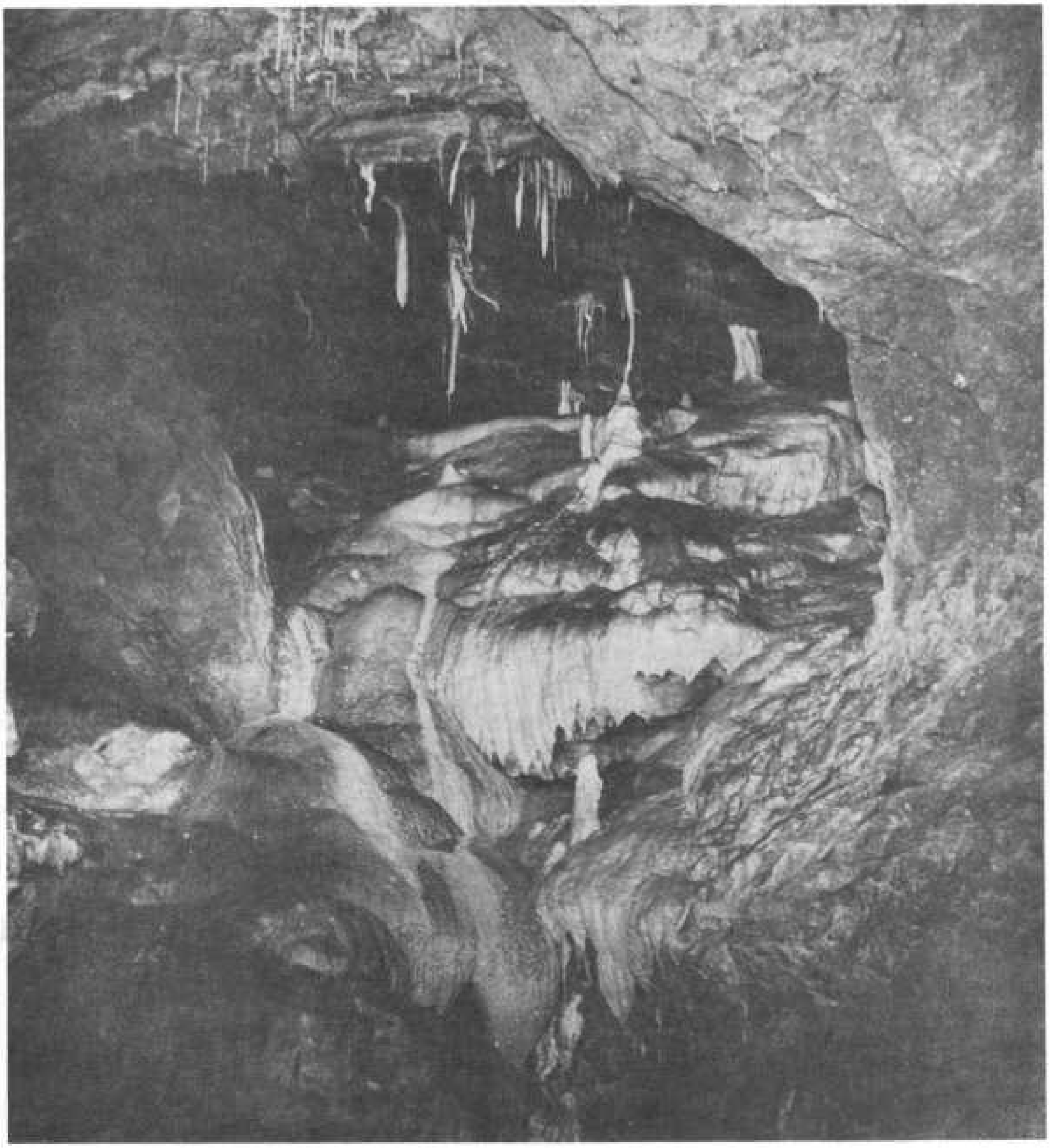
Vol. 17(2) 1979

Helictite

JOURNAL OF AUSTRALASIAN CAVE RESEARCH



Grotto in Nacimiento del Rio, Cortines, N. Spain.

photo by Guy Cox

HELICTITE

Helictite was founded by Edward A. Lane and Aola M. Richards in 1962.

This Journal was (and is) intended to be wide ranging in scope from the scientific study of caves and their contents, to the history of caves and cave areas and the technical aspects of cave study and exploration. The territory covered is Australasia in the truest sense — Australia, New Zealand, the near Pacific Islands, New Guinea and surrounding areas, Indonesia and Borneo.

In 1974 the Speleological Research Council Limited agreed to support the Journal with financial assistance and in 1976 took over full responsibility for its production.

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To explore and chart and organize and promote the exploration and charting of caves and cave systems.

To promote, support and assist in the scientific evaluation and investigation of caves and to foster speleology as a science and a sport.

To promote and to assist in the promotion of speleology and speleological Societies and to assist such Societies with grants of money and otherwise.

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REVIEW

KARSTHYDROGRAPHIE UND PHYSISCHE SPELÄOLOGIE, by Alfred Bögli. Springer, Berlin, 1978. Pp. xiii + 292

Alfred Bögli has long been a leader in karst and cave study on the continent of Europe and so this book was awaited with considerable anticipation. Its purpose is to bring together karst hydrology and physical speleology for a lay as well as a scientific audience. To this end simple examples are chosen and terms explained. One warms to the author's heartfelt recognition of his debt to his caving companions, one of whom lost his life in Hölloch Cave, which was so much Bögli's field laboratory.

The important characteristics of the rocks which give rise to karst are dealt with in the first chapter. In Australia we have a special interest in marginal materials for caves. Bögli regards karst-like features in silicate rocks as pseudokarst because they are due to irreversible decomposition rather than reversible solution and this results in so much clay and quartz sand that fissures opening up along joints get blocked up, preventing cave formation. Labertouche, Britannia and neighbouring caves in granite in Victoria don't appear to fit this theory and it becomes important to verify whether they are formed in undisturbed weathering mantle or in landslip material. If the former, the greater depths of weathering in Australia compared with Europe may explain the discrepancy. Bögli also discusses the quartzite karst problem; he combines the alternative explanations of the Venezuelans, hydrothermal alteration and long periods of exposure to extreme tropical humid weathering. The former does not apply in northern Australia and how the latter operates there, if at all, poses problems.

The second chapter, also the second longest, is on solution of karst rocks. He presents this complex matter lucidly, properly spending more space on the kinetics of the process than on saturation equilibria. It seems too early to write off oxidation of organic substances as a source of carbon dioxide for renewal of corrosion in caves. If this is important at the bottom of Bungonia Caves, it is likely to play a significant part in tropical caves. Nor does the presence of a cave at -2952 m in the Hicacos peninsula in Cuba of itself give proof of origin by mixing corrosion. The brief section on karst denudation at the end of this chapter scarcely does justice to the large body of work done on this in the last two decades and certainly some practitioners would not agree that it is impossible to separate surface and underground solution.

A review of surface karst follows, short because the central concern of the book is with the underground, but useful nevertheless. Bögli is careful about terminology showing, for instance, how Sweeting and this reviewer have confined the connotation of uvala more than European usage permits, yet one can see in his book how fresh problems arise. He describes as *'Wellenkarren'* certain minor solution features in the subsoil; this is likely to be translated as 'ripples' yet these are very different features form the two types of bare karst features already named this way in English. One wonders why rounded solution runnels *(Rundkarren)* are classed as secondary subsoil forms; in southeastern Australia there is no doubt that they occur as primary forms. Nor is it easy to understand why he states that geological organs have nothing to do with karst. Again Williams does not label dolines transitional between subsidence and solution dolines as alluvial dolines but as drift dolines. Drift in Britain includes much besides alluvium and so drift doline does not include the idea of a stream sinking in the doline as is involved in the alluvial doline of the classic Cramer paper. To say that cockpits are star-shaped because hemispheroidal residual hills interfere with doline development is to miss the point completely. Both hill and cockpit development are the product of more surface stream erosion than in temperate doline karst. One is also doubtful whether dry valleys are in fact more common in arid and seasonal climates; valleys with intermittent or ephemeral streams certainly are but these are usually marked by stream channels, less easily obscured by plant growth there. The definition of a dry valley involves absence of channels.

The short Chapter 4 is concerned with general hydrological principles relating to the development of underground karst and the hydrodynamics of underground flow are pursued comprehensively in a valuable way in the longer succeeding chapter. The author stresses the importance for cave development of height difference between the surface and the *'Vorfluter'*, the outflow points leading water away from the karst catchments, and emphasizes - it cannot be emphasized too often - that watertables in karst are in general fictitious, quoting Roglić, "In compact limestone, no groundwater tables can be formed'. He returns to this theme in Chapter 6 where he characterises the customary hydrological zones in karst, vadose, floodwater and phreatic, contrasting American and European views on the last, and once more in Chapter 7 also. There can be no doubt about the importance of this theme but perhaps there was no need to chop the material up into several short chapters.

Again the structure of the book does not seem to be the best when one passes over the contents of the five short Chapters following -(8) underground karst levels, (9) karst springs, (10) water tracers, (11) cave breakdown, (12) speleomorphology - the forms of underground erosion. The material is excellent but its organisation has no obvious rationale. There are minor points to quibble about; spigots in swirlholes are found at the surface as well as underground; scallops occur on ceilings and upper walls as well as on floors and lower walls.

The longest chapter in the book is on cave sediments, a useful introduction to a huge and growing subject. One could take up points of detail. Cauliflower is confined by Bögli to pool deposits whereas in English it is employed for certain subaerial as well as subaqueous cave decorations. In Australia we know that the longest straw in the world is not the 4.5 m one in La Clamouse Cave in France.

The following short generalising chapter on speleogensis rests soundly on the thesis that no single theory can comprise the complexity of structure and process and history which results in an immense diversity of caves.

A chapter of 13 pages on cave meteorology, though judicious in its selectivity, can scarcely do justice to this subject at the depth of treatment given to the morphology and hydrology, even if one includes the following 5 pages on ice caves, inevitably much concerned with the cave atmosphere.

A short discussion of cave classification concludes the text, through there is an appendix on cave mapping symbols. A magnificent set of black and white photographs ends the book triumphantly; Bögli could scarcely have failed to excel in this direction.

Overall there is a strong slant towards European Alpine karst, but that is particularly useful to us in Australia. The N. American karst is brought in effectively, the British not so well despite the great amount of work that has gone on there over many decades. The tropics scarcely come into consideration but that is completely justifiable given our present knowledge of tropical caves.

Some of Bögli's favourite ideas, such as mixing corrosion, perhaps loom too large. Despite these criticisms, the book is a most valuable addition to the literature and all those speleologists who can cope with clear, straight-forward German are well advised to buy it.

J.N. Jennings

SEA CAVES OF KING ISLAND

Albert Goede, Russell Harmon and Kevin Kiernan

Abstract

Investigation of two King Island sea caves developed in quartzitic rocks shows them to contain a wealth of clastic and chemical sediments.

Clastic sediments consist of wave-rounded cobbles, debris cones, and angular rock fragments produced by frost weathering and crystal wedging. Chemical deposits include a variety of calcium carbonate speleothems and also gypsum occurring as wall crusts and blisters. The latter appear to be a speleothem type of rare occurrence. Growth of gypsum is responsible for some crystal wedging of the bedrock. Three basal stalagmite samples have been dated by the Th/U method indicating Late Pleistocene as well as Holocene speleothem growth.

The caves are believed to have formed by preferential wave erosion during the Last Interglacial in altered and fractured quartzites. The evidence for pre-Holocene evolution of sea caves and geos in the Tasmanian region is summarized. Tasmania and the Bass Strait Islands provide a particularly favourable environment for the preservation of relict landforms on rocky coasts because of Late Quaternary uplift.

The potential of further studies of sea caves to test two recently advanced archaeological hypotheses is discussed.

INTRODUCTION

Geomorphological investigation, mapping and radiometric dating have been carried out in two sea caves on the SW coast of King Island between Surprise and Fitzmaurice Bays (Figure 1). The northernmost cave, Iron Monarch, is located at 143°52'0"E and 40°4'35"S and the southernmost, Blister Cave, at 143°52'5"E and 40° 5'42"S. Iron Monarch was discovered about 1960. It is the more extensive of the two and has not been previously recorded in the literature. Blister Cave was evidently known prior to 1954 as it is reported by Jennings (1956) who gave a description of both it and a small cave nearby.

The field work was done by the senior author who is responsible for the contents of the paper and for any expressed opinions. The Thu/U age determinations were carried out by Russell Harmon while Kevin Kiernan assisted with the field work and the surveys. He also compiled the cave maps.

The caves were selected for detailed study for a number of reasons:

(1) Their morphological characteristics indicate they are of considerable age and predate the Holocene (as suggested earlier by Jennings) despite their occurrence on a cliffed coast subject to strong wave attack.

(2) They contain significant deposits of both clastic and chemical sediments with the possibility of finding material suitable for dating.

(3) They contain well-developed spelothems - an unusual feature of sea caves developed in non-carbonate rocks. The presence of speleothems was explained by Jennings (1956) as being due to a source of calcium carbonate from overlying cliff-top dunes.

PHYSICAL ENVIRONMENT

The geormorphology of King Island has been examined in detail by Jennings (1959) who includes a description of the high coast from Surprise Bay to Fitzmaurice Bay. This part of King Island consists of a plateau surface which reaches its highest elevation of just over 100 metres about 2.5 km east of Blister Cave. From here it slopes down gently in all directions. To the west between Cataraqui Point and Surprise Point this surface is abruptly truncated at an elevation of approximately 40 metres by a cliffed coast subject to strong present day wave attack. In detail the coastal outline is irregular with numerous geos, inlets, natural arches and offshore rocks. The plunging nature of the cliffline was recognized by Jennings (1959) who pointed out that the real break of slope occurred at a depth of approximately 35 to 65 metres and that wave attack might be weakened because of wave

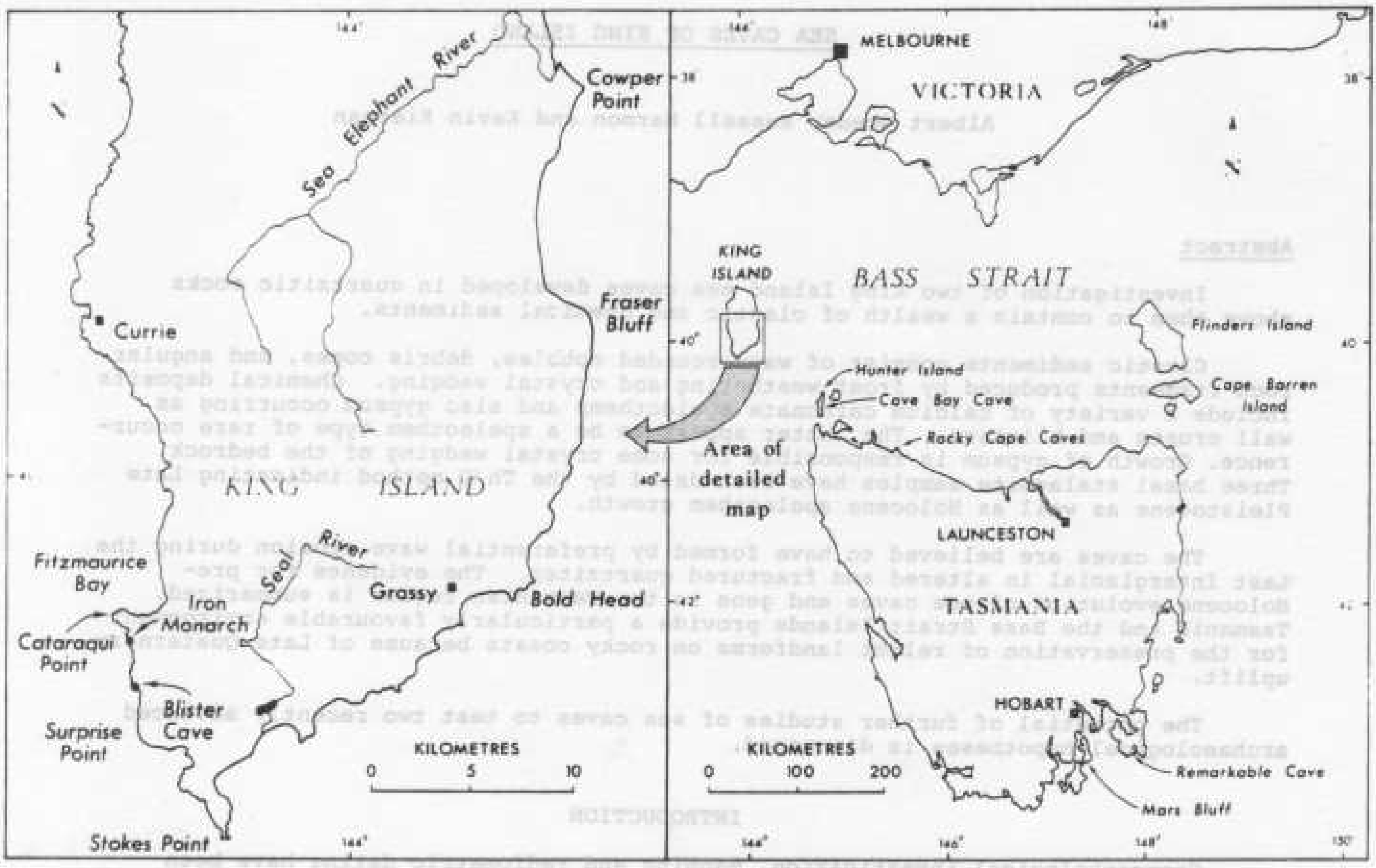

Figure 1. Locality map

reflection. This may help to account for the relict features described by him such as raised marine platforms, hanging coves, raised sea caves and slope breccias - deposits of angular rock fragments set in a sandy matrix and cemented either by sesquioxides or calcium carbonate. These deposits were locally observed to have accumulated on platform remnants. The breccias were regarded by Jennings as the product of subaerial weathering under cold climate conditions. Dated evidence of accelerated physical weathering at sea level during the last glacial period in nearby areas had recently been found by Bowdler (Hope, 1978) in a raised sea cave in Hunter Island while one of the present authors (Murray and Goede, 1977; Goede, Murray and Harmon, 1978) has found evidence of the same phenomenon in a limestone cave less than 40 metres above sealevel SSW of Montagu in NW Tasmania.

In places the plateau margin adjacent to the coastal cliffs is overlain by accumulations of calcareous dune sand. The problems posed by such accumulations in a cliff-top position have been examined by Jennings (1967) in relation to this and other areas. The relevance of these sands in the context of the present paper is as a probable source of calcium carbonate for the considerable development of speleothems that has taken place in the two sea caves described.

The rocks making up the coast in which the sea caves have developed are strongly folded metamorphic rocks such as phyllites, schists and quartzites of presumed (?) Precambrian age. They are intruded by granitic rocks at Cataraqui Point a short distance north of Iron Monarch. In the area of the caves the metamorphic rocks dip steeply in an ESE to SSE direction.

IRON MONARCH

The Iron Monarch is an impressive sea cave located nearly two kilometres SSE of Cataraqui Point. At the seaward end is a prominent geo which for the first 45 metres from the open sea is permanently water-filled and has an estimated width of up to ten metres (Plate 1). From here a steep-sided fissure extends inland for another 130 metres of which the first 52 metres is open to the sky leaving a final 80 metres of sea cave (Figure 2). The whole feature has developed by preferential erosion of a narrow sequence of beds dipping to the ESE at an angle of 80°. These beds appear to be more massive and less contorted than the surrounding beds. The trend of the fissure is closely controlled by the strike of the beds. The datum level for the cave surveys was taken to be the upper level of calcareous marine tube worms (*Serpula* sp.) and all survey heights are related to this datum. From the sea's edge the lowest part of the fissure is some eight metres wide and filled with large wave-rounded boulders. The fissure then narrows to between 3 and

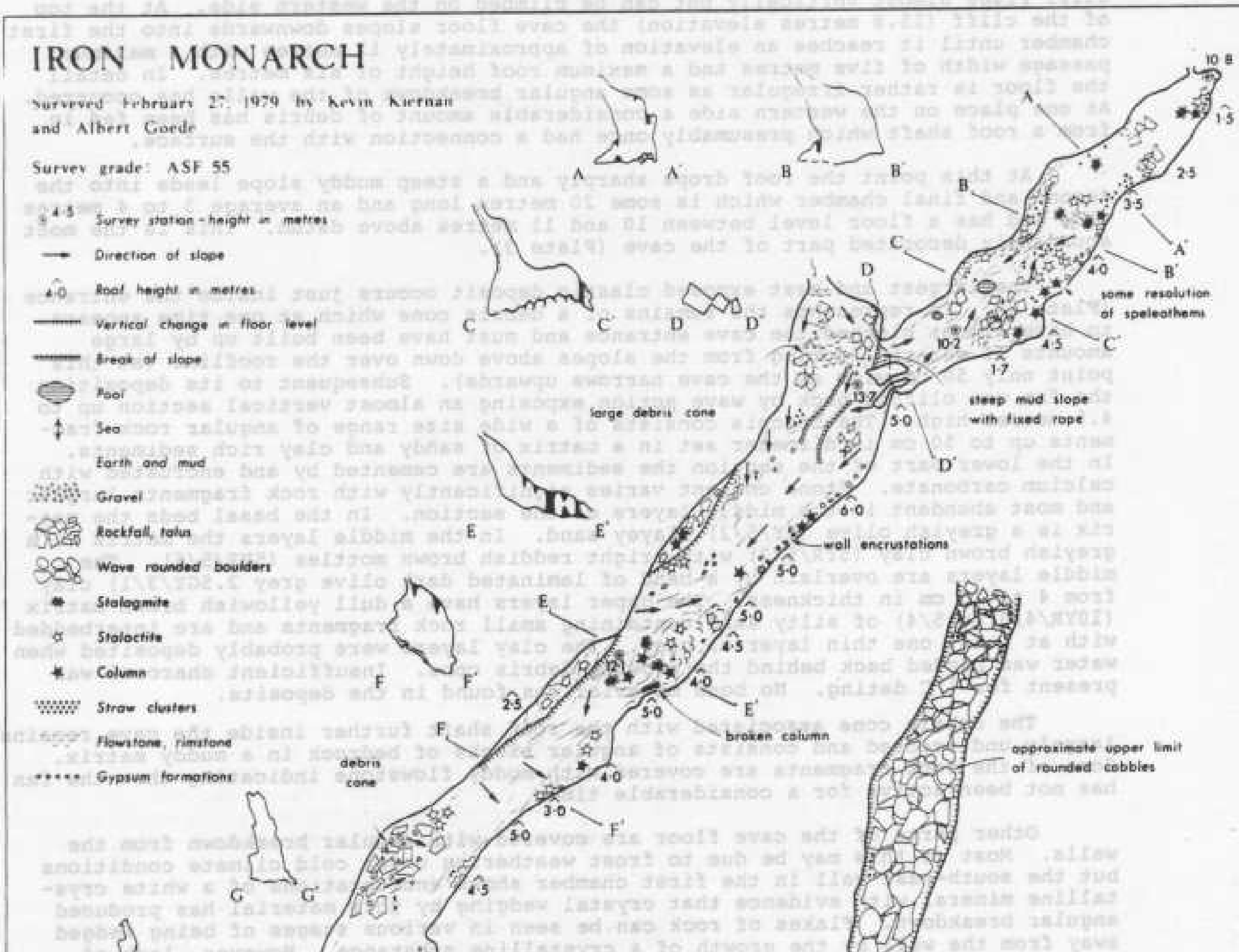

Figure 2. Survey of Iron Monarch

4 metres and angular breakdown of various sizes becomes the dominant floor material. At the roof line there is a low rampart of debris which in recent times has accumulated from the cliff above. Elevation of the crest is 11.7 metres. Ponding of water behind the rampart has formed a shallow muddy pool.

Immediately behind rises a wave eroded cliff of well compacted breccia containing blocks up to 50 cm in diameter in a fine-grained matrix (Plate 2). The cliff rises almost vertically but can be climbed on the western side. At the top of the cliff (15.8 metres elevation) the cave floor slopes downwards into the first chamber until it reaches an elevation of approximately 12 metres with a maximum passage width of five metres and a maximum roof height of six metres. In detail the floor is rather irregular as some angular breakdown of the walls has occurred. At one place on the western side a considerable amount of debris has been fed in from a roof shaft which presumably once had a connection with the surface.

At this point the roof drops sharply and a steep muddy slope leads into the second and final chamber which is some 20 metres long and an average 3 to 4 metres wide and has a floor level between 10 and 11 metres above datum. This is the most abundantly decorated part of the cave (Plate 3).

The largest and best exposed clastic deposit occurs just inside the entrance (Plate 2). It represents the remains of a debris cone which at one time appears to have almost blocked the cave entrance and must have been built up by large amounts of material coming from the slopes above down over the roofline (at this point only 50 cm wide as the cave narrows upwards). Subsequent to its deposition the fan was cliffed back by wave action exposing an almost vertical section up to 4.5 metres high. The breccia consists of a wide size range of angular rock fragments up to 50 cm in diameter set in a matrix of sandy and clay rich sediments. In the lower part of the section the sediments are cemented by and encrusted with calcium carbonate. Stone content varies significantly with rock fragments largest and most abundant in the middle layers of the section. In the basal beds the matrix is a greyish olive (5Y/5/2) clayey sand. In the middle layers the matrix is a greyish brown clay (5YR/5/2) with bright reddish brown mottles (5YR/5/6). The middle layers are overlain by a band of laminated dark olive grey 2.5GY/3/1) clay from 4 to 10 cm in thickness. The upper layers have a dull yellowish brown matrix (10YR/4/3 to 5/4) of silty sand containing small rock fragments and are interbedded with at least one thin layer of clay. The clay layers were probably deposited when water was ponded back behind the growing debris cone. Insufficient charcoal was present for ^{14}C dating. No bone material was found in the deposits.

The debris cone associated with the roof shaft further inside the cave remains largely undissected and consists of angular blocks of bedrock in a muddy matrix. Some of the rock fragments are covered with muddy flowstone indicating that the fan has not been active for a considerable time.

Other parts of the cave floor are covered with angular breakdown from the walls. Most of this may be due to frost weathering under cold climate conditions but the south-east wall in the first chamber shows encrustations of a white crystalline mineral with evidence that crystal wedging by this material has produced angular breakdown. Flakes of rock can be seen in various stages of being wedged away from the wall by the growth of a crystalline substance. However, lack of recent accumulation of angular material on the floor at this site indicates that the process is not currently active. X-ray diffraction analysis by R. Ford (pers. comm.) has shown this mineral to be gypsum. The ability of gypsum to produce angular breakdown by crystal wedging is well known.

The original floor of the cave is nowhere exposed and no wave-rounded material was found. The survey indicates that this floor could not have been higher than 10 metres above datum at the inland extremity of the cave.

The cave is particularly well decorated and has suffered little vandalism. The first chamber contains the most massive speleothems. The largest columns in the cave are found here with the tallest being over two metres in height (Plate 4). They appear to be composed of very impure calcium carbonate with iron oxides as an important impurity. Some of the columns show definite signs of being corroded at present. Impure flowstone also occurs and near the entrance is covered with a green algal mat.

The second chamber is virtually in the dark zone because of the small steeply sloping opening which connects it to the first. The formations appear to be much purer with a dense crystalline structure. Stalactites (including straws), stalagmites , flowstone and small rimstone pools are present and active deposition is taking place (Plate 3). Colours vary from white to orange brown and dark brown. Some re-solution of speleothems was observed at the far end of the chamber. A small stalagmite which had been broken from the floor was collected for possible Th/U radiometric analysis to determine its age and a sample of the wall bedrock was removed for petrological analysis.

BLISTER CAVE

This cave is briefly described by Jennings (1956) together with another smaller cave nearby with two entrances and at a lower elevation. The smaller cave was visited and roughly sketched.

The entrance to Blister Cave is located in the SE corner of a natural amphitheatre surrounded by vertical cliffs. The floor of the amphitheatre consists of an accumulation of angular rock debris of various sizes sloping down towards the sea at an average angle of 20°. Below the cave entrance, which is quite low, is a wave cliffed debris cone with an exposed face 3 metres high and rather similar in appearance to the one found inside the entrance to Iron Monarch (Plate 5). The base of the cliff is at an elevation of approximately 18.5 metres with the entrance floor at 21.5 metres. The debris cone does not appear to be subject to wave attack at the present time. Angular fragments of bedrock up to 60 cm long are imbedded in a matrix of yellow brown to reddish brown earth. No charcoal or bone was found in the sediments. The inner side of the cone slopes down into the cave at approximately 35°. The cave has developed parallel to the strike of the steeply dipping beds and is aligned on a bearing of 61° true north (Figure 3). Beyond the foot of the slope the floor is covered with angular rock fragments and again there is evidence that some of it is due to crystal wedging by gypsum - a process inactive at present. The length of the cave is only 40 metres with a bifurcation of the passage at the landward end. Wave rounded boulders, the largest 50 cm in diameter, are found on the floor of the cave near the far end where the altitude is approximately 14.4 metres. Some are capped with flowstone, one with a 20 cm high stalagmite.

Jennings (1956) described the cave as well decorated but most of the speleothems have since been removed. One column approximately 20 cm in diameter and 50 cm high remains. Two small stalagmites, already broken or disturbed, were removed for Th/U dating and a bedrock sample was collected for petrological analysis.

The most outstanding remaining decoration consists of blisters of white crystalline material on part of the ceiling and upper walls (Plate 6). The largest of these are approximately 10 cm in diameter and where broken can be seen to consist of a thin crust (2 to 4 mm) surrounding a hollow centre (Plate 7). They were also observed by Jennings (1956) who believed them to be calcite but X-ray diffraction analysis of a sample shows them to be gypsum (Ford, pers. comm.). This mineral is also responsible for crystal wedging of bedrock in both Blister Cave and Iron Monarch.

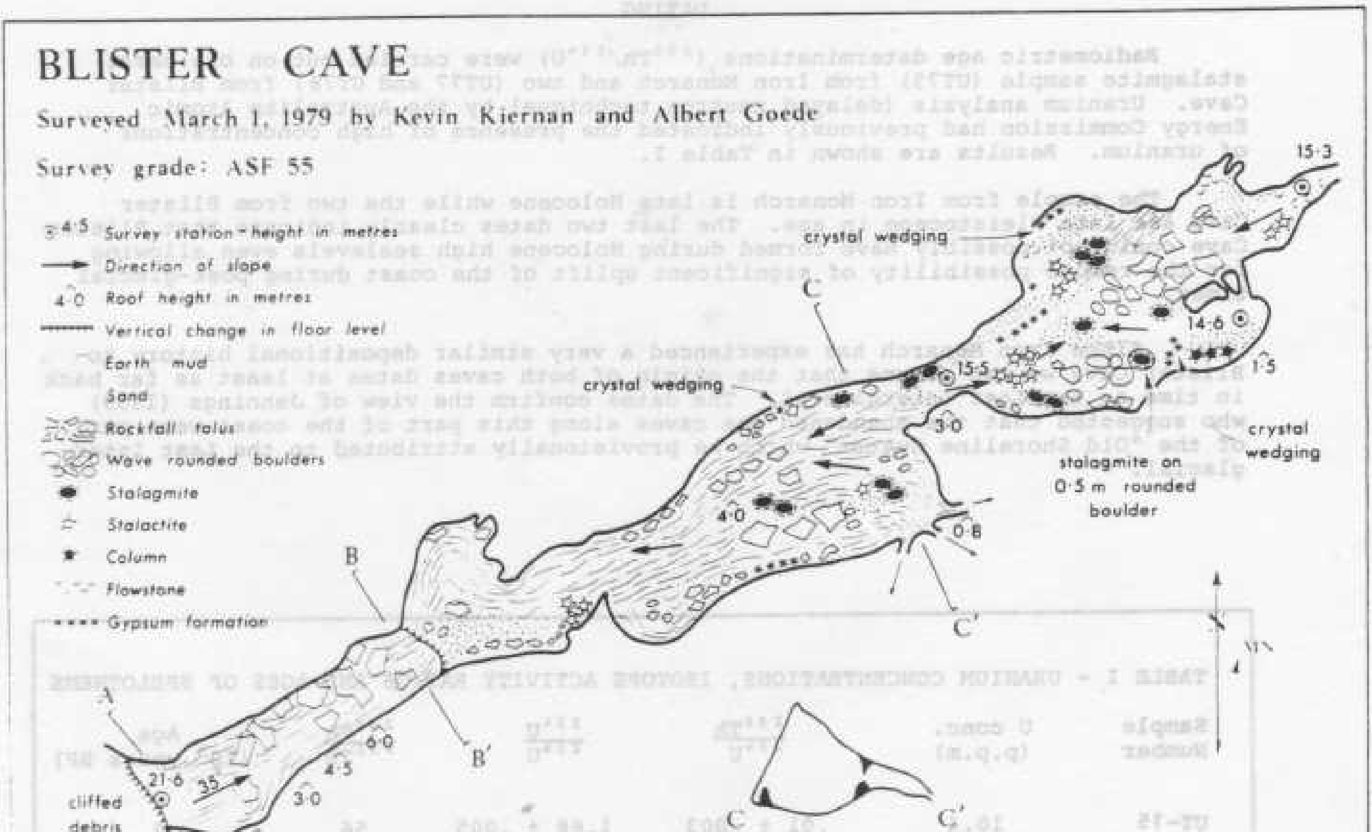

Figure 3. Survey of Blister Cave

LITHOLOGY

Iron Monarch has developed as a result of preferential wave erosion of a narrow sequence of steeply dipping massive beds of quartzite interbedded with much more highly contorted schistose beds.

A sample of wall rock collected from the cave was examined by I. Naqui (pers. comm.) who described the hand specimen as a greenish grey (5GY6/1) coloured, hard, finely crystalline rock with fractures. In thin section it is found to be a fine-grained, holocrystalline, brecciated rock composed of quartz (70%), altered plagioclase feldspar (10%), chlorite and haematite (20%). It contains numerous cavities parallel to the fracture planes and secondary silicification is common. It is quartzitic rock strongly affected by the intrusion of the Cataraqui Point granite. The granite contact occurs a short distance to the north.

The preferential erosion of quartzite is unusual and appears to be due to strong fracturing and development of cavities parallel to the fracture planes facilitating both physical and chemical weathering. The fractures also provide ready made routes for meteoric water percolating from the surface and probably account for the relative abundance of speleothems.

Blister Cave is located in a sequence of steeply dipping quartzite beds. A hand specimen from inside the cave was described by Naqui (pers. comm.) as a medium light grey (N6) coloured, hard, foliated, quartzitic rock with small cavities aligned parallel to the foliation.

In thin section it consists of interlocking quartz (70%) and some plagioclase feldspar (10%) crystals parallel to the foliation. Fractures are common both parallel to the foliation plane and at 80° to it (Plate 8). They are partly filled with chlorite, sericite (20%) and silica. The rock is a quartzite. Preferential erosion would have been encouraged by the strong fracturing of the rock and the high mica content.

DATING

Radiometric age determinations ($^{230}Th/^{234}U$) were carried out on one basal stalagmite sample (UT75) from Iron Monarch and two (UT77 and UT78) from Blister Cave. Uranium analysis (delayed neutron technique) by the Australian Atomic Energy Commission had previously indicated the presence of high concentrations of uranium. Results are shown in Table I.

The sample from Iron Monarch is late Holocene while the two from Blister Cave are late Pleistocene in age. The last two dates clearly indicate that Blister Cave could not possibly have formed during Holocene high sealevels even allowing for the remote possibility of significant uplift of the coast during post-glacial times.

Since Iron Monarch has experienced a very similar depositional history to Blister Cave we can assume that the origin of both caves dates at least as far back in time as the Last Interglacial. The dates confirm the view of Jennings (1959) who suggested that the abandoned sea caves along this part of the coast were part of the "Old Shoreline System" which he provisionally attributed to the Last Interglacial.

TABLE I - URANIUM CONCENTRATIONS, ISOTOPE ACTIVITY RATIOS AND AGES OF SPELOTHEMS

Sample Number	U conc. (p.p.m)	$\frac{^{230}Th}{^{234}U}$	$\frac{^{234}U}{^{238}U}$	$\frac{^{230}Th}{^{232}Th}$	Age (10^3 years BP)
UT-75	10.4	.01 ± .003	1.68 ± .005	56	2
UT-77	37.8	.15 ± .03	1.64 ± .01	782	17 ± 3
UT-78	8.0	.17 ± .04	1.46 ± .01	399	19 ± 4

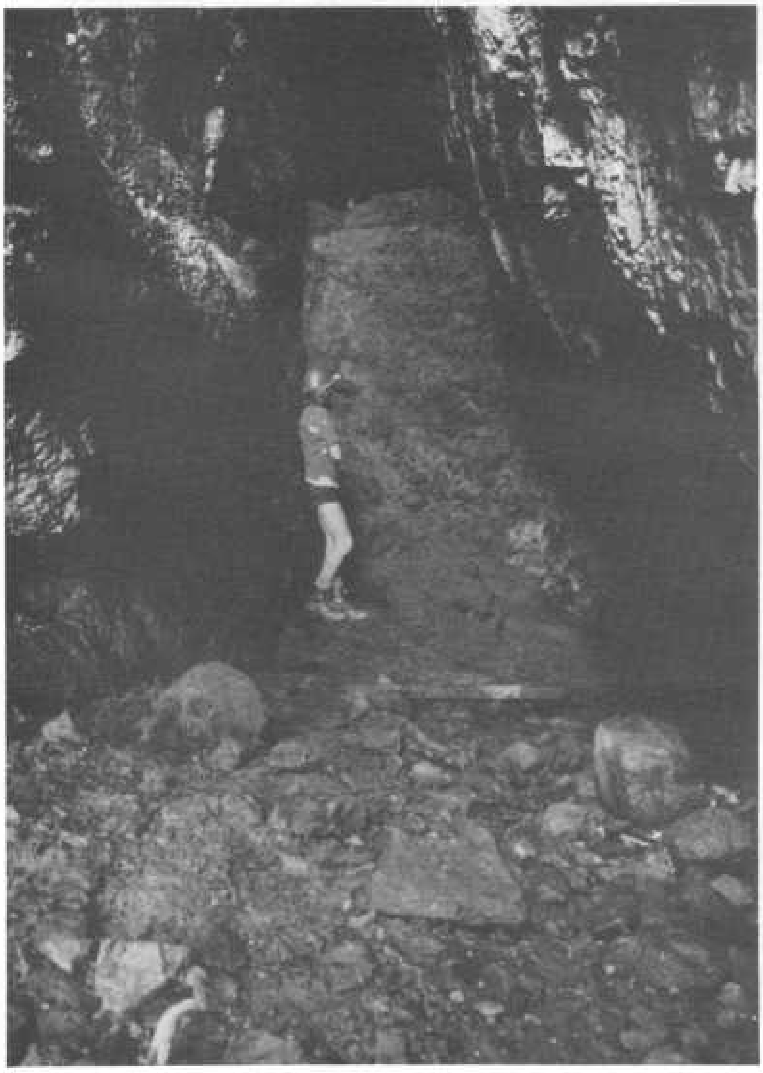

Plate 1. (above)
Geo leading to entrance of Iron Monarch viewed from cliff edge. Note figure on right.

Plate 2. (left)
Entrance to Iron Monarch with cliffed debris cone.

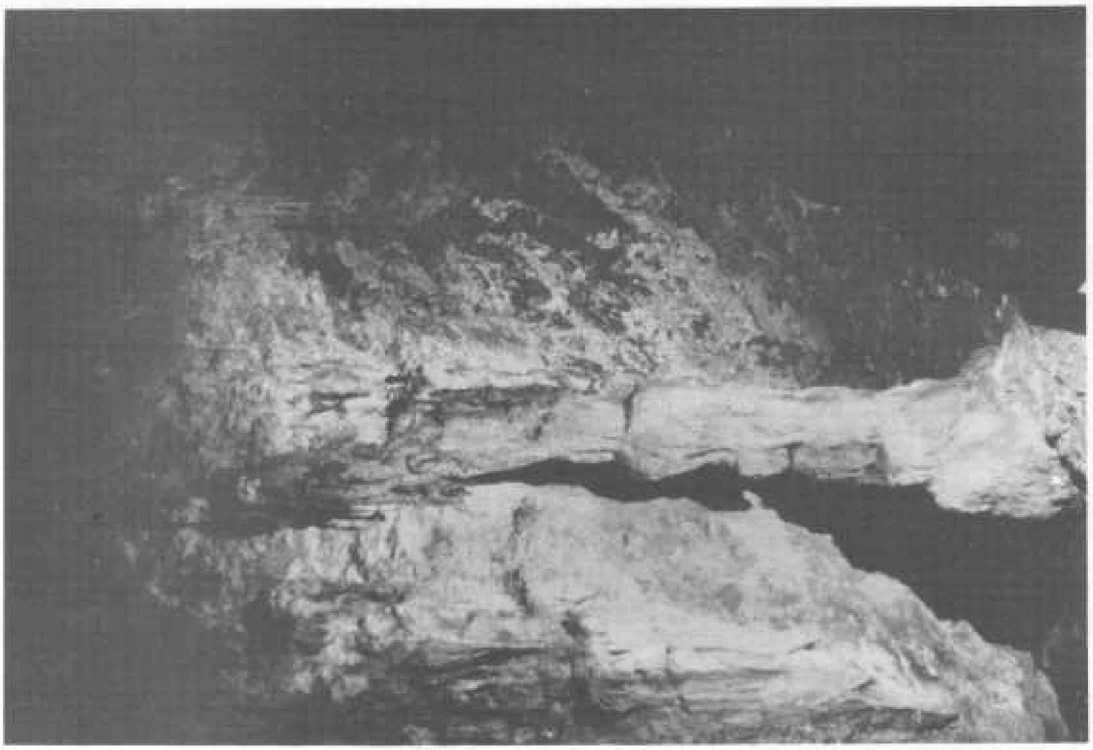

Plate 4. Large column in first chamber of Iron Monarch. Note figure for scale.

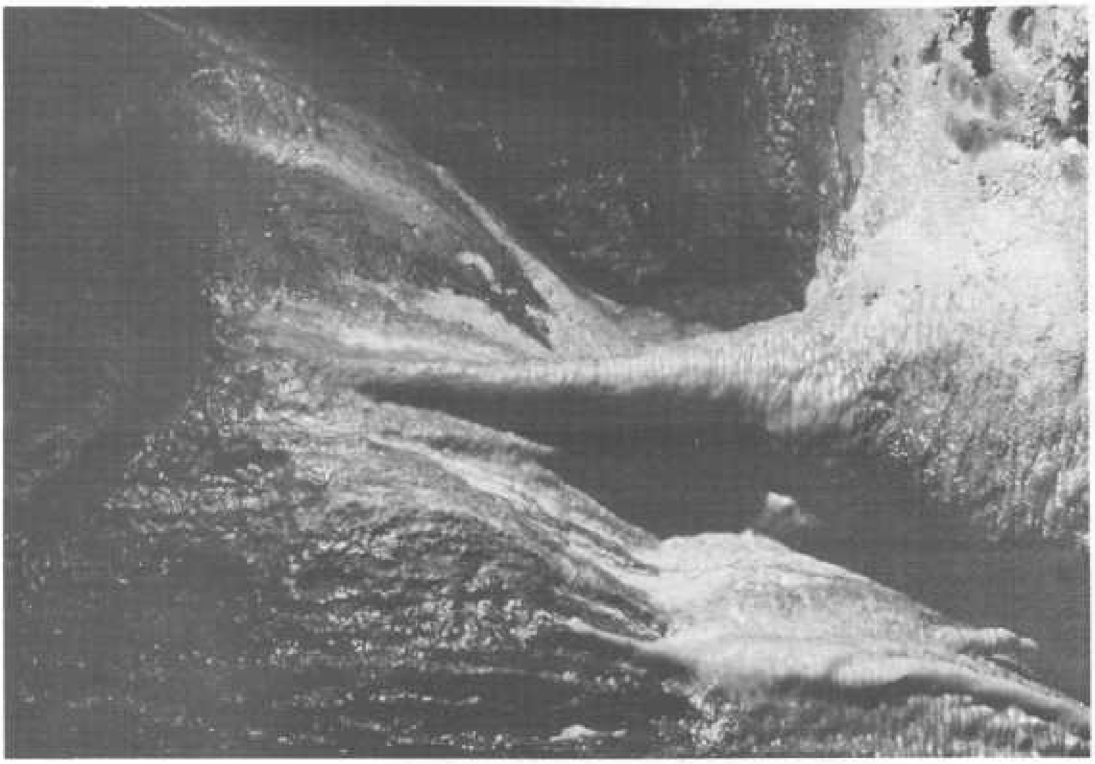

Plate 3. Decoration in second chamber of Iron Monarch.

Plate 5. Cliffed debris cone below entrance to Blister Cave

Plate 6. Gypsum blisters in Blister Cave

Plate 7. Large broken gypsum blister showing hollow interior

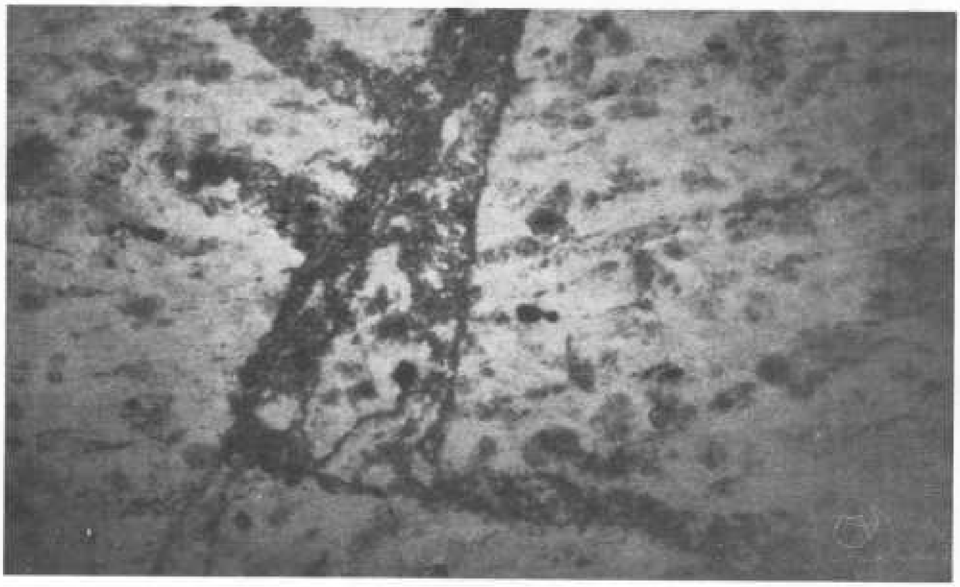

Plate 8. Quartzite bedrock in Blister Cave. Thin section under ordinary light, 30X. Note fractures. (Photo by Ikram Naqui)

EVOLUTION OF THE CAVES

The caves were formed by wave erosion, most probably during the high sealevel stands of the Last Interglacial. Their formation requires a sealevel higher than the present. Jennings (1959) concluded that the "Old Shoreline System" showed evident of a falling sequence of sealevels from 28 metres down to the present level with marked stillstands at 12 to 15 metres and 6 to 9 metres. The floor of the second chamber in Iron Monarch varies between 10 and 11 metres above sealevel, while the innermost part of Blister Cave, where wave-rounded boulders are still exposed, has an elevation of 14 to 15 metres. The inferred evolutionary history of Iron Monarch is shown in Figure 4.

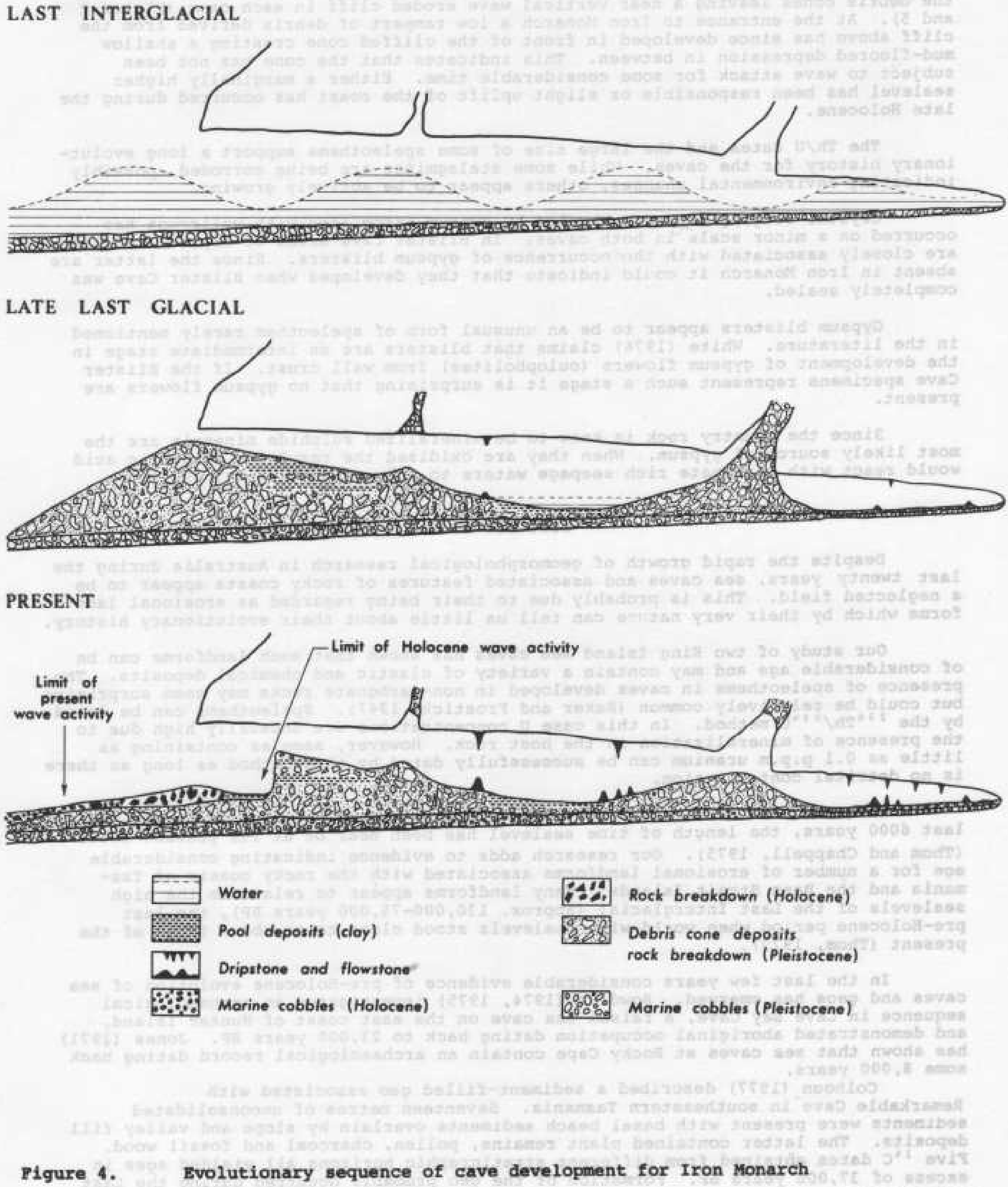

Figure 4. Evolutionary sequence of cave development for Iron Monarch

During the Last Glacial when the sea had retreated westwards to a lower level, instability of the slopes above the cliffs caused accumulation of debris cones in front of the entrances. In the case of Blister Cave it appears likely that the entrance was completely sealed. In Iron Monarch significant ponding of water occurred inside the cave as indicated by the clay layers in the sedimentary sequence. Examination of these clays for the presence of pollen and ostracods may prove rewarding.

Some physical weathering of the cave walls also appears to have taken place about this time as there are accumulations of angular breakdown which are clearly fossil in both caves. Both they and the debris cones can be expected to be contemporaneous with the slope breccias described by Jennings (1959) along that part of the coast.

In the Holocene the sealevel rose and wave erosion removed a large part of the debris cones leaving a near vertical wave eroded cliff in each case (Plates 2 and 5). At the entrance to Iron Monarch a low rampart of debris derived from the cliff above has since developed in front of the cliffed cone creating a shallow mud-floored depression in between. This indicates that the cone has not been subject to wave attack for some considerable time. Either a marginally higher sealevel has been responsible or slight uplift of the coast has occurred during the late Holocene.

The Th/U dates and the large size of some speleothems support a long evolutionary history for the caves. While some stalagmites are being corroded, possibly indicating environmental changes, others appear to be actively growing.

Crystal wedging of bedrock walls by gypsum associated with wallcrust has occurred on a minor scale in both caves. In Blister Cave areas of crystal wedging are closely associated with the occurrence of gypsum blisters. Since the latter are absent in Iron Monarch it could indicate that they developed when Blister Cave was completely sealed.

Gypsum blisters appear to be an unusual form of speleothem rarely mentioned in the literature. White (1976) claims that blisters are an intermediate stage in the development of gypsum flowers (oulopholites) from wall crust. If the Blister Cave specimens represent such a stage it is surprising that no gypsum flowers are present.

Since the country rock is know to be mineralized sulphide minerals are the most likely source of gypsum. When they are oxidized the resulting sulphuric acid would react with carbonate rich seepage waters to form gypsum (White, 1976).

DISCUSSION

Despite the rapid growth of geomorphological research in Australia during the last twenty years, sea caves and associated features of rocky coasts appear to be a neglected field. This is probably due to their being regarded as erosional landforms which by their very nature can tell us little about their evolutionary history.

Our study of two King Island sea caves has shown that such landforms can be of considerable age and may contain a variety of clastic and chemical deposits. The presence of speleothems in caves developed in non-carbonate rocks may seem surprising but could be relatively common (Baker and Frostick, 1947). Speleothems can be dated by the $^{230}Th/^{234}U$ method. In this case U concentrations are unusually high due to the presence of mineralization in the host rock. However, samples containing as little as 0.1 p.p.m uranium can be successfully dated by this method as long as there is no detrital contamination.

Holocene wave erosion of rocky coasts can only have taken place during the last 6000 years, the length of time sealevel has been near or at its present level (Thom and Chappell, 1975). Our research adds to evidence indicating considerable age for a number of erosional landforms associated with the rocky coasts of Tasmania and the Bass Strait Islands. Many landforms appear to relate to the high sealevels of the Last Interglacial (approx. 130,000-75,000 years BP), the last pre-Holocene period when world-wide sealevels stood close to or above those of the present (Thom, 1973).

In the last few years considerable evidence of pre-Holocene evolution of sea caves and geos has emerged. Bowdler (1974, 1975) investigated an archaeological sequence in Cave Bay Cave, a raised sea cave on the east coast of Hunter Island, and demonstrated aboriginal occupation dating back to 23,000 years BP. Jones (1971) has shown that sea caves at Rocky Cape contain an archaeological record dating back some 8,000 years.

Colhoun (1977) described a sediment-filled geo associated with Remarkable Cave in southeastern Tasmania. Seventeen metres of unconsolidated sediments were present with basal beach sediments overlain by slope and valley fill deposits. The latter contained plant remains, pollen, charcoal and fossil wood. Five ^{14}C dates obtained from different stratigraphic horizons all yielded ages in excess of 37,000 years BP. Formation of the geo probably occurred during the Last Interglacial.

Burns (1977) reported the occurrence of remnants of indurated slope deposits located in geos at Mars Bluff on Bruny Island off the southeast Tasmanian coast. He interprets them as having accumulated as a result of slope instability under cold climate conditions. He concludes that "the present cliffline is, at least in part, relic from the Last Interglacial, and possibly earlier higher sea levels."

The Tasmanian region may provide a particularly favourable environment for the preservation of Pleistocene landforms on rocky coasts as there is evidence of late Quaternary uplift. Marine deposits believed to be of Last Interglacial age occur at elevations of up to +20 metres along the northwest coast, +22 metres at Strahan on the west coast and also at Swansea and Mary Ann Bay in the southeast (Van de Geer, Colhoun and Bowden, 1979). In northeastern Tasmania similar deposits can be identified up to +32 metres (Bowden, 1978).

This should be compared with evidence that on stable coasts maximum heights of Last Interglacial sea levels reached only +2 to +8 metres, both within Australia (coast of N.S.W.) and on other continents (Marshall and Thom, 1976).

Sea caves should recive much closer attention as highly favourable potential locations for Pleistocene aboriginal sites. Singh, Kershaw and Clark (1979) have presented tantalizing evidence from pollen analysis of greatly increased burning around Lake George, N.S.W. during the Last Interglacial compared with earlier Interglacials, as indicated by the dominance of fire adapted taxa and the greatly increased content of charcoal particles in the lake sediments. They have speculated that this may be associated with the first arrival of man in Australia. If so, man can be expected to have arrived during the Penultimate Glacial (195,000-130,000 years BP) when sealevels were low. Another question is whether he would also have reached Tasmania during the same period.

Since many sea caves appear to have formed during the high sealevels of the Last Interglacial (125,000-100,000 years BP) they could have provided shelter at any time during the last 100,000 years of Australian prehistory.

Sea caves also warrant close examination to test the archaeological hypothesis put forward by Bowdler (1977) that Australia was colonized by people adapted to a coastal way of life. Whenever sealevels were relatively high as for example 80,000 years ago, sea caves would have provided suitable shelter within easy reach of coastal and marine food resources. They may have done so even when sealevels were low, along coasts where the continental shelf is narrow as is the case off the west coast of King Island.

If the hypotheses of Singh et al. (1979) and Bowdler (1977) are both correct we can expect to discover archaeological sites very similar to those found in raised sea caves along the coast of Southern Africa (Nelson Bay Cave and Klasies River Mouth Caves) where Middle Stone Age occupation has been found dating back to the Last Interglacial (Klein, 1975, 1979). At present the South African sites provide the oldest evidence for regular exploitation of marine resources anywhere in the world (Klein, 1975).

Sea caves, even where developed in non-carbonate rocks, frequently provide suitable preservational environments for shell and bone because of the absence of leaching by percolating water which rapidly destroys such archaeological materials at surface sites. As such they have the potential to provide most favourable sites to test one or both of the above archaeological hypotheses. Where deposition of carbonate minerals has occurred they also offer the opportunity of dating very early human occupation sites by means of Th/U analysis.

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MORPHOLOGY AND ORIGIN OF HOLY JUMP LAVA CAVE, SOUTH-EASTERN QUEENSLAND

John A. Webb

Abstract

Holy Jump Lava Cave consists of portions of lava tubes in two superimposed flows. The upper tube probably connected with the downflow section of the lower tube via a lavafall. A small upflow part of the lower tube is also preserved, and shows the original wall and roof structures. Elsewhere the cave has suffered extensive breakdown, and only small sections of the original walls are still present. The cave has been further modified by secondary silica mineralization, fine sed-ment deposition, and guano accumulation. The enclosing lava flows are early Miocene basalts of the Main Range Volcanics, making Holy Jump Lava Cave one of the oldest lava tube caves known.

INTRODUCTION

Lava caves occur commonly throughout the world, and in Australia large, well-developed examples are found in western Victoria (Ollier and Brown, 1965; Joyce, 1977) and north Queensland (Atkinson, Griffin and Stephenson, 1977). Small isolated lava caves are known from at least five other areas in Australia: in Tasmania near Mole Creek (Anonymous, 1970), in New South Wales at Mudgee (Osborne, 1979), and in Queensland at Coalstoun Lakes (Willmott, 1976), Bunya Mountains (Graham, 1971), and near Warwick. The last-mentioned, known as Holy Jump Lava Cave, is the subject of this paper.

LOCATION AND LOCAL GEOLOGY

Holy Jump Lava Cave is situated about 25 km east of Warwick, south-east Queensland (Figure 1), at grid reference 293804 on the Warwick 1:100 000 Sheet. The entrance, at an altitude of approx. 880 m, is 25 m below the crest of the ridge forming the southwest extension of Mt. Bauer. Fairly dense scrub covers the hillside and makes the entrance difficult to find, although local farmers have known of its existence for some time. "Holy Jump" is an old name for Mt. Bauer, and probably refers to an aerial cableway which once ran from near the top of the mountain to its base.

The cave is in flat-lying olivine basalt of the Main Range Volcanics, a thick sequence of basalts and occasional trachytes which makes up the Great Dividing Range from the Queensland-New South Wales border north to the Bunya Mountains (Stevens, 1965). Other, much smaller lava caves have been found in this formation, near the summits of Pinnacle Rock and Mt. Cordeaux (N. Stevens, pers. comm.; see Figure 1), and in the Bunya Mountains (Graham, 1971). All caves in the Warwick area occur near the top of the volcanic sequence, at much the same elevation (850-1 000 m).

Webb, Stevens and McDougall (1967) isotopically date a series of basalts from the Main Range Volcanics at Mt. Mitchell (see Figure 1), and obtained ages of 22-23 m.y. (early Miocene) for flows of about the same stratigraphic position as those containing Holy Jump Lava Cave. This is notable in that the majority of known lava caves are Pleistocene or younger, i.e. less than 2 m.y. old.

CAVE MORPHOLOGY

The general morphology of the cave, which has a total length of about 60 m, is shown in Figure 2. The entrance opens into a small chamber, and a tight 2 m drop at the western margin of this leads to a low-level passage which narrows and closes off after 15 m. A low crawl continues straight ahead from the first chamber into a second slightly larger one. The floor of the cave so far consists of angular basalt fragments, ranging in size from pebbles to boulders, and is thinly covered with fresh bat guano in places (Plate 1A,B). Beneath the surface of the guano a mixture of brown clay, small plates of gypsum, and white clayey taranakite has formed. Taranakite is a common phosphate mineral derived from the interaction of bat guano and clays (White, 1976).

The roof throughout the cave so far is flat or slightly arched (Plate 1A,B), and in spots displays a well-developed ropy pahoehoe surface representing the base of the overlying flow. At two places (α and β in Figure 2) the walls show an almost complete section through the basalt in which the cave has formed. Using concentrationsof vesicles and presence of pahoehoe structures, the surfaces of two flows can be recognized; one at roof level in the main part of the cave, the other 2.6 - 2.7 m below this, at roof height in the low-level passage off the entrance

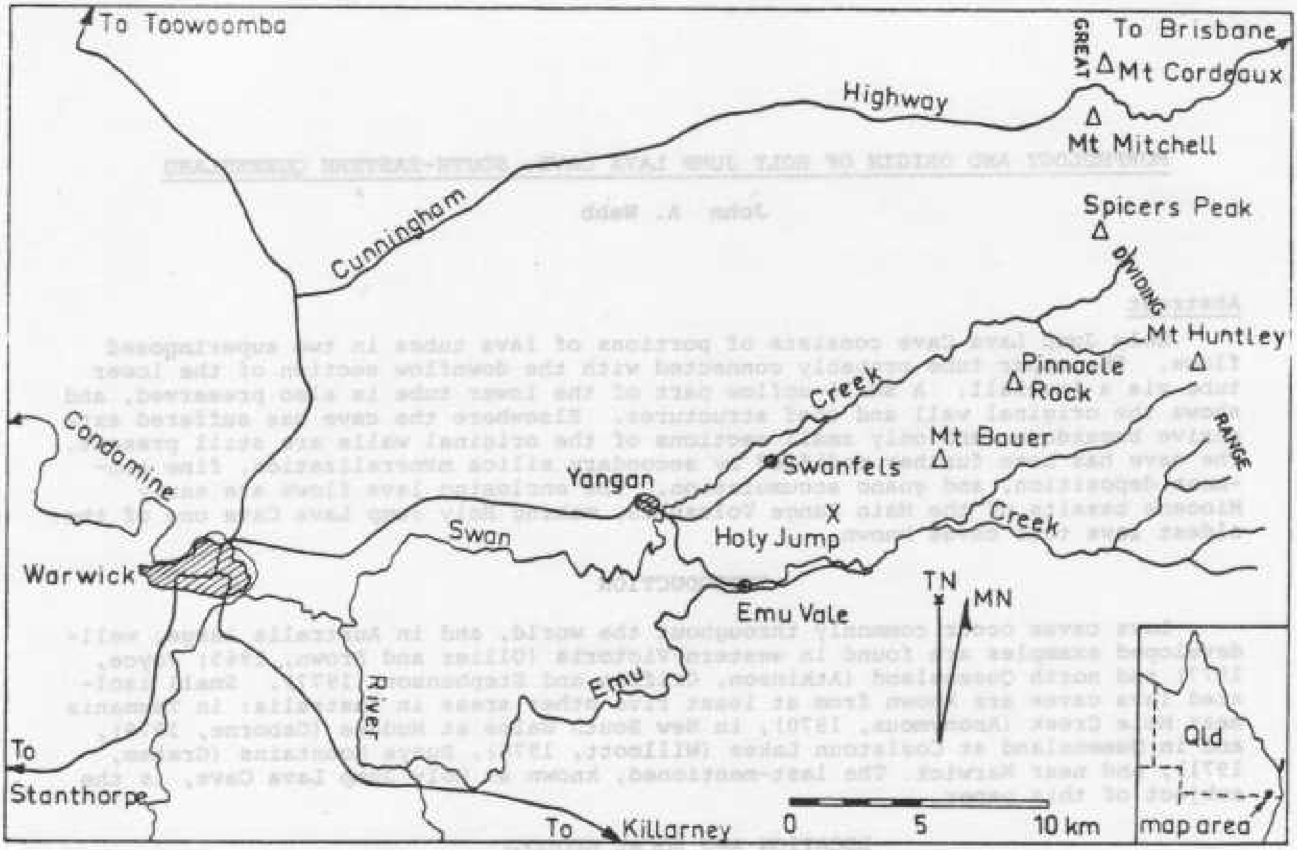

Figure 1. Location of Holy Jump Lava Cave (marked by cross).

chamber (see Figure 2). Neither flow has been affected by weathering or erosion.

Vesicles in the basalt are quite common, particularly near the tops of flows, and vary considerably in size. Small ones may be lined with a zeolite, probably chabazite. The larger vesicles are up to 50 cm in lateral extent and 10 cm high, and have treacly lava stalactites hanging from their roofs but largely smooth floors (Plate 2B). They are most frequent in the top 40 cm of the upper flow in the cave, and often form two distinct horizions 10-20 cm apart.

A squeeze down through the boulder pile on the eastern side of the second chamber leads to a very low, short passage about 1.5 m wide (Figure 2), which has a flat clay floor composed of approx. 40 cm of laminated sediment (Plate 1D). Just in front of the large fallen boulder which ends the passage is a small dome in the roof. Below this the laminated mud has been deposited around a mound of unstratified greyish-brown sediment, probably representing weathered basalt from the ceiling. Thus the stratified clay was apparently laid down some time after the formation of the cave. Similar finely laminated sediments have been recorded from both lava and limestone caves (e.g. Ollier, 1963; Bull, 1977), and were deposited in standing bodies of water.

Most of the roof in the mud-floored part of the cave is smoothly arched and has a glazed surface, with shiny lava stalactites hanging from it (Plate 1C). Elsewhere in the cave (near α and β in Figure 2), small sections of the wall have a similar appearance but with vertical ribs 1-2 cm wide (Plate 2A), some of which end in lava stalactites. At β this surface is almost obscured by rockfall, but it appears to represent the former wall of the upper level of the cave, running more or less at right angles to the mud-floored section. Near α, in the lower level of Holy Jump Lava Cave, a glazed wall with well-developed vertical ridges has been partially covered by a lava flow (Plate 2C,D). Nevertheless it shows that the original wall of this part of the cave ran approximately parallel to the present passage.

SECONDARY SILICA MINERALIZATION

A variety of silica deposits occurs on the walls of the mud-floored section of the cave and around its entrance. They include white amorphous vein infillings, opaque white or transparent botryoidal crusts, translucent stalactitic clumps and radiating masses of white acicular crystals (Plate 1E). X-ray diffraction analyses showed that the vein infillings and some of the stalactitic masses and acicular crystals are composed of chalcedony. It seems probable that the crystals represent pseudomorphs of silica after a zeolite or perhaps aragonite. One cluster of short stalactites and needles radiating from a common base proved on analysis to be made of opal-CT, one of the three structural subdivisions of opaline silica proposed by Jones and Segnit (1971). The botryoidal crusts were found to be opal-A, another of Jones and Segnit's categories. Opal-A includes hyalite, which was the hand specimen identification of the crusts.

Elsewhere in the cave secondary mineralization is restricted to occasional thin coatings of creamy flowstone on the walls and roof; some of these may still be actively forming. Although resembling calcite flowstone, the coatings proved on X-ray diffraction analysis to be composed of opal-CT.

Silica speleothems are common in lava caves (Hill, 1976), and the source of the silica is most likely the basalt itself. Hot fluids present during the eruption of the lava can hydrate feldspars, pyroxenes and volcanic glass and so form silica; this can be precipitated immediately or deposited later by the action of ground water (Swartzlow and Keller, 1937).

MODE OF FORMATION OF THE CAVE

Several kinds of caves can occur in lava flows: cavities beneath pressure ridges and spatter cones, vent and fissure caves, blister caves and lava tubes (Wood, 1976b). Only the last two are relevant to Holy Jump Lava Cave.

Blister Caves

Blister caves develop from the lifting of a lava sheet by pockets of gas trapped in the flow (Wood, 1976b). Although they may be extensive, most are less than 10 cm in height and 1 m in lateral extent, and grade into large vesicles.

Gas bubbles can accumulate beneath the crust of an active flow, particularly if the crust is thickening slowly and vesiculation is rapid (Nichols, 1946). Build-up of gas is fastest where the lava surface is domed, and sufficient pressure may develop to uplift the crust and form cavities. After the crust has thickened to include these, further vesicle accumulations may occur lower in the flow, causing superposed cavities to develop, the upper ones being the older. If the lava was fluid enough, the floors of the cavities would be flat and the roofs festooned with treacly lava stalactites.

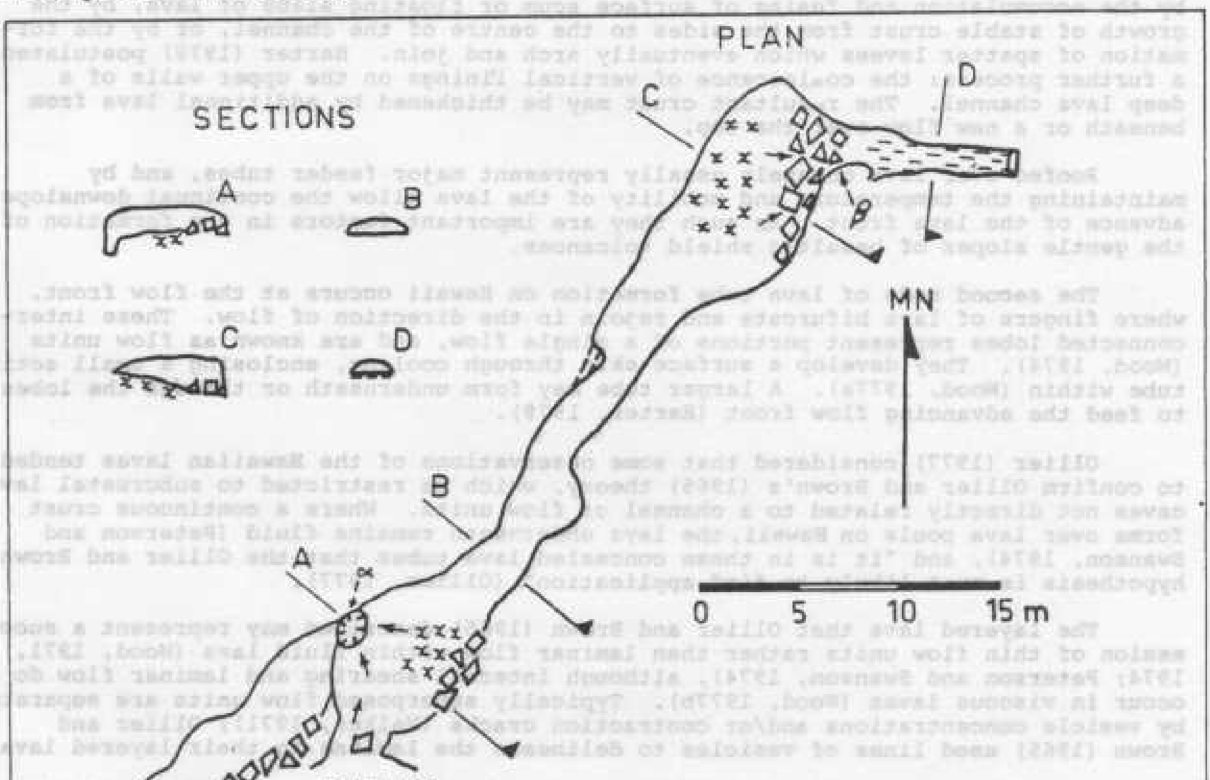

Figure 2. Map of Holy Jump Lava Cave. Unless designated otherwise, floor consists of angular pebbles and cobbles of basalt; points α and β referred to in text. Surveyed by K. Sweatman, H. Shannon and A. Watt in June, 1970, using compass, clinometer and tape.

Small caves showing these features have been described from several localities (e.g. see Nichols, 1946; Ollier and Brown, 1965), and closely resemble the larger vesicles in Holy Jump Lava Cave (Plate 2B). All probably formed in the manner outlined above, i.e. vesicle accumulation beneath the crust of a relatively fluid flow.

Lava tubes - general

Lava tubes are internal conduit systems within a lava flow. Before relating them to Holy Jump Lava Cave, it is necessary to discuss briefly their mode of formation and general morphology.

Ollier and Brown (1965) suggested that laminar flow within fluid lava produces layering, and the more liquid lava concentrated between the laminae becomes segregated as tubes. These are continuing sources of heat which can erode the surrounding lava flowing through virtually solid rock. This results in lava tubes formed beneath a continuous flat crust (Ollier, 1977).

This hypothesis has been used by several workers (e.g. Greeley, 1971a; Greeley and Hyde, 1972), and Hatheway (1977) proposed a modified version to account for long lava tubes. However, studies of actively forming lava tubes on Hawaii (Wentworth and McDonald, 1953 ; Greeley, 1971b, 1972; Cruikshank and Wood, 1972; Peterson and Swanson, 1974) have shown that they form in two ways, neither of which is directly related to Ollier and Brown's theory.

The first involves roofing of open lava channels. This can be accomplished by the accumulation and fusing of surface scum or floating slabs of lava, by the growth of stable crust from the sides to the centre of the channel, or by the formation of spatter levees which eventually arch and join. Harter (1978) postulated a further process: the coalescence of vertical linings on the upper walls of a deep lava channel. The resultant crust may be thickened by additional lava from beneath or a new flow over the top.

Roofed-over lava channels usually represent major feeder tubes, and by maintaining the temperature and mobility of the lava allow the continual downslope advance of the lava front. As such they are important factors in the formation of the gentle slopes of basaltic shield volcanoes.

The second mode of lava tube formation on Hawaii occurs at the flow front, where fingers of lava bifurcate and rejoin in the direction of flow. These interconnected lobes represent portions of a single flow, and are known as flow units (Wood, 1974). They develop a surface skin through cooling, enclosing a small active tube within (Wood, 1977a). A larger tube may form underneath or through the lobes to feed the advancing flow front (Harter, 1978).

Ollier (1977) considered that some observations of the Hawaiian lavas tended to confirm Ollier and Brown's (1965) theory, which he restricted to subcrustal lava caves not directly related to a channel or flow units. Where a continuous crust forms over lava pools on Hawaii, the lava underneath remains fluid (Peterson and Swanson, 1974), and "it is in these concealed lava tubes that the Ollier and Brown hypothesis is most likely to find application" (Ollier, 1977).

The layered lava that Ollier and Brown (1965) described may represent a succession of thin flow units rather than laminar flow within fluid lava (Wood, 1971, 1974; Peterson and Swanson, 1974), although internal shearing and laminar flow do occur in viscous lavas (Wood, 1977b). Typically superposed flow units are separated by vesicle concentrations and/or contraction cracks (Walker, 1971); Ollier and Brown (1965) used lines of vesicles to delineate the laminae in their layered lava.

Lava tube caves show a bewildering diversity, both in length and complexity. They vary in length from tens of metres to over 11 km (Simons, 1976), and Atkinson, Griffin and Stephenson (1977) reported a lava tube system that may have extended for 90 km. In morphology lava tubes range from simple conduits to braided networks of multiple level passages (e.g. Halliday, 1972). The diversity is related to the slope and topography of the land surface beneath the flow, the rate of discharge of the lava and its total amount, and the viscosity of the magma. The last-mentioned factor depends on the chemical composition, gaseous content, solid load and temperature of the lava (Knutson, 1977; Wood, 1977b). Most lava tubes form in very fluid basaltic magma, usually pahoehoe (ropy) lava, although they are found rarely in the more viscous aa (cindery) flows (Wood, 1976a). A few have been discovered in andesites (Wood, 1976b).

The general morphology of a lava tube system is also affected by erosion of the tube by the flowing lava, breakthrough from a higher to lower level (lavafall), and convergence and divergence of flow in the small tubes at the lava front (Wood, 1977a). Furthermore, only portions of a network may eventually drain of lava, either because of loss of mobility in the lava or a lack of space into which it can flow (Wood, 1977b).

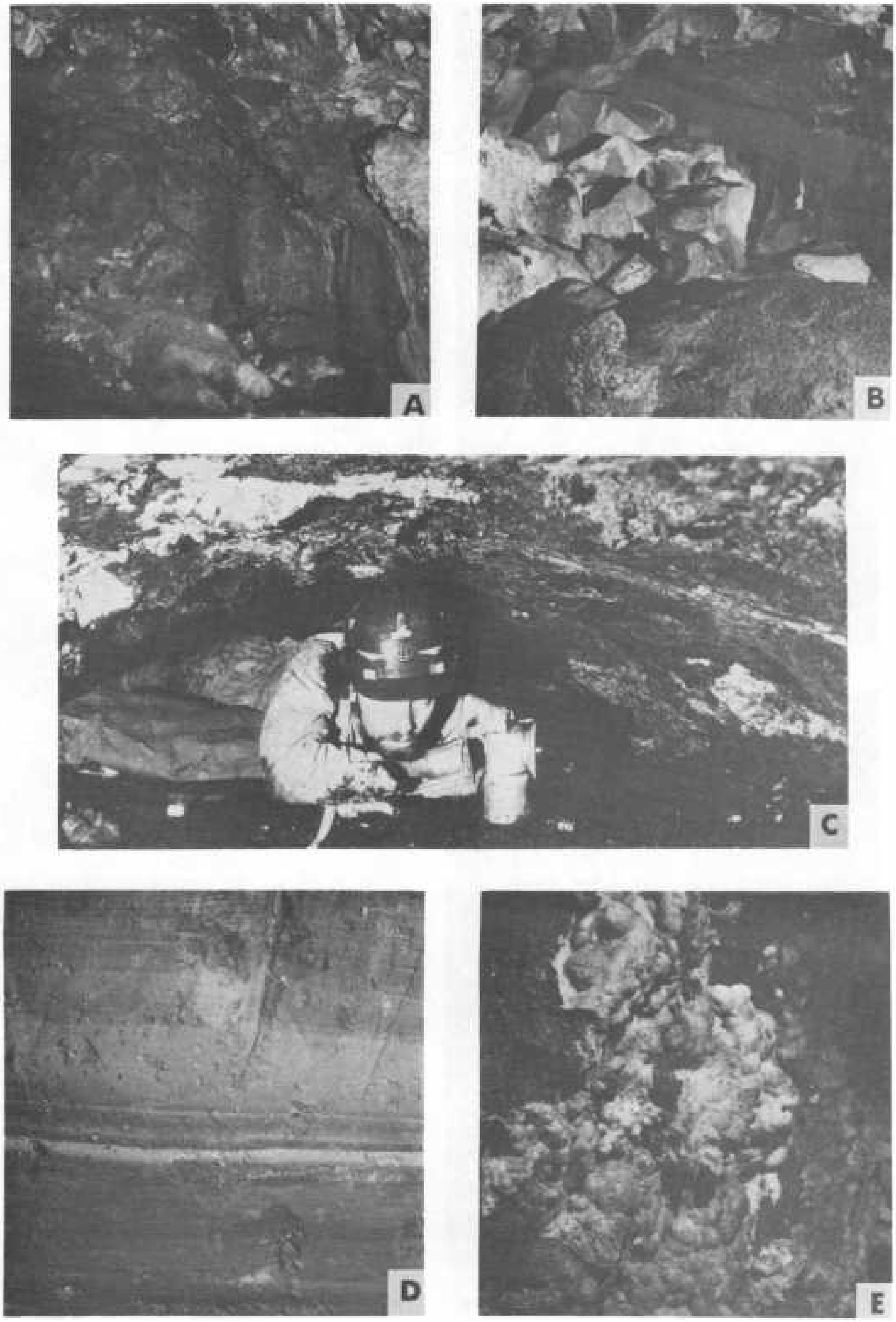

Plate 1. A: Western wall of entrance chamber, with point α to the right; roof height approximately 2 m.
B: Eastern wall of entrance chamber, with guano in foreground and rubble behind; roof height approximately 1 m.
C: Mud-floored section of cave, showing arched roof with lava stalactites and secondary silica encrustation.
D. Laminated sediment from mud-floored section of cave; narrow white layer is approximately 12 cm from top of photograph.
E. Silica deposits on wall near β; approximately 2/3 natural size.

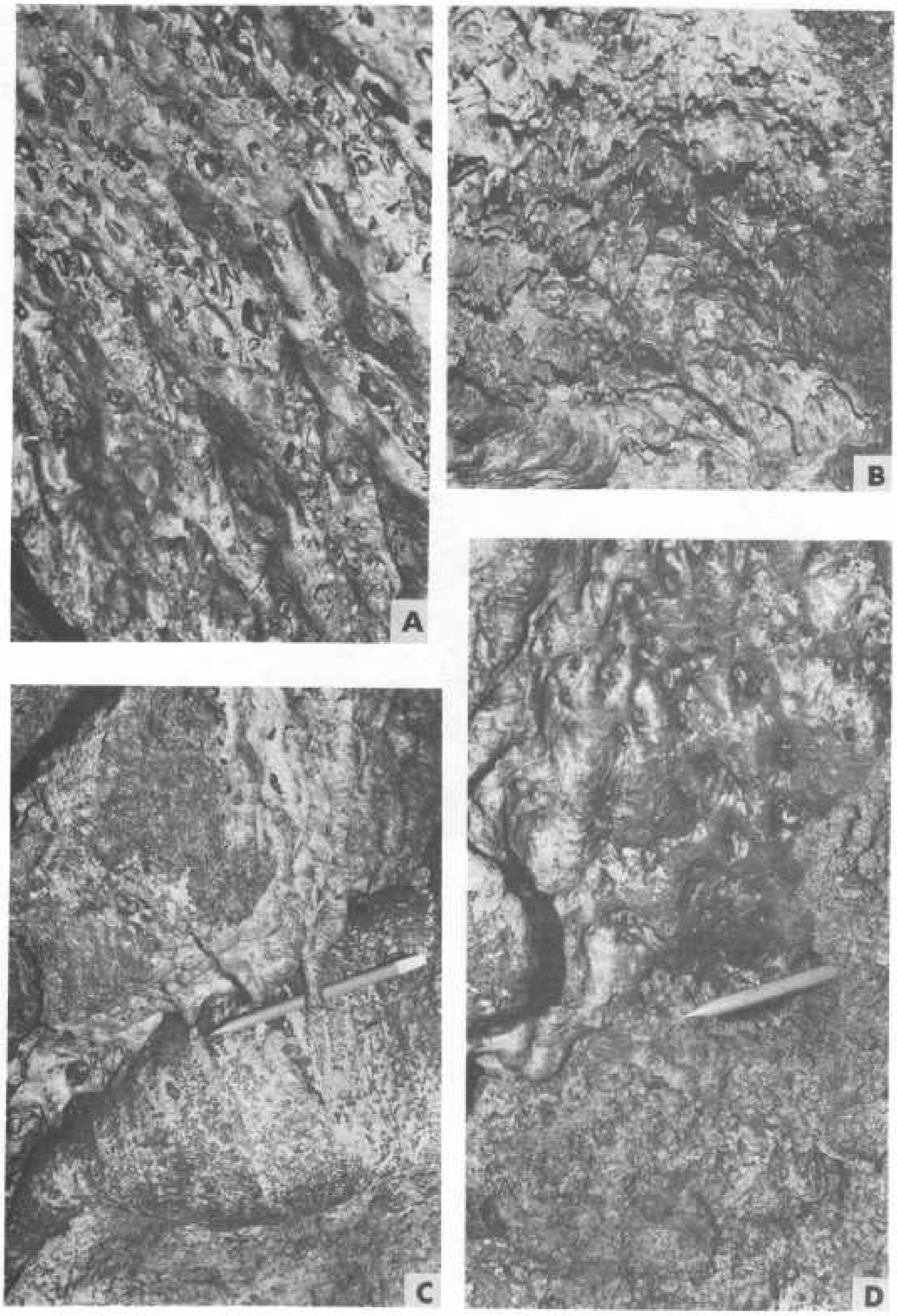

Plate 2. A: Glazed section of wall near β in figure 2, showing irregular vertical ribs, some of which have chipped to reveal vesicles underneath. Field of view about 10 cm wide.

B: Treacly lava stalactites hanging from roof of large vesicle, just beneath upper surface of upper flow in Holy Jump. Field of view approximately 30 cm.

C: Glazed section of wall in lower section of cave (near α in Figure 2), partially covered by pahoehoe lava flow, on which biro is resting.

D: As C, but photograph taken slightly to the left of that in C. Biro shows contact between glazed wall (above) and lava flow (below).

Passage shapes in lava tubes are affected by many of the above considerations, particularly volume of lava, erosion and drainage history (Wood, 1977a). During drainage the lava level may be lowered constantly or sporadically, and the accretion of cooled lava to the walls and roof can form linings, benches and shelves (Harter, 1977). If the lowering of the lava level is arrested, a crust may form over the lava stream; this may remain as a false floor when the fluid lava beneath drains away (Wood, 1976b). Rafted blocks floating on the lava flow may score the walls, or become fused to the walls or roof (Wood, 1971; Greeley and Hyde, 1972). Floors of lava tubes tend to be rough and clinkery, and may be broken into polygonal slabs (Wood, 1976b).

Due to the heat of the flowing magma and the effect of burning gases, temperatures inside lava tubes are sufficient to melt the wallrock partially and produce a black vitreous glaze (Peterson and Swanson, 1974). This can drip and form vertical ridges on the tube walls as well as delicate stalactites and stalagmites in a great variety of shapes (Wood, 1971, 1976b; Halliday, 1972). Larger stalactites, formed as a result of initial draining of the tube or from lava flung as spatter, also occur (Wood, 1976b).

Holy Jump Lava Cave - a lava tube system

The linear shape, occurrence in two superposed lava flows, and glazed roof and wall sections all indicate that Holy Jump is a lava tube cave. In particular, the mud-floored section of the cave, with its smoothly arched, glazed roof and small lava stalactites, represents a largely unmodified lava tube. Elsewhere in the cave breakdown has destroyed most of the original roof and walls, except for two small portions that are glazed and vertically ribbed (Plate 2A,C,D). The ribbing compares closely in size with that described from the walls of Icelandic lava tube caves by Wood (1971).

The two lower level sections of Holy Jump Lava Cave presumably once formed part of the same continuous tube in the lower lava flow (see Figure 2); the connecting portion has been destroyed by collapse. The upper section of the cave represents a larger tube in the overlying flow. The fissures from which the Main Range Volcanics were erupted are believed to lie along the crest of the Great Divide (Stevens, 1965), i.e. to the east of Holy Jump Lava Cave (Figure 1). As a result, the direction of lava flow within the cave was probably from north-east to south-west, although there is no direct evidence for this in the cave itself. The tube in the upper flow most likely ran directly above the lower tube for a short distance and then broke through into it via a lavafall at α on Figure 2, thereby

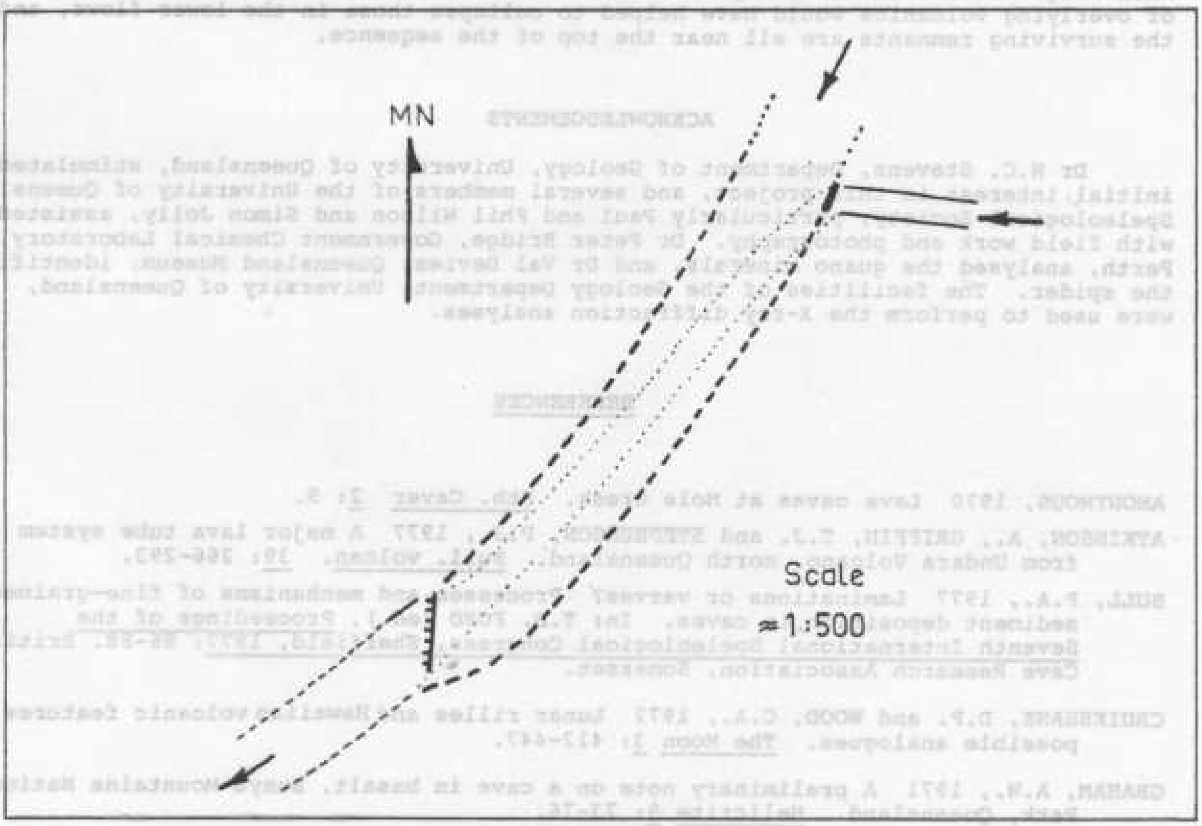

Figure 3. Diagrammatic representation of mode of formation of Holy Jump Lava Cave. Arrows indicate probable direction of lava flow; thinner lines represent lower tube and thicker lines upper tube; solid lines indicate sections of lava tube wall still preserved, dashed lines show probable courses of tubes and dotted lines show possible position of tube walls; cliff symbol represents a lavafall.

enlarging the lower tube beyond this point. This would account for the larger size of this part of the lower tube relative to the upstream mud-floored section. It would also explain the fact that the glazed wall remnant in the lower level near 3 is partially obscured by a lava flow (Plate 2C,D). The breakthrough of the upper tube into the lower would have most probably raised the lava level in the lower tube (which may have been completely drained), so covering some of the glazed wall with lava. The presumed relationships of the two tubes are shown diagrammatically in Figure 3.

Lava tube caves older than Pleistocene are rare, and Holy Jump Lava Cave is one of the oldest well-preserved examples known. The lava caves in the Bunya Mountains and Tasmania are also in Miocene basalts, but they are poorly preserved and their origin is uncertain. The basalt at Mudgee is older, Oligocene, but the caves are believed to be products of groundwater solution as they lack any lava tube features.

BREAKDOWN

Lava tubes are particularly prone to breakdown after the lava cools, because of the abundance of flow contacts, joints and partings in the parent lava. Differential cooling produces vertical joints which intersect the horizontal partings between flows, and spalling of the roof follows (Hatheway, 1977). Contacts between flows are often the principal weaknesses; the roof of a breakdown tube usually coincides with the ropy surface of a flow (Wood, 1977a). Much collapse probably occurs soon after the withdrawal of lava from a tube, but in the long term, chemical weathering caused by percolating ground water will gradually enlarge joints and cracks and finally destroy the cave altogether.

In Holy Jump Lava Cave, breakdown has severely affected the upper tube and the downstream section of the lower tube. The contact surfaces between flows provided the upper limit of collapse and gave rise to the more or less flat roofs throughout these sections of the cave (Plate 1A,B). The large vesicles present near the tops of the flows probably assisted in the breakdown.

The upstream section of the lower tube escaped destruction and remains almost intact, perhaps because of its smaller size. The entrance into it was opened by the collapse of the upper and lower tubes where the latter ran directly underneath the former.

Undoubtedly there were several lava tubes associated with the eruption of the Main Range basalts in the early Miocene, but most have been obliterated. The weight of overlying volcanics would have helped to collapse those in the lower flows, and the surviving remnants are all near the top of the sequence.

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APPENDIX : BIOLOGY OF HOLY JUMP LAVA CAVE

Two species of bats inhabit the cave. *Miniopterus schreibersii* (Eastern Bent-Winged Bat) is the most abundant, and over 300 are normally present, most of them in the second chamber. A few *Rhinolophus megaphyllus* (Eastern Horseshoe Bat) roost separately at the far end of the mud-floored section; Holy Jump Lava Cave has not previously been recorded as a roosting site for this species (see Hall, Young and Spate, 1975). In the early part of the year the *Miniopterus* may not be present, and during their absence the *Rhinolophus* are found anywhere in the cave.

The nycteribid fly *Penicillidia (Penicillidia) oceanica*, which is parasitic on *Miniopterus*, has been recorded from Holy Jump Lava Cave (Maa, 1971), and its pupae are common on the roof at bat roosting sites. Abundant mites and pseudo-scorpions can be found crawling over the guano on the floor. Spiders also occur: Gray (1973) listed *Achaeranea* sp. from the cave, and *A. extrilida* was collected from the entrance chamber in the course of this study. The latter is a surface dwelling species which occasionally lives in caves, sometimes with slight depigmentation (Gray, 1973). Roberts (1970) recorded the tick *Ixodes simplex simplex* from near the entrance of Holy Jump, and it has been subsequently collected from crevices in the roof of the second chamber (Jolly, 1979). This tick, a parasite primarily on *Miniopterus* bats, is found throughout the world but is relatively scarce (Roberts, 1970).

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INFORMATION FOR CONTRIBUTORS

SCOPE

Contributions from all fields of study related to speleology will be considered for publication. Suitable fields include Earth Sciences, Speleochemistry, Hydrology, Meteorology, Conservation, Biospeleology, History, Major Exploration (Expedition) Reports, Equipment and Techniques, Surveying and Cartography, Photography and Documentation. Comprehensive descriptive accounts of the exploration and morphology of individual caves will be welcomed, but simple trip reports and brief cave descriptions are not adequate. Papers overall should not exceed 20 printed pages in length. Contributors intending to write at greater length or requiring any advice on details of preparation are invited to correspond with the Editors. All manuscripts will be read by referees. Short 'Letters to the Editor", expressing a personal view or giving a preliminary report of interesting findings, are welcomed, and will be given preference for speedy publication.

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GRAY, M.R., 1973 Cavernicolous spiders from the Nullarbor Plain and, south-west Australia. J. Aust. ent. Soc. 12: 207-221.

VANDEL, A., 1965 Biospeleology. The Biology of the Cavernicolous Animals. Pergamon, London. Pp. xxiv, 524.

WIGLEY, T.M.L. and WOOD, I.D., 1967 Meteorology of the Nullarbor Plain caves. In: J.R. DUNKLEY and T.M.L. WIGLEY (eds), Caves of the Nullarbor. A Review of Speleological Investigations in the Nullarbor Plain. Southern Australia: 32-34. Speleological Research Council, Ltd., Sydney.

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