KARST DEVELOPMENT AND GROUNDWATER FLOW
IN THE QUATSINO FM.
NORTHERN VANCOUVER ISLAND

By

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ABSTRACT

This study attempts to determine the nature of karst developed in uplifted and intruded carbonate strata on northern Vancouver Island , Canada. Hydrochemical analyses are employed to determine the potential of the solution process imposed by geochemical constraints , and to investigate the extent to which hydrochemical character can be used to interpret the nature of groundwater flow and the spatial distribution of solution in the karst. Examination of landscape form,particularly in cave passages,is employed to further understanding of the physical constraints on groundwater flow and cave development ,and to interpret the historical record preserved there.

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This thesis was completed amid good times with the contents of the ghetto - Bob, Cecil, Deb, Emil, Joe, Simon and Rob ,who cut into his leisure time to type the manuscript.;and those of 401 -Janis, Peter, Tom and Chris. Above all I appreciate the aid,support and friendship of my best friend,Carol.

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CHAPTER 1

INTRODUCTION

1.1 Introduction

The object of this dissertation is to examine karst developed on an area of Upper Triassic limestones on northern Vancouver Island, British Columbia. The karst comprises a linear valley bottom area and an adjacent plateau. Characteristic features of the limestone are the consistent stratal dip at about 30° and the high frequency of impermeable intrusions in the unit. The dissertation contributes towards the study of the karst landscapes of Canada by providing information on a region which has been tectonically disturbed and which has a cool mesothermal climate.

The dominant erosion process in karst landform development is the mass transfer of limestone in solution. The nature of the process has been substantially quantified in theory and experiment. However, it remains difficult to define relationships between process and landform development in other than a qualitative manner in all but a few small scale situations (eg. Curl 1966). Moreover, other geomorphic processes compete with solution in the present landscape and non-karst processes (eg. glaciation) have dominated landform development in the recent past. Nevertheless, "the facts that one known process is dominant in the evolution of a karst terrain and that the manner in which this process acts upon the terrain is known in a number of simplified situations (such as sheet-flow over and pipe-flow through the rock) make the study of karst landforms one of the least inprecise in geomorphology" (Drake 1974).

Recognising the above constraints, two aspects of the solution process were studied as these were thought to provide insights of geomorphic significance. First, the controls of potential solute concentrations in waters in the region were studied as it is possible to use these to define limits to the solution process because of geochemical constraints. Second, spatial variations of certain chemical and statistical variables were employed in an attempt to determine whether these variables can yield information about the nature of groundwater flow and hence the pattern of solution.

Caves provide an historical perspective on karst development and an understanding of their sequential development provides insights into the interaction of groundwater movement and solution in a karst. The purpose of the speleological studies in this dissertation was twofold: (1) to determine the gross controls on groundwater flow and hence the spatial constraints on solution. (2) to reconstruct genetic sequences of passage development and to relate these to factors external to the cave.

This dissertation is arranged into six chapters. Chapter 2 describes the geology and the geologic history of the region. Chapter 3 describes the environmental controls on solution and the hydrochemistry of hydrologic water classes. Chapter 4 reviews aspects of the surface geomorphology of the valley bottom and the plateau. Chapter 5 considers the major controls on groundwater flow and describes the caves of the region. Chapter 6 summarises and draws conclusions from the preceding chapters.

1.2 Physiography

Karst development on northern Vancouver Island occurs primarily along 30km of the strike aligned Benson River Valley and on the adjacent Gibson Plateau in the Quatsino Fm. limestone. Relative relief in the valley exceeds 1000m but the limestone is restricted to the basal 400m. On the west side of the valley the contact with the non-karst Parson Bay Fm. is generally more than 300m above the river. However, on the east side limestone outcrops on the valley side in only two locations, both of which contain extensive cave systems. At both places the top of the limestone is less than 150m above the river. Elsewhere the eastern boundary of the limestone occurs on the valley floor in a pattern influenced by faulting.

In three locations the valley floor is uplifted into cuesta blocks which exceed 1km in length both along and normal to the strike, with a relative relief upto 400m. The Benson River and its tributary the Raging River dissect the inclined surface and function as the base level for surficial and subsurface drainage.

The Gibson Plateau is located north-west of the Benson Valley. It rises up to 900m above the surrouding base level which to the west and north is formed by two fiords, Nereutsos Inlet and the Quatsino

Narrows. The steep sided plateau reaches 7km in length along the strike and 3km normal to the strike. The surface area is approximately 15km^2.

1.3 Climate, Soil and Vegetation

The region has a cool mesothermal climate, Cfc in the Koppen classification (Strahler 1954). The mean annual temperature is about 7°C; the mean monthly temperature range is from 1°C to 14°C. Mean annual precipitation rate is about 250cm yr^{-1} but less than 50cm yr^{-1} occurs between April and September. The region has about 200 days with precipitation per year.

The regional soil type is an elluviated podzol, however, soil development in much of the karst is skeletal and influenced by local factors. The region occurs in the Pacific Coast and Interior Mid Latitude forest zone. Forest composition is dominated by Douglas Fir, Western Hemlock and Western Red Cedar. Data source - the National Atlas of Canada (1974).

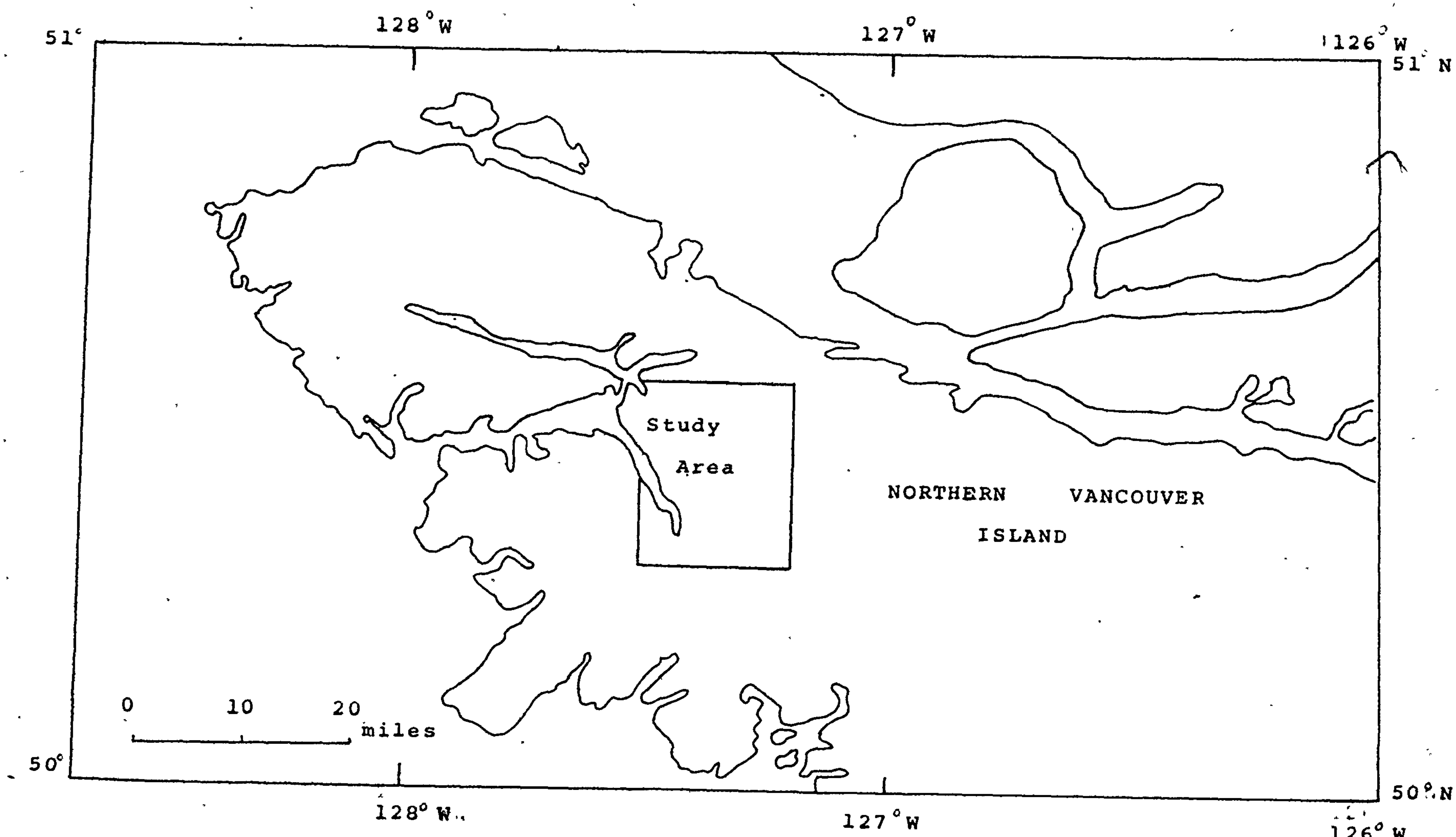


Figure 1.1 Location of the study area.

CHAPTER 2

GEOLOGY AND QUATERNARY HISTORY OF NORTHERN VANCOUVER ISLAND

2.1 Introduction

The purpose of this chapter is to outline the geology of Northern Vancouver Island. Stratigraphic and lithologic descriptions are followed by an interpretation of the structural geology and geological history of the region. The account is based largely on the recent synthesis of Muller, Northcote and Carlisle (1974). Particular emphasis is placed on the geology of the Benson Valley (Figure 2.1) as it comprises the host rock for the karst development described in subsequent chapters. Aspects of the Quaternary history are introduced in the final section.

2.2 Stratigraphy

2.2.1 Late Paleozoic

Sicker Group

The basal rock sequence on Vancouver Island, the Westcoast Crystalline Complex, is inferred to be metamorphised and dioritized oceanic crustal material. The Sicker Group overlying the basement platform represent the initiation of igneous activity on the Island. The sequence is composed of metavolcanic rocks of Pennsylvanian age overlain by minor volcaniclastic and carbonate units which are generally unmetamorphised, well bedded, coarse bio lastic reef and shelf limestones.

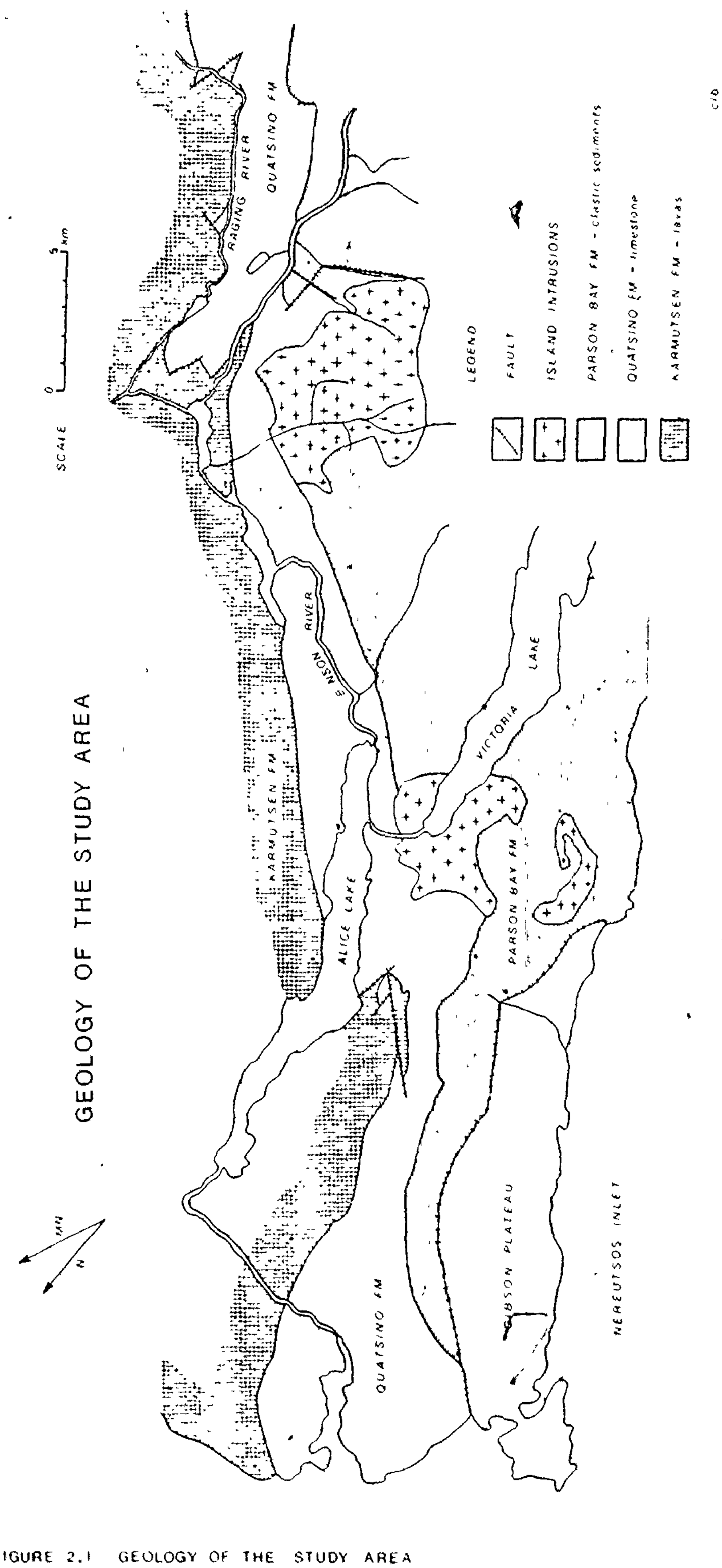


FIGURE 2.1 GEOLOGY OF THE STUDY AREA

The Sicker Group outcrops in the Alberini area of eastern Vancouver Island and is not exposed in the Benson Valley.

2.2.2 Triassic

Vancouver Group

The Vancouver Group is the aerially most extensive sequence on Vancouver Island and dominates the geology of the study area. First named by Dawson (1887) the sequence consists of basic sills and minor sediments, a thick pile of Triassic balsaltic volcanics (Karimutsen Fm.), Upper Triassic carbonate, pelitic and volcaniclastic sediments (Quatsino and Parson Bay Fms.) and a Lower Jurassic sequence of basaltic and pyroclastic volcanics with minor intercalated sediments (Bonanza sub Group). (Figure 2.2)

Sediment-Sill unit

First recognised around Buttle Lake, Carlisle (1972) has revealed the presence of a unit of basic sills and sediments between the Sicker Group and the overlying Karimutsen lavas. The sills are diabases composed of labradorite and pyroxene exhibiting a marked increase in grain size from the contacts to the central part. The sedimentary layers between the sills range from less than 1 to greater than 60m in thickness. The rocks are laminated to graded-bedded, Daonella rich, black shales and siltstones generally silicified by diagenesis and contact metamorphism by the invading sills.

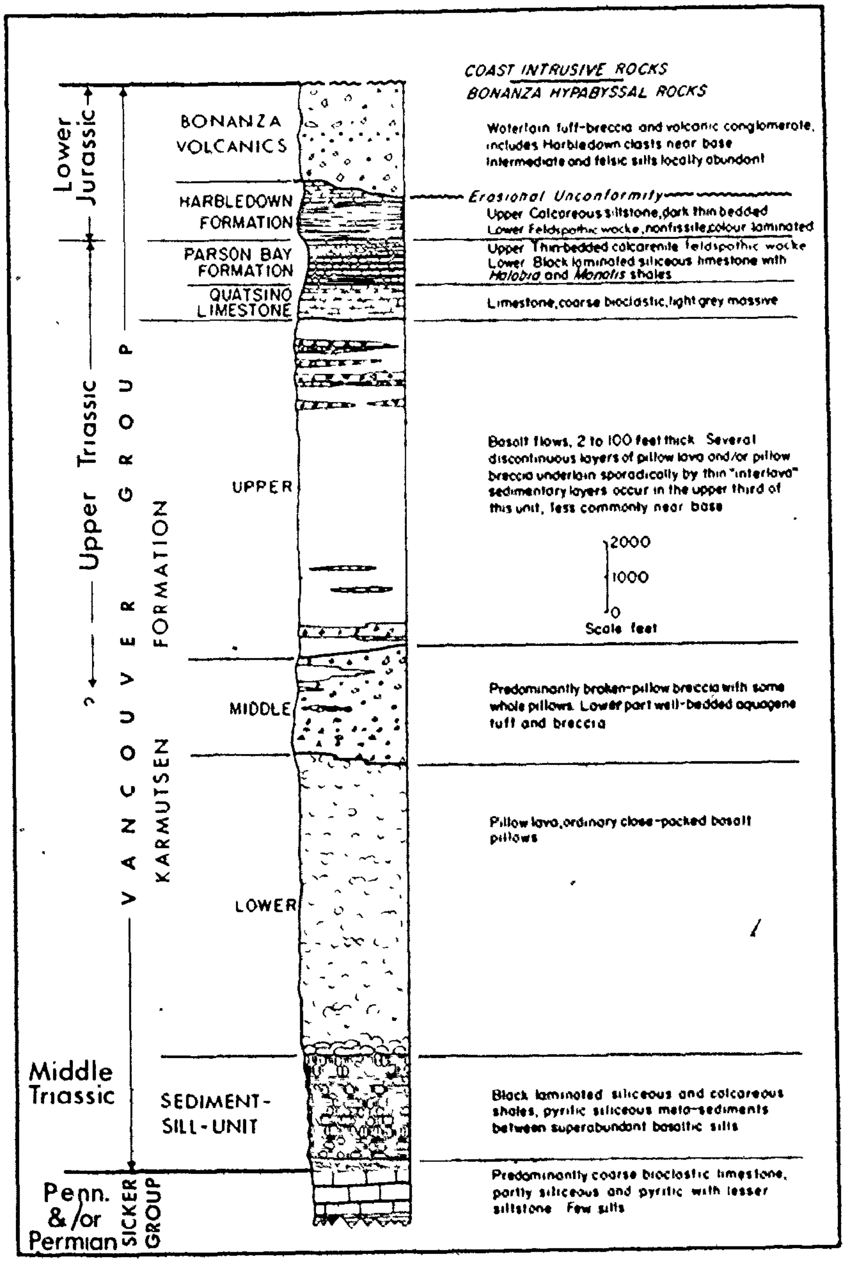


FIGURE 2.2: Stratigraphy of Sicker and Vancouver groups (Source Muller, Northcote, and Carlisle, 1974)

Karmutsen Fm.

Rifting of an old Paleozoic basin in the mid-Triassic produced lava outpourings up to 6000m in depth. The base of the Karmutsen Fm. is composed of close packed angular pillow fragments of diverse size, in a pyroclastic matrix of feldspar, augite and partly devitrified volcanic glass. The upper subdivision is composed of predominantly subaerial lava flows 1 to 30m thick, dissected by joints at angles roughly normal to the depositional layering. Columnar jointing is reportedly rare but was observed at two locations in the Benson Valley. Intervolcanic limestone, similar to the overlying Quatsino Fm. limestone but only 1 or 2m thick, has been noted near the top of the Karmutsen Fm. in several areas (Muller and Carson, 1969). In the Benson Valley, limestone sub units upto 3m thick and separated from the Quatrino Fm. by less than 6m of layered lavas, were observed.

Quatsino Fm.

The Quatsino Fm. is the principal carbonate unit of the Vancouver Group. Exposed in three, fault dissected, linear belts trending dominantly north west - south east, the outcrop covers approximately 1000 km^2. It is the most intensively investigated unit in the region, primarily because it served as the host rock for extensive mineral emplacement.

The formation consists of biomicritic limestones displaying a bimodal size distribution, dominantly composed of coral and pelecypod debris in a volumetrically dominant matrix of fine carbonate muds. The base of the sequence, which rests paraconformably on the Karmutsen lavas, consists of dark grey to black microcrystalline but locally coarsecrystalline strata. These are thick-bedded or massive and commonly display a stylolitic structure. The upper part of the Quatsino Fm. consists of medium to thick bedded limestones, interlaminated with black calcareous siltstones up to 0.3m thick. The limestone grades upwards into calcareous clastic sediments; shale intercalations increase in thickness as the depth of carbonate beds decrease. The contact with the overlying Parson Bay Fm. is diachronous and placed where black shales and arenites first predominate over pure carbonates. The thickness of the limestone unit is variable; stratigraphic thickness exceeds 700m occur south of Alice Lake in the Benson Valley (Muller, Northcote and Carlisle, 1974). However, the total thickness of limestone in this location is greater than 1000m. This may result from folding or thrust faulting.

Cherty nodules are present at several stratigraphic levels in the Quatsino Fm., but intense silicification of the limestone is only observed near granitic intrusions. Petrographic analysis of samples collected throughout the unit suggest that dolomitization of the Quatsino Fm. has been minimal (McCammon, 1969). This probably results from burial and lithification beneath Parson Bay lithologies coupled with rapid emergence during tectonic uplift, so that exposure to diagenetic inducing brines, especially those of shallow water and intertidal environments, has been minimized.

The age of the Quatsino Fm. has been documented by fossil collections examined by Tozer (1967). Deposition occured during Upper

Karmian and Norian times of the Upper Triassic.

Parson Bay and Harbledown Fms.

The Parson Bay Fm. is exposed adjacent to the Quatsino Fm. in three Triassic sedimentary belts. At the base, black to dark grey calcareous siltstones, shales and shaley limestones are interbedded with progressively thinner layers of pure Quatsino limestone. Progressively the fine grained clastic sediments are replaced by sandstones and greywackes which in turn grade into grits and breccias of limestone and volcanic composition. The sequence is continued into the Harbledown Fm. which is distinguished from the Parson Bay Fm. by its Lower Jurassic fossil assemblage.

2.2.3 Lower Jurassic

Bonanza Volcanics and Island Intrusions.

The Bonanza Volcanics are exposed in the south west of Northern Vancouver Island. The lithology is varied and heterogenous in comparison to the uniform sequence of the Karmutsen volcanics. Lavas range in composition from basaltic andesite to rhyodacite and are interbedded with maroon and green tuffs and breccias, and several clastic sedimentary units.

The Island Intrusions, produced by quartz diorite to quartz magmas, occur as variable size batholiths orientated in a north west - south east direction. Two small plutons each approximately $10km^2$ in area, outcrop at the western margins of the Benson Valley. Limited metamorphism within the surrounding limestone country rock has produced grey coloured, medium coarse even grained marbles (Dolmage, 1921).

Intrusive dykes and sills of Lower Jurassic age are widespread in the Vancouver Group. In most instances surface expression is masked by till and rugged terrain; however, the features are readily identified in river gorges and cave passages. The dykes are primarily oriented parallel to the stratal dip, but some occur parallel to the strike. Intrusive sills are less than 0.4m thick.

The frequency of occurrence of intrusive features in the Quatsino Fm. limestone is greater than that recorded in any other karst area; one 600m long strike oriented cave passage intercepted 11 dykes parallel to the dip. This frequency is many times greater than that of the much reported intrusions of the Carboniferous Limestone of Derbyshire, England (see Ford, 1977).

2.2.4 Post Jurassic

Post Jurassic deposition is limited to the coastal areas of Northern Vancouver Island and no deposits are known in the Benson Valley. Potassium-argon dating has revealed limited Tertiary tectonic and igneous activity, but only one small pluton has been positively identified in the region.

2.3 Structural Geology

The structural geology of Northern Vancouver Island is dominated by fault blocks uplifted along the medial, N.W. - S.E. trending, Victoria Arch. The Victoria Arch represents the axial high of the

Jurassic igneous period and contains the batholiths which form the backbone of the Island. The arch is flanked by outward dipping fault blocks, dissected by irregular sets of steep to vertical faults of largely unknown displacement. The predominance of tilted faulted strata rather than folded strata in the region probably reflects the inelastic behaviour of the Karmutsen basalts under deforming stress.

The Benson Valley occurs on the Karmutsen fault block to the west of the Victoria Arch. The stratal dip in the valley ranges from about 10° to 45° and is consistently to the west or south west. The principal faults in the Benson Valley occur adjacent to the intrusive plutons. Many are oriented roughly parallel to the stratal dip or strike and in some locations faults intersect other faults. In places their influence on hydrogeology is fundamental. The dominant jointing patterns in the Benson Valley limestone is a reticulate network with axes trending roughly $60^{\circ} - 240^{\circ}$ and $150^{\circ} - 330^{\circ}$. The joints parallel to the dip ($60^{\circ} - 240^{\circ}$) are densely packed, sometimes less than three centimetres apart. These joints generally do not pass from one bed to those above or below, however, some master joints occur parallel to the strike.

2.4 Geological History

It is unlikely the PreCambrian granitic crust underlies any part of Vancouver Island. Rather, the basal rock sequence (the Westcoast Crystalline Complex) is inferred to be metamorphosed and dioritized oceanic crustal material with a quartz dioritic to gabbroic composition.

Ocean floor spreading and the creation of a volcanic island arc in the Late Paleozoic produced the Sicker Volcanics. Volcanism ceased in Early Pennsylvanian times and basin infilling by volcaniclastic and limestone sedimentation followed until the early Permian. In the Middle Triassic a unit of black argillite and siltstone containing pelagic Daonella was laid down.

Rifting within the old volcanic island arc in the Upper Triassic created oceanic troughs similar to those of the present Western Pacific. The Inter Arc Basin concept developed by Karig (1971A&B) may be applicable in this situation. (Muller, Northcote and Carlisle, 1974). Karig proposed that structural basins formed in these circumstances are bounded by fault scarps, have a ridge-troughtopopgraphy and contain a shallow, medial, axial high. The basin becomes the site of major basaltic deposition. Over 6000m of basalts were extruded into the basin to form the Karmutsen Fm. The base of the sequence, composed of pillow basalts, was deposited into a marine environment. In the centre of the basin along the axial high, water shallowed sufficiently to become subject to wave action and the close packed pillow lavas were replaced by pillow breccias and aquagene tuffs. Eventually shield volcanoes emerged along the axial high and produced subaerial bedded lava flows.

In the late Karnian of the Upper Triassic rifting and related extrusive activity ceased and the prevailing tropical or sub-tropical conditions permitted the accumulation of open sea biomicritic limestones. Accumulations in excess of 700m occured where subsidence kept pace with carbonate deposition. Erosion of the surrounding arc shed debris into the basin forming the Parson Bay Fm. Initial clastic deposition occurring at the basin margins was concurrent with limestone deposition in distal waters; with time the clastic facies extended into the centre of the basin. This resulted in the diachronous boundary between the limestone and clastic units and the coarsening upwards evident in the Parson Bay Formation.

Renewed arc-type volcanism in the Early Jurassic formed the Bonanza Volcanics. The episode probably resulted from mobilization of the old Paleozoic basement (Muller, Northcote and Carlisle, 1974). Volcanism was coupled with plutonic intrusions and the uplift of the structural high along the Victoria Arch. Away from the batholiths, igneous activity was restricted to intrusion of sills and dykes.

Post-Jurassic deposition was restricted to clastic shelf sedimentation at the margins of the Island. Few events are recorded from the Tertiary era, but the region has remained tectonically active into the Quaternary.

2.5 Quaternary History

Vancouver Island has been extensively glaciated by local alpine ice bodies and intruding ice sheets of continental origin during the Pleistocene. The number of glaciations on Northern Vancouver Island is not well established. However, given the elevation of the mountains on the Island (>1750m) and the location adjacent to the West Coast mountains the number is likely to be large.

U-shaped valleys and cirques are abundent in the central mountains of the Island. This demonstrates extensive local mountain ice generation. However, the occurrence of U-shaped valleys extending through the mountain belts (Fyles 1963) and the occurrence of calcareous drift to the west of the limestone outcrops (D. Howes pers. comm. 1980) suggests that a Cordillerian ice sheet moved across Vancouver Island from the mainland during the last (Frazer) glaciation. Measurement of glacial striae indicate that all but the highest peaks were buried beneath ice (Howes, 1980). This indicates an ice depth greater than 1000m in the Benson Valley at the maximum.

Fyles (1963) and Halstead (1968) suggest that an episode of valley glaciation followed the maximum of the Frazer Glaciation in southern Vancouver Island. It is uncertain whether a similar episode occured at the north end of the Island as none of the valleys are U-shaped troughs (Muller, Northcote and Carlisle, 1974), or contain clearly defined cirques. Howes (1980) suggests that ice in the region ablated and downwasted into discrete valley glaciers which thinned and stagnated. Limited ice scouring is evident in the Benson Valley from perched and truncated cave passages and glacial striae preserved beneath carbonate tills. Glacial till occurs throughout the valley but the greatest depths occur on the valley sides and in the head waters, which suggests ice movement up the valley. Till depths exceed 6m in these areas whereas deposits on the valley floor are frequently less than one metre thick.

The late Quaternary chronology of Vancouver Island is known from work on the Island and the adjacent mainland. The oldest deposits on the Island, the Mapleguard Sediments, are thought to be pre -

Early Wisconsin in age (Fyles, 1963). Controversy has existed regarding the presence or absence, and timing, of a Mid Wisconsin glaciation in the region. A Mid Wisconsin intersrstadial is known from radiocarbon dating of sedimentary deposits (Alley 1979, Howes 1980, Armstrong and Clague 1977) and from Uranium series dating of speleothem in south-central Vancouver Island (Gascoyne 1980) and in the Benson Valley.

The late Wisconsin or Frazer Glaciation was established in the north of the Island prior to 25,000 BP, and reached its maximum after 20,600 ± 330 BP (Howes 1980). No deglaciation date was obtained from the Benson Valley but speleothem deposition had occured on an adjacent limestone ridge by 11,600 ± 300 BP.

CHAPTER 3

THE SOLUTION PROCESS IN THE QUATSINO FORMATION KARST

3.1 Introduction

The objectives of this chapter are: (1) to determine the potential equilibrium concentration of the limestone ,(2) to describe patterns of solution there in terms of the chemical character of waters routed along different hydrogeological flowpaths and the dynamic hydrochemistry of a monitored cave conduit system; and (3) to test the validity of a priori hydrogeological classes in the karst by linear discriminant analysis of class water chemistry data.

3.2 Karst Process-Response System

In karst regions the dominant erosion process is solution. Although other geomorphic processes may influence landform development in general a relationship between solution and the surficial or subsurface landforms developed is apparent. Most karst landscapes are thus products of a comparitively simple solutional process-response system.

The boundaries of the solution system can be defined by the spatial interaction of aggressive waters with limestone occuring as bedrock or in particulate form in the soil or regolith profile.

The solution process frequently alters the system boundaries, eg. the progressive leaching of calcium carbonate from the soil (Trudgill 1976) or the increasing penetration distance of solvent entering a limestone mass over time (White 1977).

Like most natural systems, karst development is controlled by the driving action of exongenous variables. Primarily, these are the climatic variables temperature and precipitation which set overall limits to the solution rate in a region through their control of mineral solubility and carbon dioxide production, and solvent availability, respectively. In the Benson Valley karst the importance of the solvent availability factor is reflected by the fact that cave systems of explorable dimensions are concentrated in zones recharged by sinking allogenic streams. Overall temperature control defines the maximum concentration of limestone removed in the water. Spatial variation of CO_2 production is reflected in the hardness distribution of saturated waters.

Endogenous factors in the karst process-response system influence the rate and distribution of solution. These are dependent primarily on the lithology and hydraulic regime of the solvent, and the framework of structural pathways available to route groundwater, respectively.

3.3 Limestone Solution

The geochemistry of limestone solution has been described in detail by Garrels and Christ (1965), Stumm and Morgan (1970), Langmuir (1971) and Drake (1974) and is not reviewed here. In this study the concentration of Ca^{2+}, Mg^{2+}, H^{+}, HCO_3^{-}, H^{+} and SO_4^{2-} and the temperature and the specific conductivity of sampled waters were measured. The carbon dioxide partial pressure and the saturation status of the water calcite (SI_c) and dolomite (SI_d) were calculated on a CDC 6400 computer using WATSPEC, a program written by T.M.L. Wigley (1977). WATSPEC was also employed to calculate the character of groundwater at equilibrium from temperature and PCO_2 data.

3.4 Specific Conductivity

The specific conductivity of a substance is a measure of its ability to conduct an electric current (measured as the ratio of the current to the applied electromotive force under standardised conditions). In an ionic solution this results from the movement of charged ionic species along an electropotential gradient and is dependent on the solution temperature, ionic concentration and specific ion mobility. In most dilute aqueous systems a linear relationship exists between ionic concentration and specific conductivity so that the latter measurement can be employed as a surrogate variable in the determination of solute concentrations. In this study measurement of specific conductivity was obtained with a Y.S.I 33 S-C-T meter. Values were standardized by conversion to the specific conductance at 25°C, by assuming a linear increase in conductance of 2% per °C by the equation:

$$SPC_{25} = (1 + (0.02(25 - T)))\, spc_T \qquad (1)$$

3.5 Hydrochemical Data Set

The sampling program ran from June to August 1980. Emphasis was placed on spot sampling from the widest range of hydrogeological environments available in an attempt to determine the spatial patterns of solution. A repeated sampling program was mounted along the Malo₅k Creek Cave system to determine the dynamic nature of the chemistry of a sinking stream conduit cave system in the region.

On-site determinations of temperature, pH and specific conductivity were made at each sampling. Temperature was determined with calibrated mercury in glass thermometers with a precision of $\pm 0.1^{\circ}$ C. pH was measured using a portable Sargent Welch meter calibrated at pH 10.0 and pH 7.0 before each measurement. The precision of the measured values is estimated to be $\pm$ 0.1 unit. Further errors may have resulted from: (1) the streaming of fluid past the electrode (2) slow instrument response with measurement made before the final equilibrium point (3) interference in the system electronics resulting from high humidity in the measuring environment. An attempt was made to reduce the influence of these factors by immersing the electrode in static water, having the buffer solutions at the water temperature and minimising the number of determinations made in cave passages. Minimising error in pH measurement is a fundamental problem since an error of 0.1 pH unit introduces an equal error into a calculated SI_C or SI_D value. Error in pH measurement is probably the largest contributing factor to error in the calculation of mineral saturation status in this study. To account for this factor and other possible errors made during chemical analysis, SI_c and SI_D values in the range $\pm$ 0.3 are considered to be at saturation in this thesis.

Analyses for Ca^{2+} and Mg^{2+} were performed within 24 hours on samples stored in NALGENE bottles. The method consisted of standard Schwartzenbach EDTA titrations using B.D.H. reagents. Titration alkalinity was determined colorimetrically with 0.05 N HCl to the Bromocresel end point, as a measure of HCO_3^-. Sulphate analyses were made using a commercially available Lamotte SO_4 kit. The accuracy of the method is estimated to be $\pm$ 5mg l^{-1}.

3.6 Controls of Limestone Solution in the Quatsino Fm.

3.6.1 Temperature

Temperature controls the solubility of the Quatsino limestone through the temperature dependence of the equilibrium constants of the reactions in the solution process, and indirectly by its supposed control of carbon dioxide production. The solubility of calcite, which dominates the composition of the Quatrino Fm. limestone is inversely dependent on temperature under a given PCO_2. For PCO_2 = $10^{-3.5}$ atm and coincident evolution the solubility of calcite increases by approximately 50% for a decrease in temperature from 25°C to 0°C. (Drake unpubl. see figure 3.1) Under these conditions the equilibrium concentration of the Quat ino Fm. limestone is approximately 27mg $l^{-1}Ca^{2+}$.

Some workers have employed the above relationship coupled with the negative temperature dependence of carbon dioxide solubility to

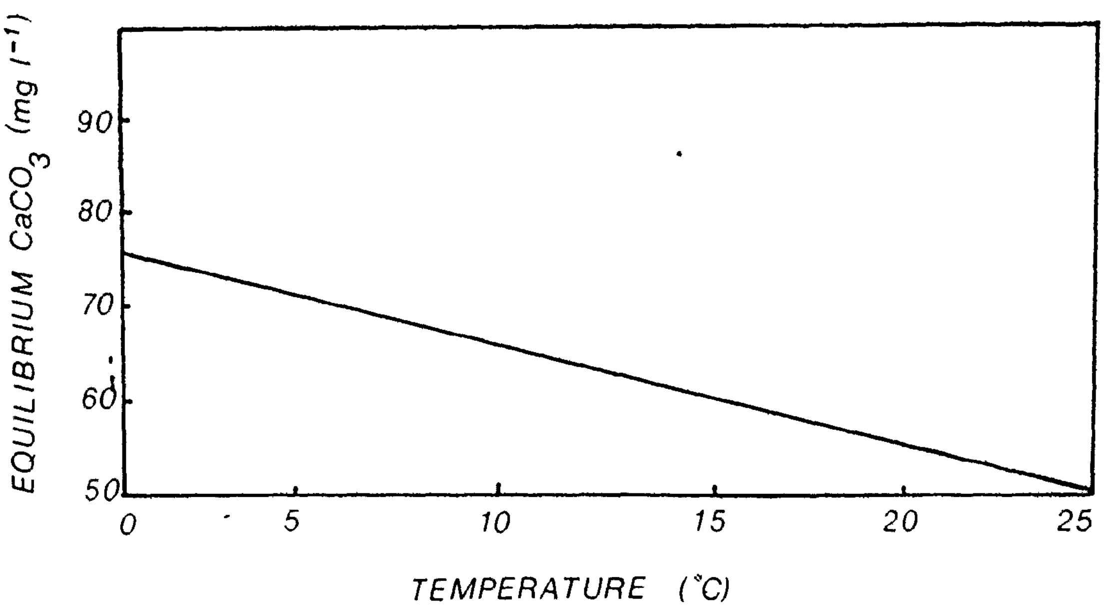


FIGURE 3.1: Solubility of calcite ($P_{CO_2}=10^{-3.5}$ atm, concordant system) as a function of temperature.
Source: DRAKE (unpubl.)

postulate maximum erosion rates in cold climates (eg. Corbel 1959). However, comparable concentrations of dissolved limestone in different climatic regions, and large variations in solute concentrations within single regions (Smith and Atkinson 1976) suggest that other factors must override the direct temperate dependance of calcite and carbon dioxide solubility in solute concentration determination.

3.6.2 Carbon Dioxide - Solute Concentration Relationships

Water enters the Benson karst via two pathways, as incident precipitation or invading allogenic stream waters. The PCO_2 of precipitation approximates $10^{-3.5}$ atm and the mean PCO_2 of allogenic stream water was determined as $10^{-3.04}$ atm. The PCO_2 of saturated groundwater in the karst was determined as $10^{-2.42}$ atm. Clearly, the carbon dioxide partial pressures of waters entering the karst are insufficient to explain the PCO_2, and hence solution potential of groundwater in the region.

Enhanced carbon dioxide partial pressures in karst waters derive primarily from biogenic CO_2 production in the soil (eg. Lehmann 1964, Jakucs 1977). Carbon enters the soil system by fixation by higher plants during photosynthesis. The principal source of soil air CO_2 is the decomposition of organic materials by microorganisms; however, higher plant respiration also releases CO_2 to the soil via CO_2 root pumps.

Carbon dioxide is lost from the soil by diffusion along the partial pressure gradient to the external atmosphere, by hydration and drainage in water under gravity flow when soil water content exceeds field capacity and by consumption during the solution of $CaCO_3$ clasts in the soil profile. This latter process is responsible for the preweathering or 'vorverwitterung' (Miotke 1974) phenomenon in calcareous soils, observed as a high hardness pulse in soil water drainage following recharge after an extended dry period.

Soil PCO_2 measurements were made in the Benson karst in an attempt to determine: (1) the potential solubility of limestone exposed to soil percolation waters, (2) the relationship between soil or bedrock PCO_2 and groundwater PCO_2 and (3) whether local factors override a simple, externally controlled pattern of CO_2 production. PCO_2 was measured at several points in each of 8 sites under varying vegetal and pedologic conditions, with Draeger tubes (Table 3.1). The sampling depth was standardised at approximately 30cm below the surface.

The mean CO_2 concentration of the 8 sites is 0.26%; site means range from 0.11% to 0.57%. The mean soil CO_2 concentration for other temperate regions determined by Smith and Atkinson (1976) from 9 data sources was 0.91% with a range in group means from 0.1% to 1.8%. Thus, at the time of sampling CO_2 production in soils in the Benson karst was comparable to the less productive sites in the Temperate data set. Contributing factors may be the low summer season mean temperature, dry antecedent conditions and the fact that soils tend to be skeletal and disturbed by logging activities.

Considerable variation is evident between the groups in Table 3.1, even between groups 5 and 6 which are located less than 5m apart. The highest group mean (0.57% or $10^{-2.24}$ atm) occured in a series of soil-filled grikes and small hollows. A similar pattern was found on the Bruce Peninsula (Cowell 1976). The mean value in thin soils between the grikes was 0.14% or $10^{-2.85}$ atm. The Mann-Whitney U test was employed to test the hypothesis that the CO_2 concentration in site 6 is significantly greater than that in site1 at the 0.005 level. The test supported the contention, which indicates that significantly different CO_2 production rates occur in soils in the Benson karst.

The calculated PCO_2 of vadose seepage water (cave roof drips) in the Bath Cave approximately 15m beneath a grike surface (site 6, table 3.1) was $10^{-1.53}$ atm. The PCO_2 of a cave drip in Haphazard Cave, approximately 5m beneath the surface of soil site 1, was $10^{-2.24}$ atm. This suggests that measured soil PCO_2 is not a good estimate of the maximum PCO_2 to which percolation waters are exposed. This probably results from underestimation of the true soil PCO_2 because of sampling in the top 30cm of the soil profile, or possibly CO_2 production in fissures in the vadose zone. It is not possible to determine whether the higher PCO_2 of the Bath Cave drip is the result of the higher PCO_2 in the overlying soil alone, or is also dependent on the bedrock depth through which the percolation water has passed.

TABLE 3.1

Soil CO_2 data

DATE	21/7	6/8	6/8	12/8	12/8	12/8	14/8	14/8
SITE	MINIGILL	TOP WINDOW	TOP WINDOW	CRAFT CREEK	DEVIL'S BATH	DEVIL'S BATH	JEUNE LANDING	JEUNE LANDING
SOIL TYPE	organic	disturbed till	till	till	thin till	grike	till (open forest)	till (fern floored forest)
SITE ID	1	2	3	4	5	6	7	8
CO_2 %(vol)	0.1	0.15	0.3	0.08	0.11	0.3	0.09	0.25
	0.1	0.15	0.3	0.1	0.15	0.3	0.1	0.28
	0.11	0.15	0.7	0.9	0.16	0.5	0.1	0.15
	0.11	0.15	0.3	0.11		1.05	0.9	0.1
	0.11	0.11		0.1		0.5	0.07	0.2
	0.1			0.1		0.55	0.9	0.2
						0.5	0.11	0.1
							0.1	0.15
							0.4	0.15
								0.15
$\bar{x}$%	0.11	0.14	0.4	0.23	0.14	0.57	0.31	0.17
SD	0/006	0.02	0.2	0.33	0.03	0.28	0.35	0.06

A similar pattern of karst water PCO_2 in excess of measured soil PCO_2 is reported in the Mendip Hills, England (Atkinson 1976). Three springs draining the central part of the Mendip plateau had groundwater PCO_2 values ranging from $10^{-1.5}$ to $10^{-1.8}$ atm. Soil PCO_2 values were less than $10^{-2.0}$ atm and thus cannot explain groundwater PCO_2. However, the mean PCO_2 of air sampled from small fissures in caves was $10^{-1.75}$ atm. Moreover, fissure PCO_2 increased with depth. This suggested (1) the presence of a vadose "ground air" source of CO_2 distinct from soil air CO_2 and (2) that maximum ground air PCO_2 is dependent on the depth of the vadose zone. Atkinson suggests that the principal source of ground air CO_2 is the decomposition of organic material washed into fissures by percolation waters. The patchy distribution of carbonate rich tills in the Benson karst, indicating extensive bedrock solution, coupled with vadose zone depths known to exceed 30m suggest that the possibility of ground air CO_2 production in vadose fissures in the Benson karst cannot be ruled out without more complete soil profile PCO_2 data.

3.6.3 Additional factors affecting the solubility of the Quatsino Limestone

The solubility of limestone is decreased in the presence of certain trace metals; eg concentrations of Cu^{2+} in the range from 10^{-4} M l^{-1} to 10^{-5} M l^{-1} were found to decrease the solubility of limestone twofold by increasing the rate of reverse reaction in the solution process (Terjesen et al 1961). Extensive mineral emplacement, primarily of copper, occurred in the Quatsino Fm. limestone during the

Tertiary tectonic episode. Commercial copper deposits have been mined at two sites in the Benson Valley. Subsidiary emplacement may have occurred in other locations in association with the dyke and sill intrusions. At present no ground water trace metal data are available and the extent of solution inhibition is unknown.

The concentration of limestone in groundwaters in certain karst regions is enhanced by the presence of ion pairs in solution (eg Fish 1977). The most significant effects in karst terrains result from the pairing of Ca^{2+} and Mg^{2+} with SO_4^{2-}, and of Mg^{2+} with HCO_3^- and CO_3^{2-} (Drake 1974) . Analyses performed on samples from a wide variety of hydrogeological environments did not detect the presence of sulphates in the Benson karst. The detection limit of the method was about $5 mgl^{-1}$.

The relationship between total hardness and SPC_{25}, determined from 67 samples is defined by the best fit (least squares) equation:

$$SPC = 11.9 + 1.698\ TH \qquad (2)$$

The equation suggests that $SPC_{25} = 11.9\ uS\ cm^{-1}$ when $TH = 0\ mgl^{-1}$. This suggests the presence of a solute load in the karst waters, derived from processes outside the karst, ie. in precipitation or allogenic waters from the Karmutsen lavas. Albutt (1977) describes a similar relationship from Yorkshire, England. In this region the predicted presence of non-carbonate hardness solutes was supported by the measurement of $8 mgl^{-1}$ Cl^- in all samples, almost certainly introduced in precipitation. Cl^-, Na^+ and K^+ may occur in the Benson karst because of the maritime location. The presence of the ions would marginally increase the solubility of the Quatsino Fm. limestone by the "salt effect", ie. by increasing the ionic strength of the solution.

3.6.4 Quatsino Fm. Limestone Solubility

The potential equilibrium concentration of the limestone in the Benson karst has been determined solely with regard to temperature and PCO_2 under coincident conditions. The contribution of the factors in 2.6.4 is ignored because the effects are either slight or impossible to predict from available data.

Table 3.2 outlines the potential equilibrium concentration of limestone in waters exposed to the various environments investigated in this study, assuming a mean ground water temperature of 7^oC.

The mean groundwater concentration of Ca^{2+} determined from 8 samples is 47 mgl^{-1} with a range from 34 mgl^{-1} to 70 mgl^{-1}. The mean potential equilibrium concentration of calcite in the soil water discharge is 52 mgl Ca^{2+} with a range from 40.1 mgl^{-1} to 74 mgl^{-1}. This suggests that the measured soil PCO_2 can account for the groundwater calcium hardness assuming fully coincident conditions. However, the patchy distribution of carbonate rich soils and the fact that cave drip PCO_2 $(\bar{x} = 10^{-1.75})$ can produce a calcium hardness of 42.6 mgl^{-1} under fully subsequent conditions suggest that (1) most

TABLE 3.2

Potential solubility of (the Quatsino Fm.) limestone in waters exposed to different P_{CO_2} environments, at $7^{o}C$

ENVIRONMENT	SITE	log P_{CO_2}	EQUILIBRIUM Ca^{2+} mgl^{-1}
atmosphere	-	-3.50	27.4
allogenic streams	-	-3.04	40.7
soil	1. Haphazard	-2.98	40.1
	2. Top Window	-2.85	45.9
	3. Top Window	-2.40	66.0
	4. Craft Creek	-2.63	54.0
	5. Devil's Bath	-2.85	45.9
	6. Jeune Landing	-2.51	60.8
	7. Jeune Landing	-2.76	50.0
grike	Devil's Bath	-2.24	74.0
vadose seep	Bath Cave	-1.53	138.5
	Haphazard Cave	-2.24	74.0

solution occurs under subsequent conditions and (2) the solution process is not controlled by shallow soil PCO_2. More detailed ground air and soil profile PCO_2 data are required to test these hypotheses.

3.6.5 Regional Solute Concentration-Climate Modelling

Limestone outcrops in every major climatic region of the world. The morphological similarity of karst landscapes in similar climates, coupled with the marked variation in morphological response between different climates has led several workers to conclude that climate must control karst landform development, and inferentially the solution process to a large degree.

Harmon et al (1975) have demonstrated that the equilibrium concentration of limestone in karst terrains throughout North America is well correlated with the mean annual groundwater temperature. The relationships between mean annual temperature and chemical parameters determined in their study are presented in Table 3.3. Although the set includes data from 7 distinct karst regions, the independent variable, mean annual temperature, distinguished only 5 groups: Canada, Virginia/West Virginia and Pennsylvania, Missouri, Kentucky, Texas and Mexico.

In order to determine the relationship between climate and potential groundwater solute concentrations it was necessary to study waters at equilibrium. Thus Harmon et al employed spring and well water data. However, the significant correlation between SI_c and

TABLE 3.3

Simple linear regression equations for temperature and chemical parameters in study of climatic control of groundwater chemistry

DEPENDENT VARIABLE Y	INDEPENDENT VARIABLE X	INTERCEPT	SLOPE	r
Ca^{2+}	water temperature (^{0}C)	-11.25	5.70	0.903
HCO_3^-	water temperature (^{o}C)	37.20	11.37	0.966
SIc	water temperature (^{o}C)	0.675	0.024	0.839
log P_{CO_2}	water temperature (^{o}C)	-3.33	0.081	0.962

Source: Harmon et al 1975

temperature indicated that many of the samples were not at equilibrium. Drake and Wigley (1975) redefined the data set to exclude samples with $SI_c > |0.1|$. The Drake-Wigley relationship between M.A.T and groundwater PCO_2 was defined by:

$$\log PCO_2 = -3.42 + 0.07\ T\ ^oC \qquad (3)$$

The predicted PCO_2 of groundwater in the Benson karst employing this relationship is $10^{-2.93}$ atm. The mean PCO_2 of four samples within the SI_c range ± 0.1 was $10^{-2.81}$. Therefore groundwater character in the Benson karst supports the hypothesis implicit in Harmon et al that equilibrium concentrations of limestone in karst groundwater can be determined from the climatic variable, mean annual temperature. The relationships between solution in the Benson karst and other carbonate terrains in North America, employing the uncorrected Harmon et al data set, are presented graphically in figure 3.2.

Drake and Wigley (1975) attempted to explain relationships between MAT and soil CO_2 production described by Harmon et al 1975, since "by isolating this co ection we can provide a vital link in the karst erosion cycle and our understanding of carbonate groundwater chemistry". Soil PCO_2 dependence on temperature was calculated indirectly from both data on soil CO_2 production rates and CO_2 transfer rates or fluxes into the atmosphere. Frequently these two quantities would be equal. However, CO_2 hydration in percolation waters or consumption during solution in the soil profile could reduce the latter. The

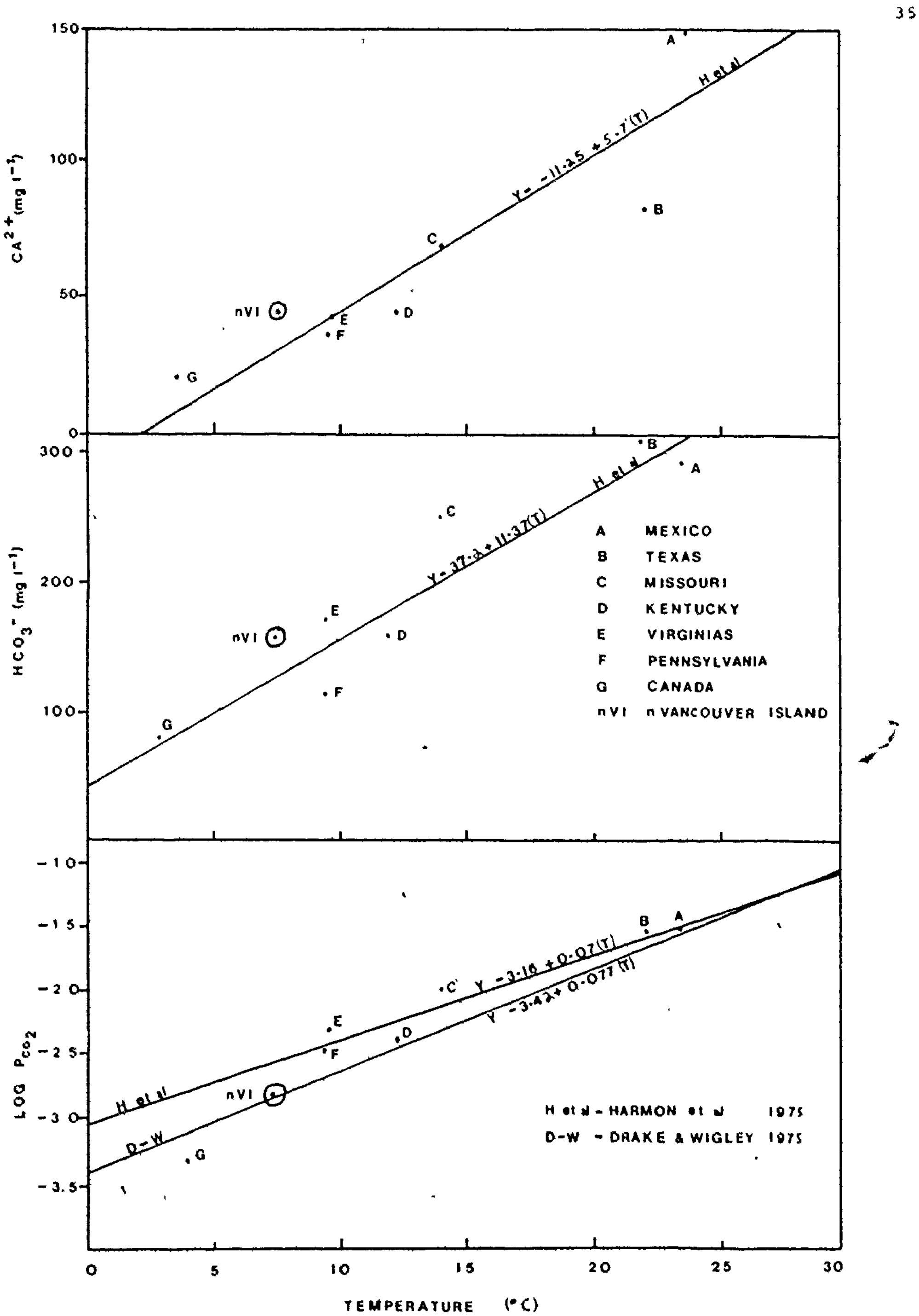


FIGURE 3.2 Groundwater chemistry v. temperature in North American carbonate terrains.

temperature coefficient of the log soil PCO_2 - temperature relationship was found to approximate $0.04/^{o}C$. This differs from the log groundwater PCO_2 - M.A.T. relationship determined to be approximately $0.07/^{o}C$. However, subsequent system evolution was used to explain the discrepency and under these conditions an initial soil PCO_2 - M.A.T. relationship of

$$\log PCO_2 = -1.97 + 0.04T \qquad (4)$$

was developed to explain the groundwater-climate relationship of Harmon et al (1975). Subsequently, Drake (1980) modified equation 4 to account for the inhibiting effect that a high PCO_2 places on respiration and hence further CO_2 production, to:

$$\text{soil } PCO_2 = \left(\frac{(0.21 - PCO_2)}{0.21}\right) PCO_2^{*} \qquad (5)$$

where PCO_2^{*} = "uninhibited" PCO_2 determined from equation 4 above.

The predicted PCO_2 of soils in the Benson karst from equation 5 is $10^{-1.73}$ atm. Assuming subsequent system solution this value predicts a solute concentration of approximately 43 mgl^{-1} Ca^{2+}. The mean PCO_2 in the region was $10^{-2.67}$ atm. Clearly equation 5 does not reflect CO_2 production in the top 30cm of the soil profile, but it <u>does</u> closely reflect the PCO_2 of vadose percolation waters.

In summary, section 3.6 has attempted to define the potential equilibrium concentration of the Quatsino Fm. limestone in water, primarily w.r.t. to carbon dioxide production and temperature. <u>Measured</u> soil PCO_2 is both spatially variable and unlikely to control groundwater PCO_2. Vadose seepage water data suggest that most CO_2 production occurs in the rooting zone at the base of the soil profile or in the bedrock ground air zone. Ground air CO_2 production is likely to occur where organic matter decays in bedrock fissures or where the soil is shallow and the base of the rooting zone occurs beneath the soil-bedrock contact.

The character of groundwater at equilibrium in the Benson karst is similar to that predicted from M.A.T. - PCO_2 models developed from data from different climatic regions. This supports the hypothesis that a process-response system has been clearly defined in karst terrains.

3.7 Aquifer Hydrochemistry

3.7.1 Hydrologic Class Chemical Characteristics

Waters routed through karst regions follow a variety of hydrologic pathways and consequently develop different chemical signatures. Variations in chemical signature reflect variations in such factors as flow regime, degree of concentration of flow, subsurface residence time and exposure to different PCO_2 environments. These result in waters having both different solution potentials and being at different stages of chemical evolution towards saturation. The distribution of conceptually defined hydrochemical environments in the Benson karst is outlined in figure 3.3. Group chemical characteristics are summarised in table 3.4.

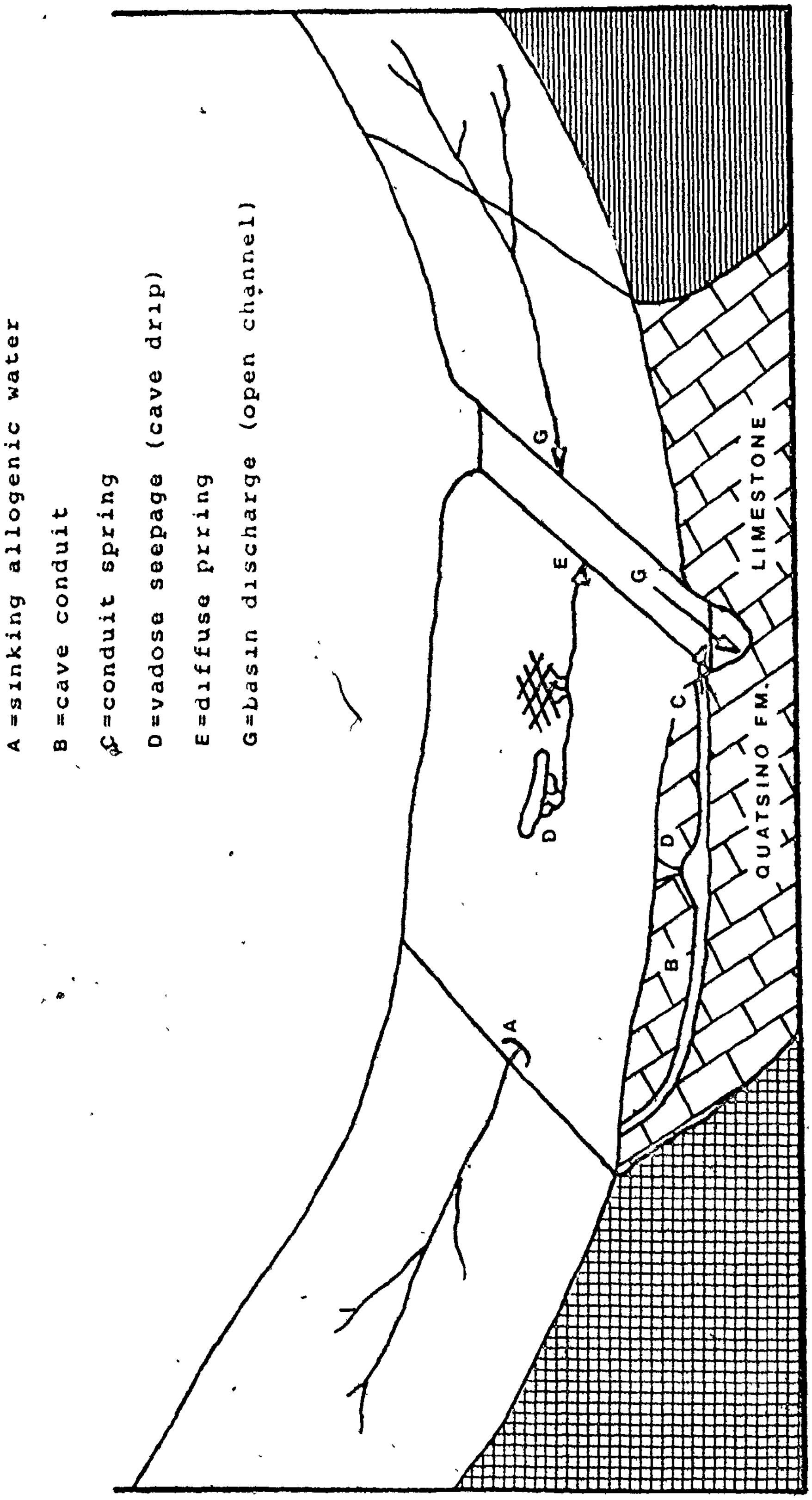


FIGURE 3.3: Conceptual hydrogeological environments in the Benson Karst

TABLE 3.4

Hydrogeologic group chemistry

		T	pH	TH ($mgl^{-1}CaCO_3$)	HCO_3^-	SIc	SI_D	log P_{CO_2}
llogenic sinks	$\bar{x}$	9.9	7.5	35.3	74.1	-1.47	-1.64	-3.04
	SD	2.5	0.67	12.6	25.4	0.91	0.87	0.57
ave conduits	$\bar{x}$	9.7	7.5	55.0	109.2	-1.03	-1.32	-2.65
	SD	3.4	0.35	13.4	28.1	0.24	0.26	0.43
onduit springs	$\bar{x}$	11.5	7.5	52.8	124.3	-1.04	-1.31	-2.59
	SD	3.4	0.36	29.6	63.8	0.30	0.25	0.50
iffuse springs	$\bar{x}$	9.0	7.5	133.7	253.8	-0.33	-0.76	-2.15
	SD	1.47	0.48	45	76.6	0.45	0.46	0.55
ave drips	$\bar{x}$	13.0	7.1	172.0	340	-0.49	-0.9	-1.75
	SD	2.8	0.28	79.2	151.3	0.04	0.36	0.50
asin discharge	$\bar{x}$	13.5	7.27	46.3	92.86	-1.43	-1.65	-2.11
	SD	2.6	0.67	10.56	21.38	0.62	0.55	0.65

3.7.2 Allogenic Stream Waters

These waters are derived from precipitation incident on the surrounding non-limestone uplands, sampled at the limestone boundary. Characteristically, the group has the lowest total hardness ($\bar{x}$ = 35.3 mgl^{-1}) and is distinctly unsaturated w.r.t. calcite ($\bar{SI}_C$= -1.47) and dolomite ($\bar{SI}_D$= -1.64). Group mean PCO_2 ($\bar{x}$ = $10^{-3.04}$ atm) indicates limited addition of soil CO_2. Streams flowing off the magnesium rich lavas to the East have Ca/Mg ratios that differ markedly from those of streams flowing off Parson Bay lithologies to the West of the karst belt (Table 3.5).

3.7.3 Cave Passages

This group comprises waters derived dominantly from sinking allogenic streams. The concentrated nature of the inputs to and the short residence time of waters within the conduits results in limited cave passage solution (TH = 55.0mgl^{-1}, SI_C = -1.03). The PCO_2 of the waters ($\bar{x}$ = $10^{-2.65}$) is slightly greater than that of the allogenic streams, reflecting limited addition of authigenic percolation waters previously exposed to soil air.

3.7.4 Cave Conduit Springs

The conduit springs discharge allogenic stream waters sinking near the geologic boundary and routed through the karst via large conduits. Five of the six studied conduit springs were connected to stream sinks by dye traces. Measured total hardness in this group

TABLE 3.5

Allogenic stream Ca/Mg ratios

	$\overline{x}$	sd.	n
Allogenic stream off Karmutsen Fm. lavas	8.88	0.49	7
Allogenic stream off Parson Bay sediments	1.43	0.69	3

($\bar{x}$ = 52.8mgl^{-1}) is similar to group 2 above, although the groups are not directly comparable because intervening cave conduit samples are not available for waters at two of the springs. The data suggests limited solution in the conduits along the mean straightline flow path length of some 960m between sink and rising point despite the aggresive nature of the spring water ($\overline{SI}_C$ = -1.04, $\overline{SI}_D$ = -1.31). A Mann-Whitney U test was employed to test the hypothesis that allogenic waters routed through the karst via large conduits exhibit little evolution towards saturation. (see Table 3.6.) No significant difference in group mean SI_C data was apparent at the 0.05 significance level which suggests that extensive solution does not occur along the underground conduit flow paths.

3.7.5 Cave Drip (Vadose Seepage) Waters

The chemical properties of percolation waters measured in two caves at different depths beneath different soil types are presented in table 3.7. The data demonstrates the CO_2 boost obtained by passage through the soil base or ground air zone. The high total hardness ($\bar{x}$ = 172mgl^{-1}) demonstrates considerable solution by percolating waters in the soil or upper region of the vadose zone.

3.7.6 Diffuse Flow Springs

Diffuse flow springs discharge autochthonous percolation waters routed in networks of small passages. These waters are character-

TABLE 3.6

Mann-Whitney U Test of allogenic stream and conduit spring water SIc data

	SINKING STREAM		CONDUIT SPRING
	SIc		SIc
	-1.34		-0.76
	-1.33		-0.76
	-0.94		-1.32
	-1.16		-1.00
	-0.68		-0.39
N=9	-2.21	N=11	-1.31
	-3.62		-1.36
	-1.07		-1.07
	-0.99		-1.17
			-1.17
			-1.18

H_o=no significant difference between groups. Significance level=0.05
U=33, critical U=23, cannot reject H_o.

TABLE 3.7

Cave drip (vadose seepage) chemical characteristics

SITE	HAPHAZARD CAVE	BATH CAVE
SOIL	organic	thin till
SOIL P_{CO_2}	$10^{-2.98}$	$10^{-2.24}$
DEPTH TO CAVE	5m	15m
CAVE DRIP P_{CO_2}	$10^{-2.24}$	$10^{-1.53}$
TOTAL HARDNESS	116 mgl^{-1}	228 mgl^{-1}
SIc	-0.52	-0.46

ised by high total hardness ($\bar{x}$ = 133.7mgl^{-1}) and considerable evolution towards saturation ($\overline{SI_c}$ = -0.33) reflecting longer subsurface residence times. The mean temperature of the diffuse spring group is 9.0 °C. However, several measurements were obtained with the SPC thermistor and are only accurate to ± 2°C. The mean groundwater temperature derived by thermometry is about 7.0 °C. A Mann-Whitney U test was employed to test the hypothesis that diffuse spring waters routed through the karst in networks of small passages have a significantly different saturation status w.r.t. calcite than the spring waters routed in large conduits (Table 3.8). A significant difference was apparent at the 0.05 level, which demonstrates that the chemical character of the groups reflect their different flow pathways.

3.7.7 Basin Discharge

This group comprises samples from the base level Benson River, its principal tributary (the Raging River), and smaller tributaries flowing off the Parson Bay Fm. lithologies at their junctions with the Benson River. Group mean SI_c ($\bar{x}$ = -1.43) and total hardness ($\bar{x}$ = 46.3mgl^{-1}) reflect both rapid flow through the karst in discrete channels and the small proportion of the total catchment area of the Benson River Valley that is composed of carbonate lithologies (less than 10%). Like the sinking streams on the eastern side of the valley (Groups 1, 2 & 3) the chemistry of the surface streams flowing off the Parson Bay Fm. lithologies to the west demonstrates only limited evolution towards saturation prior to confluence with the Benson River (Table 3.9).

TABLE 3.8

Mann-Whitney U Test of conduit flow and diffuse flow spring SIc data

	CONDUIT SPRING		DIFFUSE SPRING
	SIc		SIc
	-0.76		-0.12
	-0.76		+0.09
	-1.32		+0.25
	-1.00		-0.53
N=11	-0.39	N=7	-0.27
	-1.31		-1.01
	-1.36		-0.69
	-1.07		
	-1.17		
	-1.18		
	-1.17		

H_o=no significant difference between groups. Significance level=0.05
U=6, critical U=16, reject H_o

TABLE 3.9

Solution in surface streams crossing karst

		SI_C	SI_D	log P_{CO_2}	$CaCO_3$ (mgl^{-1})
YOOTOOK CREEK	above karst	-2.21	-2.36	-2.87	13
YOOTOOK CREEK	at Benson River	-1.73	-2.11	-2.08	43
CRAFT CREEK	above karst	-1.07	-1.30	-3.51	21
CRAFT CREEK	at Benson River (below karst)	-0.51	-0.77	-3.40	38
PINCH CREEK	above karst	-3.62	-3.67	-1.59	11
PINCH CREEK	at Benson River	-2.53	-2.45	-1.42	25

3.8 Conduit System Hydrochemistry

3.8.1 Hydrochemical Facies

The Mann-Whitney U test described in table 3.6 demonstrated that significant changes in saturation status do not occur between waters entering and leaving the conduit cave system. This suggests that there was little solution within the conduits during the sampling season and no significant addition of more saturated percolation waters to the conduits.

However, total hardness and Ca/Mg ratio data from one cave system demonstrate that solution does occur along the conduit flow paths. The Malook Creek cave system routes two streams from the Karmutsen Fm. lavas through the karst to the Benson River over a straightline distance of approximately 1.1km. The Karmutsen Fm. lavas are composed of high magnesium basalts, having a Ca/Mg ratio of approximately 1.45 (Kuno 1968). Analyses by McCammon (1969) show that the magnesium content of the Quatsino Fm. limestone can be as low as 1.15% by volume.

The mean Ca/Mg ratio of stream waters entering the Malo k Creek cave system is 1.73. Water samples from the two stream sinks close to the geologic boundary, three karst windows along route, and the system resurgence demonstrate a progressive increase in the Ca/Mg ratio along the conduit passage (Table 3.10, Column A). This demonstrates solution of low magnesium limestone in the cave passage and possibly

TABLE 3.10

Transfer of stream water from high Mg lavas through low Mg karst demonstrating solute pickup in karst conduits.

SITE	A Ca/Mg		B SIc/SI_D	
Malook Sink I & Malook Sink II	1.51	2.07	0.916	0.848
Top Window	2.21		0.855	
Eternal Fountain	2.83		0.815	
Steps Cave	3.14		0.783	
Malook Resurgence	3.02		0.784	

in the soil or vadose zone by percolation waters. (Sufficiently accurate discharge measurements are not available to determine whether percolation waters are added to the cave stream.)

It follows from the changing Ca/Mg ratio that the relative rates of evolution towards saturation w.r.t. calcite and dolomite change once waters pass from the high magnesium facies into the karst. This factor is demonstrated by the downstream decoupling of the ratio SI_C/SI_D (Table 3.10, Column B).

The mean increase in total hardness between the stream sink and spring of the Malook Creek cave system, based on twice weekly measurements was $36 mgl^{-1}$. The mean discharge from the system during the sampling season was about 0.25 $m^3\ s^{-1}$. Thus the mean dry season rate of limestone removed from the system is about 0.8 t d^{-1}.

3.8.2 Total Hardness Variation in the Malook Creek Cave System

Research in Pennsylvania, U.S.A. (Shuster and White, 1971) and Yorkshire, England (Teman, 1972) has shown that conduit flow systems may be differentiated from diffuse flow systems by consideration of spring hydrological and chemical data. In particular these authors employed the coefficient of variation of hardness V (where $V = (S.D/\bar{x}) \times 100$) to separate spring waters. In both regions springs classified as diffuse flow systems had mean V values of approximately 5% with maximum values of less than 11%, whereas V in a conduit flow system ranged between about 10% and 25%.

Total hardness determinations of waters in the Malock Creek cave system were made twice weekly between June 26 and August 14 1980 in an attempt to test this method of classification. The determinations were made from SPC_{25} data employing the relationship defined in equation 1.

The coefficient of variation of total hardness for the Malook Creek cave sites are presented in Table 3.11. The data exhibit three important features. Firstly, the value of V for the Malook Creek cave system spring occurs within the range of conduit springs in Pennsylvania and Yorkshire. Dynamic variation is evident in the data despite restriction of sampling to a single season whereas water year data was available in the other regions.

Secondly, the availability of data from intermediate sites along the conduit flow path, compared to spring data alone, demonstrates a marked reduction in V downstream. This is believed to be a consequence of phreatic storage.

Thirdly, the coefficient of total hardness variation, coupled with the hydrochemical facies data in Table 3.11 presents strong evidence that there is an addition of less saturated, high magnesium allogenic stream water to the Malook Creek cave system between the Steps Cave and Malo k Creek resurgence. These two sampling points are about 60m apart. The water is probably derived from a stream which sinks at the Karmutsen Fm. lava contact approximately 500m east of the spring.

TABLE 3.11

Coefficient of Total Hardness Variation (V) at different sites in the Malook CREEK CAVE SYSTEM

SITE	V
Malook Sink I	24.47%
Top Window	14.09%
Eternal Fountain	13.92%
Steps Cave	7.38%
Malook resurgence	16.37%

The flow retains a stronger allogenic character at its confluence with the Malook Creek cave system because of its shorter flow path in the limestone.

3.9 Statistical Analysis of Conceptually Defined Hydrochemical Environments

The water classes outlined in sections 3.7.1 to 3.7.7 were derived conceptually from a consideration of hydrogeologic factors. As such the groups were derived to reflect variations in hydrologic flow paths or positions along these flow paths. This section attempts to test the validity of this separation employing group hydrochemical data. The basis of this assessment is the hypothesis that the above mentioned hydrogeologic variations should exhibit variations in chemical character.

The conceptually defined hydrogeologic classes are presented in Table 3.12 (Column A). Corresponding hydrochemical groupings are presented in column B. The Mann Whitney U test in table 3.6 demonstrated that it was not possibly to statistically separate sinking allogenic waters at different stages along the conduits on the basis of SI_c. Thus these three classes are collapsed into a single hydrochemical grouping "Conduit Water". The hydrological class "diffuse spring water" has been separated into two hydrochemical groupings; "stable diffuse water" and "unstable diffuse water" since a subset of this group (3 samples) have an altered saturation status believed to result from degassing of CO_2 (Table 3.13).

TABLE 3.12

Conceptually defined hydrogeologic environments and corresponding hydrochemical water classes.

HYDROGEOLOGIC ENVIRONMENT		HYDROCHEMICAL CLASS
stream sink	→	conduit waters
cave passage	→	conduit waters
conduit spring	→	conduit waters
diffuse spring	→	stable diffuse
diffuse spring	→	unstable diffuse (degassed)
cave drips	→	cave drip (vadose seepage)
basin discharge	→	basin discharge

TABLE 3.13

Chemistry of 'stable' diffuse spring waters and diffuse spring waters believed to have undergone extensive degassing of CO_2.

	STABLE	DEGASSED
NO. OF CASES	7	3
TEMPERATURE	9.0	10.6
pH	7.5	8.3
TH (mgl^{-1} $CaCO_3$)	133.7	156.3
HCO_3^-	253.9	286.7
SIc	-0.33	+0.52
SI_D	-0.76	+0.34
log P_{CO_2}	-2.46	-3.14

Stepwise linear discriminant function (LDF) analysis was employed to test the significance of the a priori hydrochemical groups and determine the most powerful discriminating variables. The analysis was performed by the S.P.S.S. program using the smallest Wilks lambda method (Kleck 1975). The independent ability of four measured (temperature, pH, TH, HCO_3^-) and three calculated (SI_C, SI_D, log PCO_2) variables to separate the groupings are summarised in table 3.14.

Initially all variables were entered into LDF analysis. Their order of entry is listed in Table 3.15, column A. This yielded only 25.4% misclassification of cases. However, it is difficult to interpret the hydrochemical significance of the discriminant functions and the variables are not functionally independent.

The LDF analysis was rerun using only the calculated variables SI_C, SI_D, log PCO_2. Group separation is easier to interpret hydrochemically since the variables are all logarithmetic, functionally independent, and between them summarise both the constraints on, and extent of, the solution process. The order of entry of the variables is presented in Table 3.16 column B. A chi-square test of significance was employed to test the hypothesis that no significant difference in missclassification by the groups 'all variables' and 'calculated variables' is apparent at the 0.005 level (Table 3.16). No significant difference was apparent in the analysis.

The measured variables were entered into LDF analysis to

TABLE 3.14

Variance in hydrochemical group data

VARIABLE	F-RATIO
T	2.62
pH	2.85
TH	33.02
HCO_3	28.01
SIc	11.94
SI_D	12.22
$\log P_{CO_2}$	3.35

TABLE 3.15

Order of entry of variables into L.D.F. analysis

STEP	A ALL VARIABLES	B CALCULATED	C MEASURED
1	TH	SI_D	TH
2	TH, log P_{CO}	SI_D, log P_{CO_2}	TH, pH
3	TH, log P_{CO_2}, T	SI_D, log P_{CO_2}, SIc	TH, pH, T
4	TH, log P_{CO_2}, T, SI_D		TH, pH, T, HCO_3^-
5	TH, log P_{CO_2}, T, SI_D, SIc		
6	TH, log P_{CO_2}, T, SI_D, SIc, HCO_3^-		

Not entered pH

	A	B	C
No. of misclassified cases	14 of 55	18 of 55	16 of 55
% misclassification	25.5	32.7	29.1

TABLE 3.16

Chi-Square Test of Significance in difference of misclassification by the groups "calculated variables" and "all" variables.

	CORRECT	INCORRECT
CALCULATED VARIABLES	37	18
ALL VARIABLES	41	14

Ho=no significant difference in misclassification by groups at 0.005 level

$\chi^2 = 0.446$ critical $\chi^2 = 7.88$. Cannot regect H_o.

TABLE 3.17

Chi-Square Test of Significance in difference of misclassification by the groups "measured" and "calculated" variables.

	CORRECT	INCORRECT
MEASURED VARIABLES	39	16
CALCULATED VARIABLES	37	18

Ho=no significant difference in misclassification at 0.005 level

$\chi = 0.19$ critical $\chi^2 = 7.88$. Cannot regect H_o.

determine the statistical ability of these variables to separate the a priori groups (see Table 3.15, column C). Misclassification of samples employing these variables is 29.1%. The two variables with the highest F-ratios, TH and HCO_3^-, occur in this group. However, the amount of misclassification (29.1%) is not significantly less than that of the calculated variables (see Table 3.17). The results from the relationship M TH = M HCO_3^- / 2 that exists between pH 6.5 and 8.0 in simple limestone groundwater systems. Thus little added discriminatory power is added to the analysis with inclusion of the second variable of the pair.

The a posteriori reclassification of the hydrochemical groupings, employing the calculated variables SI_c, SI_D and log PCO_2, is presented in Table 3.18. The misclassification (overlap) between the a priori groupings evident in the reclassification reflect the spatial hydrologic and hence chemical continuums that exist in karst drainage systems. Most misclassification in Table 3.18 occurs between groups 1 and 5. This is a reflection of the fact that both groups are dominantly composed of allogenic stream waters routed through the karst in discrete channels via either cave conduits or surface rivers.

The separation of hydrochemical groups achieved by LDF analysis can be summarised graphically employing the computed variables SI_c and log PCO_2 (figure 3.4). The hydrochemical group "conduit waters" has been broken down into its three component hydrogeologic classes. The graph clearly illustrates both the hydrochemical status of the different waters and relates these to hydrological routing through the

TABLE 3.18

A Posteriori Classification

ACTUAL GROUP	I.D.	NO. OF CASES	PREDICTED GROUP MEMBERSHIP 1	2	3	4	5
Allogenic Conduit	1	36	25	1	0	9	1
Stable Diffuse	2	7	0	4	1	0	2
Degassed Diffuse	3	3	0	0	3	0	0
Basin Discharge	4	7	3	0	0	4	0
Cave Drips	5	2	0	1	0	0	1

Percent of grouped cases correctly classified 67.3%

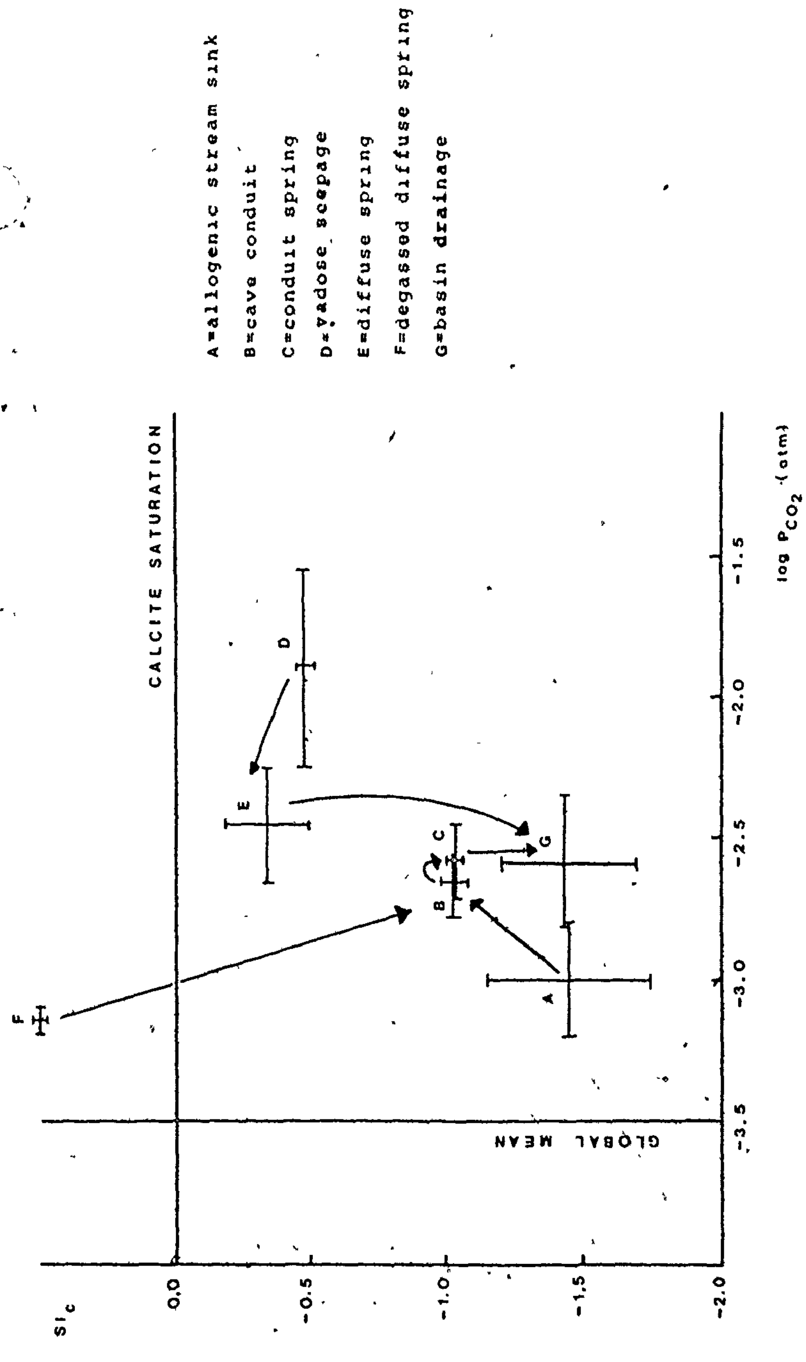


FIGURE 3.4: Geochemical relationships between the 7 water types studied. Points indicate mean values, bars denote standard error of the mean.

karst. Although routing may not always appear to be towards the saturation line, all unsaturated waters are evolving towards saturation with passage through the karst, only the effect of mixing of different waters may mask this trend. Therefore the arrows do reflect the movement of waters along both hydrologic and hydrochemical pathways.

3.10 Conclusions

Characteristics of the solution process in the Quatsino Fm. karst are outlined in chapter 3. The principal conclusions of this study are:

1. Support is given to the concept of climate control of the equilibrium groundwater solute concentrations in the karst by its control on biogenic CO_2 production (Drake and Wigley 1975). The PCO_2 in the top 30cm of the soil is less than the PCO_2 implicit in hydrochemical analyses, thus higher PCO_2 environments than those directly measured determine the solution potential. The relative importance of soil v groundair CO_2 production cannot be evaluated from the present data set.

2. Conduit flow and diffuse flow type systems can be separated by analysis of the saturation status of discharged waters.

3. The flow of water from the lavas into the limestone is reflected by changing Ca/Mg and SI_C/SI_D ratios.

4. Stepwise linear discriminant function analysis of hydrochemical data generally supports the delimitation of a priori hydro-

geologic environments (32% misclassification). 67% of the original misclassification is between two classes; cave conduits and basin discharge. This reflects the fact that both of these environments route allogenic stream waters through the karst in discrete channels which results in similar chemical signatures. No significant increase in misclassification occurred when only the 3 calculated variables (PCO_2, SI_C, SI_D) were entered into the analysis. Separation by this group was chosen in place of all the variables available because the calculated variables are functionally independent and betweem them summarise the constraints on the solution process in a more easily understood manner.

CHAPTER 4

SURFACE GEOMORPHOLOGY

4.1 Introduction

The objective of this chapter is to describe aspects of the surface geomorphology of the Quatsino Fm. karst in the Benson Valley and on the Gibson Plateau. The study employs the classification of Ford and Quinlan (1973) which subdivides features into groups according to scale; microforms - less than 10m in their greatest dimension; mesoforms - 10 to 1000m in their greatest dimension; and macroforms - over 1000m. Although these subdivisions are arbitrary they tend to fit observation. In the Quatsino Fm. the form of macroscale features is largely controlled by geologic structure and glacial processes; micro and mesoscale features are more obviously the products of solution.

The nature of this analysis of surface landforms is descriptive; only a qualitative relationship between process and form is presented. This constraint arises from a general difficulty in defining quantitative relationships between processes and form in landscapes.

4.2 Benson Valley Karst

The physiography of the Benson Valley was outlined in chapter 1.2. The outcrop of Quatsino Fm. limestone in northern Vancouver Island dominantly occurs along valley bottoms. This indicates preferential erosion of the unit since the uplift of the region during the Early Jurassic.

4.2.1 Macroforms

The physiography of the Benson Valley is dominated by the cuestas which rise up to 400m above the valley floor. Cuesta form is strongly influenced by geologic structure; steep scarp faces occur on the north-east side, gentler dip slopes on the south-west. Landform development on the cuestas is immature in comparison to the Gibson Plateau surface. This results from more extensive, and perhaps more frequent, glacial erasure in the valley situation. The geomorphology of the cuestas is dominated by shallow, principally strike oriented, dry valleys and closed depressions.

A sedimentary terrace, the Pinch Creek Delta, several km long occurs along the south-east shore of Alice Lake, down valley from the cuestas. At the southern or upvalley end the deposit exceeds 30m in depth and exhibits coarsening upwards from a silty-clay base into coarse cobbly beds. About 1km to the north (down valley) the terrace is composed of clays. The terrace is apparently a deltaic deposit built by waters flowing downvalley. It rises upto 50m above the present lake surface and may result from late glacial sedimentation. Because the deposit is above the confluence with the major Marble River Valley, sedimentation may have occurred in an ice-dammed lake. It is unknown whether the terrace is pro-glacial or sub-glacial in origin but the water was probably derived from stagnant downwasting ice masses.

There are three categories of active stream channels in the karst. The first channels water, mainly from the Karmutsen Fm. contact, to stream sinkholes close to the geologic contact. The second routes water down valley sides into the Benson River at local base level. These streams have incised vertical walled gorges up to 15m in depth (photo 4.1). Factors which control whether class one or two development occurs are examined in Chapter 5.1. The principal variable is not the size of the allogenic catchment area. The third class comprise the Benson River and its tributory, Raging River, flowing along the valley bottom. Up to 10km of the Benson channel in the study area occurs in gorge sections which attain 20m in depth. These would appear to be of late or post-glacial origin. The orientation of the second and third classes of channel is strongly controlled by joints, fractures and dykes in the limestone.

Dry valleys in the Benson Valley may be divided into two classes based on the apparent source of erosive agent. Class one are those which occur downslope from active or fossil streamsinks and were by stream water flowing off adjacent non-karst strata. These dry valleys are concentrated on the eastern side of Benson Valley because most allogenic streams which flow onto the karst from the Parson Bay Fm. remain on the surface (see Chapter 5.1). A notable exception occurs where the Merry Widow Creek sinks on a fault (Figure 4.1) on the west side of the valley.

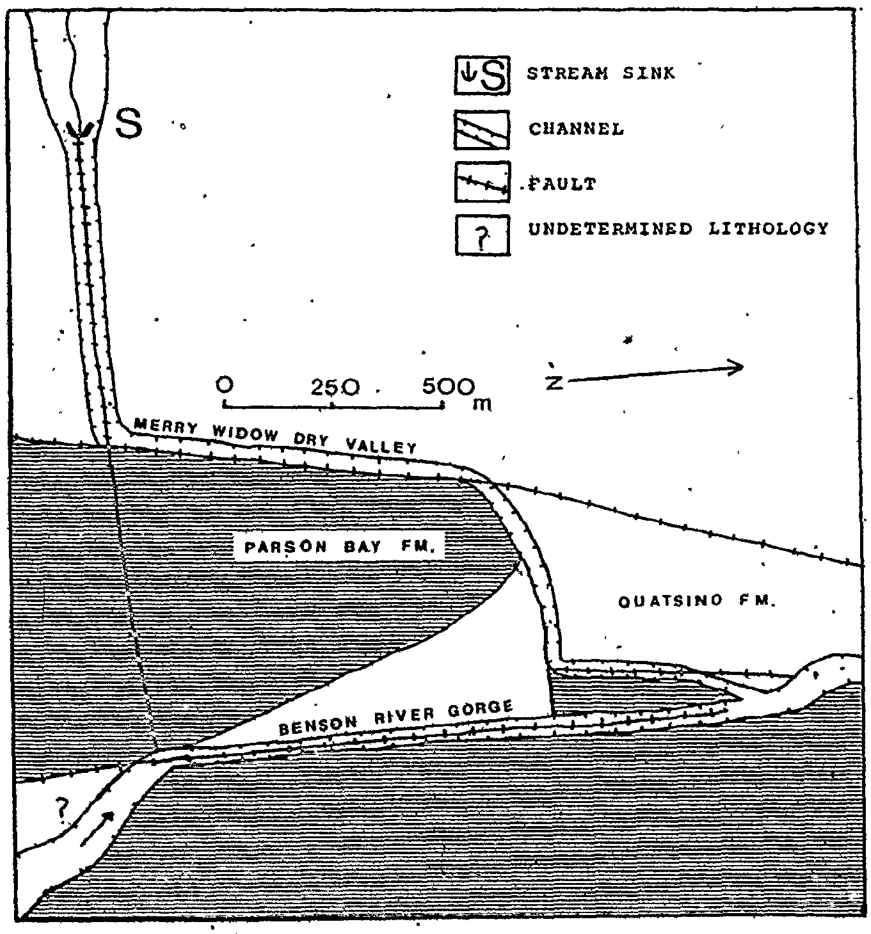


FIGURE 4.1: Influence of faulting on stream channel development

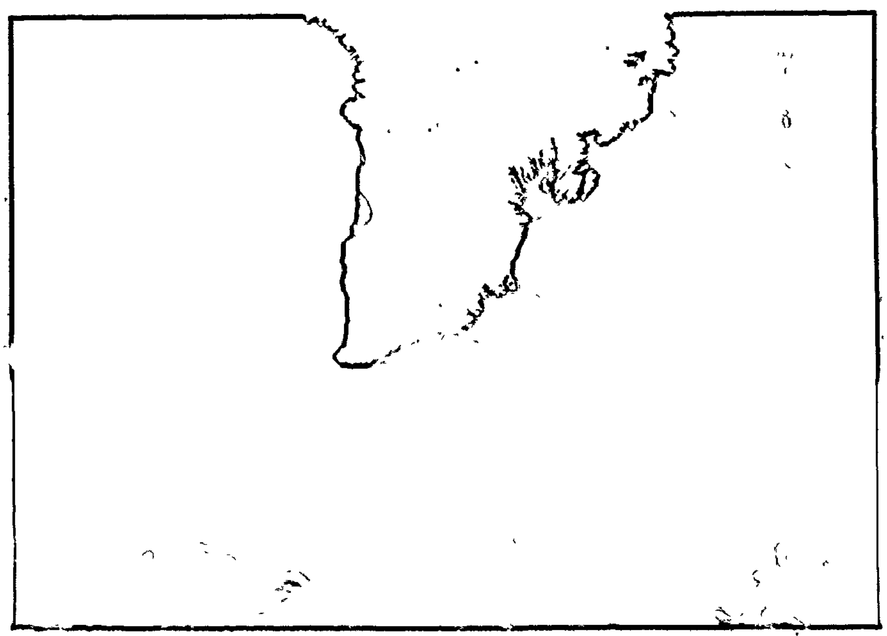


PHOTO 4.1: Yootook River gorge.

Class two dry valleys are believed to result from processes not active in the present landscape - sub-glacial melt and possibly periglacial incision. These features occur primarily on the cuestas in positions unlikely to be fed by stream water in the postglacial landscape. They do not contain stream sinkholes. The largest dry valleys occur upvalley of the Pinch Creek Delta near the Malook Creek cave system (see figure 5.4). Complex forms occur where class one and class two types apparently occur in one dry valley system.

4.2.2 Mesoforms

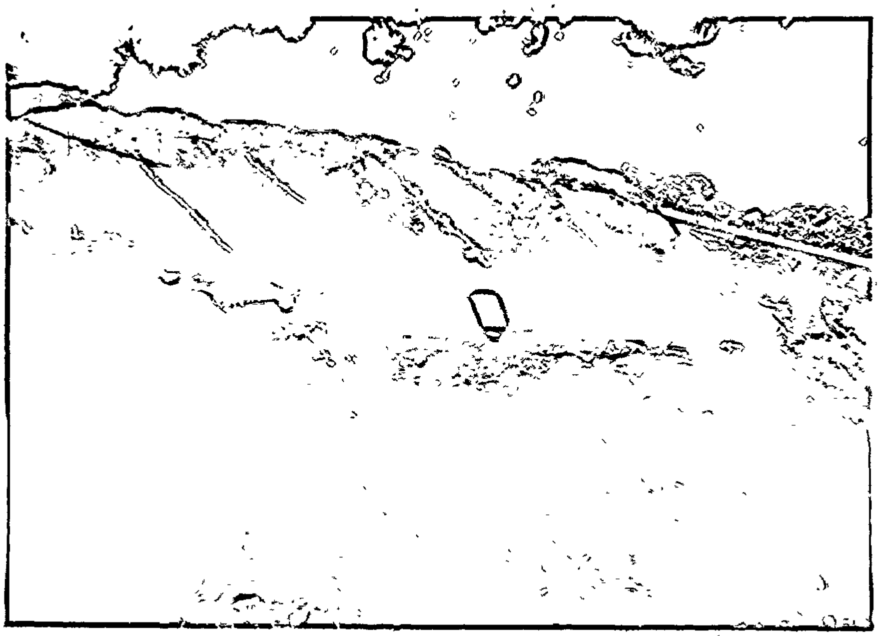
This group is dominated by dolines - i.e. closed depressions which route surficial water centripetally. Doline development is controlled principally by five processes (Quinlan 1973) which results in a large range of size and form. The morphogenetic processes believed to control doline development in the Benson Valley are solution and collapse. Solution dolines are concentrated along active or dry valley thalwegs. Downslope runoff apparently predominates over centripetal drainage on the steep gradient interfluves. Collapse dolines occur above large cave passages which route allogenic stream water. These result from cave roof collapse and are found primarily in locations where cave waters have risen up "joint chimneys" (Ford 1968). Thus they are less abundant than the solution type feature. Four collapse dolines occur along less than 1km of the Malook Creek cave system (Chapter 5.4). The largest collapse doline in the region, the Devil's Bath, measures approximately 80m x 60m in plan dimensions. It is some 45m deep and results from the removal of over $1 \cdot 5 \times 10^5$ m^3 of rock.

This vertically walled feature is similar in form to the cenotes of the Yucatan Peninsula, Mexico.

Limestone pavements are generally believed to be karstiglacial (Ford 1979) features, formed by glacial scour of limestone surfaces which are subsequently modified by solution processes (Sweeting 1972). They are poorly developed (or preserved) outside the limits of ice cover of the last glaciation.

Extensive planar pavement surfaces, typical of the Burren, Ireland; Yorkshire, England; or the European Alpine karst do not occur in the Benson karst. This results from the steep stratal dip. The form of the scoured surfaces, eg. the cuestas, are parallel series of small mesoscale ridges or cuestas, and troughs. The steep stratal dip has also prevented development of the step and terrace relief (Schichthreppenkarst, Bogli 1964) typical of near-horizontal strata in upland glaciated karsts.

The proportion of the Benson Valley karst developed into limestone pavement is small. This is because the depth of the till overlying the limestone surface has been sufficient to absorb the postglacial solution attack in most locations. At one wooded, free draining situation on carbonate tills a depth of less than 2m has been sufficient to preserve glacial striae and channels on the limestone surface (Photo 4.2). Within 10m of this feature and elsewhere on the Quatsino Fm. bedrock solution forms were observed beneath till cover now less than 1m in depth. A similar critical depth is reported on the Bruce



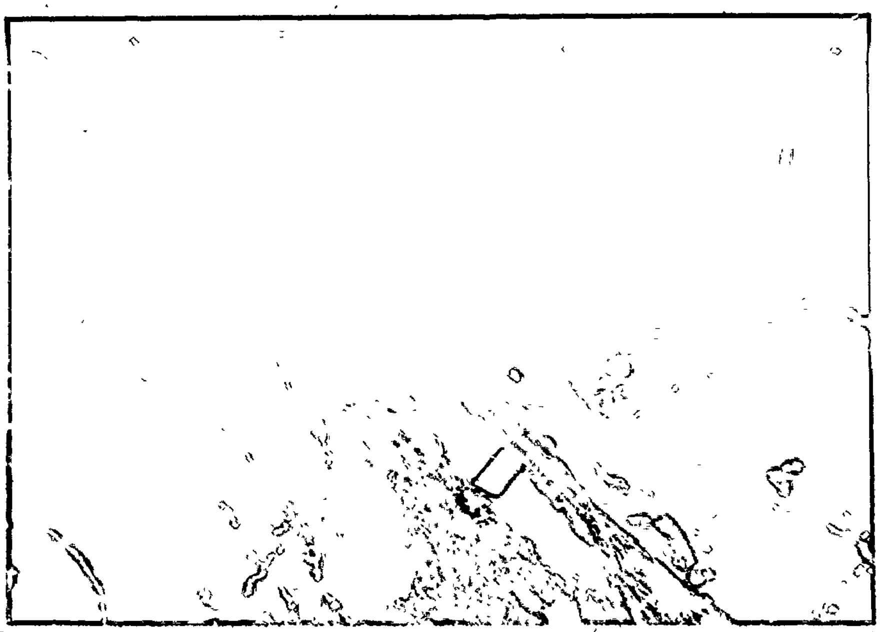


PHOTO 4.2A and 4.2B: Glacial striae and channels on recently exposed limestone surfaces.

Peninsula (Cowell 1977). However, control of till carbonate leaching and bedrock erosion rates by local factors, eg. soil drainage conditions and vegetation, have been demonstrated by Trudgill (1976). Moreover, deckenkarren develoment is possible beneath a carbonate cover if plant roots occur at the base of the profile. The critical till depth range over the Quatsino Fm. that is a consequence of these factors is undetermined.

4.2.3 Microforms

Limestone surfaces are complex features when examined in detail. They usually contain a variety of small scale solution forms collectively known as karren or lapies. The prime controls of karren morphology are geologic structure, especially angle of dip and degree of jointing; lithology; past and present chemical and biological processes; the nature (or absence) of a vegetation or soil cover; and climate (Sweeting 1972). Studies of karren are numerous but most are restricted to being contributions towards complex typologies of form; only a few have increased understanding of the genetic processes involved (eg. Bogli 1960, Glew and Ford 1980).

Much of the research on karren development has been carried out by Bogli who wrote (1980): "The multiplicity of possible karren forms makes a morphological system endless, while a genetic one allows a meaningful collection... a differentiation is made between karren forms which are created by free, unhindered water flowing off over bare limestone surfaces, those which are caused by a patchy covering of soil, and those which develop beneath a closed covering of soil". Yet even this recent classification does not consider some of the above mentioned controlling variables.

The purpose of this section is to describe the karren development at two contrasting locations in the Benson karst: (1) on steep, bare, surfaces, (2) in river gorges. The wide variety of pavement and karren forms on the Bruce Peninsula described by Cowell (1977) were not observed in the Benson karst. This may result from the greater uniformity of conditions, ie steep slope and uniform lithology, in the latter region.

<u>Karren development on steep limestone surfaces</u>: The most spectacular karren development in the Benson karst occurs on the bare small scale cuesta surfaces. These surfaces are believed to have been soil-covered until deforestation within the last 100 years since the onset of mining and lumbering activities in the region. The former depth of soil was sufficient to support large trees (Photo 4.3). The extent of subsoil karren development prior to deforestation is unknown; the present forms are sharp and rough to the touch, typical of subaerial development, but may result from modification of sub-soil forms.

The slope of the bare limestone surfaces examined range from 30^{o} - 40^{o} from the horizontal. The down slope length of the surfaces range from about 5 - 15m. A common karren form are sharp faced, sub parallel, run off channels which have a mean depth of 5cm, mean width

PHOTO 4.3: Evidence of former forest cover on steep limestone surfaces.

of 6cm and reach up to 7m in length (Photo 4.4). Linkage of channels occurs in some locations where a channel segment is developed along aslant strike joints. The channels are similar to Wandkarren (Bogli 1980) and do not resemble rillenkarren or rinnenkarren, which are the other common sub-aerial channel types. Bogli ascribes wandkarren development to subaerial solution by water derived from upslope snow-melt or soil patches. Snowmelt is unlikely to be important in the Benson situation; however, deep soil pockets may have occurred in the strike aligned troughs between the cuesta dip slopes.

On some surfaces aslant strike joints have been enlarged by solution into deep grikes or kluftkarren. In these locations wandkarren development occurs only on the upslope side of the grikes. Wandkarren development is inhibited where the frequency of solutionally enlarged joints is high. On some surfaces the maximum downslope distance overwhich water flows to enter a joint is about 6cm (Photo 4.5).

Soil erosion is an almost immediate response to deforestation on the steep surfaces; areas strip-logged and burnt exhibit bedrock outcrops in as short a time period as one year. This contrasts with deforestation on low gradient surfaces (eg. the Burren, Ireland) where soil erosion is dependant on the development of solutional sub-soil transfer routes.

River Gorge Karren Extensive karren assemblages occur along sections of the Benson River and Raging River gorges. The dominant forms are

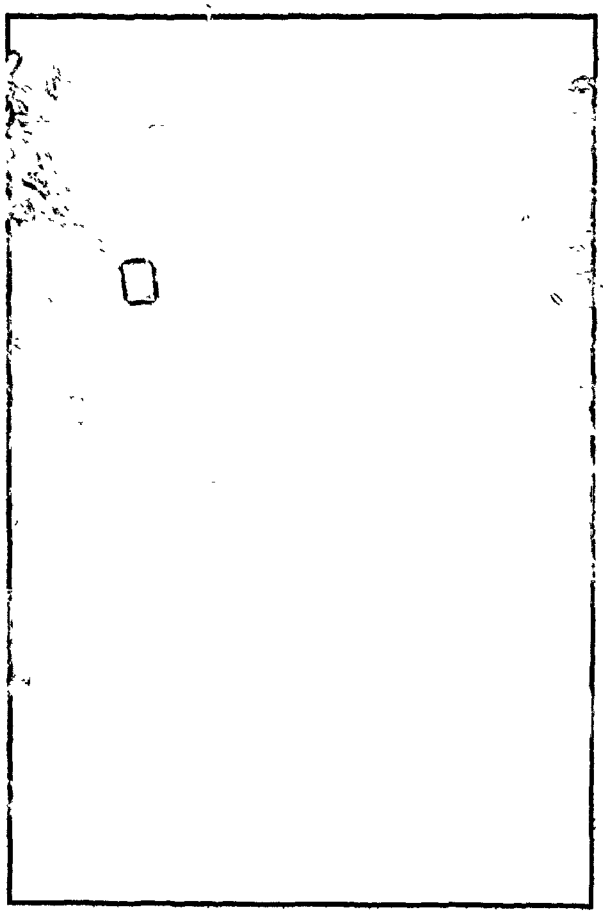


PHOTO 4.4: Wandkarren

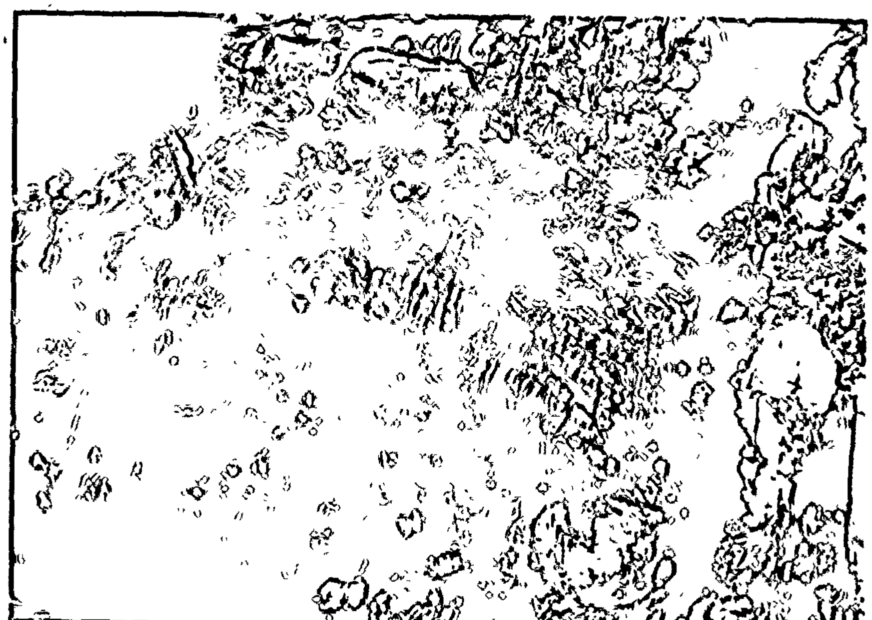


PHOTO 4.5: High frequency of solutionally enlarged joints.

pits and potholes up to 1m in depth. All forms appear to have developed free of a soil or vegetation cover.

The gorge karren have been divided into three types depending on the main source of erosive water: karren formed by authigenic seepage or drip water down the sides of the gorge, karren formed by river water, and those of mixed origin.

(1) Karren formed by authigenic water are either small pits or kamenitza, typically less than 20cm wide (Photo 4.6), or small channels similar to rinnenkarren. The kamenitza frequently occur beneath overhanging walls and are formed by drip waters. Both forms are produced by soil water or bedrock seepage water. The effect of incident precipitation is probably small.

(2) The principal karren form produced by stream water are potholes up to 1m deep (Photo 4.7 A) Stream potholes in limestone probably result from both corrasion and corrosion processes (Ford 1965). (Most contained clastic sediments of upto cobble size (Photo 4.7B)) Those examined were either fossil or seasonally inactive features. Many of the high level features have been breached, probably as a result of stream incision.

(3) This class comprises former stream potholes modified by solution by anthigenic waters. This form only occurs at the gorge margins, in a few locations.

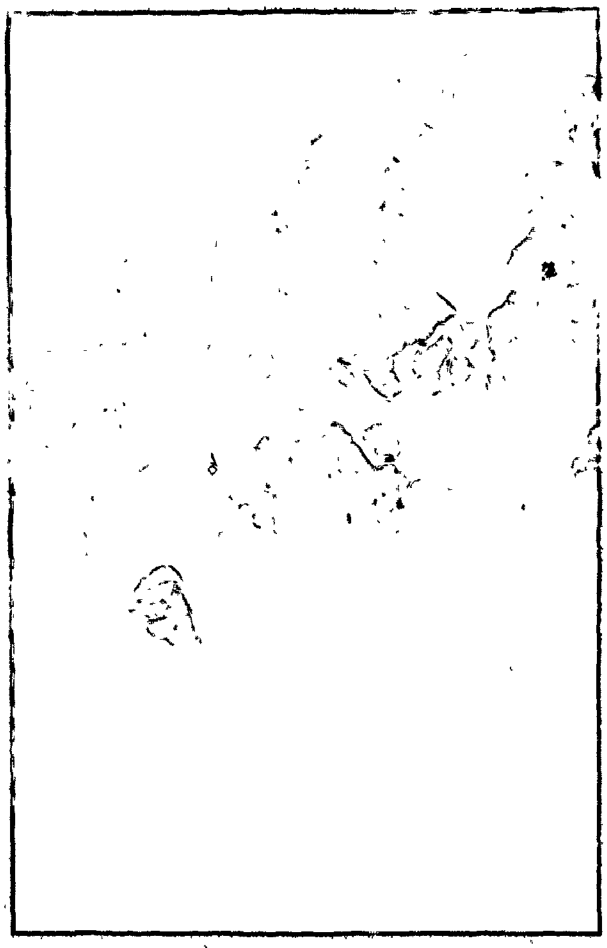
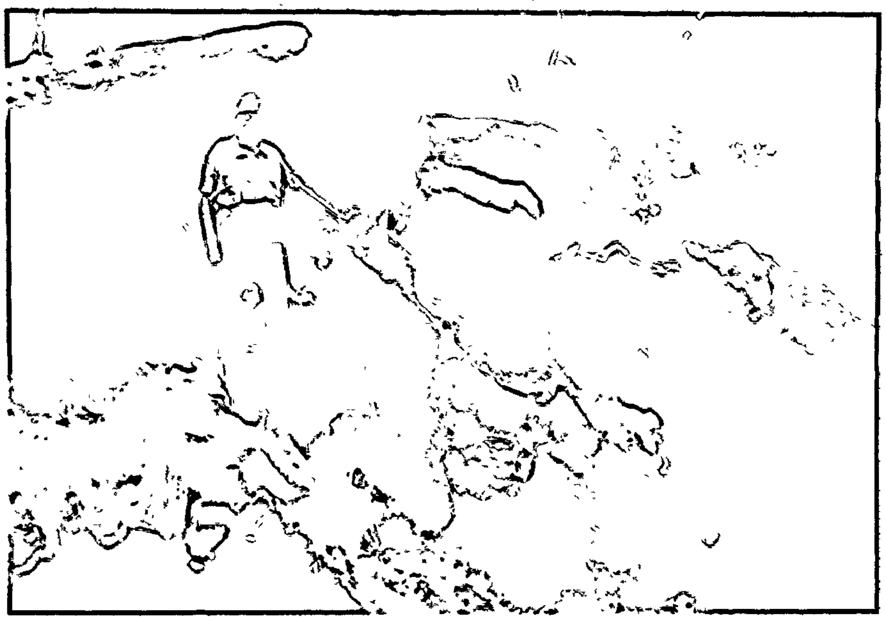


PHOTO 4.6: Karren pits (kamenitza) on surface in gorge.



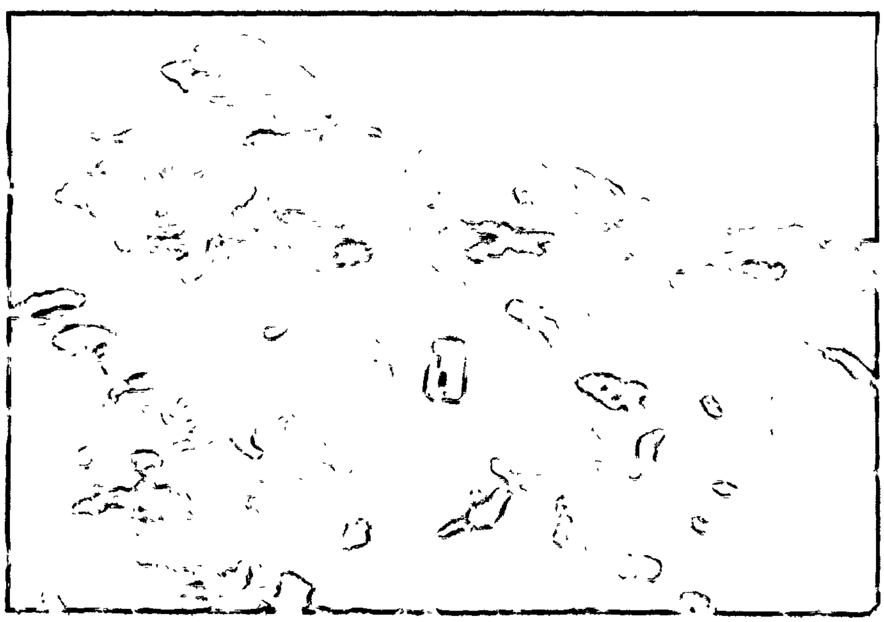


PHOTO 4.7A: Stream potholes

PHOTO 4.7B: Pothole cobbles, evidence of corrasion

4.3 Gibson Plateau

The most mature surface karst development on the Quatsino Fm. limestone in the region occurs on the Gibson Plateau. This probably results from less intense and less frequent glacial erasure of features on the plateau because it is an isolated massif which rises up to 900m above base level.

4.3.1 Macroforms

The orientation of dry valleys on the plateau is strongly controlled by geologic structure; most are strike oriented. Allogenic waters cannot flow onto the plateau surface in the present landscape. Many of the dry valleys are hypothesised to be subglacial, glacial margin or periglacial features. No flow was observed in the valleys during the summer of 1980. However, some of the valleys probably carry seasonal streams.

4.3.2 Mesoforms

The most extensive doline development known in the Quatsino Fm. occurs here. The dolines on the plateau are dominantly solutional in origin, few are believed to result from collapse which is a common process in the Benson Valley. This probably results from the steeply descending nature of cave passages draining the plateau.

The plateau dolines are divided into two classes. The first are hypothesised to result from solution by centripetally draining sub-soil waters. The features occur on near-horizontal sections of the plateau. They may be compound features, ie uvalas, several 100m in length. Along the north-east margin of the plateau the wall of a large uvala is breached by a steep walled channel cut to the edge of the plateau. This feature is hypothesised to be of mixed solutional and glacial meltwater origin.

The second class occur along dry valley thalwegs. These are steep walled features which reach upto 12m in depth. In one location three large dolines occur along some 100m of a dry valley. These dolines are old stream sinks, which have probably drained both glacial melt and interglacial runoff. The occurrence of coarse unsorted sedimentary deposits, apparently till, in several of the features suggests pre-Holocene development.

4.3.3 Microforms

Impressive karren development, such as that in parts of the Benson Valley, does not occur on the plateau. Deforestation and soil erosion are more recent phenomena on the plateau and sufficient time has not yet elapsed for extensive subaerial development. The subsoil solutional attack prior to deforestation was probably diffuse and expended in lowering the general surface rather than developing forms typical of concentrated channel flow.

4.3.4 Discussion

The first phase of solution on the plateau is an ancient phenomenon which predates surficial karst development. In several locations metamorphism of the limestone has occurred beside small intrusive zones indicated by plumes of green rock. Solution cavities containing boxwork and mineralization of unknown composition occur beside the plumes. The intrusions may date to the Tertiary tectonic episode.

The surficial geomorphology of the plateau observed in a series of quarries suggests that the plateau is a pre- last glaciation karst surface. In one quarry wall a compacted clay and conglomerate fill occurs in a solution doline about 12m in diameter. Elsewhere fragments of old solution pipes and phreatic passages infilled with compact clay sediments occur.

The Gibson Plateau is hypothesised to be a multiphase karst surface modified by periods of moderate glacial erasure and channel development by melt waters. Speleothem with a Th/U age of 11.6 ± 0.3ka was found in one plateau sinkhole cave. The cessation of glacial melt flow and the onset of the Holocene epoch probably occurred prior to this date on the plateau.

4.4 Summary

The surficial geomorphology of the Quatsino Fm. was examined in two contrasting topographic situations. In the Benson Valley the form of the surface is ascribed primarily to glacial and fluvial erosion. Landforms of solutional origin dominate at the micro and mesoscale.

Karren development is a comparatively recent phenomenon but impressive subaerial forms have developed in this period.

The Gibson Plateau is believed to be a mature karst surface. Glacial processes have enlarged rather than erased the landforms developed. The large sinkholes along the dry valleys are probably of mixed origin. It appears that the Holocene epoch commenced on the plateau prior to 11.6 ± 0.3 ka. B.P.

CHAPTER 5

SUBSURFACE GEOMORPHOLOGY

5.1 Introduction

This chapter describes groundwater flow and speleogenesis in the Quatsino Fm. The purpose of the study is to determine the factors which control groundwater flow patterns and hence the physical constraints on the distribution of solution; and to describe cave form, reconstruct sequences of passage development and relate these to factors external to the cave.

Groundwater flow can occur in any rock type once basic physical criteria are met: (1) an interconnecting or throughput routeway exists whether composed of intergranular or secondary structural pathways; (2) input sites allow fluid to flow along a pathway of decreasing potential energy to the outlet (flow direction may be reversible, eg. estavelles) and (3) the forces pushing fluid (hydraulic head, fluid velocity, fluid density) exceed the frictional forces retarding flow (fluid viscosity, channel roughness, geometry of pathways).

The characteristic features of subsurface drainage in karst regions are; (1) karst limestones typically have negligible primary permeabilities so that groundwater flow is dependent on the framework of structural elements which comprise the initial secondary permeability of the rock mass. The relative importance of different structural elements in the transmission of water is a function of their permeability, the extent of interlinkage with other elements, and generally the extent to which pathways composed of one or more elements can route water in the direction of the maximum hydraulic gradient. (2) Water flow alters the geometry of the flow path by rock solution. Karst regions are actively evolving both w.r.t. the form of the individual flow paths or caves, and also the size of drainage basins and the bulk permeability of the mass.

In the Quatsino Fm. limestone the structural framework is composed primarily of bedding planes which dip to the south west at between 25^{o} - 37^{o}, joints oriented along the strike and joints oriented normal to the strike (parallel to the dip). The joints are frequently restricted to single beds. In certain locations high angle faults of variable orientation control groundwater movement.

The different structural elements are not of equal importance in the transmission of water. Allogenic streams reach the limestone along both the east and west sides of the Benson Valley, but the pattern of sinking streams is distinctly asymmetric (Table 5.1). Streams flowing onto the limestone on the eastern flank, where the hydraulic gradient and stratal dip are roughly concordant, sink close to the Karmutsen Fm. contact and resurge at the base level springs. Streams flowing onto the limestone from the Parson Bay Fm. on the western flank where the stratal dip is into the mountainside and is roughly normal to the maximum hydraulic gradient, remain at the surface

TABLE 5.1

Pattern of sinking streams

	SINK	REMAIN ON SURFACE
ALLOGENIC STREAMS FROM WEST	2(i)	4
ALLOGENIC STREAMS FROM EAST	7	1(ii)

(i) perched on delta deposit

(ii) sink on faults

and incise gorges up to 20m in depth.

The factors which control the pattern of sinking and surficial streams are unknown. However, for a given hydraulic gradient and change in elevation between sink and spring positions, water sinking on the western flank would have to pass through a greater stratigraphic depth of limestone (Table 5.2 and Figure 5.1) and a greater proportion of the flow path would therefore be composed of joint elements, many of which are restricted to single beds. For the same reason waters in the vadose zone are primarily routed down the spatially more extensive bedding planes. On the western side of the valley vadose waters are therefore routed away from potential spring positions. Thus on a regional scale bedding planes are the most important groundwater pathways.

Groundwater flow in the Quatsino Fm. is further controlled by the occurrence of non limestone impermeable barriers in the limestone. The most common features are dykes oriented normal to the strike (down dip) but barriers also occur oriented along the strike and between beds. Most are composed of igneous material intruded during the Tertiary tectonic episode, but some of the impermeable strata are sedimentary in origin, laid down during breaks in limestone deposition.

The control exerted by the features on groundwater movement in a given location is difficult to predict because of variable spatial extent and the fact that they may be fractured and breached by pathways of secondary permeability. However, groundwater flow is commonly

TABLE 5.2

Change in stratigraphic position for given hydraulic gradient and change in elevation on east and west side of valley

REPRESENTATIVE SITE	STRATAL DIP	ELEVATION BETWEEN INPUT & OUTPUT	DISTANCE ALONG DIP BETWEEN INPUT & OUTPUT	HYDRAULIC GRADIENT	CHANGE IN STRATIGRAPHIC POSITION	CORRESPONDING CHANGE IN POSITION ON WEST SIDE OF VALLEY
Malook III	30°	75m	850m	0.088	360m	490m
Minigill	30°	120m	450m	0.27	121m	346m

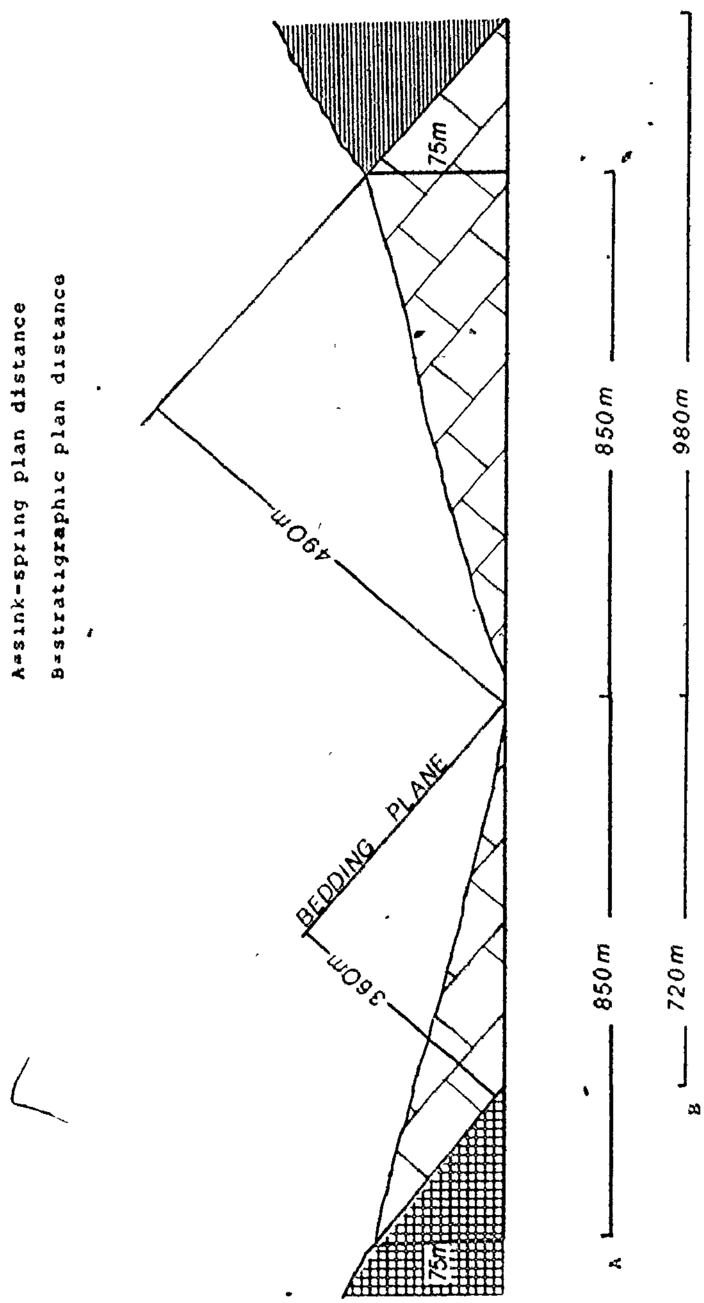


FIGURE 5.1: Effect of stratal dip on change in stratigraphic depth along a given hydraulic gradient.

confined between dykes, above, below or between impermeable beds, or by a complex combination of these situations.

5.2 Cave Development in the Quatsino Fm. Limestone

Speleogenesis in the Quatsino Fm. limestone has been divided into authigenic and allogenic types, dependent on the orographic relationship between the limestone and the adjacent non-karst strata. Variations in the relationship between structural geology, topography and the position of base level streams have resulted in a variety of different situations for cave development (Figure 5.2)

5.2.1 Authigenic Karst

Authigenic karst evolution occurs on the cuestas in the Benson Valley and on the Gibson Plateau, a limestone massif located to the north-west of the Benson Valley. Karst development on these upstanding limestone massifs is dependent on incident precipitation. However, in one location a small stream from a valley bottom lake sinks at the base of a cuesta scarp face.

The topography of the cuestas is controlled by the structure of the limestone. Steep scarp faces occur on the north-east flanks; dip slopes of lower gradient extend to the south-west. Drainage in the cuestas is strongly influenced by the framework of structural elements in the limestone. Springs occur only at the bases of the down dip or strike slopes. No active or inactive spring sites were

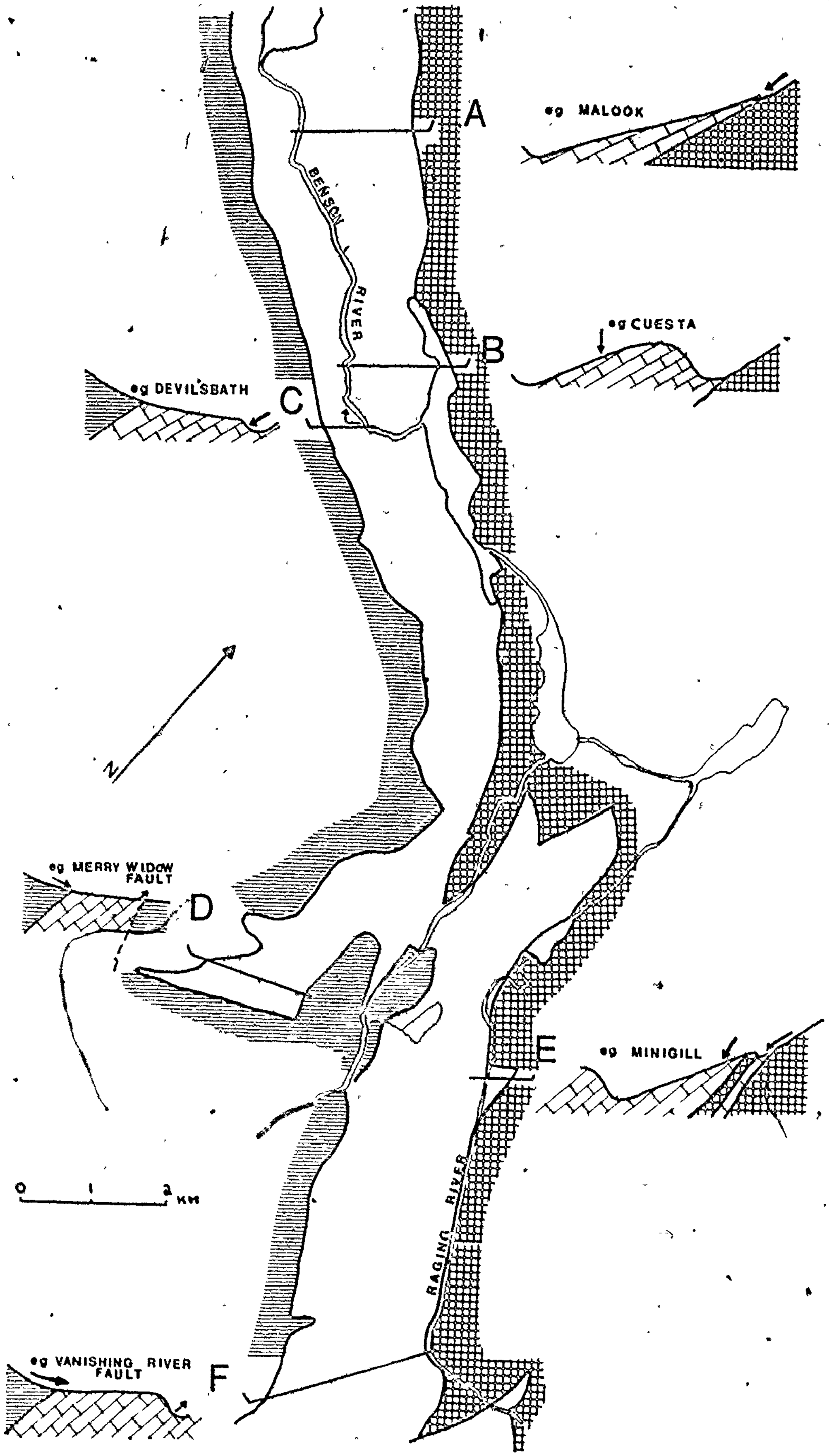


FIGURE 5.2 Cave development situations in the Benson Valley.

discovered at the bases of the scarp flanks. The pattern of discharge sites indicates that the structural framework routes water away from the steepest hydraulic gradient in certain locations. For example, the gradient from a series of closed depressions near the scarp face of Kathleen Lake cuesta to the base of the scarp is about 0.25. The only known springs draining the cuesta occur at the base of the dip slope. The gradient from the closed depressions to these springs is 0.05. The direction of flow on the most southerly cuesta is to the north-west to Lac Truit, a lake located at the down dip end of the strike flank . Lac Truit has a surface area of $0.2km^2$ and is not fed by any surface streams. On the third cuesta no springs were located; however, a bedding plane cave several metres wide has been intercepted by a road cut at an elevation 250m above the valley floor.

No caves of explorable dimensions were found in the authigenic karst in the Benson Valley. The nature of speleogenesis in the cuesta can only be interpreted w.r.t. the geological structure, the spring positions and the intercepted bedding plane passage. Incident precipitation sinks directly down enlarged joints or moves as soil throughflow to the base of closed depressions. Sinking waters apparently flow down bedding planes and use joints to rise upwards stratigraphically or along strike to their outlets. The cofiguration of the water table in the cuesta is controlled by the position of springs and variability in the permeability of the rock mass. The gradient of the water table and its elevation above the springs depends on the permeability and porosity of the rock mass and the volume and regime of incident precipitation.

The best developed authigenic karst in the Quatsino Fm. limestone occurs on the Gibson Plateau which rises 900m above the surrounding base level. The Plateau is the exception amongst the authigenic karst blocks in that it is an independent mountain rather than a block upstanding in a greater valley. The massif reaches 7km in length along the strike and 3km normal to the strike. The surface area of the plateau is approximately 15km^2. To the west and north it is bounded by two fiords, Nereutsos Inlet and the Quatsino Narrows, to the east by the Marble River lowlands and to the south by the Nequitipaalis Creek valley. The plateau is a well developed karst surface that contains systems of large closed depressions and inactive river sinks.

Caves occur at the base of most of the depressions. However, no system has been explored to a depth greater than -35m (Paul Griffiths, pers. comm*), because of boulder collapse or sediment and lumber blockage following deforestation. Despite the large surface area only two small springs have been discovered at the base of the plateau. This suggests that most waters sinking on the plateau discharge from submarine springs.

5.2.2 Dripstone Cave, Gibson Plateau

Dripstone Cave is a resurgence cave located some 40m above Nequitipaalis Creek, at an elevation of 180m O.D. The dry season discharge from the cave is less than 0.2m^3 s^{-1}. The cave has been explored for 100m to a sump (Figure 5.3). Water rises from the

* Paul Griffiths - British Columbia Speleological Federation

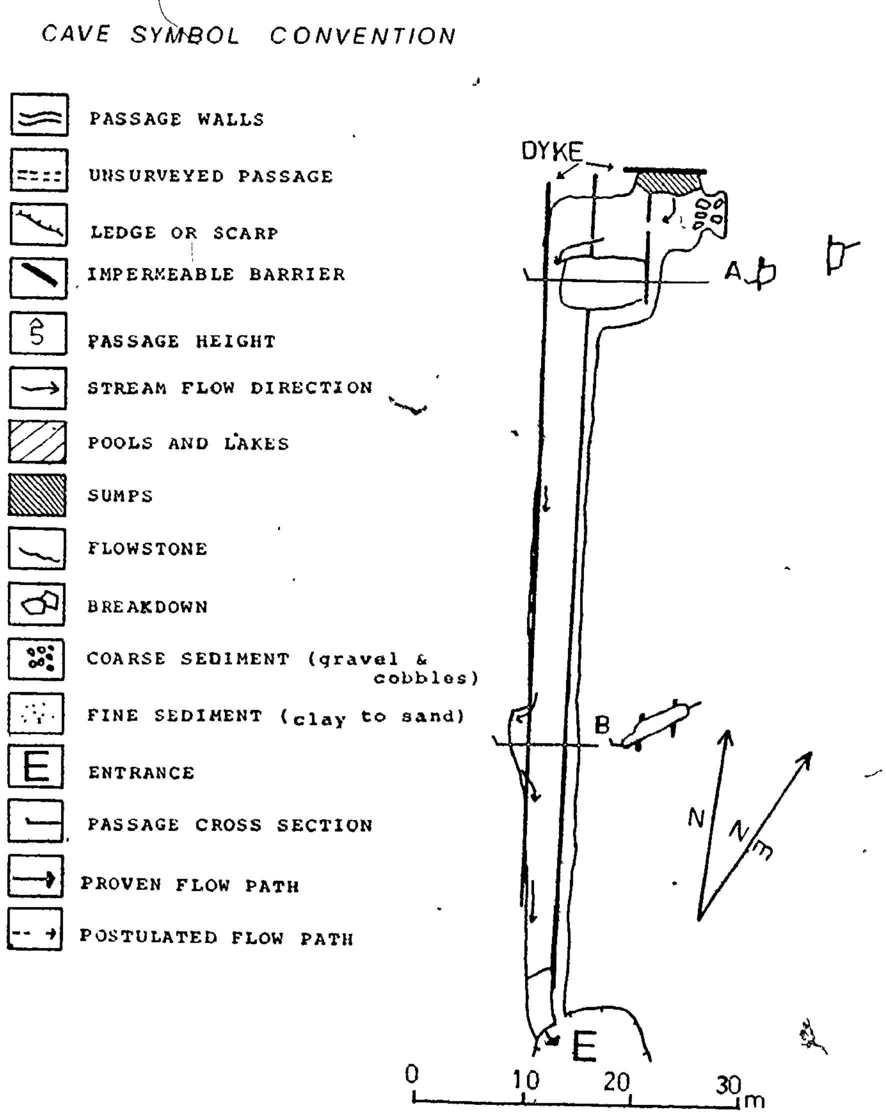


FIGURE 5.3: DRIPSTONE CAVE and cave symbol convention

sump pool located beside a dyke oriented along the dip and flows aslant strike on a steeply dipping passage to the outlet. The sump pool chamber has developed from collapse beside the dyke. The cave stream is perched on the dip side of a thin non-limestone barrier oriented aslant strike. Two other aslant strike non-limestone bands occur in the passage centre and up-dip wall of the passage. The stream has breached the barrier where a joint parallel to the dip crosses the passage. In this location passage development is eroding down dip.

The temperature and chemistry of the cave stream suggest that the spring discharges long resident percolation waters integrated into a network of conduits chaining to the strike. The distance from the cave entrance to the groundwater divide separating flow along strike to the cave and flow down dip to submarine springs is unknown. However, the small discharge from the cave suggests that the distance is not great and that the cave only drains the southern end of the Plateau.

The exsurgence is perched some 40m above the valley floor although Quatsino Fm. limestone extends to the base of the valley. This results from perching of flow above the impermeable barriers; but may also result from valley base level lowering during the last glaciation if the cave is pre-last glaciation in age.

5.3 Allogenic Karst

Allogenic karst development in the Quatsino Fm. limestone primarily occurs along the flanks of the Benson Valley. However, because of the asymmetric pattern of sinking streams, allogenic cavern development is best developed on the east side of the valley. River gorge development is the pricipal geomorphic response to allogenic stream inputs on the western flank. Allogenic karst development also occurs along the valley bottom in two locations where water in the Benson and Raging Rivers is pirated underground over distances of 0.6km and 1.3km repectively.

5.4 Malook

5.4.1 Malook Creek Cave System

The Malook Creek cave system is located on the east side of the Benson Valley, south of Alice Lake. The area is bounded to the north by the Pinch Creek delta sediments, and to the south by the Kathleen Lake cuesta. The distance from the Karmutsen Fm. lavas to the Benson River is about 0.85km, with a drop in elevation of 75m. The area is about 1.5km^2. The physiography is dominated by a series of dry valleys.

Three streams flow off the Karmutsen Fm. lavas and sink into the limestone close to the geologic boundary. Two of them (Malcok 1 and 2) drain to Malook Spring, the largest spring in the area located about 1.5km north west of the sinks. The flow paths to the spring, via three karst windows, were proven by two separate traces with flourescein dye. The third stream (Malook 3) is believed, from chemical data, to connect into the system between the bottom karst window and the spring (see Section 3.8). The position of the sinking streams, spring, karst windows and dry valleys are shown on Figure 5.4.

The stratal dip of the Quatsino Fm. limestone in the Maloak system ranges 27-33°. Waters sinking near the lava contact rise upwards in the stratigraphic sequence approximately 425m to resurge at Malook Spring.

The position of the karst windows suggests that simple point to point flow paths route water from the stream sinks to a single spring at the base of the local karst massif. However, investigations at the karst windows demonstrate that the sub surface flow pattern is complex in detail and controlled by a suite of non-limestone impermeable barriers.

<u>Malook One</u> A stream with a mean dry season discharge of less than $0.1m^3s^{-1}$ sinks in its bed 150m below the lava contact. During periods of flood discharge stream flow crosses upto 400m of limestone to larger swallet sinks.

<u>Malook Two</u> A stream with a mean dry season discharge of less than $0.1m^3\ s^{-1}$ sinks into a large swallet 80m from the geologic contact. The sinkhole occurs in till 2.5m in depth and measures 12m in diameter. The depth of water in the sinkhole exceeds 4m. Flourescein dye added to the stream was carried to the base of the down dip wall, the apparent location of the outlet. The sinkhole water level fluctuated by less than 0.5m during the two month field season, despite periodic

heavy precipitation. This suggests that the stream sinks into a pheatic loop with an unconfined outlet. The lack of an overflow channel leading from the sinkhole indicates that it can engulf the stream at all discharge and suggests that it is not a young, postglacial feature.

Malook Three A small stream sinks through coarse sediment in its bed close to the lava boundary. During high discharge periods the stream flows for a further 50m to a partially blocked swallet in the base of a large depression. A karst window about 10m deep is located 150m south-west of the sink, along a straightline path from the sink to Malook Spring. The base of the window is blocked by lumber. The window probably results from roof collapse in the cave draining Malook 3.

Top Window The Top Window is the first karst window along the cave system draining Malook 1 and 2. The window measures about 17m by 10m in plan and is 3.5m deep. The cave stream enters the window from a passage 1m in diameter, flowing up dip. The stream sinks again at the northern end of the window into limestone, about 8m stratigraphically lower. The stream flows for 13m along a strike-aligned cave passage and turns down dip into a small unexplorable passage. A fossil passage, with small dip tubes on the up dip side occurs above the stream passage. It is blocked by fine sediments (Figure 5.5).

The direction of water flow from Malook 1 and 2 to the Top Window is along strike. The flow path to the next window, the Eternal

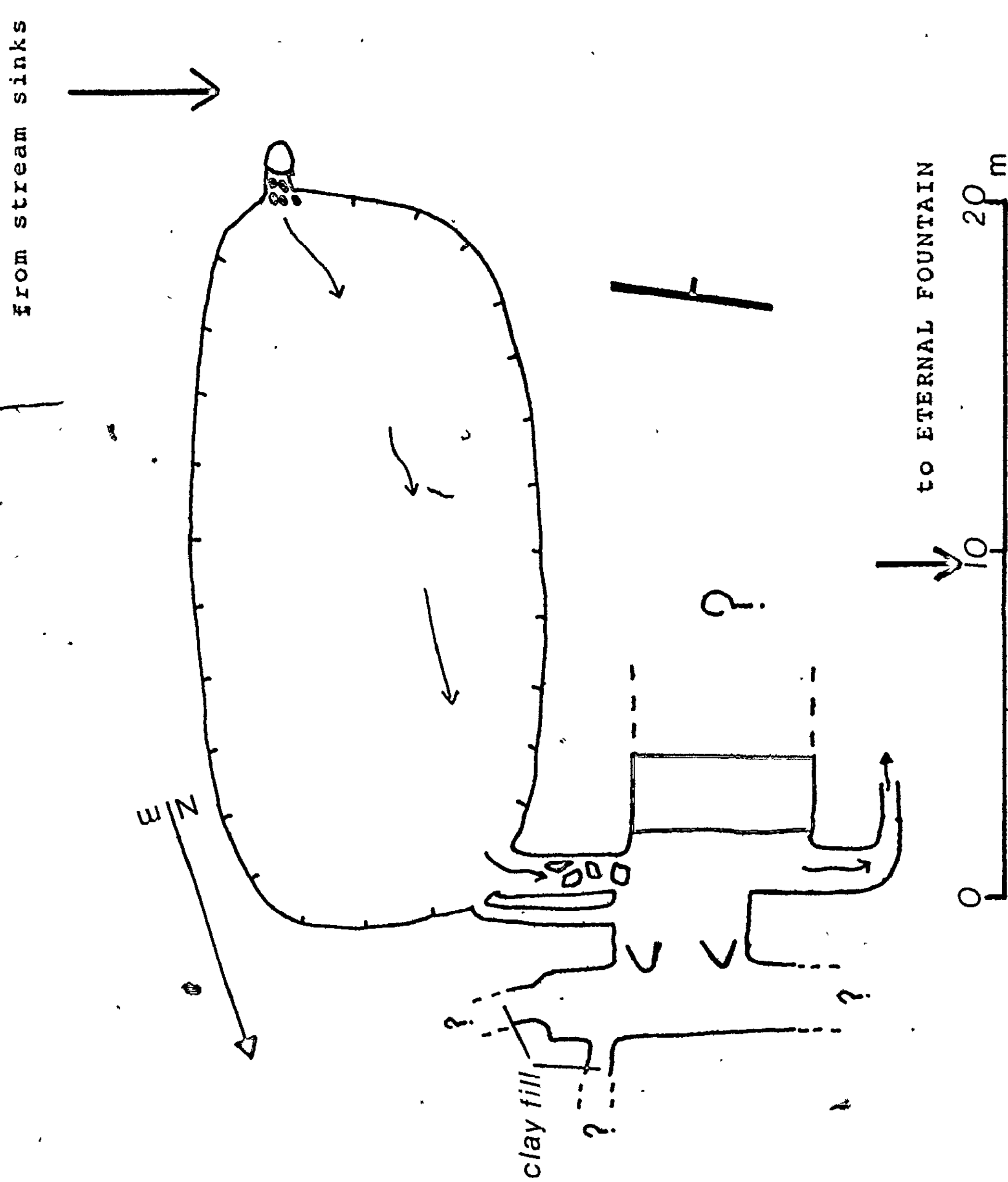


FIGURE 5.5: TOP WINDOW

Fountain, is alsant strike and rises upwards stratigraphically through many metres of limestone. However, the cave stream rises up dip into the Top Window and sinks at a stratigraphically lower level. This flow path suggests the presence of a barrier to strike drainage in the vicinity of Top Window, bypassed by stream flow at a lower stratigrapic level. The location of the barrier is uncertain but may be related to a non-limestone exposure located on the down dip side of the cave passage. The high level fossil strike passage may represent a paleo bypass developed prior to passage collapse and development of the window. It is uncertain whether the dip tubes connected to the fossil passage functioned as lifting chimneys routing water to paleosprings, or routed waters from an overlying karst surface into the cave.

Eternal Fountain The Eternal Fountain Cave occurs at the base of a 4m waterfall developed where a 30cm thick non-limestone bed, which perches the cave stream, outcrops at the surface. The cave has been explored for 110m to a sump (Figure 5.6). The passage is developed principally along joints parallel to the dip, with short strike-aligned sections. However, the gradient of the cave is less than the stratal dip and the passage rises upwards in the stratigraphic sequence some 45m between the entrance and the sump. At two locations the cave stream has punched upwards through non-limestone beds greater than 0.25m in thickness. The downstream sump occurs where the stream has breached a strike-orientated dyke.

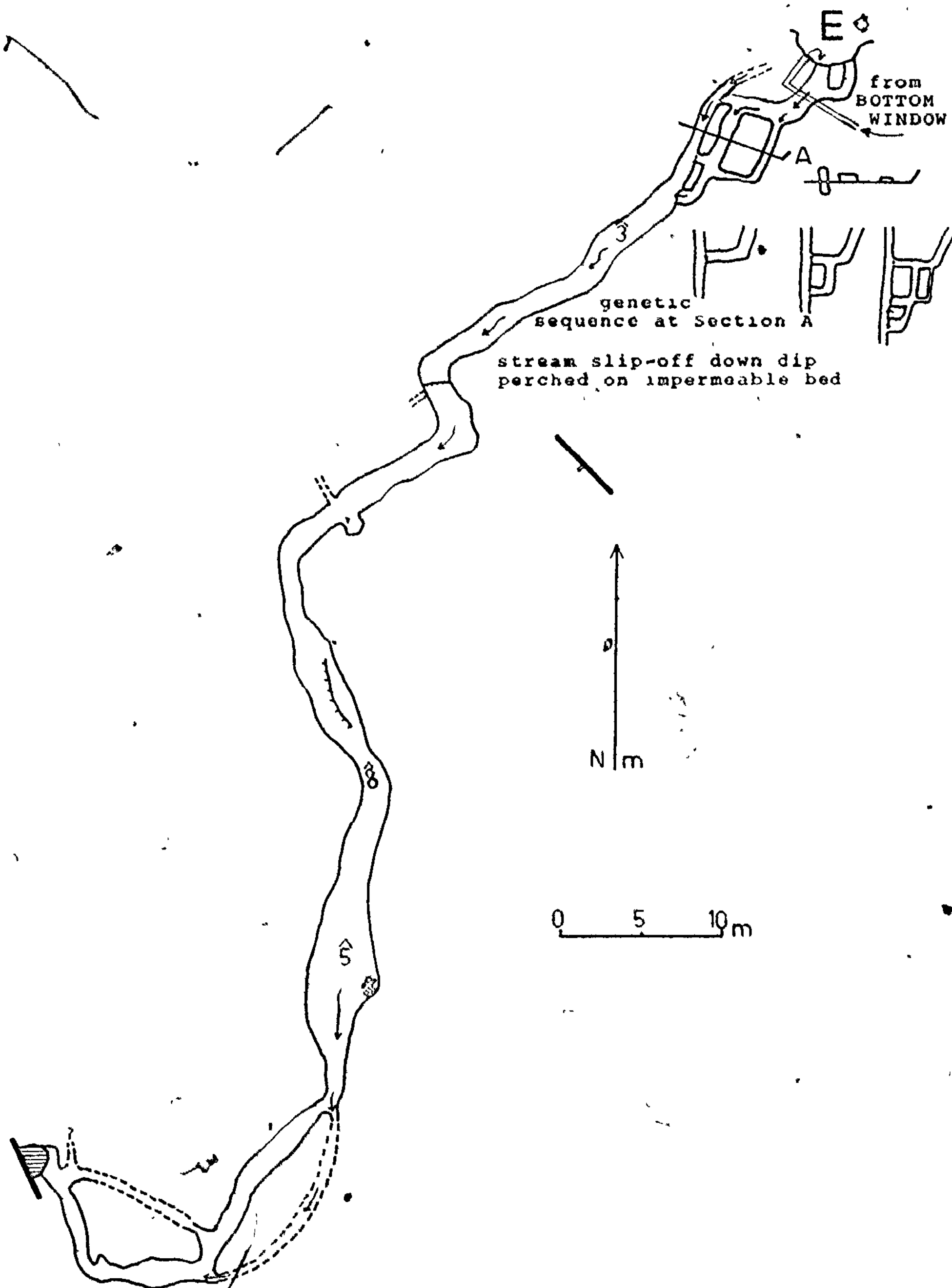


FIGURE 5.6: ETERNAL FOUNTAIN (based on survey by K. Bischoff and P. Griffiths, 1978)

The frequency of joints parallel to the dip in the Eternal Fountain cave is high. In parts of the cave this has resulted in a small scale distributary passage development. At the entrance the stream divides between two passages for 10m (Section A Figure 5.6). The passage to the west is developed along a joint parallel to the dip and has been incised to a lower level than the passage to the east, which joins the former by flow along the strike. ~~The~~ easterly passage is perched on a thin non-limestone bed. The former connection was made by a flow directly along strike; however, the stream has progressively eroded downdip and now joins the westerly passage at a lower level.

The flow direction from the Top Window to the Eternal Fountain is aslant strike; however, the cave stream flows up the dip at the cave entrance, perched on the non-limestone layer. Moreover, at the waterfall the stream passes beneath the non-limestone layer, but subsequently rises up through the same bed further downstream (Figure 5.7). This suggests the presence of a barrier to flow down the dip above the non-limestone layer which outcrops at the waterfall. The barrier is hypothesised to be a dyke oriented along strike, similar to that at the downstream sump.

<u>Steps Cave</u> Steps Cave is located at the down dip end of a karst window which measures about 30m by 10m in plan. The stratal dip of the limestone is 29° at 217°. The cave passage is developed along a dyke which is oriented aslant dip at 235°. The explored length is less than 40m (Figure 5.8).

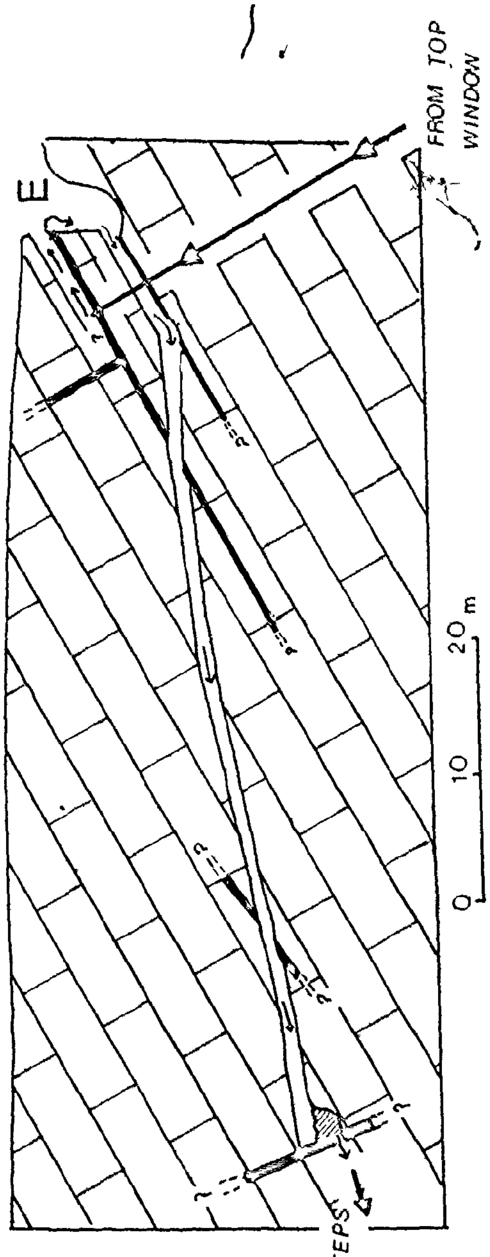


FIGURE 5.7: Influence of impermeable barriers on conduit flow patterns at ETERNAL FOUNTAIN CAVE

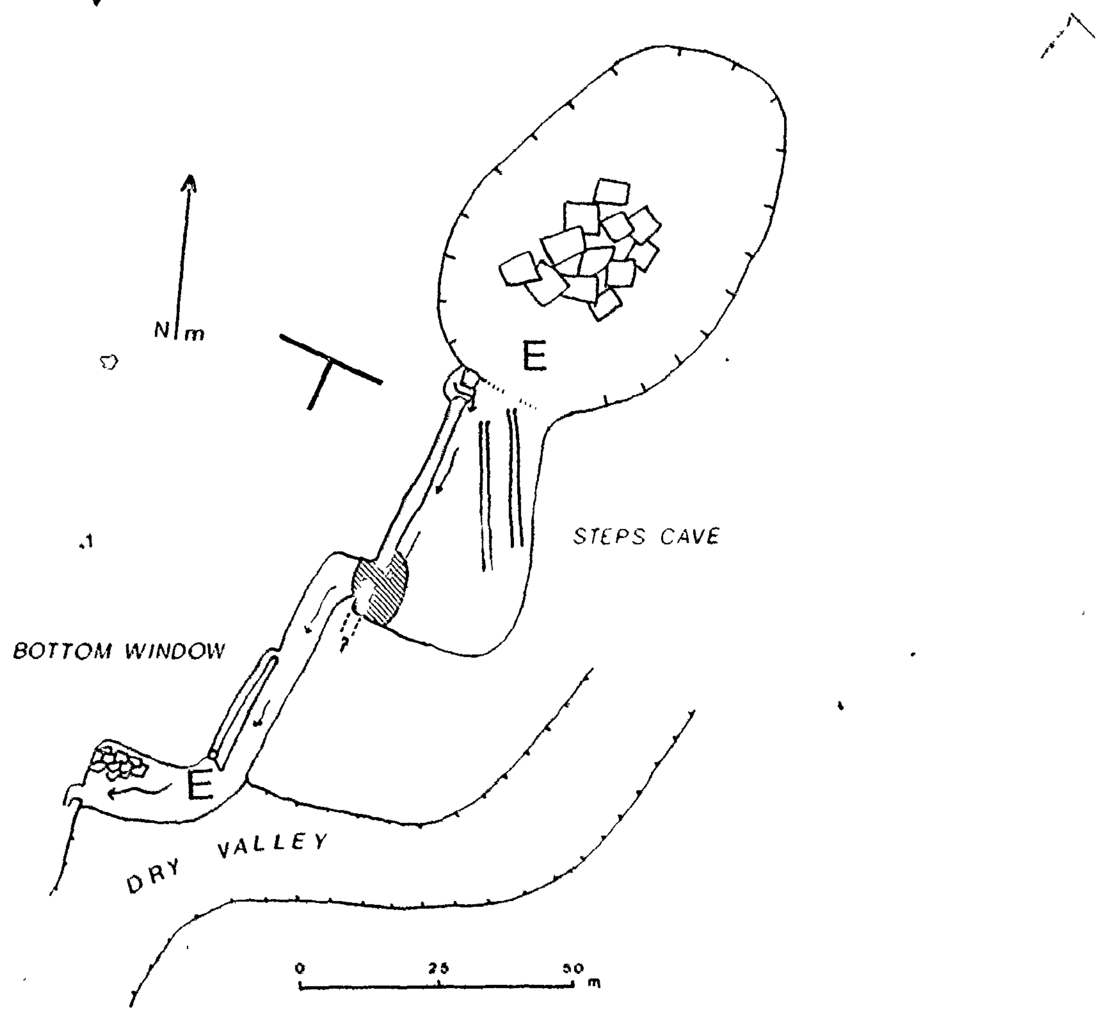


FIGURE 5.8: STEPS and BOTTOM WINDOW CAVES

Water flows to the Steps Cave from the Eternal Fountain to the South East, but enters the Steps Cave by breaching the dyke from the north west. The stream flows along the eastern side of the dyke to a sump where the dyke is breached for a second time and the stream flows to the West. This flow path is difficult to interpret, but must result from a complex compartmentalization of flow by non-limestone barriers.

Fossil dip tubes in the passage roof and an inactive strike passage at the end of the cave indicate th paleoflow occurred down dip and was linked to the route through the dyke by a strike passage. The flow path has been shortened to a more efficient route along the side of the dyke.

Water sinking against the dyke in the Steps Cave resurges into a vadose stream passage, the Bottom Window cave. Water rises several metres from the base of the sump pool and flows parallel to the dip in a vadose canyon to the entrance. Downstream from the sump the passage roof contains a 1.5m high fossil pheatic lifting tube connected to a high level dip tube. The lifting tube developed against a thin non-limestone barrier oriented along strike. The passage routed flow prior to breaching of the barrier at a lower level and the development of the vadose streamway (Figure 5.9)

The entrance of the Bottom Window cave occurs in the outer wall of a bedrock meander incised by a fossil stream. The high level phreatic passage must predate the meander incision. The dyke exposed

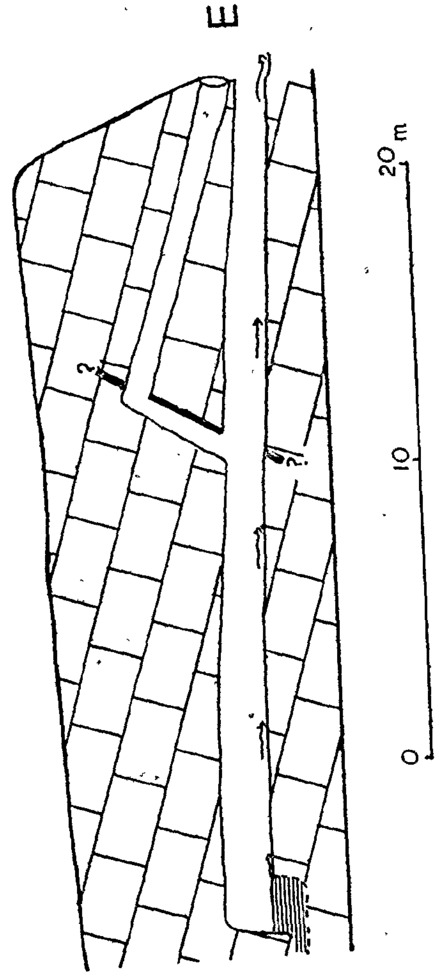


FIGURE 5.9: BOTTOM WINDOW paleoflow route

in the Steps Cave and Bottom Window sump does not outcrop at the Bottom Window cave entrance. This demonstrates its limited spatial extent.

Water resurging at the Bottom Window cave flows north west to a sink in the meander wall. The short cave passage turns down dip into a sump.

Malook Spring The Malook Creek Cave system resurgence occurs at the base of a limestone hill about 60m to the south west of the Bottom Window. The mean low season discharge of the system is about 0.25 $m^3 s^{-1}$. The water then flows in a low gradient meandering channel to the Benson River 700m to the West. Water sinking at Malook sink 3 is believed to combine with water from Malook 1 and 2 between the Steps Cave and the Malook Spring.

Suprise Well The Suprise Well cave is located 175m north of the Malook Spring at an elevation 10m above it. The cave occurs beneath the southern margin of Pinch Creek delta sediments; greater than 5m of silts and clays overlie the entrance.

The cave is developed along a major joint oriented along dip at 040 - 220° (Figure 5.10). The joint extends stratigraphically through greater than 4m of limestone. At the entrance, a shaft 2.5m in diameter ascends from a depth in excess of 3m. Water fills the shaft almost to the level of the entrance. Two metres above the pool a high level passage 0.5m in diameter leads to a second water filled shaft 15m from the entrance. The high level passage continues beyond the second pool but is too small for exploration.

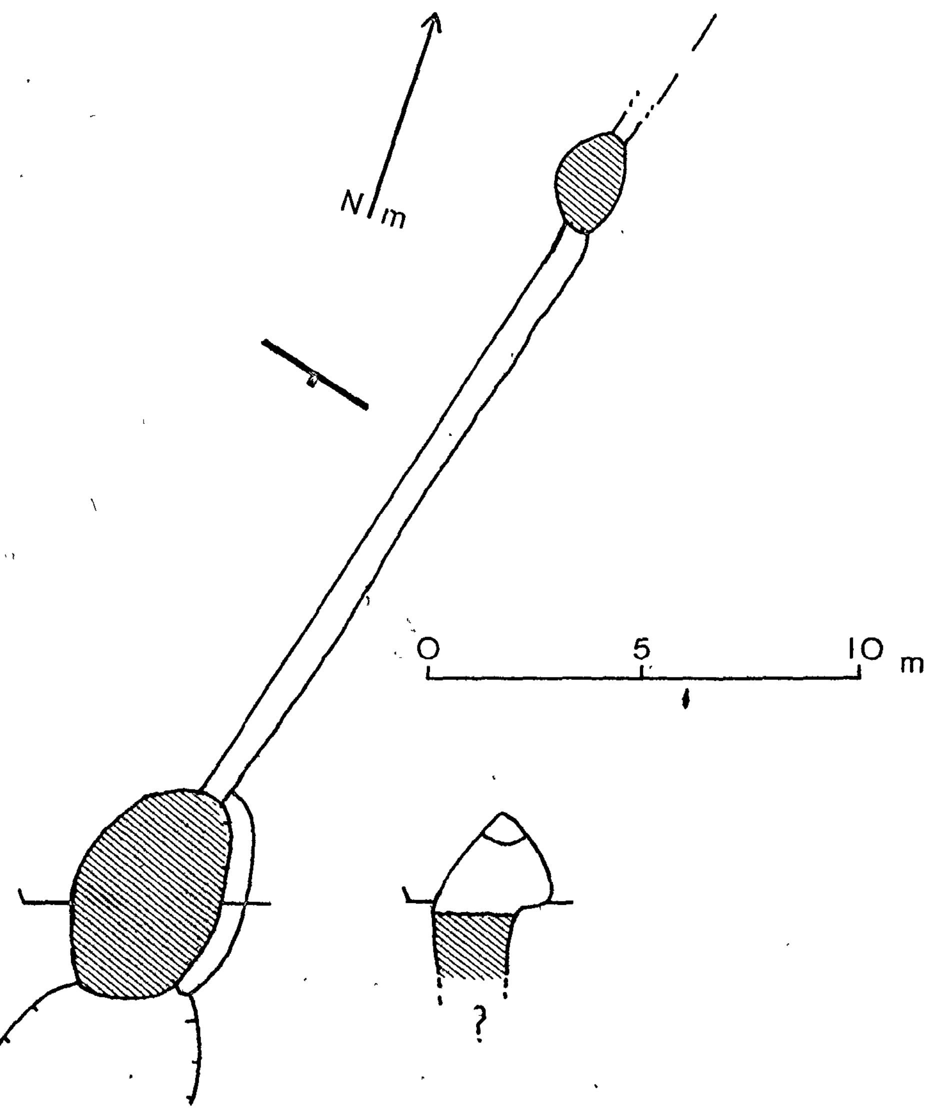


FIGURE 5.10: SUPRISE WELL

The short section of the cave explored appears to be part of a cave system composed of phreatic loops with an overlying horizontal high level passage. The high level passage probably developed as a bypass route for the phreatic loop, prior to the formation of the present entrance.

The cave is fed by percolation waters passing through the deltaic sediments. The coefficient of variation of total hardness of water in the entrance pool was 4.4%. The water level in the entrance was quasi-static during the field season and water did not discharge from the entrance.

The source of the waters which developed the Suprise Well is unknown. Small scallops in the shaft indicate that water was discharged from the entrance at a considerable velocity. The cave may once have routed a sinking stream from the Karmutsen Fm. lavas, 800m to the north east. The cave apparently predates the delta and is pre-last glaciation in age.

5.4.2 Discussion

The Malook Creek cave system is overlain by a major dry valley system (Figure 5.4). The Steps Cave and the Bottom Window occur within the dry valley system but apparently pre-date valley entrenchment because they contain phreatic passages above the dry valley floor level. The Eternal Fountain and Top Window occur outside the dry valley system. Thus it appears that (1) the cave system predates the valley entrenchment and (2) the windows are the drained upper apices of a cave system containing phreatic loops, the location of which are controlled by the framework of structural partings and impermeable barriers.

The dry valley system may be sub-glacial in origin. While 2 of the features start at the Karmutsen Fm. contact, the largest dry valleys start on the authigenic karst. This suggests that the dry valleys were activated subglacially, with possible postglacial modification of those at the lava margin. If so, the cave system must be pre-last glaciation in age. This model is supported by the form of the Malock 2 sink and the Suprise Well.

5.5 Minigill

5.5.1 Minigill Cave System

The Minigill cave system occurs on the eastern side of the strike-aligned Raging River valley in a limestone wedge with an area of about $0.4km^2$. The limestone wedge has been uplifted onto the Karmutsen Fm. lavas by faulting; the rest of the eastern side of the valley is composed of lavas. The western side of the valley is a limestone cuesta scarp. The valley is steep sided and local relief exceeds 200m on the western side and 600m on the east. The limestone wedge rises up to 120m above the river; the maximum distance down slope from the top of the outcrop to the river is less than 450m. A limestone unit with a stratigraphic thickness greater than 10m occurs within the Karmutsen Fm. outcropping about 100m up slope from the Quatsino Fm. limestone wedge.

Multiphase drainage down the stratal dip has been integrated by a strike aligned cave system at the valley base to a spring at the north end of the outcrop (Figure 5.11). The Raging River is pirated underground through the strike oriented cave to the spring along a flow path with a straight line length of about 1.3km. Water in the interlava limestone drains to a spring in Karmutsen Fm. lavas to the North of the limestone wedge.

The MInigill system comprises fragmented fossil caves developed when the valley floor stood at higher elevations (Papua Cave and Dip Cave); active caves routing water down the mountainside in the inter-lava limestone (Brutal Cave and Peat Cave) and the major valley bottom system pirating water from the Raging River. This latter system was explored in 4 sections (Window Cave, Minigill Cave, Deer Cave and Pitchford's Resurgence) because of sumps along the route. The caves are discussed in the above order.

5.5.2 Papua Cave

Papua Cave is located in the wall of a small scarp about 90m above the Raging River. The cave is a fossil phreatic passage 50m in length developed along aslant dip joints at 080-260 (Figure 5.12). The passage near the entrance is about 10m wide and 3.5m high. The rear of the cave is blocked by colluvium which has entered the passage through the thin overlying bedrock roof. Scallops with a mean long axis length of 9cm occur in the passage wall near the entrance. In this location the passage is about 5m wide and the cross-sectional area about $15m^2$. Employing the method of Curl (1973) a passage-forming flow velocity of about $0.5m^1 s^1$ is indicated. This velocity

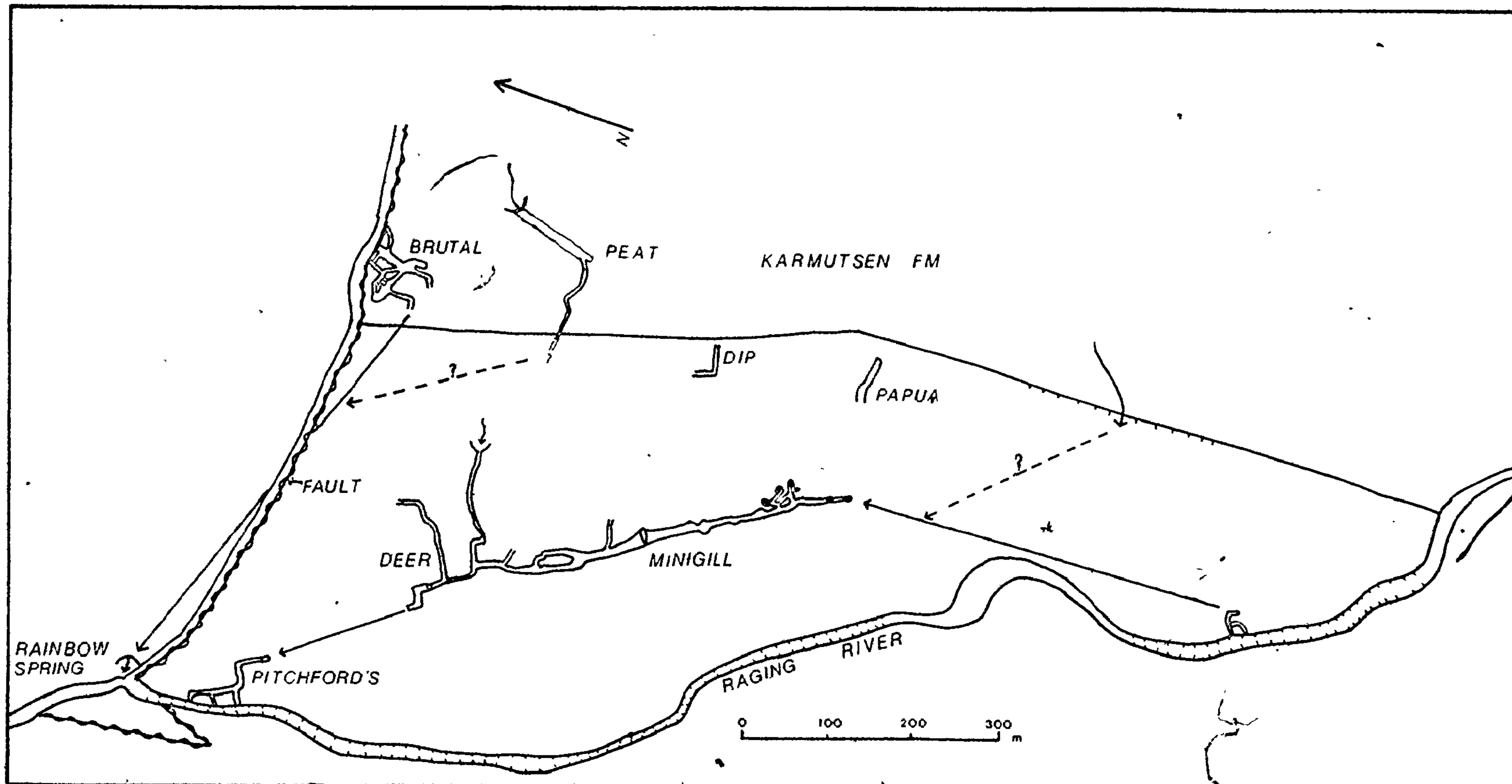


FIGURE 5.11· CAVE DEVELOPMENT IN THE MINIGILL WEDGE (Based on surveys by M. Davis & J Oldham, K. Bischoff, E. Vorkampff, & P. Griffiths)

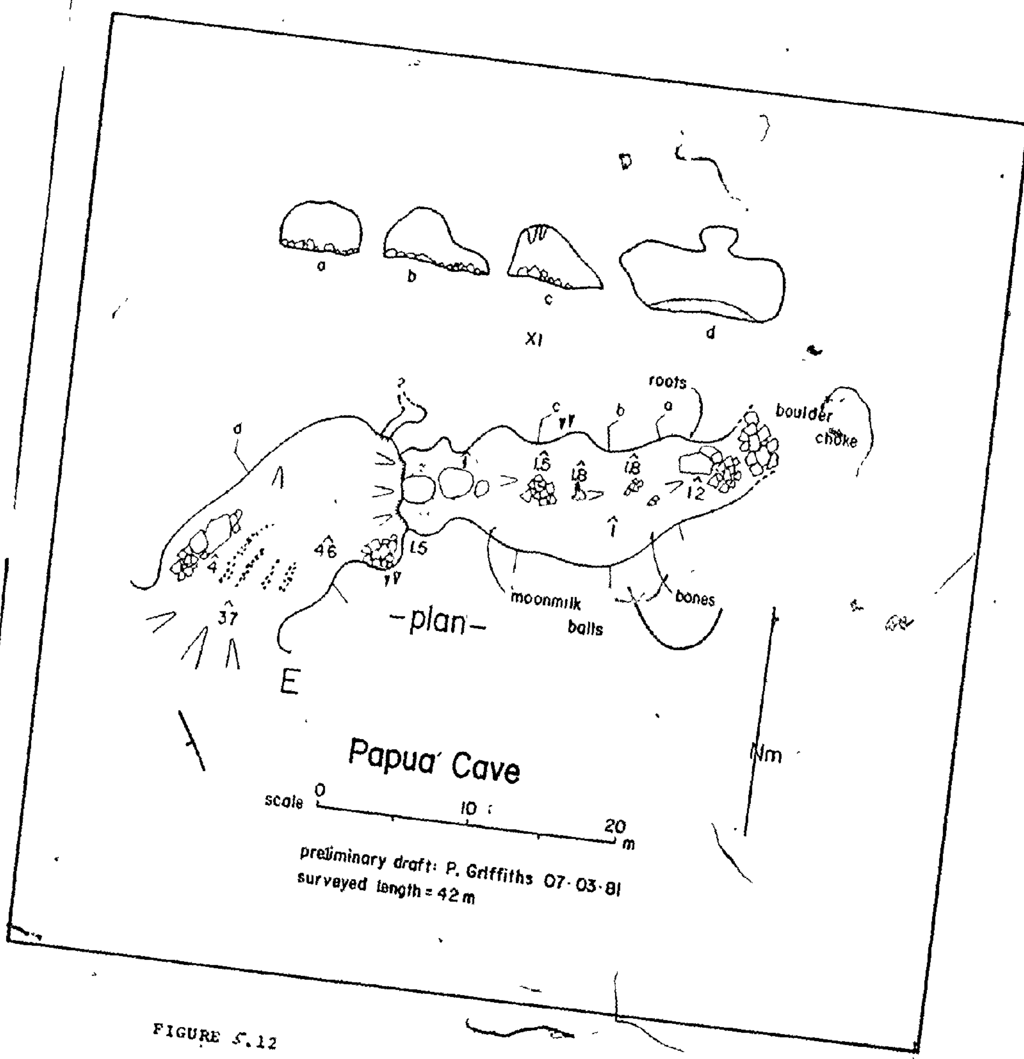

a
b
c
d
XI
roots
c
b
a
boulder
choke
d
1.5
1.8
1.8
12
1
4
46
1.5
37
moonmilk
balls
bones
-plan-
E
Nm
Papua Cave
scale
0
10
20
m
preliminary draft: P. Griffiths 07·03·81
surveyed length = 42 m

FIGURE 5.12

indicates a passage-forming discharge greater than $7m^3 s^1$. A 2m thick sediment section is exposed in the mid section of the cave, normal to the passage wall. Fifty cm of sandy silt are overlain by 60cm of coarse gravels. The gravels are overlain by up to 90cm of clays. A 50cm deep trench has been incised into the clays. The trench is partially infilled with a soft, porous calcium carbonate deposit.

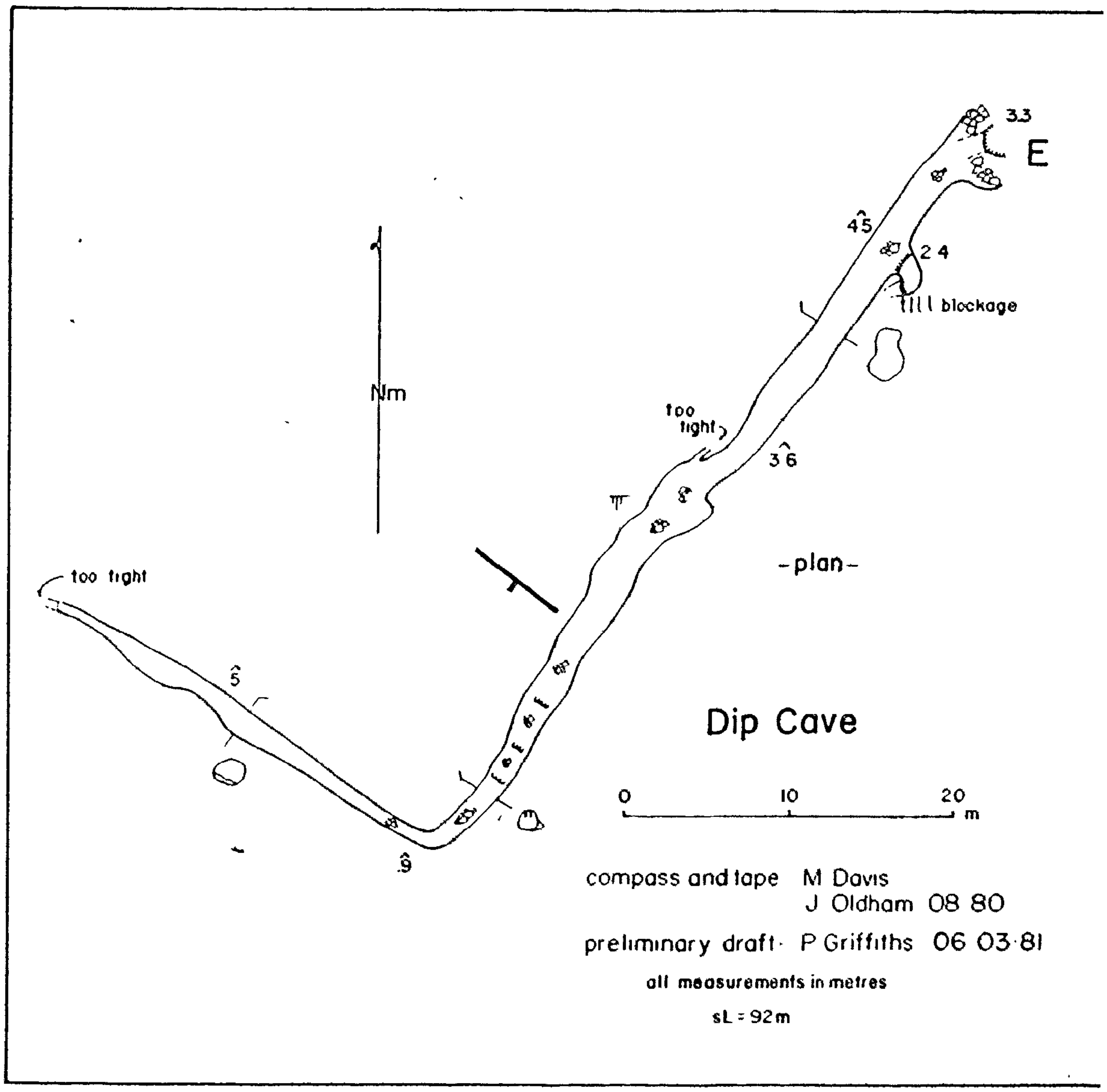
Papua Cave routed waters down the stratal dip from a higher topographic level. The flow velocity and passage discharge determined from the scallop data suggests that the cave routed a major sinking stream rather than percolation water derived from precipitation on the small area of limestone up slope from the cave. Two other possible stream water sources are: (1) glacial melt waters and (2) Raging River stream water diverted underground when the valley floor stood at a higher level. The second is thought to be possible because significant valley lowering is apparent in the area; the topographic gradient is steeper than the stratal dip at Papua Cave and the cave passage has been truncated. (It is unlikely that the stream was generated on the Karmutsen Fm. lavas directly above the cave because of the large cave discharge indicated.) The source of the coarse sediments is uncertain. However, the overlying clay deposit indicates low energy flow conditions which may result from ponding behind an ice body either in the valley or across the valley mouth.

Papua cave will be completely destroyed in a short period of geological time. The entire rock mass above the passage floor appears to be sliding down dip towards the valley bottom; limestone blocks are collapsing from the valley wall along strike joints at the entrance. The cave is being infilled bu colluvium and calcite deposits.

Dip Cave Dip Cave occurs at a similar elevation to Papua Cave, but is on a gentler slope and consequently in limestone at a lower stratigraphic level. Stratal dip is greater than the topographic gradient so that the cave passage can route water into the limestone mass.

The entrance occurs at the down dip end of a small depression above a 4m climb down through boulders. The cave passage is developed down the dip for 55m along a prominent joint oriented at 041-231^{o}. At the base of the dip tube the passage turns along strike for 25m to the limit of exploration (Figure 5.13).

The passage is almost circular in cross section except where intersected by three strike oriented joints. Water percolating down the joints has both dissolved chambers and deposited speleothem. The lower 30m of the dip tube and the strike tube are partially infilled by cobbles covered with brown flowstone. The flowstone has been partially reeroded and leached of ^{234}Uranium ($^{234}Th/^{230}U$ activity = 1.34). A second period of calcium carbonate deposition is indicated by clean white uneroded calcite overlying part of the old brown deposit. An age of $12.4^{+4.4}_{-4.3}$ ka was determined from the Th/U activity ratio of a phase 2 stalactite. Coarse gravels in a finer grey carbonate matrix, apparently till, occur in the roof of the highest level chamber near the entrance.



3.3
E
4.5
2.4
blockage
Nm
too tight
3.6
-plan-
too tight
5
Dip Cave
0
10
20
m
.9
compass and tape M Davis
J Oldham 08 80
preliminary draft P Griffiths 06 03 81
all measurements in metres
sL = 92m

FIGURE 5.13

The cave was probably developed by stream waters flowing off the Karmutsen Fm. lavas. The presence of till and leached, eroded speleothem beneath postglacial speleothem suggests that the cave is pre-last glaciation in age, albeit much younger than Papua Cave. Thus ice action may have erased cave passages above the present entrance. The absence of vegetation debris or erosion of the young floor speleothem suggests that the stream has been inactive during much of the Holocene.

5.5.3 Inter Lava Cave Development

Brutal Cave Brutal Cave occurs at the northern fault boundary of the Minigill limestone wedge. The entrance is perched 6m above a stream channel incised in the fault. The cave is developed in a limestone band within Karmutsen Fm. lavas which has a stratigraphic thickness of about 10m. Passage development is mainly along joints oriented parallel to the dip (037-227^{o}), strike joints and aslant dip joints (0-180^{o}). Stratal dip is 37^{0}.

The cave has developed by the integration of two separate passage systems (Figure 5.14). Brutal 1 is basically a series of closely spaced, steeply inclined phreatic tubes, most of which are choked with sediment. In one location three phreatic tubes up to 2m in diameter occur in a single 10m section of a bedding plane. In the lower section of the cave the passage form is a low bedding plane passage greater than 30m in width along strike.

Brutal 2 is linked to Brutal 1 near the system entrance via

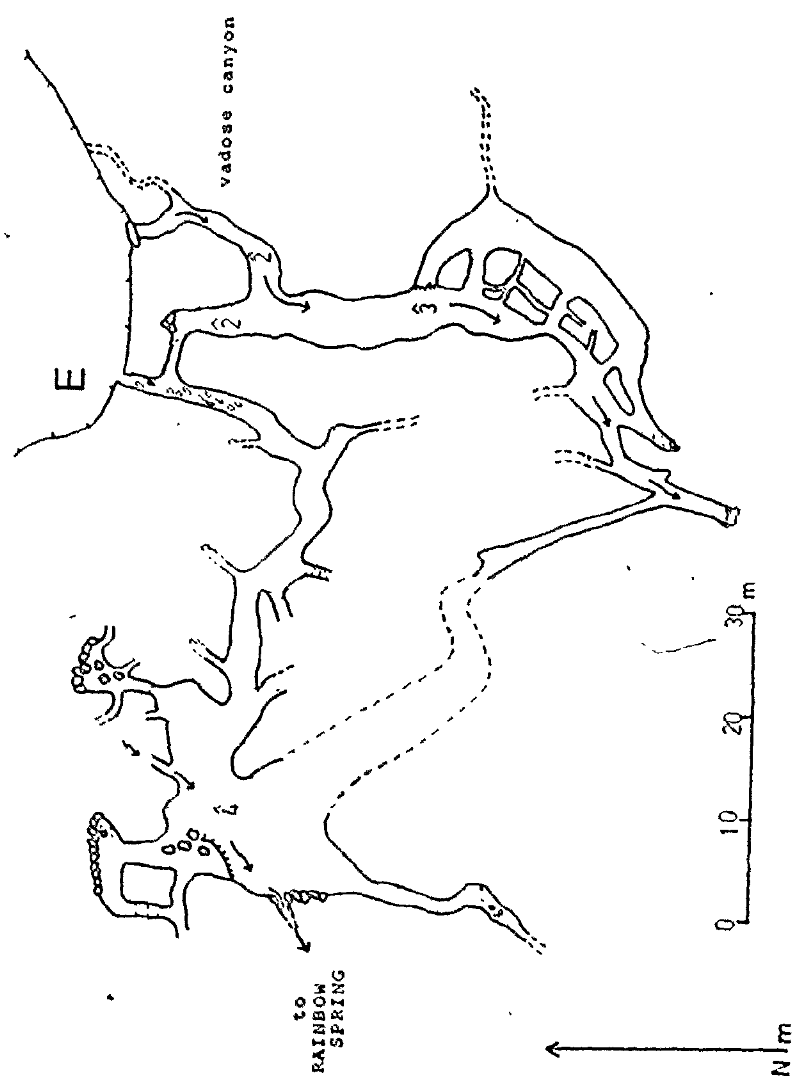


FIGURE 5.14: BRUTAL CAVE (based on survey by H. Davis and J. Oldham, 1980)

a passage developed by collapse. The cave contains a major route 5m in width developed aslant dip. In the lower section the passage is up to 15m wide. The lower section of Brutal 2 is connected to Brutal 1 by two aslant strike passages. The bottom of both Brutal 1 and 2 are too small for exploration. A narrow vadose canyon at the up dip end of Brutal 2 routes water into the cave during floods in the surface stream (M. Davis pers. comm.)

The high passage density in the Brutal cave system is typical of cavern development in confined situations. The cave probably developed by capturing water in the surface stream when the stream channel was at a higher level. The passage form suggests that both series were formed mainly under phreatic conditions. Thus both series are apparently contemporaneous in age. The direction of water flow during the creation of the (strike) connecting passages is unknown.

Flood waters entering the cave during the fall of 1980 were traced to Rainbow Spring located within Karmutsen Fm. lavas to the North of the limestone outcrop (M. Davis pers. comm.*, see Figure 5.11). This demonstrates that the overlying lavas prevent waters rising directly upwards stratigraphically into the Quatsino Fm. limestone.

Peat Cave The second interlava cave occurs at the base of a blind stream swallet greater than 10m in depth. Below the entrance the passage is oriented aslant strike. Karmutsen Fm. lavas form both the floor and roof along approximately 0.3 of the passage. About 100m from the entrance the large passage is blocked by sediments which

* Martin Davis - British Columbia Speleological Federation

fill it to the roof. The base of the fill is composed of coarse clastic sediments but, the upper few metres are dominantly fine organic material. The passage apparently continues aslant dip but is inpenetrable. (figure 5-15).

A dyke orientated at 005-185° is exposed on the down dip side of the passage beside the sediment choke. The dyke is breached by a small stream which flows in a small passage alternating along strike and down dip to a sump. The passage is less than 2m in diameter and is partially infilled by cobbles.

Peat Cave has developed by capturing a stream generated on the Karmutsen Fm. lavas. The initial passage development occurred aslant strike perched above the dyke. Dyke breaching at a later date resulted in passage development down the dip. The large aslant strike passage is probably pre-last glaciation in age. The smaller dip tube is probably post-glacial in age and may have developed after the large passage became blocked by sediment. The cave contained a small stream during the summer low flow period but extensive flooding of the young passage is evident from organic material deposited in the roof.

5.5.4 Minigill Cave

Minigill Cave is developed parallel to the Raging River at the base of the valley. The cave routes water from the Raging River along a 1.3km flow path to Pitchford's Resurgence. Most of the downstream 800m of the system has been explored. The upstream 500m is unknown except for a 50m section below the stream sink. The geomorphology of

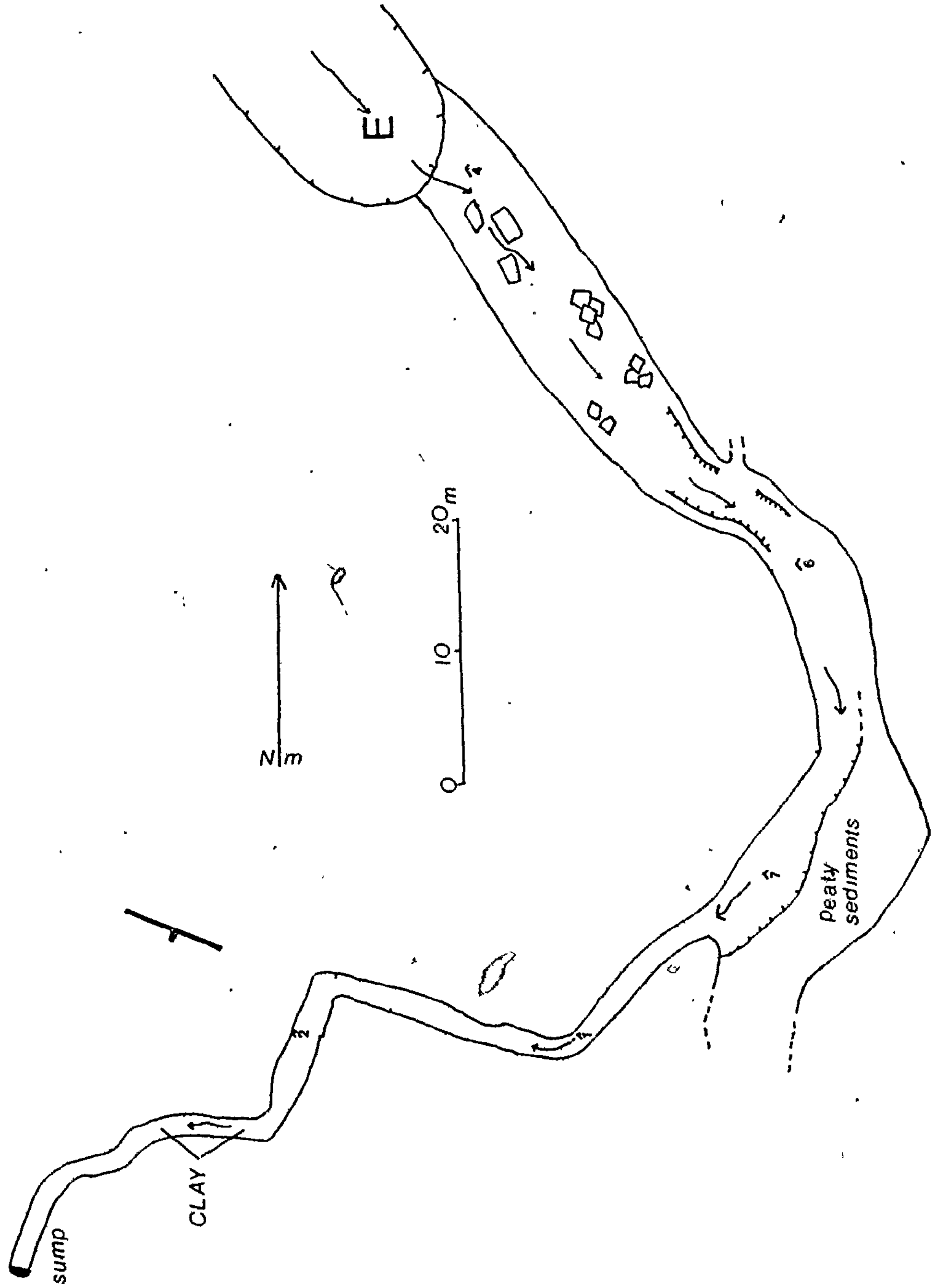
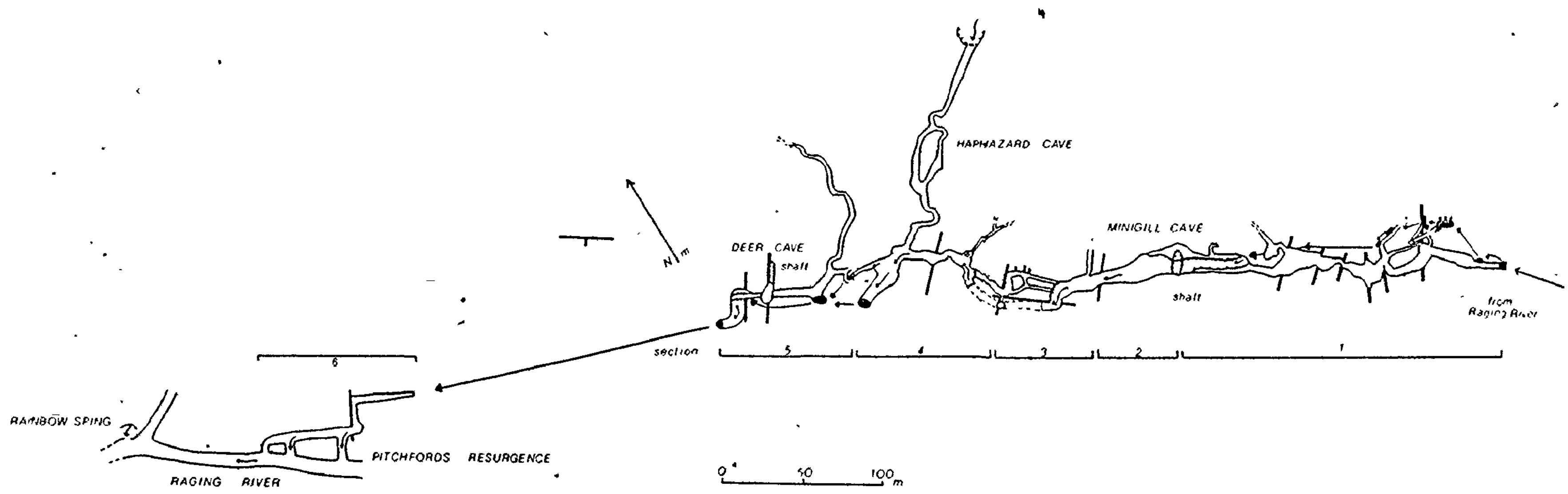


FIGURE 5.15: PEAT CAVE (based on survey by P. Griffiths
and K. Bischoff, 1979)

Minigill Cave is complex. The system ahs been divided into sections for the purpose of description (Figure 5.16). Development of the system as a whole is considered at the end of the section.

The Raging River stream sink is located at the downstream end of a small gorge. Water sinks into a strike aligned paraphreatic passage 1.5m in diameter. Twenty metres from the entrance the passage turns down the dip and the stream enters a sump after 15m. The direction of flow from the sump to the next downstream section of the cave system is roughly against the stratal dip. Thus, the flow path must turn through 180°. Flourescein dye placed in the stream sink was carried through the cave system to Pitchford's Resurgence 1.3km to the North West in less than 16 hours. This indicates a mean flow velocity of greater than $80 m \cdot hr^{-1}$ along the flow path gradient of about 0.015.

The longest segment of the cave streamway explored from one entrance exceeds 500m in length. The cave passage is developed along the strike. Section 1 comprises 220m of the cave developed upstream of the base of a 33m shaft. The shaft is developed along a fault oriented parallel to the stratal dip and measures 15m by 7m in plan. Section 1 is composed of high level fossil passages and a para-phreatic streamway with at least six sumps along its length. The fossil passage is perched about 5m above the active streamway. The fossil passage has a diameter of 3m but three chambers (First Chamber, Second Chamber and the Terminal Room) with a ceiling height of up to 12m occur along its length. The chambers have developed primarily by collapse either



MINIGILL WEDGE VALLEY BOTTOM CAVE SYSTEM (Based on surveys by P Griffiths K Bischoff E Vorkampff M Davis & J Oldham)

FIGURE 5.15.

side of dykes oriented parallel to the dip. Four dykes occur in the Second Chamber. Wall and roof pockets along the fossil passage contain the remains of an extensive infill phase of coarse sediments in a grey carbonate matrix. In a wall pocket in the Terminal Room the coarse fill is overlain by a clay deposit. Calcite with a Th/U age of 3.9 $\pm$ 5.7 ka overlies the clay deposit. South East of the Terminal Room the form changes to a low, aslant strike 10m wide bedding plane passage, which drops about 5m in elevation to a sump.

The active stream passage in Section 1 of Minigill Cave is complex (Figure 5.17). Water passes through 6 sumps in a flow path less than 100m in length. Flow in sump 1 is along strike to a window in the bedding plane passage South East of the Terminal Room. Flow in sump 2 is aslant strike. Water emerges from the sump flowing along strike perched on an inpermeable bed. Three dip tubes, developed above the impermeable bed join the passage along its updip side. The vadose stream turns downdip and enters sump 3. Flow through sump 3 is aslant strike; however, the stream emerges from the sump flowing up dip perched on the impermeable bed. The vadose stream between sumps 3 and 4 has incised through the impermeable bed and water enters sump 4 at a lower stratigraphic level than the exit from sump 3. The direction of flow through sumps 4, 5 and 6 is aslant strike. The sumps are separated by short pool chambers accessible through the floor of a high level passage. Water entering sump 6 re-emerges about 90m to the North West, 30m upstream of the entrance shaft.

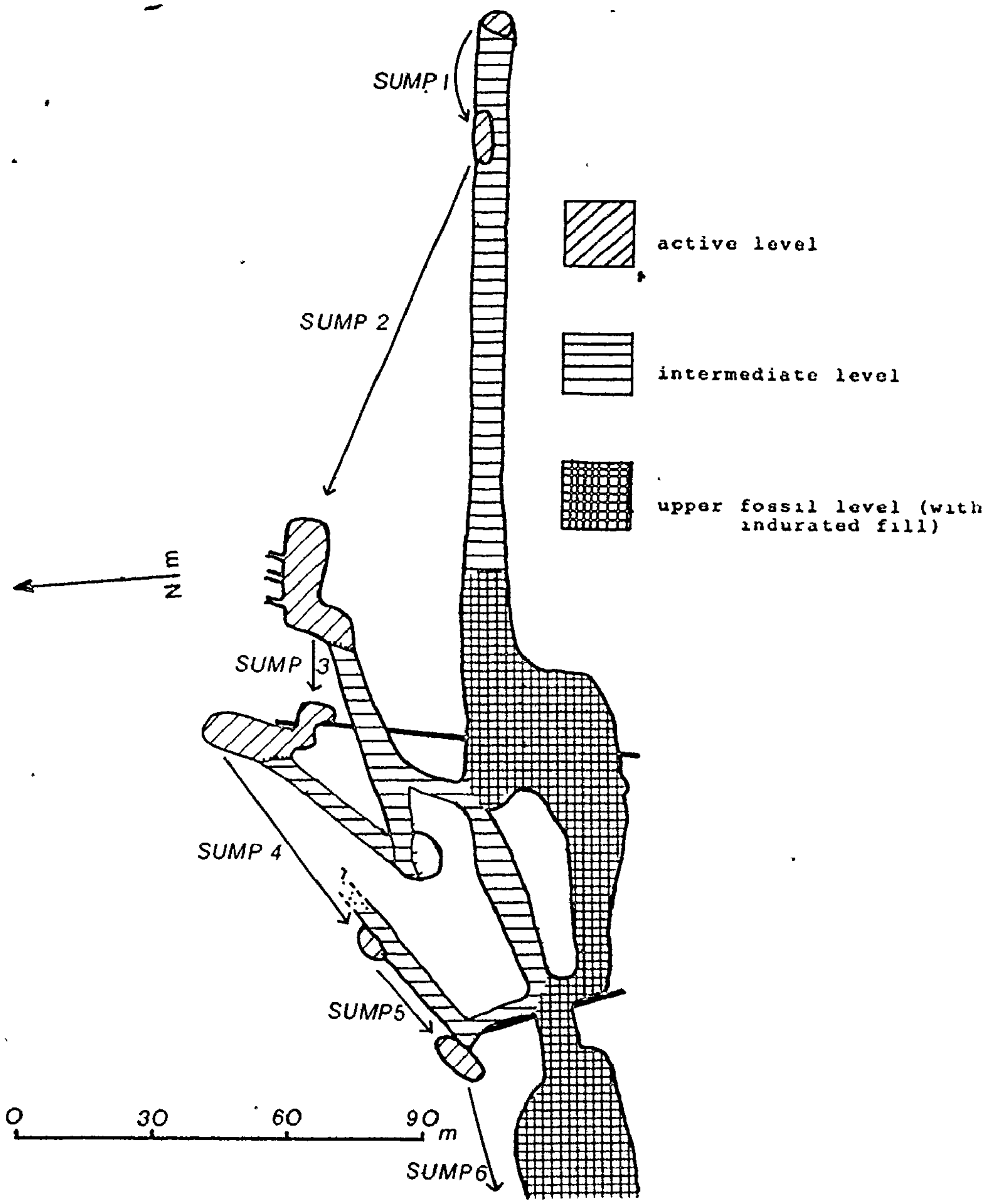


FIGURE 5.17: MINIGILL CAVE (Section One)

The stream flow path in upper Minigill Cave is difficult to interpret. The highest and apparently oldest stream level was oriented along the strike from sump 1 through the 3 large chambers to the base of the 33m shaft. However, this stream path was probably turned from a straight strike path by the dykes in the chambers. Local passage development parallel to the dip is indicated in these locations by the size of the collapse chambers. The collapse has obliterated the original passage bends leaving an almost straight, strike aligned fossil passage.

Present stream passage development is also controlled by the non-limestone barriers. Flowbetween sumps 2 and 3 and in sump 3 is perched on an impermeable bed. The flow path in sump 3 may be turned updip by a dyke which also outcrops in the Terminal Room. An intermediate level fossil passage occurs between the Second Chamber and Terminal Room and the active sumps. It appears that the flow paths through the active sumps were not developed at the same time and the intermediate level functioned to route water around poorly developed passage, notably bypassing sump 3. However, the sequence of development of all passage in this section of the cave is uncertain. Flow paths produced by non-limestone barriers such as those seen in the short cave sections of the Malook Cave system are shown in Upper Minigill to be very complex in detail.

Section 2, located on the downstream side of the entrance shaft is a simple strike oriented streamway passage 40m in length.

Downstream from Section 2 the passage form is changed by the occurrence of major dip tubes along the up dip side of the passage, and multiple development of overlying strike tubes. The cave system has breached a strike oriented dyke on its down dip side and flows in a young phreatic strike tube. The cross-sectional shape of the tubes indicates most formation under phreatic conditions. Scallops between the top two strike passages indicate some passage formation by upward flowing water.

Waters flowing down the dip in different bedding planes flowed into the strike galleries. The scallops indicate that the water may once have risen upwards through the strike series at the south-east end to discharge from the highest passage. Breaching of a dip dyke at the north-west end of the section may have drained the series leaving the stream in the lowest presently fossil level. More recently, breaching of the strike aligned dyke initiated the present phreatic stream passage.

Section 4, comprising the downstream section of Minigill Cave is a single level stream passage with a diameter greater than 5m. In the downstream 30m the streamway turns aslant dip and divides into 2 passages leading to sumps.

Haphazard Cave, carrying a small vadose stream joins Minigill Cave along the up dip side of Section 4. The cave entrance occurs in the base of a stream swallet about 120m up dip from the Minigill streamway. Part of the passage is flooded on a non-limestone bed, perhaps Karmutsen Fm. lavas. However, the passage gradient is less than the stratal dip in the lower section and the passage rises above the lower surface. About 100m from the entrance, the down dip passage turns on strike for 10m forming a narrow vadose canyon developed above a 10m pit. At the base of the pit the passage is floored on the lavas again as far as the streamway. Phreatic development in the Haphazard Cave passage is resticted to a tube approximately 0.5m in diameter. In the strike aligned passage above the 10m pitch, large slow flow phreatic scallops in the phreatic tube overlie small fast flow scallops in the vadose canyon. Haphazard Cave developed by routing a stream from the lavas into the cave drainage system at the base of the valley.

Deer Cave Deer Cave (Section 5, Figure 5.16) is the downstream continuation of the valley bottom strike aligned cave system. The entrance shaft occurs on the south east (upstream) side of a dyke oriented parallel to the dip. The 20m shaft is developed above a fossil strike passage.

The cave is composed of two passage series. A large, fossil, dip tube explored for 80m to a sediment blockage (Eric Vorkampff pers. comm.) is connected to the fossil strike tube. The 20m strike tube is connected to the base of the shaft. The active stream passage routing water from Minigill Cave underlies the fossil strike passage. The stream flows along the strike via a window in the floor of the fossil passage, through two dykes before turning parallel to the dip, for 20m into a large sump.

Formerly, water was routed down the dip from the mountainside in the fossil phreatic tube along the strike passage and rose up the shaft. At present the cave routes Raging ~~River~~ water towards Pitchford's Resurgence Cave.

Pitchford's Resurgence Cave Pitchford's Resurgence Cave discharges waters draining along the strike in the Minigill system into the Raging River channel. Water flows along strike from an upstream sump, before being turned aslant dip by a dyke oriented at 075-255°. The stream has breached the dyke in a junction chamber and drains along strike towards three excurgence passages developed along fractures parallel to the dip (see Figure 4.16). The outlet furthest from the breached dyke is totally fossil. The intermediate spring discharges water in a vadose stream. The spring closest to the breached dyke occurs at the lowest topographic level and is fully phreatic. The outlet series demonstrates route evolution towards the shortest flow path from the dyke to the river channel. The initial development was along the longest strike path but shortest path parallel to the dip. This demonstrates that cave development in this location is limited by the rate of development of the segments parallel to the dip.

5.5.5 Discussion

A complex system of multiphase cave development exists in the limestone wedge at Minigill. Papua Cave, Dip Cave and Peat Cave/ Brutal Cave indicate cave development when the valley floor stood at three different but higher elevations (Figure 5.18). This suggests cave development interupted by at least two glacial advances.

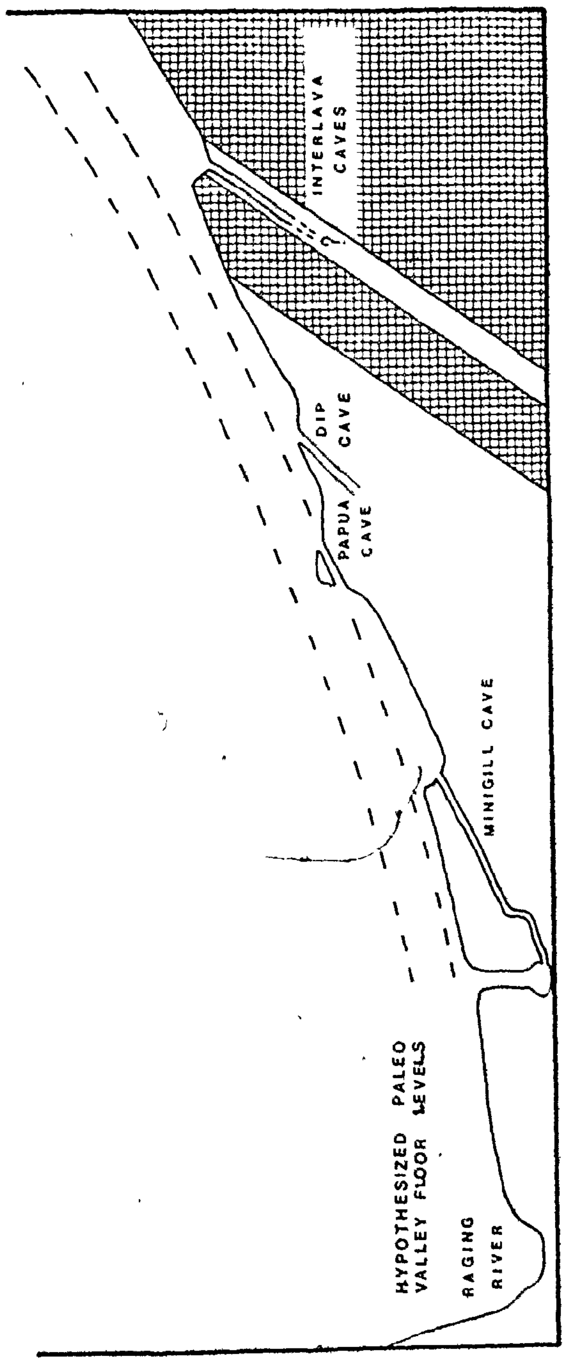


FIGURE 5.18: Multiphase cave development at MINIGILL

The valley bottom cave, at present, serves primarily to route waters from the Raging River. Most of the drainage down the mountain slope is routed through the inter lava limestone to the Rainbow Spring outside the karst. However, the valley bottom cave has apparently developed from three separate down dip cave systems (Section 1, Section 3 and Deer Cave). These routed waters down the mountainside and along the strike between dykes to discharge sites developed at the top of lifting shafts. The large entrance shaft in Minigill Cave probably discharged Raging River water, sinking at a higher level, prior to the downstream breaching of the dykes, the generation of Sections 2 and 4, and the lowering of the water table. The Raging River water was probably pirated underground into a pre-existing cave system by local down dip drainage.

Cave development, especially in Section 1, is still strongly controlled by the non-limestone impermeable barriers within the Quatsino Fm. The complex geomorphology of the valley bottom warrants further analysis to explain, in greater detail, the effects of compartmentalization between impermeable barriers, and the changing base levels. The nature of the problem has been summarised by Ford (pers. comm.):

> "(1) locally the cave has a very high frequency of dip tubes = state 4 type system but, the cave is highly compartmented by dykes and impermeable beds.
>
> (2) further, the cave exhibits multiphase development, with a major filling phase, indurated so that water has had to clear or get round the indurated fill.
>
> (3) the logic of cave building is headwards but because of 1 and 2 very complex patterns of development, not immediately ammenable to a headwards analysis a la Ewers / Ford, exist."

The indurated fill in the high levels of Minigill Cave may be glacial in origin, indicating that the cave is pre-last glaciation in age. This is supported by the height of the entrance shaft which, if it functioned to discharge river water, indicates that the valley floor stood above 33m above the present level during shaft development. The depth of post-glacial incision is probably similar to the depth of the Raging River gorge (10-15m).

The absolute age of cave development at Minigill is uncertain. However, if Minigill and Dip Cave, both of which remain beneath the general karst surface, are pre-last glaciation in age it can be suggested that Papua Cave, which is perched in a scarp wall some 90m above the Raging River, is a very ancient feature.

5.6 Devil's Bath

5.6.1 The Devil's Bath Cave System

The Devil's Bath system is located in the western part of the Benson Valley. Benson River flows in a south westerly direction from Kathleen Lake, basically parallel to the dip, to the western side of the valley where it turns sharply NW to strike to flow to Alice Lake. In the dip section the river flows in a gorge up to 20m in depth.

A large, sub-vertically walled, cenote-like depression, the Devil's Bath, occurs less than 100m from the Benson River, near the turn to strike. The Devil's Bath measures approximately 80m by 60m in plan. It is occupied by a permanent pond whose water level is about 20m below the rim of the depression. The Bath is the largest single surface karst feature in the region occupying a void space 150 by 10^5 m^3. Stratal dip at the Bath is 29^o.

The Devil's Bath is hydraulically connected to the Benson River. During July and August 1980 all flow in the Benson River was abstracted from the gorge and routed along an underground flow path to the Bath. Scuba divers report the presence of a discrete input shaft, rising from an unknown depth, at the base of the Devil's Bath 25m below the water surface (Paul Griffiths pers. comm.). Water entering the sinks in the gorge rises up through about 80 stratigraphic m of limestone to reach the present water surface in the Bath, although the height of any particular lifting chimney is unknown.

Drainage from the Devil's Bath is along the strike to the Devil's Springs, 400m to the north west, where waters debouch along a major strike-aligned fissure. The geomorphology of the area was studied in an attempt to determine (1) the genesis of the Devil's Bath and, (2) the relationship between the Bath and the adjacent Benson gorge. The situation is outlined on Figure 5.19.

About 400m upstream of the Devil's Bath the Benson River turns on strike and flows to the south east for about 50m. The river then turns through 180° and flows to the north west for 50m before flowing in a direction parallel to the dip once again (see Figure 5.19). The turn is controlled by a dyke oriented along the strike located in the river channel. A cave is perched on the up dip side of the dyke about 2m above the present stream level. Two short dip tubes are connected to a strike orientated tube about 3m in diameter. The cave shows no evidence of any development under vadose conditions.

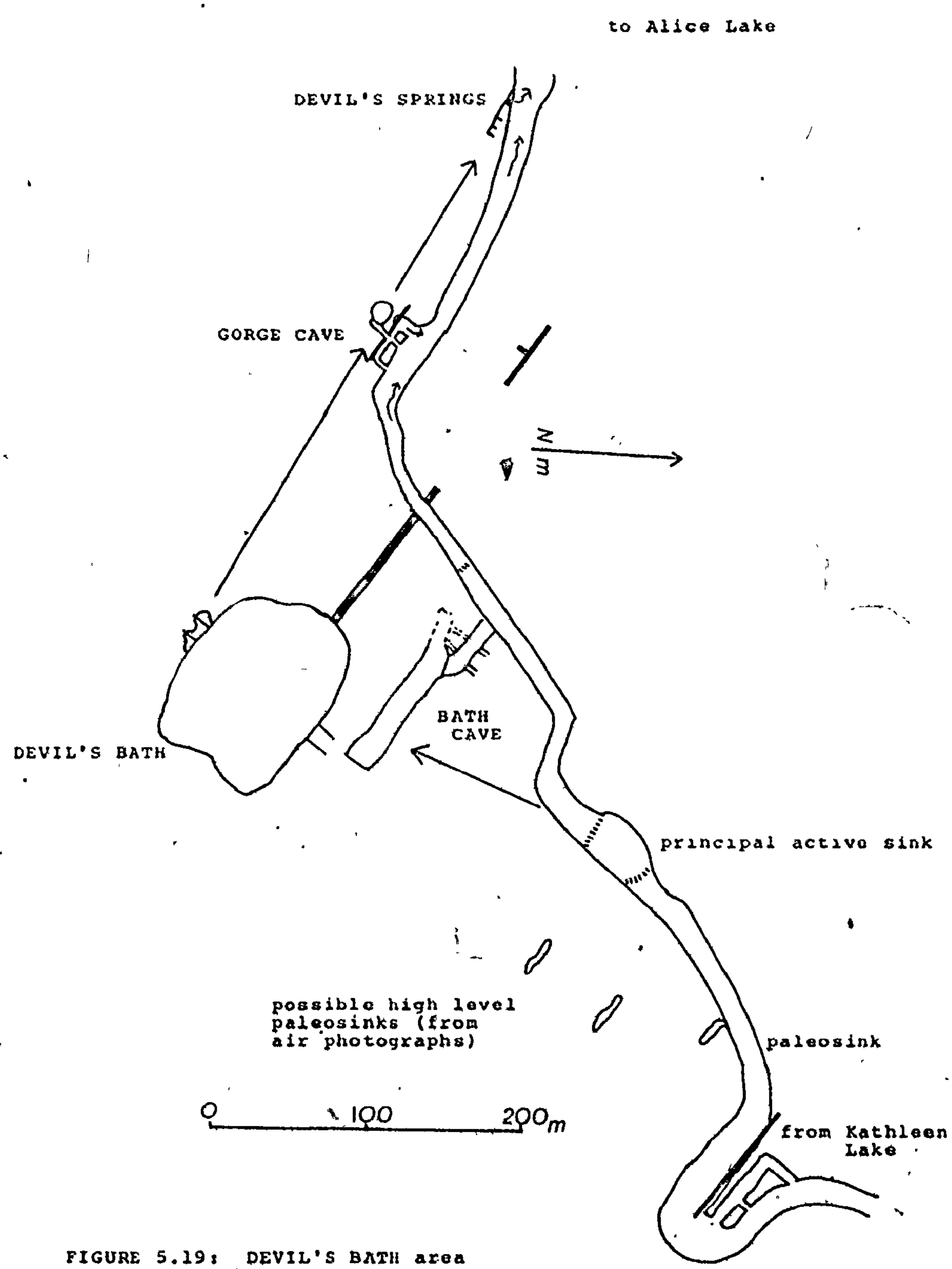


FIGURE 5.19: DEVIL'S BATH area

Benson River Cave A 120m long strike orientated cave is located up dip of the Devils Bath. The south easterly 60m of the cave consists of a large 10m wide passage containing a lake of unknown depth which sumps at either end (Figure 5.20). Small scallops occur along the length of the passage. The north westerly 60m of the cave is a phreatic tube 3m wide perched 2m above the lake. This passage has several dip tubes aligned along both its up and down dip side. The lake remained throughout the summer of 1980 despite the dry river channel outside the cave.

Bath Cave The Bath Cave is located at the down dip end of the Devils Bath. The roof of the cave, perched 12m above the water surface, is a solutionally modified bedding plane surface up to 10m wide. Two major joints about 10m apart, orientated parallel to the dip, traverse the Devil's Bath. The Bath Cave has developed by solution and bedrock collapse between the two joints. Water flows from the Devil's Bath to sump in the Bath Cave less than 10m to the south west.

Gorge Cave The Gorge Cave is located on the down dip side of Benson River gorge below the principal turn to the strike. The cave has three entrances along a 33m section of cliff (Figure 5.21). Entrances 1 and 2 are perched above the river level, but the bottom of entrance 3 is below it. The three entrances are connected to a strike oriented passage by short passages parallel to the dip. The strike tube between the first and second entrances is about 1.5m in diameter and is perched 3m above the river level. Between the second and third entrance a vadose trench has been cut into the floor of the tube and the

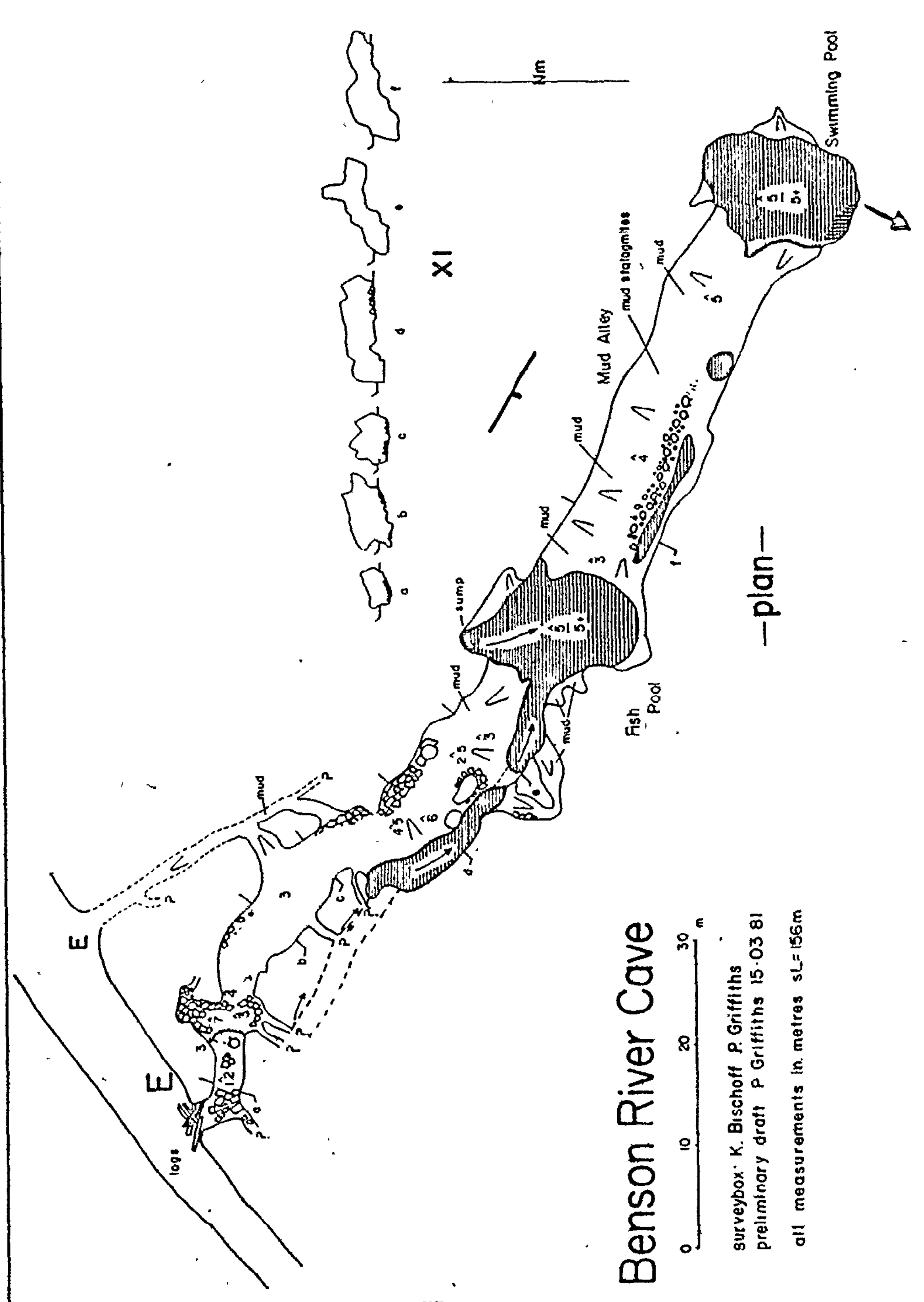


FIGURE 5.20: BENSON RIVER CAVE

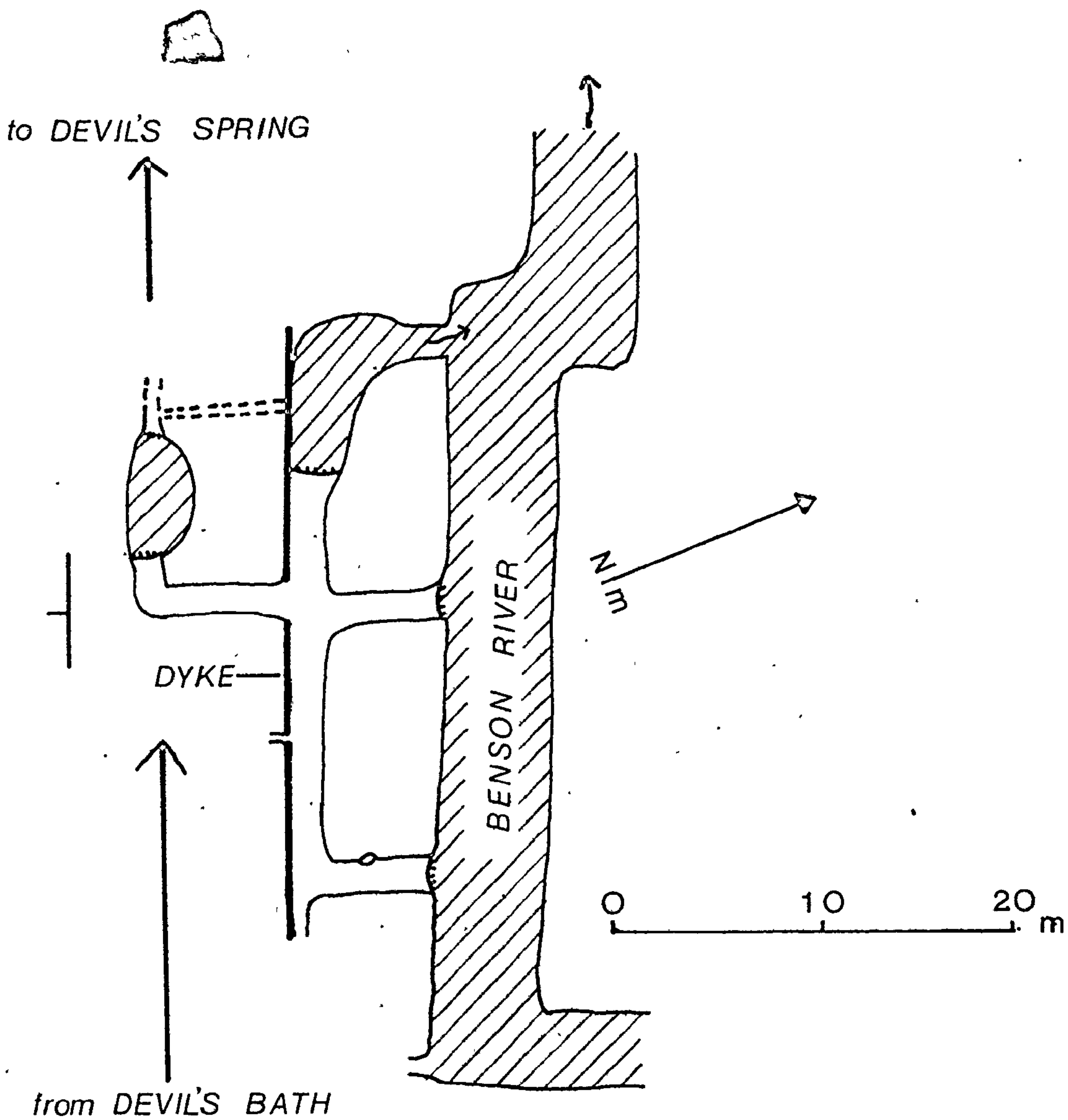


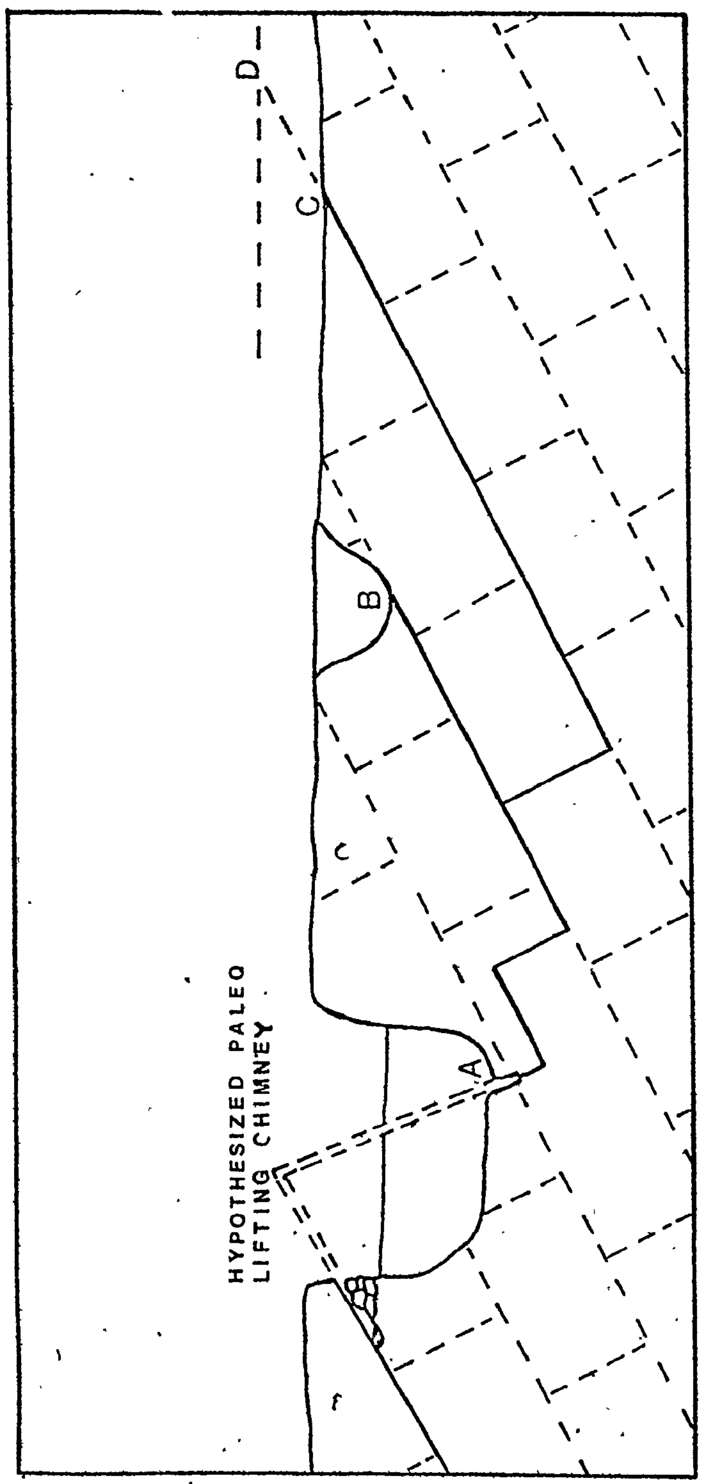
FIGURE 5.21: BENSON GORGE CAVE

passage is greater than 4m high. A strike oriented dyke is exposed on the down dip side of the passage. This is breached by a passage orientated parallel to the dip which leads to a sump pool. Scallops developed on passage walls in the cave indicate that water from the sump flowed towards the river via the three entrances.

About 250m north west of the Gorge Cave water discharges along a major strike aligned fracture at the four Devil's Springs. The dry season discharge of the springs is greater than $1.5m^2 s^1$.

5.6.2 Discussion

The Devil's Bath is a large karst depression not readily explained by development under the prevailing hydraulic conditions. Extrapolation of the solutionally modified bedding plane roof of the Bath Cave to a position above the input shaft at the base of the Devil's Bath indicates that water may have risen 40m above the present water surface here, ie. 20m above the depression river, in a joint chimney (Figure 5.22). The water was probably forced to rise upwards by a strike orientated dyke, observed in the river gorge adjacent to the Devils Bath. The Devil's Bath apparently developed by collapse around the joint chimney, primarily along the two major joints oriented parallel to the dip. The maximum elevation above the water surface of a paleosink is about 30m and thus the site of the sink which provided the hydraulic head for waters ascending the chimney is unknown. However, many metres of bedrock have probably been removed from this valley bottom position by the Pleistocene glaciations. Consequently a high level sink may have been erased.



A. Input Shaft, B. River Sink in Gorge, C. Intermediate Level Sink,
D. Paleo River Level.

FIGURE 5.22: Idealized drainage pattern in the DEVIL'S BATH cave system

The Benson River Cave routed waters along the strike in front of the dyke in a manner similar to the fossil cave 400m upstream from the Devil's Bath. However, the cave is developed at the same level as the present water table in the Devil's Bath and is perched at least 25m above the base of the input shaft. Thus, the exact relationship between the cave and the development of the Devil's Bath is uncertain.

Waters routed through the Devil's Bath cave system formerly discharged from the Gorge Cave. At present most of the water is discharged from the Devil's Springs. The outlet of the Gorge Cave probably developed at an earlier date as a result of the steeper hydraulic gradient to this site from the Devil's Bath, although ultimately all waters will discharge from the Devil's Springs because the latter are located at a lower elevation.

The narrow gorge that channels the Benson River in the vicinity of the Devil's Bath is probably younger in age than the cave system. It has developed by routing waters more efficiently along a steeper potential energy loss gradient than the cave system with the large phreatic loop. This development is similar to the tendency for multiphase cave development towards a water table type with time described by Ford and Ewers (1978):

> " (Cave systems) may be subject to gradational modifications which change their character towards that of a higher state (eg. ideal water table caves)."

5.7 The Vanishing River - Reappearing River Cave System

The function of the Vanishing River - Reappearing River system is unique in the Benson karst. The system routes water from Parson Bay Fm. rocks in the upper reaches of the Benson Valley against the dip, through a major topographic divide, to the Raging River Valley. In the Malook, Minigill and Devil's Bath systems waters discharge from stratigraphically higher positions than the stream sinks, however, waters enter this system at the top of the limestone and discharge close to the stratigraphic base. This situation results from groundwater flow along a fault zone.

Vanishing River Cave The stream sink is located in the Quatsino Fm. close to the Parson Bay Fm. contact. Water cascades down a 15m deep pothole to the base of a large chamber over 30m in diameter. The chamber has developed by collapse along a major fracture tilted about 45° from the vertical; most of the chamber floor is composed of large limestone blocks. The stream flows through the breakdown into a network of narrow passages developed along fractures parallel to the dip, to sumps. Many of the fractures are infilled with calcite which indicates an ancient fault zone. Over 300m of vadose passage exist beyond the sumps but these were not examined during this study.

Lost Richard Cave Lost Richard Cave is located south of Reappearing River Cave in the Raging River valley. The cave is believed to be part of the system from Vanishing River. The cave entrance is perched about 40m above the level of the modern system resurgence. The cave passage is predominantly developed in a single bedding plane along a series of joints orientated parallel to or aslant the dip. Below the entrance the passage is an opened bedding plane up to 20m wide but less than 1m high (Figure 5.23). Below this 'slot' passage the cave is a series of strike oriented passages connected by short dip connections. The lower strike passage is over 60m long. East of the Junction this intercepts 3 passages developed along prominent dip joints. The two easterly down dip passages continue on the lower side of the strike passage to a sump and sediment choke, respectively.

West of the Junction the strike passage is connected to the hydrologically active part of the cave, the Roman Baths. The Roman Baths are a paraphreatic network of passages containing a pool between two sumps. Water flows along the strike to the north west towards Reappearing River resurgence.

Lost Richard Cave is hydrologically inactive except for the south easterly sump and the Roman Baths. The cave is part of the system connecting Vanishing River with the Reappearing River resurgence. The cave may once have functioned as a stream sink pirating water from the Raging River when the valley was higher. However, the slot-shaped passages here may be more typical of resurgence caves where waters are forced upwards to springs. If Lost Richard developed as a resurgence cave, the water most probably entered the cave from the south-easterly sump and rose up dip in the slot passage. As the downstream resurgence at Reappearing River developed the route through the Roman Baths was developed. Thus the slot may be part of the oldest phase in the cave.

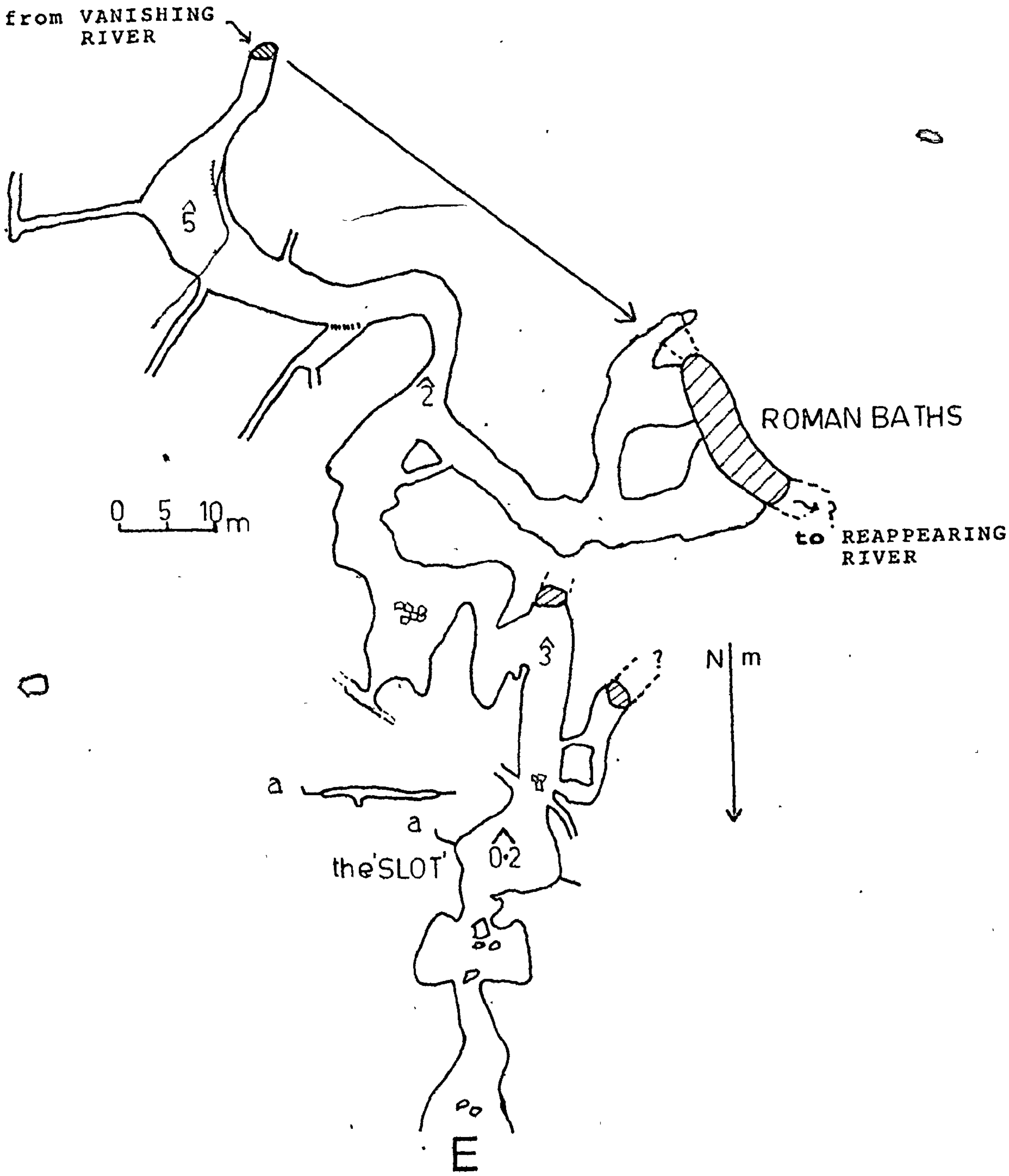


FIGURE 5.23: LOST RICHARD CAVE (based on survey by P. Griffiths, 1977)

With time, shorter and more efficient routes developed at progressively lower levels. At present the partially phreatic streamway is developed along strike at the base of the cave.

Reappearing River Cave The cave is entered at the base of a scarp in the western side of the Raging river valley. The flow path from Vanishing River cave was proven with a dye trace using Amino-G acid and raw cotton detectors. Water moves in a northerly direction along the flow path; in the 300+ m of passage examined in the Resurgence cave water flow was either along the strike or against the dip (Figure 5.24).

The upstream 30m of the streamway is a young partially phreatic passage developed along a dyke. A fossil passage containing an old phreatic loop which routed water down a joint and up a bedding plane bypasses the young streamway. Sump 3 occurs where the streamway is confined beneath an impermeable bed. The sump is bypassed by a narrow, high level passage developed above the impermeable bed along a joint parallel to the dip.

From the Junction to the entrance the cave is developed along the strike. A straight, high level passage occurs above the actual streamway. The streamway is locally turned along dip into sumps by dykes. A series of fossil, sediment choked, dip tubes occur along the up dip side of the high level strike passage. A small fossil passage, developed aslant strike, occurs above sump 3. This has been explored to an elevation of 15m above the streamway, to a sediment choke capped

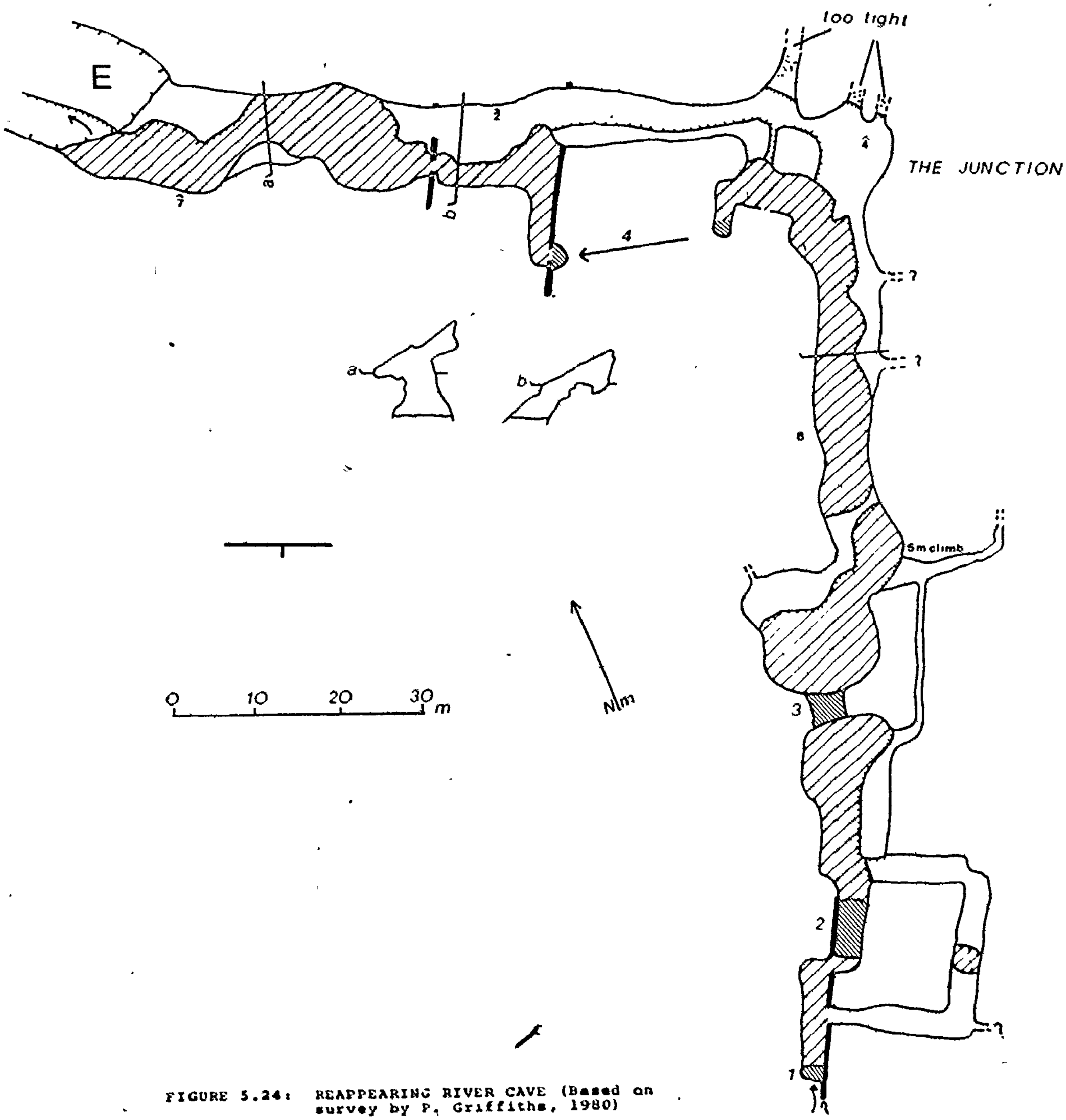


FIGURE 5.24: REAPPEARING RIVER CAVE (Based on survey by P. Griffiths, 1980)

by a thin flowstone layer with a Th/U age of $39.9 {}^{+2.7}_{-2.6}$ ka. B.P.

Two contrasting hypotheses are presented to explain the genesis of Reappearing River cave. (1) The cave developed initially to route waters sinking into the overlying limestone at a higher level. The water source may have been the Raging River when the valley base level was higher. Waters routed down the dip tubes were integrated into a single, dominantly strike-aligned passage to the outlet. This passage system later became the target for waters from the Vanishing River sink which developed the present streamway, leaving the local drainage system perched in the roof. (2) The fossil, high level bedding plane passages at the junction routed waters from Vanishing River upwards to springs prior to the breaching of the downstream dykes and the development of the strike outlet. The flowstone date indicates that the high level routes were inactive by the latter part of the Mid-Wisconsin interstadial.

5.8 Scorch Cave

Scorch Cave is located on the limestone block to the south of the Vanishing River- Reappearing River system, between the headwaters of the Benson and Raging Rivers. The cave is a classic example of the drawdown vadose type (Ford and Ewers 1978) typical of the Burren, Eire or Yorkshire, England. Its form is almost unique in the Benson karst.

The cave entrance occurs at the base of a 6m deep blind swallet. The surface stream has cut through greater than 2m of till. The cave drains two streams flowing off Karmutsen Fm. lavas. One sinks in the

the valley upstream from the blind wall; the location of the second sink is unknown but the stream enters the cave at the top of a 10m aven. The 50m cave passage is mainly a tightly meandering vadose canyon developed upstream of a 10m, 2 stage, pitch (Figure 5.25). The pitch occurs where the passage intersects a strike oriented dyke. The stream enters a sump at the base of the dyke.

The Scorch Cave passage is mainly the result of vadose stream incision; phreatic development was limited to a roof tube about 0.5m in diameter. This form is unusual in the region where most passages were developed dominantly under phreatic conditions upstream of dykes or lifting chimneys. The dyke in Scorch Cave did not prevent free drainage development and was probably breached at an early stage in the cave's development. Similar conditions prevailed above the pitch in Haphazard Cave (Section 5.5).

5.9 Caves of the Quatsino Narrows

The Quatsino Narrows are a 5km section of fiord at the northern end of the Gibson Plateau.

5.9.1 Imaginary Cave

Imaginary Cave occurs on the north side of the Quatsino Narrows. It is accessible only at lowtide. The main passage is some 40m long and oriented alongdip (Figure 5.26). A series of narrow rift passages, developed on joints oriented along strike join the main passage. Two streams enter the cave at the top of 15m and 18m avens. Sea water

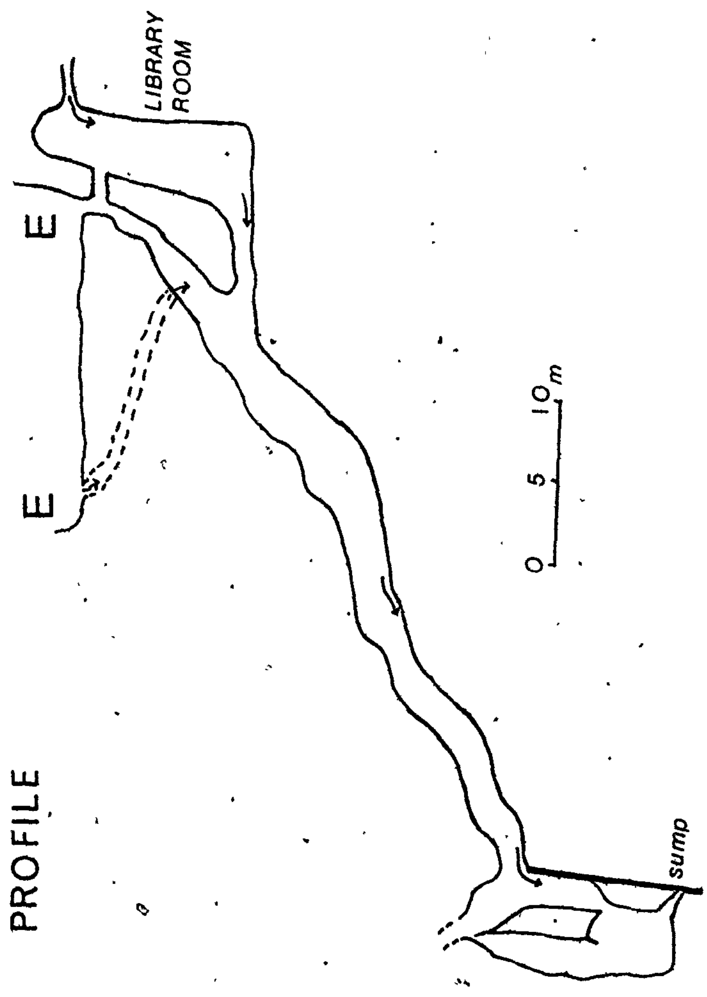


FIGURE 5.25: SCORCH CAVE (based on survey by P. Griffiths; 1980)

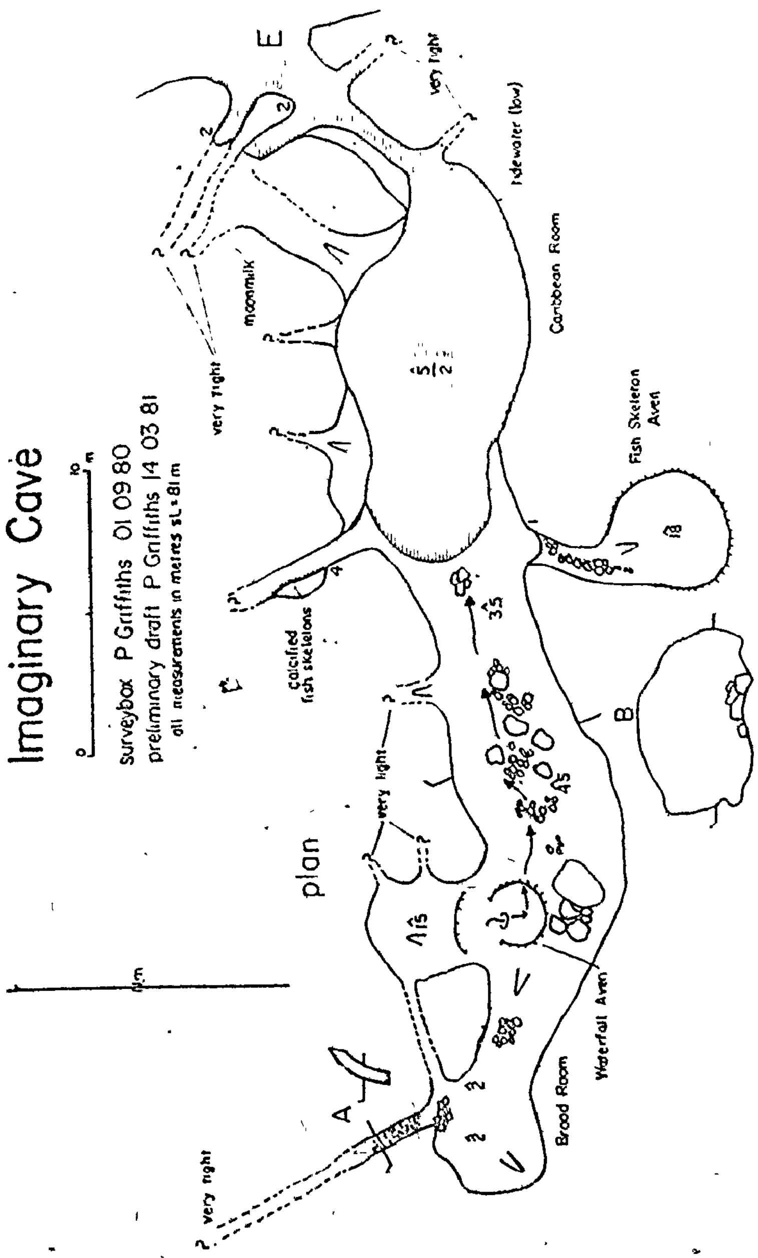

Imaginary Cave
plan
surveybox P Griffiths 01 09 80
preliminary draft P Griffiths 14 03 81
very tight
moonmilk
tidewater (low)
Caribbean Room
Fish Skeleton Aven
calcified fish skeletons
Brood Room
Waterfall Aven
A
B
E

FIGURE 5.26

pools occur inside the cave entrance even at low tide. Below the mean high tide mark the limestone passage walls are fretted. Flowstone sampled at the entrance above the high tide mark had a Th/U age of 41 ka.

Imaginary Cave developed when relative sea level in the Quatsino Narrows was lower than at present. The cave may have developed as late as the Early Wisconsin glaciation; prior to the period of speleothem deposition during the Mid Wisconsin interstadial. However, it could be much older. Imaginary Cave provides what is thought to be the first evidence of a pre-Frazer glaciation lower relative sea level stand on western Vancouver Island.

5.9.2 Burial Cave

Burial Cave is perched about 20m above sea level in the north wall of the Narrows. The cave is a fossil, strike oriented phreatic passage over 8m wide at the entrance (Figure 5.27). The back of the cave is blocked by sediment overlain by a massive, weathered calcite fill. Abundant weathered and eroded stalactite deposits occur along the passage roof.

Burial Cave is a segment of a major drainage system developed prior to the incision of the Quatsino Narrows. Speleothem samples suitable for Th/U age determination were not found. However, the cave must pre-date the development of Imaginary Cave and is probably of considerable antiquity.

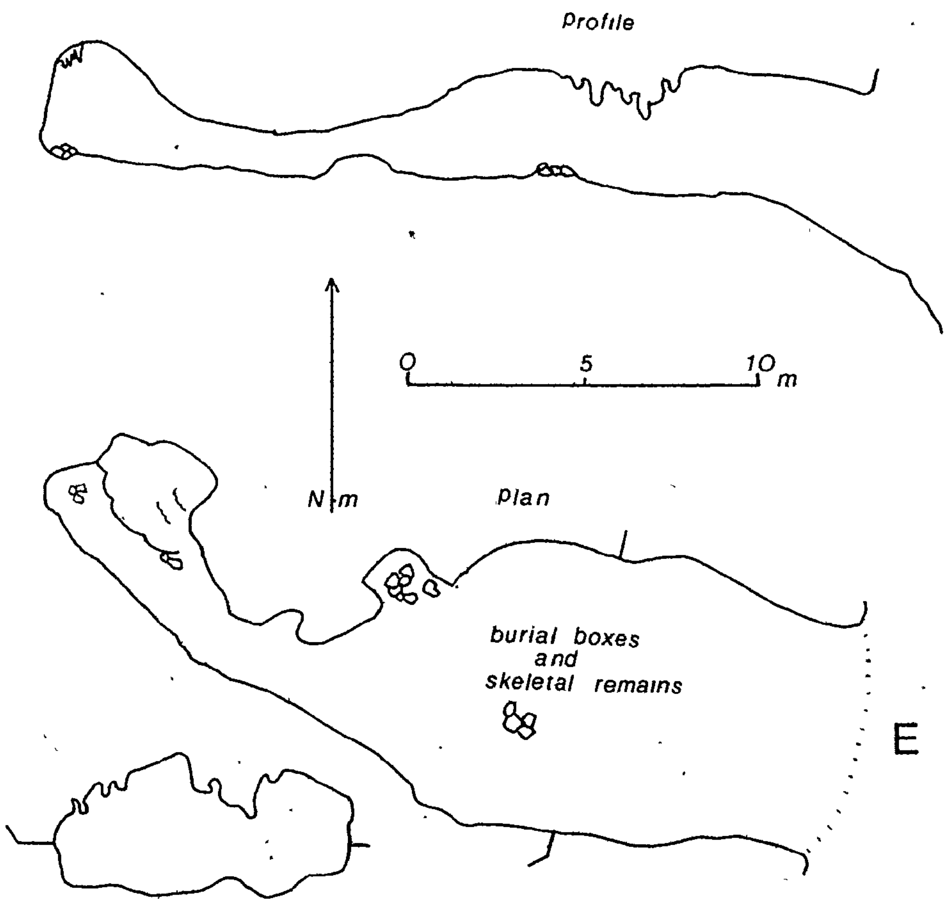


FIGURE 5.27: BURIAL CAVE

5.10 Discussion and Conclusions

Flow patterns in the steeply dipping Quatsino Fm. limestone are controlled by the framework of structural pathways available to route water and the distribution of non-limestone barriers. The structural framework is composed primarily of bedding planes and joints oriented parallel to the dip and along strike. In certain locations faults control groundwater movement. The flow barriers are mainly intrusive igneous features developed in the same planes as the structural pathways. Most frequent are dykes parallel to the dip. In Minigill Cave 11 dykes have been breached by the cave stream along a 500m section. Elsewhere drainage is observed to be compartmentalized between, or forced to rise against, the barriers. Thus, the influence of the intrusions is difficult to predict; but it decreases with time.

Recharge by sinking streams is concentrated along the eastern side of the valley where the stratal dip is roughly concordant with the hydraulic gradient. In all recharge zones groundwater flow is hypothesised to be largely in bedding planes, forming dip tubes. Near base level or in the vicinity of impermeable barriers water moves along strike or rises upwards towards springs.

Speleological studies in the allogenic karst concentrated on four systems. Exploration of much of each system was prevented as passages became phreatic or are barred by sumps. This is a consequence of the dykes or lifting chimneys and does not result from a marked downstream reduction in streamwater solvent potential. Speleological interpretations are thus based on small samples of system conduits only.

Drainage in the Malook system is aslant strike yet rises up through 450 stratigraphic m of limestone. While the direction of flow in the system is basically along the steepest hydraulic gradient the windows indicate that, locally, very complex routing patterns have developed because of the impermeable barriers. The dry valley system is hypothesised to post-date, and be independent of, the cave system.

The complex flow patterns in the Quatsino Fm. limestone indicated in the Malook windows are shown to be very complex in detail at Minigill. The Minigill system exhibits multiphase development of down-dip drainage in response to multiple stages of valley floor lowering. The valley bottom cave developed by routing the down dip drainage along strike to a series of lifting chimneys between dykes. More recent (post glacial?) breaching of the dykes lowered the water table and permitted the routing of pirated river water along the length of the valley bottom system.

The Devils Bath system occupies a position within the karst and not at the allogenic margin like Malook or Minigill. The system almost certainly originated before the last glaciation, routing stream water sinking into the karst at a higher level than the present stream sink. The Bath is the largest surface karst landform in the region directly attributed to solution (occupying a void space greater than 1.5×10^5 m^3). It is a major collapse feature at the location of a lifting chimney against a dyke. The adjacent gorge is apparently younger in age; it developed by routing water along a more efficient path in the present landscape.

The Vanishing River - Reappearing River system is unusual in the context of the Benson karst in that it routes waters eastwards from the stratigraphic top of the limestone in the Benson River Valley towards the stratigraphic base in the Raging River Valley. The system occurs in a 2km fault zone which passes through a major topographic divide. It is uncertain whether the Raging River Valley sections of the system (Lost Richard and Reappearing River) were developed initially by waters sinking into the overlying karst and became the target for drainage from the Benson Valley or, the whole system was developed by water from the latter source. If the former hypothesis is true then the capturing of the Vanishing River stream may be a post glacial development, if not the system is pre Mid Wisconsin in age.

Scorch Cave is a classic drawdown vadose type cave, unusual in the Benson karst where most passage development occurs under phreatic conditions upstream of dykes or lifting chimneys.

The complexity of flow patterns (frequency of radical change of direction of conduits and local increases in multiphase development) in the Benson karst is unparalled in the experience of D. C. Ford (pers. comm.) or the author. The complexity is attributed primarily to the high frequency of interbedded and strike and dip oriented impermeable barriers present. Indeed, the contrast between the Benson situation and Castleguard Cave in the Canadian Rockies (where individual passage elements may be guided by a single structural feature and maintain the same orientation, form and dimensions for 2km) is extreme.

It is difficult to evaluate cave system building models, eg. those of Ewers and Ford, in the Benson karst. These can only be developed by detailed reconstruction of passage genetic sequences. The difficulty arises from the above mentioned passage complexity and the small aggregate of total stream conduit length explored, which together frequently prevent the presentation of a single hypothesis of development. Moreover, certain hypotheses, eg. "the majority of systems integrate from the spring in a progression of headward steps" (Ford and Ewers 1978) cannot be accepted as a general rule of speleogenesis in the Benson karst.

CHAPTER 6

SUMMARY AND CONCLUSIONS

6.1 Summary

This study investigates karst developed on the Upper Triassic Quatsino Fm. limestone on northern Vancouver Island. The outcrop comprises a 30km long section along the base of the Benson Valley and the adjacent Gibson Plateau which rises to 900m and has an area greater than 15km^2. Characteristic features of the limestone are the consistent stratal dip to west-southwest at about 30° and the high frequency of intrusive igneous dykes and sills in the unit. The region has been glaciated extensively; evidence is preserved in till, meltwater channels on the cuestas and plateau and the Pinch Creek deltaic sediments.

Two aspects of the solution process were studied as these were thought to provide insights of geomorphic significance. Landform studies concentrated on (four) cave systems in attempts to determine the nature of factors which control groundwater flow and speleogenesis, and to interpret evidence of the development of the landscape preserved there.

6.2 Conclusions

The principal conclusions of the study are:

1. PCO_2 in the top 30cm of the soil profile does not determine the potential of soil recharge waters to dissolve limestone. Maximum

PCO_2 probably occurs in the soil profile base or fissures in the vadose "groundair" zone, at the level where plant root surface area is greatest. Measurement for the purpose of defining soil PCO_2 maximum should be made at different depths in the soil profile.

2. Groundwater PCO_2 and hence the potential equilibrium concentration of calcite in solution is apparently determined by the climatic variable, mean annual temperature. This results from the temperature control of biogenic CO_2 production. The chemical character of groundwater in the Benson Valley is close to that predicted by Harmon et al (1975) and Drake and Wigley (1975), and is thus supportive of the hypothesis that a process-response system has been clearly defined in karst terrains. The influence of other factors in setting the solution potential, ie. trace metals, could not be determined from available data.

3. While climate appears to set limits to erosion in the karst because of geochemical constraints, the fact that most water is routed through the karst in discrete channels and remains distinctly undersaturated w.r.t. calcite results in erosion rates being determined primarily by the solvent availability factor.

4. Six hydrogeologic environments were hypothesised to exist in the karst. Linear discriminant function analysis of hydrochemical data generally supports the a priori classification. No significant reduction in the ability to discriminate resulted from the introduction of only the calculated variables SI_C, SI_D and log PCO_2 into the analysis. This method was chosen because the variables are functionally independent

and between them define the potential and constraints of the solution process in a relatively easily understood manner.

5. Flow patterns and hence cave development in the limestone are dependent on the framework of structural pathways available to route water (bedding planes, joints and faults) and the distribution and orientation of intrusive barriers. The importance of individual structural elements is a function of their permeability, degree of interlinkage with other structural elements and generally the extent to which pathways composed of one or more structural elements can route water along the maximum hydraulic gradient.

6. Sinking allogenic streams are concentrated on the eastern flank of the Benson Valley where the stratal dip is roughly concordant with the hydraulic gradient. The factors which cause allogenic streams from the west to remain on the surface of the karst are unknown. However, it may result from the greater stratigraphic depth through which waters would have to flow along a given hydraulic gradient on this side of the valley and thus a greater dependence on joint elements; or the increasing clastic content towards the top of the limestone and thus the increased likelihood of the occurrence of obstructive detrital material in bedding planes.

7. Water sinking into the karst is hypothesised to flow mainly in bedding planes, forming dip tubes. The stratal dip is greater than the hydraulic gradient in most locations and waters have to rise upwards or downwards in the stratigraphic sequence or flow along strike to

reach springs. The lower sections of cave systems are mainly phreatic with sumps perched upstream of dykes or lifting chimneys.

8. Extensive Multiphase cave development occurs in three of the four pricipal cave systems studied. This is attributed to factors within the caves (eg. breaching of dykes, shortening of flow paths), and external factors (eg. lowering of base levels and altered surface stream positions) that are primarily a consequence of glaciation.

9. Uranium series dating of speleothem supports the geomorphic deduction that much of the cave passage studied is pre-last glaciation in age. Dates obtained support the hypothesis of a mid-Wisconsin interstadial on Vancouver Island (Gascoyne, 1980, and others), indicate the onset of the Holocene before 11.6 $\pm$ 0.3 ka B.P. on the Gibson Plateau and suggest the occurrence of a lower relative sea level stand on north-western Vancouver Island during the Early Wisconsin.

10. The Gibson Plateau is the most mature karsted surface in the region. The effect of glaciation on the plateau has probably been enhancement of surface form by melt water incision. This contrasts with extensive surface landform erasure by ice in the low-lying Benson Valley karst.

11. Large scale lumbering operations have deforested a significant portion of the Benson Valley karst (60%?) and the Gibson Plateau (75%?) surfaces. Soil erosion is a widespread response to these practises. Extensive sub-aerial karren development has occurred on many of the steep limestone surfaces exposed.

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