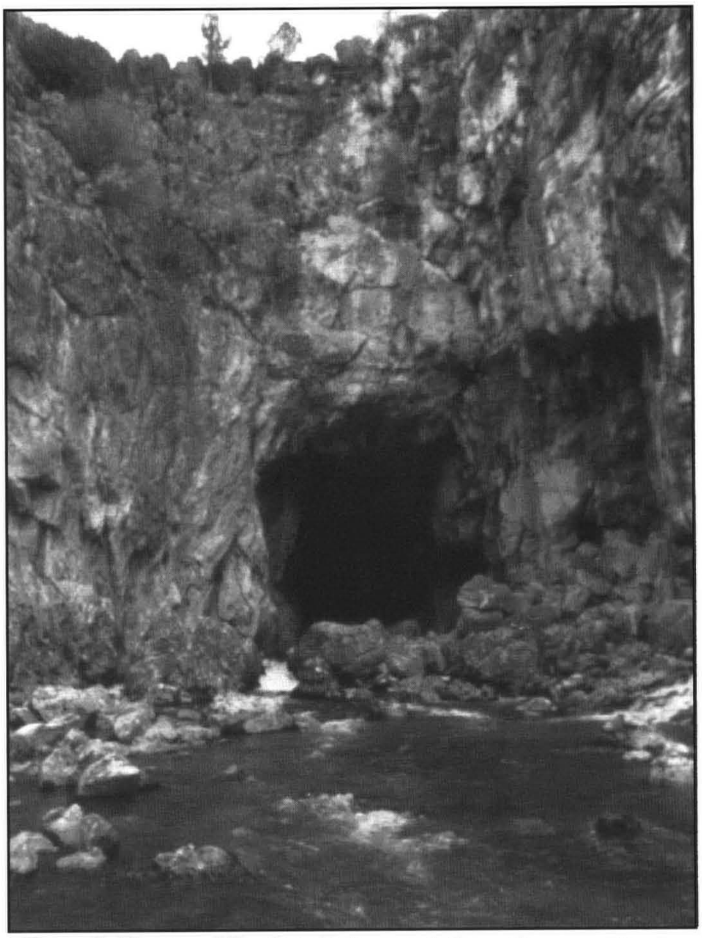

Figure 7. *The exit of Körükini (Değirmenini) Cave. The cave exit cross-section is elliptical.*

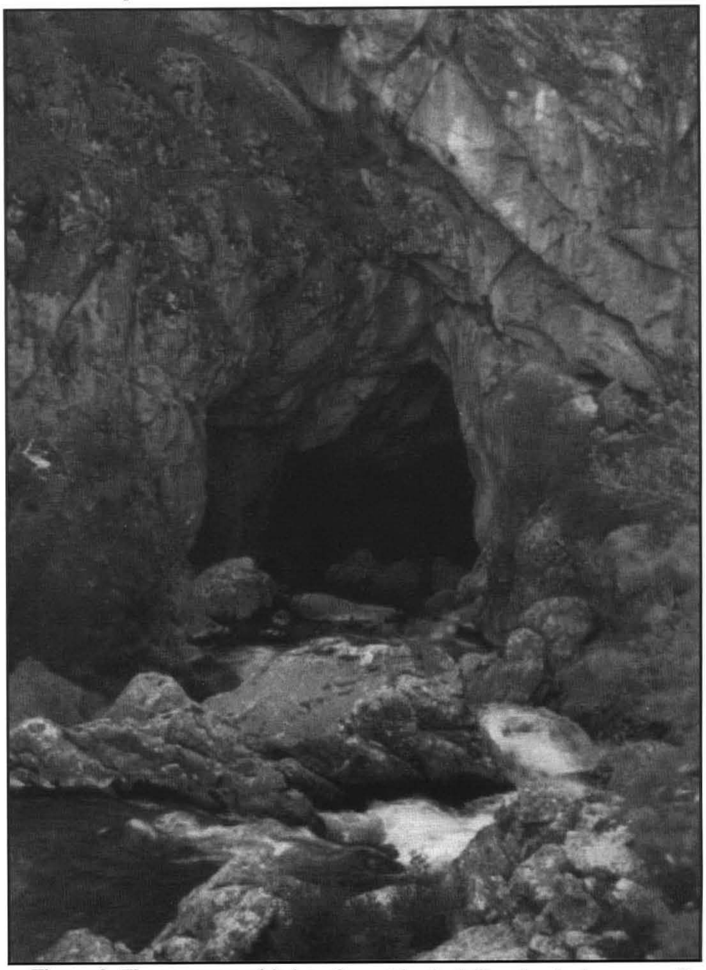

Figure 8. *The entrance of Suluin Cave. The fault line that had a great effect upon the development of the Çamlık Caves is above the entrance. The height of the cave entrance is 3m.*

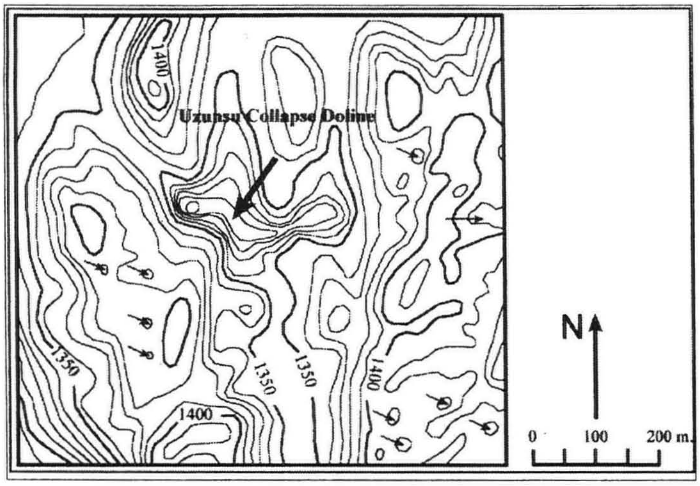

Figure 9. Topographic map of the Uzunsu Collapse Doline

Figure 10. The Uzunsu Collapse Doline. The floor of the doline is covered with blocks that are the remains of the former cave ceiling.

gallery and the Pliocene palaeo valley. The Pliocene palaeovalley passing above the cave resulted in stress relaxation within the cave ceiling. Enlarged and newly-formed fractures resulting from the stress relaxation were instrumental in the formation of the collapse doline.

The fact that the limestones in which the cave developed are not very thick in this region, and the base was underlain with impervious formations, caused the karst base level to be fairly shallow. Therefore, confined within this shallow karst, the Çamlık Caves System was unable to develop to greater depth.

When Körükini and Suluin caves are examined as regards the drainage system (gallery features) they are seen to be single conduits with sinuous galleries. The fact that Çamlık Caves were developed by a stream forced the cave gallery to have a sinuous character. Cross-sections of Çamlık Caves are of elliptical-canyon type. The canyon tube sections indicate the effect of the fault line and the elliptical tube reveals that underground water flows in the tube were full.

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CAVE AND KARST SCIENCE Vol. 30 No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Hydrodynamic behavior of the Gilan karst spring, west of the Zagros, Iran.

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Abstract: The Gilan aquifer is located in the southwest of Kermanshah, west of Iran. This aquifer discharges through the Gilan Spring, with an average discharge rate of 938 l/s. The catchment area of the Gilan Spring was determined using hydrogeological approaches. Electrical conductivity (EC), water temperature, pH, major ions and discharge were measured at two-weekly intervals within the period of September 2000 to September 2002. Water temperature, EC, and stage were measured once every two days in the wet season. Calcite, dolomite and gypsum saturation indices and partial pressures of CO_2 were calculated using the WATEQF hydrochemical model. The hydrochemical data, percentages of base flow and recession coefficients imply that the type of flow is mainly diffuse in the Gilan aquifer. The EC time Series of the Gilan Spring implies that a high percentage of the spring water is supplied by conduit flow. The peak discharge occurs when conduit water in the part of the aquifer near the spring provides the Gilan Spring discharge, whereas conduit water from more distant parts of the aquifer reaches the spring during recession periods. The contribution of conduit flow during dry periods reduces the variations of physico-chemical parameters and produces a low value of recession coefficient. This kind of behaviour occurs in aquifers with a long and narrow catchment area, in which rain duration is significantly less than the time lag so that the EC decreases during dry periods.

(Received 12th April 2003; Accepted 17th September 2003)

INTRODUCTION

During the last three decades temporal variations of physico-chemical parameters of karst springs have been used to determine aquifer characteristics. Jakucs (1959) showed that the chemical quality and discharge of karst spring water can vary with time. Garrels and Christ (1965) categorized the flow in karst regions into open and closed systems, based on the amounts of CO_2 available for dissolution. White and Schmidt (1966) and White (1969) classified the flow in karst aquifers into conduit and diffuse flow. Shuster and White (1971, 1972) concluded that the type of flow (diffuse or conduit) can be determined by its chemograph. Ternan (1972), Jacobson and Langmuir (1974), Ede (1972) and Cowell and Ford (1983) evaluated the flow systems in karst formations using other criteria, such as type of recharge, electrical conductivity and coefficient of variation of electrical conductivity (EC), discharge coefficient of variation, temperature and time variations of the different parameters.

Based on the above ideas, in a diffuse flow system, recharge is generally conducted through a network of numerous small joints and fractures that are distributed in the karst aquifer. The openings of these fractures are smaller than one centimetre and water slowly reaches the groundwater in a laminar manner. One of the main

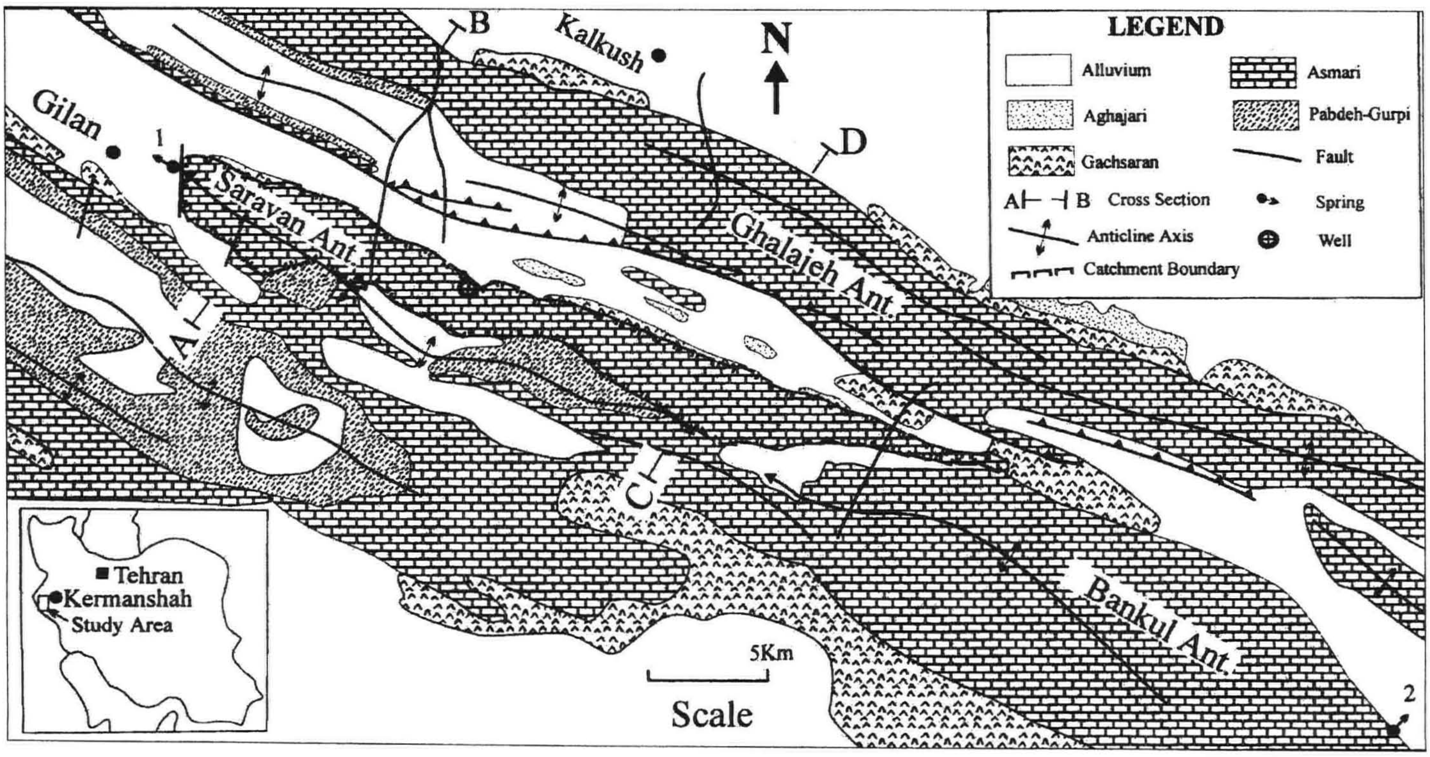


Figure 1. Hydrogeological map of the study area.

Year	Period	Volume (MCM)		Percent of volume (%)	
		Base	Quick	Base	Quick
2000-2001	Dry period	14.33	1.406	91.1	9.9
	Wet period	10.58	1.56	87.1	12.9
	Total year	24.91	2.97	89.3	10.7
2001-2002	Dry Period	17.49	0.37	97.9	2.1
	Wet period	10.50	1.28	89.1	10.9
	Total year	27.99	1.65	94.4	5.6
2000-2002	Total period	52.90	4.62	92.0	8.0

Table 1: *Amounts and percentages of base flow and quick flow of the Gilan Spring.*

peculiarities of these aquifers is the small variation of physical and chemical properties of the discharging springs. Natural discharge from such a system is usually through a large number of smaller springs and seeps. In a conduit flow system, the aquifer is fed through either large open fractures (ranging from one centimetre to more than one metre) or dolines. In such systems, water very quickly reaches the groundwater and, ultimately, the springs in a turbulent manner. Hence, the physico-chemical properties of the springwaters are non-uniform. In this type of system the discharge usually occurs through one single large spring.

Bakalowicz (1977), Atkinson (1977), Scanlon and Thrailkill (1987) and Raeisi *et al.* (1993) were not able to use the criteria proposed by previous workers to determine the flow regime, and they recorded contradictory results. The reason probably reflects the fact that purely diffuse or purely conduit flow systems rarely occur in nature, rather it is a combination of these two types of flow that usually prevails. Raeisi and Karami (1996) suggested that when the physico-chemical characteristics of a karst spring are to be used to determine the properties of the related aquifer, the first step should be the evaluation of the effects of external factors on the outflow. Lopez-Chicano *et al.* (2001) analyzed the hydrogeochemical processes in waters of the Betic Cordilleras in Spain by studying hydrography, temporal evolution of physico-chemical parameters, ionic ratios (mainly Mg/Ca) and by means of simple and multivarient statistical analysis. They concluded that the aquifer exhibits diffuse flow. Time series variations of physico-chemical parameters of springs have been inspected by different researchers, including Ashton (1966); Bakalowicz *et al.* (1974); White (1988); Williams (1983); Sauter (1992); Ryan and Meiman (1996); Raeisi and Karami (1996, 1997); Desmarais and Rojstaczer (2002) and White (2002).

In a typical karst system, after intense rainfall, discharge increases within a short period and then decreases slowly. In this period, the EC shows an increasing–decreasing–increasing trend. Based on electrical conductivity response, Desmarais and Rojstaczer (2002) divided spring response into three stages. The three stages of the model include flushing, dilution and recovery. The flushing stage marks the initial response in the spring to storms. The beginning of this stage is signalled by the increase of the slope of the conductivity curve of the spring. There are two hypotheses concerning the water source that causes this flushing of the spring:

1. The flushed water is water that has interacted or equilibrated within the soil zone, and possibly resides in small pores or fractures near the land surface, i.e. the subcutaneous zone. This water would be relatively warm and would likely contain dissolved salts (or would dissolve salts from the soil during transit), which would give the water a relatively high conductivity. The warmer, high conductivity water would be mobilized by the rainwater infiltrating into the soil and pushed toward the spring.
2. The new rainwater is able to recharge the aquifer rapidly, possibly through fractures or surface swallets, and it mobilizes older, deeper water that has been residing in smaller fractures and pores out of the aquifer. This 'old water' is at or near equilibrium with the limestone, but the new water is not. The old water, because it has resided in the aquifer for a relatively long time, would have higher conductivity than the baseflow spring water. Flushing is not typical of all carbonate springs (Ryan and Meiman, 1996; Desmarais and Rojstaczer, 2002).

The dilution phase begins with the peak in the conductivity curve and ends when the conductivity minimum is reached. The start of the conductivity decrease represents the first arrival of storm water at the spring. During this phase, the temperature commonly levels off and then remains constant until the next storm, whereas discharge continues to decrease. The area of the recharge basin is the main factor controlling the length of this phase. After that time period, the spring begins to 'recover' because there is very little recharge water remaining.

However, the system response can also be explained by a competition between the velocity at which recharge water is moving through the system, how fast this 'new' water dissolves carbonates to gain the same chemistry signature as the 'old' aquifer water, and the amount of mixing that takes place between these two water sources.

The recovery phase begins when the minimum conductivity is obtained. During this phase, conductivity increases steadily until the next storm begins. The concentrations of all the major cations and anions increase during this period. All of these changes indicate that the system is returning to equilibrium conditions. The conductivity minimum probably indicates that the last of the recharge water has been in contact with the aquifer rock long enough to begin to dissolve limestone and/or dolomite in sufficient quantities to allow the overall system to begin to recover from dilution.

The minimum electrical conductivity corresponds to the maximum dilution of groundwater by fresh recharged water. The lag time is a measure of the length of time required for the arrival of unsaturated water (minimum of EC) to reach the recording station (Hess and White, 1988).

Isotope studies indicate that there is some dilution in spring water during the rainy season because of the mixing of event water with spring water, and there is enrichment in the dry season (Stewart and Williams, 1981; Ford and Williams, 1989; Ferdrickson and Criss, 1999; Abbot *et al.*, 2000).

The objective of this study is to determine the characteristics of the Gilan aquifer, especially the type of flow, using geological setting, time series analysis of physico-chemical parameters and ^{18}O and deuterium tests of the Gilan Spring water.

Parameter	T (°C)	Ec (µS/cm)	PH	Q (l/s)	HCO_3 (Meq/l)	Cl	SO_4	Na	K	Ca	Mg	TDS (mg/l)	$LogpCO_2$	SI_c	SI_d	SI_g	Ca/Mg
Average	19.7	631.3	7.48	938.2	5.1	0.55	2.24	0.41	0.02	4.26	2.78	564.8	-1.9391	0.17	0.223	-1.56	1.5
Minimum	18.1	583	6.22	730	4.2	0.35	1.6	0.3	0.01	3.55	2.15	491	-3.2	-0.94	-2.1	-1.8	1.2
Maximum	21.1	677	8.59	1121	5.35	0.8	4.75	0.7	0.05	5.35	3.45	690	-0.77	1.4	2.6	-1.2	1.9
S.D.	0.39	18.25	0.49	108.7	0.25	0.10	0.59	0.07	0.01	0.35	0.28	41.05	0.51	0.49	1.00	0.14	0.15
C. of Var.	2.0	2.9	6.6	11.6	4.8	18.6	26.4	17.9	40.6	8.1	10.2	7.3	-26.3	282.7	448.5	-8.8	9.5

Table2: *Average, Maximum, Minimum, Standard deviation and Coefficient of variation of the Gilan Spring.*

HYDROGEOLOGICAL SETTING

The study area is situated 150km southwest of Kermanshah, in the west of Iran (Fig.1). The topography of the study area is dominated by the folded mountain structure of the Zagros thrust zone (High Zagros) (Sahabi *et al.*, 1998). Geological formations in the study area in increasing order of age are the recent alluvium, Tertiary Aghajari marl and sandstone, Tertiary Gachsaran gypsum and marl, Tertiary Asmari dolomitic limestone and dolomite and Cretaceous Pabdeh-Gurpi marl and shale with interbedded thin marly limestone (Fig.1). The stratigraphical and structural characteristics of the Zagros sedimentary sequence were described in detail by James and Wynd (1965), Falcon (1974) and Stocklin and Setudehnia (1977). The geological settings of the study area were studied by Sahabi *et al.* (1998).

The study area is the Saravan Anticline, with a length of 37km and a width of 5km. The anticline follows the general northwest–southeast trend of the Zagros Mountain Ranges. The highest elevation is about 1920m a.s.l., whereas the minimum elevation is approximately 800m a.s.l. near the Gilan Spring on the west side of the anticline. This anticline has two plunges, the southeastern plunge being connected to the Bankul Anticline, forming a saddle structure. The southeastern plunge is 550m higher than the northwestern. The southern flank of the Saravan Anticline is crushed by a thrust fault, especially near the northwestern plunge, such that the underlying Pabdeh-Gurpi Formation is at outcrop. The Asmari Formation, the main aquifer in the Saravan Anticline, is sandwiched between the impermeable underlying Pabdeh-Gurpi and overlying Gachsaran formations. A thin layer of alluvium covers most of the Gachsaran Formation in the plain adjacent to the northern flank of the Saravan Anticline. The average thickness of the Asmari Formation is approximately 300m in the study area. The karst water of the Saravan aquifer is discharged only from the Gilan (western plunge) Spring (No.1 on Fig.1). The maximum, minimum and mean annual discharge of Gilan Spring were 1121, 730 and 938 l/s respectively during the study period. The long-term (33 years) average annual precipitation in the study area is 625mm, and was 480mm during the years 2000 to 2002. The precipitation is mainly in the form of rain and it occurs from mid-autumn to mid-spring.

The mode of water entry (recharge) to the Gilan aquifer can be termed autogenic (Williams, 1983; Ford and Williams, 1989), which implies that recharge is diffuse through small joints and fractures over the entire catchment, and not as recharge via stream sinks or dolines. Surface run-off is not observed. No dolines, pits, shafts or large caves are present in the catchment area of the Gilan Spring. There are numerous joints and fractures in the Asmari limestone, most of them being filled by soil. Grikes, karren and small shelter caves are the main karst features. Trees such as oaks cover approximately 30% of the surface.

THE GILAN SPRING CATCHMENT AREA

One of the most complex and difficult problems to deal with in karst hydrology, hydrogeology or geology is the precise determination of catchment boundaries. The catchment area of the Gilan Spring was calculated by a simple balance equation during the hydrological year of 2000-2001.

$$A = V/P\ I$$

In which A is the catchment area (m^2), V is the total volume of the Gilan Spring discharge (m^3), P is the total precipitation (m) and I is the infiltration coefficient. Based on the experience at other karst sites in Iran (Rahnamaei, 1994; Pezeshkpour, 1991; Karimi *et al*, 2001) the infiltration coefficient is expected to be within the range of 0.4 to 0.6. The catchment area is between 113 and 159km², considering the uncertainty of the infiltration coefficient. The boundary of the catchment area was determined by Karimi (2003) using the 0.6 infiltration coefficient. The catchment area is limited to the northern flank (94km²) and some parts (16km²) of the southern flank of the Saravan Anticline near the Gilan Spring (Fig.1). The following factors justify the estimated boundaries of the Gilan Spring catchment area:

1. Water balance of the adjacent anticlines (Karimi, 2003) showed that most of their waters discharge through the related spring(s).

2. The hydrogeological relationship between the Saravan anticline and the parallel neighboring anticlines is disconnected (Fig.2) since: a) the karstic formations in synclines are normally buried under very thick overburden (more than 1000m). Karstification is usually very shallow. b) The elevations of underlying impermeable formations under the crest of parallel anticlines are higher than the ground water level in the adjacent alluvium aquifer, such that the alluvium water cannot flow toward the adjacent anticlines.

3. As previously mentioned, the Asmari Formation is sandwiched between two impermeable formations, so the Asmari water cannot be discharged to deeper formations.

METHOD OF STUDY

The electrical conductivity (EC), water temperature, pH, major ions (calcium, magnesium, sodium, potassium, bicarbonate, sulphate and chloride) and discharge of the Gilan Spring were measured every two weeks during the period from September 2000 to September 2002.

The stage, EC, and water temperature were measured once every two days during the wet season. The water temperature, EC, pH and discharge of the spring were determined in the field during sampling.

Calcium and magnesium were measured by titration with EDTA and Murexide and Eriochrom Black-T as indicators. Sodium and potassium concentrations were determined using flame photometry methods. Chloride and sulphate were measured by the Mohr and turbidity methods, respectively. Bicarbonate was determined by titration with HCl and Methyl orange as indicator.

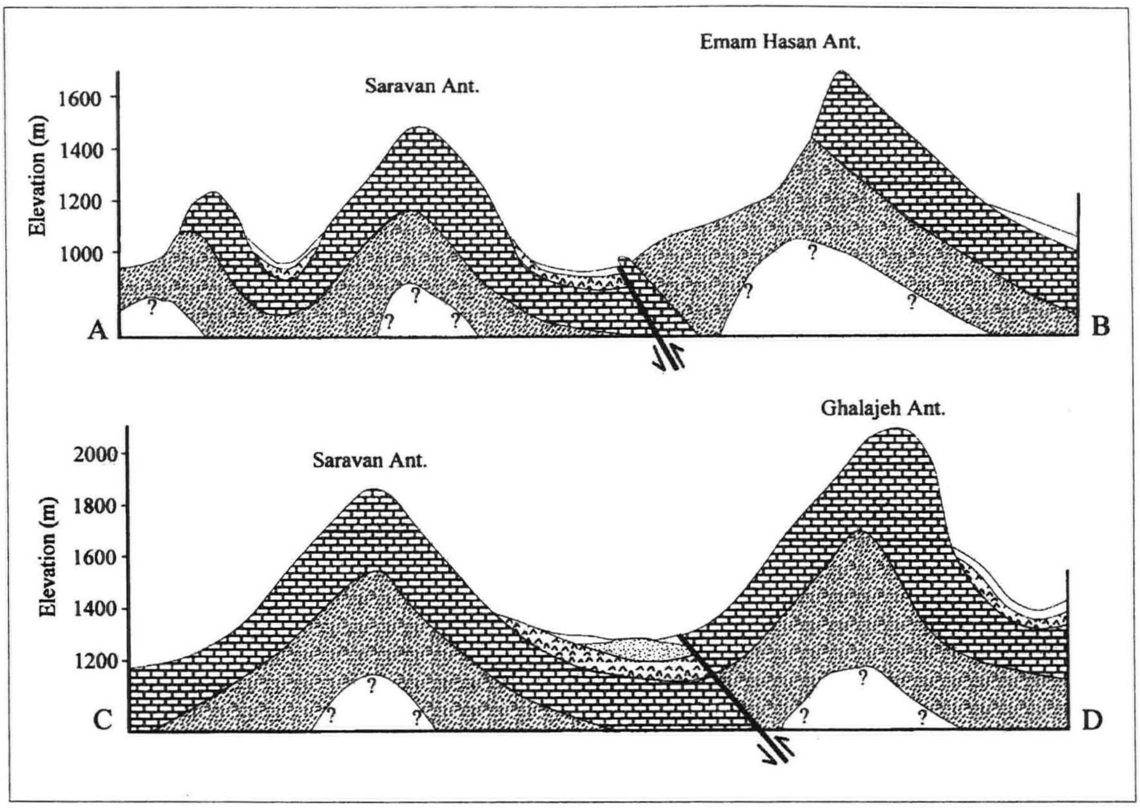
Carbon dioxide partial pressure (PCO_2), calcium saturation index (SI_c), dolomite saturation index (SI_d) and gypsum saturation index (SI_g) were calculated using the WATEQF model (Plummer *et al.*, 1978). The accuracy of analysis was estimated from the electro neutrality (E.N.) condition…

$$E.N\ (\%) = [(\text{sum cations} - \text{sum anions})/(\text{sum cations} + \text{sum anions})*100$$

... in which cations and anions are expressed as m-eq/l. These errors were less than 5% in all the samples.

In order to investigate the isotopic characteristics of the Gilan Spring, twelve monthly samples were analyzed during the 2000-

Figure 2. Geological cross sections of the study area.



2001 water year. Sampling was carried out in 150-ml glass bottles and analyzed in the laboratory of GSF (National Research Centre for Environment and Health) centre in Germany, by the group led by Professor Klaus-Peter Seiler.

HYDROGRAPH AND RECESSION COEFFICIENTS OF THE SPRING

The hydrograph of the Gilan Spring and the daily rainfall during the period of 2000 – 2002 are presented in Fig.3. The time delay between the rainfall and the observed peak discharge was approximately 13 and 20 days respectively. The base flow and quick flow of the Gilan Spring was separated by the local minimum method (Pettyjohn and Henning, 1979; Ward and Robinson, 2000) (Fig.3). The base flows are 89.3% and 94.4% in the hydrological years 2000-2001 and 2001-2002, respectively (Table 1 and Fig.3). The high percentages of base flow imply that the type flow of the Gilan aquifer is apparently mainly diffuse.

The recession curves of the Gilan Springs are presented in Fig.4. The recession coefficients were calculated by a curve-fitting algorithm (Vasileva and Komatina, 1997). The recession coefficients of the Gilan Spring are 0.0004 d^{-1}, 0.0007 d^{-1} and 0.0017 d^{-1} during the study period. The values of recession coefficients being less than 0.0017 d^{-1}, imply a fissured or diffuse system (Sauter, 1992).

HYDROCHEMISTRY OF THE GILAN SPRING

Average, minimum, maximum, standard deviation and coefficient of variation of the physico-chemical parameters of the Gilan Spring during the study period are presented in Table 2. Within the period of September 2000 to September 2002, the average, minimum, and maximum electrical conductivity (EC) are 631, 583 and 677 micro-siemens/centimetre (μS/cm), respectively. The type of water is calcium-magnesium bicarbonate. Analysis of the Gilan Spring during the study period indicated that 94% of the total dissolved ions consisted of bicarbonate (33%), calcium (28%), magnesium (18%) and sulphate (15%). The Ca/Mg ratio ranged from 1.2 to 1.5. The Ca/Mg ratio of less than 3 implies that the karst water route is through dolomite and dolomitic limestone (Kronfeld and Rosenthal, 1987; Kattan, 1997, Hess and White, 1988; White, 1988). It can be concluded that the Asmari Formation is the source of the Gilan Spring. Karst waters in the Zagrossides of Iran, which are not in direct contact with salt domes or gypsum evaporite formations, have a good quality. In general the electrical conductivity is less than 500 μS/cm (Raeisi *et al.*, 1993). The main reason for the high EC in the Gilan Spring relates to an unexpectedly high percentage of sulphate ions (15%). A direct relationship is observed between the peak values of EC and sulphate in Fig.5. The catchment area of the Gilan Spring is in direct contact with the Gachsaran Formation, such that seepage water from this formation enters the Gilan Spring aquifer, increasing the sulphate and hydraulic conductivity.

Other criteria proposed by different authors are used to evaluate the flow type in the Gilan Spring. The coefficient of variation of total hardness is less than 5% in diffuse flow, whereas this coefficient ranges from 10–24% in conduit flow (Shuster and White, 1971). Since the coefficient of variation of total hardness is 7.6% in the Gilan Spring, the type of flow is mainly diffuse. The standard deviation in the water temperature of Gilan Spring is about 0.39, which implies diffuse flow according to Cowell and Ford's (1983) criteria. Jacobson and Langmuir (1974) classified karst flow systems on the basis of EC, coefficient of variation of EC, and coefficient of variation of discharge. The average EC of 631 μS/cm, EC coefficient of variation of 2.89% and discharge coefficient of variation of 11.6% indicate that the flow system in the Gilan aquifer is diffuse.

It can be concluded that based on the percentage of base and quick flow, recession coefficients and hydrochemical data, the dominant flow regime in the Gilan Spring is diffuse. Although all the above criteria imply that the Gilan Spring has diffuse flow and the percentage of base flow is about 92%, the following discussion reasons that the percentage of base flow is less than the value calculated by the local minimum method.

SCHEMATIC MODEL OF THE GILAN KARST AQUIFER

The special behaviour of the Gilan Spring physico-chemical parameter time series cannot be justified by the conventional models. Therefore, a schematic model of groundwater flow has been proposed for the Gilan Aquifer, based on the geological setting, mode of recharge, and previous experience on some other aquifers in the Zagros region (Raeisi *et al.*, 1993; Raeisi and Karami, 1997; Karimi *et al.*, 1998). This model will be justified by the interpretation of the Gilan Spring physico-chemical parameter time series. The schematic model of the Gilan aquifer is presented in Fig.6. Part of the recharged water is transferred directly to the groundwater as conduit flow via the fractures and probable hidden shaft(s) in a short period, whereas the other part is stored in the small pores in the subcutaneous zone and gradually joins the groundwater. The steep slopes direct the water into narrow parallel conduits in the direction of the dip, preventing the collection of water in one major conduit. The main conduit develops at the foot of anticline parallel to strike where branches of small conduit join each other. The Gilan's catchment area is 4km wide and 37km long, and the recharged water emerges only from the Gilan Spring Therefore at least one conduit with maximum discharge of 1121 l/s must transport the recharged water near the spring discharging point. The hydrogeological relationship between the southern and northern flanks is disconnected in some parts of the Gilan Spring catchment area (Fig.2). Therefore two main conduits probably transport the karst water to the spring. The longest conduit is located in the northern flank, with a length of around 37km. The length of the major conduit in the southern flank is about 6km. These two conduits join each other near the spring emergence.

The following evidence indicates the existence of conduit(s) in the catchment area of Gilan Spring.

1. The catchment area of Gilan Spring is very wide and long and the water of Saravan aquifer is discharged via this spring. Some conduits in the catchment area must exist to carry water from distant locations.

2. The high discharge of the spring dictates the requirement of conduit(s) for carrying such volumes of water.

3. Spring discharge rapidly increases after very heavy rains (>40mm), indicating the existence of conduit(s), which transmit the impulse from the catchment area to the spring.

Peak flow in a surface stream occurs when the run-off water from all parts of the catchment area contributes at the measurement station. The recharged water near the Gilan Spring flows out of the system when recharged water from farther parts of the aquifer has not yet reached the spring. The main reason is that rain duration (maximum 5 days) is significantly less than the lag time between rainfall and peak discharge (13 to 20 days) in the study area. Moreover, the lag time in the karst spring is defined when the EC has its lowest value after a rainfall. In the Gilan Spring, the EC increases gradually and reaches peak value with a small increase in discharge (Fig.5) followed by a decrease in the dry period even up to seven months after the centroid of rainfalls. In most karst aquifers, the EC reaches a minimum after a rain event and then begins to rise during the dry period (Sauter, 1992; Hess and White, 1988; Ashton, 1966; Bakalowicz *et al.*, 1974; White, 1988; Williams, 1983; Ryan and Meiman, 1996; Raeisi and Karami, 1996, 1997; Desmarais and Rojstaczer, 2002; White, 2002; Raeisi *et al.*, 1993 and Karimi *et al.*, 1998). The decreasing trend of EC in Gilan Spring is not expected, especially if 90% of the water is apparently diffuse flow in dry period (Table 1). The decreasing trend of EC is justified when the fast conduit flow of far distances continues during the dry period. The average flow velocity of water at the main conduit of the Gilan Aquifer is estimated to be more than 7.3 m/h based on the conduit length of 37km and lag time of at least 7 months. Flow velocities through karst conduits for straight-line plan distances are more than 5 m/h (Aley, 1975; Bakalowicz, 1973, Kruse, 1980; Williams 1977; Ford and Williams, 1989; Hess and White, 1988; Raeisi *et al.*, 1999). Therefore it could be concluded that the conduit water from different parts of the aquifer does not reach the Gilan Spring at the same period. The peak discharge is provided only from some parts of the aquifer that are near the Gilan Spring and it needs more than 7 months for diffuse system flow to dominate the spring discharge. The time series of Oxygen 18 and Deuterium reveals that these values decrease during the dry period (Fig.7); which implies that there is lower evaporation of this water (Clark and Fritz, 1997). This condition indicates the predominance of conduit flow. An interpretation of the chemograph supports the proposed model.

TIME SERIES OF PHYSICO-CHEMICAL PARAMETERS OF THE GILAN AQUIFER

Detailed information on the response of the Gilan karst aquifer can be obtained from the time series of physico-chemical parameters (Fig.5). The first rains at the beginning of the hydrological year do not increase the spring discharge significantly. The water stored in the subcutaneous zone and soil cover is at its minimum level at the beginning of the hydrological year; therefore the first rains are mostly stored in this zone. Later rains, for example the 52mm rainfall at the end of March 2001 and 82mm rainfall during 5 continuous rainy days during April 2002, were able to increase the discharge of the Gilan Spring to 200 and 150 l/s respectively (A and B in Fig.3). The time delays between the rainfall and the observed peak discharge were approximately 13 and 20 days respectively. The first rainfall was mostly stored in the subcutaneous zone. Later rainfall could then quickly push previously stored water towards the main fractures or probable hidden shafts that convey water vertically to the water table or push stored water towards the water table by piston flow. The rise in the water table can consequently increase the spring discharge, most probably by piston flow. The discharge increase is less than 25% and the ratio of maximum to minimum is about 1.5. These low values imply that the peak discharge occurs when only a part of the spring discharge is contributed by aquifer conduit flow. This part is expected to be located near the spring emergence. Therefore the reasons that the maximum discharge does not occur by contributions from more distant parts of the aquifer are as follows:

1. The routing decreases the discharge from more distant parts of the aquifer.

2. Near the Gilan Spring, recharged water reaches the water table faster than it does in areas farther away from the spring. This is caused by the vadose zone being so thin that the water table is

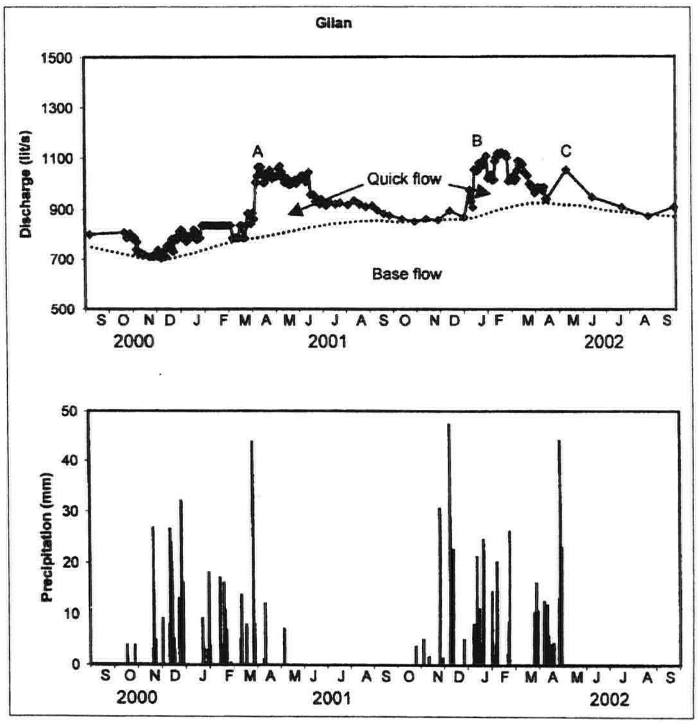
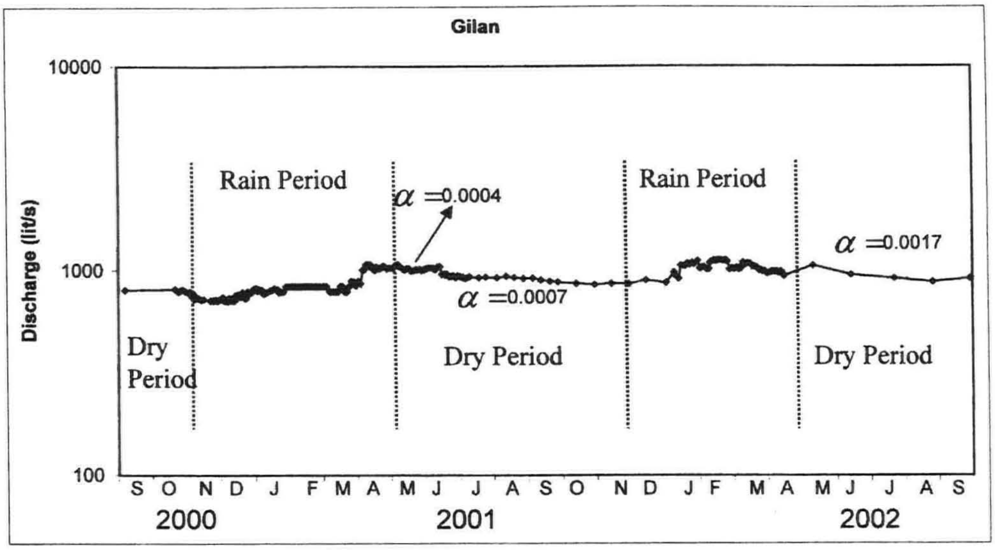


Figure 3. Hydrograph of the Gilan Spring during the study period.

Figure 4. Recession curves of the Gilan Spring.



located inside or near the epikarst zone and karstification is more developed near the surface.

3. The length of the southern flank is only 6km. Therefore the contribution of this flank to the spring discharge occurs within a short period. The peak discharge takes places when both the southern flank and the equivalent northern flank provide spring water simultaneously. When the fast flow from the southern flank and the equivalent area of the northern flank terminate, fast flow from the more distant parts of the northern flank continues to supply spring water, reducing the total discharge.

There is no rainfall in late spring and summer, consequently the spring discharge follows a decreasing trend. The EC is constant for a short period after the first rain because most of the water is stored in the unsaturated zone. The EC increases gradually and reaches a peak value, at which a small increase is observed in the discharge of Gilan Spring. This means that the area near the spring mostly contributes to discharge enhancement. The variations of EC in the Gilan Spring can be explained as follows:

1. The increase in EC at the start of precipitation is probably due to the flushing of phreatic water with a long residence time (Ashton, 1966; Bakalowicz *et al.,* 1974; Ford and Williams, 1989; Sauter, 1992).

2. For the following reasons, the increase of EC at the beginning of rainfall can be explained by the displacement of more highly mineralized water within the subcutaneous zone (Williams, 1983; Sauter, 1992), especially near the Gilan Spring:

a) The Gachsaran Formation crops out or is buried beneath thin alluvium at the foot of the northern flank and part of the southern flank near the Gilan Spring. The Gilan Spring emerges from the plunge of the Saravan anticline, therefore the dip of bedding planes and consequently the thickness of the Gachsaran Formation are less than in areas farther away from the spring. The area of the Gachsaran Formation having a small thickness is larger near the Gilan Spring than the area farther away from the spring, increasing the seepage of the rainwater to the underlying Asmari Formation. The large thickness of the Gachsaran Formation prevents the seepage of rainwater to the underlying Asmari Formation in the area away from the spring. Rainwater stored in the Gachsaran Formation from the previous year has high values of dissolved solids, especially sulphate ions. The first rains flush this stored water to the underlying Asmari Formation through the thin Gachsaran Formation.

b) The percentage of soil cover on the Asmari Formation is higher and the elevation is lower in the plunge area compared to the area farther away from the spring and, therefore, evaporation is higher in this area. Rainwater stored on this soil from the previous year evaporates from the soil surface, enhancing the total dissolved solids. The accumulated salts are flushed out by the first rains, increasing the EC of the Gilan Spring. In the area near the spring the water table is near the surface and within the subcutaneous zone. Thus, it takes a short time for stored water in the Gachsaran Formation or the soil cover to reach the large conduit near the Gilan Spring. This confirms the priority of peak EC to peak discharge.

The EC decreases smoothly during the wet period with no specific minimum, even during periods of heavier rain and at peak discharge. Parts of the rainwater from distant parts of the catchment area contribute to the aquifer as conduit flow. It takes at least seven months for conduit flow from the farthest parts of the aquifer to reach the Gilan Spring. The water quality of this conduit flow is expected to be good. The diffuse flow from the vadose zone joins the main conduit, which transports the water to the Gilan Spring. As discussed previously, water quality of the diffuse flow from the farther parts of the aquifer is better than that in the vicinity of the spring. In addition, as the rain continues, the shares of low quality vadose and phreatic flushing water are reduced.

The time series of sulphate, and more or less calcium and magnesium, approximately coincide with the time series of EC. This is because of the existence of gypsum in the evaporites within the soil and also within the Gachsaran Formation on the periphery of the aquifer.

Water temperature is relatively constant and has no distinct trend. Standard deviation and coefficient of variation of temperature are low. Thus, the water has enough time to reach thermal equilibrium within the karst system.

Calcite and dolomite saturation indices are oversaturated in most of the samples, indicating the chemical equilibrium of water within the aquifer. These indices are undersaturated during the receding of discharge and decrease of EC, implying that the new conduit waters from remote areas have insufficient time to reach equilibrium with the karst system or are flowing through open joints and conduits.

CONCLUSIONS

The base flow separation method reveals that 90% of the Gilan Spring output is base flow (Fig.3). The high percentages of base flow apparently implies that the flow is mainly diffuse. The recession coefficients, the percentage of base and quick flow, the standard deviation of temperature and the hydrochemical data of the Gilan Spring imply that the Gilan aquifer is mainly dominated by diffuse flow. The EC time series of the Gilan Spring reveals that the conventional method is not applicable for determining base and quick flow for a spring with the characteristics such as displayed by the Gilan Spring. The EC has an increasing trend in most of the karst springs in the dry period, but decreases gradually during the dry period, indicating that conduit flow continues to the spring discharge. The main reason is that the catchment area is long and narrow, such that it takes a long time for more distant conduit flow to reach the spring. Rain duration is significantly shorter than the lag time of the karst spring. Recharge water near the Gilan Spring flows out of the system before recharge water from more distant parts of the aquifer has reached the spring.

Peak discharge is provided mainly by conduit flow within parts of the aquifer in the vicinity of the Gilan Spring. Conduit flows from more distant parts of the aquifer gradually reach the spring during the recession period only after the peak has been reached. It may be concluded that part of the supposed base flow, estimated erroneously by the conventional method, is in fact conduit (quick) flow.

The recession coefficient is in the range of diffuse flow, in spite of the contribution of conduit flow during the recession period. The main reason is that when the hydrograph start its recession, the arrival of conduit flow from remote areas prevents a quick recession. The term "pseudo-diffuse flow" is proposed for this type of flow regime. The variations of physico-chemical parameters are damped out due to the contribution of conduit flow during the recession period.

This type of behavior may occur in aquifers with the following characteristics:

1. The catchment area of the spring is long and narrow.
2. The rain duration is significantly shorter than the lag time of the spring.
3. The EC has a decreasing trend during the dry period and conduit flow contributes to the spring flow as long as an increase in EC is not occurred.

In such conditions, presentation of lag time is ambiguous.

ACKNOWLEDGEMENTS

The authors thank the Research Council of Shiraz University for their financial support. We especially thank Professor Klaus-Peter Seiler of the National Reasearch Center for Environment and Health (GSF) for Oxygen 18 and Deuterium measurements.

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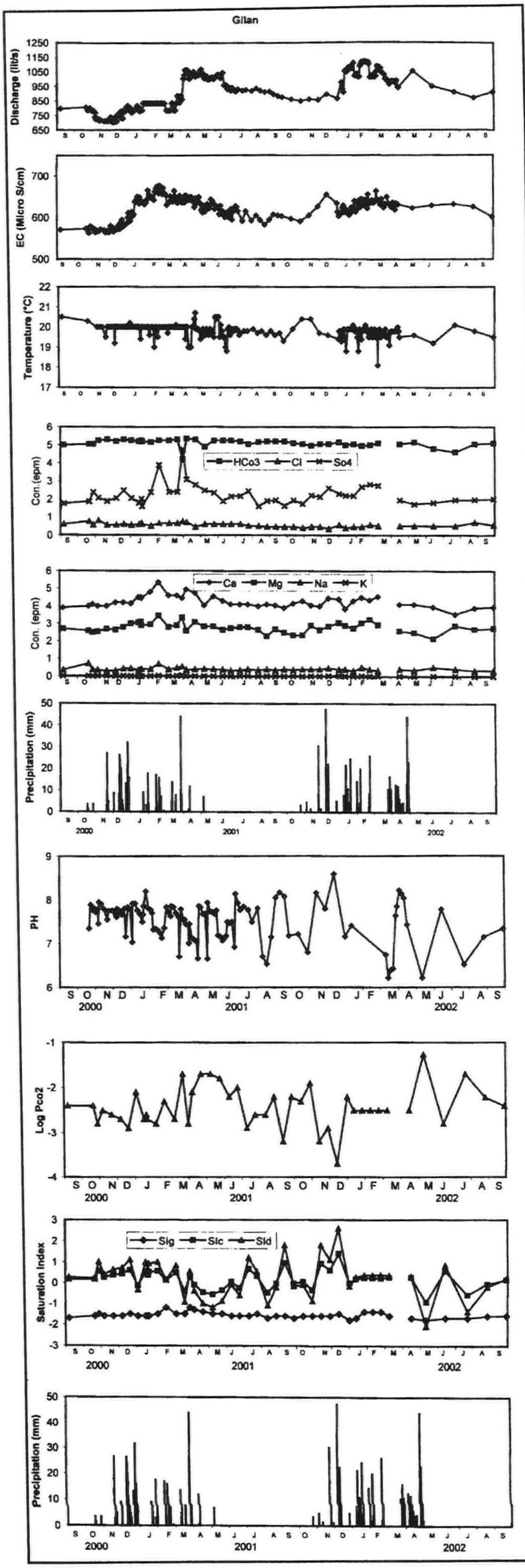

Figure 5. Time series variations of the Gilan Spring parameters.

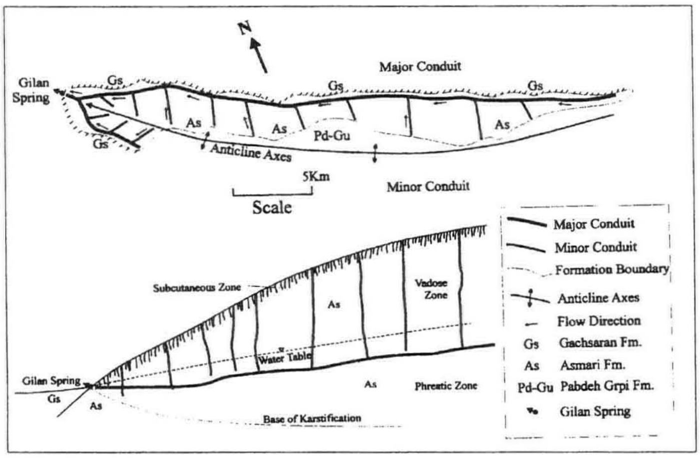

***Figure 6.** Schematic model of groundwater flow in the Gilan aquifer.*

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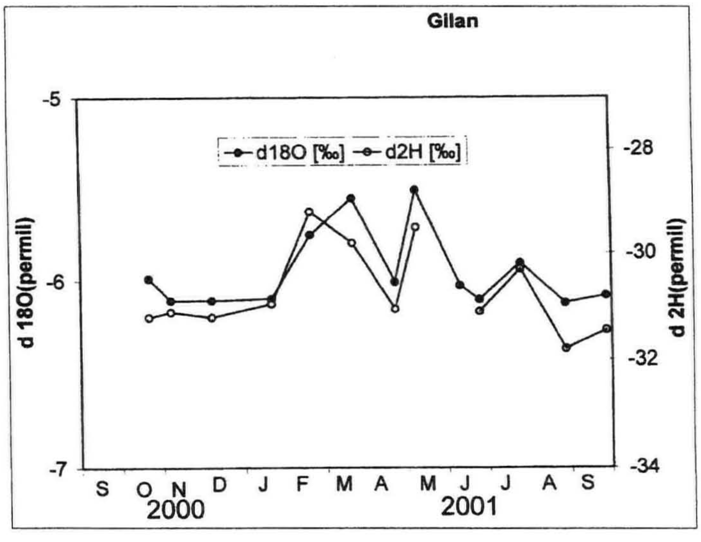

***Figure 7.** Time series variations of environmental isotopes in the Gilan Spring.*

CAVE AND KARST SCIENCE Vol. 30, No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Speleothem-like calcite and aragonite deposits on a tropical carbonate coast

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Abstract: Vadose precipitation of calcite and aragonite can occur in marine notches and similar coastal overhangs in the tropics. The resultant deposits are reminiscent of speleothems and have been termed "littoral dripstone" and "littoral flowstone." They lack the luster and crystallinity of cave analogues, and are composed of highly porous layered microcrystalline calcite and aragonite, commonly containing inclusions of calcareous grains of marine origin. These deposits appear to be a distinct coastal variety of tufaceous stalactites reported from entrances of tropical caves. Aside from providing valuable insights into the previously unrecognized carbonate precipitation in the modern supratidal zone on tropical coasts, these deposits are of practical interest to geomorphologists. If not recognized as not being true speleothems, they can cause misinterpretation of wave cut or bioeroded landforms as remnants of solution cavities.

Key Words: marine notch, littoral, intertidal, supratidal, flowstone, dripstone

(Received 15th September 2003; Accepted 5th November 2003)

INTRODUCTION

Speleothems are secondary cave deposits (Jennings, 1985). Their formation requires a humid, enclosed cave atmosphere and typically does not occur under normal surface conditions, where it is limited by evaporative processes (Hill and Forti, 1997). Speleothems can, however, be found at the land surface due to breakdown of caves, and in such cases they represent evidence of former dissolution voids. This is particularly common in littoral areas, where karst caves are breached by erosion and incorporated into the modern coastline (Mylroie and Carew, 1995). In such settings, the presence of speleothems is an important factor in identifying former cave walls integrated into coastal scarps; distinguishing between littoral (erosional) caves and true karst (dissolution) caves; and even differentiating between marine (bioerosional) notches and breached dissolution cavities (Mylroie and Carew, 1991).

There are, however, secondary carbonate precipitates that form in the modern subaerial littoral environment. On the coast of Tinian, Mariana Islands, calcite and aragonite deposits reminiscent of speleothems are not uncommon in marine notches, coastal scarps and overhangs, and littoral caves. Shaped as draperies, irregular patches, and bulbous stalactitic formations attached to bedrock overhangs and walls, these coastal, non-spelean, vadose precipitates are apparently associated with groundwater seepage and sea spray. They have not been reported previously. However, analogous "lumpy porous stalactitic features" (Mylroie and Carew, 1991), and aragonite glazes, a few mm thin sheet deposit on coastal rocks thought to precipitate from evaporating sea spray and groundwater seeps, have been observed in the Bahamas (A C Neumann, pers. comm; J Wilber, pers. comm.). These deposits also appear to be related to, and may represent a distinct coastal variety of, subaerial tufa deposits commonly observed growing at the entrances of caves, or plastered on cliffs, in humid tropical karst areas (Viles and Goudie, 1990).

Aside from providing valuable insights into the previously unrecognized carbonate precipitation in the modern supratidal zone on tropical coasts, these deposits are of practical interest to geomorphologists and karst hydrologists. If not recognized as not being true speleothems, they can cause misinterpretation of wave cut or bioeroded landforms as remnants of dissolution cavities.

The purpose of this paper is to expand on the first field report of these coastal speleothem-like precipitates from Tinian, Mariana Islands (Taboroši and Stafford, *in press*), further document their morphology and depositional environments, and present petrological and mineralogical data corroborating their formation outside of caves.

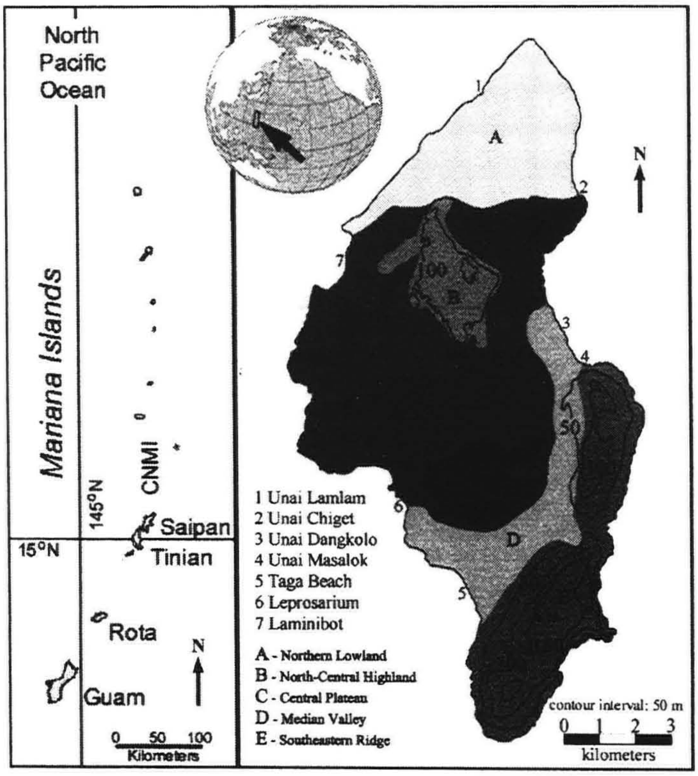

Figure 1. Location and simplified physiographical map of Tinian, showing sites mentioned in text. Physiographical regions are adapted from Gingerich and Yeatts (2000).

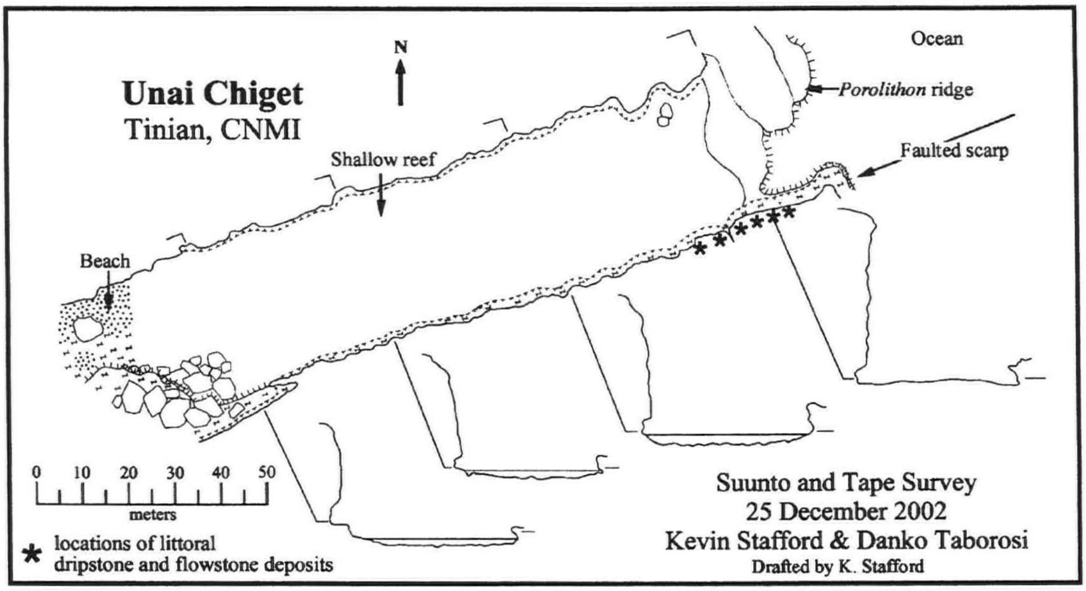

Figure 2. Plan and selected profiles of Unai Chiget inlet. Note the highly localized incidence of speleothem-like secondary deposits in the marine notch.

STUDY AREA

Tinian, a tropical carbonate island, lies in the western Pacific Ocean at 15°00′N, 145°40′W. It is the second largest (105km^2) island in the Commonwealth of the Northern Mariana Islands (CNMI), located 160km northnortheast of Guam (Fig.1). Tinian is somewhat rhomboid in shape, approximately 19km long, and up to 10km wide. Intricate systems of horizontal and tilted plateaus and terraces dominate the land surface, forming five physiographical regions (Fig.1). They are separated by steep, rocky scarps, like those that make up most of the coastline. The island is a Palaeogene volcanic edifice mantled by younger carbonate rocks. Volcanic units are exposed at only about two percent of the land surface, while the rest is covered by early Miocene Tagpochau Limestone and Plio-Pleistocene Mariana Limestone, eogenetic limestones (Vacher and Mylroie, 2002) composed mostly of coralliferous and algal lithologies. Outcrops of Mariana Limestone occur at elevations up to 175m on the southeastern ridge, which indicate that regions have been uplifted more than 100m (>0.05 mm/yr) during the late Neogene and Quaternary. Uplifted beach deposits and reefs along the coastline and elevated marine notches indicate that active tectonism is still occurring, with approximately 1.8m of uplift during the Holocene (Dickinson, 2000). The principal geological reference for Tinian is the U.S. Army report by Doan *et al.* (1960), which includes a 1:25,000 scale geological map. Current geological work on Tinian is focused on the island's karst features (Stafford *et al.*, 2003a; Stafford *et al*, 2003b).

The climate of Tinian is uniformly warm and humid, with a mean annual temperature of 27°C and a relative humidity range from 60 to 100%. The estimated annual rainfall average of 200 cm/year exhibits marked seasonality. Approximately half of this precipitation occurs during the wet season (July through October), and only about 10% falls during the dry season (February through April). Trade winds are dominant throughout the year, and annual variations in insolation and illumination are slight. Tinian has no perennial streams and thick shrub brush and limestone forests form the dominant vegetation. The mean tidal range is c.45cm and the spring tide range is c.52cm. Both the swell and waves generated by local winds generally come from the east, with the roughest seas occurring between November and February.

METHODS

Fieldwork on Tinian was carried out in December 2002 and May 2003. A total of twenty-three speleothem-like samples were removed by hammer and chisel from the marine notches at Unai Lamlam, Unai Chiget, Unai Dangkolo, and Unai Masalok (Fig.1). Similar settings at Taga Beach, the historical Leprosarium site, and Laminibot (Fig.1) were explored and found to contain no comparable deposits. Because many of the sampling locations are exposed to nearly constant high surf, much of the work could only be carried out during low tides, while exercising extreme caution. After drying, cutting, and macroscopic and binocular microscope examination, a detailed study of resin-impregnated petrographical thin sections was conducted through conventional transmitted-light microscopy. Additionally, small fragments were glued to aluminium stubs, sputter-coated with platinum, and observed with a Hitachi S-3000H Scanning Electron Microscope (SEM), under operating conditions at 20kV and 60μA. X-ray diffraction (XRD) analyses were carried out on a Bruker AXS MX-Labo powder diffractometer with Cu radiation at 40kV and 20mA. Powdered samples mounted on glass slides were rotated at 60rpm, and radiation counts were measured at stepped diffraction angles of 0.02°2θ from 2.02 to 70.00°. Peak characteristics were analyzed using MacDiff 4.2.5 software.

RESULTS

Field occurrence

Most of the coastline of Tinian is comprised of rocky scarps of Plio-Pleistocene Mariana Limestone. The bases of the scarps contain a Holocene uplifted marine notch associated with a narrow sea-level bench. In places where the ocean is not in direct contact with the tall, nearly vertical coastal cliffs, the coast is dominated by large, angular blocks, broken loose from the steep cliffs above, low terraces of extremely rough, dissolution-pitted emergent limestone, wide constructional algal benches and rimmed pools, and littoral caves and fractures, widened by wave erosion. There are a few well-developed beaches that are protected by fringing reefs, and several small beaches within rocky coves.

The distribution of speleothem-like calcite and aragonite precipitates along the coast is highly localized. They are commonly found only in the marine notches of coves and other protected areas regularly sprayed by but not directly hit by waves, and are rare in the notches in sea cliffs elsewhere. Even within such specialized settings as coastal coves, interesting distribution patterns are apparent on the metre-scale. At Unai Chiget (Fig.2), a slot-like inlet formed by the lateral retreat of a fault scarp, speleothem-like secondary deposits are numerous but restricted to a small area. The inlet extends at approximately 90° to the coastline and contains an exemplary marine notch, 2.2 to 2.5m high and 1.2 to 2.4m deep (Figs 3a; 3b). Although located in northeast Tinian, where rough seas are exceedingly common, the notch is largely isolated from wave action by its right-angle orientation to the coastline and a well-developed *Porolithon* algal ridge that breaks the waves (Fig.2). Calcite and aragonite precipitates are observed in the back wall and the roof of the marine notch, and are limited to the vicinity of the algal ridge, from the point 3m seaward to the point 17m inland of it (Fig.2). This area is regularly, but indirectly, splashed by waves. No similar deposits are found in other parts of the marine notch, both seaward of the

Porolithon ridge, where the wave impact is more direct, as well as inland, where sea spray does not reach except during storm events.

Farther south along the coast, the coves at Unai Dangkolo (Fig.4) and Unai Masalok (Fig.5a) exhibit numerous analogous precipitates. Both sites are characterized by a series of adjacent coves, formed by a locally-recessed coastal scarp and separated by narrow rocky headlands. The coves are partially protected from surf by a narrow fringing reef, and contain small carbonate sand beaches, large boulders, and beachrock deposits. The coves appear to have formed by the breaching of flank margin caves (Mylroie and Carew, 1990), and represent former dissolution cavities intersected by coastal erosion. Consequently, the seaward parts of the coves are largely indicative of modern coastal processes, whereas the more landward positions retain ample evidence of karst features (Fig.4). Because of their dual origin by encroachment of coastal processes on karst topography, the coves at Masalok and Dangkolo actually contain both true speleothems and speleothem-like modern precipitates in the immediate vicinity of each other. The former are found in inland portions of the coves and are not a contemporary part of the coastal environment, but a remnant of former caves now exposed at the coast. The latter are modern coastal deposits, unrelated to caves. They are found in the roof and back wall of the coastal notch, which is 1.8 to 3.0m and 1.0 to 4.0m in vertical and horizontal dimensions, respectively, and was produced by wave and bioerosion locally superimposed on pre-existing karst features. Interestingly, these deposits occur exclusively on the flanks of the rocky scarps separating the coves (where wave splashing is intermittent), but are absent in the headlands (exposed to surf), as well as the more inland areas (Fig.4).

The incidence of littoral calcite and aragonite precipitates is variable with regards to both horizontal and vertical positions and is apparently restricted to sites wetted by sea spray but protected from direct wave action. They are common only in coastal locations fronted by fringing reefs and beaches, where they are generally limited to uplifted Holocene marine notches and elevations of 1.8 to 3m. They have not been observed within higher Pleistocene features or at elevations lower than 1.8m above sea level. They are also absent from inland sites permanently isolated from the impact of seawater, and are only rarely observed in areas of more direct exposure to surf, where fringing reefs are lacking. One example of the latter are sea cliffs 150-300m south of Unai Masalok, where littoral calcite and aragonite precipitates are found in overhanging bedrock 4 to 6m above sea level, which is higher than elsewhere. Presumably, at this elevation, the deposits are not hit by many direct waves, while still regularly being wetted by sea spray from the surf breaking against the base of the cliff.

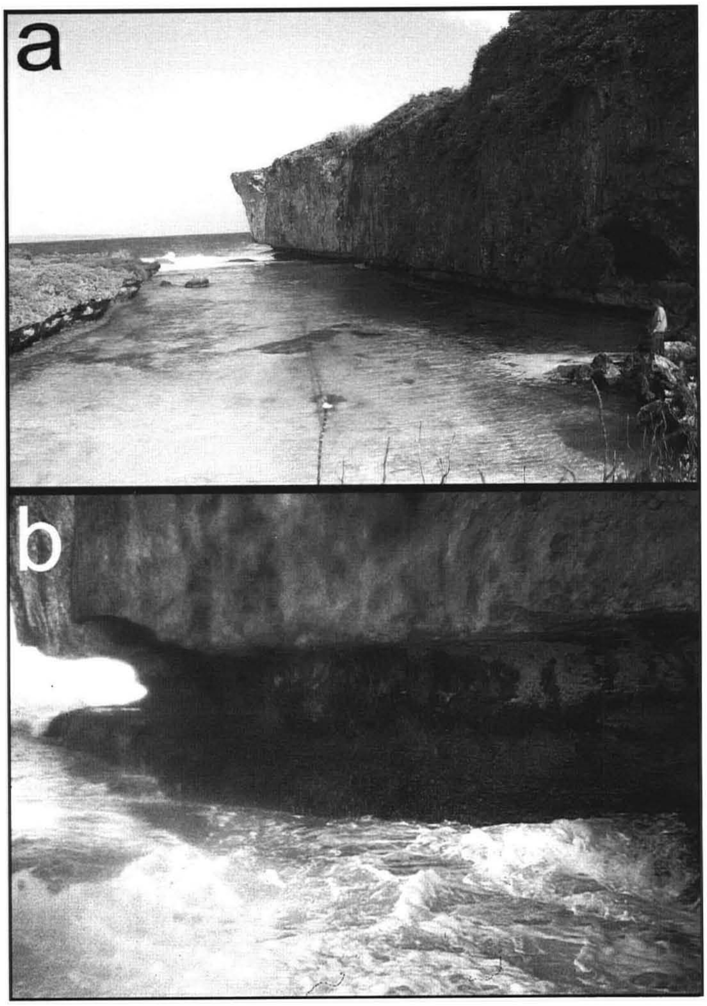

Figure 3. *Unai Chiget. a) A view of the inlet. Note the prominent marine notch and the algal ridge (breaking waves). Person for scale. b) The portion of the marine notch adjacent to the algal ridge. The notch is about 2.2 m tall.*

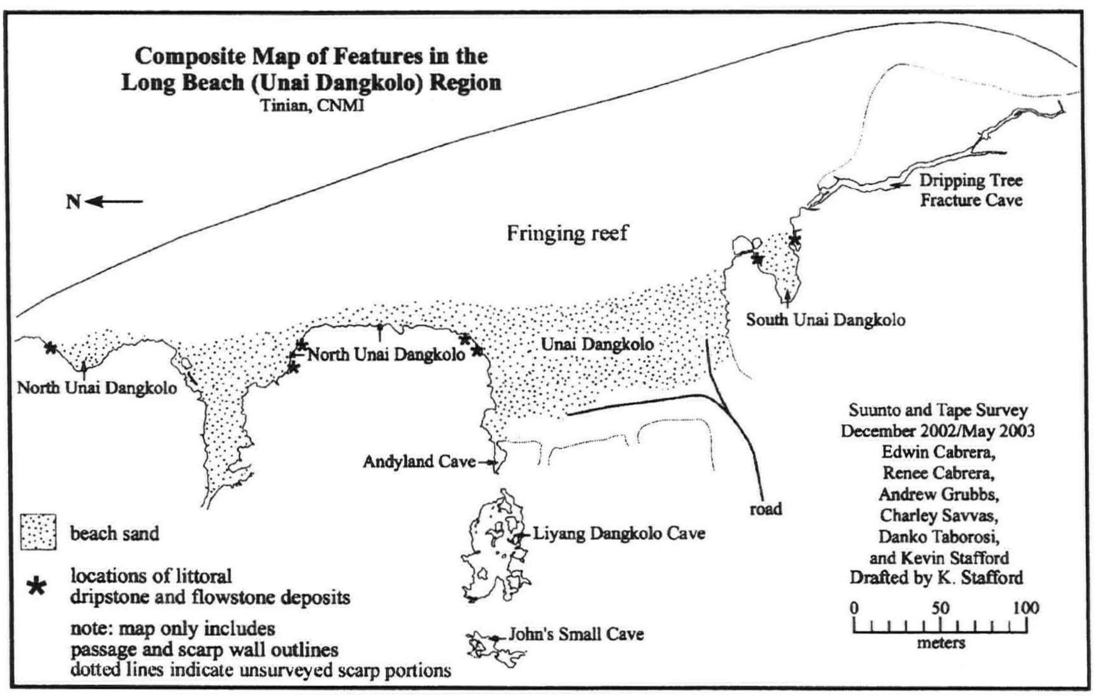

Figure 4. *Plan of Unai Dangkolo area. Note the series of adjacent coves and the localized distribution of speleothem-like secondary deposits.*

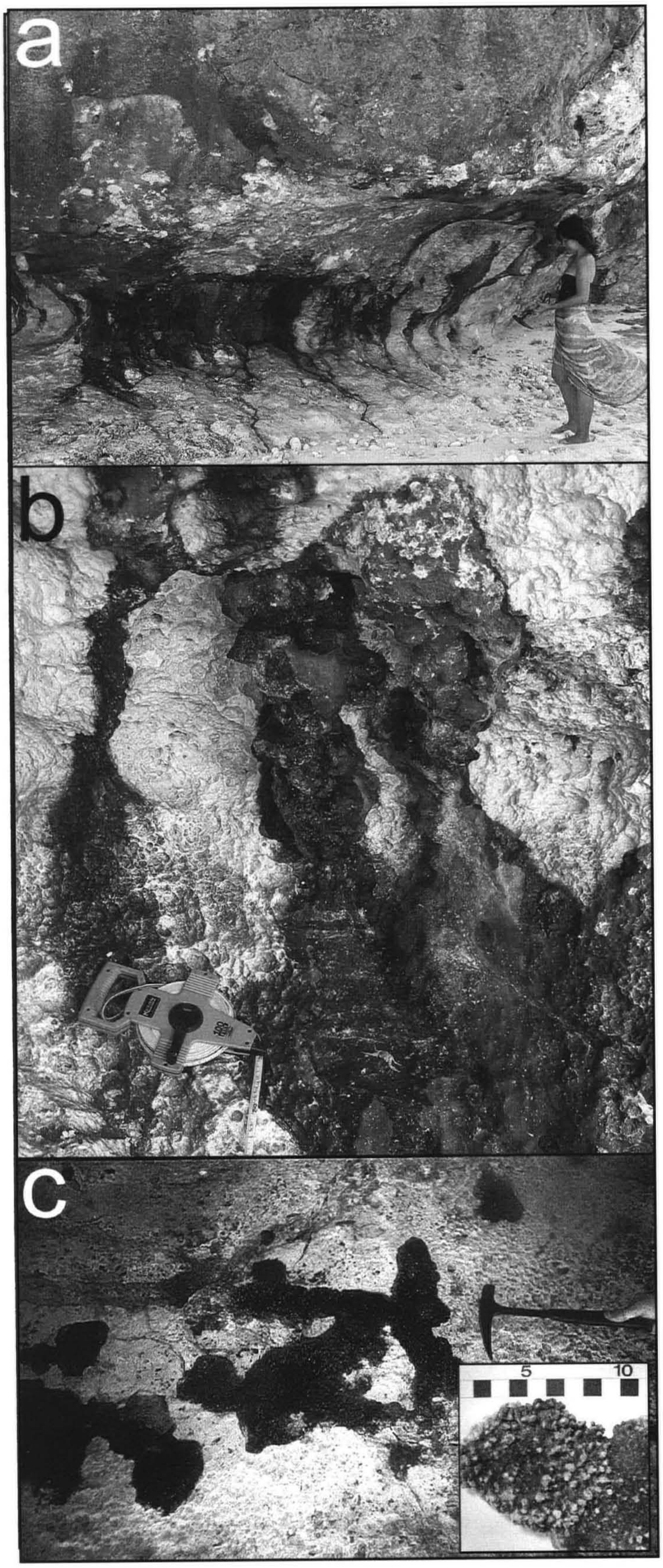

Figure 5. Deposits in situ. a) Notch circumscribing a coastal rock outcrop at Unai Masalok. Dark material is secondary deposits. Person for scale. b) Irregular draperies and bumpy stalactitic features, hanging from the roof and partly plastered to the back wall of the coastal notch at Unai Dangkolo. Tape reel (14.5 cm in diameter) for scale; photo was taken looking at the back wall of the notch. c) Patchy deposits attached to the roof of the marine notch at Unai Chiget. Hammer for scale; photo was taken looking up at the notch roof. Note the coralliform texture in an inset close-up photo (scale in centimeters; the knobby surface was oriented downward in situ).

Macromorphology

The calcite and aragonite speleothem-like precipitates found in the modern supratidal zone on Tinian are morphologically varied and distinct from true speleothems, but can be akin to tufaceous stalactites found in cave entrances in the tropics. Their overall forms reflect the basic speleothem types, such as draperies, stalactites, and cave coral, but lack the lustre and crystallinity of cave analogues, and are not nearly as well-formed or dense. They are grey to black in colour due to colonization by epilithic organisms. When broken, they reveal much paler, even white, interiors. Drapery-like deposits are similar to cave draperies, but are never quite so thin and wavy (Figs 5a; 5b). They can be very noticeable in marine notches, where they typically occur as a series of vertical deposits lining the back walls, with their dark coloration in stark contrast with the bedrock (Fig.5). Stalactitic formations are locally found hanging from the roofs of coastal notches. They lack the elegant, smooth and linear appearance of cave stalactites, and are contorted, globular and craggy, commonly partially affixed to notch walls (Fig.5b). Retaining the vertical, hanging aspect of stalactites, their maximum length of about 20cm does not greatly exceed their width, and the overall appearance is bulbous rather than cylindrical. This makes them distinct from subaerial tufa deposits from cave entrances and tropical cliffs, which are elongated, massive, and much more convincingly stalactite-like. The third common morphology is irregular, flat patches attached to notch roofs (Fig.5c). Typically several centimetres thick, they are usually covered by small knobs reminiscent of cave coral (Fig.5c – inset). This unique morphotype is the least similar to tufaceous stalactites seen on cliffs and in cave entrances in the tropics.

When hit by a hammer, most of these deposits break readily at the contact with the bedrock, and removal of a sample with a piece of bedrock can be difficult. Both active (dripping) and apparently inactive specimens have been documented, and whereas many of the active deposits are evidently fed by epikarstic water, this is not always obvious, especially in places exposed to vigorous wave action where the formations drip splashed-up seawater nearly continuously.

The layering is typically not as concentric and tight as in speleothems, but rather sinuous and irregular, commonly with patches of non-layered microcrystalline micrite and many cracks and open spaces. Many such cracks extend across deposited layers and are caused by desiccation, but some, particularly in the large, bulbous "stalactites", are primary, apparently caused by deposition of outer layers in the form of a bumpy crust (Fig.6a). Although exceedingly common, such prominent open spaces should not be taken as a diagnostic feature of littoral zone precipitates, as internal megaporosity has also been observed in speleothems from deep inside karst caves on islands like the Bahamas (J E Mylroie, pers. comm.). The high porosity is not always macroscopically apparent, however, and finely laminated deposits have also been found. The laminations can be extremely convoluted, and even polycentric. Many specimens, particularly the coralliform patches and "draperies", include numerous pisolith-like bodies, seen on cross-sections as clusters of concentric circles up to 15mm across (Fig.6b).

The sampled deposits exhibit various level of firmness, as became evident during the collection process. Most formations could easily be broken off by a single nudge of a hammer but the removal of a few required laborious cutting by a chisel. Although most specimens consist of porous and crumbly tufaceous material, a few samples were surprisingly hard and finely layered calcite, and would be indistinguishable from true cave speleothems were it not for their obvious setting superimposed on bioerosional scars and karren (Fig.7), and inclusions of marine skeletal grains (figures in next section). In such fringe cases, a careful examination of the contact with the bedrock (and the relationship to karren and bioerosional scars) and petrological observations might be necessary to avoid mistaking them for true speleothems. Fortunately, such ambiguous specimens are rare, and have most likely formed in relatively enclosed settings, such as might develop by partial filling or obstruction of marine notches and littoral caves by collapsed blocks or beach sand.

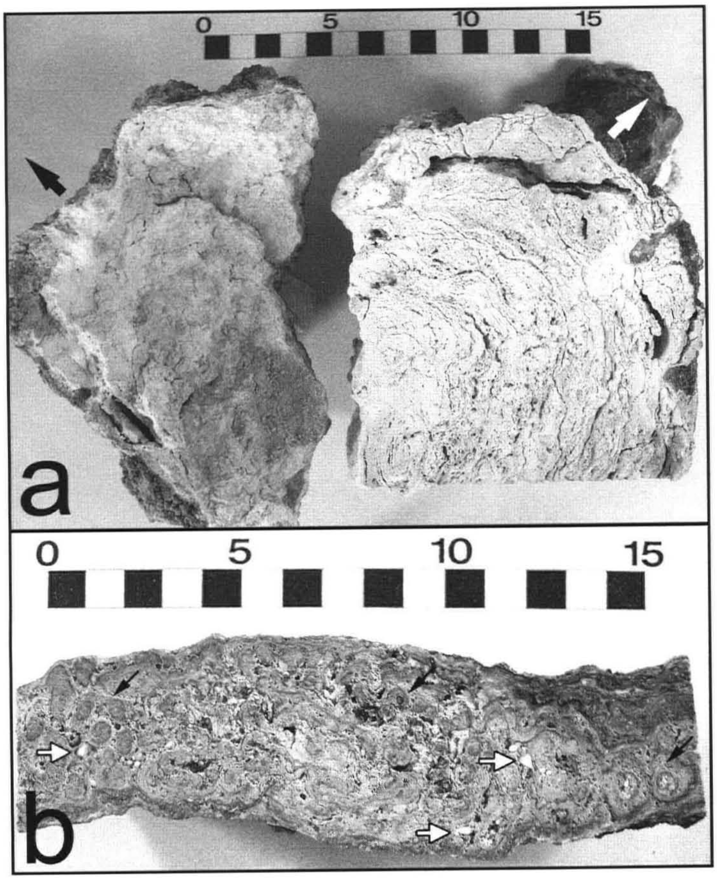

Figure 6. Cross-sections of secondary deposits from the roof of the marine notch at Unai Chiget. a) A bulbous stalactitic feature, broken on the left and cut on the right. Arrows indicate growth directions. Note the irregular banding and prominent open spaces between the layers. This specimen is from the site illustrated in Fig. 3. Scale in centimeters. b) An elongate patchy deposit. Note the convoluted layers and pisolith-like concentric circles (indicated by black arrows) and incorporated coral and mollusk shell fragments (indicated by white arrows). This specimen is from the site illustrated in Fig. 5a. Scale in centimeters.

Figure 7. An extremely hard and dense secondary deposit on the wall of a shelter cave at Unai Masalok. Note that it is prominently superimposed on bioerosional karren. Hammer for scale.

Petrology and mineralogy

Speleothem-like deposits from the modern supratidal zone are composed predominantly of layered microcrystalline calcium carbonate (micrite). Lacking the sparry calcite of true speleothems, they range in composition from pure calcite to almost pure aragonite, with most samples containing both phases in various proportions (Fig.8). Also prominent are inclusions of marine skeletal grains and various structures of organic origin.

The deposits are colonized by a great variety of microorganisms, as evidenced by numerous epilithic organisms (Fig.9a) and biofilms (Fig.9b), and ubiquitous marks of boring activity (Fig.9c). Most crystals making up the deposits are less than ten microns in size and randomly oriented, presumably due to relatively rapid precipitation (Fig.10a). Coarser crystals are notably present, but are never dominant, and include both blocky calcite (Fig.10b) and acicular aragonite (Fig.10c).

Laminations are generally conspicuous and consist of alternating layers of a dark material of apparent microbial origin and microcrystalline calcium carbonate (Fig.11a). Layers of blocky calcite and acicular aragonite can be present but are not extensive (Fig.11b). Similar sequences of micrite and coarser crystals have also been observed in beachrock (Bernier and Dalongeville, 1996). The layering is locally surprisingly regular and similar to true speleothems, but inclusions of tests of benthic foraminifera and other calcareous grains clearly indicate development in close contact with the marine environment (Fig.11d). Conversely, some deposits are made of a microcrystalline aggregate so highly porous and unorganized that layering is barely perceptible (Fig.11e). Evidence of constructional microbial activity is visible in many deposits (Figs 11a; b; c).

Coarse crystals, when present, most commonly grow within pore spaces, gradually reducing the original high porosity of the deposits. Pores can be occluded completely or partially (Fig.11f) and are sometimes filled by both calcite and aragonite, deposited in remarkable succession (Fig.11g). The relationship between crystals and micrite is not everywhere clear and gradual, and various size crystals of blocky calcite are locally found surrounded by microcrystalline matrix (Fig.11h). This seems to indicate that larger crystals can decay and break down to a size unresolvable by

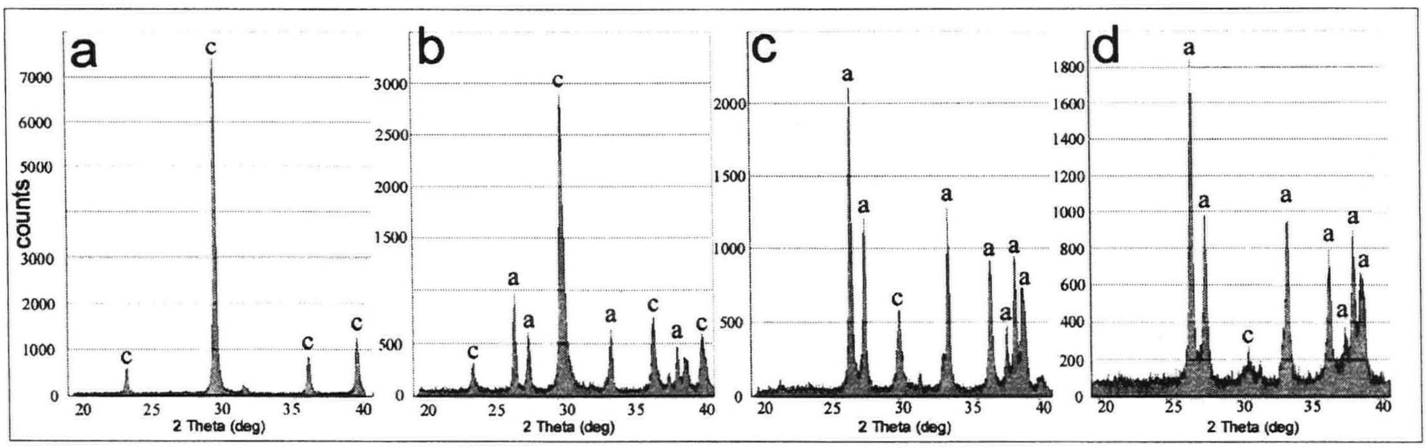

Figure 8. Series of X-ray diffractograms of powdered samples from Unai Chiget (only angles around the main peak of calcite are shown): a) A drapery-like deposit composed of pure calcite; b) A stalactitic deposit composed of mostly calcite with some aragonite; c) A patchy deposit composed of mostly aragonite with some calcite; d) A coralliform patch deposit composed of almost pure aragonite.

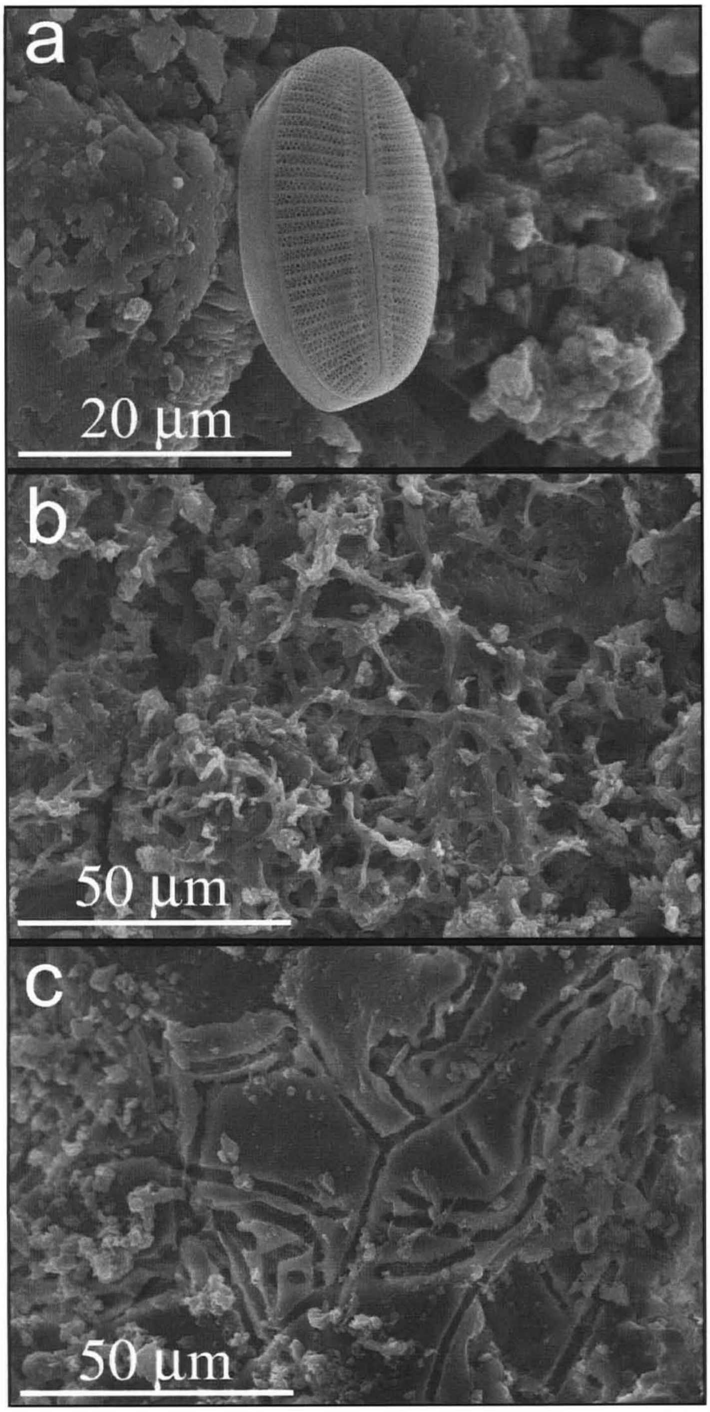

Figure 9. SEM micrographs of biological structures and traces. a) A frequently encountered epilithic diatom frustule. b) Substrate covered with a complex microbial coating. c) Common traces of microbial boring.

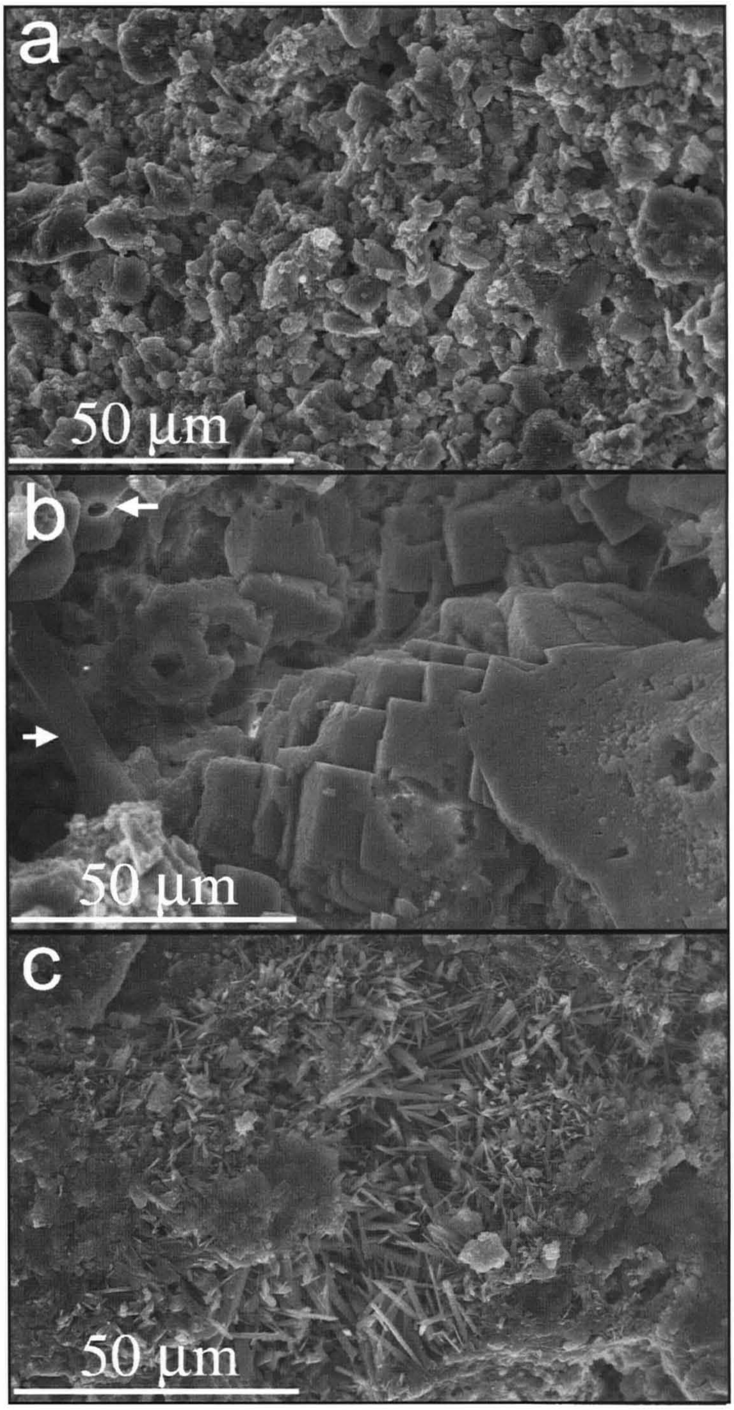

Figure 10. SEM micrographs of three types of crystal morphologies encountered. a) Microcrystalline calcite forming the bulk of all examined samples. Note the highly variable size of the crystallites, and lack of any organization. b) Blocky crystals of calcite. Note the organic rod-like structure and a bore hole (indicated by arrows). c) Acicular crystals of aragonite. Note the well developed crystal form of individual aragonite needles growing towards a pore space in the microcrystalline matrix.

petrographical microscopy. The extent of this phenomenon appears to be minimal, however, and there is no evidence that much of the observed microcrystalline material could have formed by decay of coarsely-crystalline microfabrics. Even in places where calcite crystals seem to be transitioning to a microcrystalline mass, there is evidence that they originally formed within pore spaces (Fig.11i).

Various pre-existing particles embedded in the precipitated material are exceedingly common, but have not been found in all samples. The grains are generally skeletal fragments of coral, calcareous algae, and molluscs, and whole tests of benthic foraminifera. They vary widely in incidence, and can appear as isolated single grains floating in the precipitated matrix (Fig. 11e) or large groups of grains cemented together. The latter are sometimes cemented by an isopachous layer of acicular aragonite, as in beachrock (Fig.11j), but are more commonly held together by microcrystalline material (Fig.11k).

In contrast, true cave speleothems exposed in coastal environments by erosion and collapse of caves lack the features discussed above (except evidence of bioerosion), and are uniformly composed of sparry calcite (Fig.11l).

DISCUSSION

The deposition of $CaCO_3$ inside caves is usually governed by the outgassing of CO_2 from vadose water, which typically carries a high load of dissolved Ca^{2+} and HCO_3^-. This increases the local pH, causes the removal of carbonate ion species from solution, and precipitates solids that form speleothems (Dreybrodt, 1988). Correspondingly, epikarstic water, dripping from the roofs of cave

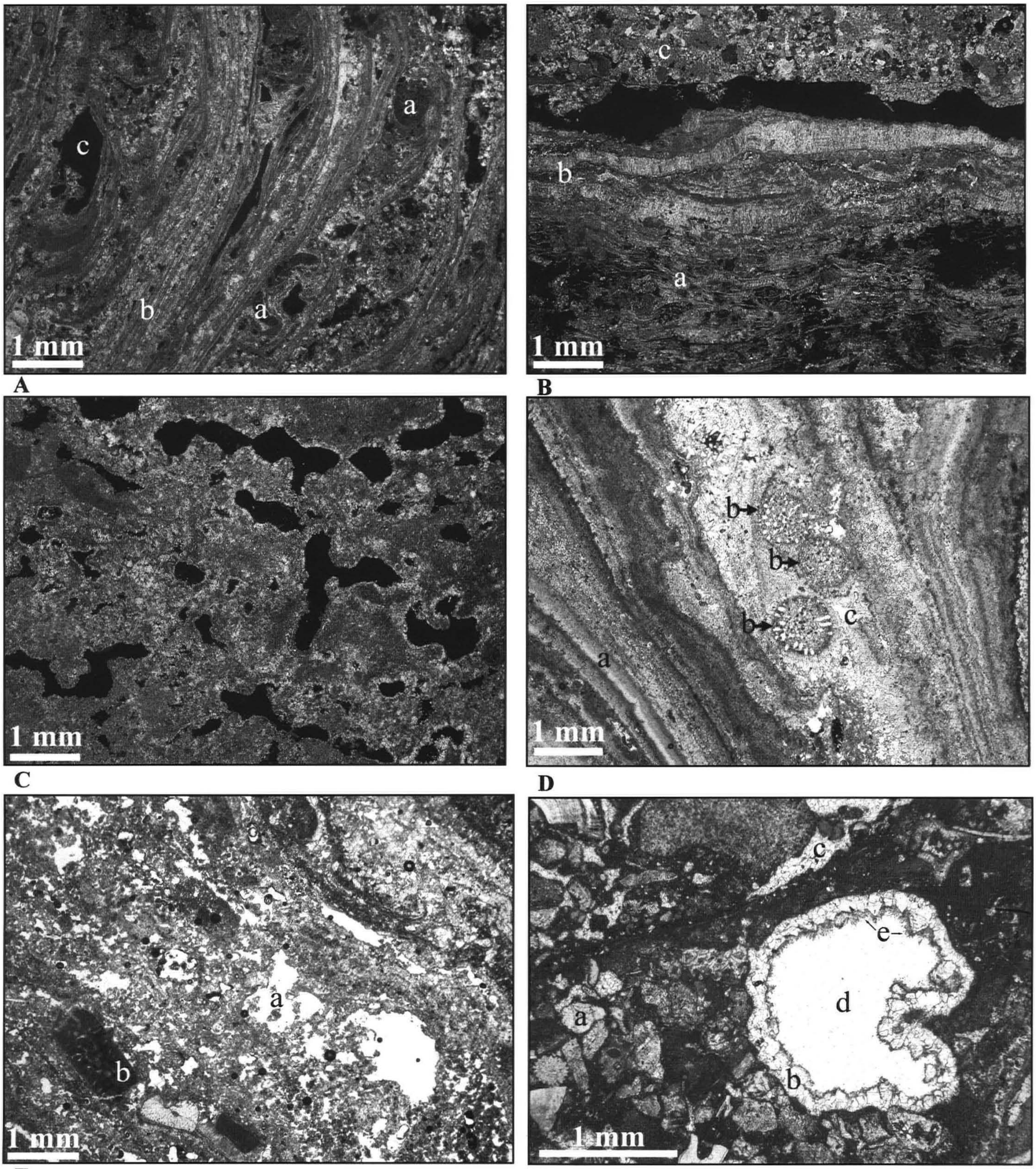

Figure 11. (Part 1) *Thin section photo micrographs of littoral dripstone and flowstone deposits (a-k) and a true speleothem exposed in a coastal setting (l), viewed in cross-polarized light (a, b, c, g, i, l) and plain light (d, e, f, h, j, k).*

a) Cross-section a few millimeters below the outside surface of a patchy deposit (shown in Fig.6b). Note the pronounced, but irregular, layering convoluting around several pre-existing grains (a); the alternation of dark, organic material and lighter, microcrystalline layers (b); prominent empty pore spaces (c); and the lack of any distinct crystals.

b) Cross-section of a coralliform patch deposit (shown in Fig.5c). Note the irregular, largely discontinuous layering of dark organic material and micrite (a); two pronounced layers of columnar crystals of aragonite extending across the micrograph (b); and non-layered crystals of calcite interspersed with micrite in the top of the micrograph (c).

c) Transverse section of a coralliform patch deposit (similar to the one shown in Fig.5c). Note the striking constructional pattern of the deposit and the dominance of micrite of organic origin (dark material); crystals of calcite interspersed within the micritic matrix (light); and isopachous layers of acicular aragonite lining pore spaces (light linings around some voids).

d) Cross-section of a hard flowstone from the walls of a coastal shelter cave (shown in Fig.7). Note the dense layering characteristic of true speleothems (a); the tests of benthic foraminifera encased between the layers (b); and infilling of pore space by micrite and an isopachous layer of acicular aragonite growing radially from the foraminiferal tests (c).

e) Cross-section of a drapery-like deposit from the wall of a marine notch (from the site shown in Fig.5a). Note the extremely porous and unorganized nature of the deposit (a); and pre-existing skeletal grains embedded in microcrystalline matrix (b).

f) Calcareous grains (a) cemented by microcrystalline matrix, and crystals of calcite partially (b) and completely (c) infilling pore spaces (d). Note the layer of acicular aragonite growing on calcite crystals (e).

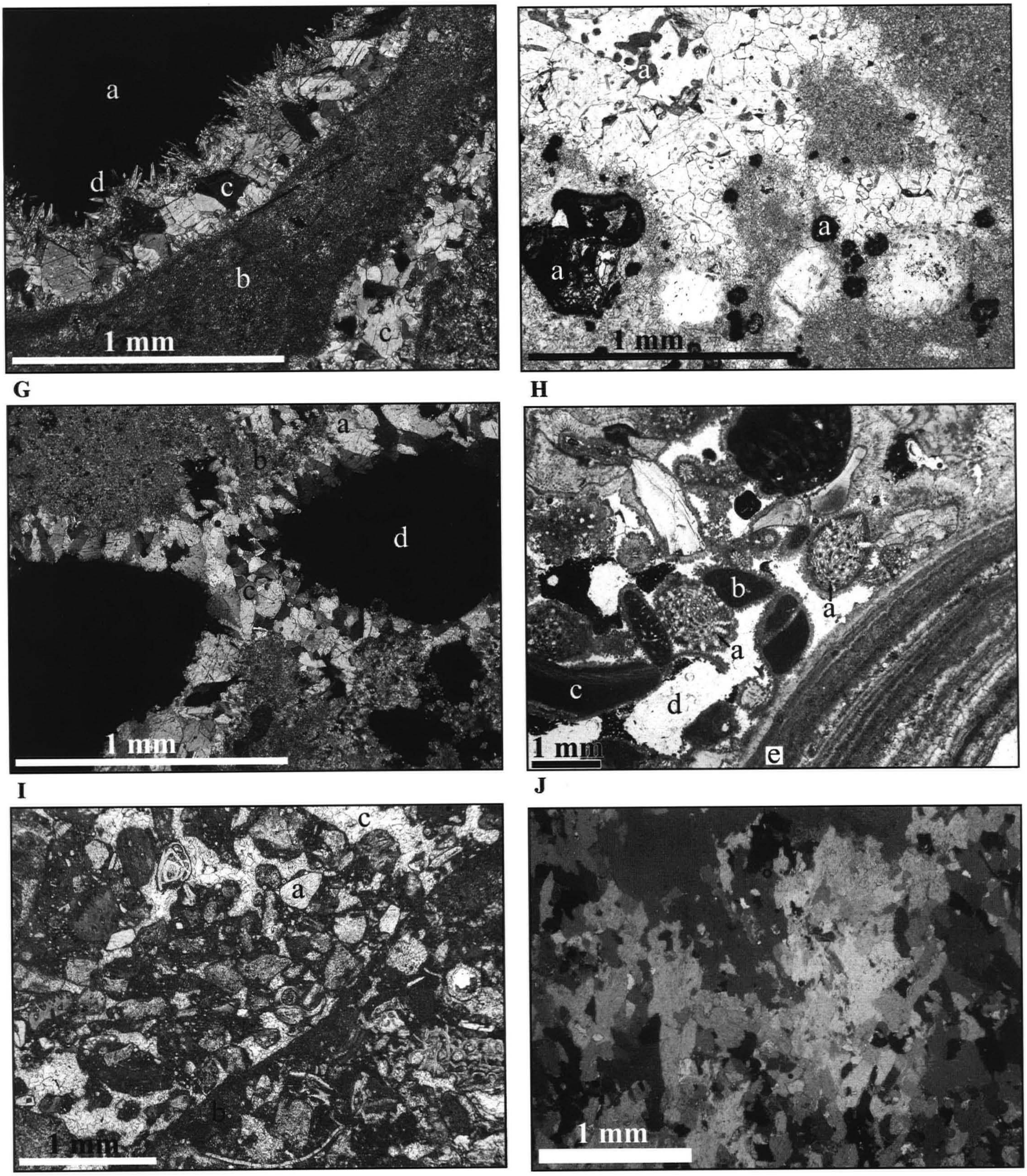

Figure 11. (Part 2)

g) Crystals infilling pore space (a) in microcrystalline matrix (b). Note the lining of pore space by equant crystals of calcite (c), followed by a layer of acicular aragonite (d).

h) Interspersed crystals and microcrystalline calcite (in a specimen similar to those shown in Fig.5c). Note the prominent dark bodies (a), no doubt of organic origin.

i) Equant crystals of calcite (a), which appear to be breaking down into microcrystalline mass (b) (sample shown in Fig.6a). Note the meniscus form (c), indicating that calcite grew filling pore spaces (d) under vadose conditions.

j) Littoral flowstone shown in Fig.7. Note the various calcareous grains, benthic foraminifera (a), fragments of calcareous algae (b), bivalve fragments (c), etc., cemented in a beachrock-like fashion, mostly by an isopachous layer of acicular crystals of aragonite. Note the very incomplete filling of pore spaces (d). Also note the relationship between the cemented material and the precipitated, layered calcite (e). This striking arrangement is a result of a littoral flowstone deposit coming into brief contact with beach sand, which was incorporated.

k) Skeletal grains of marine origin (a) cemented by micrite (b). Note the complete infilling of pore space by calcite crystals (c), and the visually striking contrast and sharp contact between the top left and the bottom right parts, characterized by different amounts of dark organic material.

l) Microfabric of a true cave speleothem exposed near Unai Dangkolo by breaching of its host cave. Note the uniform structure composed of sparry calcite and the lack of pore spaces.

entrances and relatively open, non-spelean, but overhung environments, is in equilibrium with the high partial pressure of CO_2 in the soils above. As it drips, it is not surprising that some degassing of CO_2 and precipitation of $CaCO_3$ should occur. However, while the enclosed atmosphere of caves exhibits minimal evaporation due to high humidity, the evaporation effects in the more open settings are much greater. This induces rapid deposition of calcite, causing a more random orientation and smaller size of crystals than those precipitated inside caves. This phenomenon is commonly observed in springs and seeps, as well as cave entrances and cliff faces in the tropics, where calcareous tufa is deposited, commonly in the form of crumbly and porous tufaceous stalactites deflected toward the direction of light (Viles and Goudie, 1990).

On tropical coasts, interplay of additional climatic, hydrological and biological factors impresses a unique character on these deposits, resulting in distinct, littoral varieties of subaerial tufas known from elsewhere. Whereas the macromorphology of deposits discussed in this paper can be comparable to other speleothem-like tufas in the tropics, their globular and coralliform morphology, very dark to black colour, small size, and lack of any light orientation generally sets these littoral deposits apart from the massive stalactitic tufas characteristic of inland settings. It should be noted that the latter can occur on coastal cliffs in certain locations (e.g. Krabi, Thailand), but have never been reported from sites regularly affected by waves and sea spray. The lack of any detailed mineralogical and petrological studies on tropical tufaceous stalactites makes any further comparison difficult, but the authors consider the following additional characteristics to be unique to littoral zone deposits:

- presence of aragonite in addition to calcite;
- inclusions of marine skeletal grains in addition to precipitated material;
- incidence of convoluted layers and pisolith-like bodies; and
- prevalence of intertidal epilithic biofilms instead of lichens, mosses, and higher plants characteristic of inland tufas.

In the light of their unique depositional environment, morphology, and peculiarities outlined above, the authors consider these littoral deposits to be a distinct type of subaerial tufa. They are related to speleothems, and it is proposed that the terms "littoral dripstone" and "littoral flowstone" be used to refer to these precipitates. The terms "dripstone" and "flowstone" are generic terms for the secondary chemical deposits that form by dripping (e.g. stalactites) or flowing (e.g. draperies) vadose water (Jennings, 1985), and are suggested here to draw a parallel with speleothems, whereas the adjective "littoral" is used to evoke the unique depositional environment and make a clear distinction from the archetypal and genuine dripstone and flowstone that form inside caves. Such terminology would also encompass the coastal aragonitic glazes previously recognized in the Bahamas (A C Neumann, pers. comm.; J Wilber, pers. comm.), since precipitated wall coatings have already been classified as forms of dripstone and flowstone (White, 1988).

The exact precipitation mechanisms of these deposits are not known and are the subject of an ongoing study. The authors consider the mechanisms to be similar to those involved in the formation of tufaceous stalactites common in tropical karst areas. They are thought to be precipitated from vadose water dripping in sheltered sites, such as cave entrances and cliff faces, where evaporation effects outweigh the CO_2 degassing (Viles, 1988), although no studies have yet confirmed this. In the deposition of littoral flowstone and littoral dripstone, seawater seems to play an additional role, as is apparent from the presence of acicular aragonite and the horizontal and vertical distribution patterns of the deposits. They are abundant in locations that are regularly, but indirectly splashed by waves; but are absent in places exposed to more direct surf, as well as areas completely protected from the surf. A likely situation involves sea spray entering fractures in the rock and mixing with downward percolating vadose meteoric water within the overhanging rock ledge. Undersaturation due to mixing dissolution could cause the calcite to become mobilized, resulting in its precipitation when the water comes out on the underside of the overhang, where it both degasses and evaporates. Thus, the proportion of fresh to seawater controls the mineralogy. Geochemistry is expected to vary with surf and weather conditions, and seems to be reflected in uneven, sometimes sequential deposition of calcite and aragonite. In addition, the effects of living organisms, considered crucial in many entrance zone tufaceous speleothems (Forti, 2001), appear to be extensive but not instrumental in the formation of littoral dripstone and flowstone. At least some of the many biologically-mediated mechanisms suggested in the precipitation of carbonate cements in beachrock (see Gischler, 2003 for review) could be involved. Growth of photosynthetic organisms, however, thought to cause "phototropism" in many tufaceous stalactites, appears less relevant in this case.

Finally, it is important to note that these deposits are somewhat ephemeral. Because they lack sturdiness, they could be removed on a routine basis when typhoons turn their depositional environments into high wave energy localities. Growing back quickly and being episodically stripped away by major storm events is compatible with, firstly their location in such specialized littoral settings (suggesting they do not survive long in energy rich environments), secondly the lack of many big accretions (suggesting a short life span), and, thirdly characteristic "scars" observed on bedrock in the vicinity of extant deposits (indicating locations of former broken specimens). Despite being newly recognized and somewhat unusual deposits, the authors believe that future research in tropical coastal karst areas will reveal that they are more common than it may appear.

CONCLUSION

"Littoral dripstone" and "littoral flowstone" are hereby defined as secondary chemical deposits precipitated subaerially in the supratidal zone. They can be reminiscent of speleothems, but are not nearly as well-formed, crystalline, and dense. They also show similarities to tufaceous stalactites commonly seen on cliffs and in cave entrances in the tropics, and may be a variety thereof, modified by unique climatic, hydrological and biological factors operative in their supratidal depositional environment.

Ranging from tufaceous to more firm deposits, they are generally composed of microcrystalline calcite, with various amounts of blocky calcite, acicular aragonite, organic material and skeletal grains of marine origin. Their morphology, dark colour and texture, and their position superimposed on karren and bioerosional etching, make them easily recognizable as secondary deposits and not a part of the bedrock. In most cases, they readily break at the contact when hit by a hammer.

Because true speleothems, remnants of dissolution voids breached by coastal erosion, are also commonly present on tropical coasts, care must be taken not to confuse them with the littoral dripstone and flowstone deposits. The distinction between the two is crucial, because true speleothems are inactive remnants and indicators of karst cave paleoenvironments, whereas littoral dripstone and flowstone are actively growing deposits and contemporary parts of the modern coastal environment. This explains why the former are generally etched and eroded, whereas the latter are not, despite being exposed in the same general settings.

The depositional mechanisms of these littoral vadose precipitates are presently under active study. Future research will focus on geochemical work, aimed at understanding the dripwater chemistry of littoral dripstone and flowstone deposits and comparing them with those of genuine speleothems and of tufaceous stalactites from tropical cliffs and cave entrances.

ACKNOWLEDGEMENTS

The authors are grateful to Dr John Jenson and Dr John Mylroie for the opportunity to work in Tinian. Fieldwork was funded by the US Geological Survey, through the National Institutes for Water

Resources Research program, award No. 01HQGR0134. Assistance from the Tinian Mayor's Office and the Tinian office of the CNMI Department of Wildlife and Natural Resources was critical in organizing and conducting fieldwork. Joey Charfauros assisted with expedition logistics. Dr Tomohisa Irino graciously ran the XRD analyses. Dr Mitsuhiro Nakagawa and Ms Akiko Matsumoto facilitated the use of a petrographical microscope. Dr Hisatake Okada provided the SEM and his students Mr Kesiuke Nagai and Ms Chieko Aizawa kindly assisted with its use. The authors are especially thankful to Mr Edwin Cabrera, of the Tinian Mayor's Office, and his family, for their immeasurable kindness and help. The paper benefited greatly from the critical comments provided by Dr John Mylroie, Dr Ken Grimes and two anonymous reviewers.

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CAVE AND KARST SCIENCE Vol. 30, No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Dinantian vertebrate (fish) remains in the walls of Ogof Draenen, South Wales, UK.

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(Received 11 June 2003; Accepted 02 October 2003)

INTRODUCTION

Unusually well preserved, and preferentially eroded, vertebrate remains have been recorded in many sites in the walls of Ogof Draenen (NGR SO 2463 1178), South Wales. These have previously been reported in the briefest of details only (Kendall, 1995). They are described more fully here, with some consideration of their palaeontological and stratigraphical significance, and their relationships to some passage names within the cave. Brief descriptions of the fish taxa and morphology are also provided. The nature of the preservation of the fossils, and their location within the cave, may provide valuable indicators of water flow régime during the formation of the passages in which the fossils have been found. This document also forms a permanent record of some sampling from the cave.

PRESERVATION

All of the known fish remains in Ogof Draenen are composed of apatite and are well preserved. They appear as black, slightly brittle but generally competent and shiny, well defined and easily recognizable fossils against a background lithology of pale cream oolite. Commonly differential erosion has left them protruding, sometimes by many centimetres, from the surface of the cave passage walls.

DESCRIPTIONS OF FOSSILS

A range of types of fish remains are found within Ogof Draenen, including small pieces of unidentifiable bone, scales, teeth and large pieces of fin spines. Descriptions of each (except bones) are given below.

Scales

These are distributed fairly widely through the Craig-y-Gaer Coral Bed and can commonly be seen glinting as small dark shapes in the creamy white limestone. Scales of a number of shapes and sizes can be seen ranging from <1mm to around 5mm. It has not been possible to attribute these to any particular taxa. Excellent examples can be seen in the Big Bang Pitch area, and in Indiana Highway, but these latter examples are now less visible due to the passage of cavers adding a layer of mud to the walls.

Teeth

Teeth (grinding plates) from both carnivorous and bradyodont fish are exposed in the walls of the cave. Bradyodont teeth were first reported by Kendall (1995) from the Psammodus Loop passage in the Big Bang Pitch area of the cave. Teeth of carnivorous fish are also present in this area. A single loose bradyodont tooth-plate was removed from the northern side of Big Bang Chamber and donated to the National Museum and Galleries of Wales (NMW) (NMW 95.32G.2). This tooth-plate was identified as being from a species of *Psammodus* (Holocephali: Psammodontiformes). It was the discovery of this specimen that gave rise to the passage name of Psammodus Loop.

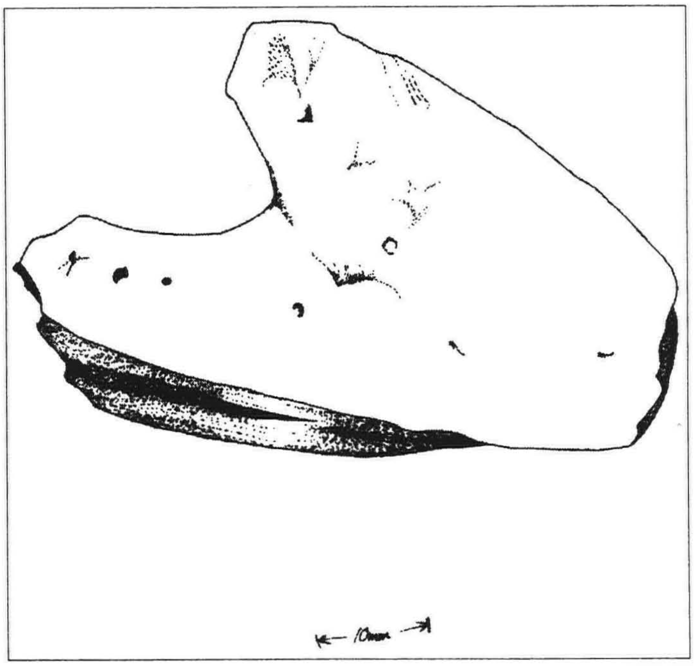

***Figure 1.** Transverse section of a fin spine showing blood vessels (Drawing by Rhian Hicks).*

Spines

The most spectacular and most readily identifiable of the fish remains are dorsal and ventral fin spines. These are from species of cartilaginous fishes (Chondrichthyes). The spines are seen in both transverse and longitudinal section. The spines have a U shaped transverse section and typically vary in width from 12 to 36mm.

Blood vessels

The fossil preservation is sufficiently fine for internal structures as well as some external ornamentation to be preserved. This is clearly seen, for example, as blood vessel holes, typically 1 to 2mm across and numbering 4 to10 per sample among those specimens examined to date.

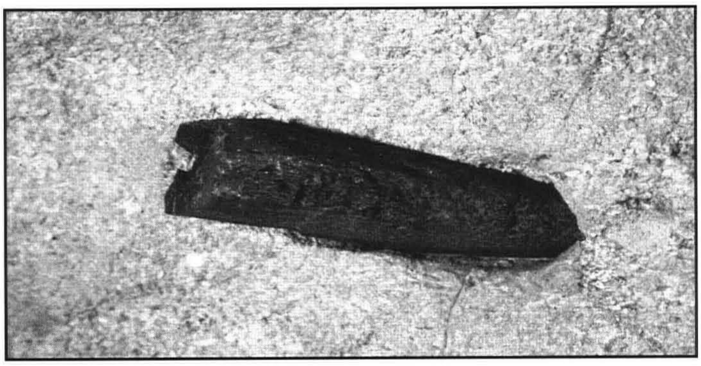

***Plate 1.** Ctenacanthus – longitudinal view from front of spine section*

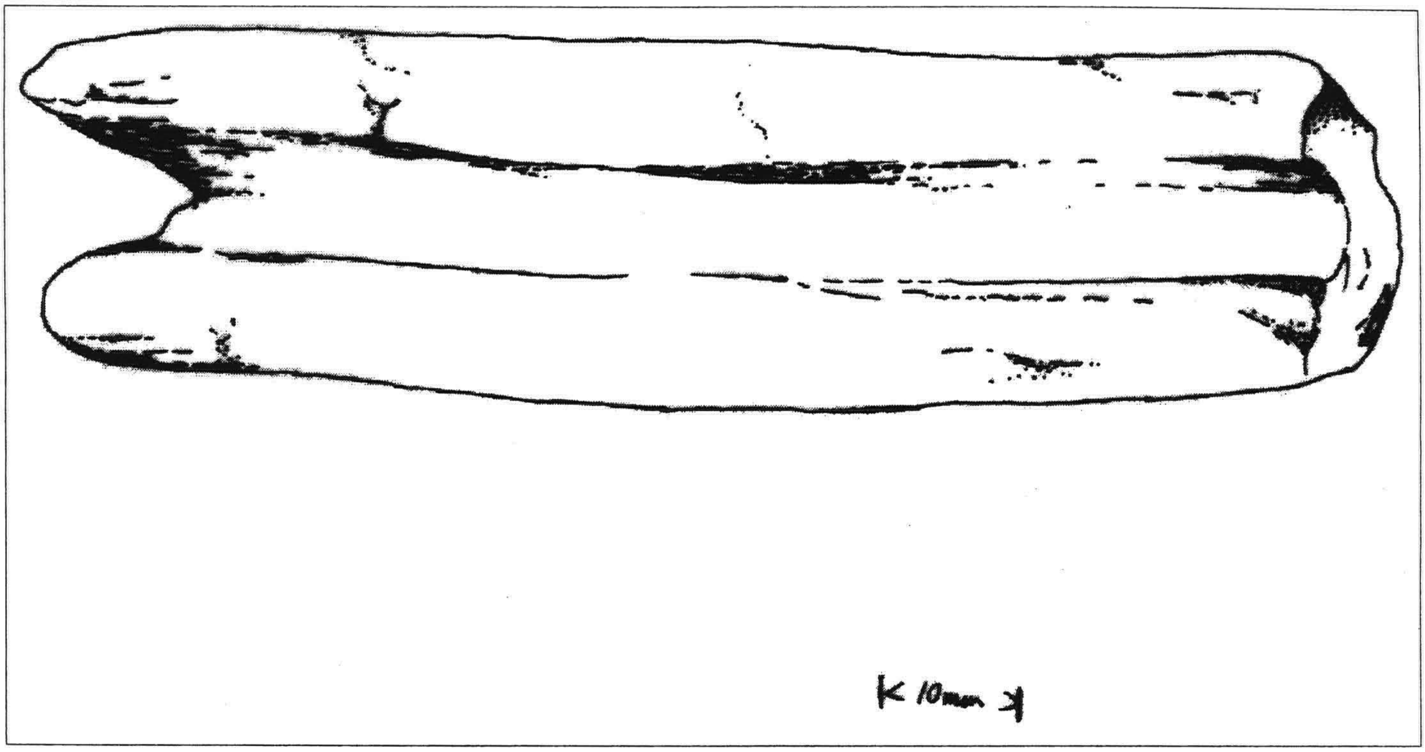

Figure 2. Longitudinal view of spine (Drawing by Rhian Hicks).

However, ornamentation that would enable a more precise identification is commonly lacking. This may be a reflection of the high-energy environment in which the Craig-y-Gaer Oolite was deposited. Ornamentation would be eroded away at the time of deposition, but fine preservation of internal structures would be possible. Such conditions would also explain the lack of other bones and the disarticulated nature of the fossils that have been recorded.

A well exposed longitudinal specimen of a fin spine in Gyracanthus Loop is 80mm long. Part of the spine is missing from the wall, but is represented by a well-developed cast that is only slightly weathered, and which is a further 40mm long. This specimen displays a finely striated ornament aligned along its length.

A single specimen of a small spine (from a species of *Gyracanthus*) has been removed from the streamway where it was found loose and is now housed in the NMW (NMW 95.32G.1). A passage in this area was named Gyracanthus Loop after this and other fine specimens.This passage name stands even though the these are now known to be another genus as described below.

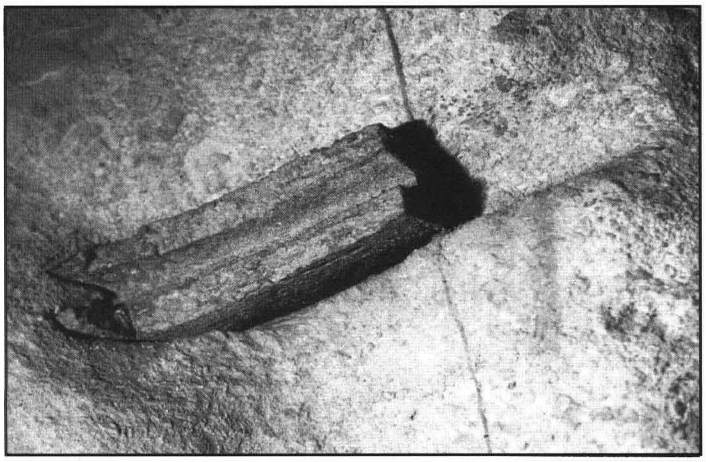

Plate 2. Ctenacanthus – longitudinal view from rear of spine section and eroded mould where specimen would have been. Specimen is 80mm long and 27mm high, The mould is a further 40mm long.

INTERPRETATION AND RECONSTRUCTION OF THE FISH REMAINS

The Ctenacanthidae had two fin spines, one in each of 2 dorsal fins. The anterior one was fairly low angled to the body, whereas the posterior one was held more vertically with a fin supported by it.

IDENTIFICATION, DISTRIBUTION AND STRATIGRAPHICAL USEFULNESS

Tentatively the spines have been assigned to the genera *Gyracanthus* and *Ctenacanthus* (Elasmobranchii: Ctenocanthidae) on the basis of information provided by Maisey (1978, 1981, 1982, 1984). Identification to species level has not been possible because of the state of preservation and the restricted viewing angles that the specimens present.

Identification of fish spines is complex. Morphological variation at different points along the spine has created taxonomic difficulties, with a tendency to erect superfluous palaeontological names. Maisey

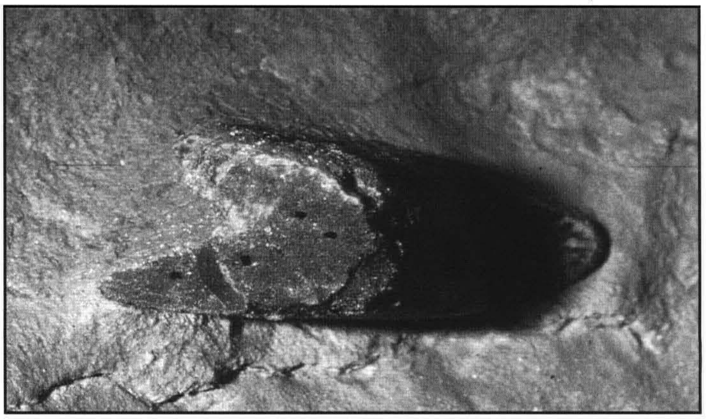

Plate 3. Ctenacanthus – transverse section showing blood vessels. Specimen is 70mm across and 50mm high. Blood vessel holes are 2 to 3mm across.

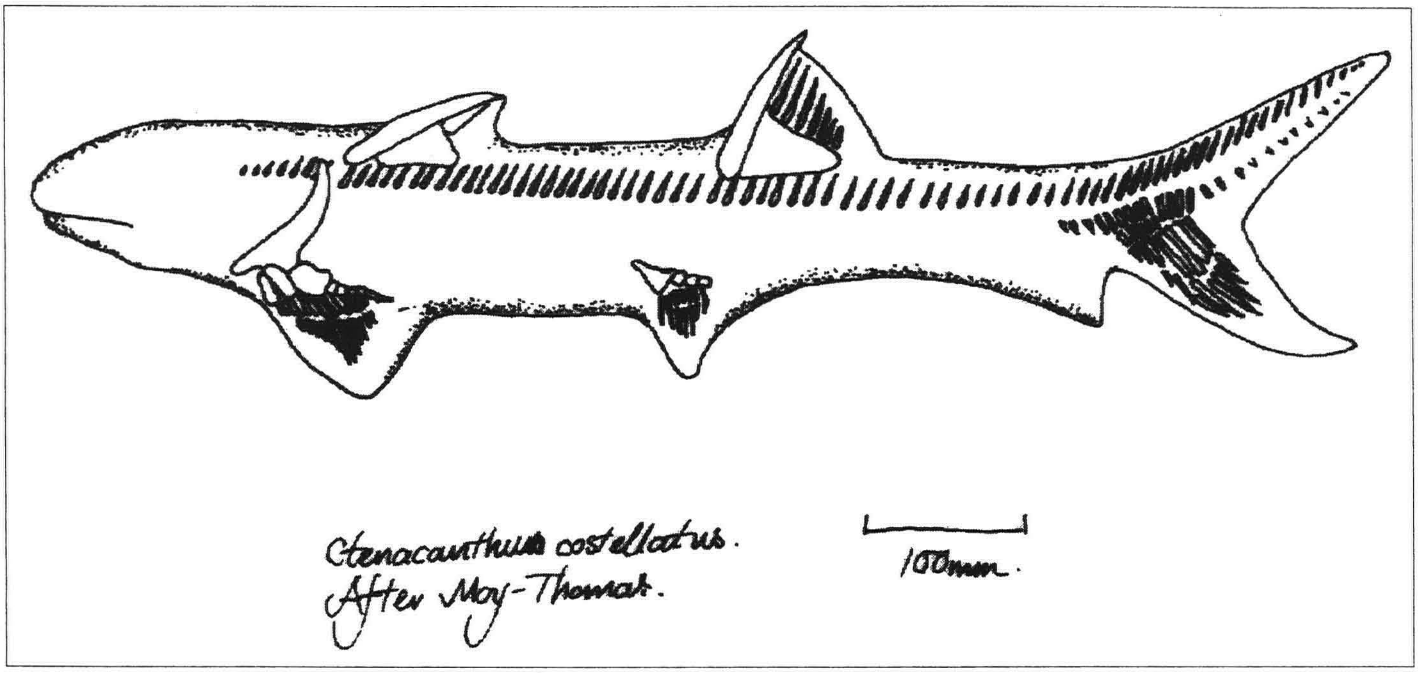

Figure 3. Diagram showing fish reconstruction (Drawing by Rhian Hicks).

(1981, 1982, 1984) has reconsidered available evidence and shown that many species were formerly assigned incorrectly to the genus *Ctenacanthus*. Another problem is that several superficially similar genera are distinguished by ornamentation and profile details that are not visible in the specimens so far observed in the cave.

Fish remains have been seen in a number of stratigraphical units within the cave, but appear to be most common in the Craig-y-Gaer Coral Bed (Member) of the Gilwern Oolite Formation (?Chadian Stage). It should be noted, however, that the fossils are particularly striking in this unit, as they are almost black in colour and stand out against the pale cream oolite. Similar fossil material may be present, but less obvious, in other rock units or hidden by the dark manganese-stained mud covering the walls in other parts of the system, such as the streamway.

Examples of fish remains may be seen in several places including the Gyracanthus and Psammodus loops, Pitch Bypass, Raiders Passage and in sections of Indiana Highway and Megadrive.

The Craig-y-Gaer Coral Bed is commonly developed as a cave horizon as it is a relatively soft and easily eroded oolitic limestone, and is usually cut by narrow rift passages. The presence of the fish remains and their relief from the cave walls facilitate recognition of this horizon within the caves, but the fish remains do not confirm identification of the Bed, because similar material occurs locally in other formations.

The fish remains themselves do not provide particularly precise stratigraphical markers. Examples of fossils from the Family Ctenacanthidae appear first Mid Devonian (Givetian) times and persist well into the Viséan rocks of the Lower Carboniferous (Dinantian). Neither are these fish a particularly good environmental indicator as the principally marine Chondrichthians are also found in brackish and even in freshwater environments. Better evidence for the environmental conditions under which the Craig-y-Gaer Coral Bed was deposited are provided by the abundant marine invertebrate fossils that it contains.

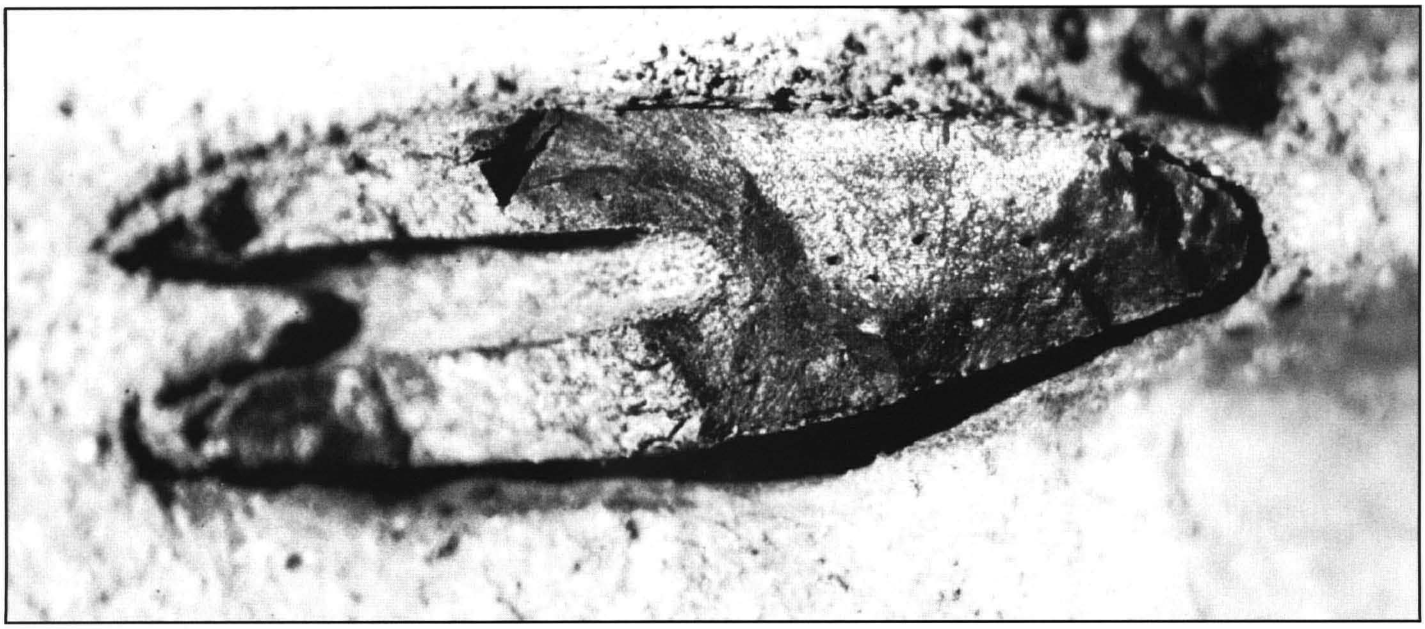

***Plate 4.** Ctenacanthus – longitudinal view from rear of now damaged spine section.*

CAVE DEVELOPMENT INDICATOR

Spines and teeth are extremely well displayed in the walls of the cave. The spines typically stand out 10 to 20mm from the wall, in one case reaching 75mm. This type of presentation is not commonly found in cave walls where a clean truncation of fossil material against the passage wall is the norm. In part this is probably explained by the difference in hardness between apatite and calcite, but this does not explain either the degree that the specimens stand proud of the walls or the relatively well preserved casts that remain within the wall rock.

The spines seem to show little or none of the rounding that might be expected in a highly aggressive, sediment-rich, erosional environment. Typically they have broken along straight fracture lines with crisp corners, such as would occur if they broke due to the unsupported weight of the eroded-out specimen.

These features probably indicate that the water in these passages at the time of their major enlargement was gently flowing with a low clastic sediment load. The fact that this type of water flow led to narrow rift passage development probably relates to the softness of the limestone promoting relatively rapid incision of the passage floor in preference to sideways erosion.

CONCLUSION

The presence of so many fish remains in the limestones of the area is not unexpected, as Ogof Draenen is little more than 50km from the Avon Gorge where the type material of *Ctenacanthus* was first described by Agassiz (1833 – 1844). However, both the degree of preservation and the manner in which they are presented displayedare unusually good.

However, the fossils should be considered as being more informative about the possible karst environment rather than being of particular stratigraphical or palaeontological significance. Further examples should therefore not be removed from the cave.

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POSTSCRIPT

Since this report was drafted the authors have revisited the Big Bang area of the cave and must sadly report that the specimen shown in Figure 2 has been damaged. Despite a careful search the broken end of the specimen was not found.

If anyone reading this knows the whereabouts of the missing part of the specimen the authors would appreciated being contacted. It is important to ensure that any material removed from Ogof Draenen is included within a single collection at the National Museums and Galleries of Wales. Such items will then be part of a single, scientifically useful, collection rather than being distributed and possibly unrecorded as individual specimens.

CAVE AND KARST SCIENCE Vol. 30, No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Dolomitization of the Great Scar Limestone Group in the Black Keld catchment, North Yorkshire, UK.

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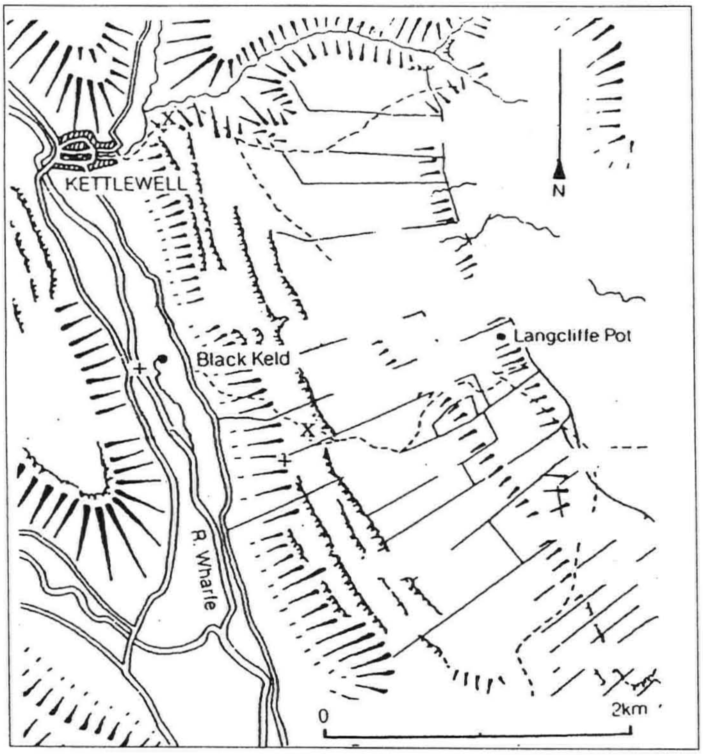
(Received 4th September 2003; Accepted 22nd September 2003)

Black Keld is the rising for water sinking on an18km^2 area of moorland on the east side of Wharfedale including drainage from Langcliffe Pot (9.6km long, 116m deep) and Mossdale Caverns (10km long, 60m deep)(Fig.1). Diving exploration began in the 1949, when divers reached an underwater pot 50m from the entrance (Davies, 1950). At the limit of the exploration the passage walls were observed to change from smooth to *"...being corroded to a fantastic three dimensional maze"*. Progress beyond this point was made in the late 1960s and early 1970s, when a further 100m of passage was explored. This passage is unlike that explored in 1949, being much smaller, and the diver has to manoeuvre between interleaved fingers of friable rock (Fig.2). The friable nature of the rock in this region led to some divers describing the area as a boulder choke (e.g. Brook *et al.*, 1988, pp18-19). The way on was lost 150m from the entrance. A description of the passages up to this point was provided by Monico (1995).

No further progress was made until the late 1990s where a further 2km of large passage was explored (Judd, 2000). The new passage is entered through two very low sections, where the diver has to squeeze along low bedding-oriented passages. These low areas are again associated with friable rock.



\+ Outcrops of dolomitised limestone from Warren 2001
X Outcrops of dolomitised limestone noted during fieldwork fo this paper.

***Figure 1.** Map of the study area. Reproduced from Brook et al 1988 with permission.*

X-ray diffraction analysis revealed that the friable rock is dolomitic rather than calcitic limestone (Fig.3). Thus the reason for the small passage size beyond the 1949 limit is due to the lower solubility of dolomite compared to that of calcite. Examination of thin sections reveals that dolomitization has caused a near complete loss of the rock's depositional fabric. The crystals have subhedral to anhedral boundaries and contain many inclusions (Fig.4).

Dolomitization of the limestone in the bed of the River Wharfe adjacent to and at the same altitude as Black Keld (Fig.1) was noted by Warren (2001). The dolomitization occurs in irregularly shaped pods. Warren also identified a bed of dolomitized limestone southeast of Black Keld at an altitude of 320m (Fig.1). During fieldwork in 2003 the author located dolomitized limestone northeast of Kettlewell at an altitude of 330m (Fig.1) and southeast of Black Keld at an altitude of 340m (Fig.1). The dolomitic rock is recognizable in the field by its pale brown colour, crumbly weathering texture and hollow sound when tapped with a hammer. The dolomitization appears to be stratiform, with one level at around 200m altitude and a higher level at 320–340m altitude.

No reference to dolomitization of rocks within the Great Scar Limestone Group of the Kettlewell region is known prior to that of Warren (2001), though dolomitization is described to the south by Arthurton *et al.* (1988), associated with the Craven Fault Zone. In this latter region the dolomitizing fluids are thought to have been introduced via fractures. Published geological maps show no significant faulting present in the Black Keld area. However direct observation from caves shows this not to be the case. The apparently stratiform nature of the dolomitization in the Black Keld area suggests environmental rather than a structural control, though the distribution may be due to dolomitizing fluids having migrated laterally from fractures into beds of favourable lithology. Where dolomite is encountered within the cave system behind Black Keld it clearly has an inhibiting effect on cave development. Langcliffe Pot is entered in the Middle Limestone of the Yoredale Group and can be followed down the succession to where the way on is lost in a small perched sump within the Hardraw Scar Limestone. Despite

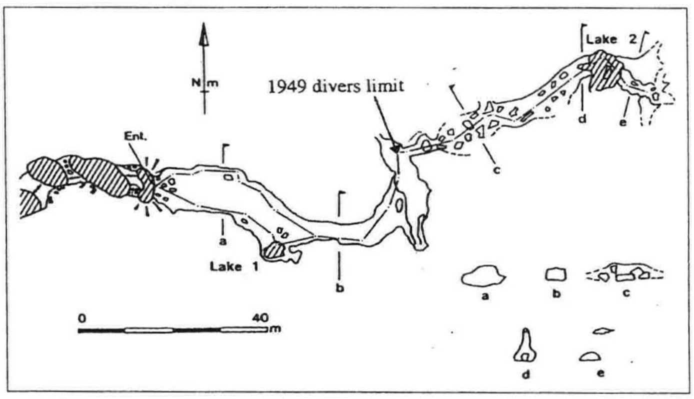


***Figure 2.** Survey of the entrance area of Black Keld. From Monico 1995. Above water passages omitted for clarity.*

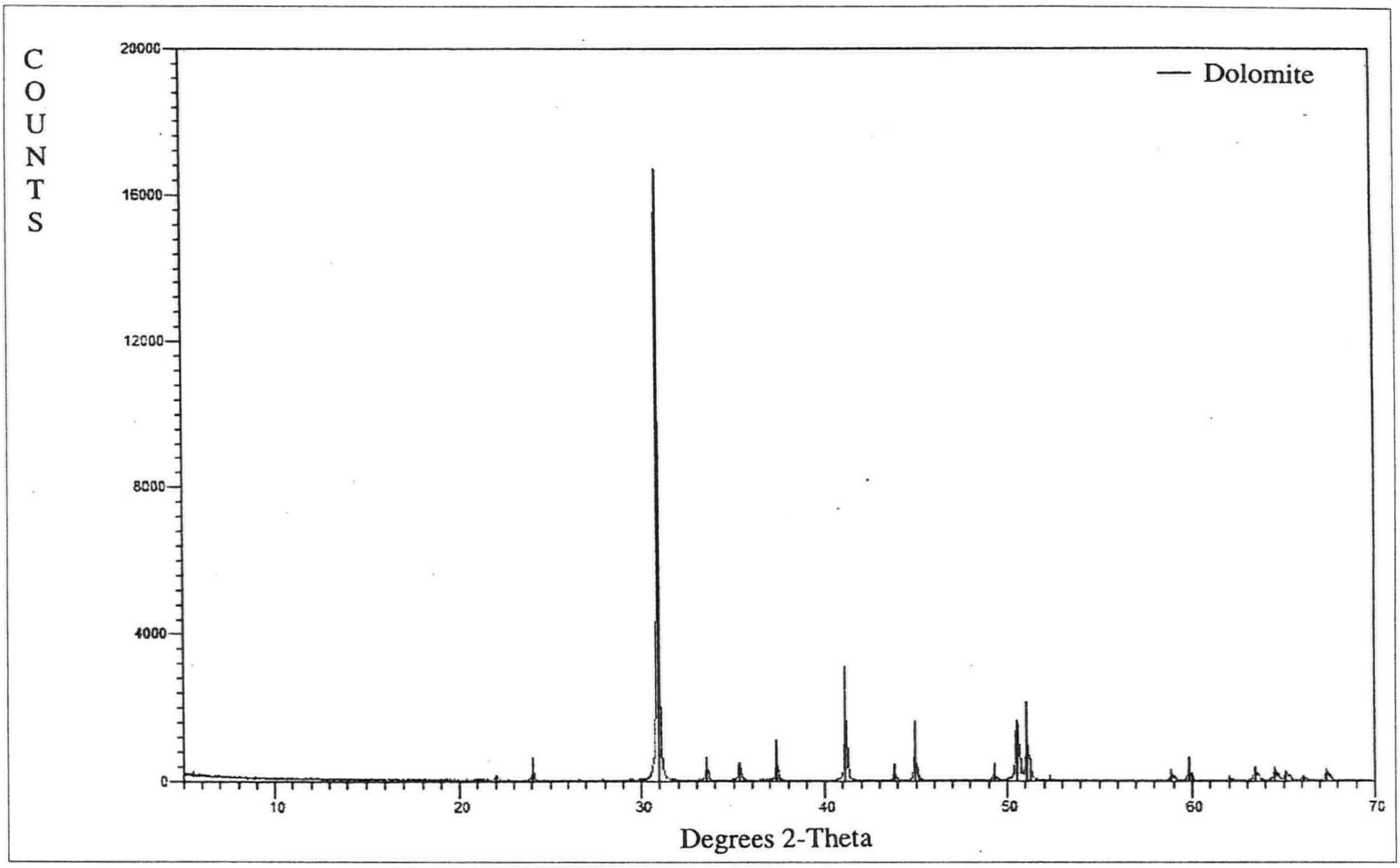


Figure 3. *X-ray diffraction trace of dolomitised limestone from Black Keld*

much work by cavers no progress has been made beyond this point. One possible but unconfirmed explanation for this perched phreas forming a barrier to cave exploration is that a higher level of dolomitization has been encountered than the two described in this paper. Chubb and Hudson (1952) describe slight dolomitization of the Undersett Limestone northeast of Kettlewell, confirming that dolomitic rock does occur in the Yoredale Group.

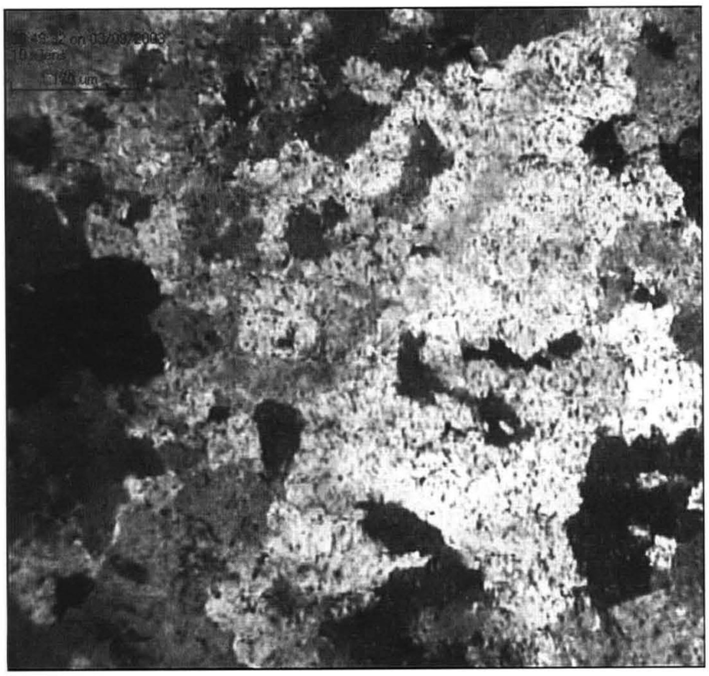


Figure 4. *Thin section of dolomitised limestone from Black Keld taken under crossed polars.*

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ACKNOWLEDGEMENTS.

Mr Steve Warren of the White Rose Pothole Club for sharing his geological knowledge of the area.

The Lambert Family, owners of Black Keld Farm for allowing access.

Brian Judd and Phil Howson of the Cave Diving Group for allowing me to dive with them in Black Keld.

Please Note: Black Keld is on private land. There is no public access.

CAVE AND KARST SCIENCE Vol. 30 No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Sub–sea level speleothems from the Andaman coast of southern Thailand, and sea level change in Southeast Asia.

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Abstract: Recent exploration of a freshwater, vauclusian spring near the Andaman coast of southern Thailand has discovered submerged, calcite speleothems of sub–aerial origin. These occur between depths of –15m and –90m, equating to 0m to 75m below mean sea level (b.m.s.l.). A sample from –59m (= 44m b.m.s.l.) has been dated using radiocarbon and found to be 34.2 ± 4.5ka old, implying that the spring was dry at least to this depth at this time. This situation arose as a result of deeper water rest levels in response to a relatively lower sea level.

Keywords: Sea level change, speleothem, radiocarbon dating, Krabi, Thailand, Andaman Sea, Southeast Asia

(Received December 2001; Revised August 2003)

INTRODUCTION

Krabi Province is located on the Andaman coast of Peninsular Thailand (Fig.1) and has arguably some of the most spectacular karst scenery on earth. The area is characterised by broad alluvial plains punctuated by numerous, precipitous towers formed of massive Permian limestone. Tower summits reach heights of several hundred metres and are draped with dense, evergreen vegetation. The towers also extend out into the sea where they form gravity–defying islands (Odell and Odell, 1984; Maffre *et al.*, 1986; Kiernan, 1986, 1988; Deharveng and Bedos, 1988). This landscape has evolved over a long period of time and results from a complex interplay between physical, geological, climatological and biological influences and processes (Kiernan, 1988). Of relevance here are the marine processes that have played an important role in steepening tower sides by undercutting their bases. The coming together of the sea and the karst has been brought about through changes in relative sea level, evidence for which is summarised below.

During the last glacial maximum, eustatic sea levels dropped to an estimated 120m below mean sea level (b.m.s.l.) and large areas of now submerged land in Southeast Asia became emergent. The so–called Sunda Shelf connected the islands of Borneo, Sumatra and Java to the mainland (Medway, 1972). Maps of the sea floor on the Sunda Shelf show the existence of a dendritic pattern of drainage channels formed during periods of emergence (Verstrappen, 1975).

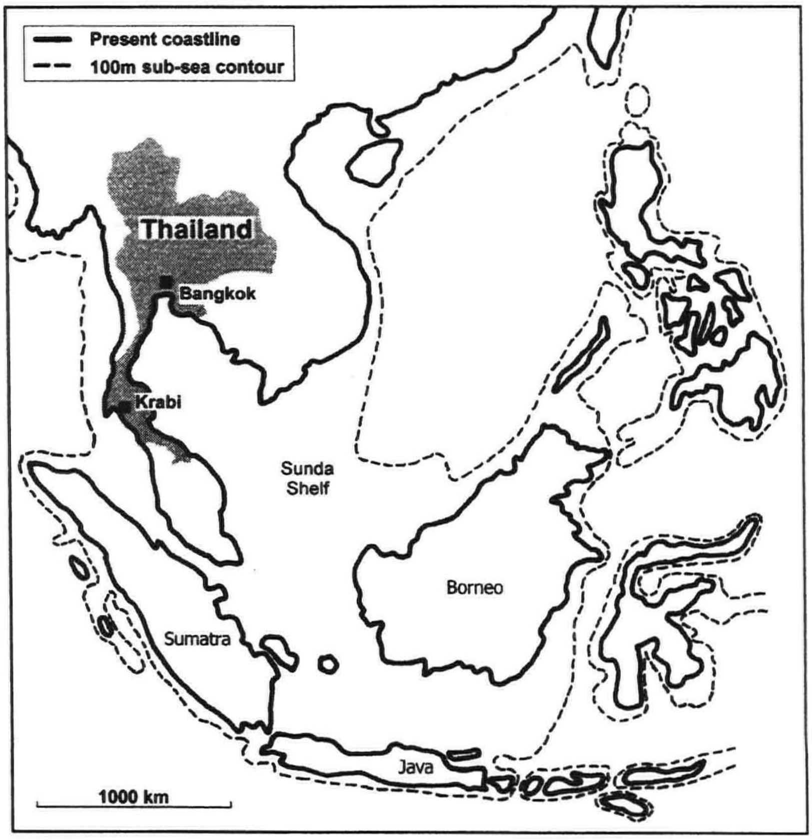

Figure 1. Map of SE Asia showing the location of Krabi Province, Thailand. The 100m sub–sea contour gives an indication of the extent of land (the Sunda Shelf) likely to have been emergent during periods of relatively lower sea level (After Medway, 1972).

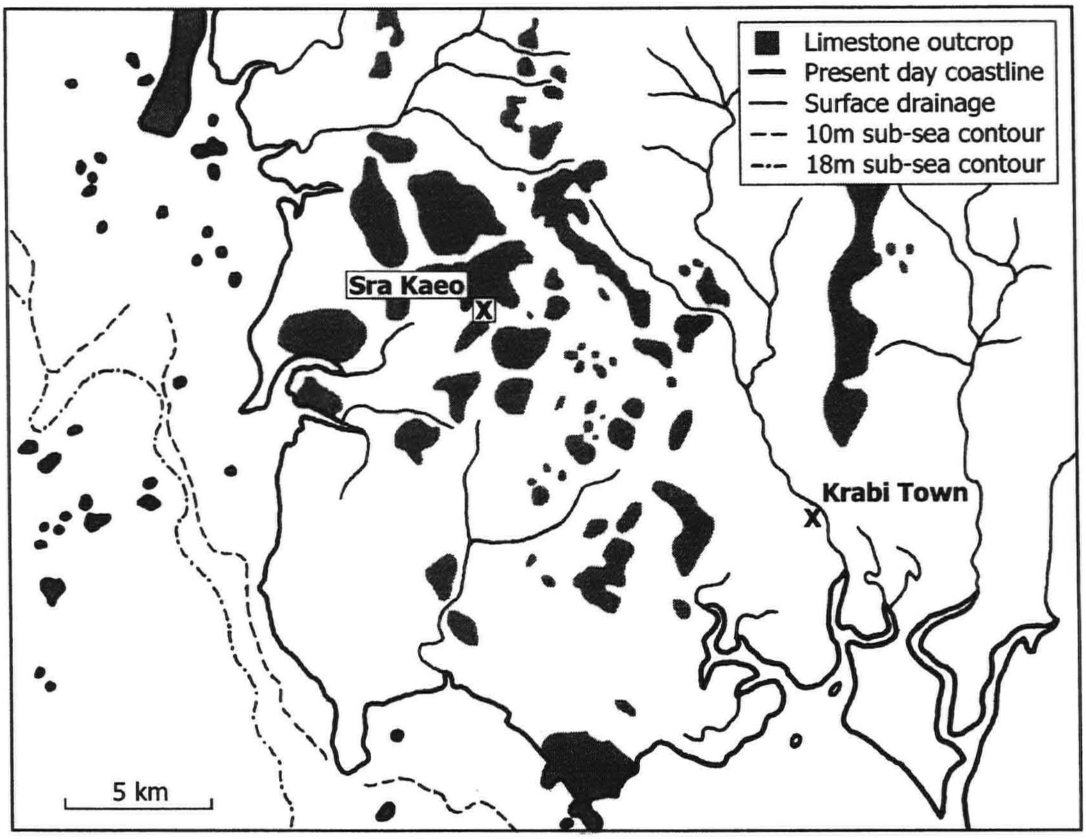

Figure 2. *Location of Sra Kaeo in relation to the present coastline and Permian limestone outcrop.*

Submerged peat deposits also occur here (Aleva *et al.*, 1973; Biswas, 1973) and drowned shorelines have been identified up to 90m b.m.s.l. (Tjia *et al.*, 1977). At Krabi, shallow–submarine caves in the offshore towers commonly contain speleothems such as stalagmites (Mathew London, pers. com.). The towers also have two prominent tidal notches, one at present sea level and the other 4 to 6m above. The upper notch formed during a eustatic sea level high around 6.0 ka BP (Kiernan, 1988). Kiernan also suggests that there is possible concordance between overhangs, caves, breaks of slope and tower summits that may have been formed or modified by marine influences. This implies relative sea levels of up to 120m above today's level and maybe even higher at some time in the past. Archaeological studies have provided indirect evidence for sea level change. Excavations at two sites in Krabi – Lang Rongrien Rockshelter (Rockshelter Behind the School) and Tham Moh Khiew (Green Doctor Cave) – found a shift in faunal composition from earlier rainforest species such as deer, pig and buffalo (Pleistocene) to later shellfish and crabs (early to mid Holocene). This shift was a consequence of the coastline moving from in excess of 60km away to as near as 2km (Anderson, 1990; Pookajorn, 1991).

Uplift rates across the Southeast Asia region vary, with the southwestern margin experiencing greatest tectonic activity and the northern margin the least. Tjia (1996) estimated Holocene uplift rates of between 10.0m/ka and 0.03m/ka, although rates have not been consistent through time. Peninsular Thailand, including Krabi, is thought to have been relatively stable for the past 300.0ka (Kiernan, 1988). Studies of raised terraces on the tectonically active Huon Peninsular, Papua New Guinea, have revealed a chronological sequence of sea level high–stands during the Quaternary (Chappell, 1974, 1983; Ota and Chappell, 1999). However, this approach is subject to uncertainties concerning the rate and constancy of uplift over time (Bloom and Yonekura, 1985). Stable coastlines offer less uncertainty, but may not provide complete records, e.g. the Quaternary reefs of the Bahamas (Neumann and Moore, 1975). Speleothems from flooded caves have the potential for greater altitudinal range and the timing of growth commencement (draining of the cave) and final growth cessation (flooding) can be dated accurately. Harmon *et al.* (1983) and Smart *et al.* (1998) have studied submerged speleothems from the stable platforms of Bermuda and South Andros respectively and demonstrated the practicality of the method.

METHODS

Nestled amongst the inland towers of Krabi are a number of freshwater pools. Most of these pools are springs with small discharge rates. A few act as sinks. The largest and deepest of the pools is a spring called Sra Kaeo (Crystal Pool), which is located 10km NW of Krabi town (Fig.2). Sra Kaeo is 7km from the present coastline and the water surface is at 15m above mean sea level. Water rises up out of two pools 80m apart (Fig.3) and flows a short distance across the surface to a sink. Flow rate is small (<0.05 cumecs) and little seasonal change has been observed. The pools are connected at a depth of –70m and continue down to the present maximum explored depth of –120m where the passage descends further. Calcite speleothems have been found in Sra Kaeo at depths starting at –15m and extending down to –90m (= 0m to 75m b.m.s.l.). They have the form of large stalactites and flowstone indicative of growth under sub–aerial conditions.

The exploration of Sra Kaeo and other sites in the area is ongoing and being carried out by Matthew London and the Thailand Cave Diving Project. Diving to such great depths necessitates the use of mixed gases and long underwater decompression times. Visibility in Sra Kaeo is normally around 5m. Although by no means insurmountable, these logistic complexities make it difficult to collect speleothem samples. To date, only one sample (DA525) has been brought out.

The DA525 sample was collected from a depth of –59m (= 44m b.m.s.l.) and dated using radiocarbon methods at the Office of Atomic Energy for Peace, Bangkok. This is currently the only readily accessible, radiogenic dating method available in Thailand. The sample had a porous outer layer showing that some surface dissolution had taken place, though the calcite inside was solid and free of impurities. All porous and impure material was removed manually, leaving 28.1g of calcite for testing. This was ground, washed in 25% HCl, rinsed in de–ionised water and dried in an oven. Addition of 250ml of 50% HCl to the pre–treated sample released 9.43g or 4.8 litres of carbon dioxide gas. The gas was purified through a series of traps containing KI/I_2, $AgNO_3$, $K_2Cr_2O_7/H_2SO_4$, dry ice/alcohol and liquid N_2 and then absorbed into a solution of permofluor V/carbosorb. 20ml of this solution was

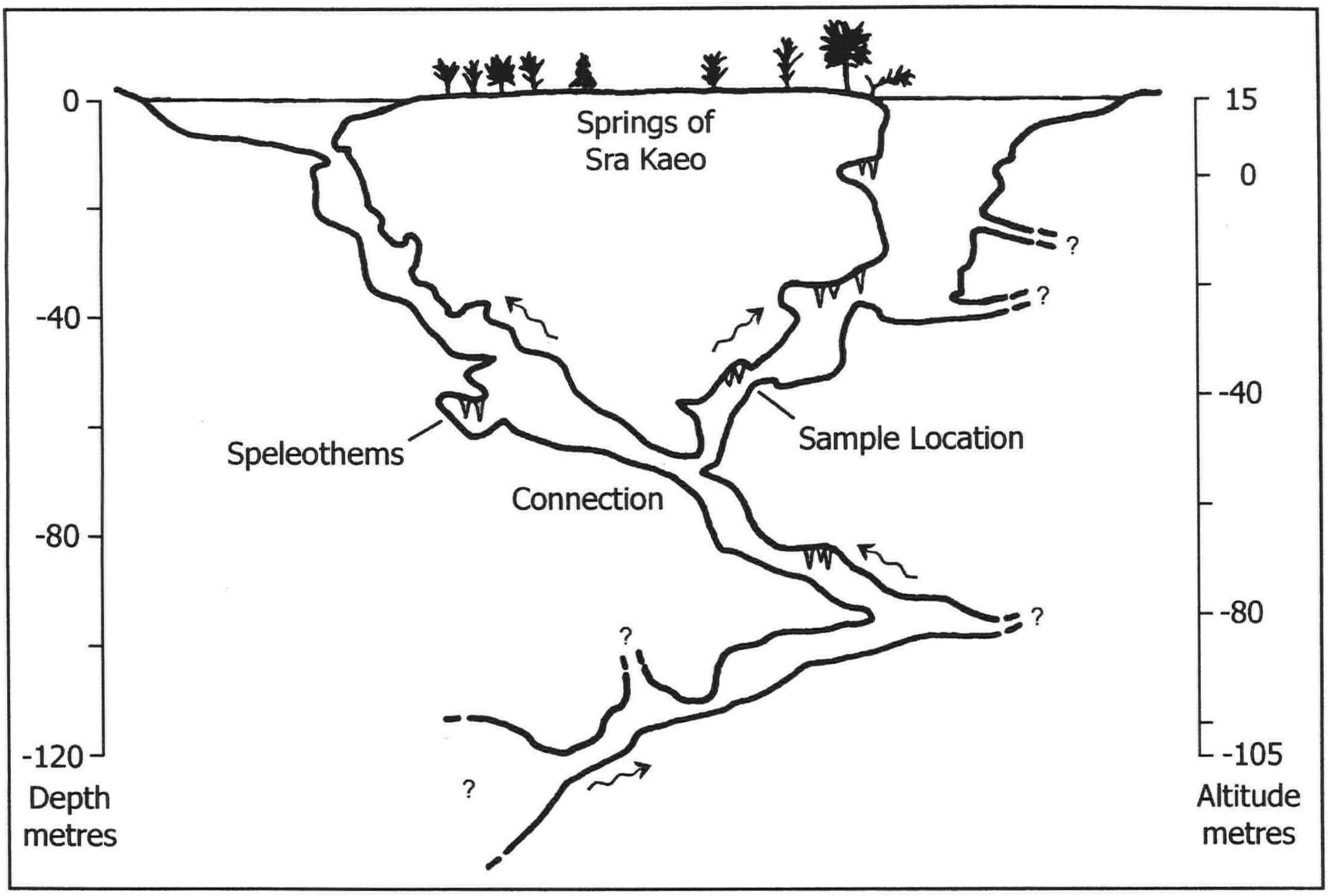

Figure 3. Cross–section through Sra Kaeo and the speleothem sample location (map by Mathew London). Vertical and horizontal scales are identical.

placed in a liquid scintillation counter for testing together with similarly prepared solutions of ANU sucrose standard and a marble background sample. Counting lasted 987 minutes.

RESULTS

Recent exploration of Sra Kaeo, a freshwater spring near the Andaman coast of southern Thailand, has discovered submerged calcite speleothems of sub–aerial origin. The speleothems occur between depths of –15m and –90m (= 0m to 75m b.m.s.l.) and a sample from –59m (= 44m b.m.s.l.) has been dated using radiocarbon. The result obtained shows an age of formation for the calcite of 34.2 ± 4.5 ka BP. Isotopic fractionation (C–13/C–12) was 0.00 ± 2.00 permil.

DISCUSSION

It is clear that the vauclusian spring of Sra Kaeo has been dry for considerable periods in the past allowing large, sub–aerial speleothems to form at depths of up to –90m (= 75m b.m.s.l.). Today the spring is active and emits a small underfit flow with the water rest level lying at the surface of the pool. In order for the spring to have been filled with air to at least –90m, the water rest level in the area must have been even deeper. This situation could only arise if the relative level of the Andaman Sea was at least 75m lower than it is today. Therefore the submerged speleothems of Sra Kaeo provide evidence for sea level change.

The date obtained here agrees with the findings of other studies of sea level change falling within a period when the spring would be expected to be dry (Fig.4). At 34.2 ka BP, the eustatic sea level was around 60m lower than today, i.e. 16m below the sample location. However, any correlation at this stage can only be regarded as speculative due to having dated just one sample.

The morphology of Sra Kaeo and the other springs and sinks suggest they have a phreatic origin. Present day water flows are small and underfit and have invaded after the time when the caves were dry. It is likely this invasion took place as recently as 6.0 to 4.0 ka BP when the water rest level rose to near to its current altitude. For cave development to have taken place phreatically, the water rest level (and relative sea level) would need to be at or above its present altitude. U–series dating of speleothems in Bermuda by Harmon *et al.* (1978) inferred higher eustatic sea levels for two short periods around 125.0 ka BP and 200 ka BP. Other than these periods and more recently at 6.0 to 4.0 ka BP, sea levels have been lower than today for at least the past 250.0 ka. If tectonism has been stable in Peninsular Thailand for at least 300.0 ka, as Kiernan (1988) suggests, these caves could prove to be very old indeed and the speleothems they contain could provide data well beyond the chronological limit of the work by Harmon *et al.*

The speleothems of Sra Kaeo also extend down to depths previously unknown in submerged caves (75m b.m.s.l. compared to 57m b.m.s.l. for South Andros) suggesting a very low sea level stand some time in the past. This may be represented by the stand seen at 23.0 to 14.0 ka BP or possibly another, even older stand.

A potential difficulty with the evidence provided by Sra Kaeo arises in the fact that as sea level drops the coastline moves away. At –75m the coast would be more than 100km away diminishing the relationship between sea level and the altitude of the water rest level. Care would be needed before any accurate correlations could be made between the speleothems and past sea levels. The problem might be reconciled through dating a large number of samples from a number of different localities or perhaps creating a groundwater model for the area.

Further research is needed to confirm or disprove the tentative speculations given here. It is apparent though that Sra Kaeo provides an opportunity for collecting additional data on sea level change.

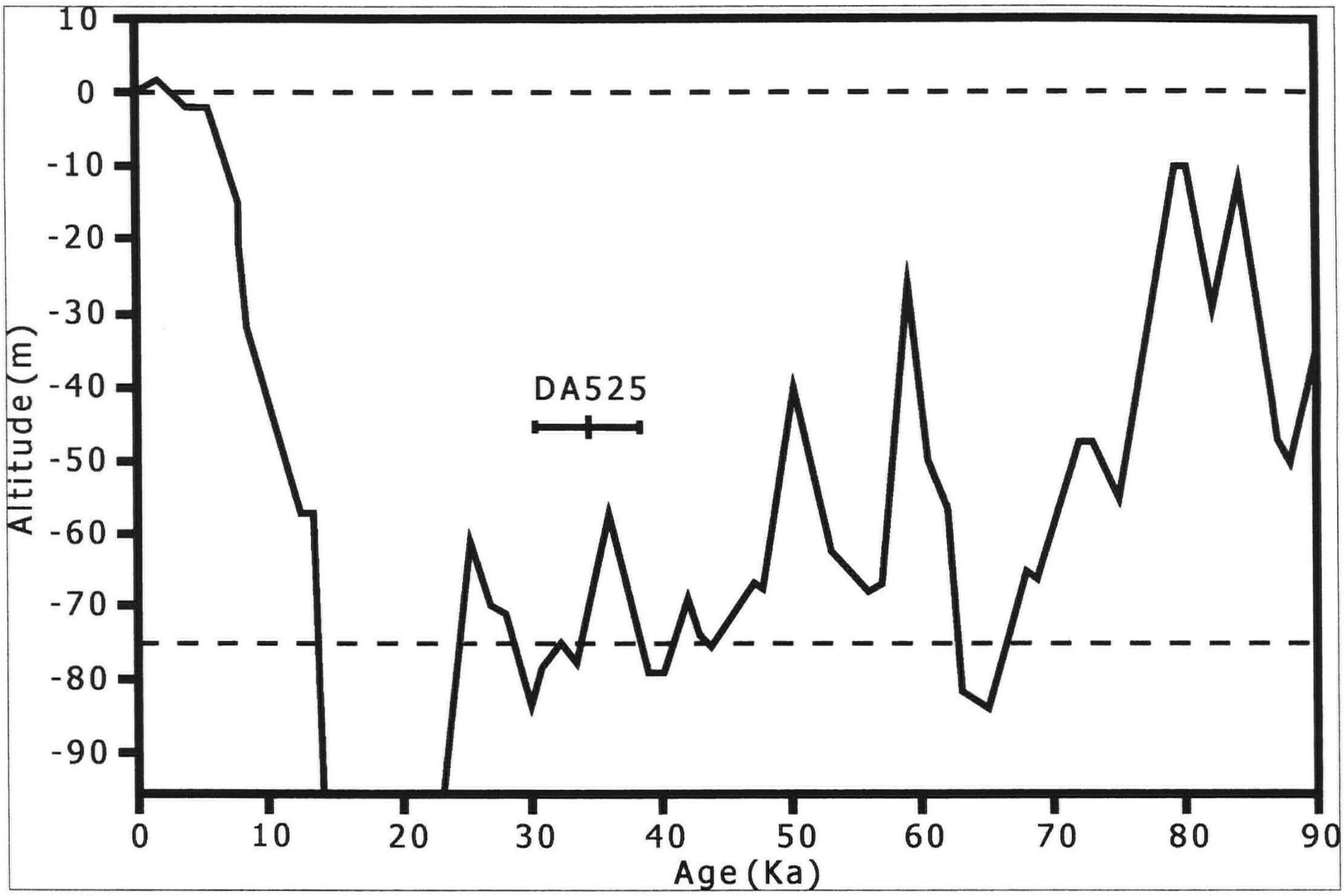

Figure 4. Altitude (44m b.m.s.l.) and determined age (34.2 ±4.5 ka BP) of speleothem sample DA525 from Sra Kaeo in relation to sea level curve predicted from ice volume record from the Norwegian Sea (Smart et al., 1998 after Labeyrie et al., 1987). Dashed lines are upper and lower depth limits for speleothems in Sra Kaeo.

ACKNOWLEDGMENTS

Matthew London has been extremely helpful in carrying out the diving, sharing information and collecting the specimen. I am also very grateful to Mr Manit Sonsuk and Mr Kiattipong Kamdee of the Office for Atomic Energy for Peace, Bangkok, who performed the dating. The Department of National Parks, Fauna and Flora Conservation provided the opportunity for this work, and logistical support.

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CAVE AND KARST SCIENCE Vol. 30, No. 1, 2003
Transactions of the British Cave Research Association

ISSN 1356-191X

Forum

Readers are invited to offer thesis abstracts, review articles, scientific notes, comments on previously published papers and discussions of general interest for publication in the Forum of *Cave and Karst* Science.

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CORRESPONDENCE

Dear Editors,

ENERGY REPLENISHERS

Rino Bregani's report, "Energy replenishers in speleological practice..." in *Cave and Karst Science* Vol.29(3) states, of the experiments, that "... most results do not achieve statistical significance". It then goes on to explain that, nevertheless, the data showed an improvement in parameters and that with more data "... the results could achieve statistical significance".

Perhaps I have missed something here but surely it is complete nonsense to attempt to extrapolate from results that are not statistically significant? I mean: either the results **are** significant, or they **are not** – you surely cannot have it both ways? In fact, I would go so far as to say that the abstract of this report is misleading, with the author doing his best to proclaim a physiological effect that is not borne out by the experiments.

There is a second fault in this report, which is the implied accuracy of the measurements. The easiest way to describe this problem is by an analogy, as follows. A caving guidebook lists a pitch length as 30 ft, and this is translated into metric units as 9.144m. The original measurement has an implied accuracy of perhaps six inches (150mm), but the metric conversion implies an accuracy of 1mm, which is ludicrous. Clearly the correct translation of (about) 30 ft is (about) 9m.

Bregani seems to have fallen foul of a similar problem caused, perhaps, by the process of averaging his data. In his tables 4 and 5 there are too many significant figures for the parameters to be plausible or, at least, for the figures to be readily digestible by the reader. Blood pressure measurement is not easily repeatable at the best of times, so to quote it to hundredths of a mmHg is bizarre, to say the least.

Given that Bregani's experiment actually showed that there was **no** advantage in his proposed dietary supplement, may I suggest an alternative, but related, line of research? Rather than considering dietary supplements for energy replenishment, it may be worth investigating supplements for improving tolerance to cold conditions.

The xanthene group of compounds (as found in tea, coffee and cocoa) contains adenosine receptor antagonists or deactivators that have the effect of boosting the body's tolerance of cold conditions. According to papers I have read, (including European patent application 88312107.1, date of filing 21/12/88) the preferred compound is theobromine, optionally in conjunction with theophylline or aminophylline, and accompanied by an appropriate energy supplement.

I reported on this work at the BCRA conferences in 1999 and 2000 and I also described my attempts to concoct a suitable dietary supplement in my own kitchen. But, given that continental chocolate can be 70% cocoa, and cocoa is 1 to 2% theobromine, I wonder whether the simple advice to *"Eat a bar of dark chocolate when you feel cold"* might be adequate?

The above-mentioned patent is a good starting point for a literature search on the subject, which was also discussed on-line in the Cavers' Digest in 1996. I can provide a summary of this discussion if anyone is interested.

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Dear Editors,

ENERGY REPLENISHERS

I thank David Gibson very much for his observations.

The problem related to the "apparent nonsense" can be explained as follows. The statistical significance is related to probabilistic considerations in numerical terms, whereas sometimes reality is completely different. In some cases such significance can be reached simply with a small increase in the number of observations. There are lots of examples of real phenomena that can't be demonstrated with statistical tests, and a great many cases that appear to display statistical significance, but are completely absurd. Statistical analysis of the single dataset from this experiment commonly doesn't reach the level of statistical significance, but the statistical test doesn't take account of all other data and of the global trend of observations.

So, from the statistical point of view, data cannot always demonstrate significance. However, logical consideration of all the data, together with the evidence of other statistically significant results allows the view that "probably", with more data, statistical significance "could" (conditional) also be reached.

Simple statistical results are sometimes insufficient and, in such cases, human intelligence and common sense can override the limits of mathematical rules. I mean that a statistical test takes into consideration only one result at a time, while it seems unlikely that data of CPK and LDH that **are** significant, **and** data of respiratory and heart rate showing better recovery in the treated group **and** better time performances in the meander in tired treated subjects, all together, are all simply due to casual effects.

So, even if statistical analysis is significant only in few cases, there is the possibility (in my opinion also the probability) that a real advantage can be provided by creatine, if not in performance times, at least in recovery capacity and muscular damage limitation. More data are surely needed to allow this hypothesis to be demonstrated by statistical methods.

As for the second question, I'm not sure I understand the problem, but perhaps this reflects my understanding of the English language. The measurements in Figure 4 are real, regardless of considerations of their plausibility. It would have been wrong to change the data to make them appear more plausible. The differences are not discussed because that was not the aim of the study. In Figure 5 the significant difference are due, in my opinion, to the effect of creatine, even if they are not significant statistically. Blood pressure was measured with an electronic instrument, and in this way such measurements are easily repeatable; moreover the values are not quoted in "hundredths" of a millimetre of mercury, but in millimetres of mercury.

The experiment shows that **there is** an advantage (statistically proven) in muscular protection. It would be interesting to repeat the experiment with more volunteers, to verify whether the improvement of heart and respiratory rates observed in the treated subjects is related to an effect of the dietary supplement (as has been demonstrated in other sports, with different supplements) or whether this is simply a casual effect.

I hope that my comments are sufficiently clear but, if not, I will be glad to continue the discussion.

Sincerely yours

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✉

Dear Editors

PEAK CAVERN

The paper by J Pringle *et al.* (*Cave and Karst Science*, Vol.29, No.2) on their geophysical investigation of the Peak Cavern entrance is welcome, but it only tells half the story and has some matters that require further consideration.

The diagrammatic plan (Fig.1) shows a dip of 30° to the northeast on the east wall of the entrance, but the photo on page 68 shows horizontal to gently undulating bedding planes along that wall. The Swine Hole passage is a classic phreatic tube following a bedding plane dipping at around 8° to the east, and that bedding plane has been followed by divers at shallow depth all the way to the Resurgence. The bedding planes visible on the east wall can be followed outwards along the gorge wall and inwards round the back of the Vestibule to Lumbago Walk and the west wall where they meet a series of avens rising above the footpath on Level 6. It thus seems likely that the original of the "hall" cavern lay in inception horizons along those bedding planes and the fracture now represented by the avens, followed by development as the intervening beds were removed. Until the Halfway House to Resurgence stream course became available the Peak Cavern stream exit must have been via Lumbago Walk under epiphreatic conditions, and the cave mouth and gorge would have been something like a Vauclusian spring, drained only as the valley floor outside was eroded down.

The nature of the sediment bank with the Rope Walk terraces (levels 1 – 6) is a problem that will only be solved if one or more trenches can be excavated to investigate the stratigraphy of the sediments therein. Pringle *et al.* made the surprising statement that the sediments under levels 5 and 6 were the oldest, without offering any evidence. It is worth considering how the 6m or so of sediment fill could have been deposited. There could be three fill components. Frost-shattered scree in the entrance is likely to have come from the gorge walls but only from the cave roof for a limited distance inwards, as the roof's rounded contours show little sign of frost-damage. The second component would have come from Peak Caverns itself as sand, silt and mud washed out of the epiphreatic Lumbago Walk, perhaps under higher run-off conditions during the waning of the last ice age. Until the Swine Hole became operative, sheets of these sediments would spread right across the Vestibule, with the oldest at the bottom and not confined to beneath levels 5 and 6. Towards the entrance they would be interleaved with frost-shattered scree. The third expected component would be cave earths from occupation by animals and primitive man, perhaps as far back as Palaeolithic times. This is an obvious site for human occupation, but only when the drainage system from the Halfway House to the Resurgence had been established leaving the cave mouth relatively dry. That, in turn, meant that the flood overflow channel along the east wall was at least intermittently operative and its stream had cut down through the sediment bank.

It is a great pity that no cave archaeologist has been to dig a test trench through the Rope Walk levels. Years ago I found a boar tooth in the Swine Hole entrance, perhaps suggestive of the mammal bones that might lie within the terraces. Occasional collapses of the walling along the backs of the higher terraces have revealed layers of scree and cave earth.

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✉

Dear Editors

PEAK CAVERN – response to correspondence from Dr T D Ford.

We appreciate and value Dr Ford's comments. We readily agree that our results are only "half the story", if that. If they stimulate the further consideration he recommends then we shall have an ample reward. We applied new methods that need to be placed in context. His (1999) plea for cave archaeological excavation justified our use of geophysics. Since excavation is prohibited for conservation reasons, non-invasive methods are the only option. We are in the exquisite situation that geologists and speleologists may propose, but geophysicists must dispose!

We projected the generalised 30°dip of the fore-reef beds from Figure 5 in Ford (1999). We do plan to analyse the bedding plane dips in The Vestibule in more detail. Dr Graham Hunter of 3D Laser Mapping Ltd., Nottingham, has recently acquired a 16.8M-point, 1cm-resolution, ground LIDAR survey of The Vestibule. Preliminary, unpublished results suggest that the apex of The Vestibule roof lies along an anticlinal axis. The cave follows the bedding down a northeasterly fold-limb. Dr Ford's observation of an 8° dip in The Swinehole would then be from an intermediate position between the lower dips of the anticlinal axis and the higher dip of the apron-reef. We agree with him that *"... the original of the "hall" cavern lies in inception horizons along those bedding planes and the fracture represented by the avens, followed by development as the intervening beds were removed"*. We now suggest that the aven-fractures may be axial-planar strike-slip and tension joints developed along the anticlinal axis.

Our "surprising statement" follows from the argument we developed. The underlying evidence surprised us too! Our argument is based upon observations we made on our 3D-CAD multi-data model. For re-evaluation purposes, the key elements are:

1 **Cave morphology:**

a Bedding dips are up to 30°, rather than the *"... greater than approximately thirty degrees"* of Osborne (2001). However, the morphologies of The Vestibule and The Swinehole seem to match Osborne's *"Hall and Narrow"* classes of cave in their form and relative orientation to bedding. Of course, the aven-fractures may be the controlling surfaces, and these dip at more than 30°

b The system is still developing, as the stream is in the process of leaving The Vestibule *"Hall"* in favour of eroding The Swinehole *"Narrow"*.

c Our combined results showed that the rockhead beneath the cave fill is roughly parallel to the curved and inclined roof of The Vestibule. Specific beds seem to guide roof, floor and lateral development. A cross-section of the rock surface surrounding The Vestibule is not circular but a rough, bent rectangle, concave downwards.

2 **Cave sediments:**

a As the streamway enters The Vestibule, there is a bank of cave earth on the outside of the bend in the passageway between the steps and Lumbago Walk. Our GPR section there appears to show lateral accretion surfaces. We infer that layers of cave earth built up successively in the external bend. However, the external bend of a normal stream would be subject to erosion, not deposition.

b We note Dr Ford's observations in The Vestibule of *"layers of scree and cave earth"* where walling has occasionally collapsed. Our GPR sections in The Vestibule also show contrasting layers. These GPR image-layers are cut by distinctive patterns that we can match or confidently attribute to human excavation. Since our 2D-GPR profiles lay along the benches, parallel to the cave axis, we are not yet able to demonstrate lateral accretion surfaces under the ropewalks in the main part of The Vestibule. However, if the result (2a) from the smaller bank of cave earth between the steps and Lumbago Walk were applicable to the rest of The Vestibule, then the whole of the cave fill would appear to have been accreted laterally.

3 **Cave sediments in the context of cave morphology:**

a As the streamway enters The Vestibule (2a), the bank of cave earth lies up-dip and the stream lies down-dip, albeit with considerable re-engineering (the channel is now confined by a man-made drainage system). So, clearly, the Peak Cavern stream is eroding down-dip and depositing up-dip as it enters The Vestibule. Geological dip and sub-parallel cave formation appear to have dominated the depositional processes. Although the natural flow of a meandering current normally controls depositional processes, streamway sinuosity appears to have been overruled by substrate dip in this case.

b If that applies to The Vestibule as a whole (2b), then the oldest cave fill would lie up-dip, beneath the anticlinal crest and the cavern-initiating avens. The youngest cave fill would lie down-dip, towards the modern Swinehole on the northeastern flank of the anticline and towards the apron-reef.

So, were the cave erosion and cave earth deposition processes linked or independent? Did cave earth deposition follow erosion down-dip, or was the cave fully developed before cave earth deposition commenced?

We had supposed, as does Dr Ford in his comment, that: "Until the Swine Hole became operative, sheets of these sediments would spread right across the Vestibule, with the oldest at the bottom and not confined to beneath levels 5 and 6." Behind that supposition is the implicit assumption that the rockhead beneath the cave earth was roughly level. Contrast observation (1c). Since we found the rockhead of the true cave floor to dip sub-parallel to the roof, sub-horizontal sedimentary layers would be constrained to either onlap or off-lap the floor. They would not have been able to "...spread right across the Vestibule".

From the observed lateral accretion, (2a), we infer that the cave earth deposits off-lapped the rock floor. Two off-lapping mechanisms could be considered, the first being dependent on, the second independent of the progress of down-dip cave erosion:

- Either the sediments were deposited successively, remaining behind and accreting, as The Vestibule was excavated down-dip.
- Or the sediments were deposited in an existing cavern and accreted as successive stream flood levels waned.

Either mechanism could be related to a larger scale, external, erosional event. Ford (1996) suggests that, as the Edale Shales of the Hope Valley were progressively eroded during the Pleistocene, *"... water levels fell within the limestone and streams ran along the floors of old phreatic tubes."* Similarly, the walls of his fascinating Vauclusian spring (op. cit.) could have eroded to form the present gorge. That too would have lowered water levels progressively within The Vestibule *"Hall"*. As is common in geology, our proposed mechanisms may be plausible and sufficient, but they are not necessarily mutually exclusive, closed or complete.

We think that we have found an important result, even though it cannot yet be demonstrated conclusively. The surprise comes because; with Dr Ford, we would normally expect lower sedimentary levels to be older. On the contrary, we now predict that the lowest deposits in The Vestibule should be the youngest and the highest should be oldest. Consider the fact that river terraces behave in a similar top-down manner. Ford (1999) briefly reviews the work that has been done on the relationship between glacial tills; river terraces, drainage and the evolution of the Castleton cave system. The importance of our result would come from a redirection of investigative effort. We are delighted that such a redirection would be compatible with several of Dr Ford's own earlier observations, results and suggestions.

Aside from the above discussion, Dr Ford mentions finding a boar's tooth at the entrance to The Swinehole. Is it possible to discriminate between the tooth of a medieval farm animal and that of a wild boar? Peak Cavern cave guides say that The Swinehole is so named because the resident rope-makers kept their pigs there, see for example Tomlinson (1991, 2001).

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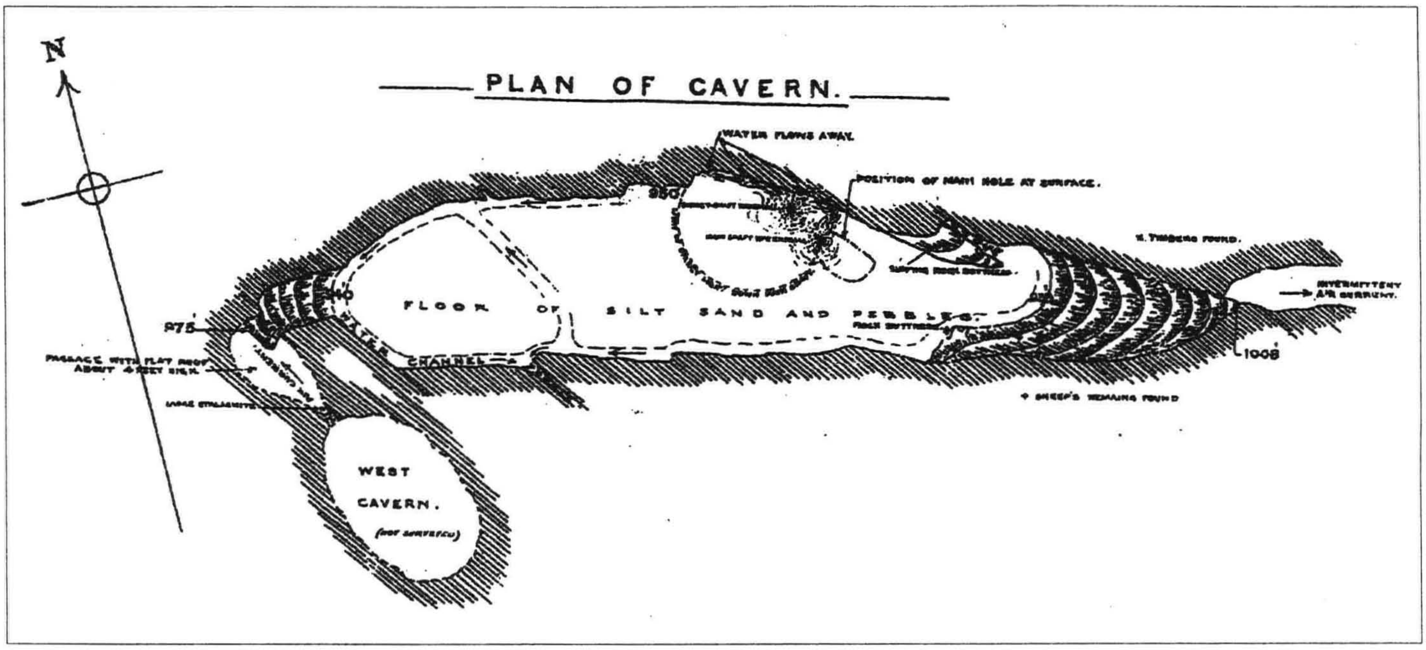

Figure 1. *From Calvert 1900*

Dear Editors

In Dr Craven's recent letter [*Cave and Karst Science*, Volume 29, No.3, p.135], in which he listed depths given for Gaping Gill in publications dating from 1872 to the present day, and apparent decrease in the shaft depth is recorded. Dr Craven suggests the possibility of flood-borne boulders raising the floor of the Main Chamber during the last century. If, however, changes in the sediment distribution in the Main Chamber floor during the last century are considered (Figs 1–3) the main change is the loss of large volumes of fine sediment. Rather than a decrease in the shaft depth this loss of fine-grained material suggests that a probable increase in depth has occurred. Dr Craven's letter illustrates the pitfalls of relying on the accuracy of historical references, commonly written to promote tourism in the area, rather than measuring the depth yourself!

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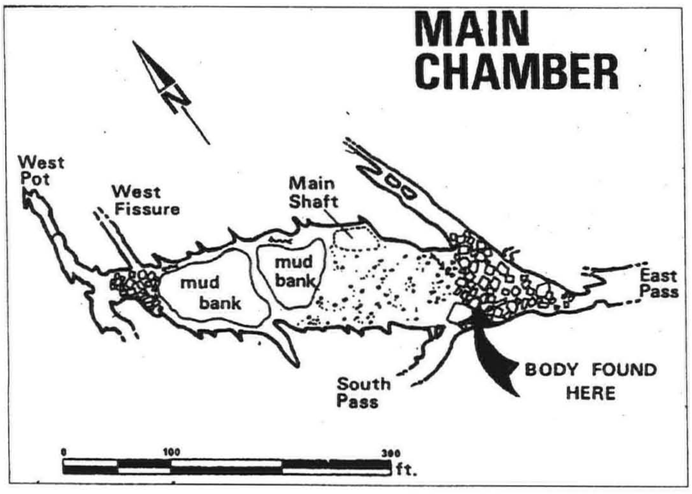

Figure 2. From Beck 1984

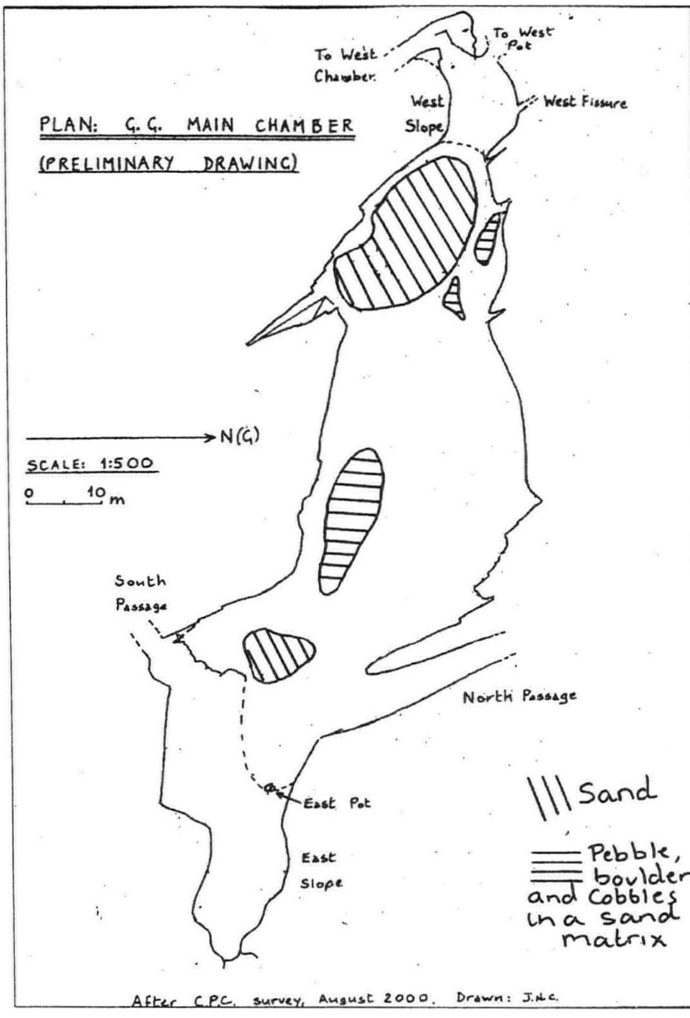

Figure 3. *Survey from Cordingley 2001 Sedimentological details by Murphy and Allshorn 2003.*

ABSTRACTS FROM KARSTOLOGIA FOR 2002

Karstologia 39 (2002)

Mapping of a palaeokarst: The example of the "Clypot" quarry (Hainaut Province, Belgium)

Méthodes et éléments de cartographie d'un paléokarst. L'exemple de la Carrière du Clypot (Hainaut, Belgique), p. 1-8
Yves and Gilles Quinif

Palaeokarsts have been often considered like geological objects different from present karst systems, which can be explored partially by speleological means. But it is obvious that their genesis has not been different from the genesis of the Neogene karst systems, in same environmental conditions. The study of palaeokarsts has a great importance for the comparison with present systems. Moreover, they conserve continental sediments that generally disappear, but with the possibility of dating by marine trangressive series which cover the palaeokarsts or by absolute dating like K-Ar or Ar-Ar on glauconite or ferriferous illite. We present here an interesting example of palaeokarstic features in a quarry where the workings permit mapping of those features. This map constitutes the basis for future studies; it has shown different types of morphological features and deposits, their geometrical relations and their genetic links. We have (i) ghosts-rocks and pseudo-endokarsts, which result from the alteration *in situ* of the host-rock with formation of residual alterite. Those features organise like linear channels along tectonic fractures. Some channels can joint together in great "pockets". At the top of the limestone formation, (ii) palaeo-clints develop under the transgressive cover where we find pebbles and sands. Finally, (iii) endokarstic galleries can come from an autonomous hollowing ("classical" karst) or from old ghost-rocks, which become partially empty due to new hydrological activity.

Karsts of New-Zealand

Les karsts de Nouvelle-Zélande, p. 9-22

Jean-Noël SALOMON

New-Zealand presents numerous karsts developed as well in ancient rocks (Palaeozoic) as in recent ones (Oligocene). The stretching in latitude of the land, the high vegetal biodiversity and the strong rainfalls explain the importance of the karst development and the variety of the morphologies. Endokarsts are well developed, but many are still to explore. The polygonal karst of the King Country (North Island) on one hand, and the karsts of the Marble Mountains (South Island) on another, are the most interesting. The possibilities of crossing of numerous datings (dendrochronology, speleothems, volcanism, etc) and the location of these karsts in the Southern Hemisphere provide to these last exceptional palaeo-environmental recording systems.

Contribution to Calern karst hydrogeology knowledge (Alpes-Maritimes, France). Dye tracing in Moustique aven

Contribution à la connaissance de l'hydrogéologie karstique du plateau de Calern (Alpes-Maritimes, France).Traçage à l'aven du Moustique, p. 23-26 + map

Marie Martin and Philippe Audra

Within the framework of a hydrogeological study of the plateau of Calern, an artificial tracer has been put into the underground river of the Moustique aven. The dye has reappeared at a single outlet: the Bramafan spring. This emergence lies to the east of the Calern plateau, in the Loup's valley, at the contact between Jurassic limestones and Triassic impervious formations. It appears to be the main outlet of this vast karst area.

Quaternary sedimentary archives of the Sous les Sangles Cave (Lower Bugey, Southern Jura, France); paleo-climatic and seismo-tectonic evidences

Les archives sédimentaires quaternaires de la grotte sous les Sangles (Bas-Bugey, Jura méridional, France). Indices paléo-climatiques et sismo-tectoniques, p. 27-46

Vincent LIGNIER and Marc DESMET

The "Sous les Sangles" Cave is located in the southern part of the Jura Mountains at the front part of the northwestern alpine tectonically active massifs. This region was covered by alpine and jurassian glaciers during the Last Glacial Maximum. An old gallery contains stratified fluvial and moraine injection, covered by a 3.5m-thick deposit of finely laminated silty carbonate and clays. Sedimentological investigation reveals several periods of different water flow depending on glacial and interglacial periods. The upper finely laminated sediments correspond to the end of the last glacial maximum according to the exokarstic equivalent of the Cerin lake and the U/Th ages obtained with speleothems. Spectral analysis (using Fourier methods and pass-band mapping techniques) on grey-level raw data have been used on the "Sous les Sangles" sediment. The main result shows evidence of a cyclic deposition according to the three main periodicities recognised through the 1.5m top sequence. The laminated material is affected by plastic and brittle deformations. The entire deposit is characterised by (1) vertical faulting without apparent dominant relative movement, which can be interpreted as tension faults; (2) an associated soft and brittle deformation similar to thin skin tectonic at centimetre scale affecting the base of the deposit and testified to gravity reworking, which could correspond to discrete seismotectonic activity; (3) brittle deformations associated with fluid escape patterns occurring at two specific levels along the vertical faults, confirming the earth tremor existence suggested by several broken speleothems. These observations are well supported by the geodynamic and tectonic framework of this part of the Jura Massif, which reveal an actual uplift of several millimetres/year, especially in this part of the "Cluse des Hôpitaux" cross valley. Numerous historical earthquakes have been documented in this area. Microtectonic study of the cave shows dominant inverse and strike-slip structures correlated to the general tectonic framework.

Noodsberg Cave, a cave in Kwazulu-Natal Province quartzites (South-Africa)

Noodsberg cave, une grotte dans les quartzites de la province du Kwazulu-Natal, Afrique du Sud, p. 47-50

J E J MARTINI and C G A MARSHALL

In the Kwazulu-Natal Province, South Africa, a wet cave of karst origin is described in Ordovician quartzite. Like in a number of other caves of the same type in South Africa and in the rest of the world, the development proceeded in two phases: (1) weathering of hard quartzite into a friable rock by quartz solution along crystal boundaries; (2) speleogenesis proper by piping in the loose material produced in phase 1. A particularity of this cave is the presence of a conglomerate with a sandy-clay matrix at the hanging wall, more resistant mechanically, which prevented the early collapse of the cavities.

Pamukkale (Hierapolis): An outstanding site of hydrothermal travertines in Turkey

Pamukkale (Hiérapolis) : un site de travertins hydrothermaux exceptionnel de Turquie, p. 51-54

Jean NICOD

These travertines result from the deposit of carbonates near the hydrothermal springs, on the main active fault zone on the northern border of the Denizli Basin (W Turkey). Their high mineralised water, rich in CO_2 of geothermal origin, accumulate limestone in the fissure ridges and in the cascades on the front of the old travertines balcony, building up in it flowstone and rimstone dams. This site is particularly important as much for the archaeological and palaeoenvironmental researches as the palaeoseismic and neotectonic regional data.

Karstologia 40 (2002)

Inventory of coastal and submarine springs in the Alpes-Maritimes (France)

Les karsts littoraux des Alpes-Maritimes : inventaire des émergences sous-marines et captage expérimental de Cabbé, p. 1-12

Éric GILLI

Experimental catchment at the Cabbé spring. Several submarine freshwater springs are present on the karst shore in the Alpes-Maritimes (France). Salinity and conductivity measuring coupled with GPS location has permitted to inventory these springs. Three main springs have an average flow around 500 l/s. A balance on inland and offshore springs allows to explain the deficit observed on karst units of the Arc de Nice area. A dam was built in the submarine karst spring of Cabbé Massolin (Roquebrune-Cap-Martin, France) to study the effects of an artificial augmentation of the pressure on the salinity of a karst aquifer. Trials in low and high water levels show the impossibility to increase the pressure. The presence of several springs and the important jointing of limestone don't allow a sufficient impermeability of the dam site. Nevertheless, the salinity decreases, due to the physical separation between the two kinds of water.

Montfat cave as a stage in the entrenchment of the Meuse Valley

La grotte de Montfat : un jalon dans l'évolution de la vallée de la Meuse, p. 13-18

Yves Quinif

Perched caves in the side of valleys are precious indicators for the speed of the rivers incision. Clastic and chemical deposits give information about the sedimentation in relation with the tributary rivers. In particular, speleothems are younger than the drying of the galleries and give a limiting age to the palaeothalweg situated at this altitude. Moreover, clastic deposits are also precious indicators about the processes linked to the incision. The cave grotte de Montfat in Dinant, along the Belgian Meuse river, constitutes a part of an under thalweg karst system and contains sedimentary evidence which proves that the Meuse river had left the altitude of 45m above the present altitude before 400,000 years BP.

Topographical and electromagnetic positioning of a deep sump: Qattine Azar (Lebanon)

Modes de positionnement topographique et électromagnétique d'un siphon. Exemple du Qattine Azar (Liban), p. 19-26

Paul COURBON

Wasting, pollution, climatic modifications, urbanisation and demographic explosion will generate a serious water scarcity in many countries, among them, the Middle East. In Lebanon, after the discovery of an important underground river, is established a water impounding project, needing a 280 meters drilling through limestone. The author describes the topographic method used to survey the cave and to set up the future drilling. The survey has been confirmed by an electromagnetic positioning ARCAS (radiolocation) perfected by Joan Erra. The accuracy of the two determinations is estimated.

The karst and the underground quarries of Barrois: 150 years of relationship Man/Nature

Le karst et les carrières souterraines du Barrois : un siècle et demi de relations Hommes / Milieu, p. 27-38

Stéphane JAILLET, Jean-Pierre DEPAQUIS and Claude HERBILLON

The exploitation of the "Savonnières stones" in the Barrois area (Lorraine/ Champagne) was made during the last 150 years by underground quarries. The development of these quarries is estimated between 300 and 350km today. During this exploitation the cutting of a palaeokarst ("viailles") and an active karst (pits), disturb the exploitation, obliging the quarrymen to fill or to avoid the karst. Karst water added a supplementary constraint but gave a good contribution for the exploitation of the quarry. In the final, karst and waters are for the quarry a minor set of constraints and potentialities, which structure a spatial organisation of this anthropological underground space. Quarry, karst, geology and fauna constitute a unique underground patrimony in Lorraine, which merits interest and adapted protection.

Turbidity, as an indicator of perturbed functioning of karst geosystem of Beni Mellal Atlas (Southern Middle Atlas, Morocco)

Turbidité, indicateur du fonctionnement perturbé du géosystème karstique de l'Atlas de Beni Mellal (Moyen Atlas méridional, Maroc), p. 39-44

Yahia EL KHALKI and Abdellatif HAFID

The turbidity of karst springs is a general phenomenon of anthropized Mediterranean limestone mountains. But actually we note that it has been somewhat neglected by karstologists. In other respects, it represents a pertinent indicator of the dysfunctioning of karst geosystems that have undergone a strong impact of human action. Cloudy waters distinguish Asserdoune spring, as the main outlet of the Liassic aquifer. After dry periods, followed by violent storms, water turbidity can exceed 2000 JTU (Jackson Turbidity Unit). Three factors are united to favour the turbidity water: 1) The climate: autumn rain storms, which follow a long dry season, fall on soils that are dry and badly protected by vegetation. They cause an important mobilisation of fine particles, which infiltrate the epikarst. 2) Inner karst structure: the advance of cloudy water toward Aïn Asserdoune spring is made easier by developed drains. Dye tracings have shown that the time of transit waters is comprised between 24 and 72 hours. 3) Human impact: the agriculture modernisation of Tadla plain (140 000 ha irrigated) was accompanied with a violent action on forest resources of Beni Mellal Atlas. Clearings induced a large stripping of soils, which caused the excavation of covered karrens. Furthermore, the dysfunctioning of the karst geosystem of Beni Mellal Atlas is translated by the hindrance of present travertinisation downstream of springs.

Measures of atmospheric O_2 and CO_2 in the cave system of Fanges-Paradet (French Eastern Pyrénées)

Anomalies des teneurs en oxygène atmosphérique mesurées dans le réseau Fanges-Paradet (Aude / Pyrénées-Orientales), p. 45-50

Bernard OURNIÉ and Jean-Michel OSTERMANN

In the cave system of Fanges-Paradet and in several caves of Region Languedoc-Roussillon (in Corbières mountains, High Aude Valley and Larzac Causse), successive measuring of atmospheric O_2 and CO_2 contents have been taken in 1997, 1998 and 2001. The aim of the paper consists in getting more accurate composition of endokarstic atmosphere. Abnormally low contents in O_2 appear, especially in the deep parts of "puits de l'Ours", "grotte de la Coume dels Adoutx" and "grotte TM 71". Several explanations may justify these "anomalies".

Identification of surface karst features on a XVIth century Gran Sasso d'Italia map (Abruzzes, Central Italy)

Identification de formes karstiques de surface sur une représentation cartographique du Gran Sasso d'Italia du XVIe siècle (Abruzzes, Italie centrale), p. 51-54

Ezio BURRI

In August 1573, Francesco De Marchi climbed for the first time the Corno Grande, the highest Gran Sasso peak, which is the highest limestone massif of the Apennines. From this exploration, he let a detailed description that was published after his death in 1599. A map, where some karst features are clearly mentioned and particularly some small karst lakes illustrated this description. Some other karst features, such as the large Campo Imperatore polje are described. In the same time of this climb he also explored Grotta a Male, one of the main caves of the area and gave a scientific description of it.

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.

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, STI7 9BT, UK.