Cave Science

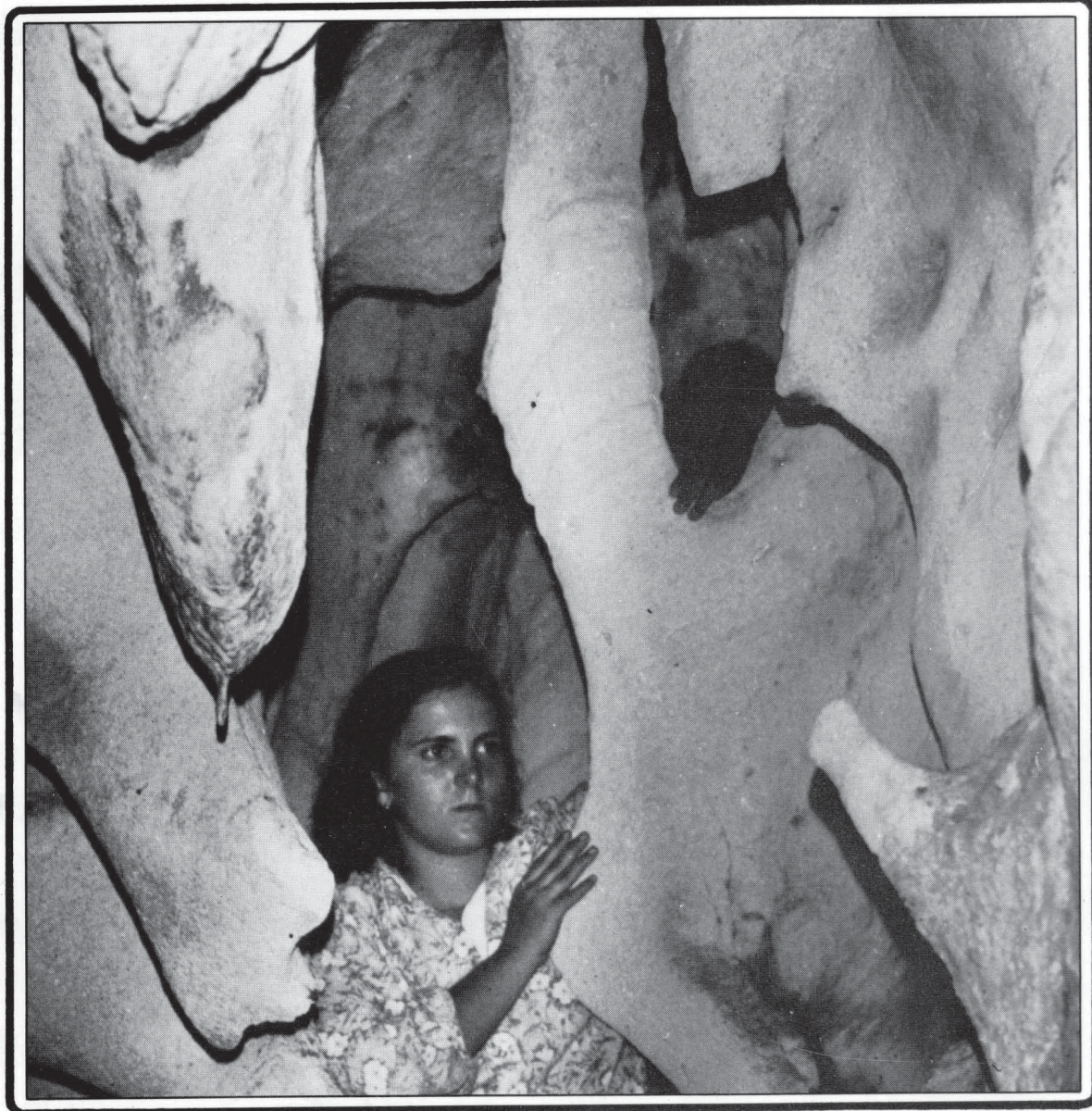The Transactions of the British Cave Research Association covers all aspects of speleological science, including geology, geomorphology, hydrology, chemistry, physics, archaeology and biology in their application to caves. It also publishes articles on technical matters such as exploration, equipment, diving, surveying, photography and documentation, as well as expedition reports and historical or biographical studies. Papers may be read at meetings held in various parts of Britain, but they may be submitted for publication without being read. Manuscripts should be sent to the Editor, Dr. T. D. Ford, at 21 Elizabeth Drive, Oadby, Leicester LE2 4RD. Intending authors are welcome to contact either the Editor or the Production Editor who will be pleased to advise in any cases of doubt concerning the preparation of manuscripts.

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Acknowledgements: Anyone who has given a grant or helped with the investigation, or with the preparation of the article, should be acknowledged briefly. Contributors in universities and other institutions are reminded that grants towards the cost of publication may be available and they should make the appropriate enquiries as early as possible. Expedition budgets should include an element to help publication, and the editor should be informed at the time of submission.

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Cave Science

Contents

Cover: Complex phreatic roof pendants in Lighthouse Cave, on San Salvador Island in the Bahamas. By James Carew.

Speleogenesis in the Limestone Outcrop North of the South Wales Coalfield: the Role of Micro-organisms in the Oxidation of Sulphides and Hydrocarbons

T. K. BALL and J. C. JONES

Abstract: The controls on speleogenesis are reviewed and it is argued that bicarbonate dissolution alone is inadequate to explain the tight stratigraphical positioning of primary solution tubes. Geochemical studies of confined aquifers indicate that phreatic fissure water is vertically zoned with an oxidised upper portion underlain by a reducing zone. Flow rates are greater in the oxidised zone. Mechanisms for limestone dissolution are considered, and it is concluded that oxidation reactions involving sulphide minerals and hydrocarbons, catalysed by micro-organisms, and producing both $CaSO_4$ and CO_2, are operative in opening up the shallow (oxidised) phreas. Sulphide is inhomogeneously distributed in the limestone succession and controls the positioning of the main passages. These are largely confined to the impure sulphidic limestones and seldom found in the purer oölites.

The general controls of cave development in the limestone outcrop around the northern edge of the South Wales coalfield are well recognised in descriptive terms, although the reasons are only imperfectly understood genetically. Cave passages tend to align along trends which accord to those joint and fault directions which approximate most closely to the dip and to others which are aligned most closely to the strike (Ball 1961, 1980; O'Reilly and others 1969, Weaver, 1973; Coase, 1977; Smart and Gardener, 1989). Sometimes, over very large areas, it can be recognised that the primary solution tube for passages is confined to specific beds, and although the same bed may change in character from place to place (lateral facies variation), often the bedding control is stronger than the facies changes (Glennie 1948, 1950; Ball, 1961, 1980; Jenkins, 1963, Coase, 1977; Christopher and Charity, 1978; Smart and Gardener, 1989). There is evidence for water table control in the extensive cave series that are located over a restricted height range (Railton, 1953, 1958; Ball, 1961, 1980; Smart and Gardener, 1989).

Table 1 and Figure 1 summarise the rock succession over the north crop (Barclay, Taylor and Thomas, 1988; Barclay, 1989, Lowe, 1989). The main limestone of the area is confined between two relatively impermeable sandstones; the underlying Old Red Sandstone of Devonian age and the overlying Namurian Basal Grit. The caves at the Tawe Valley and westward are virtually confined to the Dowlais (=Cil-yr-ychen) Limestone. In this area

the overlying Penderyn Oölite is thick, massively bedded and relatively pure Further east in the Nedd/Mellte area much of the cave development is in the lower beds of this oölitic formation where the sequence is much more impure, being characterised at outcrop by rubbly and shaley tops to the massive individual oölitic beds. Most of the Nedd Fechan system however is contained in the Dowlais Limestone. Even further east the Agen Allwedd/Eglwys Faen system shows initial development mostly along a particular bed in the Blaen Onneu Oölite (the Concretionary Band of Jenkins, 1963). The Blaen Onneu Oölite is a sub-member of the Abercriban Oölite in this area.

If carbonate dissolution were the only controlling feature then one would expect that passages would tend to be uniformly distributed throughout the sequence, allowing for mechanical control due to faulting, and channelling of solutional activity by impervious beds and at contacts. The location of passages in the more impure beds almost to the exclusion of the purer massive oölites, implies that there must be some other feature of the limestones that controls the positions of the passages within the sequence, other than simple carbonate dissolution.

The chemistry of the genesis of cave passages in the deep phreatic zone is one of the greatest problems confronting speleology. Whereas it is a fairly simple matter to explain passage development in the vadose and shallow phreatic environments by standard explanations invoking the dissolution of calcium

Table 1

|	Formation	Rocktype	Thickness
NAMURIAN	Basal Grit	Quartzite	30m
	unconformity		
VISEAN	Upper Limestone Shales	Calc mudstones thin limestones	0-3m
	Penwyllt limestone	Dark cherty, sandy, bituminous limestones	0-20m
	Penderyn Oölite	Light coloured oölite	0-33m
	Cil-yr-ychen Limestone (=Dowlais Limestone)	Variable, mostly bituminous, bioclastic	0-120m
	unconformity		
	Llanelly Formation	Variable, sandstones oölites and clays.	0-8m
	unconformity		
	Abercriban Oölite	Pale grey oölites with micrites	0-28m
	Lower Limestone Shales	Thin basal conglomerate limestones and shales	5-23m
	unconformity		
DEVONIAN	Grey Grits and Conglomerates		
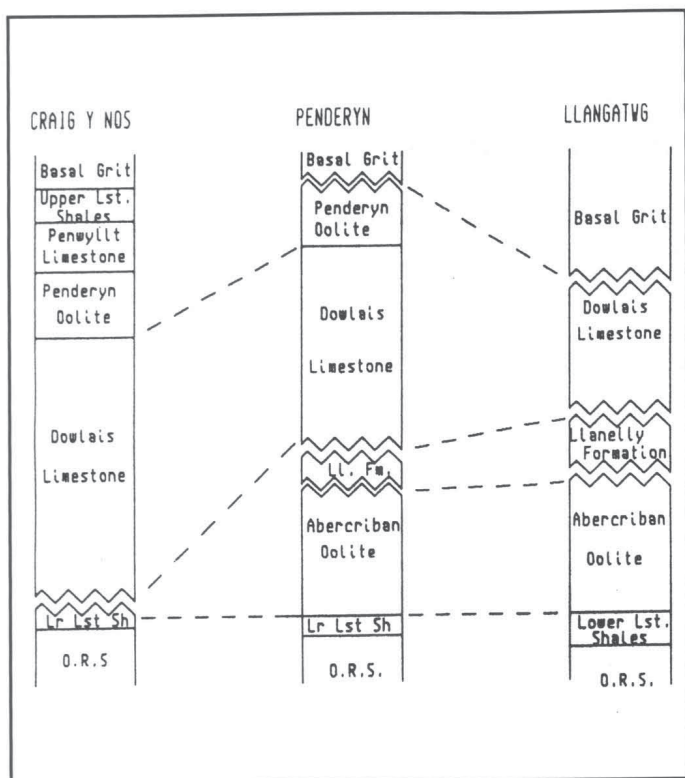
Figure 1. Schematic figure showing the changes in stratigraphical detail for localities on the North Crop of the South Wales coalfield. Not to scale.

carbonate by carbonated groundwaters, the initiation of solution in the deep phreas can be difficult to envisage.

MECHANISMS OF LIMESTONE DISSOLUTION

Explanations concerning limestone dissolution are based firstly upon the solution of carbon dioxide in rain or soil water, and the reaction between the dilute carbonic acid and limestone to form calcium bicarbonate, which is much more soluble than the original limestone. The degree of dissolution is therefore high in the vadose zone but the waters become increasingly saturated as they make their way down to the water table. Saturated waters can become aggressive to limestone again if the physico-chemical conditions change e.g. there is a temperature change or the carbon dioxide content of the cave air increases, but the major increase in aggression is probably due to the mixing of waters which have become saturated with calcium bicarbonate in response to different starting conditions (the "Mixing Corrosion" of Bögli, 1963, 1980).

Figure 2 gives the saturation curve for calcium bicarbonate in water at various temperatures (after Holland and others, 1964). Above the curves the waters are oversaturated with CO_2. The figure shows the change in composition of carbonated water during equilibriation with calcium carbonate, at various temperatures, in the presence (A) or absence (B) of a vapour phase and for air in which there is e.g. 1% and 10% carbon dioxide. Consider water having equilibriated with the vapour phase and then being brought into contact with calcium carbonate, depending upon whether it remains in contact with the vapour phase (as it may in the vadose and the upper phreatic zone) or is immediately removed from contact (as it would if plunged into the phreas or is otherwise confined), the composition of the resulting aqueous phase is represented by path A or path B. Clearly it is in the vadose environment that corrosion of limestone is most efficient.

Consider two saturated compositions represented at "C" and "D" on the 10°C curve. C and D compositions can only mix on a straight line and the mixture will therefore be oversaturated with CO_2 which would then tend to be liberated and corrode more limestone, whilst retaining the inherited $Ca(HCO_3)_2$ in solution. This process of Mixing Corrosion takes place mainly in the lowest vadose or upper phreatic zone where mixing is most prevalent and the availability of water is greatest. The enhanced corrosion at the level of the water table explains many of the features observed in South Wales caves where there are extensive cave series developed over a relatively constant height. It is difficult to envisage such reactions taking place deep within the phreas, for one thing there is by definition no air surface by which the carbon dioxide may be replenished in the water, and any increased level of aggression due to the mix corrosion processes will have diminished.

The geomorphology of South Wales indicates that there has been a stepwise (pulsatory) decline in the erosional (base) levels for the resurgences in the Tawe Valley area (Jones,1939). With the exception of Ogof Ffynnon Ddu, where it is possible that the stream has invaded an already existing system, there is no evidence that the water table has migrated upwards, only declined. Furthermore there is strong evidence that there are certain beds within the limestone sequence which are more conducive to cave development than others. Much demonstrable vadose development of cave passages is by trenching (canyon development), and by collapse, the extent of which is independent of stratigraphical position except in that the position of the primary solution tube is bedding controlled. The initiation of cave passages in the phreatic zone, by the development of bedding controlled solution tubes, must thus be postulated before further development in the shallow phreas or the vadose zone of the larger cave passages. The problem is therefore to consider which mechanisms may give rise to such initial cave passages, the positions of which are controlled by the stratigraphy. Three mechanisms by which cave passages may be initiated are considered:
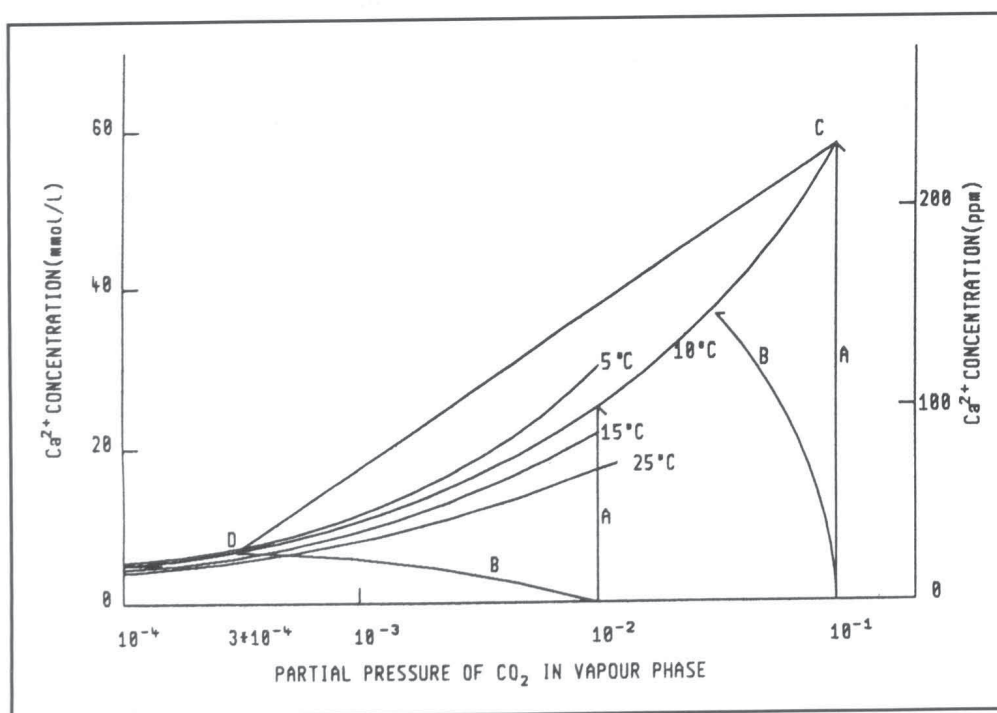
Figure 2. After Holland and others (1964), this figure shows the changes in composition of carbonated water during equilibriation with calcite at various temperatures and in the presence (A) and in the absence (B) of a vapour phase. For further details see text.
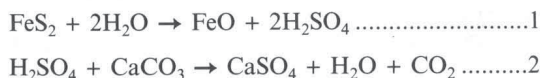
1. Calcium bicarbonate solution. There is the situation in the deep phreas where the waters are saturated with respect to calcium bicarbonate. This does not mean that there is no dissolution but that dissolution and deposition are in balance. In general however it is not envisaged that there would be much cave development as a result, nor would there be any reason why dissolution would be selective to certain beds.

2. Dolomitisation. Dolomite is a double carbonate of calcium and magnesium. This rock type may be primary, in which case it was precipitated either on the sea floor as dolomite, or as the result of dolomitisation of calcium limestone shortly after deposition. There is a pair of dolomite beds, separated by a normal limestone, extensively developed in the Tawe Valley area near the base of the Dowlais Limestone, and this outcrops in Ogof Ffynnon Ddu and in Dan yr Ogof. In general dolomite seems to be more resistant to erosion than the calcite limestones (see e.g. Charity and Christopher, 1977).

Secondary dolomitisation affects the hardened limestone and when complete (i.e. 1:1 Ca:Mg) there is a volume reduction of some 14%. The production of voids by this process can provide channels for moving groundwater and passages may develop. Cave passages spatially related to secondary dolomitisation along faults and joints can be observed in the Llygad Llwchwr/Mynydd Du area but in the Tawe Valley neighbourhood this process does not seem to have been operative on a large scale. In the calcite limestone sandwiched between the double dolomite mentioned above, there are narrow dolomitised joints along which corrosion has been more intense, but these cannot be regarded as a major contributor to cave development.

3. Gypsum formation. In two of the largest caves in South Wales (Ogof Ffynnon Ddu and Agen Allwedd) there is extensive deposition of gypsum in the passages. Gypsum (and its clear variety selenite) is a hydrated calcium sulphate. It is much more soluble than calcite (or dolomite) and is only found in passages that are now dry. In both caves the main concentration of gypsum is found mainly, but not exclusively, in the vicinity of shale horizons and the association of gypsum or selenite with iron oxides gives an indication as to the formation of the mineral.

A commonly accepted mechanism for the formation of gypsum is the oxidation of pyrite (FeS_2) by a variety of mechanisms, eventually to give iron oxides and sulphuric acid. The acid then reacts with the ubiquitous calcite to give gypsum (Figure 3).

$$FeS_2 + 2H_2O \rightarrow FeO + 2H_2SO_4 \dots\dots\dots\dots\dots 1$$

$$H_2SO_4 + CaCO_3 \rightarrow CaSO_4 + H_2O + CO_2 \dots\dots 2$$

There is a strong spatial association in support of this mechanism with, in Agen Allwedd, a frequent thin glaze of selenite over a thin wall covering of red iron oxides. In the Rawl Series of Ogof Ffynnon Ddu the association is equally clear but the colours are not so bright.

One of the reaction products in equation 2 is CO_2 which is of course available for other reactions, including further dissolution of calcite by producing calcium bicarbonate. When consideration is taken of the mass balances and of the different densities of the minerals, one volume of totally oxidised pyrite can result in the dissolution of a further six volumes of calcite (see for example, Pohl and White, 1965).

The reaction of sulphuric acid with limestone to produce caves has been discussed by a number of authors who all suggest an extrinsic source for the sulphur. For example the caves of the Guadalupe Mountains in the U.S.A., which contains the famous Carlsbad Caverns, have been attributed to such a mechanism. Hill (1987) proposed that H_2S, a common component of natural gases in the adjoining oil-producing sedimentary basin, migrated into the limestone. There it was oxidised to sulphuric acid, and produced gypsum and CO_2. These reactions were proposed to explain the evolution of such large exotic chambers as those in Carlsbad Cavern. The very large deposits of gypsum in these caves are thought to result from precipitation from warm saturated gypsiferous solutions on being cooled in the cave environment (Bretz, 1949 and Good, 1957). However Palmer et al. (1977) have shown that at least some of the gypsum had been produced by metasomatism of the limestone prior to the development of the caves.

Egemeier (1981) proposed similar reactions to account for the Kane Caves, Wyoming, but deduced that the sulphide originated from a thermal spring. Using oxygen and sulphur isotope geochemical studies, Van Everdingen and others (1985)
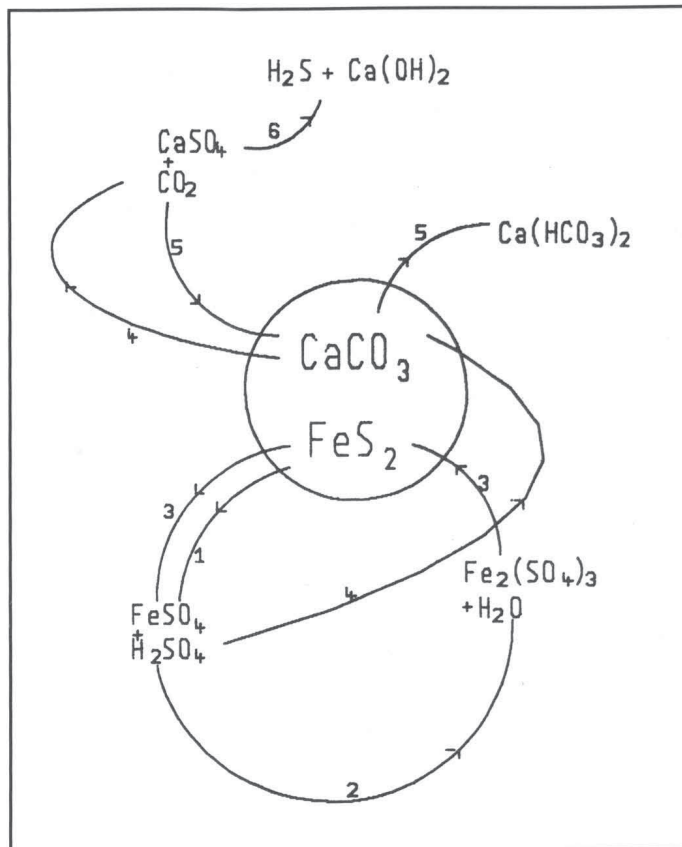

Figure 3. Schematic diagram showing reactions and reaction products for oxidation of pyrite in a calcareous matrix.

emphasised the importance of chemical corrosion processes by which H_2SO_4 produced by oxidation of reduced sulphur species reacted with a hot spring travertine deposit in Alberta, Canada. They concluded that microbiological reduction of dissolved sulphate in the hot springs, which also gave rise to the travertine, produced sulphide species which were subsequently re-oxidised to sulphuric acid and a considerable fractionation of the oxygen and oxygen isotopes resulted. Yonge and Krouse (1987) carried out sulphur isotopic analysis for minerals from the Castleguard Cave, Canada and concluded that bed-rock pyrite was a possible source for most of the gypsum.

Krothe and Libra (1983) concluded that microbial activity was implicated in the fractionation of sulphur isotopes during the deep flow of water in a limestone aquifer in Indiana. Sulphate (from gypsum) was reduced to sulphide and the S-34 isotopic content increased over that found for the local gypsum deposits.

Morehouse (1968) proposed the sulphuric acid reaction to account for caves in the heavily sulphide mineralised Galena Dolomite of Iowa. Stenner (1969) ascribed some carbonate aggressiveness to sulphuric acid from shale sources.

Since the sulphuric acid reaction is largely independent of carbonate dissolution controls it may provide a very potent method for initiating cave passages deep within the phreatic zone, and with the strong probability of the mechanism having been active in our area it is therefore worth considering whether there is sufficient sulphide present and consider likely controls over its oxidation.

AQUIFER HYDROCHEMISTRY

Modern studies of the chemistry of limestone aquifers, with data derived from water extraction boreholes, indicates that the phreatic water is zoned (Edmunds and others, 1984, 1987). Water near the vadose/phreatic interface is often still aggressive to limestone as one might expect. The water however soon becomes saturated with calcium bicarbonate with depth although incongruent solution can occur. Underlying this calcium bicarbonate saturation front is a zone of variable thickness where there is still dissolved oxygen. This is in turn underlain by a zone of oxygen depletion (Figure 4). The interface of the two zones is taken at the oxidation reaction potential of the ferric to ferrous transformation. Above the interface therefore pyrite or marcasite (FeS_2) will tend to be unstable or metastable whilst below, it will tend to be quite stable.
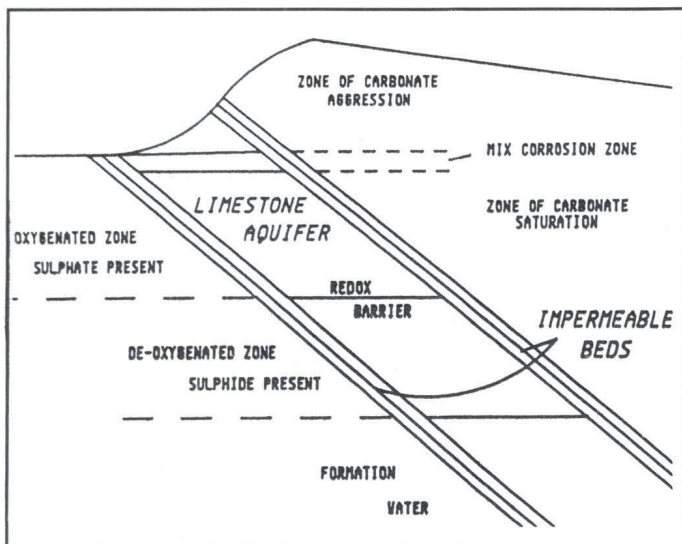
Figure 4. Illustration of chemical zonation in fissure held water in confined aquifers. Not to scale.

Edmunds (1973) and Edmunds and Walton (1983) have carried out a survey of the Lincolnshire Limestone aquifer which provides a useful model for the North Crop of the South Wales Coalfield. In both areas the limestone is sandwiched between relatively impermeable beds. In South Wales a stepwise (pulsatory) decline in base level is deduced by considerations of longitudinal stream profiles (thalwegs) in neighbouring stream sections, and by cave passage frequency in accord with these horizons (Ball, 1980). In Lincolnshire pumping of the aquifer has led to a decline in the water table, and data from the wells gives information on the chemical changes which have occurred during the period of the century or so that the pumping has taken place. Groundwater movement is almost entirely by fissure flow along bedding and joints. Transmission rates are high and especially so near outcrop, where true karstic conditions are found, and in the well developed fissure system which is important in the top few metres of the aquifer (Rushton and Redshaw, 1979). As a result of the oxidation processes that have occurred in geological and historic times the permeability of the oxidised limestone is slightly greater than that in the reduced zone. Lawrence and Foster (1986) have shown that in the oxidised zone the groundwater can be divided into two regimes with the fissure water being mobile and oxidising, and with diffusive exchange of chemical components between the stationary, more reducing, micropore water of the limestone matrix. The oxidised zone surrounding the fissures is characterised by a marked reduction in the non-carbonate organic matter. These features seem to be general in confined aquifers elsewhere, e.g. in the chalk and the Permo-Triassic aquifers of Central England (Edmunds, Miles and Cook, 1984).

At even greater depth in the limestone, the remnants of formation (or connate) waters have been observed and are thought to represent the sea water which remained in the rocks following lithification but which has reacted with the limestone. The presence of $NaHCO_3$ and $CaCl_2$ in solution suggests an exchange reaction involving $NaCl$ from sea water and $CaCO_3$ from the limestone.

BIOGEOCHEMICAL ASPECTS

In a very comprehensive survey of the biospeleology of both Dan yr Ogof and Ogof Ffynnon Ddu, Edington (1977) noted the presence of a wide range of bacteria in water, including nitrogen fixing and nitrifying bacteria and especially *Thiobacillus sp.*, which in certain areas was recorded as being common. Microbes can also occur deep under the water table, for example, Edmunds (1973) and Lawrence and Foster (1986) record the presence of denitrifying bacteria in aquifers down dip to 23km from outcrop.

Certain types of bacteria or micro-organisms can oxidise sulphur compounds and deposit the sulphur outside the cell. The generic name *Thiobacillus* has been given to these microbes. They are primarily characterised by their ability to oxidise reduced inorganic sulphur, although certain types of *Thiobacilli* can also oxidise ferrous iron. They are capable of growing in the absence of organic nutrient by reducing carbon dioxide to provide nutrition, deriving their energy from reduced inorganic sulphur compounds (Pooley, 1986).

The energy released in the enzymatically catalysed transformations is converted into biologically usable energy for cell growth and maintenance. *Thiobacilli* require other nutrients for their metabolic function, such as nitrogen from ammonia and nitrates (produced by reactions in soils and carried into the groundwater), phosphorus (present as phosphates in the limestone), calcium, sulphur (from pyrite) and, of course, carbon dioxide an important source of carbon. Carbon compounds are absorbed from dissolved carbon dioxide and, as the microbes are aerobic, a quantity of dissolved molecular oxygen is also required. *T. ferrooxidans* is the major microbe to concern us here and will oxidise ferrous iron and sulphides. The rate of ferrous iron oxidation is more rapid than that for sulphur compounds. *T. thiooxidans* is another bacterium which may be involved in the oxidation of elemental sulphur.

Most of the information concerning the microbial reactions with minerals derives from laboratory studies of reactions concerned with leaching of ore minerals. The controls governing these reactions have been investigated empirically, but in order that these studies may be completed in a reasonable time schedule long term experimentation has been avoided. For our purposes we may well be interested in reactions which may take thousands of years to reach any significance. However laboratory controls are often very useful markers as to the potential of certain reactions in the natural environment.

The following factors are usually considered in laboratory experiments to assess the potential leaching rates for ore minerals (Pooley, 1986):

1) Nature of the mineral substrate.
2) pH control.
3) Microbial nutrients.
4) Temperature control.
5) Particle size.
6) The cycle of the leaching process.

Mineral or Bacterial Substrate

There are many minerals that can be oxidised by *T. ferrooxidans*. In an air/water medium it has been suggested that the controlling factor (primary determinant) of the reaction kinetics is controlled by the electrical potential of the mineral and by the electrical currents generated between minerals having different potentials. In neutral solution pyrite, sphalerite (ZnS), galena (PbS), and chalcopyrite ($CuFeS_2$) are all readily oxidisable. These minerals all occur in veins in the area accompanied by barite ($BaSO_4$) and fluorite (CaF_4) and of course by calcite Pyrite, however, is very common in the host rock, being found in small amounts in most limestone specimens from the Dowlais Limestone, one of the main cave bearing formations in the area. It is particularly common in the calcareous shale bands which are found throughout the sequence of this limestone, and especially so in two carbonaceous shale horizons which frequently sandwich the Honeycombed Sandstone, which forms a marker horizon near the base of the overlying Penderyn Oölite. Preliminary investigations indicate that organo-sulphide compounds are also present.

The *T. ferrooxidans* and related species are extremely common in soils and may be carried into the cave environment from this primary source.

pH Control

Most individual grains of pyrite are enclosed within a calcite matrix. Oxidation of pyrite alone results in the production of a basic ferric sulphate. Further precipitation of iron results in the pH dropping to values of 1.5 or below. As this is close to the pH value which limits bacterial activity the leaching will automatically come to a stop. In limestones, reaction with the ubiquitous calcite will help to stabilise the pH at a very much higher value although there are likely to be transient microdomains close to the pyrite surface where a low pH is to be found.

T. ferrooxidans works most efficiently at pH levels of 2.5 to 5 but it can operate with reasonable efficiency in the range 1.4 to 6.

Bacterial Nutrients

The major nutrient requirements have been dealt with above where it can be shown that most components can be supplied from the limestone or from the water.

The most important nutrient is nitrogen which is the basis for protein production. Ammonia ions are the best source of nitrogen for *Thiobacilli*. It is believed that *T. ferrooxidans* can also fix nitrogen directly, in the absence of a free air surface there is likely to be a copious supply of nitrogen from the air dissolved in the groundwater.

Nutrients are vital determinants of reaction rates as the solubilisation of metal is a function of cell division, which is in turn associated with the growth of the bacterial population. When growth ceases so does metal solubilisation. In commercial bacterial leaching experiments reactions may virtually cease if the supply of one of the nutrients ceases, however in the cave environment we have almost unimaginable time scales at our disposal.

Another source of nutrients for the sulphur active microbes may be derived from the light hydrocarbons (C_1-C_5) which are present universally in limestones as a result of the diagenetic reactions during rock formation (Goodman, 1987). Enhanced values may be present in the neighbourhood of mineral deposits (e.g. Ferguson, 1984). It is considered that these are generated at the site of ore deposition as a result of bacterial action on organic matter within the host carbonate and introduced into the mineralising brines (Goodman, 1987). Microbial oxidation of these light hydrocarbons according to reactions involving complementary sulphate reduction can provide more nutrients either directly or secondarily via the CO_2 that may be produced (Figure 5). The microbes *Desulphovibrio sp.* and *Desulphotomaculum* are probably involved (Ehrlich, 1982).

Temperature Control

Bacteria are most active at temperatures of about 35°C. This is high for this particular area. At lower temperatures activity diminishes, but active bacteria of the *Thiobacillus* type have been found at temperatures as low as 3°C, which is probably below the range that would be of concern here.

Particles Size Effects

The finer the grain size of the particles the faster the rate of decomposition since there is a greater surface area for the bacteria to act upon. We have no systematic information on the grain size of the pyrite particles but preliminary data suggests that there are few pyrite grain cubes with edges greater than 200 μm.

The Geomicrobial Cycle

Since it can be envisaged that there would be the necessary bacteria and nutrients present in the phreatic zone then the reaction products can be considered as part of a cycle of reactions. Once the microbes are emplaced within the phreatic zone all the necessary nutrients are present.

If the microbial oxidation of pyrite is considered then the following reactions have been identified or postulated (Figure 5):
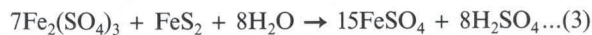
Pyrite in the presence of oxygen and water is slowly oxidised microbially to ferrous sulphate and sulphuric acid:

$$2FeS_2 + 7O_2 + 2H_2O \rightarrow 2FeSO_4 + 2H_2SO_2 \dots\dots\dots(1)$$

The iron oxidising bacteria then oxidise the ferrous sulphate in the presence of oxygen and sulphuric acid.

$$4FeSO_4 + 2H_2SO_4 + O_2 \rightarrow 2Fe_2(SO_4)_3 + 2H_2O \dots(2)$$
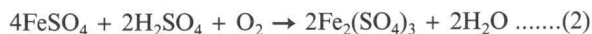
The ferric sulphate formed can react with more pyrite:

$$7Fe_2(SO_4)_3 + FeS_2 + 8H_2O \rightarrow 15FeSO_4 + 8H_2SO_4 \dots(3)$$

The ferrous sulphate produced in these reactions is oxidised by the iron oxidising bacteria to form more ferric sulphate and the entire cycle is repeated.

One of the products of the reaction of sulphuric acid with the ever present calcite is CO_2 (4), which is an essential nutrient for the bacterium. Therefore there appears to be a runaway chain reaction which will end only when one of the components in the system is consumed. The CO_2 will either form more feed stock for the bacteria or will be available for enhanced dissolution of the calcite (5). The least abundant component in the cycle is likely to be the pyrite. When this is completely consumed locally the reaction zone will move elsewhere. In the meantime channelways will have been opened up for a greater degree of water movement and to expose more pyrite to oxidation.

Microbial oxidation of light hydrocarbons according to the reactions e.g.:

$$CaSO_4 + 2(C_nH_{2n+2}) \rightarrow CaS + 2CO_2 + 2H_2O \dots(6) \text{ and}$$

$$CaS + H_2O \rightarrow Ca(OH)_2 + H_2S \text{ or mercaptans} \dots\dots(7)$$

and carried out by sulphate reducing bacteria such as *Desulphovibrio sp.* and *Desulphotomaculum sp.* is another series of mechanisms which can lead to enhanced solubilisation of
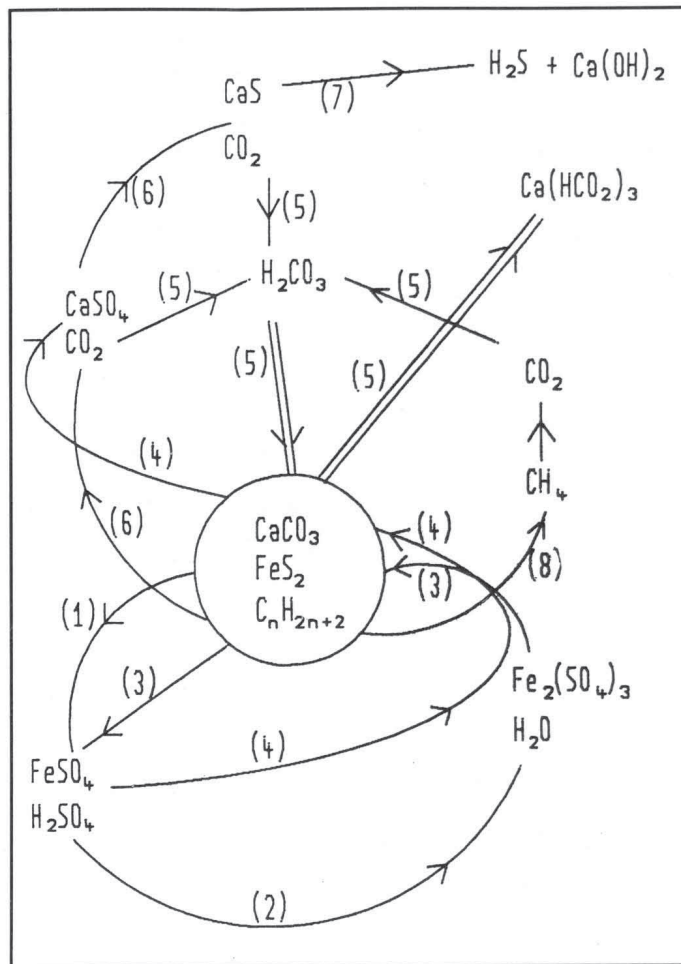

Figure 5. Generalised flow diagram illustrating the microbially aided oxidation of pyrite and hydrocarbons in a calcareous environment. For further explanation see text.

limestone (Figure 5). Some of the CO_2 produced will of course react with the $Ca(OH)_2$ but the remainder is available for the production of bicarbonate by reaction with calcite (5). The H_2S is also then available for oxidation to H_2SO_4.

Breakdown of the intrinsic higher hydrocarbons to methane is also possible (8), which, in the presence of sulphate, may be oxidised microbially to give more CO_2. Another source of CO_2 ultimately may be from the poly-saccharides produced as metabolic products of other bacteria, especially *Thiobacilli*.

Whilst it is suggested that these related mechanisms are operative at opening up the phreatic zone to circulating groundwaters they may well be far more effective in the shallow phreatic and vadose zones where the supply of nutrients is much more abundant. Is it a coincidence that the largest cave passages in South Wales are found in rocks where the content of sulphide is apparently the highest? In the Derbyshire dome the largest caves are found in the north where there is extensive metalliferous sulphide mineralisation.

The common presence of sulphate in limestone spring waters from South Wales (personal communication W. Gascoine) also supports the suggested mechanisms.

CONCLUSIONS

Several mechanisms have been proposed for cave development in the deep (unoxygenated) phreas but none of these are hydrochemically convincing unless there is input of aggressive chemicals from an external source (e.g. CO_2 from volcanic activity). There must therefore be some doubt as to whether there can be significant cave development in this zone of the phreas although very slowly moving groundwater over an extremely long period of time could be effective.

The accepted wisdom is that caves are formed by the action of aggressive, carbon dioxide charged, water upon calcite limestones. We propose an allied series of mechanisms involving a range of bacteria deriving their primary energy from sulphur. This explains many puzzling aspects of cave development, for example; why do cave passages tend to favour the relatively impure sulphide-rich limestones or limestone sequences, often to the exclusion of the purer members of the succession? These microbially triggered

reactions enhance the solubility of limestone by producing soluble $CaSO_4$, increase the concentration of CO_2 and bicarbonate ions, promote dissolution of calcite and thereby increase the permeability of the limestone.

A sequence of hydrochemical events can therefore be envisaged either in response to a drop in the base level of erosion or even as a zonation in a more static groundwater regime:

1) In the very deep phreas and in the presence of saline connate or formation waters, there is likely to be a uniform but very small degree of dissolution of limestone controlled by exchange reactions between $NaCl$ and $CaCO_3$. Flow may be channelled by fissures.

2) With the decline in base levels the formation waters are exchanged for fresh rainfall-derived (meteoric) waters. The waters are still likely to have no free oxygen with little impetus for oxidative biological activity. The waters are likely to be saturated with $CaHCO_3$.

3) Further decline in base levels introduces oxygen into the phreatic waters. Microbially induced oxidation of sulphides and hydrocarbons proceeds with the production of soluble sulphates and increased levels of carbon dioxide, which produce more bicarbonate ions, and result in increased dissolution of the limestone. These reactions proceed fastest in zones with the highest concentrations of sulphide minerals. The concentrations of cave passages in particular stratigraphical horizons or zones and along mineralised faults and joints originates at this stage.

4) Near the vadose/phreatic interface maximum erosion occurs. This is a function of several factors. Oxygenated water is at its most abundant and consequently organic and inorganic oxidation at its maximum. Because of the physical nature of the interface zone "Mixing Corrosion" is also at its maximum. However conduits resulting from phase 3 exert an important control over the initial position of solution tubes and normal features such as passage piracy and migration of cave passages result from the controlling effects of the original solution tubes.

5) In the vadose zone the corrosion is largely confined to water channels but oxidative reactions are at their maximum and as a result, where water is present the degree of dissolution is very high. Confined groundwater flowing along joints can carry sulphate and precipitate gypsum and selenite, especially in drier parts of cave systems.

It is not suggested that the reactions noted above have much quantitative significance in the development, as distinct from the initiation of cave passages. The concentration of sulphide minerals is usually far too small to result in more than limited dissolution of limestone. However the sulphide minerals are not homogeneously distributed and concentrations in certain beds or near bedding planes and faults, can open up the limestone to water flow and accelerated corrosion. Neither should it be thought that this simplified outline of potential reactions is exclusive. It is necessary to seek further understanding of the nature of the various organisms, of the detailed petrography and geochemistry of the limestone, and the chemistry and biochemistry of the groundwaters.

ACKNOWLEDGEMENTS

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Geomorphic Evolution of the Ingleborough Karst

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Abstract: The karst landscapes of Ingleborough have evolved under the influence of the Pleistocene glaciations. A paleogeographic reconstruction of the landscape about 500,000 years ago reveals the youthfulness of the karst around Ribblehead, and explains the contrasts between it and the older surface and underground landforms along the southern and southwestern flanks of Ingleborough. Most details of Ingleborough's modern surface landscape date from the Devensian glaciation and subsequent solutional processes, while the cave systems evolved through a longer sequence of phases of interglacial underground drainage.

Lying in the heart of the Yorkshire Dales glaciokarst, the limestone plateau which surrounds the summit mass of Ingleborough (figures 1 and 2) is well known for its splendid interrelated karst landforms. Most of the karst features lie in or on the strong and relatively pure Dinantian limestone, locally known by its facies designation, Great Scar Limestone (Arthurton et al, 1988). The statistics of the Ingleborough karst tell their own story: there are over 3000 sinkholes and dolines of various types and more than 300 ha of mature limestone pavement, more than 250 cave entrances lead to over 54 km of mapped cave passages, and there is also a range of fluvioglacial features including dry valleys and gorges. Karst on the thin limestones within the Yoredale rocks of the summit mass is limited to lines of shakeholes, some limited areas of pavement and just a few small caves.

Ingleborough was overrun by ice in a number of Pleistocene stages, with major flows on each occasion down Ribblesdale and Chapel-le-Dale; much of its landscape has therefore developed into spectacular glaciokarst. However, this part of the Pennine uplands was a dominantly erosional environment, with alternating cycles of both glacial and fluvial erosion. Devensian landforms therefore dominate the modern surface topography, but the cave systems contain a higher proportion of features remnant from the earlier stages of the geomorphic evolution. The pre-glacial ancestry of the Ingleborough surface survives only in the pattern of the largest features, notably the two flanking dales and the southern fault line scarp.

This review draws together the evidence from the karst and caves pertinent to the overall geomorphic evolution of Ingleborough. It starts with a landscape reconstruction of 500,000 years ago, reflects briefly on earlier events, and then follows the changing processes through to man's influence. The total picture so obtained explains some of the contrasts in the details of the karst around the various sectors of the Ingleborough limestone plateau.

THE PRE-ANGLIAN LANDSCAPE AND GEOLOGY

The earliest date for which it is practical to reconstruct the Ingleborough landscape is around 500,000 years B.P. This predates the known Anglian glaciation, currently estimated to have started around 480,000 years ago (Imbrie et al, 1984); it is also before the 350 ka limit of stalagmite dating using the uranium series technique (Gascoyne et al, 1978), which is the only method to have yielded, so far, absolute dates of any Ingleborough landforms older than Devensian.

A tentative reconstruction of the Ingleborough paleogeography is shown in figure 3, which covers the same area as figure 1. The shale boundary below the summit mass has been drawn through the ancient sinkholes of Alum Pot, Gaping Gill, Newby Moss Cave, Braithwaite Wife Hole and Great Douk Cave. Newby Moss Cave has a stalagmite dated to >350 ka (Gascoyne & Ford, 1984) just inside its entrance, thereby establishing its antiquity. The other four sinkholes are interpreted as being very old on the
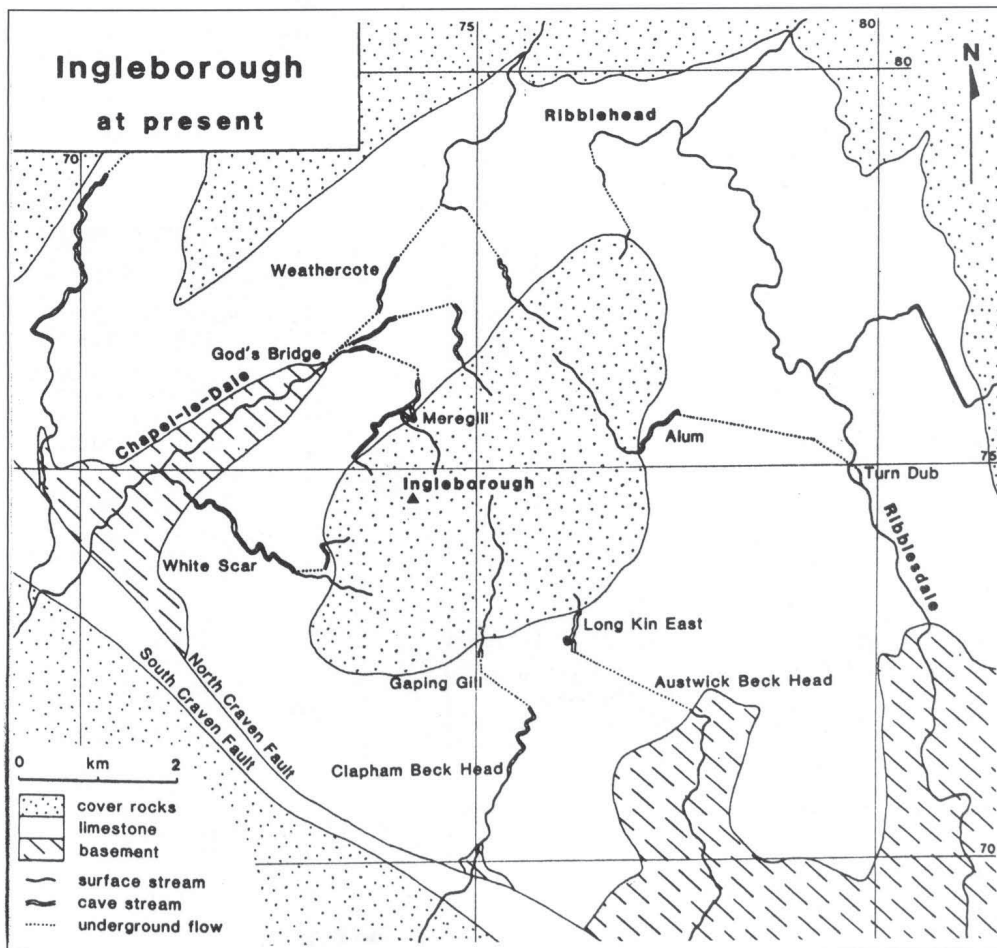

Figure 1. Geological map of Ingleborough showing the extent of the limestone and just some of the drainage routes which cross it above and below ground. The underground flows have been proved by dye-testing where the cave streams have not yet been completely followed and mapped.
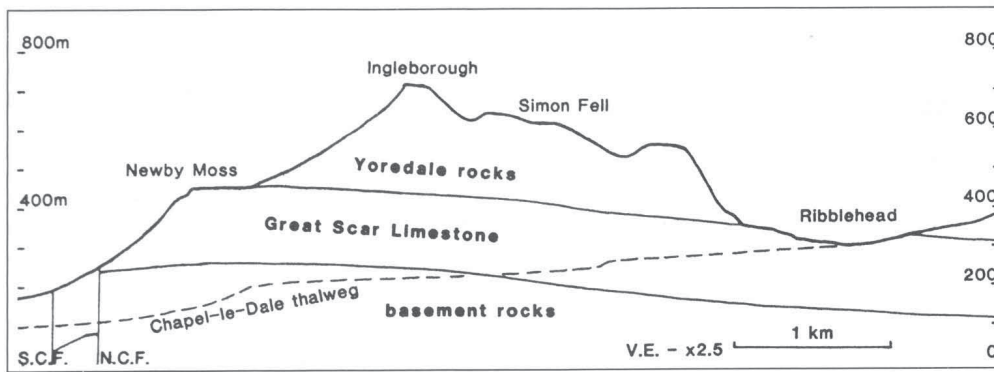
Figure 2. Profile through Ingleborough showing the main features of the geology, and also the Chapel-le-Dale thalweg which indicates the scale of local relief.

evidence of their relationship to ancient cave passages, their clay fills of glacial sediment and/or their large size. However, only Gaping Gill feeds to a cave passage dated to >350 ka (Gascoyne & Ford, 1984); the others could conceivably date from only an Ipswichian or Hoxnian shale margin. Questions over the precise contemporaneity of all the features on figure 3 must remain unanswered until further absolute dates are available.

The morphology of the valley floors in pre-Anglian times has been deduced from reconstructed profiles based on successive dates of cave drainage (Waltham, 1986). The dates of stalagmite deposition are indicated by the proportions of uranium series isotopes which now exist in the calcite; isotope ratios have changed due to radioactive decay within the closed system of the stalagmite since its deposition (Harmon et al, 1975; Gascoyne et al, 1978). As stalagmites can form only in air-filled caves, their ages provide latest possible dates for the drainage of the cave passages which contain them. The cave passages are then correlated with the paleo-valley profiles by geomorphological interpretation of the cave patterns, and by using the important assumption that the main resurgences of the caves are at the contemporary valley floor levels which represent the lowest drainage outlets from the karst aquifers. At present, very few significantly old stalagmite dates have been obtained in the Ingleborough caves, but these already indicate some latest possible dates for certain levels of valley floor erosion. When more dates are obtained, a more precise estimation of the progress of surface lowering will be possible.

Some paleo-profiles of Chapel-le-Dale have been reconstructed by the above method (Waltham, 1986), and suggest 80-100 m of surface lowering of the valley floor over the last 400-500 ka (figure 4). From this figure the pre-Anglian shale outcrop across the Chapel-le-Dale thalweg has been identified and added to figure 3. The same figure has then been applied in the same way to Ribblesdale.

Clearly, the rates of surface lowering cannot have been the same over the whole area, and maxima are likely to have been achieved in the glaciated troughs of the Dales. Valley floor lowering in the Ease Gill valley, west of Gragareth and protected by Great Coum and Crag Hill from the main ice flow, has only been at about half the rate that applied in Chapel-le-Dale (Waltham, 1986). Even more conspicuously, the shale boundary at Newby Moss Cave on the protected lee of Ingleborough (figure 5), has remained almost unchanged since Anglian times, as indicated by its ancient stalagmite (Gascoyne & Ford, 1984). Until more precise local erosion rates are deduced, or until additional pre-Anglian sinkholes are recognised, the reconstructed shale boundary on figure 3 can only be considered as an estimate between identifiable extremes.

The basement inlier on figure 3 has been reconstructed on the same principles as applied to the shale boundary, but more reliably so as it only occurs in the valley floor where the erosion rate is better assessed The Chapel-le-Dale inlier did not exist in pre-Anglian times (figure 4) Conversely the Ribblesdale inlier is little different from its present form due to the steep north-facing slope of the unconformity off the basement ridge, now so well exposed in both flanks of Crummack Dale. Since the Craven Faults are steeply dipping, they are hardly displaced in their paleo-outcrops.
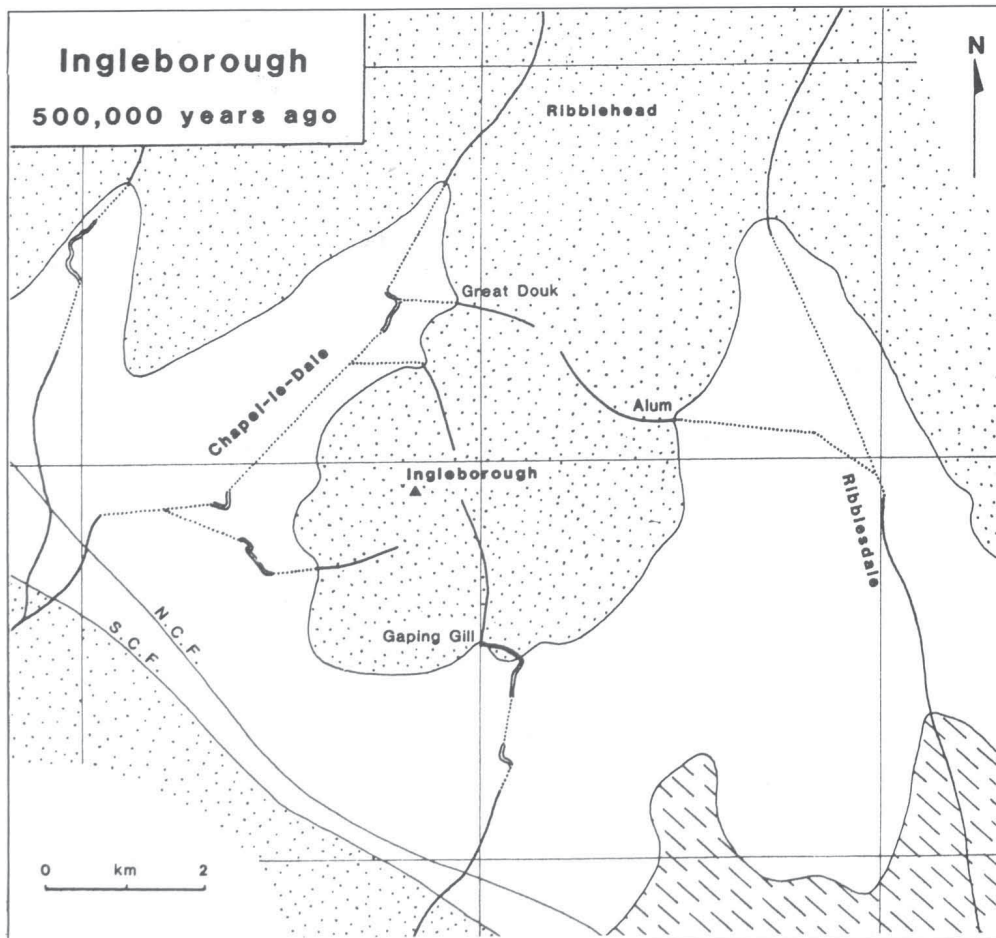
Figure 3. Geology and major drainage features of Ingleborough about 500,000 years ago. The frame, grid lines and key are the same as in figure 1. This map can only be regarded as an approximate interpretation; the evidence for its compilation is described in the text.

Figure 4. Comparison of profiles down Chapel-le-Dale today and about 500,000 years ago, showing the upstream shift of the karst in the valley floor and the exposure of the basement inlier.
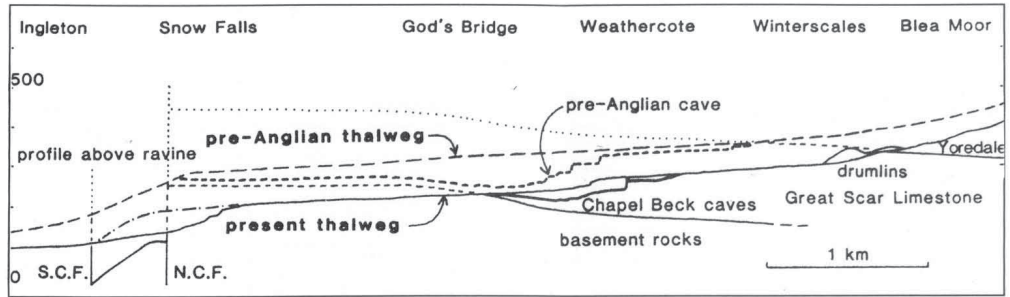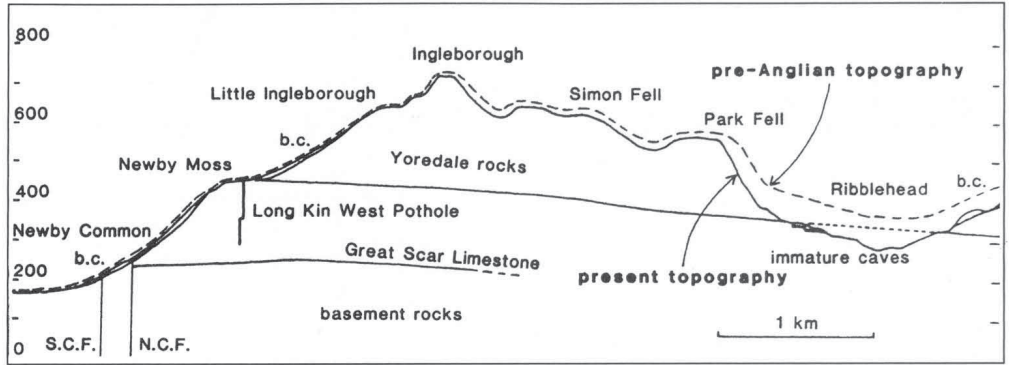

Figure 5. Comparison of profiles over the Ingleborough summit today and about 500,000 years ago, showing the minimal erosion in the lee over Newby Moss, and the recent exposure of the limestone at Ribblehead.

Overall relief of the Ingleborough area was less in pre-Anglian times than it is now. With the valley thalwegs still not approaching base level, the subsequent half million years would have seen more valley excavation than summit lowering, especially with the added impact of multiple erosional glacial episodes. The limestone plateau has been a conspicuous topographic feature throughout this time due to differential erosion of the shales and limestones, and the plateau edges would have always given a U-shape profile to the dales, even before the glacial trimming which created the splendid troughs of to-day.

The pre-Anglian karst and caves

Figure 6 compares the present and pre-Anglian limestone outcrops, and hence the lateral extent of the karst development, as they changed over time. It is significant that there was no pre-Anglian limestone outcrop in the Ribblehead area; shale covered the limestone and consequently there are no fossil caves in this area.

Conversely the major fossil cave systems are at the south end of Ingleborough where the limestone has been long exposed. White Scar Cave contains two segments of ancient trunk passage (figure 3), both of which have yielded stalagmites older than 350 ka (Gascoyne & Ford, 1984). Gaping Gill contains a complex of very old passages, not yet completely interpreted, but with some of similarly proven age. In both these caves, the ancient passages are just segments of trunk routes whose continuations are now choked by sediment and breakdown or have been eroded away. By analogy there should be more fossil cave awaiting discovery beneath Newby Moss, where there has been minimal glacial erosion, and almost no shale retreat, in the lee of Ingleborough summit, and where the active and fossil caves should be superimposed.

Mature limestone pavement at Southerscales Scars on the glacially scoured rock terraces fringing Chapel-le-Dale. *(Photo: A. Tillotson)*

The Hurnel Moss sink at the end of a short blind valley cut in the boulder clay on the main limestone plateau. Penyghent Hill is in the distance, on the far side of Ribblesdale.
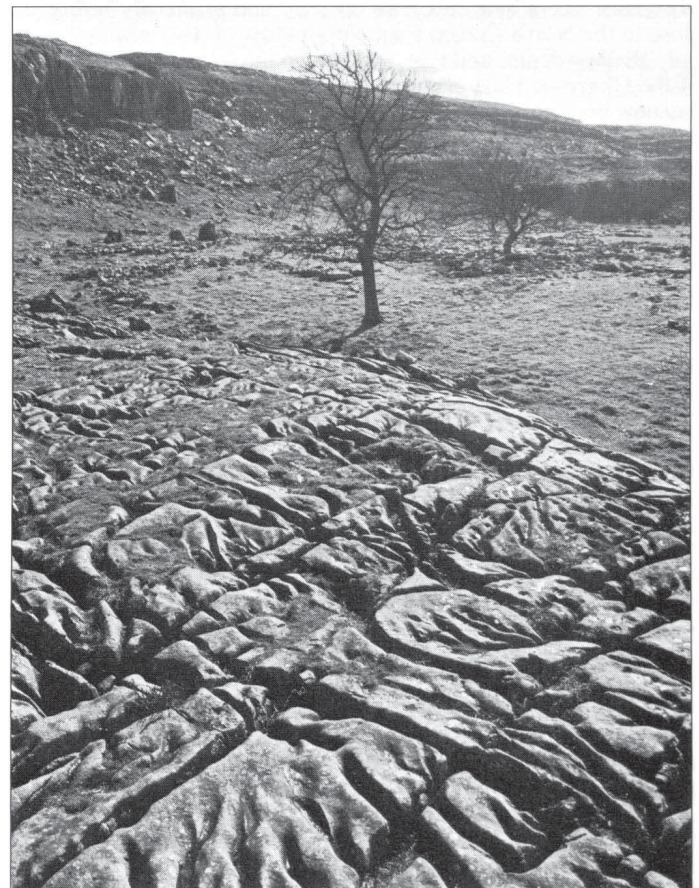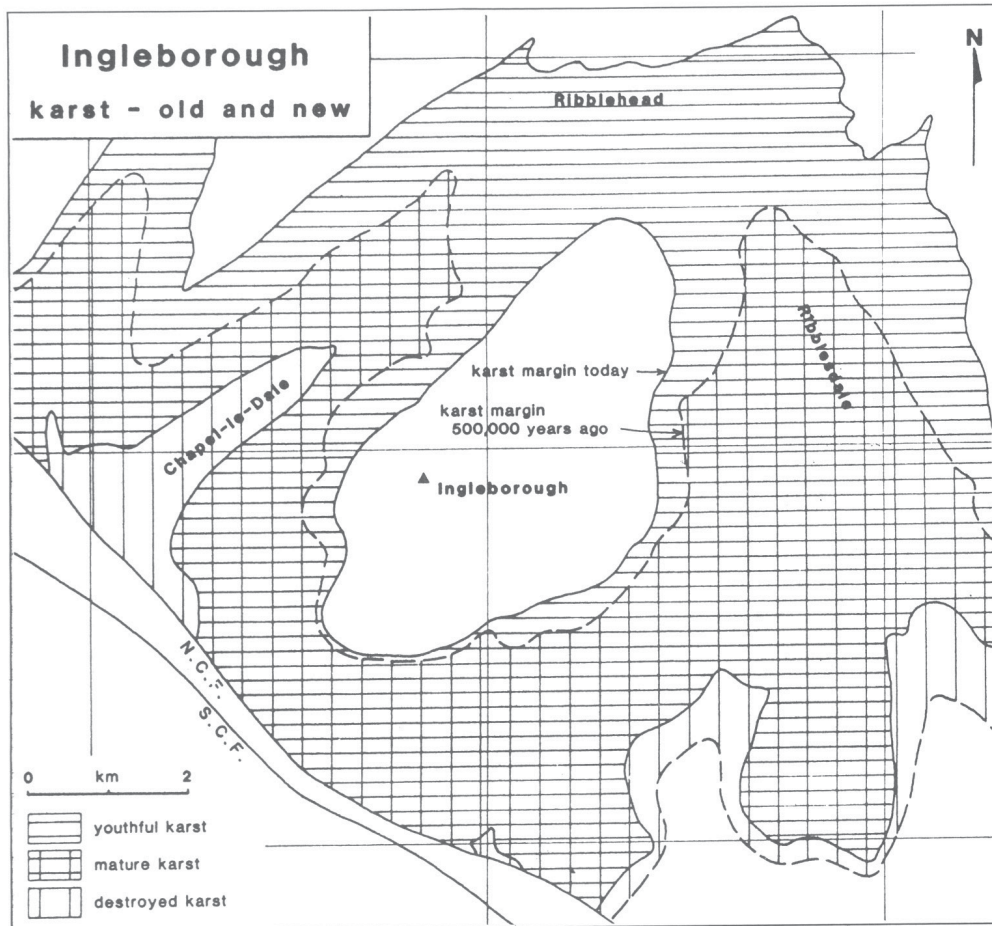
Figure 6. Subdivision of the Ingleborough karst on the basis of maturity, utilising the geological boundaries 500,000 years ago, taken from figure 3. The frame and grid lines are the same as in figures 1 and 3. Limestone outcrops less than 500,000 years old define the youthful karst, notably around Ribblehead. The mature karst is on limestone exposed for longer, though considerable surface lowering has occurred and cannot be shown on this map. The destroyed karst is on outcrops since removed by erosion, exposing the Chapel-le-Dale inlier but having less effect in Ribblesdale where the unconformity dips more steeply. The limestone between the Craven Faults is hydrologically isolated, and its karst is not subdivided on this map.

A pre-Anglian cave system probably extended just below or beside much of the length of the contemporary floor of Chapel-le-Dale. It could have compared with the modern caves mapped beneath Chapel Beck upstream of the God's Bridge resurgence and also with the caves beneath Kingsdale (Waltham & Brook, 1980). Remnants of this system may perhaps survive in the multi-level flooded passages of Joint Hole, and more probably in the Battlefield Series of White Scar Cave (figure 3). It was probably joined by tributary passages from Great Douk Cave, Braithwaite Wife Hole and a sink in Crina Bottom, and ultimately resurged close to the North Craven Fault. With most of its course lying at only shallow depth, and then exposed to the full erosional impact of the Chapel-le-Dale glaciers, it is likely that much of this cave has now been eroded away. Segments may survive beneath Raven Scars, comparable to that in White Scar Cave whose northern end is merely choked with debris. The precise age of the active phase of this cave is unknown. From the stalagmite dating, all that is known is that the cave was already drained, fossilised and being partially refilled with stalagmite more than 350,000 years ago. An age of 500 ka for the trunk drainage of Chapel-le-Dale to be utilising this route can therefore be only a crude approximation.

There are no recognisable remnants of any comparable trunk cave that may have existed beneath an ancient floor of Ribblesdale. However, the present route of the water from Alum Pot to Turn Dub, which is an anomaly in the modern drainage pattern, may follow an ancient passage. There is also probably more fossil cave awaiting discovery in the area between Alum Pot and Gaping Gill. The Moughton plateau was already remote from the shale boundary and allogenic drainage, so it is likely that only percolation-fed caves, and no major cave passages, underlie its expanse. At the earlier times, when the shale boundary did retreat across the Moughton plateau, the lack of significant gradient would have severely restricted the scale of cave development.

THE EARLIER HISTORY

Little is known of events in the Ingleborough area preceding these just previous to the Anglian glaciation. The deep-sea sediment record (Shackleton & Opdyke, 1973) reveals 10 cold stages in the Pleistocene, each cold enough to promote glacial conditions. Three glaciations in northern England are known. The extent of any earlier ice advances remains in the realm of speculation (Catt, 1981), and it is not known whether the Ingleborough area was subjected to ice cover or merely to

periglacial conditions. Similarly it is not known for what proportion of earlier Pleistocene time temperate conditions prevailed when karst processes could develop.

In these earlier times, caves may have formed and been removed by subsequent surface lowering; some appear to have survived and it would be remarkable if more had not existed. Development of the Gaping Gill complex of ancient phreatic passages could span a considerable length of time, and the East Passage, which appears to be the oldest, probably dates back substantially more than 500 ka. Proof of the age and understanding of these older caves awaits the application to the cave sediments of the dating methods (election spin resonance, paleomagnetism and thermoluminescence) which can reach back more than 350 ka.

The major valleys are the oldest surviving features of the landscape, and within their broadest confines it is locally possible to recognise significant past planation surfaces. A conspicuous surface close to the 400 m level has long been recognised in the Yorkshire Dales (Sweeting, 1950), though in many areas it is unduly emphasized because it coincides with the stratimorphic surface created on the top of the Great Scar Limestone (figure 2). There may also be a 600 m surface traceable in the Dales area (Clayton, 1981), but it is hardly recognisable on Ingleborough. If the 400 m surface is ascribed to the early Quaternary, and the 600 m surface to the late Tertiary (Clayton, 1981), the overall rate of surface lowering roughly matches that recorded through the last half million years.

This then means that the first exposure of the Ingleborough limestone would have been about two million years ago around the area now occupied by the lower part of Ribblesdale.

Attempts at reconstructing some of the earlier land surfaces, within the late Tertiary, (King, 1969; Clayton, 1981) have yielded little of significance with reference to the karst development. They have however emphasised the long-standing existence of an area of high ground north of Ingleborough which attracted maximum snow accumulation and became a centre for ice dispersal during the Pleistocene glaciations.

THROUGH THE ICE AGES

Within the last 500 ka, Ingleborough has been subjected to two or three major glaciations with two long temperate interglacial stages and intervening periods of periglacial conditions. The Anglian glaciation lasted roughly the time of Oxygen Isotope Stage 12, from 478-423 ka (Shackleton and Opdyke, 1973; Imbrie

The large depression north of Clapdale Scars partly filled with boulder clay.

et al, 1984) before the climate warmed into the Hoxnian interglacial.

A subsequent cold stage, in Oxygen Isotope Stage 6, from 186-128 ka (Martinson *et al,* 1987) has long been referred to as the Wolstonian glaciation. The sediment record in the English Midlands, on which this nomenclature was based, is now held in serious doubt, and the evidence for a third extensive glaciation is inconclusive (Bowen *et al,* 1986). However, sediments along the East coast, in and around Yorkshire, do provide evidence of a Wolstonian glaciation (Catt and Penny, 1966; Bowen *et al,* 1986). The question remains as to whether the Yorkshire Dales were also glaciated at this time. The absolute dating record from cave stalagmites does indicate three periods of reduced calcite deposition at around 20, 160 and 260 ka (Gascoyne *et al,* 1983b; Atkinson *et al,* 1978; Hennig *et al,* 1983), though the record is noticeably weaker through lack of data before 140 ka (Gordon *et al,* 1989). This confirms the existence of a "Wolstonian" cold stage, but until more cave dates are obtained and correlated with sediment and erosional features, there is no firm evidence of a Wolstonian glaciation in the Dales. Details of the geomorphic history of Ingleborough through this period cannot therefore be identified, and the karstic evolution can only be appraised in broad terms. But whether or not ice reoccupied the Ingleborough area during a Wolstonian advance, the stalagmite record does indicate a reduction of karst processes under very cold climates at that time.

The Devensian cold stage developed slowly around 100,000 years ago after the Ipswichian interglacial, and matured into a maximum ice extent little over 20,000 years ago; the ice finally retreated from the Dales between 13,000 and 12,000 years ago (Gascoyne *et al,* 1983b). No evidence has yet been found for any ice re-establishment contemporary with the Loch Lomond advance, when Ingleborough probably sustained 1000 years of periglacial conditions. The high-level bowls of Clapham Bents, Borrins Moor and Humphrey Bottom could have been well-placed to accumulate snow and firn. The last 10,000 years have been the post-glacial, or perhaps a third interglacial, stage with climates not unlike that of today.

In each glacial maximum, ice covered the entire area, reaching a thickness of perhaps 300 m over the summit of Ingleborough (Boulton *et al,* 1977). The ice moved outwards from Scottish and Lake District source areas, with an added component from local snow accumulation; the latter may have produced ice dispersal, perhaps at different times, from the area around the head of Wensleydale (King, 1976) and/or from the high ground around Dodd Fell and Widdale Fell (Clayton, 1981; Rose and Mitchell, 1989). Over and around Ingleborough, the overall ice movement was always to the south, with Chapel-le-Dale and Ribblesdale acting as iceways beneath ice sheets of very uneven flow which invaded landscapes of fluvial and glacially modified hills and valleys.

Little remains from the ice maxima. Any boulder clay once left on the summit ridges of Ingleborough has been removed by subsequent erosion. Most of the glacial landforms are from the Devensian retreat stages.

Each ice advance modified the Ingleborough landforms. It scoured and deepened the marginal dales, trimmed the flanking limestone scars, scoured and lowered the plateaus and caused retreat of the shale margins. The exception was the Newby Moss area which suffered minimal erosion in the protected lee of the summit mass (figure 5). Overall rates of surface lowering during the past 350,000 years have been estimated at around 0.12 m per 1000 years, and perhaps three times that rate in the Chapel-le-Dale glacial channel, on the evidence of cave drainage dating (Waltham, 1986). However lowering of the limestone plateau surfaces appears to be no more than about 0.04 m per 1000 years, on the evidence of both the height of the protected pedestals beneath the Norber erratics, and also direct measurements and measured solute removal from the area (Trudgill, 1986; Sweeting, 1966). These estimates of limestone erosion account for only solutional activity in non-glacial conditions, and the glacial erosion of the limestone can only be assessed from the retreat of the shale margin; it appears to be similarly low.

The caves and karst of Ingleborough developed largely during the warm interglacial stages of the Pleistocene. During the cold phases, both glacial and periglacial, solutional karst processes were greatly reduced. This postulate benefits from the analogy of the modern Castleguard karst in Canada, whose structural geology and morphology are remarkably similar to a Pleistocene glacial Ingleborough (Waltham, 1974c; Ford, 1983).

Solutional activity and major underground drainage restarted at the onset of each interglacial. New caves were formed in limestone newly exposed from beneath retreating shale margins. Some old caves were re-invaded, while others were left completely fossil, and yet others had been truncated or totally removed by glacial erosion of the limestone. The extensive shale loss at Ribblehead has allowed wholly youthful caves to dominate there today; most of these are shallow and sub-horizontal, restricted by the low gradients of the area and also perhaps by the geology, with many shale beds and few deeper fractures. Though caves can develop beneath a shale cover, they generally only do so where a favourable geological structure permits efficient drainage through the limestone; this was not the case in the buried limestone at Ribblehead.

With a comprehensive framework of absolute stalagmite dates not yet available for Ingleborough, little can be elucidated of the successive interglacial stages of the karst development. Undoubtedly some of the cave passages originated in these stages; these may include parts of Ingleborough Cave which appear to be later than the pre-Anglian passages of Gaping Gill yet are abandoned by the present drainage.

Each climatic cycle also produced variation in the cave deposits. Calcite deposition, as stalagmite and other forms, occurred mainly in the warm interglacial stages when a vegetation cover generated carbon dioxide to increase solutional activity. Conversely, the cold periglacial stages increased clastic deposition
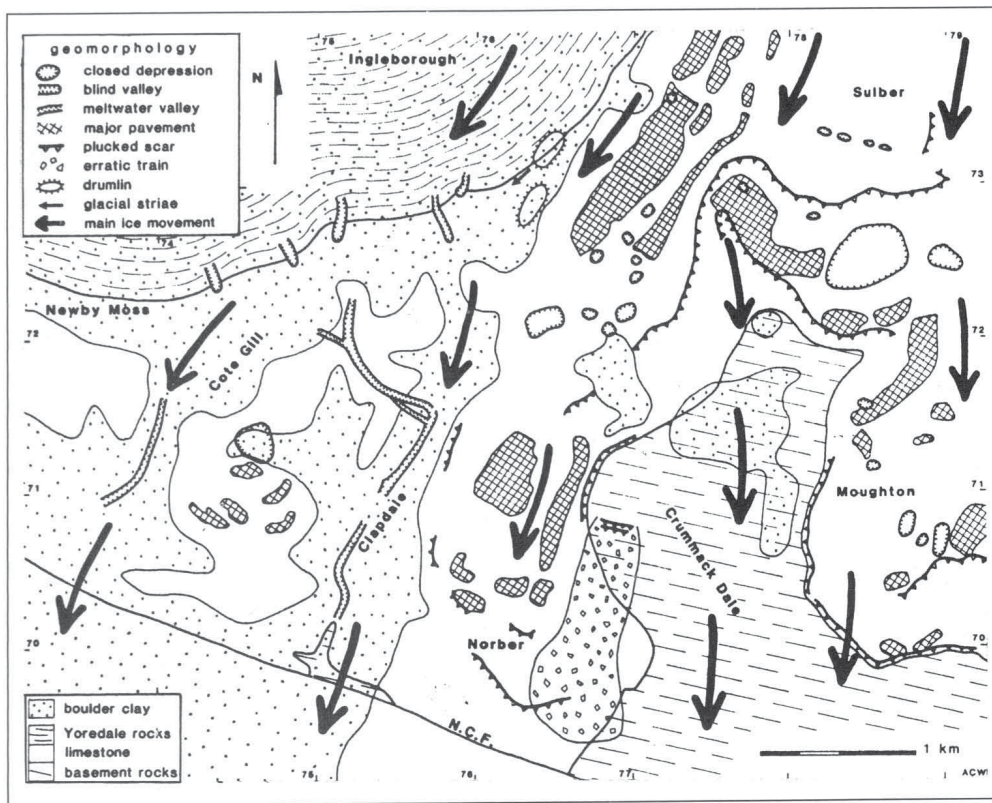
Figure 7. Glacial and karstic features of part of southeastern Ingleborough. The directions of ice movement only refer to the later part of the Devensian. Within the area of the map there is a marked contrast between the erosional environments, towards the east nearer to the Ribblesdale iceway, and the depositional environment, in the lee of Ingleborough towards the west.

within the caves. Glacial cover caused either a cessation of underground drainage, or the plugging of cave entrances with boulder clay, along with some reworking of boulder clay and meltwater deposition inside the caves. Intervening phases of cave stream flow eroded many of the earlier deposits, but enough material remains to provide a valuable data source for future research on Pleistocene chronology and environments. The Battlefield passages in White Scar Cave (Waltham, 1977b), and Mud Hall and other chambers in Gaping Gill (Glover, 1974), are notably important sites in this context.

Many of the modern surface features of Ingleborough date from the Devensian ice retreat. Then, over a period of a few thousand years, the ice sheet shrank into the valleys, and the summit mass was for a time a nunatak. Finally the ice formed just the two flanking valley glaciers, before melting away entirely.

The Chapel-le-Dale glacier was the smaller, but was powerfully erosive due to its steeper gradient down to the Craven Lowlands. As it declined from the level where it had over-ridden the summit ridge, it first scoured the stratimorphic benches on the top of the limestone, creating the clean rock surfaces which were the proto-pavements. Later, it shrank into the glaciated trough, trimming the sides, cleaning the excellent limestone scars, and enhancing the U-shape profile which is still so clear today.

The Ribblesdale glacier was larger. From around Selside the ice floor had to rise only slightly to spread all over the wide benches of southeastern Ingleborough, scouring the limestone surfaces which now include the extensive pavements of Sulber and Moughton. Flows of ice were directed into Crummack Dale, Clapdale and Cote Gill (figure 7). The first was powerful enough to scour the floor of its dale, and also to pluck the long line of limestone crags that form Moughton Scars. The Crummack ice also plucked greywacke blocks from the basement slope now just southwest of Crummack farm and spread them along the valley side to be left behind as the Norber erratics; the most famous of these now stand on the stratigraphically higher limestone where it has dipped down south of the basement ridge, and some of them have even been carried slightly uphill. The Clapdale ice was weaker, and left abundant boulder clay in Clapham Bottoms after a narrow tongue of ice had retreated from lower down the dale. Only decaying ice reached Cote Gill, which is now blanketed in boulder clay.

The Devensian ice retreat left the glacial troughs, and also the bare limestone plateaus subsequently to be fretted into pavements. It also left the splendid drumlin field at Ribblehead where ice over-rode and streamlined the boulder clay. More boulder clay appears to have filled a number of large dolines, of interglacial and pre-glacial age, which now only appear as grassy hollows in the bare limestone plateaus The finest of these lie due east of Gaping Gill, between the Allotment and Crummack Dale;

their sub-surface structure awaits investigation.

Glacial meltwater also contributed to the surface erosion of Ingleborough, carving a number of small valleys and ravines which are now mostly dry. The best known feature is the dry valley below Gaping Gill, which develops into the rocky gorge of Trow Gill (figure 8), and its older, parallel neighbour, where it cuts through a strong band of limestone in its steep descent into Clapdale. Whether the meltwater that carved these was sub-glacial or pro-glacial is open to debate (Pitty *et al*, 1986), but much of their excavation probably originates from during and/or just after the Devensian ice-melt phase. The periglacial conditions of that time were probably also largely responsible for various solifluction features still recognisable, such as the debris flow in Crina Bottom, and also the landslips on the summit slopes of Ingleborough.

CAVE DEVELOPMENT

When the limestone was first exposed at outcrop, the groundwater flow in its bedding planes and fractures was very restricted. Solutional enlargement of the fissures was therefore extremely slow in the initial stages; furthermore the fissure water was saturated with respect to calcite soon after entry. Sulphuric acid, perhaps derived from shale pyrite, may have played a significant role in these early stages. But as the fissures did inevitably slowly enlarge, the flow increased, and more biogenic carbon dioxide could enter the system, where it appears to have dominated the solution chemistry through all later stages. Fissures enlarged even more when turbulent flow was established in them (White, 1988).

By this stage, flow was concentrated into favourable routes linking the higher level input areas with resurgences at the lowest available outcrops. These routes developed into proto-caves, and further natural selection, based on hydrological efficiency, determined which should become large enough to warrant being called caves (nominally, those large enough to admit a man). Caves therefore inherited geological controls; they follow bedding planes, shale beds, joint/bedding intersections, and some horizons of lithological contrast within the limestone (Glover, 1973); they follow joints and faults to change stratigraphic level, downwards or upwards. The Ingleborough caves are dominated by staircase profiles of roughly level bedding passages and roughly vertical, fracture-guided shafts, though some caves do cut obliquely up or down fractures.

The initial caves were largely phreatic due to the restricted flow in the low permeability aquifer at that stage. As caves developed and improved in efficiency, the aquifer generated a water table close to the contemporary resurgence levels, which were in turn determined by the available surface outlets. The caves then
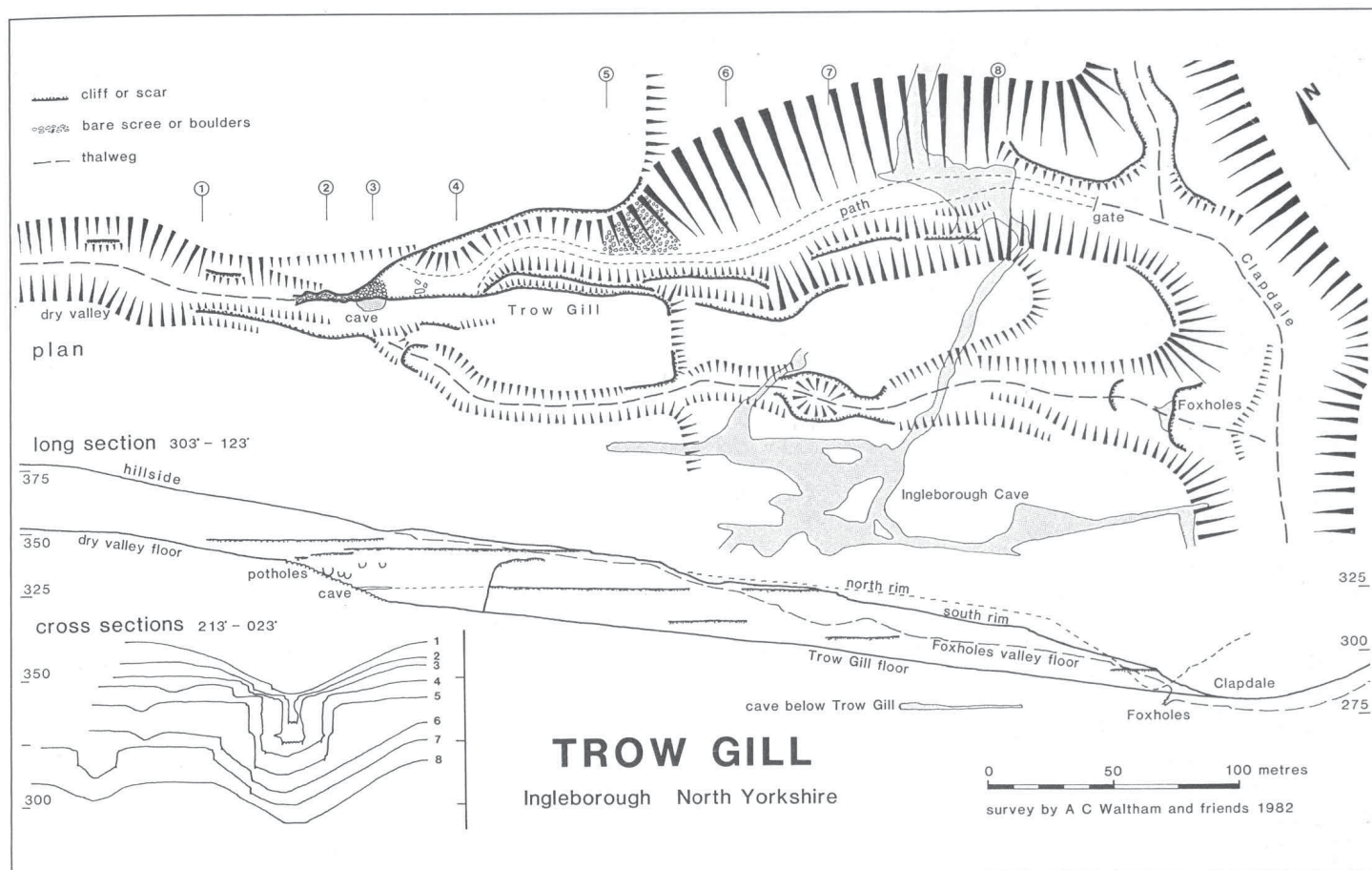
Figure 8. The karst gorge of Trow Gill, cut by periglacial meltwater entrenching through strong beds of limestone as it cascaded down into Clapdale. Some cave passages have been omitted to improve clarity.

matured in the vadose environment above the water table, or the phreatic environment below. But the great contrast between the overall aquifer permeability and the individual conduit permeability is such that a cave conduit creates its own local water table, or succession of water tables, to fit its geologically controlled profile. Perched sections of flooded cave are common where the initial phreatic route climbed up a fracture or was directed up-dip. As the caves evolve, these flooded sections may be drained by passage floor entrenchment.

Within the phreas, most of the Ingleborough caves developed, and still develop, into tubes of elliptical or circular cross-section, generally along bedding horizons, with tall rift chambers often developed on faults or major joints. In the vadose zone the majority of caves develop into deep meandering canyons, below bedding plane roofs, with their long profiles broken by fracture-guided vertical shafts; these are either rounded due to spray corrosion, or elongated along waterfall-retreat slots. These patterns are typical of Yorkshire Dales caves (Waltham, 1970), of which the type example may be regarded as Swinsto Hole in Kingsdale (Waltham *et al*, 1981); this has both vadose and phreatic components, together with segments of older invaded passage, reflecting the frequency of multiple-age development in the Ingleborough cave systems.

Rates of entrenchment of vadose cave canyons have been measured at only a few localities, where they are in the order of 0.02-0.05 m/1000 years (Gascoyne *et al*, 1983a). Higher overall rates are undoubtedly achieved in steeply graded stream caves where waterfall retreat makes a greater contribution to canyon incision. It appears that a timescale of 10 to 100 ka is appropriate for the formation of most Ingleborough cave passages, and this is compatible with the overall chronology of karst development.

The dominant pattern within the Ingleborough caves is for a vadose bedding-controlled passage to be oriented down-dip, until it finds joints or faults which permit it to descend (Waltham, 1974b). Eventually it finds the resurgence level where it enters the phreas, and, free of gravity control, it may then change direction to head for the resurgence still controlled by the geological structures but no longer constrained to follow the dip. The critical laws of control are that vadose flow takes the easiest immediate route, downhill, while phreatic flow takes the easiest overall route to the outlet (Brook, 1974), and the easiest, most efficient route is the one which develops from the initial network of fissures into a cave passage.

The pre-Anglian phreatic roof tunnel in White Scar Cave with the vadose Far Streamway deeply entrenched in its floor.

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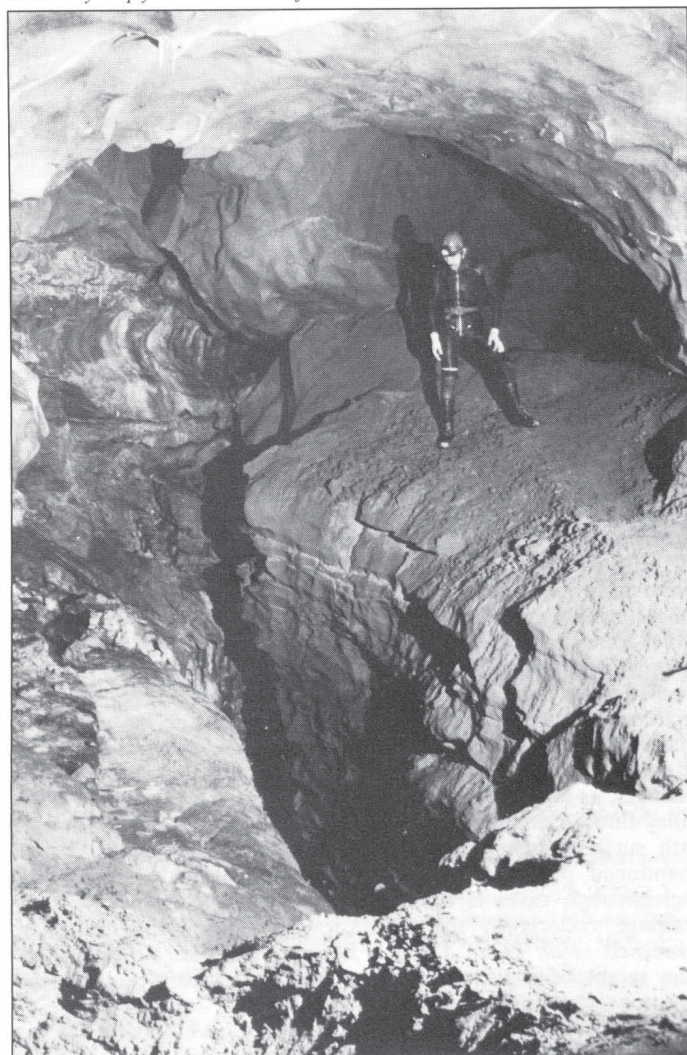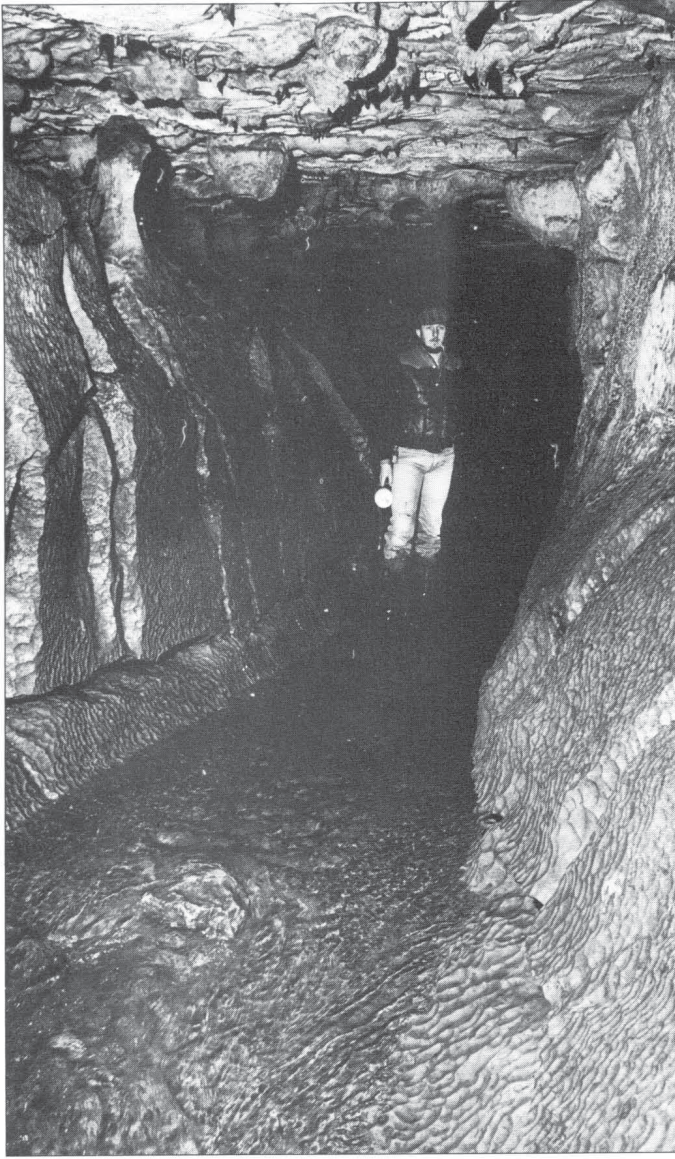
The clean scalloped vadose canyon in the Long Churn Cave appears to have originated before the last glaciation as it is too deep to have formed since then at measured rates of entrenchment.
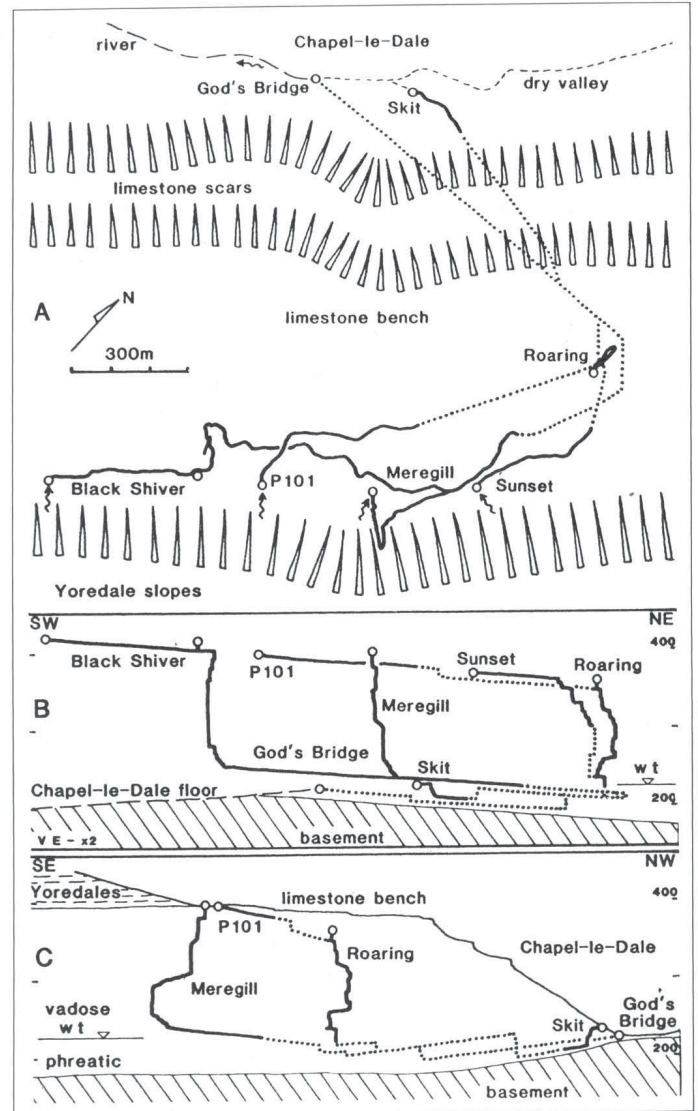

Figure 9. Geological control of the caves around Meregill. A = plan. B = dip section. C = strike section. The solid lines represent mapped caves passages; the dotted lines are only diagrammatic representations of caves known to exist by their dye-tested drainage connections. All the vadose caves follow bedding planes and shale beds downdip to the northeast, except where they drop down fault-guided shafts. The water table is determined by the position of the God's Bridge resurgence, and the phreatic caves loop through the bedding and joints downdip of the rising. Meregill Skit is a flood overflow resurgence. There is no syncline visible in section C; the phreatic passage slopes (and the parallel basal unconformity) are oriented obliquely up or down dip. Recent diving explorations suggest the cave downstream of the confluence below Roaring Hole, may take a longer, more northerly route to join a gently ascending phreatic tube in Joint Hole, much closer to the line of the dry valley. This route still feeds to God's Bridge, with overflow through Meregill Skit, but confirmation of its role awaits further explorations. Such a cave loop would hardly effect the profiles B and C, and the principles of the geological control remain unchanged.

In Ingleborough, the regional dip of the limestone is very gently to the north, while the overall surface drainage is to the south. This conflict provides some contrasts in cave patterns around the different sides of the mountain. White Scar Cave has a long vadose canyon, almost to its resurgence, as it can drain obliquely downdip into Chapel-le-Dale. In contrast, Gaping Gill drains south into Clapdale, against the dip, and is dominated by phreatic passage development. Around Meregill Hole, the vadose caves follow the bedding downdip to the north, at various levels within the limestone; but they then turn to the west within the phreas to drain to the God's Bridge resurgence via a series of shallow phreatic loops following the bedding and joints (figure 9).

Additions to these simple patterns are created where major faults guide the drainage more effectively. The deep shaft caves of the Allotment are all on faults, and the first section of Meregill Hole follows obliquely down a fault to the southeast, in completely the opposite direction to its ultimate destination at the God's Bridge rising (figure 9). Cave streams may also invade older sections of fossil cave, which are only extensive around the southern end of Ingleborough where the limestone was exposed in earlier times. These fossil caves are dominated by phreatic passages, as more of the limestone was below their contemporary valley-floor resurgences. The falling water table, commensurate with surface erosion, has permitted vadose invasion of many abandoned phreatic caves. An extra control on some of the Ingleborough caves is exercised by basement relief, as vadose drainage collects in the depressions in the limestone floor (Halliwell *et al*, 1975). There is however less cave development than might be expected right on the impermeable base of the limestone, as shale beds and lithological contrasts within the basal beds perch many passages some metres above the basement (Waltham, 1977a).

POST-GLACIAL DEVELOPMENT

Since the Devensian ice retreat, the Ingleborough region has become a fully karstic environment superimposed on inherited glacial, fluvial and karstic landforms. All drainage across the Ingleborough plateau is now underground for at least part of its course. There are numerous stream caves, but measured floor erosion rates (Gascoyne *et al*, 1983) suggest that all but the smallest have origins which predate the 13,000 years of post-glacial time. The present karst hydrological regime includes both conduit flow and percolation fissure flow (Halliwell, 1980). The former washes clean the eroding rock walls of the stream caves such as those in the splendid Long Churn system feeding Alum Pot. The percolation flow is responsible for the deposition of a new generation of stalactites and stalagmites, as yet small but clean and active, before it joins the conduit flow. Some resurging percolation water has deposited small banks of tufa at a few sites on Ingleborough (Pentecost & Lord, 1988).

On the surface, dolines of solutional and/or rock collapse origins are rare; the active sinkholes, fed by allogenic water around the shale margin, have mostly not yet flared out to achieve the conical profile of the conventional doline. In contrast, Ingleborough is pocked by many hundreds of subsidence sinkholes, locally known as shakeholes. These are all formed by

Re-establishment of vegetation on the sheep-free limestone pavements in the Scar Close nature reserve.

slumping, collapse and downwashing of the boulder clay into fissures in the underlying limestone, in many cases choking the entrances of cave passages. The rate of development of the shakeholes is low, and only a few have been recorded in historical times; a major slump occurred at Marble Pot in 1980, and new shakeholes have been observed on Newby Moss since then.

Outside the areas of boulder clay, the limestone plateau surfaces of Ingleborough are dominated by limestone pavements, including many of outstanding morphological quality. These have all formed by solutional weathering of the bare rock surfaces left by the Devensian ice retreat. Where the glaciers did not strip the rock bare, the modern outcrops are generally covered by more broken rock and soil. In some places there is so much frost-shattered debris, known as shillow, that it may be described as a felsenmeer (literally a stone sea), and it also forms screes below the higher scars. On the pavements, solutional widening of intially narrow fractures has created the deep linear fissures known locally as grykes, and internationally as kluftkarren. The remnant blocks of bedrock are locally known as clints. These may be 10 m or more across, separated by grykes of some metres depth, and may also be incised by solutional runnels which represent another form of karren.

Within the Ingleborough pavements, the dominant karren are the rounded rundkarren, which form under a soil or vegetation cover (Jones, 1965; Sweeting, 1966; Goldie, 1973; Trudgill, 1986). The sharper-edged runnels known as rillenkarren, which form under either direct rainfall or beneath a snow cover (Bogli, 1960) are restricted to minor forms within some of the larger rundkarren. It therefore appears that much of the Ingleborough limestone benches have had a more extensive cover of soil and vegetation than now exists.

Some of the pavements matured under a soil or peat cover which is now being eroded away, and the excellent karren forms occurring close to actively retreating soil margins are evidence of this process. Others may have developed under permeable boulder clay, but this is likely to have been limited, as much of the boulder clay is impermeable enough to prevent any solution of the limestone rockhead; glacial striae survive beneath the boulder clay near the Long Kin East Cave. Since the ice retreat, and previous to the last few thousand years, the natural vegetation on the Ingleborough limestone plateau has been a mixture of moss banks, grass hummocks, bushes and 'scrub vegetation, all contributing to an organic soil. This now only survives in the nature reserves of Scar Close and Colt Park. Under these conditions, rundkarren mature well. On soil-free rock surfaces the role of the ubiquitous lichen cover may be significant in the rounding of the karren forms.

A process important to the Ingleborough karst has been the removal of the woodland by man in a long series of clearances spread over the last 5000 years. The loss of trees promoted soil loss, and vegetation was reduced even more with the introduction of sheep. The pavements were laid bare, and now have no chance of regaining any plant cover except in sheep-free nature reserves. Footpath erosion, drainage gripping with its consequent downstream flood scouring, highly destructive limestone pavement removal for walling (Goldie, 1986) and cave entrance excavation or infilling by cavers or farmers complete the geomorphic evolution of the Ingleborough karst.

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Limestone Petrology and Cave Morphology on San Salvador Island, Bahamas

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Abstract: The horizontal dissolution caves of San Salvador Island are characterized by a unique phreatic morphology of large globular chambers and dead-end dissolution tubes. The walls are smooth and curvilinear. Analyses of 36 wall rock samples reveal that the caves have developed in rocks that are primarily eolian calcarenites, but a few penetrate into marine facies. Allochems found in the wall rock samples include ooids, peloids, aggregate grains, lithoclasts, forams, algae, and other bioclasts. Grain size ranges from micritic mud to very coarse sand, but the majority of samples are well sorted and have an average size of 0.32 mm (medium sand). Based on amino acid racemization analyses of gastropods, and stratigraphic relationships with uranium/thorium dated fossil coral reefs, the rocks in which the caves are developed are 150,000 years old or less. The cements found in these rocks show the influence of diagenesis in the marine phreatic, halocline mixing zone, freshwater phreatic, and vadose environments. Petrographic analyses reveal the presence of aragonite, high and low-magnesium calcite, and minor amounts of gypsum and halite. X-ray diffraction analyses confirm the petrographic determinations, and indicate the presence of trace amounts of dolomite not confirmed in thin section. The gypsum, halite, and traces of dolomite in the San Salvador cave wall rocks appear to be restricted to secondary crusts precipitated by the evaporation of vadose groundwater contaminated by sea-spray. Porosity, ranging from 1% to 49%, is both fabric and non-fabric selective. San Salvador caves are developed primarily in geologically young eolian deposits that formed in a few thousand years, have undergone minimal overprinting by glacio-eustatic sea level fluctuations and associated freshwater lens migration, and are currently in the vadose zone.

San Salvador Island is part of the Bahamian Archipelago which is located along the isostatically subsiding eastern margin of North America (Figures 1 and 4). The island lies on a small isolated bank at the eastern edge of the Bahamas, and is composed of Pleistocene and Holocene limestones which include fossil coral reefs and other subtidal facies, beach facies, beachrock, eolianites and paleosols.

In the Bahamas, many abandoned horizontal dissolution caves are found at elevations between 0 and 7 m above present sea level. Located both along the perimeter and in the interior of the islands, these relict caves are related to past high sea level events associated with deep-sea oxygen isotope stage 5 that extended from approximately 125,000 to 85,000 years ago (Shackleton and Opdyke, 1973).

## Previous investigations of Carbonate Island Karst Hydrology

The volume and geometry of the fresh groundwater lens below carbonate oceanic islands is largely determined by climate and hydrologic characteristics of the aquifer. In homogeneous relatively unconsolidated porous limestones, such as those found in the Bahamas, the downward infiltrating freshwater will displace the denser saltwater, and a lens-shaped fresh to brackish groundwater body may develop in accordance with the idealized Ghyben-Herzberg relationship (Figure 2).

The shape of such a lens is determined by the difference in density between the fresh and saline water, the amount and rate of recharge, and the porosity and permeability of the aquifer. If the aquifer is extremely permeable, the freshwater will spread upon the surface of and mix with the salt water to form a thin brackish layer; however, if the aquifer lacks sufficient lateral permeability, the freshwater moves downward because of the hydraulic gradient more rapidly than it can disperse horizontally. This process creates a mound of freshwater which extends above sea level. The resulting interface, or halocline, between the freshwater and saltwater at the base of the lens will have a depth below sea level of approximately 40 times that of the height of the water table above sea level (see Figure 2). Under steady state conditions, the amount of water flowing out at the margin of the lens is equal to the amount of water infiltrating from the surface, resulting in a stable but dynamic lens (Tarbox, 1987).

Hydrologic studies of Pacific atolls (Ayers and Vacher, 1986; Wheatcraft and Buddemeier, 1981) suggest that Ghyben-Herzberg lenses are hypothetical and highly idealized, and do not pertain to real atoll-island systems. Those studies have found that the

Figure 1. Location map of San Salvador Island in the Bahamian Archipelago.
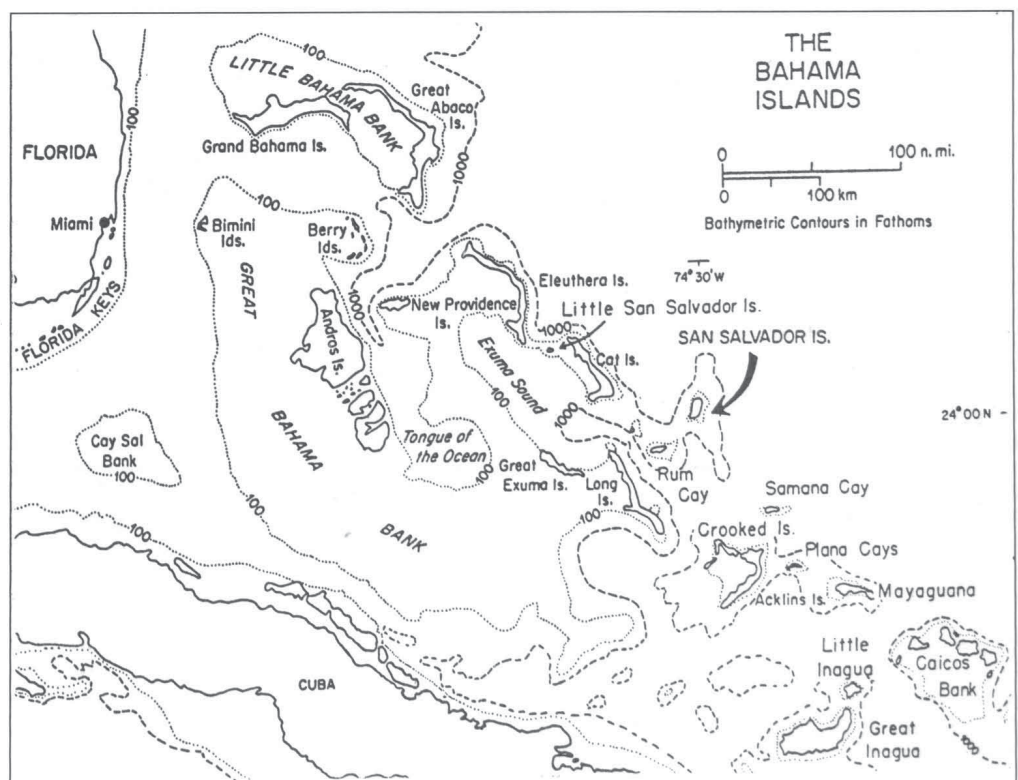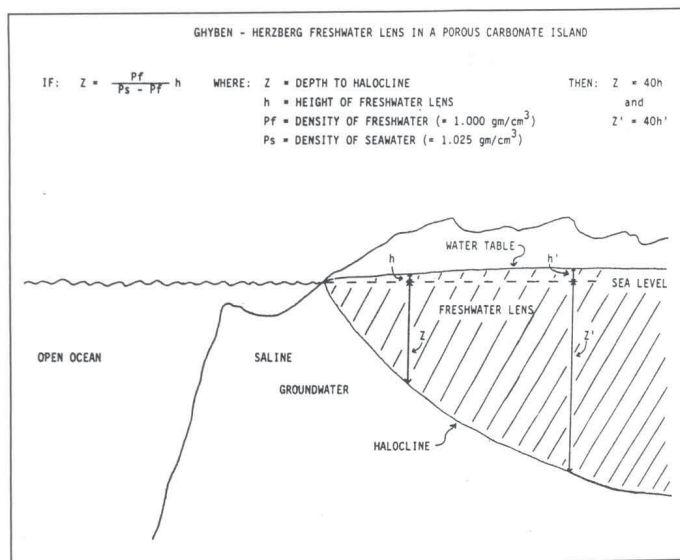
Figure 2. Idealized representation of a freshwater lens in a carbonate island, demonstrating the Ghyben-Herzberg principle.

freshwater lens, which may be partitioned into many smaller lenses, is partially confined and occurs within interfingering geologic units with differing pore-space characteristics (Vacher and Bengtsson, in press). Also, the lenses are subject to recharge events of variable magnitude and duration, and are greatly affected by sea level changes of tidal and shorter-frequency periodicities (Ayers and Vacher, 1986). Less pronounced fluctuations of freshwater lens configurations are associated with atmospheric pressure and temperature and salinity (density) of the water bodies (Vacher, 1978). Despite the departure of the idealized Ghyben-Herzberg model from what is actually found in the field, it remains a useful construct to describe the broad parameters of freshwater lens development in carbonate islands. The boundary between the marine and freshwater, the halocline, may be a sharp boundary or a broad zone of changing salinity resulting from mixing of the two waters.

CAVE DEVELOPMENT

Dissolution caves in an aquifer can change the shape of an ideal lens by changing the effective hydraulic conductivity of the aquifer in preferred directions. In the Bahamas, many of the dissolution features have pronounced horizontal orientations, and their origins are related to changes in sea level. The development of caves in the Bahamas, and their relationship to the position of sea level at the time of cave formation, has been discussed by Mylroie (1983b; 1988), R. Palmer and Williams (1984), and Mylroie and Carew (1988). Mylroie and Carew (1988) have used the elevation of horizontal dissolution caves on San Salvador Island, Bahamas to estimate Late Quaternary sea level positions.

Although most caves in calcium carbonate rocks are formed by the dissolution activity of carbonic acid, sulphuric acid mixing-dissolution and some poorly understood mechanisms of dissolution occurring at depths of thousands of metres below base level have also been recognized (White, 1988). Bögli (1980) demonstrated that a mixture of two saturated waters has the potential to produce undersaturated water, a process called mixing corrosion. In addition to mixing dissolution caused by the nonlinear relationship between calcite solubility and CO_2 concentration, changes in the state of saturation of mixed waters can result due to other nonlinear effects such as differing ion strengths, temperatures, and magnesium concentrations (White, 1988).

Recent work by cave divers on Andros Island (R. Palmer and Williams, 1984; R. Palmer, 1986) and Cat Island, Bahamas (R. Palmer, et al, 1986); Yucatan, Mexico (Back, et al., 1986); and Bermuda (A. Palmer, et al., 1977), has shown that the halocline mixing zone is a site of intense dissolution. The notion of enhanced dissolution in the halocline region is supported by direct observation, chemical measurements, and theoretical calculations. Direct observations of the walls of halocline-produced caves have revealed intense dissolution, exposure of delicate calcite fossils by dissolution of the aragonite matrix, and the development of megaporosity (Back, et al., 1986; Smart, et al., 1988). Cave passages have been observed to follow the halocline from near sea level on the platform margins to significant depths

beneath the island interior (R. Palmer and Williams, 1984).

While most Bahamian cave development is clearly related to dissolution along the halocline, other evidence demonstrates that caves also develop along the vadose/phreatic contact where corrosion is produced by descending vadose water mixing with the phreatic freshwater lens (Mylroie, 1983a; 1984). Throughout the Bahamas, large subaerial cave passages exist in eolianites from 0 to 7 m above current sea level (Mylroie and Carew, 1986; Mylroie, 1988). Amino acid racemization analyses of fossil land snails (*Cerion* sp) in the eolianites and stratigraphic relationships with uranium series dated fossil reefs indicate that the caves are developed in rocks younger than 150,000 years (Carew and Mylroie, 1987). These caves must have formed when a past higher sea level lifted the freshwater lens to about +7m elevation (Mylroie, 1983a; 1984; 1988; Mylroie and Carew, 1988). In general, Pleistocene sea level fluctuations may have repeatedly caused Bahamian cave passages to lie within the aggressive dissolutional zones associated with the halocline and the top of the freshwater lens. Such cave passages, especially those that are presently submerged, may be polygenetic.

The subaerial conduits examined in this study, however, could have developed only during the highest sea level event(s) of the Late Pleistocene, and the problems of interpretation associated with overprinting by many sea level events are minimized. Data from San Salvador Island (Carew and Mylroie, 1987; Mylroie and Carew, 1988) indicate that sea level was at present levels during the oxygen isotope substages 5e, 5c, and 5a of Shackleton and Opdyke (1973), centered around 125,000, 105,000 and 85,000 years ago, respectively. Only the 125,000 year highstand is generally accepted to have reached significantly (+6 m) above current sea level. The total available time for high sea level events in the Late Pleistocene is certainly less than 25,000 years out of the last 150,000 years, indicating that a rapid dissolutional process formed the caves.

Caves on the platform margin represent halocline controlled features coupled to sea level position. Because sea level has rarely been at +6 m elevation during the last 150,000 years, it is unlikely that halocline and freshwater effects have been superimposed more than once at elevations between +2 m and +6 m. Hunt's Cave on New Providence Island, Bahamas, may be an exception. The evidence there suggests a sea level of about +7 m inundated the cave after its initial formation and decoration by subaerial speleothems, as the speleothems show subsequent partial phreatic dissolution (Garrett and Gould, 1984). Neither the dunes nor the stalagmites of Hunt's Cave have been dated, and the cave could be older than those on San Salvador. Jennings (1968) made a strong argument for syngenetic karst development (karst development concurrent with lithification) in Australian eolian calcarenites.

FLANK MARGIN MODEL OF CAVE DEVELOPMENT

Recently the Flank Margin Model for cave development in carbonate islands has been proposed by Mylroie (1988). This model explains the morphology, position, and scale of the horizontal dissolution conduits found at 0 to +7 m elevation in the Bahamas. These caves have a consistent and recognizable morphology that usually consists of a large central room within a Pleistocene dune, and a series of smaller passages that radiate toward the interior of the dune (Figure 3). These divergent passages often end abruptly in bedrock walls, or they may feather out into tiny bedding plane voids. Secondary vadose modifications, such as intersecting pits and canyons found in some horizontal caves, provide evidence of the transition from phreatic to vadose conditions as these caves were drained by a drop in sea level (Mylroie, 1988). Study of the primary bedding structures of the eolian calcarenites (dunes) that enclose these caves suggests that the caves have been breached by modest lateral erosion of the original depositional margin of the dunes. Because the main cave chamber is near the dune margin, the entrance to these subaerial Bahamian caves usually opens into that chamber. The horizontal extent of these caves is always much greater than their vertical range. In addition to San Salvador, this pattern of cave development has been seen on Eleuthera, Great Inagua, New Providence, North Andros, and South Andros Islands, Bahamas (Mylroie, 1988); and on Cat Island (R. Palmer, et al., 1986).

The flank margin model suggests that these caves were formed rapidly by dissolution during past high stands of sea level, as a result of chemically undersaturated water produced by the interaction of a small fresh or brackish water lens and sea water along the margins of the lens. Dissolution within the dune was

Figure 3. Diagrammatic representation of the flank margin model of cave development. 3A shows initial conditions in plan and section. As discharge of the lens begins, and mixing with marine waters occurs under the flank of the dune, a series of small phreatic pockets develop, shown in 3B. If the dune discharge can be concentrated, as in 3C, larger phreatic pockets develop. Through time these pockets enlarge into the dune as mixing corrosion dissolves routes along each contributing flow path from the dune interior. This headward migration of the dissolution front produces caves with large chambers under the margin of the dune, and smaller passages leading into the dune which end abruptly, as shown in 3D. Compare with Figures 6-10.
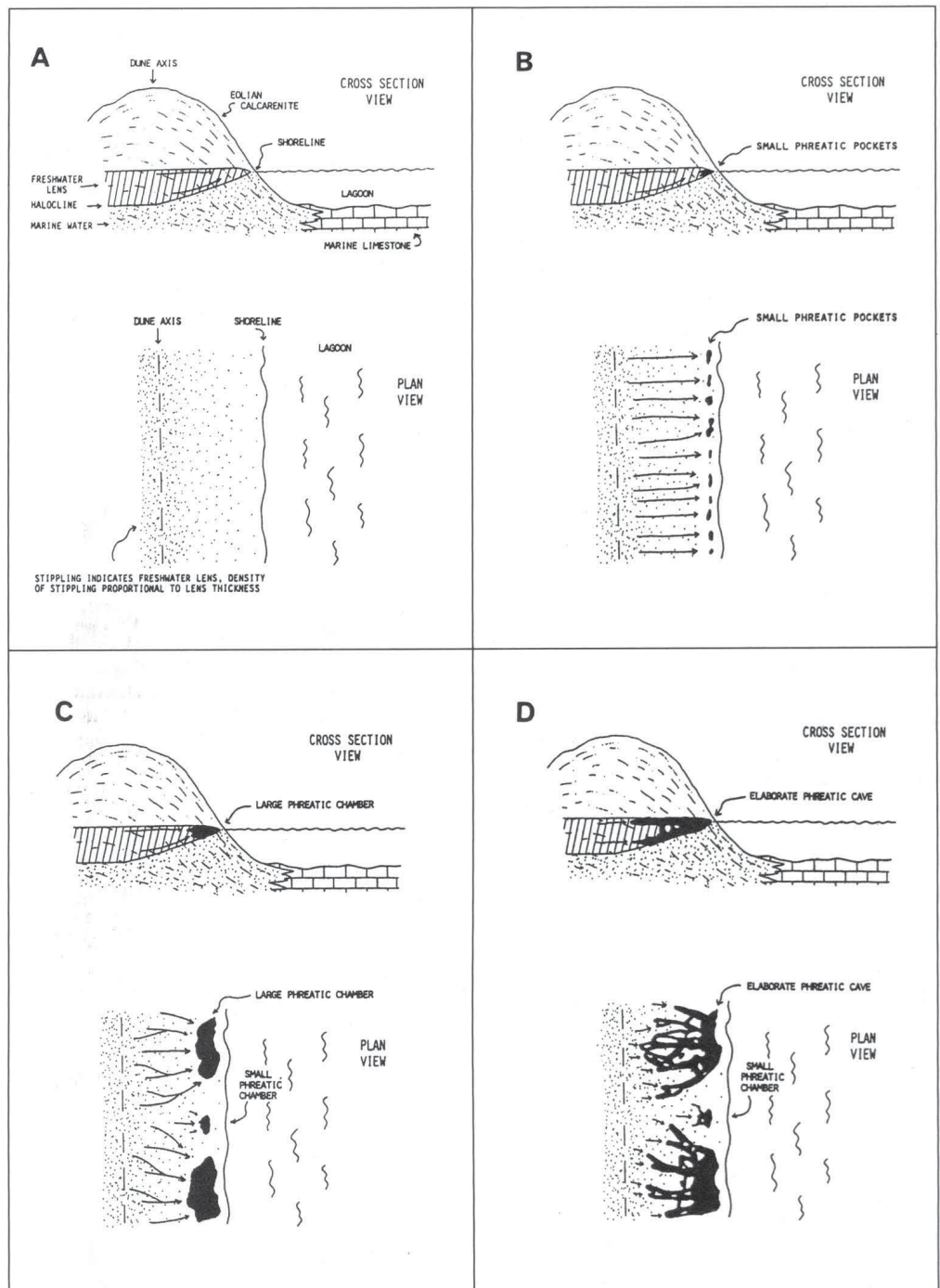
enhanced where the sea water mixed with the fresh or brackish water being discharged from the interior of the dunes. There may also be undersaturation created by the mixing of descending CO_2-charged vadose water. The absence of flow scallops in San Salvador caves supports the flank margin model.

It was the aim of this study to analyze the relative position, general morphology, and wall rock petrology of representative horizontal dissolution conduits on San Salvador Island, Bahamas, to determine whether any recognizable signature of dissolution and diagenesis related to halocline mixing zones on carbonate islands could be identified.

METHODS OF STUDY

Seventeen caves were surveyed using methods described in Daunt-Mergens (1981). The surveys determined the geographic location of caves on the island (Figure 4), their orientation to surrounding features such as the eolian calcarenite ridges, their position relative to sea level, their passage patterns, and general cave wall morphology. Thirty-six hand samples of cave wall rock were taken from eleven individual caves. Samples were usually taken from the upper and lower portions and front and rear of the caves. The horizontal and vertical position of each sample and its original orientation in the cave wall was noted.

Cave maps and cross-sections were constructed using data obtained from the cave surveys. Prominent breakdown blocks,

speleothems, and other significant features were also recorded on the maps. Thirty-eight thin sections of cave wall samples were analyzed. Allochem and orthochem percentages were determined. Thirty-eight samples of cave wall rock were crushed, sieved, and analyzed by X-ray diffraction. A summary of the petrologic data and XRD data is shown in Table 1. A complete review of the raw data is available in Vogel (1988).

CAVE MORPHOLOGY

Most of the caves of San Salvador can be assigned to one of two types: depression karst (pits and banana holes) or horizontal dissolution caves (Mylroie, 1988). Descriptions of five representative horizontal dissolution caves found on San Salvador are given below.

Stouts Lake Caves

Stouts Lake is an hypersaline lake located at the southeastern end of San Salvador Island. Four caves are known along its shoreline at elevations of roughly 2 to 6 m above present sea level (lake level). Their locations are shown in Figures 4 and 5. The four caves have both similar morphologies and similar positions along steep slopes adjacent to the lake. George Storr's Cave (Figure 6), is the largest of the caves. It is entered through a small horizontally elongate opening on the flank of a lithified eolian
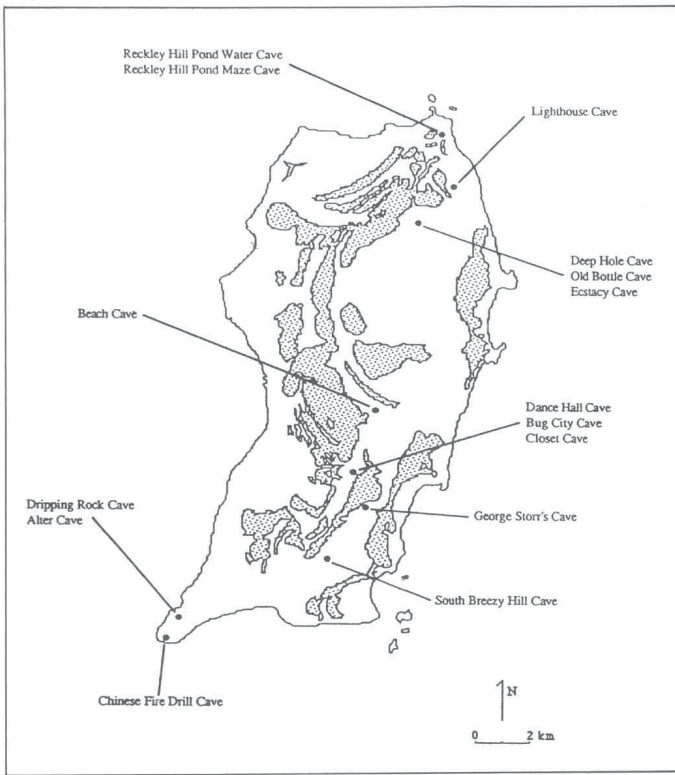
dune. The breach into the cave has been caused by slope retreat. The cave consists of a large main chamber that measures 20 m across and up to 3.5 m high, and smaller passages that lead back into the dune. These smaller passages end abruptly. The west side of the main chamber contains numerous large breakdown blocks, which is an unusual occurrence among the dissolution caves on the island.

Dance Hall Cave, Bug City Cave and Closet Cave on the northwest shore of Stouts Lake lie on the flank of one of the highest dune ridges on the island. These caves are associated with a small escarpment about 2 to 3 m above present sea level. Dance Hall Cave (Figure 7) is the best example of a flank-margin cave. As with many of these flank-margin caves, the entrance is low and elongate (0.6 m high x 9 m long) and opens into the central chamber (17 m wide, 7 m deep, and 1 to 1.5 m high) at 2 to 3 m above current lake level (sea level). Numerous smaller passages radiate from this main chamber into the interior of the dune ridge. The largest of these passages is a 1.5 m diameter tube which can be traversed roughly 20 m northwest into the dune. This passage abruptly ends with a few tiny tubes that feather-out into the dune. All other traversable passages also feather-out or end abruptly.

Bug City (Figure 8) and Closet Cave (Figure 9) are very similar in morphology to Dance Hall Cave, although on a smaller scale. Both have small entrances that open into a main chamber with small conduits that radiate duneward.
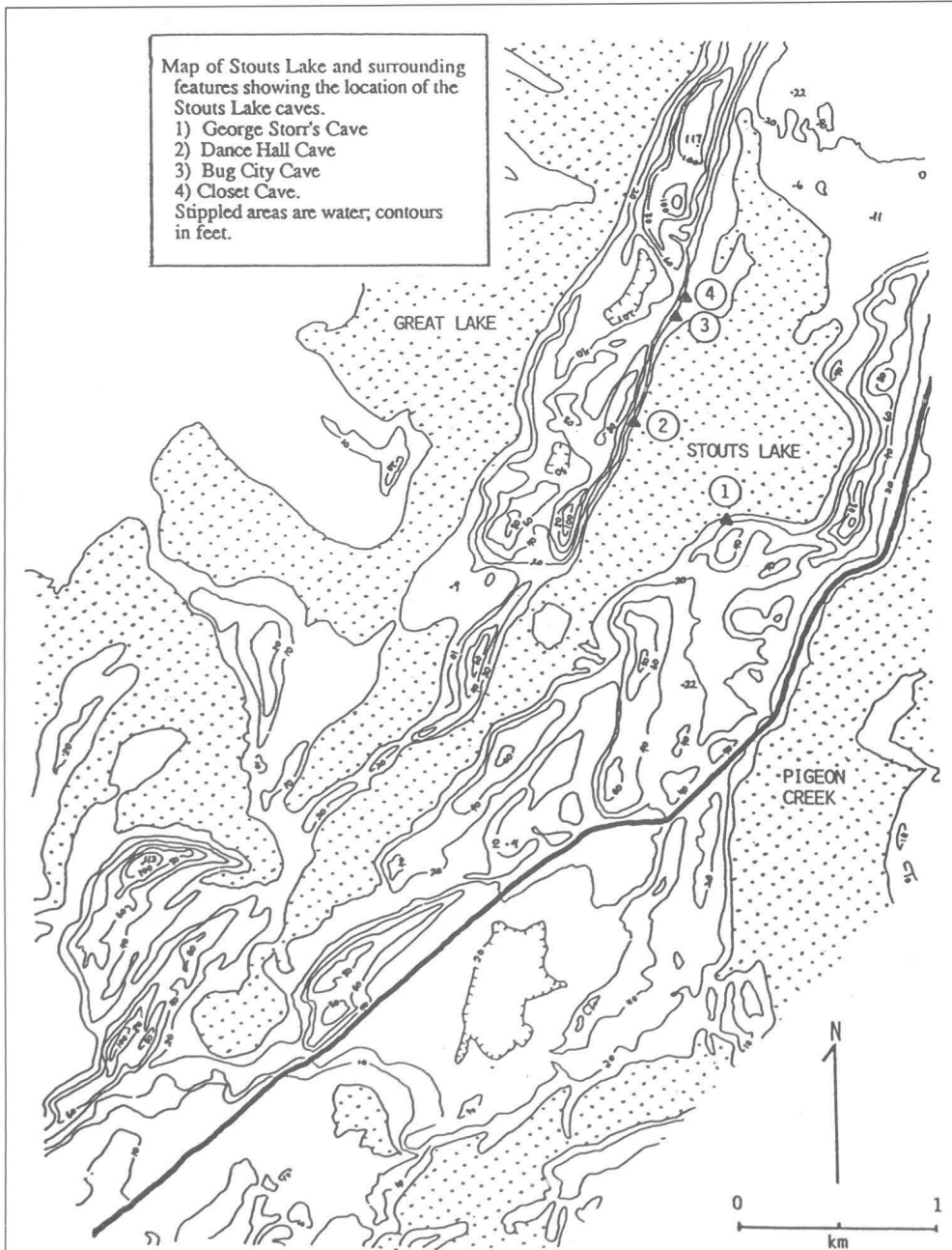
Figure 5. Detailed map of the Stouts Lake area, showing the relationship between four caves and the surrounding topography of lakes and dune ridges. Under a past, +6 m high sea level, the dune ridges would have become separate islands and the caves would have been in the freshwater lens.

Map of Stouts Lake and surrounding features showing the location of the Stouts Lake caves.
1) George Storr's Cave
2) Dance Hall Cave
3) Bug City Cave
4) Closet Cave.
Stippled areas are water; contours in feet.

GREAT LAKE

STOUTS LAKE

PIGEON CREEK

Figure 6. Map of George Storr's Cave, San Salvador Island.

Figure 7. Map of Dance Hall Cave, San Salvador Island.

Figure 9. Map of Closet Cave, San Salvador Island.

Figure 8. Map of Bug City Cave, San Salvador Island.
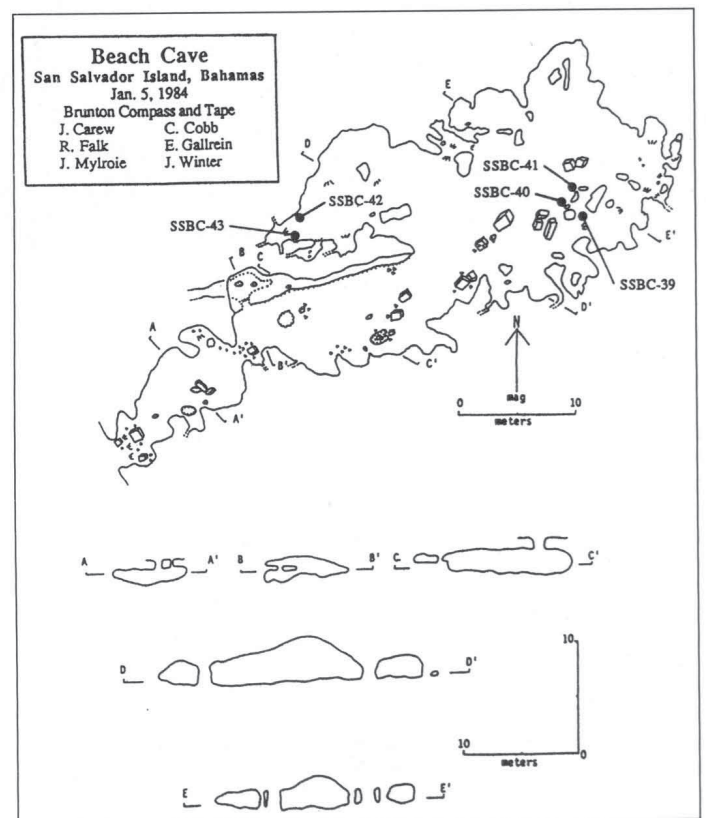
Figure 10. Map of Beach Cave, San Salvador Island.

Beach Cave

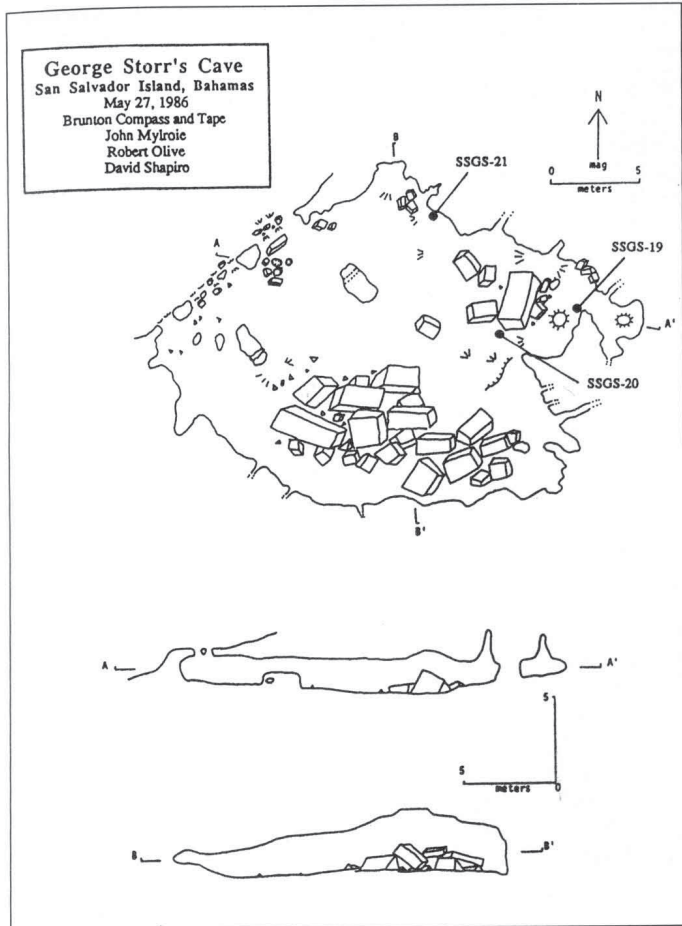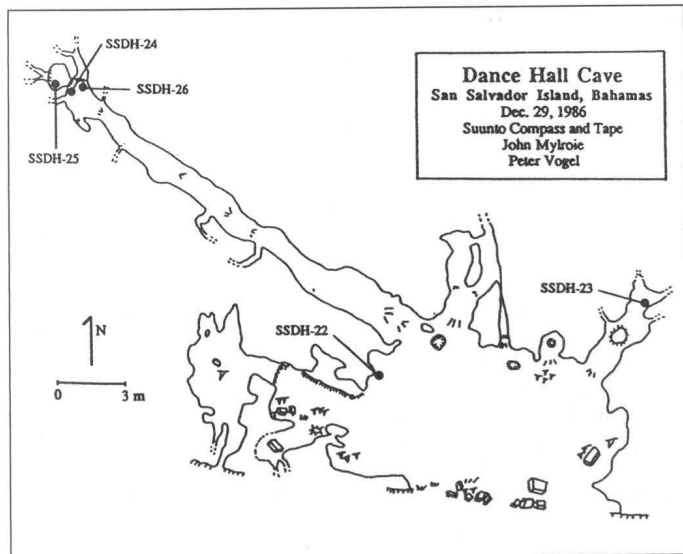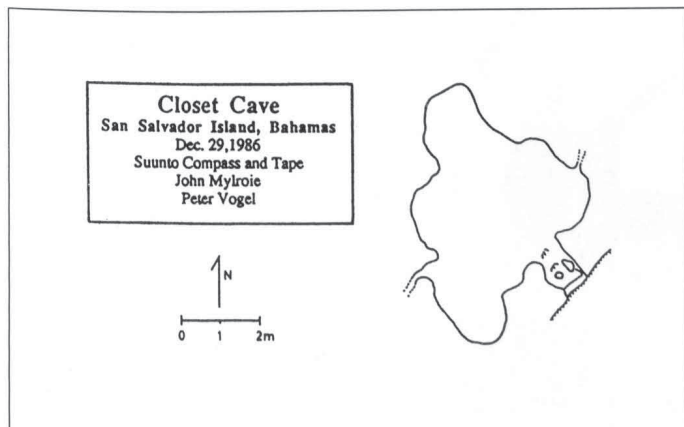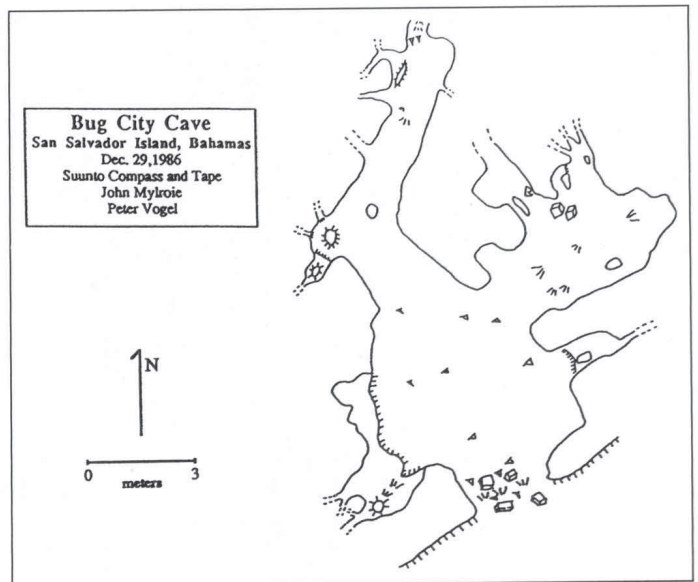Beach cave (Figure 10) is the second largest known cave on San Salvador, and exhibits flank margin cave morphology. The main entrance opens through the ceiling at approximately 7 m above present sea level. There are other minor horizontal entrances at the southwest end of the cave at the dune margin. The slightly undulating bedrock floor of the main chamber lies about 3 m below the main entrance. The cave is elongate parallel to the axis of the dune. It is approximately 57 m long, while its width varies from 1.5 m to 25 m. The cave consists of several interconnected chambers with ceiling heights reaching 3.5 m. The undulating roof is supported by bedrock pillars which divide the cave into chambers that are arranged in a "beads on a string" pattern. Small solution conduits radiate from most of the walls in the cave. Some collapse material is present, but it does not appear that collapse was a significant factor in the cave's development.

General Cave Morphology

Compared to karst regions in interior continental settings, the morphological features of subaerial Bahamian caves described above, such as the general cave passage pattern and wall rock

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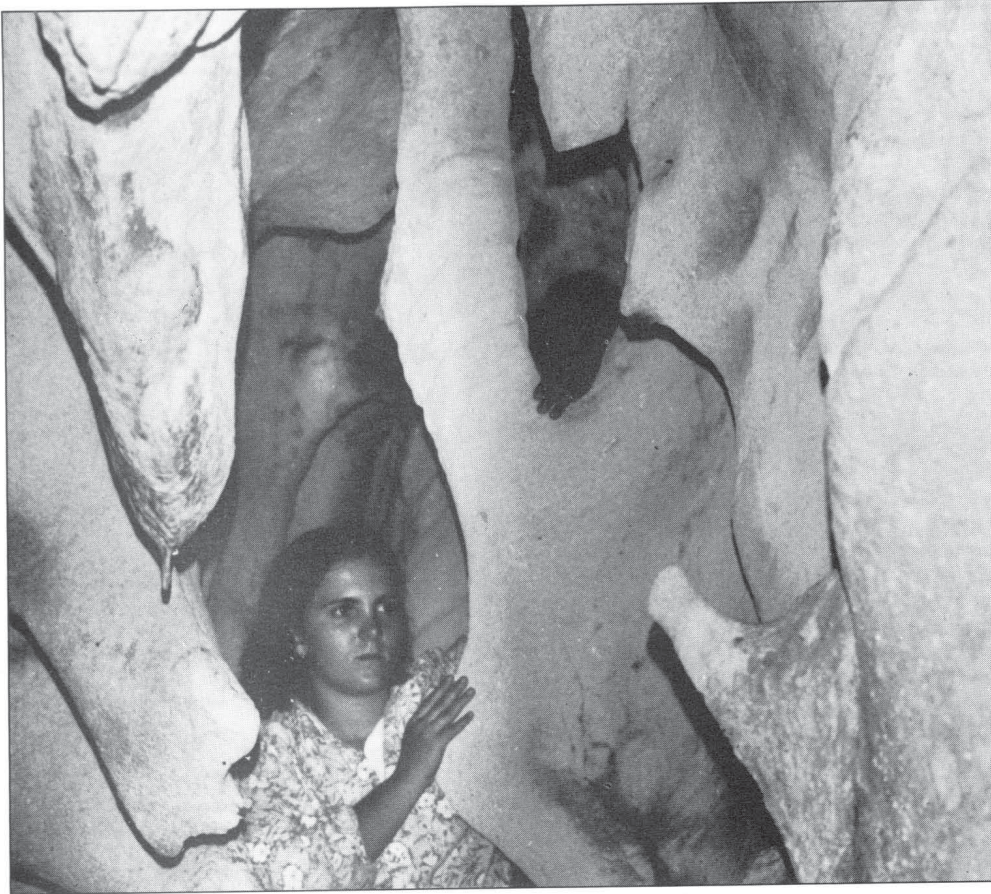
morphology, are distinctive. This morphology is probably caused by the discharge pattern of the freshwater lens on such carbonate islands. Despite the individual characteristics that each horizontal dissolution conduit may display, these caves generally have a central chamber from which numerous smaller tubes radiate into the interior of the dune. Despite minor differential dissolution of the wall rocks, the overall pattern of cave passage development does not appear to be lithologically, stratigraphically, or structurally controlled. Flow scallops, which are generally a prominent feature on the cave walls in older and more indurated limestones, are obscure or absent in these Bahamian caves. The caves fit the morphological pattern explained by the flank margin model (Mylroie, 1988).

Macroscopic wall rock texture varies between the upper and lower portions of many of these caves. Rocks near the ceiling are relatively smooth and contain widely spaced pores that range from 4 to 17 mm in diameter. Rocks near the floor are also smooth, but commonly contain closely spaced pores that range from 1 to 4 mm in diameter and are often associated with bedding planes (Figure 11). Differential relief on the dissolution surfaces in the caves tends to be greater low in the caves passages than it is in the upper regions. Similar, but more highly developed, sponge-

like dissolution patterns have been found in active mixing zones along the coast of Yucatan, Mexico (Back, *et al.,* 1986), and in blue holes on Andros Island, Bahamas (Smart, *et al.,* 1988).

PETROLOGY

Wall Rock Composition

The rocks in which the caves have formed are composed of a variety of allochems that include ooids, peloids, aggregate grains, lithoclasts, forams, algae, and other bioclasts (Table 1). Grain size ranges from micrite to very coarse sand, but the majority are well sorted and have an average size of 0.32 mm (medium sand). Composition studies of the dune rocks reveal generally well sorted fine-to-medium sand composed of skeletal fragments, ooids, pellets, and peloids. With the single exception of George Storr's Cave, all other studied caves have formed in rocks that were deposited as eolianites and associated beach facies. The lower portion of George Storr's Cave is developed in low porosity matrix-supported rock interpreted as a lagoonal mud. This low porosity rock should have restricted both fluid circulation and dissolution conduit development. As the cave passage
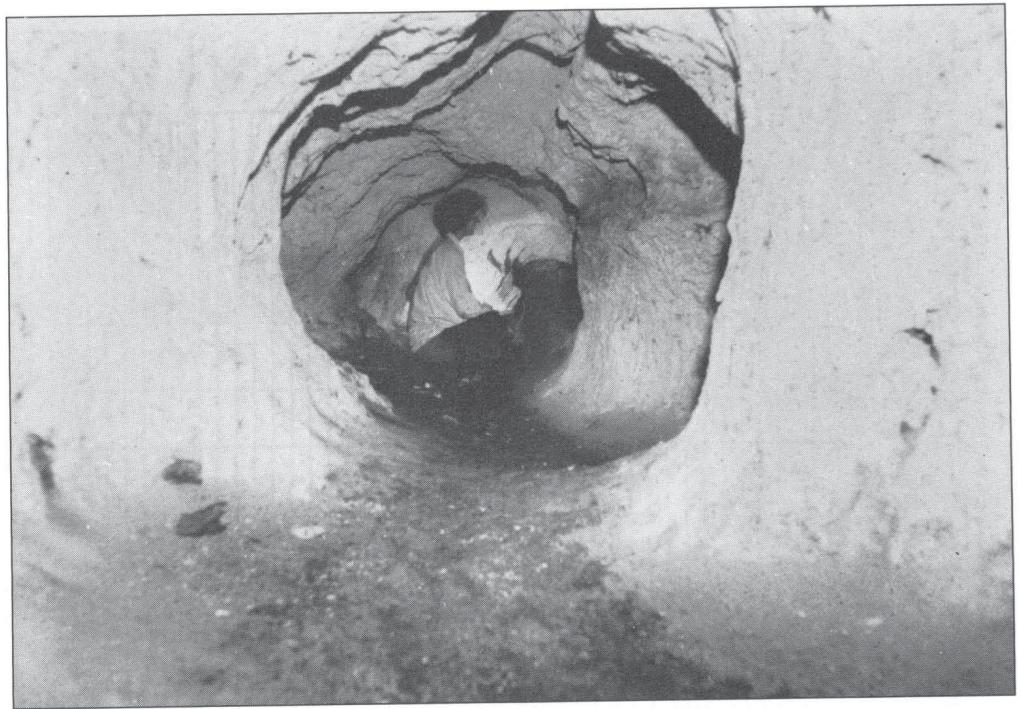
morphology of George Storr's Cave does not change when it crosses into the lower porosity rock, it may be that the cave enlarged by ablation of the rock surface, rather than by enlargement and interconnection of intergranular porosity. The features in George Storr's Cave indicate that cave location is not dependent on lithology, but rather on the position of the freshwater lens and halocline mixing zone.

Diagenesis

Several factors play a role in cementation in carbonates: a) the rate and volume of water movement; b) fluid chemistry; and c) the effects of pH, CO_2 activity, temperature, and pressure on mineral solubilities. Four major diagenetic environments can have affected the wall rock of caves located at 0 to 7 m above present sea level. These include the marine phreatic, halocline mixing zone, freshwater phreatic, and vadose diagenetic environments. Each of those diagenetic environments produces unique dissolution and precipitation characteristics.

The primary diagenetic effects of marine water are micritization and grain overgrowth. Fibrous and equant micritic aragonite; and fibrous, micritic, and bladed high magnesium calcite cements are stable under marine conditions (Harris, *et al.,* 1985). The diagenetic environment in the mixing zone promotes neomorphism of aragonite to calcite as well as enhanced dissolution of calcite (Back, *et al.,* 1986). The freshwater phreatic diagenetic environment promotes leaching of aragonite and high-Mg calcite, and neomorphism of metastable grains to low-Mg calcite. Cementation by equant and isopachous bladed low-Mg calcite is common. Syntaxial overgrowths on echinoderms and dolomitization can also occur (Harris, *et al.,* 1985; Longman, 1980). Freshwater vadose diagenetic environments are characterized by dissolution of aragonite and high-Mg calcite, and precipitation of low-Mg meniscus, pendant, equant, and whisker crystal cement. A wide variety of cement types and complex cementation patterns are indicative of non-marine cementation (Chafetz, *et al.,* 1985).

Cementation within each cave varies in both type and abundance, from layer to layer and from locality to locality. Micrite, micro and pseudospar, radial bladed, isopachous fibrous, meniscus, and whisker calcite cement are common among the wall rock samples studied. Twenty-two percent of the wall rock samples contain microspar and pseudospar cement.

Micrite cement is found as a rim cement that coats grains within the rocks, and as thin surficial crusts on cave walls and as linings in small dissolution tubes in the rock. Minor amounts of radial bladed, isopachous fibrous, and meniscus cements occur in some samples. These cements usually occur as relict primary cements that are normally obscured by second generation cements. Only in samples from Lighthouse Cave have isopachous fibrous aragonite cements been seen. Sample No. 55 from Lighthouse Cave (Table 1) contains a primary isopachous fibrous fringe which has a secondary micrite rim cement deposited on its
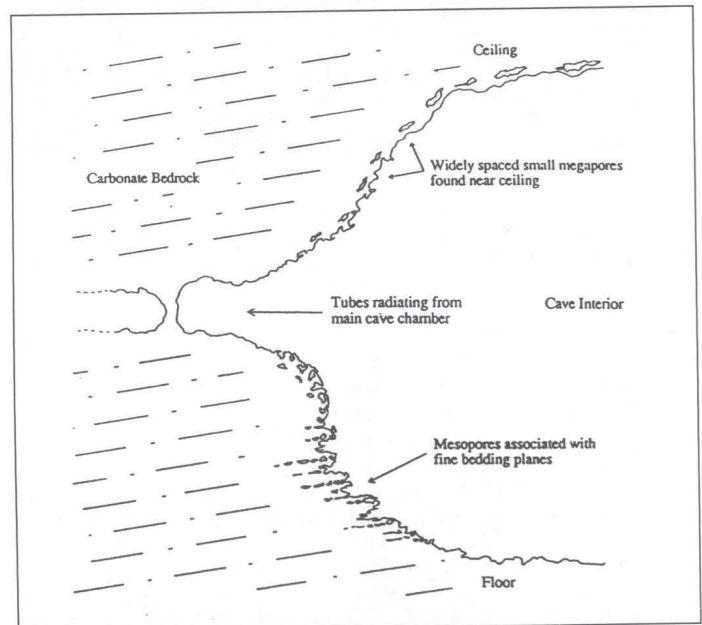
Figure 11. Schematic cross section of a typical cave wall on San Salvador Island, showing change in wall rock morphology from ceiling to floor. The ceiling to floor distance ranges from just under 1m to over 3m, depending on the cave.

surface. According to Scholle (1978) the isopachous fibrous cement is characteristic of lower intertidal and submarine cementation.

Calcite whisker crystal cement, also called needle fibre cement, is associated with vadose diagenesis (Scholle, 1978) and is usually found among these samples as low-Mg calcite crystals in intragrain voids.

No simple or obvious relationship exists between cement type or abundance and cave location or sample position within a cave. The cements indicate a complex diagenetic history. If a clear relationship exists, it may require a more detailed sampling regime than was used in this preliminary study.

Micritization of cave wall rocks seems to have been an important diagenetic process. While most samples exhibit at least some micritization, the degree of micritization ranges from micrite envelopes to total micritization of allochems and orthochems. Micritization decreases inward from the cave wall surface. Micritization of carbonate grains and cements has been associated with boring endolithic algae, bacteria, and fungi within marine sediments and on rock surfaces (Bathurst, 1975). Although not addressed in the literature, nonbiological micritization through retrograding neomorphism may occur as a result of the intense chemical activity within the halocline mixing zone.

Sample	Avg. Grain Size	Porosity	Micrite	Spar	Ooids	Peloids	Forams	Algae	Other Bio	Intraclasts	Position	Name	calcite	gypsum	aragonite	halite	dolomite	vugs	inter	intra	moldic	oomoldic	Primary Cement
SSBH-1	.16 mm	11%	-	23%	47%	11%	1%	-	1%	3%	high	peloosparite	60		40			√	+			+	microspar
SSBH-2	.15 mm	6%	-	13%	48%	9%	-	-	-	-	low	oosparite	62		38			√	+			+	microspar
SSBH-3	.40 mm*	18%	12%	14%	11%	39%	+	1%	2%	1%	back/high	oopelsparite	86		14			√			√		microspar & pseudospar
SSAC-4	.25 mm*	4%	-	18%	45%	19%	+	-	+	10%	high	intrapeloosparite	58		38	4		+	√		√		microspar & pseudospar
SSAC-5	.20 mm	35%	-	15%	36%	11%	+	-	+	2%	low	peloosparite	60		40			√					pseudospar
SSAC-6	.30 mm	7%	-	31%	45%	9%	-	-	+	5%	back/middle	oosparite	57		43			√	+		+		pseudospar
SSGS-19	.20 mm	15%	56%	19%	-	5%	2%	-	+	-	back/middle	bio/pel bearing micrite	43	4	11	30	12	√	+	+			original micrite matrix
SSGS-20	.30 mm	31%	32%	4%	1%	21%	3%	3%	3%	-	high	pelmicrite	79		21				+	+			micrite
SSGS-21	.20 mm	minor	20%	29%	5%	11%	12%	2%	11%	-	low	pelbiosparite	72		19		9	+	+				microspar matrix
SSDH-22	.25 mm	25%	2%	12%	1%	6%	12%	9%	10%	-	high	biosparite	100					√	+	+			microspar
SSDH-23	.30 mm	42%	19%	16%	-	12%	5%	2%	1%	-	back/middle	pelmicrite	68		15	17		+	√	√			micrite & microspar
SSDH-24	.30 mm	26%	2%	30%	2%	9%	9%	7%	1%	-	tube/middle	biosparite	100					+	√	√			microspar
SSDH-25	.25 mm	35%	2%	14%	1%	7%	16%	4%	2%	-	tube/low	biosparite	100					√	√	√	+		microspar
SSDH-26	.25 mm	14%	-	23%	-	13%	13%	6%	8%	-	tube/high	pelbiosparite	85		15			√	√	√	+		microspar
SSEC-38	.25 mm*	18%	53%	20%	4%	1%	1%	+	-	6%	middle	intramicrite	60		40			√	+	+			high micritization
SSBC-39	.40 mm	45%	4%	12%	-	4%	17%	+	6%	-	middle	biosparite	94		6			+	√	√	+		microspar
SSBC-40	.40 mm*	17%	54%	26%	-	-	2%	-	1%	-	low	fossiliferous micrite	68	10		22		√	+	√			micrite & microspar
SSBC-41	.50 mm	44%	16%	18%	-	6%	-	3%	3%	-	high	pelsparite/micrite	87	5	2	6		√	√	√	+		microspar & micrite
SSBC-42	.40 mm	31%	10%	38%	-	4%	7%	1%	1%	-	back/high	pelbiosparite	97		3			√	√	√	+		microspar
SSBC-43	.40 mm	31%	2%	12%	-	11%	18%	6%	3%	3%	back/low	pelbiosparite	100					+	√	√			microspar
SSCF-49	.25 mm	14%	1%	19%	55%	10%	-	1%	2%	+	high/back	peloosparite	61		39			√	+	+	+	+	microspar
SSCF-50	.25 mm	9%	-	18%	60%	11%	-	-	-	1%	medium/back	peloosparite	64		36			√	+	+	+	+	microspar
SSCF-51	.20 mm	14%	-	22%	50%	12%	-	1%	+	1%	medium/low	peloosparite	52	30	15	3		+	√	+			pseudospar
SSCF-52	.35 mm	31%	3%	26%	8%	28%	-	1%	2%	+		low	pelsparite	61		39			√	+	+			micro & pseudospar
SSCF-53	.30 mm	14%	1%	19%	30%	27%	-	-	-	3%	low/outside	peloosparite	49	8	43			+	√		+		pseudospar
SSLH-54	.30 mm	32%	-	15%	-	25%	2%	17%	6%	-	high	pelbiosparite	83		17			+	√	+			microspar & fib rim cement
SSLH-55	.80 mm	36%	1%	-	-	12%	7%	20%	10%	2%	medium/high	pelbiosparite	63		17		20	√	√		+		micro & micrite
SSLH-56	.50 mm	30%	1%	10%	-	20%	2%	28%	7%	-	medium/low	pelbiosparite	85		15			+	√		√		microspar
SSLH-57	.60 mm	42%	2%	7%	-	12%	1%	25%	7%	-	low(subaqueous)	pelbiosparite	85		5	10		+	√		√		microspar
SSRW-58	.20 mm	42%	1%	12%	-	28%	6%	5%	1%	-	high	biopelsparite	100					+	√		√		microspar
SSRW-59	.25 mm	39%	5%	28%	-	20%	3%	2%	-	-	middle	pelsparite	100					+	√		√		microspar
SSRW-60	.30 mm	49%	2%	8%	-	33%	4%	2%	-	-	low/ceiling	pelsparite	100					√	√	+	√		microspar
SSRW-61	.30 mm	47%	8%	18%	-	18%	3%	4%	1%	-	low (intertidal)	pelsparite	66			34		√	√	+	√		microspar
SSRM-62	.30 mm	15%	10%	73%	-	-	+	+	+	-	middle	fossiliferous sparite	100					√			+		microspar
SSRM-63	.30 mm	6%	13%	68%	-	-	6%	2%	1%	-	high	fossiliferous sparite	100					√	+	+	√		microspar
SSRM-64	.30 mm	29%	18%	45%	-	2%	4%	+	-	-	low	fossiliferous sparite	100					√	+	+			microspar

* = poorly sorted   - = N/A   + = minor amounts   √ = large amounts

Table 1. Petrographic and X-ray diffraction data from 36 samples from 10 caves on San Salvador Island, Bahamas. Sample codes are: SSBH = South Breezy Hill Cave; SSAC = Altar Cave; SSGS = George Storr's Cave; SSDH = Dance Hall Cave; SSEC = Ecstasy Cave; SSBC = Beach Cave; SSCF = Chinese Fire Drill Cave; SSLH = Lighthouse Cave; SSRW = Reckley Hill Pond Water Cave; SSRM = Reckley Hill Pond Maze Cave. See Figure 4 for cave locations.
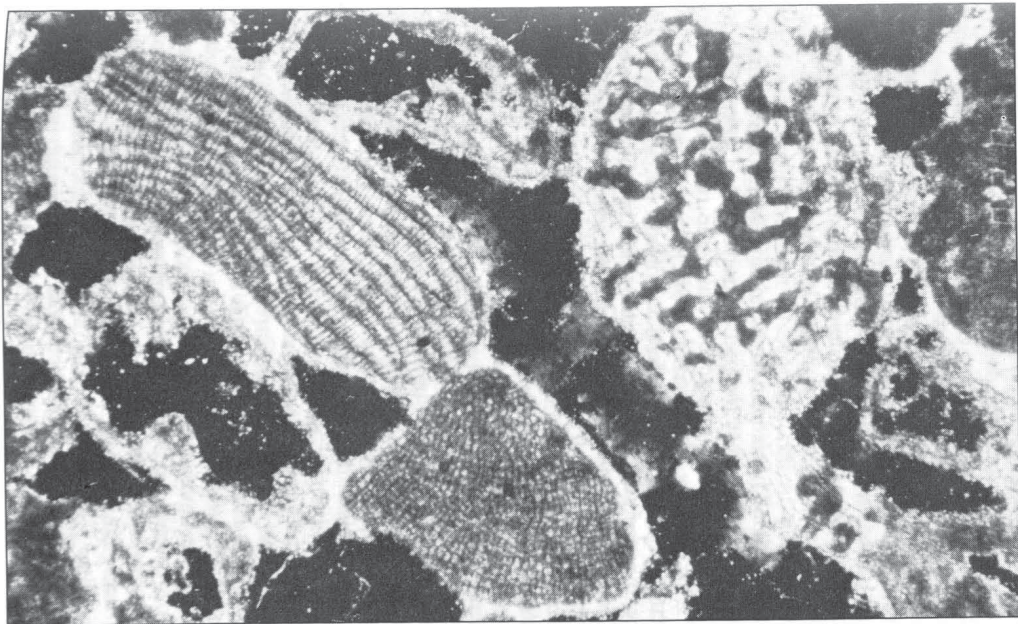
Photomicrograph of thin section showing fringing meniscus-style low-Mg calcite cementation indicative of freshwater vadose conditions. Magnification 50X.
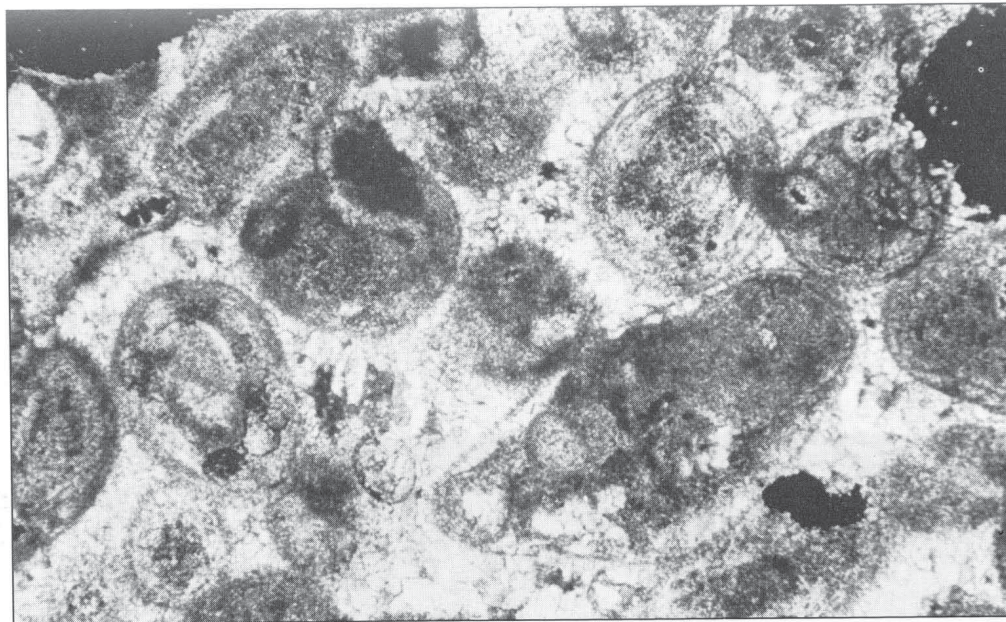
Photomicrograph showing evidence of an initial thin fringing cement followed by equant calcite pore fill, indicative of possible early marine cementation followed by freshwater phreatic cementation. Magnification 100X.
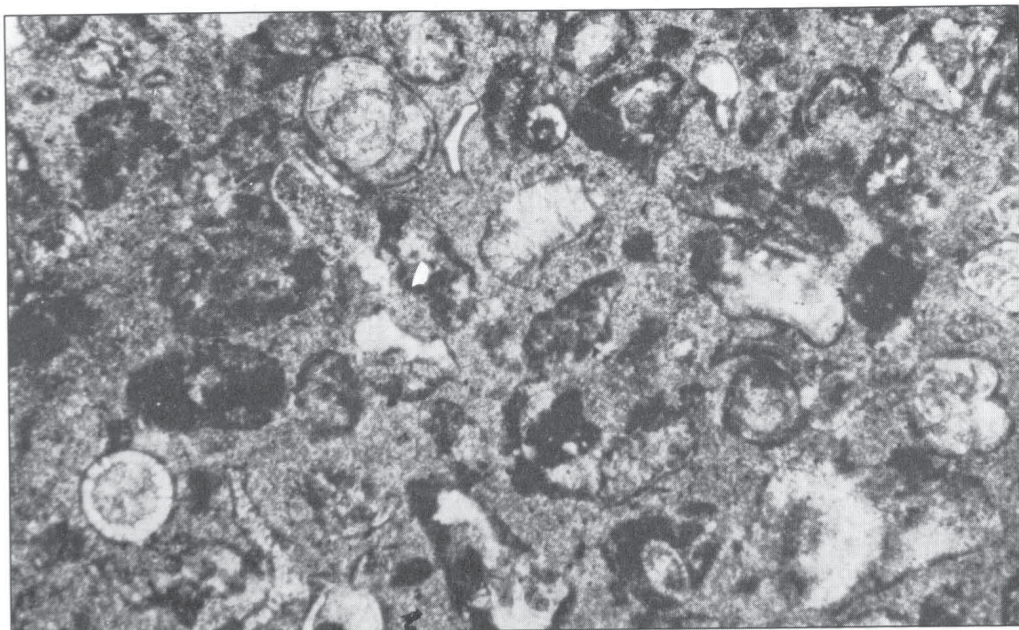
Photomicrograph of thin section of a micritic rock with little primary of secondary porosity. Magnification 50X.
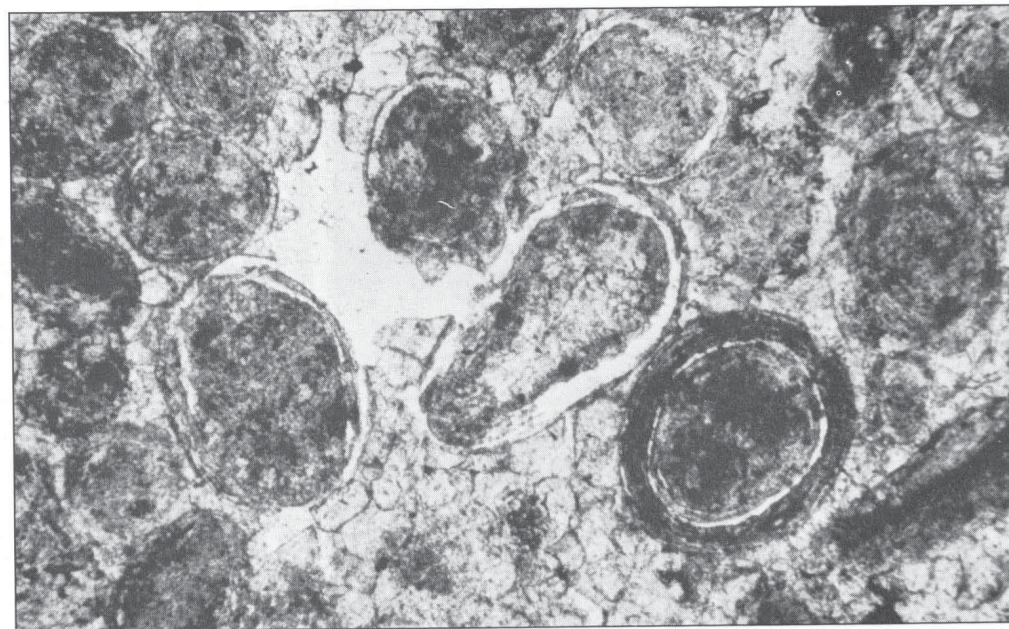
Photomicrograph of cave wall rock showing partial replacement of ooid laminae by gypsum. Magnification 100X.

Mineral Assemblages

As might be expected, low magnesium calcite is the dominant mineral constituent in all of the wall rock samples, and its abundance ranges from 45% to 100%. The calcite occurs primarily as originally low-Mg calcite cement, but neomorphism of metastable aragonite and high-Mg calcite allochems has resulted in the formation of additional amounts of low-Mg calcite.

The next most abundant mineral is aragonite, which is present in a majority of the wall rock samples. Two caves containing a biosparite and a biomicrite, respectively, show no petrographic or X-ray evidence of aragonite. Elsewhere, the aragonite is present as ooids, green algae, scleractinian coral, molluscs, and as intertidal and submarine cements.

X-ray diffraction analyses revealed that gypsum was present in 7 of the 38 samples examined (Table 1). Analysis of sequential subsamples taken from the exterior to the interior of the same sample indicates that the majority of the gypsum occurs as a finely crystalline crust on the walls of the cave. XRD analysis of one of these crusts indicates that it contains 49% gypsum. Petrographic analysis has revealed gypsum within the concentric layers of ooids. It has not been determined whether the original aragonitic ooids were partially replaced by gypsum, or whether the concentric rings were leached out and gypsum later filled the cavities. Meyers and McClain (1988) found similar material on Gun Cay, in the northwest Bahamas. There the infilling mineral was bassanite.

X-ray analyses have indicated minor occurrences of halite among the cave rock samples, but the only petrographic evidence of halite consists of a single halite lamination within a surficial crust. Samples which were collected subaqueous or in the intertidal zone contain halite, but these can be explained as a result of evaporation of the saline water in which they were immersed when collected.

Dolomite precipitation in the halocline mixing zone has been theoretically suggested (Back, et al., 1986), and thought to be indicative of halocline mixing zone diagenesis (Swart, et al., 1987; Smart, et al., 1988). Dolomite has been identified along the halocline (Land, 1970; Badiozamani, 1973; Ward and Halley, 1985). Measurements of 87Sr/86Sr ratios and 234U/238U and 230Th/238U activity ratios in dolomites from a core taken on San Salvador Island indicate that dolomitization of Pliocene rocks resulted from exposure to mixing-zone geochemistry during Pleistocene sea level changes, (Swart, et al., 1987). Even though studies of modern mixing zones and cave wall rocks reveal that the water is supersaturated with respect to dolomite, no evidence has been found of modern dolomite formation (Back, et al., 1986; Smart, et al., 1988; Stoessell, et al., 1989).

Considerable effort was undertaken to determine whether dolomite was present in any of the wall rocks of the caves on San Salvador, but X-ray analyses revealed dolomite in only two samples (Table 1). A single rhombohedron that may be dolomite was seen in a thin section.

The limited occurrence of dolomite might be regarded as evidence that these caves did not form in the halocline mixing zone, but work by Smart, et al., (1988) revealed no dolomite in cave wall rocks that lie within a modern Bahamian halocline mixing zone. Dedolomitization by the action of meteoric waters may occur (Adams et al., 1984), but no evidence of this was seen among our samples.

Our work and that of Smart, et al. (1988) suggest that dolomite should not be regarded as positive evidence for or against the implication of halocline mixing zone diagenesis. Bathurst (1975) suggested that the growth of dolomite is a slow process in any natural environment. Estimated rates of deposition are about 0.1mm/1,000 years, and that exceedingly slow rate probably accounts for the rarity of recent deposits. It is possible that sea level highstands were long enough for dissolution conduit development, but were too short for significant dolomitization to occur.

The minor occurrence of gypsum, halite, and dolomite is sporadic. Suggestions for their origin include infiltration from aerosols and salt spray (Gardner, 1983), seepage from the overlying rocks and soils, or flooding of the cave with saline or hypersaline waters during a previous higher sea level. The capability of hypersaline water to promote the replacement of aragonite and calcite by dolomite, and to precipitate gypsum, anhydrite and halite has been documented by Harris, et al. (1985).

The mineral deposits found within and encrusting upon the wall rocks of San Salvador's horizontal dissolution caves do not provide distinctive evidence of cave formation along the halocline. We suggest that after sea level fall and cave abandonment, the gypsum, halite, and dolomite probably precipitated on the dry cave walls from evaporation of downward percolating meteoric water carrying dissolved minerals. In dry subaerial caves with relative humidity below 75 to 90%, water soluble minerals are deposited. The most common form of gypsum in the caves of San Salvador is as a granular crust probably formed by water seeping uniformly from pores in the wall rock. Precipitation of gypsum from vadose water contaminated by sea spray may have resulted in the concentration of small amounts of magnesium that allowed the formation of rare dolomite rhombs. The evaporite minerals remain in solution and are carried out of the system in wet caves and caves with nearly 100% humidity (White, 1988). Despite the complexity of the cements and the secondary vadose precipitates, it is interesting to note that metastable aragonite is still present in most caves, and may be as much as 43% of the cave wall rock (Table 1). The aragonite does not show association with the evaporite minerals, but appears to be present mostly as original allochems.

Porosity

In addition to mechanical and chemical compaction, the main factors that control porosity in carbonates are dissolution, cementation, and dolomitization. (Harris, et al., 1985). The diagenetic processes require both a permeable rock and a mechanism for flushing chemically active water through the rock. On a small island this flushing is controlled by tidal pumping, hydrostatic head, sediment compaction, thermal expansion, and regional rock fabric.

Two general observations can be made about the macroscopic porosity within the horizontal caves on San Salvador:

1. Macroporosity sharply decreases from the cave wall into the bedrock.

2. The amount of differential dissolution of, and macroporosity in, the cave wall increases from ceiling to floor.

As determined from thin section, porosity of our samples ranges from less than 1% to 49%, and both fabric-selective and nonfabric-selective porosity is present. The fabric-selective porosity includes interparticle, intraparticle, and moldic porosity. The nonfabric-selective porosity is mostly vugs. Borings, which may be considered either fabric-selective or not, are also present.

Porosity from vug development is the most common type found within the wall rock samples. At least a few vugs are found in each sample. According to Longman (1980), vugs are associated with vadose, freshwater phreatic, and mixing zone diagenesis.

Although less abundant than vugs, interparticle porosity occurs in most samples. Intraparticle and moldic porosity are less abundant. Leaching of bioclastic grains, peloids, and ooids has produced abundant empty micrite envelopes. Porosity produced by borings is occasionally found within orthochems and allochems.

Within nearly all horizontal caves on San Salvador, the samples taken low in the cave have the highest porosity. In George Storr's Cave that condition is inverted because the rock in the lower portions of the cave is matrix supported, and thus started its diagenetic history with little effective primary porosity. As such, porosity enhancement was hindered by the difficulty of moving dissolutionally aggressive water through the sediment. The curved solution surfaces of George Storr's Cave pass through the porous upper eolianite into the tight lower lagoonal mud facies without any change in gross morphology, implicating ablative dissolution as the mechanism of wall rock destruction, as opposed to coalescence of small scale porosity. It is surprising that cave development has penetrated into that rock without the influence of jointing or lithological boundaries; yet the fact that it does has important implications for the dissolutional aggressivity of the water which formed the cave, and the importance of the position of the water table and halocline over other potential controlling factors.

Petrographic studies of surface samples on San Salvador Island have been done by Hutto and Carew (1984), Sims (1988), and Stowers (1988). Hutto and Carew (1984) found that the upper portions of Pleistocene dune ridges on San Salvador are well sorted and are composed primarily of ooids with only minor skeletal fragments and pellets. As one would expect, vadose diagenetic features dominated those samples. Sims (1988) reported Pleistocene eolianites from south-eastern San Salvador that were predominantly oosparites, peloid-bearing oosparites and biosparites. They were all characterized by meniscus, pendulous, and whisker cements. Compositional similarities and differences

between eolian ridges are attributed to variations in sediment sources and times of formation. Stowers (1988) found that surface samples of Late Pleistocene dunes from south-western San Salvador contain either equant, meniscus, meniscus with equant overprinting, or micrite cements. In his samples porosity averaged 15%, and did not exceed 26%. These various studies show that the upper dune samples, which have been in the vadose environment since their deposition, demonstrate markedly less diagenesis than similar rocks located in the 0 to 7 m elevation range.

SUMMARY AND CONCLUSIONS

San Salvador Island, Bahamas, possesses numerous horizontal caves approximately 0 to 7 m above present sea level that are typically found along the flanks of Late Pleistocene eolian dune ridges and their underlying marine foundations. Caves located within these vertical constraints are associated with past higher sea level(s) associated with Late Pleistocene interglacials that occurred approximately 125,000 to 85,000 years ago. The caves were probably developed by halocline associated geochemical dissolution along the margins of freshwater or brackish lenses which were perched on the high stand of sea level during that time. The location, vertical position, and morphology of these caves are consistent with present theoretical models of cave formation associated with a freshwater lens (Back, *et al.*, 1986; Mylroie, 1988; Smart, *et al.*, 1988). Except for the general absence of dolomite, the San Salvador caves exhibit most of the characteristics expected from halocline cave formation. Cave morphology is not lithologically, stratigraphically or structurally controlled, and the morphology suggests a process unique to caves in the Bahamas and geologically similar carbonate islands and coasts.

Petrologic analyses indicate that the caves are developed primarily in eolian facies, but some more poorly sorted wall rocks are interpreted as beach, intertidal, and subtidal lagoon facies. Wall rock samples are composed predominantly of ooids and bioclastic grains, with cements of marine phreatic, freshwater phreatic, halocline mixing zone and vadose diagenetic origins. Most samples show evidence of micritization, and porosities vary from 49% to less than 1%. The wall rocks are composed primarily of calcite, but also contain aragonite and minor amounts of gypsum, halite, and dolomite; the minor minerals are evaporites attributed to current vadose conditions. The survival of significant amounts of primary aragonite is important in assessing the duration of the various diagenetic environments. The amount of diagenesis within the caves is substantially greater, however, than that in similar rocks at higher elevations that have experienced only the vadose environment since deposition.

The rocks in which the caves are developed are less than 150,000 years old. The sea level highstands that could have formed the caves occurred around 125,000, 105,000 and 85,000 years ago, and represent a maximum of 25,000 years of time for speleogenesis. Cave formation must have been rapid, and as a type of diagenesis,

proceeded more rapidly than dolomitization or complete loss of aragonite.

Studies such as this one may increase the ability to predict porosity distribution, especially in relation to potable water and hydrocarbon reservoirs. Further investigations of halocline mixing zone diagenesis, especially that associated with eolianites, which have excellent potential as oil and gas reservoirs, may provide better understanding of the hydrologic controls which determine porosity development.

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Tertiary Caves in Norway: a Matter of Relief and Size

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Abstract: The previously postulated Miocene age of Hammarnesgrotta, Rana, North Norway was tested by numerical modelling of valley entrenchment rates and wall retreat rates of cave conduits under various environments of chemistry and hydrology (sub-glacial, interglacial and relict). Erosion rates were deduced from Uranium series dating of speleothems and from sediment yield in off-shore basins. The model demonstrates that valley erosion rates are the most crucial parameter for determining the maximum possible age of a cave. Wall retreat rates under interglacial, phreatic conditions are always large enough to create a 3-4 m diameter cave in less than 50 ka. Even the most conservative rate estimates fail to make Hammarnesgrotta older than 1.2 Ma. The most probable timing for speleogenesis is the interval 1200-320ka, which is in accordance with speleothem dates from the cave. A potentially Tertiary cave would have to be larger than 3 m diameter and be situated more than 240 m above the present-day base level of erosion. Four caves seem so far to satisfy these criteria, Stordalsgrotta, Salthölene, Svarthammarhola and Solvikhulen.

The very existence of caves and the atmosphere of their environment is intellectually stimulating to most visitors. Theoretical speculations on the age and origin of these dark cavities may become even more fascinating, as we are faced with the long recognized problem of dating a hole rather than a deposit. Hence, unique conclusions are generally difficult to present. The best we can do is to investigate the relative probability for different genetic modes, although radiometric dating of cave deposits are the best proofs for minimum age of the cavity. On the contrary, the age of the host rock, the age of the guiding fractures, the timing of uplift episodes or other crucial conditions may provide a conceptual maximum age. However, such maximum constraints are often off-scale by several orders of magnitude compared to the minimum criteria. Meaningful concordance is only documented in rare cases, like Carew et al. (1983). Karstification is a continuous process, lacking intrinsic thresholds. Consequently, the formation of karst cavities is recognized from most geological periods (Ford and Williams 1989; James and Choquette 1988; Bosak et. al. 1990).

Recently, Haugane and Grönlie (1989) advocated the view that several of the major relict caves in Nordland, Norway were formed in Tertiary times. By considering the data of off-shore Cenozoic sediment deposits and plate tectonics off the Norwegian coast, combined with assumed uplift rates, Haugane and Grönlie arrived at the conclusion that the major development of Hammarnesgrotta (Horn 1947) must have taken place in the Miocene (5 to 22 Ma B.P.), and that little alteration has occurred since. Moreover, they stated that the previous discussion on the speleogenesis of the Norwegian caves has been restricted to Pleistocene conditions, and that the implied limitation has never been documented, basically because the previous workers cited (Oxaal 1914; Horn 1947; St. Pierre 1967) lacked "geological perspective". This statement is neither correct, nor is it fair to those previous workers. Therefore, the purpose of this paper is twofold: First, to review the speleogenetic discussions with respect to Tertiary speleogenesis in Norway. Second, to continue the discussion of Haugane and Grönlie's (1989) major case for Tertiary speleogenesis, the Hammarneset caves, under the geological perspectives of geomorphology, rate processes and time.

The definition of a cave and its existence in time

Before we decide what age or period of genesis we may attribute to a conduit or cave system, we would have to define what we actually mean by a "cave". This is critical for the timing of when the cavity came into existence as such. We may either define a cave hydrochemically, i.e. when the conduit can convey aggressive water along its total length, and thereby attain a maximum rate of growth. This is the distinction between a proto-conduit and a proper cave (White 1988; Ford and Williams 1989). Alternatively, we may use an anthropomorphic definition, i.e. a cave has to be passable to man (Curl 1964). Both definitions have some merit; the first out of hydrological reasons, because the activity of the aquifer can be related to proto-cave generation, and the anthropomorphic view allows us to inspect all proper caves. In the following, we shall only accept a cave as existing when its size is above that of the proto-stage of 1-10 cm diameter. If we wish to classify a cave as "Tertiary" it must have attained at least that size at that time. This definition is obvious and it would, for instance, be meaningless to extend the cave "system" into imaginary, no

longer existing levels of the limestone formation. These older parts of the aquifer are gone beyond study. Our concern here is the probability that some of the presently existing caves as such may be relics of Tertiary karstification.

Pre-glacial or pre-Pleistocene?

When discussing the timing of speleogenesis beyond the range of the dating methods that are presently available to us (i.e. U-series, paleomagnetism), we are confined to discuss the processes and the timing with respect to environmental and geomorphic parameters. In areas that have suffered glaciations in the past, it is useful to discuss whether speleogenesis occurred within a glacial regime (i.e. sub-glacial, interglacial) or within a pre-glacial regime. This concept involves climatic change and requires that the geomorphological processes were different for the two regimes and also that they produced different forms or sizes of cave. During the last 5-10 million years (Ma) the northern latitudes have experienced a gradual cooling which has terminated in a series of extreme glacial/interglacial oscillations. The first evidence of ice-rafted material in the Norwegian Sea occurred in the late Miocene (5.45 Ma). These glacial events were of a small magnitude compared to the large scale glacial cyclicity after 2.57 Ma which were further amplified after 1.2 Ma (Jansen et. al 1989). The existence of Pliocene tillites reported from Iceland (3.1 Ma) points in the same direction (Saemundson 1979). When comparing these dates with the formal Plio-Pleistocene boundary as defined in the Vrica type section in Italy at 1.6 Ma (Backman et al. 1983), we realize that the "Pre-Glacial" and "Glacial" climatic boundary is time-transgressive and still open for debate (see for instance the discussion of the 2.5 Ma alternative in Nilsson 1983).

With respect to our discussion, we must keep in mind that the Neogene/Pleistocene boundary implies some sort of formalism, and that this boundary does not necessarily involve a corresponding climatic change on the northern latitudes that would be relevant for the classification of caves.

GEOLOGIC AND GEOMORPHIC SETTING

A characteristic and spectacular feature of the relict, phreatic karst caves of Norway is that they are hanging relatively high in glacial topographies, separated from the catchments that once supplied water and created them. Many caves are also situated so remote from glacial valley floors that they may be included into the supposedly Tertiary ("paleic") landforms of Reusch (1900) and Gjessing (1967), (Figure 1).

Previous discussions of the possibility of a Tertiary age

A vast literature exists on bedrock geomorphology in Norway; see for instance Gjessing (1967) and references therein. A few of them deserve closer comment:

Pre-glacial or Tertiary peneplanation was identified by Reusch (1900), although peneplains were recognized 80 years earlier (Keilhau 1820). These ideas were further developed by later workers, like Ström (1948) who advocated two main stages of peneplanation in South Norway, a Precambrian and a Miocene peneplain. The former is a major unconformity, and the latter makes up large, mountainous plains ("vidder"), in altitude somewhere between the enveloping, alpine summit surface and the shoulders of glacial, U-shaped valleys. The association of the
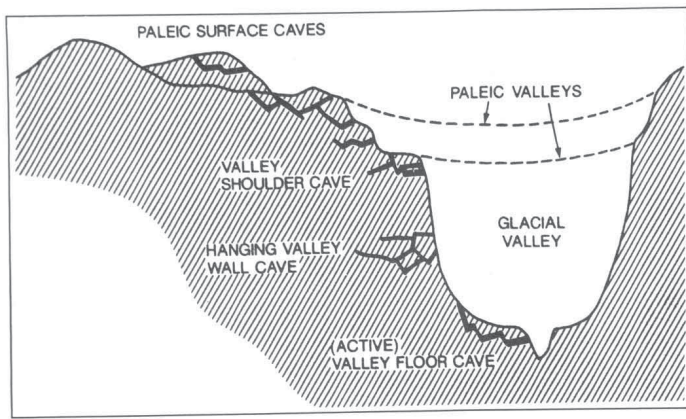
Figure 1. Phreatic caves classified according to their topographic position and the assumed type of landform they are associated with. Adapted from Lauritzen (1981a).

peneplains to the Miocene was based on analogy with central Europe where such peneplanation had been dated from deposits on them. Other workers, like Budel (1978, 1982) postulate the existence of inselbergs and widespread etchplains in northern Norway. Whether we may agree with these hypotheses or not, it is evident that the ideas of a pre-glacial geomorphology in Norway are quite old.

With respect to Quaternary and Tertiary caves, Horn (1947, see also the English translation by McCrady 1978) presented the original discussion of the speleogenesis of Hammarnesgrotta. Horn's discussion considered 5 possible cases of genetic modes: 1) Post-glacial. 2) Subglacial. 3) Interglacial. 4) Pre-glacial, i.e. Tertiary. 5) Polygenetic, including all of 1-4.

Horn (1947, p 58) stated explicitly that he considered the modes 2) and 3) to include any glacial and interglacial period. In order to distinguish between each of the 5 possibilities, Horn considered two crucial aspects; the size of the caves (i.e. size as a measure of age), and the availability of water (i.e. the rate of the speleogenetic processes). Water availability was discussed with respect to glacier hydrology and with respect to possible interglacial and Tertiary topographies. The crucial step in Horn's argumentation is that he considered the Norwegian caves to be *too small to account for a pre-glacial or Tertiary age*. This was based on comparison with the much larger and older caves that had been documented from southern, basically non-glaciated areas.

Moreover, Horn also estimated that the topographic position of phreatic caves, like Hammarnesgrotta, would correspond to 500 m or deeper below the surface topography of a Tertiary landscape. Horn also quite correctly recognized that karst (and other fractured aquifers) would become activated by interaction with englacial groundwater systems (see for instance recent studies of this phenomenon by Smart 1983). A flow of water through incipient fracture systems is in principle a sufficient requisite for

mass transport and thereby speleogenesis. This statement also holds true for very sluggish kinetics in near-saturated groundwaters, i.e. state 3 of White (1984). Horn's conclusion was that sub-glacial water flow was a sufficient requisite to explain both the size and the topographic position of the caves. By counting backwards from the present, he apparently judged the last glacial cycle as sufficient to account for the size of the caves. With respect to time, this is a conservative view, but with no chronology or rate data at hand, it is the only scientifically correct conclusion. When seen within its context of time, Horn's thinking stands out as a lucid and ingenious contribution to the science of speleology.

When discussing the speleogenesis of Pikhaaggrottene, Jenkins (1959) considered them as mainly developed in a glacial environment, but he considered a Tertiary *initiation* as possible. In her review of cave studies in Norway, St Pierre (1967) also referred to these views. Skundberg (1967) and Grönlie (1976) basically followed Horn's established discussion. Lauritzen (1981a) classified fossil, phreatic caves after their terrain position, whether they were hanging in valley walls, on valley shoulders or were situated in paleic landforms, (Figure 1). Uranium series dating of speleothems made it possible to *prove* that large phreatic conduits must be several glacial cycles old (Lauritzen and Gascoyne 1980, Lauritzen 1984a) and the concepts were extended to consider pre-glacial and Tertiary possibilities (Lauritzen 1981b, 1986a). The potential maximum constraints for Tertiary speleogenesis were discussed by Haugane and Grönlie (1989), and Lauritzen (1989a) suggested that a few, exceptionally large paleo-conduits could possibly be ascribed to Tertiary genetic regimes.

The geomorphology of Hammarnesgrotta

The cave was surveyed by Horn (1947) who suggested a sub-glacial genesis for the cave. The cave consists of a hanging, truncated system of essentially phreatic tubes that attain diameters of 3-4 m in places. The system is developed within a sloping lithologic boundary at the base of the aquifer. In vertical projection, the caves comprise an inclined looping system connected with a series of shafts that could either be phreatic chimneys and/or vadose invasion shafts, (Lauritzen 1981b) (Figure 2). This geometry is quite similar to other, larger, obliquely sloping phreatic cave systems (like the Holloch cave in Switzerland). The conduits are well scalloped in places, and the system has received input of substantial amounts of allogenic, apparently glacifluvial boulders and gravel and also hosted deposition of fine-grained laminates. This is a similar situation as seen in many other caves in glaciated regions, like Castleguard Cave of the Canadian Rockies (Schroeder and Ford 1983). Speleothem dating in Hammarnesgrotta has revealed bulk-corroded stalagmites older than 350 ka, but younger than 1.25 Ma (Lauritzen *in prep*).
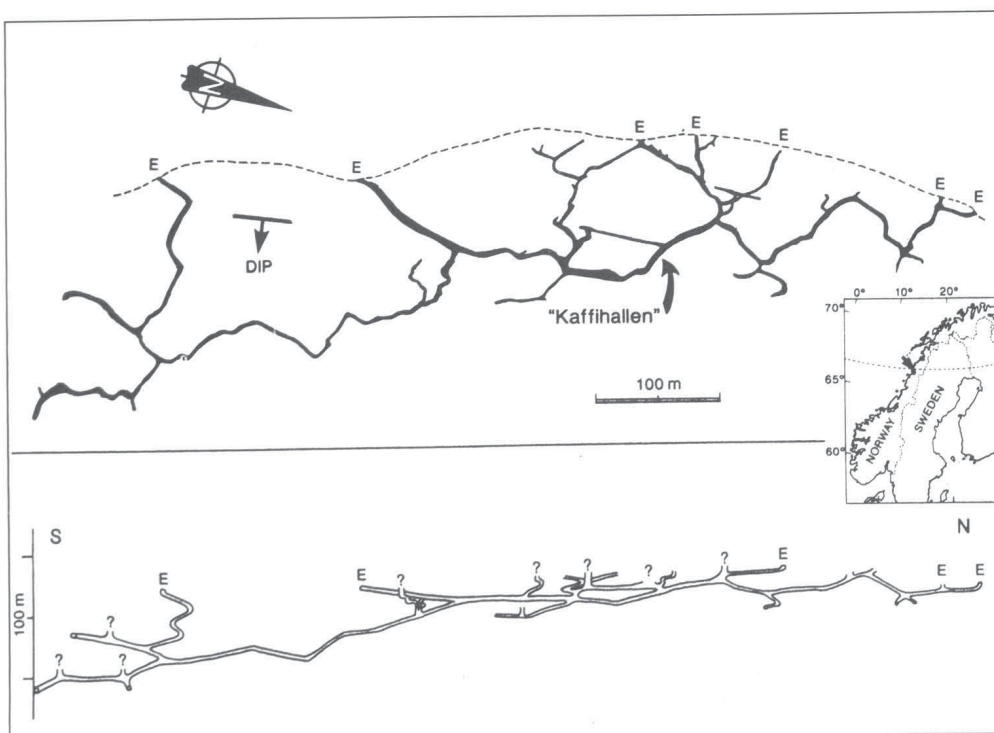

Figure 2. Plan and vertical projection of Hammarnesgrotta. The cave consists of a series of phreatic loops confined within a sloping plane. E: truncated entrances; vertical chimneys are marked with question marks. See text for further discussion. From Lauritzen (1981b).

Based on field observations and mapping during the last few years, three new points may be added to the cave morphology outlined in past writing (Lauritzen 1981b). First, scallop morphometry strongly suggests that some of the chimneys have acted as bypass loops above sections where the main conduit has been totally blocked or severely constricted by the glacially derived fill. Hence, considerable amounts of the conduit volumes appear to have developed as a response to glacial impact, (Figure 3). Second, paleocurrent estimates in Hammarnesgrotta (calculated as outlined by Lauritzen, 1982) demonstrate a significant, down-valley (i.e. strike-aligned) hydraulic gradient. Moreover, hydraulic continuity suggests a total scallop dominant discharge for the system of approximately 1 m³sec⁻¹. Using Darcy-Weissbach friction factors evaluated elsewhere (Lauritzen et al. 1985), the discharge and passage morphology convert to an hydraulic gradient of 4×10^{-3} to 4×10^{-4}. As the scallop dominant discharge corresponds to the upper (flood) flow regimes, an even lower hydraulic gradient would be sufficient to drive water through the system and enlarge it. Third, the scallop dominant discharge suggests that the known conduits had a probable drainage area of 5-7 km². This number is based on calibration against presently active conduits with known drainage areas and where the scallop assemblage is known (Lauritzen 1989b). The drainage area is approximately the same size as the area of limestone that is exposed above the cave today. Therefore, in order to explain the paleo-recharge of the cave, it is not necessary to extrapolate the aquifer very much in up-dip direction: it is sufficient to raise the water-table above the cave.

These points have important speleogenetic implications. First, the system comprises a set of interconnected phreatic loops, confined within a sloping plane. This geometry directly suggests that the cave *system* might have been generated by headward linking of phreatic loops in the conventional way (Ford and Williams 1989). "Headward" is in this case along the strike to the north. Each loop segment had a length of 100-200 m. The linking of conduits may then be sufficiently explained by water penetrating into the guiding plane, either from the valley side, or vertically, through fractures (Lauritzen 1981a). The pattern of loops is also compatible with the observed southward paleoflow hydraulic gradient, i.e. down-valley.

Second, the stratigraphic position of the phreatic loops is rather peculiar for a system that should have developed under deep phreatic conditions. The cave conduits occupy the *lowest possible* position within the aquifer. In spite of an overall high permeability of the whole thickness of the aquifer (i.e. the numerous vertical shafts), major water flows were apparently confined to the lower aquiclude contact. This suggests a *perched rather than a confined* situation, i.e. the local water table was dictated by the underlying aquiclude contact somewhere downstream of the cave. Hence, a shallow phreatic development, controlled either by a spillover point on mica schist or by repetitive englacial flooding, seem like the most probable explanation for the observed cave geomorphology.

Third, when we consider the last phase of corrosive water flow through the caves, (which was most probably superimposed by ice-contact, see Figure 3), we find that a low hydraulic gradient was sufficient for the paleoflow. The paleodischarge was also rather low, and the corresponding small watershed, could well have been upheld by simple, englacial or pro-glacial ice-damming of the cave system at a level somewhat above the present-day entrances of 220 m a.s.l. In other words, high discharges are not necessary to explain the speleogenesis. The transit of glaciofluvial material through the system may well have been episodic, although longer duration of high discharges in the past would have shortened the growth time for the conduits.

In conclusion, the morphology of the cave system suggests that it could well have been developed under a gentle, southward hydraulic gradient, fed by a rather small catchment. The phreatic zone was perched upon a base level that was controlled by the lower aquiclude contact somewhere downstream of the cave.

Bedrock base levels for Hammarnesgrotta

Downstream of Hammarnesgrotta, two independent base levels are possible, (Figure 4). The main valley (Glomdalen) follow the regional strike. A paleic valley system, which we should expect to be largely dictated by lithologic boundaries, would continue southward and gradually turn west. However, strong E-W lineaments (fracture zones) in Nordland comprise several shortcuts of the originally strike-directed drainage. These shortcuts are probably younger than the paleic strike-valleys; two examples are Ranafjorden and Langvatnet. Indeed, the present-
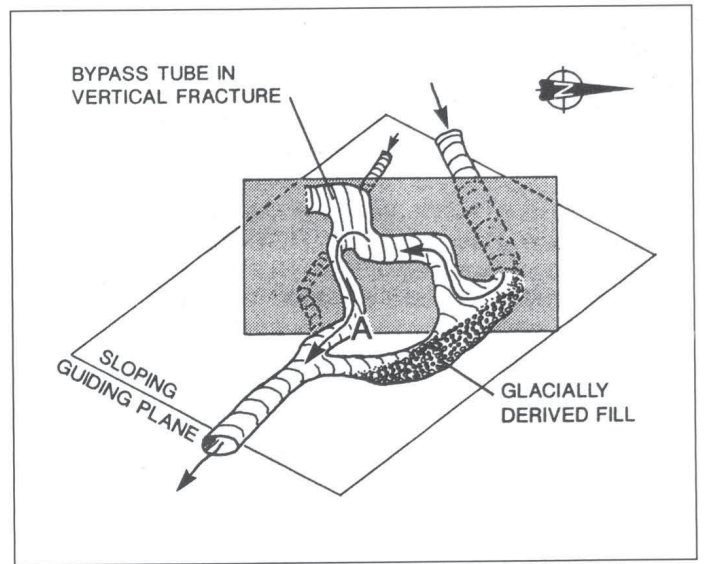

Figure 3. Principal sketch of the glacially derived bypass loop above "Kaffihallen" ("Coffee Hall") in Hammarnesgrotta, see Figure 2. Scallop morphometry demonstrates that the glacially derived clastic fills has obstructed the passage enough to establish the bypass loop through a vertical rift above the main conduit. The conduit (A) of about 3m diameter may therefore have gained most of its size in response to the bypass flow.

day drainage is diverted eastward by the Langvatnet lineament, a glacially molded trough, (Figure 4). The present-day valley base level on bedrock downstream of Hammarnesgrotta is Reinfossen at 50 m a.s.l. The secondary, possible base level control would have been through the col at Alteren, at approx 100 m a.s.l., (Figure 4).

MODELLING SIZE AS A MEASURE OF AGE

Dissolution kinetics

Cave conduits develop from pre-existing voids through solutional enlargements of the initially most efficient pathways. Conduit initiation, on a scale of 1 mm to 10 cm diameter, is a relatively slow process, depending on water chemistry and the length of the initial fracture system (Dreybrodt 1988). This timespan is generally considered to last 10^4 to 10^5 years. After kinetic breakthrough, conduit enlargement is proportional to discharge, attaining an asymptotic, maximum rate of wall retreat. The maximum rate is generally regarded close to 1 mm a⁻¹ (Palmer 1981; Dreybrodt 1988). This raises the prospect, already addressed by Horn (1947), that size could become a quantitative measure of age.

The model: growth rates in reversal

Principally, there are two ways of raising the watertable to recreate the phreatic conditions under which the caves must have formed, either by filling up the valley with ice or with rock. These conditions occurred at different times and were also of different duration (Lauritzen 1982). Therefore, the potential age of the cavity is dependent of its size and elevation above base level. The magnitude of these measures is the integral with respect to time of a set of three principal rate processes. Firstly, the valley entrenchment rate through glacial cycles; secondly, the passage growth rate (wall retreat rate) under sub-glacial conditions and thirdly, wall retreat rate under interglacial conditions. When considering speleogenesis *sensu stricto*, i.e. the entire formation of a conduit from pre-existing fractures, we would also have to consider the penetration rate for first-order kinetics in joints, *the breakthrough time* of Dreybrodt (1988). These factors are summarized in Figure 5, where the potential age of a cavity is dependent of a critical, present-day relative elevation, H_t and of a critical, present day diameter D_t. Hence,

$$H_t = \int_{t_0}^{t_1} (VR_s)\, dt \qquad (1)$$

and

$$D_t = \int_{t_0}^{t_1} (WR_s)\,dt + \int_{t_1}^{t_2} (WR_s)\,dt + \int_{t_1}^{t_2} (WR_s)\,dt + \int_{t_2}^{t_3} (L^*BR)\,dt \qquad (2)$$
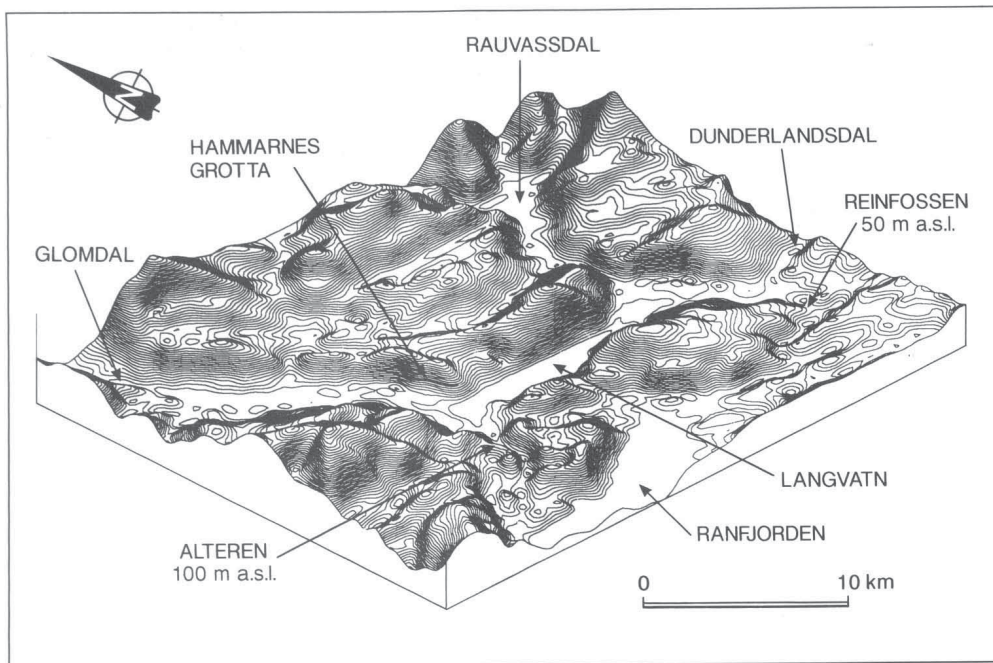
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Figure 4. Oblique view of Glomdalen, Langvatnet and the topographic situation of Hammarnesgrotta. The Paleo-Glomdal valley went along the strike, across Alteren into Ranfjorden. The Langvatn trough is probably a later, glacial feature, controlled by EW fractures. The Langvatn trough has captured the drainage eastward to Rauvassdalen into the main drainage from Dunderlandsdalen into Ranfjorden. The present-day bedrock base level of erosion for Hammarnesgrotta is on Reinfossen (50m a.s.l.), and a paleo-base level would be the col at Alteren (100 m a.s.l.).
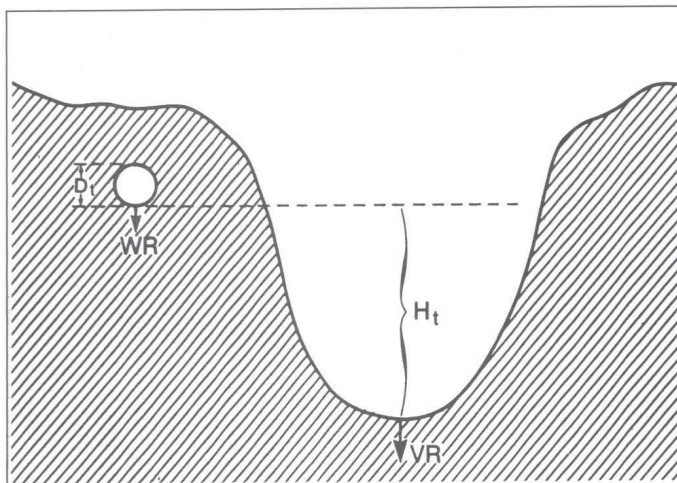

Figure 5. The concept of the present-day, critical relative elevation (H_t) of a relict, phreatic cave conduit, and its critical diameter (D_t). The probable age of the conduit is a function of these two measures, and of the rates of the processes that increase them. Those processes are the rates of valley entrenchment (VR) and the rates of wall retreat (WR) under the different environmental conditions that have prevailed during the lifetime of the cave. Running these processes backwards, we may imagine the cave to close and the valley floor to raise beyond the cave level.

where VR_e is the average valley entrenchment rate, WR_g is the wall retreat rate under subglacial conditions, WR_i is the wall retreat rate under interglacial conditions, when the cave is an integrated part of a fluvial system, and BR is the breakthrough rate for first order kinetics in fractures that are initially 10^{-2} cm width and of length L. The time t_0-t_1 is the time elapsed since the cave was raised above the phreatic zone, t_1-t_2 is the accumulated growth time for the conduit to attain the size it had when it was lifted above the phreatic zone, t_2-t_3 is the time needed for first order kinetics to penetrate the full length of the guiding fracture system of length L.

By starting with D_t and H_t, the model can be run backwards in time to determine the total, cumulative time needed to (a) submerge the cave underneath a rising valley floor, and (b) to close the cavity by subtracting the size it gained subglacially as a relict, above the watertable, and by the size it gained during its phreatic phase, developing from a proto-conduit some 10 cm diameter. Finally, the time needed to integrate pre-existing guiding fractures into the system of proto-conduits may then be added. The total sum of all these times would then represent the time required for speleogenesis *sensu stricto*. This model has been implemented on a microcomputer and will be discussed elsewhere (Lauritzen *in prep*). A linear version of the model is used here as a sensitivity analysis to determine which of the measures listed above are the most crucial for determining the maximum age of the cave conduit in question. Obviously, the input parameters in such a model are very important.

Valley entrenchment rates (R_e)

Maximum valley entrenchment rates may be estimated from speleothem ages and their present-day elevation above the base level of erosion. This has been performed by Ford *et. al* (1981), Gascoyen *et al.* (1982), Lauritzen and Gascoyne (1980) and Lauritzen *(in prep)*. Maximum rates for Norwegian glacial valleys attain about 0.35-0.55 m/ka. British and Canadian rates are within the same order of magnitude (0.13-0.5 m/ka).

Average glacial denudation rates may be inferred from the integrated volume of difference between the present-day topography and an estimate of the fraction of time occupied by glaciations during the Pleistocene. This has been done for a part of Nordland, Norway, yielding an estimate of 0.44-0.45 m/ka (Nesje *et al.* 1989).

Minimum glacial denudation rates may be assessed through integrating the accumulated sediment yield as measured by the thickness of the Pleistocene sediments on the off-shore continental shelf (Rokoengen *et al.* 1988). From a paleolandscape model, the catchment area of such deposits may be assessed, yielding minimum rates, as the trap efficiency is unknown but significantly less than 100%. Minimum rates of erosion during the Pleistocene has been estimated to 0.15-0.18 m/ka (Nesje *et. al* 1989).

These different erosion rates correspond well, and it is reasonable to assess the average valley entrenchment rates (for full glacial/interglacial cycles) to be in the range 0.15-0.55 m/ka.

Subglacial wall retreat rates of conduits (WR_g)

Evidence of significant subglacial wall retreat is presented elsewhere (Lauritzen 1984b, 1986). Moreover, the average rates may be assessed from bulk re-solution of interglacial speleothems. In two cases, the corroded speleothem was part of a scalloped wall texture. Speleothem dating and estimates of the total wall retreat since the deposition of the speleothem yield rates in the order of 5-10 cm/100 ka, i.e. 1 mm/ka or less (Lauritzen *in prep*). A reasonable model input would then be 0.5 to 1.0 mm/ka for the rate of subglacial conduit modification. This is probably a gross underestimate, when we compare it to the reported weathering rate of quartz which is some 4 mm/ka (Ford and Williams 1989).

Interglacial rates of wall retreat (WR_i)

The possibility of using size as a measure of age in a quantitative way was first invoked by Renwick (1958) at a Norwegian site. This idea has been further developed on the same site (Lauritzen *et. al.* 1985; Lauritzen 1986b, 1989b) yielding present-day rates of wall retreat of phreatic conduits in marble in the range of 0.2 to 0.6 mm/a, or 200 to 600 mm/ka.

Breakthrough time for first-order kinetics (proto-cave generation), BR_t were discussed by Dreybrodt (1988) as the time required for widening of joints from an aperture of 0.1 mm to 10 cm, i.e. the generation of a proto-cave system from joints. The critical process is that the dissolution rates decrease with time and
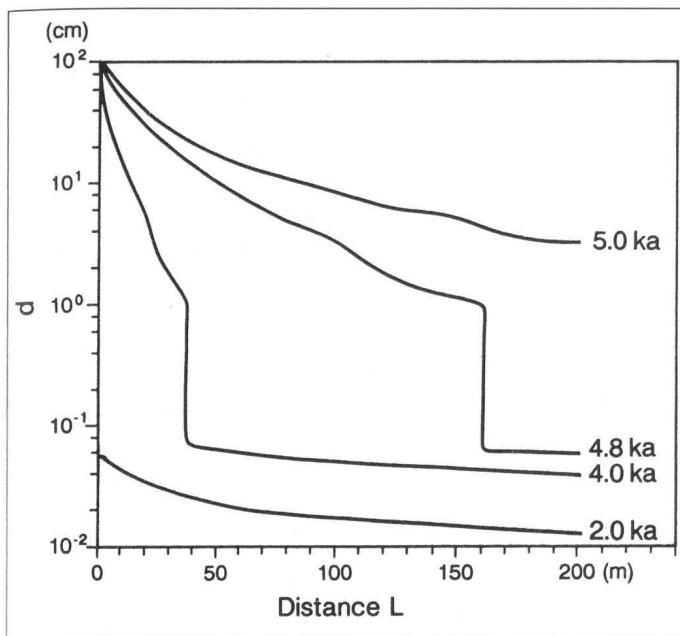
Figure 6. The breakthrough time for first-order kinetics in a 200m long fracture that had an initial aperture of 10^{-2}mm. First-order kinetics (i.e. maximum wall retreat rates) will break through after about 4800 years. This process corresponds to the transition from the photo-conduit stage into a proper cave conduit stage where water flow may be turbulent and where the water emerging from the system is still undersaturated with respect to calcite. Adapted from Dreybrodt (1988).

porous varnish, consisting of quartz, mica and iron oxides. In abandoned galleries, the residue of this weathering comes loose and may be the origin of the light, powdery "eolian" deposits that cover many horizontal surfaces. Observations in Hammarnesgrotta indicate that the *in situ* fraction of these crusts are about 2 mm thick. Moreover, in such galleries, ancient scallops have usually lost their sharp crests, suggesting that the weathering might have been substantial. As most caves bear evidence of severe flooding during deglaciations, we may infer that the crusts might have formed during the Holocene, i.e. the wall retreat rate from condensation and winnowing is at least 0.2 mm/ka.

Provided that the passages of Hammarnesgrotta were only subject to weak draughts and condensation as seen today, then the largest passages (1.5 m radius) would have gained 80% of their size (i.e. since they were 50 cm diameter) in some 6.25 Ma. This is of course a rather unrealistic example, but it serves the purpose of illustrating that if "most of the development" of Hammarnesgrotta took place in the Miocene, the average rate of wall retreat during the Pliocene and Pleistocene epochs would have to be considerably less than that caused by condensation water alone.
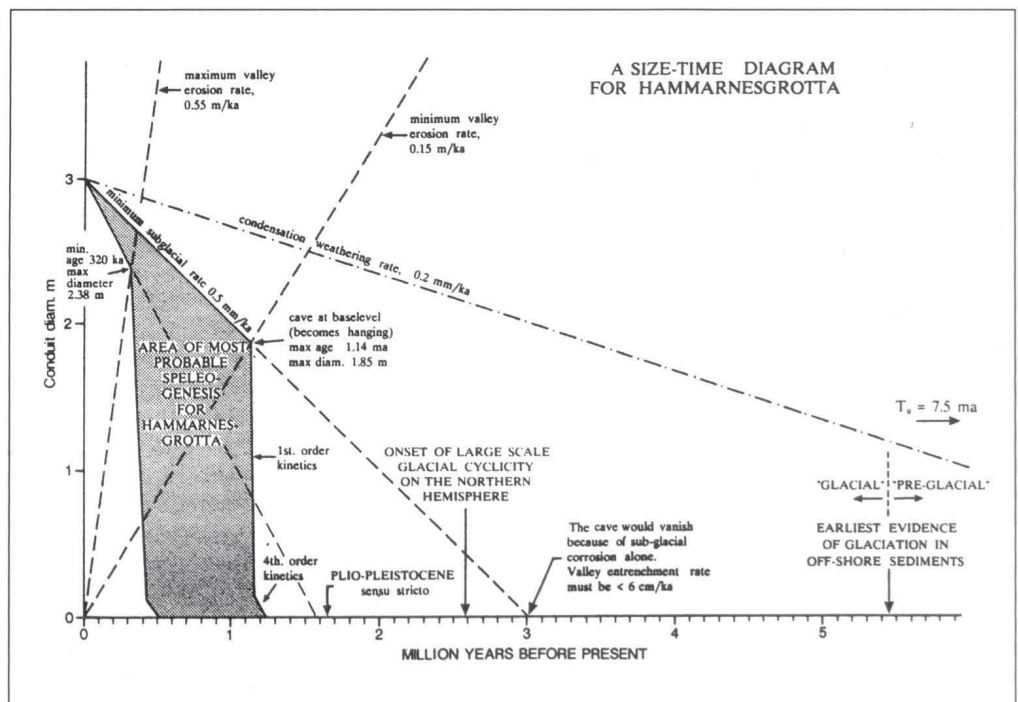
hence with penetration into a joint. Rapid, first-order kinetics is soon replaced with more sluggish higher-order kinetics. At a given aperture and hydraulic gradient, first-order kinetics possess an Effective Penetration Distance (EPD). If EPD is less than the length of the joint, speleogenesis is very slow, but as soon as EPD is longer than the extent of the joint segment in question, the wall retreat rate may attain a maximum value of up to 1 mm/a as mentioned above (Palmer 1981).

Figure 6 shows the modelling of first-order kinetics in a 200 m long joint that had an original aperture of 0.1 mm (Dreybrodt 1988). Breakthrough took place after some 4800 years. As mentioned, the length of individual looping segments of Hammarnesgrotta are approximately 200 m each, but even if we consider the total extent of the cave (2 km) as one loop, breakthrough time would only be about 50 ka. As a conservative estimate, the radial growth rate of the conduit below 100 mm diameter was set to 1 mm/ka.

Rate of weathering by condensation water

Sub-aerial weathering of the wall is seen in most relict caves. On Norwegian marbles, this is usually expressed as a rusty,

RESULTS AND DISCUSSION

The numerical model (eqn 1 and 2) was run for a cave diameter of 3 m (D_t), at an elevation of 172 m above base level. These numbers correspond to the largest diameter reported from phreatic tubes in Hammarnesgrotta, and to the level of the entrances, i.e. all the conduits that make up the known cave are situated below this level. The input rates taken were the maxima and minima as discussed above. The results are summarized in the size-time diagram of Figure 7.

Subglacial rates dominate until the base level reaches the level of the cave. This would happen between 320 ka and 1.1 Ma ago. The diameter of the cave would then have been between 0.8 and 2.4 m. Before this state, the cave is waterfilled and phreatic when the growth rate is dominated by first order kinetics under interglacial P_{CO2} conditions. This means that the cave grew to its size in much less than 100 ka. Proto-cave speleogenesis took place during the preceding 50 ka. By using the extreme estimates of valley entrenchment rates, the model is unable to make the cave older than 1.1 million years. If the base level lowering was little enough (< 6 cm/ka) to make the average subglacial growth rate dominate the whole life history of the cave, the conduit would still vanish if we go back 3 Ma.

The potential age of a relict cave conduit is much more sensitive to its terrain position and the valley entrenchment rate than to the growth rate of the passage itself. Our experiments clearly demonstrate that, as soon as a cave becomes an active part of a phreatic system, the growth is almost instantaneous when seen from the Cenozoic timescale of millions of years.

Figure 7. Size-time diagram for Hammarnesgrotta. The most probable time interval for speleogenesis is between 320 ka and 1.2 Ma. The valley entrenchment rates are not plotted to scale, but with respect to the passage diameter at t_1 (eqn. 1). See text for further discussion.
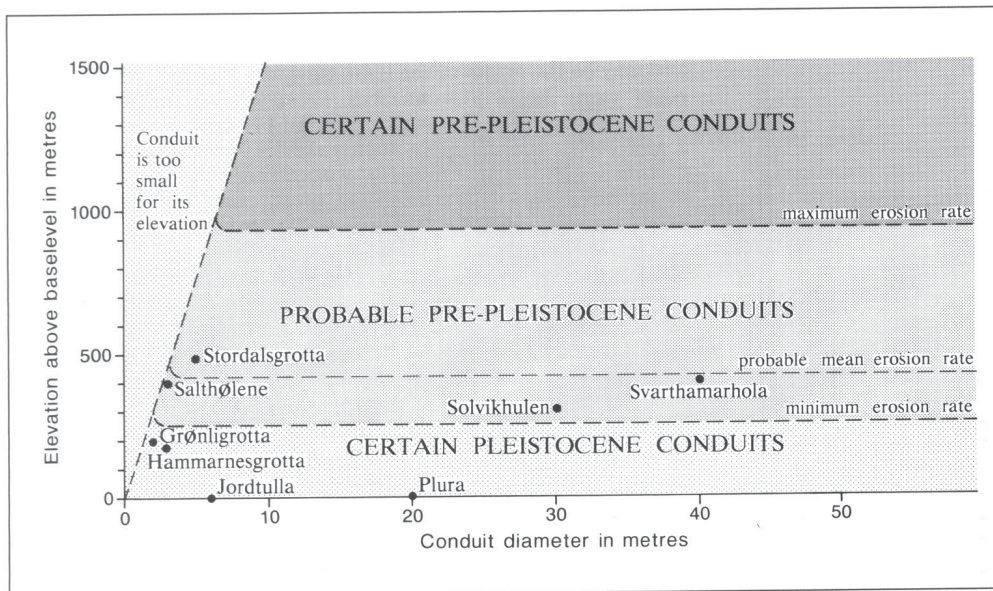
Figure 8. Critical altitude and size for some Norwegian caves. The diagram shows the critical relationship between elevation above base level and conduit diameter that would dictate the probability of survival through the Pleistocene glaciations. The criterion is set at the formal Plio-Pleistocene boundary at 1.6 Ma. Hammarnesgrotta and Grönligrotta do not satisfy these criteria. Svarthammarhola and Solvikhulen are exceptionally large conduits situated closely below the paleic surface and therefore strong candidates for a Tertiary age (Lauritzen 1989a). Note that quite large conduits (Jordtulla and Plura) exist at or below the base-level, demonstrating that active conduits may grow to great sizes while they still are hydrologically active.

The most probable interval of speleogenesis for Hammarnesgrotta as we see it today, is within the interval of 320 ka to 1.1 Ma. The previous dates of speleothems that were in the range of [350 ka $<$ x $<$ 1.25 Ma] are fully compatible with the result of the modelling.

Other caves: candidates for the Tertiary case

As Hammarnesgrotta appears to be too small and situated too low in the glacial topography to account for a pre-Pleistocene age, other caves may still satisfy these criteria. If we set the Plio-Pleistocene boundary *sensu stricto* at 1.6 Ma as a formal threshold, a potential candidate must be larger than about 3 m diameter and possess an elevation greater than 240 m above base level. Few caves in Norway satisfy these criteria, even fewer do so if we set the threshold at the environmentally more relevant transition, the onset of large scale glacial cyclicity at 2.57 Ma (Jansen *et al* 1989).

In Figure 8, some other well-documented Norwegian caves are plotted in a diameter-altitude diagram which illustrate their probability of having survived the Plio-Pleistocene boundary at 1.6 Ma. Only 4 of the caves shown do plot outside the boundary, Stordalsgrotta (5 m diameter, 480 m relief, Lauritzen unpublished mapping 1988), Salthölene (3m diameter, 400m relief, unpublished mapping 1985), Svarthammarhola (40 m diameter, 400 m relief, Heap 1970) and Solvikhulen (30 m diameter, 300 m relief). The size of the latter two deviate strongly from all other known caves in the country and make them into obvious candidates for a preglacial age (Lauritzen 1989a).

CONCLUSIONS

The possibility that most of the development of Hammarnesgrotta took place in the Miocene, as stated by Haugane and Grönlie (1989) has been tested by numerical modelling of valley entrenchment rates and of wall retreat rates for cave conduits under sub-glacial, interglacial (phreatic) and proto-cave generation. The modelling has demonstrated that the valley entrenchment rate is the limiting factor for the potential age of the caves; wall retreat rates of active conduits are always large enough to create large caves almost instantaneously in the timescale of millions of years. Even the most conservative rate estimates fail to make the cave older than about 1.1 Ma. Most probably, the speleogenesis *sensu stricto* of Hammarnesgrotta commenced between 320 ka and 1.2 Ma. This timespan is fully compatible with the geomorphology of the cave.

We may therefore conclude that, under the present knowledge of rate processes, the speleogenesis of Hammarnesgrotta can hardly be ascribed to pre-glacial conditions. The postulated Miocene genesis of the cave seem out of range by more than an order of magnitude.

Salthölene and Stordalsgrotta both display passages of a sufficient size and possess a sufficient elevation above base level to account for a pre-Pleistocene age. It is, however, important to keep in mind that this only implies that the caves might have been established beyond the proto-cave stage at that time. Much of the passage widening must necessarily have taken place during the Pleistocene. If we shall have any chance of finding real *Tertiary*

relics, we would have to go for very large conduits. The most spectacular examples of that kind, Svarthammarhola and Solvikhulen, have attained sizes of more than an order of magnitude larger than Hammarnesgrotta. As outlined elsewhere (Lauritzen 1989a), this distinguishes them from all other known caves in the country, and their topographic positions give them the highest probability of being Tertiary relics.

The present analysis has pointed to certain critical characteristics of the size and elevation of a cave passage with respect to its age. A crucial test for the hypotheses outlined in this paper, would be to find and identify *in situ* Tertiary deposits in the caves. Future studies should aim at these problems.

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The Caves of the Cloud Mountains, Irian Jaya: the 1988 Expedition

Dick WILLIS, Editor

Abstract: During the months of September and October 1988, a group of 10 British speleologists visited the Baliem valley in the highlands of Irian Jaya. This venture followed two earlier reconnaissance trips which had established the general speleological potential of this area. The primarly aims for 1988 were to reconnoitre the slopes of the Gunung Trikora range at 3000 m. and above and to identify the logistical problems associated with cave exploration at this altitude. Various areas were visited, a number of caves explored and surveyed and the groundwork carried out for a future expedition.

The island of New Guinea is divided into two political units, independent Papua New Guinea to the east and Irian Jaya, part of Indonesia, to the west. In 1975 a major British speleological expedition visited Papua New Guinea (Brook *et al*, 1976) and, in doing so, realised the likely speleological significance of the areas westward, across the border in Irian Jaya. As a result, small reconnaissance trips were eventually undertaken to the Wamena area in 1984 and 1985. (See Fig. 1).

As a result of the 1985 reconnaissance, a team of 9 cavers visited the Wamena region in September and October 1988, intending to continue the work which had been carried out in 1985 and to reconnoitre other nearby areas.

The island of New Guinea has a central chain of east-west mountains composed largely of limestone interbedded with sandstones and shales. Peaks in the Wamena area rise to over 4700 m. and the Baliem river, flowing in the "Grand Valley" cuts down through them in a spectacular gorge and flows south to the Arafura Sea. The Grand Valley is at 1600 m. and surrounded by limestone mountains. For a distance of 100 km. all drainage is into either the west or east Baliem Rivers which meet and flow into the northern end of the Grand Valley. The landscape of the area is discussed in greater detail in Tony White's account of the 1985 reconnaissance (White, 1986).

The 1988 team divided into small groups and visited the ranges south of Habbema, south and south west of Kwiawogwi, the cone karst east of Wamena, the Wolo/Ilugwa valley, Usilimo and the area around Korupun (east of the Baliem Gorge). Each of these areas is discussed in turn, and descriptions are given for the caves which were explored. These are followed by accounts of the logistical problems facing the expedition and, finally, a discussion of the potential of the area for future exploration. (See Figs 2 & 3).

HABBEMA

The area around Lake Habbema was one of the most interesting of those identified on the maps during the planning of the expedition. It seemed to offer a way into the Trikora mountain range which lies to the south of Wamena and runs north west - south east, rising at the highest to the peak of Gunong Trikora itself at 4730 m. Lake Habbema lies beneath G. Trikora at the eastern end of a wide bench of land at a height of about 3000 m.

It had been intended to spend several weeks in this region, up on the high limestone, looking for shafts and caves. Unfortunately the local police would not permit this because of "armed rebels" (but more probably because of their embarrassment at a serious
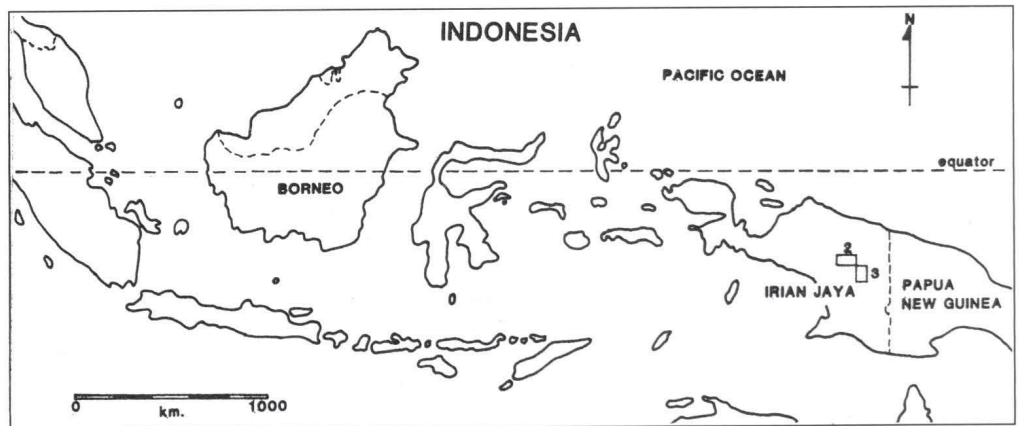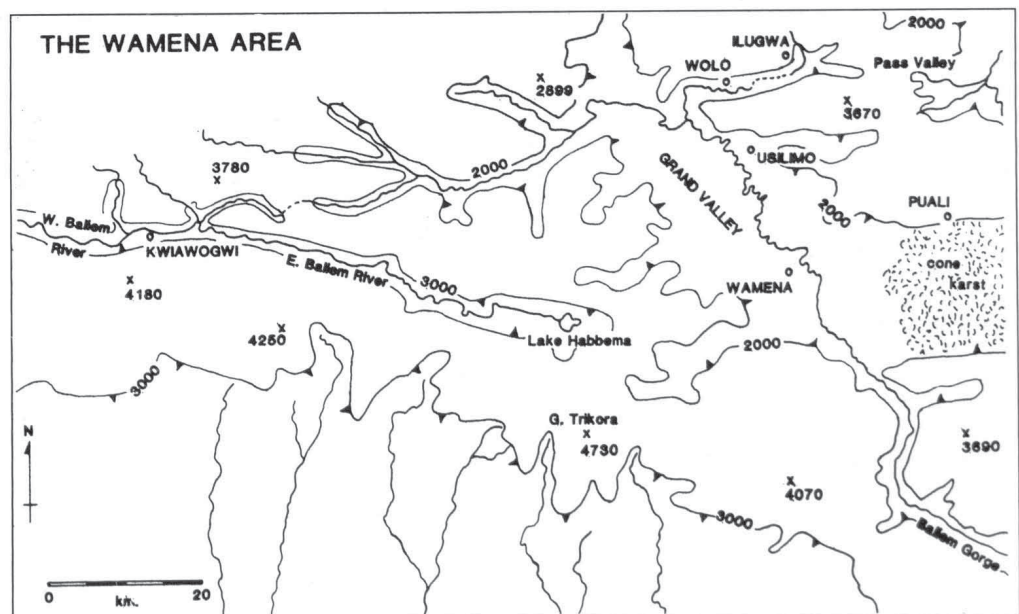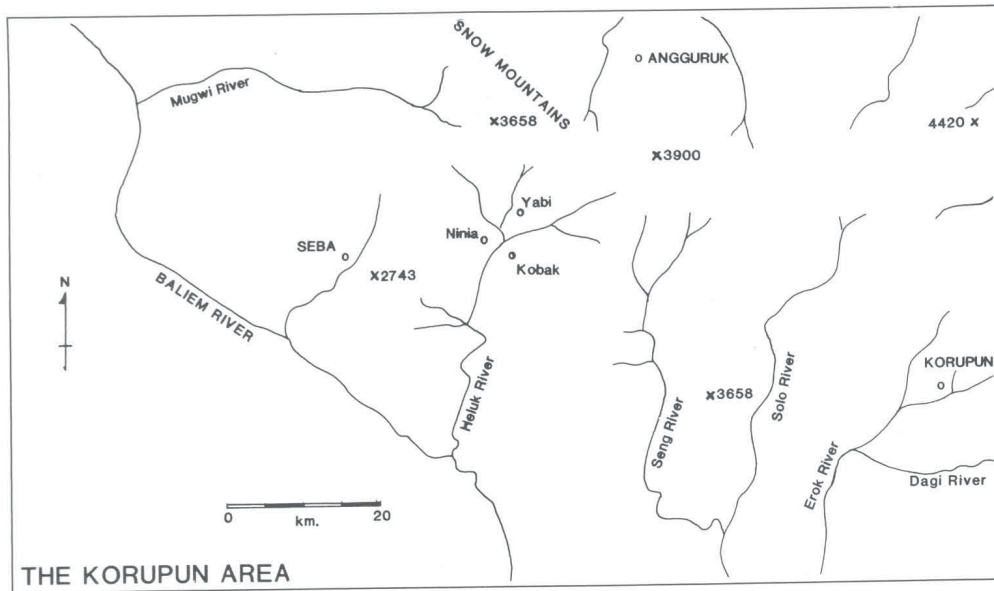
Figure 1.

Figure 2.

THE KORUPUN AREA

Figure 3.

LAKE SINK

HABBEMA, IRIAN JAYA
Cloud Mountains Expedition 1988

BCRA 5b

N

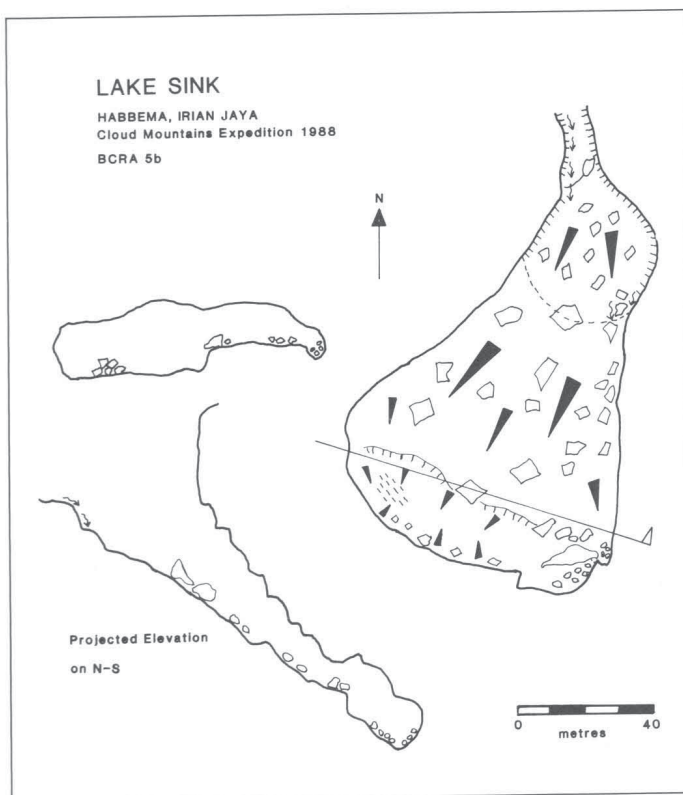Projected Elevation
on N-S

0   metres   40

Figure 4.

accident suffered on the mountain by an Austrian para-glider). After two days of persistent negotiation a compromise was reached, the visit was allowed to go ahead providing that we promised not to climb Trikora itself, that we undertook to take and feed 5 armed policemen and, worst of all, that we agreed to spend only 2 days at Habbema. The visit was almost abandoned, the walks in and out took 2 days each, but in view of the potential for deep caves the restrictions were finally accepted.

The walk-in to Habbema is spectacular and follows a winding river through cultivated valleys before starting to climb steeply and continuously through forest. After several hours a flat grassland plain is reached, with Lake Habbema in sight. On the south side of the lake is a small rise beyond which lies a huge peat bog behind which the mountains rise steeply. The mountain slopes are generally covered with long grass with patches of tree ferns or dense, almost inpenetrable, moss-forest - neither is an easy terrain to cross; above this bare rock is reached.

The limit of only two days restricted the team to visiting merely the main features which had been identified by an aerial reconnaissance. One of these was a sink which takes water flowing from another smaller lake south-west of Habbema. Despite a large entrance, Lake Sink closed down into a boulder collapse after 60 m. (See Fig 4). A further sink beyond this was not visited and may provide a way into the same system.

Three other sinks were visited but none could be entered. One, a large doline at the edge of the peat bog, had a 5 m. shaft, blocked at the bottom; another took a 1-2 cumec. river with a likely resurgence several kilometres away; the third had a 600 l/sec. river which sank into the ground up-valley from a large old collapsed cave, but had no way on.

KWIAWOGWI

Working from a base in the village of Kwiawogwi (which has the highest airstrip in Irian Jaya) two visits were made to the ranges south of the Baliem and one visit to the ranges south of the East Baliem. Kwiawogwi lies a few kilometres west of the junction of the East and West Baliem Rivers.

The first of these followed an attempt to walk west along the river valley before cutting up onto the mountains to examine an area marked on the map as having "numerous sinkholes". This excursion provided some team members with an introduction to Dani negotiating tactics and the intention of cutting up the mountain side was abandoned after numerous protests from the labour force: these included a lack of tracks, cold, danger and enemies. The group returned along the south bank of the river looking for resurgences but, with one tiny exception, none were found.

A more successful attempt to reach the heights was then made via a locally well known path running up through moss forest from a point south-east of Kwiawogwi. A camp was established under a rock overhang in a sinkhole on a bench at about 3300 m. An adjacent doline field was searched and revealed a choked 100 m. shaft. Although there is ample evidence of huge volumes of water going underground, the area is covered in extensive glacial deposits which are, in turn, obscured by thick layers of peat. Access into any cave systems is thus very difficult.

Local guides told of a major cave two hours walk to the west, near the summit of G.Minic. This was visited and found to be a tiny remnant phreatic passage. The walk was arduous but through beautiful country. Despite the assertion that it would only take two hours, and a series of "only 20 minutes more" promises, the walk in fact took about eight hours and necessitated an overnight bivouac in a hunting hut.

The second area visited from the Kwiawogwi base was the mountain area to the south-east of the village, close to Lake Kulip. This area was approached by walking along the main valley for about 10 km. to a small village. At this point the main path deteriorated into a track running through dense forest and clearings. A smaller track southwards up a thinly vegetated ridge to a grassy basin at about 3500 m.

The area around L.Kulip was examined and so too was the grassy basin. Immediately east and above Kulip was an extensive doline field where a 30 m. blind shaft was descended. The small valley in which the lake was situated was followed southwards into impressive pinnacle karst and moss forest. Walking was difficult although the moss forest could often be avoided by keeping to the grassland or ridge tops.

One Yorkshire-style sink cave was found but exploration was quickly abandoned at the head of a very wet pitch which required more equipment than the reconnaissance group was carrying.

PUGIMA AND PUALI

Due east of Wamena is an area of cone karst, cleary visible on the maps. This is approached by crossing the Baliem on the impressive suspension bridge and walking to Pugima. Past this village and over a ridge is a resurgence, Gua Guam, which is flowing at over 2 cumecs. which was explored for 160 m. in deep water to a sump (see Fig. 5). The position of the cave suggests that it is one of the main resurgences for the cone karst area. The river is crossed either on a single pole bridge or by jumping boulders in the entrance.

In view of the anticipated difficulties of traversing the cone karst, time was spent following a series of clear paths along the northern sandstone/limestone margin. At the limit of travel a line of high limestone cliffs were seen in the distance NNE from the vantage point. Previous to this, a lack of clear topographical features had created difficulties in locating the precise position on the map; this situation was not assisted by inaccuracies in the maps which show some villages up to 50 km. out of position! Limited time made a visit to the distant massif impossible but the area would clearly merit closer examination.

During the return walk, a major trading path was encountered which crossed the cone karst directly, back to Pugima. This provided access for the exploration of shafts of Lubang Putella and Lubang Wambigmo, both of which were found to choke. In the cone karst area, one hour's walk beyond Lubang Putella going out from Wamena, a sandstone cap of 0.6 m. had been pierced at eroded joints allowing the development of two adjacent unnamed shafts of 24 m. and 30 m. respectively; both of these quickly became too tight. A further shaft of 20 m. diameter was observed but not descended, since the valley floor was less than 100 m. below

WOLO AND ILUGWA

The 1985 reconnaissance visited this area and examined some of the most obvious features such as the massive blind valley and arch of Yogoluk. At present a road is being extended northwards

GUA GUAM
ANELAGA, IRIAN JAYA
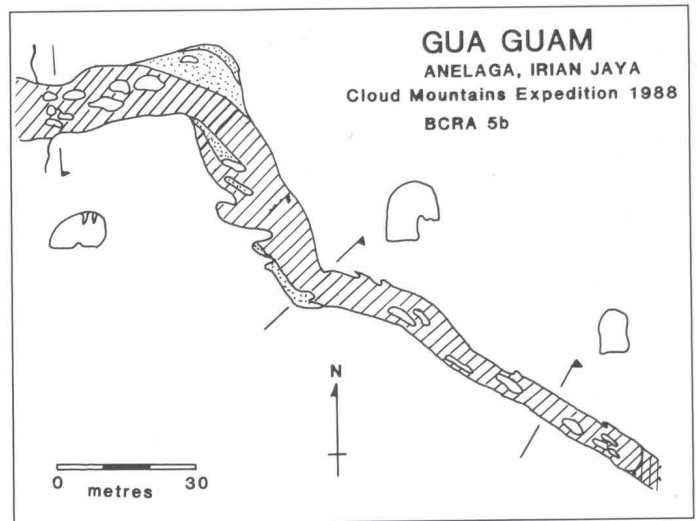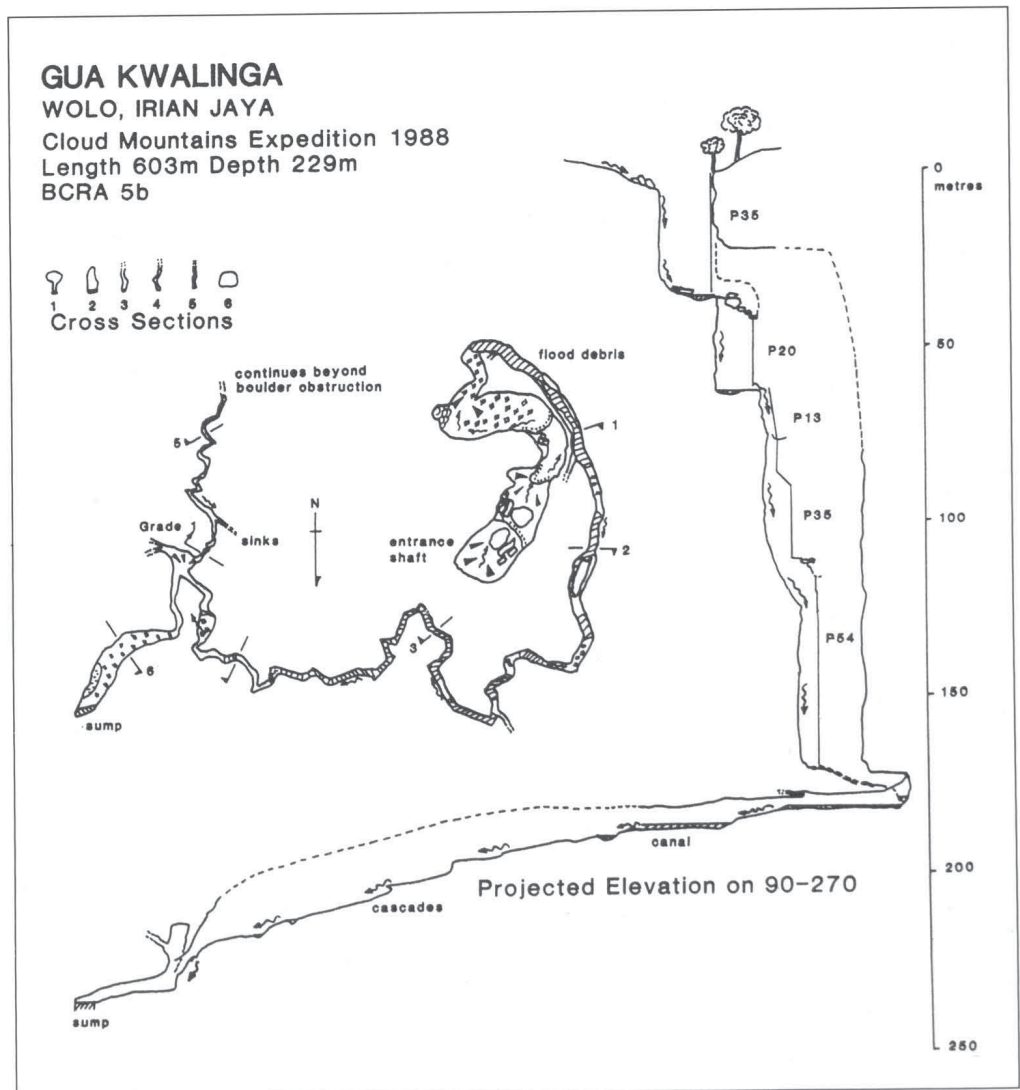Cloud Mountains Expedition 1988
BCRA 5b

Figure 5.

from Wamena and it is suggested that this will eventually go through to Wolo and Ilugwa, although this seems far from certain. The 1988 team took a minibus ('bemo') from Wamena to the roadhead at Wagawaga and walked for 6 hours, past the village of Usilimo, to Wolo. The track runs through beautiful country along a wooded river valley bordered by steep limestone cliffs. Wolo itself is superbly sited, with a well organised layout resulting, presumably, from the rebuilding necessitated by military bombing in the 1970s. The village has a small hydroelectric plant supplying electricity and an airstrip which, pointing steeply downhill and directly at a mountain wall, causes acute anxiety in spectators, let alone passengers or pilots.

Approximately one hour beyond Wolo, towards Ilugwa, there are three large sinkholes south of the path. The first of these, the impressive Gua Kwalinga, carries a sizeable stream and was descended to a sump at —229 m. (see Fig 6). One local man stated
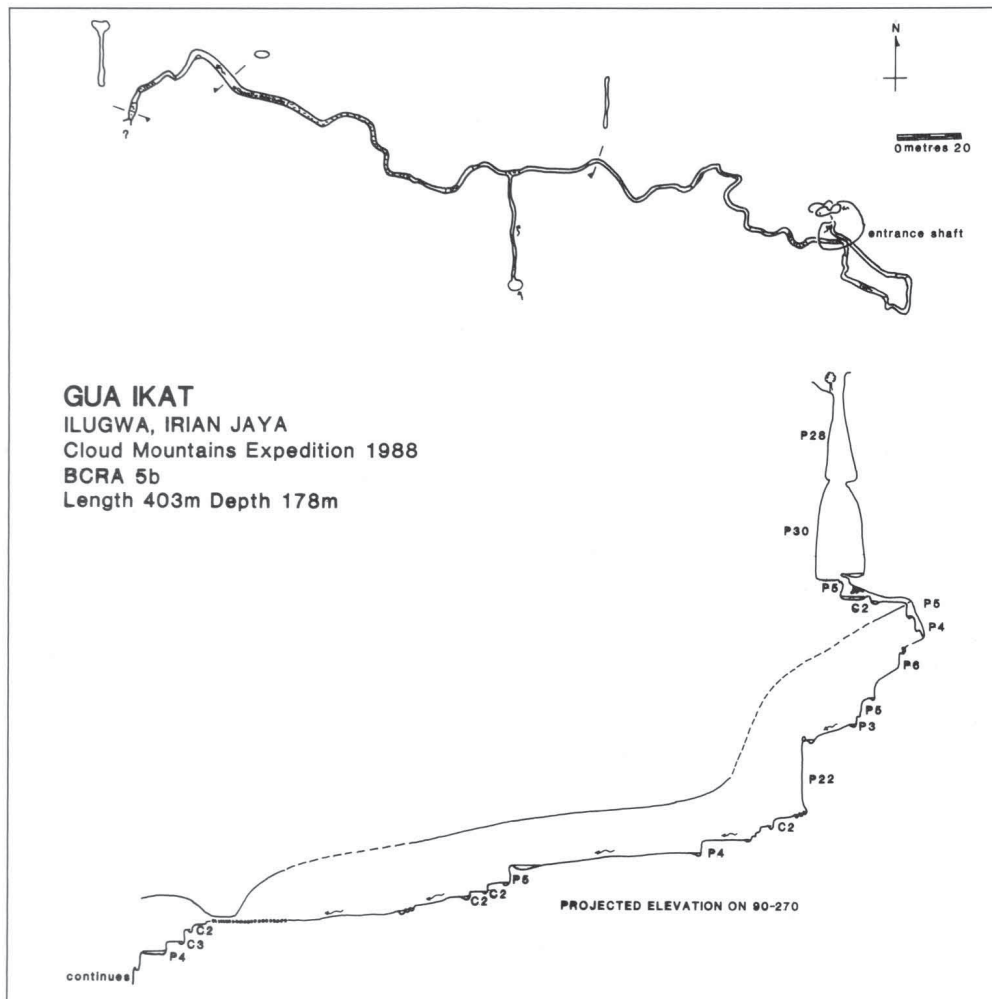
Figure 6.

GUA KWALINGA
WOLO, IRIAN JAYA
Cloud Mountains Expedition 1988
Length 603m Depth 229m
BCRA 5b

Cross Sections

Projected Elevation on 90-270

GUA IKAT
ILUGWA, IRIAN JAYA
Cloud Mountains Expedition 1988
BCRA 5b
Length 403m Depth 178m

PROJECTED ELEVATION ON 90-270

Figure 7.

GUA LUGDAK
ILUGWA, IRIAN JAYA
Cloud Mountains Expedition 1988
BCRA 5b

Projected Elevation
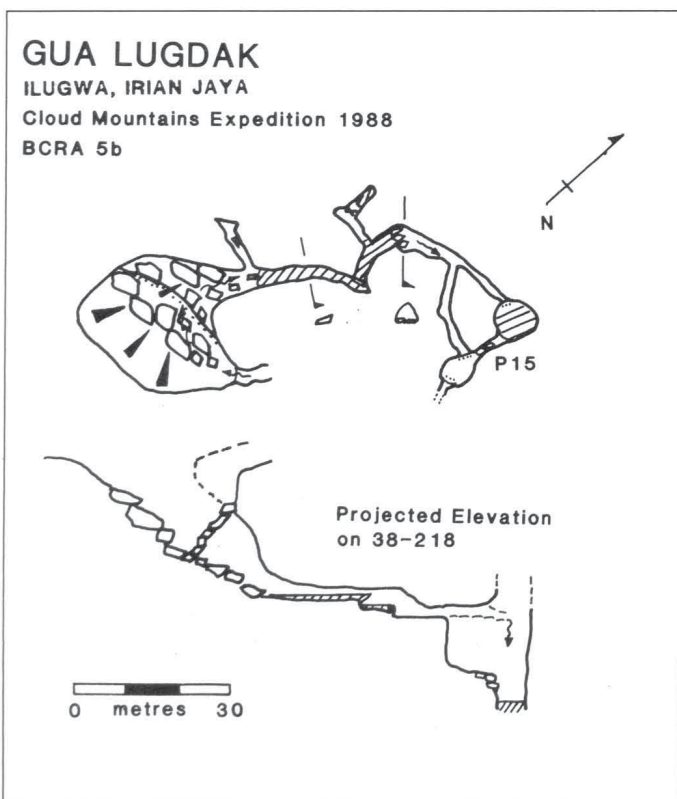on 38-218

0   metres   30

Figure 8.

that water flows out of its entrance in the wet season. The second, close by, carries a smaller stream and is almost certainly part of the same system. The local people demanded a fee to descend both of these shafts. The third, Gua Ikat, a half hour further along the track, takes a small stream running off shale and was not fully explored due to a lack of rope; it provides excellent caving (see Fig 7).

The Ilu river resurgence cave is the main resurgence for the area and can be reached by following the steep path down into the gorge past Gua Kwalinga, No further work was carried out in that or adjoining systems due to a combination of relatively high water conditions and access disputes with the local people.

The likely principal sink for the Ilu is 40 minutes walk south east of Ilugwa village, at the entrance to the gorge leading to Yogaluk. The sink is frighteningly impressive: at the first visit a river of 8 cumecs was sinking and, even when the flow had reduced to 4-5 cumecs, the cave itself could not be entered for more than about 30 m. A clear flood level marks the cliff about 5 m. above the top of the entrance and the 1985 team reported an overnight rise of over 10 m. A few shafts were visited in Danda, a nearby abandoned high level, but these led down to static water. The best hope of entering the Ilu river system would seem to be from the resurgence end.

On the path down to the Ilu sink from Ilugwa, just below the airstrip, is another sink cave, Gua Lugdak. This is heavily choked with vegetation and flood debris and floods to the roof in wet weather. It ends with a pitch leading to a sump (see Fig 8). The water presumably also drains to the Ilu resurgence.

The Wolo/Ilugwa area appears to have large areas of shale overlying limestone, with sink caves found at the margin. Although many obvious features have now been checked, the 1988 explorations were hampered by wet conditions and there is undoubtedly more work to be done in this area. Rainfall data (see Fig. 9) suggests that the ideal time to visit the area is July and a visit to the Ilu resurgence in dry weather might pay dividends. One very disturbed night's sleep confirmed that the area is seismically active.

USILIMO

This area was visited for 2 days during the return from the Wolo/Ilugwa valley. Usilimo lies an hour's walk beyond Wagawaga, the point at which the road then ended. The village itself is sited at the eastern edge of the Baliem Valley, north-east of Wamena. Behind the village the mountains rise up to 3000 m. From this base, several shafts and caves were explored, the entrances having been identified by local people.

The highest shaft, (Lubang Saknerah) lies 200 m. above the valley bottom and is spectacular, 30 m. wide and 80 m. deep; unfortunately it is choked at the bottom. Another, nearby,
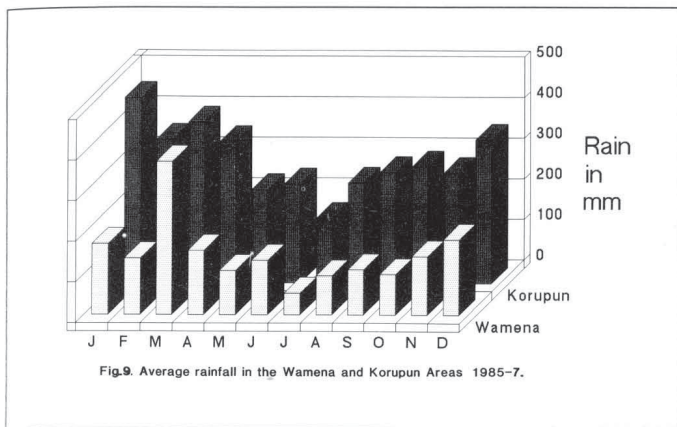
Fig.9. Average rainfall in the Wamena and Korupun Areas 1985-7.

Figure 9.

(L.Yiluaria) was descended for 30 m. but was again choked. The most significant cave in the area was Gua Wikuda, located at valley level, with an unimpessive stooping entrance leading to 1.7 km. of well-decorated passage (see Fig 10). This cave is obviously well known and, thanks to a visit by 2 American tourists, the 'owner' now insists on an access payment. Above Gua Wikuda is the Wikuda shaft, unfortunately blocked, despite its 50 m. wide dimensions. On the opposite side of the valley is Gua Hubak, a short cave ending in an opening which is, quite literally, a rat-hole (see Fig 11).

The potential may exist for the discovery of other shafts and caves higher up in the valley, or in the mountain to the north of the village, but it seems unlikely that any obvious features near the valley base remain to be explored.

KORUPUN

Korupun lies in an enclosed valley on the southern slopes of the main range, 90 km. south-east of Wamena; by air it is less than one hour's flying time. The flight-path goes along the Baliem and through the south gap, then across the grain of the country over Ninia and the Heluk, Seng, and Solo River valleys (the latter devastated by earthquakes) and finally along the Erok River gorge to the Korupun airstrip.

Despite the breathtaking scenery, the only speleological features seen in the dense forest were a stream emerging from a point half way up the east wall of the Seng and cascading several hundred metres to the river, and a large open shaft just before clearing the ridge above.

At Korupun the mission station is staffed by Elinor Young, a courageous American who, despite partial disablement, had travelled the rugged paths in the surrounding mountains, borne aloft on a sedan chair; she speaks the local Kimyal dialect and was extremely hospitable. She had previously visited two local caves known, imaginatively, as "Wet Cave" and "Dry Cave" (see Figs 12 & 13) and had arranged local guides. Both caves had previously been explored as far as pitches and rumour had it that a family had entered one and emerged in the next valley.

Despite Elinor's efficient preparation in Korupun, the team discovered that once out in the field they were subject to the capricious whims of the local chiefs. Communications were difficult, with interpreters switching between English, Indonesian, Yali and Kimyal and, like their colleagues elsewhere, the group's progress was often interrupted by protracted financial negotiations.

The two caves were finally explored, after much debate and negotiation. They were sited about 2-2.5 hours walk east of the village on the south side of the valley and both had walk-in entrances. Wet Cave had a small stream and some pleasant pitches to a depth of 100 m. before becoming choked. Dry Cave was reached very late in the day and after considerable negotiation. It was explored in haste and was larger, with a loose dry pitch leading to a constricted series of rifts from which issued the tantalising sound of a small stream.

Local people spoke of other caves to the east of Korupun within a few hours walk of Sela. Carriers were employed and the
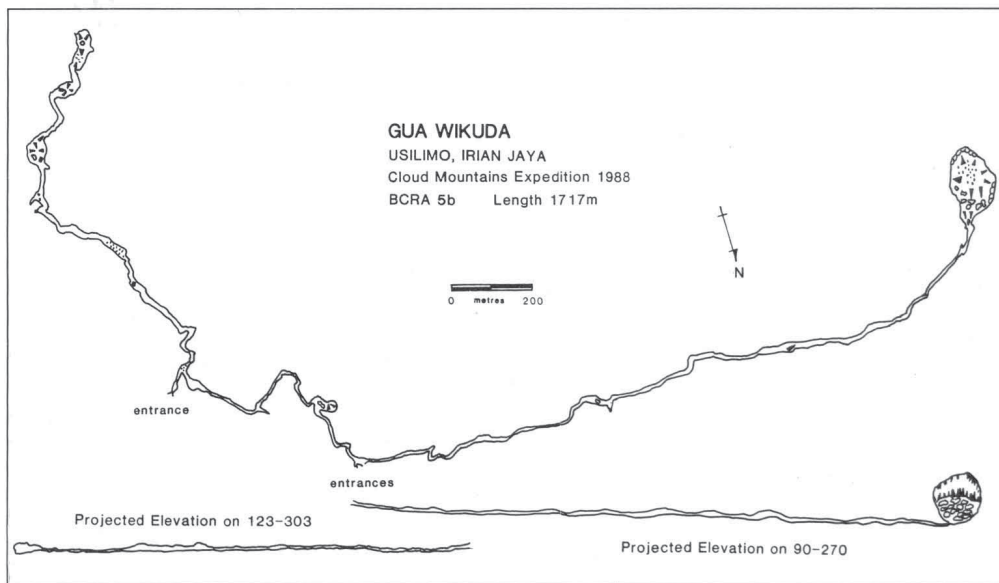
Figure 10.

GUA WIKUDA

USILIMO, IRIAN JAYA
Cloud Mountains Expedition 1988
BCRA 5b   Length 1717m

0 metres 200

entrance

entrances

Projected Elevation on 123-303

Projected Elevation on 90-270

Figure 11.

GUA HUBAK

USILIMO,IRIAN JAYA
Cloud Mountains Expedition 1988
BCRA 5b

0 metres 30

Figure 12.

WET CAVE

KORUPUN, IRIAN JAYA
Cloud Mountains Expedition 1988
Length 312m Depth 104m
BCRA 5b

Cross Sections

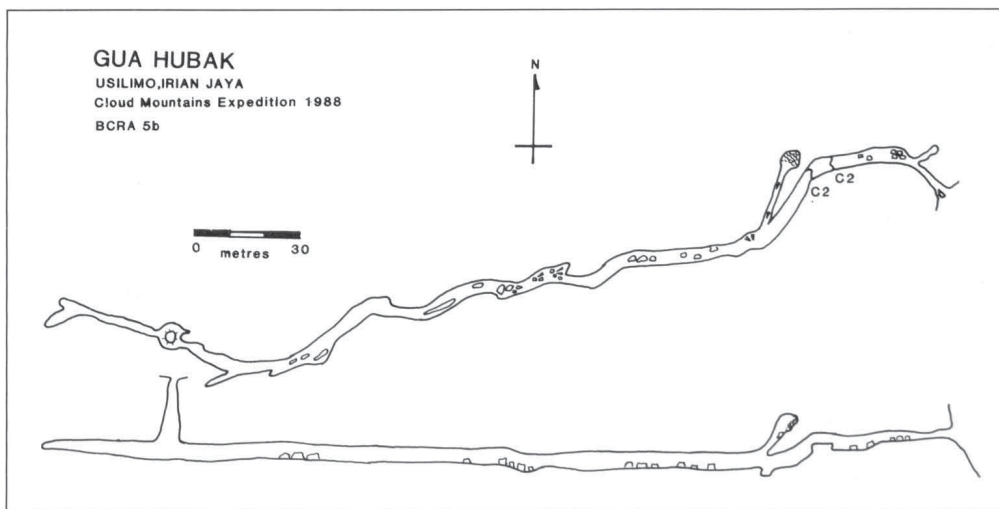**DRY CAVE**

KORUPUN, IRIAN JAYA

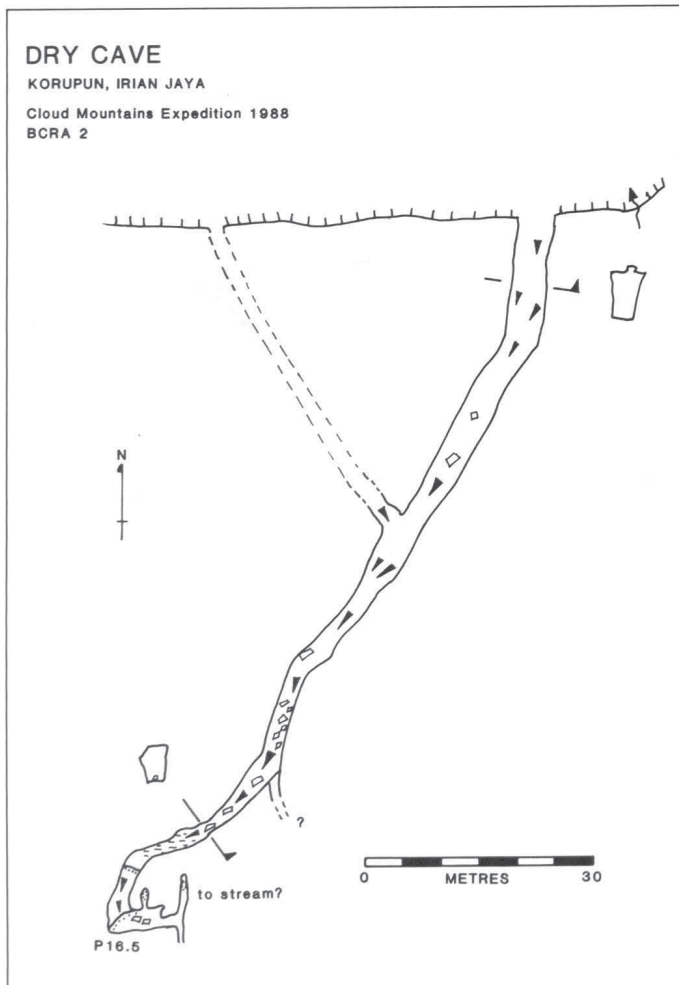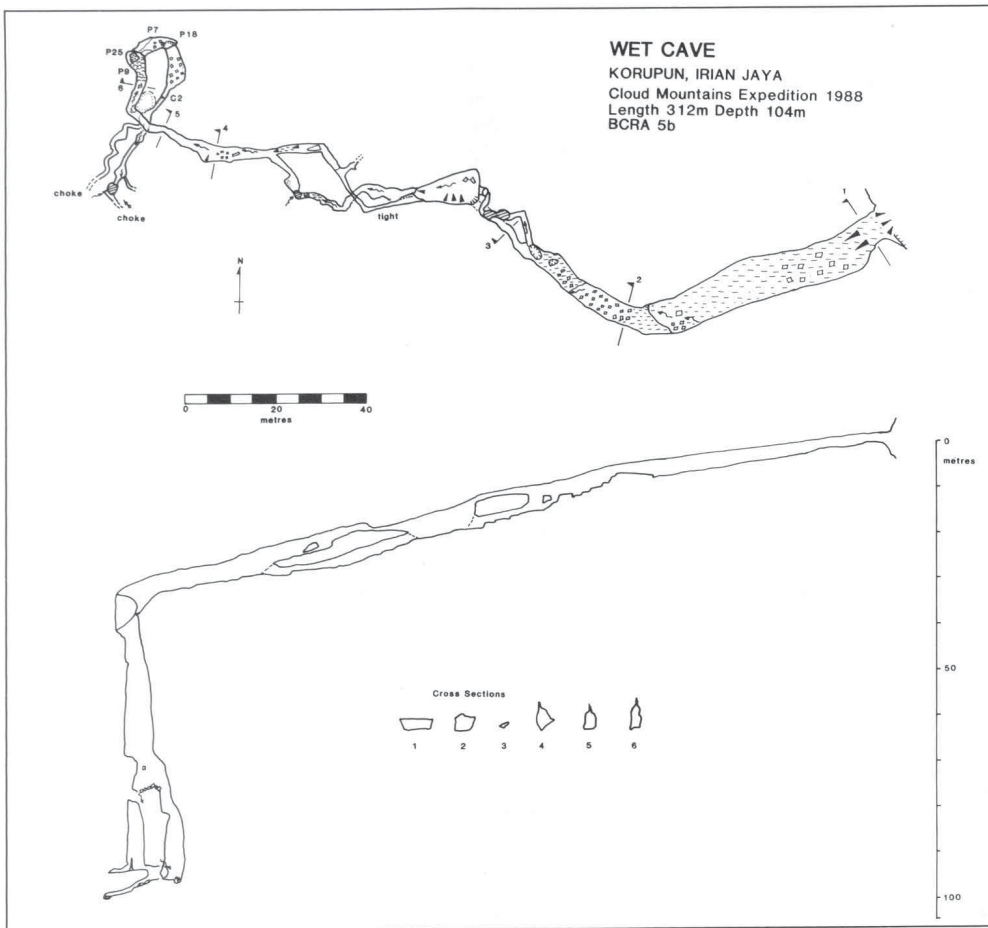Cloud Mountains Expedition 1988
BCRA 2

Figure 13.

team set off to look at caves in the next valley near Durum. One of these, 1 km. above the village, was just a collapsed rock shelter. Another, Gua Diringam, close to the track and half an hour's walk from Durum, was a resurgence cave in which one entrance led to a high aven down which the stream was falling (see Fig 14). The other entrance led to a small stream crawl of decreasing size.

Despite much questioning, no other caves appear to exist in the area. This may, however, be related to the fact that neither the guide nor the carriers were interested in visiting caves.

The team then followed a path for 3 days towards Wamena, crossing the main range at 3750 m. and walking down to Angguruk. En route, some small holes were seen in limestone pavement, but no features of real speleological signficance were observed.

During the return flight from Angguruk to Wamena, soon after crossing the main range, a promising area was seen with clear evidence of shafts. Unfortunately, time did not permit a visit.

POTENTIAL FOR FUTURE EXPLORATION

It is clear that the remote, uninhabited mountains in the Wamena area have considerable potential for cave development.

Early in the trip an aerial reconnaissance was made with the assistance of the Mission Aviation Fellowship, whose willingness to fit expedition flights into an already busy schedule was a major contribution to the success of the expedition. The flight went over the region between G.Trikora and a point 60 km. to the west, just beyond Lake Kulip. Visibility was good and enabled examination of the northern slopes of the mountain range, between the high peaks and the Baliem River. The southern slopes of the range were not covered since they are difficult to reach and were largely cloud covered on the day.

The high mountain areas are completely uninhabited. Vegetation varies from open grassland with tree ferns to bare limestone pavements, craggy outcrops and moss forest. The only big animals appear to be large rats, possums and a few birds. The limited wildlife is presumably the result of the fairly recent retreat of the glaciers. Relief varies from 3000 m. at the East Baliem to 4720 m. at the summit of Trikora, with the majority of the land lying between 3350 m. and 4000 m. It is cold and wet, and because of the altitude and vegetation, very difficult terrain.

During the reconnaissance flight a number of speleological features were seen. Subsequent work on the ground resulted in the largest three sinks being visited, together with one of two observed regions of dolines. All three sinks were choked although one had a large but short cave associated with it and would be well worth a dig! The doline area produced a 30m. deep shaft but shortage of time meant that the area was not fully investigated.

Figure 14.

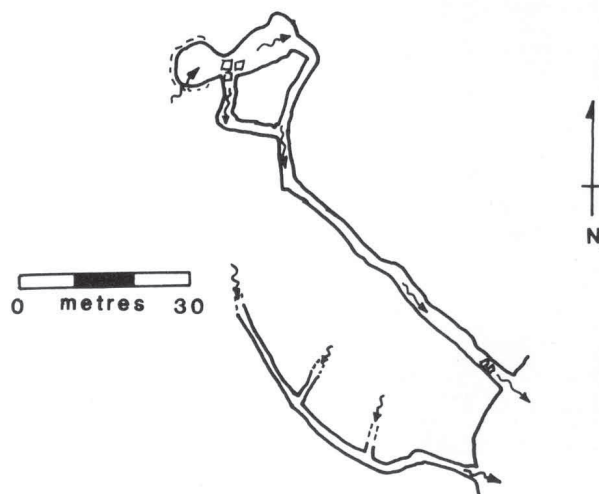Decorated old phreatic passage in the Balien valley close to Wamena. Photo: Dave Checkley.

One small sink cave, Kulip Sink, was visited but abandoned at the head of a very wet shaft. The entrance to this sink was obscure and had not been seen from the air, it was situated at the low point of a large grassy basin and several of these had been seen on the recce flight. These features and the others observed from the air remain to be visited and our experiences show that other entrances will be discovered by careful searching on the ground.

The potential of the region depends very much upon the direction of the drainage. It is clear that water is running off impervious beds and sinking into the limestone at heights of up to 4000 m. The question thus arises - where does the water resurge? The only resurgence seen from the air was at the far eastern end of the range at an altitude of approximately 1850m. This may well be simply related to the choked river sink close to Habbema on the edge of the extensive peat bog. If this is the case, then the associated cave may be quite localised and about 1000 m. deep, carrying a stream of between 1 and 2 cumecs. This hypothetical link would suggest that drainage is along the strike, since the bedding generally dips north.

The depth potential presented by such a drainage pattern, where resurgences will be along the spectacular Baliem Gorge below Wamena, will be in the order of 2-2500m. from the highest sinks. The absence of resurgences higher up the Baliem, in the area visited between Habbema and Kwiawogwi, points either to very long caves developed along the strike or more probably, in view of the complex folding evident in the high peaks, some drainage to the south. The southern edge of the range is very steep, with height differences of 4700 m. over only a few kilometres in certain areas. Resurgences on this side of the range would clearly give the area world depth potential but the difficulties of exploration in this remote and inaccessable area will be immense.

Needless to say, there are plans to return to the Trikora range in 1990 . . .

CAVE DESCRIPTIONS

LAKE SINK — Habbema (Fig. 4)

Length: 80 m.   Depth: 50 m.
Altitude: 3160 m.   Latitude: 4 11   Longitude: 138 36

Location

From a camp in the southerly branch of the east Baliem valley, follow a track for 2 km. south and over the ridge from Lake Habbema. Various Dani hunting tracks lead to the enclosed valley in which the Lake is situated. The tracks are easily missed and the consequence is hard walking for a considerable distance. A stream leaves the Lake on its south-eastern point and flows 1 km before entering an impressive doline. The flow was about 0.25 cu.mec. with a temperature of 16°C. at noon.

Description

The stream sinks in boulders at the bottom of a small waterfall in the doline and can be heard flowing at various points along the east wall to the end of the cave. The entrance slope is made up of large boulders and sediment which are stabilised in the daylight zone by vegetation. The boulder choke at the lower end was searched in vain for access to the audible stream below. The west wall of the cave is a massive unstable sediment bank.

KULIP SINK CAVE — Kwiawogwi

Length: 120 m.   Depth: +30 m.
Altitude: 3353 m.   Latitude: 4 9   Longitude: 138 19

Location

Danau (Lake) Kulip is the lake furthest west from Habbema and that closest to Kwiawogwi. It is a good day's walk from Kwiawogwi to Kulip and the easiest route is to walk along the Baliem Valley east from the village, through a number of wooded areas to a point beyond Kulip. An easy ascent up a steep open ridge can then be made. Once over the col the path descends slightly and 1 km. beyond, down the valley left, leads to a large grassy, flat-floored basin. (The path carries on up right, over a series of small ridges towards the lake). In the southern most and lowest point of the basin an obscure stream sinks into a narrow canyon which leads to the entrance.
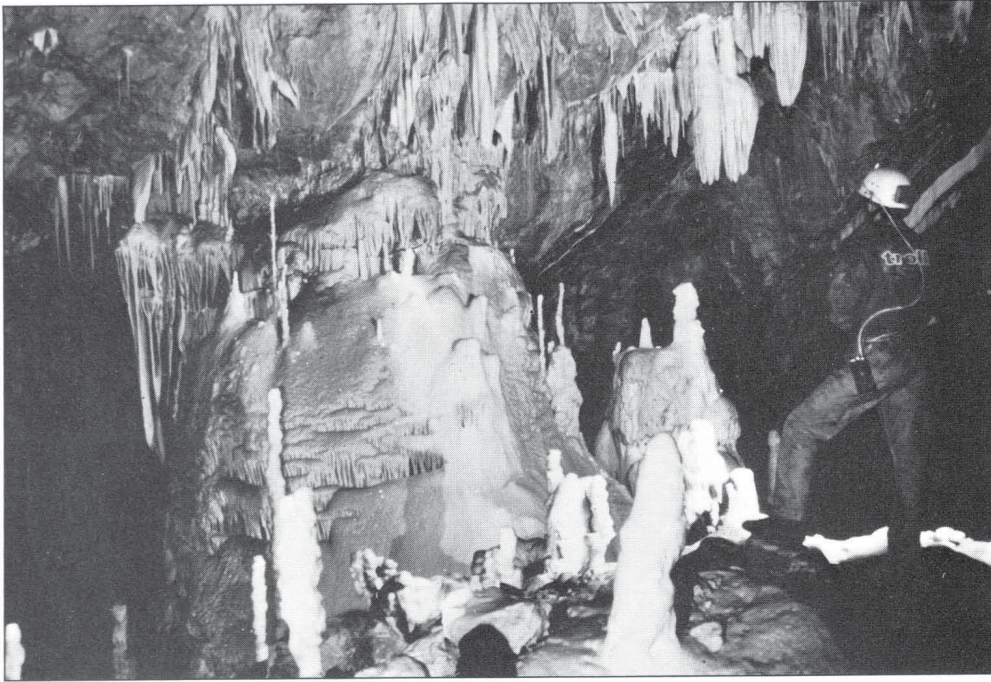
Description

The 1 m. wide, high vadose canyon descends to the head of a shaft. This was descended for 10 m. to a point where it was impossible to stay out of the stream without bolts. The flow was approximately 200 l/sec. The shaft bottom was not visible and the cave draughts strongly.

SHAFT NEAR DANAU KULIP — Kwiawogwi

Depth: 30 m.
Altitude: 3660 m.   Latitude: 4 9   Longitude: 138 18

Location

Follow the path up the ridge from Kwaiowgwi to the col and, having dropped slightly for 500 m., the faint path climbs up right and continues to the west. Several small north - south valleys are passed before a wooded area of limestone pinnacles and deep depressions is reached. The most westerly edge of this area is on the ridge top with a spectacular view down to Kulip. From the path on the ridge, follow the shallow valley down northwards away from the lake. 300 m. further on at the end of the valley is the slightly obscure shaft entrance.

Decorated old phreatic passage in the Balien valley close to Wamena. Photo: Dave Checkley.

Description

The shaft is overhung by trees and vegetation. It is approximately 4 m. by 6 m. and descends to a calcite choke.

GUA LUGDAK — Ilugwa (Fig. 8)

Length: 93 m.　　　Depth: 35 m.
Altitude: 1830 m.　　Latitude: 3 51　　　Longitude: 138 55

Location

Walk down the airstrip at Ilugwa and follow the path at the end which drops down towards the Ilu river. On the left, two small valleys join in a tree lined shakehole.

Description

The entrance is at the bottom of a steep-sided shakehole and is 5 m. high and 4 m. wide. The river flows down a boulder slope to a level section with a small blind passage to the left. Ahead, the passage is almost completely blocked by large tree trunks and flood debris. Beyond this logjam, a small side passage on the left ends in a sump. The main passage soon divides. The right branch leads to a 15 m. pitch ending in a sump above and beyond which is an unexplored high level. The left branch also drops to a sump.

GUA KWALINGA — Wolo. (Fig. 6)

Length: 603 m.　　　Depth: 229 m.
Altitude: 1980 m.　　Latitude: 3 51　　　Longitude: 138 54

Location

Three quarters of an hour's walk beyond Wolo towards Ilukwa, a steep gorge drops down on the right towards the main valley bottom. Two streams flow across the path into this gorge and first of these sinks in Gua Kwalinga, 500 m. from the path. The second stream sinks in an adjacent shaft which was not descended but is likely to be the first inlet near the bottom of Kwalinga.

Description

This spectacular shaft is filled with spray from the sizeable stream. In wet weather the system fills completely. Bolting and judicious use of natural belays enable a moderately dry descent of the left hand wall of this 175 m. shaft in dry weather. At the bottom, a fine sporting cave leads via pools, canals and cascades to a sump. This may be passable but is risky in view of the effects of daily rain.

GUA IKAT — Ilugwa (Fig. 7)

Length : 403 m.　　　Depth : 177.8 m.
Altitude : 2050 m.　　Latitude : 3 51　　　Longitude : 138 56

Location

Twenty minutes walk towards Wolo from Ilugwa, the track descends a hill to a small stream running on shale before rising again to the next ridge. 15 m. SE of the track the stream drops into the head of a shaft with extensive slumping of soil on the NE side.

Looking west towards the Carstenz from the benches at 3000 m. above Kwiawogwi.

One of the many lakes in the Trikora range.
Photo: Dave Checkley.

Description

An obvious tree belay on the NE side of the shaft has been used for a local ladder made of vine and sticks. This descends the first pitch of 26 m. to a spray filled ledge with a large quantity of rotten vegetation. A rebelay on bolts gives a clear hang to the floor of the shaft 30 m. below. The water sinks in one corner but an obvious flood overflow can be followed to join the water again at the head of a short pitch. At the time of exploration the stream passed through the carcass of a pig before cascading over the drop. (Attempts to recover the pig accounted for the ladder; one local man asked if we had recovered the meat: it was several months old . . .)

At the foot of this drop a dig in water has created a low duck which leads via a wet stoop to a canyon passage. This descends in a series of short awkward pitches with difficult belays. One side passage brings in a small stream which originates from an aven. Exploration ended at a wet pitch, estimated at 30 m., for which the team had insufficient rope. The caving is excellent but flood debris on the walls shows that the passage fills to the roof and a return through the duck would become impossible with only a small rise in water.

LUBANG PUTELLA — Pugima

Depth: 50 m.
Altidude: 1840 m.   Latitude: 4 5   Longitude: 139 3

Location

A major track leads from Wamena over the Baliem river to Pugima. From this village take the high path towards the village of Puali. The shaft is one hour from Pugima and 200 m. directly above Simokolek Kampong.

Description

This 40 m. wide impressive shaft can be descended without equipment on the northern side. A number of short sections of cave lead off around the rim, but no significant way on was found.

LUBANG WAMBIGMO — Pugima

Depth: 30 m.
Altitude: 1640 m.   Latitude: 4 5   Longitude: 139 3

Location

As for L.Putella take the track for Puali. This obscure entrance was hidden by bracken 3 m. off the track just below Simokolek Kampong.

Description

The 2 m. wide entrance is overhung by a boulder. The shaft ends convincingly at a boulder floor.

GUA GUAM — Anelaga, Pugima (Fig. 5)

Altitude : 1590 m.   Latitude: 4 6   Longitude: 138 57

Location

From Pugima take the obvious track NE over a small ridge. On the far side is the resurgence of Gua Guam. A single pole bridge crosses the stream to continue the track to the adjacent village of Anelaga.

Description

A stream of about 2 cumecs flows out of an arched entrance about 10 m. across. Boulders in the stream provide an alternative crossing, slippery but safer than the pole. The passage continues for approximately 160 m. in cold deep water to what appeared to be a sump. Due inadequate equipment this was not confirmed.

GUA WIKUDA — Usilimo (Fig. 10)

Length: 1717 m.
Altitude: 1676 m.   Latitude: 3 57   Longitude: 138 50

Location

Beside the airstrip at Usilimo is a wooden church. Take the path behind the church which heads directly towards the cliff, in line with the largest shaft on the hillside. The two small entrances are insignificant holes at the base of the cliff, 1 m. wide and 2 m. high. The cave is best found with the help of local people as the cave is well known and occasionally visited by tourists.

Description

A climb down through either entrance drops into a passage. The right-hand branch intitially runs parallel to the cliff face and is surprisingly similar for almost its entire length. The passage is narrow, with a flat mud floor, and the left wall is a calcite slope. It continues like this for about 1 km. with occasional boulders in the floor and some attractive calcite and stalagmite sections. Finally the passage closes down in a 50 m. wide, 100 m. high chamber. The roof of the chamber and the top of the boulder pile are covered in speleothems and no way on was found.

The left-hand branch from the entrance is of similar dimensions, continuing past a small boulder-filled chamber to the edge of an 8 m. wide circular shaft which is 5 m. deep with a mud floor. Water can be heard flowing but is not accessible. Across the shaft the passage continues to a T-junction. To the left is a small entrance, after about 30 m. To the right is a further 500 m. to a boulder filled chamber with no way on.

LUBANG SAKNERAH — Usilimo

Depth: 80 m.
Altitude: 1850 m.   Latitude: 3 56   Longitude: 138 51

Location

From the village of Usilimo, walk north-east up the valley for 2 km. until the obvious path climbs up the northern side of the valley. Once up the first steep section, the path climbs west above the cliffs. The entrance is 500 m. past a stream on the left of the path and is well known to local people.

Description

A spectacular 30 m. wide shaft inhabited by a colony of large fruit bats. A vegetated descent leads to a large ledge and continuation to a choked floor.

LUBANG YILUARIA — Usilimo

Depth: 30 m.
Altitude: 1750 m.   Latitude: 3 56   Longitude: 138 51

Location

The entrance is in the cliff on the north side of the valley, 2 km. east of Usilimo.

Description

The 8 m. by 3 m. entrance is in a bay in the cliff and gives a clean 30 m. descent to a mud and guano choked floor.

GUA HUBAK — Usilimo (Fig. 11)

Length: 250 m.
Altitude: 1690 m.　　　Latitude: 3 56　　　Longitude: 138 50

Location

The entrance is in a low cliff 500 m. north of the northern end of the village of Usilimo. It is essential to obtain a guide.

Description

The 1 m. by 0.5 m. entrance leads via a stooping passage to a T-junction with a larger passage. Left is 250 m. of walking and scrambling along a 3x5 m. passage passing under a skylight and ending in a choke. A tiny hole draughts strongly and is obviously used by wildlife.

WIKUDA SHAFT — Usilimo

Depth: 40 m. approximately.

Location

The shaft is situated in the hillside above Usilimo, directly above Gua Wikuda; it is obvious from a considerable distance.

Description

The shaft is 50 m. wide and about 40 m. deep; it appears to be a large collapsed cavern. At the bottom are several small passages but all choke.

WET CAVE — Korupun (Fig. 12)

Length: 315 m.　　　Depth: 100 m.
Altitude: 2000 m.　　Latitude: 4 28　　　Longitude: 139 40

Location

From Korupon, cross the Desul river and climb 300 m. to the village of Laabon then follow the Aso river up its north side. The track then crosses the river about 4 km. further up the valley and climbs to the "upper gardens" on a bench, bounded to the south by cliffs. The entrance is at the base of the cliff in forest and next to an impressive waterfall.

Description

An 80 m. entrance slope leads either to a muddy climb down to a canyon or a climb up to a crawl. Both ways lead to a chamber. Once again, upper and lower passages lead off, rejoining in a large joint contolled passage to a pitch head. A small stream flows into the entrance passage 50 m. into the cave and flows via the lower route to the pitch. The shaft of 70 m. starts with a ramp and, after 40 m., a loose ledge gives access to the final drop. From the bottom there are two leads in the same joint, one high (passing a fine aven with a 2 m. stalagmite) and one low. Both end in calcite blockages after only 20 m.

DRY CAVE — Korupun (Fig. 13)

Length: 150 m.　　　Depth: 50 m.
Altitude: 2000 m.　　Latitude: 4 28　　　Longitude: 139 40

Location

1 km. east of the "Wet Cave" and in a similar situation, in forest at the bottom of the cliff, is "Dry Cave". A small hut has been built at the entrance to this nearly perfect rectangular passage. It appears almost man-made but for a fossil phreatic half-tube in the roof.

Description

120 m. of sloping passage leads to a steepening slope and a 16 m. pitch. The chamber at the bottom has no obvious way on but a stream can be heard.

GUA DIRINGAM — Korupun

Length: 120 m.
Altitude: 1915 m.　　Latitude: 4 26　　　Longitude: 139 37

Location

This resurgence cave is just off the track between Duram and Subilil on the west side of the Erok valley.

Description

The main passage is a small streamway leading 80 m. to the bottom of a 20 m. aven with a waterfall. 5 m. to the south of the main entrance is another 40 m. of passage, fed by various small inlets on its right wall, which becomes a flat out crawl.

FIELD LOGISTICS

A. General

Permission to enter Irian must be obtained in advance and visitors are required to have the necessary documentation (surat jalan) which is carefully checked by the police and all losmen owners are required to furnish the police, daily, with a list of tourists staying in their establishments. A surat jalan will normally be issued to citizens of most western nations provided they have a ticket out of the country and enter and leave via certain designated air or sea ports.

Accommodation in Wamena is in small hotels (losmens). There is a limited choice and any tourist arriving at the airport will be assailed by local people who vie with each other to attract you to their establishment. The 1988 team stayed in Losmen Syaharial which was basic but comfortable and where the owner allowed use of one room as a tackle store. The cost per person was US$10 per night.

A range of western foods are available in the shops in Wamena and the local market supplies an extensive choice of vegetables and fruit. There are several small restaurants serving a limited range of Indonesian foods and a meal for 9 cost, on average, £10. Rice is available and, in the villages ubi (sweet potato) and other vegetables and fruit could be purchased.

Carbide was not available in Wamena although it can be purchased on the coast and air-freighted.

Labour can be hired easily, guides cost about £5 per day and normal carriers cost about £1 (they provide their own food, in theory). There are, however, problems and the difficulties of constant renegotiation are discussed below. In addition, labourers hired in Wamena tend not to be familiar with the localities into which you may walk. The area has a long history of inter-tribal fighting and local animosities are hard to avoid. Once you have engaged a guide (and visitors are besieged at the airport by numerous hopefuls) you may find that you are committed to using that individual for the remainder of your stay, even if you find him to be completely useless. Any attempt to change guides is likely to be thwarted by the tribal support system - your guides' relatives and friends will warn-off any potential usurper. The best advice is not to hire any labour in Wamena until you have had the opportunity to take advice from as many sources as possible.

Accommodation in the villages can be arranged either in unoccupied (or occupied) missionary houses. Negotiations for these may be conducted with the missionaries in Wamena or the local Pastor. Payment may not be requested but a donation to the Church funds is welcome. Accommodation may also be provided in the men's houses in small villages. This is very comfortable, providing you do not mind an audience and fleas, and is the best way to meet local people with the minimum influence from European culture.

Each village will require separate negotiation, effectively for bed and breakfast, as you are likely to be charged for the place to sleep, ubi and fire-wood. Overnight accommodation at Puali, for 3, cost about £3 including food and fuel; on the return walk the rate had risen to £5, for 2 people! Very often, every possible attempt will be made to extract money from visitors. Many local people will demand a fee in order to be the subject of a photograph, another indication of the insidious effects of mass tourism.

B. High Peaks

The 1988 trip revealed a number of logistical mistakes in trying to explore the caves high in the Trikora range. Although Wamena is the easiest place to reach from the coast, it is not a good place from which to mount an exploration of the high peaks. The people from Wamena do not travel to the high country and even those from the outlying villages in the East and West Baliem rarely go up on top. They therefore have little idea of the topography or of the problems which are likely to be encountered.

An improvement in this situation could be achieved by flying, not to Wamena, but to the villages south of the main range. From here, people regularly walk up to Wamena to trade and they have a much better understanding of the country and its difficulties.

The reason why the Baliem people are reluctant to go on top is not hard to see — it is very cold. Frost was not uncommon at night on the benches at 3300 m. and the local men traditionally wear only a penis-gourd; although this garment is normally lined with possum-fur it provides little protection against extremes of temperature. Even at Kwiawogwi frosts are not unknown and the population there almost starved when their crops were destroyed by two successive bad frosts. Their neighbouring tribe took no action to alert MAF to the Kwiawogwi predicament and it was only a chance landing by one of the pilots which saved the village.

Short camps at altitude can be maintained by having constant fires to provide the carriers with overnight warmth but any sustained work would require them to be supplied with waterproofs, clothing, tentage and, possibly, footwear. The Dani normally go barefoot and the exposed limestone of the pavements is extraordinarily sharp.

All food must be carried or airdropped to high camps since there is almost nothing in the vicinity to eat. The only food source is possum and large rat (cat sized). These are normally prepared by burning off the fur, gutting and throwing the carcass onto a fire until it is charred at which point it is dismembered and eaten. This process often takes place after dark and it is rather hard to ascertain if the meat is thoroughly cooked or not. Although the carriers were supposed to provide their own food they tended to consume the entire supply during the first night during extended singing sessions.

High altitude work will be made more productive by using air reconnaissance. This clearly provides problems both in terms of cost and timing - it is not always possible to obtain flights due to weather and MAF scheduling. Some areas will only be realistically explored with helicopter support.

The high mountain benches receive considerable rainfall and all evidence from caves around Habbema, Ilugwa and above Kwiawogwi indicates that underground flooding is extensive. Timing of any exploration must therefore take due account of the rainfall patterns for the area. The prize, of course, will be some of the deepest caves in the world.

TRAVEL

Five members of the expedition travelled to Indonesia from the UK, three from Australia and one was resident in Jakarta. From Jakarta the team flew to Irian Jaya using a Garuda Indonesian Air Pass - the most economical way of travelling for either 20 or 40 day periods.

Flights from Jakarta travel via Ujung Pandang (Sulawesi) and Biak to Jayapura, where the Sentani airport is 50 km. from the capital. Sentani is the HQ of the Mission Aviation Fellowship. Flights from here to Wamena go over seemingly endless lowland forest before entering the sheer limestone walls of the Pass Valley and descending to the Grand Valley; the change in scenery is astonishing. Wamena is claimed to be the largest town in the world supplied only by air; there are plans to build a road link but this will be a major engineering feat and will take many years.

A limited road network runs out from Wamena and has been driven north to Wagawaga. Outlying settlements, missions and villages are supplied routinely by the MAF. The relaxed confidence of their pilots belies the danger of flying in this region and, unfortunately, serious accidents occasionally occur. Whilst in Wamena, members of the team trained and equipped some MAF staff for SRT. This will enable the pilots to mount a rescue/recovery mission in the event of a future accident, using a helicopter as an abseil platform.

Vehicles can be chartered for transport along the road network and flights can either be infilled or chartered for more distant destinations The MAF schedule is, however, very tight and constantly disrupted by weather and emergencies. Charter for fixed-wing aircraft from MAF cost £80 per hour, and £275 per hour for a helicopter.

Although the MAF staff were incredibly supportive of the expedition, anyone visiting the Grand Valley must realise that MAF's normal operations, in support of the missions and the local people, must come first. Without air travel, all transport is by foot.

The area is well served by high quality footpaths. In the main valley these are often constructed between drainage ditches and are frequently wide enough for two or three people to walk abreast. The area is superb for trekking. Journeys outside the main valley are strictly controlled by the police, with whom all visitors have to register, and will require special permission.

THE DANI

The people of the Baliem area are normally grouped together under the name of the "Dani", although this is in fact not their name for themselves but a name given by their neighbours which has slightly abusive overtones.

The Baliem valley has been continuously inhabited for 25-30,000 years with pig-raising and horticulture well established by about 5000 years ago. The Dani have a highly sophisticated form of agriculture and the first explorers to visit the Grand Valley, in 1938, commented, "From the air, the gardens and ditches and native-built walls appeared like the farming country of Central Europe". Indeed, the pattern of irregular fields and ditches, with apparently neat settlements, still present an attractive and unique landscape.

Traditionally the people live in villages of round thatched, 'bee-hive' huts. These are formed by a skeleton of poles supporting a low ceiling and with a central hearth. Between the ceiling and the thatch is a sleeping platform. The ground floor is covered in dry grass and the entire inside is polished black from wood smoke and greasy bodies (Dani men by tradition covered their bodies in pig-fat). The open fires and low doors give the impression that the inside will be intolerable with smoke but in fact this is not so; when the fire is lit, air circulation takes the smoke out of the door and the hut is warm and extremely comfortable, the best description is that of a "human nest": fleas also find it ideal. Cool by day and warm by night they are well suited to the climate. In keeping with much of the rest of New Guinea, the women and men sleep separately and our experience was that the standard of accommodation for the women was often far lower than that for the men.

A controversial census of elders in 1969 unanimously agreed to Irian becoming part of Indonesia to which it now represents an important asset. Its timber and mineral reserves are immense and rapidly being exploited and its vast land area has been seen as an opportunity for settlement by ethnic Javanese under the "Transmigration" programme. The Wamena area has recently been designated as an area for transmigration, under which families from the overcrowded central Indonesian islands are moved, given land and a house, tools and seeds and are expected to blossom into model communities in their new host country. This programme causes considerable resentment on the part of the native population and a small guerilla movement, the OPM, has been active in parts of Irian Jaya.

Changing patterns of agriculture and housing are also evident. Under Indonesian influence, traditional crops are being displaced by rice and the Dani field-pattern is being broken by the regular outlines of wet rice paddy. Little work seems to have been done to assess the suitability of cultivation of this new crop in the area and fears have been expressed that the constant flooding will increase soil salinity and poison the land. Dani huts are also now discouraged and the valley is dotted with Indonesian-style tin-roofed shacks. These are hot by day and freezing by night; the Dani prefer their huts.

The influx of Indonesians and tourists have led to an increase in the amount of cash crops being grown and Wamena market has a wide selection of vegetables. This is resulting in the best agricultural land being used for this purpose and more marginal land being cleared for the growth of ubi. On one walk, where there was a clear view of the main valley near Pyramid, smoke plumes marked 27 forest clearance fires. The steep relief and poor soils will not long support this activity and there is already extensive evidence of soil erosion.

Increasing tourism brings another insidious effect. The Baliem has become an important location for solo travellers and adventure holidays. Westerners pay large sums of money to visit the area for a short period. The Dani are astute businessmen and have realised the income potential of this situation. On almost every occasion when local people were hired as guides or carriers, the agreed rate was settled but, within a short period, there would be a sit-down, discussion and a demand for an increase in the pay. The tourists presumably pay, reinforcing what seems to be a trend in Dani negotiation techniques (the last case of human cannibalism, in the late '60s, resulted from a similar argument), but the wastage of time and irritation can be immense.

An excellent summary of the human, physical and political geography of Irian is given in "Indonesia, a Travel Survival Kit" which must be recommended reading for anyone intending to visit Irian. (Bruce, 1986).

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B.C.R.A. Research Funds and Grants

THE JEFF JEFFERSON RESEARCH FUND

The British Cave Research Association has established the Jeff Jefferson Research Fund to promote research into all aspects of speleology in Britain and abroad. Initially, a total of £500 per year will be made 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 which 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 which 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(s) must be the principal investigator(s), and must be members of the BCRA in order to qualify. Grants may be made to individuals or small groups, who need not be employed in universities, polytechnics or research establishments. Information and applications for Research Awards should be made on a form available from S. A. Moore, 27 Parc Gwelfor, Dyserth, Clwyd LL18 6LN.

GHAR PARAU FOUNDATION EXPEDITION AWARDS

An award, or awards, with a maximum 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 exploration in remote or little known areas. Application forms are available from the GPF Secretary, David Judson, Rowlands House, Summerseat, Bury, Lancs. BL9 5NF. Closing date 1st February.

SPORTS COUNCIL GRANT-AID IN SUPPORT OF CAVING EXPEDITIONS ABROAD

Grants are given annually to all types of caving expeditions going overseas from the U.K. (including cave diving), for the purpose of furthering cave exploration, survey, photography and training. Application forms and advice sheets are obtainable from the GPF Secretary, David Judson, Rowlands House, Summerseat, Bury, Lancs. BL9 5NF and must be returned to him for both GPF and Sports Council Awards not later than 1st February each year for the succeeding period, April to March.

Expedition organisers living in Wales, Scotland or Northern Ireland, or from caving clubs based in these regions should contact their own regional Sports Council directly in the first instance (N.B. the closing date for Sports Council for Wales Awards applications is 31st December).

THE E. K. TRATMAN AWARD

An annual award, currently £25, 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, not later than 1st February each year.

BRITISH CAVE RESEARCH ASSOCIATION PUBLICATIONS

CAVE SCIENCE — published three times annually, a scientific journal comprising original research papers, reviews and discussion forum, on all aspects of speleological investigation, geology and geomorphology related to karst and caves, archaeology, biospeleology, exploration and expedition reports.
Editor: Dr. Trevor D. Ford, 21 Elizabeth Drive, Oadby, Leicester LE2 4RD. (0533-715265).

CAVES & CAVING — quarterly news magazine of current events in caving, with brief reports of latest explorations and expeditions, news of new techniques and equipment, Association personalia etc.
Editor: A. Hall, 342 The Green, Eccleston, Chorley, Lancashire PR7 5TP. (0257-452763).

CAVE STUDIES SERIES — occasional series of booklets on various speleological or karst subjects.
Editor: Tony Waltham, Civil Engineering Department, Trent Polytechnic, Nottingham NG1 4BU. (0602-418418, ext. 2133).
No. 1 Caves & Karst of the Yorkshire Dales; by Tony Waltham & Martin Davies, 1987.
No. 2 An Introduction to Cave Surveying; by Bryan Ellis, 1988.
No. 3 Caves & Karst of the Peak District; by Trevor Ford & John Gunn, 1990.

CURRENT TITLES IN SPELEOLOGY — annual listings of international publications.
Editor: Ray Mansfield, Downhead Cottage, Downhead, Shepton Mallet, Somerset BA4 4LG.

CAVING PRACTICE AND EQUIPMENT, edited by David Judson, 1984.

LIMESTONES AND CAVES OF NORTHWEST ENGLAND, edited by A. C. Waltham, 1974. (out of print)

LIMESTONES AND CAVES OF THE MENDIP HILLS, edited by D. I. Smith, 1975. (out of print)

LIMESTONES AND CAVES OF THE PEAK DISTRICT, edited by T. D. Ford, 1977. (out of print)

LIMESTONES AND CAVES OF WALES, edited by T. D. Ford, 1989.

Obtainable from B.C.R.A. Sales
B. M. Ellis, 20 Woodland Avenue, Westonzoyland, Bridgwater, Somerset TA7 0LQ.

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