Karst and caves of Great Britain

THE GEOLOGICAL CONSERVATION REVIEW SERIES

The comparatively small land area of Great Britain contains an unrivalled sequence of rocks, mineral and fossil deposits, and a variety of landforms that span much of the earth's long history. Well-documented ancient volcanic episodes, famous fossil sites and sedimentary rock sections used internationally as comparative standards, have given these islands an importance out of all proportion to their size. The long sequences of strata and their organic and inorganic contents have been studied by generations of leading geologists, thus giving Britain a unique status in the development of the science. Many of the divisions of geological time used throughout the world are named after British sites or areas, for instance, the Cambrian, Ordovician and Devonian systems, the Ludlow Series and the Kimmeridgian and Portlandian stages.

The Geological Conservation Review (GCR) was initiated by the Nature Conservancy Council in 1977 to assess, document and ultimately publish accounts of the most important parts of this rich heritage. Since 1991, the task of publication has been assumed by the Joint Nature Conservation Committee on behalf of the three country agencies, English Nature, Scottish Natural Heritage and the Countryside Council for Wales. The GCR series of volumes will review the current state of knowledge of the key earth-science sites in Great Britain and provide a firm basis on which site conservation can be founded in years to come. Each GCR volume will describe and assess networks of sites of national or international importance in the context of a portion of the geological column, or a geological, palaeontological or mineralogical topic. The full series of 42 volumes will be published by the year 2000.

Within each individual volume, every GCR locality is described in detail in a self-contained account, consisting of highlights (a précis of the special interest of the site), an introduction (with a concise history of previous work), a description, an interpretation (assessing the fundamentals of the site's scientific interest and importance), and a conclusion (written in simpler terms for the non-specialist). Each site report is a justification of a particular scientific interest at a locality, of its importance in a British or international setting and ultimately of its worthiness for conservation.

The aim of the Geological Conservation Review series is to provide a public record of the features of interest in sites being considered for notification as Sites of Special Scientific Interest (SSSIs). It is written to the highest scientific standards but in such a way that the assessment and conservation value of the site is clear. It is a public statement of the value given to our geological and geomorphological heritage by the earth-science community which has participated in its production, and it will be used by the Joint Nature Conservation Committee, English Nature, the Countryside Council for Wales and Scottish Natural Heritage in carrying out their conservation functions. The three country agencies are also active in helping to establish sites of local and regional importance. Regionally Important Geological/Geomorphological Sites (RIGS) augment the SSSI coverage, with local groups identifying and conserving sites which have educational, historical, research or aesthetic value, enhancing the wider earth science conservation perspective.

All the sites in this volume have been proposed for notification as SSSIs; the final decision to notify, or renotify, lies with the governing Councils of the appropriate country conservation agency.

Information about the GCR publication programme may be obtained from:

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Front cover: The cover illustration shows Gordale Scar in the Yorkshire Dales karst. A deep gorge was cut through horizontal Carboniferous limestone by waterfall retreat in a powerful meltwater stream from a Pleistocene glacier; it now carries an underfit stream, because most water in the reduced catchment sinks underground. (Photo: A.C. Waltham.)

Karst and Caves of Great Britain

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Work on the GCR cave sites started in the 1970s with the initial identification of sites in a widespread consultation exercise co-ordinated by David Judson and George Black. By 1982 the first version of the cave site descriptions had been completed by Tony Waltham, who also then identified and produced the first descriptions of the karst sites. Subsequently all the site descriptions were rewritten into the designated format for this publication series. Mike Simms rewrote the cave site descriptions, except for that of the Slaughter Stream Cave which was later added by Dave Lowe. Andy Farrant rewrote the main karst site descriptions. Helen Goldie produced the descriptions of the limestone pavement sites in the northern Pennines, with assistance from Simon Webb who also wrote the section on Helbeck Scars. Some site descriptions and all the sections of introductory text were written by Tony Waltham, who also edited and unified the material from his co-authors, and produced the figures.

The first stage of data collection and compilation of the maps of the cave sites involved the extensive co-operation of many members of the British Cave Research Association. Without their assistance and the great efforts of many, mainly sporting cavers who mapped the caves in the first instance, this volume could never have been produced. Consultations on the karst and cave sites extended to a broad spectrum of those studying limestone karst throughout Britain. These included academic scientists pursuing geomorphological research, and also active cavers with unparalleled knowledge of the underground sites. Many made valuable contributions with information, advice, comment and assistance; Pete Smart was particularly helpful with comment and guidance during preparation of the manuscript.

Within this volume, the descriptions and interpretations of individual sites lean heavily on the observations and research of many individuals. Published source material is all referenced, the authors of the volume have contributed their own personal knowledge of many of the sites, and numerous extra notes and concepts have been incorporated from unpublished thoughts and discussions. This text is a synthesis of understanding where the credits reach far wider than the names on the title page. The various topographical and geological maps, which constitute many of the figures in the volume, have been compiled from numerous sources, and have inevitably extracted bits of information from the many high-quality maps produced for this country by the British Geological Survey and the Ordnance Survey. The cave surveys which form many other figures are credited to the main caving clubs who produced them, but thanks are due to many others who contributed above or below ground and yet remain anonymous.

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Access to the countryside

This volume is not intended for use as a field guide. The description or mention of any site should not be taken as an indication that access to a site is open or that a right of way exists. Most sites described are in private ownership, and their inclusion herein is solely for the purpose of justifying their conservation. Their description or appearance on a map in this work should in no way be construed as an invitation to visit. Prior consent for visits should always be obtained from the landowner and/or occupier.

Information on conservation matters, including site ownership, relating to Sites of Special Scientific Interest (SSSIs) or National Nature Reserves (NNRs) in particular counties or districts may be obtained from the relevant country conservation agency headquarters listed below:

English Nature,
Northminster House,
Peterborough PE1 1UA.

Scottish Natural Heritage,
12 Hope Terrace,
Edinburgh EH9 2AS.

Countryside Council for Wales,
Plas Penrhos,
Ffordd Penrhos,
Bangor,
Gwynedd LL57 2LQ.

Preface

This volume summarizes the results of surveys of Britain's karst regions, undertaken between 1978 and 1990 as part of the Geological Conservation Review (GCR). The GCR was the first attempt to assess the scientific significance of Britain's geological conservation strategy.

The surveys of the karst geomorphology were carried out in two parts. The first was of the cave sites and the second was of the surface landforms, and they were amalgamated at a later date. In each part, the first stage was to produce a provisional list of potentially significant sites, and this was circulated to all relevant specialists in the country. All the sites were visited, or were already familiar to the authors. New features revealed by new explorations underground were discussed with the local cavers making the discoveries, and many of the new caves were visited to assess their significance. The comments made by the specialists and the field observations were used to produce a modified site list, and this was then slightly adjusted during preparation of the site descriptions. The GCR sites finally listed are therefore those of national scientific importance, and they include some of recognized international importance.

The list of GCR sites has been used to establish a new set of Sites of Special Scientific Interest (SSSIs). Where there is no other significant interest at or adjacent to the site, a proposal was made to establish an SSSI on the karst geomorphological interest alone. Many sites contain other significant features, or adjoin another site of non-karstic significance; a composite SSSI has then been constructed from a set of GCR sites. Despite the heterogeneous nature of such sites, it is important to remember that the geomorphological interest is sufficient on its own to justify the site conservation. Most of the SSSI proposals that have arisen from this survey have already been translated into site designations by the appropriate country conservation agencies (English Nature, Countryside Council for Wales, Scottish Natural Heritage).

This volume is not a field guide to karst sites, nor does it cover the practical problems of their future conservation. Its remit is to put on record the scientific justification for conserving the sites, discussing the interest of the surface and underground landforms within them, and placing them in a wider geomorphological context. Each site is documented in a self-contained account, starting with the highlights (a précis of its special scientific interest) and a general introduction (with a note of investigation and research literature concerning the site). A morphological description of the various features of the site, also places them in the context of an assemblage of interrelated landforms. An interpretation of the site geomorpholgy and its significance then follows. The depth of the interpretation varies considerably across the sites, as some have been thoroughly researched over many years, and others are barely investigated in detail. Some of the cave sites are excellent examples of their morphology and are scientifically valuable for the erosional and sedimentary records that they contain, but their difficulties of access and the hostile working environment underground have precluded detailed study to date. The interpretative sections have involved some use of technical language, but each account ends with a brief summary of the interest framed in less technical language, in order to help the non-specialist.

This volume does not provide a fixed list of the important karst and cave sites in Britain. Geomorphological science continues to progress, and increased or hitherto unrecognized significance may be seen in new sites. The limestone caves provide a special case where new explorations reveal cave passages that were totally unknown previously. During the progress of this survey and documentation, three new sites have been added to the GCR list, and it is inevitable that further sites worthy of conservation will be discovered in future years. There is also the problem of potential site loss, and part of one cave site has already been removed by an expanding quarry.

This volume deals with our knowledge of the sites available at the time of writing, in 1995, and must be seen in this context. The data within these pages clearly demonstrate the value of British karst sites, and their important place in Britain's scientific and natural heritage.

Chapter 1

Introduction

KARST AND CAVES

Karst may be defined as a distinctive terrain created by erosion of a soluble rock where the topography and landforms are a consequence of efficient underground drainage. Its characteristic features therefore include disrupted surface drainage, closed depressions, dry valleys and caves. The essential underground drainage means that caves are an integral component of a karst landscape; however, caves are commonly defined as natural cavities large enough to be entered by humans, and some karst landscapes on the softer rocks are drained by fissures too narrow to be described as true caves. Limestone is the only common rock that is highly soluble in natural surface waters, so nearly all karst is formed on limestone. Dolomite may have karstic landforms, generally less well developed than those on limestone. Some karst features are formed by solution of gypsum or salt, but pseudokarsts on basalt or ice are not due to rock solution.

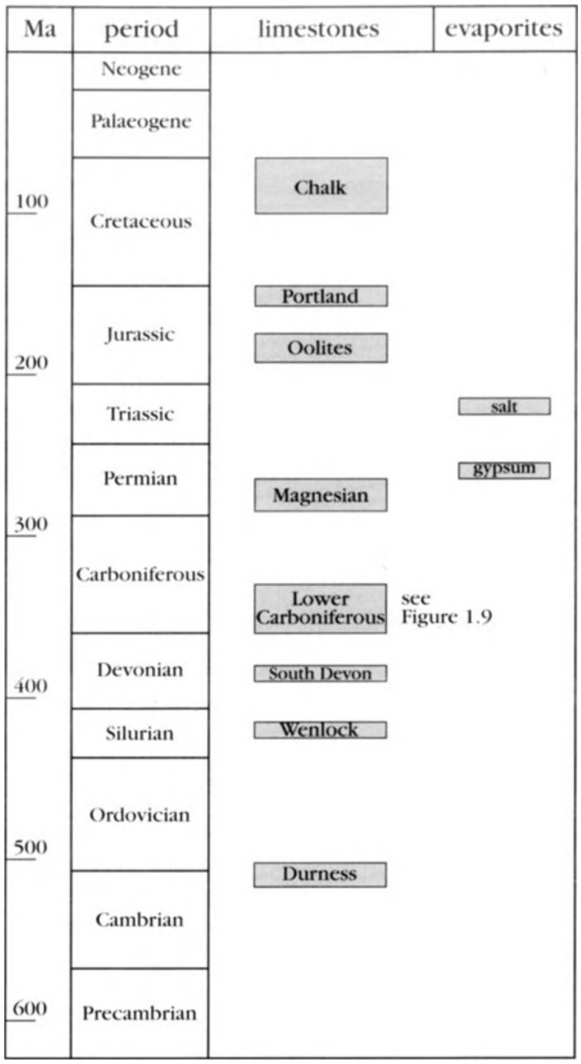


Figure 1.1 The main limestones and evaporites which have karstic features within Great Britain.

Most cavernous limestones have an unconfined compressive strength of around 100 MPa. They are strong rocks, capable of spanning large underground voids and forming stable cliffs; they are also massive, with widely spaced fractures, some of which are enlarged by solution to form discrete conduits. Chalk is the best known of the weaker, porous limestones, which form a type of karst with few caves large enough to be entered by humans. Because limestone contains little insoluble residue, soil generation is limited, but many karst areas have some cover of mineral soil derived from adjacent non-carbonate outcrops, organic debris, or residual soil accumulated over a very long time; bare rock outcrops are common features of karst.

Karst lands tend to provide some of the more spectacular natural landscapes, with hills and mountains of white crags and bare rock pavements, pitted by sinkholes and caves. Though limestones are soft (in that they are easily abraded), many are mechanically strong due to their microstructure of interlocking crystals, and all are topographically resistant because much of their erosion is underground; even the weakest limestones, such as chalk, survive as the high ground – 'because they devour their own agents of erosion'. For the same reason, limestone karst contains many deep and spectacular gorges; formed in climatic environments of the past, they are preserved because their walls are not eroded by water which has sunk underground.

The preservation of landforms within a karst landscape is most significant beneath the ground surface. A complex cave system is the only erosional environment where each phase of erosion does not remove the features of earlier phases. Capture and diversion of drainage, rejuvenations and steady downcutting create new cave conduits at lower levels, and preserve the products of earlier erosion and deposition that have been abandoned in high-level cave passages – on a scale which can never be achieved in an evolving, eroding and lowering surface topography. Caves are therefore especially significant to geomorphological studies, as their erosional features and accumulated sediments are unique records of past environments in upland regions.

The karst of Great Britain is formed on a number of limestones spaced through the stratigraphical column, and also on some units of gypsum and salt (Figure 1.1). The outcrops of these rocks therefore define the areas of karst landscape, which are widely distributed across England and Wales, but are rather sparse in Scotland (Figure 1.2). They include all the well known karst sites much visited by students and researchers of geomorphology: the cliffs at Malham, the gorge at Cheddar, the caves of Dan-yr-Ogof, the pavements and potholes of Ingleborough, and the dry valleys of the Peak District and Chalk Downs. Most of these sites are on or in the massive limestone of Lower

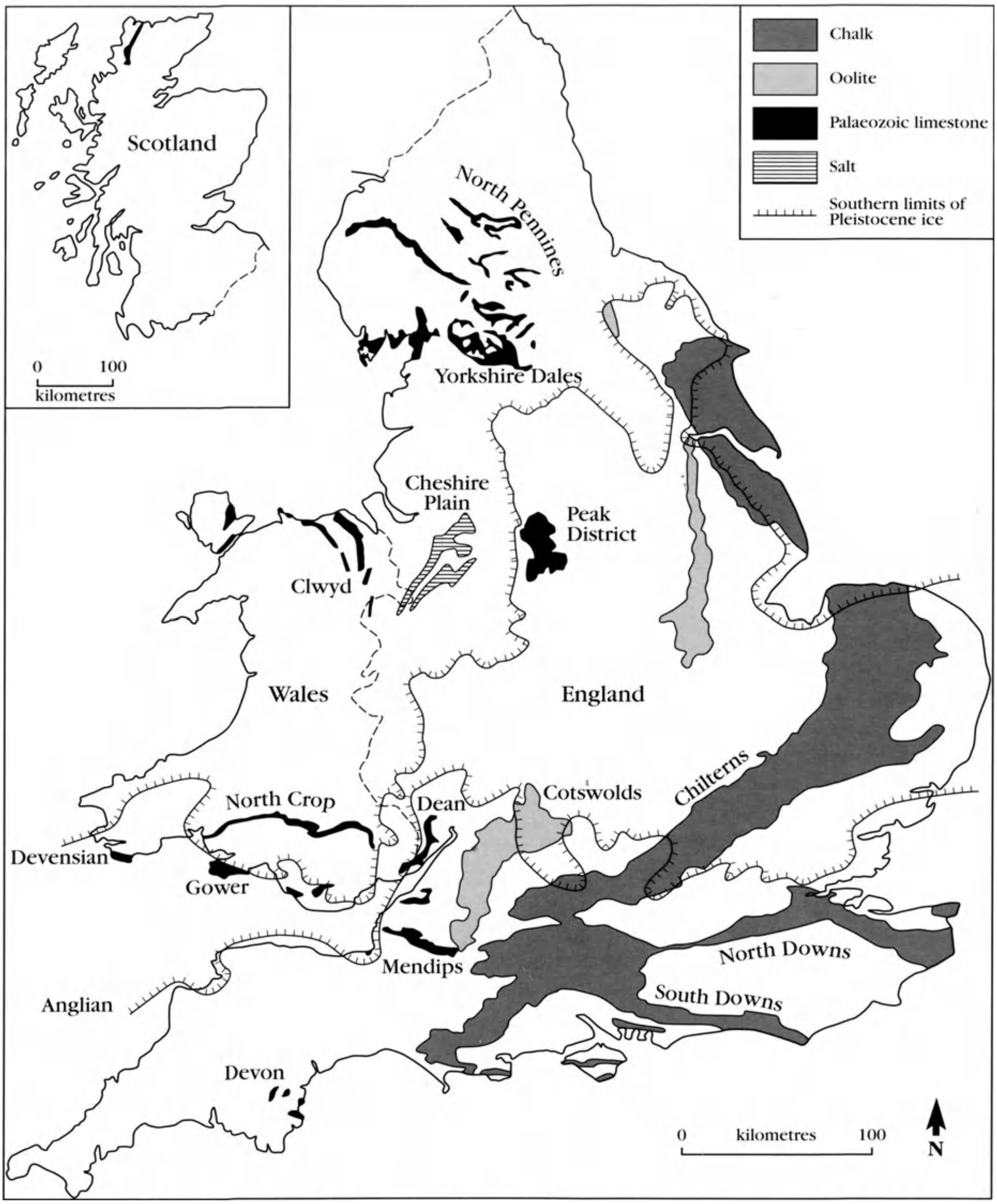


Figure 1.2 Outline map of the main areas of karst in Great Britain. The Palaeozoic limestones are of Lower Carboniferous age, except for the Devonian limestone in Devon, and the Cambrian-Ordovician limestone in Scotland.

Carboniferous; this is by far the most important karstic rock in Britain, though the largest area of karst is on the Cretaceous Chalk, which has very limited cave development. The sites included in this volume of the Geological Conservation Review cover a wide sample of the surface and underground landforms in these areas of Britain's karst.

SOLUTIONAL PROCESSES

Limestone consists largely of the mineral calcite, which is composed of calcium carbonate, which is only slightly soluble in pure water. Limestones are however much more soluble in acids, and the most important process in the overall development of surface karst landforms and caves is solution by carbonic acid, that is produced by the introduction of carbon dioxide. The process of dissolving the rock, to create the liquid solution of calcium and bicarbonate ions in water, may be referred to as either solution or dissolution; the term 'solution' is in common use, and is retained in this text. The many factors that influence the complex chemistry of limestone solution are reviewed by Ford and Williams (1989) and White (1988).

The solutional capability of water with respect to limestone is related directly to its carbon dioxide content. All rainwater absorbs small amounts of carbon dioxide from the atmosphere, but soil water contains biogenic carbon dioxide at far higher concentrations due to its production by plant roots. Biological activity increases greatly with temperature, and the development of karst landforms is therefore maximized in hot, wet, tropical environments where there is a dense cover of plants and organic soils. Although carbon dioxide is slightly more soluble in cold water than in hot water, karstic processes in cold climates are severely restricted by the reduced biogenic production of the gas.

Rates of limestone reaction and removal are a function of molecular diffusion that is determined by the kinetics of the calcite solution process (Dreybrodt, 1981a, 1988). Solutional erosion is distributed through a karst partly in relation to the source of biogenic carbon dioxide in the soil cover. The maximum increase in solutional load occurs at the soil/limestone interface within the zone of epikarst (Williams, 1983, 1985), and thus contributes directly to surface lowering. Caves receive flows of saturated percolation water, and of allogenic, sinking (or swallet) stream water which are also low in carbon dioxide and solutional capability. The passages are enlarged by waters of minimal aggressiveness, because the flows are concentrated in fissures and conduits with wall areas far smaller than the area of soil-covered rockhead which receives only diffuse flows. Water in the phreatic, or flooded, zone is normally saturated, in equilibrium with the limestone walls of the conduits; it is capable of further solution due to the process of mixing corrosion. Two waters, each saturated to different levels of carbon dioxide and calcite, form an unsaturated, aggressive water when they combine, due to the non-linear equilibrium of carbon dioxide and calcite in solution (Bögli, 1964, 1971, 1980; Dreybrodt, 1981b).

Percolation water descending through a limestone is normally saturated with calcium carbonate in equilibrium with its high content of soil carbon dioxide; subsequent diffusion of the gas into a cave air causes precipitation of some of its calcium carbonate load, to form stalactites and other cave speleothems. Comparable loss of carbon dioxide from a surface stream, normally associated with algal growth, causes the precipitation of calcite as tufa or travertine (Ford, 1989b).

Limestone solution can also occur in the presence of acids other than the carbonic acid generated by carbon dioxide. Sulphuric acid occurs naturally, by oxidation of pyrite, and is extremely corrosive. Its effects in karst are subordinate to those of carbonic acid, except in the earliest stages of cave inception (Lowe, 1992b; Worthington and Ford, 1995); sulphuric acid generated by oxidation of sulphide minerals in the rock is probably important in opening the initial voids in a limestone mass, so allowing the input of increased flows of surface water charged with carbon dioxide. Organic acids from soils play a solutional role far subordinate to that of the carbonic acid. Solution of gypsum and salt is not dependent on acids, as both rocks are highly soluble in pure water.

KARST GEOMORPHOLOGY

The geomorphology of karst is widely described and reviewed, and three of the major texts in the English language are by authors who originated from, and worked extensively in, Britain (Sweeting, 1972; Jennings, 1985; Ford and Williams, 1989). These volumes are the best guides to the very substantial literature through which karst geomorphology and cave science have evolved to their present levels of understanding.

Both the broadest structure and the topographic texture of a karst terrain are determined by the lithology, strength, porosity and structure of the exposed carbonate succession. In contrast, the main landforms, the karst types and many of the smaller solutional features are all functions of process, and can be identified in karst regions on all types of limestone, as well as on many other types of soluble rock.

Karst landforms

Dolines

A closed depression in the land surface, with no drainage outlet except underground, may be regarded as the diagnostic landform of karst topography. They are generally known as dolines, after the Slovene term for a surface depression in the classical karst landscape where there are no continuous valleys; they are also known as sinkholes, especially in the engineering and American literature. Dolines may form by a variety of mechanisms, and most are the product of multiple processes. There are no size constraints on dolines; most of those in Britain are 1-50 m in diameter, with width:depth ratios varying from 2:1 to 4:1.

Solution dolines are formed by localized surface lowering through chemical erosion of the limestone (or other karstic rock). Solution is dominantly at the subsoil rock surface, but may be subaerial, by rainwater converging on a bedrock fissure, which acts as a drainage outlet. The doline therefore slowly deepens over time, and the evolution of its cross profile is analogous to that of a non-karstic valley; it contrasts with the normal valley only in its closed long profile. The internal slope gradients are largely a function of the rock strength and fracture patterns, and of any degradation which has taken place after solutional processes have diminished. There is a spectrum of doline profiles, from the broadest of saucer-shaped depressions, through to the potholes and shafts which are the entrances to many underground drainage conduits.

Collapse dolines form by failure of the limestone into underlying caves; they are commonly identified by their internal rock walls and scars, which are the remnants of the failed rock spans. Dolines formed purely by collapse are extremely rare, but nearly all large dolines contain some elements of collapse processes. On a small scale, subsoil solution within a solution doline inevitably involves some bedrock fissure opening; this leaves residual blocks of unsupported rock, which ultimately fail and fall within the soil profile. On a larger scale, rock walls may collapse into an expanding doline; progressive failure of limestone ribs left between wide fissures may create the large quarry-like dolines, such as Hull Pot in the Yorkshire Dales.

Subsidence dolines form by the failure, sagging or collapse of an insoluble soil or cover rock into solutional voids in a buried limestone. The most common are those formed in unconsolidated soil, as infiltrating rainwater washes the soil into fissures within the limestone. Fines within the soil are washed away first, from beneath, followed by removal of the coarser particles, and then subsidence of the upper soil and surface. This process is known as suffosion, which is a type of piping failure of the soil, and the surface depressions may therefore be known as suffosion dolines. Many thousands of this type of doline have formed in the glacial till on the limestone of the northern Pennines, where they are known locally as shakeholes (Figure 1.3). Most are 1-15 m across, and are

Figure 1.3 A small subsidence doline, or shakehole, recently formed where the glacial till and soil cover have been washed into a fissure in the underlying limestone; on Ingleborough, in the Yorkshire Dales karst. (Photo: A.C.Waltham.)

no deeper than the till cover; many have slumped so that the limestone is not exposed, but others contain open fissures and cave entrances. The formation of subsidence dolines by the failure of a strong cover rock into limestone voids is a feature of interstratal karst. The hundreds of large dolines in the Namurian sandstone outcrops of South Wales are of this type, and the mechanism of formation is similar to that of collapse dolines, wholly in limestone, except that there is no solutional weakening of the rock span over the void before collapse.

Poljes are larger forms of karstic closed depressions, with sharply defined rock slopes around the perimeter of wide flat floors, which are commonly alluviated (Gams, 1978). They form by lateral planation on the sediment floor or at the water table, and both their inflows and outflows are underground. There are no true poljes in Britain, though the karstic depressions at Hale Moss and the turlough of Pant-y-llyn have many features similar to those of a polje.

Continued enlargement of closely spaced dolines produces a landscape of increasingly disordered relief, where the residual hills ultimately become the dominant landforms. These positive features of a karst landscape include the cones, towers and similar hills in the mature tropical karst terrains, but none of these landforms occurs in Britain's limestone, due to the climatic constraints on the karstic evolution through the Quaternary. The conical hills in the Peak District, once thought to be remnants of Tertiary tropical karst, are exhumed reef knolls.

Karst valleys

The loss of surface drainage into sinkholes precludes the development of most surface valleys in a fully mature karst where dolines become the dominant landform. There are, however, many situations where valleys are or have been formed on limestone.

Blind valleys carry a surface stream but terminate where the water sinks underground. Most form where valleys extend from a non-karstic outcrop and end where the streams find fissure routes into caves beneath their limestone floors. Ingleborough's Fell Beck flows in a blind valley cut into the shale and drift as far as the sink into Gaping Gill.

Headless or pocket valleys form in complementary style where a substantial stream or river emerges from a limestone aquifer and cuts a valley downstream of its resurgence, as at Wookey Hole.

Allogenic or through valleys form where a surface river enters a karst with a flow too large to sink underground in the available fissures. They develop into larger features, and commonly into karst gorges, where the rate of entrenchment exceeds the local rate of fissure enlargement in the maturing karst; this precludes subsequent underground capture, especially where the hydraulic gradient in the limestone is low beneath a gently graded valley. The River Wye flows in an allogenic valley through the Peak District, where the opportunity for underground capture is also reduced by impermeable units within the limestone sequence. The Rivers Tawe and Wharfe are two of those which cross their limestone outcrops in inherited glacial valleys with very low gradients. Valleys with ephemeral streams, active only in flood conditions, are common both on cavernous limestones and on the weakly cavernous chalk.

Dry valleys are features of many karst terrains. They are fluvial landforms, cut by subaerial streams and rivers, and then abandoned and left dry when their flows were lost to underground captures (Smith, 1975b). Some were formed by surface flows on limestone when it was first exposed and was only minimally permeable, before secondary permeability was increased by solutional fissure enlargement. Many dry valleys in Britain were excavated or enlarged under periglacial conditions when ground ice of the permafrost sealed the limestone fissures during cold stages of the Pleistocene. Also in the cold conditions, surface valleys were deepened by annual snowmelt floods which exceeded the capacity of unfrozen sinks and choked them with sediment.

Karst gorges were formed by subaerial, fluvial entrenchment in limestones strong enough to stand in stable steep faces. Most were formed where river incision was accelerated in descents off upland blocks. Gordale Scar is one of many that was cut rapidly by powerful flows of proglacial or subglacial meltwater; Cheddar Gorge is one of those cut by seasonal flows of snow melt during the cold stages of the Pleistocene. The gorge at Matlock Bath is one of those cut by a large allogenic river; it is essentially superimposed and was never the site of a major river descent. Common to them all is their preservation in the limestone, even after the first two types were abandoned by the streams and rivers which cut them. At Gordale and Cheddar, surface erosion and degradation of the gorge walls were almost eliminated by underground capture of the surface drainage; this occurred when the permafrost melted as the climates ameliorated at the end of the Pleistocene cold stages. Surface water flows on the sides of the Matlock gorge have always been insignificant in comparison to the flow of the through river. Gorge formation by cavern collapse is extremely rare; in Britain it is limited to short sections of gorge immediately outside large cave entrances or exits whose roof arches are retreating, as at Wookey Hole and Porth-yr-Ogof. Small-scale collapse and unroofing of small cave passages are normal features which contribute to valley deepening in a limestone.

Limestone pavements and karren

Subaerial and subsoil limestone surfaces are etched by solution into a variety of small features. Dominant are the solution runnels, which are better known by the German term, karren. These are most conspicuous on the bare limestone pavements which were scraped clean by the Pleistocene glaciers in the northern Pennines (Parry, 1960; Sweeting, 1966) (see Chapters 2 and 3). Postglacial solution by rainwater enlarging the bedrock fissures has left a pavement of *in situ* limestone blocks, each locally known as a clint and separated from its neighbours by grikes. The clints are fretted by solution runnels, and some of the grikes are partly relicts of pre-Devensian erosion. Some of the modern pavement features may be inherited from Carboniferous palaeokarsts developed on intraformational calcretes (Vincent, 1995). The largest clints and the most extensive pavements lie where bedding planes have been scoured on top of strong beds of limestone. Where the surface steps across a sequence of beds and bedding planes, the terrain is known as staircase karst (from the German *Schichttreppenkarst*).

An exhaustive classification of karren forms uses German terms (Bögli, 1960; Sweeting, 1972); there are no equivalent terms in the English language. Only some of the types are widespread in Britain's karst, and rillenkarren and rundkarren are the most significant.

Rillenkarren have sharp crests between channels which are normally 10–20 mm wide and deep, and are aligned down gentle or steep rock faces, usually with some degree of channel convergence. They are the normal solutional features on bare limestone which is exposed to direct rainfall or snowfall, but good examples are rare in Britain.

Rundkarren have rounded crests and troughs and are generally 100–400 mm wide and deep, cut into almost level or sloping limestone surfaces. They are normally formed under a cover of dense vegetation or organic soil which retains rainwater over the ridges between the solution runnels, so that surface solution creates a rounded profile. Rundkarren are dominant on the bare pavement surfaces of the northern Pennines, suggesting that much of their development took place beneath a permeable, organic soil cover. Most of the original plant cover has been lost due to artificial clearance of the protecting trees, and sheep grazing has precluded regrowth of anything except grass. There has been almost no subsequent development of rillenkarren; this appears to be due to the very complete coverage of the limestone surfaces by lichen, which provides the cover beneath which rounded rundkarren evolve.

Rinnenkarren have rounded channels similar in size to rundkarren, but form with sharp rims cut into bare, sloping surfaces. They deepen downstream, and Britain's classic examples are those on the Rakes of Hutton Roof.

Kamenitzas are solution basins, or pans, generally 50–800 mm across, cut into level rock surfaces; they are rounded or elongate and commonly have an overflow channel from them. Once established in random hollows, they are self-deepening due to solution by the regularly recharged rainwater, commonly aided by organic acids produced by plants and peat trapped in the basins.

Trittkarren have stepped profiles on bare surfaces with slopes too gentle to support rillenkarren. Each step is typically 10–50 mm high, between wider flat treads; they are rare in Britain.

Kluftkarren are the deep, open fissures formed by solutional enlargement of primary tectonic fractures within the limestone; in northern England they are commonly known as grikes, and they separate the clints of remaining limestone. Grikes are typically 50–500 mm wide and can reach depths of many metres. Clint sizes and shapes, and grike spacing, are features of the limestone structure.

Spitzkarren are pinnacles or blades of limestone, with sharp or rounded crests. They are residuals of limestone left by deep subsoil or subaerial solution down closely spaced kluftkarren, and include some of the narrow or knife-edge clints in the more densely jointed Pennine pavements.

Karst types

Different assemblages of limestone landforms create identifiable karst types which are largely related to the present and past climatic environments in

Figure 1.4 The Yorkshire Dales glaciokarst, with bare cliffs and scars, limestone pavement and a fossil meltwater channel at Comb Scar, above Malham. (Photo: A.C. Waltham.)

which they have evolved. Within each type, the geological structure of the host soluble rock determines the patterns of underground drainage and also influences the surface topography. The contrasting karst types in the different regions of Britain are functions of both Pleistocene history and local geology.

Glaciokarst is characterized by the inheritance of glacial landforms, and is distinguished by the bare rock surfaces scoured by Pleistocene (or more recent) glaciers. Limestone pavements and scars form on the tops and edges of the outcrops of stronger beds; they are fretted by postglacial karren, and there is minimal development of postglacial soil cover. Deep karst gorges were formed by temporary meltwater rivers, but generally there are few dry valleys. The Yorkshire Dales contain Britain's finest glaciokarst (Figure 1.4).

Fluviokarst is characterized by dendritic systems of dry valleys. The finest area in Britain is the Peak District (Figure 1.5), where the valleys were largely excavated under periglacial conditions during the Pleistocene. Karst gorges are developed where the valleys entrenched into steeper slopes, but there are few rock scars; most outcrops of the limestone are covered by soils of solutional residue and aeolian loessic silt.

Polygonal karst is a more mature karstic terrain where dolines have replaced valleys as the main form, and a polygonal network of topographical divides has replaced the dendritic systems of interfluves. This type of karst is poorly represented in Britain, where the Pleistocene climatic fluctuations and glaciations repeatedly interrupted solutional erosion; it is better developed in Mediterranean climatic regimes (Gams, 1969, 1974).

Tropical karsts are the climatic extreme, where the negative landforms of valleys and dolines are replaced by the positive, residual landforms of cones and towers. There is no trace of these karst types in Britain's modern landscape; the classic areas of cone karst include those in Java (Lehmann, 1936) and Jamaica (Sweeting, 1958), and tower

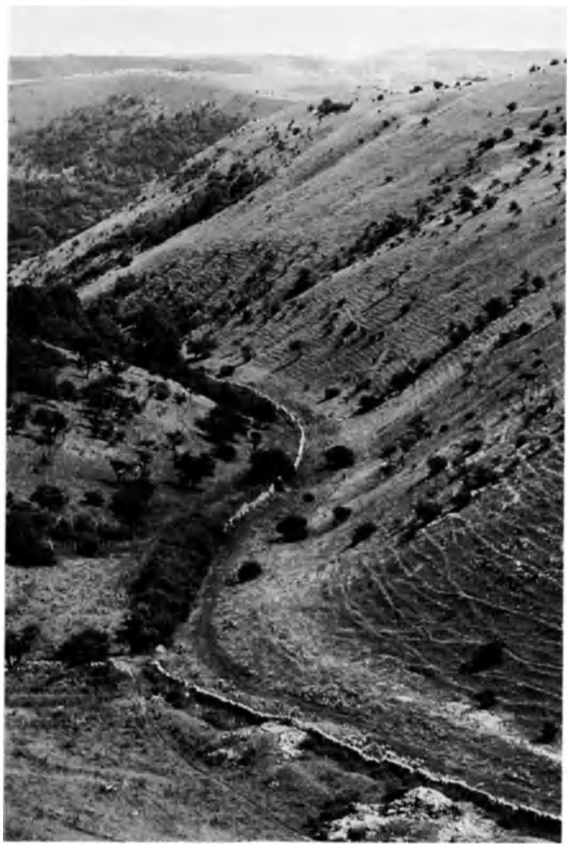


Figure 1.5 The dry valley of Deep Dale, in the Peak District fluviokarst. (Photo: A.C. Waltham.)

karst is largely restricted to the limestones of southern China (Zhang, 1980; Smart *et al.*, 1986).

Fossil karst or palaeokarst has its solutional landforms buried by later sediments of either clastic or carbonate composition. It includes features as old as the intra-depositional structures within the Carboniferous limestone sequence (Ford, 1984), and the many fissures filled with Triassic sediments (Simms, 1990). It also includes the many buried and filled dolines containing Tertiary and Quaternary sediments in the Carboniferous limestones of the Peak District and in the Chalk of south-east England; the latter include the many buried and steep-sided features commonly known as pipes.

Chalk karst is a very distinctive style of topography, developed on the mechanically weak, porous, very permeable and only mildly cavernous chalk; it extends across large outcrops in south-east England (see Chapter 7). It has extensive dry valley systems, which were enlarged under periglacial conditions, and numerous subsidence dolines formed in weak cover rocks. Soil cover is complete, and there are no scars or crags in the weak rock. Underground drainage is efficient, but there are few caves large enough to be accessible (Lowe, 1992a).

Salt karst and gypsum karst are formed on the respective evaporite rocks, both of which are extremely soluble in water. Surface landforms are dominated by solution dolines and broad depressions too shallow to be described as true dolines; these occur on both rock types in Britain (see Chapter 7). Gypsum karst may also have large cave systems, but there are only a few small caves in Britain's gypsum.

EVOLUTION OF CAVES

Cave passages form through a limestone karst where there is an available flow of water, with chemical potential to dissolve the limestone, with an adequate hydraulic gradient between a sink and a rising, in a favourable geological structure. Extensive cave development therefore depends on a combination of geological and topographical factors, and on a climate which provides meteoric water charged with biogenic carbon dioxide.

Any single cave passage evolves through three distinct stages. Initiation creates the openings through the rock, which permit the flow of groundwater and allow the accelerated erosion of the next stage. Enlargement is the main stage of cave development, when the small, initial fissures are enlarged to reach and pass the size limit of accessibility by humans, that defines a cave. Degradation is the terminal phase of destruction, where the cave either collapses, is filled with sediment or is removed by surface lowering. In a complex cave system, all three processes take place simultaneously in passages at different depths and positions in the limestone; solutional enlargement and sediment infilling can take place at the same time in a single passage.

The time-scales of cave evolution are long. The enlargement and degradation stages can be observed and monitored. An order of magnitude for their combined completion in Britain's karst is about a million years. A cave passage can develop to a diameter of a metre inside 10 000 years (Mylroie and Carew, 1987). Caves can evolve over many millions of years in karsts of greater depth, thicker limestone and slower surface lowering (without interruption by glaciations) than are found in Britain. In all situations, the initial stage of cave evolution may take vastly longer; it is nearly impossible to observe and assess, but it should be viewed on a time-scale of tens of millions of years.

Cave passage dimensions vary greatly. The smallest are those which can just be entered by a human, but there is a continuity from these down to the solutionally enlarged fissures and protocaves which are abundant in karst aquifers. The largest caves are about 100 m in diameter, but these are restricted to warmer climatic zones with longer records of rapid limestone solution. In Britain, a cave passage 5 m high and wide is described as large, and there are few which exceed 15 m in diameter. Cave systems may have complex passage networks which reach great lengths. Any figures quoted are for the mapped lengths only, and most caves extend beyond the flooded, choked or constricted passage sections which are the contemporary limits of exploration. Currently there are in the world 28 cave systems which have been explored for longer than 50 km, and two of these are in Britain (Ease Gill Cave System and Ogof Ffynnon Ddu). Cave depths are limited by the heights of the limestone mountains which contain them, and Britain cannot have very deep caves. Throughout the world, there are 50 caves known with depths exceeding 1000 m, but Ogof Ffynnon Ddu is the deepest in Britain and reaches only 308 m.

Cave initiation

The longest stage in cave evolution involves the creation of the initial opening through a solid rock mass. Only after completion of a route through the limestone can groundwater flow and solutional erosion progress. Some initial openings are provided by the primary porosity of the rock; highly porous limestones such as chalk have more diffuse groundwater flow, and hence less tendency to direct solutional effort into conduit, or cave, enlargement. Tectonic fractures and bedding planes constitute the main initial openings in the less porous limestones. These are present to some extent in all limestones in older terrains and structural blocks, including those of Britain. They may be opened by unloading in response to erosional reduction of the cover; alternatively, they may be tight in zones of high tectonic compression, plastic deformation and metamorphism, accounting for minimal cave development in some limestones of the younger mountain chains outside Britain.

At depth in a limestone mass, solutional enlargement of the initial openings is independent of meteoric water which has no access to them. Available connate waters and mineral acids are probably dominated by sulphuric acid that is generated by oxidation of sulphides (Ball and Jones, 1990; Hill, 1987). The limestone solution process is influenced and directed by the smallest of chemical contrasts within the rock sequence. An inception horizon is a locus of cave initiation (Lowe, 1992b); it may be a shale parting within the limestone, perhaps containing pyrite as a source of sulphuric acid, or it may be no more than a bedding plane between limestones of slightly contrasting lithologies. Faults and joints may provide links between inception horizons and are themselves enlarged by solution, but most cave passages are initiated on the bedding or at bedding/fracture intersections.

The geological control of cave inception is absolute, but hydraulic factors start to influence the cave evolution after the initial openings are established. Flow is laminar until the fissure is wide enough to permit turbulent flow; the breakthrough dimension is about 5 mm (White and Longyear, 1962; Atkinson, 1968a; White, 1988). Turbulent flow in wider fissures permits far higher rates of solution and erosion, which are enhanced by throughflows of meteoric water, charged with biogenic carbonic acid (Thrailkill, 1968). The cave then enters its second stage of evolution.

Cave enlargement

As the limestone fissures are enlarged, the permeability of the rock mass increases, and the hydraulic gradients decrease. The upper zone of the aquifer drains and some form of water table is established. Above it, the fissures and caves are drained so that they contain free air surfaces within the vadose zone. Below it, all openings remain full of water within the phreatic zone, also known as the saturated zone or phreas. The water table does not have a uniform slope, as in a diffuse aquifer, but is a complex stepped surface partly related to, and changing with, the pattern of cave conduits (Drew, 1966). The onset of turbulent flow permits increased erosion by mechanical abrasion in cave streams that carry surface sediment. Hydraulic advantages also become apparent, and many protocaves are abandoned while fewer favoured routes continue to be enlarged. All caves are initiated under phreatic conditions, but are enlarged in styles which contrast above and below the water table.

Vadose cave passages, enlarged above the water table, are dominated by canyons or trenches. They are either cut downwards by freely flowing streams or headwards by waterfall retreat, and they maintain downstream gradients. They may meander due to erosional exaggeration of bends, or may be straight rift passages where guided by rock fractures; canyons at different levels are connected by waterfall shafts enlarged by spray corrosion. Undercutting of passage walls occurs along geological weaknesses or where the stream is deflected laterally by sediment accumulation that prevents downcutting.

Phreatic cave passages that are full of water are enlarged by solution of their floor, walls and roof, and can therefore have very complex shapes. The dominant form of a conduit is a rounded tube, but this may have an elliptical cross-section extended along a bedding plane, or may enlarge into a high rift on a fracture plane. Roof hollows and cross-rifts are common, and some are the product of mixing corrosion. Phreatic caves have looping or irregular long profiles with phreatic lifts and reverse gradients, as their flow is maintained by hydrostatic pressure. They are commonly formed in a shallow phreatic situation, following the shortest available routes just below the water table, as in the Kingsdale caves, but many are guided by inclined geological structures into deep phreatic loops, as in Wookey Hole.

Many phreatic cave passages are left in the vadose zone when the local or regional water table falls to a lower level. A passage may be abandoned and left completely dry. Alternatively, any stream still in it cuts a vadose trench in the floor of the old phreatic tube, and creates a passage with a keyhole profile (Figure 1.6). This very distinctive cross-section is always evidence of rejuvenation, or a slow lowering of base level, in response to surface lowering between two phases in the cave's evolution.

A special type of phreatic cave develops in a confined aquifer containing slowly moving water. All potential flow routes are enlarged equally, so that a maze cave is formed on the network of available fractures (Figure 3.23). There is no focusing of flow on input or outlet points, as where conduits develop at the expense of abandoned proto-caves (Palmer, 1975). A similar effect is produced where water enters the limestone at many points from an adjacent diffuse aquifer, and maze caves may also be formed by backflooding.

Active cave entrances at stream sinks are either at the heads of gently graded cave passages, or are vertical shafts or potholes. Active resurgences may be outlets of vadose cave passages. Far more resurgence passages rise from limestone which extends below the outlet level, and are therefore flooded; deep phreatic lifts of this type are known as vauclusian risings.

Figure 1.6 The classic keyhole profile of a cave, where a drained phreatic tube has a vadose canyon cut in its floor; above the Far Streamway of White Scar Cave in the Yorkshire Dales karst. (Photo: A.C. Waltham.)

Cave systems evolve with increasing complexity as individual passages are abandoned in favour of new routes at lower levels. The progressive abandonment is partly due to the availability of new resurgences at lower sites in a topography subjected to surface lowering, but is also due to the karst system becoming more mature. Rejuvenation may affect the whole cave system or just an individual passage. Phreatic up-loops are eliminated by vadose entrenchment through their crests; down-loops are reduced by paragenesis, where the cave roof in the trough is preferentially eroded while solution of the floor is hindered by a protective cover of clastic sediment. New shorter routes are initiated in both the vadose and phreatic zones. The trunk drain through the cave system progressively approaches a graded profile, where the initial stepped water table is replaced by one which slopes gently and lies close to the level of the resurgence (Figure 1.7).

Cave degradation

Degradation and destruction of cave passages involve filling and choking by sediments, collapse, and total removal, in most cases after abandonment by their formative streams. Within a large cave system, individual passages are at all stages of initiation, enlargement and degradation at any one time. Abandoned passages are largely left in the vadose zone, and their degradation may commence while they are still being enlarged by an underfit stream cutting a floor trench.

Clastic stream sediments are carried into caves from allogenic surface sources (Ford, 1976; Ford and Williams, 1989). Some are washed through the system, but others accumulate underground. Massive influxes of sediment from glacier melt streams commonly choked cave passages at a time when solutional activity and cave enlargement were at a minimum, as happened widely in Britain's caves during the Pleistocene Ice Ages. Many cave entrances and exits are also blocked or buried by glacial till and other clastic sediments.

Calcite deposition is common in caves where saturated percolation water issues from fissures, loses carbon dioxide to the cave air and deposits

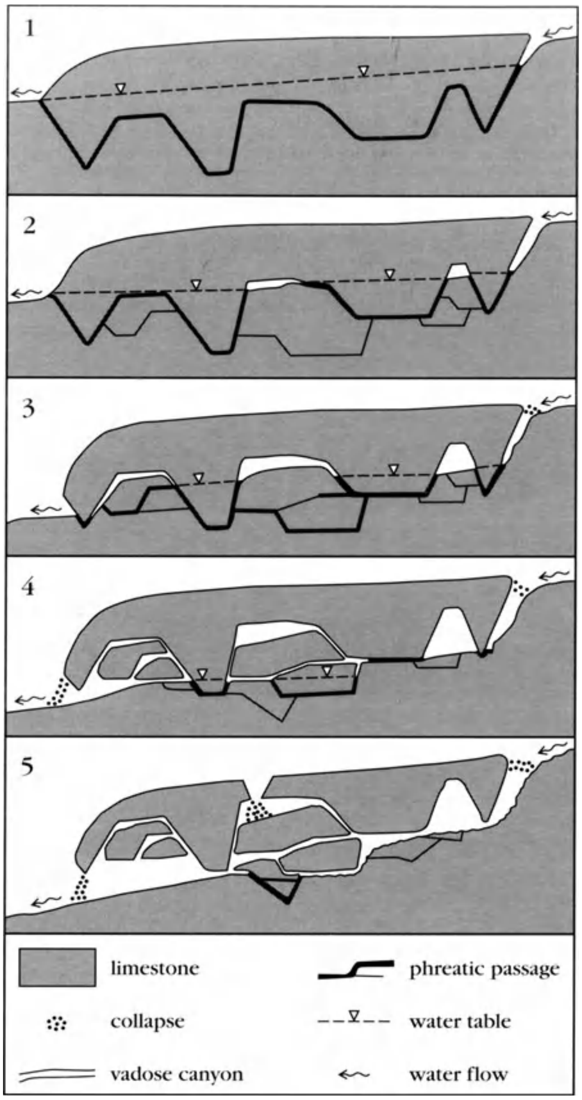


Figure 1.7 Schematic vertical sections which demonstrate five stages in the evolution of a cave system in response to time and a falling base level. The early stages are mainly of phreatic re-routing and captures; the middle stages are dominated by the entrenchment of vadose canyons through the crests of the phreatic loops; the later stages continue the deepening of the vadose canyons. The model is based on Ogof Ffynnon Ddu, which is developed close to the strike direction in dipping limestones. The principles could apply to many other cave systems if the geometry of the passages was adapted to the local geological structure. (After Smart and Christopher, 1989.)

calcium carbonate to regain equilibrium. The variety of calcite deposits, or speleothems, is immense (White, 1976; Ford and Williams, 1989). Stalactites hang from the cave roof, and many in Britain retain the thin hollow structure of the straw stalactite without external thickening; others are distorted by crystal growth patterns into the complex shapes of helictites (Figure 1.8), and others extend into curtains where dripwater flows laterally down a sloping roof. Deposition on the cave floor creates tall stalagmites or rounded bosses, whose profiles relate to saturation levels and drip rates. A flowstone floor is formed by seepage water, commonly over layers of clastic sediment, which subsequently may be removed to leave a false floor; gour dams, crystal pool linings and cave pearls are additional types of floor deposit.

Other secondary mineral deposits in caves are less abundant than calcite, and barely contribute to the blocking of a cave. Crystals, or anthodites, of aragonite grow by deposition from surface film water and condensate, and gypsum, or selenite, crystals can grow by sulphate generation within some clastic cave sediments; both these crystal forms are well developed in the Llangattwg caves.

Wall and roof collapse is a widespread feature in caves; it modifies passage profiles, and contributes to cave enlargement where fallen blocks expose new surfaces to solutional attack. Extensive collapse ultimately blocks cave passages, because the fallen material occupies larger volumes than the undisturbed rock. Cave roof failure occurs where the passage widths exceed the stable span, which

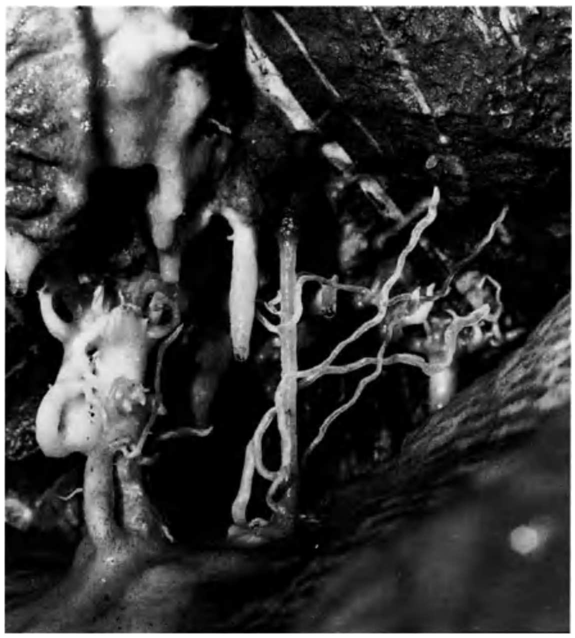


Figure 1.8 Secondary deposition of calcite in a cave passage, forming straw stalactites, small stalagmites, a sloping flowstone floor and delicate helictites; in Withyhill Cave in the Mendip Hills karst. (Photo: J.R.Wooldridge.)

is dictated by the local bed thickness and fracture density (White, 1988); the Time Machine in Daren Cilau, and other large cave passages are typified by extensive block collapse. Total failure is not common, and many cave boulder chokes are composed of inwashed debris.

A cave is totally destroyed when it is overtaken by surface lowering. More common is the fragmentation of a cave system where surface valleys cut down through it, leaving truncated passages in their walls. The extensive cave unroofing and collapse, characteristic of the final stage in karstic evolution, has not been reached in Britain, mainly because of interruptions and rejuvenations due to Pleistocene climatic changes.

RESEARCH IN LIMESTONE GEOMORPHOLOGY

Concepts and theories on the origin and development of karst landforms, above and below ground, have matured internationally, with parallel and frequently overlapping sequences of ideas evolving in different countries. The level of activity in scientific research in any one country has been dictated in part by the extent, and therefore the parochial relevance, of limestone karst within the national borders; China, the USA and the nations which once formed Yugoslavia have all been world leaders in their time. By comparisons with other countries, Britain has only small areas of cavernous karst (thereby excluding the distinctive chalk karst), but these include the internationally renowned glaciokarst of the Yorkshire Dales, and they contain some of the world's longest known cave systems. On international scales, the level of karst research in Britain has probably exceeded that which would be expected on the basis of the extent of its limestone outcrops, and has certainly included some contributions of international significance.

Development of karst research

An early, partial understanding of the landforms and processes within karst regions evolved progressively as part of wider geomorphological research, and the role of solution by carbonic acid was recognized before 1800. The benchmark studies originated from the Dinaric karst, led by Cvijić (1893, 1918) and Grund (1903, 1914) who documented the critical roles of dolines, underground drainage, caves and collapse in the evolution of limestone landscapes. The disorganized relief, discontinuous valleys and large closed basins were seen as the products of solutional erosion, and the special case of poljes was further described by Roglić (1938). Limestone pavements are better developed in the Alpine karsts than in the Dinaric karst, and the classic description of their karren features originated from Switzerland (Bögli, 1960).

Parallel studies of karst landscapes in the warmer climatic zones within China had little influence on Western thought. Lehmann (1936) described the cone karst of Java, and the limestone landscapes of the wet Tropics were then recognized as extreme forms of solutional erosion; the tower karst of China largely remained an enigma to Western geomorphologists until the political barriers to access and communications were lowered in the 1970s.

Early karst research in Britain largely followed the lead from the far greater limestone lands of Europe, though Reid (1887) was ahead of his time when he recognized the formation of dry valleys in the chalk during bygone periods of periglacial conditions. Among numerous regional studies of Britain's karst, research reports of wider significance include those concerning the dry valleys in the Peak District (Warwick, 1964), the pavements of the Yorkshire Dales (Sweeting, 1966), the depressions and valleys of the Mendip Hills (Ford and Stanton, 1968) and the dolines of South Wales (Thomas, 1974).

Britain's contributions to limestone research were summarized by Sweeting (1972). Modern studies have diversified from pure geomorphology, and have emphasized research into processes (Smith and Atkinson, 1977; Trudgill, 1985a). They have also included applied aspects of the science, relevant to both engineering (Waltham, 1989) and groundwater resources (Atkinson and Smith, 1974), and have expanded into all aspects of karst research in the limestone terrains of foreign lands.

Development of cave research

The earliest theories on cave origins suffered from a shortage of geomorphological data on underground features, and are now of only historical interest (Halliwell, 1974; Shaw, 1992). Archaeological excavations of near-surface cave sediments dominated cave studies in Britain in the nineteenth century, when the origin of the caves was largely ignored.

Cvijić (1893) recognized the role played by underground drainage in the evolution of karst topography, but it was Katzer (1909) and Martel (1921) who expounded the special case of karst drainage by discrete conduits and cave passages. Subsequently, a series of American papers established contemporary understanding of cave genesis in the English language. Davis (1930) favoured deep phreatic cave development when he tried to fit karst processes into his cyclic pattern of landscape evolution (Davis 1899), but his ideas lacked a foundation of underground observations. Swinnerton (1932) backed the concept of shallow phreatic cave development, at or just below the water table, and a dominance of vadose cave development was postulated by Gardner (1935) and Malott (1937). Bretz (1942) interpreted the features of cave morphology to identify early phreatic development and subsequent vadose development in most caves. Each of these authors was only partly correct in his understanding of the complex initiation of caves and the polygenetic nature of subsequent cave development (Lowe, 1992c).

A hydrological approach to the environment of cave development (Rhoades and Sinacori, 1941), combined with an early concept of the role of the water table, promoted papers on the levels of cave development (with respect to altitude), including studies of the Yorkshire caves by Sweeting (1950) and the Mendip caves by Ford (1965b). The concept of the water table in a complex karst aquifer was later questioned (Drew, 1966), and cave levels have been re-assessed in the light of modern knowledge by Palmer (1987).

Geological controls on the Yorkshire caves were identified by Simpson (1935), Myers (1948), Atkinson (1963) and Waltham (1970), and were then more widely recognized as the dominant influence on cave development (Rauch and White, 1970, 1977; Waltham, 1971a). The Mendip caves were the model for a wider theory on cave development established by Ford (1965b) and then modified to take increasing account of geology (Ford, 1971). Ultimately this matured into a general theory by Ford and Ewers (1978) which respected geological guidance within vadose, shallow phreatic and deep phreatic environments. The special case of the evolution of maze caves was described by Palmer (1975) and related to British examples by Ryder (1975).

The essential feature of the environments, processes and controls of cave development are now well established (Palmer, 1984, 1991; Ford, 1988; White, 1988; Ford and Williams, 1989), and general theories have been reviewed by Lowe (1992c). Modern cave studies are evolving on lines beyond pure geomorphology. The earliest stages of cave initiation are viewed with reference to the role of solution by sulphuric acid (Lowe, 1992b; Worthington and Ford, 1995). Research into the increased secondary permeability due to circulation of saline groundwater is based on fieldwork in submarine karstic caves in the Bahamas, and has important implications with respect to hydrocarbon circulation and storage in oil reservoirs (Smart *et al.*, 1988a; Whitaker and Smart, 1993). Cave systems in subaerial karst regions are providing unique evidence of the times-scales of landscape evolution, because many of them contain the longest sequences of dated sediments - which can be related to surface events on geomorphological principles.

Dating of cave sediments

There is no means of directly measuring the age of a cave. As the early stages of initiation may take millions of years, the timing of a cave's origin is arbitrary, but the main phase of enlargement from a narrow fissure to a large cave passage is commonly a recognizable event within its erosional history. This event cannot be dated, but any sediments within the cave must post-date the erosional enlargement, and there are various methods available for dating the sediments (Ford and Williams, 1989; Smart and Frances, 1991).

Absolute or radiometric dating relies on measurement of some process which has taken place in the material at a known rate from a known starting level. Radiocarbon dating is based on the decay of the unstable carbon-14 isotope since it was trapped in a sediment at the time of deposition; the initial isotopic ratio is that which is constant in the atmosphere, and the half-life is known. Dates are only reliable back to 45 000 years (45 ka), so the application to cave sediments is limited.

Uranium-series dating is based on the decay of unstable uranium-234 to produce thorium-230 where uranium-234 is trapped in the calcite lattice on deposition but the daughter thorium isotope is absent. Thorium-230 is produced by decay of the uranium-234, and measurement of the isotope ratio indicates time from the initial deposition. If the calcite is subjected to re-solution, or is contaminated by detrital thorium, the technique does not yield reliable ages. The normal age limit for this technique is 350 ka, though a stalagmite may be positively identified as being older than the limit;

mass spectrometric techniques, and the measurement of other uranium isotopes, can extend this limit to around 500 ka in some cases. Uranium-series analyses have provided the most numerous cave sediment dates (Gascoyne *et al.*, 1978; Ford and Williams, 1989).

Electron spin resonance (ESR) and thermoluminescence provide measures of the exposure of any calcite speleothem material to environmental radiation. The accumulated radiation dose received by a sample is determined by laboratory irradiation and deterioration of the thermoluminescence or ESR signal strength, while *in situ* measurement of the site-specific radiation dose rates allows calculation of the sample age. It is necessary to assume that the dose rate has been constant since deposition, but ages up to about 900 ka can be achieved with about 15% accuracy.

Comparative dating methods rely on correlation of a recognizable parameter with an externally calibrated chronology. Fossil and artefact records provide the conventional methods, but have limited application in most caves. Palaeomagnetic stratigraphy is based on identifying the polarity and orientation of the natural remanent magnetism recorded at the time of deposition by ferric minerals trapped in the sediments. It can be applied to sequences of clastic cave sediments, and also to calcite speleothems that carry weaker magnetic signals from their impurities but are not prone to post-depositional disturbance (Latham *et al.*, 1979). Sediment ages up to 2 million years (Ma) have been recorded from caves.

Determinations of cave sediment ages provide minimum ages for the cave passages in which they lie. Passage sequences may then be correlated, by their geomorphology, with surface palaeo-topographies, and this constitutes a powerful tool in the elucidation of landscape evolution. Stalagmites are subaerial deposits formed in caves that are at least partially drained, and their positions therefore indicate the maximum elevations of the contemporary local resurgences and valley floors; dated stalagmites provide important evidence of the rates of valley excavation and surface lowering through much of the Pleistocene. Speleothem sequences may also yield valuable information on palaeo-climates, as the trace element contents and the ratios of stable isotopes, notably oxygen-18, may relate to the temperature at the time of deposition.

Radiometric cave sediment ages are conventionally expressed in ka, which represents thousands of years before the date of analysis. Only radiocarbon dates are expressed in years BP, measured before an arbitrary 'present' at 1950 AD. All ages are defined with error bars, which are commonly around ± 5–10%, increasing slightly for the older material; these are all published in the primary records of the data, but are not included in this review. Cave sediment dates are particularly valuable sources of geomorphological data. Published data of major significance in Britain include those referring to the Mendip caves (Atkinson *et al.*, 1984; Smart *et al.*, 1988b; Farrant, 1995), the Yorkshire caves (Gascoyne *et al.*, 1981, 1983a, b; Gascoyne and Ford, 1984; Baker *et al.*, 1995b, 1996), the Peak District caves (Ford *et al.*, 1983; Rowe *et al.*, 1989b), caves in Devon (Proctor and Smart, 1991), caves throughout Britain (Atkinson *et al.*, 1978, 1986; Gordon *et al.*, 1989) and caves over a wider area (Hennig *et al.*, 1983; Baker *et al.*, 1993).

BRITISH KARST REGIONS

Most of Britain's caves and karst landforms occur on the thick and strong limestones of the Lower Carboniferous succession. The submarine palaeogeography of the Dinantian seas varied considerably across the area now occupied by Britain. Consequently, there is substantial lateral variation within the Carboniferous succession; the main limestone units in the main karst regions are correlated in outline in Figure 1.9. Most of the karst and caves are formed in the more massively bedded facies of the carbonates, which were deposited on slowly subsiding shelf, lagoonal and ramp areas. Contemporaneous clastic sediments accumulated in adjacent troughs, and interrupted the carbonate deposition when they extended over the shelf areas.

The major regions of cavernous karst are therefore defined by the major outcrops of the massive facies of the Carboniferous limestones – in the two parts of the Yorkshire Pennines, the Peak District, the Mendip Hills and South Wales (Figure 1.2). The finest limestone landscapes and the greatest extent of cave development lie in the glaciokarst of the Yorkshire Dales, formed on the thick Great Scar Limestone in the area around Ingleborough and Malham. The peripheral zone of the northern Pennines includes all the karst on the thin Yoredale limestones, and also on outcrops of the thinner and faulted equivalents of the Great Scar fringing the adjacent Lake District and Morecambe Bay. Both the White Peak limestone area of the Derbyshire Peak District and the Mendip Hills are upland karsts

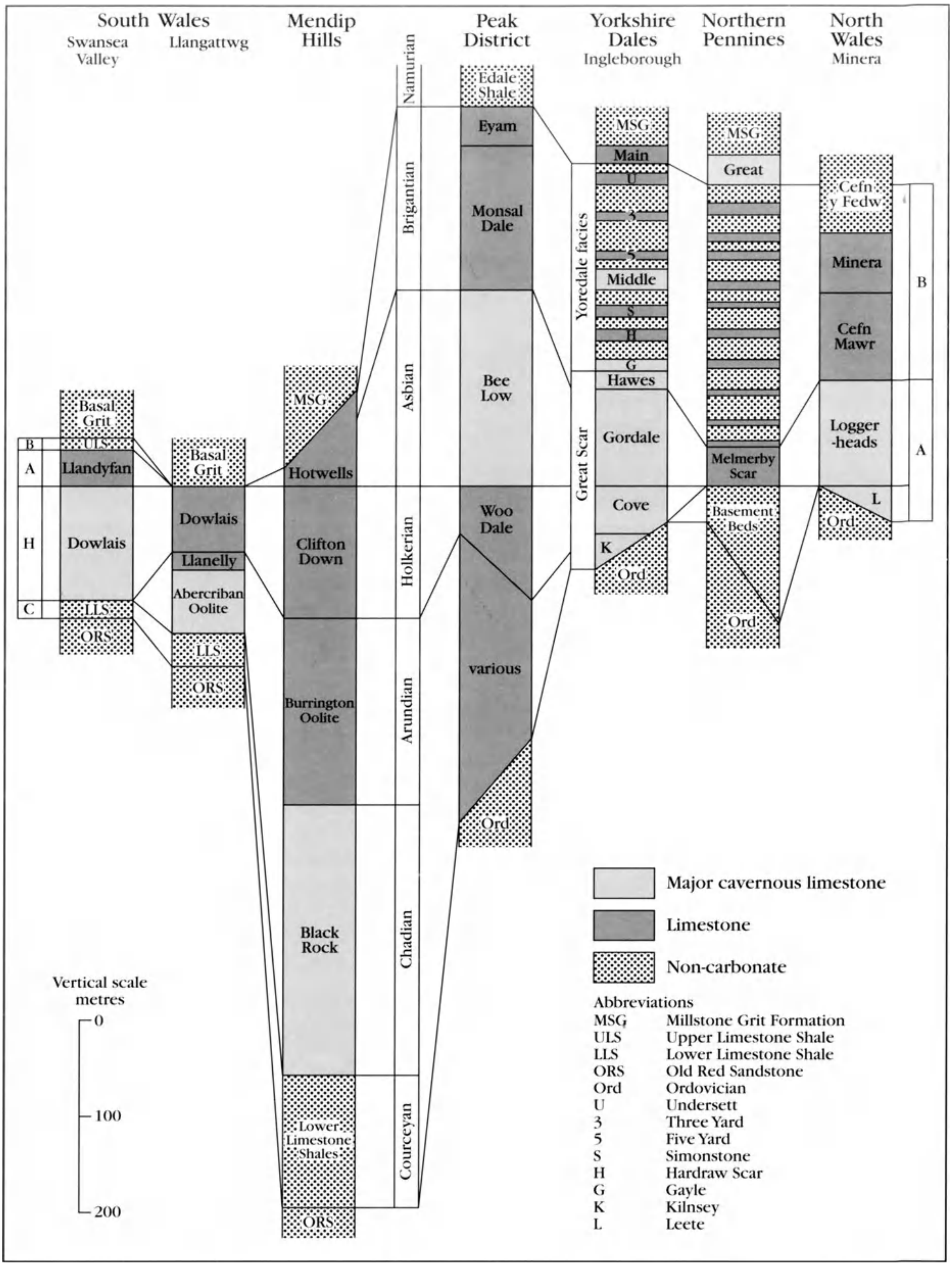


Figure 1.9 The main limestone units of the Lower Carboniferous within the major karst region of Britain. Thicknesses are generalized as there are considerable lateral variations. All the limestones are Dinantian, except for the Namurian Main and Great Limestones of the Pennines. In the Yorkshire Dales karst, the Great Scar Limestone is the massive carbonate facies developed on the Askrigg Block, and the Yoredale facies belongs to the Brigantian Wensleydale Group. In South Wales the Abercriban Oolite Group includes the Blaen Onneu Oolite. The main cover and basement rocks are identified; the Cefn y Fedw Sandstone extends across the Brigantian/Namurian boundary. All the named limestones are karstified to some extent, but the major cavernous units are distinguished. (Largely after George *et al.*, 1976; Arthurton *et al.*, 1988; Lowe, 1989a.)

which are clearly defined by geology and topography. The South Wales karst is spread along the limestone outcrop which fringes the coalfield syncline; it is not a conspicuous feature of the regional topography but it does contain many long, deep and important cave systems.

Each of the five main karst regions has suites of landforms and cave systems with their own distinctive characteristics (Table 1.1). The regional individualities are largely imposed by the geological structure, the relationships between geology and topography, and the local Pleistocene history of fluvial, periglacial and glacial stages. The geomorphology of the caves in the four main cavernous karsts have been comprehensively reviewed in the British Cave Research Association series of books on the limestones and caves of Britain (Northwest England - Waltham, 1974a; Mendip - Smith, 1975a; Peak District - Ford, 1977a; Wales - Ford, 1989a).

Outside the main areas of Carboniferous limestone, Britain's karst is dominated by the large area of chalk outcrop (Figure 1.2); this has a distinctive landscape of rolling downland and dry valleys, but contains very few caves. There are more caves in the smaller outcrops of older limestones, notably in North Wales, the Forest of Dean, Devon and Scotland. Individually, these lesser karst regions are often overlooked, but they form important components of Britain's landscape; for statistical purposes they are grouped together in Table 1.1. The Jurassic limestones, and other less extensive carbonates, have limited development of karst landforms and very few caves; they are briefly reviewed in Chapter 7.

Karst in the Quaternary

Most of Britain's landforms are the products of erosion and deposition during the Quaternary. The broad pattern of highlands and lowlands is a function of geological structure, with origins that reach back to Tertiary and earlier times. There are also remnants of uplifted, deformed and dissected erosion surfaces which pre-date the Pleistocene. But most individual landforms, and all the details of the landscapes, evolved within the Pleistocene and Holocene - when the cyclic climatic variations exercised great influence over the karst processes. Solutional activity was at a maximum during each warm phase. Conversely, it was greatly reduced in most cold phases; it ceased completely in most areas during periods of total ice cover, though glacial meltwater poured through the caves in some limestone blocks.

The main features of the later half of the Quaternary are outlined in Figure 1.10. The most conspicuous single event was the Anglian glaciation in the Middle Pleistocene, during which ice sheets extended across the whole of Britain north of a line roughly through London and the Bristol Channel (Figure 1.2); glaciers covered all the main karst regions, except for the Mendip Hills and the southern half of the chalk karst. Earlier and subsequent glacial advances covered lesser areas of Britain.

Little is known of the pre-Anglian cold phases, but there is evidence for at least five stages of glaciation in the northern parts of Britain (Bowen *et al.*, 1986). The chronology of these is uncertain, but sediments in the Mendip caves (Figure 5.7) identify multiple cold phases both before and after the Matuyama/Brunhes magnetic reversal, which is dated to 780 ka (Baksi *et al.*, 1992). The end of the Early Pleistocene is not defined within Britain, and is variously ascribed to the 780 ka magnetic event or the base of the Beestonian (Bowen *et al.*, 1986).

Following the Anglian glaciation, the warm interglacials of the Hoxnian and Ipswichian were important periods of renewed karstic activity. Glacial tills were deposited by a limited ice advance in north-eastern England during the intervening cold stage of Oxygen Isotope Stage 6 (Bowen *et al.*, 1986). This stage is widely referred to as the Wolstonian glaciation, but the age of the sediments at the Wolston section are open to question (Rose, 1987). The Wolstonian label may therefore be regarded as inappropriate, but it is still in use until a substitute name is accepted for this glacial event. There is similar uncertainty over events in the Early Devensian, where the Upton Warren and Chelford interstadials cannot be reliably correlated with the warm phases of 5a and 5c in the oxygen isotope record. Although these climatic oscillations can be recognized and dated in cave sediment sequences, the critical sediment profiles on the surface cannot yet be correlated to those underground.

The major ice expansion of the Late Devensian extended across Scotland, north and north-east England and most of Wales (Figure 1.2). It is referred to as the Dimlington glaciation, and is distinguished from the lesser Loch Lomond re-advance which post-dates it. Both the ice cover and the climatic change had massive impacts on many of the karst landscapes. The limestone outcrops of the Yorkshire Dales were scoured to form the basis of Britain's finest glaciokarst. Periglacial

Table 1.1 A comparison of the major features which give the individual character to each main karst region of Britain

Region	Yorkshire Dales[1]	Northern Pennines[2]	Peak District	Mendip Hills	South Wales	Rest of Britain[3]
Geology						
Karst area[4]	320 km^2	220	420 km^2	110 km^2	220 km^2	9000 km^2 (mostly chalk)
Karst relief[5]	270 m	70 m	260 m	260 m	330 m	200 m (chalk)
Limestone thickness[6]	200 m	40 m	400 m	700 m	150 m	200 m (chalk)
Typical dip	1°	1°	5°	30°	10°	Varies between areas
Last glaciation	Devensian	Devensian	Anglian[7]	None	Devensian	Varies between areas
Karst[8]						
Glaciokarst	• • •	• •				• (Scotland)[9]
Fluviokarst			• •	• •		• • (chalk)
Interstratal karst			•		• •	
Pavement area[10]	677 ha	613 ha	0	0	8 ha	28 ha (Scotland, North Wales)
Dry valleys	•		• •	•	•	• • (chalk)
Karst gorges	• •	•	•	• •		
Collapse features	•		•			
Doline fields	• •				• •	• • (covered chalk)
Ephemeral lakes					•	• (chalk)
Polygonal karst	•			•		
Famous sites	Malham Cove Gaping Gill	Hutton Roof Crags	Dove Dale Peak Cavern	Cheddar Gorge Wookey Hole	Dan-yr-Ogof Porth-yr-Ogof	
Caves						
Major passage types	Vadose joint shafts, phreatic on bedding	Joint mazes	Phreatic on veins and bedding	Downdip phreatic loops	Downdip vadose, strike phreatic	Vary between areas
Number of caves[11]	1420	620	210	220	270	410
Total cave length[11]	325 km	65 km	50 km	55 km	195 km	45 km
Caves over 1 km long	50	9	9	10	12	6
Longest caves[12] (km)	Ease Gill System 71	Goyden Pot 6	Peak-Speedwell System 14	Swildon's Hole 9	Ogof Ffynnon Ddu 50	Slaughter Cave 11 (Forest of Dean)
	Kingsdale System 24	Knock Fell Caverns 5	Giants Hole 5	St Cuthbert's Swallet 7	Ogof Draenen 48	Ogof Llyn Parc 4 (North Wales)
	Gaping Gill System 18	Fairy Hole 4	Bagshaw Cavern 4	Wookey Hole 4	Ogof Agen Allwedd 34	Uamh an Claonaite 3 (Scotland)
	Ireby-Notts System 12	Devis Hole 2	Carlswark Cavern 2	Gough's Cave 2	Ogof Daren Cilau 30	Ogof Llyn Du 2 (North Wales)
Deepest caves[12] (m)	Ease Gill System 211	Goyden Pot 61	Giants Hole 214	Eastwater Cavern 180	Ogof Ffynnon Ddu 308	Ogof Llyn Parc 115 (North Wales)
	Meregill Hole 206	Scrafton Pot 44	Masson Cavern 1 90	Longwood Swallet 175	Ogof Daren Cilau 217	Slaughter Cave 99 (Forest of Dean)
	Pen y ghent Pot 196	Pate Hole 33	Peak-Speedwell System 184	Swildon's Hole 167	Ogof Agen Allwedd 177	Cnoc nan Uamh 90 (Scotland)
	Gaping Gill System 195	Ayleburn Mine Cave 30	Nettle Pot 180	Manor Farm 151	Dan-yr-Ogof 140	Ogof Hesp Alyn 90 (North Wales)

1 The main southern Dales area on the Askrigg Block, including Dentdale, and excluding Nidderdale.
2 Including Nidderdale, the karst east of Morecambe Bay, and the eastern fringe of the Lake District.
3 Mostly the weakly cavernous karst of the chalk and oolitic limestones; including the cavernous karst of Devon, Forest of Dean, North Wales and Scotland.
4 Approximate area of karstic landscapes; does not include all the limestone outcrops.
5 Approximate values for the local relief within the limestone, which dictates the maximum descent from sink to rising, added to any depth of karstification beneath the resurgence level.
6 Geological data are generalized for purposes of comparison.
7 Or possibly Wolstonian - see text.
8 Most karst features are found to some extent in all the main karst regions, but their importance is assessed in relative terms:
• = significant, but minor;
• • = important and widespread;
• • • = internationally important.
9 Location of the major features noted in parentheses.
10 From Ward and Evans (1976).
11 Recorded caves longer or deeper than 5 m; figures rounded to nearest 10 caves and 5 km of passage; from unpublished database of Limestone Research Group, University of Huddersfield.
12 Subject to continuous revision, as lengths (and less frequently depths) are increased by newly discovered passages or by links found between known caves.

climate	stage		oxygen isotope stage	ka	period	culture
warm	Flandrian		1	3.8	Holocene	
wet				5		Neolithic
warm				10		Mesolithic
cold	Late Devensian	Loch Lomond	2	11	Late Pleistocene	Palaeolithic
warm		Windermere		13		
cold		Dimlington		24		
warm	Middle Devensian		3	59		
cold	Early Devensian			116		
warm	Ipswichian		5e	128		
cold	'Wolstonian'		6		Middle Pleistocene	
warm	Hoxnian			423		
cold	Anglian		12	478		
warm	Cromerian		13	524		
cold	Beestonian					
warm	Pastonian				? Early Pleistocene	
4 earlier cold phases						

Figure 1.10 The major glaciations and climatic variations, stages and subdivisions, and cultural phases, of the later parts of the Quaternary. The chronology is based on terrestrial material since about 120 ka, and on correlation with the earlier oxygen isotope stages in the marine sediment record. A more complex pattern of climatic variations is known to exist; both they and other debatable correlations are omitted. The problems of the Early Devensian subdivisions, the 'Wolstonian' glaciation and the Early/Middle Pleistocene boundary are referred to in the text. (After Imbrie *et al.*, 1984; Bowen *et al.*, 1986; Martinson *et al.*, 1987; Campbell & Bowen, 1989; Shackleton *et al.*, 1990.)

conditions were imposed on the Peak District, Mendip Hills and most of the chalk karst, during which time most of the modern dry valleys were deepened by subaerial fluvial activity. The Devensian ice cover retreated when the late glacial Windermere interstadial opened at about 13 ka; evidence of the warmer climate from fossil beetles in subaerial sediments (Coope, 1977; Gordon and Sutherland, 1993) correlates closely with the renewed cave stalagmite growth (Gascoyne *et al.*, 1983b; Atkinson *et al.*, 1986). A subsequent, short, cold stage was marked by the Loch Lomond Stadial glaciation in the Scottish Highlands, and stalagmite growth recommenced in many caves only at the beginning of the Flandrian (Gascoyne *et al.*, 1983b). From then until the present day, solutional processes have been dominant in the continued evolution of Britain's karst.

SELECTION OF GCR SITES

The karstlands of Britain contain a great range of surface and underground landforms, many of which are worthy of conservation in order to maintain the integrity of the nation's geological heritage. They include famous landmarks such as Cheddar Gorge and Malham Cove, some of which already receive a measure of protection within the National Parks. They also include remote cave passages, only ever visited by a handful of cave explorers. Their scientific values do not correspond to their fame or their accessibility, but they have been assessed on a national basis for the purpose of the Geological Conservation Review (GCR).

The aim of the GCR has been to compile a list of karstic landforms suitable for designation as Sites of Special Scientific Interest (SSSI). These represent all the important aspects of Britain's karst, and include the sites which have scientific significance for both the present and the future. The criteria for selection have therefore been any one of four factors:

1. The finest example of any particular landform or cave type, such as Cheddar Gorge, the Kingsdale cave system and the pavements of Asby Scar.
2. Unique sites, such as Malham Cove, Langcliffe Pot, the Pant-y-llyn turlough and Beachy Head Cave.
3. Sites important for teaching and research, such as the glaciokarst of Malham and Ingleborough and the Dove Dale fluviokarst.
4. Important assemblages of landforms, such as the Ingleborough karst and some of the more extensive cave systems.

By these criteria in their simplest form, 22 of the sites are included as the best examples of the features listed in Table 1.2. The values of the sites in the secondary listings of Table 1.2 are not diminished, as many are the best examples of important subtypes; the dolines of Wurt Pit, Sandpit and High Mark have structures and origins that are totally different from each other and from those of Ingleborough. There are also many local features of the geology which define important contrasts between the karst landforms of the different regions; geological controls are particularly conspicuous in the morphology of caves, and sites are included in the GCR to establish the regional characteristics.

Among the karst landforms, the cave systems and the limestone pavements constitute special cases with regard to their needs of conservation (Glasser and Barber, 1995; Webb, 1995; Bennet *et al.*, 1995).

Beside their values to karst geomorphology, many caves have great importance and value in the stratigraphy of the sediments that they contain. In erosional upland environments, caves constitute unique preservation sites where sediments can accumulate in stable conditions and remain safe from destruction by continued surface denudation. The value of these sediments is enhanced by the climatic sensitivity of solution processes and karst hydrology, and also by the chronological record that is deduced from their radiometric dating. The cave sediments provide a record of events with implications for research into the evolution of landscapes far beyond the confines of the karst. Many of the known cave sediment sequences have not yet been studied in detail, but as a data source for Pleistocene research, caves have unparalleled value. Remnants of sediments can survive in so many obscure corners of cave systems that a large proportion of caves could be considered as scientific resources worthy of conservation. Not every site can be sensibly protected, but this factor has been taken into account in the selection of GCR sites, which therefore include the most important known cave sediment sequences.

Limestone pavements form a special case for conservation because of their sensitivity to rapid and total destruction under the guise of small-scale stone extraction. The solutionally fretted clints forming the surface layer of many pavements have been in demand as a weird form of decorative stone, and many areas of very fine natural rock outcrop have been destroyed in the past. The pavements of the Yorkshire Dales and northern Pennines are of international repute, and are also of very special botanical value (Ward and Evans, 1976). The more important sites are designated as SSSIs, and separate protection measures have been created for many other valuable sites which cannot be included in the GCR (Webb, 1995).

The list of sites in the GCR can never be taken as a complete citation of Britain's karst geomorphology. New discoveries underground, and new research results from above and below ground, continually add to the geomorphological record. Three potential GCR sites have been identified during the documentation phase. The importance of Slaughter Stream Cave and Ogof Draenen were recognized when cavers removed sediment chokes from the entrances, and in each case discovered many kilometres of cave passage that have considerable geomorphological importance. Helbeck Scars is also a potential GCR site, reflecting the improving perception of the values of the lime-

Table 1.2 The finest examples of individual karst and cave features within the GCR sites of Britain. The listing of features is in the order of their description in Chapter 1. The tabulated data are recognized as being subjective, especially among the important secondary examples, which are not presented in any sequence of merit and are referred to by short versions of their full site titles.

Feature	Prime example	Important examples
Limestone karst		
Dolines	Ingleborough karst	Wurt Pit, Sandpit, High Mark
Dry valley	Lathkill Dale	Cave Dale, Conistone, Malham & Gordale
Karst gorge	Cheddar Gorge	Malham and Gordale, Hell Gill, Winnats
Collapsed cave	Penyghent Gill	God's Bridge, Porth-yr-Ogof
Limestone pavement	Great Asby Scar	Scales Moor, Ingleborough, Gait Barrows
Glaciokarst	Malham & Gordale	Ingleborough, Traligill
Fluviokarst	Manifold Valley	Lathkill Dale, Dove Dale
Polygonal karst	High Mark	Brimble & Cross
Interstratal karst	Mynydd Llangynidr	Draenen, Nidderdale, Llangattwg
Fossil karst	Green Lane Pits	Masson Hill, Pikedaw
Limestone caves		
Deep phreatic	Wookey Hole	Ease Gill, Cheddar
Shallow phreatic	Kingsdale caves	Ingleborough
Abandoned phreatic	Dan-yr-Ogof	Alyn Gorge, Ingleborough, Castleton, Llangattwg, Minera, Priddy, Sleets Gill
Maze cave	Knock Fell Caverns	Mossdale and Langcliffe, Hale Moss
Vadose canyons	Ease Gill Caves	Ogof Ffynnon Ddu, Castleton
Vadose shafts	Ingleborough caves	Ease Gill, Brants Gill, Buttertubs
Calcite deposits	Otter Hole	St Dunstan's, Boreham, Dan-yr-Ogof
Dated sediments	Charterhouse caves	Cheddar, Traligill, Ease Gill
Chalk karst		
Dolines	Cull-pepper's Dish	Devil's Punchbowl, Castle Lime Quarry
Dry valleys	Millington Pastures	Manger, Devil's Dyke
Cave	Beachy Head Cave	Water End
Salt karst		
Subsidence	Moston Long Flash	Rostherne

stone pavements in the wilder areas of the Pennines. These three sites are proposed for SSSI status, but have not been designated at the time of writing.

This volume of the GCR embraces the various aspects of karst geomorphology, but there are some notable omissions. Many cave entrances or passages with immediate access from the surface have been used as animal lairs or have become natural pitfall traps. These bone caves have therefore accumulated valuable records of past faunas (Stuart, 1983; Andrews, 1990; Simms, 1994), but their importance is to Pleistocene palaeontology rather than karst geomorphology. Some famous British cave sites, including Victoria Cave, Kirkdale Cave, Creswell Crags, Minchin Hole and Kent's Cavern, are therefore excluded from this volume, but are described in other parts of the GCR. Similarly, the major tufa deposits of Caerwys are of karstic significance, but are included with the Pleistocene sites of Wales. Many small caves are protected because they are important bat roosts, but they are not included in this review of the geomorphologically valuable sites. Britain does have some areas of gypsum karst, worthy of geomorphological study, but the subsiding dolines

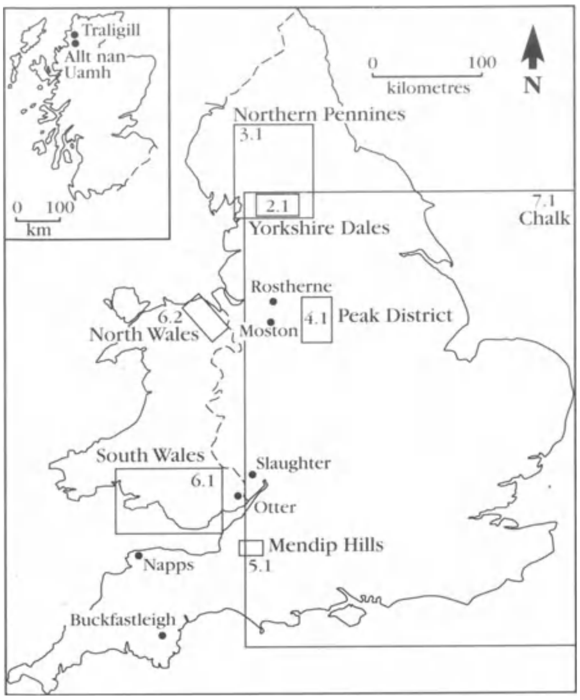


Figure 1.11 Key map showing the coverage of location maps in each chapter, identified by their figure numbers, and also the location of sites which are documented in the text but fall outside the chapter location maps.

of the Ripon area are currently of more significance to the construction industry, and are not included in the GCR.

The GCR karst and cave sites are distributed across most of the country (Figure 1.11). The structure adopted in this volume is to document the karst and caves of each main limestone region within their own chapter, covered by Chapters 2 to 6. The lesser karst regions are reviewed, and their more important sites are described in Chapter 7, except for the Scottish karst (reviewed in Chapter 8) and the North Wales karst (included in Chapter 6).

Chapter 2

The Yorkshire Dales karst

INTRODUCTION

The main part of the Yorkshire Dales karst lies on the largest outcrop of the Great Scar Limestone across the southern dales. The finest of the karst landscapes are around Ingleborough and Malham, but many of the major caves lie to the east and west, and the karst is continuous from the Dent Fault, in the west, to the eastern watershed of Wharfedale - a distance of 40 km (Figure 2.1). The topography is dominated by the massive unit of nearly horizontal limestone, whose top surface forms a series of plateaus and benches at around the 400 m level. Outliers, formed largely of shale, rise to summits at around 700 m, and the glaciated troughs of the Dales cut through the limestone to expose basement inliers. The limestone landscapes are the most spectacular in Britain, and have the country's finest glaciokarst landforms, while the geology has proved ideal for the development of large caves.

The Great Scar Limestone Group is the unit of strong Dinantian carbonates that is so conspicuous in the topography of the Yorkshire Dales karst. It consists of limestone beds of massive facies, formed through the Arundian, Holkerian and Asbian stages, locally subdivided into the Kilnsey, Cove and Gordale Formations (Arthurton *et al.*, 1988). The facies includes the Hawes Limestone of the lower Brigantian (Figure 1.9). The Great Scar Limestone is mainly formed of very pure, cream or pale grey, thickly bedded, bioclastic, sparites and micrites; these were a shallow water facies formed on the Askrigg Block, a shelf area partly bounded by faults and surrounded by deep water in the Dinantian sea (Ramsbottom, 1973, 1974). The Porcellanous Band is a fine, cream micrite at the Holkerian/Asbian boundary at the top of the Cove Formation; it is generally only 1 m thick, but may split into multiple units. Thin shale beds throughout the limestone succession greatly influence the cave development (Waltham, 1971b), but reef facies at Malham and in lower Wharfedale are of little significance to the wider karst geomorphology. The limestone is 160-220 m thick, and the variation is almost entirely due to transgression across over 50 m of local relief on the basal unconformity. Beneath the

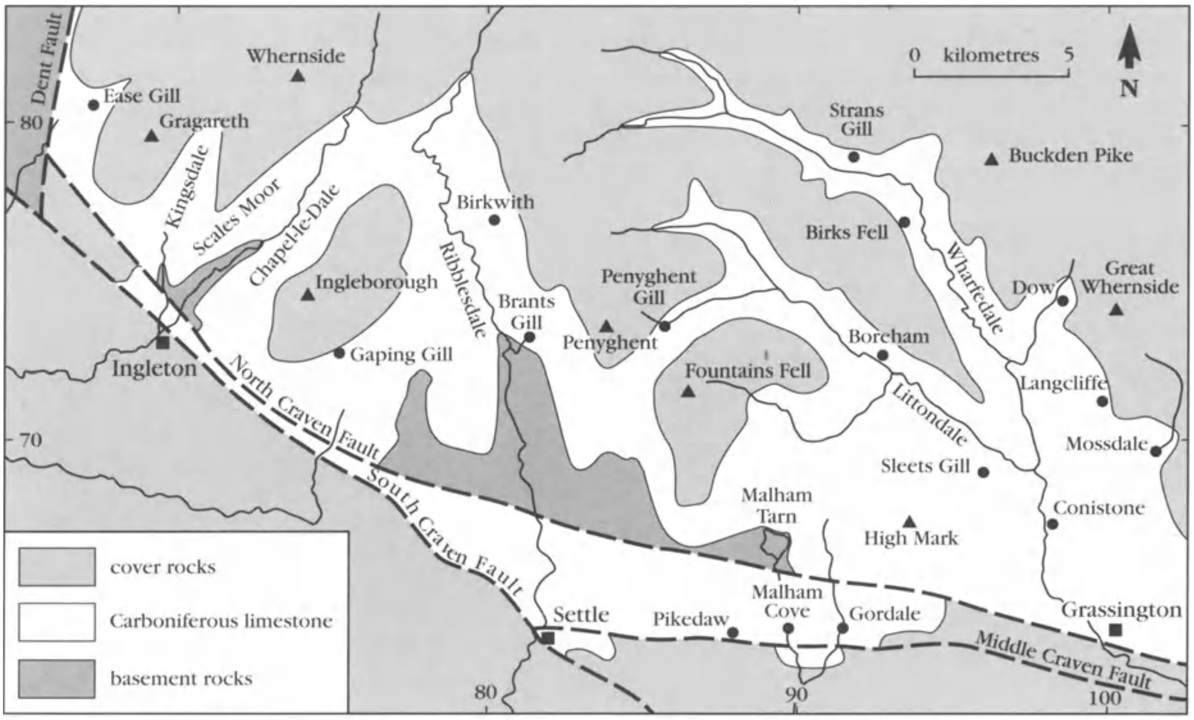


Figure 2.1 Outline map of the Yorkshire Dales karst, with locations referred to in the text. The Carboniferous limestone shown includes all the Great Scar Limestone (Kilnsey, Cove and Gordale Formations) and also the lower Yoredale limestones (of the Wensleydale Group) where they are hydrologically linked to the Great Scar and are therefore part of the same karst unit. Higher limestones within the Yoredale Series are not marked. Basement rocks are Palaeozoic slates and greywackes. Cover rocks are the Yoredale facies of the middle and late Brigantian Wensleydale Formation and various Upper Carboniferous and Permian clastic formations.

limestone, Lower Palaeozoic greywackes and slates are totally impermeable, and are exposed in the floors of most of the dales; their buried ridges provided the clastic debris for the discontinuous conglomerates at and near the base of the limestone.

The basement inliers in the dale floors are truncated to the south by the North Craven Fault (Figure 2.1). The southern limit of the karst is along the South and Middle Craven Faults, which downthrow to the south by many hundreds of metres. The slice of limestone between the faults is widest where it forms the splendid karst above Malham. The Dent Fault bounds basement rocks to the west, and forms the western edge of the karst. Both the northern and eastern limits of the main karst are formed where the Great Scar Limestone dips gently beneath its cover rocks. The regional dip is just a few degrees north, splaying to the north-west and north-east off the axis of the Pennine anticline. Minor faults occur all across the limestone outcrop.

Above the Great Scar Limestone, an alternating series of thin shales, limestones and sandstones forms the Brigantian Wensleydale Group (Figure 1.9). These were formerly known as the Yoredale Series (Hicks, 1959), but are now described as the Yoredale facies of the Brigantian. They are up to 300 m thick, and have considerable lateral variation; all the limestones have some karstic features, mostly on a small scale. The clastic units between the five lower limestones are locally absent, and the Gayle and Hawes Limestones are inseparable from the Great Scar across much of the Dales area. The Girvanella Band is a nodular, algal limestone about 1 m thick within the Hawes Limestone; it is often regarded as the top of the Great Scar facies. The entrance bedding passages of some of the major caves are within the Hawes Limestone, and east of Wharfedale, the Mossdale and Langcliffe caves in the Middle Limestone drain underground right through to the Great Scar. More significant to the karst than the local stratigraphy is the ubiquitous situation, where the higher slopes of the Yoredale shales provide surface streams that drain onto the main limestone benches.

The karst

Ice sheets scoured the entire Yorkshire Dales karst, during at least four of the Pleistocene cold stages. They interrupted the karstic processes of warmer climates, and their phases of glacial erosion alternated with those of fluvial erosion to impose a sequence of rejuvenations on the pattern of geomorphic evolution. The effects of the Devensian glaciation are most conspicuous within the present landforms. At its maximum, Devensian ice covered the entire area; during its retreat, summit nunataks appeared while the ice still swept over the limestone plateaus, and the final retreat stage saw only shrinking valley glaciers in the dales beneath the limestone benches. Ice flowed from the north, and its impact on the dales varied with the ice catchment as defined by the topography; Wharfedale and Ribblesdale carried the largest glaciers, but the Ease Gill and Malham valleys were both sheltered from major ice scour (Figure 2.1). Except for those two valleys, all the dales are deep glaciated troughs flanked by limestone scars.

All the streams and rivers have dry sections in their surface courses across the limestone. The Ribble and Wharfe maintain their surface flows in all but very dry weather, while most small streams off the shale outliers sink into caves and potholes under all conditions. The limestone high ground is therefore normally streamless. Its fine glaciokarst is best developed on the wider plateaus south-east of Ingleborough and north-east of Malham. The bare outcrops of limestone have great expanses of pavement, deeply incised by solution runnels. Where the fissured limestone is veneered with glacial till, doline fields with thousands of subsidence dolines (locally known as shakeholes) have formed, and are still active. Fluvial erosion of the mature karst has been limited to short periods of periglacial conditions, most significantly during the Devensian ice retreat. There are few dry valleys, but some were formed by subglacial and proglacial meltwater, and include the spectacular gorges of Gordale and Trow Gill, whose walls are now largely dry and therefore preserved. Malham Cove has a more complex origin, but the evolution and survival of its great limestone cliff are further consequences of the changing karstic processes. The large solutional dolines of High Mark (Figure 2.1) are probably the largest relics of interglacial karstic development in the region.

The Yorkshire Dales karst owes its spectacular geomorphology to the combination of so many landforms: the sinks which take all the drainage, the expanses of pavement, the long scars, the deep gorges and the innumerable dolines. The area is strictly a glaciokarst, but Ingleborough and Malham provide the finest limestone landscapes in Britain.

The caves

Nearly half of all Britain's known caves lie in the Yorkshire Dales karst (Table 1.1). This is because the geology presents an ideal cavernous environment: allogenic streams from the shale cover provide input to the top of the limestone, and this drains through to resurgences at or near the base of the limestone exposed in the dale floors (Waltham, 1974a). Most underground stream routes therefore have a simple staircase profile. Shafts are formed on joints or faults which are close to vertical, and nearly horizontal caves lie along the bedding planes and shale horizons within the limestone. Vadose canyons follow the bedding down the gentle dips, while phreatic routes are directed towards the available resurgence sites, regardless of geological structure. Hence looping cave plans are created where passage directions change in response to the hydrology, and patterns are further complicated where faults divert the underground drainage by overriding the bedding influence.

The geology imposes local detail on cave profiles, notably because vadose water descends the first available tectonic fracture. The deep daylight shafts, including the famous Gaping Gill, therefore drain into long sub-horizontal conduits at depth. Many other sinking streams find and follow shale horizons high in the limestone sequence, and therefore drain through long caves at shallow depth; Mossdale Caverns provide the extreme case, but their water does eventually descend to depth, and the Birkwith caves provide the grandest exception by draining out of a perched resurgence.

Vadose flow in the caves is mainly downdip to the north. Phreatic flow is then updip to the south, towards the lower surface levels. This accounts for the long flooded zones in the lower levels of nearly all the Yorkshire Dales caves; Keld Head is the finest example with over 7 km of flooded cave behind the resurgence. The phreatic conduits can also loop up and down between submerged bedding horizons; the route from Ireby Fell Cavern to Leck Beck Head has at least five phreatic lifts, of which the highest carries water more than 60 m up a vertical shaft on a joint or minor fault. The only long vadose streamway out to a resurgence is White Scar Cave, draining downdip into Chapel-le-Dale.

Previous to successive rejuvenations and surface lowering during the Pleistocene, higher levels of phreatic caves developed where the aquifer was impounded behind the impermeable rocks south of the Craven Faults. These caves were then abandoned as the entrenching dales created new resurgence sites, close to where they breached the fault barrier. The old phreatic caves were also developed largely along the bedding, and now form the high-levels, abandoned, invaded or intercepted by the modern, rejuvenated stream caves. The Gaping Gill Cave System has a long system of old sub-horizontal caves at depth beneath the famous daylight shaft, and Sleets Gill Cave has the best examples of abandoned phreatic lifts. These old caves contain extensive calcite and clastic sediment sequences, which record the Pleistocene environments and rejuvenations, but dating of the material has so far been on a modest scale, and much remains to be evaluated (Atkinson *et al.*, 1978; Gascoyne *et al.*, 1983a, b).

The combination of large dendritic systems of active cave passages and intercepted networks of abandoned conduits produces very long caves. The Ease Gill Cave System is the longest in Britain, and its links through to the Kingsdale caves are known to exist. It is only a matter of time and exploration effort before these links are found, and a single cave system over 100 km long will extend the whole way round the southern flank of Gragareth (Figure 2.1).

EASE GILL CAVE SYSTEM
(GCR name: Leck Beck Head Catchment Area)

Highlights

The caves under and around Ease Gill form the longest and most complex system in Britain, which ranks among the ten longest caves in the world. They contain an exceptional range of cave geomorphological features, and the passages, sediments and speleothems provide an unparalleled record of karstic evolution and associated surface development through the Pleistocene.

Introduction

The Ease Gill Cave System extends beneath Casterton, Leck and Ireby Fells, around the western and southern flanks of Gragareth Hill (Figure 2.1); it is bisected by the county boundary of Cumbria and Lancashire. A limestone upland is drained by the most extensive and most complex series of caves in Britain. The Dinantian Great Scar Limestone is about 200 m thick and contains numerous shale bands up to 2 m thick; it has a very gentle dip to the north-west. The Dent Fault juxtaposes the limestone against impermeable Lower Palaeozoic greywackes to the west; adjacent to the fault the limestone is folded into steep dips, with several small folds and tear faults in a disturbance zone up to 200 m wide. The south-west margin of the limestone is a clean break along the North Craven Fault, where the limestone has been thrown down to the south by about 200 m; thick glacial till masks the outcrop of the fault and the Yoredale shales beyond it. Away from the marginal faults, the limestone is broken by many minor faults and major joint sets dominated by trends almost north–south and north-west–south-east. A shallow syncline plunging downdip on Leck Fell is one of many minor flexures not yet mapped in detail.

Allogenic recharge of the limestone is derived from shale sequences of the overlying Yoredale facies forming the upper slopes of Gragareth and Crag Hill; all the streams sink close to the limestone margin. The surface topography and hydrology are complicated by thick deposits of glacial till over much of the limestone. This cover collects the runoff and directs it into the numerous subsidence dolines developed where the till is washed into the underlying limestone fissures.

The main surface feature of the area is the narrow valley of Ease Gill Beck (Figure 2.2). In normal weather conditions the beck carries no surface stream between the shale margin and the resurgence at Leck Beck Head, but flood flows reach various sinks down the limestone valley. Leck Beck Head lies just east of the Dent Fault, and is the rising for all the underground drainage between Aygill Caverns in the north and Ireby Fell Cavern in the south-east.

Despite the scientific importance of an integrated multi-level cave system with more than 80 km of passages, containing extensive clastic and speleothem deposits, published work on the underground geomorphology is limited. Accounts of the geomorphology of the caves beneath Casterton Fell have been given by Ashmead (1967, 1974b), and of those beneath Leck Fell by Waltham (1974d). Descriptions of all of the caves are given by Brook *et al.* (1994), and additional reports on particular caves include those by Foley (1930), Atkinson (1950), Simpson (1950), Eyre and Ashmead (1967), Waltham and Hatherley (1983), Yeadon (1985), Eyre (1989) and Monico (1995). A survey of all of the caves then known in the Three Counties System was published by Waltham and Brook (1980a). Aspects of the geology and its guidance of cave development in this area have been further discussed by Waltham (1970, 1971a, b, 1974a) and Lowe (1992b). Speleothem dates have been obtained from several caves (Waltham and Harmon, 1977; Atkinson *et al.*, 1978, 1986; Gascoyne *et al.*; 1983a, b; Gascoyne and Ford, 1984; Baker *et al.*, 1995b, 1996), and the relationship between the caves and landscape evolution has been discussed by Waltham (1986).

Description

The catchment of Leck Beck Head contains more than 80 km of known cave passages, most of which are integrated into the Ease Gill Cave System. The caves are sensibly divided into three sectors. Ease Gill Caverns contain about 58 km of passages, largely beneath Casterton Fell and linked beneath Ease Gill to Link Pot and Pippikin Pot (Figure 2.2). The caves of southern Leck Fell have about 12 km of connected passages (Figure 2.2); they also have a flooded connection with Pippikin Pot, making a single interconnected cave system 71 km long. The caves of Ireby Fell form another system nearly 12 km long which drains into the Leck Fell caves, though the connection has not yet been explored (see Figure 2.6).

Casterton Fell and Ease Gill Caverns

Ease Gill Caverns is a huge dendritic cave system containing more than 58 km of explored passages (Figure 2.2), which include virtually every type and feature of cave passage morphology encountered within the Yorkshire Dales karst. The caves are accessible through more than a dozen entrances, most of which lie in the Ease Gill valley, though four are located high on the fells to the north and south.

The main feature of the cave system is the massive, abandoned phreatic trunk passage which extends east to west under Casterton Fell. It originates beneath the upper reaches of Ease Gill and heads west to the complex of old passages in Lancaster Hole, where it is joined by similar large old passages from Bull Pot of the Witches and turns south to end at sediment chokes in multiple outlets. Parts of the trunk passage are 10 m in diameter; some sections retain their phreatic mor-

Figure 2.2 Outline map of the Ease Gill Cave System. Limestone is the Great Scar Limestone. Basement rocks are Silurian clastics only exposed west of the Dent Fault. Cover rocks are Yoredale and Namurian shales and sandstones. Notts Pot drains into the lower reaches of Gavel Pot (see Figure 2.5) (from surveys by Red Rose and Happy Wanderers Caving and Potholing Clubs, Northern Pennine Club, Cave Diving Group and many others.)

phology of rounded tubes and enlarged cross-rifts, but other sections are greatly modified by roof breakdown. There are extensive clastic infills, completely choking some passages, and calcite speleothems of spectacular variety decorate some very beautiful chambers (Figure 2.3). This old, high-level passage lies almost directly above the present, active main streamway. This is a splendid vadose canyon, mostly 2–4 m wide and up to 25 m deep, with clean limestone walls rising above stretches of gently graded streamway interspersed with water chutes and deep moulin pools. Parts of the canyon are entrenched in the floor of the old high-level tunnel, but other sections have bedding plane roofs or turn into high rifts. There is extensive undercutting, particularly along shale bands, and some has led to massive collapse from the passages above. The canyon ends below Lancaster Hole in a high rift containing the terminal sump pool; this drains into a large phreatic conduit 30 m below water level and heading straight for the Leck Beck Head resurgence.

Numerous smaller vadose streamways drain down the dip from the Ease Gill valley as inlets to the main streamway; several provide entrances, notably through Top Sink, Boundary Pot, Pool Sink and County Pot. Most of these are meandering canyon passages incised beneath shale beds and broken by small cascades and shafts mainly where joints and faults provided routes to lower shale beds. Repeated entrenchments, captures and re-routings into pre-existing passages have left complex intertwinings of active and abandoned passages on a scale unparalleled in Britain. Most inlets descend about 50 m to reach the main drain, and many lie below ancestral routes graded to the abandoned high-level trunk route. This further adds to the passage complexity, and some of the upper levels are cut by the Ease Gill valley where their choked upstream continuations are still visible in the south bank.

Cow Pot is the only significant sink on the shale margin high on Casterton Fell. Its stream descends an open vadose shaft to a washed-out shale bed along which it meanders before dropping 45 m through the roof of Fall Pot, a massive chamber formed where the old Ease Gill trunk passage has been enlarged by collapse into the streamway below. Close by, the entrance shaft of Lancaster Hole is a rift dropping 35 m into the high-level passages, and is now dry except in wet weather.

Immediately south of Ease Gill Beck, the northern end of Leck Fell is drained by the cave streams of Link Pot and Pippikin Pot (Figure 2.2). This part of the system is another complex of high-level and low-level passages connected by shafts and some descending inlets. Abandoned high-level routes of complex morphology originate under the Ease Gill valley and just south of the old trunk route through Ease Gill Caverns. They head south-west

Figure 2.3 Straw stalactites and short stalagmites form a spectacular display in Easter Grotto, part of the high-level passages in Ease Gill Caverns. (Photo: A.C. Waltham.)

and converge on the trunk route through Pippikin Pot which is finally lost beneath the breakdown in the spacious Gour Hall (east of Witches Cave). Like its contemporary in Ease Gill Caverns, the trunk route has abandoned phreatic tubes over 5 m in diameter, collapse modified chambers, thick clastic fills and areas of spectacular calcite decorations; some parts are totally blocked by sediment and collapse. The western limb of the high levels has been breached by downcutting of the Ease Gill valley, but a second abandoned passage provides an accessible route just beneath Ease Gill Beck where it is joined by the incidental entrance shaft of Link Pot. South of the Gill, four streams drain through young, low-level canyon passages, mostly less than 1 m wide, incised beneath the gently dipping shale horizons. The largest stream in Pippikin is the Cigalère inlet, with its long passage from sinks near the shale margin and an entrance high on the fell. This and the smaller streams find routes down joints to join the stream route from Link Pot which passes underneath on a lower shale bed. Where this water flows south-east against the dip, it is ponded into shallow phreatic loops, but it eventually descends a flooded rift to join the phreatic tunnel from Gavel Pot more than 20 m below resurgence level.

Immediately behind the Leck Beck Head resurgence, a complex area of flooded collapse and phreatic lifts obscures the confluence of the drainage from the Casterton and Leck Fell sides of the catchment (Figure 2.2). The main Leck Fell conduit appears to lie in Witches Cave, where a vertical shaft descends to a massive flooded passage which extends south at an average depth of 30 m. The nearby Pegleg Pot (Figure 2.2) contains sections of large old passage whose relationships to the other abandoned routes is unclear.

Bull Pot of the Witches contains a complex series of old high-level passages (Figure 2.2), which are abandoned phreatic tunnels extensively choked with clastic sediment. They largely follow shale beds along the axes of folds within the Dent Fault disturbance zone. Beneath them a younger vadose canyon drains into a phreatic loop through to the high-levels of Lancaster Hole; downstream of the loop, the stream cascades down a younger rift passage to reach the Ease Gill main drain just above the terminal sump. Aygill Caverns are an upstream continuation of the active and abandoned passages in Bull Pot of the Witches, but the connecting links are either choked or flooded.

Calcite speleothems from several parts of the Ease Gill caves have been dated by uranium-series analysis (Waltham and Harmon, 1977; Gascoyne *et al.*, 1983a, b; Gascoyne and Ford, 1984; Atkinson *et al.*, 1986; Baker, *et al.*, 1995a, b). Relatively few speleothems are older than 140 ka, and many post-date the main Devensian glaciation of the area, with ages less than 13 ka.

The caves of Leck Fell

A small group of deep caves are fed by streams draining off the shale slopes of Gragareth, and link into a single system of nearly 12 km of passages beneath the southern slopes of Leck Fell. The core of this system is the Leck Fell Master Cave, 1500 m long (Figures 2.2 and 2.6). Most of it is a large vadose canyon, with a phreatic roof half-tube in its downstream section, developed at a single stratigraphic horizon and draining downdip to the north and north-west. Its furthest upstream inlets are Lost Pot, with a sequence of vertical shafts, and the adjacent Box Head Pot, where a single shaft 110 m deep drops from a small shakehole directly to the streamway level. West of these, the complex Lyle Cavern High Level Series has large, old abandoned passages, choked with sediment and well decorated with secondary calcite, extending south-east and south-west towards passages in Notts Pot (Figure 2.6). The Lost Johns entrance series has a single stream canyon entrenched below a shale bed feeding a complex of rifts, canyons and deep shafts (Figure 2.5). These represent a sequence of stream routes invading a complex of tectonic fractures initially enlarged by phreatic solution. All routes rejoin at the fine waterfall shaft of Wet Pitch (Figure 2.4), from where bedding guided canyons lead to the Master Cave. Downstream, the main conduit collects further inlets, and drains through the Long Pool, a perched flooded section which retains the phreatic tube morphology; it then descends another vadose trench to the terminal sump pool, where an active phreatic tube continues to an underwater junction with the trunk conduit from Notts Pot.

Just 100 m directly above the Leck Fell Master Cave, the main streamway of Short Drop Cave is a meandering vadose canyon fed by inlets from sinks along the edge of the limestone (Figure 2.2). It contains sections of large, old canyon partly choked by sediment and collapse, but since reinvaded and undercut by the much smaller and younger vadose stream. The old vadose passage descends into a contemporary phreatic tube at the Gavel Pot entrance. It is almost choked by sedi-

Figure 2.4 A few horizontal bedding planes score the cleanly washed walls of the waterfall shaft at Wet Pitch in Lost Johns Cave. (Photo: A.C. Waltham.)

ment and collapse where it passes beneath the large dolines of Gavel Pot and Ashtree Hole. The stream flows through the chokes and continues as an underfit in the drained phreatic tunnels beyond; it then descends a series of vadose shafts to a sump which is a window into the phreatic conduit from Notts Pot.

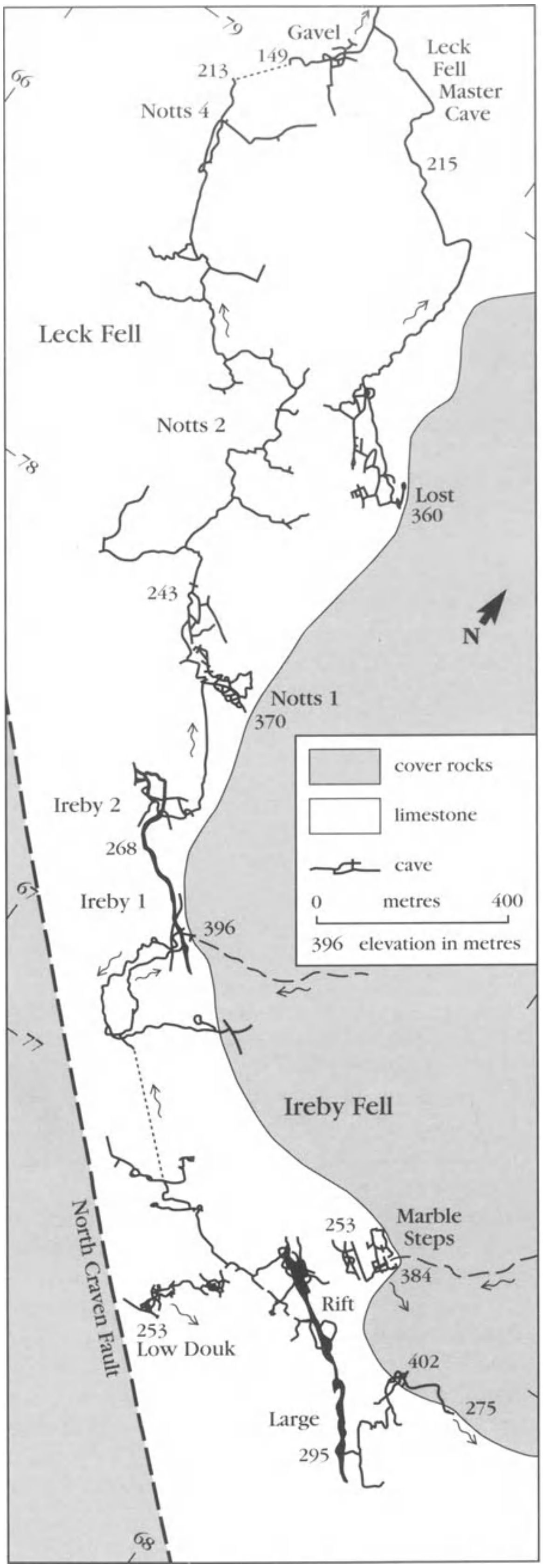
North of the Leck Fell Master Cave, the deep shafts of Big Meanie, Death's Head Hole and Rumbling Hole are developed on a single vertical fault. In Rumbling Hole the stream descends a series of cascades towards the east, to reach a long, narrow canyon tributary to, and cut below the same inception horizon as, the Master Cave. The shafts of Death's Head Hole and Big Meanie intercept an abandoned phreatic tube aligned on the fault. Extensive clastic fills block its old inlet from Rumbling Hole and its old outlet to the abandoned high levels of Gavel Pot. A small stream flows through it, from Long Drop Cave to a choked link to its inlet in the Master Cave below.

Absolute ages have been determined for calcite stalagmites and flowstone from Gavel Pot and Lost Johns Cave (Waltham and Harmon, 1977; Atkinson *et al.*, 1978, 1986; Gascoyne *et al.*, 1983a, b; Gascoyne and Ford, 1984). All sampled material proved to be either postglacial, <13 ka, or from a period of 140–85 ka spanning the Ipswichian interglacial and the early part of the Devensian.

Ireby Fell Cavern and Notts Pot

A major cave system with nearly 12 km of passages lies beneath Ireby Fell (Figure 2.6). It swallows two small streams at the entrances of Ireby Fell Cavern and Notts Pot, collects a large flow of percolation water, and drains out under Leck Fell. Ireby Fell Cavern has a fine meandering vadose canyon which descends a series of small shafts for more than 100 m to enter a large abandoned phreatic trunk passage more than 6 m in diameter. This is choked by sediment at both ends. Its stream is an underfit, which is ponded in the shallow sump into Ireby 2, before leaving the large old tunnel in a younger passage to the north. This follows a single bedding downdip, down a vadose canyon and then down a submerged tube to a phreatic lift up a fault into the lower galleries of Notts Pot. Higher levels of abandoned phreatic passages are known in both Ireby 1 and Ireby 2, connected to the lower levels by invading inlets, including the entrance passage. One small inlet has been dye tested from the abandoned high-levels in Rift Pot (Figure 2.6).

Notts Pot contains the most complex vertical maze of closely spaced, parallel shafts in any British cave. A small stream descends 120 m through this maze, cascading down routes which change when the flow is diverted by accumulations of clastic debris. Some high-level chambers and abandoned outlets are at similar levels to the Lyle Cavern high levels above the Leck Fell Master Cave. The shafts form the entrance series, known as Notts 1, which is merely an inlet to the drainage route beneath from Ireby Fell Cavern into Notts 2 (Figure 2.6). From the terminal sump of Notts 1, a phreatic loop drains into the long stream passage of Notts 2. The upstream end of this is a fine phreatic tube which can be followed for 700 m to a knickpoint, beyond which the stream descends in a narrow vadose canyon; the whole passage follows a single inception horizon downdip to the north-west. There are many tributaries, both active and abandoned, and the cave continues through further shallow phreatic loops into Notts 3 and Notts 4.



Underwater sediment chokes prevent progress downstream from Notts 4, into the continuation reached in the bottom of Gavel Pot. At least some of the flow appears to descend to depth, guided by unknown structural features, and then rise up a vertical flooded shaft in the Gavel Pot sump, forming a spectacular phreatic lift from a depth of 64 m. The top of the lift is on a bedding horizon just below water level, and a phreatic tube continues downdip to the north (Figure 2.2). The Gavel Pot stream joins it through a shaft in its roof, and the Leck Fell Master Cave joins it in a tube on the same horizon. It continues to a depth of 25 m where it is joined by the inlet from Pippikin Pot, and the continuation to Witches Cave remains unexplored.

Interpretation

Inception and development of the caves of Leck and Casterton Fells have proceeded within a framework of close controls imposed by geological structure and lithology. The role of bedding planes in directing underground flow is seen in the predominance of vadose passages draining down the dip to the north-west; these include the numerous inlet passages draining from sinks in the Ease Gill valley into the trunk drain of Ease Gill Caverns. The Leck Fell Master Cave, the main streamway in Notts 2–4, and the flooded conduit which takes both their flows to the submerged link with the Pippikin drain, all appear to be on the same bedding horizon. This is a spectacular example of a major inception horizon; the cave only leaves it in the deep, fracture-guided, phreatic loop which has its rising segment in the Gavel Pot sump (Figure 2.7).

Beneath Leck Fell many of the vadose stream caves follow constant stratigraphical horizons, and are guided in plan where they converge on trunk routes down the axial zone of a shallow, gently plunging syncline. The control which this fold has exerted on cave development is clearly seen

Figure 2.5 Outline map of the cave systems under the southern flank of Gragareth. Ireby Cavern and Notts Pot drain north into the Ease Gill Cave System, joining the water from the Leck Fell Master Cave, where the high-level passages in Lost Johns and Short Drop Caves have been omitted for clarity (see Figure 2.2); Marble Steps and Large Pot drain east to Keld Head (see Figure 2.8) (from surveys by Northern Pennine Club, Northern Cave Club and others).

where the Short Drop cave system lies directly above the Leck Fell Master Cave; both drain along the fold axis, aligned just north of west, and the drainage only converges after the high-level Short Drop stream has descended deep shafts on the joints in Gavel Pot. Where the Leck Fell Master Cave originally entered its contemporary phreas, it turned north, obliquely updip out of the synclinal trough. This created a shallow phreatic loop, only partially drained by subsequent rejuvenation and vadose entrenchment through the crest of the loop, creating the low airspace which exists today through the lake (Waltham, 1974d).

In similar style, a second shallow syncline appears to have collected the cave streams into its trough, along the line of both the Ease Gill Master Cave and the abandoned trunk passage directly above it. This is adjacent to an anticline down the line of the surface gill, south-east of which another syncline collects the drainage in Pippikin. Deeper synclinal troughs exert comparable controls over some passage locations in Bull Pot of the Witches and Aygill Cavern, adjacent to the Dent Fault. Elsewhere, the shale bands within the almost horizontal limestone influenced the development of the caves on distinct levels (Waltham, 1970), which were considered by Sweeting (1950) and Ashmead (1974b) to correlate with earlier base levels.

Fracture control of the passages and shafts is seen clearly in many parts of the cave system. Joint control is particularly well developed in the entrance series of Lost Johns Cave; the upper and lower streamways are developed along shale horizons, and a complex of active and abandoned rifts descend over 100 m between them, developed almost entirely on two intersecting systems of joints (Figure 2.6). The rifts were first partially opened by phreatic solution on the joints, far below the initial stream route which followed the upper streamway and then headed north, still following the shale beds in a continuation now totally blocked by sediment and flowstone. Rejuvenation prompted vadose drainage down through these rifts into a new streamway on the lower shale bed, and this captured the stream from above. The multitude of joints allowed frequent re-routing to form the parallel shafts and rifts seen today. The complex vertical maze of Notts Pot has also developed at the intersection of a number of joints and small faults. The influence of larger faults is clearly seen in the shafts and linear phreatic rifts and tubes along the fault through Rumbling Hole, Death's Head Hole and Big Meanie (Figure 2.2). The deep phreatic lift between Notts and Gavel Pots is on a major fracture which is probably a minor fault.

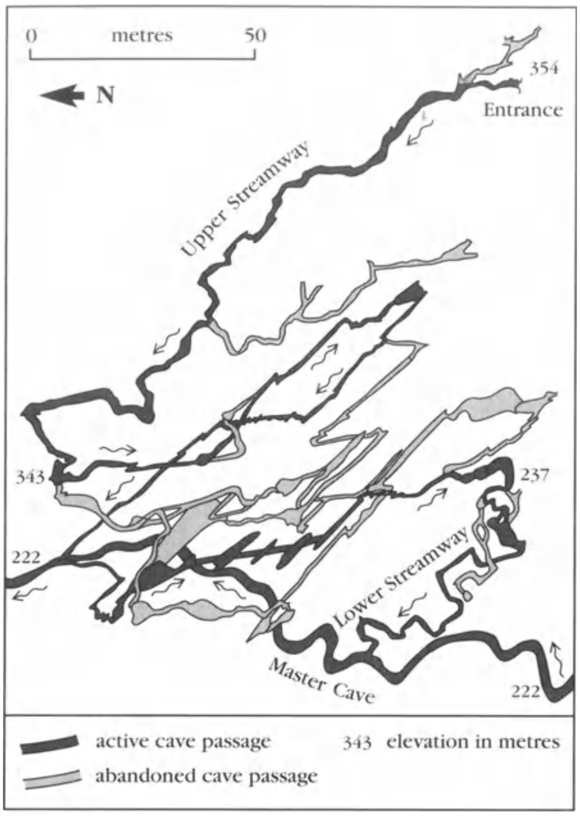


Figure 2.6 Map of the Entrance Series of Lost Johns Cave showing contrasts in passage morphology on shale horizons and joints, as described in the text (from surveys by London University Caving Clubs and University of Leeds Speleological Association).

The complexity of the cave system behind Leck Beck Head, and the presence of numerous high-level phreatic passages up to 100 m above the present resurgence, reflects a long and complex history extending well back into the Pleistocene. The Ease Gill valley appears to have been protected from deep glacial excavation by high ground to the north which prevented any major iceway developing towards the south, comparable to those in most of the Dales through the Yorkshire karst. This has restricted the proportion of old cave passages lost to glacial erosion. In the Leck Fell caves, three phases of erosion have been identified, separated by two episodes of sediment infilling (Waltham, 1974d). Each phase was characterized by vadose erosion to a greater depth in response to successively lower resurgence levels. Within each major phase, successive events can be recognized as underground captures evolved within the karst. The first erosive phase had a con-

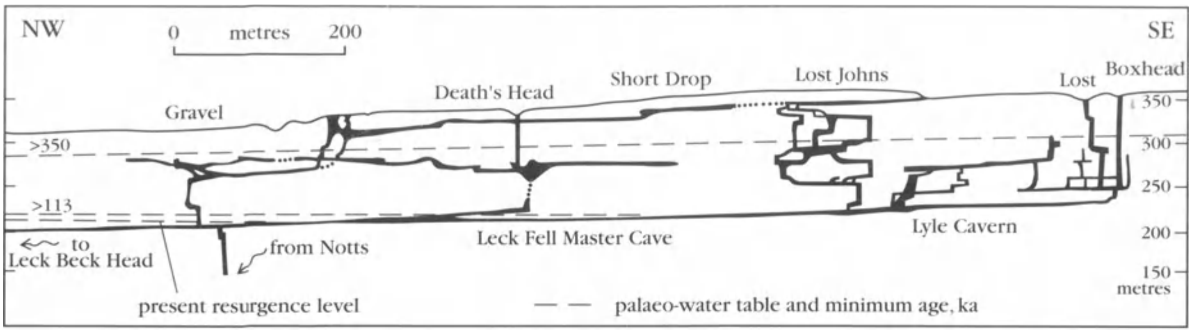


Figure 2.7 Profile through the main caves under the southern part of Leck Fell, with many passages omitted to improve clarity. The two palaeo-water tables are recognized from the cave morphology, and are ascribed minimum ages based on dated cave sediments, as described in the text. The horizontal scale is approximately the same as the vertical scale, but is distorted by changes in the direction of projection. (After Waltham and Hatherley, 1983; Waltham, 1986.)

temporary water table at an altitude of 290 m, identified by the vadose to phreatic transition down the Short Drop conduit into Gavel Pot (Figure 2.7). The second erosive phase correlated with a resurgence level at 225 m, only 12 m above that of the modern phase. A similar sequence could be anticipated in the caves of Casterton Fell, but Ashmead (1974b) identified only two main cycles of cave development, related to water levels found in the first and third phases recorded in Leck Fell. The intermediate level close to the present resurgence level may not easily be seen in the deep canyons at the lower end of the vadose main drain under Lancaster Hole. Evidence for the middle phase almost certainly lies in the complex of the Ease Gill inlets, but may be masked by the numerous captures and diversions controlled purely by the local geology.

The older passages include the large, abandoned, high-level, trunk route through Ease Gill Caverns, the ancient phreatic tubes forming the core of Pippikin Pot (Figure 2.2), the trunk route along the Death's Head fault, and the ancient stream canyon in Short Drop Cave which drained to the abandoned phreatic tubes of Gavel Pot (Figure 2.7). Another ancient conduit now forms the large, partially drained tunnel through Ireby Fell Cavern; this appears to have carried the main drainage from an old, high-level, ancestral Kingsdale through the large tunnels of Large Pot, and also from the old Marble Steps sink (Figure 2.5). The outlet of this route, north-west of its choke in Notts 2, was probably to a rising close to the Dent Fault, now obscured beneath the glacial till south of Leck Beck Head. It appears to have been abandoned when the more rapid Pleistocene deepening of Kingsdale, where it crosses the Craven Faults, diverted the flow to Keld Head.

By analogy with other sites, the abandoned, high-level, trunk caves under both Leck and Casterton Fells are likely to be at least half a million years old. Most of the dated stalagmites are much younger, from a major phase of deposition spanning 60-140 ka, though this included several long interruptions. Two flowstones from the high level in Ease Gill Caverns show only that this route was abandoned by 290-350 ka, and it was probably drained long before this. More significant are the flowstones in the Leck Fell Master Cave, which show that the resurgence was no more than 12 m above its present level at an age of about 115 ka. Since about 300 ka, the maximum rate of mean surface lowering in the Ease Gill Valley appears to have been only 0.2 m/ka (Waltham, 1986).

There are clear gaps in the recorded periods of calcite deposition within the caves, and there are also interruptions to the growth patterns in many sectioned samples. Calcite deposition was widespread during the Ipswichian interglacial, between 140 and 85 ka, but speleothem hiatuses show that the high-level main passage of Ease Gill Caverns was filled or flooded at least once between 200 and 140 ka. Detrital layers also show that deposition was interrupted, probably by flooding, in the Lyle Cavern high-levels between 128 and 123 ka. Baker *et al.* (1995b) have found a remarkably good correlation between the timing of speleothem growth in Lancaster Hole and periods of maximum solar insolation during the last 130 ka; the main periods of deposition were at about 130, 85 and 60 ka. Erosion of speleothems indicates that the lower part of the caves was again in the phreatic zone between 85 and 38 ka, and there was major flooding and reworking of the cave sediments during the latter part of the

Devensian, at 35–12 ka. These details of the cave geomorphology are evidence of the climatic fluctuations of the late Pleistocene; in this case, these appear to have influenced the stalagmite deposition largely by variations in recharge rates through the vadose aquifer (Baker *et al.*, 1995b).

Conclusion

This site contains the largest and most complex dendritic cave system in Britain, containing splendid examples of almost every type of cave morphology. The caves exhibit a clear influence by a variety of geological factors. The evidence from cave morphology, configuration, sediment and speleothem content is extremely important in studies of the Pleistocene history of the northern Pennines.

KINGSDALE CAVES

Highlights

The long and deep, dendritic cave system under Kingsdale contains four drainage routes which have been followed and mapped for the whole way from their sinks to the single resurgence. These include influent caves on both sides of the deep glaciated valley, and one trunk route passes completely beneath the valley floor. The lowest level of the cave system include the longest series of submerged passages known in Britain.

Introduction

Kingsdale is one of the smaller of the Yorkshire Dales, cut into the limestone due north of Ingleton (Figure 2.1). It is notable for the very fine caves which lie under the limestone benches of both flanks and also beneath the valley floor, many of which are connected into a single underground drainage system feeding the resurgence of Keld Head. Kingsdale is a straight, glaciated trough descending gradually from the north towards the scarp of the Craven Faults. The valley is floored by clastic sediment up to 20 m thick, deposited as an alluvial sheet grading into the lake sediments which accumulated behind a late Devensian retreat moraine; this barrier lies at the southern end of the dale, and is breached by a postglacial ravine.

The Kingsdale trough is cut into the Great Scar Limestone, most of which dips north at about 3°, except where shallow synclines locally reverse the dip. Drainage from the shale slopes of Gragareth and Whernside sinks almost immediately on reaching the limestone bench. The flow in the Kingsdale Beck is maintained on thick glacial and alluvial cover over the limestone as far as sinks below Kingsdale Head, below which the beck is dry as far as Keld Head except under flood conditions. Keld Head is the sole resurgence for the main cave system; it lies on the western side of the alluviated valley floor, close to the base of the limestone.

Comprehensive descriptions of the Kingsdale caves include those of West Kingsdale by Brook and Crabtree (1969a) and Brook (1971b), of the flooded passages behind Keld Head in Monico (1995), of some of the East Kingsdale caves by Gascoyne (1973), of Rift Pot by Davies (1984) and of all the cave passages in Brook *et al.* (1994). The development of Kingsdale and its caves has been discussed by Brook (1969, 1971b, 1974a), and Waltham *et al.* (1981). The chronology of the cave and valley development was discussed in the light of dated cave sediments by Waltham and Harmon (1977), Atkinson *et al.* (1978) and Waltham (1986), and further dates were published by Gascoyne *et al.* (1983a, b) and Gascoyne and Ford (1984). Aspects of the geology and geomorphology of the caves have been discussed by Waltham (1970, 1971a, 1974c), Lowe (1992b) and Halliwell (1979b).

Description

More than 35 km of passages have been mapped in the cave systems beside and beneath Kingsdale. The cave passages fall into four groups: the largely integrated caves of West Kingsdale, the influent caves under the eastern bench of Kingsdale, the submerged conduits in the phreas behind Keld Head, and the more isolated caves to the west around Marble Steps Pot.

West Kingsdale Cave System

The dry weather flow of Kingsdale Beck is lost into choked sinks near the head of the glacial trough, and is next seen in an active phreatic conduit east of Rowten Pot (Figure 2.8). This follows the bedding updip, into a partially drained series of canals, and then into the head of the West Kingsdale Master Cave. This is a splendid vadose

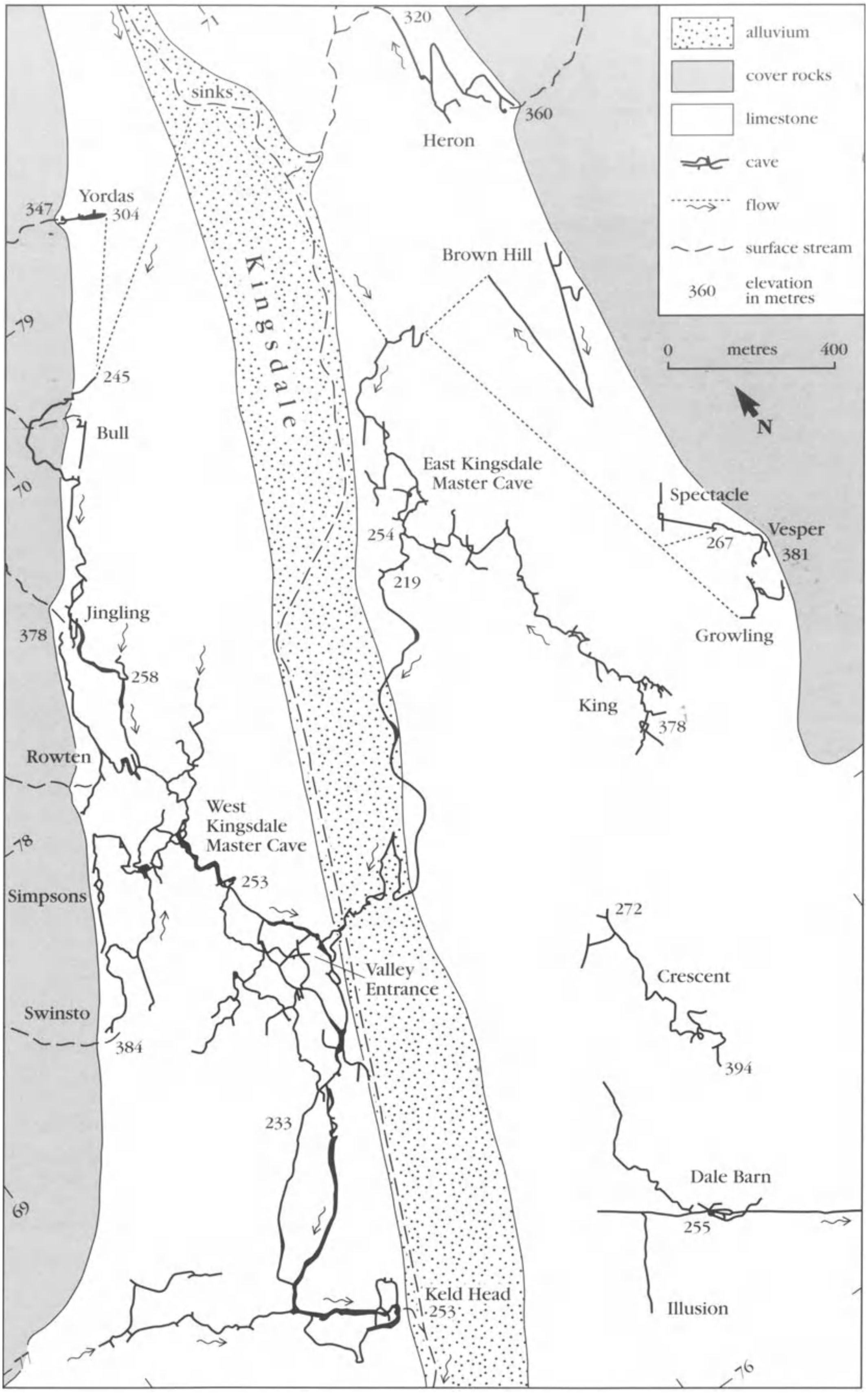


Figure 2.8 Outline map of the caves of Kingsdale. These include 7.5 km of caves behind Keld Head, at the southern end of the system and beneath the valley floor, which are totally flooded (from surveys by University of Leeds Speleological Association, Cave Diving Group and others).

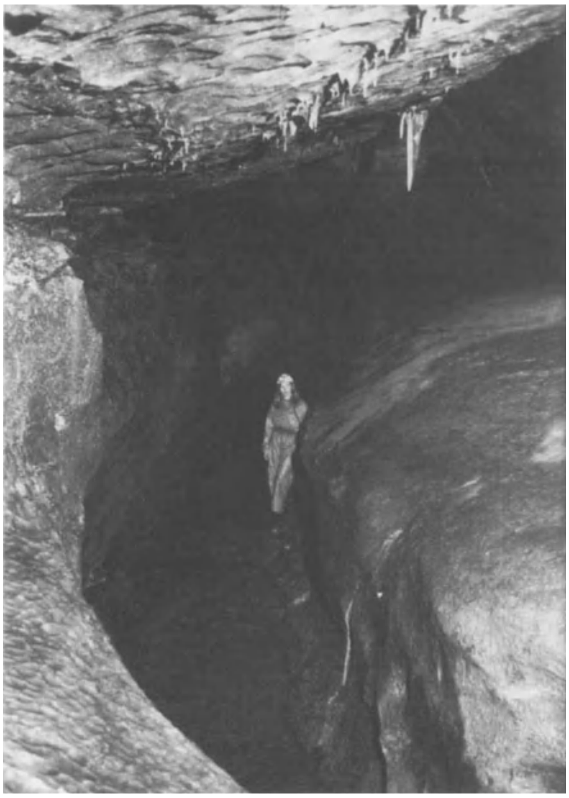


Figure 2.9 The vadose trench cut in the floor of the broad phreatic tunnel in the West Kingsdale Master Cave. (Photo: A.C. Waltham.)

canyon up to 7 m deep and 2 m wide below a phreatic roof tube 5 m in diameter (Figure 2.9). The roof tube follows the bedding updip so that the vadose canyon, descending over a series of small cascades, becomes more deeply entrenched downstream. Where the stream enters a sump pool at the start of the flooded conduits behind Keld Head, the roof tube continues south as an abandoned tunnel, to where it has been truncated by the glaciated flank of Kingsdale at Valley Entrance. Flowstone from the Roof Tunnel has been dated at 320–168 ka, but not yet with enough detail and accuracy to ascribe it to a sequence of Pleistocene stages. Divergent, abandoned tubes extend west of Valley Entrance, and intercept one vadose inlet well decorated with calcite speleothems, before becoming choked.

The main conduit beneath West Kingsdale is joined by inlets from a line of sinks along the shale margin on the limestone bench which stands 60–130 m above the dale floor. The most northerly of these is Yordas Cave, where a stream cascades down a series of rifts into the Main Chamber, 55 m long and 15 m wide, containing remnants of calcite-cemented sediments high on its walls. Its water joins some flow from sinks in Kingsdale Beck and drains into a long active phreatic tube passing beneath Bull, Jingling and Rowten Pots, before emerging in canals at the head of the Master Cave (Figure 2.8). Bull Pot has a series of narrow rifts choked at their southern outlet. Jingling Pot is a large open vadose shaft, 67 m deep, choked close to a high aven in a roof window to the phreatic conduit beneath. The high-level vadose canyons of Jingling Cave and Rowten Cave converge in the trough of a shallow syncline, where they drain into Rowten Pot. Developed along major north–south fractures, this is an open shaft system 105 m deep; it joins the low-level conduit almost at the point where drainage by the knickpoint at the head of the Master Cave canyon has created an airspace in the phreatic tube. Simpson's Pot has shallow vadose canyons on high-level shale beds draining to a shaft system, which descends to a partially drained phreatic tube tributary to the Master Cave. Swinsto Hole also has a long passage on the high-level shales, before a staircase of waterfall shafts and a descending rift lead to the partly drained tube to the Master Cave shared with the Simpson's water (Figure 2.10).

Caves of East Kingsdale

Most drainage from the limestone bench on the east side of Kingsdale collects underground in the East Kingsdale Master Cave (Figure 2.8). This is very similar to its parallel west of the dale, as it has a long phreatic tube, following the bedding and draining updip from the north, into the head of a vadose canyon up to 5 m deep and wide beneath a roof tube. The canyon descends to a sump pool where the roof tube drops down joints into the phreatic zone behind Keld Head.

On the limestone bench above, Brown Hill Pot has descending rift passages draining to the south, against the dip, along a major fracture; a flooded tube then drains north from a sump pool level with its outlet in the Master Cave (Figure 2.8). The flooded conduit upstream from the Master Cave has a branch carrying water from choked sinks in the floor of Kingsdale. Spectacle, Vesper and Growling Pots all have short passages on high-level shale beds, and deep shafts on fractures which descend to chokes close to the Master Cave level. King Pot has a long sequence of small shale-guided, downdip canyons, shafts and narrow rifts descending 120 m to a drained phreatic tube which joins the roof tube of the Master Cave.

Crescent Pot is another immature system of small stream passages whose link to the submerged main drain is unknown.

Away from the central group of potholes, Heron Pot has stream canyons entrenched below shale beds and draining downdip, linked by joint rifts which drain back to the south as they cut down through the bedding (Figure 2.8). Dale Barn Cave has a series of partly drained phreatic tubes and small vadose canyons, draining east at low level beneath Scales Moor into Chapel-le-Dale (Figure 2.11); sediment chokes block the old passages beneath the eastern flanks of Kingsdale, and Illusion Pot provides an entrance through joint rifts.

Keld Head Cave System

The phreatic zone behind Keld Head has a system of 7.5 km of converging and looping passages all below water level (Figure 2.8). The trunk conduit from the West Kingsdale Master Cave follows the bedding updip, except at three points where it steps down joint-guided shafts (Figure 2.10); the final section before the resurgence has developed along a series of parallel calcite veins. Passage morphology varies from elliptical phreatic tubes, up to 8 m wide and 3 m high, to wide and low bedding plane passages. Fine silts cover the floor in many places; elsewhere the floor is bare rock or scattered with cobbles derived from choked inlets. Close behind the Keld Head resurgence, tributaries from the west carry the drainage from Marble Steps Pot and its adjacent caves; these phreatic tubes form a series of loops not yet fully explored.

The flooded shaft below the East Kingsdale Master Cave drops 35 m to reach the bedding plane which the conduit then follows to the south-west, to connect with the main Keld Head phreas (Figure 2.8). This passes beneath Kingsdale in the 20–30 m of limestone that remains between the glaciated rockhead and the underlying basement rocks.

Caves of the Marble Steps area

The southern slopes of Gragareth, west of Kingsdale (Figure 2.1) are underlain by cave passages including those of Marble Steps Pot, whose streamways drain to Keld Head. Marble Steps Pot is a massive sinkhole which swallows a large moorland stream in times of flood (Figure 2.5). An upper series of chambers, rifts and shafts are developed on a series of hading fractures, and contain extensive fluvioglacial sediments which choke their outlet. A series of smaller, joint-guided rifts and shafts further to the south-west drop to a deep flooded rift at the same level as Keld Head.

Large Pot has a small streamway descending a series of immature vadose rifts and vertical shafts to a sump, which also drains to Keld Head. An abandoned distributary extends south to meet another streamway, which drains through a series of narrow rifts to a magnificent circular shaft 46 m deep into the large, old chamber of Necropolis. This is a section of abandoned, phreatic, trunk passage containing thick banks of clastic sediments. It continues north-west, beyond a series of boulder chokes, into another large chamber which can be reached by a 60 m deep shaft from the narrow entrance rifts of Rift Pot (Figure 2.5). Passages heavily choked with sediment and collapse extend westwards to intercept another old phreatic trunk route 3–4 m in diameter. This is choked in both directions, but a stream sinking in its floor has been dye tested to an inlet in Ireby Fell Cavern.

Low Douk Cave has a meandering vadose canyon draining to a sump pool level with Keld Head (Figure 2.5). It has intercepted several sections of large, abandoned vadose and phreatic passage, one of which terminates at a choke close to the old passages in Rift Pot.

Interpretation

The Kingsdale caves clearly show the geological control on their inception and development. The vadose canyons were initiated on bedding planes and shale beds, and therefore drain down the dip. This is roughly to the north except where local folding is stronger than the regional dip around Rowten and Simpson's Pots. Canyon streamways converge in the troughs of two shallow synclines, before finding fractures which allow them to drop to lower levels (Waltham, 1970). Most phreatic trunk passages are also developed along the bedding, except where they gain depth by dropping down joints to reach lower bedding planes. They then drain gently updip, following the hydraulic gradient to the south towards the lower ground and lower resurgence sites. Most of the bedding planes, which were the inception horizons for the caves, contain thin beds of shale; these are seldom seen in surface exposures, but their stratigraphical distribution underground adds an extra component to interpretations of cyclicity in the limestone deposition, and also accounts for levels of cave development unrelated to erosion levels (Waltham, 1971a, b).

Joints have determined the location of many cave passages, including the series of rifts in Marble Steps Pot and the zig-zag course of Heron Pot. The large vadose shafts and rifts in the potholes of East Kingsdale are developed on faults; many other shafts, including the sequence in Rowten Pot, are aligned on joints, which allow streams collected on the higher shale beds to drop down to the level of the main phreatic trunk passage. Joints have also influenced the phreatic flows, notably within the Keld Head phreas, where the passages drop down joint-guided shafts to lower bedding planes.

The section of the West Kingsdale Cave System between Swinsto Hole and Keld Head has been proposed as the type site for cave development in the Yorkshire Dales karst (Waltham *et al.*, 1981). The route from sink to rising is completely explored and mapped, and contains all the main types of cave passage: a vadose canyon drains downdip on shale beds, to shafts and a rift on joints, through old caves just below an ancient water table, down into a partially drained phreatic zone with canyons entrenched in tube floors, and into an active phreatic conduit largely updip on the bedding (Figure 2.10).

The sequences of abandoned high-level passages show that the Kingsdale caves have had a long history extending well back into the Pleistocene. Their development is linked to that of the adjacent Ease Gill caves, and to the fluvial and glacial excavation of the Kingsdale valley.

The oldest cave passage in Kingsdale appears to be the large abandoned phreatic conduit through the lower levels of Large and Rift Pots; it is likely that Marble Steps Pot was a major tributary sink into these passages. For much or all of their history, they drained to the north-west, through Ireby Fell Cavern, to an ancient resurgence in the lower Ease Gill valley. During subsequent glacial episodes, these passages and their resurgence have been largely or wholly choked with glaciofluvial clastic sediment. Parts of this trunk route were later invaded by smaller vadose streams, which now drain through lower outlets to both Keld Head and Leck Beck Head. This cave carried water sinking in an immature proto-Kingsdale whose floor was still well above the 300 m level of the passage in Large Pot. There were also inlet caves from sinks along the shale margin around Kingsdale, but only fragments of these abandoned passages are now seen at high levels intersected by the modern stream caves. It is conceivable that an even earlier stage had underground drainage from Gragareth feeding to a resurgence in Chapel-le-Dale, which is a much lower and older valley. If this route existed, the Rift–Large passage first carried a flow to the south-east, and the downstream conduits have either been eroded away by surface retreat on Twisleton Scar End, or remain undiscovered beneath Scales Moor.

During the Pleistocene Ice Ages, Kingsdale carried a major ice flow from the north and was deepened much more rapidly than the sheltered Ease Gill valley. The outlet to the Ease Gill resurgence was therefore abandoned in favour of new resurgences where the entrenched Kingsdale approached the Craven Fault scarp (Figure 2.1). The enlarged limestone catchment around the deeper Kingsdale supplied drainage to the developing low-level trunk routes, which are still at the core of the cave system. The sequence and levels of the resurgences in Kingsdale are largely unknown, but one old outlet was subsequently truncated at the Valley Entrance by deepening of the glacial trough. A later, lower route was out through Keld Head. Either or both of these truncated conduits could have continued beneath Scales Moor, perhaps through parts of Dale Barn Cave (Figure 2.11). Truncation by the side of the dale at Keld Head rejuvenated the first fractures upstream to rise above the new lower outlet level – and the West and East Master Caves were entrenched by knickpoint retreat in the two main conduits.

Calcite flowstone from the Roof Tunnel, inside the West Kingsdale Valley Entrance, has been dated to 168, 230 and 239 ka (Gascoyne and Ford, 1984; Waltham, 1986). These indicate that this passage was abandoned and the outlet to Keld Head was active by late Hoxnian times, if not before. This then implies that there was very little glacial deepening of Kingsdale in the post-Hoxnian and Devensian stages. The glacial trough is fresh and uneroded, but its large lateral and retreat moraines may indicate that there was more deposition than erosion during the Devensian. The caves would have been flooded or inactive while ice occupied Kingsdale, and the low levels were temporarily flooded when the lake was impounded behind the retreat moraine. A more detailed chronology of the successive deepening of Kingsdale and the evolution of its caves cannot yet be established.

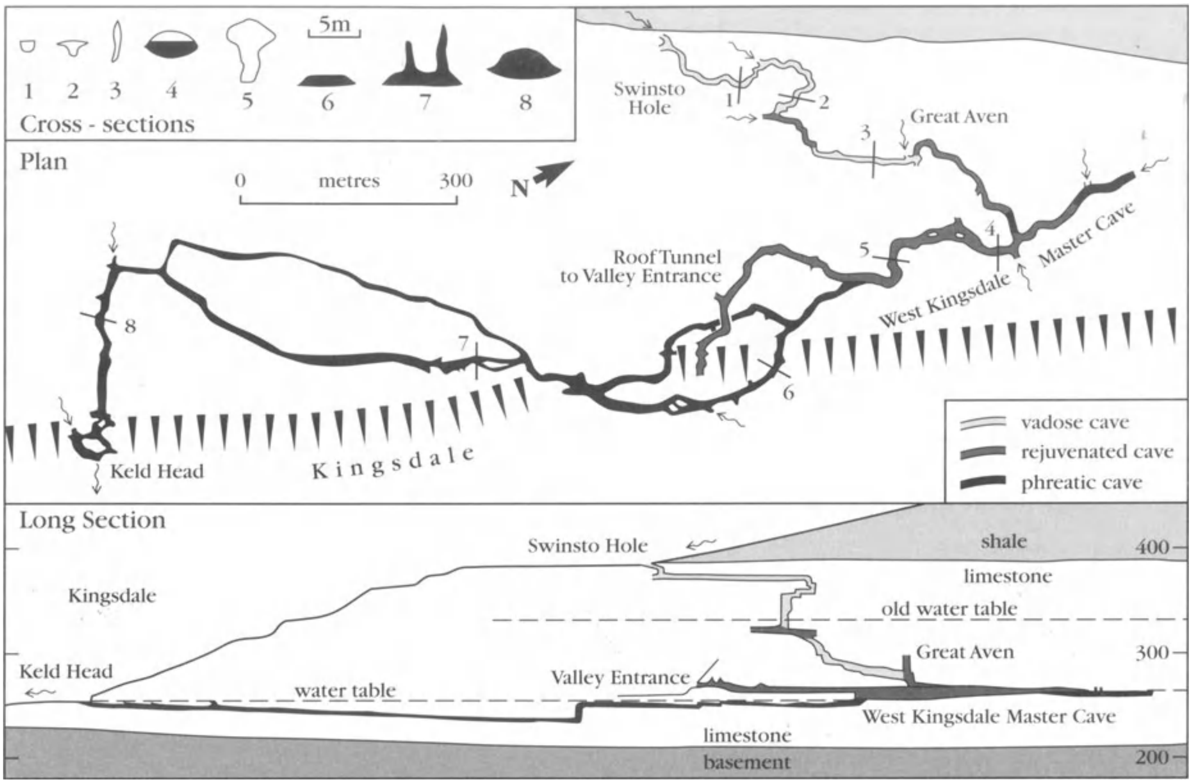


Figure 2.10 Swinsto Hole to Keld Head, the type example of a Yorkshire Dales cave, and part of the Kingsdale cave system. Only the main passages along the underground drainage route are shown; there are additional vadose inlets, abandoned passages and phreatic loops. The vertical scale of the long section is exaggerated by a factor of 1.5. (Mainly after Waltham *et al.*, 1981.)

The history of Dale Barn Cave is unclear, and there are old abandoned passages at both ends of the system which relate to early phases of the drainage of both dales. The main passage is at low level, and appears to represent recent drainage from Kingsdale towards a lower resurgence in Chapel-le-Dale, when the latter was deepened more rapidly by larger ice flows in the Devensian and perhaps earlier glaciations.

Conclusion

The caves of Kingsdale include the drainage route from Swinsto Hole through to Keld Head; this contains elements of vadose, phreatic and rejuvenated passages, and is completely mapped from sink to rising; it is the type example of cave development in the Yorkshire Dales. The 24 km of cave passages include a conduit which passes beneath a major valley, and the 7.5 km within the active phreas constitute the longest flooded cave system known in Britain.

The caves also represent past and present drainage links between Kingsdale and its neighbours, Chapel-le-Dale and the Ease Gill valley, and these provide evidence of the contrasting glacial histories of the three valleys.

SCALES MOOR

Highlights

Extensive limestone pavements on the limestone bench of Scales Moor, overlooking Chapel-le-Dale, are among the finest examples of horizontal pavements in Britain, and include some massive, undissected clints. Below the bare limestone plateau, Twisleton Scars form excellent sequences of terraces, scars, screes and pavements. Deep beneath the plateau, Dale Barn Cave carries drainage from Kingsdale through to Chapel-le-Dale.

Introduction

The Scales Moor pavements lie on the main limestone bench between the southern shoulder of

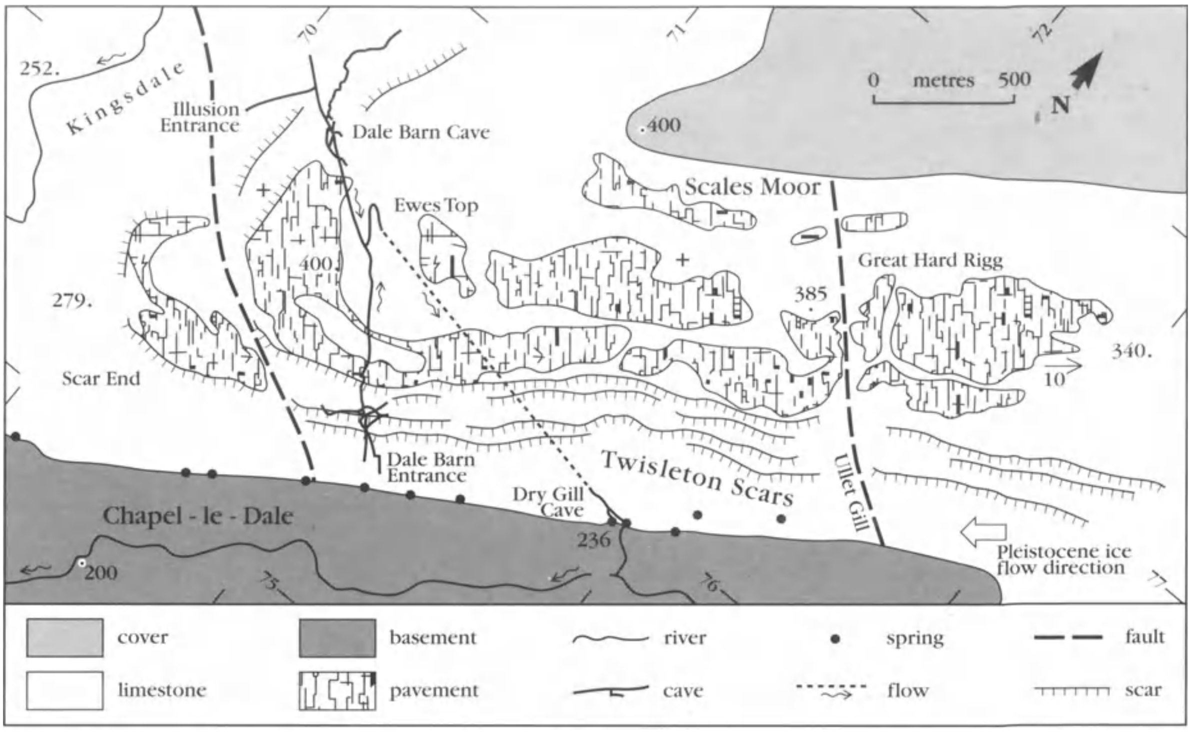


Figure 2.11 Geological map of Scales Moor. The limestone is the Great Scar Limestone, including the Hawes Limestone. Cover rocks are mainly clastic units in the Wensleydale Group. Basement rocks are Palaeozoic slates and greywackes. Only the larger areas of pavement are marked, and there are thin strips of pavement along the crest of nearly all the scars. Dale Barn Cave lies close to the base of the limestone, about 150 m below the main limestone pavements (cave survey from Northern Cave Club).

Whernside and the glaciated trough of Chapel-le-Dale (Figure 2.1). With a width of 800 m over a length of nearly 4 km, the bench constitutes one of the largest areas of nearly level limestone outcrop in the Yorkshire Dales karst. Most of its top surface stands at an elevation of just under 400 m, where it is formed on the top beds of the Great Scar Limestone. These are strong, massively bedded, sparry, bioclastic limestones; they are almost horizontal across most of the site, but north of the Ullet Gill fault they dip about 10° north-east (Figure 2.11). Drainage is entirely underground but the only known caves are close to the base of the limestone.

The morphology and solution processes on both the pavements of Scales Moor and the terraces of Twisleton Scars were studied and described at length by Sweeting (1966). Subsequent morphometric studies were carried out by Goldie (1973), and the glacial history of Chapel-le-Dale has been assessed in the light of dated sediments from its adjacent caves (Atkinson *et al.*, 1978; Waltham, 1986, 1990). Dale Barn and the various other caves are described by Brook *et al.* (1994) and Gascoyne (1973).

Description

From Ewes Top to Great Hard Rigg, Scales Moor contains wide expanses of bare limestone, littered with erratic boulders of locally derived limestone and sandstone; there are belts of acidic grassland, but the plateau is devoid of tree or shrub vegetation. On the main pavements, this empty rock landscape is very dramatic (Figure 2.12), but the panoramas soften towards the southern tip at Scar End, where terraced pavements drop to lower levels and are slightly better vegetated in the shelter below the main plateau.

Variations in the pavement surfaces were mapped by Sweeting, who recorded much of the northern part of Scales Moor as a very flat, glacially scoured pavement, dissected by few grikes and very few solution runnels (Sweeting, 1966). Further south much of the bench surface has more mature limestone pavement, with its surface morphology dominated by large clints, deep grikes and excellent runnel development of large rounded rundkarren. Many clints have centripetal runnel systems converging on small potholes which puncture the horizontal limestone surfaces.

Figure 2.12 The wide expanse of massive limestone pavement on Scales Moor, broken only by the deep rundkarren runnels and erratic blocks of sandstone and limestone. (Photo: A.C. Waltham.)

There are some very large clints on Scales Moor, with the average lengths in some sample sites exceeding 5 m (Goldie, 1986). The largest clints occur in the pavements just north and south of the Ullet Gill fault, and smaller clint sizes typify the areas south of Ewes Top. This is clearly a function of the spacing of tectonic joints in the limestone, and the southerly decrease may be related to the approach towards disturbed ground along the North Craven Fault (Figure 2.1). Not all these clints are truly massive, as some have a distinctly flaggy top clint while retaining large lateral dimensions. Some pavements at Ewes Top show lamellar weathering of the clints, giving them a flakey appearance, and other clints display honeycomb weathering, especially on their most exposed aspect (Sweeting, 1966).

Twisleton Scars forms a remarkable staircase of rock terraces, descending 150 m from the plateau rim to the base of the limestone, just above the valley floor (Figure 2.11). They form a spectacular sequence of limestone scars, each with an apron of scree along its foot; each scree overlaps onto a veneer of discontinuous glacial till, which has a soil cover and is punctured by small subsidence dolines (shakeholes). The till thins out onto a strip of pavement along the crest of the next scar (Sweeting, 1966). The scars are between 2 m and 15 m high, probably reflecting the spacing between thin shale beds which eased glacial plucking of the overlying beds. Alternatively, the terraces and pavements may have formed simply on the stronger limestone beds, but these commonly underlie the thin shale horizons within the lithologically varied limestone sequence (Schwarzacher, 1958; Waltham, 1971b; Ramsbottom, 1973). Clint dimensions on the terraces are generally 0.8–2.9 m (Goldie, 1976), and there is some tree and shrub cover in the more sheltered sites.

There is no surface drainage on Scales Moor. Rainfall runs into the deep grikes, but after heavy rain streams can be heard and occasionally seen flowing down the bedding planes 1–3 m below the surface of the inclined pavements of Great Hard Rigg. Ultimately all the water finds joints to descend to greater depths. Much of it resurges from the well defined spring line along the base of the limestone (Figure 2.11), but water in the dipping limestone probably flows north-east to join the main drains feeding God's Bridge (Figure 2.13). Dale Barn Cave lies very close to the base of the limestone. Streamways from the two entrance areas carry water from both Kingsdale and Chapel-le-Dale to a confluence almost directly beneath the topographic divide, and then drain to the Dry Gill resurgence (Figure 2.11). Abandoned passages occur above all three entrance areas, mostly at levels no more than 15 m above the streamways.

Interpretation

The main plateau surface of Scales Moor is formed on the top of the Great Scar Limestone at an altitude of about 400 m. It is clear that much of the surface is a stratimorph (Waltham, 1970), as it follows the dip down the inclined pavements north of the Ullet Gill fault. The main horizontal bench also forms part of a conspicuous erosion surface, widely recognized in the Yorkshire Dales (Sweeting, 1950). Present opinion tends towards the lithological explanation of the gross form of the limestone benches in the Dales, and glacial scour has been clearly the dominant process in the most recent stripping of the Scales Moor pavements (Waltham, 1990). Ice flowed from the north-east and was powerfully erosive as it swept up the dipping slabs and then across the horizontal limestone, even though the glaciated trough of Chapel-le-Dale must have acted as an adjacent ice-way.

The most striking areas of pavement on Scales Moor are on the more massive limestones where they were subjected to the most intensive glacial scour, but the solutional features are dominated by rounded rundkarren, whose origins relate to past covers of soil and vegetation. Over many years the position of the contact line between bare limestone and vegetation cover was monitored on a terrace-top site on Twisleton Scars (Sweeting, 1966). After 13 years the grass cover had advanced slightly over the inner part of the pavement, outwards from the till and scree below the upper scar. Rock exposure in the centre of the pavement was unchanged, but the grikes had increased vegetation, and soil depth within them had locally increased by as much as 100 mm.

On Scales Moor, Sweeting (1966) observed limestone surfaces cleared of drift with rock hollows over 800 mm in diameter containing erratics 500–600 mm across. These depressions appear to have existed before the erratics were deposited in them, and were therefore formed before the last glaciation; the implication is that some of the grikes may have a preglacial component to them, and merely continued to evolve since the Devensian glaciation. There is no evidence to support the alternative concept that the hollows were enlarged by solution beneath the drift. An area of limestone freshly exposed in 1947 was re-examined in 1960 having been subject to solutional attack by peaty water (Sweeting, 1966). Some parts of the exposed surface had been lowered by 30–50 mm in this period. At another test site on Scales Moor, solution runnels 70–150 mm deep had been cut into the limestone by waters coming off a peat slope for 13 years. Mean solution rates measured in the area over a wide range of local conditions extrapolate to indicate about 500 mm of surface lowering in the last 12 000 years; the enormous local variations within this average account for the variety of pavement morphology.

The pavements of Scales Moor suffered modification through the mechanized removal of clints during the 1960s, and surface limestone was used to build sheep folds and other structures in earlier periods (Goldie, 1976). Known areas of clint removal on Scales Moor became well vegetated with moorland grasses within the following ten years. Scales Moor is also heavily grazed by sheep, which help to confine tree and shrub vegetation to the grikes.

Little is known of the cave drainage beneath most of Scales Moor. Infiltration through the pavements probably descends rapidly to lateral conduits near the base of the limestone; high avens do feed water into the Dale Barn streamways. Patterns of flow are difficult to predict in the horizontal limestones, but drainage occurs from Kingsdale because its floor is nearly 50 m above the level of the base of the limestone exposed in Chapel-le-Dale. Except for some truncated fragments in Ullet Gill, there are no known relict caves at high levels under the fell, though some may have been removed when Twisleton Scars were cut back by the Chapel-le-Dale glaciers.

Conclusions

Scales Moor has some of the finest examples of level and gently inclined, little dissected limestone pavements in Britain. Both on the plateau and in the scar sequences, there is a wide variety of pavement types related to glacial plucking and varying joint densities. The karst has been an important research site, yielding data on solution rates, erosion processes and vegetation changes, with wide implications on glaciokarst studies in Britain and elsewhere.

INGLEBOROUGH KARST

Highlights

A broad limestone bench surrounding the summit mass of Ingleborough constitutes Britain's finest single area of glaciokarst. It contains spectacular limestone landscapes, which have an unparalleled scale of surface and underground karstic development. Ingleborough has virtually every type of karst feature, and is one of the best documented and most visited karst areas in Britain.

Introduction

The Ingleborough benches form a triangular block of limestone nearly 10 km across, bounded by the glaciated troughs of Ribblesdale and Chapel-le-Dale and by the Craven Fault scarp across the south-west (Figure 2.1). An outlier of sedimentary rocks dominated by shale forms the summit mass, which rises as one of the well known Three Peaks above the limestone plateau of the Craven Uplands. The summit reaches an altitude of 723 m, the top surface of the gently sloping limestone bench lies at 350–440 m, and base level in the dale floors is at about 220 m. Ingleborough forms a magnificent limestone landscape of wild open country, containing an impressive range of splendid glaciokarstic and fluviokarstic landforms. Its conservation values are regarded as a special case within the Yorkshire Dales National Park, and the whole mountain is designated as a Site of Special Scientific Interest for its geological and biological features.

Ingleborough is remarkable for the excellence of both its surface and underground karst landforms. Within any mature karst landscape, the processes and evolution of the cave drainage systems are totally interrelated with the progressive development of the surface topography, and they should be viewed together. The sheer scale and significance of the geomorphological interest in the Ingleborough karst dictates that its description and interpretation are subdivided, and separate reviews of the surface karst and the caves are more appropriate than a geographical subdivision. The caves are described in the next section.

Geologically, Ingleborough lies on the upstanding southern edge of the Askrigg Block, bounded to the south by the Craven Fault system which separates it from the Craven lowlands. The main karst is formed on the Dinantian Great Scar Limestone, consisting of nearly 200 m of pale grey, fine-grained, bioclastic limestones (Hughes, 1909; Moseley, 1973; Doughty, 1968). There is considerable local variation in the carbonate lithology, and individual beds are mostly 0.5–5.0 m thick, commonly separated by thin partings of shale. The limestone is well jointed, has a high secondary permeability and is well karstified; it generally dips a few degrees to the north. Above the limestone, the summit outlier consists of the cyclic sequences of interbedded shales, sandstones and thin limestones of the Yoredale facies of the Brigantian Wensleydale Group. Ordovician and Silurian slates, mudstones and greywackes form the basement to the karst aquifer. They lie beneath a strong unconformity, which is exposed in the floors of Chapel-le-Dale, Clapdale, Crummack Dale and Ribblesdale.

The diversity of karst features on Ingleborough has prompted a long history of research, of which the earlier work is reviewed by Halliwell (1974). The geology of the area has been described by Garwood and Goodyear (1924), Dunham *et al.* (1953), Rayner (1953), Wilson (1974), Waltham (1974b), Arthurton *et al.* (1988) and many others. Further research on the Great Scar Limestone includes that by Schwarzacher (1958), Sweeting and Sweeting (1969), and Waltham (1971b), and the Yoredale beds were described by Hicks (1959). The karst geomorphology of Ingleborough is reviewed by Sweeting (1950, 1966, 1974) and Waltham (1970, 1990), Waltham and Davies (1987) and Waltham and Tillotson (1989); the various erosion levels were further described by Trotter (1929), Hudson (1933), King (1969) and Clayton (1966, 1981). Dating of stalagmites from some of the caves has provided limited evidence for the rates of valley entrenchment and the chronology of geomorphic evolution of Ingleborough (Atkinson *et al.*, 1978; Gascoyne *et al.*, 1983a, b; Gascoyne and Ford, 1984; Waltham, 1986).

Description

The diversity of the karst landforms on Ingleborough is due in large part to geomorphic contrasts produced by the patterns of the Pleistocene glaciations; the ice scoured and eroded some of the limestone outcrops, while protecting and burying others beneath till. During each cold stage of the Pleistocene, ice moved from the north, and the main flows diverged around Ingleborough to continue down Chapel-le-Dale and Ribblesdale. During the glacial maxima, ice covered the entire landscape, and the two dales were iceways beneath an ice sheet which spread over and scoured the limestone benches. During advance

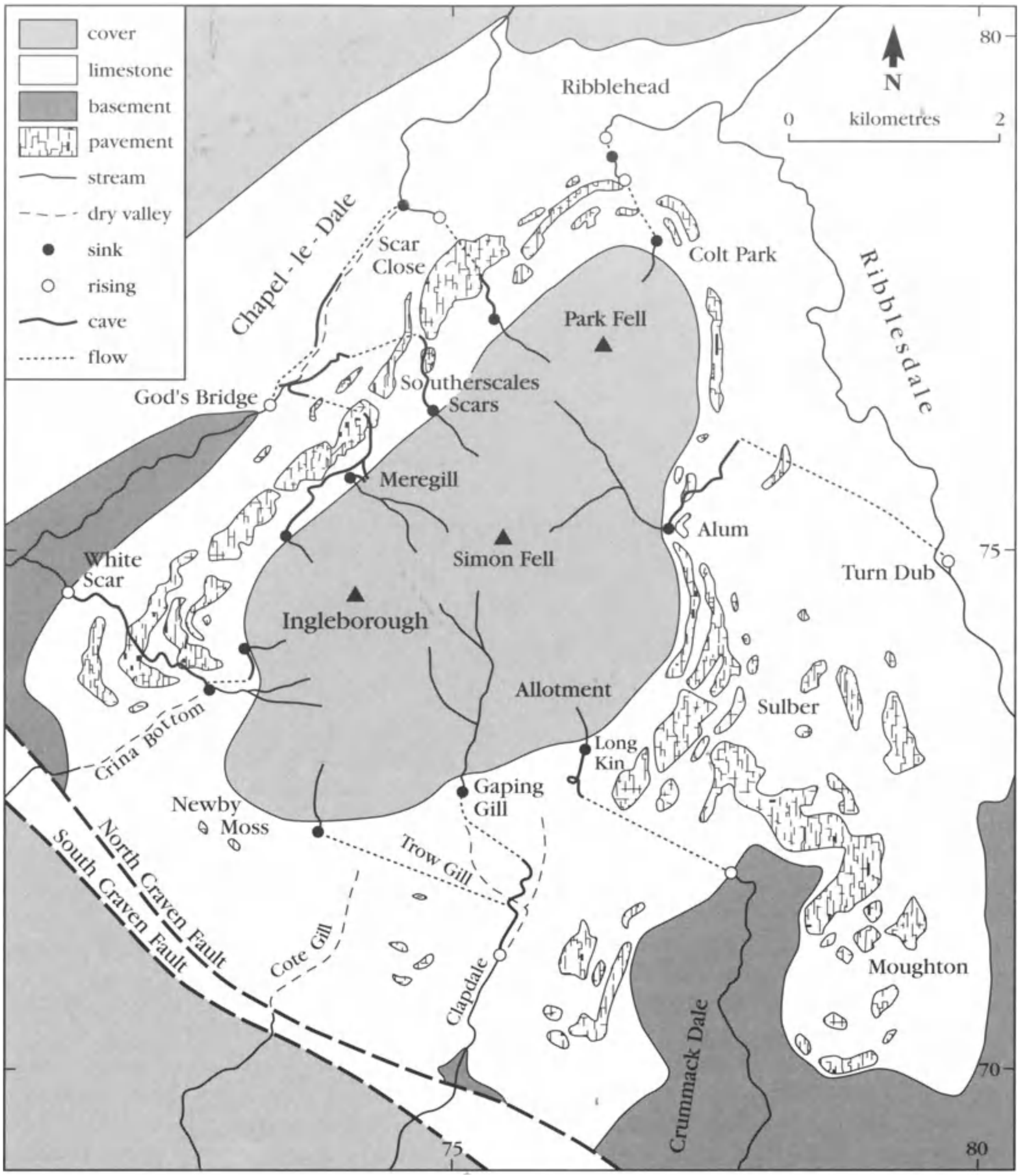


Figure 2.13 Geological map of Ingleborough, with the main areas of limestone pavement, the larger dry valleys and some of the main underground drainage routes. The limestone is the Great Scar Limestone, including the Hawes Limestone. Cover rocks are various clastic units and thin limestones in the Wensleydale Group and the Namurian Millstone Grit Group, and Upper Carboniferous clastics south of the Craven Faults. Basement rocks are Palaeozoic slates and greywackes. The only pavements marked are those of good or excellent quality (as defined by Waltham and Tillotson, 1989).

and retreat phases, the dales were occupied by valley glaciers which deepened the troughs below the limestone benches. A significant distributary ice flow left the Ribblesdale iceway and crossed the eastern benches of Ingleborough to drop into Crummack Dale and Clapdale (Figures 2.13, 2.14). Areas of lesser ice flow became the main zones of glacial deposition, and the thickest till mantles the limestone outcrop on Newby Moss, lying in the protected lee of the Ingleborough summit mass.

The limestone plateau around Ribblehead now carries a splendid drumlin field, left by Devensian ice on the broad upland before it was constrained southwards between the Three Peaks.

Stream sinks, dolines and shakeholes

Ingleborough is commonly cited as a textbook example of karst landscape, due to the huge number of closed depressions on the limestone benches. These include deep potholes, open cave entrances, active sinkholes, blind valleys, large solutional and collapse dolines, drained structural depressions, and the thousands of shakeholes which are subsidence dolines in the till cover.

Streams draining off the summit slopes of Ingleborough sink into the top of the limestone all around the shale outlier (Figure 2.13). More than 250 cave entrances are known, providing access to over 54 km of mapped caves. Some lead into almost level cave passages formed along shale horizons between the upper beds of limestone. Many are vertical potholes, with fluted limestone walls disappearing out of daylight as they reach depths of 10–100 m; the large open shafts of Gaping Gill, Alum Pot and Meregill Hole are just the better known of many impressive shafts. The cave systems fed by these sinks are described below, but they have developed almost radially to resurgences in the floors of all the adjacent dales (Figure 2.13). Some of the larger potholes, including Great Douk Cave, Alum Pot and Gaping Gill, lie out on the limestone benches, and may be close to past positions of the retreating shale margins (Sweeting, 1974; Waltham, 1990).

The few blind valleys on Ingleborough are cut only into the glacial till where streams have found routes into the buried limestone, so none exceeds

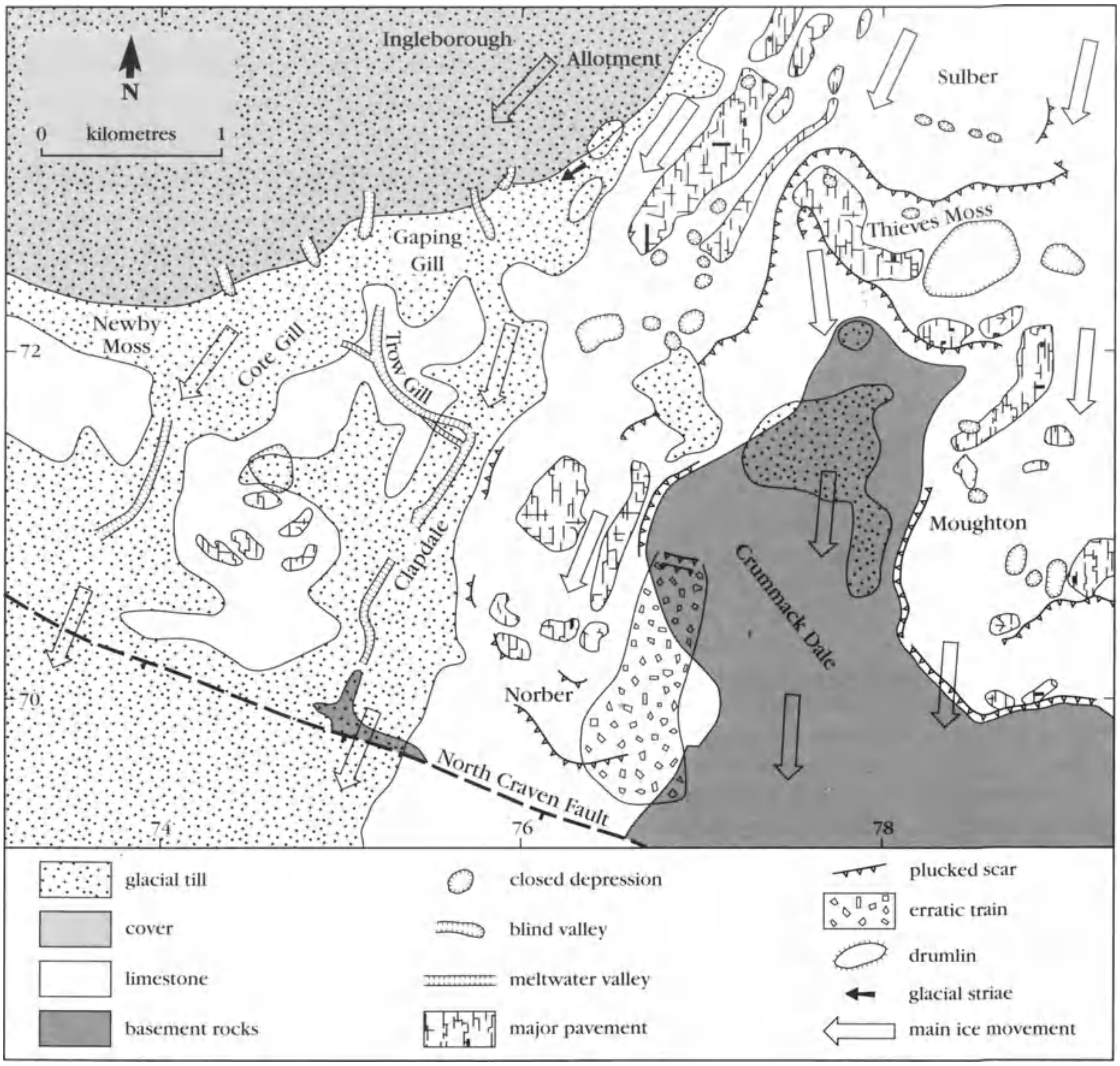


Figure 2.14 Geomorphological map of the southern sector of Ingleborough. The main pavements in the eastern half of the map area were scoured by ice moving down Ribblesdale, while the limestone in the western half is extensively veneered by glacial till deposited in the lee of the Ingleborough summit mass (from Waltham, 1990).

a depth of about 10 m; Gaping Gill is the best known example, and there are others along the buried shale boundary to the east and west (Figure 2.14).

Most of the limestone benches were scoured by Devensian ice, and large solutional dolines have not developed in the short period of postglacial time. The only solutional features larger than the main open shafts are some preglacial depressions now wholly or partly filled with clastic sediment. Braithwaite Wife Hole, on the bench north of Meregill (Figure 2.21) is a conical hole 60 m across and 25 m deep, with sides of slumped till which mask most of the rock profile. Wider but not as deep is the partially filled solutional doline above the west side of Clapdale. A group of dolines on the limestone bench north-west of Crummack Dale (Figure 2.14) have almost level grass-covered floors only about a metre below their rims of bedrock limestone. These appear to be preglacial dolines, filled with till and perhaps truncated by surface lowering, but their subsurface structure has not been investigated; shakeholes within them reveal soil fills a few metres deep with no exposure of rock. These and other, similar large depressions are of largely solutional origin, but collapse modification is almost inevitable; it is known to have occurred in Braithwaite Wife Hole where a tributary cave passage gives access to a zone of collapse reaching 30 m below the floor of the surface depression.

Large, shallow closed depressions on Thieves Moss (Figure 2.14) are structural basins within the limestone, each excavated to a single bedding plane by glaciers. They are not karstic, except that they remain dry due to underground drainage.

Where the limestone benches of Ingleborough are covered with glacial till or any other clastic soils, the ubiquitous feature of the karst is the shakehole. This style of subsidence doline is formed by the soil cover ravelling into the fissures in the buried limestone as rainwater filters through and creates small piping failures. There are around 3000 shakeholes on Ingleborough. Most are 1–10 m across, no deeper than the till thickness, and with sides sloping at 10–40°. The deepest are therefore in the thick till blanket on Newby Moss (Figure 2.14), but the greatest densities of shakeholes lie along the strip of till over the shale boundary along the bench above Chapel-le-Dale. Only a small proportion expose limestone in their floors; some have open shafts or cave entrances, and others are blocked so that they now contain small ponds. The shakeholes have formed and enlarged over the 10 000 years of postglacial time, and they continue to evolve

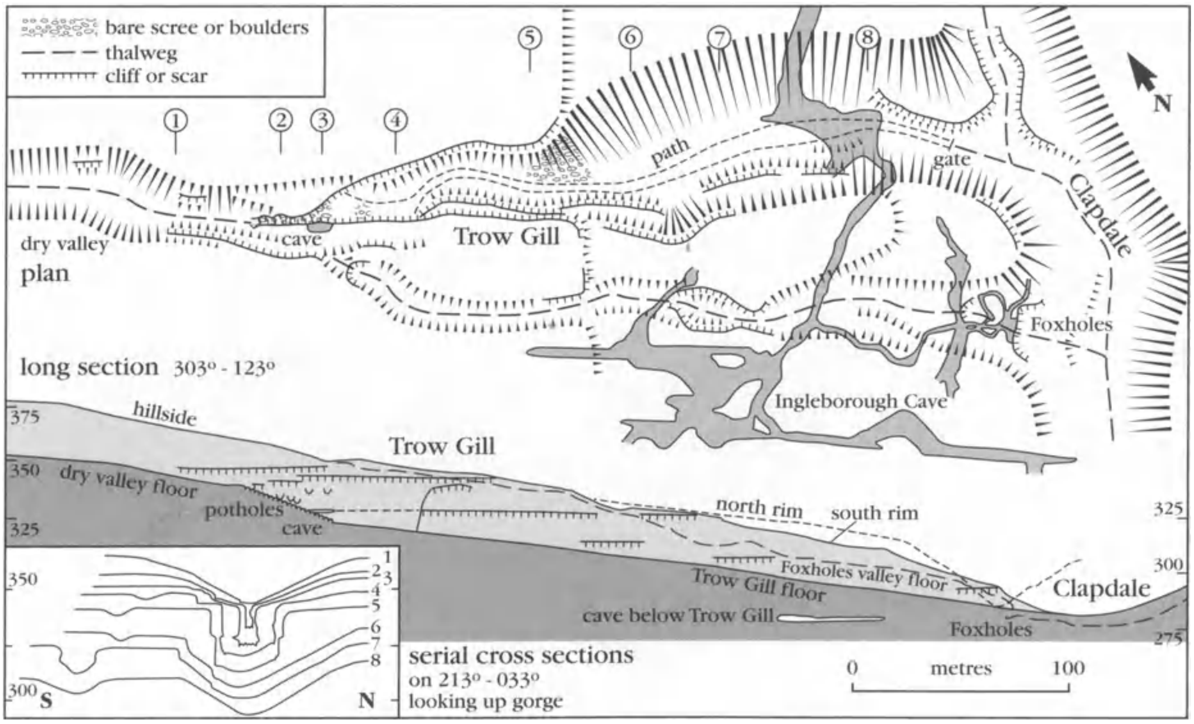


Figure 2.15 Topographic map and projected long profiles of Trow Gill and the underlying caves. Some cave passages have been omitted to improve clarity, and all the caves lie below the level reached on the serial cross-sections; the thalweg down Trow Gill lies along the centreline of the path (from Waltham, 1990).

today. New shakeholes are occasionally recorded, but many more old shakeholes periodically deepen or widen as their soil fills slump into hidden limestone fissures. In 1980, a massive failure enlarged the shakehole containing Marble Pot when hundreds of tons of soil and debris ran into the chamber below and blocked the outlet cave passage (Waltham, 1989).

Dry valleys and gorges

The largest dry valleys on Ingleborough are all cut into the southern slopes (Figure 2.13). Crina Bottom is the longest and deepest, and has a thin alluvial fill along part of its floor; its headwaters have been captured by White Scar Cave, draining beneath the scars to the north-west, but flood flows reach down the normally dry valley to a series of sinks which feed parallel drainage routes to Chapel-le-Dale. Cote Gill is a smaller dry channel, largely cut into the glacial till which fills a broader valley in the bedrock.

Trow Gill is the finest dry valley feature. It forms a trench across the plateau south of Gaping Gill, and then descends 70 m through a narrow rocky gorge less than 3 m wide at its narrowest point (Figure 2.15). The gorge is incised more than 20 m into the strong limestone, and has vertical walls in its upper section. It descends into Clapdale, where it is joined by a dry valley through the retreat moraine in Clapham Bottoms, and it forms the upper reaches of a fluvial channel that is almost continuous down the floor of the glaciated trough of Clapdale. Trow Gill beheads an older dry valley which is less deeply entrenched on its southern side and ends over the low scars that contain the Foxholes cave.

Apart from these four, the only dry valleys on Ingleborough are shallow rocky ravines which descend through some of the limestone scars. Sulber Nick is the longest of a number aligned on fault zones, and others are associated with caves which have captured their drainage.

Limestone pavements and scars

Glaciation has sculpted many of the landforms present on Ingleborough, and the extensive pavements and long scars of bare white limestone dominate many of the landscape vistas. The spectacular limestone pavements extend over nearly 400 ha (Ward and Evans, 1976), and lie on the plateaus and benches which received the full impact of glacial scour by ice moving down both the west and east sides of the Ingleborough summit mass (Figures 2.13, 2.14). The very variable quality and morphology of the Ingleborough pavements were assessed in a survey by Waltham and Tillotson (1989). They found that the pavements of the finest quality all lie in two narrow belts, around Scar Close and between Alum Pot and Thieves Moss, both at the cores of the main pavement zones west and east of Ingleborough.

The Ingleborough pavements are extremely varied, presenting suites of pavement landform types, from great undissected sheets of scoured limestone, to closely fractured linear clints. The most massive pavements lie in two smaller areas, on Scar Close and west of Thieves Moss, and these contain the largest clint blocks, with some of nearly 2 ha on Scar Close. In contrast, the linear pavements, with knife-edge clints and larger blocks with length/width ratios greater than 4, lie mostly in the limestone nearer the Craven Faults, on White Scar, above Clapdale and on southern Moughton (Figure 2.13). These broad patterns of pavement distribution are superimposed on a wealth of detail at individual sites.

On the terraces and benches of Raven Scar, overlooking Chapel-le-Dale, many large clints, particularly near the inner edge of the terraces, have centripetal systems of deep runnels. There are also many lamellar clints with bedding planes etched into the walls of the perimeter grikes; smaller clints in flaggy limestone show flakey weathering which produces more loose debris. Scattered sandstone boulders are glacial erratics.

Some of the finest of Ingleborough's pavements lie below Meregill Hole, across the top of the Southerscales Scars and into Scar Close (Figure 2.16). The exposed limestone bedding is horizontal or dips 1–5° north-east, and many of the clints are more than 10 m across. There is an identifiable sequence of variation across the pavements on each terrace, with a tendency for the large clints to be etched by very smoothly rounded rundkarren near the inner boundaries where strips of glacial till survive. Towards the central sections of the terraces, the larger clints still have deep runnels, many of which are individual, sharply edged rillenkarren not forming dendritic systems. The outer edges of the terraces are broken into smaller clint blocks which cannot support large systems of deep runnels. On the larger clints on Southerscales Scar, some of the large single runnels are tadpole-shaped, and each drains downdip from the remains of a kamenitza.

The main pavement in Scar Close has been pro-

Figure 2.16 The limestone pavements of Southerscales Scars, on the north-western bench of Ingleborough. (Photo: A.C. Waltham.)

tected from grazing sheep since 1960; it is therefore developing a new natural plant cover of mosses, heather, juniper, yew, ferns and flowering shrubs. Very large rectangular clints commonly have a central hump down their length, etched by short rundkarren draining to either side. There are also some extensive, undissected surfaces which carry rich vegetation on islands of remnant till and peat (Gosden, 1968). Dissection of the limestone is by dendritic runnels originating from drainage off these islands, and also off the marginal shale cover, as well as by grikes on the rectangular networks of tectonic joints.

At the head of Ribblesdale, Colt Park contains a stretch of pavement retaining a mature, natural vegetation cover; ash dominates a thick woodland which hides a dense undergrowth with deep moss over a massive pavement with very deep grikes. Borrins Moor Rocks, south of Alum Pot, has some excellent, massive clints with deep rounded runnels. Massive pavements continue to the south along the outcrop of the strong beds of limestone near the top of the Great Scar sequence; they extend west of Thieves Moss, where the largest clints occur, and on to the scars east of Clapdale. Extensive pavements are developed on the lower limestone beds which extend over the gently sloping Moughton plateau (Figure 2.14); there is great variety in the karren morphology, but the clints are generally smaller to the south where the Craven Faults are approached.

To the west, the margin of the shale cover is largely obscured by drift, but a small pavement by Long Kin East Cave (Figure 2.18) is notable for its glacial striae. These are preserved on limestone sealed beneath an impermeable till cover which is slowly retreating. Newly exposed striae are removed by rainfall solution within about 10 years, and the limestone is then etched by new solution runnels (Tiddeman, 1872; Sweeting, 1966, 1974). Further south, the Norber ridge is famous for its perched glacial erratics of greywacke (Figure 2.17). These were derived from outcrops in Crummack Dale and carried obliquely upwards onto the limestone outcrop (Figure 2.14). They now stand on pedestals of limestone which have been protected from corrosion by direct rainfall while the surrounding limestone surface has been lowered by subaerial and subsoil solution. As the protected pedestals are mostly 400-500 mm high, the mean rate of lowering of the exposed surface has been about 30-40 mm/ka since the Devensian ice retreat (Sweeting, 1966). The narrowness of the pedestals, and their incision by solution grooves, may be accounted for by dripwater flowing down the underside of the boulders.

Figure 2.17 Glacial erratic of Silurian greywacke on the Norber bench of southern Ingleborough. The erratic is 2 m across and stands on a plinth of limestone which has been protected from solution by direct rainfall. (Photo: A.C. Waltham.)

The limestone scars are best developed where ice moved obliquely along them or down over them, maximizing the scale of glacial plucking. The scars south of Moughton and Sulber (Figure 2.14) are vertical cliffs, while the Raven Scars along the south side of Chapel-le-Dale form a series of terraces each capped by a stronger bed of limestone. There are few scars along the western side of Ribblesdale, where the ice was moving obliquely up the limestone slope onto the Sulber plateau.

Interpretation

The summit of Ingleborough has been regarded as a remnant of one of a series of old erosion surfaces (Trotter, 1929; Hudson, 1933; King, 1969). The regional drainage pattern was probably initiated on the earliest of these, with rivers originally flowing east, before tectonic warping diverted some of the drainage south. Subsequent denudation was ascribed to four phases by Sweeting (1950); the first was the formation of the 400 m surface, during which the limestone suffered a widespread planation, and this was followed by the 'First Rejuvenation Stage', causing dissection of the 400 m surface, during which underground drainage was initiated; the subsequent 'Dales Stage' was one of relative stability, where the major rivers were able to grade to their base level, after which there was another rejuvenation, forming the master caves of the area. Modern interpretations suggest that large parts of the 400 m surface are stratimorphs, developed on the top of the resistant limestone, and the cave levels ascribed to the successive stages may be a feature of the distribution of inception horizons on shale beds unevenly distributed through the limestone (Waltham, 1970). Though fragments of erosion surfaces can be recognized cutting across the limestone bedding, geological influences on the karstic landforms above and below are strong and tend to mask the effects of past erosion levels. Attempted reconstructions of palaeosurfaces of the late Tertiary and early Quaternary (King, 1969; Clayton, 1981) have little bearing on the modern karst features for which Ingleborough is renowned, other than to outline the broadest geomorphic patterns.

The earliest palaeogeography of Ingleborough for which a tentative reconstruction has been proposed is one with an age of 500 ka, just prior to the Anglian glaciation (Waltham, 1990). The interpreted position of the former shale margin is based on the presence of old cave passages older than 350 ka in Gaping Gill and Newby Moss Cave, and the sites of the major old sinkholes of Alum

Pot, Gaping Gill, Braithwaite Wife Hole and Great Douk Cave. The lack of old caves in the Ribblehead area suggests that the limestone was not exposed there in pre-Anglian times.

Stalagmites from White Scar Cave were formed after passages at successive levels were drained in response to the deepening of Chapel-le-Dale, and include material older than 350 ka from the upper levels (Atkinson *et al.*, 1978; Gascoyne *et al.*, 1983a, b; Gascoyne and Ford, 1984); these indicate a maximum mean rate of incision of 0.2 m/ka over this period. Reconstruction of pre-Anglian valley floor profiles for Chapel-le-Dale, based on the stalagmite dates, suggests that 80-100 m of surface lowering has taken place over the last 400-500 ka, both in the glacial trough and around Ribblehead (Waltham, 1986, 1990). Whether the major surface lowering was by fluvial or glacial processes, in their respective climatic stages, is open to debate. The minimal retreat of the shale margin on the steep slopes of Newby Moss (Waltham, 1990) reflects overall low rates of fluvial surface lowering, and also local protection from glacial excavation in the lee of Ingleborough; the implication is that ice erosion accounted for much of the surface lowering over the larger part of the area, which lacked the protection.

Over the last 500 ka, Ingleborough has been subjected to two or three major glaciations, two long interglacial periods of fluvial environments, and several intervening periglacial phases. In each glacial maximum, ice covered the entire area, perhaps reaching thicknesses of 300 m over Ingleborough, and modified the older surface profiles. Few except the largest surface landforms remain from pre-Devensian times, and many of the modern surface features can be attributed to advance and retreat of the Devensian ice.

The dry valleys of Ingleborough are essentially the product of meltwater erosion, and probably all date largely to periglacial environments during the retreat phase of the Devensian ice. Crina Bottom may have been a marginal channel carrying snowmelt water off Ingleborough beside the Chapel-le-Dale glacier. The smaller features were probably only active for a short period, carrying surface drainage over the exposed frozen limestone until underground capture caused their abandonment as the climate ameliorated. Trow Gill may have a more complex history. The narrow profile of its gorge section has led to speculation that it may be a collapsed cavern (Waltham, 1970). This concept is now not accepted, as there is no positive evidence of collapse, and remnants of stream moulins are visible high on the walls of the narrowest section (Figure 2.15); the gorge is merely the steepest, and therefore most entrenched, section of a fluvially excavated subaerial valley (Waltham, 1990). It was cut when the ground was frozen and impermeable, and lost its stream when karstic drainage was re-established. There is scope for debate as to whether the dry valleys and gorges were carved by subglacial or proglacial meltwater (Pitty *et al.*, 1986). Trow Gill is fed by no channel of significant size from the higher slopes of Ingleborough; its source could have been the snout of an ice lobe from the north-east ending on the limestone bench, or could have been crevasses and glacier moulins in a subglacial situation.

Since the Devensian ice retreated, karstic processes have become dominant once more, largely superimposing texture onto the inherited glacial landscape. Underground drainage has been re-established, enlarging the modern stream caves. The limestone pavements have matured on the stronger beds exposed in the bare rock surfaces left by the glacial retreat. Tectonic fractures were widened by solution of their walls to form the grikes around the clints, and karren runnels drained into them. The rounded form of the rundkarren suggest that much of their development took place beneath a permeable, organic soil cover; this may have been of the type now present in Colt Park, or like that now expanding on Scar Close. Most of the original plant cover was then lost due to artificial clearance of the protecting trees, and sheep grazing has precluded regrowth of anything except grass. The lack of rillenkarren may be due largely to the ubiquitous lichen cover which acts as a substitute for a soil cover in facilitating solution over the ridges between the runnels. The preserved ice striae near Long Kin East Cave demonstrate that there has been almost no limestone solution beneath the impermeable mineral soils of glacial till.

Conclusions

The glaciokarst of Ingleborough constitutes some of Britain's finest limestone landscape. It is of international reputation and importance, and is probably the most used karst teaching example in the country. The dry gorge of Trow Gill, the dry valley of Crina Bottom, the sinkhole of Gaping Gill, the open shaft of Alum Pot, and the perched erratics of Norber are just some of the widely known landforms which are classics of their types. Where the limestone is veneered with glacial till, around 3000 subsidence dolines have formed, and the bare limestone outcrops have nearly 400 ha of spectacular limestone pavements. These include the massive clints of Southerscales Scars and Thieves Moss, and also the protected area of Scar Close with its new plant colonization. The limestone plateaus scoured by ice, the adjacent glaciated troughs and the potholes around the retreating shale margin combine to provide a record of glacial erosion through the late Pleistocene.

INGLEBOROUGH CAVES

Highlights

The caves of Ingleborough include examples of almost every type of cave morphology. They include the finest group of deep potholes and shafts in the country, as well as the largest cave chamber and the highest waterfall in Britain. The caves form a complex and varied system of underground karstic drainage, which is an essential component of the spectacular glaciokarst of the Ingleborough surface landscape.

Introduction

Ingleborough forms a magnificent block of karst between Chapel-le-Dale, Ribblesdale and the Craven Fault scarp (Figure 2.1). It has a plateau, nearly 10 km across, formed on the top surface of the Great Scar Limestone at an altitude of about 400 m, with a central summit mass of shales, sandstones and thin limestones rising to 723 m (Figure 2.13). The summit rocks belong to the Yoredale facies of the Wensleydale Group, except for the Namurian grit cap. The Great Scar Limestone is a strong, fine-grained, massive carbonate with bedding planes mostly 0.5–5.0 m apart and commonly marked by thin shale horizons; it ranges in age from Arundian to Asbian, and is about 200 m thick. Most of the Ingleborough limestone dips 1–3° north, but the south-eastern sector has many shallow folds and local dip variations; it is well jointed and there are many small faults with mainly horizontal displacements. The base of the limestone is a marked unconformity over the folded and faulted, impermeable rocks of the Lower Palaeozoic; ridges and valleys on the buried pre-Carboniferous surface create over 30 m of local relief on the unconformity.

The drainage of Ingleborough is essentially radial. Streams off the shale outlier sink into the limestone all round the margin, and over 250 caves are recorded. Underground drainage from the larger sinks continues the radial pattern, except for deflections of the cave conduits in response to the immediate geology. There is some convergence of underground drainage, most of which feeds to ten major resurgences; most of these lie on or close to the basal unconformity, and the largest are marked on Figure 2.13.

Though the caves of Ingleborough are important components of an important karst, the inhospitable nature of the deep, cold shafts has limited the extent of detailed scientific studies in them. Descriptions of nearly all of the known caves on Ingleborough are given in Brook *et al.* (1991), and are summarized in Waltham (1974a). The more comprehensive descriptions of individual caves and areas include those of White Scar Cave (Waltham, 1977b), the Meregill area (Brook and Crabtree, 1969b), Chapel Beck (Monico, 1995), Alum Pot (Milner, 1972), the Allotment potholes (Booth, 1905; Brodrick, 1905; Griffiths, 1927) and the Gaping Gill area (Brindle, 1949; Patchett, 1953; Glover, 1974; Ford, 1975; Beck, 1984). The geology and geomorphology of the Gaping Gill caves were considered in detail by Glover (1974), and the geomorphology of the Ingleborough karst was reviewed by Waltham and Tillotson (1989) and Waltham (1990). Particular features of the cave geology have been discussed by Halliwell (1979b), Halliwell *et al.* (1975) and Waltham (1970, 1971b, 1977a). The chronology of cave development, based on speleothem dates, and its relationship to landscape evolution has been discussed by Waltham and Harmon (1977), Atkinson *et al.* (1978), Gascoyne and Ford (1984), Gascoyne *et al.* (1983a, b) and Waltham (1986, 1990). The underground drainage was determined by a programme of water tracing by Carter and Dwerryhouse (1904), and further aspects of the cave hydrology were discussed by Pitty (1974), Richardson (1974), Waltham (1977b) and Halliwell (1980).

Description

More than 55 km of cave passages are known within the Ingleborough limestone. The radial

Figure 2.18 Outline map of Ingleborough, with locations of the main caves referred to in the text. Geology as in Figure 2.13.

drainage pattern has caused these to form a series of discrete caves, and there is no integration into large dendritic systems comparable to those of Ease Gill and Kingsdale. Furthermore, the caves within the different sectors around the Ingleborough summit mass have considerable contrasts between them, imposed by local variations in the geology and topography. The influent caves lie all round the perimeter of the shale outlier (Figure 2.18), and fall into groups which are largely defined by their shared resurgences. No long drainage routes have yet been followed from sink to rising, and most known caves reach only from a sink to a sump at the head of a phreatic conduit.

Caves of White Scars

The limestone bench between Chapel-le-Dale and Crina Bottom is named after its extensive, bare pavements and scars (Figure 2.13). Much of it is drained by White Scar Cave, which carries water from sinks in Crina Bottom to the resurgence exit in Chapel-le-Dale (Figure 2.19); it has 6500 m of mapped passages, and is operated as a show cave as far as the Battlefield Chamber. Except for a few high avens, all the known cave is developed in the lowest 30 m of the Great Scar Limestone; Ordovician slates form the floor of the passages to the resurgence and the show cave entrance, and also at a small cascade up the Far Streamway. The main stream emerges from the flooded tubes of the Phreatic Series, and drains over a knickpoint cascade into a fine streamway canyon which is continuous to the cave exit. Deep lakes are ponded behind sediment banks and collapse debris from avens, and there are numerous small inlets. The tube from the Phreatic Series continues as a roof tube over part of Far Streamway, creating a splendid keyhole cross-section; it then turns away to the west where the dry passage is heavily choked. Sleepwalker is an old phreatic tributary, now drained and carrying an underfit stream from choked sinks in Crina Bottom. Just downstream of the Sleepwalker junction, the old phreatic tunnel diverges from the line of the streamway; it continues on the west side for 150 m, beyond which it is choked with sediment. Further north, the Battlefield is a large old chamber whose collapsed floor has been partly undermined by the modern streamway; it is part of an isolated segment of abandoned trunk cave 20 m above stream level. All the abandoned passages contain thick deposits of sand and mud; straw stalactites, stalagmites and flowstone are spectacular in many parts and also in sections of the streamway (Figure 2.20). Stalagmites from the Battlefield and Sleepwalker passages have been dated to over 350 ka (Gascoyne and Ford, 1984), while flowstone from the roof of Far Streamway has an age of 225 ka (Atkinson *et al.*, 1978).

Floodwaters which overflow the Sleepwalker sinks in Crina Bottom go underground at Rantry Hole, and are joined by some drainage from Newby Moss before resurging at Skirwith Cave (Figure 2.18). They pass through old, choked tunnels beneath Crina Bottom and then into the kilometre of small streamway known in Skirwith Cave.

At the head of Crina Bottom, Quaking Pot has a series of narrow, immature, vadose canyons and shafts, aligned along a fault. This drains into a meandering streamway which reaches a depth of 143 m, where it has invaded an older, largely choked, chamber on a fault; the water is next seen in White Scar Cave. The next large sink to the north is Tatham Wife Hole, where a deep, meandering, vadose canyon intercepts the Tatham Wife Fault; the rest of the cave consists of tall, inclined rifts developed along the fault, and it drains to Granite Quarry Rising.

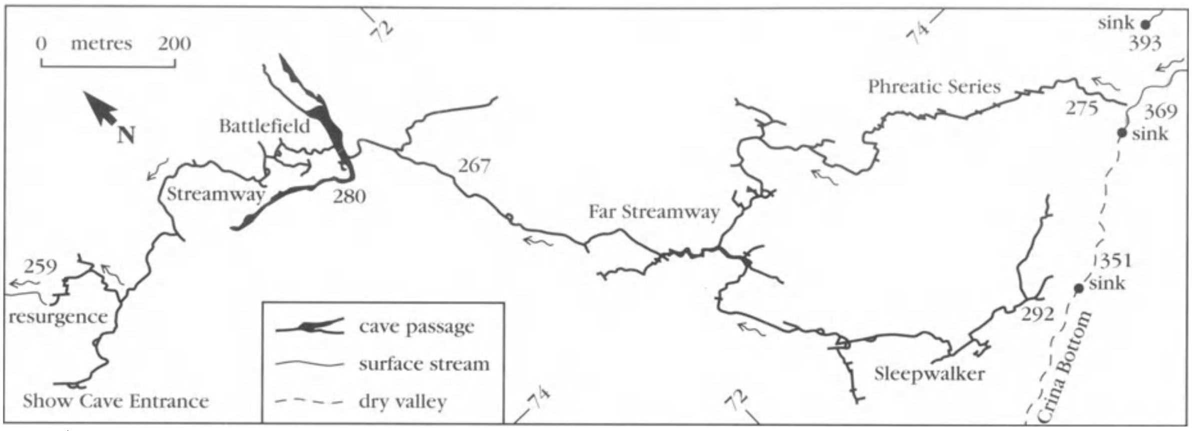


Figure 2.19 Outline map of White Scar Cave (from survey by Happy Wanderers Cave and Pothole Club).

Figure 2.20 Long calcite straw stalactites in the Far Streamway of White Scar Cave. (Photo: A.C. Waltham.)

Caves around Meregill Hole

Meregill Hole takes the largest sinking stream on the Chapel-le-Dale flank of Ingleborough (Figure 2.13). Its entrance is a fault-guided rift 50 m long and 5 m wide, with vertical walls dropping 15 m to The Mere, a perched lake 25 m deep. The lake outlet is into a tall rift passage, which extends south-east under the shale margin and drops 120 m down a series of shafts developed along a fault (Figure 2.21). The stream canyon then drains north, down the dip of its bedding plane roof. At a level of 225 m, this drains into bedding controlled phreatic tubes which rejoin and continue through the lower, flooded passages of Roaring Hole. Black Shiver Pot has a long upper series of low bedding-plane passages, leading to joint-guided cascades and a massive vadose shaft dropping 90 m down a fault. From the foot of the shaft, the streamway follows the bedding planes again, passes through a flooded section, and becomes a tributary to the lower streamway in Meregill Hole. Roaring Hole has a sequence of rifts and canyons descending steeply to join the flooded conduit from Meregill Hole.

Directly above the deep bedding plane caves of Meregill's lower level, a parallel series of caves drain downdip along the bedding planes and shale beds near the top of the limestone. These include Hallam Moss Cave, Sweetwater Hole and Sunset

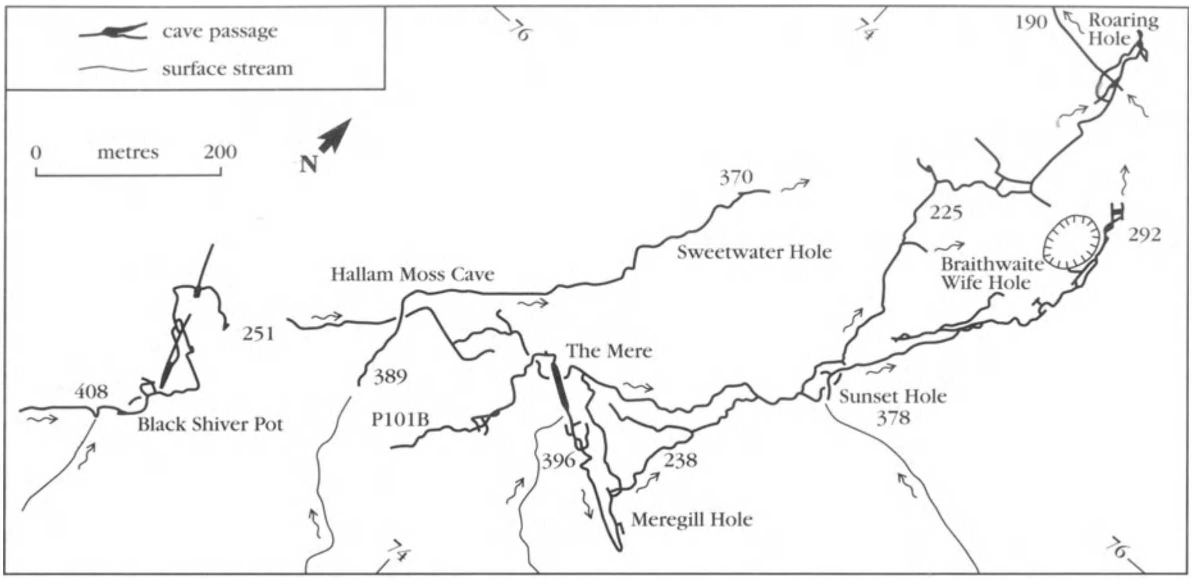


Figure 2.21 Outline map of the cave systems of Meregill Hole; the flooded passage downstream of Roaring Hole is known to continue for another 300 m. Numbers given refer to elevation in metres (from surveys by University of Leeds Speleological Association).

Hole (Figure 2.21). The Sunset stream canyon can be followed into a complex of chambers modified by collapse and fill beneath the large old doline of Braithwaite Wife Hole.

A series of influent caves lies further north along the shale margin (Figure 2.18). Knacker Trapper Hole has narrow stream canyons leading to a large fault-controlled rift which descends to the south-west, against the dip, to a perched sump at a depth of 98 m; some high-level chambers are well decorated with calcite stalactites. Another stream cave drains into the shaft of Hardrawkin Pot. Sinks at Middle Washfold Cave and Southerscales Pot feed long streamways which converge on the large collapsed pothole of Great Douk Cave, where the downstream route is lost in narrow fissures and collapse debris. Keld Bank Sink is one of a series of small, shallow caves which eventually drain to the perched risings at Eller Keld. Round the northern tip of Ingleborough, the caves are young and immature; Gauber Pot feeds Batty Wife Cave through passages too small to be followed.

Caves of Chapel Beck

Chapel Beck crosses the limestone at the head of Chapel-le-Dale and its course is normally dry between the main sink at Haws Gill Wheel and the resurgence at God's Bridge (Figure 2.18). An almost completely flooded cave system has over 4 km of active phreatic conduits, beneath a surface channel which carries only floodwater. Weathercote Cave, Jingle Pot and Hurtle Pot are all shafts in or beside the river bed. In Weathercote Cave, the main flow drops from a high-level bedding cave, down a waterfall and into rifts and collapsed bedding caves descending to sump pools; the other two are normally dry windows into the phreas, whose water surface is at 225 m. The phreatic conduits downstream of Jingle Pot mainly follow the limestone bedding, which rises gently to the south; joints provide the alignment of some sections, cause enlargements on cross rifts, and guide some sections into phreatic loops which reach depths of 30 m.

The main Chapel-le-Dale conduit is known from Jingle Pot to Midge Hole, south of which it may continue under the west bank (Figure 2.22). A complex overflow route lies at shallow depth in the northern part of Joint Hole, where it joins a deeper route from the east carrying water from sinks high on the limestone bench. Chapman's Rising and Meregill Skit are flood outlets from the main submerged trunk routes, which continue to the perennial God's Bridge resurgence.

Caves around Alum Pot

Alum Pot is a massive open shaft, 30 m long, 10 m wide and 70 m deep, developed on a minor fault. From its foot a tall rift follows the fault north into a chamber, with a sump pool in its floor. The flooded passage has been followed for 385 m, along the fault and then north-east at depths of up to 25 m. The resurgence is through the floor of

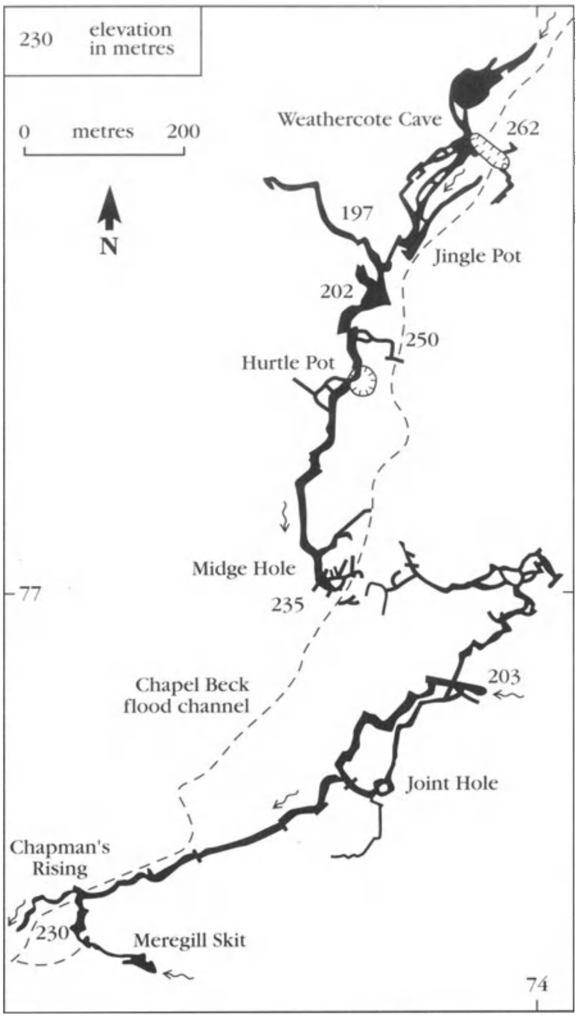


Figure 2.22 Outline map of the mainly flooded cave passages beneath Chapel Beck (from surveys by Cave Diving Group).

Turn Dub, a sediment-choked pool on the far side of the River Ribble (Figure 2.13). Footnaw's Hole is a pool in the alluviated valley floor west of the Ribble, and appears to be a flooded window into the conduit from Alum Pot to Turn Dub; it normally swallows minor local drainage, but it produces large outflows in flood conditions.

Except after heavy rain, little water enters the open shaft of Alum Pot. The main streams of the area drain into youthful vadose caves less than 10 m below the surface. Borrins Moor Cave is a shallow dendritic system fed by several sinks, which drains into the Upper and Lower Long Churn Caves, and then into Diccan Pot (Figure 2.23); the four caves are separated by short unroofed and collapsed sections of the stream course. The main passages are splendid vadose canyons with clean, scalloped walls and floors of pale limestone; these are cut beneath the wide roofs of initial, shallow elliptical openings etched out of thin shale beds (Figure 2.24). They drain downdip until they meet the Alum Pot fault; in Diccan Pot, the active streamway drops 100 m down a series of spectacular waterfall shafts into the lower chamber of Alum Pot.

Washfold Pot lies further north (Figure 2.18) and also drains beneath the river to Turn Dub. A stream sinking at the shale margin flows north-east, down the dip, for more than 300 m in a shallow vadose canyon, before descending a

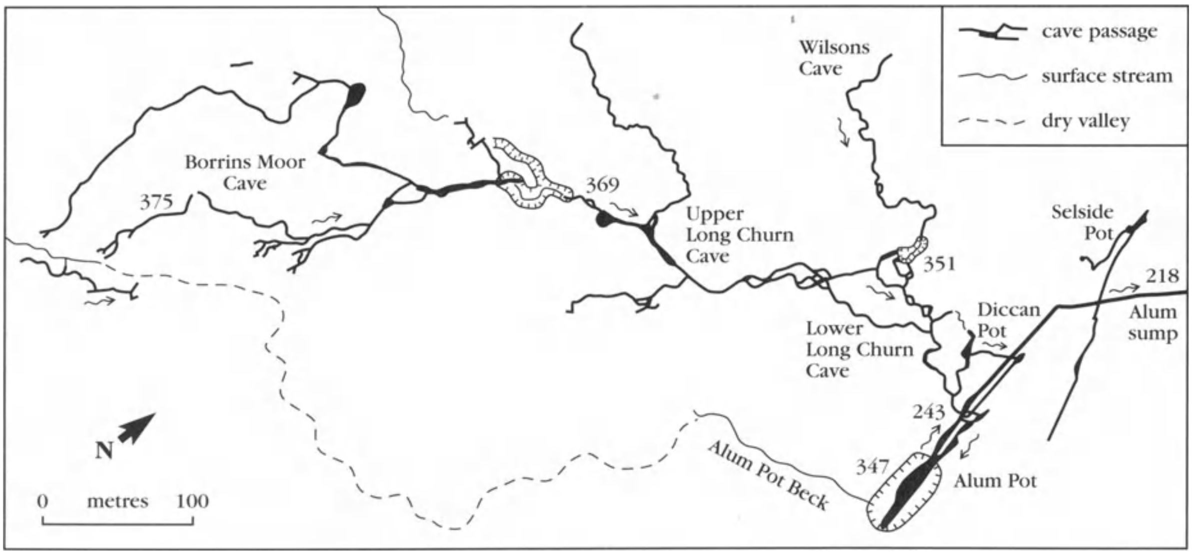


Figure 2.23 Outline map of the Alum Pot cave system; the Alum Pot sump is known to continue for another 220 m beyond the margin of this map (from survey by University of Leeds Speleological Association).

Figure 2.24 Sharply scalloped limestone forms the walls of the vadose canyon cut beneath the bedding plane roof in the streamway of Upper Long Churn Cave. (Photo: A.C. Waltham.)

series of vadose shafts; these are developed obliquely down a major fracture, as far as a flooded shaft 10 m above resurgence level.

Potholes of the Allotment

On the eastern flank of Ingleborough, the Allotment contains a group of deep and spectacular potholes, largely developed on minor faults in the limestone close to the shale margin (Figure 2.18). They all descend steeply to sumps at or close to the level of Austwick Beck Head, their common resurgence on the base of the limestone in Crummack Dale. Flow to the resurgence is obliquely updip, largely along flooded bedding caves.

Nick Pot lies on the Sulber Nick Fault; it has a complex of short entrance rifts which descend 20 m, and then open onto a shaft 10 m in diameter and 100 m deep dropping to a sump. In the adjacent shakehole, Hangman's Hole has a series of rifts and shafts descending the same fault obliquely to the east. Juniper Gulf is a magnificent pothole with a long fissure entrance, deep rifts, massive shafts and fine underground waterfalls, all developed on a vertical fault. Slasher Hole is an adjacent system of narrower rifts on another fault.

Rift Pot has a sequence of deep shafts developed on a brecciated fault zone with slickensides visible on the rift walls. Long Kin East Cave has a long, meandering, vadose canyon with a shale bed roof just below ground level; this ends at a 60 m underground waterfall into the main chamber of Rift Pot. The outlet from the chamber is a series of bedding-plane passages extending east of the fault to a sump perched 14 m above the level of Austwick Beck Head. Also on the Rift Pot fault, Jockey Hole is a dry 75 m deep shaft from daylight, and Lizard Pot is a rift system reaching a depth of 90 m.

Marble Pot lies at the end of a spectacular blind valley cut in thick glacial till. The cave below consists of a series of partially choked rifts and shafts connected by bedding caves, but was blocked in 1980 by a massive collapse of the till slopes rising over the entrance. Marble Sink is entered from an adjacent shakehole, and its very narrow rifts and canyons lead to some tall chambers extensively choked with collapse debris and glaciofluvial fill; these are probably part of the old choked passages below Marble Pot.

Gaping Gill Cave System

Gaping Gill Hole is the best known pothole in Britain, and is the sink for Fell Beck, the largest stream on Ingleborough. Its water falls 110 m down the Main Shaft, forming Britain's highest waterfall. The shaft is about 10 m in diameter, until half way down where it breaks through the roof of the Gaping Gill Main Chamber; this is the largest

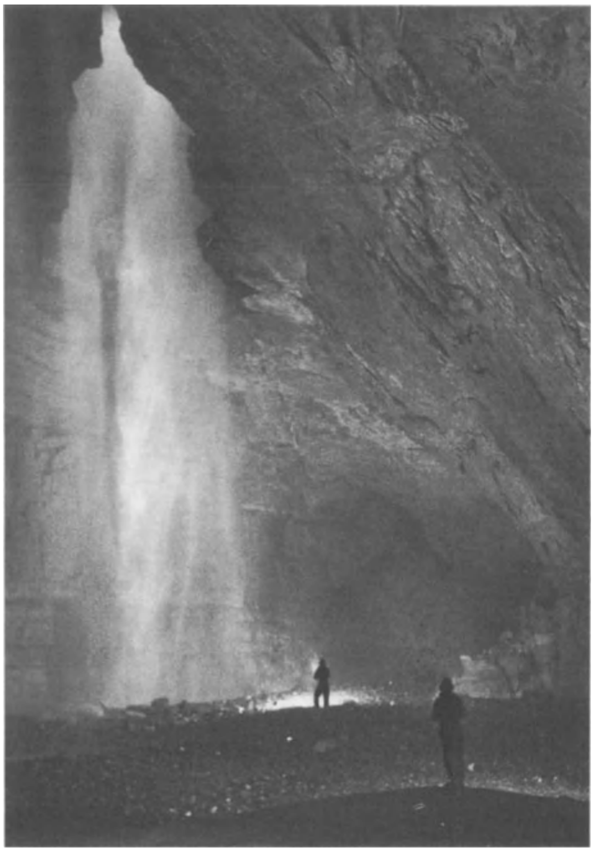


Figure 2.25 The Main Chamber of Gaping Gill, with the 110 m waterfall lit by daylight from the pothole which swallows Fell Beck. (Photo: A.C. Waltham.)

known cave chamber in Britain, 140 m long and 30 m high and wide (Figure 2.25). Under normal flow conditions the water sinks into the floor of the chamber, and is not seen again in the northern part of the cave system, but a network of largely abandoned passages extend to the south and east.

The cave system of Gaping Gill has over 16 km of mapped passages, including inlets from five other entrances and a route through to the old and new resurgences at Ingleborough Cave and Clapham Beck Head (Figure 2.26). The main passages form a low-level distributary system spreading out from the Main Chamber; nearly all are abandoned phreatic tubes, developed into rifts along some of the faults in the limestone. Probably the oldest passage is the highest route, through the large muddy tunnels of East Passage, Mud Hall and part of Car Pot; a branch at a lower level continues through Whitsun Series, and may relate to the tubes of Far Waters. Another old route heads south and joins fault guided rifts and tubes including Sand Cavern; clastic sediment chokes many of the outlets, but they may once have continued through to Mountain Hall. Between these two old trunk routes, Hensler's Passage starts as a remarkable bedding cave; it is over 300 m long, mostly 5 m wide and all less than 1 m high. This leads to a large entrenched vadose canyon, and an abandoned high-level which continues as the Far Country phreatic network.

A complex of shafts and chokes in the Far Country drops to a series of active phreatic tubes carrying the water from the Main Chamber. These drain into the Inauguration Series of Ingleborough Cave, then into partially flooded bedding caves and totally flooded rifts through to the gently descending streamway in Beck Head Stream Cave, and out to the Clapham Beck Head resurgence. One abandoned distributary leads to daylight through the old phreatic tubes of Ingleborough Cave, now accessible as a show cave, and another emerges at Foxholes.

There are few passages above or below the main network which forms a series of levels stepping down only 70 m in total from the Main Chamber to the resurgence. Stream Passage and Disappointment Pots have fine streamways and shafts from active stream sinks. Car, Bar and Flood Entrance Pots are largely abandoned inlet systems. Various shafts reach below the main cave level, but are flooded windows into the phreas that hides the main conduit at an unknown depth for most of its route from the Main Chamber.

Some of the larger chambers and fault rifts have been modified by collapse and are choked with breakdown debris. Many of the old, abandoned, phreatic tubes contain thick clastic deposits, and some sections are well decorated with calcite flowstone and dripstone. Stalagmites from East Passage have been dated to periods of deposition at about 230->350 ka, 37-50 ka and 10-15 ka, and material from Far Country dates from 114-135 ka (Gascoyne *et al.*, 1983a, b; Gascoyne and Ford, 1984). Stalagmite on top of the varved sediments in Sand Cavern is only 800 years old.

Grange Rigg Pot has a small, descending streamway unconnected to Gaping Gill but also draining to Clapham Beck Head. Above it, the blind valley of P5 feeds a shaft system into Grange Rigg, and also a separate small streamway into a complex of abandoned phreatic tunnels (Figure 2.26). Hurnell Moss Pot has a vadose shaft dropping 65 m into a section of very large, ancient, phreatic tunnel along a fault (Figure 2.18).

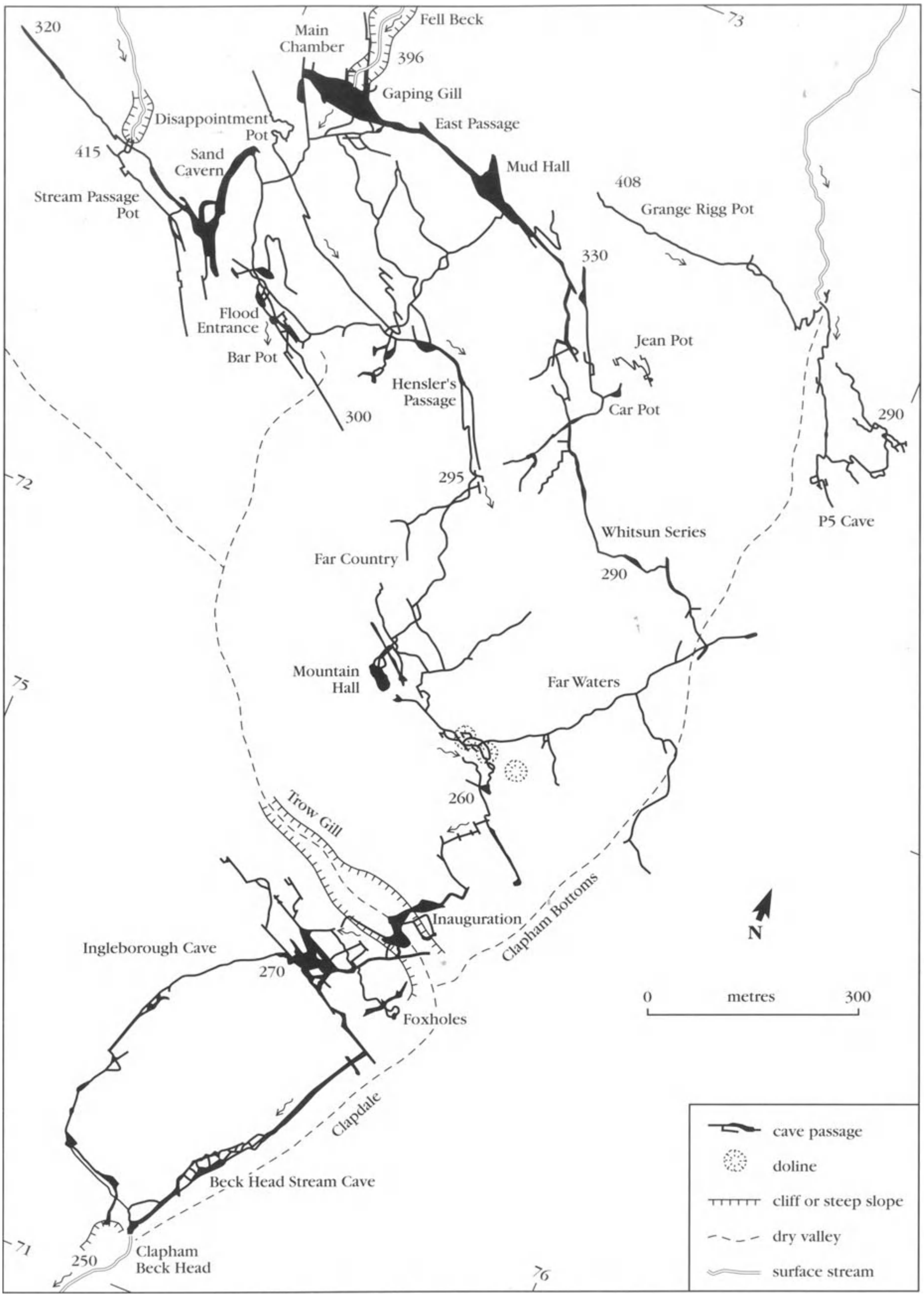


Figure 2.26 Outline map of the Gaping Gill Cave System, including the passages of Ingleborough Cave and Car Pot. Figures given represent elevation in metres. (From surveys by Bradford Pothole Club and many others.)

Potholes of Newby Moss

The south-west corner of Ingleborough contains over 20 caves characterized by deep vertical shafts developed on faults, with almost no horizontal passages (Figure 2.18). Long Kin West is the deepest, and the foot of its second great shaft is choked by breakdown at a depth of 168 m. Grey Wife Hole and Newby Moss Pot contain short lengths of canyon passage, but neither reaches a depth of 100 m. Lying further east on Hurnell Moss, the short passage fragment of Newby Moss Cave contains flowstone dated to over 350 ka (Gascoyne *et al.*, 1983a, b; Gascoyne and Ford 1984). Most water sinking on Newby Moss resurges from Moses Well, but flood flows emerge from Cat Hole (Figure 2.18), and some appears in Ingleborough Cave; sinks west of Long Kin West drain into White Scar Cave.

Interpretation

The cave systems of Ingleborough are many and varied, and between them provide fine examples of almost every feature of underground morphology. Most notable are the many cave features that demonstrate the clear influences of geological guidance.

Stratigraphic control of cave inception and development is evident in both the vadose and phreatic passages in many of the caves. Many bedding planes within the limestone succession contain thin partings of shale, and the downdip vadose streamways guided by these are major elements in many cave systems. Meregill Hole provides the finest example, in the parallel formation of vadose canyons close to the surface and more than 100 m down in the main drain. The Long Churn streamways into Alum Pot provide some of the best and most easily accessible examples of shallow vadose canyon passages in Britain. The only long vadose streamway out to a resurgence is in White Scar Cave, because the limestone basement is exposed in a dale side downdip of the sinks, and there is no high basement ridge to create ponding within the aquifer. The active phreatic tubes below Chapel Beck largely follow bedding horizons updip, and the old drained phreatic tubes in White Scar Cave can be seen to follow the bedding, but also step up joints to change horizons.

Some bedding planes contain no shale, and purely lithological contrasts within the carbonate sequence can provide horizons of cave inception. The most conspicuous of these is the Porcellanous Band, formed by less than 1 m of very fine-grained micritic limestone. Many of the old trunk passages in Gaping Gill lie just above this band, and Hensler's Passage lies directly on top of it, forming the longest and most spectacular bedding cave in Britain.

The buried topography of impermeable rocks, expressed in the local relief on the unconformable base of the Great Scar Limestone, has influenced groundwater flow and cave development in the lowest beds of the karst aquifer. Most of the current main resurgences lie at or close to the basal unconformity, and the flooded caves behind God's Bridge are a feature of both the updip drainage direction and the ponding behind a ridge in the basement rocks (Waltham, 1990). The resurgence from White Scar Cave is on the unconformity, but the cave passages inside are nearly all perched on shales and bedding planes in the limestone above the basement (Waltham, 1977a, b); phreatic inception of nearly all the caves minimizes the gravitational flow of groundwater to the limestone floor except where the aquifer is fully drained close to the valley resurgence site. Where underground drainage has to pass over impermeable basement ridges, horizons which just clear these become the favoured inception lines; a basement ridge beneath Clapham Bottoms probably accounts for the perching of the Clapham Beck Head and Ingleborough Cave passages (Glover, 1974).

The limestone of Ingleborough is broken by numerous minor faults, some of which contain breccia zones up to 1 m thick; there are also many major joints with no recognizable displacement. Both types of fracture influence the morphology of the caves. Vertical shafts are conspicuous on the fractures; they include those of Alum, Nick and Rift Pots, Long Kin West and the Main Shaft of Gaping Gill. A steeply dipping fault also appears to guide the sloping roof of the Gaping Gill Main Chamber where it cuts away south from the Main Shaft. Other caves are developed obliquely down the faults along sequences of rifts and shafts; Juniper Gulf and Meregill Hole are the best examples of this type. Many horizontal passages are formed along fracture/bedding intersections; the lower reaches of Tatham Wife Hole are a clear example, and many of the old conduits in Gaping Gill are guided in this style.

An understanding of the chronology of cave development under Ingleborough is complicated by the contemporaneous development of many discrete conduits; these drained to disparate resurgence sites, each subject to its own cycles of surface lowering and karstic rejuvenation. The Pleistocene climatic cycles and glaciations provided common links, but the geomorphic histories of many caves are independent of their neighbours. A number of older caves may be recognized by their sink entrances away from the retreating shale margin. These include Alum Pot and Great Douk Cave, both 500 m from the shale cover. Newby Moss Cave contains stalagmite dated to >350 ka (Gascoyne and Ford, 1984); it lies almost at the modern shale margin, and is unlikely to have formed beneath the shale cover. These sites therefore indicate greater erosion and slope retreat on the north side of Ingleborough, exposed to the Pleistocene ice flows from the north; the lack of old cave passages in the limestone around Ribblehead may be a consequence of a complete shale cover until removal by glaciers in the late Pleistocene (Waltham, 1990). The clean vadose canyons are the youngest caves, but many of them appear to have origins which predate the last glaciation, as modern rates of passage entrenchment (Gascoyne *et al.*, 1983a) are too low for their development entirely within the last 13 000 years.

The Battlefield passages in White Scar Cave may be remnants of an ancient trunk cave beneath an ancestral Chapel-le-Dale at higher level, perhaps analogous to the modern Chapel Beck caves, and resurging against the North Craven Fault. Stalagmites, dated to over 350 ka, were formed after the phreatic route was largely drained and abandoned. Flowstone from the roof of the main streamway has an age of 225 ka, and must have been above the contemporary resurgence level in Chapel-le-Dale. The levels of the old passages in White Scar Cave indicate a maximum of 0.35 m/ka of valley floor lowering in Chapel-le-Dale, since the caves were drained (Waltham, 1986).

The very old abandoned phreatic passages of the Gaping Gill Cave System, with their thick sediment sequences and stalagmite layers, may prove to contain the most complete record of Pleistocene climatic change and landscape modification in the Yorkshire Dales. However, the evolution of this complex network of abandoned passages has only been assessed in outline (Glover, 1974), and stalagmite dating has only shown that some passages are older than 350 ka (Gascoyne *et al.*, 1983a, b; Gascoyne and Ford 1984), when they are probably much older. Interpretation of the Gaping Gill geomorphology is made more difficult by the strong geological controls, by faults and a thin band of stratigraphical levels, so that past resurgence levels are not easily recognized in a complex profile of deep phreatic loops. Old outlets from the Gaping Gill caves may include Bar Pot (as a vauclusian rising), the roof passage over Mountain Hall (Figure 2.26), a depression in the Foxholes valley (Figure 2.15) and the floors of Clapham Bottoms and Trow Gill, but all are obscured by debris and collapse. Gaping Gill appears to be one of the older caves in the Yorkshire Dales karst, but its history remains largely unknown.

Conclusion

Ingleborough provides Britain's finest example of cavernous karst, as it has not only a spectacular suite of surface landforms but also an excellent range of associated caves. As a teaching site it is without parallel, and many of the individual features are classics of their type. The cave morphology is strongly influenced by many geological factors, and the many deep shafts are the clearest expression of fracture control of cave development. Gaping Gill is among the best known karst landforms in Britain, and its cave system may span a time range longer than any other in the Yorkshire Dales karst.

BIRKWITH CAVES

Highlights

The caves of Birkwith are unusual in that the entire karst drainage remains perched high above the adjacent valley floor. They demonstrate the importance of shale beds as cave inception horizons which can guide a perched conduit within a well fractured limestone aquifer.

Introduction

The Birkwith caves lie in the upper part of the Great Scar Limestone, under the flank of Birkwith Moor where it forms the eastern slopes of Ribblesdale (Figure 2.1). All the streams flowing west from the shale slopes of Birkwith Moor sink into the limestone, but the surface topography and hydrology is complicated by a spectacular drumlin field which lies across the terraced limestone outcrop. This ice-moulded till was left on the retreat of the broad, Devensian glacier which flowed south down Ribblesdale. The Great Scar Limestone dips very gently to the north-west, and is broken by strong joint sets aligned roughly to the NNW and north. All the cave passages are developed within the top 40 m of the Great Scar Limestone, which has a full thickness of over 150 m. The resurgences are perched more than 100 m above the base of the limestone, and the streams issuing from them follow a surface course descending 75 m over limestone and glacial drift to the alluviated floor of Ribblesdale.

The cave passages are all described briefly by Brook *et al.* (1991), and Red Moss Pot was documented by its explorers (Hartley, 1972).

Description

A major cave system with over 4500 m of known passages extends between the sinks of Red Moss Pot and the resurgence at Birkwith Cave (Figure 2.27). Its main feature is the remarkably linear main streamway, draining almost due north along the main fractures from Canal Cavern to the junction inside Birkwith Cave. Over most of its length, this main streamway is a vadose rift passage 2 m wide and up to 10 m high, largely occupied by ponded, slowly moving water. The floor and roof levels vary along the rift, partly under the influence of the low bedding dip; parts are therefore deep canals, and there are four submerged sections of deep flooded fissures. Some parts of the rift above water level are well decorated with calcite speleothems. The cave only descends at two sequences of cascades, one at the upstream end, and one below the Old Ing inlet; the latter cascades are formed where the cave sidesteps between parallel joint fissures.

Access to the main rift streamway is gained via inlet passages from the east. Allogenic streams flow off the shale cover, whose buried boundary is close to the eastern margin of Figure 2.27, and sink where they breach the thinner till between the drumlins. From the entrances of Red Moss Pot, twisting vadose canyons descend over cascades and through small collapse chambers, uniting in a passage which then descends to the main streamway. The longest inlet is well decorated with calcite straws and stalagmite, but cannot be followed to daylight. From its shakehole entrance, Old Ing Cave has a series of cleanly washed vadose canyons and rifts which carry a stream to a

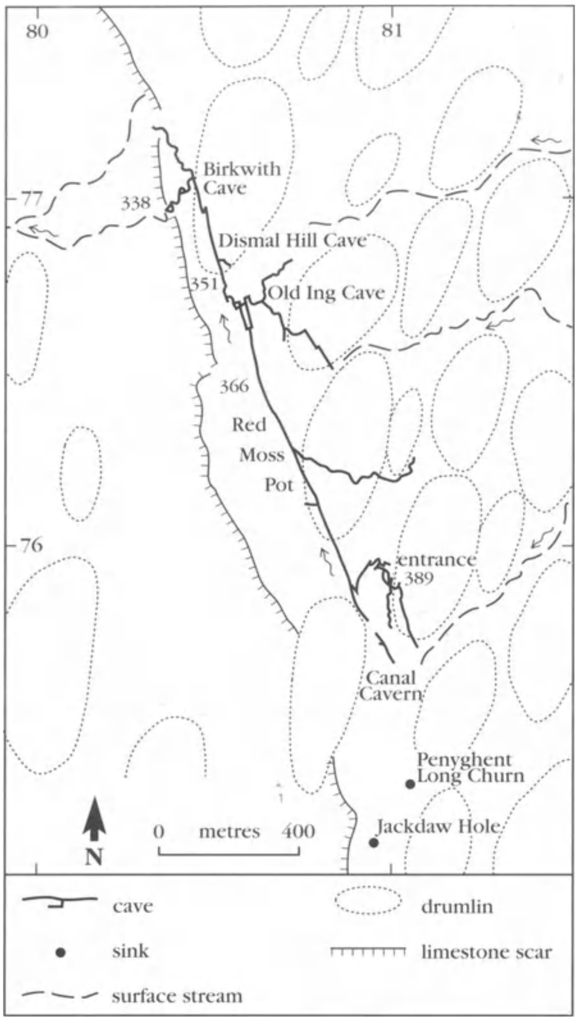


Figure 2.27 Outline map of the caves of Red Moss and Birkwith, draining the limestone bench beneath part of the Ribblesdale drumlin field. Figures given represent elevation in metres (from surveys by Burnley Caving Club and others).

flooded confluence with the Red Moss water. Dismal Hill Cave is another inlet series of narrow rifts and low bedding plane passages. Beyond a last flooded section the main rift streamway drains south-west and descends through sections of bedding-controlled gallery to reach the resurgence exit of Birkwith Cave (Figure 2.27). The abandoned passage north of the junction extends through low bedding planes partly choked with clastic sediment to a small exit below the scar north of the resurgence.

South of Red Moss Pot, Jackdaw Hole is an old choked shaft, and Penyghent Long Churn is an active rift cave which drains south to the New Houses rising. North of Birkwith Cave (and just north of Figure 2.27), Calf Holes is an open waterfall shaft 11 m deep into a fine stream cave, 850 m long; this can be followed through bedding plane passages and then down joint-guided shafts and rifts to a larger streamway out to the Browgill Cave resurgence. Like the Birkwith system, this cave is perched in the top 30 m of the limestone, and its outlet follows a surface course to the floor of Ribblesdale except for the section 15 m long beneath the 2 m thick bed of limestone known as God's Bridge.

Interpretation

Strong geological control is conspicuous in the morphology of the Birkwith caves. They are developed entirely within the top 40 m of the Great Scar Limestone, perched above another 100 m of massive limestone and 75 m above the nearby valley floor, and demonstrate the role of thin shale beds in creating a perched aquifer within a fractured karstic limestone. A narrow zone of joint/shale intersections has provided a single trunk route for cave inception, which has subsequently been enlarged into a mature conduit. The joints extend below the inception horizon, as seen in the deeper canals and flooded sections; this contrasts with sites elsewhere in the Yorkshire Dales, where individual joints fail to breach the shale bands in the upper part of the Great Scar Limestone, causing caves to be perched above continuous shale beds (Waltham, 1971a). Capture of the drainage, into lower fractures and bedding planes, is minimal at present, even though the cave now stands far above a base level determined by potential resurgence sites in the floor of Ribblesdale; some leakage may be taking place to Low Birkwith Cave, a rising 600 m down the beck, whose flow is only partly accounted for by known sinks.

The Birkwith caves are perched and immature. The lower parts of the Great Scar Limestone have no significant cave development where they are traversed by the surface streams flowing from the resurgences. The inlet passages of the cave system all drain from between the drumlins, and there are no known passage terminations at chokes underneath the drumlins. All the evidence points to the caves being comparatively young, and many of the passages may be post-Devensian. The natural drainage of the fracture limestone is west towards the scar edge with a descent into Ribblesdale. The main cave is therefore an anomaly, developed along the joints and downdip until the scar was intersected north of the present resurgence.

Prior to the excavation of Ribblesdale, and also when the glacial trough was occupied by ice, the groundwater drainage would have been towards the lowland to the south. Phreatic initiation of the main rift passage may date back to these conditions, but no morphological evidence of such an early phase has yet been recognized. Abandoned high-level passages in the Old Ing and Birkwith sections are features of local rejuvenation through phreatic uploops, and the outlet passage may have developed in response to a retreat phase of the scar in which the resurgence now lies. Dates of calcite speleothems from the caves may provide evidence for the evolution of both the caves and the local surface morphology, but they are not yet available.

Conclusion

The cave system at Birkwith consists of relatively immature stream caves which clearly demonstrate the significance of shale beds in cave development. Despite their linear, joint-controlled plans, the cave's trunk conduit remains perched at shale horizons far above base level, and drains downdip against the pattern of surface drainage.

BRANTS GILL CATCHMENT CAVES

Highlights

The karst drainage of the Brants Gill catchment is uniquely complex, transmitting water both updip and downdip from widely separate sinks to a single perennial resurgence and two flood risings. Complex series of active and abandoned cave passages include flood overflow routes and a site where the underground drainage has been seen to be diverted in response to evolution of the cave.

Introduction

The caves of the Brants Gill catchment lie beneath the western slopes of Penyghent Hill and Fountains Fell, along the east side of Ribblesdale (Figure 2.1). The Great Scar Limestone forms an extensive outcrop along the middle and lower benches, below outliers which are dominantly shales of the Yoredale facies and form the higher slopes of both Penyghent and Fountains Fell. Lower Palaeozoic basement rocks are exposed in the floor of

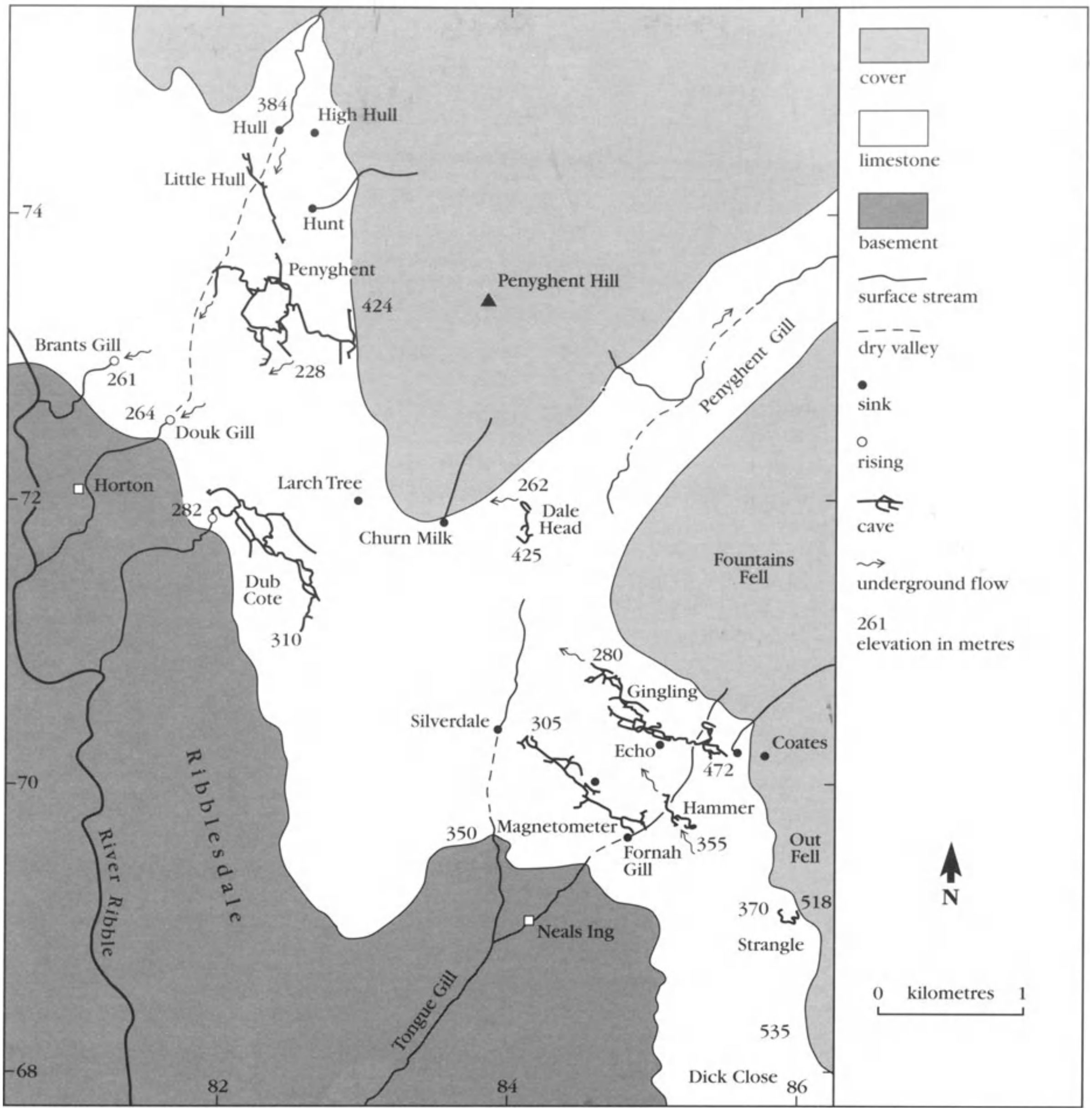


Figure 2.28 Outline map of the cave systems within the Brants Gill catchment. The limestone includes the Great Scar, Hawes and Gayle Limestones. Cover rocks are the shales and higher limestones of the Wensleydale Group. Basement rocks are Palaeozoic slates and greywackes. Only the main cave passages are marked. Flow from all the main sinks is to Brants Gill Head, except when floodwaters emerge from Douk Gill Head (from surveys by University of Leeds Speleological Association, Northern Pennine Club and Cave Diving Group).

Ribblesdale (Figure 2.28). The unconformable base of the limestone, on these greywackes and slates, has considerable relief, reaching a maximum in the lower part of Silverdale where a basement ridge almost 100 m high is exposed. The general dip of the limestone is about 1° north, across many small faults but relatively undisturbed by minor folding. All the upland streams sink on reaching the top of the Great Scar Limestone, and there are more than a dozen, known, major cave systems, together with many smaller caves and choked sinkholes. More than 25 km of passages have been mapped in the catchment. The cave waters all emerge at three risings, of which only Brants Gill Head has a permanent flow. All three risings are located in the basal beds of the limestone sequence, and they lie over 270 m below the highest sinks.

Descriptions of most of the cave passages are given by Brook *et al.* (1991) and were briefly reviewed by Waltham (1974e). The more significant descriptions of the caves by their original explorers include those of Penyghent Pot (Monico, 1989c), Dub Cote Cave (Monico, 1995), Hammer Pot (Batty, 1957; Heys, 1957) and Gingling Hole (Batty, 1967; Monico, 1995).

Description

Although linked into a single integrated underground drainage system, only a comparatively small proportion of the total length of the cave passages that must exist within the Brants Gill catchment has yet been explored. The known caves form two main groups of sinks, on Penyghent Hill and Fountains Fell, and a group of resurgence caves (Figure 2.28).

Influent caves on Penyghent Hill

Hull Pot is the largest single sink feeding the Brants Gill system (Figure 2.28). Its entrance is a huge quarry-like pothole, 90 m long and 20 m wide and deep, aligned WNW–ESE along a minor fault. In flood, a river cascades into it from the north, but under normal flow conditions the water sinks in the riverbed at various points up to 100 m upstream. Beneath the eastern end of the open pot, a series of immature bedding plane passages and vadose shafts descend steeply to a depth of 60 m, where the main route on appears to be obscured by a massive collapse pile which may be continuous up to the floor of the surface pothole. High Hull Pot is a series of vadose rifts and shafts descending joints for 65 m to a choke of collapsed blocks (Figure 2.27).

Hunt Pot has a classic rift entrance 25 m long and 4 m wide, developed on a minor north–south fault (Figure 2.30). A stream from the east cascades down it, to enter a perched sump in a choked and very narrow rift 61 m below the surface and 75 m above the Brants Gill Head rising (Figure 2.28).

Little Hull Pot has a meandering vadose canyon which leads to two vadose shafts dropping 60 m to a rift passage developed along a minor fault (Figure 2.28). The stream flows south-east, obliquely against the dip, into a series of phreatic loops up to 12 m deep; these are perched 24 m above the Brants Gill Head rising, as they drain into the lower passages of Penyghent Pot. Probably all of the water from Hunt Pot, High Hull Pot and Little Hull Pot, and at least part of the flow from Hull Pot, passes through Penyghent Pot on its route to Brants Gill Head.

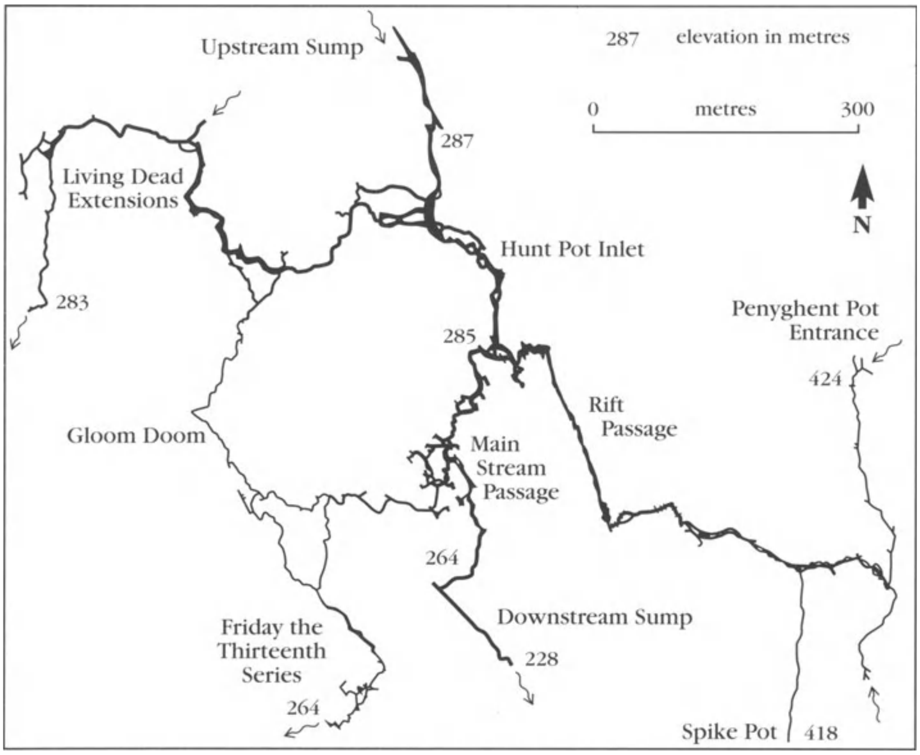


Figure 2.29 Outline map of the cave passages in Penyghent Pot; the Spike Pot entrance passage is only sketched in (from survey by University of Leeds Speleological Association).

Penyghent Pot is the most extensive cave system yet known in the catchment, with more than 5500 m of mapped passages (Figure 2.29). From the small sink at the entrance, the stream follows a low, bedding-guided, vadose canyon southwards, before it turns to the west and is joined by a tributary inlet from the Spike Pot entrance. Rift Passage is 1-3 m wide, developed on a minor fault, with the stream descending 90 m in its series of cascades and vadose shafts. At the foot of the rift, the stream turns west into bedding plane caves, and drops into a major streamway draining from north to south. Upstream this forms the Hunt Pot Inlet, where a rejuvenated phreatic passage leads to a series of flooded rifts, which have been explored to a point very close to similar flooded rifts in Little Hull Pot. Downstream, the same phreatic tube is interrupted by two waterfall shafts in its descent to a sump at the level of Brants Gill Head; the flooded passage continues beyond a depth of 36 m. West of the Main Stream Passage, an extensive and complex series of caves are nearly all developed on a shale bed at an altitude of about 283 m, 22 m above the Brants Gill Head resurgence. Most of these passages are small, half-flooded, phreatic tubes with limited vadose modification, but they intersect some short sections of larger, and probably older, phreatic tubes whose continuations are blocked by sediment. Drainage from Hunt Pot, Little Hull Pot and High Hull Pot used to enter the Hunt Pot inlet and flow into the Main Stream Passage but, since 1986, it has taken a new route through the Living Dead Extensions to the west. A separate small stream flows through the Friday the Thirteenth Series and drops down shafts to a sump at the same level as the main downstream sump.

South of Penyghent Pot there are several small caves at sinks just below the shale margin. Larch Tree Hole and Churn Milk Hole are large old sinks choked with boulders; Churn Milk swallows a small stream which resurges at Brants Gill Head (Figure 2.28). Dale Head Pot is the only cave on this section of fell yet explored to great depth (Figure 2.28). A small, meandering vadose canyon follows the dip to the north, to the head of a series of vadose shafts and narrow rifts, developed on a series of closely spaced joints; these enter a sump at the same level as Brants Gill Head.

Influent caves on Fountains Fell

The two largest streams on Fountains Fell converge on the Gingling Wet Sinks, where they drain through a series of constricted bedding planes and rifts to a perched sump at a depth of only 52 m (Figure 2.31). An adjacent sinkhole contains the entrance to Gingling Hole, which contains over 5200 m of known passages, reaching a depth of 192 m (Figure 2.31). Vadose canyons and small shafts descend to a series of large chambers and old phreatic tunnels at a depth of 50 m; these include Stalactite Chamber and Fool's Paradise, both exceptionally well decorated with calcite straws and dripstone (Figure 2.32). The main passage continues north down a series of narrow rifts and shafts to a junction, where two sets of rifts and shafts descend north-west-south-east joints in parallel for nearly 90 m to sumps at the same level. The sump in the southern rift is a perched phreatic loop which is an inlet to the complex of passages forming the Fountains Fell Main Drain. A large stream flows to the north-west, emerging from a flooded link from the Wet Sinks, and finally flowing into the remote Terminal Sump, about 20 m above the level of Brants Gill Head. The main passages are phreatic tubes 2-3 m in diameter and vadose canyons 2-5 m deep. Small

Figure 2.30 Hunt Pot with a small stream tumbling 27 m down a vertical, solutionally enlarged fissure. (Photo: A.C. Waltham.)

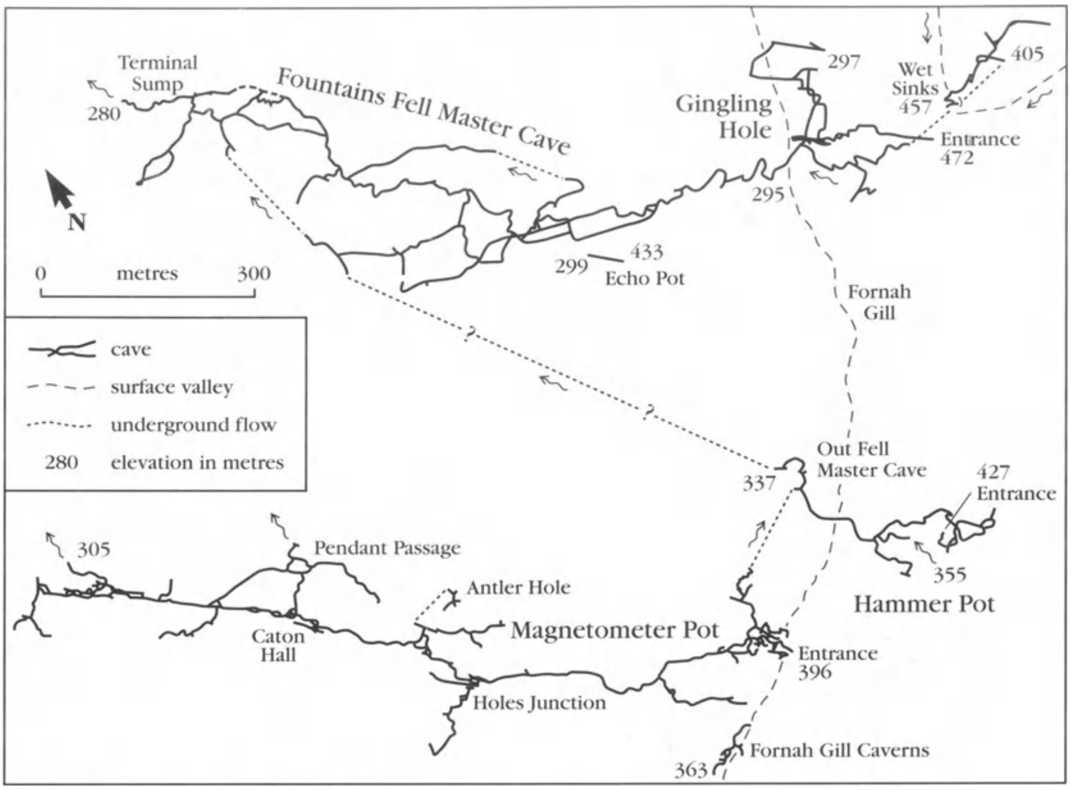


Figure 2.31 Outline map of the main caves of Fountains Fell; the drainage links into the Fountains Fell Master Cave are not proven (from surveys by University of Leeds Speleological Association, Northern Pennine Club and Cave Diving Group).

cascades break the gentle overall gradient, and some sections of tube remain flooded through shallow loops. The largest inlet appears to carry the water from Hammer Pot; a series of abandoned passages, partly choked with collapse and clastic sediments, lies between the two streamways and provides dry by-passes to the flooded sections.

Hammer Pot has a small streamway in its entrance series, joining a much larger passage at depth (Figure 2.31). From the entrance, tightly meandering, vadose canyons, mostly less than 0.5 m wide but about 5 m high, link wider shafts in the descent to a low bedding plane cave which emerges in the Out Fell Master Cave. The large stream in this rises from a deep flooded shaft, and can be followed downstream for 200 m to a waterfall into a chamber just before a sump, which probably drains into the Fountains Fell Master Cave (Figure 2.31).

Magnetometer Pot has a complex entrance series of canyons and shafts, which probably drain east to Hammer Pot, but also intersect an abandoned phreatic cave heading north-west (Figure 2.31). The old trunk passage as far as Caton Hall is up to 5 m in diameter, and there are numerous smaller side passages. Many of these are choked with sediment, but there are at least 20 sumps preventing progress along inlet and outlet passages. Although Magnetometer Pot is largely abandoned beyond its entrance series, the relict galleries transmit large flows in flood conditions.

The large stream in the lower passage of Hammer Pot appears to be the drainage from the many small sinks on Out Fell and Dick Close (Figure 2.28). Few of these can be followed beyond their short entrance shafts. Strangle Pot is the notable exception. It has two sections of phreatic passage, guided by joints along almost level shale beds, incised by small vadose trenches and broken by seven small shafts; these lead to the head of deep rifts which descend more than 70 m. The lower rift is a massive old feature with a vertical wall of boulders opposite the stream entry. Its impassably narrow outlet is about 15 m above the likely destination of the water, in Hammer Pot.

Resurgence caves

There are three separate risings within the Brants Gill catchment (Figure 2.28). All lie close to the base of the limestone, but regional dip and the

Figure 2.32 Fool's Paradise in Gingling Hole - a beautifully decorated phreatic tube and entrenched vadose canyon. (Photo: J.C. Cunningham.)

relief on the unconformity give them an altitude range of 21 m. Under normal flow conditions all water from the influent caves of Penyghent and Fountains Fell resurges at Brants Gill Head. The water emerges from narrow fissures between blocks in a large collapse zone, and it is significant that flood flows from Brants Gill Head never rise above about double the base flow. In flood conditions Douk Gill Head, 500 m to the south-east and 3 m higher in altitude, becomes active; normally dry, this produces flows of more than 0.5 m^3/s in wet weather. Narrow beddings and large dropped blocks prevent access to the extensive cave system which lies behind Douk Gill Head. Dub Cote Cave is the third resurgence, lying further south and updip. Normally this cave carries only a tiny stream, but it acts as a major resurgence under extreme flood conditions. Over 4000 m of passages have been mapped behind the sumps which guard the entrance (Figure 2.28). The main stream route along the north side of the system is largely flooded through active phreatic tubes. A second small stream flows in small vadose canyons through the main series to the south; this consists largely of abandoned phreatic tubes and bedding plane caves, all partly choked by clastic sediment.

Interpretation

The Brants Gill Head catchment contains a major karst drainage system linking numerous sinks with one permanent rising and two flood overflow risings. The hydrology is complex, with convergent and divergent conduits, together with numerous perched phreatic loops influenced by a variety of geological factors.

On the western side of Penyghent Hill, all the drainage from Hunt Pot, High Hull Pot and Little Hull Pot enters the lower reaches of Penyghent Pot, where it flows into an extensive series of passages developed on a single shale horizon about 22 m above the Brants Gill Head resurgence. Perched sumps between these influent caves and Penyghent Pot have been created by the drainage against the northerly dip. The Main Stream Passage of Penyghent Pot is the only long section of cave which can be followed below this shale horizon, to enter a phreatic loop down a flooded shaft and then probably largely up the dip of bedding caves to the resurgence. Continuations of the known passages on the higher level are all choked with sediment, and it is likely that the main water flows diverge to the various flood resurgences due to obstructions within passages on this shale horizon. Some of the Hull Pot water may also flow through Penyghent Pot, but the remainder takes a separate, unknown, route to Brants Gill Head. The ephemeral nature of karstic drainage routes was seen in 1986 when the Hunt Pot Inlet in Penyghent Pot suddenly dried up; the water found a new route through the Living Dead passages further to the west. Such individual events are likely to be related to the creation or removal of sediment blockages, as solutional modification of the limestone is a much slower process.

The hydrological relationships of the three risings clearly show that their distributary drainage is active only in flood conditions, and also provide fine examples of intermittent flow in karstic conduits. Douk Gill is clearly the overflow passage for Brants Gill Head, and must diverge from the base flow conduit downstream of the confluence of the Penyghent and Fountains Fell drainage. The relatively constant flow rate at Brants Gill Head is due to constriction of its conduit downstream of the divergence. Dub Cote Cave floods rarely but rapidly, and appears to be a higher overflow route that probably only extends off the Fountains Fell conduit.

The sinking streams of Fountains Fell all drain to Brants Gill Head, taking an underground route of up to 6 km rather than the shorter westerly course to the base of the limestone in Ribblesdale. Past and present drainage routes descend to depth in the limestone, and then head obliquely and gently downdip in a direction about 20° from the strike, to pass behind the high basement ridge exposed in Silverdale. The plan positions and levels of each of these routes were then controlled by the locations of the contemporary sinks and their rising. The older route lay from sinks on a shale cover margin west of its present position, through Magnetometer Pot and out to Dub Cote Cave, where the modern cave exit has been slightly truncated by surface lowering (Figure 2.28). The later route developed from new sinks, on a shale margin retreating towards the east, to a lower resurgence at Brants Gill Head in a deepened Ribblesdale. This created the main passages of Gingling Hole on a course again at a low stratigraphical level in the limestone, parallel to, 40 m below, and nearly 1 km further downdip to the north-east from the Magnetometer route.

On both phases of development of the Fountains Fell drainage routes, a deep vadose zone appears to have been established rapidly in the cavernous limestone. There are no known traces of abandoned caves on a graded profile associated with a sloping water table commensurate with an immature karst aquifer. The route from the Gingling Wet Sinks descends 160 m in its first 100 m of plan length, and then falls only 40 m in the following 5 km to the resurgence. Though the lower passages are phreatic, and the palaeo-water tables have not yet been recognized from the cave morphology, the steep initial profile of the conduit indicates largely vadose drainage to a deep water table. The abandoned passages in both Gingling and Magnetometer are largely at levels only a few metres above the active drains, and represent minor adjustments and rejuvenations as the cave matured towards a graded profile established by its resurgence level.

Though stratigraphical features have provided the greatest guides to cave inception, the joint patterns have exerted influence on the details of the cave morphology. The deep shafts of Gingling Hole and Dale Head Pot are all aligned on tectonic fractures, and rift passages have developed along the faults in Little Hull and Penyghent Pots, both with and against the dip. The entrances of Hull and Hunt Pots are classic examples of fault control, and the former has enlarged by the progressive solution and collapse of narrow limestone blocks between closely spaced fractures.

The sequence of development of the caves draining the northern sector of Penyghent Hill is not clear. Douk Gill Cave appears to be an older resurgence, now truncated by valley deepening and relegated to an overflow role. The extensive passage development at around the 283 m level in Penyghent Pot suggests that there may be an abandoned rising hidden beneath the soil profile at a level higher than Douk Gill; an old link to the Dub Cote Cave outlet could exist, but no evidence of it has been found. The influent caves contain various calcite and clastic sediments which indicate that the passages are old, and the caves appear to have developed through the interglacial stages of the Pleistocene. No sediments, from either the Penyghent or Fountains Fell caves have yet been dated, and further comment on the evolution of both the caves and the surface topography is therefore speculative.

Conclusion

The caves of the Brants Gill Head catchment form a large dendritic system of karstic conduits which gather water from widely scattered sinks, and drain to one or more resurgences depending on flow conditions. The known caves reveal the rerouting of floodwaters in a karst better than any other site in Britain. Many of the cave passages which must exist within this catchment are yet to be discovered, but it is clear that their morphology has been influenced by the stratigraphy and fracturing of the limestone and also by a ridge in the impermeable basement.

PIKEDAW CALAMINE CAVERNS

Highlights

Pikedaw Calamine Caverns is the larger of two accessible caves in the limestone of the Malham area with secondary minerals of base metals deposited on the cave walls.

Introduction

Pikedaw Calamine Caverns lie beneath the limestone plateau west of Malham Cove (Figure 2.33). An isolated segment of relict phreatic cave contains just over 1000 m of mapped passage, now only accessible via a 25 m mine shaft. The mining activity was reviewed by Simpson (1967), and the cave passages are described by Brook *et al.* (1991).

Description

The mine shaft from the surface enters through the roof of a chamber from which three passages radiate. A phreatic passage up to 10 m wide and 5m high extends over 200 m west and then south to end in a choke. South and east of the entrance shaft, large and small phreatic tunnels radiate and each extend about 100 m to chokes; there is also a small stream passage which can be followed to an upstream sump. All the large relict passages contain thick sand and mud deposits, some of which have been removed by miners. In some chambers there are thin green and blue wall coatings of secondary minerals of zinc, copper and lead. The abandoned tunnels are largely horizontal and lie about 250 m above the resurgence of Malham Cove Rising; the destination of the modern, underfit stream is unknown.

Interpretation

Pikedaw Calamine Caverns are the only abandoned, high-level, phreatic cave passages of any significant length known in the Malham area. Their great elevation above the present resurgence indicates that they are of considerable age; they long predate the evolution of Malham Cove and the other major features of the local karst. The phreatic tunnels appear to be fragments of major relict conduits, but there is no indication of where the sinks and resurgence were located; they were probably close to contemporary boundaries of the cover rocks which have since been stripped away.

Within the Yorkshire Pennines, the hydrated carbonates of zinc, copper and lead are only known in the Calamine Caverns and in the Grollit, a small cave 500 m to the north-east from which miners have also removed the mineralized sediment. The secondary mineralization may have been due to the redistribution by solution of metal ions from primary sulphides in hydrothermal veins, during the phreatic phase of the caves' history (Raistrick, 1938, 1954); this has implications for the potential generation of sulphuric acid, which could have played a significant role in the processes of karstic solution.

Conclusion

Pikedaw Calamine Caverns are a fragment of an ancient phreatic system which must relate to a former drainage and topography. They are distinguished by the coatings of secondary base metal carbonates on the walls of the ancient phreatic caves.

MALHAM COVE AND GORDALE SCAR

Highlights

Malham Cove and Gordale Scar are two of Britain's best known karst landforms. Malham Cove is a spectacular amphitheatre of limestone cliffs which in part represents a dry waterfall, while the neighbouring Gordale Scar is a fine gorge still carrying a small stream. The area is also renowned for its dry valleys, limestone pavements, tufa deposits and underground drainage patterns.

Introduction

The spectacular, concave, crescentic limestone cliff of Malham Cove and the impressive limestone gorge of Gordale Scar are located on the southern margins of the Yorkshire Dales karst plateau (Figure 2.33). Both are impressive examples of fluviokarstic landforms which demonstrate the intertwining of fluvial knick point retreat, glacial scour and meltwater excavation with karstic processes. The area has many other features of note including the fine Watlowes dry valley above Malham Cove, the stream sinks at the southern end of Malham Tarn, which feed to the Malham Cove and Aire Head risings, and the massive tufa deposits in Gordale Scar. Limestone pavements are prominent features of the plateau landscape, and those at the top of Malham Cove are especially well known for their fine morphology. The underground drainage of the area is a classic demonstration of the complex nature of limestone hydrology, with convergent and divergent flow patterns. The area contains some of Britain's finest glaciokarst, much of which is clearly related to the local geological structure.

The geomorphology of the area is summarized by Sweeting (1974) and Clayton (1981) and features in numerous textbooks. The denudation chronology of the area was deduced by Sweeting (1950), while regional denudation studies were carried out by Hudson (1933) and King (1969). Moisley (1955) describes many of the other karstic feature in the area while Pitty *et al.* (1986) discuss the formation of Malham Cove and Gordale Scar. A descriptive account of the geology of the Malham area is outlined by Arthurton *et al.* (1988), O'Connor (1964) and Shaw, J. (1983). The hydrogeology of the area was investigated by dye tracing (Smith and Atkinson, 1977) and by stable isotope analysis (Brown *et al.*, 1986). The Malham area has been the focus for limestone surface solution studies, such as those of Sweeting (1966) and Trudgill (1985b), while its many glacial features were studied by Clark (1967). The tufa deposits of Gordale Scar were discussed by Pentecost (1981).

Description

The limestone geology at Malham is complicated by the major facies variations across the line of the Middle Craven Fault, which was active in Carboniferous times when it formed the boundary between a shallow submarine shelf to the north and a subsiding basin to the south. The central outcrop of massive limestone belongs to the Malham Formation, the upper part of the Great Scar Limestone; it is bounded by the North and Middle Craven Faults. The same beds form the higher ground north of the North Craven Fault, where erosion levels have reached just low enough to expose the basement of impermeable Silurian siltstones, on which most of Malham Tarn lies. A block of reef limestone, contemporaneous with the bedded Malham Formation, lies just south of the Middle Craven Fault and is overlain unconformably by the Upper Bowland Shales (Arthurton *et al.*, 1988). Within the shale sequence, a thin basinal limestone crops out at Aire Head (Figure 2.33) and has a buried contact with the reef limestones to provide a routeway for the unseen, underground karst drainage. Glacial till is significant only on the lowland areas to the south, and there is an extensive glaciofluvial kame complex across the limestone plateau along the line of the North Craven Fault (Clayton, 1981). Malham Cove and Gordale Scar both lie about 500 m north of the Middle Craven Fault.

Malham Cove

The Cove is a massive, concave crescentic, vertical cliff cut into the Great Scar Limestone (Figure 2.34). Its rim is 80 m above the Malham Cove Rising, where water resurges at the foot of the cliff, and the curved walls extend over 100 m on each side of the resurgence. Above the Cove, the Watlowes dry valley extends up past Comb Scar, but loses most of its depth before its head zone on the plateau below Malham Tarn. The valley is entrenched up to 50 m deep into the massive horizontal limestone, more resistant bands of which form scars along the valley sides. Malham Tarn covers 62 ha, most of which is less than 3 m in depth, and owes its existence to the small inlier of impermeable Silurian rocks to the north of the North Craven Fault (Figure 2.33). Water from the Tarn flows out at its southern end and sinks shortly after crossing the fault onto the limestone; a number of choked fissures lie along the streambed, various of which are active at different times and stages. After exceptional rainfall, the sinks may overflow so that the dry waterfall at Comb Scar becomes temporarily active; the water then soaks away in the floor of Watlowes. On rare occasions, water gathered from seepage springs and direct rainfall in the lower Watlowes can cascade over the Cove. Continuous surface flow, from Tarn to Cove, is recorded in the past, but not in this century (Halliwell, 1979a). West of the Tarn, Smelt Mill Beck also sinks shortly after crossing the fault.

The hydrology of the Malham area is commonly regarded as a fine example of the unseen complexities of underground drainage in limestone. There are two major sinks: Water Sinks fed by the outflow of Malham Tarn, and the smaller Smelt Mill Beck sink to the west. There are two resurgences: Malham Cove Rising at the foot of the Cove, and Aire Head Rising south of Malham village. Dye tracing has proved that water from both sinks flows to both resurgences (O'Connor *et al.*, 1974; Smith and Atkinson, 1977). Most of the water from Malham Tarn resurges at Aire Head, taking 13–24 hours, while a smaller amount resurges at Malham Cove, taking 24–29 hours to arrive. The waters of Smelt Mill Beck arrive at the Cove in 2–7 hours, and at Aire Head in 6–10 hours. Both the sinks and Aire head are choked by

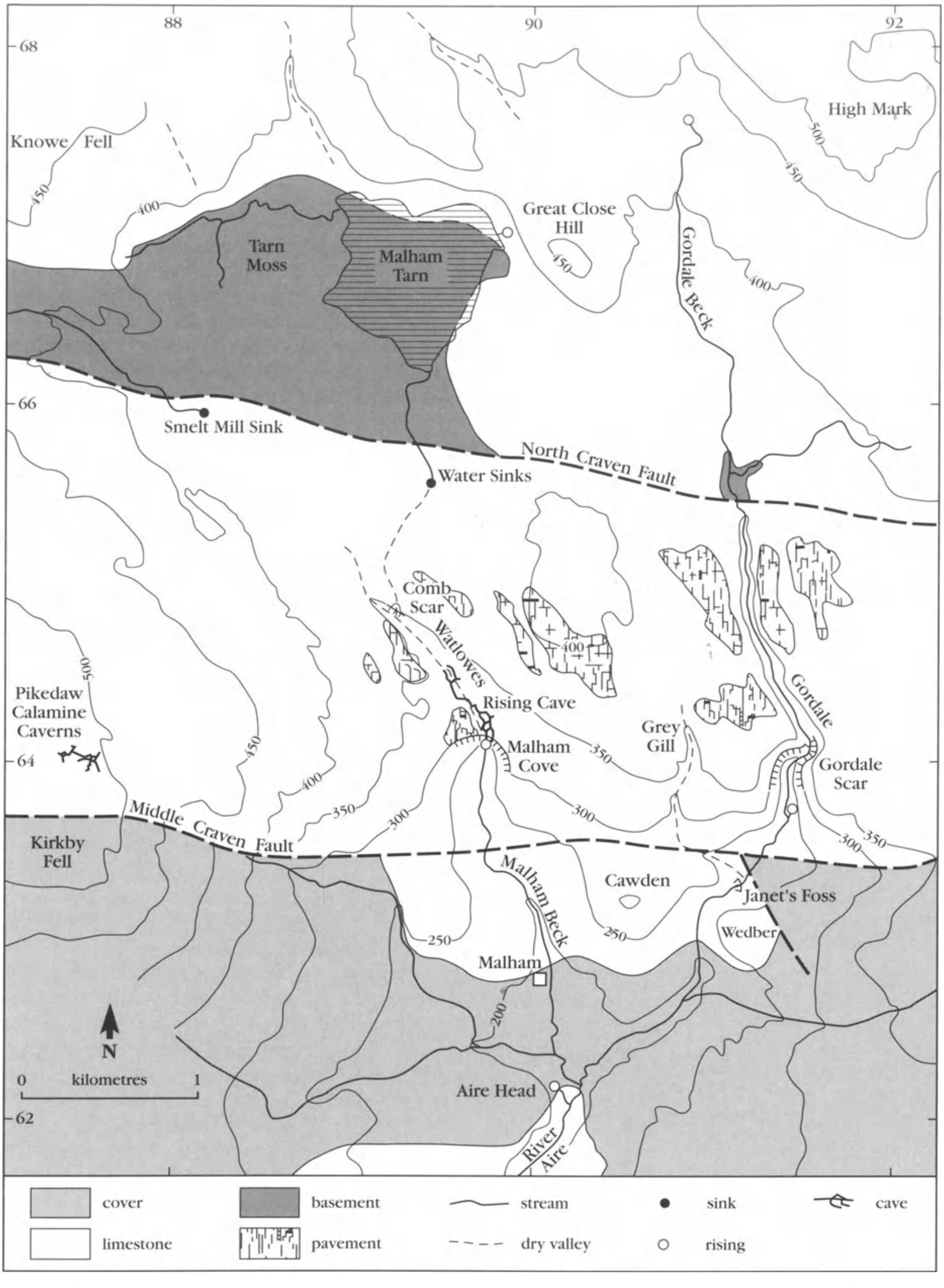


Figure 2.33 Geological map of the area around Malham Cove and Gordale Scar. The limestone at Aire Head is a thin basinal facies, distinct from the reef and shelf limestones north of Malham village. Cover rocks are Bowland Shales. Basement rocks are Silurian siltstones. There are many minor faults, mostly orientated NW–SE between the North and Middle Craven Faults. Only the main areas of well formed limestone pavement are marked.

Figure 2.34 The vertical limestone cliffs, 70–80 m high, of Malham Cove. (Photo: A.C. Waltham.)

boulders that prevent any access. At the Cove Rising, the water flows up the very gentle dip in low, wide and totally flooded bedding plane caves, which have been mapped for 600 m behind the Cove (Figure 2.33). Somewhere in the unexplored zone beneath Comb Scar, a complex of caves and partially choked fissures allows the waters to merge, diverge and overflow into separate flood routes.

Gordale Scar

Two kilometres east of Malham Cove, Gordale Scar is a deep and narrow gorge emerging through the headwall of a massive limestone amphitheatre (Figure 2.35). The cliffs are higher than those of Malham Cove, though they are more broken and less vertical. The gorge cut into them is the most spectacular feature, and it descends steeply from the foot of Gordale via two waterfalls; the upper fall emerges from beneath a rock arch, effectively a short cave passage breaching a high rib of limestone bounded by minor faults. This route through the eyehole has only existed since 1730. Before then, the thin rock rib may have been intact, or the eyehole may have been blocked by coarse debris; the ravine behind was choked with sediment, and the stream cascaded through a slot just north-west of its present route. Gordale is entrenched across the limestone plateau parallel, and in similar style, to the Watlowes valley, but it has a narrower, box canyon profile, and, unlike the Watlowes, it always carries a surface stream. The waters of Gordale Beck rise between Great Close Scar and High Mark, and then flow across the impermeable Silurian inlier, across the North Craven Fault, through the box canyon and into the gorge.

Gordale Beck is saturated with calcium carbonate, and deposits a considerable amount of tufa, especially in the lower reaches. Here, there are extensive tufa banks and terraces, many a little way above the present stream level. The upper waterfall within the Scar gorge is actively building an apron of algae-shrouded tufa where it lands below its hole in the rock rib. A similar apron of tufa stands on the right bank of the cascade, where the stream landed from the slot through the rock rib until the change of course in 1730. The lower falls, formed over a partially re-eroded bed of strongly banded tufa, cascade onto a flatter gravel floored section, where the flow is aug-

Figure 2.35 The limestone cliffs of Gordale Scar. The tufa waterfalls are lost in the shadows in the meltwater gorge which opens into the glacially excavated amphitheatre. (Photo: A.C. Waltham.)

mented by a number of springs fed by percolation and also by small sinks higher in Gordale. Downstream of the Scar, Janet's Foss is another small waterfall over a fine, moss-covered, screen of massive tufa (Figure 2.33).

The interfluves between Gordale and Watlowes have areas of excellent limestone pavements. Rundkarren and kluftkarren are well developed on many beds separated by low ice-plucked scars. The single best known pavement is that at the top of Malham Cove. It is a classic of its type, even though it is now unnaturally polished by the passage of boots and shoes. Kluftkarren grykes form a rectilinear grid, and the clints are scored by deep rundkarren runnels which are deeper and more rounded close to where they can be traced beneath a cover of organic soil.

Interpretation

The origins of both Malham Cove and Gordale Scar have long histories of debate and speculation, widely mentioned in textbooks but little documented in research literature. Popular concepts of the Cove as a dry waterfall and the Scar as a collapsed cave contain only fragments of the more complex histories of development (Clayton, 1981; Pitty *et al.*, 1986; Waltham and Davies, 1987).

The morphologies of Malham Cove and Gordale Scar have many similarities (Figures 2.33 and 2.36); both have large rock amphitheatres upstream of the Middle Craven Fault and downstream of gently graded, narrow, fluvial valleys. The main contrast lies in the steeper gradient of Gordale, its permanent stream, and the incision of its gorge at the Scar. At each site, the amphitheatre and the valley appear to have separate origins.

Malham Cove is conspicuously wider than the Watlowes valley at its head. It was excavated largely beneath an ice sheet moving south over the scarp of the Middle Craven Fault. A concentration of ice flow in an iceway, probably guided by an earlier and smaller fluvial feature, formed much of the Cove by locally enhanced plucking and wall retreat. The iceway has no expression on the plateau above the Cove; the Watlowes arrives obliquely from the north-west. Ice overdeepened the wide valley below the Cove to leave a reverse gradient on the bedrock floor (Figure 2.36); the rock hump has now been entrenched by the post-glacial stream, but is clearly recognizable where the stream turns briefly to the east. The Gordale amphitheatre was formed by similar glacial head-wall retreat, and it too has no valley on the plateau above.

Both the Watlowes valley and Gordale, above the Scar, are clearly fluvial features. Their excavation was largely by seasonal meltwater when ground ice reduced the permeability of the cav-

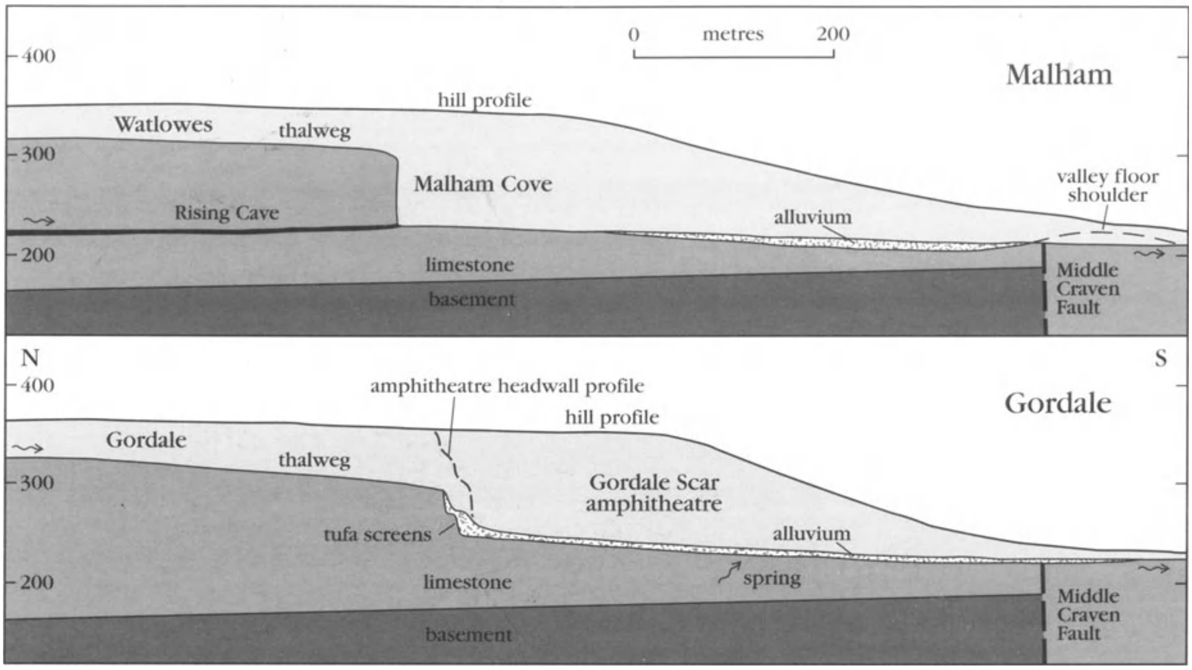


Figure 2.36 Long profiles through Malham Cove and Gordale Scar showing the prominent thalweg steps which have retreated from the Middle Craven Fault scarp.

ernous limestone. This took place subglacially at first, and then downstream of glacier snouts as the ice retreated; it is difficult to estimate the relative contributions of the subglacial and proglacial flow phases. Gordale is Britain's finest example of a meltwater channel cut into limestone under periglacial conditions, with a morphology closely comparable to modern examples in Arctic Canada (Smith, D.I., 1972); Watlowes is more degraded due to its more complete abandonment following underground capture of its drainage (Figure 2.37).

Pitty *et al.* (1986) suggested that Watlowes valley and the Cove were cut by repeated jökulhlaup outbursts of subglacial waters draining from the Malham Tarn basin; current groundwater temperatures and documented Devensian cooling indicate the likelihood of water ponding beneath a melting ice sheet in the Malham basin until it was released in the style of the jökulhlaup floods seen in Iceland today. They also suggested that these floods may have contributed to the scale of excavation at the site of a Cove waterfall. Alternatively, jökulhlaup events may have enhanced wall retreat at the Cove beneath an ice cover. In Holocene times, the steepness of the Cove cliffs has been maintained and assisted by spring sapping at the base, with associated undercutting and periodic collapse.

The meltwater incision of the Gordale Scar gorge may be ascribed to more rapid excavation along the line of minor faults, which are not present at Malham Cove. Solution of the limestone along and between fissures beneath the ravine bed created cavities which were subsequently exposed, and thereby accelerated downcutting in advance of wall retreat. One cave survives to form the eyehole above the top waterfall; cavern collapse is only a minor contributory feature in the gorge excavation by waterfall retreat. Pitty *et al.* (1986) point out that abrasive scour by sediment particles would have accelerated incision and knick point retreat at Gordale Scar, but was lacking in the Watlowes, where the Tarn basin was an effective sediment trap.

Grey Gill is a steep narrow limestone gorge in the slope between Malham and Gordale (Figure 2.33). It demonstrates the form of a gorge cut by meltwater into the fault scarp away from any larger, glacially excavated headscar.

The surface stream in Gordale is something of an enigma. Moisley (1955) surmised that carbonate precipitation may be more important here than solution, and drift deposits may have blocked former sinks. Springs at the foot of the Scar are fed largely by percolation water, and there are no cave passages comparable in size with those beneath the Watlowes. Moisley (1955) drew attention to a series of lacustrine deposits upstream of the Scar, through which the stream has cut leaving a series of terraces. He suggests that they were laid down in lakes held back by tufa dams or

Figure 2.37 Watlowes, the dry valley excavated by meltwater which leads down to the top of Malham Cove. (Photo: A.C. Waltham.)

moraine barriers. The unusually extensive tufa dams and screens of Gordale Scar were described by Pentecost (1981) as 'probably the best example of a tufa depositing stream in the British Isles'. The formation of tufa is aided by the presence of certain mosses and algae, which extract carbon dioxide and therefore cause precipitation from spring water saturated with calcite. This is aided by an increase in water turbulence, which is why many active tufa screens in Gordale are situated in the lower steeper section.

Dye-tracing in the Malham area has had a long history. The first unsuccessful attempts were made around 1870 (Tate, 1879) using a variety of crude tracers, and successful flood pulse traces were conducted in about 1879. Further traces in 1899, used both flood pulse methods and chemical tracers (Howarth *et al.*, 1900) to demonstrate the major links between Smelt Mill Beck and Malham Cove, and between Malham Tarn and the Aire Head Risings (Figure 2.33). A review of these early experiments is presented in Smith and Atkinson (1977), who then conducted rigorous dye-tracing experiments using *Lycopodium* spores, Rhodamine WT and flood pulse methods. These confirmed the underground links, and indicated that the two sinking streams joined before bifurcating to go to the two risings. The high velocity of underground flow suggested that a conduit drainage system existed, as opposed to a diffuse fissure network. Analysis of the flood pulses implied that most of the passage between Water Sinks and Aire Head is phreatic. Dye budgeting at varying discharges established that a greater proportion of the Tarn water emerges at the Cove risings under low flow than under high flow conditions; this suggests that a bottleneck in the cave system behind the Cove restricts the discharge at high flow thus diverting a greater percentage of the flood water to Aire Head. The waters of Malham Tarn also have a distinctive label in their stable oxygen isotope profile, which can be clearly identified in the Aire Head springwaters (Brown *et al.*, 1986).

The geochronology of the Malham area is poorly constrained. The broad relief was probably established during the late Tertiary, either by differential erosion or differential uplift across the Craven Fault system (Clayton, 1981). Both Clayton (1987) and Sweeting (1950) recognized erosion surfaces at 600 and 400 m, although in many places these erosion levels may be confused with stratimorphic surfaces (Waltham, 1970, 1990). Clayton (1981) advocated a Miocene age for his 600 m surface and an early Quaternary age for his 400 m surface. Glaciation has since modified the landscape, although when most of the valley incision took place is unclear. Short-term erosion rates of 0.1-0.6 m/ka were calculated for the Malham area by Trudgill (1985b) using limestone pill data; he showed that the composition of the soils and drift influences differential erosion, as the most rapid erosion occurs under acid soils. Many of the landforms seen today were probably formed during the Devensian glaciation (Clark, 1967; Clayton, 1981).

The valleys of Watlowes and Gordale were cut by meltwater, probably first beneath and then beyond the snout of a waning Devensian glacier (Waltham and Davies, 1987). The tufa deposits are all almost certainly Holocene in age (Pentecost, 1981).

Conclusions

The area bounded by the Middle and North Craven Faults of the Malham district has long been recognized to include some of the country's best developed glaciokarst landforms and shows a spectacular relationship to the regional geological structure. Malham Cove and Gordale Scar especially provide two excellent examples of glaciokarstic landforms, both formed in part by erosional retreat from a fault scarp. Malham Cove is a unique feature, partly an old waterfall, partly a glacial step, whose origins are complex and highly debated. Gordale Scar is a spectacular karst gorge entrenched between very high limestone cliffs. It was cut largely by meltwater; its active and fossil tufa deposits are the best exposed and among the most massive in Britain. The karst also contains many other impressive features, including the Watlowes dry valley and several expanses of well developed limestone pavement. The Malham drainage system is a classic example of the complexity of karstic hydrogeology.

HIGH MARK

Highlights

The dolines and closed basins of the High Mark plateau, east of Malham Tarn, are some of the best developed in Britain. The series of large closed basins, dolines and dry valleys constitute one of the country's finest examples of polygonal karst.

Introduction

The high ground of High Mark, between Littondale and Malham Tarn, forms a complex dissected plateau. Its surface is broken by closed depressions, dolines, rocky scars and shallow dry valleys. These include some of the largest closed basins in Britain. It constitutes an area of fine polygonal karst, and appears to be one of the most mature karst landscapes in Britain. It therefore has important implications for reconstructing the early evolution of this part of the Yorkshire Dales. The Great Scar Limestone reaches its maximum elevation in this area, which would have been one of the first areas of limestone to be exposed in the Yorkshire Dales, and it seems to have escaped the worst erosional effects of the Pleistocene glaciations.

The dolines of High Mark have been the subject of much discussion (Moisley, 1955; Clayton, 1966, 1981; O'Connor *et al.*, 1974; Sweeting, 1966, 1974). Trudgill (1985b) studied limestone solution rates in the area. The regional geology is documented by O'Connor (1964), Waltham (1974a), Shaw, J. (1983) and Arthurton *et al.* (1988).

Description

The dolines and closed basins of High Mark occupy the highest parts of the limestone outcrop a few kilometres north-east of Malham Tarn, at altitudes around 500 m. They are developed on a continuous succession of Carboniferous limestones; these are the Gordale Limestone from the upper part of the Great Scar, and the Hawes, Gayle and Hardraw Scar Limestones from the Wensleydale Group (Arthurton *et al.*, 1988). A very small outlier of Wensleydale Group shale and sandstone survives at the summit of Parson's Pulpit (Figure. 2.38).

The complex dissected plateau of High Mark includes over 20 large closed depressions. Most lie in two clusters: a group of 11 on Parson's Pulpit and Clapham High Mark, and another group of 9 on Proctor High Mark (Figure 2.38). Dendritic dry valleys lie between and around these clusters, but within them the entire land surface consists of basins, and forms two areas of polygonal karst. Within the depressions there are numerous small dolines and shakeholes in the cover soils and drift.

The area is devoid of surface drainage except after heavy rain, when temporary lakes may form in some of the outlying depressions around Middle House Farm (Clayton, 1981). The largest depression, south of Parson's Pulpit, is 800 m across and almost 100 m deep (Figure 2.39). Most of the depressions are saucer shaped, though some have deeper profiles with low scars on their steeper sides. Most of the dolines lie in undisturbed strata, though some west of Parson's Pulpit have more linear shapes dictated by small faults. They have varying amounts of fill whose depth is unknown, and none has exposed bedrock in its

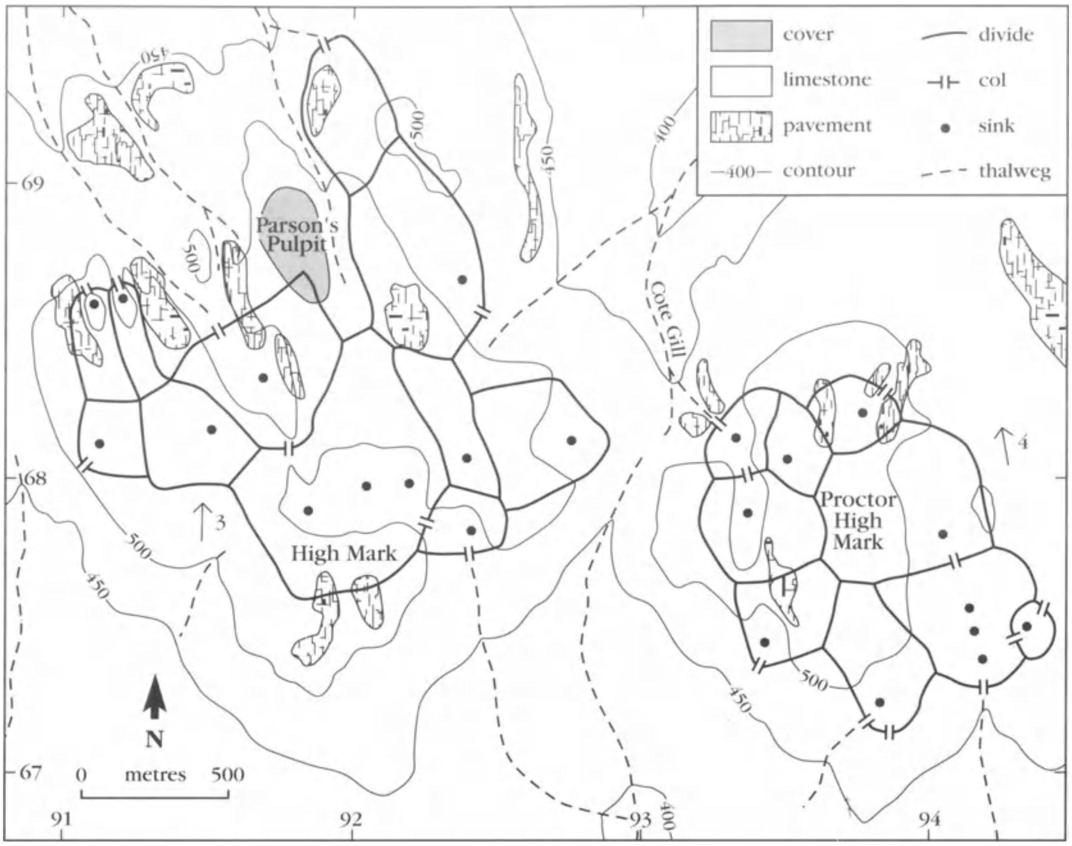


Figure 2.38 Outline geomorphological map of the polygonal karst developed on the limestone crests around High Mark.

floor. The fill includes clastic debris from the eroded cover of Wensleydale Group rocks, remnants of old soil horizons, glacial till, and ponded silts and clays deposited in periglacial lakes. The latter are comparable with sediments in the closed basins of Brimble Pit and Cross Swallet in the Mendip Hills.

The watersheds and cols separating the basins are, in general, poorly defined across ridges of gentle relief. Some of the depression rims are broken by low cols, some of which lie at the heads of shallow dry valleys down the outer slopes. On the east side of Proctor High Mark, two depressions form a more linear feature emphasized by the line of three low points in the southern basin (Figure 2.38); these appear to represent the dissected remains of a once continuous dry valley, which continues southwards outside the polygonal karst. The intervening ridges are generally mantled with thin soils, though patches of limestone pavement are distributed through the polygonal karst independent of the depression relief patterns (Figure 2.38). Some small shafts and hollows are the remains of lead mining about 200 years ago.

Interpretation

The dolines are solutional forms which appear to have developed over a long period of time, since the impermeable cover became thin enough to allow extensive solution to commence in the underlying limestone. The absence of any collapse features and marginal faults refutes any suggestion that the depressions were collapse features. Clayton (1966, 1981) proposed that they are a result of limestone solution, intensified by a layer of sediment holding water and generating carbon dioxide. He suggested that the high rainfall, low evaporation, acidic soils and peaty sediment created an 'acid sponge'. Solution of the limestone in the floors of the depressions occurs more rapidly than it does under the thin soils of the interfluves, thus continuing to deepen the depressions. Limestone pill experiments by Trudgill (1985b) support this hypothesis. A depression is self-perpetuating, once this 'acid sponge' effect is initiated within it. The inward sloping sides of the depression ensure a continued supply of water and reduce solution on the interfluves. In addi-

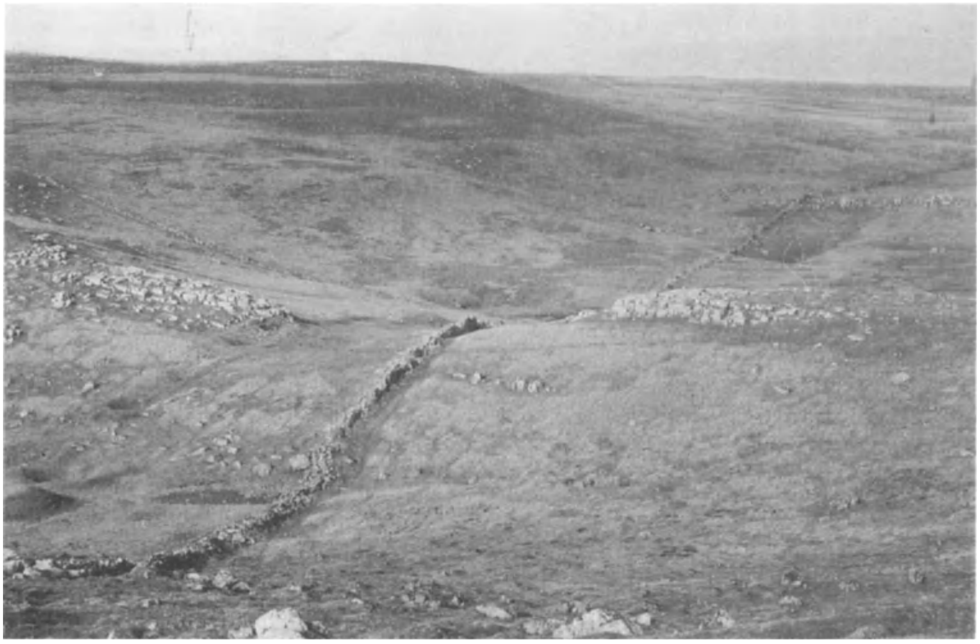


Figure 2.39 The large closed depression between High Mark and Parson's Pulpit. (Photo: A.C. Waltham.)

tion, transfer of material by creep, flow and soil wash into the depression further enlarges the 'acid sponge', increasing its solutional capability. As a basin enlarges and deepens, small limestone scars form around the rim, which then retreat as free-faces, widening the basin further.

The chronology of these features' development is difficult to constrain. Size and depth may indicate the relative age of dolines, but modifications by glaciation and meltwater erosion make these data of limited value. Current estimates of limestone solution rates in the area suggest that these depressions may predate the Pleistocene glaciations (Sweeting, 1966; Clayton, 1981; Trudgill, 1985b), and Sweeting speculated that the marginal plains around the high limestone masses are also very old erosional features. They are located where the Great Scar Limestone of the Yorkshire Dales reaches its greatest elevation and was probably exposed earliest, and the karst subsequently attained the greatest maturity. Although the age evidence is circumstantial, Clayton (1981) maintains that other hypotheses are hard to erect; he argues that the form of the depressions is unlike those such as the Malham Tarn basin produced by glacial erosion, and that areas such as Ingleborough have been eroded by ice and have no comparable features.

The age of the depressions is also relevant to their unusually large sizes. Either they formed over a very long period of time, or they evolved in environmental conditions that were different from those of today. Clayton (1981) suggests that the warmer climate of the Late Tertiary was more favourable for the formation of large dolines and polygonal karst; similar conditions could also date from the earlier Pleistocene interglacials. A pre-Devensian age is undeniable, and it appears that glaciations only caused minimal modification, by ice or meltwater, to the karst depressions. They lie away from the zones of powerful glacial scour, as the main ice flow from the north was obliquely deflected into the Littondale iceway to the east (Figure 2.1). It is significant that these areas of polygonal karst occur on the only limestones in the Yorkshire Dales which form a topographic summit area.

Conclusions

The high limestone plateau of High Mark features an unusual range of karst landforms, including dry valleys, rocky scars and dolines, with some of Britain's largest closed basins. The area provides a rare example of polygonal karst. The closed depressions were formed by solutional processes operating on the limestone underneath a blanket of wet acidic soil, which accentuates erosion in the centre of the basin compared to the higher drier interfluves. Their age is unclear, but they appear to be some of the oldest features in the Yorkshire Dales karst. The polygonal karst housing the depressions constitutes one of the most mature limestone landscapes in Britain.

PENYGHENT GILL

Highlights

The Penyghent Gill valley contains the finest illustration of cavern unroofing and collapse in Britain. The collapsed cave system shows clear relationships to both the geological structure and the valley rejuvenation.

Introduction

Penyghent Gill is a major tributary to the River Skirfare, which flows down Littondale to Wharfedale. The lower section of the gill valley is deeply entrenched in response to glacial overdeepening of Littondale. The Giant's Grave cave system lies at the head of the gill, just above a major knickpoint. This cave system contains over 700 m of passages, all at shallow depths beneath the valley floor. At its lower end there is an extensive area of collapse where the structure of the subsided limestone blocks is clearly recognizable on the surface. Although the site is a spectacular example of cave formation, collapse and unroofing, all in response to rejuvenation, it is poorly documented. The caves are briefly described by Long (1974) and Brook *et al.* (1988).

Description

Streams draining from the Blishmire peat bogs sink into caves formed within the Girvanella Nodular Band of the Hawes Limestone. The Giant's Grave caves have nearly 700 m of passages, broken by unroofed sections where the stream flows in daylight. The Main Cave carries the stream along a single bedding plane in a passage mostly about 1 m high and 2–12 m wide, with braided channels looping round bedrock columns (Figure 2.40). The stream emerges to daylight where the bedding plane is exposed at the head of the incised section of Penyghent Gill. It cascades into an area of collapse and feeds multiple routes either into the Lower Cave or into the surface course. The Lower Cave resurges lower down the gill, where its gradient increases and the valley sides steepen.

The almost horizontal limestone at Giant's Grave has two major bedding planes, about 2 m apart, on each of which there are small areas of well dissected limestone pavement (Figure 2.40). The Main Cave and the lower pavement are formed on the lower bedding plane. The zone of collapse is about 50 m across, and involves the

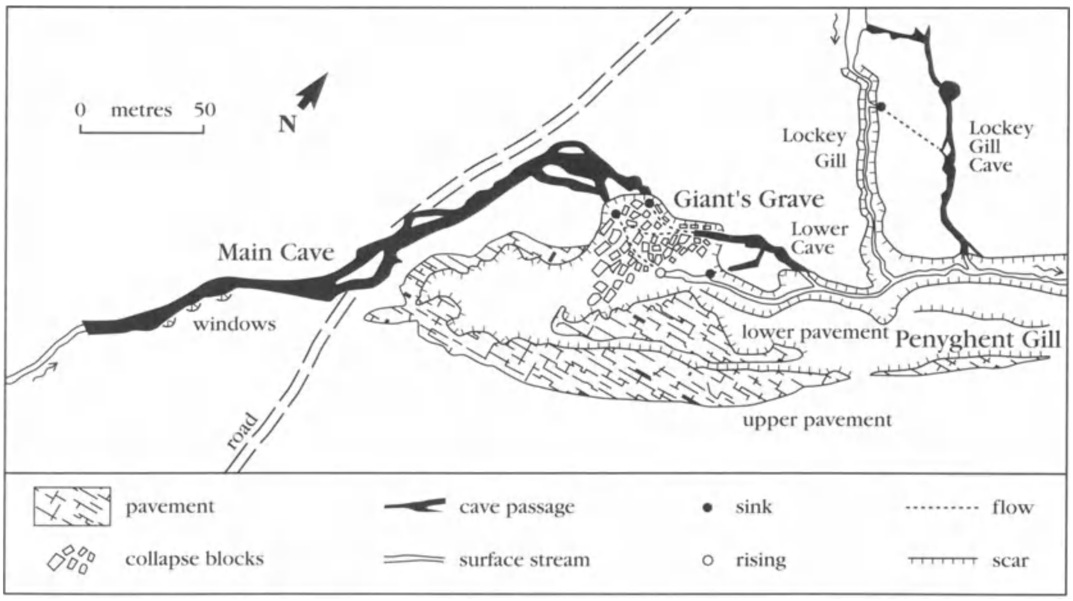


Figure 2.40 Geomorphological map of the collapse area and associated caves at Giant's Grave at the head of Penyghent Gill.

Figure 2.41 The foundered blocks of limestone in the collapse area at Giant's Grave, Penyghent Gill. (Photo: A.C. Waltham.)

beds immediately beneath both bedding planes. The limestone has been undercut by the Main Cave and its downstream, unroofed extension, and also by flow along a third, lower bedding plane which now floors the collapse zone between the Main and Lower Caves. A mass of jumbled tilted blocks constitute the collapse zone (Figure 2.41); minor bedding planes have been etched into the walls of the detached blocks, whose original relationships to each other can still be recognized.

At Giant's Grave the surface drainage has been captured underground and then re-exposed. Many other tributary streams to Penyghent Gill also exhibit underground capture, cave formation and cliff undercutting in response to valley rejuvenation. Lockey Gill has been diverted underground, but this is through a narrow fissure cave which is stable and very far from a state of collapse.

Interpretation

The main development of the Giant's Grave cave system followed the rejuvenation of Penyghent Gill, as a consequence of the glacial deepening of Littondale, which constitutes the local base level. The formation of the wide, braided cave passages has been aided by the presence of the conspicuous bedding plane, and the lack of deeper cave drainage is partially due to the lack of vertically extensive joints in the less disturbed limestone far from the Craven Faults; single joints rarely extend through more than one bed of limestone (Waltham, 1974b). Following the rejuvenation of Penyghent Gill, numerous tributary streams sank into joint fissures in response to the increased hydraulic gradients, but they then followed shallow, nearly horizontal courses perched on the bedding planes. As the limestone above the shallow caves is further thinned by surface lowering, the wide roof spans progressively collapse, creating the masses of subsided blocks, of which Giant's Grave is the best example. This process clearly contributes to the karstic development and entrenchment of the valley, and the process may be repeated many times in the evolution of a deep limestone valley.

Other caves downstream in Penyghent Gill also show the effects of rejuvenation; many have been truncated by downcutting of the Gill, leaving their entrances perched several metres above the valley floor (Long, 1974). The age of the caves is unknown, but their clean-washed youthfulness and the absence of any significant fills suggest that they are entirely postglacial (Long, 1974).

Conclusions

Giant's Grave is the largest area of undisturbed cavern collapse in Britain. The remaining shallow caves demonstrate the role of bedding planes and joints in cave formation. The site clearly demonstrates the scale and role of cavern collapse in karst valley formation; single massive collapses instantly creating valleys or gorges are not involved. The very small scale of the Giant's Grave collapse stands in marked contrast to the larger karst valleys of Cheddar Gorge, Trow Gill and the Winnats Pass; these sites are not collapsed caverns, but cave collapse has played a role in their progressive excavation.

SLEETS GILL CAVE

Highlights

Sleets Gill Cave and the adjacent Dowkabottom Cave constitute fragments of an ancient, deep phreatic cave system truncated by surface lowering and now exposed out of equilibrium with the present geomorphology. The sediments within both caves contain clasts derived from strata no longer surviving at outcrop. Sleets Gill Cave contains the finest examples in Britain of both abandoned and intermittently active phreatic lifts.

Introduction

Sleets Gill Cave and Dowkabottom Cave lie on the western side of Littondale, just above its confluence with Wharfedale (Figure 2.1). The plateau above the caves is developed largely on the Great Scar Limestone, with only two very small outliers of Yoredale facies shales surviving within the catchment. The water from both caves resurges under normal flow conditions at Moss Beck Rising, 1 km east of the Sleets Gill entrance. The caves have remained largely undocumented. Sleets Gill Cave was described by Monico (1989b), and all the caves were described briefly by Long (1974) and Brook *et al.* (1988); Long (1992) made further comment on the unusual hydrology and flooding regimes.

Description

Sleets Gill Cave (Figure 2.42) is entered through a 3 m wide arch at the head of a normally dry stream bed, 60 m above the floor of Littondale. The entrance passage is an inclined tube descending to a depth of 24 m, where it levels out into the Main Gallery, a 4 m diameter phreatic tube which can be followed for 370 m to a choke. An immature active streamway lies below and just south of the old phreatic conduit, and provides access beneath the choke into a continuation of the large, relict phreatic tunnel. The Ramp is a large phreatic tube ascending over 60 m at an angle of 35° to a calcited boulder choke.

The choked ends of Sleets Gill Cave lie more than 100 m below Dowkabottom, a large, shallow closed depression cut in the limestone just below the main plateau level. Low rock scars line most of its perimeter, but its floor is grass on a thick soil. Behind the western wall of the depression, Dowkabottom Cave is another fragment of largely abandoned phreatic conduit, accessible through a

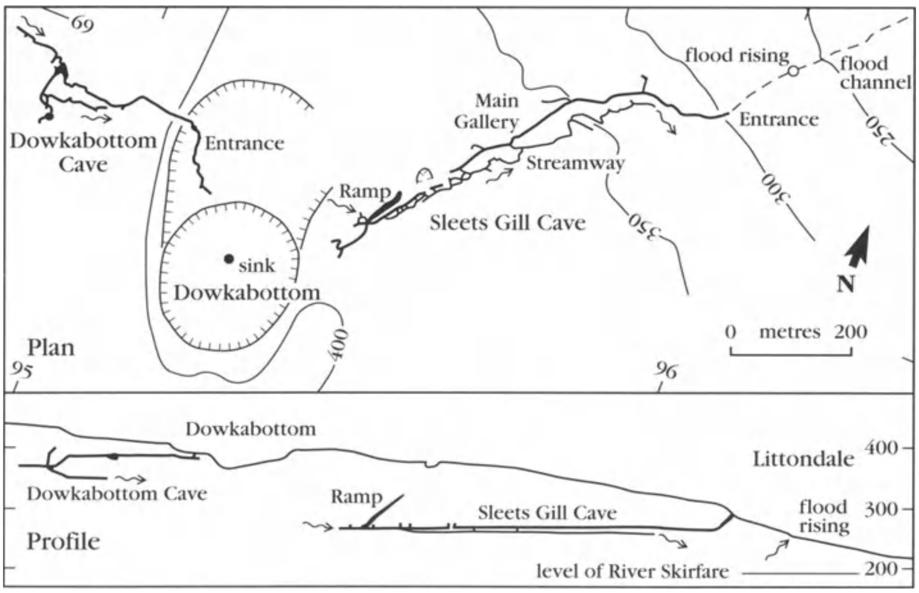


Figure 2.42 Outline map and profile of Sleets Gill Cave and its associated karstic features (from surveys by University of Leeds Speleological Association and others).

rift in the flat floor of an outer, higher basin (Figure 2.42). Most of its passages are large phreatic tunnels, rectangular in section due to block fall; one section is developed along a fault as a narrow rift 20 m high. This cave contains more extensive sediment deposits than does Sleets Gill Cave, and they include calcite flowstones and clastics with pebbles of Yoredale sandstone.

Interpretation

The Sleets Gill entrance passage and the Ramp both carried water upwards and are spectacular examples of relict phreatic lifts. They both ascend at an angle of 35°, cutting through the horizontal limestone with no visible sign of structural control. Their locations appear to have been guided purely by hydraulic factors where water was escaping from deep contemporary phreatic zones towards resurgences in the floor of Littondale when this was at much higher levels early in the Pleistocene. There is no evidence of the relative ages, except that the Ramp's position further into the hill suggests that it is the older, whose flow was subsequently captured by a new lower route along the Main Gallery. The entrance passage is truncated by the modern hillside, and the Ramp is choked at its upper end, so that it is unclear whether the inclined passages were the lower sections of deep vauclusian risings, or were merely phreatic lifts midway along deep karstic conduits. The passages of Dowkabottom Cave appear to be upstream fragments of the same old phreatic cave system, but their higher level suggests that they could represent earlier phases of the cave development.

The altitude of the cave system, now in the flanks of Littondale and with the main passage perched 65 m above the adjacent valley floor, indicates that it is of considerable age. The cave may encompass a geomorphological history spanning the entire glacial modification of Littondale, since the phreatic top of the Ramp is at an altitude at least 130 m above the present dale floor. The mean rate of valley floor excavation in the western Yorkshire Dales is 0.12 m/ka (Waltham, 1986), which suggests that Sleets Gill Cave is at least 1.1 Ma old. Sandstone pebbles in some of the clastic deposits of Dowkabottom Cave were derived from rocks of the Yoredale facies, now stripped from the plateau of Great Scar Limestone; their presence suggests that either the cave formed when the Yoredale cover was more extensive, or the Yoredale material was derived from a partial cover of glacial debris, subsequently largely eroded away. The sediment and speleothem sequence within these caves constitute a stratigraphical record of the evolution of Littondale, yet to be elucidated in full.

The present hydrology is complex, with percolation water derived largely from distant autogenic sources draining via immature caves to a resurgence near the valley floor. Response to heavy rainfall is delayed, but the outlets from Sleets Gill Cave, to both the permanent Moss Beck resurgence and also a flood rising below the entrance (Figure 2.42) appear to be constricted. Flood waters back up in the cave's large tunnels, find various fissure outlets to the rocky channel of Sleets Gill Beck, and in high flood create a temporary vauclusian rising when they overflow the lip at the cave entrance.

Conclusion

The two caves are fragments of a very old system which once carried deep phreatic flow from a large limestone plateau towards Littondale prior to its glacial entrenchment. Sleets Gill Cave contains Britain's finest examples of steeply inclined phreatic tubes, which may have fed ancient vauclusian risings. An immature modern phreas, fed by autogenic input and draining to a valley floor resurgence, lies alongside the larger abandoned passages, and the ancient phreatic conduits may still be utilized during periods of high flow.

BOREHAM CAVE

Highlights

Boreham Cave represents a classic example of a phreatic system which has experienced rejuvenation and reversal of flow direction. It contains superb examples of phreatic tubes, which in places contain straw stalactites in a profusion unparalleled in Britain.

Introduction

Boreham Cave lies in the north-east side of Littondale, 1 km up the valley from Arncliffe (Figure 2.1). The cave entrance is at valley floor level, with the passages extending to the east and north under Old Cote Moor. The slopes and floor of Littondale are cut into Great Scar Limestone, dipping very gently to the north-east. Boreham Cave represents the longest cave system yet explored on the northern side of the dale, with active and abandoned phreatic passages extending for a total length of 3100 m. The cave passages are described by Brook *et al.* (1988) and Yeadon (1975), and the latter includes discussion of the local karst hydrology.

Description

From the entrance more than 500 m of low-level passage extends north-east, before swinging round to the north for a further 1000 m (Figure 2.43). Most of this passage consists of a phreatic tube about 2 m in diameter. It is developed largely at one stratigraphical level, and long sections are permanently flooded, with water flowing slowly to the north, then down a shaft to a lower flooded passage. One half-flooded chamber in the low-level tunnel provides access to 1500 m of high-level caves extending mainly to the east. The main passage is a relict phreatic tube up to 4 m in diameter, which is joined by several small inlet passages, both active and abandoned. Some sections of the high-levels contain thousands of closely packed calcite straws each up to 3 m long, with the finest display in the China Shop (Figure 2.44). There are thick sediment banks in some of the passages, all of which terminate in boulder chokes.

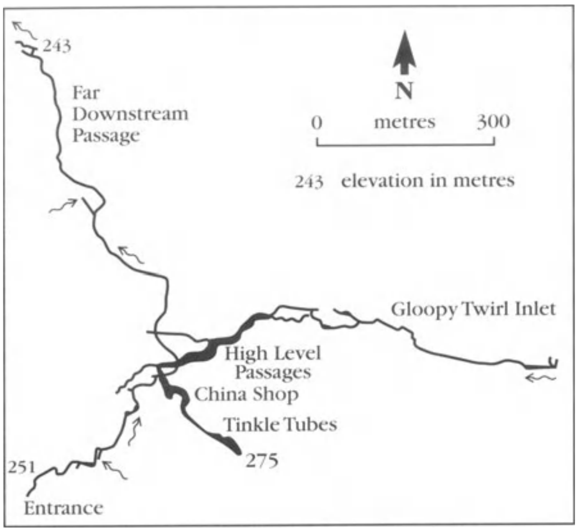


Figure 2.43 Outline map of Boreham Cave (from survey by Cave Diving Group).

Figure 2.44 Delicate straw stalactites hang down into standing water in the old phreatic tube of the China Shop in Boreham Cave. (Photo: T.G. Yeadon.)

Interpretation

The upper and lower series of passages in Boreham Cave show a clear sequence of abandonment and partial rejuvenation of old phreatic conduits, with an associated reversal of flow direction. In the first phase of Boreham Cave, the large phreatic tubes of the high-level series formed the main conduit taking water southwards, via the Tinkle Tubes, towards a resurgence, now choked, in the valley side above the present entrance. Subsequently, the lower passage developed as a phreatic conduit which carried water south for part of its history, and the present entrance represents a truncated resurgence; the partially drained tunnel now carries water to the north. Both upper and lower series have since been invaded by vadose inlet streams and the percolation water responsible for the spectacular straw stalactites. Invading waters in the upper level have cut a vadose trench in part of the old phreatic passage, and have cut down to intercept the flooded lower passages.

Passages in the lower series are still largely flooded, and water in the first section, from the entrance to the junction with the high-level series, is virtually static, being largely ponded percolation water. North of the junction, water draining in from the high-level series flows very slowly to the north, into a vadose canyon which ends at a waterfall shaft down to a lower flooded level; this probably drains to the Litton Risings, 850 m to the south-west. The flooded passages represent a perched phreas in progress of being drained as the downstream vadose canyon cuts back into them. The two levels of passages, both close to the floor of Littondale, contain in their morphology, their fluvioglacial sediment and their calcite speleothem contents an important, but as yet unstudied, record of the Devensian glaciation of the eastern Dales.

Conclusion

Boreham Cave is notable for its phreatic tubes, both active and relict, which follow the bedding at two levels in the limestone close to the level of the modern valley floor. The abandoned high-level passages contain the most beautiful assemblages of straw stalactites found in a British cave; their preservation from accidental damage will be ensured by the natural access restrictions imposed by the flooded passages in the lower series.

STRANS GILL POT

Highlights

Within Strans Gill Pot a series of fault-guided vadose shafts descends to a large, well developed, old phreatic tube. This tunnel has thick mud banks, a spectacular display of calcite straws, and a deep, vadose, floor trench incised in response to rejuvenation. In only a short length of passage, Strans Gill has all the classic features of a multiphase Dales cave.

Introduction

Strans Gill is a shallow ravine cut into the northern side of Langstrothdale, north-west of Buckden (Figure 2.1). Below the slopes of Yoredale shale, the stream sinks on a narrow stratimorphic bench at the top of the limestone, and under normal flow conditions the gill is dry down its steeper, lower course. A single rift in the streambed gives access to Strans Gill Pot, which underlies the surface gill. The entrance to the pothole lies in the Hardraw Scar Limestone, which is contiguous with the underlying Great Scar Limestone where most of the cave passages are formed. The cave has been described by Long (1969), and more briefly by Brook *et al.* (1988) and Long (1974).

Description

Beneath the entrance fissure of Strans Gill Pot a series of short and very constricted rifts lead to a large shaft 50 m deep developed on a north–south tear fault (Figure 2.45). The stream pours down to the boulder floor of a chamber aligned on the fault, and drains into a vadose canyon developed along the intersection of a horizontal shale bed and the vertical fault. This ends at a cascade into a chamber modified by collapse at the intersection of two faults, and lower rifts descend to a sump 105 m below the entrance. Three relict phreatic passages, formed at the levels of shale beds within the limestone, radiate from the collapse chamber. To the south, a narrow, vadose, floor slot descends the fault for 23 m to standing water, but a bedding cave above it enlarges into the Passage of Time - an elliptical phreatic tube, 10 m wide and 2–3 m high over broad mudbanks. This phreatic tunnel is notable for its magnificent array of calcite speleothems, including stalactites, sta-

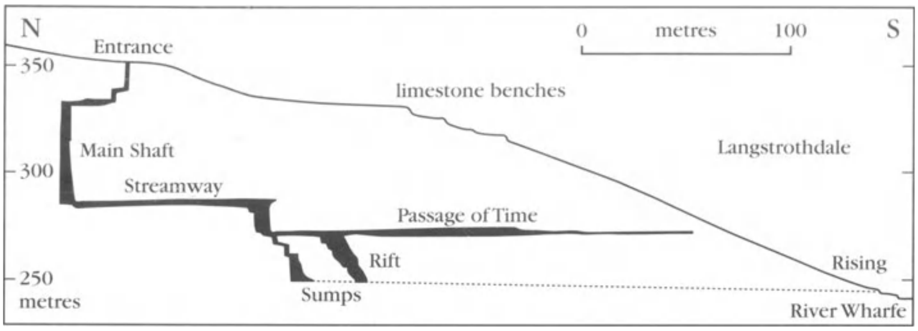


Figure 2.45 Long section through Strans Gill Pot (from survey by British Speleological Association).

lagmites, gour pools and straws up to 3 m long; some large individual calcite crystals are stained green. The ancient passage was truncated by glacial deepening of Langstrothdale, but the cave is choked to the roof with inwashed mud and sand, and surface sediments now bury the exit.

Interpretation

The vadose shafts of the Strans Gill Pot entrance series are in a fault zone, while the relict phreatic tubes are at the levels of shale beds, demonstrating the importance of these beds to cave inception within the limestone. It is notable that the vadose rifts and shafts, both above and below the Passage of Time, are all formed on the tectonic weaknesses in the limestone; in contrast, the large phreatic tunnel curves away from the fault line, while maintaining the stratigraphical level which was favourable to cave inception. Drainage of the lower levels of phreatic cave passages, and deposition of the abundant clastic sediments and calcite speleothems, were the consequences of rejuvenation in response to the adjacent deepening of Langstrothdale by Pleistocene glaciations. Sediments in the cave and around the partly inactive surface gill record the erosion and modification of the karst through past climatic changes, whose chronology has not yet been determined.

Conclusion

Strans Gill Pot is a textbook example of a cave system, with fault-controlled vadose rifts and shafts descending into abandoned phreatic passages. The Passage of Time contains an exceptional and very beautiful display of calcite decorations in a dramatic location.

BIRKS FELL CAVES

Highlights

Two remarkably linear, active, cave systems lie parallel to each other under the edge of Birks Fell, and are almost parallel to the adjacent, steep side of Wharfedale. They demonstrate the overriding influence of a dominant joint set, causing underground drainage to flow for a considerable distance against the dip of the limestone and parallel to the valley side.

Introduction

The caves are located under the eastern slopes of Birks Fell, overlooking Wharfedale south-west of Buckden (Figure 2.1). Both caves are developed in the Great Scar Limestone, with their sink entrances just below the mixed shale and limestone sequences of the Yoredale facies. The limestone dips very gently to the north and is broken by a fault and a major joint set, both trending north-west–south-east. Birks Fell Cave was described initially by Coe (1968), and the passages in all the caves are described by Brook *et al.* (1988).

Description

Birks Fell Cave is the eastern system, containing more than 3600 m of passages extending to a depth of 142 m (Figure 2.46). The entrance lies where a stream sinks through the Girvanella Nodular Band, the distinctive horizon of nodular algal limestone within the Hardraw Limestone, marking the top of the Great Scar facies. Narrow, joint-controlled rifts alternate with low, bedding-controlled sections in the entrance series. Beyond

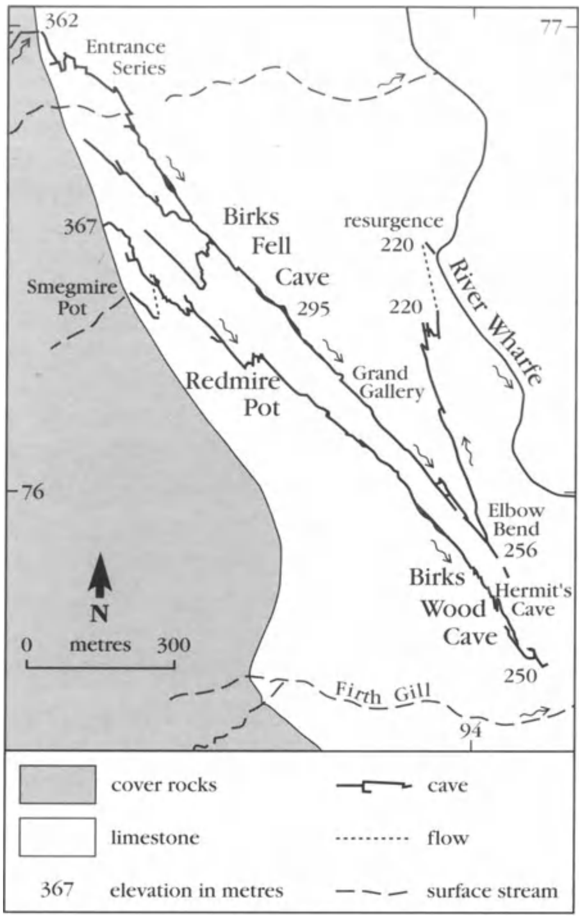


Figure 2.46 Outline map of the Birks Fell and Birks Wood Caves (from surveys by Craven Pothole Club and Cambridge University Caving Club).

two inlets draining from the west, the main passage heads south-east as a large rift developed along a fault. Beyond two avens, also with inlets draining from the west, the passage becomes larger and contains extensive collapse within its tall rifts; parts are well decorated with secondary calcite. Grand Gallery is a rift passage 20 m high leading to a series of high-level oxbow passages, decorated with stalactites and gour pools. The main streamway continues to Elbow Bend where the dry passage to the south-east ends in a choked rift 20 m from the end of Hermit's Cave (Figure 2.46). The active stream passage turns to the NNW, and a low passage draining downdip along a thick shale bed lies between two sections of rift passage. The final narrow rift ends at a sump only 120 m from the resurgence.

Parallel to Birks Fell Cave, and a short distance west into the hillside, a cave system with 1600 m of passage, developed over a depth of 117 m, links a sink at Redmire Pot to a resurgence at Birks Wood Cave (Figure 2.46). From the sink, a narrow rift streamway leads down several cascades and through some flooded sections, before joining the larger Main Streamway, where the water enters from Smegmire Pot. A large, vadose, rift continues past inlets from the west, to a short section decorated with remarkable helictites up to 300 mm long. Further rift passages, sequences of cascades, chambers modified by collapse, and more sections well decorated with helictites continue to a series of low bedding-plane passages with several flooded sections, which emerge at the resurgence exit of Birks Wood Cave.

Interpretation

The structural geology of the limestone has clearly influenced the development of both caves. The exceptionally linear form and the tall rift passages of Birks Fell Cave reflect the development along a fault for most of its length. The parallel Redmire/Birks Wood cave system, as well as Smegmire Pot and the two main inlets of Birks Fell Cave, have developed along the joints, and probably some minor faults, of a set parallel to the main fault. Both the Birks Fell caves have developed obliquely down their limestone fractures, with remarkably gentle gradients and only small shafts along their courses. This is in marked contrast to most of the fault or joint guided potholes in the Yorkshire Dales karst, which drop rapidly down vertical shafts to base level. The influence of a second joint set, trending NNW–SSE, can be seen in both caves, notably in the passage downstream of Elbow Bend in Birks Fell Cave (Figure 2.46).

Dominance of the fractures in guiding the initial drainage through the limestone is reflected in the fact that, over most of their length, the caves drain obliquely against the dip. The passages in Birks Wood Cave are developed largely along bedding/joint intersections; shallow phreatic loops are developed by ponding where the cave drains updip, and are interspersed with cascades where the streamway breaks through to lower stratigraphical levels. This contrasts with the type of phreatic loop prevalent in the Mendip caves at both Cheddar and Wookey, where the water flows downdip before ascending to stratigraphically higher levels up phreatic lifts developed along joints.

Hermit's Cave represents a former outlet for Birks Fell Cave, now lying 35 m above the valley floor. A later phase of Birks Fell Cave evolved with the lower passage draining north, downdip but still along major joints to the modern resurgence in the valley floor; this was probably a consequence of a glacial deepening of Wharfedale, wholly or partly in the Devensian. Secondary calcite deposits in the lower caves are not yet dated, but could provide a time-scale for the excavation of Wharfedale.

Though both caves are structurally guided in plan, their gently graded profiles ignore the potential influences of fractures and bedding, except on short sections of bedding cave. The uniformly graded profiles appear to have developed in a single phase along a hydraulically favourable, straight line path from the sinks to the contemporary resurgences in the slope north of Firth Gill. These paths would have been features of the steep water table in a youthful karst, and the conduits have retained their drainage roles, even though their upper parts are now effectively perched in the mature karst where the stable water table is close to the level of the nearby resurgences in the dale floor.

The development of two adjacent, independent, parallel, linear caves along the hillside that they drain has had a distinctive and unusual influence on their hydrology. Much of the allogenic drainage from the shale cover is captured by the Birks Wood Cave, and the only inlets to enter Birks Fell Cave drain from sinks which lie beyond the northern limit of Redmire Pot. None enters the lower part of Birks Fell Cave.

Conclusion

The two parallel and adjacent caves of Birks Fell provide a classic example of the manner in which geological controls on underground drainage and cave development can override the prevailing surface drainage patterns. Both caves are excellent examples of rift development along joints and faults, and Redmire Pot is also notable for some exceptionally large calcite helictites.

DOW CAVE

Highlights

Dow Cave contains a remarkably linear joint-controlled inlet passage draining from sink to rising beneath a surface interfluve. The main passage intersects a massive choke of debris in a large interglacial shaft.

Introduction

Dow Cave lies under the western slopes of Great Whernside, north-east of Kettlewell, in Wharfedale (Figure 2.1). It is a resurgence cave with about 3000 m of mapped passage, dominated by the straight and narrow Dowbergill Passage, which feeds in from Providence Pot in the next valley to the south. This provides a small flow, but the main drainage enters from choked sinks in Caseker Gill, above the much shorter main passage.

The cave was initially described by Brindle (1954, 1955) and Powell (1954). Its geomorphology is briefly discussed by Long (1974), and the passages are described by Brook *et al.* (1988).

Description

The resurgence entrance to Dow Cave is over a large debris pile where roof collapse has retreated from its exposure in the flank of Caseker Gill. Behind the fallen blocks, the stream flows in a fine keyhole passage up to 4 m wide and 8 m high. Upstream the vadose slot diminishes and the phreatic tube emerges from a wide bedding-controlled cave. This can be followed for only a short distance beyond a large chamber 400 m from the entrance to a major collapse feature, the boulder choke of Hobson's Choice. Above are higher chambers roofed with ill-sorted limestone and sandstone blocks in a matrix of mud. A smaller upstream passage has several waterfalls in rifts up to 25 m high, before a final boulder choke from which the main stream emerges. Some of the old roof channels in this part of the cave are well decorated with speleothems.

The Dowbergill Passage inlet enters Dow Cave low in the south wall a short distance downstream of the Hobson's Choice choke. From its junction with Dow Cave it extends SSE in a remarkably straight line for more than 1300 m to an upstream sump. A series of small, muddy rifts and fragments of larger passages in Providence Pot, are entered through an excavated shaft in the streambed of Dowber Gill, and link into Dowbergill Passage 90 m downstream of the sump. Through much of its length, Dowbergill Passage is a vertical rift, 10–25 m high but rarely more than a metre wide (Figure 2.47). Wedged blocks create sections of false floor, high above the gently graded stream. There are very few roof inlets or speleothems, since the passage lies beneath the Yoredale shale outcrop.

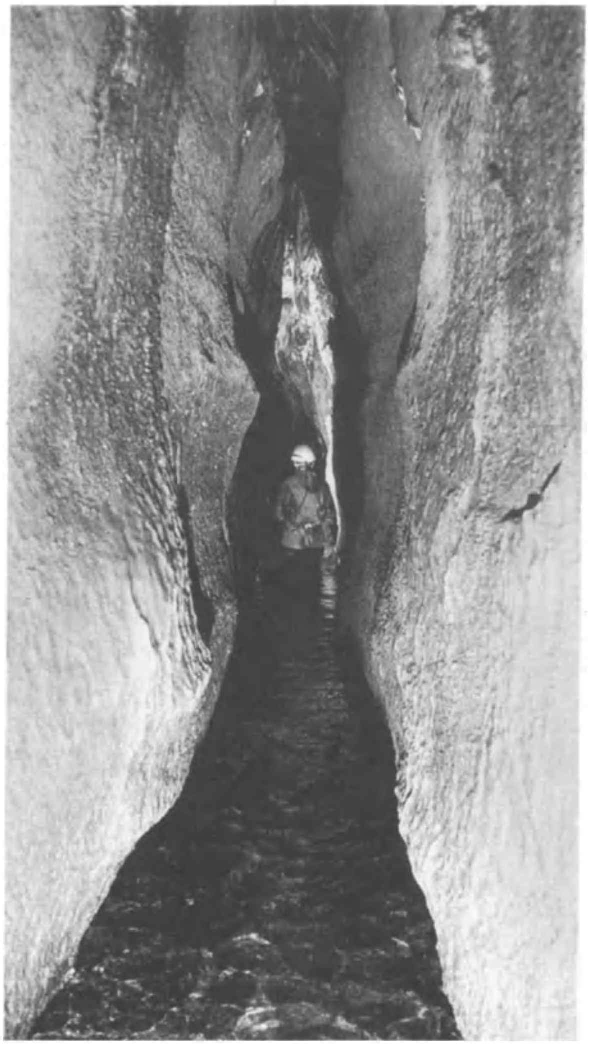


Figure 2.47 The tall, narrow rift of Dowber Gill Passage in Dow Cave (Photo: M.H. Long.)

Interpretation

The large main passages of Dow Cave represent part of a major, old, phreatic conduit, since modified by vadose entrenchment and truncated by valley downcutting. The mix of very large limestone and sandstone blocks in the Hobson's Choice choke suggests that it is a fill of both collapsed and inwashed material in a large pothole, extending to the surface and formed prior to the Devensian glaciation. This fill has since been partly removed from below by the Dow Cave stream.

The Dowbergill Passage inlet is almost dead straight, and is formed along a single major joint. It is a spectacular example of both control by a tectonic joint (Halliwell, 1979b), and a drainage route passing beneath a surface interfluve. Cave passages this straight over such a great distance are generally formed on faults, but Dowbergill Passage appears to be on a simple joint, with no visible sign of fault displacement. The phreatic origins of the rift cave are no longer discernible. Similarly, there is now no evidence of the extent of tectonic opening on the fissure, prior to its solutional enlargement; the site of the joint, parallel to the hillside, would have favoured opening by de-stressing, probably after a glacial retreat. Vadose incision along the fissure was probably very rapid, and has now produced a smoothed profile graded to the level of the Dow Cave streamway. This has been aided by capture of the drainage in Dowber Gill, where Providence Pot appears to be an old sink; this is now active only in flood, as most of the water joins Dowbergill Passage from sinks further up the gill.

Conclusion

The main passage of Dow Cave demonstrates the progressive destruction of a phreatic cave, by drainage, vadose entrenchment, truncation, collapse and choking by debris. It also provides unusually easy access to the base of a large, choked, interglacial sinkhole. Dowbergill Passage is a rift passage of unusual straightness and length, formed along a joint and capturing the drainage from an adjacent valley.

BLACK KELD CATCHMENT AREA

Highlights

Mossdale Caverns and Langcliffe Pot are the two most extensive cave systems developed in the Yoredale limestones. The known cave passages in Langcliffe Pot breach the sandstones and shales into a lower Yoredale limestone, and both caves drain through relatively impermeable beds into the underlying Great Scar Limestone.

Introduction

The two long cave systems lie beneath the southern slopes of the Conistone and Grassington Moors, east of Wharfedale; between them they drain a large part of the slopes of Great Whernside (Figure 2.1). Each cave system is accessible through its stream sink entrance, and can be followed only as far as flooded or choked passages well before their confluence. Both caves drain to the Black Keld resurgence, in the floor of Wharfedale. Most of the accessible cave passage lies within the Brigantian Middle Limestone, which is about 30 m thick, dipping very gently south-east. Shale and sandstone separate this from the underlying Simonstone Limestone, less than 20 m thick; this is underlain by another thin shale separating it from the Brigantian Hardraw Scar and Gayle Limestones which are locally contiguous with the Asbian Great Scar Limestone. The cave drainage traverses a total stratigraphic thickness of over 280 m, to reach the Black Keld resurgence low in the Great Scar Limestone. Namurian Grassington Grit lies unconformably over the Brigantian sequence, and rests directly on the Middle Limestone at Mossdale Scar.

Many of the cave passages were originally described by Leakey (1947), Grandison (1965) and Monico (1989a). Further descriptions of the caves are given by Brook *et al.* (1988), and the hydrology and geomorphology of the area have been discussed by Brook (1971a, 1974b).

Description

Mossdale Caverns

Mossdale Caverns are entered where Mossdale Beck flows into the largest sink in the area. Over 10.5 km of cave passages have been mapped (Figure 2.48), all developed at or near the base of the Middle Limestone. The beds dip south-east at 1–5°, so that the cave reaches a depth of about 60 m. The system has an unusual branching morphology with various outlets for its water. From the descent through the collapse zone near the entrance, the main streamway continues east, only 2 m high but up to 5 m wide, floored by the sandstone underlying the Middle Limestone. Two distributary passages, one perched 2 m above the base of the limestone, branch off the south side before becoming floored with sandstone, subdividing, narrowing and ultimately becoming too tight

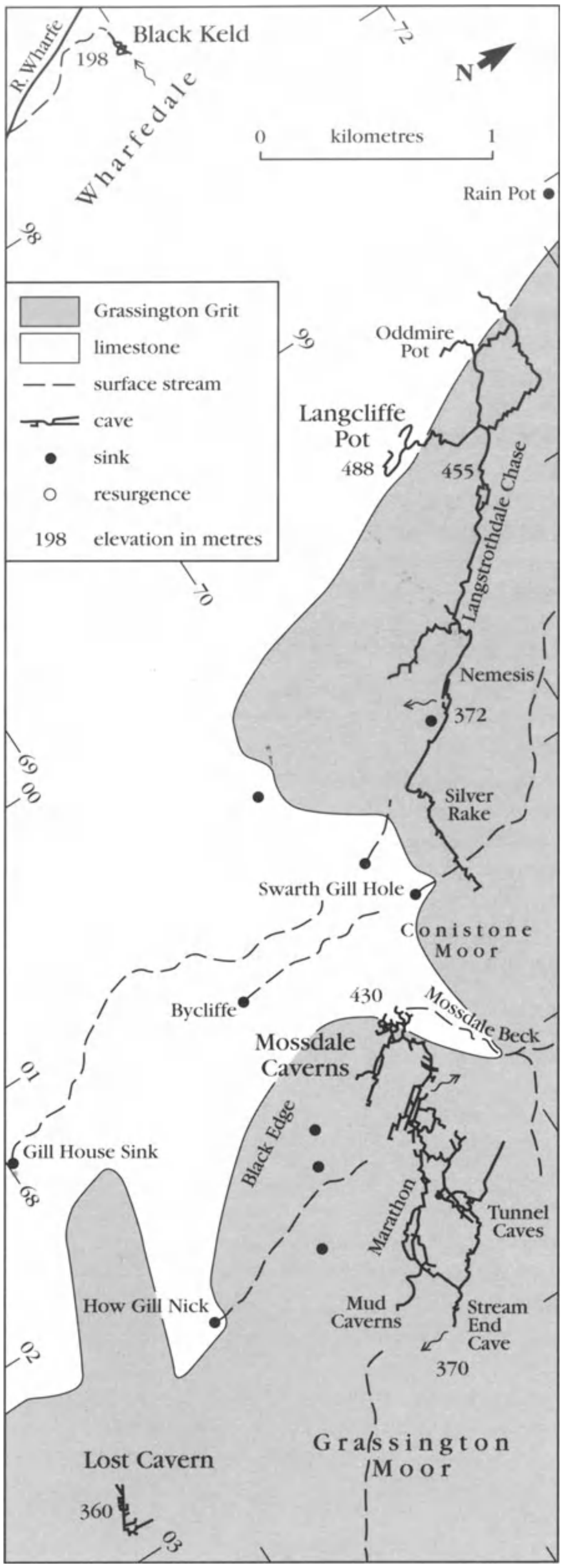


Figure 2.48 Outline map of Mossdale Caverns and Langcliffe Pot, which both drain to Black Keld. The limestone includes the Great Scar Limestone and the Yoredale facies limestones of the overlying Brigantian Wensleydale Group; the latter are separated by thin shales and sandstones that are not marked (from surveys by University of Leeds Speleological Association and others).

to follow. In a second area of branching passages, the main stream turns north into low rifts too narrow to follow. The caves further to the south-east are downdip, overflow passages that normally take only small streams but can be almost filled by large flows generated by floods or sediment choking of the normal outlet rift.

The Marathon passages consist of long, angular series of small joint guided rifts. They are nearly all less than 1 m high, with a narrow triangular cross-section only widening close to the sandstone floor. Some merely form loops and some end in narrow fissures, but Kneewrecker leads into the Tunnel Caves and Far Marathon eventually meets the larger Stream End Cave. The high levels of Tunnel Caves and the Mud Caverns are largely abandoned passages about 5 m high and wide, heavily choked with sand, mud and collapse debris; they lie 5–10 m above the lower streamways. Inlets from the north join the water which flows into Stream End Cave. Both this and all the ends of the high-level passages are finally blocked by impenetrable collapses and boulder chokes. In flood conditions most of the known cave fills to the roof. The small, lower streamways are swept clean by turbulent flows, while ponded water deposits fine sediment in the larger, higher-level passages.

South of the known ends of Mossdale Caverns, the Lost Cavern of Grassington Moor is another fragment of large, abandoned cave passage in the Middle Limestone. It was entered by miners at a depth of 55 m below the Old Turf Pits Shaft, early in the nineteenth century. They found a large dry tunnel, over 230 m long, up to 10 m high and of variable width. This and the larger side passages were mapped (Figure 2.48), but many other passages were left unrecorded, and all the caves have been inaccessible since the entrance shaft collapsed after the mining ceased (NCMRS, 1980).

Langcliffe Pot

Langcliffe Pot is very different from Mossdale Caverns, with five sinks draining to a single underground streamway, in a cave system over 9600 m long (Figure 2.48). Entrance shafts at two small stream sinks each drop about 25 m to the base of the Middle Limestone. Their outlet passages, and many tributaries, are little over 1 m high and wide; they are floored by sandstone, locally eroded through to the shale beneath. Individual segments of the passages are nearly all aligned on joints, but the overall cave pattern is of passages coalescing in the main streamway, which follows the very gentle dip to the south-east. Langstrothdale Chase carries the main stream in a square-cut canyon mostly 2 m wide and 5 m high, its floor rising stratigraphically until it is 4 m above the base of the limestone. In Boireau Falls Chamber the stream cuts down through 0.5 m of sandstone and 4 m of underlying shale to cascade down a 22 m shaft through the entire thickness of the Simonstone Limestone.

The stream then enters a vast zone of blockfall and collapse, estimated to have total dimensions of 100 m by 50 m and 40 m high. Beyond, the stream flows in the Hardraw Limestone but enters a perched phreas, preventing further exploration, before the Great Scar Limestone is entered. A large abandoned inlet passage enters from the south-east before the sump; this extends eastwards as a series of tall rifts along the Silver Rake mineral vein, eventually terminating in a choke.

Dye tests of the Swarth Gill sinks into the Middle Limestone almost above the Silver Rake passages were not conclusive, but suggested that the sinks do not drain into the known parts of Langcliffe Pot; the intervening sandstone and shale appear to be an effective aquiclude in the immediate area, but all the sinking water eventually flows to the Black Keld resurgence (Figure 2.48). The flooded passage at Black Keld has been followed upstream only to a choked area after 150 m. A short dry series reached from airbells has no open continuation.

Interpretation

The underground drainage system which feeds Black Keld is one of the largest and deepest in Britain, though only a small proportion of its cave passages are accessible at present. It is also unique, in that the cave drainage is constrained initially within the Yoredale limestones but then breaks through into the underlying Great Scar Limestone, and resurges about 160 m lower down from near the base of the carbonate succession. The caves therefore breach the intervening shales and sandstones which are normally hydrological barriers. Two of these aquiclude breaches are visible in Langcliffe Pot. The breakthrough from the Middle Limestone into the Simonstone Limestone occurs via a vadose canyon through undisturbed shales and sandstones which then leads to a vertical shaft in the underlying limestone. There is no known continuation of an older passage in the limestone above the breach; the initial route through the non-carbonates was probably via an open tectonic fissure or small fault. The vast area of collapse immediately below the Nemesis shaft may be located in a major zone of fractures which could have guided the initial drainage route through the shale and sandstone between the Simonstone and Hardraw Limestones. The choke now represents a type of interstratal karst, with major collapse of the beds above the limestone, though this has not yet worked through to create a surface depression. The Mossdale water must also pass through several shale and sandstone aquicludes before entering the Great Scar Limestone. This may occur along one of the faults in the area, which have throws sufficient to bring the Middle Limestone and Great Scar Limestone into juxtaposition. The extent to which the Mossdale cave stream backs up in flood conditions suggests that its drainage route into the Great Scar Limestone is via constricted or extensively choked passages.

The complex plans of both the Mossdale and Langcliffe caves are unusual for sites in the Yoredale limestones, which more typically have a single stream passage, as typified by Fairy Holes. Mossdale Caverns has the more complex system, partly due to multiple sink points along the Mossdale valley. Its divergent, branching pattern approaches that of phreatic maze caves such as Knock Fell Caverns, and may reflect development under conditions of frequent back-flooding caused by its restricted drainage outlet. Langstrothdale Chase and Marathon Passage are the longest stream passages in the upper limestone beds in the two caves; they are almost entirely developed on joints, but they are constrained to drain downdip to the south-east within the vadose zone, though the resurgence of Black Keld lies to the north-west.

Mossdale Caverns has an older series of large relict passages along its north-eastern sector through Tunnel Passage, Stream End Cave and the High Level Mud Caverns. This may have developed from sinks higher in Mossdale Beck which have been subsequently choked, probably by an input of glacial debris. The large passages below the modern sink in Mossdale Scar may constitute an old phase which has been reinvaded, and these now drain into the small, relatively immature passages on the base of the limestone. There are other abandoned and choked passages within the entrance complex and Western Passages. Abandoned sinks occur further down the valley, but floor deposits of alluvium and peat now prevent Mossdale Beck from continuing down to Bycliffe Sink. The cave passages show very little vadose development other than small grooves in their sandstone floors. The known parts of Mossdale Caverns probably developed initially within a shallow perched phreas above the sandstone, which was unaffected by regional rejuvenation. A long history of intermittent flooding and epiphreatic development has etched the cave passages out of the joint network to produce a rather open version of a phreatic maze cave (Palmer, 1975).

Langcliffe Pot is morphologically simpler than Mossdale Caverns. The initial phreatic development of the passages occurred up to several metres above the base of the limestone. It is only the active inlets which have cut down to the sandstone, and then rise above it further downstream where the dip is steeper than the cave gradient. The lower series of largely abandoned passages entering from the east in the Hardraw Scar Limestone may represent an earlier phase of development associated with old sinks in the Mossdale valley, but their point of origin is unknown and there is now no flow though from the sinks in the Middle Limestone.

Conclusion

The two influent caves perched high above their Black Keld resurgence are the longest known systems in the Yoredale limestones. They are unique in that they drain through shale and sandstone sequences into the Great Scar Limestone beneath; two of the hydrological breaches which have permitted this are already visible in Langcliffe Pot, as a cave canyon entrenched into the non-carbonates and a major collapse zone.

CONISTONE OLD PASTURE

Highlights

The area around Conistone Old Pasture, on the east side of Wharfedale near Conistone, contains two fine meltwater valleys, one with an excellent dry waterfall and the other a classic narrow gorge section. Between these two valleys is a superb expanse of well developed limestone pavement.

Introduction

The two dry valleys on Conistone Old Pasture are fine examples of channels incised by meltwater at the end of the last glaciation. They lie in the eastern flank of the glacial trough of Wharfedale, beneath a limestone bench which carries areas of well developed pavement. Conistone Dib provides a spectacular illustration of fluvial incision in a limestone area, and admirably demonstrates the role of jointing in determining the form of its gorge section. Dib Scar is an impressive dry waterfall that invokes debate on the role of cave undercutting in its formation. Although little has been written specifically about these sites, they are important examples of their representative landforms. The valleys are briefly described by Sweeting (1974) and Waltham (1984) and the pavements are referred to by Goldie (1976, 1981, 1993).

Description

The dry valley of Dib Scar lies south of Conistone village and the Old Pasture plateau (Figure 2.49). It is less than 1 km long, descending from a broad col out of a broad and shallow closed depression. It is entrenched up to 30 m into the strong Carboniferous Great Scar Limestone. Its main feature is Dib Scar, a fine dry waterfall that now forms an overhanging limestone cliff, 20 m high, at the head of a short gorge with vertical sides. Immediately east of its namesake village, Conistone Dib is a longer and larger dry valley. It is incised by up to 60 m as it descends 120 m through the Great Scar Limestone. The upper end of the valley has a broad grassy floor where it collects three short tributaries; its sides are scored by solifluction terracettes (Figure 2.50). Downstream it narrows into an impressive ravine, often referred to as the Gurling Trough. This is cut into two of the stronger limestone beds, with its form influenced by the joints. The walls pinch in to less than a metre apart at one point, and have fluvial potholes and old swirl pool sites on the bends. The gorge descends gradually until it widens to die out at the edge of Conistone village. The overhanging limestone cliff of Kilnsey Crag rises on the opposite side of Wharfedale; it is a glacially truncated spur, and its only karstic feature may be that its foot was undercut by lateral solution when a postglacial, moraine-dammed lake lapped against it.

The limestone plateau above the dry valleys lacks any surface drainage. It is a broad feature with disorganized low relief, with shallow, broken troughs aligned parallel to Wharfedale. Close to the dry valleys there are fine expanses of lime-

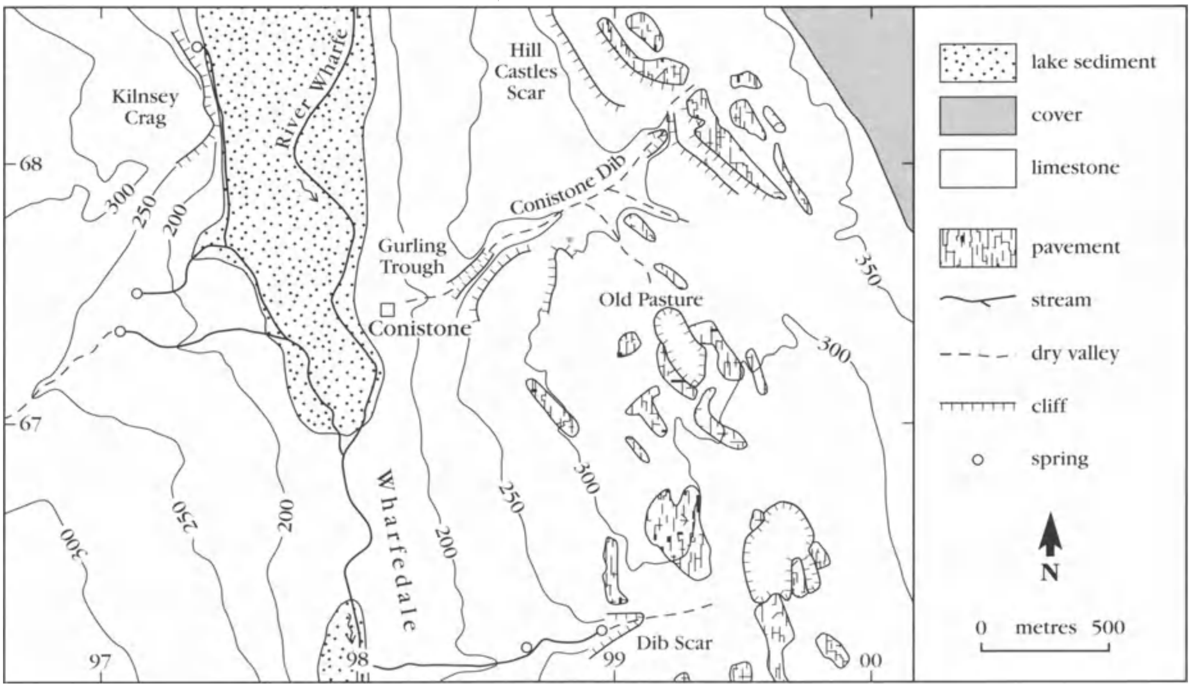


Figure 2.49 Outline map of the dry valleys, scars and limestone pavements of Conistone Old Pasture. Cover rocks are the shales and limestones of the Yoredale facies, above the Great Scar Limestone.

Figure 2.50 The dry valley of Conistone Dib, seen from the limestone scars with Wharfedale in the distance. (Photo: A.C. Waltham.)

stone pavement and small rocky scars in the stronger and more massive beds. Above Hill Castles Scar the pavements are horizontal with blocky clints less than 1 m by 2 m on average, and shallow grikes with a mean depth of 0.7 m (Goldie, 1976). Solution runnels are mainly smooth rundkarren, and there is great variety in morphological detail with excellent kamenitzas and small-scale rippling and pitting. Nearer to Dib Scar, the limestone slabs dip as much as 15°, and longer solution runnels are orientated down their slopes.

Interpretation

The fluvial gorges have been interpreted as glacial meltwater channels active when underground drainage was impeded by permafrost and glacial debris (Sweeting, 1974; Waltham, 1984). The dry gorge at Dib Scar is a classic example of gorge formation by waterfall retreat. It was incised at the tail end of the Devensian glaciation by meltwater flowing down a channel from the fells above, perhaps from or under the snout of a retreating glacier. At the location of Dib Scar this flow crossed a resistant bed of limestone, whose face retreated beneath the sediment-laden water to form the tapering gorge seen today. The waterfall was subsequently left dry by the climatic amelioration when underground drainage resumed. Cliff undercutting and cavern formation may have played minor roles in the development of the waterfall, but modern weathering and frost action are degrading its form.

Conistone Dib was a second meltwater channel, which once carried a large stream. This may have been fed from a retreating glacier lobe on the plateau above, but its source could have been related to ice margin meltwater flows which may have excavated the linear depressions parallel to Wharfedale on the limestone plateau (Raistrick, 1931). The upper gorge section occurs where the meltwaters incised through the stronger limestone beds of Hill Castles Scar, and the Gurling Trough gorge is cut into the same strong beds of limestone as is Dib Scar. The whole feature was probably incised very rapidly at the end of the Devensian glaciation, and was abandoned by the resumption of underground drainage in the more temperate Holocene climate.

The limestone pavements are restricted to the ice scoured bench less than 400 m wide along the rim of the Wharfedale glaciated trough. They occur on the outcrops of the more massive beds of limestone, away from the meltwater channels which score the bench and the slopes below. The rundkarren are well developed, but clint sizes are restricted by the closely spaced rock joints. The shallow grike depths suggest that the top layer of clint blocks may have been extensively removed (Goldie, 1981), perhaps during construction of the many early Celtic settlements on this bench.

Conclusions

The dry valleys of Conistone provide two excellent examples of fluvial erosion processes on a karst terrain. Dib Scar is a classic illustration of a tapering retreat gorge with a dry waterfall at its head, and is comparable to Malham Cove without the involvement of glacial excavation. The larger Conistone Dib, with its fine gorge section, is a superb demonstration of the role of fluvial erosion by glacial meltwater under periglacial conditions in a karst terrain. The well developed limestone pavements offer a wide variety of morphological detail and show clear response to structural control, valley and soil formation.

Chapter 3

Outlying karst areas of the northern Pennines

INTRODUCTION

The northern Pennines contain Britain's most extensive and spectacular karst landscapes, spread across a range of uplands formed in mixed sequences of sandstones and limestones interbedded with shales. Most of the region is dissected by deep glaciated valleys cut between numerous residual summits which rise to over 700 m. The karst is all formed on the Carboniferous limestones which form bands across the whole region (Figure 3.1). The best known and most extensive outcrops are in the southern Yorkshire Dales, and these are referred to in Chapter 2.

Outside the main Yorkshire Dales karst, the karst landforms of northern England are widely scattered on various Dinantian and Namurian limestones which exhibit considerable lithological and structural diversity (George *et al.*, 1976; Dunham and Wilson, 1985). Both the modern structures and the Carboniferous environments are best viewed as features of three positive areas and some intervening troughs. The Pennines form a massive escarpment; the Askrigg and Alston Blocks, separated by the shallow trough of Stainmore, both have limestones dipping gently east from their upper margins along the Dent and Pennine Faults respectively (Figure 3.1). The third positive area is the denuded dome of the Lake District, where the limestone survives as part of an annular escarpment facing the Lower Palaeozoic inlier; these outcrops now lie on the edge of the Vale of Eden and around Morecambe Bay, where, though topographically outside the Pennines, they are geologically related.

The single most important limestone in the northern Pennines is the Great Scar, largely of Asbian and Holkerian age. This massive and pure facies forms most of the karst in the southern Dales (Chapter 2), and extends with very similar lithology into the faulted blocks just east of Morecambe Bay and also into the Stump Cross area just north of its bounding Craven Fault Zone. Northwards the Great Scar Limestone thins considerably, and thick shales occupy much of the Asbian succession. It forms the low escarpment with its many pavements between the Vale of Eden and the Lake District, but its extension onto the Alston Block, as the Melmerby Scar Limestone, has very little exposure beyond the pavements at Helbeck.

The Brigantian rocks are dominated by the Yoredale facies of cyclic shales, sandstones and limestones; these form the Wensleydale Group on the Askrigg Block, and the Alston Group on the block of the same name. The cave systems of Nidderdale are cut in the Middle Limestone, and are the largest karst landforms in the many limestones within these sequences (Figure 1.9). Cyclic sedimentary sequences continue with little change into the Namurian, where the lowest unit is a limestone. Known as the Main Limestone on the Askrigg Block, and as the Great Limestone on the Alston Block, this contains all the largest caves and karst landforms in the Pennines north of Wensleydale. All the Carboniferous limestones are very similar with respect to their karstic erosion; they are strong and generally massively bedded and are broken by thin shale partings; the Brigantian and Namurian limestones are generally darker than the Asbian Great Scar.

Lower Palaeozoic rocks underlie the unconformity which occurs throughout the region at the base of the Dinantian, but have almost no direct influence on the limestone aquifers and their karst morphology. In the Vale of Eden, the Dinantian limestones are underlain by the Holkerian Orton Group of mainly clastic sediments, which were deposited in the contemporary Ravenstonedale Trough.

Most of the limestones in the Pennines are structurally uncomplicated, with dips of 1–3°, widely spaced faults and generally two, well developed, conjugate joint systems. The notable exception is along the major fault zones bounding the Pennine blocks; the Stump Cross and Short Gill caves and the Clouds and Helbeck pavements all owe their character to strong folding of the limestone within these disturbance zones (Figure 3.1). Dips of 10–20° are common in the limestones around the fringes of the Lake District, and block faulting dictates the overall morphology of the hills east of Morecambe Bay.

The karst

Pleistocene ice sheets repeatedly scoured the northern Pennines, and the Devensian ice covered the entire area. The limestone landscapes are therefore dominantly glaciokarsts, except where glacial till totally blankets the outcrops or where landforms on the thinner limestones are lost in a broader topography of glaciated features. Ice sheets flowed south from Scotland, east and south off the Lake District, and almost radially from centres of ice accumulation on both the Pennine blocks, near the head of Wensleydale and over

Cross Fell (King, 1976). The Stainmore Gap was a major iceway across the Pennines, and all the Dales were modified to some extent, so that the glaciated troughs now epitomize the Pennine landscape. Scouring of the shale and soil cover and plucking of the limestone beds occurred across all the limestone outcrops, with varying intensity related to aspect and exposure to the flow of the glaciers and ice sheets.

On the large scale, the main valleys are all influenced by geological structure. The Vale of Eden is a major structural low, the Dent Fault has glaciated troughs along most of its length, and all the main valleys north of Wharfedale drain east with the dip, though at lower gradients. Nearly all the glaciated Dales have low longitudinal gradients, and most of them contain permanent surface rivers. Nidderdale and Dentdale have sections of limestone floor where the drainage is underground except for flood flows. Wensleydale may have an unmeasured underflow through its limestone floor, but this route offers no significant hydraulic advantage and a permanent surface flow is therefore maintained over a long limestone outcrop. The northern Dales cross only narrow outcrops of limestone, and only the River Greta goes underground for a few metres at God's Bridge (Figure 3.1). There are no long dry valleys, but How Stean Gorge in Nidderdale and Hell Gill are two of the more spectacular karst gorges cut through the narrower limestone outcrops by streams in valleys tributary to the main dales.

Limestone scars and pavements form the most conspicuous elements of the northern Pennine karst regions. Low white scars, plucked clean by the glaciers, fringe narrow rock terraces which almost trace the contours in most of the northern Dales, but they are overshadowed by the larger scars in the Great Scar Limestone of the southern Dales. The limestone pavements are much more dramatic, and form huge expanses of bare rock wherever the stronger limestone beds lay in the right attitudes to be stripped clean by Pleistocene ice. The pavements of Morecambe Bay and the Lake District fringe constitute some of the finest glaciokarst landforms in northern Europe (Goldie, 1981, 1993).

Where the Pennine limestones are covered by glacial till, they are distinguished by huge numbers of subsidence dolines. Locally known as shakeholes, these are classic karst features formed by suffosion and ravelling where percolation water washes the unconsolidated till into underlying limestone fissures. Each shakehole is typically 2–10 m across, with sloping sides and a depth limited by the till thickness. Almost all the Pennine limestone outcrops are marked by either zones of shakeholes in a till cover, or strips of pavement on scoured rock surfaces. Even the thinnest of the Yoredale limestones is commonly defined by a line of shakeholes round a hillside. Some larger dolines swallow allogenic stream flows, with open cave entrances or shafts descending into solid limestone, and there are innumerable risings on all the Dales hillsides. The larger are resurgences of sinking streams, many draining from open cave passages, but most are small flows of percolation water emerging from impenetrable fissures.

The caves

Nearly all the Carboniferous limestones have underground drainage; they are massively bedded, strong limestones whose high mass permeability is entirely due to fissure flow with secondary, solutional enlargement along tectonic joints. Most of the limestones are ideal hosts for karstic development, and caves are widespread at favourable sites; a few units of thinly bedded and heavily fractured limestones behave as diffuse aquifers and have no accessible caves.

The relatively thin scatter of known caves across the northern Pennines, in contrast to the very cavernous karst around Ingleborough (Chapter 2), is a function of both geology and drainage patterns. Most of the northern Pennine limestones form units less than 50 m thick and lie almost horizontally, thereby precluding the development of deep caves. Convergence of the underground drainage, or the sinking of large allogenic streams, is necessary to create stream caves of humanly accessible dimensions, and these criteria are generally not fulfilled within the thin limestones of the region. The typical situation has narrow hillside outcrops of the thin limestones, which swallow innumerable tiny streams; each flows underground for a very short distance to a rising at the base of the limestone directly down the hillside, forming one of many, roughly parallel, very small caves. Furthermore, most of the large areas of limestone pavement lie on high ground so that they receive no input of allogenic drainage; rainfall drains into all their fissures and emerges from numerous small risings fed by seepage flows and inaccessibly small cave passages.

Where the geology, topography and drainage

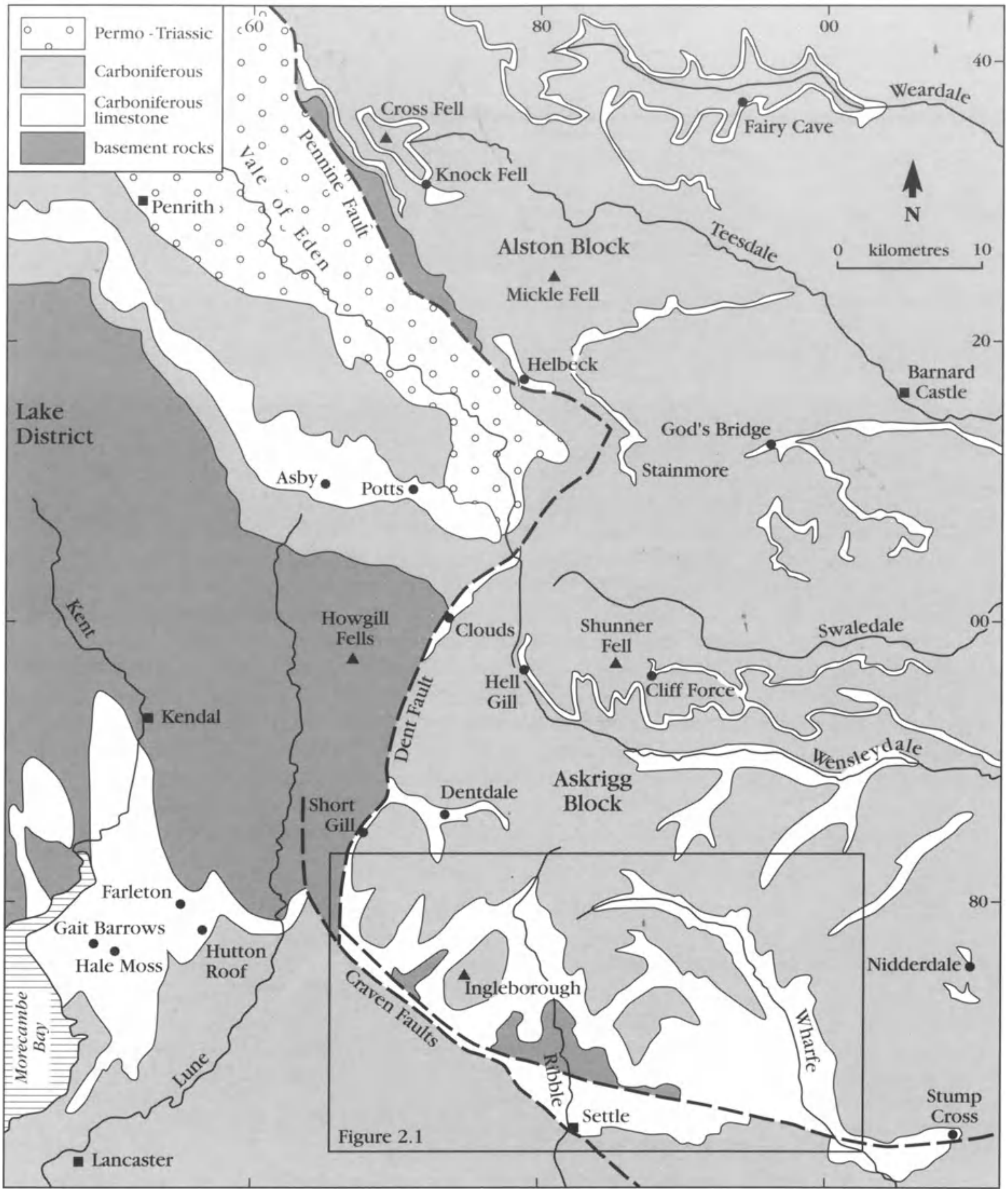


Figure 3.1 Outline map of the karst regions in the northern Pennines, with locations referred to in the text. The other Carboniferous rocks are the non-carbonates of the Orton Group and Yoredale facies of the Dinantian, and the Namurian, but they include thin bands of limestone with lesser karst features not shown on this map. The Carboniferous limestone includes the Dinantian Great Scar Limestone, the Yoredale limestones with significant karst, and the Main or Great Limestone of Namurian age. The basement rocks are Lower Palaeozoic non-carbonates. Details and locations in the southern Dales are shown in Figure 2.1.

are favourable, significant caves are formed, and this region houses three very distinctive and very different cave environments, each almost restricted to its own unit of geological structure.

Within the structural and environmental unit of the northern Pennines, two types of cave morphology are a function of the thin limestones which house them. These limestones are interbedded with clastic sedimentary rocks of low permeability, and therefore formed confined or artesian aquifers, before they were drained and rejuvenated as valley floors were excavated through them. Under these conditions of slow phreatic flow, the systems of tectonic joints were opened by solution to form networks of fissure caves. Developed to maturity, these form the first type of northern Pennine cave - spectacular maze systems; a number of these have been intersected by mine workings in the Main Limestone in Swaledale (Ryder, 1975), and further north Knock Fell Caverns is Britain's largest known maze cave (Figure 3.1).

Rejuvenation of the immature fissure networks permitted vadose stream caves to develop though them, forming the second distinctive cave type within the northern Pennine environment. These vadose canyons are distinguished by zigzag plans where they have followed and enlarged routes through existing interconnected fissures on the conjugate joint systems. The major caves are formed where topography creates a stream sink site at a significant distance updip from a potential resurgence site - in any of three situations. Where a major valley floor exposes limestone, its river drains underground beneath a surface flood route, as in Nidderdale. Where a hillside outcrop follows the gentle dip, a stream can drain through a cave parallel to the surface slope, as at Fairy Holes. Where a limestone dips gently through an interfluve, beneath an impermeable cap, stream sinks in one valley drain to risings in another; a number of sinks in Wensleydale drain to risings, including Cliff Force Cave, in Swaledale (Ryder, 1975).

A second cave environment is provided by the fault blocks of Great Scar Limestone east of Morecambe Bay. There is no allogenic drainage onto the limestone hills, but shallow phreatic flow has created horizontal caves in the hill margins adjacent to Pleistocene lake flats, including Hale Moss (Ashmead, 1969).

The third cave environment is within the strongly folded limestones in the disturbance zones along the major Pennine block faults. Phreatic drainage along the strike dominates the cave morphology, and Short Gill Cavern is the finest example in limestone folded into nearly vertical dips.

HUTTON ROOF

Highlights

Hutton Roof Crags form an extensive area of limestone pavements which are notable for their variety of structural form, including some areas of very well runnelled clints. The site is famous for its steeply sloping pavement with a diamond pattern of joint fissures and excellent long karren runnels, making a dramatic landform in an area of fine karst scenery.

Introduction

Hutton Roof Crags is a southern extension to the prominent limestone hill of Farleton Knott rising above the lowlands east of Morecambe Bay (Figure 3.1). The limestone forms an anticline plunging to the south, and faulted across its northern end; the Kendal Fault bounds it on the west, and the beds steepen into the Hutton Roof Monocline on the south-east (Figure 3.2). The Holkerian and Asbian limestones dip beneath a Brigantian cover east of the monocline, and their base is not exposed. The Hutton Roof limestones mostly dip at 10-20° away from the anticline, but dip more than 30° on the east flank, where the steeply sloping pavements of the Rakes are formed.

Although the steeply inclined limestone pavements of the Rakes are classic karst landforms which are widely known, Hutton Roof has been the focus of little detailed research. Morphometric work on the clints and grikes has compared the main pavements with others in Cumbria and elsewhere (Goldie, 1976, 1981), and the steeply dipping Rakes pavements have been referred to more widely (Williams, 1966; Sweeting 1966, 1974). Specific aspects of the karst morphology of the site have been described by Corbel (1957), Gale (1981a, b), Vincent and Lee (1981) and Pfeiffer (1991). The site is also rated as the second most valuable in Britain for the richness and variety of its flora (Ward and Evans, 1976).

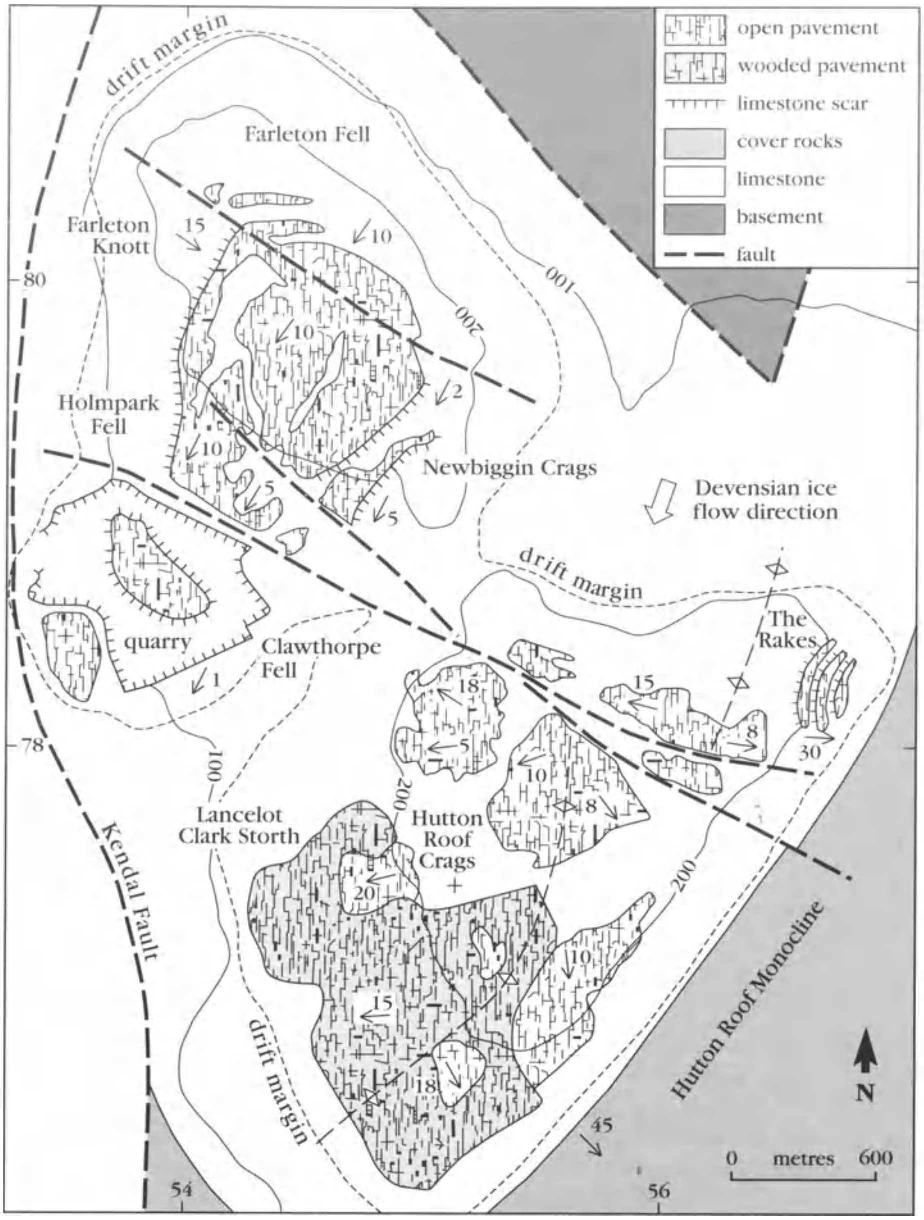


Figure 3.2 Outline map of the limestone hills of Farleton Knott and Hutton Roof Crags. Basement rocks are Silurian mudstones. Cover rocks are the Brigantian and Namurian Bowland Series. The drift margin marks the edge of the thicker glacial till which covers most of the lowland around the limestone hills (partly after Moseley, 1972).

Description

The greater part of the Hutton Roof karst is represented by the many tracts of good, well-dissected pavement, dipping south-west at less than 10°, on the higher parts of the hill west of the monoclinal crest. These gently inclined pavements are broken into small patches by minor structural undulations and faults, but they still display a great variety of surface solution features including kamenitzas, rundkarren and rinnenkarren. They include fine examples of stepped pavement (*Schichttreppenkarst*) - a stripped karst with exposed bedding planes on limestone beds which are truncated to create a stepped profile where the dip is not parallel to the surface slope. A partial vegetation cover only partly disguises this morphology.

The western outcrops, at Lancelot Clark Storth (Figure 3.2), are well preserved dipping pavement, some of which have unusual undulations with a wavelength close to 5 m; most of these appear to be stratimorphs over small, local folds, but some may be a direct product of large-scale glacial scour. Much of this area is bare pavement, but upslope to the east are more broken outcrops, and further south are the wooded pavements of Dalton Crags, straddling the plunging crest of the anticline.

The Rakes lie on the monoclinal flexure on the eastern edge of the site (Figure 3.2). They consist of three parallel escarpments, with their dip slopes formed on adjacent bedding plane in the massive limestone; they dip at 25–35°, and the dip slope bedding planes are scored by the famous assemblages of joint fissures and karren grooves (Figure 3.3). They are dissected by two major joint sets at about 90° to each other, orientated so that their intersection is bisected by the direction of dip. They were scoured by ice from the north, along the direction of strike. The joints have been opened by solution to form a rectilinear pattern of deep kluftkarren, and these delimit diamond-shaped clints on the steep limestone slabs. Each clint is deeply scored by almost straight, parallel rinnenkarren - solution runnels with rounded troughs and sharp edges to the clint surface, which lie symmetrically downslope across the diamond clints. The runnels vary in length, up to 5 or 10 m, according to their position across each clint, starting not far short of the clint crest and draining rainwater into the grykes crossing at the lower edge. The Rakes are lightly vegetated, except in the small strike valleys at the base of each slab. A fourth limestone bed, beneath the three massive beds of the Rakes, is densely fractured and frost shattered, and its outcrop has a surface veneer of felsenmeer instead of clean pavement.

Interpretation

The gross morphology of Hutton Roof is defined by geological structure. Early suggestions that the isolated limestone hills around Morecambe Bay represent a relict tropical karst of Tertiary age (Corbel, 1957) have been positively refuted (Gale, 1981b; Vincent and Lee, 1981).

The smaller karst landforms have all been imposed by Holocene solution onto limestones exposed and scoured by Devensian ice. Scar edges on single limestone beds were ice-scoured, leaving smooth, rounded scar fronts which still have only limited runnel development. The ice also moved and then dumped the many erratic limestone blocks perched on the pavements at Hutton Roof Crags. Glacial scour may also be responsible for the long, parallel, rounded undulations on the gently dipping pavements of Lancelot Clark Storth.

The limestone outcrop has a patchy cover of

Figure 3.3 The distinctive inclined limestone pavements of the Rakes above Hutton Roof, with the deep rinnenkarren raking down the diamond-shaped slabs between the joint-guided kluftkarren. (Photo: A.C. Waltham.)

soil and vegetation, which has retreated locally to reveal the rounded rundkarren features typical of subsoil solution. The time-scale of development and removal of the soil and vegetation cover is not known. The rinnenkarren on the inclined Rakes are excellent examples of these large solution runnels formed by high subaerial flows of rainwater on sloping pavements. They are far larger than the little, sharp-edged rillenkarren found on many outcrops of bare limestone; they have large rainfall catchments on the large sloping clints, and develop into Hortonian channels which are only slightly convergent on the steep slabs of limestone. They retain their sharp upper rims, because the steep slabs have not retained a soil cover which could blanket them and round their features into the normal subsoil rundkarren.

The morphology of some of the pavements, notably on and around Lancelot Clark Storth has been modified by the recent removal of limestone for garden rockery stone, and on a lesser scale to feed limekilns in earlier times. The results vary from the occasional lack of the top bed of solutionally runnelled clints, to areas systematically stripped of these features leaving a surface of rough bedding planes (Goldie, 1976; Ward and Evans, 1976). There has been virtually no damage on the eastern part of Hutton Roof Crags, and the Rakes remain pristine.

Conclusions

Hutton Roof Crags contains diverse and unusual limestone pavement features of national and international importance. The steeply dipping pavements at the Rakes contain the finest rinnenkarren in Britain, and their diamond patterns of deep kluftkarren are uniquely spectacular with their diagonal fluting by the rinnenkarren.

FARLETON KNOTT

Highlights

Farleton Knott is a prominent limestone hill with large expanses of spectacular pavements across its summit and flanks. These have a great variety of limestone pavement types and a range of solutional features reflecting the different aspects, slopes, minor structural features and sparse drift cover on the site.

Introduction

Farleton Knott stands 200 m above its surrounding lowland, east of Morecambe Bay (Figure 3.1). The hill is a fault bounded inlier of limestone which lies within the block faulted zone of limestone outcrops on the south side of the Lake District uplift. The massive, fossiliferous limestones are Holkerian and Asbian, equivalent to the Great Scar Limestone in the Pennines. There is no caprock on the hill, and the cover of drift and soil is thin and variable; the outcrop was all scoured by Pleistocene ice from the north.

The pavements of Farleton Knott have been referred to in much of the literature on the northern glaciokarst (Sweeting, 1966, 1972, 1974), the clints and grikes have been subjected to morphometric analysis (Goldie, 1981; Rose and Vincent, 1986c), and there have been further studies on details of the pavement morphology (Vincent and Lee, 1981; Vincent, 1981, 1982).

Description

The limestone landforms vary considerably across Farleton Knott, reflecting both the changing geological structure and the geomorphic history, especially beneath the invading Pleistocene ice sheets. There is no surface drainage on the hill, as all rainfall sinks almost immediately, to emerge from springs around the perimeter; there are no known caves large enough to enter.

Farleton Fell forms the northern end of the Knott (Figure 3.2), and its north-facing slope is well fractured beneath a blanket of talus. Pavements are formed on the south-facing slope, and those near the crest are steeply inclined, with small clints, on beds which dip 14–20° south into a small syncline. South of the fold axis, the pavement is nearly horizontal, and well developed with good rectangular clints and numerous transported limestone boulders on protected pedestals. Clints average 2.75 m long and 1.05 m wide, reflecting the fracture patterns of the folded and faulted limestone, and grikes average 1.2 m deep, reflecting bed thickness (Goldie, 1981). The solutional details on the sloping limestone include kamenitzas and solutional ripple marks, some akin to trittkarren. Runnels are poorly aligned on the smaller clints, but larger clints to the west have runnels with stronger downslope alignment.

The limestones at Holmepark Fell, on the southwest side of Farleton Knott (Figure 3.2), dip at

3-8° south-west, and were strongly scoured as Pleistocene ice swept downhill. These pavements are the most smoothly scoured of all on the Knott; they are prominently runnelled by large rundkarren with rounded floors and sharp rim contacts to the pavement surface. The downdip edges of the pavements are more closely runnelled by smooth rundkarren, characteristic of a pavement edge which was once covered by soil and vegetation. They are the least dissected of the Farleton Knott pavements, with average clint dimensions of 3.15 m by 2.32 m (Goldie, 1981). This area also has many transported limestone boulders which are the remains of glacially plucked scars. At its southern end, much of the outcrop on Clawthorpe Fell has been quarried away, but an 'island' of pavement survives with excellent large clints, deep convergent rundkarren and many kamenitzas; it is gaining a new cover of vegetation now that it cannot be grazed by sheep (Figure 3.2). Very large clints on surviving pavements south-west of the quarry have gently sloping tops scored by rinnenkarren runnels up to 15 m long.

On the south-east side of the Knott, Newbiggin Crags (Figure 3.2) is important for its beautiful, nearly horizontal pavements, with outstanding networks of rundkarren (Figure 3.4). The more massive limestone beds form very striking edge scars, 2-3 m high, scored by fine vertical solution grooves and with fallen blocks below. The large rectangular clints, many up to 2 m across, are scored by spectacular rundkarren systems with deep runnels converging down the gentle dip. North of Newbiggin Crags, the outcrops are poorly runnelled as much of the original fretted surface bed has been artificially removed.

Figure 3.4 The excellent pavements with square clints deeply scored by rundkarren on Newbiggin Crags. (Photo: A.C. Waltham.)

Interpretation

Standing well above surrounding lowland, Farleton Knott received the full impact of Pleistocene ice flowing south from the Lake District. On the north face the limestone was broken and ground down by the ice under pressure, while the more gentle lee slopes facing south were plucked and scoured - to leave the bare rock slabs subsequently fretted by solution. Some blocks of limestone were transported and dumped as erratic boulders on the pavements; many of these now stand on pedestals of limestone, which have been sheltered from direct rainfall. It is unlikely that any features survive unmodified from before the Devensian glaciation.

Geological structural has influenced much of the geomorphic variety at Farleton Knott. Several faults extend across the site, and are responsible for topographic breaks including low scars, small structural depressions and dry valleys. The largest structural valley lay along the fault on the southern margin of Holmepark Fell, and was partly floored by pavements, until it was completely removed by the quarry (Figure 3.2). The pavements at Newbiggin Crags are the best of the many on Farleton Knott which show a rectangular pattern (Figure 3.3), influenced by the dominant north-west and north-east orientated joint sets (Moseley, 1972). A third set of north-south joints creates some triangular clints, and influences some runnel patterns.

The great variety of runnel types, dimensions and patterns on the Farleton Knott pavements reflects contrasts in the limestone lithology, structure, slope, aspect, glacial history and vegetation history between individual locations. Grike morphometry at Holmepark Fell revealed a bimodal distribution in histograms of grike widths, suggesting that the group of narrower grikes may be postglacial, while the group of wider grikes inherited a component of preglacial opening (Rose and Vincent, 1986c); it was estimated that about 72 mm of grike opening has taken place since the Devensian glaciation. The same data revealed lower proportions of wide grikes than at comparable pavement sites at Underlaid and Longtail Woods, near Morecambe Bay, which may indicate less glacial scouring at Holmepark Fell than at the other sites. However, morphometric data for the whole of Farleton Knott (Goldie, 1981) suggest that Holmepark Fell was probably the most scoured part of this particular hill. Trittkarren occur on sloping pavements which have probably remained free of soil and vegetation since deglaciation (Vincent, 1983). Around snow patches on Farleton Fell, contemporary processes are largely confined to intermittent freeze-thaw action on the clitter-strewn slopes. Meltwater infiltrates through the limestone clitter, and this helps the karstic hollows to deepen beneath the snow; the hollows are thus polygenetic (Vincent, 1982).

The landforms on Farleton Knott have been extensively affected by human activities. Large areas of Newbiggin Crags, the central part of the limestone pavement area, and parts of Holmepark Fell have displaced clints, rough bedding plane surfaces and veneers of rubbly debris, all of which result from the removal of the top layer of solutionally fretted clints (Goldie, 1981). On the low plain east of Newbiggin Crags, grass regrowth has been encouraged on the rough, artificially stripped limestone surfaces, and only small isolated clints now remain exposed.

Conclusions

The surface of Farleton Knott has a number of excellent limestone pavements, whose morphology exhibits considerable variety. This reflects contrasts in surface slope, geological structure and exposure to scour by Pleistocene glaciers. The spectacular, square cut, clint fields and deep runnels on the pavements of Newbiggin Crags are of national importance and international repute.

GAIT BARROWS

Highlights

Gait Barrows is an extremely important limestone pavement site, being the finest of the many pavements on the low limestone hills east of Morecambe Bay. It is distinguished for its botanical and zoological features, as well as for its wide range of karstic surface landforms.

Introduction

The pavements at Gait Barrows are on the gentle southern slope of a low limestone hill south-east of Arnside (Figure 3.1). They are developed on a fault block of Carboniferous limestone of Asbian age which has a southerly dip of about 3°. The limestone is thickly bedded, and some sparite beds reach 3 m in thickness. The whole site lies at altitudes under 50 m, and was scoured by Devensian ice from the Lake District (Rose and Vincent, 1986c).

An extensive literature about Gait Barrows is primarily concerned with the botany, with passing references to the geomorphology. The site has been described as the single most important limestone pavement in Britain for its richly varied flora (Ratcliffe, 1977; Ward and Evans, 1976). The quality of the pavements gives Gait Barrows almost equal geomorphological importance (Ashmead, 1974a; Goldie, 1986), and Rose and Vincent have studied the spring waters (1986a) and the kamenitzas (1986b) of the site.

Description

The Gait Barrows pavements contain an exceptional range of morphologies within an area of less than a square kilometre. The density of vegetation cover varies considerably and defines three zones within the site (Figure 3.5). The outer zone is thickly wooded, and covers about half the area of the limestone outcrop; the two inner zones contain the open pavements, and have shrubs and trees scattered thinly over them, rooted in the grikes or on patches of soil on the limestone surface.

The most important part of the site is the central exposure (Figure 3.5), with three expanses of massive, open pavement, undisturbed by any clint removal, on a few beds of limestone locally as thick as 3 m. Massive clints are up to 30 m long, with clean surfaces on limestone scored by thin mineral veins (Figure 3.6). They slope gently to the south, and slight flexures in the regional dip produce broad undulations in the pavement. The small-scale karst landforms include a great range

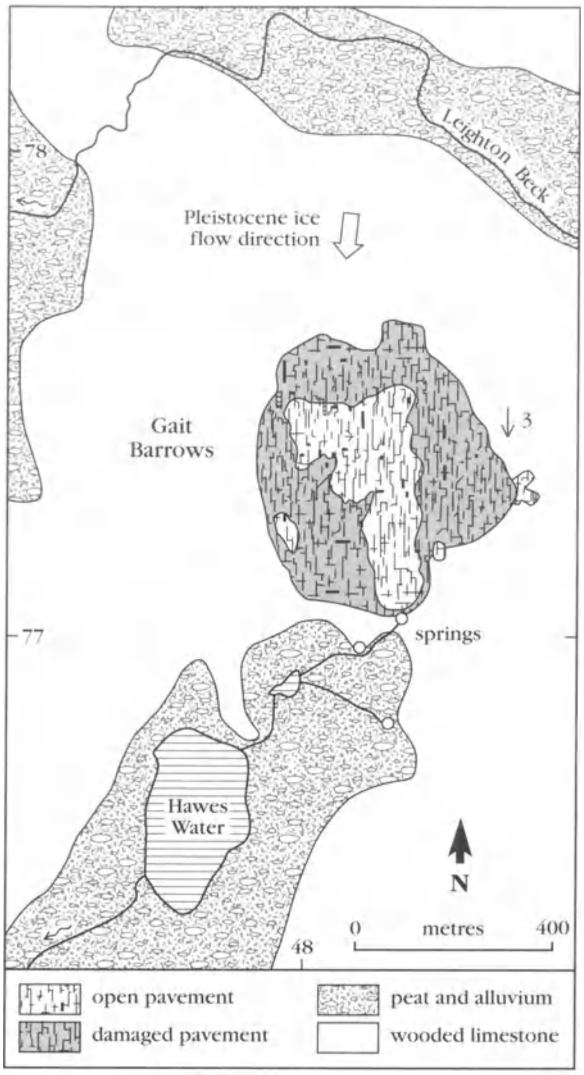


Figure 3.5 Outline map of the limestone pavements on Gait Barrows.

of kamenitzas, at varying stages of development, immature but deep grikes which commonly do not intersect, and pedestals of protected limestone beneath erratic boulders. Some kluftkarren grikes are more extensive, and some are concentrated along zones of tectonic fractures. There are also many of the small, sharp-edged rillenkarren solution grooves on the edges of some of the clints. The pavement has many small erratics of Namurian sandstone, including some wedged in the grikes.

Around the central area of open, massive pavement, there are several areas of more broken and dissected pavement (Figure 3.5). Some of these are in their natural state on more fractured limestone, while others are artificially stripped; these expose bedding planes from which the top bed of clints has been removed within the last 100 years, revealing some of the subsurface morphology. There are also areas where removal of the original pavement was distinctly scrappy, and clint blocks were left loose, tilted or overturned, providing different faces exposed to modern subaerial erosion.

The third zone of the Gait Barrows limestone is densely vegetated. The rock surfaces are completely covered by ground vegetation, including mosses and low flowering plants, so that their morphology is extremely difficult to see. Close examination shows that these areas are well dissected pavement with large clints; beneath the organic cover, the dominant morphology is that of well rounded rundkarren. Good, intact pavement with a largely rectilinear pattern of grikes lies east and west of the central area. Many clints are a few metres in length, though they are often narrow due to the local dominance of any one major joint set. Some areas towards the southern boundary with the peatlands have clints which are smaller and produce many naturally loose blocks of limestone.

Interpretation

These pavements are formed on limestones close to sea level which were overrun by sediment-laden Pleistocene glaciers spreading over lowland plains. Any erosional effects on the limestone landforms by high sea levels during the Pleistocene have not survived the Devensian glaciation. The excellent quality of the Gait Barrows pavements is a function of both the massive limestone beds at surface level and also the aspect which the site presented to ice erosion. The gentle southerly dip combined with the southward flow of the glaciers to maximize the scale of ice plucking on the lee of the hill, so leaving the cleanest of rock surfaces ripe for subsequent solutional fretting. The main pavements have large, bare clints on the massive limestone beds, separated by kluftkarren grikes which are largely postglacial in origin. Some grikes have significantly greater widths, which suggest that they have a component of preglacial solutional opening, and these contain large numbers of small erratic blocks of Namurian sandstone.

Kamenitzas are numerous on the central area of massive pavement. They probably started to form soon after the Devensian ice retreat, but their water chemistry shows that they could form within only a fraction of Holocene time (Rose and

Figure 3.6 The very large clints in the central open pavements on Gait Barrows. (Photo: A.C. Waltham.)

Vincent, 1986b). Their solution environment, and consequent morphology varies under the influence of plant growth, ice formation and other factors. All the kamenitzas at Gait Barrows are formed over calcite veins in the limestone (Rose and Vincent, 1986b). It is not clear how the veins have become the loci for kamenitza development; they may have provided mechanical weaknesses scoured into hollows by overriding ice, or their mineralogical contrast may have become the focus of solution and cavity inception. Ultimately further solution on the veins beneath the pools creates fissures which drain the ponds and terminate the kamenitza enlargement. The pavements are also penetrated by small, deep potholes and elongate fissures, again on the mineral veins. Their shapes contrast those of the circular kamenitzas, but they also collect pool water and plant material which enhances their deepening; they ultimately coalesce into kluftkarren.

On the massive central pavements, islands of shrub and tree vegetation are rooted in organic soils; beneath and around these the limestone is scored by the rounded runnels of rundkarren which are typical of subsoil solution. Elsewhere on the site, less permeable, inorganic soils lie on uncorroded limestone which they have protected from solution. The influence of plants on limestone solution is also shown by the spring waters on the site. The solute loads of these vary directly with the proportion of soil and vegetation cover within their catchments (Rose and Vincent, 1986a); the lowest solute load is in the spring issuing from the downdip end of the main open, bare pavements (Figure 3.5). The springs do not discharge clastic sediments and no large conduits are known under the site.

Around the main intact pavements, large areas of broken rock outcrop are the result of removal of the top bed of limestone clints for the garden rockery trade (Ward and Evans, 1976; Goldie, 1986). The freshly exposed bedding planes are developing new solution features, where they are not being covered over by evolving vegetation and soil. Shallow depressions, some scalloping and sharp rillenkarren are forming on some of the stripped surfaces which have remained bare.

Conclusions

Gait Barrows has limestone pavements of exceptional quality, lying at low altitude and partly covered by vegetation; the large expanses of bare clints are of national and international repute. The coastal lowland environment is in contrast to that of the many pavements at high altitude in the Yorkshire Dales karst, but the morphological contrasts are slight. Within the site, the small-scale karst landforms show considerable variation, which is influenced by lithology, aspect and post-glacial evolution. Notable are the many kamenitzas at various stages of development, which are all located on mineral veins in the limestone.

HALE MOSS CAVES

Highlights

The caves of Hale Moss are the best examples in Britain of network caves which may have formed in narrow zones marginal to former lakes and within the range of their water table fluctuations.

Introduction

Network caves are a feature of the low limestone scars adjacent to the peat mosses in the lowland karst east of Morecambe Bay (Figure 3.1). The mosses occupy broad depressions in the limestone which may have originated as poljes enlarged by base level solution. They were modified by Pleistocene ice transgressions, and now have outwash fills (Oldfield, 1960) and subaerial drainage outlets across the limestone. Along the western margin of Hale Moss, bluffs of Dinantian limestone form the high ground adjacent to the peat bog; this is massively bedded but well jointed, and dips at up to 6°. There are ten short cave systems in the limestone; all are almost horizontal and lie within a narrow altitude range of 23–27 m.

The caves and karst of the area have been discussed by Ashmead (1969, 1974a) and Gale (1981a, b), and the caves are described by Brook *et al.* (1994).

Description

The main passage style in the known caves at Hale Moss is a joint controlled maze in the limestone adjacent to the margin of the peat moss (Figure 3.7). A few larger trunk passages extend further into the limestone bluffs, and some sections have developed as low, wide bedding plane passages. Most of the cave passages are less than 1 m high, and those formed on the joints are generally less than 1 m wide. Because of their small size and consequent inaccessibility, many of the joint networks have not been entered, and the trunk passages occupy an unduly large proportion of the mapped caves.

The ten known caves at Hale Moss have a total of over 1 km of passages, but this can represent only a fraction of the inaccessible, choked, unknown or undersized fissure networks in this zone of the karst. Hale Moss Cave is typical, with over 200 m of joint maze, joint guided trunk route and bedding passages, all on the same level in the dipping limestone (Figure 3.7). Hazel Grove Cave is a longer system of passages of similar style, and is formed on two levels, 2 m apart (Ashmead, 1969). All the other caves are shorter, and many are just fragments exposed by degradation of higher benches in the limestone.

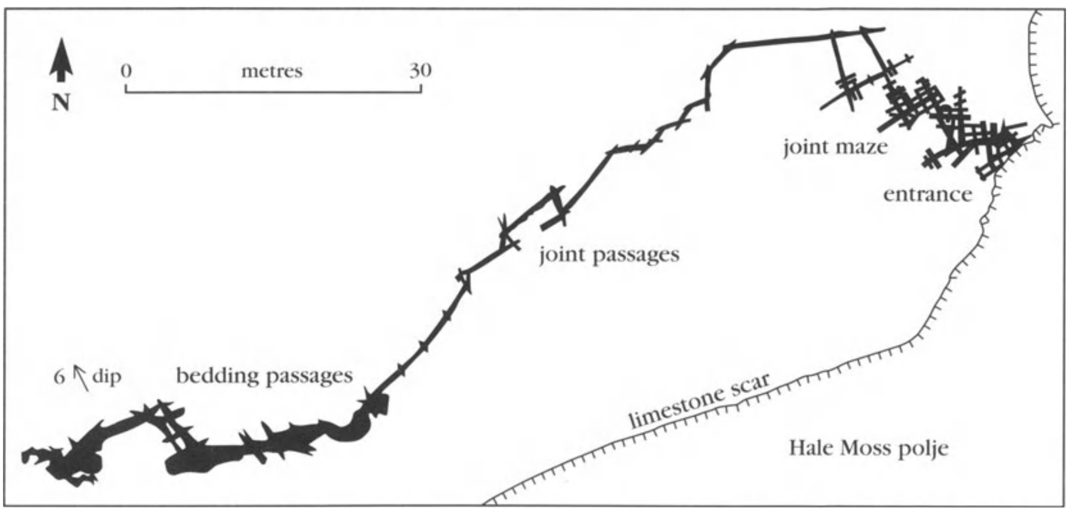


Figure 3.7 Outline map of Hale Moss Cave (from survey by Red Rose Cave and Pothole Club).

Interpretation

The caves of Hale Moss include excellent phreatic maze caves of the type normally developed by slowly moving water (Palmer, 1975). They are true three-dimensional mazes in that their rift openings have developed by solution of multiple intersecting joint systems and also of the bedding planes. Their macro-appearance as two dimensional mazes is only due to the solution being confined to narrow zones of altitude by external hydrological factors. Most of the cave passages lie at an altitude of 25–27 m, just below a poorly defined limestone bench; some passages in Hazel Grove Cave lie at a separate level about 2 m lower, which is very close to the top surface of the sediment fill within the Moss (Ashmead, 1974a).

There is debate over the environment of development of these cave systems. Ashmead (1969, 1974a) interpreted the Hale Moss caves as having developed close to the top of the phreatic zone by very slowly circulating water around the margins of the lake which formerly occupied Hale Moss. This accounts for the horizontal development of each cave, where preferential flow within the shallow phreatic network led to the development of the more linear trunk passages. Gale (1981a) considered the phreatic network caves to represent mere fragments of typical karst drainage systems formed by water flowing under hydrostatic pressure. Wall scallops are evidence of flowing water, small phreatic avens indicate higher contemporary water tables, and the horizontal development of the caves is related to flow along the strike.

The two hypotheses only differ in emphasis. The maze caves may be the only true examples in Britain of cave development at the water table adjacent to a lake margin, with their altitudes correlating with former lake levels. The linear caves associated with them indicate higher flow regimes away from the lake margins, in a more normal environment of efficient karst drainage. Hale Moss Cave has both types of passage (Figure 3.7). The morphology of these caves stands comparison with other sites. Maze caves, foot caves and trunk routes are all developed at water table levels in most areas of tropical limestone, perhaps typified by the Mulu karst in Borneo (Waltham and Brook, 1980a, b), but this has no implications of a tropical palaeokarst at Hale Moss. Horizontal networks of blind passages, without associated trunk caves, are developed in steeply dipping Carboniferous limestone around the margins of the Killarney lakes in Ireland (Priesnitz, 1985).

The ages of the inland terraces and perched lake levels east of Morecambe Bay are not yet known, but the presence of glaciofluvial fills indicates the pre-Devensian origins of the rock basins. The caves must have a similar age, and the multiple levels at Hazel Grove suggest a sequence of stages in their development. Glacial modification of the surface topography may have been slight in this lowland karst (Gale, 1984), but it has created access to the caves, while rejuvenation has led to their abandonment and fossilization.

Conclusion

The caves of Hale Moss contain mazes of phreatic passages which appear to be the result of solution by groundwater in the limestone margins of subaerial lakes, now filled with sediment and peat. They also have linear conduits which developed by faster karstic drainage away from the lakes. They are the finest of the many small caves in the Morecambe Bay karst, which are the only ones in Britain whose origins are directly related to past water levels in adjacent lakes.

SHORT GILL CAVERN

Highlights

The caves in Short Gill are the finest and most easily accessible of those formed in the nearly vertical limestone adjacent to the Dent Fault. Their tall rift passages along bedding planes are an unusual expression of stratigraphic controls on cave development.

Introduction

Short Gill Cavern lies under the eastern slopes of Barbondale, south-west of Dent (Figure 3.1). Short Gill is a tributary stream to Barkin Beck, which drains the glaciated trough of Barbondale along the line of the Dent Fault. Silurian mudstones and slates form the hills west of the fault. Carboniferous limestone crops out on the east side, and the upper slopes of Barbon High Fell consist of almost horizontal Yoredale limestones and shales. Drainage from these slopes feeds the streams, including Short Gill, which drain onto the narrow outcrop of Great Scar Limestone adjacent to the fault; nearly all of them have short associated cave systems. The Great Scar Limestone is steeply inclined against the fault, and the steepest dips of 75–90° are exposed in the narrow gorge of Short Gill. The Barbondale caves have been described by Sutcliffe (1974) and very briefly by Brook *et al.* (1994).

Description

The entrance to Short Gill Cave is through a narrow fissure in the floor of the gill, which is active only in times of flood. The entrance fissure drops into an abandoned phreatic tube, which is choked beneath the gill, but continues north only partially blocked by sediment banks, deep gour pools and a fine stalagmite false floor. It joins the main stream passage, carrying water from sinks at Short Gill Pot and south of the gill. This continues as a high narrow rift along the strike of the nearly vertical bedding. A high-level passage in the roof is choked with speleothems, and the stream enters a vadose canyon which zigzags to cut across the bedding. This descends to a sump where the flooded passage must head south, cutting further across the bedding and passing beneath the lower course of Short Gill to reach the resurgence in the floor of Barbondale.

The other significant cave in this site is Short Gill Pot, a short distance to the east, a 34 m deep rift developed along a washed-out vertical shale bed. Over most of its length it is 1–2 m wide, with some wider sections due to solutional enlargement. Water sinking here must subsequently utilize joints to cross through the bedding, to emerge at the upstream sump in Short Gill Cave.

Interpretation

These caves have developed in a structural setting, unusual in Britain, where water has drained down the vertical bedding to enter rift conduits along the strike. Drainage has then escaped via joint fissures cutting across the strike, to reach a resurgence in the valley floor which is also aligned on the strike. Shale beds lie on the main bedding planes utilized by the cave, and their mechanical removal has contributed to passage enlargement. Short Gill Pot is a nearly vertical rift 34 m deep in the bed of Short Gill, and is formed in a washed out shale bed with only limited solutional excavation of the limestone walls.

The relict phreatic tube in the entrance series of Short Gill Cave lies 20 m above the level of the Barbondale floor. Its alternating clastic and calcite sediment sequence, though undated, suggests that its history extends back into the Pleistocene, predating valley floor excavation to the present depth. Alternatively, it could have developed within a phreas perched behind the vertical limestone beds which clearly have lower transmissivity across their bedding than along the strike.

Conclusion

The caves and potholes of Short Gill are the finest of a small group in Barbondale, which are the only caves in Britain developed in nearly vertical limestone. Their morphology of vertical rifts along bedding planes and shale beds is therefore distinctive, and the shafts and strike conduits, both developed on the bedding, are an unusual expression of stratigraphic control over cave development.

UPPER DENTDALE CAVES

Highlights

The caves of Upper Dentdale constitute an excellent example of a karstic drainage system which has been developed and partly intersected beneath a limestone valley floor. The passages show strong geological controls in their structure, including cavern modification by collapse.

Introduction

The glaciated trough of Dentdale is cut just deep enough to expose the Great Scar Limestone in a stretch of its floor east of the village of Dent (Figure 3.1). The River Dee traverses the limestone outcrop in a shallow rocky gorge which is dry in normal weather for 1500 m, when all the flow is underground. Sinks and open cave entrances swallow the water into a major sub-valley conduit, where the main flow is joined by tributary streams sinking along the southern side of the dale (Figure 3.8). As the main caves, below river level, are frequently flooded, many passages and entrances are choked by debris, and only a fraction of them have yet been entered. The limestone outcrop lies across a very gentle anticline,

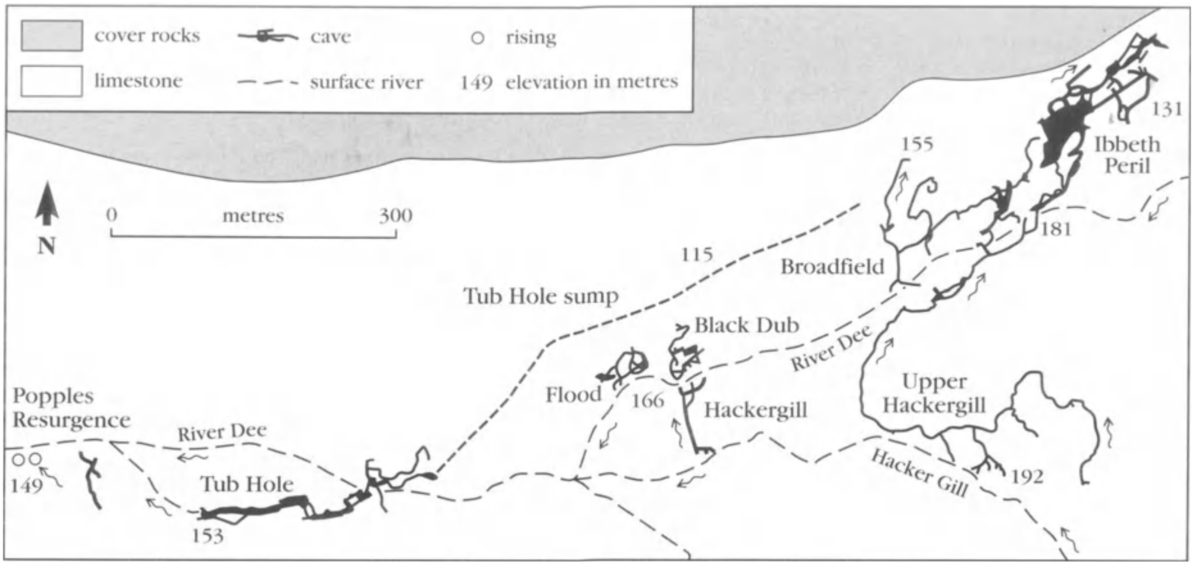


Figure 3.8 Outline map of the caves of Upper Dentdale. The line of the flooded section in the upstream sump of Tub Hole is only approximate (from surveys by Kendal Caving Club, British Speleological Association, Cave Diving Group and others).

plunging with the regional dip to the north, and is locally disrupted by steep dips in shatter zones; the carbonate succession is broken by shale beds up to 2 m thick, and includes the Gayle and Hawes Limestones which are contiguous with the Great Scar.

The geomorphology of the main caves in Dentdale was first described by Long (1971) and Lyon (1974) before further significant discoveries were made (Monico, 1992, 1995; Allwright *et al.*, 1993; Brook *et al.*, 1994; Holmes, 1994) and a number of speleothem dates were obtained from the site (Gascoyne *et al.*, 1983a, b; Gascoyne and Ford 1984).

Description

There are numerous sinks into rifts and bedding planes along the course of the River Dee. The furthest upstream are 1 km after it crosses onto the limestone outcrop, and they are spread over the next 2 km of riverbed as far as the Ibbeth Peril plunge pool, below a small waterfall which is normally dry. Water from all these sinks enters a conduit which is only partially explored and appears to be largely flooded; the sinks are at altitudes of 220 down to 185 m, the caves at Ibbeth Peril reach water levels at 155 m, and the Popples resurgence is at 148 m.

Most of the known caves in Dentdale are tributaries or flood distributaries of this main conduit, which has only been reached in the flooded section of Tub Hole (Figure 3.8). The Ibbeth Peril Caves have a main entrance beside the waterfall and plunge pool, from where a downstream passage heads north-east into the Main Chamber. This is one of the largest cave chambers in the Pennines, covering 30 m by 60 m with a height of up to 14 m; its floor is a chaos of massive limestone blocks which have fallen away from the roof. Several stream passages converge on the chamber, and the combined outlet drains north-east to a sump and a complex flooded zone. Flowstone overlying fluvial or fluvio-glacial sediments in an almost choked side passage is postglacial, giving dates of 6–29 ka (Gascoyne and Ford, 1984).

The pattern of cave streams draining downdip to the north-east, almost directly opposite to the surface valley flow, is repeated in the other active inlets. The aptly named Upstream Downstream Passage lies directly beneath the surface stream, yet the vadose cave stream flows in the opposite direction. A very old phreatic bedding passage, under the north side of the riverbed connects the inlets in the Ibbeth Peril and Broadfield Caves. Drainage from the upper sinks in Hacker Gill flows north through the Upper Hackergill Caves, which continue downstream as the Upstream Downstream Passage in Ibbeth Peril (Figure 3.8); a distributary takes some of the water under the river, into the small streamways in Broadfield Caves (Figure 3.9). Water from the lower sinks in

Figure 3.9 Tributary passage in Broadfield Cave with calcite deposits in a shallow vadose canyon cut beneath a bedding plane. (Photo: M.H. Long.)

Hacker Gill drains to risings in the south bank of the Dee riverbed and then sinks again into short caves on the north side, which are assumed to drain into the main conduit.

Just upstream of the impenetrable bedding planes of the Popples resurgence, a flood channel joins the River Dee from the Tub Hole rising. Feeding this flood resurgence, a wide cave passage has a flat roof left by extensive collapse of bedding slabs. Where the dry cave passes under the surface riverbed, holes lead down into a complex zone of flooded passages. The main phreatic tunnel extends downdip to reach depths of 34 m at the exploration limit almost beneath Broadfield Caves (Figure 3.8). In low flow conditions the water in this passage drains to the west, but in normal conditions the flow is to the east (Monico, 1992); and in flood conditions a massive flow pours from Tub Hole. The explored cave all appears to be part of a series of loops which form only part of the main conduit system; the complex flow patterns suggest that there are more, parallel conduits further downdip to the north.

Interpretation

The modern drainage of the floor of Dentdale feeds sinking streams which drain downdip through vadose caves, until they meet flooded conduits which carry the flow along the strike to a single resurgence. In this respect, the valley provides a perfect example of karstic drainage influenced by geological structure. Bedding planes, mostly marked by thin shale beds, have provided the main horizons of cave inception. Most of the tributary stream caves follow the limestone dip to the north-east, whereas the surface drainage is to the west. The main sub-valley conduit is unexplored, but it probably takes the line of a series of shallow phreatic loops following bedding planes under the northern, downdip side of the valley. Thick shale beds, and local zones of contorted and fractured limestone, in the Ibbeth Peril Caves created areas of weakness which were exploited by solutional undermining and collapse to create the large chambers.

In detail, the situation is more complex as many of the drained, vadose passages have features of phreatic morphology, suggesting that they developed before the riverbed cut down to its present level. Speleothem dates of up to 29 ka (Gascoyne and Ford, 1984) indicate that the shallower parts of the system had largely attained their present size, and had been drained, before the Devensian glaciation. The wide phreatic bedding passage connecting the two main inlets of the Broadfield Caves probably represents an earlier sub-valley drainage route towards the resurgence.

The large flooded passage in the Tub Hole sump appears to be the present main conduit, but it is a rising phreatic loop which is now being undercut by development of a lower route direct to the permanent resurgence. Base-level lowering progresses up the valley by successive erosion through the rising phreatic loops and complete or partial draining of the downloops. The gentle gradient of Dentdale ensures that this process is slow. Perched sump levels in and upstream of Broadfield Caves may correspond to ponding behind rising phreatic loops which have not yet been eliminated by vadose entrenchment.

Conclusion

The caves of Upper Dentdale provide clear examples of all stages of development of karstic drainage beneath a major valley floor. Both vadose and phreatic parts of the cave system show clearly the influence of geological controls on drainage routes, and earlier phases of phreatic cave have been drained, incised and partially filled by calcite, clastic sediment and collapse in the vadose environment.

STUMP CROSS CAVES

Highlights

The Stump Cross and Mongo Gill Caves contain a complex of active and abandoned passages closely related to geological structure within a plunging anticline crossed by faults and mineral veins. Calcite flowstones and fossiliferous clastic sediments record climate fluctuations during the late Pleistocene.

Introduction

The two cave systems of Stump Cross Caverns and Mongo Gill Hole are connected into a network with 5800 m of mapped passage. These all lie beneath the western end of Greenhow Hill, between Wharfedale and Nidderdale (Figure 3.1). The Dinantian limestone is heavily faulted and mineralized in a small inlier on the crest of an anticline, which is orientated almost east-west, immediately north of the North Craven Fault (Dunham and Stubblefield, 1945). The caves lie under the northern limb of the fold, where the limestone dips north and north-east at 15-30°. From the north and east, streams off the higher outcrop of the Namurian Grassington Grit sink into the limestone, and drain through the caves to the Timpony Joint resurgence in the floor of Dry Gill west of Stump Cross. The caves are largely accessible through mined shafts which intersected them when veins rich in lead ore were worked early in the last century.

The cave geomorphology is briefly reviewed by O'Connor *et al.* (1974), and subsequent speleothem dates have been obtained by Sutcliffe *et al.* (1985), Atkinson *et al.* (1986), Baker (1993) and Baker *et al.* (1996). The caves of Stump Cross were described by Cook (1950) and those of Mongo Gill by Judson (1964), and all the passages are documented in Brook *et al.* (1988).

Description

The combined cave system of Stump Cross Caverns and Mongo Gill Hole has a complex of passages (Figure 3.10) which roughly fall into three major levels of development, all within an altitude range of 370-317 m. Part of the upper level is operated as a show cave at the Stump Cross end of the system.

The upper, show cave, level has short sections of large phreatic tube containing extensive sediment deposits and a fine array of speleothems. Most of these fragments of old passage end in major collapses, some of which contain debris run in from the surface. Reindeer Cavern and the Wolverine Cave are named after the many bones of *Rangifer tarandus* and *Gulo gulo*, respectively, found in the sediments washed into them from surface fissures now impenetrably choked. Calcite stalagmites and flowstones from this upper level range in age up to 170 000 years (Sutcliffe *et al.* 1985; Atkinson *et al.* 1986; Baker *et al.* 1995c). Keep Chamber is an isolated fragment of this same passage level, well decorated with calcite and now only reached through a mined shaft. An upstream continuation of the abandoned level heads east through the Heaven passage, through some major chokes containing slumped clay, and into larger passages, with another mined entrance through Shockle Shaft. The upper level continues through Mongo Gill Hole, which has some of the largest tunnels and old phreatic chambers in the system. These are locally modified by roof collapse, and contain extensive clastic sediment fills; they also have

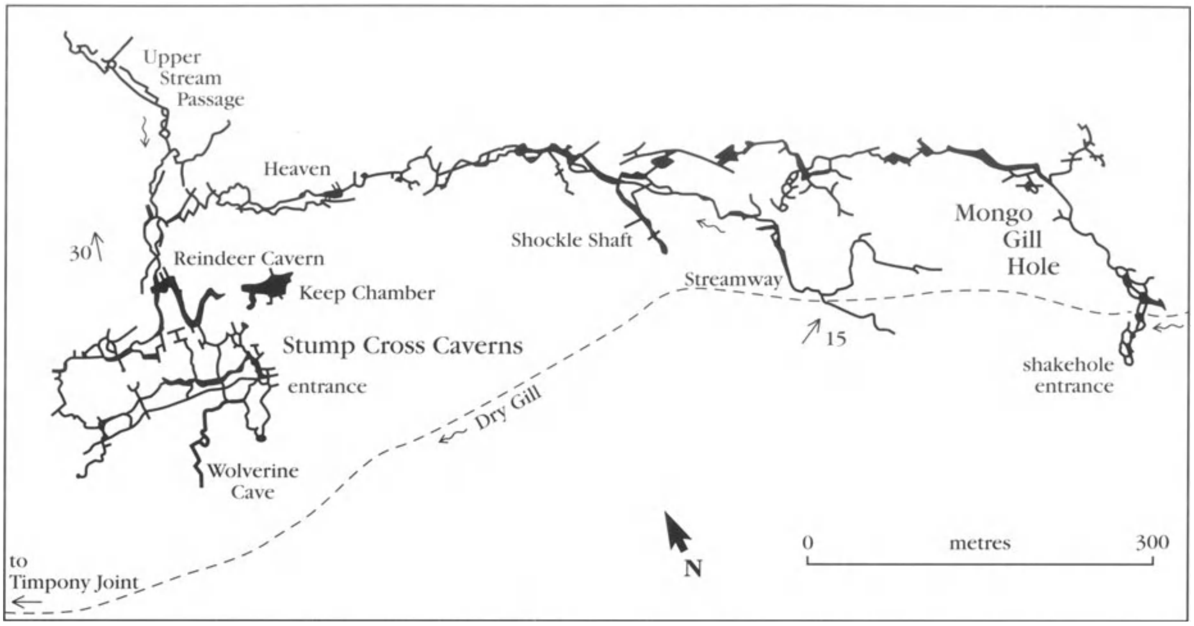


Figure 3.10 Outline map of the cave passages in Stump Cross Caverns and Mongo Gill Hole (from survey by Craven Pothole Club).

very beautiful calcite deposits, which were more abundant before the caves were invaded by the lead miners. The eastern end of Mongo Gill Hole swings round with the strike, to pass beneath Dry Gill to the only natural entrance in a doline shakehole.

The middle level of the cave system is a much more extensive, abandoned phreatic network of rifts, developed along joints and mineral veins, with sections of phreatic tube on some beddings. Most of these passages are at least half filled with fluvioglacial sediments, which are locally covered and interlayered with stalagmite.

The lower level of Stump Cross Caverns consists of constricted streamway canyons interrupted by short flooded sections. It drains from the Upper Stream Passage, south beneath the western end of the high level networks, and is fed by impenetrable sinks along the northern edge of the limestone outcrop. Water sinking in the upper reaches of Dry Gill drains through a low level streamway in Mongo Gill Hole (Figure 3.10); this also has alternating vadose and phreatic sections, but is now only active in flood conditions since the lead miners drove adits to lower the water table. The downstream end of the active system, behind the rising of Timpony Joint, remains permanently flooded.

Interpretation

The Stump Cross caves closely reflect the geological structure. The main abandoned trunk passages carried drainage to the west by following the strike of the bedding in a sweep around the anticline, which plunges east; this pattern is still recognizable in the remnants of choked and truncated passages shown on the cave survey (Figure 3.10). The main original cave drainage was through Mongo Gill Hole, but the old inlets and the earliest resurgence passages at high level were removed by glaciation (Judson, 1964). The modern drainage route through the lower levels of Mongo Gill has sections of vadose streamway between shallow phreatic loops aligned on joints and bedding planes.

Most of the phreatic passages follow the bedding but are aligned on the closely spaced fractures, and the mineral veins, within the limestone. The middle level of Stump Cross Caverns was formed partly by drainage from sinks in the north, which was ponded as it flowed against the dip to reach the main conduits and the outlets in Dry Gill. The main network of joint rifts was formed by this slow moving water, and rejuvenated sections are being entrenched by the modern streams between downloops which are

Figure 3.11 Thick flowstone deposits in a suite dated to 83 000 ka in the Wolverine Cave in Stump Cross Caverns. (Photo: A.C. Waltham.)

still flooded. At the resurgence, the flow is up the dip from a phreatic loop.

The sequence of passage levels and their subsequent modification indicate a long history of cave development, dated by speleothems from the early Devensian, and with older phreatic phases which must date back at least as far as the Hoxnian. Further dating of the older flowstones and their intercalated fluvioglacial sediments should be most significant at this site where the caves lie under the interfluve between two of the glaciated dales - Wharfedale and Nidderdale.

The speleothem dates already obtained imply that the expansion of the Devensian ice sheet into the area did not occur until after 26 000 years ago, which is contemporary with the ice advance over the Assynt karst. Stalagmite growth through the last 170 000 years, in the Wolverine Cave of Stump Cross, occurred only during interstadial phases, and ceased during full glacials and also during the interglacials (Figure 3.11). These interruptions are attributed to permafrost expansion during the glacial stages, and to flooding of the system during the warm stages; the latter are unusual, as most other sites have calcite deposition correlating with times of maximum solar insolation (Baker *et al.*, 1996).

Conclusion

The caves of Stump Cross and Mongo Gill are part of a complex, largely phreatic system, developed in folded Great Scar Limestone with close control by the geological structure. Some passages were drained only recently by mining activities. Others are much older and rejuvenated, and have considerable geomorphic significance for their interfluve location. Dated flowstones in Stump Cross Caverns record an unusual history of intermittent growth during the late Pleistocene, interrupted by both freezing and flooding.

NIDDERDALE CAVES

Highlights

The limestone inliers of the upper part of Nidderdale provide windows into a major cave system largely developed beneath the sandstone which forms the outcrop along the valley floor. The positions of shallow phreatic loops in the flooded zone of Goyden Pot are constrained within thin beds of limestone which cross a number of faults. Tributary caves exhibit further geological controls, and include some associated with the subaerial limestone gorge at How Stean.

Introduction

An important group of caves lies in the upper valley of Nidderdale, and its tributaries, upstream of Lofthouse (Figure 3.1). The easterly dip off the Pennine anticline carries the Great Scar Limestone well below the floor of Nidderdale, although it lies at the same altitude as Wharfedale, which is cut 150 m deep into the Great Scar. Nidderdale is cut largely into the Namurian Grassington Grit, which locally oversteps and cuts out much of the Brigantian succession of Yoredale beds. The valley floor reaches through the Grit to expose Yoredale limestone in a sequence of three inliers (Figure 3.11). In the north, the Limley inlier is an anticline confined within a triangular fault block. Lesser faults cross the valley downstream and a southerly upthrow on the Dry Wath fault returns the limestone to outcrop in the Thrope inlier, until it again dips very gently to the south beneath the Grit. The Lofthouse inlier is the largest of the three, has a gentle dip to the east, and is faulted along its southern margin. The limestone in Nidderdale forms a unit 40 m thick; most of this is the Middle Limestone, which is locally contiguous with the overlying Five Yard and Three Yard Limestones (Wilson, 1983).

Allogenic water from the Grassington Grit catchment to the north and west flows along the River Nidd until it sinks into the fractured limestone of the Limley inlier. Mild flood flows reach on the surface to the sink into Goyden Pot, but the river bed downstream is normally dry as far as the resurgence inflows in the Lofthouse inlier. From Manchester Hole to Nidd Heads the underground drainage route is within the limestone, which is about 40 m thick. The caves of How Stean Beck and Blayshaw Gill lie in the same limestone in the Lofthouse inlier.

The geology and geomorphology of the Nidderdale caves have been described by T.D. Ford (1964b) and Davies (1974a), and passage development in New Goyden Pot was further discussed by Davies (1974b). Descriptions of the cave passages are given by Yates (1934), Brindle (1956) and Brook *et al.* (1988), and of the flooded caves between Goyden and Nidd Heads in Monico (1995).

Description

The caves of the River Nidd

Goyden Pot is the main cave on the River Nidd and has flooded connections to Manchester Hole and New Goyden Pot, creating a single system downstream from the sink (Figure 3.12). This has a mapped length of more than 6.3 km, descending 61 m, and another 1.2 km of flooded passage has been explored upstream from the Nidd Heads resurgence.

Under normal flow conditions the River Nidd sinks in fissures in the Middle Limestone on the northern side of the Limley anticline to enter the main river passage of Manchester Hole. This is a single large canyon, up to 12 m high and 6 m wide, extending south for 500 m to a sump. Over the crest of the Limley anticline, the cave river has breached the base of the Middle Limestone exposing the underlying shale and the Simonstone Limestone in the canyon walls. The main chamber of the cave is heavily modified by massive block collapse.

The downstream sump is a short phreatic loop which ends at multiple outlets into the open cave passages of Goyden Pot, which are joined by a large flood route from the gaping entrance in the surface river channel. Following the western edge of the Goyden Pot network, the River Passage is a splendid, wide canyon strewn with sandstone boulders. This descends 35 m by following the bedding obliquely down dip; fractures guide it around sweeping loops, with chert nodules projecting from the walls, beneath a sloping bedding roof, and then downdip into a sump. East of the River Passage, the Labyrinth is a sloping network of small phreatic tubes, chambers and tall rifts, lying updip in the same bedding planes (Figure 3.12). This area is now largely inactive, except where an underfit stream flows in the New Stream Passage before entering a flooded phreatic loop through to New Goyden Pot. Many of the Labyrinth passages are choked with clastic sediments, and there are some calcite speleothems which show evidence of erosion and re-solution. Upstream of the sump in the River Passage, high-level rifts provide a dry route into another short section of canyon passage, and then into a series of shallow phreatic loops.

New Goyden Pot contains another section of the underground River Nidd flowing through large passages, which are reached by two shafts dropping down a fault from a small entrance in the surface river bed (Figures 3.12 and 3.13). The river emerges from shallow phreatic loops where the cave passage follows almost horizontal bedding horizons within each fault block (Figure 3.13), and steps in their profile lie where each fault is crossed (Davies, 1974b). The long crest of a loop between the Dry Wath and Thrope Edge Faults has the gently graded stream flowing along the floor of rejuvenated phreatic tunnels which take a large double bend towards the east, collecting the tributary flow from the Goyden Pot New Stream (Figure 3.12).

South of the Thrope Edge Fault the underground River Nidd enters further shallow phreatic loops, separated by very short lengths of vadose cave over some of the loop crests. Banks of gravel and cobbles are common in the downloops, and almost block the flooded passage at the present limit of exploration. It is likely that the continuing passage is mostly underwater, but sections of descending open streamway must occur as the water levels in downstream New Goyden are a few metres above the level in Nidd Heads. Upstream from the resurgence, a complex of passages to the twin risings are distributaries from a trunk conduit, which has been followed for over 700 m without meeting airspace. Most of this flooded cave is nearly level, but in at least two places the flow rises steeply up rifts from depths of more than 30 m.

The caves of How Stean Beck

How Stean Beck is a tributary of the River Nidd, which flows through a spectacular limestone gorge up to 15 m deep, currently operated as a commercial tourist attraction. On the north side of the gorge, a group of caves contains almost 2.5 km of mapped passages, traversed by underfit streams. Eglin's Hole lies upstream of Low Eglin's Hole (only the latter appears on Figure 3.12), to form a single, linear, vadose system draining down the limestone dip, partly guided by chert beds. Both caves have low bedding plane passages, locally up to 15 m wide where loops and inlets coalesce. Some of the

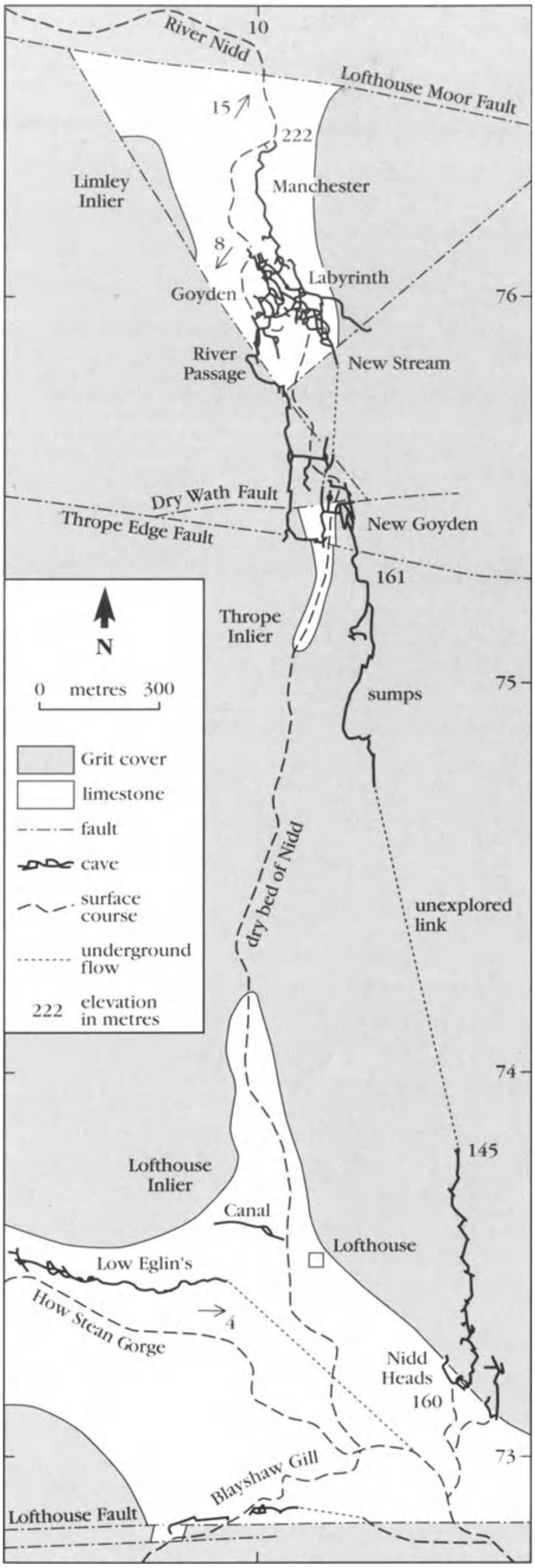


Figure 3.12 Geological map of the caves in the upper part of Nidderdale. Limestone includes the Middle, Five Yard and Three Yard Limestones. Grit cover is the Grassington Grit and some overlying Namurian beds on the higher slopes. Eglin's Cave extends off the map to the west. (Outcrop geology after Wilson, 1983; cave surveys from Yorkshire Underground Research Team, Cave Diving Group and others.)

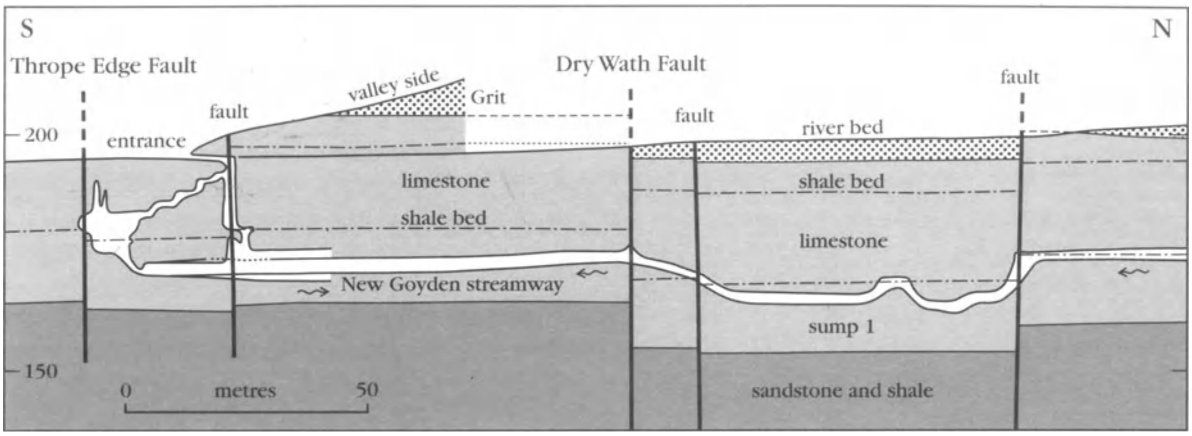


Figure 3.13 Profile of the geology and cave passages in the western part of New Goyden Pot. The main streamway flows south towards the Thrope Edge Fault, and then turns into a loop back to the north which is not shown in this profile downstream of the entrance shaft. The grit is the Grassington Grit; the limestone includes the Middle, Five Yard and Three Yard Limestones; the sandstones and shales are of the lower Brigantian and include the Simonstone Limestone at an unknown depth. (After Davies, 1974b.)

bedding openings have shallow vadose floor trenches, and others are choked with boulders or blocked by roof collapse. The main water enters from choked sinks which are minor leaks from How Stean Beck. The vadose caves descend nearly 50 m to a sump where the bedding plane passage continues below the level of the resurgence, which is a rising through the Nidd alluvium.

The How Stean Gorge is a subaerial canyon which carries a larger stream flow than the parallel caves; its floor lies about 10 m below the caves. A single tributary on its south side flows for 50 m through How Stean Tunnel, and Tom Taylor's Cave is a rift on the north side carrying flood flows from the Eglin's caves (Waltham, 1984; Brook *et al.*, 1988).

The caves of Blayshaw Gill

Adjacent to the mineralized faults on the south side of the Lofthouse inlier, the two potholes in Blayshaw Gill enter fragments of an old phreatic cave intersected by the modern valley (Figure 3.12). Multiple levels add to a total of more than 950 m of mapped passage, much of it now choked with ochreous clay and boulder falls. Blayshaw Beck sinks into the Five Yard Limestone and a well developed cave passage crosses a fault directly into the Middle Limestone, where it continues following bedding planes downdip to the east. The modern stream has cut a vadose canyon in the floor of the older cave, as far as a sump at the level of the alluviated resurgence 30 m below the sinks.

Interpretation

The caves between Goyden Pot and Nidd Heads provide a classic example of karstic drainage beneath a major valley floor which only carries a surface flow in high flood conditions. Much of the cave system lies beneath the outcrop of the Grassington Grit, between limestone inliers at the sink and resurgence, but the geology of Nidderdale is unlike that of the other Yorkshire dales. Cave inception in the buried limestone was probably accelerated by seepage of aggressive water from the surface soils down though the permeable sandstone, as the unconformity at the base of the Grassington Grit has locally cut out the intervening Brigantian shales.

All the caves in the valley show strong control by bedding planes within the limestone unit which is only 40 m thick. The extensive phreatic Labyrinth in Goyden Pot is formed on dipping bedding planes, and the vadose caves beside How Stean Beck drain directly downdip. The major cave passage carrying the River Nidd drains obliquely down the dip in Goyden Pot, and takes a circuitous course through almost horizontal limestone in New Goyden Pot. In both cases, segments of the cave passages are aligned on joints which are enlarged into tall rift fissures and chambers.

The many faults across the line of the underground drainage have been utilized to link sections of nearly horizontal cave conduit formed on just a few inception horizons within the limestone. This has created a stepped profile in the shallow, phreatic loops across downfaulted blocks (Figure

3.13), in a pattern rarely seen because of the intrinsic inaccessibility of these caves within the phreas: it is in marked contrast to the deep loops in the steeply dipping limestone off the Mendip Hills. Along the underground River Nidd some of the fault planes have proved to be sites of cave inception, linking the bedding horizons. In contrast, the caves of Blayshaw Gill cut cleanly across a fault plane, whose only role has been to bring two limestone beds into hydrological continuity.

The history of development of the Nidderdale caves is clearly long, as there are many series of abandoned and rejuvenated phreatic passages, but cannot be tied to a chronology of absolute dates. Surface modification during the climatic fluctuations of the Pleistocene included the rejuvenation of How Stean Beck. This appears to represent an unusual case where an underground drainage route, through the Eglin's caves, was abandoned in favour of a surface course. How Stean Gorge is a splendid limestone trench, which is effectively a vadose canyon passage without a roof. It may have origins from periglacial regimes when high flows of meltwater were constrained from sinking by permafrost sealing of the limestone fissures. Its continued subaerial entrenchment in postglacial conditions was probably aided by a gorge thalweg mostly steeper than the limestone dip, so that bedding planes could not offer hydrologically favourable routes for underground capture.

Conclusion

Nidderdale contains a major cave system carrying the entire valley drainage. Large vadose canyons descend from the main sinks to the head of a long phreatic series, where a staircase of flooded down-loops is interrupted by short sections of vadose cave over loop crests dictated by geological structure. The influence of faults on the pattern of conduit development through the karstic aquifer is seen more clearly than anywhere else in Britain. The tributary of How Stean Beck has a series of vadose caves which have lost their drainage to a subaerial gorge, in a reversal of the usual role of underground capture in a youthful karst landscape.

HELL GILL

Highlights

The section of Hell Gill cut into the limestone is one of the finest examples in Britain of a gorge formed entirely by subaerial fluvial action. It admirably demonstrates the role of subaerial erosion in the formation of gorges in karst areas.

Introduction

Hell Gill Beck is the largest of the streams draining off the Namurian shales and sandstone of Mallerstang Common, and is the highest head-stream of the River Eden (Figure 3.1). It cuts through the horizontal Main Limestone in a deep and narrow ravine. This is clearly of subaerial origin, and yet lies adjacent to caves formed by parallel streams. Except for its lack of roof, the gorge is very similar in morphology to vadose canyons in cave systems, and clearly shows the similarity between subaerial limestone gorges and underground cave passages. Although very impressive and locally well known (Waltham, 1984), Hell Gill has not been studied and documented in any serious investigation. The adjacent caves are recorded by Brook *et al.* (1994).

Description

The Hell Gill gorge is cut into the Namurian Main Limestone by a stream draining off a hillside of horizontal, interbedded shales, sandstones and thinner limestones within the Carboniferous Alston Group (Figure 3.14). The Main Limestone is the thickest limestone exposed on the fell, but is only 20 m thick. Hell Gill descends between altitudes of 425 and 395 m, following the limestone dip; it is a narrow, twisting, rock gorge 500 m long, mostly less than 5 m wide, and up to 15 m deep. The stream descends steadily through moulins and connecting trenches, and cascades over three small falls into deep, round plunge pools. The sides of the gorge are mainly vertical, smooth and polished, with the stream occupying the entire floor width, locally undercutting the walls in deep swirls. Immature cave development is represented by various short rifts, and some of the flow passes through short parallel loops which are intersected phreatic fissures; a rock bridge stands across the gorge where the stream has enlarged an underground short cut. At the lower end of the gorge, the stream flows onto the underlying sandstone, and then breaks out from the low limestone scar into the valley.

Many of the neighbouring, parallel streams on the fellside sink underground where they reach

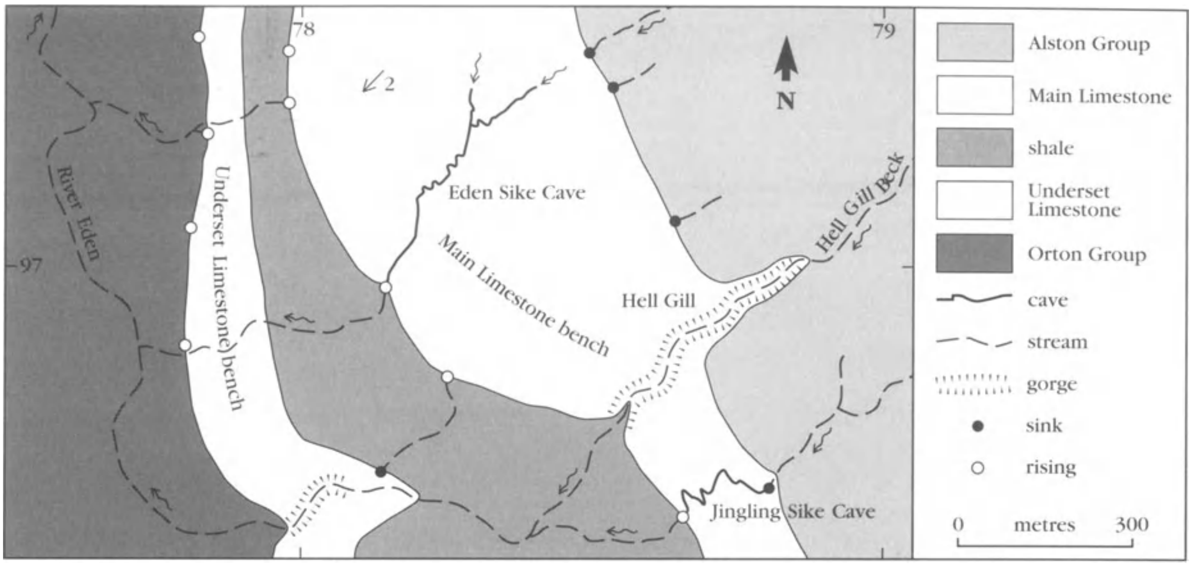


Figure 3.14 Outline map of the limestone bench containing the Hell Gill gorge and various sinkholes, risings and cave on the adjacent streams.

the top of the Main Limestone, and resurge several hundred metres to the south-west at the base of the same limestone outcrop (Figure 3.14). Jingling Sike goes underground through 300 m of cave just to the south, and streams feeding Eden Sike flow through 770 m of cave just to the north, both in the Main Limestone. The streams have cut subaerial ravines through most of the thinner limestone bands on the fell, and Hell Gill Beck flows through a shallow rock canyon cut in the Underset Limestone (Figure 3.14). Percolation water feeds small risings at the base of all the limestones.

Interpretation

There is no evidence in the Hell Gill gorge of any cave roof or wall collapse, currently or in the past. Cave development in the gorge is limited to flow through short fissures in the immediate walls and floor, which is part of the normal mechanism of entrenchment in a limestone river bed. The location, and the dimensions consistent with the modern flow of Hell Gill Beck, suggest that the ravine is entirely a surface feature cut by the stream which it still contains. Its youthfulness, the absence of fill, and the lack of deep weathering in its walls suggest that it was excavated during the Pleistocene, though it may have been initiated by meltwater flow as the Devensian glaciers retreated from the area.

Apart from details of plant colonization and minor weathering on its upper walls, the gorge is very similar to many large vadose canyons in the Pennine cave systems. The ability of Hell Gill Beck to maintain its surface course across the limestone outcrop is the result of its high discharge. This ensures that the floor of the gorge is lowered by solution and mechanical abrasion fast enough to unroof, expose and incorporate fissure openings developed in its floor by slow solution alone. The smaller parallel streams of Eden and Jingling Sikes have not been able to entrench their beds fast enough, and have subsequently been captured by underground drainage forming small cave systems of joint rifts and bedding passages with little or no collapse.

Conclusions

The gorge on Hell Gill Beck provides an excellent example of subaerial fluvial action in a karst terrain. It is especially significant as it can clearly be demonstrated to be of subaerial origin, and yet lies adjacent to caves cut by smaller, parallel streams through the same limestone. It also demonstrates the similarity between surface fluvial gorges and underground vadose canyons, and has important implications for the understanding of process in limestone gorges.

CLIFF FORCE CAVE AND THE BUTTERTUBS

Highlights

Cliff Force Cave is the finest known example of an inter-dales cave, in which geological structure has allowed the underground drainage to cross beneath a major surface watershed. The Buttertubs are a series of spectacular potholes developed by small sinking streams meeting a thin limestone, with open shafts descending directly to the underlying aquiclude.

Introduction

The well known Buttertubs potholes and the entrance to Cliff Force Cave lie within 150 m of each other on opposite sides of the Cliff Beck valley. This is cut into the steep fells of Muker Common, descending from the Buttertubs Pass on the south side of Swaledale (Figure 3.1). Cliff force Cave has a long stream passage with drainage from the Wensleydale slopes passing beneath the topographic divide. The Buttertubs swallow small streams from the valley sides into large entrance shafts which belie the very short distances to their resurgence. Both features are developed where the Namurian Main Limestone is exposed in the valley sides (Figure 3.15); this forms a bed about 25 m thick, with clastic sedimentary rocks both above and below. The sequence dips very gently to the north-west and is broken by mineralized faults orientated north-east.

Cliff Force Cave has been documented briefly by Langthorne (1976), Clough and Clough (1981), Ryder (1981) and Brook *et al.* (1988). The Buttertubs are widely cited as examples of limestone potholes, but further comment on their morphology is only brief (Waltham, 1984).

Description

Cliff Force Cave contains about 2000 m of passages reached through an abandoned exit just above the present resurgence (Figure 3.15). The Entrance Series consists of narrow, joint-guided, phreatic rifts and avens, partly blocked by collapse and fluvioglacial fill; they lie above a short inaccessible section of the stream route. Beyond them the Lower Streamway is a vadose canyon, initially more than 3 m high and wide, developing into a narrow rift with undercuts at stream level, oxbow loops and a few collapse chambers. Above the streamway, the Spar Shop Series is an abandoned gallery largely formed as a sequence of interconnected rifts and now partly filled with clastic sediment. The two levels join in Fault Hall, a chamber 15 m in diameter developed on a mineralized fault. The stream cascades into Fault Hall

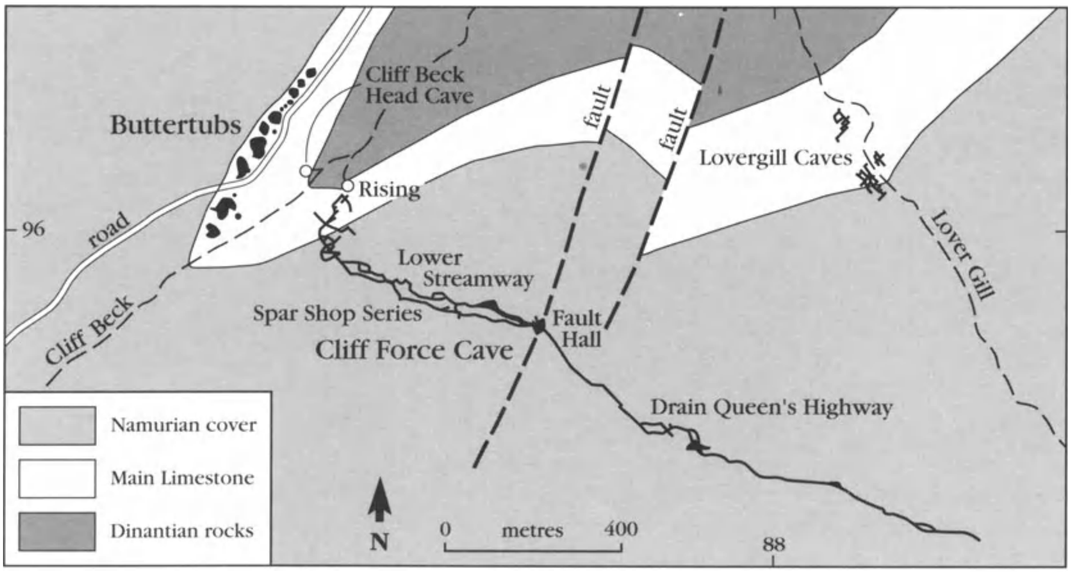


Figure 3.15 Geological map of Cliff Force Cave and the Buttertubs. Both the Namurian cover and the underlying Dinantian rocks include thin limestones not shown and not connected to the Main Limestone (cave survey from Moldywarps Speleological Group).

Figure 3.16 The fluted potholes and limestone pinnacles at the Buttertubs. (Photo: A.C. Waltham.)

from the Drain Queen's Highway, a phreatic passage mostly 3 m high and wide with prominent rock flakes projecting from the walls. The Room of Dangling Doom is the start of a long zone of heavily collapsed phreatic passage developed just below a thin mudstone which is exposed in parts of the roof. Further passages are flooded, but the water has been traced from a group of sinks in the Main Limestone at Sargill, nearly 3 km south-east of their resurgence. These lie in valleys whose surface thalwegs descend into Wensleydale.

Where Lover Gill crosses the Main Limestone, north of Cliff Force Cave (Figure 3.15), an old phreatic maze cave has been intersected, and some drainage now passes through truncated fragments of the maze preserved in both banks of the gill (Brook *et al.*, 1988).

The Buttertubs are a series of open vadose shafts in a narrow bench formed on the top of the Main Limestone along the west side of the Cliff Beck valley (Figure 3.15). In wet weather they take small streams from the peat-covered shale slopes above, but their catchments are very small. There are five main shafts, with clean, fluted walls; these have been deeply crenulated by waterfall retreat, to form interconnected series of shafts and slots between remnant pinnacles of limestone (Figure 3.16). They are all 15-20 m deep, to floors of fallen limestone blocks and clastic sediment, and their streams drain either through the floor debris or into impenetrable wall fissures. All this water resurges from Cliff Beck Head Cave, which lies directly down the bank from the Buttertubs (Figure 3.15), about 25 m lower down. This has small converging stream passages formed on joint/bedding intersections close to the base of the Main Limestone, and sandstone is exposed in the beck just below the rising.

Interpretation

Cliff Force Cave is one of a number of underground drainage routes carrying flow beneath the major surface watershed between Swaledale and Wensleydale (Ryder, 1975). It demonstrates clearly the fundamental control by geological structure on karst drainage routes; the Sargill streams flow south-east on the surface towards Wensleydale, until they sink underground and flow north down the dip towards Swaledale. Geological controls have influenced the cave in many ways. A few bedding planes and shale bands were the inception horizons for the whole cave, but the extensive collapse in the upstream sections has occurred beneath a thin, incompetent mudstone band; this was clearly not an inception horizon as it lies above the solutional features within the roof. Joint fissures are a part of the cave morphology, but the stream route has not encountered any old phreatic networks comparable to the maze fragments breached by Lover Gill. Phreatic solution and collapse along the mineralized fault which crosses the drainage line has produced the large Fault Hall; however, the cave has continued to follow the bedding, perhaps because the fault has none of the older phreatic rifts on the scale of those on faults in the karst of north Wales and the Peak District.

Unlike many caves in the thin limestones of the northern Pennines, Cliff Force Cave has evolved through several levels of passage development. Upstream of Fault Hall there is only a single phreatic passage, barely modified since it was partially drained. Downstream of Fault Hall there are three levels, of which only the active streamway is lower than the upstream passage. For much of the cave's history, water in the upstream passage was ponded behind a small phreatic lift on the fault. The highest level is only represented by passages close to the resurgence entrance; these may have been reached by a second phreatic lift, perhaps part of an old vauclusian resurgence system now obscured by glacial debris on the valley side. The second level is represented by the Spar Shop Series feeding to the dry passages and the present entrance; this was also active via the Fault Hall phreatic lift. The active streamway has a third level of initial phreatic development, followed by rejuvenation and vadose entrenchment.

The three cave levels are all primarily phreatic, and appear to reflect successive stages of phreatic flow on lower inception horizons within a maturing aquifer. Only the last stage of vadose incision can be directly related to lowering of the resurgence. A scarcity of calcite speleothems in the cave is a consequence of the location beneath an impermeable shale cover, and no evidence of absolute ages is yet available. Though the stages may relate to the glacial excavation of Swaledale, it would be premature to relate rejuvenations to valley floor positions, when the resurgence position may have been so easily influenced by details of geological structure in such a narrow limestone outcrop. The inception stage for the development of this cave may have been exceptionally long, as it lies in a thin limestone with non-carbonate rocks both above and below; these would have excluded infiltration flows of soil water, though the underlying sandstone may have carried some primary groundwater flow. The lithology of the Main Limestone may include features very pertinent to cave inception and karstic evolution in carbonates not directly exposed to the surface.

The Buttertubs present a striking contrast to Cliff Force Cave. They are classic examples of invasion vadose shafts developed close to a steep valley side, with horizontal cave development only at the base of the limestone where the underlying aquiclude has perched the groundwater flow. Relaxation opening of the limestone fractures, towards the destressed hillside, may have accelerated early development of the shafts, whose position and lack of fill are essentially post-Devensian. The deeply fluted shafts demonstrate the role of small flows of corrosive water rich in carbon dioxide and organic acids from soil and peat. In both these respects, the Buttertubs are not typical of the many Pennine potholes which are at large stream sinks feeding cave drainage routes to distant resurgences. They are, however, spectacular and very accessible karstic shafts.

Conclusion

Cliff Force Cave is the prime example of a cave developed in a thin limestone by drainage passing beneath a major surface watershed. Geological controls, both structural and lithological, have influenced the configuration and morphology of the cave, producing a cave system on multiple phreatic levels, which is unusual in the thinner limestones of the northern Pennines. The Buttertubs are the product of entirely vadose development in the same limestone, with simple vertical shafts descending to the underlying aquiclude and draining rapidly into the adjacent valley.

THE CLOUDS

Highlights

Three limestone pavements at Stennerskeugh and Fell End Clouds are notable for their location on strongly folded limestones. The structural variety and the consequent range of aspects are reflected in the varied morphologies of the well dissected and well runnelled pavement landscape.

Introduction

The pavements of the Clouds lie at altitudes of 350–470 m on outcrops of Dinantian Limestone on the north-western slopes of Wild Boar Fell, adjacent to the Dent Fault (Figure 3.1). The exposed limestones are the top beds of the Asbian Great Scar Limestone and also the Robinson Limestone, which is the lowest in the Brigantian shale and limestone sequence. The folding is all within the narrow belt where the western edge of the Askrigg Block is crumpled against the Lake District block across the Dent Fault (Underhill *et al.*, 1988).

Description

Most of the pavements are on the eastern limb of a minor anticline (Figure 3.17), where the limestones dip into the hill at 10–30°. The Great Scar Limestone has the wider outcrop, and is overlain by a thin shale, followed by the Robinson Limestone. The stratigraphy shows some variation along the strike outcrops, but these two limestones form the main pavements, which ease in gradient over the anticline in the Great Scar Limestone. The folded limestones support pavements on many bedding planes with varying aspects to the Devensian ice which swept over them from the south. The lower slopes are masked in scree, head and peat soils. The slopes above the main pavements are formed in the overlying mixed Brigantian rocks of the Alston Group, with low scars, thin strips of pavement, lines of

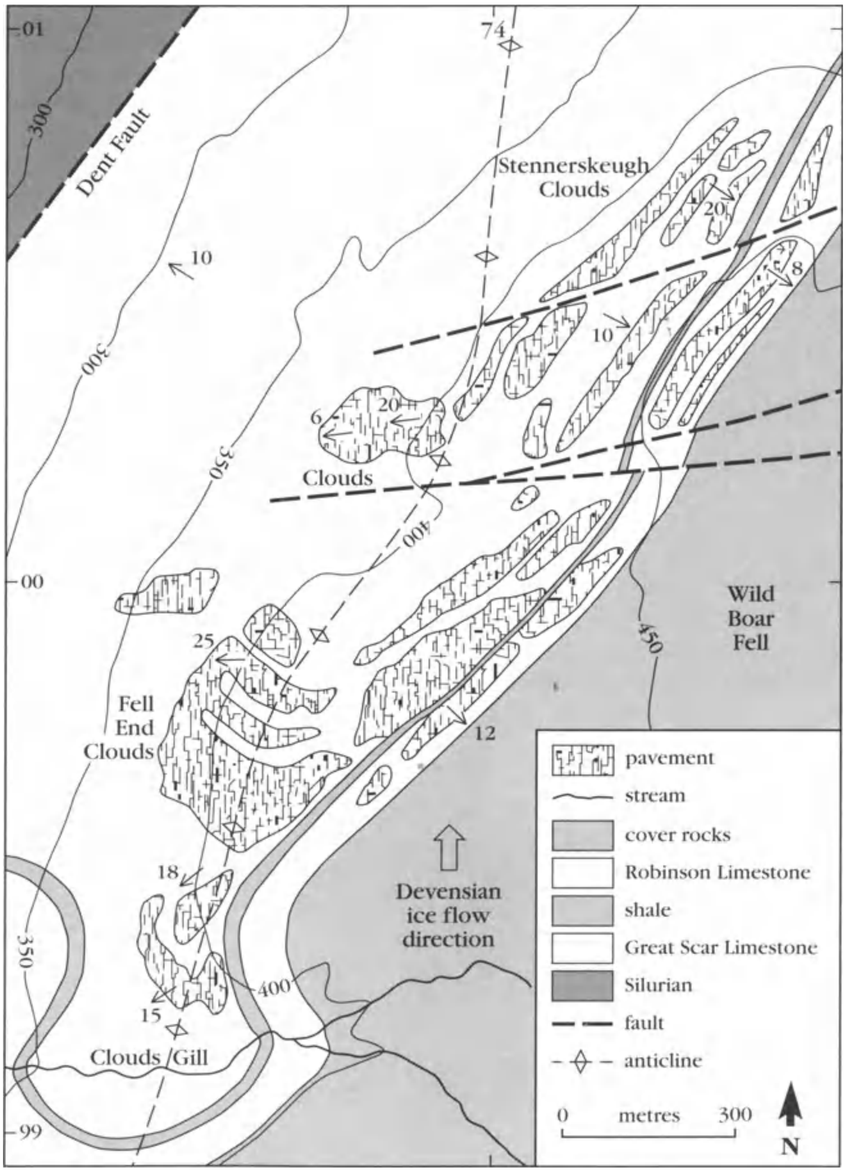


Figure 3.17 Outline map of the limestone pavements of the Clouds. The position of the anticline axis is only approximate. Cover rocks are the alternating sequence of shales and thin limestones which follow in the Alston Group.

Figure 3.18 Crescentic scars in strong beds of limestone folded over the anticline crest on Fell End Clouds. Pleistocene ice moved from right to left, leaving deeply runnelled pavement in the immediate lee of the scars. (Photo: H.S. Goldie.)

sinkholes and small caves in the thin limestones.

A distinctive form of stepped pavement (*Schichttreppenkarst*) is well developed on Stennerskeugh Clouds where steeply dipping limestones in the higher outcrops form long ridges of narrow pavements sloping to the south-east. Clints are large and diamond shaped at the northern end but decrease in size to knife-edge ridges further south. Deep narrow grikes separate the larger clints, and surface solution features include converging rundkarren systems and kamenitzas which vary in shape from elliptical to round. The small, knife-edge clints are separated by shallow grikes, and have both laminar and honeycomb weathering. A few erratic blocks of sandstone lie wedged in the grikes. Further east on Stennerskeugh Clouds, the stepped pavements dip at angles up to 30°, with considerable local variation across the small-scale folds. Long rectangular clints in the north are replaced southwards by narrower knife-edge clints.

Most of the limestone pavements on the west side of the Clouds are on bedding planes dipping west at up to 30°, with joints aligned diagonally to the slope. The clints are generally less than 1 m long and 0.3 m wide. Higher on the fell, the dips lessen over the anticline and some of the exposed limestone beds are more massive, so forming larger clints.

At Fell End Clouds, the strike and outcrops of the beds swing round the anticline which plunges to the south-west (Figure 3.17). The more massive limestones are higher in altitude and lower in the sequence to the north-east. Some of the more thinly bedded limestones produces a shattered surface of felsenmeer, in place of a pavement. The steeper fold limbs have stepped pavements dipping as much as 30°, with small, knife-edge clints between narrow, shallow, V-shaped grikes containing wedged sandstone erratics. Near the highest part of Fell End Clouds, there are some striking embayments in gently dipping limestone over the crest of the anticline. Small scars, 2–3 m high, mark the inner edges of crescentic arcs of pavement with massive, well runnelled clints (Figure 3.18). The inner zones of these pavements, below the succeeding scar, are scored by large and deep rundkarren; the outer zones, at the top of the scar to the next bed below, are less well runnelled and are merely well fractured. There is no significant cover of drift or vegetation, and the dominant joints are orientated NNE parallel to the Dent Fault.

Interpretation

Pavement morphologies at the Clouds are influenced by both lithology and geological structure. Thinly bedded limestones produce shattered surfaces with thin, easily broken clint tops; debris from these fills the grikes and extends into scree aprons and sheets of felsenmeer. In contrast, thick beds of massive limestone form better pavements; even these are generally well dissected, with relatively small clints, due to the closely spaced tectonic jointing in the disturbance zone adjacent to the Dent Fault. This limits complex runnel development, though many small clints are incised by deep rundkarren of the type inherited from solution beneath a soil and vegetation cover. Low scars, narrow benches, knife-edge clints and areas of stepped pavement are all features dictated by the steep local dips.

An important influence on the morphology of the Clouds pavements was their location close to a Pleistocene ice centre on Wild Boar Fell, immediately to the south-east. The scale of scour from such a nearby and small ice source may not have been great (Mitchell, 1994), and the Clouds pavements may have escaped severe glacial scour. This would explain the presence of the very mature rundkarren in front of the small scars near the top of Fell End Clouds. At this site the pavement edges have not been plucked and scoured to form the smooth scar edges typical of much of the Pennine glaciokarst; instead they have partly bevelled slopes from one bed to the next, with weathered and runnelled limestone preserved on the inclined bedding planes in the protected troughs below the next scar. This karren distribution contrasts with the typical Pennine case where the deeper karren are in the more exposed sites above the scar edges. The Fell End Clouds scars face north, so that the ice advanced down over them, and the well dissected pavements in their lee may retain some elements of early Pleistocene erosion not removed by Devensian ice.

Conclusions

The Clouds contains a fine range of well dissected pavements demonstrating the influences of structure and lithology on the details of karst morphology. The combination of tight folds and closely spaced fractures provides a geological environment for the pavements which is unique within Britain. The site was close to a basal ice shed in the Pleistocene, which resulted in some elements of the landform features surviving from before the Devensian.

GREAT ASBY SCAR

Highlights

Spectacular expanses of limestone pavement lie on the dip slope of Carboniferous limestone forming the Orton–Asby escarpment. The pavements are the most extensive outside the Ingleborough karst, and have a wide variety of well developed pavement morphologies.

Introduction

Asbian limestones from the Lower Carboniferous succession form a long, rounded escarpment north of the Howgill Fells, where they dip off the Lake District dome beneath younger rocks flooring the Vale of Eden (Figure 3.1). The crest of the south-facing escarpment reaches an altitude of only 412 m; the main pavements, separated by grassland grazed by sheep, extend down into farmland to the north. The regional dip of 5–10° north is interrupted by shallow flexures with axes aligned roughly downdip. All the pavements are formed on the massively bedded limestones of the Asbian Great Scar. They are underlain by the mixed sedimentary sequence of the Holkerian Orton Group; limestone beds within the Orton Group are the cause of the numerous shakeholes in the lowland below and south of the Great Scar escarpment, but they form no pavements. Brigantian shales and limestones of the Yoredale facies overlie the Great Scar; the lowest of these limestones is the Robinson, but its karstic expression is minimal.

There has been little research into the geomorphology of the Asby pavements, though they are briefly described by Goldie (1986, 1993) with reference to their damage by exploitation for rockery stone. The best developed and least damaged pavements are now protected within a National Nature reserve, and this also reflects the high botanical value of the pavements (Ward and Evans, 1975, 1976; Ratcliffe, 1977).

Description

Extensive and varied pavements cover much of the dip slope along the escarpment east of Orton. The highest and most spectacular area is known as Great Asby Scar, rising to a summit near Castle Fold (Figure 3.19). The limestone extends east

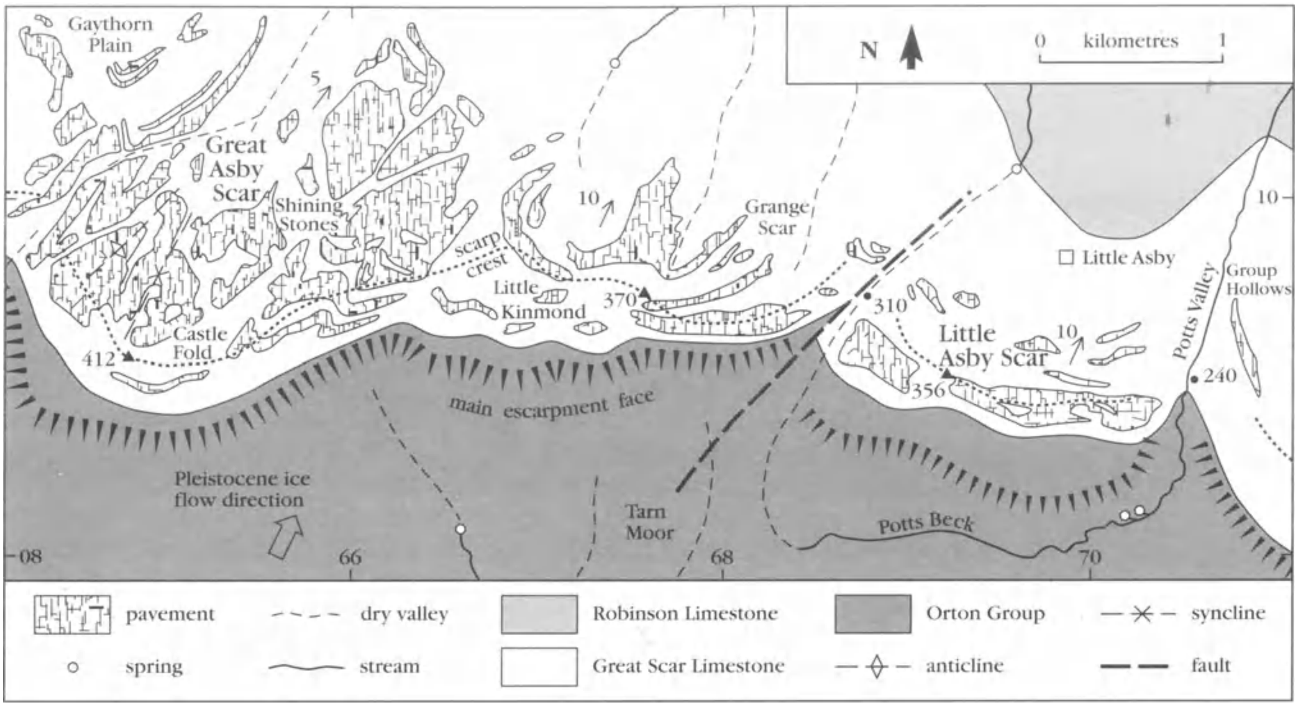


Figure 3.19 Outline map of the karst features on the limestone escarpment between Great Asby Scar and Potts Valley. The Robinson Limestone includes a thin shale separating it from the Great Scar.

through Little Kinmond to poorer pavements on Grange Scar, beyond which a shallow col breaks the escarpment along the line of a fault. East of the fault, the pavements of Little Asby Scar reach to Potts Beck, in a deeper valley through the escarpment. Pavement morphologies vary along the outcrop, and are best reviewed in sequence from west to east.

Gaythorne Plain lies low on the dip slope, with a large outcrop of thinly bedded limestone, of which about 80% of the original surface rock has been removed or disturbed. Clints are small, and runnels are poorly developed; there are scattered sandstone erratics. Just to the south, a higher and stronger bed forms a more massive pavement, with large clints and deep runnels, sloping northwest off a stratimorphic ridge on a gently plunging anticline. Grikes are mostly less than 1 m deep, but some are up to 2 m wide where close, parallel joints originally enclosed a blade of limestone or a calcite vein now totally removed by solution. Some clints within these pavements stand on pedestals about 100 mm high, which they overhang by about 100 mm. There are also areas of cleanly scoured bedding slabs with simple systems of large, sharp-edged rinnenkarren runnels.

Great Asby Scar has the most varied and best preserved pavements of the escarpment, with the greatest variety of solution features on the outcrops near Shining Stones and Castle Fold. Variations in slope, aspect, joint density, lithology and length of exposure to subsoil or subaerial solution are responsible for the enormous range of small karstic landforms. Long clints between joint grikes aligned on the strike have spectacular sets of deep parallel runnels cut into their downslope edges (Figure 3.20). Across the centre of the area, a synclinal valley contains the best of the pavements, and the anticline to its west provides structural variety within them. From the trough of the plunging syncline, excellent pavements rise towards Castle Fold where an ancient settlement stood on a low knoll formed by a remnant of a higher limestone bed. Massive, large rounded clints are underlain by pedestals of well fractured limestone to form mushroom-shaped features. These change eastwards through a development sequence, as the pedestals become less developed, the top clint less runnelled, and the grikes narrower, until a pavement of massive clints disappears under the debris cover around the Castle Fold outlier. The west side of the synclinal valley has pavements with deep rinnenkarren sloping east at about 15°, while gently sloping clints on the anticlinal crest are cut by deep, convergent rundkarren runnels. Further down the main dip

Figure 3.20 Deep parallel rundkarren on very long clints aligned on the strike of the gently dipping limestone pavements on Great Asby Scar. (Photo: S. Webb.)

slope, an area of very large clints with smooth surfaces, little scored by solutional runnels, is aptly known as the Shining Stones.

Little Kinmond and Grange Scar have less continuous pavements in bands along the crests of bedding scars continuing east along the escarpment. North of Little Kinmond, there are areas of massive pavement sloping north at 3–10° with some undulations over shallow plunging folds. The larger clints have kamenitzas and convergent runnel systems.

Interpretation

Pleistocene ice flowed from the Howgill Fells north-east across the limestone outcrop (Mitchell, 1991, 1994). Its flow against, then up and over the escarpment precluded the development by ice plucking of terraced limestone scars and capping pavements on the scarp face, which presents a very rounded profile. Ice flow down the dip slope efficiently stripped the exposures down to bedding planes on the stronger beds, where postglacial solutional fretting has created the finest and largest pavements. An iceway over the low point on the scarp crest produced deeper scouring along the fault line trough east of Grange Scar, while more selective scouring in the lee of the high point on the escarpment created the stratimorphic topography on Great Asby Scar. Pavements survive right across the synclinal valley floors which have never carried subaerial stream drainage.

Some of the larger runnels on the pavements appear to exceed the dimensions that can be realistically attributed to solution at modern rates throughout the Holocene. The survival of pre-Devensian relics appears to be incompatible with the site's glacial history and exposure to scouring, and the rounded forms of the karren ridges can not be attributed to rapid and deep excavation by meltwater during glacial retreat. The well rounded rundkarren are indicative of slow evolution beneath a cover of acidic soil and vegetation, but could have been formed by the postglacial rounding of older, subaerial features with sharper profiles. Past solution rates could have been higher in an environment of enhanced levels of biogenic carbon dioxide in soil waters from beneath a denser cover of shrubs and trees. The presence of early settlements at Castle Fold and other sites on the limestone suggest that there has been a richer soil and vegetation cover in the recent past, and pollen profiles from Sunbiggin Moor, south of the escarpment, support this concept (Webster, 1969).

The pavements east of Castle Fold are broken by numerous pits, each up to 7 m in diameter and several metres deep. They are unrelated to the tectonic joint patterns, and may be exhumed palaeokarstic features of Carboniferous age; these would have formed on calcretes which developed in sabkha environments during short periods of emergence from the shallow shelf seas (Vincent, 1995). The origin of their clay fills is not yet known, but may be Carboniferous soil evolved from volcanic ash, rather than Permo-Triassic loess or Devensian till.

Many clints stand on pedestals which are about a quarter of the height of those beneath the Norber glacial erratics on Ingleborough (Sweeting, 1966). This may suggest that solutional lowering of the surface has only lasted about 2500 years, if Holocene erosion rates are comparable at the two sites, and therefore conflicts with the environmental evidence of the large runnels. However, the evidence of some of the pedestals is invalid where they are lithologically defined by undercutting in bands of weaker, rubbly limestone. The extent and role of past soil covers on the Great Asby pavements are not yet fully defined.

Conclusions

The pavements at Great Asby Scar are nationally outstanding for their very large expanses of bare solutionally fretted limestone, and their wide range of morphologies related to the gently folded limestone structure. Past environments of contrasting soil and vegetation cover have also influenced the details of the solution features. Despite a sad record of destructive clint removal from large parts of the outcrop, much very beautiful and very varied pavement remains intact.

LITTLE ASBY SCAR AND POTTS VALLEY

Highlights

Little Asby Scar and the adjacent Potts Valley contain a distinctive area of limestone pavements, cliffs, scree and rough grassland which has origins traceable to pre-Devensian glacial erosion. Many of the pavement features appear to have preglacial elements.

Introduction

The Potts Valley is a major breach through the eastern end of the limestone escarpment which extends between Orton and Kirkby Stephen, north of the Howgill Fells (Figure 3.1). The south-facing escarpment is formed of the Great Scar Limestone, whose structure and lithology broadly continue those of the Great Asby Scar site immediately to the west. Little Asby Scar forms the high ground just west of the Potts Valley, whose western flank provides the type sequence for the Asbian stage of the Lower Carboniferous (George *et al.*, 1976), where the Holkerian Ashfell Sandstone meets the Asbian limestones and shales passing up into the Great Scar Limestone. Published research on the karst geomorphology is minimal and the site literature is incidental (Ward and Evans, 1975).

Description

South of the main scarp, Potts Beck drains a lowland formed on the shales and impure limestones of the Orton Group. The beck then flows north as an underfit on an alluviated floor through the narrower, rocky valley which breaches the scarp (Figure 3.19). Springs from the Orton limestones add to the flow of the beck, but infiltration to the Great Scar Limestone drains to various small risings on the dip slope to the north. Limestone screes and inclined scars form the western slope of the valley, but are largely obscured by a bank of glacial till on the eastern side.

The major part of the gently graded scarp face is formed in the mixed sedimentary sequence of the Orton Group, where the thin limestones form a few low scars. Only the crest of the escarpment is formed by the Great Scar Limestone. The pavements are formed only on the more massive beds within this unit, and lie in the terraced, sloping benches above small scars on the upper part of the scarp face. The northern flank of the escarpment is hardly a dip slope, as the dip is steeper than the surface profile, and the pavements again form on only narrow outcrops.

On Little Asby Scar the main pavements slope north at 10–18°, and are very well dissected. The more massive beds have some large clints, scored by mature rundkarren runnels and shallow kamenitzas. Most grikes are deep and 100–200 mm wide, but those in a distinctive sub-set are 1–2 m wide, shallow and commonly infilled. These wider grikes are common near the ridge crest and are formed on the same systems of tectonic joints as the deep and narrow kluftkarren; they are mostly spaced 10–30 m apart. The pavements are noticeably discontinuous, forming strips only a few metres in width along the tops of some of the scars. East of Potts Valley, the limestone has a generally thicker cover of soil and grass. There is no significant surface drainage.

Interpretation

Potts Beck originally had a much larger catchment, before capture of its headwaters by the River Lune. The valley through the limestone escarpment was cut by a larger river draining much of the northern slopes of the Howgill Fells (McConnel, 1939; King, 1976), and was probably incised to close to its present depth early in the Pleistocene. Subsequently, the valley may have acted as an iceway, or carried subglacial meltwater, but modification by glacial scour was probably very limited, as the site lay at a basal ice shed (Mitchell, 1994). This protection from glacial erosion may account for the survival of many older landforms on the higher parts of the escarpment, and it is likely that some components of the limestone pavements predate the Devensian glaciation, though there are no absolute dates to confirm this.

The bimodal distribution of grike widths provides evidence for some inheritance of older features. The narrow grikes are clearly postglacial, but those of the wider sub-set appear to be pre-Devensian relics, as solution rates recorded widely on the Pennine limestones (Sweeting, 1966; Rose and Vincent, 1986c) could not account for kluftkarren 2 m wide within the 10 000 years of the Holocene. The very dissected nature of the pavements also reflects their considerable age, but the bulk of the rundkarren runnels are less than 400 mm deep and could therefore be entirely postglacial. This suggests the possibility that glacial scour removed a top bed of limestone, with its older runnels, and the surviving grikes are just the lower parts of interglacial kluftkarren which reached down through two or more beds.

Conclusions

The limestone outcrops on Little Asby Scar have well dissected limestone pavements within an area which lay on a basal ice shed through much of the Pleistocene glaciations. It is likely that many of the pavement landforms were inherited from preglacial features which escaped complete removal by glacial erosion. The site also contains the type section of the Asbian stage of the Carboniferous.

HELBECK SCARS

This is a proposed GCR site, not yet designated as an SSSI

Highlights

Helbeck Scars form an extensive area of open limestone pavement high on the Pennine escarpment. The limestones are well folded, and consequently the pavements contain a wide range of solution forms. The larger clints have kamenitza and rundkarren, and more jointed areas are reduced to outcrops of knife-edged clints.

Introduction

Helbeck Scars is the collective name given to a series of limestone outcrops just below and west of the crest of the Pennine escarpment overlooking the Vale of Eden, north of Brough (Figure 3.1). They range across altitudes of 350–600 m. The pavements form an almost continuous band 300 m wide and 4 km long, from Helbeck Intake, north-west across Key Scar, Musgrave Scar and Middle Fell, to Long Fell. They are all formed on the Dinantian Great Scar Limestone, exposed between the Swindale Beck and Barnarm Faults near the top of the Pennine scarp face. Mixed sequences of shale, sandstone and limestone of the Alston group overlie the Great Scar, and similar rocks of the Orton Group lie below. The regional dip is to the east, but local folding and block faulting produce considerable dip variation across the limestone outcrops. The limestone outcrops therefore include both wide pavements and narrow scars.

Most of the site lies within the Warcop military ranges, which probably accounts for the small degree of human damage to the pavements. There has been almost no geomorphological research on these little known but very spectacular pavements. The botanical values were assessed by Ward and Evans (1975) and in recent unpublished reports for English Nature.

Description

Pavement morphologies vary considerably across the site, in response to geological structure, outcrop pattern and aspect. They are best reviewed sequentially from south-east to north-west (Figure 3.21).

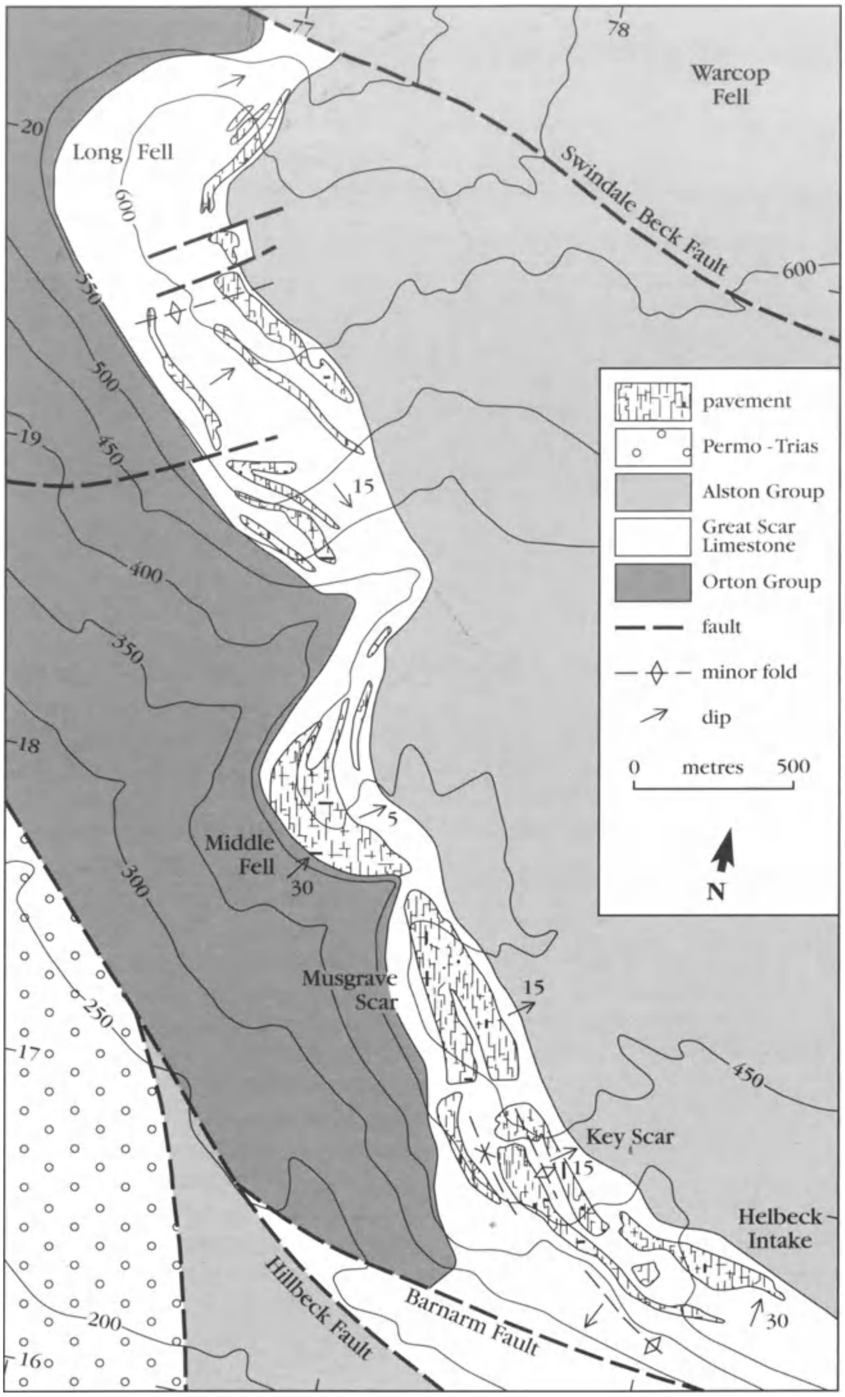


Figure 3.21 Geological map of the pavements on the Helbeck Scars. The Alston Group includes thin limestones with low scars and narrow pavements which are not marked.

Helbeck Intake has outcrops across its southern slope with narrow bands of scar top pavement dipping north at 30°. Deep and narrow grikes separate elongated clints which are reduced to linear, knife-edge blades as narrow as 20–30 mm in areas of close jointing. At the western end of the scars, the beds dip over an eroded anticline, where parallel runnels develop on the larger and steeper clints. A large expanse of undulating pavement in the north of the Intake stands on limestone bedding planes dipping 10–25°, with curved grikes between clints with well developed rundkarren, kamenitzas and solution pits.

Key Scar has stepped pavements and small scars up the hill side. Local joint patterns dictate the shapes of rhomboid clints on the higher beds and more elongate beds lower down. The large outcrop at the western end of the scar lies across a monocline, so that a belt of the pavement slopes at 15–20° to the west, between almost level pavements above and below (Figure 3.21). The level areas have scattered kamenitzas 20–200 mm across, but virtually no runnel development. The steeper pavements have rhomboid clints with solution basins which have lost their front rim to form trittkarren. They also have sequences of kamenitzas, in which each overflow into its lower neighbour, and the stepped systems drain into rundkarren lower down the clint.

Musgrave Scar has narrow bands of dipping pavement separated by strips of acidic, grass-covered soil forming on the bottom of each dip slope below scars less than 1 m high onto the next bed. Along the crests of the scars, the grikes are deeper and narrower than they are lower down the same bedding planes. Natural breakdown is degrading the scar edges, most rapidly in areas of closer jointing and thinner bedding. Runnels are broad, shallow and convergent, but steepen and deepen into the grikes. Some grikes, inclined at 30° from the vertical, have well developed flutes on their lower surfaces. Kamenitzas are shallow and locally stepped.

Middle Fell has wide pavements formed across a synclinal flexure, where dips of 30° east at the scar edge ease to 10° further into the slope (Figure 3.21). Clints are up to 1 m across but are smaller and increasingly knife-edged towards the scar edges. Deep rundkarren score the clints, and some start as stepped series of kamenitzas. Protogrikes form along lines of weakness, commonly parallel with mature grikes. Further north, a group of larger depressions, 2–10 m across, have formed in a band within the stepped and dipping pavements, and a line of active sinks along the rear of the pavements swallows drainage from the overlying sandstone.

Long Fell has the most northerly and highest of the pavements. Narrow outcrops and small scars have deep, narrow grikes between small rhomboid and triangular clints with poor runnels. Further west the highest limestones of the Great Scar form dip slope pavements with larger clints on the massive beds between zones of broken rock on the outcrops of more rubbly beds.

Interpretation

The Helbeck limestones were subjected to intensive glacial scour by Pleistocene ice moving the length of the outcrops south-east towards the ice-way through the Stainmore gap. Solution features on the pavements are mostly on a small scale commensurate with formation since the Devensian ice retreat. Reaching altitudes of 600 m, the Helbeck pavements are higher than any others in England, except for the small features on the Yoredale limestones high on Simon Fell, Ingleborough. No direct influence of this altitude, and its climatic impact, can be recognized in the pavement morphology, though the modern flora is certainly restricted (Ward and Evans, 1975).

The structural variety in the limestone has produced a full range of bed scars, bare pavements on the dip slopes, and soil-covered pavements in sheltered sites. The Middle Fell pavements stretch undisturbed from scar top open pavement down to a soil cover and the scree of the next scar. On Musgrave Scar, narrow dipping pavement grades downslope into bands of acidic grassland dominated by *Nardus stricta*, probably growing on accumulations of windblown loessic silt at the lowest point of each dip slope. Breakdown of the narrow bands of pavement is accelerated by unloading fractures close to the scar edges. Closely spaced tectonic joints create the linear knife-edged clints of Helbeck Intake and Musgrave Scar.

Limestone lithology also influences pavement morphology. On Long Fell, adjacent beds at outcrop include an upper pseudobrecciated, rubbly limestone with no recognizable karren forms, over a lower massive limestone developing a smooth pavement surface with gently rounded solution features. The upper bed degrades and retreats to reveal the fresh surface beneath, but there is no clear pattern related to the retreating cover in the morphology of the lower pavement. Some pedestal clints have developed on Middle Fell where a massive bed overlies an easily degraded rubbly limestone.

Solutional features on the massive beds of limestone are well developed and show morphological response to the local dips and clint slopes, which vary so much across the folded limestone. Rundkarren develop below smooth areas of clint which are not runnelled. The relationship between dip and morphology is clear, with parallel, straight runnels on the steeper dips and branching, meandering forms on the more gentle slopes. Catchments on the steeper slopes include broad, tapering, shallow depressions, and runnels steepen into flutes down the sloping walls of grikes.

Kamenitzas are abundant, and many link to form cuspate elongated forms. Some kamenitzas eventually drain through rock fissures, but others overflow, to create stepped basins, and some appear to evolve into trittkarren. On Key Scar, many of the runnels on the sloping clints appear to originate as sequences of linked, overflowing kamenitzas. This may conflict with the wider evidence that rundkarren develop beneath a soil cover, whereas kamenitzas form on bare pavements where they catch water and organic debris. The relationship adds support to the concept of rounding the rundkarren crests merely beneath a lichen cover.

Eight circular depressions, in a group on the north of Middle Fell, are each 2–20 m across and up to 2 m deep, with level, grassed floors. These are distinctly larger than any other features on the Helbeck Scars, and appear to be relics of pre-Devensian landforms. An annular zone of pavement about 50 m wide around these basins has more mature karren morphologies, with runnels up to 400 mm deep in large, smooth, rounded rundkarren.

Conclusion

Helbeck Scars have the only extensive pavements on the limestones of the Alston Block, at higher altitude than any other large pavements in Britain. They are formed on folded limestones which support a wide range of pavement types and features in response to variations in structural dip and lithology. Abundant kamenitzas appear to be genetically related to rundkarren, and there are very fine linear, knife-edged clints in the densely jointed limestones.

GOD'S BRIDGE

Highlights

The natural limestone span crossing the River Greta at God's Bridge is the best example of a natural limestone bridge in Britain. Cave development in a thin limestone in the valley floor has now captured much of the river flow.

Introduction

The River Greta drains a large area of the fells of Stainmore Forest in the northern Pennines (Figure 3.1). Its headwaters lie on impermeable rocks, but 4 km west of Bowes, the river crosses a thin bed of Carboniferous limestone, where several generations of caves have developed. The progressive development and subsequent collapse of a sub-valley floor cave system has produced a natural limestone bridge spanning the river. A lower system of caves is still active and captures much of the river flow. There is no published study of the site geomorphology, but the caves are described in Brook *et al.* (1988).

Description

Nearly horizontal sequences of Carboniferous shales, sandstones and limestones form the high fells around the Stainmore saddle over the northern Pennines. Karst landforms are limited in the thin limestones, of which many outcrops are hidden beneath glacial till. The River Greta is an underfit in a broad valley which carried substantial flows of Pleistocene ice through the Stainmore gap. God's Bridge is developed where the River Greta crosses the outcrop of the Great Limestone, which is about 20 m thick and lies at the base of the Namurian succession.

The river flows onto limestone a few hundred metres above the Bridge, and has developed a series of caves below the valley floor. God's Bridge is a bedrock span over a cave 12 m long, 2 m high and about 4 m wide through which the river flows, and is large enough to accommodate almost all the modern flood flows. The Bridge is made from two beds of limestone and is only about 2 m thick (Figure 3.22). There is limited block collapse at the upstream end, while a shallow rocky gorge represents the unroofed continuation of the cave on the downstream side. The upper surface of the bridge is bare rock, exposed to weathering and ultimately destined to collapse by a combination of thinning, fissuring and undercutting. Part of the river flow now passes through a lower cave system, extending 500 m from sinks upstream of the Bridge to resurgences downstream. Most of this cave is a series of low, wide, bedding passages with oxbow loops, and parts of the route are permanently flooded (Brook *et al.*, 1988).

Figure 3.22 The upstream side of the limestone span of God's Bridge across the River Greta. (Photo: A.C. Waltham.)

Interpretation

God's Bridge is the last surviving relic of a valley floor cave system. Early solution by the River Greta of the limestone at outcrop produced a cave system below the valley floor, which was subsequently unroofed and dissected as valley lowering proceeded. The natural bridge is part of this earlier generation of sub-valley floor caves, and the rocky gorge downstream is an unroofed section of the same cave. Continued solution has created a younger and lower cave system extending parallel to the Bridge site from new sinks upstream. This new cave has developed along bedding planes downdip of the surface river bed, so that it forms a drainage loop beneath the north bank. Ultimately, this new cave will suffer the same fate as its predecessor and will be unroofed, leaving temporary fragments spanning the river.

There is no positive evidence of the age of the caves. The river bed location is compatible with youthful caves, but abandoned loop passages in the lower cave system north of the surface river suggest that even this may not be entirely Holocene. The caves can only have developed where the limestone was exposed in the valley floor with a downstream outlet for their drainage, and so cannot be older than the time taken for surface lowering to pass through the 20 m thickness of the limestone. However, this is greater than the 15 m of valley floor lowering attributed to a single glacial episode in some of the Pennine limestone valleys (Waltham, 1986). It is therefore possible that the caves were overridden by Devensian ice. The section of river channel downstream of God's Bridge is a cave which may have been unroofed by glacial plucking; this would have been greater downflow of the old cave exit than where the ice overrode the upstream entrance.

Conclusions

There are at least three sites in the limestone Pennines known as God's Bridge. All are cave remnants which provide convenient natural routes over rivers or streams. The God's Bridge of Stainmore is the finest of them. It is a truncated fragment of a formerly more extensive valley floor cave passage, and has a similar, but newer, cave system now developing parallel to it.

KNOCK FELL CAVERNS

Highlights

Knock Fell Caverns is the finest example in Britain of a joint-guided phreatic maze cave, with more than 4500 m of passages known within a single thin limestone in an area of less than 3 ha.

Introduction

Knock Fell Caverns lies at an altitude of 750 m directly beneath the surface watershed along the crest of the Pennine escarpment north of Knock

Fell (Figure 3.1). The cave is developed in the Namurian Great Limestone, which is about 20 m thick and dips very gently to the north-east; mixed sequences of sandstones and shales lie both above and below the limestone. There are numerous shakeholes in the soil and drift over the limestone outcrop, and one contains the 7 m deep entrance shaft to the Caverns. Underground drainage within the limestone resurges at a strong spring near the head of Knock Ore Gill. The cave was mapped and described by Sutcliffe (1985).

Description

The shaft entrance to the cave lies on a joint intersection modified by collapse which has broken through to the surface. All the main passages are formed at one level within the Main Limestone. They are vertical phreatic rifts, all formed along joints, and they intersect in a maze of spectacular complexity (Figure 3.23). Most are less than a metre wide, and narrower joint fissures with fretted walls extend above and below the main solutional enlargement to give total passage heights of 5-10 m. Horizontal rock ribs and blades protrude from many of the passage walls, left by selective solution of closely spaced lithological contrasts within the limestone (Figure 3.24). Most of the known cave system, which has a total passage length of more than 4500 m in an area roughly 320 m by 120 m, lies beneath the cover of shales and sandstones. Passages to the west extend under the shakeholes on the limestone bench; these are largely choked by boulder falls and inwashed gravels, which now fill the floor rifts in adjacent passages. The eastern extremities of the cave reach towards the Teesdale flank of the ridge, and are also choked by sediment.

The entire cave is formed on joint fissures, and the tectonic fracture patterns therefore control the maze topography. Joints are more closely spaced in the southern half of the cave, while a more open maze has formed on more widely spaced fractures to the north. Some wider passages and chambers with rectangular profiles have formed by the breakdown of narrow blades of rock between solutional fissures on close, parallel joints. Part of the northern end of the main cave is underlain by a discrete lower level, the Inferiority Complex, with smaller phreatic passages forming a denser network than those in the main maze about 5 m above. Below this may lie younger, active caves draining towards Knock Ore Gill Head. The known cave is dry, apart from percolation water entering from roof fissures.

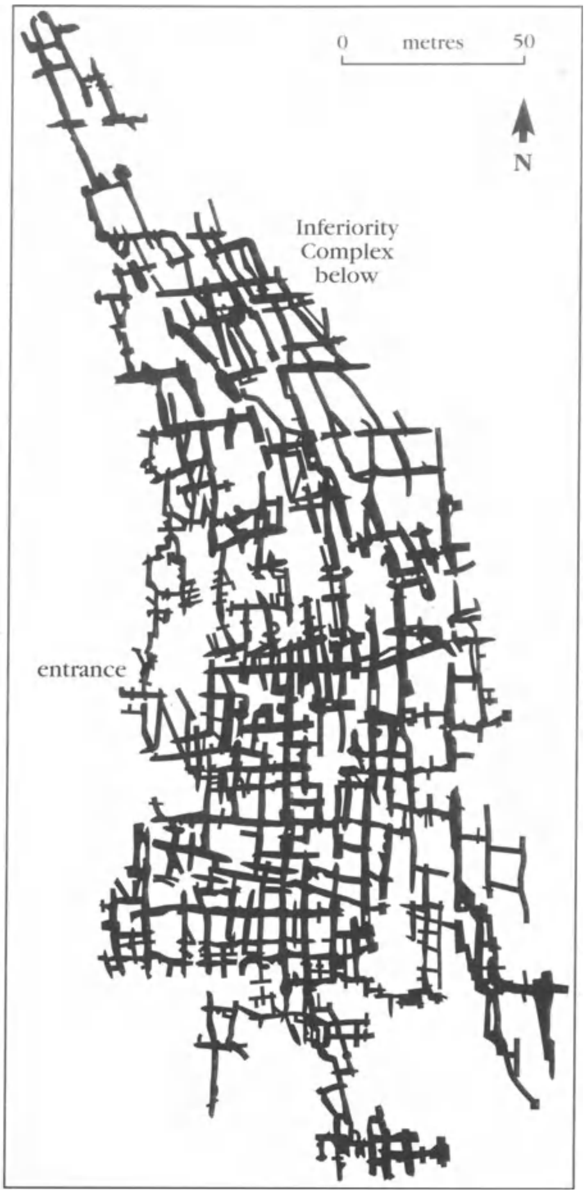


Figure 3.23 Outline map of Knock Fell Caverns, without the much shorter lower series which are omitted for clarity (from survey by Gritstone Club).

Some roof fissures reach to the top of the limestone, and the undermined shale has partially failed. Several wider avens on joint intersections reveal the sandstone roof which overlies the thin shale. The sandstone is fissured sufficiently to allow acidic water to percolate down from the blanket peat above. This water is mostly aggressive as it etches the cave walls, but small secondary calcite deposits have formed in a few places. The fossil corals of the Frosterley Band are

Figure 3.24 One of the rift passages in Knock Fell Caverns, where dissolution by the slowly moving phreatic water etched out lithological contrasts in the limestone walls. (Photo: A.C. Waltham.)

conspicuous in the walls of many parts of the caves, and are locally spectacular where the limestone has been etched from around them by the aggressive percolation water.

Interpretation

Knock Fell Caverns represent the finest of the complex phreatic maze caves which are a feature of the thin Yoredale limestones in the northern Pennines. It is more extensive than the comparable mazes intersected by mine workings in Swaledale (Ryder, 1975), and all of these have much denser passage networks than the rectilinear stream caves of Mossdale Caverns and other comparable sites. Knock Fell Caverns is typical of the dense mazes of cave passages formed by slowly moving water in confined aquifers (Palmer, 1975); solution takes place along all the fractures without selective enlargement on those fissures with hydraulic advantage in an environment of high flow rates. No flow patterns have been recognized in the cave.

Permeable, jointed sandstones lie both above and below the cavernous limestone, in each case separated by only thin shale beds which are seen to be breached in some of the roof shafts. These sandstone aquifers could have provided, via the fractures, a diffuse input of aggressive water into the limestone, in the style recognized in many other maze caves (Palmer, 1975). This could have taken place with either upward or downward flow when the limestone was deeply buried in an artesian phreas. Alternatively, it may be much later, with downward flow through an exposed sandstone cap into a limestone phreas perched on shale, and unable to drain across the low dip into distant surface valleys. In either case, the phreatic development was terminated when surface lowering left the cave perched just beneath the watershed cap. Vadose modification has been minor.

The cave lies at very high altitude, close to both the Pennine fault scarp and a long dipslope down to the Milburn Forest. Hence much of the phreatic passage development may be very old, substantially predating the surface landforms. Clastic and calcite deposits within the cave represent the only material suitable for absolute dating in this part of the Pennines, and hence may provide a valuable record of the valley incision and geomorphological history of the area.

Conclusion

The scale and complexity of the phreatic maze of Knock Fell Caverns are unparalleled in Britain. Its configuration and position, with a joint network enlarged by solution in limestone beneath a permeable sandstone, suggest that it was probably formed by diffuse recharge to a confined aquifer with very slow drainage.

FAIRY HOLES

Highlights

The long, almost horizontal cave passage in Fairy Holes carries a small stream and has few tributaries. It is the finest and longest of the linear

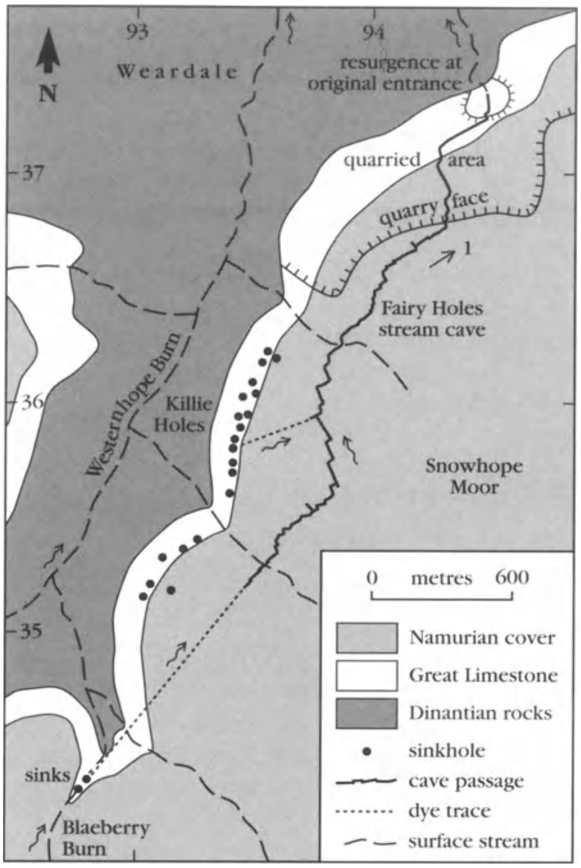


Figure 3.25 Outline map of Fairy Holes and the limestone bench which it drains. The cave and outcrops are shown in their original form, previous to development of the quarry; the limestone has been largely removed from its outcrop southwards to the quarry face, which has also cut into part of the non-carbonate cover. Except for a tiny fragment behind the original resurgence entrance, all the cave passage north of the quarry face has been destroyed (from cave survey by University of Leeds Speleological Association).

caves with the simple underground drainage patterns which are common in the thin Yoredale limestones exposed in the hillsides of the northern Pennines.

Introduction

Fairy Hole lies under the southern flanks of upper Weardale, between Eastgate and Westgate (Figure 3.1). It is the prime example of cave development in the Yoredale limestones of the northern Pennine dales, where one or very few sinks feed a rising via a single passage with few tributaries. The single cave stream passage has been explored from the rising for most of the 3.5 km to the main sinks, in Blaeberry Burn (Figure 3.25). It is developed in the Carboniferous Great Limestone, which is locally about 18 m thick. Mixed sequences of shales, sandstones and thin limestones lie both above and below the Great Limestone, which forms the lowest unit in the Namurian.

The passages are described by Jones (1957) and Brook *et al.* (1988), but lack of access to the cave has precluded any scientific studies.

Description

The original cave entrance above the resurgence survives in a remnant of limestone, encircled by a quarry which has completely removed about 600 m of stream passage immediately upstream. South of the quarry face, the truncated cave still has 3200 m of passages, nearly all forming the one streamway (Figure 3.25). Most of this is a clean vadose canyon over a metre wide, but most sections are aligned on joints which were initially opened by phreatic solution. An abandoned upper level of the cave survives partly as a series of loops where it is offset from the active cave, and partly as roof sections of the streamway. There are sections modified by blockfall and wall collapse, with some chambers up to 30 m long and 6 m wide, mainly formed where the active streamway intersects and undercuts wider parts of the abandoned roof passage. Some of the passage walls have protruding fossil rugose corals etched out of the Frosterley Band. The streamway collects small flows from sinks in the Killie Holes along the limestone bench, but inlet passages are small and not fully explored.

Interpretation

The thin Yoredale limestones offer limited scope for the development of complex multi-level cave systems, and most underground drainage in them is simple and direct. Fairy Holes is typical in that it has a single, youthful, vadose streamway between sinks and a rising on the edges of the modern outcrop. The cave stream has invaded, linked and modified an earlier generation of phreatic rifts. These are widespread in the Yoredale limestones, and were formed by solution in a confined aquifer before it was drained by incision of the adjacent valley, probably in the late Pleistocene. The vadose stream drains downdip through the fissured limestone, which carries it from the sinks parallel to the Westernhope Burn. This accounts for the large distance to the downdip rising within a very narrow limestone outcrop. Aggressive percolation water sinking into the exposed limestone along its hillside bench enhances cave excavation by solution in the rock immediately below; the Fairy Holes stream cave, just behind the outcrop bench, is therefore larger and more accessible than is normal in these limestones.

Conclusion

Fairy Holes is a cave typical of those in the thin Yoredale limestones, having a long, gently graded stream passage with few tributaries, but its passage sizes are unusually large due to its alignment parallel to the narrow limestone outcrop.

Chapter 4

The Peak District karst

INTRODUCTION

The southern end of the Pennines contains the largest unbroken area of cavernous karst in the Carboniferous limestones in Britain (Figure 4.1). A limestone upland roughly 35 km by 15 km lies between Matlock and Chapel-en-le-Frith; it may be referred to as the White Peak, to distinguish it from the Dark Peak of the Millstone Grit moors to the north, the two combining to form the Peak District.

The limestone is of Dinantian age with most of the outcrops, karst and caves lying within rocks of the Asbian and Brigantian stages. Over 500 m of limestones are exposed across the Peak District karst, and another 1000 m of carbonates underlie these but have no outcrop. Thickly bedded, pure limestones of shelf lagoon origins dominate the heart of the Peak District, locally interrupted by dark, thinly bedded, impure carbonates of basinal facies. Massive reef limestones are important. They form a zone of marginal reefs, complete with their own debris slopes of the fore reef facies, which fringe a long-standing positive area; in Carboniferous times they bordered a shelf area of shallow water, and today they lie close to the

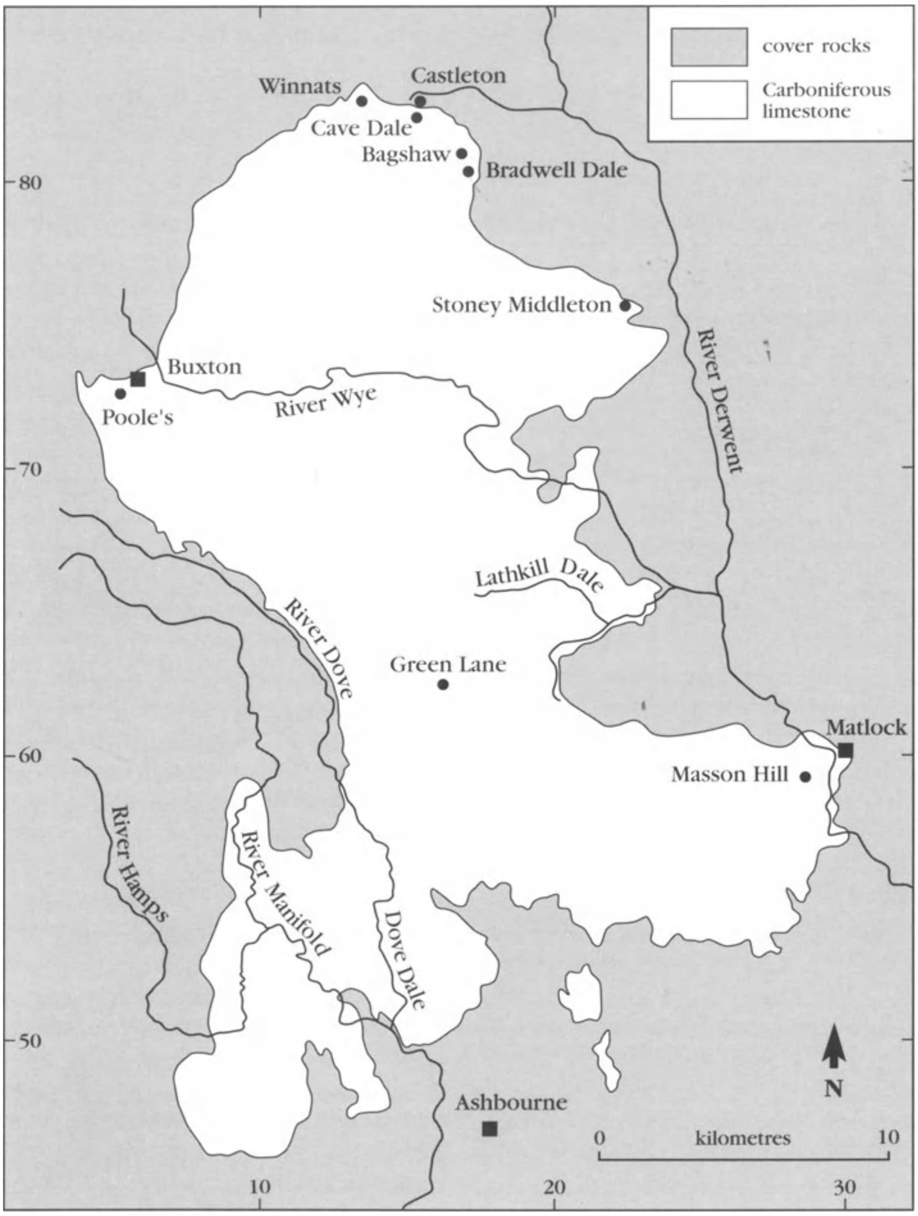


Figure 4.1 Outline map of the Peak District karst, with locations referred to in the text. The cover rocks are Namurian shales and sandstones, and younger stratigraphic units.

perimeter of the main limestone outcrop. Patch reefs occur inside this fringe, and form low reef knolls on the karst plateau. The limestone succession is interrupted by discontinuous basaltic lavas and pyroclastic horizons; many of the latter are heavily altered and weathered, and are known locally as toadstones and wayboards. Secondary dolomitization is associated with some of the mineralized areas in the southern third of the karst outcrop.

Impermeable basement rocks are known only in a few deep boreholes. The cover rocks, forming an annular outcrop around the karst, are shales mostly of Namurian age, which are followed by the strong sandstones in the Millstone Grit Series. Only along a short section of the southern perimeter of the karst do Triassic sandstones overstep onto the limestone. Neogene sands and clays accumulated on the southern part of the karst as the adjacent Triassic cover retreated, but most were removed during the Pliocene. They only survive as the fills in about 60 solutional depressions, dolines and karstic collapses, where they are known as the Pocket Deposits of the Brassington Formation (Ford and King, 1969; Walsh *et al.*, 1972); overlain by Pleistocene till, these sites are features of a Tertiary palaeokarst (Ford, 1984).

Structurally the area may be known as the Derbyshire Dome; this is an oversimplified description of the distorted positive area on the southern end of the Pennine anticline. There are numerous crossing folds, but, except in some marginal areas, the limestone dips are mostly low. Faults are widespread across the limestone, and are extensively mineralized. The hydrothermal mineralization is of the Mississippi Valley type generated by the migration of connate fluids into the domed area from adjacent basins in late Carboniferous times (Ineson and Ford, 1982; Mostaghel and Ford, 1986; Quirk, 1986, 1993; Ixer and Vaughan, 1993). The main ore deposits are in the long faults across the karst, aligned roughly east–west and known as rakes; ores also form bedded flats and cavity infills in palaeokarstic openings. Mineral working in the Peak District has been continuous for over a thousand years, firstly for the lead ores and latterly for the fluorspar resources. Miners were the first to encounter and explore many of the caves, and they significantly modified much of the karst drainage as they endeavoured to lower the water table around their mines by cutting long drainage adits, known as soughs, close to base level.

The karst

The limestone forms a dissected plateau with local relief generally less than 200 m and nowhere more than 300 m, and the entire karst was overrun by ice sheets during the Anglian stage of the Pleistocene. The extent of glacial cover of the plateau by a 'Wolstonian' ice expansion is unknown, but till was introduced into some of the valleys by ice lobes from the west (Burek, 1991). The Peak District karst was not covered by ice during the Devensian, when it was subjected to a long period of periglacial conditions.

The Peak District plateau is essentially a fluviokarst, incised by dendritic systems of dry valleys which are the dominant feature of the landscapes. These were probably superimposed from a retreating cover of Namurian shales, and were initially desiccated by the maturing of the karstification (Warwick, 1964; Pitty, 1968). They were subsequently reactivated and deepened by summer meltwater flows when groundwater was frozen under the periglacial conditions of the cold Pleistocene stages.

The Rivers Wye and Dove maintain their surface flows right across the limestone outcrop, and a few other rivers have shorter surface courses within the karst. Most of this surface flow is in gently graded valleys at base level, aided by local perching of the water tables on impermeable volcanic horizons, and some stretches where the river beds have been artificially puddled with clay. Conversely, some stretches of valley are now dry only since their flow was captured by the miners' drainage soughs. Tufa barriers have formed below some of the karstic springs, and survive on a larger scale than at most other sites in Britain. Some valleys steepen into gorges where they descend the plateau margins through strong reef limestones; other gorges have been entrenched behind stratimorphic reef mounds which trapped valleys whose uniclinal shift was curtailed (Ford and Burek 1976). Throughout the Peak District karst, the landforms are strongly influenced by geological structure and lithology.

Limestone pavements are rare in the Peak District; those which originated from the Anglian glaciation have either been destroyed by later frost action or have been buried beneath younger soils. Many of the deeper dry valleys are lined by rock scars which mark the outcrops of stronger beds of the limestone and especially the massive reef units. Rocky tors have formed by Pleistocene frost action in some of the dolomite outcrops

(Ford, 1963). Most of the karst lies under a veneer of soil, derived from frost action and loess accumulation, both largely in the Devensian periglacial environments (Pigott, 1962, 1965). On the steeper valley sides this is widely soliflucted, and soil creep has initiated the formation of extensive areas of terracettes. Only isolated patches of glacial till have survived the erosion since the Anglian ice cover retreated.

Most of the limestone outcrop stands topographically higher than the surrounding outcrops of the stratigraphically younger shales, which are more easily eroded. Surface drainage is therefore predominantly towards the shale, and allogenic waters draining onto the limestone are limited in extent. Only where the escarpments of the Millstone Grit - the Edges of the Dark Peak - are close enough to the karst and high enough, do the surface streams cross the shale and sink into the limestone. Rushup Edge overlooks the northern tip of the karst perimeter, and supplies the water to the cave systems behind Castleton. Most water enters the karst aquifer from direct rainfall, constituting percolation input with little flow concentration.

Consequently, there are relatively few open sinkhole caves in the Peak District karst. Solutional dolines, subsidence dolines and collapse features do occur, but closed depressions are generally only details within the fluviokarst landscape. With no basement rocks exposed, many of the resurgences are vauclusian, offering only difficult access to submerged passages, and open cave passages at resurgences are few in number.

The caves

The great majority of the known Peak District caves are located close to the marginal shale outcrops (Ford, 1977b), where allogenic drainage creates stream flows large enough to create sinkholes of enterable dimensions. By far the most extensive cave systems lie behind Castleton, where the high sandstone scarp of Rushup Edge supplies drainage across a narrow shale outcrop onto the limestone; from the sinkholes there is a steep hydraulic gradient through the limestone into the deep glaciated trough of the Hope Valley. Between the Rushup Edge sinks and the Castleton risings, more than 20 km of cave passages have been mapped. They constitute an excellent karst drainage system, which exhibits a close control by the geology and reveals a long evolution through successive rejuvenation stages during the Pleistocene (Ford, 1986a). Dated stalagmites and sediments in successive cave levels yield a chronological framework which can be correlated with surface erosion from Anglian times to the present (Ford *et al.*, 1983).

The Castleton caves have developed in all the major environments found in the Peak District karst. They traverse rocks of both the reef and lagoonal facies of the Carboniferous Limestone. In the reef masses, the caves are mostly irregular complexes of chambers, with some extending into the fore-reef boulder beds, notably at Treak Cliff Cavern. In the bedded lagoonal limestones, the caves follow bedding planes for considerable distances, except where deep phreatic loops developed on faults and mineral veins, some of which may have had solutional cavities relict from Tertiary or earlier events. The deep meandering vadose canyon of Giant's Hole and the relict phreatic tubes of Peak Cavern are classics of cave morphology. More than any other site in Britain, the Peak–Speedwell Cave System clearly demonstrates a complex hydrology with flood diversions through parallel routes; this typifies the deep drainage of the Peak District karst (Christopher *et al.*, 1981).

Little is known about cave development in the centre of the karst plateau, beneath outcrops remote from allogenic drainage supplies. Upper Lathkill Dale contains the only large segments of cave passage yet revealed. Lathkill Head Cave is a perched flood route with small passages feeding to a natural resurgence, but the other caves in the site have been revealed only through chance intersection by old mine workings. The large abandoned phreatic tubes of Water Icicle Close Cavern originated from substantial flows of underground water, concentrated by sizeable upstream catchments. These were either allogenic supplies provided previous to extensive removal of the overlying shales, or distant coalescences of percolation water. The caves are known to be pre-Anglian (Ford *et al.*, 1983) but further sediment data are needed to determine their exact origin and the role they have played in the plateau hydrology, especially during periods of Pleistocene meltwater activity.

A special feature of the Peak District caves is their relationship to the hydrothermal mineral deposits in the limestone. Some palaeokarst features predate the late Carboniferous mineralization, which has subsequently influenced the Tertiary and Pleistocene development of the modern caves and karst; large solutional rifts lie along the mineral veins (the rakes) in many cave systems. In the Matlock area, caves intersect and follow some of the mineral orebodies, where mining has re-exposed some of the palaeokarst features (Ford, 1984). Combined with sediment sequences which include fluvioglacial material more than 730 000 years old (Noel, 1987), the Masson Hill caves offer fragmentary evidence covering an exceptionally long timespan of karstic evolution. A second equally long record of Pleistocene events is provided by the fossil caves and sediment fills exposed by the quarry in Eldon Hill (Farrant, 1995).

CASTLETON CAVES

Highlights

The Castleton area contains the most extensive and complex karst drainage system in the Peak District of Derbyshire. It exemplifies a style of drainage unique to this area; draining from both stream sinks and percolation sources, via deep phreatic conduits guided by mineral veins, to linked vauclusian risings. Within the confines of this site are contained a record of nearly one million years of landscape development.

Introduction

The limestone plateau around Eldon Hill is drained by the most extensive cave systems in the Peak District; the resurgences are in the Peak Cavern gorge behind Castleton. The Carboniferous limestone, of Asbian and Brigantian age, forms a reef complex along the northern boundary of the site, with lagoonal carbonates and thin interbedded basalt lava flows to the south. Large east-west mineral veins, or rakes, traverse the limestone and have, in the past, been worked for lead and zinc ores. Both the reef and lagoonal facies are penetrated by the caves, and the mineral veins are also utilized by the underground drainage. Streams flowing from the impermeable shale slopes of Rushup Edge feed a line of sinkholes along the north-western edge of the limestone outcrop. This allogenic water, joined by autogenic input, passes beneath the topographic divide and resurges from vauclusian risings in the Peak-Speedwell cave system in the floor of Hope Valley.

An extensive literature has been published on the caves and hydrology of this area (T.D. Ford, 1966b, 1969, 1977b, 1986a, b; Christopher, 1980, 1984; Christopher *et al.*, 1981; Christopher and Wilcock, 1991; Ford and Gunn, 1992), with other work by Pitty (1971) and Johnson (1967). A thematic issue on the Peak-Speedwell Cave System was published as Volume 18, Number 1 of *Cave Science* (Ford, 1991). Descriptions of all of the caves are given by Gill and Beck (1991), with more detailed accounts of specific parts of the system in Cordingley (1986, 1988, 1989), Ford (1956), Proudlove (1985), Salmon (1956, 1959), Salmon and Boldock (1951a, b), Shaw, R. P. (1983), Smith and Waltham (1973), Westlake (1967) and Wright (1987). Recent major discoveries have been described by Nixon (1991, 1992) and Beck (1991). Data on speleothems and sediments have been published by Ford *et al.* (1983), Thistlewood and Noel (1991) and Murphy (1993), and the underground hydrology is summarized by Gunn (1991) and Bottrell and Gunn (1991).

Description

The caves of the Castleton area represent parts of a single, though complex, hydrological unit (Figure 4.2). The accessible sections of cave passage fall into three distinct groups, which are linked by inaccessible passages, flooded or choked with sediment.

Sinkhole caves below Rushup Edge

Twelve sinks lie along the limestone-shale boundary to the east of Perryfoot. Short sections of accessible cave passage are associated with some of these but only at two can access be gained to more extensive systems. The westerly of these is P8 (Jackpot), with more than 1500 m of passage, while to the east the Giant's-Oxlow system has more than 4700 m of known passage.

The P8 cave (Figure 4.2) is developed mainly in rocks of the Asbian lagoonal facies. It consists of a complex of tall vadose canyons cut into the floor of large, high-level, phreatic passages which are partly fault-controlled (Smith and Waltham, 1973). These lead to a perched phreatic zone where flooded passages are separated by short sections of drained, old phreatic tubes abandoned by the present main stream. Several sumps have been explored in this part of the cave but the main stream is lost beyond sump 6. The explored pas-

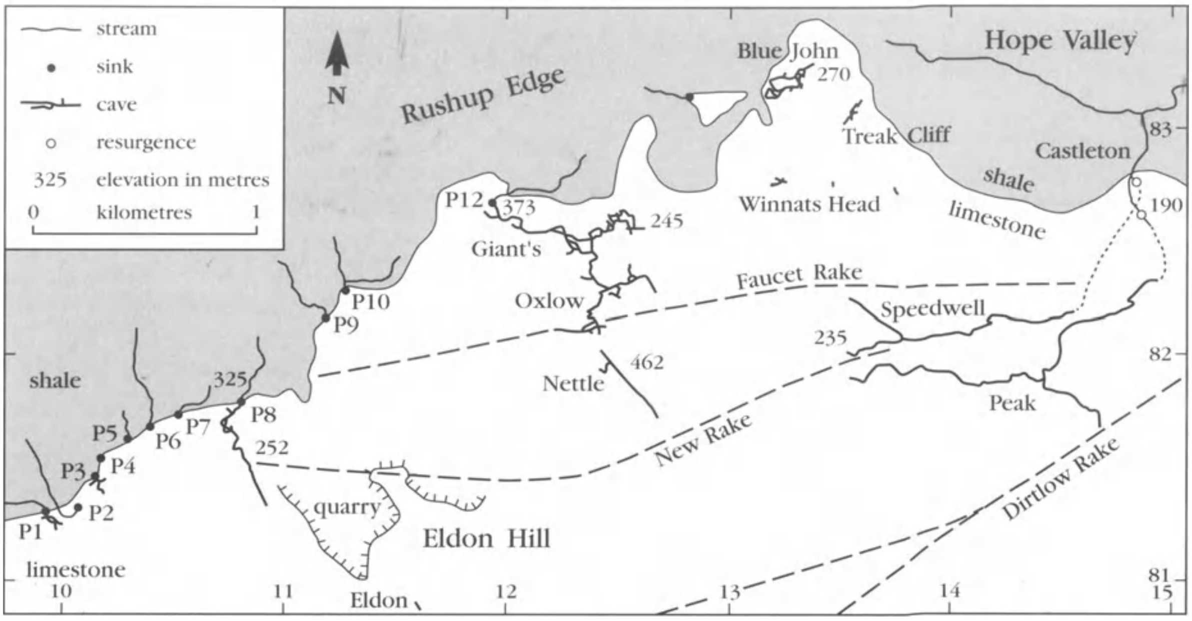


Figure 4.2 Outline map of the Castleton caves, with only the main streamways shown in the Peak and Speedwell caves at the resurgence end of the system. The rakes are mineral veins which carry some of the karst drainage though their fissure systems.

sage has been followed upstream and into a high-level flood route, where sumps 7, 8 and 9 have been passed by divers to enter a short dry passage with two shafts descending to sump 10. This is unexplored; it lies at a level about 18 m above Speedwell Main Rising. At least three separate streams now occupy the passages of P8. The entrance stream sinks into gravel in the upper part of the cave under normal conditions and is not seen again in the known cave. A second stream rises from a sump in the lower part of the cave and flows into the main sumps at the end of the lower streamway. This water is derived from sinks 2-7 (Figure 4.2) and is entirely independent of the entrance stream (Gunn, 1991). A third independent stream is met in the passage between sumps 5 and 6.

The Giant's Hole–Oxlow Caverns system (Figure 4.3) is the deepest in Derbyshire, at 214 m, and the second deepest in Britain (Westlake, 1967). The main stream in Giant's Hole flows eastwards through the reef limestones and into rocks of the bedded lagoonal facies. Here it has cut a meandering vadose canyon, the Crabwalk, more than 1 km long and up to 20 m high, though rarely more than 0.5 m wide; development of the meanders has been influenced by many small mineral veins. Beyond the Crabwalk a complex of phreatic rifts and incised canyons, formed along faults and veins, are now partly flooded and extend below present water level. Abandoned passages in the entrance series contain thick sediment sequences. Other passages, developed along bedding planes in the nearly horizontal lagoonal limestones, extend southwards from the roof of the Crabwalk and from the downstream complex to link with Oxlow Caverns, accessible via a mined shaft from the surface. Oxlow Caverns are a stacked series of phreatically enlarged vein cavities, linked by low or narrow passages. They extend over a vertical range of 150 m on Faucet Rake, and cut through the bedding plane on which lies the abandoned phreatic outlet from Giant's Hole.

Less than 100 m separates the southern end of Oxlow Caverns from the northern end of Nettle Pot (Figure 4.3), a 180 m deep, old phreatic fissure system developed largely along a post-mineralization fault. The deepest point is in parallel rifts in the Red River Series. An extensive series of low, wide chambers is formed by phreatic solution adjacent to thin toadstone lavas, at two levels about 50 m and 60 m from the surface; these extend along the fault, to Gour Passage, reaching towards New Rake, the probable route for water from P8 to the Speedwell stream cave.

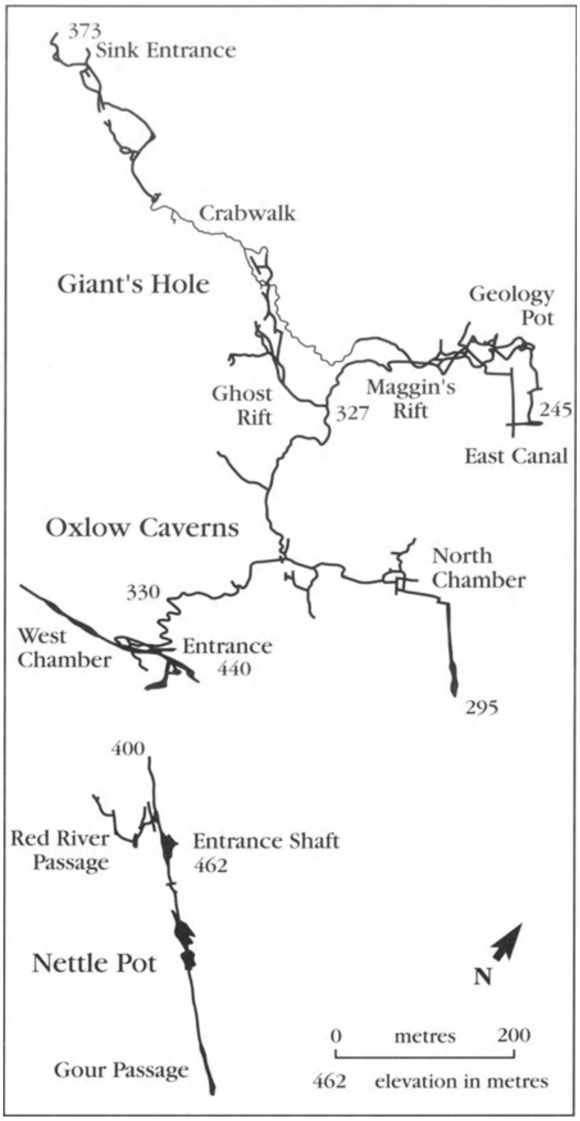


Figure 4.3 Outline map of the cave passages in Giant's Hole, Oxlow Caverns and Nettle Pot (from surveys by Eldon Pothole Club).

Relict caves of Treak Cliff and Eldon Hill

Several almost completely abandoned caves lie in the northern tip of the reef limestone where it forms Treak Cliff, a steep bank at the head of the Hope Valley (Figure 4.2).

Winnats Head Cave, contains high level, phreatic chambers up to 50 m long and 20 m high and wide, from where stalagmite has been dated to 176–191 ka BP (Ford *et al.*, 1983). Small passages and rifts connect with a series of collapse chambers and vadose shafts containing a small stream. These have been descended to a silt choked sump, at a depth of 136 m.

Blue John Cavern, now a show cave, contains a network of abandoned phreatic passages and a wide vadose canyon 25 m high. The cave intersects several much older, pre-Mesozoic, phreatic tubes filled with hydrothermal fluorspar and since partly re-excavated (Ford, 1984).

In Treak Cliff Cavern, also a show cave, phreatic chambers cut through the reef limestones and also through the reef front Boulder Bed. The latter shows mid-Carboniferous karst features in the form of solution fissures and small phreatic tubes, and contains spectacular 'Blue John' fluorspar mineralization within them. The inner chambers contain excellent speleothems, some of which have been dated at 125 ka BP, overlying ochreous clay perhaps derived from early Pleistocene periglacial loess.

Windy Knoll Cave is truncated and choked fragment of a very old sinkhole, and a fissure at its entrance has yielded a suite of Pleistocene mammal bones (Dawkins, 1875; Bramwell, 1977).

A series of sediment-choked swallet caves have been exposed in the faces of Eldon Hill Quarry (Figure 4.2). They contain a variety of facies which can be broadly grouped into four types: (1) very coarse, poorly sorted gravels dominated by clasts of arkosic sandstone derived from the Millstone Grit; (2) finer gravels dominated by well-sorted, well-rounded, decalcified chert pebbles; (3) quartzose sand; and (4) laminated cap muds. The quarry face has also broken into a decorated phreatic tube, roughly parallel to New Rake.

Further to the south, Eldon Hole is an open shaft 30 m long, 5 m wide and 60 m deep enlarged by wall-collapse of a solution chamber bounded by two subparallel joints. A large parallel chamber, decorated with speleothems in the roof, can be entered via a short passage. A downward continuation with a stream at the base, reported by Lloyd and King (1780), is now blocked by debris.

The Peak–Speedwell Cave System

More than 15 km of passages have been mapped in the connected caves of Peak Cavern and Speedwell Cavern (Figure 4.4). All the collected underground drainage of the area flows through the two main streamways towards the resurgences of Russett Well and Peak Cavern gorge. Most of the water from the Rushup Edge sinks reappears at the Main Rising in Speedwell Cavern, while the stream in Peak Cavern is derived largely from percolation water and from flood overflows from Speedwell.

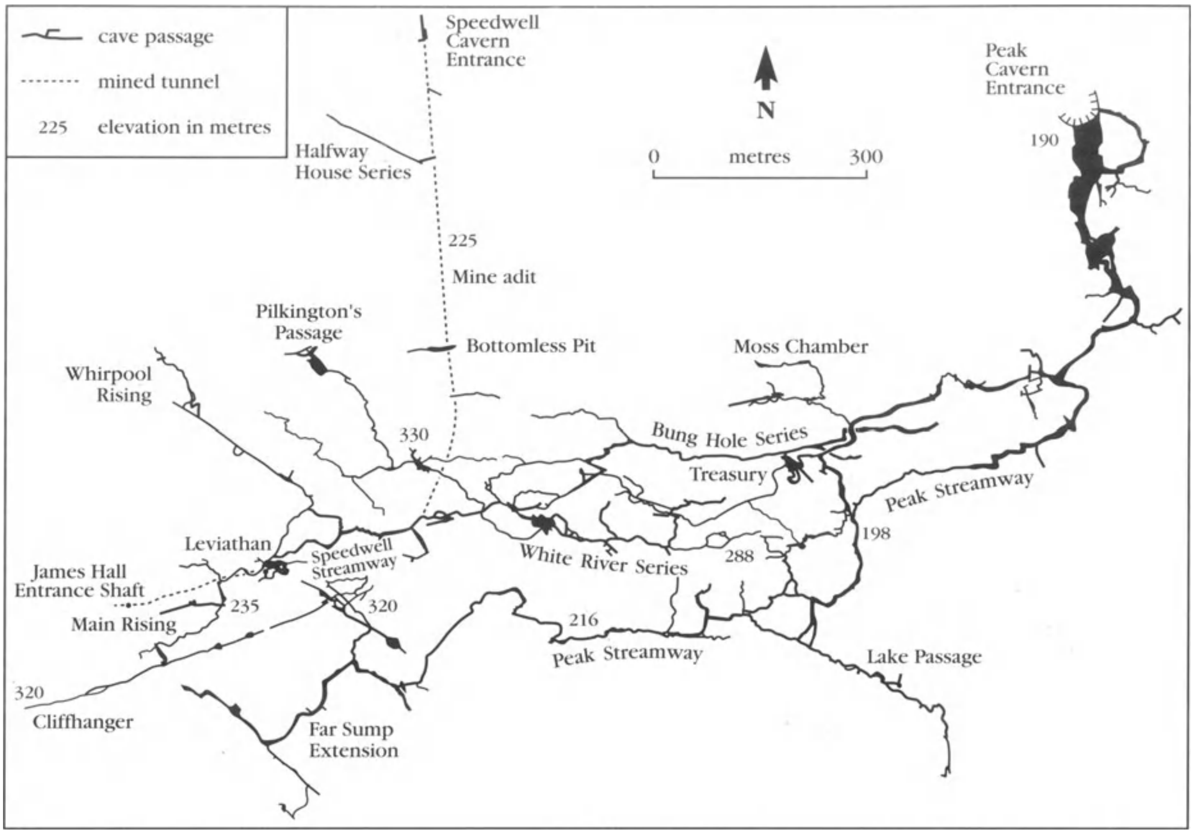


Figure 4.4 Outline map of the Peak-Speedwell Cave System (from surveys by Technical Speleological Group and Cave Diving Group).

Peak Cavern, now in part a show cave, has the largest natural entrance of any cave in Britain, more than 30 m wide and 10 m high. In the passage beyond, several avens have formed by upward solution along joints (Pitty, 1971). The Main Stream Passage, developed entirely in a single bedding plane without significant influence from joints, is a magnificent phreatic tube up to 7 m in diameter (Figure 4.5); it has several sections of vadose canyon, each entrenched where the gradient of bedding and tube is locally steeper. Within the same bedding plane, a series of smaller clay-filled tubes and joints are slowly being re-excavated. From one of these tubes, a series of avens rises 70 m into the White River Series, with more than a kilometre of old phreatic passages (Nixon 1991, 1992). These passages cut through the mineral vein of New Rake, and contain the most extensive and beautiful development of speleothems in the Peak District. At several points the old phreatic passages are breached by vadose shafts which descend 70 m to chokes on the main bedding plane. Lake Passage is a major inlet fed mainly by percolation water derived from the limestone outcrop around Dirtlow Rake (Figure 4.2). At the western end of Main Passage, the Far Sump Extension has several rifts rising more than 130 m above stream level (Cordingley, 1988). Much of the system lies beneath the Lower Lava, an aquiclude which has kept most percolation water and calcite deposition away from the lower levels.

Speedwell Cavern connects to Peak Cavern via several flooded or abandoned routes, but is more readily accessible through a mined shaft and an underground canal, which are now developed as a show cave. The tourist section ends at the Bottomless Pit, a solution cavern developed on Faucet Rake, but the canal tunnel continues south to intercept the main streamway. Most of the water from the sinks to the west usually reappears at the Main Rising, but at times the principal flow may come from Whirlpool Rising; only in flood conditions does it rise from both (Christopher, 1984; Bottrell and Gunn, 1991). Main Rising is the top of a 35 m deep phreatic loop developed along solution vein cavities in New Rake; upstream of the phreatic loop, there are two vertical phreatic

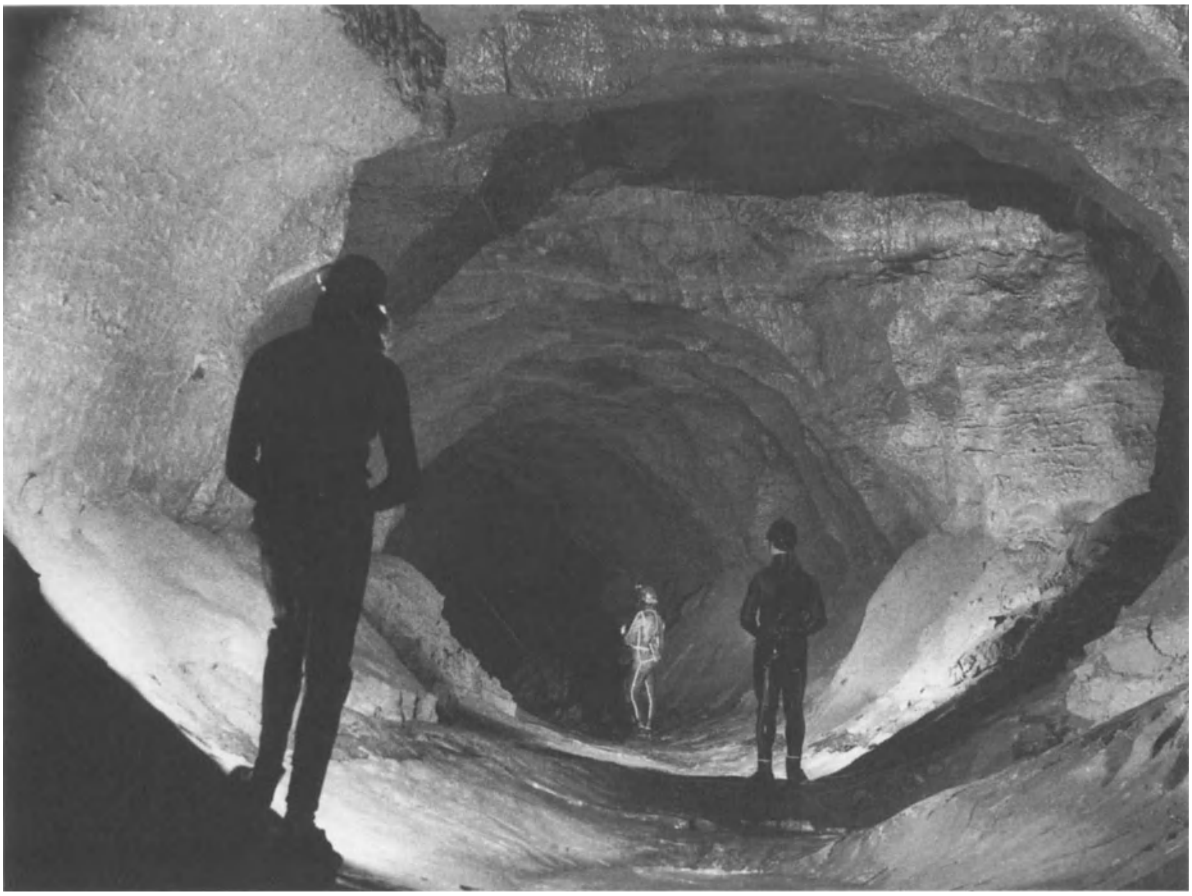


Figure 4.5 The phreatic tube which forms the main part of the stream cave in Peak Cavern. The inception bedding plane is marked by the wall niches, and this section has no vadose trench yet cut in its floor. (Photo: J.R. Wooldridge.)

lifts, and the flow emerges from a flooded rift 70 m below water level. The outlet water flows eastwards and is joined by water from Whirlpool Rising and two other inlets, before entering the long, immature passage to the Downstream Sump. It finally resurges at Russett Well, having passed beneath the Peak Cavern gorge and stream. Several older passages also enter the main upstream passage; some are partly filled with clay while others, originally entered by miners, were once steeply descending vadose inlets (Shaw, 1983). Under flood conditions some of the water from Speedwell Cavern overflows into Peak Cavern through Overspill Passage and Treasury Sump. The main passages of Speedwell are developed on a single bedding plane about 14 m below that which contains most of the Peak Cavern passages. At the western end of the system, Cliffhanger is a high-level phreatic tube, and the Leviathan is a massive and complex vein cavity above the streamway level, with a mined access to its top from James Hall's Over Engine Shaft.

Interpretation

The caves of the Castleton area constitute a complex integrated karst drainage system; this has been traced from multiple sinks to two adjacent resurgences with parallel feeders at different stratigraphic horizons. The underground drainage penetrates both reef and lagoonal limestones and has utilized mineral veins throughout its evolution. Dye-testing has established that water travelling between the various sinks and the two resurgences follows convergent, divergent, crossing and flood-related drainage routes (Christopher, 1980, 1984; Christopher *et al.*, 1981) and passes beneath the surface interfluve. There is a very long history of karstic development in the area, commencing in the mid-Carboniferous with the solution fissures carved in Treak Cliff beneath the Boulder Bed (Ford, 1984). Deep phreatic caves, probably also of considerable antiquity, developed along mineral veins; they subsequently guided the through drainage from new sinkholes. The caves have been influenced by the distribution of reef and lagoonal facies within the limestone, notably by the extensive development on the bedding planes of the lagoonal facies. The presently accessible swallet and resurgence caves show a history of development, at least as far back as the Hoxnian (Ford *et al.*, 1983). This history must be related to the episodic water table lowering and rejuvenations in response to the incision of Hope Valley through the Pleistocene.

The evolution of the Castleton cave systems is long and complex; it has been discussed in detail by Ford (1986b), and the evolution of individual cave systems has also been reviewed by Smith and Waltham (1973) and by Westlake (1967). The limestone of the Castleton area was first exposed during the mid-Carboniferous, when solution fissures in the bedrock and the Treak Cliff Boulder Bed were formed. Faulting and mineralization in the late Carboniferous and Permian produced a series of east–west mineral veins across the area. A deep, slow phreatic circulation along mineral vein cavities may have been initiated shortly after this, enhanced by the stripping of the late Palaeozoic and Mesozoic cover during Plio-Pleistocene times. With increased runoff, associated with changes of climate in the Pleistocene, a shallower system of swallet and resurgence caves developed along prominent bedding planes, though still draining via the deep mineral vein conduits. Their subsequent evolution was influenced by the incision of the major surface drainage of the area, which controlled local base levels within the limestone. Treak Cliff Cavern, Blue John Cavern and Winnats Head Cave represent former swallet caves draining off a more extensive Millstone Grit cover. The large size of the vadose canyon in Blue John Cavern indicates that it was a major sink at this time. Uranium-series dates from these sites indicate ages in excess of 190 000 BP (Ford *et al.*, 1983).

The modern swallet caves lie along the shale–limestone boundary below Rushup Edge. Their stalagmites give generally younger uranium-series dates, though the complex morphology and abandoned passages, in both Giant's Hole and P8 Cavern, indicate that they are of considerable age and have undergone extensive modification since their initial formation.

The sand infills preserved in the filled caves of the Eldon Hill quarry indicate episodes of aeolian reworking of glaciogenic sediment (Farrant, 1995). By analogy with the currently active swallet caves, the coarse and fine gravel facies may reflect sites of deposition which lie respectively proximally and distally to these ancient swallets. Uranium-series and palaeomagnetic dating of these sediments and the intercalated speleothems indicates a history of development extending back at least 780 000 years; the earliest sediments may date back to more than 910 000 BP, as an episode of normal magnetic polarity precedes the last period of reversed polarity. The caves evidently predate the valley below Rushup Edge, and are probably of early Pleistocene age.

The unusual size of the entrance chambers in Peak Cavern is due to solution and collapse in a lenticular development of back-reef shoal limestones, influenced by major joints. This was further aided by vadose entrenchment through a phreatic lift, which originally fed a vauclusian rising at the site of the modern entrance gorge (Ford, 1986b). The development of the phreatic drainage system in Peak Cavern, feeding to this vauclusian rising, probably predates an episode of incision of surface drainage in the Hoxnian, or in the Anglian glacial, which led to vadose entrenchment of the passages. The main passages in Peak Cavern are too large to have been formed solely by the percolation water which now drains through them; at some time in the past, the main drainage from the Rushup Edge sinkholes flowed through Peak Cavern, before underground capture took the water to the Speedwell Cavern route. A further similar capture appears to be developing now, as seen in the switching of flows between Main Rising and Whirlpool Rising, within Speedwell Cavern (Bottrell and Gunn, 1991). Further incision, probably in the Ipswichian, was responsible for the final draining of many of the phreatic tubes in the Peak–Speedwell system. Subsequent modification has been restricted to the infilling of some passages, by clay derived perhaps from periglacial loess, and by minor phreatic solutional enlargement of some parts of the system as a result of water dammed up by debris.

Conclusion

The Castleton limestone houses a large and important integrated cave system which shows evidence for a history of development longer than at most British karst sites. Caves have formed in different limestone facies and are closely linked with mineralized faults. The scarcity of calcite speleothems in the parts of the Peak–Speedwell cave system underneath an interbedded lava demonstrates the influence on autogenic drainage of minor aquicludes within the limestone aquifer. Speleothems and sediments within the caves have already provided evidence for a history extending back nearly a million years. The great depth range of passages within the system further increases the value of the evolutionary record of the cave and its surrounding landscape through the Pleistocene.

WINNATS PASS

Highlights

The Winnats Pass, often known just as Winnats, is the most spectacular, deeply incised karst gorge in the Peak District and has a complex origin which dates back to the Carboniferous. It is Derbyshire's best example of a fluvially excavated gorge and one of its most famous karst sites. It provides a superb transect through the Lower Carboniferous marginal reef belt.

Introduction

Incised into the northern margin of the limestone plateau 1 km west of Castleton, the Winnats Pass is regarded as one of the finest karst gorges in Derbyshire. It displays evidence of fluvial incision during periglacial events, while additional interest is provided by its complex origins which involves Pleistocene modification to a Carboniferous submarine ravine. The gorge displays a relatively clean section through the Lower Carboniferous (Dinantian) reef belt.

The origin of the gorge has been discussed by several authors, often with little supporting evidence (Sadler, 1964; Warwick, 1964; Broadhurst, 1972; Millward and Robinson, 1975; Ford, 1977a, 1986a), but no comprehensive geomorphological or chronological study of the Winnats Pass had been published until Ford presented a detailed account (1987). The geology is discussed in Broadhurst and Simpson (1973). Several caves exposed in the gorge walls and on the plateau nearby provide additional information on the evolution of the gorge (Beck, 1980; Shaw, 1983). Their relationship to the Winnats is discussed in Ford (1986a).

Description

The Winnats is a narrow steeply graded gorge cut into the steep slope on the edge of the limestone massif at the head of the Hope Valley (Figure 4.6). It drains a relatively small area of the limestone plateau at an elevation of about 400 m near Winnats Head Farm, and debouches onto the floor of the Hope Valley at 250 m altitude about 1 km west of Castleton (Figure 4.7). Less than a kilometre long, the gorge is bounded by cliffs up to

Figure 4.6 The limestone gorge of Winnats Pass, seen from the Hope Valley. (Photo: T.D. Ford.)

100 m high. Its floor is dry, as all the drainage sinks underground, and scree slopes mantle most lower parts of the sides.

Due to its position on the edge of the limestone plateau, the gorge is entrenched into, and reveals a profile through, the Carboniferous reef belt. Behind the reef to the south are the horizontally bedded lagoonal mudstones, while the reef itself is made up of thick algal bioherms. The fore-reef is dominated by two separate facies. The Beach Beds are submarine debris slopes of material transported across the reef by tidal and wave scour. The Boulder Beds are fossil talus slopes derived from pre-Namurian uplift and erosion of the reef, and postdate the beach beds (Simpson and Broadhurst, 1969). The head of the gorge is incised into the back-reef lagoonal limestones, the Bee Low Limestones, while the bulk of the gorge exposes the main algal apron reef, and the boulder and beach beds of the fore-reef.

Several caves occur in the sides of the gorge. Winnats Head Cave contains some old, high-level phreatic chambers, while Suicide Cave consists of largely abandoned passages near the foot of the gorge; an abandoned inlet system to Speedwell Cavern, the Pilkington's Cavern series, lies about 200 m south of the Winnats.

Interpretation

The origin of the Winnats Pass has proved to be controversial and enigmatic. Various ideas have been put forward to explain its origin, most of which were summarized by Ford (1986, 1987). The main theories have involved:

1. Exhumation of an inter-reef channel of mid-Dinantian age, contemporaneous with the deposition of the reef belt.
2. Recent exhumation of an erosional channel cut through the reef belt during a period of uplift in very late Dinantian or early Namurian times, and subsequently infilled with Namurian shales.
3. A collapsed cavern.
4. Superimposition of a drainage network, initiated on the Namurian shale cover, and subsequently incised into the limestone.
5. Fluvial excavation, during stages of periglacial climate within the Pleistocene, followed by underground capture of the drainage to leave the gorge dry.

The first hypothesis (suggested by Broadhurst, 1972), that the gorge was a resurrected Lower Carboniferous sea-floor channel, was discounted by Ford (1987) as he and others (e.g. Parkinson, 1953) noted that the three major lithofacies, the lagoonal, reef and fore-reef facies, strike across the pass in such a way as to preclude the possibility of a significant inter-reef channel having been present. This evidence also precluded Sadler's idea (1964) that the pass was a submarine channel in Asbian times. The Beach Beds survive up to an elevation of at least 300 m, so any channel could not have extended any deeper than that, if it was to be the source of a submarine fan; however, a very shallow channel may have existed. The presence of outcrops of the Boulder Beds in the upper part of the gorge (Figure 4.7) led Ford to suggest (1987) that the site of the Winnats was a moderately shallow channel, eroded during a period of pre-Namurian uplift and subsequently infilled with Namurian shales. The concept of the Winnats gorge originating as a collapsed cavern has been refuted by many authors. Warwick (1964) preferred the superimposed drainage hypothesis, although there is no direct evidence for it at this site.

Both Ford (1987) and Millward and Robinson (1975) advocated the Pleistocene periglacial hypothesis after comparison with other dry valleys in the area; the latter described the pass as 'cut by swift torrents of water passing down during certain pluvial phases at the end of the Ice Age'. The major problem with this hypothesis was the tiny catchment area feeding into the gorge. Ford (1987) suggested that this problem could be overcome if a large mass of stagnant ice filled Rushup Vale and fed meltwater into the valley, thus vastly increasing its catchment. The meltwater runoff during periglacial periods would have accelerated erosion of any shales which may once have filled a pre-Namurian valley. The timing of this must have occurred after the Hope Valley had cut down to the level of the Hope Terrace, as the floor of the Winnats is graded to this level, and the Namurian shale cover has been stripped back. This is interpreted as having taken place during the retreat stages of the Wolstonian glaciation, with perhaps some later modification during the Devensian. The steep initial gradient down the reef front, possibly aided by an easily excavated shale infill, contributed to the deep incision and spectacular morphology of the gorge.

Speleothem dating of several of the Castleton caves (Ford *et al.*, 1983; Ford, 1986a, 1987) has

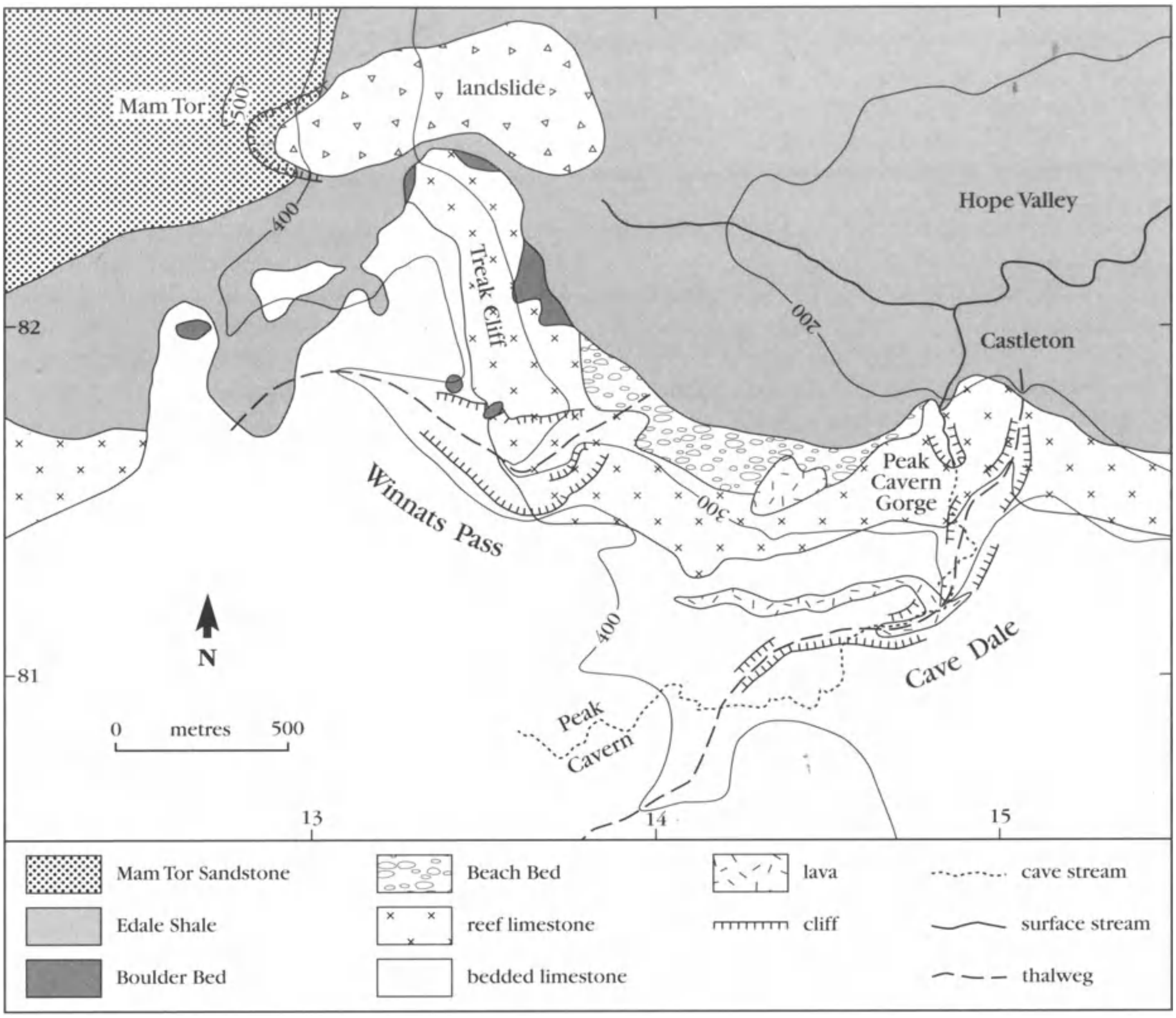


Figure 4.7 Geological map of the Castleton reef belt containing Winnats Pass and Cave Dale.

shed some light on the timing of incision in the gorge. The re-discovery of Pilkington's Cavern (Pilkington, 1789; Shaw, 1983) in Speedwell Cavern, the top of which is only some 200 m south of Winnats Pass, provided key evidence for the timing of incision. To function as an active swallet, Ford argued that the cave must have had a significant catchment area, which almost certainly extended into the area now occupied by the pass. The underground morphology demonstrates that the cave system was active before the Winnats had been cleared of shale and re-established as a gorge, and provides evidence that the shale cover had only been partially stripped back. Pilkington's Cavern therefore pre-dates the gorge incision, and has tentatively been assigned to the Cromerian interglacial (Ford, 1987). However, recent work on some infilled caves at Eldon Hill quarry (Farrant, 1995) suggests the shale cover had been stripped back earlier than previously thought and that significant cave development had begun in the area over 780 ka ago.

Ford concluded that the Winnats originated in series of stages, beginning in the Dinantian as an inter-reef hollow. It was then uplifted and excavated subaerially, to form a moderately deep channel during pre-Namurian or early Namurian times, before it was resubmerged and infilled with shales. It was exhumed and reactivated in the mid-Pleistocene when meltwater scoured out the shale fill to deepen the valley, and was further trimmed and modified during the Devensian.

Conclusions

Winnats Pass is the finest meltwater gorge in the Peak District. Its evolutionary history is both long and controversial, with early stages dating back as far as the Carboniferous. Speleothem dating and morphological studies of nearby caves suggests the gorge was mainly excavated by meltwater draining from a stagnant ice mass in the Rushup Vale during a retreat phase of the Wolstonian glaciation. The gorge appears to be located on the line of a Carboniferous ravine which was the product of both submarine and subaerial erosion.

CAVE DALE

Highlights

The deep karst valley of Cave Dale is significant in that it is a deep limestone gorge immediately underlain by a major cave system with which it has no evident genetic link. It was carved by fluvial erosion under periglacial conditions, and its lower end narrows into a rocky gorge. Another, totally separate, gorge has been formed by cavern collapse at the outlet of the underlying cave passage. The juxtaposition of these two gorges, unconnected and of contrasting origins, is unmatched elsewhere in Britain's karst.

Introduction

Cave Dale is a fine example of a dry karst valley, narrowing to a rocky gorge in its lower reaches (Figure 4.8). Immediately adjacent to the valley and underlying it is Peak Cavern, at the resurgence entrance of which is the Peak Cavern gorge. Both gorges are incised into the northern flank of the limestone plateau immediately south of Castleton and provide sections through the Lower Carboniferous reef belt.

The entrance gorge and cave system of Peak Cavern have been described in numerous publications (reviewed by Nash, 1991), but only Ford (1986a, b) has fully described the genesis of the gorge. The formation of Cave Dale has received scant attention, but its origins were debated in a wider argument over the formation of dry valleys in Derbyshire by a number of authors (Warwick, 1964; Knighton, 1975; Ford, 1986a, 1987). Dating of the underlying Peak Cavern was undertaken as part of a larger study of the Castleton caves (Ford *et al.*, 1983).

Description

The Cave Dale valley begins on the limestone plateau near Rowter Farm at an elevation of 440 m and feeds down the steep outer slope of the exhumed limestone reef front before debouching into the Hope Valley at Castleton, 240 m lower (Figure 4.7). In its upper reaches it is a shallow, wide, open grassy valley with one small tributary, incised about 10 m into the plateau. Its floor and sides are pitted with old mining depressions and

Figure 4.8 The lower part of Cave Dale looking downstream. Peveril Castle, on the left, overlooks the head of the adjacent Peak Cavern gorge. (Photo: A.C. Waltham.)

dolines. Lower down the gradient steepens (Figure 4.8), eventually forming a fine karstic gorge at the foot with cliffs over 30 m high. Resistant bands of limestone form scars along other lengths of the valley sides.

The lower reaches of the dale are graded to the Hope Valley floor which is the level of the Hope Terrace (Waters and Johnson, 1958). An outcrop of basaltic lava (the Cave Dale Lava) occurs in the middle section, creating a positive irregularity in the long profile. A spring occurs where the lava outcrops, as downward drainage through the limestone is impeded. The resulting stream flows a short distance down the valley before sinking into the limestone below the lava flow, to reappear in Peak Cavern almost directly below. Apart from this, the valley is totally dry. Some of the cliffs at the downstream end have been modified by small-scale quarrying.

Underlying much of Cave Dale are the main streamway and tributary passages of Peak Cavern. This has the largest cave entrance in Britain, sited at the head of a short narrow gorge with cliffs over 50 m high. This is also cut into the side of the limestone hill, and its floor is breached by the various resurgences which carry most of the water sinking on the plateau above. The cross-section, long profile and overall dimensions of the Peak Cavern gorge are all in marked contrast to those of Cave Dale.

Interpretation

The origins of Cave Dale were discussed by Warwick (1953, 1964) who suggested that, in common with the other dry valleys in the area, it developed through superimposition of a complex drainage network initiated on a Namurian shale cover. Rejuvenation led to the desiccation of the tributary valleys, after the formation of knickpoints in their floors. Knighton (1975) put forward an alternative interpretation for these knickpoints, advocating that the steepening of the thalwegs was a response to maintain flow continuity where geology imposed constraints on the adjustability of width. Ford (1986a) noted that the step in the Cave Dale profile was probably a structural feature caused by the outcrop of the basalt lavas rather than a true knickpoint. In a series of publications dealing with the limestone geomorphology of the Castleton area, Ford (1977b, 1986a, b, 1987) identified the role of periglacial meltwaters in the formation of the Cave Dale. He suggested an Ipswichian age for the main period of incision based on dating evidence from the underlying Peak Cavern (Ford *et al.*, 1983).

The origins of the Peak Cavern gorge are also discussed by Ford (1977a) and Ford *et al.* (1983), who noted that the gorge showed evidence of having been a vauclusian spring, which was initiated during the phreatic development of Peak Cavern during the Hoxnian interglacial. Downcutting through the lip of the vauclusian spring coupled with roof collapse has created the spectacular gorge seen today. Its roofed-over continuation can be seen in the entrance chamber of Peak Cavern.

The lower part of Cave Dale overlies one of the largest chambers in Peak Cavern. The only connection between the dale and the cave is through a very narrow fissure (now blocked); this is not an old sink, but is a phreatic rift in the cave roof which has been intersected by the valley. This lack of relationship between Cave Dale and Peak Cavern supports the view that Cave Dale is a young valley, excavated when the ground was frozen and the cave below was temporarily inactive (Ford, 1986a).

Conclusions

Cave Dale and Peak Cavern provide a valuable and exemplary site with two genetically unrelated types of limestone gorge, one formed largely by cave unroofing, the other by subaerial fluvial erosion during a periglacial period. Each gorge is a fine example of its type in its own right. Cave Dale is also significant in being a deep limestone gorge immediately overlying a major uncollapsed cave system, to which it is genetically unrelated.

BRADWELL DALE

Highlights

Bradwell Dale and its upstream continuation, Stanlow Dale, lie along the margin of the karst, south-east of Castleton. They are the product of gorge incision caused by reef knolls interrupting the uniclinal shift of a valley excavated along a shale-limestone interface. Within Britain, this type of gorge is unique to the Peak District karst, where these two dales exhibit the clearest morphology of the type.

Introduction

Bradwell Dale and Stanlow Dale (Figure 4.9) form part of a dry valley network which drains north along the east dipping margin of the limestone plateau, immediately south of Bradwell. The valley is incised into the Lower Carboniferous limestones of the Eyam Group, which consist of both bedded, back-reef, lagoonal carbonates and also mounded reef knoll limestones. The gorge is important in demonstrating the role of the reefs in obstructing the downdip uniclinal shift of the valley to the east, instead forcing the valley to incise vertically, creating the gorge seen today. Modern drainage is now underground, so that the gorges, which were developed under periglacial conditions in the Pleistocene, are now dry.

The geomorphic evolution of Bradwell Dale and its associated caves were comprehensively described by Ford *et al.* (1975). These and several other anomalous gorges in the Peak District were further assessed by Ford and Burek (1976) who outlined the role of the reef knolls in the gorge formation, and by Ford *et al.* (1977a) as part of an overview of the karst geomorphology of the Bradwell area. The chronology of Bagshaw cavern, and its implications for Bradwell Dale, was outlined by Ford *et al.* (1983).

Description

The gorges and dry valleys are cut into the Carboniferous limestone, which dips east at 5–10°. The limestone then disappears under the Edale shales a few hundred metres to the east, where a broad strike valley draining to the north has developed along the limestone–shale boundary. The main Bradwell and Stanlow gorges are developed slightly updip along a prominent line of knoll reefs (Figure 4.9). They are incised up to a depth of 40 m into massive limestones, forming steep cliffs and craggy outcrops. The gorges extend some 2 km from Nether Water Farm in the south, north along strike to Bradwell village. Tributary to these are the dry valleys of Hartle Dale and Intake Dale which drain east down the limestone dipslope, meeting the gorge at Hazlebadge Farm. All the valleys are now dry, as modern drainage is underground. Several stream sinks are present along the shale boundary and these feed to a major resurgence in Bradwell village. The largest cave system associated with the gorge is Bagshaw Cavern (Figure 4.9); there are

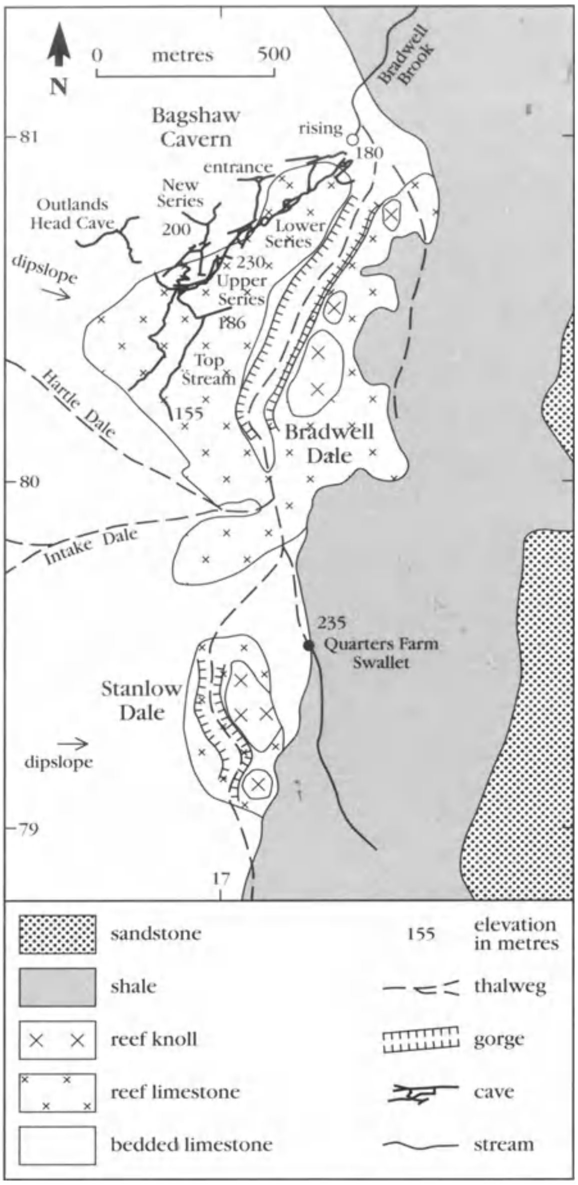


Figure 4.9 Geological map of Bradwell Dale and Stanlow Dale. Bagshaw Cavern is shown in outline, and lies mainly in the bedded limestones beneath the reef knolls (from survey by Eyam Exploration Group).

also other small sinks and cave fragments within the gorge and its tributaries, including fissures which have yielded Pleistocene mammal remains (Ford *et al.*, 1977a).

Interpretation

Ford and his co-workers (1975, 1976) described the role of the knoll reefs in the formation of Bradwell Dale, Stanlow Dale and the various other deeply entrenched valleys, or anomalous gorges, in the Peak District karst. They recognized that Bradwell Dale evolved along the shale/limestone boundary, and gradually shifted uniclinally eastwards and downdip as the shale margin was eroded back. Eventually, the original river draining the base of the dip slope was trapped by a series of reef knolls and prevented from migrating any further east; from then on, vertical incision predominated. The gorges of Bradwell Dale and Stanlow Dale were thus incised immediately updip of the main reefs. The shale cover continued to be stripped back, forming the broad shallow strike valley to the east of the gorge along the shale–limestone contact.

The chronology of the area was discussed by Ford *et al.* (1983). They noted that the gorge was graded to the level of the Hope terrace, like Cave Dale and the Winnats, and concluded that the gorges were mainly incised just following the Anglian or pre-Anglian glaciations. All authors have recognized that both gorges, and their associated dry valleys, were incised into the limestone by the action of subaerial fluvial erosion during periglacial periods (Ford, 1977a). Some erosion of the gorge may have predated adequate development of underground drainage; however, the truncation of old high-level phreatic passages provides evidence that at least part, if not most, of the gorge was incised during the Pleistocene cold periods when underground drainage was restricted.

Further work on the dating, by uranium-series and other methods, of both the major cave systems and some of the isolated high-level phreatic fragments may allow the rates of valley incision and uniclinal shift to be deduced; this could provide important evidence on the evolution of the area.

Conclusions

The site encloses part of a fine karst valley system with two gorge sections, which are the clearest examples in the Peak District where incision is due to the prevention of a valley migrating uniclinally downdip by reef knolls in the limestone sequence. British examples of this phenomenon are found only in the Peak District. The tributary valleys are also good examples of dry karstic valleys in their own right, abandoned by the modern drainage which is underground.

BAGSHAW CAVERN

Highlights

Bagshaw Cavern lies immediately behind a large resurgence, and is developed at a very high stratigraphic level within the limestone sequence. It has a less complex configuration than other cave systems in the Peak District, reflecting its more recent development.

Introduction

Bagshaw Cavern lies west of Bradwell Dale (Figure 4.9), cutting through a gentle anticline plunging with the easterly dip. It lies in thin-bedded limestones of upper Brigantian age, at a stratigraphically higher level in the Carboniferous Limestone than any other cave in Derbyshire. Its development has been influenced by the incision of Bradwell Dale and the presence of reef mounds to the east. The cave stream has a large catchment area of autogenic input from limestone moors, updip to the west, and this is joined by allogenic waters entering sinkholes along the shale boundary east of Bradwell Dale. The Bagshaw resurgence, at the head of Bradwell Brook, is one of the largest risings in the Peak District karst.

The evolution of Bagshaw Cavern and Bradwell Dale has been discussed by Ford *et al.* (1975, 1977a), and aspects of its hydrology have been discussed by Christopher and Wilcock (1991). Description of the cave passages are given by Gill and Beck (1991), and in a brief account by Ford and Gunn (1992).

Description

Bagshaw Cavern is entered via a staircase in an old mine working. At the foot, a phreatic tube partly full of sediment leads south to join the main cave at the Dungeon, a 6 m deep pothole. From the bottom of this, a lower series of intermittently active passages trends north-east towards the resurgence (Figure 4.9). Above the Dungeon, an upper series of abandoned passages trends south-west along the strike. Both upper and lower series display fine vadose downcutting in phreatic bedding tubes up to 4 m high and wide. The system is developed in thinly bedded limestones, and many passages show good examples of tabular collapse of roof blocks. The active stream is normally seen only for short distances at the two ends of the system between flooded sections of its route; upstream in Top Stream Passage, the water rises for over 30 m through flooded caverns. Much of the flow comes from the Quarters Farm Swallet (Figure 4.9).

The western end of the main passages approaches the abandoned phreatic tubes and rifts of Outlands Head Cave. The New Series of Bagshaw, discovered in the 1930s, extends into a calcite pipe vein and contains chambers decorated with straw stalactites up to 2 m long. An aven rises from it into Batham Gate, a high-level passage almost directly above the main series and extending south parallel to Top Stream.

Interpretation

The Bagshaw Cavern drainage system is unusual in crossing a limestone anticline close to a major dry valley. Its development was closely linked to the formation of Bradwell Dale and the erosion of the limestone plateau to the west. As the shale cover was progressively removed eastwards, unroofing of the limestone allowed entry of sinking water and the development of phreatic circulation. Surface drainage experienced a uniclinal shift downdip until trapped by reef mounds, at which point Bradwell Dale began to form immediately to their west; it therefore cut down through a gentle eastward-plunging anticline, instead of following the strike round the nose of the fold. Most of Bagshaw Cavern was a late development, largely postdating the initial excavation of Bradwell Dale during the Hoxnian or Ipswichian interglacial. The cave is a young system, of relatively simple geometry, which has not developed to the same extent as those around Castleton or Stoney Middleton, despite having a comparable throughflow of water. The main passages lie at altitudes of 190–210 m, and the older conduit of Batham Gate is at the 230 m level, maintaining its level by following the strike. With continued incision of Bradwell Dale the underground drainage has migrated downdip; downstream of its phreatic lift in the Top Stream sump, the present streamway lies close to the 180 m level, just above the resurgence. Both the upper and lower series of Bagshaw Cavern exhibit classic vadose trenching of phreatic bedding passages, while permanently flooded passages represent the youngest part of the system.

Conclusion

Bagshaw Cavern shows close control both by geological structure and also by the evolution of the adjacent dale. It has a simpler geometry than other cave systems of comparable drainage capacity, due to its youth and short history. It is developed in thinly bedded limestones which permit extensive tabular roof collapse; both this and the long straw stalactites are unusual features in the Peak District caves.

STONEY MIDDLETON CAVES

Highlights

The caves of Stoney Middleton show, with exceptional clarity, the development of a series of phreatic cave levels in response to base-level lowering and the presence of aquicludes and bioclastic horizons. They provide a valuable record of landscape modification in this area of the Peak District through the Pleistocene.

Introduction

Stoney Middleton Dale is a deep limestone gorge draining eastwards to the River Derwent (Figure 4.1). Allogenic recharge into the karst aquifer occurs at the Waterfall Swallets, north of the Dale head, where streams sink off the Namurian shale under the Millstone Grit escarpment, and also by a sinking stream from a shale outlier at Wardlow Mires, 3 km west of the gorge. The Dale is the thalweg out of a wide, shallow, topographic basin, which reflects the structure of the Wardlow syncline with the shale outlier at its centre. The catchment is bounded to the west of the basin by the outcrop of the Litton Tuff, which acts as an aquiclude, maintaining a large groundwater reservoir within the basin. Several large phreatic cavities, possibly of considerable age (Beck, 1977), lie beneath the southern and eastern flanks of the basin. The catchment area for the risings at Stoney Middleton covers 17 km^2, with 60% on limestone. The discharge is now entirely by mined drainage soughs, except in flood conditions, when the estavelles at Wardlow Mires and Carlswark Cavern discharge large streams.

The geological setting, evolution and hydrology of the caves of this site have been discussed by Beck (1975, 1977), Christopher and Beck (1977)

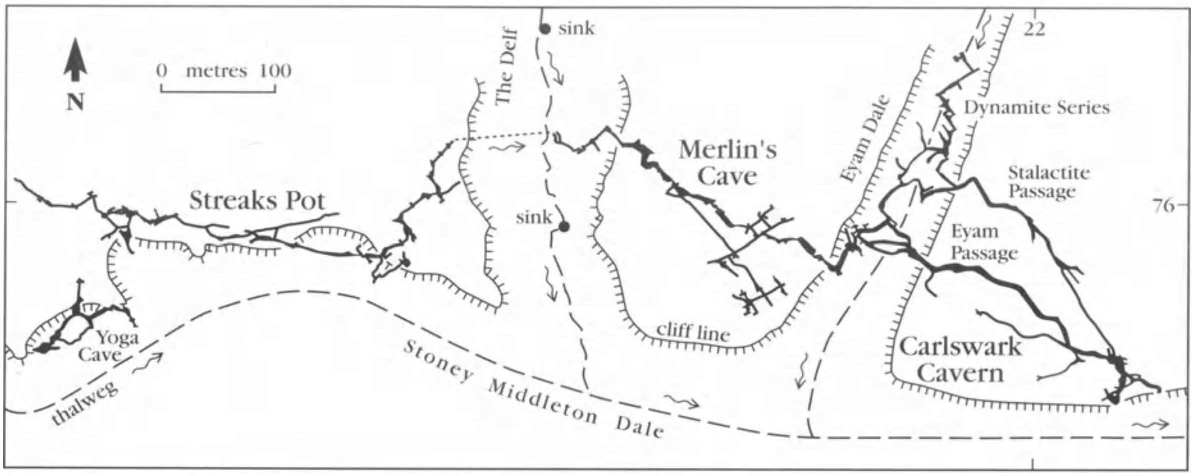


Figure 4.10 Outline map of the cave systems under the northern flank of Stoney Middleton Dale (from survey by Technical Speleological Group).

and Ford and Gunn (1992). The caves are described by Gill and Beck (1991), with an account of Streaks Pot and Merlin Mine in Beck (1990).

Description

Waterfall Swallet is the largest of the sinks on Eyam Edge. It lies in a large doline which fills in flood and overflows into the adjacent cave system of Waterfall Hole. This cave reaches 43 m deep in a series of rift caverns, extensively modified by collapse, which enlarge beneath the wayboard at the base of the Eyam Limestone. Little Waterfall Swallet lies on the same fracture system a short distance to the north-east. Sinkholes at Eyam are largely hidden by the culverts which carry the Jumber and Hollow Brooks through the village, but one has been followed to a depth of 100 m in vein cavities. These brooks continue southwards via the Delf and Eyam Dale respectively, seeping into their valley floors in dry weather, or joining the Dale Brook in wet conditions.

Stoney Middleton Dale exposes an almost continuous section, 3 km long, through the Brigantian Monsal Dale limestones of the Carboniferous. Within this sequence several important speleogenic horizons have been recognized (Beck, 1975). A number of caves are developed, mainly along the strike on these horizons (Figure 4.10); they form a series of levels which reflect external erosional events.

The highest and oldest part of the cave system is the First Remnant Complex, represented by a series of tubes at levels of 210–216 m (Figure 4.11). Vadose feeders to the system are represented by Cucklet Church Cave and The Saltpan with their isolated fragments of passage in crags west of the Delf. The Second Remnant Complex is seen only as a large phreatic tube in the Dynamite Series of Carlswark Cavern; it lies directly beneath the First Remnant tube, which it clearly postdates.

The Carlswark Complex is the most extensive level of the system, with the majority of Carlswark Cavern, Streaks Pot and Yoga Cave developed at a level of about 180 m. Carlswark Cavern has two main relict phreatic tubes. Eyam Passage lies to the south and Streaks Pot represents its truncated

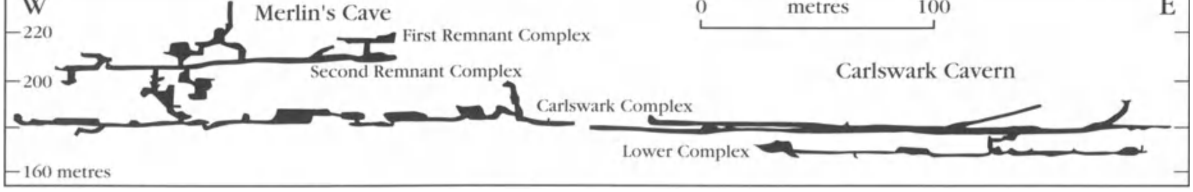


Figure 4.11 Long profile through Merlin's Cave and Carlswark Cavern showing the development on four levels (after Christopher and Beck, 1977).

continuation west of the Delf. Stalactite Passage is the northern tube downstream of the joint-controlled phreatic rifts of the Dynamite Series. The entire Carlswark Complex is developed at the base of a limestone bed crowded with silicified *Gigantoproductus*, where a thin clay bed has arrested vertical percolation and initiated development of the network of tubes. Eyam Passage has a spectacular roof formed in the Lower Shell Bed, and also reveals excellent examples of bedding plane anastamoses. The Lower Complex is little known since it lies mostly below the thalweg and is largely flooded.

Interpretation

Cave development in the Stoney Middleton area has been influenced to an unusual degree by a combination of stratigraphic and surface topographic controls. The Lower Shell Bed is a bioclastic horizon, directly underlain by a clay wayboard aquiclude, which acted as an important inception horizon for the phreatic caves. Joint control is also conspicuous within the caves, both at the speleogenic horizons and as rifts linking them vertically.

The sequence of four cave levels represents a succession of shallow phreatic networks developed in response to intermittent rejuvenation and incision of the River Derwent upstream of the Matlock knickpoint (Beck, 1977). Each cave level was formed where favourable inception horizons lay just below the contemporary water table which was gently graded towards base level at the river. The minimal vadose trenching within the phreatic tubes suggests that each new level captured the entire drainage and fossilized the upper levels very rapidly (Ford *et al.*, 1983). The higher-level passages tend to be larger than those lower down, perhaps reflecting the formerly greater extent of the shale cover, and hence larger catchment area of allogenic water, at the time that they were active. Similarly, the position and vadose character of Cucklet Church Cave also suggest a more extensive shale cover at the time of initiation. Sedimentary fills in the abandoned levels demonstrate several stages of infilling and re-excavation; correlations with terrace levels suggest that the old highest level in the caves was active in pre-Anglian times, but stalagmite dating has not yet provided a chronological framework for all the levels (Ford *et al.*, 1983).

Conclusion

The Stoney Middleton caves provide excellent examples of the influence of aquicludes on the level of passage development. The successive levels of passage development also record, with exceptional clarity, the effect of lowering of surface drainage on underground drainage levels.

POOLE'S CAVERN

Highlights

Poole's Cavern is a large section of cave passage with an underfit stream fed by karst drainage from the south. A proven hydrological link with nearby thermal springs provides valuable information on the nature of recharge to such phenomena. Stalagmites which have developed on gas pipes within the cave provide important data on the growth rate of speleothems.

Introduction

Poole's Cavern lies north-east of Stanley Moor, and represents the only significant length of accessible cave passage within the catchment of this upland karst (Figure 4.1). Several sinkholes lie close to the boundary of the overlying shale on Stanley Moor. All drain to a series of resurgences in the floor of the Wye Valley at Buxton, and several drain via Poole's Cavern in all but very low flow conditions. Although the flow route is confirmed by dye tests, passage sizes at the sinkholes are small and none of the cave streams can be followed far underground. Poole's Cavern is the only large cave in the area. Coal fines were formerly stored in the Grinlow quarries, midway between Stanley Moor and Poole's Cavern, and have appeared in Poole's Cavern and at the Buxton hot springs, suggesting that some of the cave stream joins the hydrothermal system after leaving the cave.

The caves and hydrology of the area have been discussed by Ford (1977c), Gunn and Edmans (1989) and Ford and Gunn (1992), and the passages in Poole's Cavern were described by Gill and Beck (1991).

Description

Poole's Cavern is currently operated as a show cave. Though only 240 m long to a boulder choke, it has an impressive main passage up to 20 m high and wide; this is an excellent example of solutional enlargement in a dense system of beddings and joints, with solutional undermining and collapse. Subsequent vadose erosion has removed most of the fallen blocks. The cave contains good examples of stalactites and a massive bank of flowstone with large gour terraces. Pitty (1969) has used this site to ascertain the difference in response times of percolation water and stream water in order to distinguish contrasting residence and through flow time of different components of the karst groundwater. The entrance passages contain thick sediment sequences, yet to be documented in detail, which have yielded Pleistocene mammal bones and include undisturbed stalagmite layers; Romano-British material lies on the Pleistocene silts. Further sediments were excavated from the cave entrance by Victorian archaeologists, but the finds were poorly recorded and preserved.

Interpretation

The phreatic origins of Poole's Cavern and its position adjacent to the modern Buxton valley suggest a considerable age for the cave and a history which may extend as far back as the Anglian glaciation. The undisturbed bone deposits and stalagmite layers provide a record of Pleistocene events and climatic change in this area, with the possibility of absolute dates being obtainable from the stalagmite layers. An unusual feature is the stalagmite columns, which have been deposited up to 100 mm high on a century-old gas pipe; they provide data on the growth rate of speleothems, but their development may have been influenced by lime-burning in Grin Woods above. Some of the stalagmites have unusual colouring distinguished by their resemblance to poached eggs.

Poole's Cavern represents a former resurgence for the area. The main underground drainage now takes a different route, probably via Green Lane Pot (SK 050726), to the resurgences at Otter Hole (SK 046733) and Wye Head (SK 050751), both about 45 m lower. Although the hydrology of the area appears fairly simple, the connection with the Buxton hot springs suggests that some of it follows a much deeper phreatic route which is probably fault-guided. Alternatively, the shallow drainage from the Stanley Moor sinkholes has intercepted an independent, deep phreatic system. Tritium contents of the Buxton spring water shows that it has been underground only for 15–20 years; it appears to be meteoric water which has passed through unusually deep systems of karstic fissures (Ford, 1977c).

Conclusion

Poole's Cavern is an isolated segment of large cave passage containing clastic sediment deposits which incorporate both vertebrate remains and stalagmite layers. This sequence preserves a valuable record of climatic and geomorphological change in this area through the Pleistocene. The karst hydrology of the area encompasses both the shallow drainage from moorland sinkholes and also the deep recharge to thermal springs.

LATHKILL DALE

Highlights

Lathkill Dale is a dendritic dry valley system deeply entrenched into a karst plateau; it is the best developed in the Peak District, and among the finest in Britain. The River Lathkill emerges from several springs at the lower end of the Dale, and has some of the most important examples of barrage and sheet tufas in Britain. The complex hydrology has been considerably affected by mine drainage.

Introduction

A dendritic network of shallow dry valleys incised into the limestone plateau drains eastwards from around the village of Monyash, and feeds into the head of Lathkill Dale (Figure 4.12). The upper part of the valley network is guided by the synclinal geological structure and is dry above Lathkill Head Cave, which is an active resurgence in wet weather. Below this cave, the surface stream is intermittent and seasonal until Pudding Springs are reached below the junction with Cales Dale, where the stream is permanent except in extreme drought. The surface flow is maintained in part due to toadstones (Carboniferous lavas) and less permeable limestones exposed in the valley floor.

The dry valley network was studied by Warwick (1953, 1964), while Ford and Beck (1977) expanded on his work, studying the chronological relationship of the downcutting of the dale to

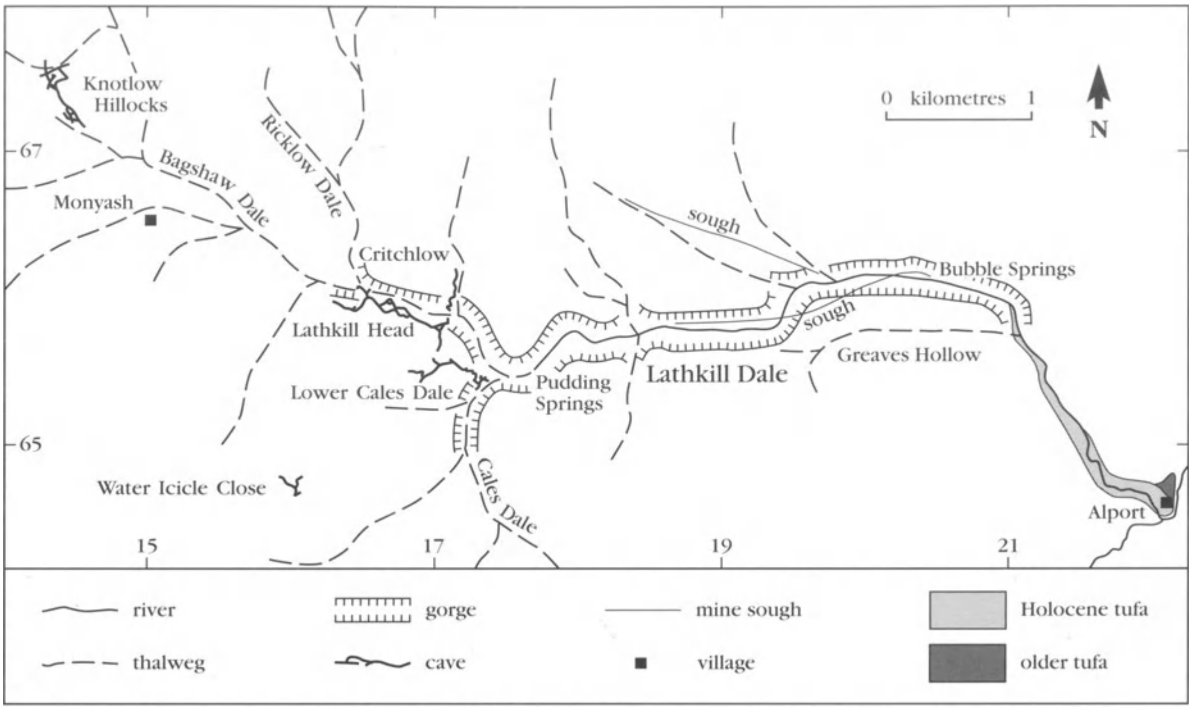


Figure 4.12 Outline map of Lathkill Dale, its tributary dry valleys and its associated cave systems.

glaciation. In the lower part of the dale, thick deposits of tufa occur, some of which are still actively growing and form a series of barrages across the river (Towler, 1977; Aitkenhead *et al.*, 1985; Ford, 1989b; Pedley, 1993). These have enabled Pedley (1993) to deduce part of the Quaternary history of Lathkill Dale, but this is complicated by the considerable modifications imposed on the dale by past measures to affect mine drainage (Bamber, 1951; Robey, 1965). A comprehensive study of the dale's geomorphological history is awaited, though Ford *et al.* (1983) outlined a tentative chronology based on cave levels related to the Derwent terraces. The many caves in the dale are described by Gill and Beck (1991).

Description

The upper part of the Lathkill catchment consists of an elongate bowl centred on the village of Monyash, with a group of shallow dry valleys leading out to the east. These coalesce into a dry gorge with steep rocky sides incised up to 75 m below the level of the plateau surface. The gradient of the valley steepens markedly into the gorge section, and the steep rocky sides are fringed by coarse screes (Figure 4.13). At the lower end of this gorge, Lathkill Head Cave is the wet weather resurgence of the River Lathkill; in flood conditions this discharges a very large flow (Gill and Beck, 1991). Directly opposite is Critchlow cave which also discharges water in flood. Below this several more springs add to the stream, depending on the stage of flow, with water emerging from a spring at Holme Grove, from the Lower Cales Dale Cave, from Pudding Springs a kilometre downstream, and from Bubble Springs 3 km from Lathkill Head Cave. During severe drought, there may be no flow above Bubble Springs.

The general trend of Lathkill Dale is eastwards along the line of a gently plunging syncline. The dale exposes the Monsal Dale Limestones; these are mainly pure calcarenites with coral bands, but they include a lower facies of dark limestones which are rich in shale partings, thinly bedded and less permeable. Ford and Beck (1977) suggest that the dark limestones would have helped to maintain a surface flow along the dale, if the mine drainage had not artificially lowered the water table and captured much of the tributary input. At Bubble Springs, faults bring a bed of lava to the surface.

For much of its length below Pudding Springs, the floor of Lathkill Dale is covered partly by the remains of artificial dams, placed to improve the fishing, and partly by a large sheet of tufa. The tufa

Figure 4.13 The entrenched and normally dry section of Lathkill Dale just above the Lathkill Head flood resurgence. (Photo: A.C. Waltham.)

is most extensive at the lower end, and two tufa phytoherm barrages occur between Bubble Springs and Alport (Figure 4.12), with pool deposits in between (Pedley, 1993). The fossil tufa deposits further upstream are now being eroded by the stream or grassed over, and that at Pudding Springs has been modified by quarrying. An older, massive tufa sheet forms a cliff up to 8 m high and 150 m long immediately north-east of Alport (Figure 4.12).

Interpretation

Warwick (1964) described the Lathkill valley network in some detail and was convinced that the majority of the valley evolved from a complex drainage pattern initiated on overlying, impermeable shales, and that rejuvenation had led to progressive elimination of the tributary valleys. Ford and Beck (1977) suggested the main development of Lathkill Dale took place during the Pleistocene cold phases when periglacial conditions allowed surface flow of glacial meltwater and valley incision. They suggest that incision was initiated during the Last Interglacial, and followed the axis of the syncline; remnants of glacial till survive on the plateau. Initial incision was greatest at the downstream end and was via Greaves Hollow, which is now truncated and dry. Subsequently, the river was offset to the north (Figure 4.12), probably by some form of river capture (Ford and Beck, 1977), to follow the line of the mineral veins.

The role of rejuvenation appears to have been more limited in Lathkill Dale than it was in the Dovedale and Manifold valleys. None of the tributary valleys in the upper section hangs above the main valley floor (Warwick, 1964). Only lower down do some of the tributaries hang, and then they do so only by a few metres. One hanging tributary lies immediately south of Lathkill Head Cave (Warwick, 1964), but this may be a structural feature where contrasting limestone lithologies are juxtaposed across a mineralized fault (Ford and Beck, 1977). In other cases the tributary valleys may be partially infilled, thus apparently hanging above the main valley from where any infill has been removed.

The hydrology of Lathkill Dale is complex, has been much affected by mine drainage (Bamber, 1951; Ford and Beck, 1977), and is far from being fully understood. A natural phreatic cave system drains the upper part of the dale; Lathkill Head, Critchlow and the Lower Cales Dale caves represent the flood overflow or epiphreatic parts of the system. Below the Cales Dale junction, the Lathkill Dale and Mandale soughs were cut to drain the mines and have altered the flow regime, so that the main water now reappears at Bubble Springs.

Considerable deposits of tufa blanket the valley floor between Bubble Springs and Alport (Figure 4.12). Reference has been made to the now inactive tufas above Bubble Springs (Aitkenhead *et al.*, 1985; Ford, 1989), the most detailed account being in Towler (1977), while Pedley (1993) produced a detailed study of the active tufas downstream of Conksbury. These modern tufas are minor in extent compared to their Holocene counterparts, due in part to the ponding of the river to improve fishing and the general lowering of the water table by mine drainage in the eighteenth and nineteenth centuries. The restriction on active tufa formation is due to both the falling water tables drying up the river and also man-made pollution inhibiting the algal growth. Pedley concluded that the pre-tufa Lathkill gorge was deepened during the earlier Devensian, with active phytoherm development causing ponding during a late Devensian interglacial. Tufa then accumulated through much of the Holocene, and isotope analysis has yielded a dated record of climatic and environmental changes from 10 000 to 4000 BP (Andrews *et al.*, 1994; Taylor *et al.*, 1994). Mean temperatures reached a maximum around 8000 BP, and the forest cover was largely cleared in two stages by 5000 and 4000 BP. The tufa in the cliff above Alport was probably deposited during the Ipswichian interglacial.

Ford *et al.* (1983) suggest that the earliest phase of cave development was that associated with Water Icicle Close Mine, which has been dated in excess of 350 ka. The next phase was the development of Upper Cales Dale Cave, followed by renewed incision and the formation of Lathkill Head Cave. Limited incision of the dale occurred during the Devensian, followed by calcite deposition in the caves. Further dating of the speleothem sequences within the Lathkill caves is needed to confirm this chronology, and dating of the travertines could provide further data on Quaternary environments. Further dye tracing is required to resolve the complex hydrology of the karst.

Conclusions

Lathkill Dale is one of Britain's finest examples of a dendritic dry valley system. It was largely developed under periglacial conditions during Pleistocene cold phases. Surface flow occurs over differing lengths of the river bed depending on stage, with partial parallel drainage through an immature phreatic cave system. Flood waters discharge through a higher-level cave system. The hydrology is complicated by the effect of mine drainage in the eighteenth and nineteenth centuries. The detailed morphology of the valley shows a close relationship to the limestone lithology and structure. Several fine tufa barrages and sheets occur in the lower part of the dale and provide some evidence for the Late Pleistocene and Holocene development of the valley.

UPPER LATHKILL DALE CAVES

Highlights

The caves of Upper Lathkill Dale are features of abandoned or intermittently active karst drainage, which now lie isolated from any significant surface catchment. The high-level caves are relics of early Pleistocene landscapes, and contain examples of sediments no longer found on the surface here. The intermittently active caves demonstrate the nature of karst drainage fed entirely by percolation sources.

Introduction

The limestone basin at the head of Lathkill Dale contains isolated fragments of large phreatic cave passages at Lathkill Head, Water Icicle Close and Knotlow Mine (Figure 4.12). All the caves lie in the bedded, lagoonal facies of the Monsal Dale Limestones, and are far from any allogenic drainage sources; they are believed to be fed solely by autogenic input. The River Lathkill flows from one of the largest areas of riverless terrain in the Peak District karst, and it represents a classic example of a river flowing on or beneath the limestone valley floor depending upon the amount of preceding rainfall. After periods of wet weather, the river flows from Lathkill Head Cave, though at other times the dale may be dry as far down as Bubble Springs (Figure 4.12). This situation is no longer entirely natural, as the regional water table has been lowered by Hillcarr Sough to the south and also by Lathkill Dale Sough driven below the valley floor and Mandale Sough beneath its northern flank (Oakman, 1979). Whether the river flowed permanently on the surface before the construction of these drainage adits is uncertain. In historical times the river flowed from as far up as Monyash, and was restricted to flowing from

Lathkill Head Cave only in times of drought (Bamber, 1951; Oakman, 1979).

The caves and karst of Upper Lathkill Dale have been discussed by Bamber (1948, 1951), Ford and Beck (1977) and Ford and Gunn (1992); the cave passages are described by Gill and Beck (1991).

Description

The large open entrance of Lathkill Head Cave discharges a powerful stream in winter, but dries out allowing access to the cave system in summer. The main passage is a bedding-guided phreatic tube, typically 1 m high and 5 m wide, which can be followed upstream to high, solution-enlarged joints, with a series of large chambers well decorated with speleothems. Further low passages have been followed north-west, and a series of small phreatic tubes pass beneath the dale floor to link with the lower levels of Ricklow Cave, with its entrance on the north side. A low, sediment-filled distributary extends downstream to within 150 m of Lower Calesdale Cave. Critchlow Cave lies directly opposite Lathkill Head, and has 800 m of low, partly sand-filled, phreatic passage, with several small chambers decorated with speleothems. Lower Cales Dale Cave is the third major cave in the immediate area; it has more than a kilometre of passage trending north-west through low phreatic tunnels and small chambers. Like Lathkill Head and Critchlow, it acts as a flood resurgence during prolonged periods of wet weather. Two other small cave fragments in Cales Dale have yielded late Palaeolithic remains (Jackson and Storrs Fox, 1913).

Water Icicle Close Cavern is entered only via a 32 m mine shaft which intersects the junction of three drained phreatic tubes, each up to 3 m in diameter. They have only very minor vadose entrenchment. All are blocked after short distances by collapse or fluvioglacial deposits partly derived from the Millstone Grit, the nearest outcrops of which are now 7 km to the west. Stalagmite from these phreatic passages has been dated to earlier than 350 000 BP (Ford *et al.*, 1983).

The limestone north of Monyash contains natural caverns which are accessible through the artificial shafts of Knotlow and Hillocks mines (Figure 4.12). Constricted phreatic tube complexes and some large, fracture-guided caverns have been partly modified by mining, but still retain their main morphological features. Large banks of fluvioglacial sand are present in the Hillocks chambers. The caves are largely relicts and lie at a height intermediate between those of Water Icicle Close and Lathkill Head. There has been some invasion of the passages by vadose water, which flows into sumps before reappearing at Lathkill Head.

Interpretation

The Upper Lathkill Dale caves encompass successive levels of partly drained and totally flooded phreatic caves situated beneath the central part of the Derbyshire karst plateau. The origin and destination of the water which formed Water Icicle Close Cavern is unknown; the passages clearly represent a dissected remnant of a former high-level, phreatic system which pre-dates the present cave drainage at Lathkill Head, and is probably older than the incision of Lathkill Dale. Speleothems dated by the uranium-series method can only indicate abandonment more than 350 000 years ago, and the cave is probably very much older than this. These old, high-level cave passages, in Water Icicle Close and Knotlow, appear to have been invaded and modified by glacial meltwater streams; these flowed from more extensive, contemporary outcrops of the impermeable cover, or from spreads of glacial till since removed. Either bedrock or till could have been the source of the Millstone Grit sand and gravel preserved in Water Icicle Close Cavern.

The active sections of Lathkill Head Cave, Critchlow Cave and Lower Cales Dale Cave are fed mostly by, and may have been formed partly by, percolation water; there has been little allogenic input other than seepage from the few small lava outcrops in the area. They represent fine examples of epiphreatic systems, formed within the zone of water table fluctuation, which have become accessible only in recent times through lowering of the water table by sough drainage.

The ancient and active drainage patterns within the caves therefore relate to the progressive incision of Lathkill Dale well back into the Pleistocene, when the impervious cover may have extended further across the limestone plateau. A tentative chronology by Ford *et al.* (1983) recognizes sporadic new cave development from the early Pleistocene through to the Holocene, but awaits confirmation by further dating of cave sediments. Data from Lathkill Dale may provide an evolutionary model for dry valleys in the Peak

District karst where associated caves are not accessible. The sequences of clastic and speleothem deposits within the caves provide an important, if incomplete, record of Pleistocene events in the southern Pennines.

Conclusion

Lathkill Dale has an important series of isolated cave fragments preserved in the heart of the limestone plateau. These are significant as indications of the extent of karst drainage development remote from the present shale margin. The truncated relict caves, and the underlying active phreatic drainage, far from any impermeable catchments, make Lathkill Dale unique within Britain's karst.

GREEN LANE PITS

Highlights

The Green Lane Pits are four collapse dolines which were infilled with Pliocene, or very late Miocene, sands. These large-scale Tertiary karst features are unique to Derbyshire, and the Green Lane Pits are notable as they admirably show the geomorphology of the depressions. The deposits are of major importance in elucidating the Tertiary history of the area.

Introduction

Over 60 solution collapse dolines occur across the southern end of the Derbyshire limestone plateau. Many of these contained sediments of the Tertiary Brassington Formation, and have been worked for the manufacture of refractory bricks. The quarrying of the sand has revealed the limestone morphology of these dolines (Figure 4.14). The Green Lane Pits are notable in that the rock walls may now be seen with uncommon clarity, and little backfilling has taken place. The dolines all occur in dolomitized Carboniferous limestone, and are infilled with clays, sands and gravels from a fluvial depositional environment. These deposits have been preserved because solution of the underlying limestone caused collapses, into which the sediments sagged or slumped.

The dolines and their deposits are of major importance in elucidating the Tertiary history of upland Britain, and also provide evidence of the scale of Pliocene uplift in the southern Pennines. The 'Pocket Deposits' (Howe, 1897) were worked at least as early as the eighteenth century (Pilkington, 1789). They were once considered to be Triassic palaeokarstic features (Kent, 1957), but more recent studies have revealed the true nature of the deposits (Ford and King, 1969; Boulter, 1971; Boulter *et al.*, 1971; Walsh *et al.*, 1972, 1980; Wilson, 1979; Ford, 1984). The earlier work is reviewed by Ford (1977a).

Figure 4.14 Limestone walls and some remnants of the Pliocene sediment fill left after quarrying of the northwestern of the Green Lane Pits. (Photo: T.D. Ford.)

Description

The four dolines at Green Lane, in the centre of the Peak District karst (Figure 4.1), have had most of their sand infillings removed by commercial operations. All lie in dolomitized facies of the Carboniferous limestone, at an altitude of 335 m, adjacent to the floor of a dry valley which feeds into Long Dale. The southern doline is a circular pit 90 m across and 25 m deep, with almost vertical sides in limestone. The largest of the northern dolines is elliptical in plan, over 150 m long and 12 m deep; the two smaller dolines adjacent to it are also about 12 m deep. All four dolines have been almost totally excavated, to reveal rock walls, with rock saddles and pinnacles exposed between them (Figure 4.14). Some of the original Tertiary sands and the overlying Quaternary loess deposits remain preserved in the walls of the southern doline and along the margins of the northern dolines. The floors of the dolines are now obscured by vegetation, tyre dumps or slumped sediment.

Boulter *et al.* (1971) examined the Tertiary deposits preserved in the solution hollows of south Derbyshire and termed them the Brassington Formation. This was subdivided into three members:

Kenslow Member	plant-bearing clays	c. 6 m
Bees Nest Member	coloured clays	c. 7 m
Kirkham Member	sand and gravel	c. 30 m

The Kirkham Member consists largely of white, fluvial, cross-bedded sands with many quartzite pebbles, reworked from the conglomerates of the Triassic Sherwood Sandstone Group (formerly known as the Bunter Pebble Beds). The Bees Nest Member is dominated by red, yellow and white clays, and the Kenslow Member is mainly grey clays, with abundant fossil plant debris. The sediments are generally folded into small synclines, as a result of sagging into the collapsing dolines in the limestone. Commonly these fluvial deposits are underlain by remnants of the Namurian Edale shales and up to 5 m of angular chert gravels, derived from solution of the chert-rich limestones. Parts of the Brassington Formation are present in at least 60 of the limestone depressions, but there is no complete sequence in the Green Lane Pits. A small thickness of glacial till covers the Pliocene sands in some of the Brassington pits, and this shows evidence of sagging through continued subsurface solution.

Interpretation

The doline deposits were long regarded as features of a fossil karst surface with Triassic sands unconformably overlain by Tertiary clays (Kent, 1957), until Ford and King (1969) recognized that the Kirkham Member was a Tertiary deposit derived from Triassic conglomerates. The stratigraphy and paleobotany of the deposits were examined by

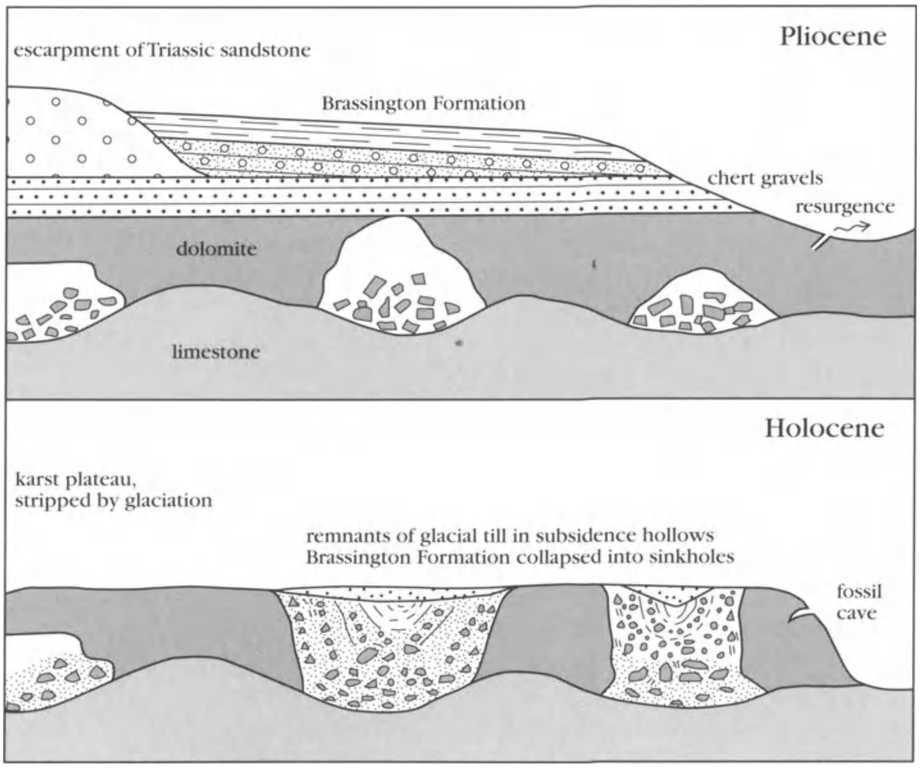


Figure 4.15 Diagrammatic sections of two stages in the formation of the Brassington Formation and their preservation in the collapse dolines in the limestone (after Ford, 1984).

Boulter (1971) and Boulter *et al.* (1971), who recorded 60 species of plant from the Kenslow Member, including *Sphagnum* and logs of *Sequoia*; they inferred an early Pliocene environment of a sandy heathland with scattered ponds.

The implications of the doline deposits for the paleogeographic history of upland Britain were recognized by Walsh *et al.* (1972), who regarded the subsidence outliers as small relics of a once continuous sheet of sands and clays. They calculated that subsidence of the Brassington Formation into collapse dolines, such as those at Green Lane, was in the order of 200 m. This indicated that the highest beds of the Brassington Formation were deposited at an altitude around 460 m. Thus the limestone block has been uplifted, during the Pliocene, by up to 250 m relative to the Triassic source areas at elevations around 240 m to the south; the uplift was probably much less than 250 m as the source could have been Triassic rocks once overlapped onto higher parts of the limestone upland. Paleocurrent structures in the Kirkham Member confirm the southerly provenance of the sands (Walsh *et al.*, 1980), while SEM analysis of the quartz grains suggested a short distance, low-energy fluvial regime with little chemical weathering (Wilson, 1979).

The synclinal bedding in the sediments preserved in the Green Lane Pits indicates that most of the limestone solution was underground, and was followed by progressive collapse and upward stoping of the voids (Figure 4.15). The Neogene sediments subsided into the dolines when the cavity roofs finally failed and dropped onto the accumulated piles of fallen debris. At some sites, they were later covered by glacial till which has been slightly disturbed by subsidence, indicating continued solution at depth. The collapse must postdate the initiation of a major karst drainage system, which produced the solution cavities. This was probably initiated following the incision of a major valley which provided the hydraulic head needed to start underground circulation. The dolines at Green Lane admirably show the nature of the solution during late Tertiary times.

Conclusions

The dolines exposed in the Green Lane Pits show the limestone morphology better than any other similar feature in Derbyshire. They provide an excellent example of this type of large-scale Tertiary solution and collapse feature. The sediments preserved in the dolines, and in 60 other similar pocket deposits, represent an important component in the Tertiary geomorphic evolution of Derbyshire. They provide evidence of Pliocene rivers draining a receding Triassic scarp in the south, and indicate that the limestone block has subsequently undergone perhaps as much as 250 m of relative uplift.

MASSON HILL CAVES

Highlights

The Masson Hill caves pre-date and post-date mineralization in a deep phreatic zone, subsequently drained by incision of the Derwent Valley. They are critical to an understanding of vein-guided karst drainage elsewhere in the Peak District. Sediment within the caves contains evidence for some of the earliest Pleistocene glacial episodes in Britain.

Introduction

The caves lie beneath the northern and eastern slopes of Masson Hill, immediately west of the anomalous limestone gorge at Matlock Bath where the River Derwent has entrenched updip of the limestone reefs (Figure 4.16). Most are sections of very ancient, partially choked, phreatic passages. These have been modified to some extent by lead and fluorspar mining, which has destroyed some natural features, but has allowed access to many more. The geology and mineralogy of the caves are uniquely complex within the Pennines. The hydrothermal mineralization is directly related to the cave development, as the mineralizing fluids both utilized and created solutional cavities within the Carboniferous limestone.

Descriptions of various parts of the cave system are given by Flindall *et al.* (1981), Gill and Beck (1991) and Warriner *et al.* (1981). The karst and cave development is discussed by T.D. Ford (1964a, 1984), Ford *et al.* (1977b) and Worley and Nash (1977), and the processes of mineralization are reviewed by Ineson and Ford (1982) and Quirk (1993).

Description

The Carboniferous limestone exposed at Masson Hill forms a gentle flexure over an anticline which

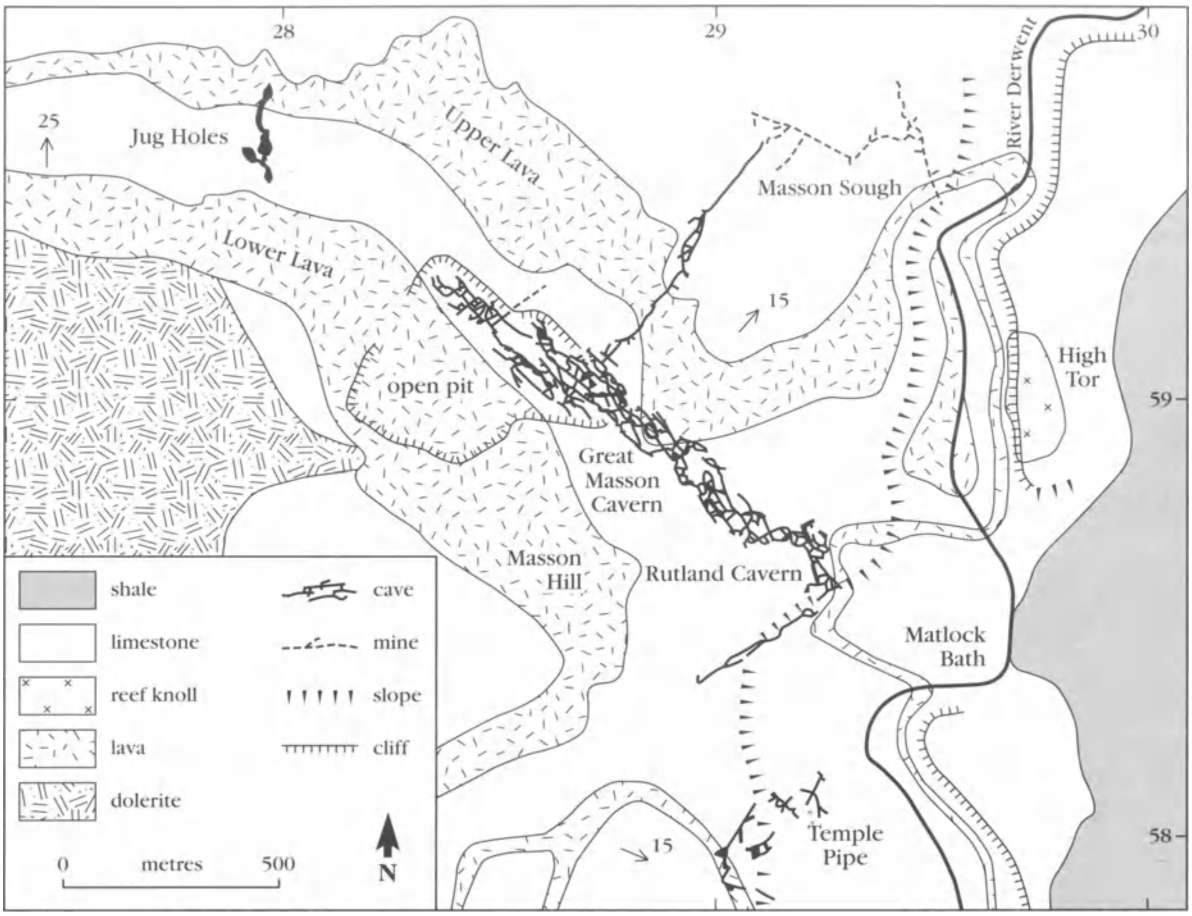


Figure 4.16 Geological map of Masson Hill and its cave passages, in relation to the Matlock Bath gorge. The mine workings in solid rock and the re-excavated natural caves are complexly interwoven; the symbols for cave and mine are generalized. The caves within the open pit have all been destroyed.

plunges steeply with the regional dip to the east. The carbonate sequence is broken by two basalt lavas - the Matlock Lower Lava and the Matlock Upper Lava - and by several thin wayboards of volcanogenic clay. Between the lavas, about 40 m of limestones contain the major caves.

Most of the known cave passages interconnect to form the Masson Cavern system, where pre- and post-mineralization phreatic solution caverns are linked by old mine workings into a single underground complex with more than 2000 m of natural passage (Figure 4.16). The north-western part of the system has been destroyed by surface mining on the hilltop, leaving two elongate networks, just separated by the open pit. One descends south-east, obliquely downdip, through Great Masson Cavern (Figure 4.17), and on to the disconnected fragment of Rutland Cavern; the other descends almost straight downdip, north-east to the old Masson and Ringing Rake Soughs at river level. Individual cave chambers are up to 20 m high or wide, and the whole cave system has a vertical range approaching 200 m.

Pre-mineralization caves are filled or lined with hydrothermal fluorspar and other minerals. Further solutional enlargement of these old caves by meteoric groundwater occurred in the late Tertiary and early Pleistocene, at the same time as new caves were formed. Much of the phreatic network has been filled with complex sediment sequences, comprising both locally derived vein minerals and also inwashed glaciofluvial material. Interbedded stalagmite layers are sparse, and evidence for a major vadose episode of cave excavation is lacking. Magnetostratigraphic analysis of sediment sequences in three separate parts of the system has established that fluvioglacial material was deposited during an interval of reversed magnetic polarity, indicating an age in excess of 780 000 years (Noel, 1987; Noel *et al.*, 1984).

Temple Pipe (Figure 4.16) shows similar pre-

Figure 4.17 Ribs of limestone left around solution cavities which were filled and then re-excavated by miners in the Black Ox Mine workings in Great Masson Cavern. (Photo: T.D. Ford.)

and post-mineralization cavern development to that seen in Masson Cavern. Some excellent sediment sequences are preserved in its two fossil phreatic chambers; it is now a show cave. A number of other cave passages are intersected by the other mines in the area, of which the longest are in Devonshire Cavern.

Jug Holes (Figure 4.16) consists of an isolated series of phreatic cave chambers. These were formed by limestone solution both before and after the hydrothermal mineral infilling. Factors controlling cave development are clearly recognizable, and the site is typical of a Derbyshire pipe vein system. The cave was developed downdip in about 40 m of Asbian limestone sandwiched between the Upper and Lower Matlock Lavas; subsequently the cave has been partly filled with sediments, which were derived from the mineralized cave walls, from the lava flows, and from inwashed glaciofluvial material. There are extensive stalagmite deposits in parts of the cave; some of these rest directly on the altered top of the Matlock Lower Lava which forms the cave floor.

Interpretation

The three cave systems reveal a complex history, with the development of pre- and post-mineralization phreatic solution caverns separated by an episode of hydrothermal mineralization and a considerable timespan (T.D. Ford, 1964a, 1984; Ford *et al.*, 1977b; Worley and Nash, 1977). The earliest episode of cavern development was associated with the initial phase of hydrothermal mineralization in the late Carboniferous. Solution voids within the limestone were excavated by meteoric water which had travelled distances up to 100 km through Namurian clastic rocks at depth, where it was enriched in minerals before rising into the limestones on the Peak District block (Quirk, 1993). There is no definite evidence for limestone solution by locally derived meteoric water influence at this time. Many of these cavities were then filled partly or wholly with fluorspar and other hydrothermal minerals. Voids remaining within these veins, pipes and flats were then utilized by meteoric karst water as the limestone was exposed by erosion in late Tertiary and Pleistocene times; this new phase of solutional activity both enlarged some of the old caves and also developed new ones.

Cavern development was strongly influenced by the geology. The lavas acted as confining aquicludes, and the patterns of solutional opening were determined by the irregular dolomite/limestone interface, early diagenetic solution of the lower part of the limestone, the presence of several thin wayboard tuffs, and a NNE–SSW joint system. The presence of sulphide minerals may have contributed to cave development, by acting as a source for the generation of sulphuric acid, in the style now widely recognized as inception in karst limestones (Ball and Jones, 1990; Lowe, 1992).

Most of the caves were filled with both autochthonous and allochthonous sediment, prior to the downcutting of the River Derwent which caused the change from phreatic to vadose conditions. Chatter marks on the sand grains indicate a glacial meltwater origin for some of the allochthonous material, and the age of 780 ka indicates that they date from one of the pre-Cromerian cold stages (Noel *et al.*, 1984). There is little evidence for significant vadose modification following draining of the phreas. This suggests that the vadose stream phase was brief and overloaded with fluvioglacial sediment, quickly choking the system, perhaps indicating that it was directly associated with an episode of glacial incision. Since then the surface catchment has been insufficient to allow significant drainage into the cave system. The scarcity of speleothems probably reflects the position of the system beneath the Upper Lava aquiclude.

Conclusion

The Masson Hill site is the best example in Britain for demonstrating the relationships between mineralization of the limestone and cave development. It is important for understanding aspects of the hydrology of other cave systems associated with mineral veins, where the phreas remains largely inaccessible. Sediments in Masson Cavern include fluvioglacial material deposited during an episode of reversed magnetic polarity, more than 780 000 years ago. They are of comparable age to the material in the fissure caves of the Eldon Hill quarry, and are considerably older than most Pleistocene glacial deposits proven on the surface in Britain; these caves therefore provide incomparable evidence for the Pleistocene history of the area extending back beyond the Anglian glaciation.

DOVE DALE

Highlights

Dove Dale is perhaps the finest example in Britain of an allogenic river cutting through a limestone massif. It forms an extensive and spectacular gorge with many notable karst features, and admirably demonstrates the nature of fluvial erosion within a limestone terrain.

Introduction

The River Dove crosses the Carboniferous limestone outcrop in a deeply entrenched gorge. It maintains its flow over the entire limestone outcrop, a distance of 10 km, and provides a clear example of river superimposition onto a limestone outcrop. The river, fed by a large shale catchment, is too large to sink underground, and erosion has continued until the present. The gorge morphology shows excellent adjustment to the different lithologies and structures within the limestone, and the walls contain various karst features, including Dove Holes and Reynard's Cave.

Warwick (1953, 1964) first studied the geomorphology of the Dove and Manifold valleys. Ford and Burek (1976) discussed the importance of the geological structure and the position of reef-knolls in determining the river course. An overview is documented in Ford (1977a). Rowe *et al.* (1988b) and Atkinson and Rowe (1992) discussed the geomorphic evolution of the area and commented on the age of the relief using uranium-series dating of caves in the neighbouring Manifold valley. The caves in the valley sides are described in Gill and Beck (1991), and the area is described in a field guide by Ford and Gunn (1992).

Description

The River Dove maintains its course across the south-eastern corner of the limestone outcrop (Figure 4.1), cutting a meandering valley north to south up to 150 m below the limestone plateau level. The headwaters of the river lie on the Millstone Grit, and it flows onto the limestone at the head of Wolfscote Dale. For the next 10 km the river is deeply entrenched into the undulating limestone plateau, descending some 70 m to emerge from the mouth of Dove Dale (Figure 4.18). A series of tributary valleys, notably Biggin Dale and Hall Dale, feed from the limestone plateau into the main valley; all are now dry, and many hang above the main valley floor. The valley rim is characterized by sharp breaks of slope at levels around 300 m. The lower section of the valley sides are thickly wooded, but are broken by

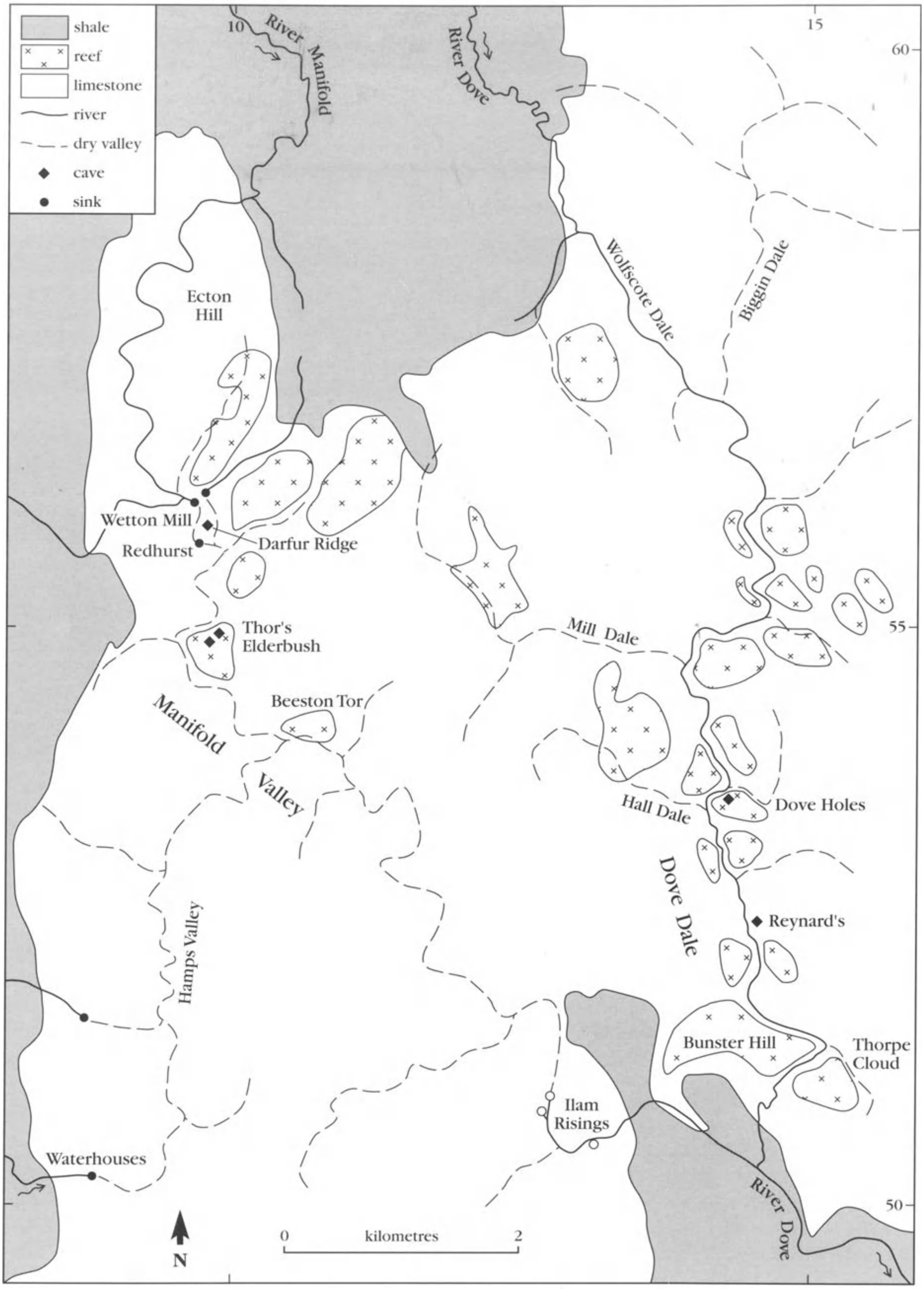


Figure 4.18 Geological map of the active and dry valley systems of Dove Dale and the Manifold River in relation to the reef knolls in the Carboniferous limestone (partly after Ford and Burek, 1976).

Figure 4.19 The limestone arch of Reynard's Cave, in the side of Dove Dale, looking through to the cave remnant beyond the breached section. (Photo: A.C. Waltham.)

lines of cliffs, some forming impressive vertical crags. Ilam Rock is an isolated limestone pillar breached by a cave which contains extensive tufa.

Several caves open into the valley sides, but most are small fissures penetrable for only a few metres; all show phreatic features. Dove Holes have two large unconnected cave entrances on the east bank, but only extend back a few metres. This site and several of the smaller caves have yielded Devensian and Holocene mammal remains as well as human artefacts (Spencer and Melville, 1974). A number of small risings lie along the banks of the Dove, most emerging just above river level on the east bank; these feed percolation water from the adjacent limestone plateau. The modern underground drainage appears to be immature and poorly integrated.

Interpretation

The gorge section of Dove Dale owes its origins to the superimposition of the River Dove onto the limestone outcrop by progressive erosion of the Namurian shale cover, largely during the Pleistocene. Discharge of the river is sufficient to prevent it all sinking underground, and it maintains surface flow across the entire limestone outcrop; this is now aided by artificial ponding to improve the fishing in its downstream reaches. Knickpoints in the dry tributary valleys indicate multiple rejuvenations, caused by erosion and base-level lowering in the valleys to the south on Triassic mudstone (Burek, 1977). The dry valleys were incised during periglacial periods, when underground drainage was prevented, only to be abandoned during each warm phase. Continued incision in the main valley, due to maintenance of the Dove flow, has left the tributary valleys hanging.

The sinuous form of Dove Dale is not a pattern of meanders inherited from when the drainage course superimposed onto the limestone; it was determined by the outcrops of the reef knolls within the limestone (Ford and Burek, 1976). As it cut into the carbonates, the river was forced to take the lowest available course between the biohermal masses of strong reef limestone. The river course tends to wind in between the reef knolls, and is deflected around several of them (Figure 4.18). The river has trimmed the edge of some reef masses, rather than cut through them. Where the river cuts between adjacent reef knolls, it forms steeper sides in its gorge, notably where it passes between Bunster Hill and Thorpe Cloud at the lower end of the Dale. East of Thorpe Cloud, an abandoned meander is perched 60 m above river level.

None of the caves in the valley sides is of any significant length; most are small old phreatic systems within the reef knolls. Dove Holes have large entrances where the river cliff has breached phreatic rifts. Reynard's Cave is another old phreatic tunnel, truncated in front of a joint-guided breach to leave a rock arch (Figure 4.19). Cave intersection and collapse has contributed little to the formation of the gorge. Nearly all the high cliffs, towers and crags are controlled by the limestone lithology; only their preservation is a function of karstic processes.

No absolute chronology is yet available from cave sediment dating. By analogy with the Manifold Valley close to the west, it appears that Dove Dale began to incise into the plateau surface about 3.5 Ma ago (Rowe *et al.*, 1988b; Atkinson and Rowe, 1992). Breaks of slope at levels of 265-300 m mark the start of the latest phase of incision into an older landscape of low relief (the '1000-Foot surface' of Clayton, 1979).

Conclusions

Dove Dale is the most spectacular and longest allogenic limestone gorge in Britain, and its allogenic flow contrasts with the dry tributary valleys where rainfall is directly absorbed into the limestone. The winding nature of the river course is guided by the position of reef masses within the limestone, and indicates the importance of lithological control. Cliffs, truncated caves and natural bridges demonstrate the nature of fluvial incision in a karstic terrain. The gorge was probably incised into the plateau surface as a result of renewed uplift about 3.5 Ma ago.

MANIFOLD VALLEY

Highlights

The Manifold and Hamps valleys are allogenic river gorges cut into the Carboniferous limestone. Both rivers are being progressively captured by underground drainage, and the length of active riverbed varies with the amount of run-off. The multiple sinks and risings demonstrate a complex underground hydrology. Abandoned caves in the valley sides have enabled the valley age and rate of incision to be estimated.

Introduction

The Manifold River has cut a deeply entrenched valley for 9 km, from Wetton Mill downstream to Ilam Hall. The tributary Hamps valley is similarly entrenched, downstream of the limestone-shale boundary at Waterhouses. The valleys lie across the south-western corner of the Carboniferous limestone outcrop, where it contains many reef knolls. The headwaters of both the Hamps and Manifold rivers lie on the Namurian shales and sandstones, and both flow onto the limestone where they progressively sink underground. At high stage, water may flow on the surface all the way to the main Ilam rising, though the Hamps rarely flows over its full course. In dry weather the water sinks further and further upstream. The Manifold Valley has truncated a number of caves, some of which contain important archaeological remains.

The area was studied by Warwick (1953, 1964), and Ford and Burek (1976) examined the role of the reef knolls in determining the form of the valley. The caves are discussed by Potts (1977), and are described by Gill and Beck (1991). Excavations at Elderbush Cave, among others, are described by Bramwell (1964, 1977). Sediments from the same cave and also Darfur Ridge Cave have been dated using uranium-series and paleomagnetic techniques (Rowe *et al.*, 1989b; Atkinson and Rowe, 1992) which give estimates for the rate of incision and the age of the valley. Ford and Gunn (1992) provide a field guide to the Manifold Valley.

Description

The headwaters of the Manifold drain an area of Namurian shales below the Millstone Grit escarpments. The river flows onto the limestone, and in dry conditions sinks at Wetton Mill, shortly after it encounters the first reef knoll. Downstream, the river bed may be dry for 9 km, as far as the Ilam risings (Figure 4.18). Under slightly wetter conditions the water continues to flow over the surface to sink by the Darfur bridge or at Redhurst Swallet, about 400 m below Wetton Mill. Under progressively higher stages, the flow continues downstream to a further series of sinks. Only under very wet conditions does the river flow above ground all the way to the main risings at Ilam. Similarly, the Hamps sinks into its bed shortly after contact with the limestone at Waterhouses, but in wetter weather sinks progressively further downstream; it rarely flows on the surface to meet the Manifold. Within the shales, the valleys have gentle slopes and the tributaries are graded to the main valley floor. Where the valleys are cut through the limestone, the sides are much steeper, often forming impressive vertical crags such as those at Beeston Tor. Many of the short tributary valleys are permanently dry and hang above the main valley, forming knickpoints indicative of successive rejuvenations. The main flanks of both valleys are evenly graded and covered in vegetation, but cliffs and crags are common in the reef limestones.

Downcutting of the Manifold Valley has truncated a number of old, high-level caves, most of which were small phreatic systems. Elderbush Cave is a truncated phreatic tube located close to the valley rim at 275 m OD, while the lower Darfur Ridge Cave is a small phreatic passage extending 100 m. Both of these caves contain stalagmites which have proved suitable for dating (Rowe *et al.*, 1988; Atkinson and Rowe, 1992).

Thor's Cave has a massive entrance, but its phreatic rifts reach back less than 50 m. There are more remnant rift caves in Beeston Tor. A new cave system is presently developing under the valley floor, gradually capturing the surface flow. Fragments of this system can be entered at a number of locations. Darfur Pot, just downstream of the main sink at Wetton Mill has 360 m of very flood-prone passage, while further downstream, Redhurst Swallet extends for some 280 m in a series of tight joint and tube passages (Potts, 1977). Ladyside Pot is another fragment where 450 m of passage extends under the river bed. All the water resurges at a series of large springs at Ilam Hall; diving in the main rising has revealed only 250 m of submerged passage reaching a depth of 54 m.

Interpretation

Remnants of former valley floors can be identified from knickpoints in the dry tributary valleys, Warwick (1953, 1964) used these to deduce that the valley had been deepened in six successive stages. This was almost certainly in response to base-level lowering in the valleys on the Triassic mudstone to the south, aided by more vigorous phases of downcutting during the Pleistocene cold phases. The influence of the limestone lithology is important, as the river course has been dictated by the position of the reef knolls (Ford and Burek, 1976). The river has been diverted round each reef, as at Beeston Tor and Thor's Cave Crag; other bends on the river, around Ecton Hill, have been influenced by the strong folding of the limestones. The sinuous course of the river was not superimposed from a former shale cover, as suggested by Warwick (1953); it developed as a result of lithological contrasts as incision progressed, concomitant with falling base levels to the south.

Nearly all the high-level caves are developed within the reef knolls, and they provide evidence of karst development which predates incision of the gorge. They represent earlier generations of caves developed at or below the contemporaneous valley floor, and are comparable to the phreatic system that is developing today beneath the river bed. To investigate the chronological development of the Manifold Valley, Rowe *et al.* (1988b) dated a suite of speleothems from Darfur Ridge and Elderbush caves using magnetostratigraphic and uranium-series methods. Uranium-series dating proved the stalagmite from Elderbush Cave to be older than 350 ka, while the presence of reversed-polarity stalagmite overlying normally magnetized stalagmite indicated a minimum age of 1.87 Ma. Elderbush Cave appears to have been drained by downcutting in the Manifold Valley, by or soon after 2.0 Ma. The cave lies 110 m above the present river bed, giving a maximum rate of valley incision of 5.5 cm ka^{-1}. Similarly, Darfur Ridge Cave was shown to be about 300 ka old, and gave an incision rate of 4.1 cm ka^{-1}. Extrapolation of these rates to the plateau surface at 265–300 m OD leads to the tentative estimate that valley incision began in the Pliocene about 3.5 Ma ago (Atkinson and Rowe, 1992).

The modern hydrology of the Hamps and Manifold valleys is complicated. The multiplicity of sinks and risings makes the underground flow patterns very complex. The resurgences at Ilam consist of 12 separate springs, which appear to have different catchment areas. The lowest two springs discharge only autogenic percolation water, while the next three are fed by the Manifold sinks. The Hamps water emerges from the Upper rising (taking about 3–6 days to flow from Waterhouses). Dye tracing has shown that the flow is transmitted via a complex conduit system, but with no mixing of the Hamps and Manifold waters (Ford and Gunn, 1990). The Hamps drainage system falls about 70 m over 4 km, which suggests there may be a significant vadose component, and is close to total underground capture. The underground Manifold drainage has a lower gradient, is more likely to be phreatic, and is less mature in that it is less able to transmit flood flows.

Conclusions

The two limestone valleys of the Manifold and Hamps provide fine examples of allogenic river gorges undergoing progressive capture by a developing underground drainage network. The river beds can be active or dry depending on stage, and the karst hydrology of the area is complex with a multiplicity of sinks and risings. Knolls of strong reef limestone defined the sinuous course of the Manifold Valley as it was entrenched into the karst plateau. The valley has truncated earlier cave development, recording a history of incision spanning about 3.5 Ma.

Chapter 5

The Mendip Hills karst

INTRODUCTION

The Mendip Hills rise east of the Bristol Channel as an elongate plateau roughly 30 km long and 8 km wide, composed largely, but not entirely, of cavernous limestone (Figure 5.1). At their western end they stand 200 m above the Somerset Levels, but their eastern end declines gently and is buried beneath the Jurassic scarplands.

The carbonate succession includes almost the entire Dinantian sequence of the Lower Carboniferous, and reaches over 800 m in thickness. The main karstic rocks are the strong, fine-grained shelf limestones; they include beds of very fine-grained calcite mudstone, known as chinastone, and also beds which are conspicuously bioclastic or oolitic. Some of the carbonates are dolomitized and there are many thin clastic horizons within the main sequence. The lowest unit in the succession is the Black Rock Limestone, a dark well bedded series in which most of the swallet caves are formed. Above this, the Burrington Oolite, Clifton Down Limestone and Hotwells Limestone are all pale grey but weather to the white patina seen at outcrop, notably in the white cliffs of Cheddar Gorge.

The limestone is underlain by Dinantian calcareous shales with thin interbedded limestones; these are known as the Lower Limestone Shales and are transitional from the Devonian Old Red Sandstone. The sandstones crop out in four anticlinal cores to form hills rising 30-60 m above the limestone plateau (Figure 5.1). On the flanks of the Mendip plateau, the limestones are overlain unconformably by Triassic screes and fan deposits, formed largely of limestone blocks and known locally as the Dolomitic Conglomerate: these are also cavernous at some locations, notably at Wookey Hole, where they constitute an integral component of the single karst aquifer. On the lower, eastern part of the plateau, outliers of Mercia Mudstone, silicified limestones of the Harptree Beds and Liassic limestone lie unconformably on the Carboniferous limestone. Triassic palaeokarst was restricted by the contemporary desert environment, and most of the cave infills

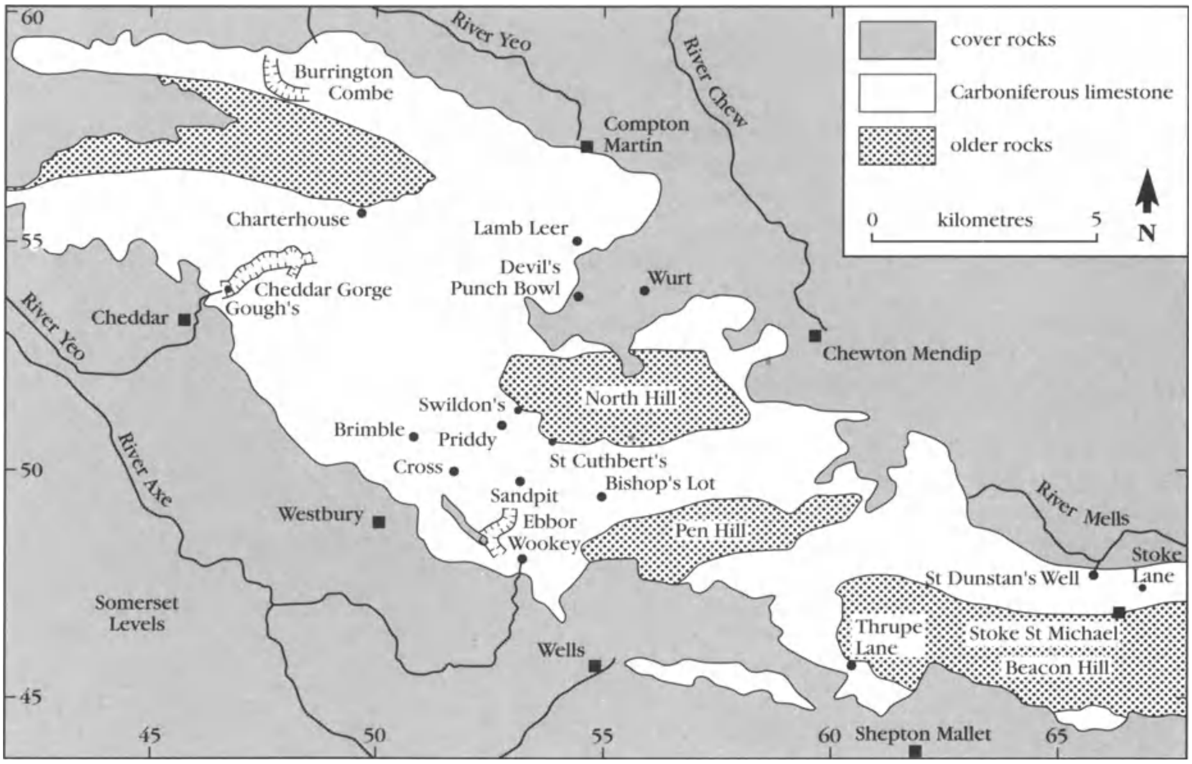


Figure 5.1 Outline map of the Mendip Hills karst, with locations referred to in the text. Cover rocks are mostly the Triassic and Jurassic mudstones and limestones; Upper Carboniferous rocks form the thrusted outlier on the east side of Ebbor Gorge. The Triassic Dolomitic Conglomerate is included with the Carboniferous limestone where it is composed of blocks of the limestone and is an integral part of the karst. Older rocks are the Devonian Old Red Sandstone and the Dinantian Lower Limestone Shale.

appear to be in tectonic fissures and associated with Jurassic neptunian dykes (Ford, 1984; Stanton, 1991).

Structurally the Mendip Hills are the most complex of Britain's four major regions of cavernous karst. They are essentially formed of four *en-echelon*, periclinal anticlines. Marginal dips are mostly between 20° and 70°; they are steepest and at some locations overturned on the northern limbs. Extensive faulting, overthrusting and overfolding complicates the geology (Smith, 1975a). In some areas, the limestone contains hydrothermal mineralization, but this has not influenced the karst processes in the way that it has in the Peak District.

The karst

The Mendip Hills form an upland exhumed from beneath a Mesozoic cover, and traces of Triassic features survive, mainly in the marginal slopes. Though the limestone hills may have been stratimorphic early in the Trias, the summit surface is now eroded and planed to reveal the anticlinal cores of sandstone; the main plateau surface is regarded as a result of subaerial planation largely during the Pliocene (Ford and Stanton, 1968). This was followed by exhumation of the plateau sides from beneath a Triassic cover; this was a gradual process stripping the cover progressively towards the east, during the course of about a million years. This has left the highest parts of the plateau and the most mature karst at the western end of the Mendips, around Cheddar Gorge, while the younger, eastern end of the limestone plateau is only partially exhumed and barely rises above the level of the Triassic plain in the Mells Valley (Figure 5.1). There is no evidence that the Mendips were glaciated at any time during the Pleistocene, though ice did occupy parts of the surrounding lowland (Hawkins and Kellaway, 1971; Smith, 1975a).

The dominant surface landforms on the Mendip limestones are fluviokarstic. Some of these may be superimposed from an earlier impermeable cover, and were modified by subsequent subaerial evolution, before karstification reached a stage at which drainage was entirely underground. Surface erosion was then temporarily re-established under periglacial conditions during the Pleistocene, when karstic drainage was hindered by the ground ice (Smith, 1977). Dry valleys are entrenched into much of the upland limestone surface, and these steepen into gorges in their descents of the plateau margins (Ford and Stanton, 1968); the best known is Cheddar Gorge, which has the largest feeder system of dry valleys.

On the interfluves and the plateau margins, the topography is more disorganized and less fluvial; large, shallow, closed depressions create some limited areas of polygonal karst. Solutional dolines of various morphologies and origins are scattered across the karst plateau, but many have been modified by past mining activities. Subsidence dolines are numerous in some outcrops of the Harptree Beds, but the absence of till precludes the development of extensive fields of subsidence dolines such as are found in Britain's glaciated karstlands. There are no limestone pavements. Rock scars are not a feature of the karst, except along the steeper flanks of the deeper sections of the gorges. Thick soils on the plateau have a significant loessic component, but are replaced by thin stoney soils on steeper slopes; except for the steeper valley and marginal slopes, the whole area is now farmland.

The caves

The erosional resistance of the sandstone in the anticlinal cores has created low hills which supply allogenic drainage onto the limestone in the heart of the Mendip karst. Nearly all these streams sink where they reach the limestone. Further allogenic stream sinks are provided by drainage from the Mesozoic outliers. The sinking streams are joined underground by percolating autogenic recharge, and flow through numerous stream caves which feed to resurgences around the foot of the plateau marginal slopes. This situation is best seen at the western end of the Mendip plateau, which is the oldest section with the highest local relief; stripping of the Mesozoic cover has progressed towards the east, where relief is lower and the karst is less mature.

The typical Mendip cave has a stream sinking at the stratigraphic base of the Black Rock Limestone (Smith, 1975a). A vadose passage descends rapidly with the steep dip, until the local base level (or notional water table) is reached on a profile gently, and roughly, graded to the contemporary resurgence level. From there, a phreatic cave continues with a looping profile following down the bedding planes and up the joints, with sections of shallow, sub-horizontal loops along the strike. Continued erosion within the looping phreatic passages cuts trenches through the loop crests and raises the loop troughs by sedimentation and paragenetic roof solution; together these processes lead to a more uniformly graded passage profile. The Mendip Hills have long been cited as the type area of cave development in steeply dipping limestones (Ford 1968; Ford and Ewers, 1978).

Long-term base-level lowering and episodic rejuvenation fossilize the phreatic systems, commonly before they have achieved graded profiles above their contemporary risings. Major rejuvenations are generally followed by utilization of completely new phreatic routes at lower level, but inception and initiation of the lower routes had already been started deep within the earlier phreas. Though the geological controls on cave development are strong, the inclined networks of intersecting fractures in the limestone permit rapid responses to rejuvenation, and there is a minimum of widespread water table perching within the modern aquifer.

Due to passage constrictions within the phreas, the depth of the active phreatic loops, and sediment accumulation in both active and abandoned phreatic loops, no cave system in the Mendip Hills has yet been found to be accessible over its entire path from sink to rising. Most Mendip caves are therefore either swallet systems with influent vadose streamways descending to active and abandoned phreatic levels, or resurgence caves with active and abandoned deep phreatic passages.

The cave systems at Priddy and Charterhouse are all of the swallet type. Between them they show considerable morphological variety in response to contrasting details of geological structure. The two main groups of sinkhole caves all drain to the resurgence caves at Cheddar and Wookey; these contain old phreatic passages notable for some of their secondary calcite deposits, and active caves with spectacularly deep phreatic loops.

The Mendip Hills are also noted for karst and cave development in lithologies other than the Carboniferous limestone. Of the caves formed by solution in the Triassic Dolomitic Conglomerate, the outer parts of Wookey Hole are the largest and most extensive. The Mesozoic limestones have their own miniature cave systems, in addition to the piping failures and collapse stoping induced by drainage into the underlying Carboniferous.

The absence of glacial cover enjoyed by the Mendip karst throughout the Pleistocene permitted a very complete record of sediments to accumulate in the caves. These sequences of both clastic detritus and calcite flowstone are exceptionally valuable to Pleistocene stratigraphy (Atkinson *et al.*, 1978, 1986); correlations with surface evolution are facilitated by their survival in the successions of rejuvenated phreatic cave passages which evolved in close accord with falling resurgence levels.

BURRINGTON COMBE

Highlights

Burrington Combe is a fine example of a fluvial karst gorge, which was cut largely under periglacial conditions across the narrow steeply dipping limestone outcrop on the northern side of the Mendips. Relict and active caves exposed in the limestone flanks provide evidence of a long history of solutional erosion predating the formation of the valley. The lower part of the gorge has exposed part of an infilled Triassic gorge or wadi. An almost complete succession through the Carboniferous limestone sequence of the Mendips is exposed in the Combe.

Introduction

The Combe is a dry karst gorge immediately south of Burrington village, which cuts thorough the northern flank of the Mendip Hills (Figure 5.2). The gorge is in many ways very similar to Cheddar Gorge; however, its walls contain smaller cliff sections, it intersects two fossil Triassic valleys, and it has a large alluvial fan at its mouth. Burrington Combe is entrenched through limestones which dip at about 60° north, on the northern side of the Black Down pericline. Being a less spectacular feature than the nearby Cheddar Gorge, little has been written specifically on the Combe, although the arguments about the formation of Cheddar Gorge equally apply.

A description of the general geomorphology and hydrology of the area was published by Tratman (1963), while the gravels associated with the alluvial fan at the foot of the gorge were the subject of work by Clayden and Findlay (1960). The geological succession exposed in the Combe is described in detail by Green and Welch (1965). Many caves occur in the side of the Combe, documented by Barrington and Stanton (1977) and by Irwin and Jarratt (1992). Although most are very

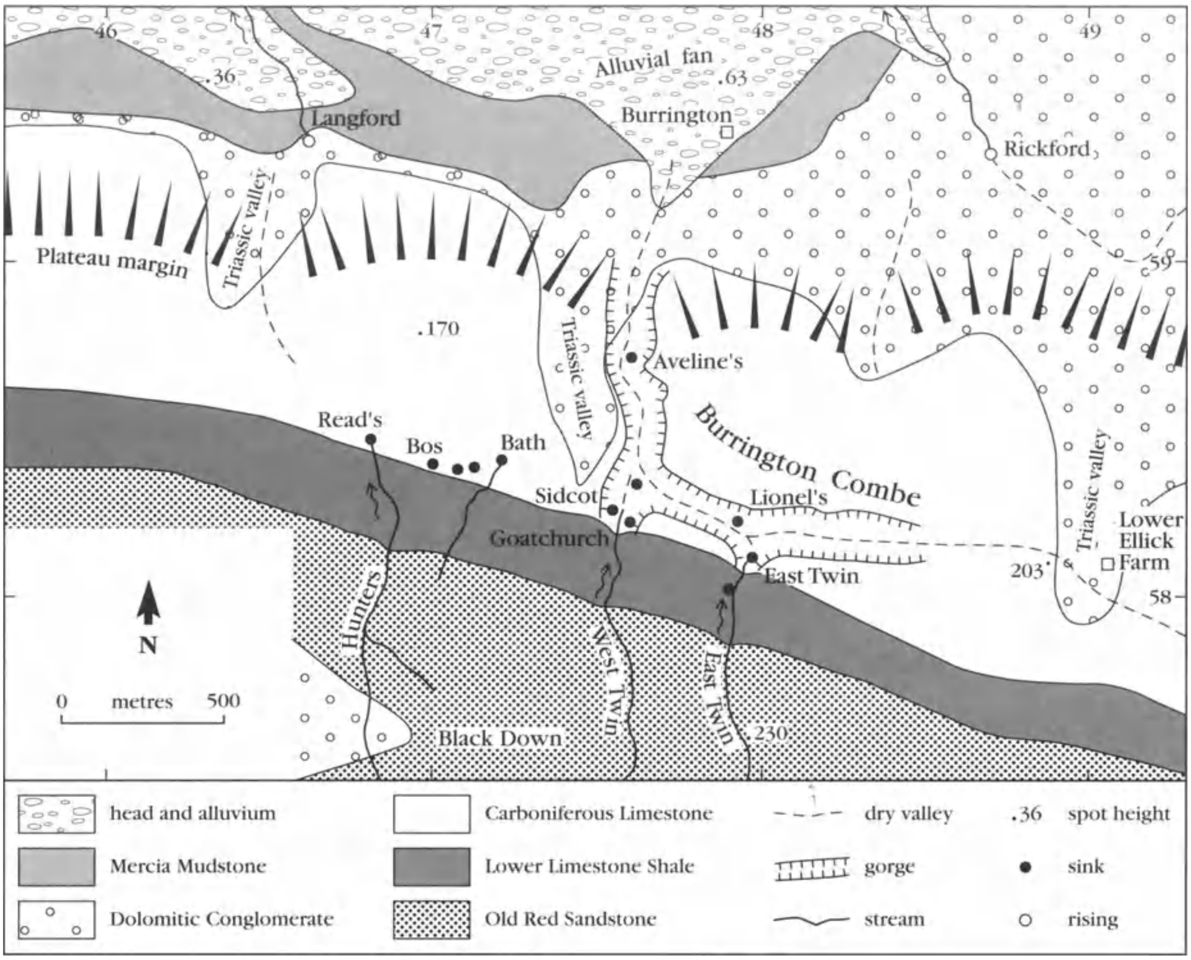


Figure 5.2 Geological map of Burrington Combe and the infilled Triassic valleys cut into the northern slope of the Mendip Hills (after Williams and Farrant, 1992).

small, they provide evidence of former stages in the evolution of the gorge.

Description

The upper part of the gorge is developed along the strike of the Black Rock Limestone, before turning sharply north and cutting through the rest of the Carboniferous limestone sequence. Burrington Combe partially intersects two infilled fossil Triassic valleys, which can be recognized by the outcrops of Triassic Dolomitic Conglomerate which infill them. One is located near the head of the combe at Lower Ellick Farm, while the second is exposed where the combe beaches its eastern flank near its lower end (Figure 5.2).

The combe is a steeply graded feature, descending from 200 m at Lower Ellick Farm to debouch at around 80 m onto the lowland excavated in Mercia Mudstone to the north. A well developed alluvial fan (marked as head on most geological maps) is developed at the foot of the gorge and spreads out into the Vale of Wrington to the north. Two tributary valleys, the East and West Twin valleys, descend steeply off Blackdown's northern slopes to join the main combe about halfway down. The upper parts of these tributary valleys are developed on the Old Red Sandstone and have surface streams which disappear underground at the contact with the Carboniferous limestone, with some small sinks in the East Twin into the Lower Limestone Shales. Solutional voids in the Shale group are also revealed in a drainage adit driven through them from the floor of the West Twin valley. The combe is dry, except in extreme floods, including that of 1968 which cut a trench through the clastic fill in the East Twin valley (Hanwell and Newson, 1970); normal drainage is now underground, and resurges at the Rickford and Langford Risings (Newson, 1972; Crabtree, 1979), east and west of the mouth of the combe (Figure 5.2).

Many caves have been exposed by the downcutting of the combe, of which the most extensive are the inclined mazes of Goatchurch Cavern and Lionel's Hole (Figure 5.2). These are mainly abandoned phreatic systems, and thus predate valley incision; they represent fragments of earlier generations of swallet caves and the complex phreatic networks which they fed (Bull and Carpenter, 1978). A newer generation of small sinks has developed in the valley floor. Aveline's Hole, located near the foot of the combe is the remains of a major phreatic tube, which was once part of a resurgence cave, but now acts as a sink for road drainage. Another set of abandoned and active stream sinks is located on the hillslope west of the combe. From east to west these are Bath Swallet, Rod's Pot, Drunkards Hole, Bos Swallet and Read's Cavern, forming a series of steeply descending caves developed downdip; they are almost certainly linked, although no connections have been found yet (Williams and Farrant, 1992). Their descending, dip-orientated passages contrast with the level strike-oriented rifts in Lionel's Hole, Goatchurch Cavern and the other caves at lower levels in the combe.

Interpretation

Burrington combe clearly shows the features associated with fluvial erosion during periglacial periods, and as such is very similar to Cheddar Gorge. However, the smaller catchment and the more limited relief has created a less dramatic gorge than that at Cheddar. This has meant that cavern collapse has not so commonly been invoked to explain its development. Both Reynolds (1927) and Tratman (1963) recognized the fluvial origin of the combe. Tratman went further and suggested the valley was eroded during periods of periglacial spring snowmelt, when the ground was frozen and underground drainage impeded. The large alluvial fan below the mouth of the gorge consists largely of gravel, produced by intense frost action and transported by periodic torrents of water flowing down the combe (Clayden and Findlay, 1960; Findlay, 1965; Stanton, 1977). The relationship between the Combe and the alluvial fan is clearer here than at any other site on Mendip. Deposition of the fan may also have been responsible for diverting karstic drainage to the two modern Risings on either side of it (Figure 5.2), in a style reflecting the diversion of surface streams off the crest of an aggrading fan. Although much of the combe was excavated under periglacial conditions, its early stages may have pre-dated significant karstification, and could have been cut by normal surface drainage.

The caves exposed in the side of the gorge provide evidence of underground solutional erosion which pre-dates the surface excavation of much of the gorge. The modern active cave system appears to be fairly immature. This is shown by the bifurcation of the drainage to two separate springs and the small nature of the active streamways. As yet no dating studies have been completed, but an absolute chronology for the caves would provide a time-scale for the geomorphological evolution of the combe.

It has been suggested that many of the modern valleys on Mendip follow earlier infilled Triassic valleys (Ford and Stanton, 1968). This is not the case in Burrington. The combe truncates the head of one such valley at Lower Ellick farm (Figure 5.2), but does not follow it. Similarly, the combe intersects another Triassic valley near the foot of the modern gorge, but, instead of following it, runs parallel to its eastern flank, cutting partly into the Carboniferous limestone. The Pleistocene erosion of the combe is independent of earlier Triassic development, and the Dolomitic Conglomerate does not appear to have offered a line of least resistance.

Conclusions

Burrington Combe is an excellent example of a fluvially eroded valley cutting through steeply dipping limestone, and has a long history of development through the Pleistocene. It partially intersects two earlier infilled Triassic valleys, but unlike many other valleys on Mendip, is not directly influenced by them. Ancient caves which have been truncated by downcutting of the combe provide evidence of a long history of solutional development. The combe also provides an excellent exposure of virtually the entire Carboniferous limestone sequence.

CHARTERHOUSE CAVES

Highlights

The Charterhouse caves encompass classic examples of vadose swallet caves in steeply dipping limestones. Their varied and complex morphologies, and extensive sediment and speleothem deposits, provide a valuable record of the Pleistocene development of the Mendip plateau and adjacent lowlands. These caves have been, and continue to be, the scene of intensive scientific research, often pioneering new techniques and methodologies. As a basis for so much karst research they are of international importance.

Introduction

Close to the centre of the Mendip limestone plateau, a group of ten caves lies along the southern flank of Black Down, 3 km north-east of Cheddar (Figures 5.1 and 5.8). Four of these are major influent cave systems, fed by allogenic streams draining south from the Old Red Sandstone outcrop of the Black Down pericline. The sinks are all close to the base of the Black Rock Limestone, and the known cave passages are in the lower beds of this unit, which dips to the south at 15–30°. The limestone is fractured by a number of faults, which may be associated with local steepening of the dip and have brecciated zones up to 6 m wide; it is also well jointed, with the dominant set having a roughly north–south trend. All the water from the caves resurges at Cheddar Rising, at the foot of the gorge.

The Charterhouse caves have been intensively studied, partly as a consequence of their proximity to the very active karst research unit in Bristol University. Detailed accounts of the main caves have been published by Goddard (1944), Stride, R.D. and Stride, A.H. (1946, 1949), Atkinson (1967), Atkinson *et al.* (1986), Ford (1964), Norton (1966), Smart and Stanton (1974) and Smart *et al.* (1984). Descriptions of the caves can be found in Barrington and Stanton (1977) and Irwin and Jarratt (1992). Further accounts of the geomorphology and development of the systems are given in Drew (1975b), Donovan (1969) and Ford (1965b, 1968). Aspects of the hydrology have been investigated by Atkinson (1968b), Atkinson *et al.* (1967), Drew (1975a), Stenner (1973), Smart and Hodge (1979, 1980), Smith and Mead (1962), Stanton and Smart (1981), Friederich (1981) and Friederich and Smart (1981, 1982). Effects above and below ground of the major floods in July 1968 were described by Hanwell and Newson (1969, 1970), Newson (1969) and Savage (1969). Uranium-series dates for some of the sites have been published by Atkinson *et al.* (1978, 1984); others remain unpublished, while dates derived from uranium-series decay and electron spin resonance, and studies of sediment remnant magnetism are recorded by Farrant (1995).

Description

The most westerly cave in the group is Tyning's Barrows Swallet (Figure 5.8). The cave consists of an initial series of narrow vadose passages descending steeply downdip and through several large rift chambers. These tributaries converge before entering a much larger vadose canyon, which is up to 5 m high and wide. The passage then swings round to the west, following a predominantly strike-orientated course, with many minor offsets on cross joints, as far as a sediment choke. The whole system shows very close joint control of its passages. There are extensive breakdown deposits but very few speleothems.

The swallet system of GB Cave lies at the foot of a short blind valley which ends on the limestone boundary (Figure 5.8). It contains almost 2000 m of passages (Figure 5.3) extending to a depth of 135 m. A number of small inlets near the entrance converge on the head of the main streamway. The passage from the Gorge to the Main Chamber is the largest in a Mendip cave, a vadose canyon in places more than 10 m high and wide. It contains massive banks and terraces of sediment and breakdown debris (Figure 5.4), and descends steeply to a sediment choke. Extensive inlet passages and several oxbows on its western side are smaller but mimic the overall morphology. Rhumba Alley and some of the inlet passages near the entrance show clear evidence of initial phreatic development, and a phreatic half-tube is visible in the roof in parts of the Gorge. From the south end of Main Chamber several much smaller distributary passages branch off. These have a much more gentle gradient and show clear morphological evidence of phreatic development below the water table. Above the downstream choke one of these abandoned distributary passages extends to further chokes and the Great Chamber, 50 m in diameter and extensively modified by roof collapse and upward stoping.

Many of the passages in GB Cave show close geological control, with the dip of 25° influencing the profile, and joints and faults dictating the plan relationship of the various passages. Throughout the cave, speleothems of various types are abun-

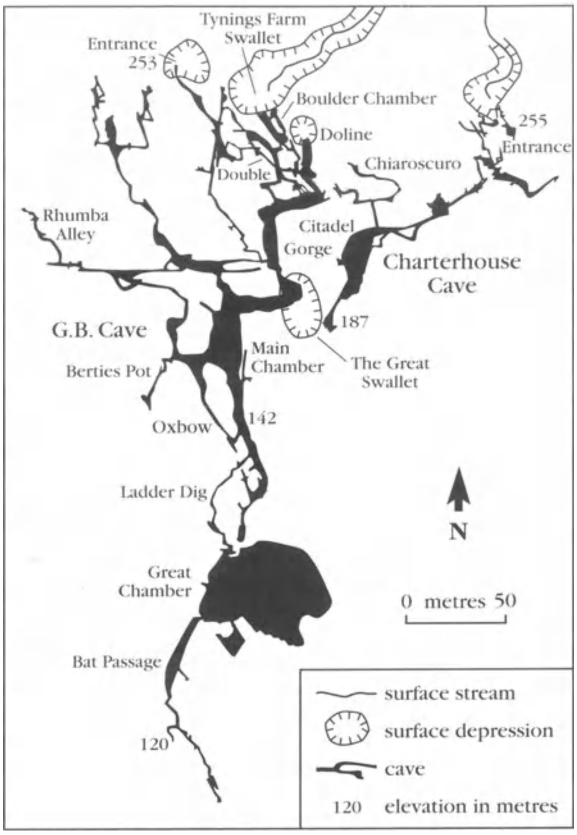


Figure 5.3 Outline map of GB Cave, Charterhouse Cave and the main surface features above them (from survey by University of Bristol Speleological Society).

dant, some of them formed of aragonite or interlayered calcite and aragonite. In addition there are extensive clastic deposits interbedded with stalagmite layers and recording episodes of erosion and deposition. Major flooding in July 1968 modified parts of the system considerably, blocking at least one of the inlet passages, causing surface collapse into the northern end of the Gorge and causing extensive scouring and redeposition of some sediment sequences. Dating of stalagmite and sediment sequences (Atkinson *et al.*, 1978; Farrant, 1995) has revealed evidence of at least four phases of speleothem growth timed at about >330, 170–120, 63 and <13 ka.

Adjacent to GB Cave lies Charterhouse Cave (Figure 5.3). This resembles GB in having a series of small inlet phreatic tubes leading into a main streamway, the Citadel, which has been greatly enlarged by vadose erosion. The cave is essentially part of GB, but there is no passable connection between the two. It also contains thick clastic sediment deposits and is exceptionally well decorated with speleothems. Both GB and Charterhouse Caves lie below a shallow dry valley where successive stages of fill have choked old sinks and thereby generated multiple sink passages on diversionary routes which coalesce underground.

Longwood Swallet lies in the floor of one of the main valleys which are tributary to Cheddar Gorge

Figure 5.4 Massive banks, terraces and false floors of coarse breakdown, clastics and stalagmite flowstone in the Gorge of GB Cave. (Photo; A.C. Waltham.)

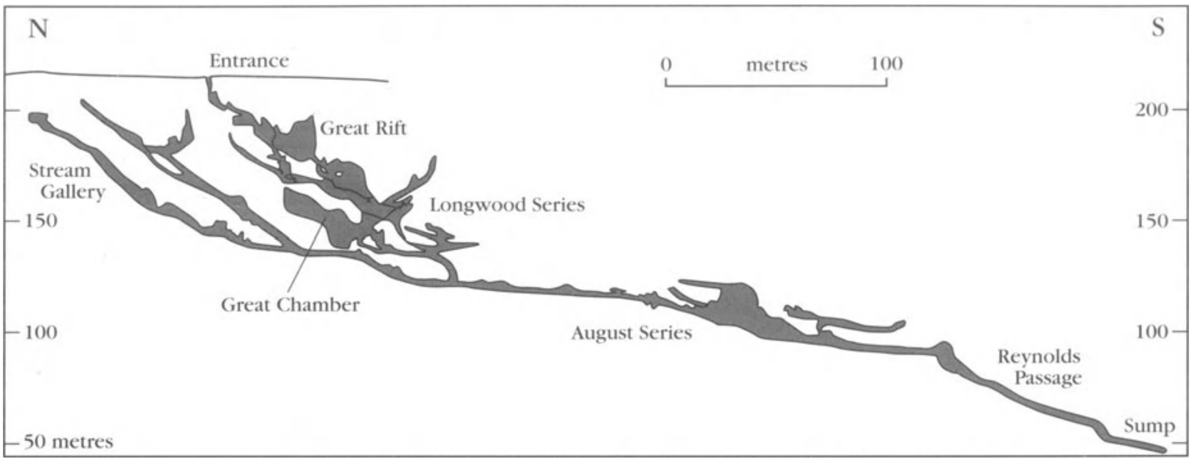


Figure 5.5 Extended profile of Longwood Swallet (from survey by University of Bristol Speleological Society).

(Figure 5.8). It contains over 1600 m of cave passages, reaching to a depth of 175 m. A natural shaft in the valley floor leads to a complex series of passages with several large chambers developed on faults and enlarged by collapse (Figure 5.5). Evidence of an extended phase of early high-level phreatic development is seen in the presence of tubes, loops and avens with crests at the same height. Lower down the cave these passages converge on a steep fault-controlled rift descending to the main streamway. This extends for more than 500 m along mainly vadose canyon passage often only a metre or so wide but of considerable height and with phreatic remnants preserved in places. Upstream, separate tributary passages have been followed to chokes a short distance below the surface sinks. Downstream the known cave ends at a sump beyond a series of very narrow joint-guided rift passages. Although the system contains relatively few speleothems there are extensive clastic sediment deposits containing interbedded stalagmite layers.

Rhino Rift consists essentially of a single series of vadose shafts developed along a fault and adjacent joints and descending to a small, choked phreatic tube at a depth of 145 m. The entrance lies in a tributary valley below Longwood Swallet, some distance south of the boundary between the Black Rock Limestone and the underlying shales (Figure 5.8). The cave contains extensive collapse debris but relatively little finer clastic material or speleothems. Dating of flowstone indicates that the cave has been in existence for at least 75 000 years (Atkinson *et al.*, 1984).

Manor Farm Swallet lies at the next sink to the east (Figure 5.8), and has over 900 m of passage (Figure 5.6). A series of fairly small, steeply descending, vadose inlet passages include one choked by the debris from the Great Shaft, which collapsed in the 1968 flood. These all unite before entering NHASA Gallery, which is a section of old phreatic passage extensively modified by collapse; it is up to 10 m wide and 3 m high, with a dipping bedding plane roof and a floor of mud and breakdown blocks. Parts of the cave contain thick clastic sediments and false floors, locally overlying massive gour barriers and flowstone. There are many stalactite curtains and banks of active flowstone.

A large stream sinks in the Blackmoor Valley (Figure 5.8), but the associated cave system has yet to be discovered. Several smaller cave systems have been found, including Blackmoor Flood Swallet, Waterwheel Swallet (Stanton, 1987) and Grebe Swallet. The latter is important as it contains evidence for the origin and emplacement of the lead ores in the Mendip Hills (Stanton, 1991).

Charterhouse Warren Farm Swallet (Levitan *et al.*, 1989), lies to the south of the Velvet Bottom valley (Figure 5.8). It is entered via a narrow shaft which drops into a series of phreatic passages, much modified by collapse, speleothem deposition and the influx of clastic material. The site is an important archaeological site and its position is intermediate between the Charterhouse swallet caves and the Cheddar resurgence.

Interpretation

The caves developed on the southern flank of the Blackdown pericline include classic examples of vadose caves developed in dipping limestone. They show a wide range of morphologies from the massive canyon passage in GB Cave to the

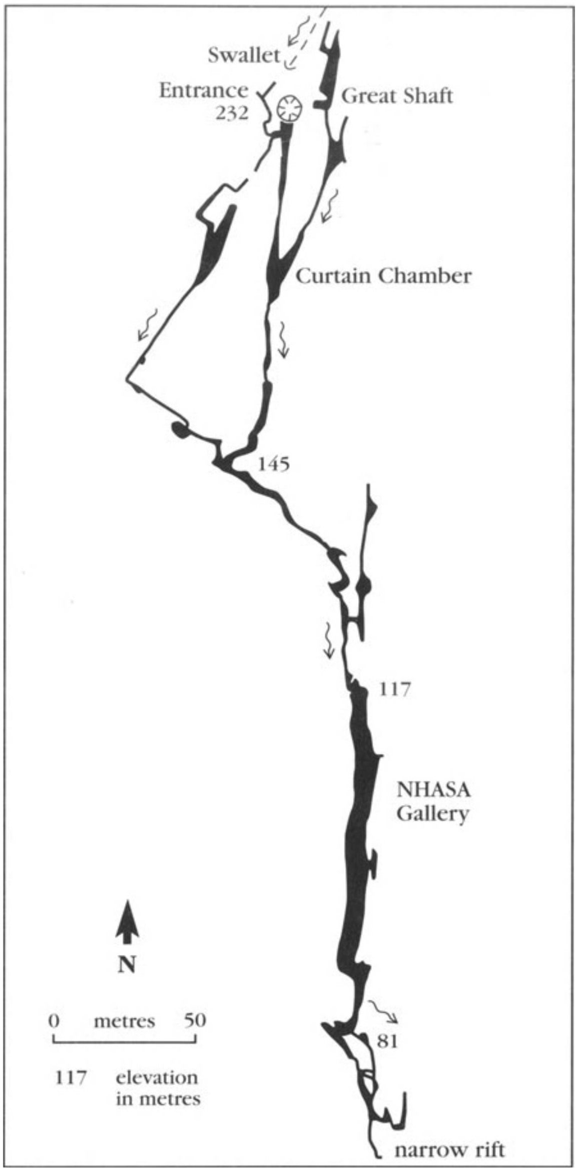


Figure 5.6 Outline map of Manor Farm Swallet (from survey by University of Bristol Speleological Society).

fault-guided rifts in Longwood Swallet. The smaller cave of Charterhouse Warren Farm Swallet is the only one known on the Mendip Hills which is intermediate between the predominantly vadose swallet caves and the largely phreatic caves at the resurgence.

All the major swallet streams were dye tested to the Cheddar Risings with travel times between 16 and 48 hours (Atkinson *et al.*, 1967; Drew, 1975a). Pulse wave tests at Longwood Swallet (Smart and Hodge, 1979) indicate that at low flow, only 9% of the Longwood-Cheddar conduit is vadose. Repeated dye tests at several sites on the Mendips included Longwood Swallet to demonstrate that the travel time is inversely proportional to resurgence output (Stanton and Smart, 1981). The presence of up to three fluorescein peaks for each trace suggests the possibility of several alternative conduits to the resurgence. The identification of a common resurgence at Cheddar has enabled the evolutionary history of the swallet caves to be compared and their response to base-level change at the resurgence examined.

Geomorphology

The Charterhouse caves display many of the features common to the typical Mendip cave. All the swallet caves exhibit complex passage networks developed predominantly downdip and extensively modified by vadose erosion, sedimentation and collapse. Many of the caves provide excellent evidence of structural control. Joint direction is the dominant control, and is clearly recognizable on cave plans (Figures 5.3, 5.6). Over 80% of the passages in Manor Farm Swallet are joint controlled (Smart and Stanton, 1974). Bedding plane control is shown by the downdip orientation of many of the passages while faulting is especially important in the formation of large chambers, notably in Longwood Swallet and vadose shafts such as Rhino Rift. All the swallet caves are developed at the base of the Black Rock Limestone.

The most westerly of the caves is Tyning's Barrow Cave. Although not a true stream sink, its morphology is similar to the swallet caves. Admirable structural control is shown in the lower streamway where downdip joint-controlled segments are linked by strike-orientated passages.

The most complex of the swallet caves and one of the most intensively studied is GB Cave, genetically related to the adjacent Charterhouse Cave. The geomorphology of GB Cave was first studied in detail by Ford (1964), who elevated it to his type example of a vadose drawdown swallet cave. He envisaged an initial period of phreatic erosion forming a complete passage network. This was followed by alternating phases of vadose drawdown, erosion, clastic sedimentation and speleothem deposition along the outline plan established during the initial phreatic phase. The water tables were initially controlled by the lack of cave development, but fell rapidly to a stable base level once a mature cave system had been established. Thus vadose cave development took place in a vertically extensive vadose zone, the sequence of captures and trenches being unrelated to base-level lowering.

The discovery of the neighbouring Charterhouse Cave enabled Smart *et al.* (1984) to test Ford's hypothesis. They concluded that rather than an initial period of phreatic development, followed by rapid base-level lowering and vadose drawdown, base-level probably fell slowly and intermittently, thus deepening the vadose zone slowly through time. During this period, passage morphologies reflect the transition from phreatic, through paraphreatic to entirely vadose conditions, which ultimately led to the abandonment of the strike-orientated pressure-fed phreatic conduits in favour of free-draining vadose dip tubes and joint-guided rifts. Initially the only mature phreatic conduits were along the Double Passage–Chiaroscuro Passage–Citadel route and the Rhumba Alley–Berties Pot–Ladder Dig route. The sequence of trenches in Charterhouse can be related to the declining water table level. Both Ford (1964) and Smart *et al.* (1984) recognized two major phreatic rest levels in the GB–Charterhouse system; in Double Passage at 238 m and a second just above Ladder Dig at 137 m. The multiplicity of inlet passages reflects the large number of sinks the stream has utilized through time, caused by the infilling of former sinks by clastic material and the opening of new ones when the loess cover was eroded.

The hydrology of GB Cave has been studied in some detail. D.C. Ford (1966) found that the calcium hardness of dripwaters varied by up to 50 ppm. Stenner (1973) showed that the increase in solute load in the GB Cave stream was not due to direct solution of calcite by the stream water, but was the result of admixtures of waters with higher calcium contents. Friederich and Smart (1981, 1982) studied the water in the vadose zone and based their classification of autogenic percolation waters on samples taken from GB Cave.

Longwood Swallet was studied in detail by Atkinson (1967). He suggested that initial phreatic erosion was followed by a fall in the water table by 56 m, thus initiating vadose erosion. He identified three further aggradation and two renewed vadose incision stages corresponding to further drops in base level, compared to the two identified by Ford (1964) in GB Cave. On this evidence, he concluded that Longwood was older than GB. Atkinson identified three phreatic rest levels in Longwood Swallet, at 138–141 m, 120–123 m and 90–93 m. There is slender evidence for a fourth at 70 m. The modern phase of vadose erosion is graded to a water table at 40 m. It is clear from their contrasting morphologies that the Longwood stream is capturing water from GB Cave. In GB, the large size of the gorge in comparison to the stream suggests that the cave once had a much larger catchment than at present, whereas the opposite is true in Longwood Swallet where a large stream flows through some small and immature passages. The incision of the deep Longwood valley has enabled headward erosion along the strike of the Lower Limestone Shales, beheading the GB catchment area.

In Manor Farm Swallet, Smart and Stanton (1974) identified an initial phreatic dip-tube network, which was later entrenched under vadose conditions as the phreas fell to a stable level at 120 m OD, shown by the excellent vadose trench graded to the floor of the NHASA Gallery phreatic tube. Two phases of vadose erosion followed by clastic sedimentation and speleothem deposition occurred, as the phreas dropped to 92 m and below 81 m. They concluded from the smaller size and relative simplicity of the cave that it was younger than GB Cave and Longwood Swallet.

The two non-swallet caves show markedly differing morphologies. Rhino Rift is a classic vadose invasion cave (Ford, 1965b), comprising of five vadose shafts descending to a small phreatic passage at the 75 m level. Stanton (1972) argued that Rhino Rift was an earlier sink for the GB stream and thus predated GB Cave. Atkinson *et al.* (1984) refuted this hypothesis, and suggested that the cave was formed by local run-off from snow-patches sinking along the line of a prominent fault. Similar modern examples can be seen in alpine karst areas, and it is commonly found that vadose shafts can develop to large dimensions with a comparatively small stream (Pohl, 1955). In contrast, Charterhouse Warren Farm Swallet (Levitan *et al.*, 1989) is dominantly phreatic and represents an important link between the Charterhouse swallet caves and the resurgence in Cheddar. It consists of a remnant of phreatic passage which functioned as a major strike integrator when the regional base level was at or above 227 m. Three types of sediment fill in the cave include a siliceous allochthonous gravel derived from the Blackdown pericline, several calcareous allochthonous fills, and limestone breakdown. The calcareous fills are especially important as they contain profuse archaeological remains (Levitan *et al.*, 1989).

Geochronology

Dating of the cave sediments, using uranium-series, electron spin resonance (ESR) and palaeomagnetic techniques (Atkinson *et al.*, 1978, 1986; Atkinson and Smart, 1982; Levitan *et al.*, 1989; Farrant, 1995; Smart *et al.*, unpublished data), has revealed much about the evolution of the Charterhouse caves; the data have enabled comparisons to be made between the swallet caves, and to the sequence of caves at the resurgence in Cheddar, and thus to changes in external base level. Early work with uranium-series dates showed that GB Cave was older than 350 ka. The chronology was extended using ESR and palaeomagnetic methods back as far as 900 ka (Farrant, 1995). These dates demonstrate that the early phreatic conduit in GB and Charterhouse, along the Double Passage–Chiaroscuro Passage–Citadel route, was probably established before about 900 ka and certainly prior to 780 ka.

The levels of the phreatic still-stands (at about 238, 138, 120, 90 and 70 m) show a good correlation between all four of the major swallet caves (Smart and Stanton, 1974; Farrant, 1995). The timing of these major phreatic still-stands has been estimated; the highest at 238 m is about 900 ka, with the lower levels at about 480, 350–380, 200–225 and 95–100 ka, respectively. Similar distinct levels have been found in Gough's Cave at the resurgence, inviting correlation with the swallet caves (Figure 5.10). Ford (1964) and Atkinson *et al.* (1978) correlated the Ladder Dig water table level at 138 m to that of Great Oone's Hole in Cheddar on stratigraphic and geomorphic grounds. This was challenged by Farrant (1995) who suggested that the 120 m level drained to Great Oone's Hole based on evidence from uranium-series dates (Figure 5.10).

The good correlation of water table levels between the major swallet caves at Charterhouse suggests they underwent a uniform response to changes at the resurgence. This response is driven by progressive base-level changes at the resurgence which propagate up the conduit. The rate of propagation is controlled by the abandonment and capture of phreatic links at the resurgence (Smart *et al.*, 1984). This correlation is not so clear in the Priddy caves, where the swallet caves have markedly contrasting morphologies. This is probably because the Priddy-Wookey system responds slower to base-level lowering; it has yet to fully respond to the last phase of base-level lowering, as active vadose incision of the earlier phreatic loop crests is still progressing.

Thick, coarse, angular, sandstone and limestone gravel fills occur in all the swallet caves. At many places, stream erosion has undercut these cemented gravels leaving perched false floors. The gravels were emplaced under periglacial conditions by the transport of frost-shattered surface material into the cave by solifuction and debris flow events. The associated speleothem was deposited in the intervening warmer periods when increased biogenic soil activity raised the carbon dioxide levels in the soil, causing saturation of the percolation groundwater and stalagmite deposition. In GB Cave, several generations of fills are recognized, interbedded or capped with calcite flowstones which have been dated. At least eight major gravel fills have been identified in a complex sequence of gravel emplacement and speleothem deposition (Figure 5.7). Within the limits of the available dating, the phases of gravel emplacement appear to correspond with the cold stages of the Pleistocene. Stratigraphical relationships show that similar, complex sequences occur in Charterhouse Cave,

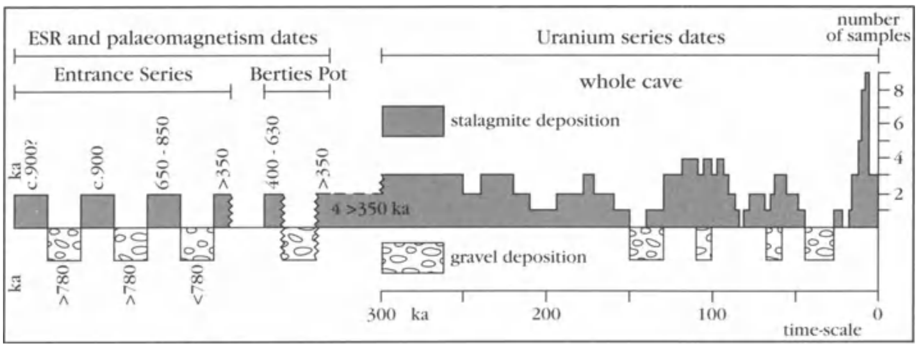


Figure 5.7 Phases of stalagmite and gravel deposition in GB Cave, with a chronology based on stalagmite dates obtained from uranium-series, ESR and palaeomagnetic techniques (after Farrant, 1995). Stalagmite ages are represented covering the error bars on the dated samples; actual time spans of the deposition phases may be smaller, but data from stalagmites as yet undated may increase the lengths of the deposition phases.

Longwood Swallet and Manor Farm Swallet, but have yet to be dated. However, the dangers of trying to elucidate the developmental history of the cave from the clastic sequences alone were highlighted by the profound changes wrought by the 1968 flood (Hanwell and Newson, 1970).

Conclusions

The Charterhouse caves provide some of the finest examples of cave development in dipping limestones. The cave morphology, clastic sediments and speleothems are the most intensively studied in Britain, and have far-reaching implications for the study of cave development, karst hydrology and Pleistocene chronology. The wealth of dated sediment and speleothem has enabled the construction of a remarkably long chronology and an elucidation of the geomorphic history of the Mendip Hills. The correlation of water table levels in the swallet caves with those at the Cheddar caves demonstrates the role of base-level control at the resurgence on the whole conduit system. Although the designation of some of the caves as type sites has been challenged, the pioneering nature of the underground work renders the Charterhouse caves of international importance.

CHEDDAR GORGE

Highlights

Cheddar Gorge is perhaps the single best known karstic feature in Britain and provides a spectacular example of a limestone gorge fed by a system of dry feeder valleys. The morphology of the gorge and the associated well-dated caves, provides a unique insight to its geomorphic evolution over the last million years. It demonstrates the results of episodic fluvial erosion in a karst terrain, which was left dry when the surface drainage disappeared underground into caves.

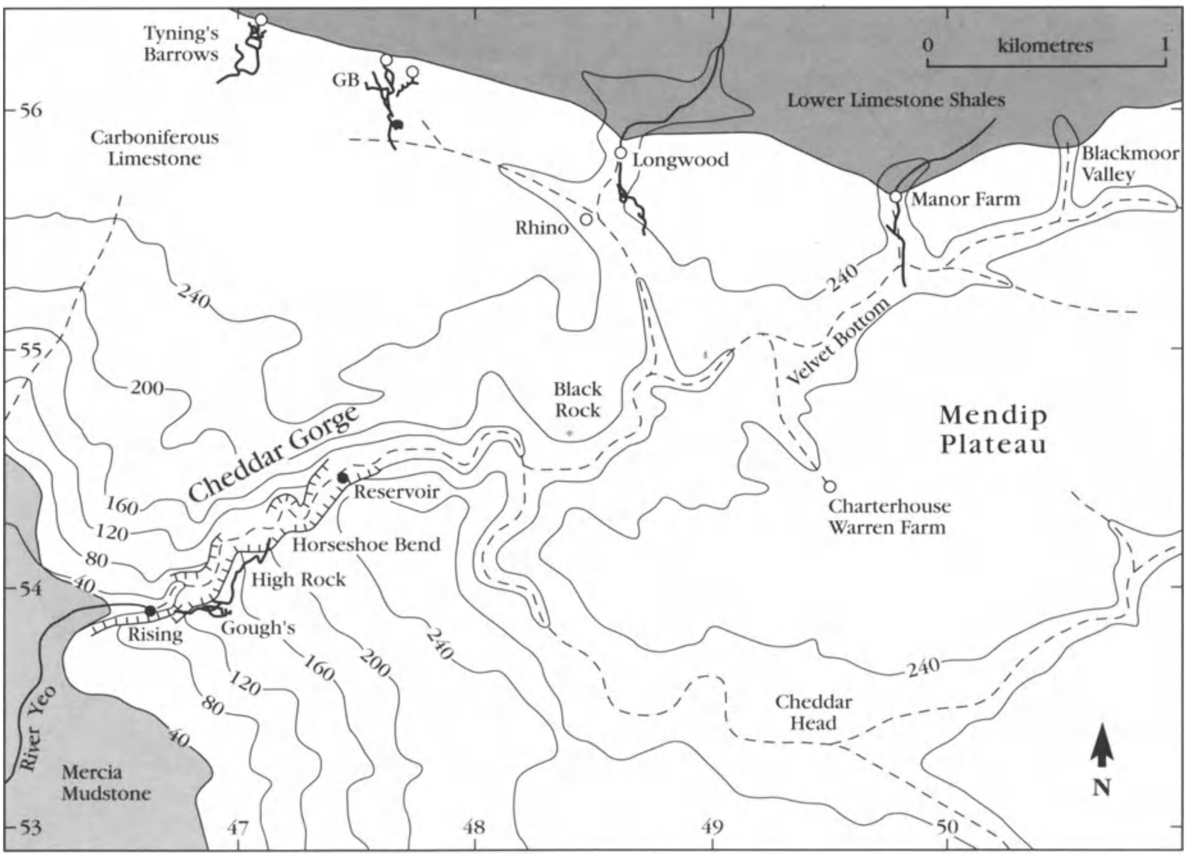


Figure 5.8 Map of Cheddar Gorge and the lower part of its dry valley system reaching across the karst to the edge of the Mendip Plateau.

Figure 5.9 Cheddar Gorge, looking upstream from the northern rim opposite High Rock. The limestone dips to the right, ensuring the stability of the cliffs on the right, while the left slope is cut back almost to the dip of the bedding planes. (Photo: A.C. Waltham.)

Introduction

Cheddar Gorge extends approximately 2 km eastwards from Cheddar village and forms the downstream portion of an extensive dry valley network which once drained most of the Mendip plateau (Figure 5.8). Above Black Rock three dry valleys form part of the dendritic network feeding into Cheddar Gorge. These represent particularly good examples of dry valleys incised into the Mendip plateau. Below Black Rock the gorge is far more precipitous and is entrenched to 120 m deep into the limestone where it descends steeply towards the flank of the Mendip Hills at Cheddar village. Many caves exist in the side of the gorge and are described in Barrington and Stanton (1977) and Irwin and Jarratt (1992).

The origins of the gorge have been the subject of much discussion since the early 1800s (Dawkins, 1862; Winwood and Woodward, 1891; Callaway, 1902; Reynolds, 1927; Stride, A.H. and Stride, R.D. 1949; Ford and Stanton, 1968; Trudgill, 1977; Smith, 1975a, b, 1977), with several theories including cavern collapse, earthquake activity, and incision by a periglacial meltwater river being put forward. Only the last of these now carries any credence. Recent work by Atkinson *et al.* (1978), Atkinson *et al.* (1986), Smart *et al.* (1988b) and Farrant (1995) has focused on the morphology and dating of both the swallet caves and the caves exposed in the gorge, which has led to a better understanding of the geomorphic history of the gorge, and enabled it to be set in a chronological framework.

Description

Cheddar Gorge is entirely incised into the Carboniferous Limestone succession on the southern flank of the Mendips. Above the gorge the main valley continues south-east from Black Rock, towards Cheddar Head where it widens, becomes more open and divides again. From Black Rock, the northern tributary splits and extends up Velvet Bottom and the Longwood valley, as far the modern stream sinks at Longwood swallet and the other Charterhouse caves, which lie at the contact of the limestone with the Lower Limestone Shales on the south side of the Black Down pericline.

The Longwood–Cheddar valley is the best example of the relationship between the stream sink, dry valley and resurgence (Figure 5.8). The upper dry valleys descend gently from the plateau surface at an elevation of around 245 m to Black Rock, where the gradient increases markedly. Over the next 2 km the floor descends over 130 m to an elevation of 20 m at the resurgence (Figure 5.10).

The rock exposed in the gorge is mainly the Clifton Down Limestone, although some Hotwells Limestone crops out in the cliffs of the downstream part. The limestone dips at about 20° to the south and is cut by a major joint set trending NNW–SSE which has a strong influence on the cliff morphology and on the caves. The cliffs on the south side reach heights of 120 m, and are dominantly vertical, mainly aligned on the strong joints, and broadly stable because the dip is away from the gorge and into the cliffs. The northern side is less steep, because the southerly dip facilitates bedding plane slip and the formation of steep bedding plane slabs. The cross-section of the gorge is therefore an asymmetric V-shape, typical of fluvial excavation (Figure 5.9). Its floor meanders between rock buttresses whose pattern is at least in part orientated by the major fractures which also determine the profiles of the main limestone walls. The gorge is now dry, except in major floods such as that of July 1968 (Hanwell and Newson, 1970), and the drainage is entirely underground, the water resurging at the Cheddar Risings at the foot of the gorge.

In the gorge walls are a number of caves, the most important of which are the Gough's Cave complex, clustered around the resurgence at the foot of the gorge, and Reservoir Hole. Many of these contain stalagmites which have been dated by uranium-series and electron spin resonance methods. Stalagmites in the upper parts of Gough's Cave give uranium-series ages around 235 ka (Farrant, 1995). A scalloped flowstone in Great Oones Hole above yielded uranium-series ages around 375 ka (Farrant, 1995), but an electron spin resonance date for some of the same flowstone yielded an age of 1060 ka (Smart *et al.*, 1988); the latter is probably an overestimate, reflecting uncertainties in the dosimetry. It is apparent that the Gough's main bore was abandoned by around 120 000 years ago. Uranium dates from Reservoir Hole show the upper cave levels had been abandoned by 350 ka. The gorge shows many classic features of a fluvially excavated valley. These include a steep long profile, a recognizable V-shaped cross-section, a clear relationship to normal fluvial valleys, lack of any collapse debris, knickpoints including a conspicuous one at Horseshoe Bend, and a large alluvial fan dissected by the modern streams at the foot of the gorge.

Interpretation

Cheddar Gorge is Britain's largest and most spectacular karst gorge. It provides a particularly fine and easily accessible example of fluvial erosion in a karst landscape, although this was not always thought to be the case. The earliest theories on the formation of the gorge were put forward by Dawkins (1862) who suggested that it was formed as a result of cavern collapse. This was also the view held by Winwood and Woodward (1891) in an account of a Geologists' Association field excursion to Mendip. Later, Callaway (1902) noted the joint-controlled nature of the gorge and concluded that it must have been formed by a subterranean stream. Reynolds (1927) was one of the first to suggest that the gorge was cut by a surface river. The last advocates of the collapsed-cavern theory were Stride, A.H. and Stride, R.D. (1949), although this myth is still often perpetuated in many modern geological texts. Thus Cheddar Gorge is probably Britain's most frequently misinterpreted geomorphic feature.

It is possible that some subaerial fluvial excavation took place before karstification had developed sufficiently to divert drainage underground. However, this process cannot have played a major role as the gorge truncates older high-level caves. Ford and Stanton (1968) demonstrated that the gorge must have been formed by a subaerial stream, pointing to the difference between the gorge's long profile and that seen in the stream caves. They also noted that the gorge has often cut cleanly through existing cave passages. Additionally, they stressed the disparity between the immense volume of the gorge and that of even the largest Mendip caves such as GB and Lamb Leer. Cavern collapse can only have played a minor, if not trivial role in the gorge formation. A useful summary of the different theories on the origins of the gorge is published in Smith (1975a).

The commonly accepted view is that Cheddar Gorge was incised over the last million years by a subaerial meltwater river during periglacial periods, when underground drainage was restricted

(Smith, 1975a). Extensive mass movement and solifluction, coupled with development of permafrost during glacial periods led to the blocking of the swallet caves with ice and frozen mud and the establishment of surface drainage. Due to the nature of the Mendip plateau, the steepest stretches were at the valley mouth, where incision was therefore the greatest. Cheddar Gorge is the largest of the gorges on the Mendips because it drained the bulk of the plateau and because its lower end was lower than those elsewhere, thus maximizing its erosive power. Each successive periglacial episode caused renewed incision of the gorge, while during the interglacial periods, underground drainage was renewed and the gorge became dry, except under conditions of major flood. A late Pleistocene fauna in several of the valley floor caves (Tratman, 1975; Currant, 1987) shows that the gorge had reached almost to its present floor level by early Devensian times.

Erosion of the softer Jurassic and Triassic rocks along the southern flank of the Mendips during interglacials enabled each successive reactivation of the valley to work from a lower level, thus creating a series of knickpoints in the gorge which receded upstream through time (Barrington and Stanton, 1977). The knickpoints have been correlated on geomorphic grounds with a series of erosional benches along the southern flank of Mendip (Ford and Stanton, 1968). However, these erosional benches may not accord with the former positions of base level; Stanton (1985) reinterprets them as random associations of stratimorphic flats. These knickpoints may also correlate with a series of abandoned cave levels in the Gough's Cave system (Ford, 1965b; Ford and Stanton, 1968; Stanton, 1985; Farrant, 1991). The cave levels relate to a succession of past stable resurgence positions (Figure 5.10); they may also correlate with levels in the swallet caves to the north, but more dating evidence is needed before conclusions can be drawn.

Recent work has concentrated on defining the rate of excavation of the gorge, its development through time, and how it relates to the climate fluctuations during the Pleistocene. By dating the caves and relating their morphology to former stages in the evolution of the gorge, Atkinson *et al.* (1978) and Farrant (1991, 1995) have shown that the lower section of the gorge has been incised at an average rate of $0.25\ \mathrm{m\ ka^{-1}}$. Extrapolation of this rate up to the plateau surface suggests incision of the gorge began approximately a million years ago. With refinement of a chronology it may prove possible to ascertain the relative importance of periglacial erosion and temperate fluvial erosion, but low resolution of the older dates makes this very difficult for the earlier phases of the gorge's history. The large number of dated stalagmite samples from extensive associated caves makes this one of Britain's best documented karst gorges.

Conclusions

Cheddar Gorge is the largest and most spectacular karst gorge in Britain, and is unique in that a series of well-dated caves in the gorge walls have enabled its geomorphic evolution to be deduced. Although often wrongly cited as a collapsed cavern, it is a fine example of fluvial erosion in a karst landscape, left dry by the onset of underground drainage. Cheddar clearly shows the relationship between the gorge and the dry valleys which feed it, while the abandoned and active caves show the relationship between surface and underground features in a limestone karst.

CHEDDAR CAVES

Highlights

The Cheddar caves show the development of successively lower passages to a sequence of resurgences; these formed at positions dictated by the lowering of a periglacial surface drainage route which constituted the local base level. The main upstream river cave is aligned with the dip, and provides a fine example of phreatic loops developed on joints and bedding planes in dipping limestone. Downstream of this, distributary passages, aligned on the strike, demonstrate the role of local geological structures.

Introduction

The Cheddar caves are located in the walls and beneath the floor of the lower end of Cheddar Gorge (Figure 5.1). A number of caves represent fragments of a single extensive system. Cheddar Rising, the outlet for the active cave system associated with the Cheddar Gorge, is the largest resurgence in the Mendip Hills. Allogenic water drains off the Old Red Sandstone slopes on the south side of Blackdown Hill, entering sinks in the

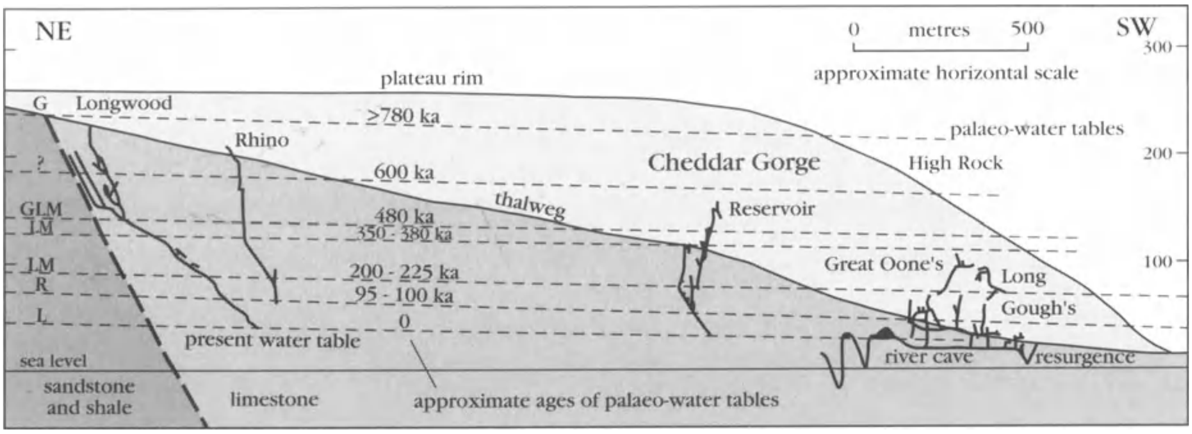


Figure 5.10 Long profile of Cheddar Gorge up into the Longwood Valley, with the caves beneath. Each palaeo-water table is recognized from cave and surface morphology, and is dated from the sediments in associated cave passages at both the swallet and resurgence ends of the system. The water tables steepen greatly in the sandstone and shale, but are marked beyond the limestone only to label the caves in which each is recorded (G = GB Cave; L = Longwood Swallet; M = Manor Farm Swallet; R = Rhino Rift). The horizontal scale is distorted by the projection, and the vertical scale is exaggerated three times (largely after Stanton, 1985; Farrant, 1995).

Charterhouse area, 3 km to the north. A second group of sinks feeding to the Cheddar resurgence lies up to 11 km to the east, draining the northern limb of the North Hill pericline (Figure 5.1). Atkinson (1977) identifies the drainage system as an important example of conduit flow with diffuse input and storage. The explored caves at the resurgence end of the system are developed largely along the southern flank of the gorge, with the most extensive section, Gough's Cave, located at the lower end of the gorge immediately adjacent to the active risings.

The Cheddar caves have been the subject of many popular and scientific publications. Descriptions of the caves are given in Barrington and Stanton (1977) and Irwin and Jarratt (1992), with a more detailed account of discoveries in the river cave given by Palmer (1988) and Stevenson and Palmer (1986). Aspects of the hydrology are discussed by Atkinson (1977), Drew (1975a), Drew *et al.* (1968), Smart (1981), Smart and Hodge (1980) and Smith and Newson (1974). The history of development of the cave systems has been discussed at length by Ford (1965b, 1968), Drew (1975b), Stanton (1985) and Farrant (1991). Aspects of the sediments contained within Gough's Cave have been the subject of papers by Collcutt (1985) and Leroi-Gourhan (1985). There have been many publications concerned with the Pleistocene fauna recovered from excavations in the Cheddar caves; these have been reviewed by Jacobi (1985).

Description

The most extensive caves in Cheddar Gorge lie close to the present resurgence. Numerous other caves have been explored, most of which probably are connected with Gough's Cave in some way. The most important of these are Great Oone's Hole, Long Hole and Reservoir Hole (Figure 5.10).

Gough's Cave, together with Great Oone's Hole and Long Hole (Figures 5.10 and 5.11), contains more than 2200 m of explored passage developed over a total vertical range of more than 180 m. Part of the system is currently operated as a show cave. The lowest part of Gough's Cave is the active river cave, a phreatic tube typically 5 m wide and 3 m high. Upstream, the River Cave forms a series of deep phreatic loops on a north–south alignment (Figures 5.10 and 5.11). In each loop, the river flows under pressure almost straight down the dip of bedding planes and then rises through vertical rifts aligned on joints. The loops reach depths of up to 58 m, more than 30 m below present sea level. Vadose incision by the river has cut a loop crest to leave the Bishop's Palace chamber, which rises to almost 30 m above the present river level, partly due to upward stoping of the roof. The flooded passages are up to 7 m wide and 2 m high, and contain laminated mud sediments which are being re-excavated by the river. Downstream, the river passes through Lloyd Hall, a rift chamber 20 m long and 12 m

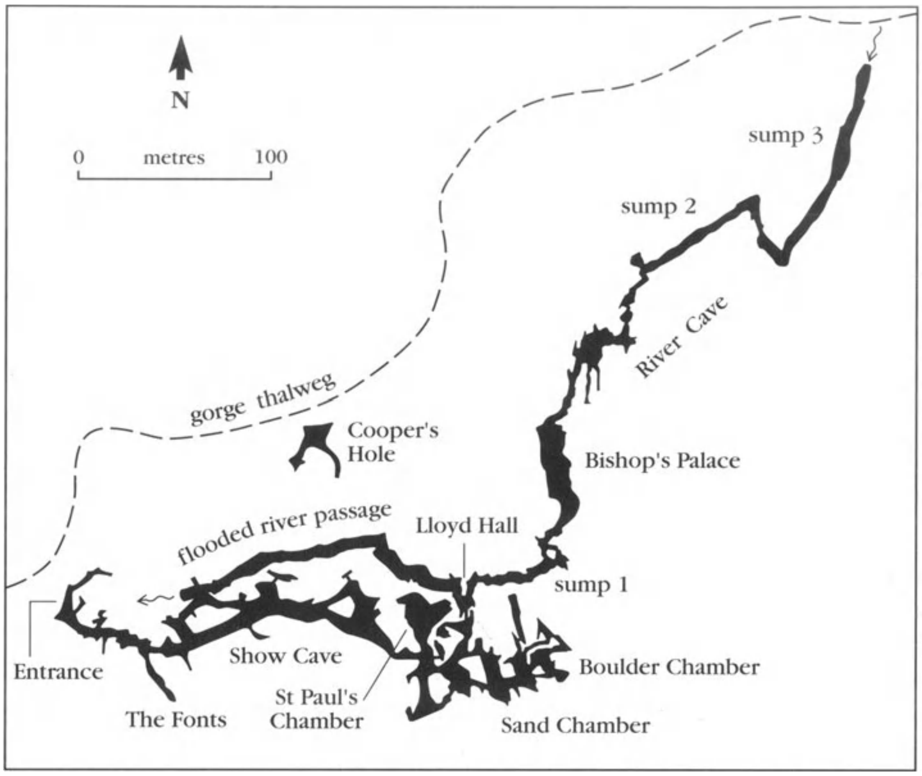


Figure 5.11 Outline map of Gough's Cave, Cheddar. Long Hole and Great Oone's Hole lie partly over the show cave section and are omitted for clarity (from surveys by Wessex Cave Club and Cave Diving Group).

high with a roof connection to the cave level above; it then continues to a choke close to the resurgence.

The lowest of the abandoned levels lies at about 45 m OD, about 18 m above the river cave, and includes much of the show cave (Figures 5.10 and 5.11). Part of the tourist route follows a magnificent phreatic tube of similar dimensions to the active river passage almost directly below. Phreatic solution along the NNW–SSE joints, has created cross rifts and avens up to 30 m high. The large gour pools of the Fonts occupy a side passage, and speleothems are abundant at several points within this level. Speleothem dates from this level (Farrant, 1995) indicate an age in excess of 120 ka. The passage terminates at a large chamber above Lloyd Hall, but further caves at this level are known to extend west from the lower part of Boulder Chamber.

Boulder Chamber has high-level, tall chambers and rifts, well decorated with speleothems in places, linked by largely sediment-filled passages at about 60 m above sea level; speleothem dates indicate an age of more than 230 ka (Farrant, 1995). Excavations in the floor of Boulder Chamber revealed a narrow shaft, filled with boulders and clastic sediment, which once formed an inlet to the system (Stanton, 1965).

The highest level in the Gough's Cave system is represented by Great Oone's Hole and Long Hole, both of which lie almost directly above parts of the lower levels. Flowstone with a uranium-series date of 380 ka (unpublished) lies in Great Oone's Hole, which has 150 m of abandoned phreatic passage opening in the side of the gorge more than 60 m above the show cave entrance. Beyond a level section the passage gently descends to a choke close above one of the chambers in the back of Gough's. Long Hole, with 260 m of passages, probably represents a downstream continuation of Great Oone's Hole. Descending rifts connect to chambers in Gough's Cave, while the main cave ascends to passages on the same level as Great Oone's Hole choked close to the present surface.

Reservoir Hole lies on the southern side of the gorge upstream and north-west of Gough's Cave (Figure 5.10); it covers a vertical range of 123 m and descends to within 8 m of resurgence level. A descending series of small tubes and tall vertical rifts intercepts an abandoned stream passage 6 m wide containing mud formations and thick sediments. This continues downstream beneath tall avens to a choke. A side passage has a decorated collapse chamber and avens including the Great Aven, a wide rift 60 m high on a fault; it may be a truncated downstream phreatic loop. A high-level passage contains speleothems dated in excess of 350 ka (Farrant, 1995).

Interpretation

With the recent discovery of the river cave the Gough's Cave system now shows, more clearly than any other cave in Mendip, the characteristic geological controls exercised by the dipping limestone. The downstream section of the river cave, and all the old passages developed at the three levels above, are developed along the strike and hence maintain fairly constant levels throughout their lengths. The flattened elliptical shapes of the main passages reflect a strong bedding plane control, though narrow rifts have developed at prominent joints. In contrast, the upstream section of the river cave displays classic dip-orientated development in inclined limestone, with deep phreatic loops caused by the passage following the bedding downdip before rising vertically up joint-guided rifts. The close superimposition of the three levels of old strike passage above the present active river passage suggests that their position has been controlled by a major east–west zone of fracturing and minor folding which has coincided with the available outlet position in the gorge floor.

Ford (1965b, 1968) and Farrant (1991) have identified at least four distinct levels within the strike-orientated part of the Gough's Cave system. These have been interpreted as zones of mature, shallow phreatic cave development formed sequentially below four base levels, possibly recognizable also as surface terraces (Ford and Stanton 1968; Stanton 1985). No evidence has been found for any higher levels of development in the dip-orientated upstream section of the river cave, suggesting that these upper three levels were all fed by phreatic lifts towards the eastern end of the strike passages. There are at least two abandoned phreatic lifts from the River Cave into the upper levels of Gough's Cave, but the size of these appears inconsistent with the flow which they would have had to transmit, and the ancient strike passages may have been fed by other, unknown, dip passages (Farrant, 1991, 1995).

Great Oone's Hole and Long Hole represent parts of a single conduit which was the first phase outlet (Figure 5.10); this may be correlated with the middle level of Longwood Swallet and Manor Farm Swallet at Charterhouse (Farrant, 1995). The connection between Long Hole and Gough's Cave is much more recent and due to chance interception by later passages. The second phase is represented by the high-level chambers in Gough's Cave, perhaps fed by a phreatic lift along Damocles' Rift. It may correlate with the lower levels in Longwood Swallet and Manor Farm Swallet (Farrant, 1995), and the main resurgence at the time may have been Cooper's Hole, now largely truncated by incision of the gorge. Gough's Old Cave may represent a late diversion of this phase or else a southerly derived inlet. The modern tourist route in Gough's Cave, and the isolated fragment of Cox's Cave further to the west, were formed in the third phase of phreatic cave development, possibly fed by a phreatic lift at Boulder Chamber, since bypassed by a lower connection. The abandoned passages are all phreatic with little vadose modification, and the fourth phase is the modern river passage still largely within the phreas. The four levels represent phases of adjustment to successive resurgences (Figure 5.10), whose positions were determined by surface lowering of the lowlands to the south of the Mendip Hills. Incision of the gorge may also have exerted a significant influence through breaching of phreatic drainage routes. The vadose entrenchment at the crests of the phreatic loops in the active river passage, particularly at Bishop's Palace, also reflects the lowering of the modern resurgence level.

Reservoir Hole is a complex system showing very strong control by faulting and associated fractures. The main passage has an inclined profile and appears to represent the downdip sector of an old phreatic loop on a tributary to the main Cheddar cave system.

Speleothems and thick clastic sediment deposits in the various cave levels of Cheddar Gorge may enable a chronology to be constructed for the sequence of events in the development of the cave system. Since this sequence reflects the history of the surface landscape through the Pleistocene, such an investigation will have a fundamental bearing on any future study of the geomorphological evolution of this area of Somerset.

Conclusion

The Cheddar caves clearly show phreatic cave development at the resurgence end of a major cave system extending through the dipping limestones of the Mendip Hills. Both active and abandoned passages exist, formed both along the strike, and with a dip-orientated, joint-guided, looping profile. The several levels of old cave record the Pleistocene entrenchment of Cheddar Gorge and the adjacent lowlands; they contain stalagmite which provides an absolute chronology, based largely on uranium-series dates. The relative simplicity of the dip-orientated section of the river passage, apparently lacking any tributaries or distributaries, contrasts with the converging passages in an analogous position behind Wookey Hole.

PRIDDY CAVES

Highlights

The Priddy caves represent excellent examples of predominantly phreatic swallet cave systems which have been rejuvenated, or abandoned, as a result of base-level lowering. The three main caves show evidence for significant differences in the duration of the initial phreatic phase and their ensuing vadose histories, despite all draining to the same resurgence. They provide a striking contrast with the dominantly vadose swallet caves of the Charterhouse area.

Introduction

The caves lie under the limestone plateau on the south and south-west slopes of North Hill, around the village of Priddy (Figure 5.1). Swildon's Hole, St Cuthbert's Swallet, Eastwater Cavern and Hunter's Hole are all major influent cave systems, though the latter two are now largely abandoned. Allogenic streams flowing from the Old Red Sandstone outcrop of the North Hill pericline cross the Lower Limestone Shales to sink near the base of the Black Rock Limestone, which dips south at 20–40°. All of the water draining through these caves resurges at Wookey Hole (Figure 5.1). The cave systems are formed mostly downdip and their accessible portions are developed largely within the Black Rock Limestone, locally about 280 m thick. The limestone in this area is broken by two important faults and several minor ones. The Priddy Fault runs roughly east–west across the northern part of the site, passing through the middle of the Swildon's Hole system where a brecciated zone up to 8 m wide is developed. Towards the eastern side of the site a smaller NNE–SSW fault lies very close to the Eastwater Cavern system.

The caves around Priddy have been intensively studied. Various aspects of cave geomorphology have been discussed by Drew (1975b), Ford (1963, 1965a, 1968), Newson (1969) and Irwin (1991). The hydrology has been discussed by Atkinson (1968b), Atkinson *et al.*, (1967), Drew (1975a) and Stenner (1968, 1978). Descriptions of the caves are given in Barrington and Stanton (1977), Irwin and Jarratt (1992) and Irwin (1991).

Description

Swildon's Hole is the most extensive of the caves at Priddy, and the streamway passes directly beneath the village (Figure 5.1). It has 9100 m of mapped passages, forming a complex dendritic system with many crossing links provided by high-level galleries (Figure 5.12). The main streamway takes a course westwards from the entrance to beyond Sump 1, and then turns south along the western margin of the system. Both legs of the streamway are oblique to the south-west dip. The first portion of the cave, as far as Sump 1, descends fairly steeply in a large vadose canyon (Figure 5.13). One section near the entrance was formerly filled almost entirely with clastic sediment, creating the 40 Foot Pot, but this was scoured out by the catastrophic floods of July 1968 (Hanwell and Newson, 1970). Deep rounded potholes, excavated by both solutional and mechanical action, are a notable feature of the steeper sections (Ford 1965a). Beyond Sump 1 the gradient of the cave is much lower (Figure 5.15), and the stream meanders over a floor of clastic sediment fill in a vadose canyon entrenched in the floor of a phreatic passage.

Along the course of the Swildon's streamway there are 12 flooded sections of passage where phreatic loops have been created by the obliquely downdip and up-joint route of the stream. Isolated sections of vadose canyon along the streamway, and elsewhere in the system, have formed by entrenchment through the crests of these phreatic loops, while the troughs have been infilled by clastic sediment, their ceilings migrating upwards by paragenesis. Sump 12, the present limit of explo-

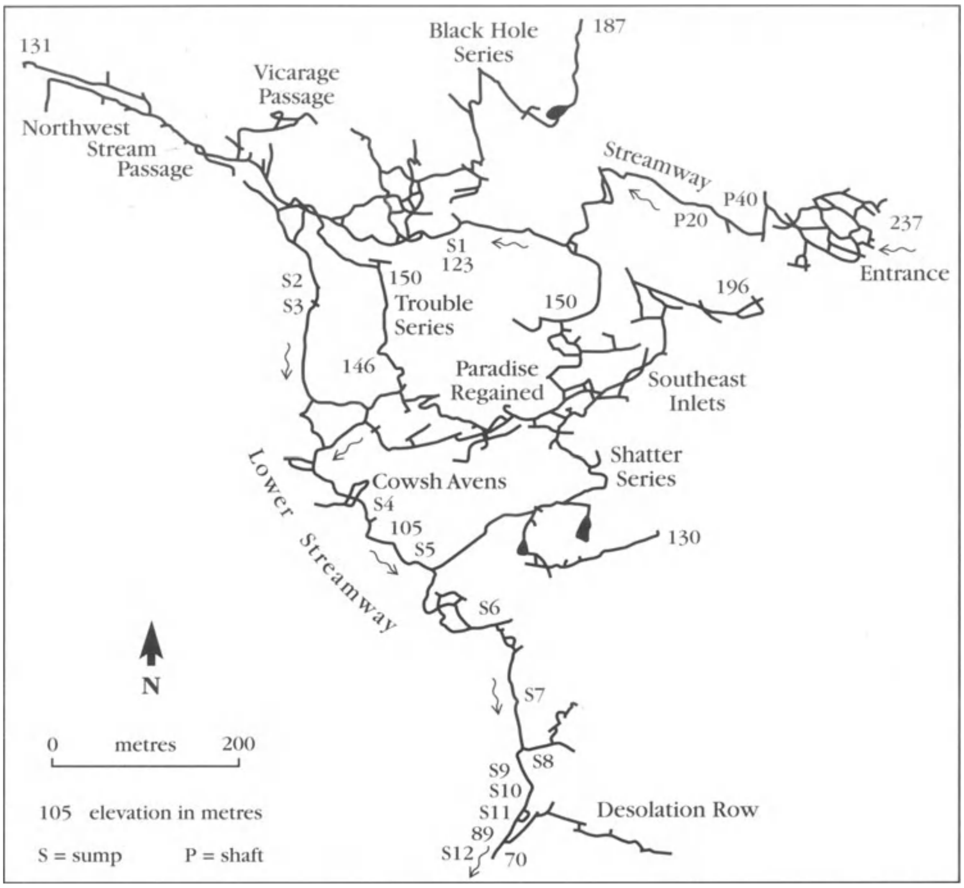


Figure 5.12 Outline map of Swildon's Hole (from survey by Wessex Cave Club).

Figure 5.13 The cascading streamway in thinly bedded Black Rock Limestone in Swildon's Hole. (Photo: J.R. Wooldridge.)

ration, has been dived to a depth of 20 m, 167 m below the entrance. North of the upper limb of the main streamway lie the old rejuvenated inlets of Vicarage Passage and Black Hole Series, while in the area enclosed by the two limbs of the main streamway lies a complex series of abandoned passages at several levels. Some sections are blocked by collapse, but elsewhere they contain extensive clastic sediment deposits and are locally well decorated with speleothems. The Priddy Fault cuts across the cave, and Cowsh Avens (from Priddy Green Sink), Shatter Series and Southeast Inlets developed in its fracture zone.

St Cuthbert's Swallet lies east of Priddy (Figure 5.1), and is the most complex system on Mendip. It contains 7100 m of mapped passages, largely developed over a minor anticline plunging SSE (Figure 5.14). From the entrance, the streamway descends steeply for more than 100 m, beneath a multi-level series of inclined bedding-plane mazes of abandoned phreatic passages and chambers (Irwin, 1991). A roughly linear series of chambers and passages defines the south-western margin of the main part of the system and is developed along a minor fault, the Gour-Lake Fault (Figure

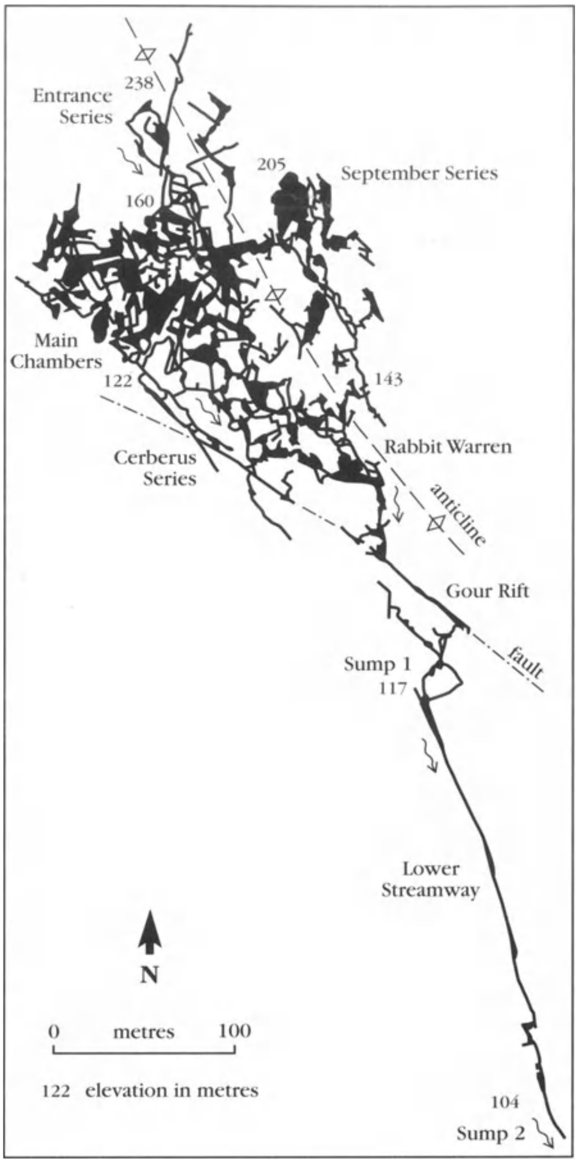


Figure 5.14 Outline map of St Cuthbert's Swallet (from survey by Bristol Exploration Club).

5.14). Caves cross this fault in at least two places. At the eastern crossing, the streamway continues as a gently sloping passage to Sump 1, which is perched. Beyond lies a further 300 m of tall, joint-guided, vadose canyon, partly entrenched beneath a gently ascending phreatic tube, leading to Sump 2. The abandoned phreatic passages and chambers have been extensively modified by collapse. They also contain abundant clastic sediment deposits and exceptionally fine speleothems of many types; these include some notable calcite curtains formed on the overhanging walls and some clusters of cave pearls in shallow pools beneath high shafts. Within one sediment sequence at least nine successive cycles of clastic sedimentation, vadose erosion and stalagmite deposition have been recognized.

Eastwater Cavern lies 400 m west of St Cuthbert's, and contains nearly 2500 m of explored passages reaching to a depth of 180 m. The upper part of the cave is mainly a steeply inclined phreatic maze formed on a bedding plane; there is only minor vadose trenching, as the cave has a very small catchment in its modern phase of development. Series of vertical vadose shafts, developed on fractures associated with the nearby fault, drop to lower levels with short sections of streamways; most passages end in small choked rifts. The cave system is largely abandoned but lacks significant speleothem development.

Hunter's Hole lies 600 m south-east of St Cuthbert's, and contains less than 300 m of passage. It drains a closed depression and has no allogenic stream input from North Hill. A vadose shaft 20 m deep drops directly into a remnant of large phreatic passage descending steeply to the south-east but largely choked with sediment and collapse debris.

Interpretation

The caves of Priddy are classic examples of predominantly phreatic cave development in steeply dipping limestone (Ford 1965b, 1968). Hence they complement the largely vadose cave systems in the Charterhouse area. Swildon's Hole is a fine example of underground dendritic drainage and has been cited as the type example of a shallow phreatic, influent cave system. This contrasts with St Cuthbert's Swallet which is a deep phreatic influent system (Ford 1965b, 1968; Irwin, 1991). In both caves, phreatic loops have developed by the stream flowing downdip and then rising up joints or faults. This is most clearly seen in the alternation of open vadose passage with short flooded sections in the Swildon's Hole main streamway. Rejuvenation of the streamway has led to vadose entrenchment into the crests of the phreatic loops, while the troughs have acted as sediment traps and have become infilled. Hence, not all of the loops are as clearly defined as those in the active phreatic sections of Wookey Hole and Gough's Cave, Cheddar.

In St Cuthbert's Swallet only a single ancient phreatic loop has been identified, but this extends to a depth of more than 80 m along the dipping shale-limestone boundary before rising obliquely up a fault. The presence of a minor plunging anticline has had a strong influence on the initial development of the St Cuthbert's cave system. In the first phase of development, passages were formed along the western flank of this anticline but, as swallets opened further upstream, the water sank and flowed along and down the eastern flank of the anticline. Later these networks coalesced to become a single system. In both this cave and Eastwater Cavern, complex inclined phreatic mazes have developed on bedding planes. Joints are limited in extent and appear to have played a relatively minor role in cave inception in the Priddy area, other than in the rising component of phreatic loops. Faults have exerted a significant influence in the development of rifts and vadose shafts, particularly in Eastwater Cavern.

The complex sequence of passages in the three main systems clearly reflects a long and complex history which probably extends well back into the Pleistocene and must relate closely to the history of the resurgence at Wookey Hole. Drew (1975b) and Ford (1965c) have identified a sequence of events for cave development in this area on the basis of morphological criteria but no speleothem dates are yet available to test their tentative hypotheses. Hunter's Hole was a major depression drain early in the Pleistocene (Drew, 1975b; Smith, 1977), though how it relates to the three swallet caves is unclear.

In Swildon's Hole, Ford (1963, 1965c) identified three main stages of development which he interpreted as responses to successively lower water tables, which then remained static for some time before dropping rapidly to the next level. Although Drew (1975b) questions the concept of a single water table in a limestone aquifer, he does accept that Swildon's experienced at least two or three rest levels during its development. The oldest section of passage appears to be the Shatter Series (Figure 5.12), draining south-west along the fault from a sink 150 m south of the present entrance. With the lowering of base level, the stream sink moved close to its present position to enlarge the second phase of passages, with flow along St Paul's Series to be joined by water from the Black Hole Series, flowing along Trouble Series and Paradise Regained. Further lowering of base level produced the system as seen today by development of new streamway passages (the third phase), vadose modification, capture of strike drainage by dip tubes and rifts, and rejuvenation of existing passages. The cave therefore developed while the local water table fell in at least two stages from an altitude of about 183 to 100 m; details of the cave morphology may allow these stages to be divided into a total of four periods of rapid water table decline (Ford, 1963, 1965c).

In contrast, St Cuthbert's Swallet lacks any clear expression of these episodes of base-level lowering and instead shows evidence for prolonged solutional enlargement in the phreatic zone. Ford (1968) attributed the persistence of this deep phreatic loop to ponding behind an aquiclude at the Wookey resurgence; this suggests that St Cuthbert's Swallet is considerably older than Swildon's Hole, where evidence of uninterrupted phreatic development is not apparent. Eventual breaching of this aquiclude led to rapid draining of the St Cuthbert's phreas, when the local water table fell from an elevation around 200 m to one at about 117 m. The hypothesized aquiclude may be the Ebbor Thrust, in which a thin wedge of Upper Carboniferous sandstones and shales has been preserved, up to an altitude of 190 m, beneath an overthrust block of the Lower Carboniferous limestone, exposed west of Ebbor Gorge (Figure 5.1). An alternative is that the St Cuthbert's passages were merely ponded behind a large phreatic lift, which was formed on one of the widely spaced major fractures in the dipping limestone; the cave was drained when a new route was opened beneath the phreatic uploop. This concept is more applicable to a karst aquifer whose high transmissivity is so dependent on conduit flow. The eventual rejuvenation of the cave resulted in vadose entrenchment of the present streamway canyon through the phreatic network as well as forming some high-level inlet passages. At least nine episodes of vadose erosion, clastic sedimentation and stalagmite deposition have been identified (D.C. Ford, 1964).

In Eastwater Cavern there appears to have been only minor vadose trenching, following draining of the phreas. It appears that, unlike the other systems, lowering of base level caused the main routes to be abandoned in favour of swallets further up the valley. The drainage from these now flows largely beneath the explored passages of Eastwater Cavern.

The individual histories of each of these three systems appear to differ considerably. Hydrological investigations by Atkinson *et al.*, (1967) suggested that the paths of the streams from the three main systems are discrete for the greater part of their lengths, uniting only a short distance behind the resurgence. This view has since been challenged by Irwin (1991) but appears to account for some of the differences between the adjacent cave systems. Nonetheless, since all now drain to a common resurgence it might be anticipated that they should share at least some features related to the evolution of the resurgence system. This notion is destroyed by recognizing that the main drainage route from each swallet was a discrete conduit looping through the phreas by following inclined bedding planes and fractures. The evolution of the separate cave systems at the swallet end of each conduit was therefore dependent on the bypassing or incision of the phreatic uploops, and subsequent drainage of the passages upstream of the loop crests. The pattern of loops is dependent on the local geology, and each conduit therefore has its own particular initial profile. This allows for the deep phreatic development of St Cuthbert's, at the same time as Swildon's cave was progressively drained when a sequence of shallow phreatic loops were successively breached. Eastwater appears to have only one drainage phase, comparable to St Cuthbert's, which is its nearest neighbour.

In the absence of dates for any of these events, it is impossible to correlate phases of development identified in one cave with those in another. However, all the Priddy caves contain abundant clastic sediments which are commonly interbedded with stalagmite layers. These present an ideal opportunity not only to correlate events between caves, and so investigate the relationship between the development of different sinkhole systems and their common resurgence, but also to document the climatic history of the area through the Pleistocene.

Conclusion

The Priddy site contains a series of sinkhole caves which show varying degrees of development in ponded phreatic conditions within the steeply dipping limestones. They were subsequently rejuvenated in response to surface lowering on the Somerset Levels, and show contrasting styles of evolution into the vadose environment. Sediment deposits and speleothems within the cave provide an exceptionally valuable record of Pleistocene environmental changes, whose full elucidation awaits analysis of both the radioactive and stable isotopes within the calcite speleothems.

WOOKEY HOLE

Highlights

Wookey Hole is a large resurgence cave developed in a unique geological situation, passing from the Carboniferous limestone into the cemented scree of limestone debris represented by the Triassic Dolomitic Conglomerate. The upstream reaches of the cave system display classic examples of deep phreatic circulation in a dipping aquifer, with successive passage levels developed in response to downward migration of the resurgence.

Introduction

The cave of Wookey Hole, located just north of the village of Wookey Hole, is operated in part as a show cave. It is a major resurgence lying on the southern margin of the Mendip limestone plateau (Figure 5.1) with a mean flow only exceeded by that of the Chedder Rising. It is the outlet for allogenic water draining off the North Hill sandstone inlier into the swallet caves of Swildon's Hole and St Cuthbert's Swallet, as well as much of the remaining subterranean drainage derived from the southern flanks of North Hill and Pen Hill. The Ebbor Thrust extends north-west-south-east only a short distance south of the mouth of Wookey Hole and Ebbor Gorge, and has preserved a narrow slice of Upper Carboniferous sandstones and shales between two masses of limestone. West of the Ebbor Gorge this potential aquiclude extends to an altitude of up to 190 m but to the east, near Wookey Hole resurgence itself, it has been breached by a Triassic valley. The show cave is developed entirely within this ancient ravine which is filled up to 100 m of Dolomitic Conglomerate, a poorly sorted Triassic breccia of limestone fragments in a calcareous silt matrix. The upstream portion of the cave is developed largely in Carboniferous limestone which has a south-west dip of 10–15°. The Dolomitic Conglomerate is crudely bedded and is crossed by a number of fractures aligned north-west-south-east.

Wookey Hole has an extensive literature covering aspects of cave development (Drew 1975b;

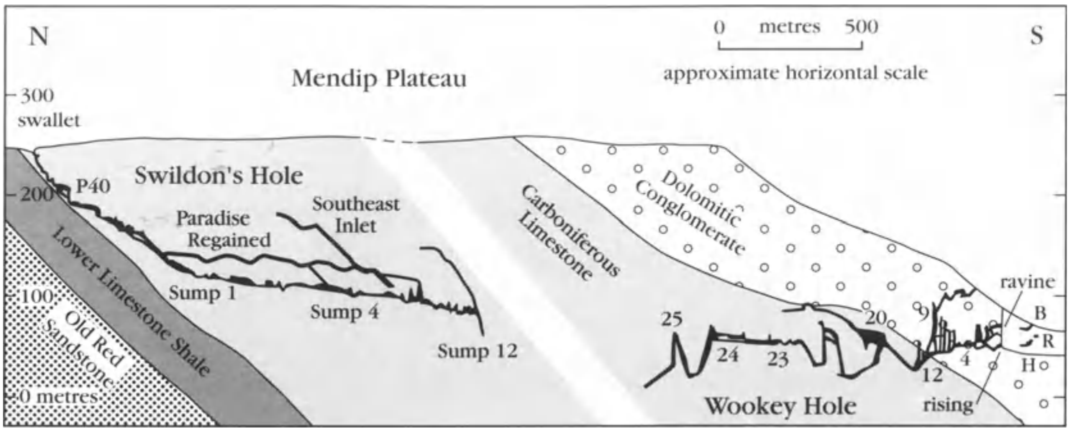


Figure 5.15 Semi-extended profile through the cave system from Swildon's Hole to Wookey Hole. The gap in the middle has not yet been reached by underground explorations; the distance between the explored limits of the two caves is about 2.3 km, and the vertical scale is exaggerated by five. The small caves in the ravine are keyed as: B = Badger Hole; R = Rhinoceros Hole; H = Hyaena Den (after drawings by W.I. Stanton).

Ford 1965b, 1968; Donovan 1988) and hydrology (Atkinson 1978; Atkinson *et al.*, 1967). Macfadyen (1970), Gatacre *et al.* (1980) and Duff *et al.* (1985) provide general accounts, and Barrington and Stanton (1977) and Irwin and Jarratt (1992) describe the cave passages.

Description

The present Wookey Hole resurgence is located near the base of the Dolomitic Conglomerate at the head of a short gorge, created by headward retreat of the cliff face over the active and abandoned cave exits. The lower part of the streamway, as far as Wookey 12, is developed entirely in Dolomitic Conglomerate (Figure 5.15). Largely flooded passages, typically 5 m across, link low bedding chambers and tall, narrow rift chambers developed along vertical fractures. From the roofs of some of these outer chambers, old outlet passages extend to the surface. Between the resurgence and Wookey 4 the modern streamway is almost level, but beyond this it descends in a major loop, emerging from the Carboniferous limestone at Wookey 12.

The inner streamway continues in a series of deep phreatic loops linking chambers which have been formed by vadose incision and modification of the loop crests. A larger section of vadose passage up to 15 m high, 6 m wide and more than 120 m long is developed at Wookey 23. The furthest point yet reached in Sump 25 is a gravel constriction 60 m below water level. Above the streamway at Wookey 20 an inclined rift, with many speleothems, ascends over a distance of 600 m to enter a boulder collapse at the junction of the Carboniferous limestone with the Dolomitic Conglomerate, beyond which the passage divides. One branch continues to ascend through the Dolomitic Conglomerate while the other re-enters the limestone before ending in a choke. An abandoned phreatic lift forms a truncated passage linking Wookey 9 to the hillside (Figure 5.15).

Three small cave remnants survive in the Dolomitic Conglomerate on the east side of the ravine below Wookey Hole cave. Badger Hole is the largest of these, with a 13 m wide entrance and almost 60 m of excavated passages. Close by lies the Hyaena Den containing some 45 m of passages. Rhinoceros Hole is another small fragment of phreatic passage 13 m long. All of these sites contain rich mammalian faunas, including Devensian mammoth, reindeer and hyaena, and Ipswichian rhinoceros and hippopotamus, together with Middle and Upper Palaeolithic human artefacts. These indicate that sediment deposition commenced at least 100 000 years ago (Donovan, 1988; Tratman *et al.*, 1971; Tratman, 1975).

Interpretation

Wookey Hole is the only large cave in Britain developed in both steeply dipping Carboniferous limestone and Triassic Dolomitic Conglomerate, a well-cemented fossil scree. The different influences which these two rock types have exerted on cave development is clearly seen in the contrasting passage morphologies between the outer part of the system, developed in Dolomitic Conglomerate, and the inner part developed in Carboniferous limestone. In the Dolomitic Conglomerate, the streamway flows through a series of shallow loops linking low, bedding plane chambers, or through tall, narrow rifts developed by solutional enlargement of vertical fractures under phreatic conditions. In the Carboniferous limestone, the cave forms phreatic loops over 60 m deep, with the stream flowing downdip along bedding planes, before rising through rifts on the joints. The cave represents the finest example in Britain of deep phreatic development in steeply dipping limestones, and also shows excellent examples of vadose incision through the loop crests.

Ford (1965b, 1968) considered that much of the phreatic character of the cave, and of the swallet caves at Priddy, developed through ponding behind a major aquiclude, perhaps of sandstone east of the Ebbor Thrust or of Triassic Mercia Mudstone, producing a considerable hydrostatic head. However, the Ebbor Thrust was already breached by Triassic times, while the resurgence stream would have rapidly incised into the soft Mercia Mudstone, preventing the development of a perched phreas for any length of time. The ascending rifts above Wookey 20 and 9 may represent feeders to relict vauclusian risings in the flank of the Mendip Hills. The high-level passages from the outer chambers may represent a series of distributary passages which developed, at successively lower levels, of approximately 80, 72 and 65 m down to the present water table at 60 m. These levels can be correlated with a sequence of altitudes and episodes of vadose incision on the crests of the phreatic loops. The successive lowering of the phreas overflow was controlled by the resurgence positions which developed in response to removal of the aquiclude confining the limestone to the south; ultimately this was a function of surface lowering of the plains to the south of the Mendip Hills during the Pleistocene (Macklin, 1985). The clastic and speleothem deposits within the cave offer the prospect of establishing an absolute chronology for this sequence of events, which can then be used in reconstructing the geomorphological evolution of the landscape in this area.

Conclusion

Wookey Hole is a major resurgence cave with the finest example of deep phreatic cave development in Britain. It is unique in being developed in both the Carboniferous Limestone and in the Triassic Dolomitic Conglomerate, and therefore demonstrates the different controls on karst drainage within these two important aquifers. The deep phreatic loops, controlled by the bedding and joints, include active and abandoned conduits in a configuration more complex than in the river cave at Cheddar.

BRIMBLE PIT AND CROSS SWALLET

Highlights

Brimble Pit and Cross Swallet are two of the finest closed drainage basins on Mendip, and together exhibit all the geomorphological features characteristic of Mendip closed basins. Both basins provide evidence of the periglacial development of lakes and overflow channels on the Mendip plateau during the last glaciation.

Introduction

A belt of twelve drainage basins extends along the southern rim of the Mendip plateau from Cheddar Gorge to Ebbor Gorge; they constitute a zone of polygonal karst (Figure 5.16). The Brimble Pit basin is one of the largest of the chain, while the adjacent Cross Swallet basin is smaller, but has a very distinctive internal morphology (Ford and Stanton, 1968). Both depressions once contained lakes which drained via a low col into associated overflow channels. The geomorphic significance of the closed basins was recognized by Ford and Stanton (1968), further elaborated on by Barrington and Stanton (1977), and briefly described by Duff *et al.* (1985).

Description

Brimble Pit is a pool at the lowest point of a shallow depression 10 m deep, over 1000 m long and 500 m wide (Figure 5.16). The floor of the basin is covered in a thick layer of horizontally stratified loessic silty clay, pitted with small sinkholes, one

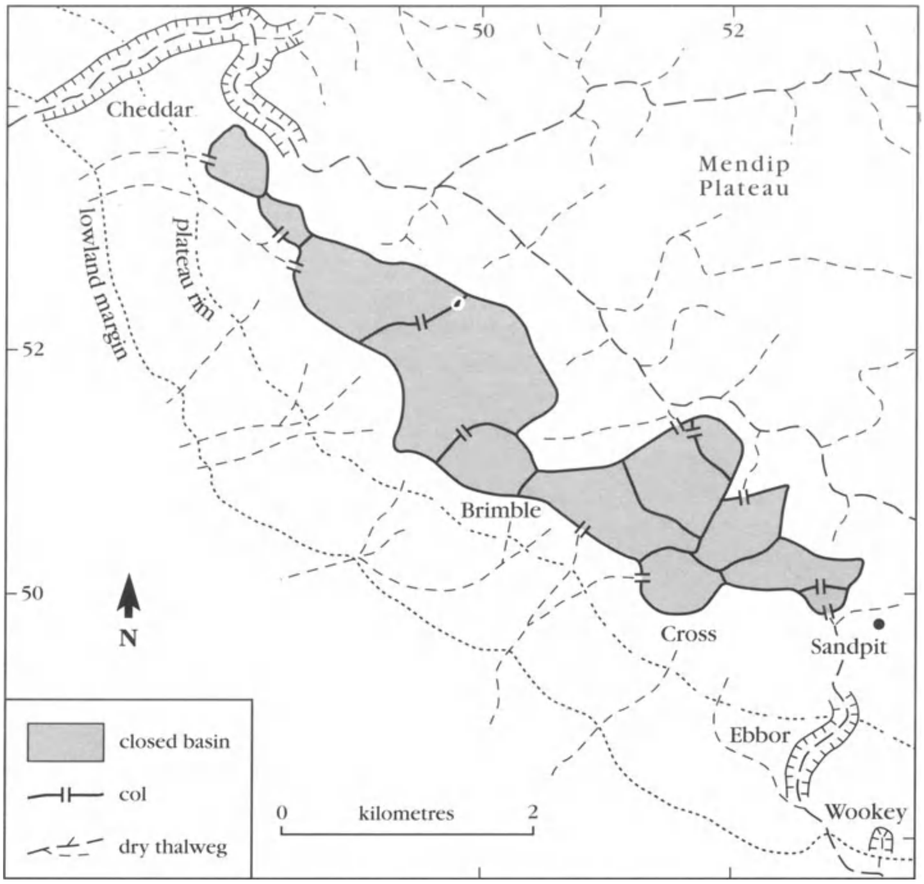


Figure 5.16 Topographic map of the group of closed depressions forming the zone of polygonal karst on the edge of the Mendip Plateau (after Ford and Stanton, 1968).

of which contains Brimble Pit. The pool is artificial; it originated as a sinkhole whose sides were puddled with silty clay to provide drinking water for cattle, and is now fed by road drainage. The basin is bounded by very gently graded slopes, dividing it from neighbouring depressions and valleys. The south-eastern margin of the basin is marked by a low col which feeds into an overflow channel incised several metres into the flanks of the plateau. Brimble Pit Swallet is a cave excavated to a depth of 20 m beneath one of the sinkholes in the basin; it is developed along the line of a major fault zone occupied by vein calcite, and is infilled with Triassic and Jurassic neptunian dyke sediments. Water draining into this swallet from an adjacent reservoir has been traced to Rodney Stoke rising. Locke's Hole is another excavated sinkhole, which yielded siliceous gravels similar to those seen in the Westbury Quarry deposits a few hundred metres to the south.

The Cross Swallet basin is similar in depth, but is only 500 m in diameter (Figure 5.16). It also has a marginal col and overflow channel, but not as well defined as that at Brimble Pit. A clearly defined corrosion terrace has formed at the level of the col, and extends all the way around the basin, locally extending to 23 m in width. At one point, an undercut limestone bluff rises above it. The main basin floor is formed on an infill of horizontally laminated yellow-brown silty clay at a level 5 m below the edge of the terrace. The clays are over 7 m thick, and within them a closed depression is cut 8 m deep at the centre of the basin. Fissures in the limestone floor of this have been penetrated for about 10 m depth before they become impassably narrow (Figure 5.17).

Interpretation

Ford and Stanton (1968) argued that the basins were initially formed by solutional activity during warm phases of the Pleistocene, and the sinks were blocked by permafrost during the ensuing cold periods. Meltwater became ponded during the brief summers until it spilled over the cols to

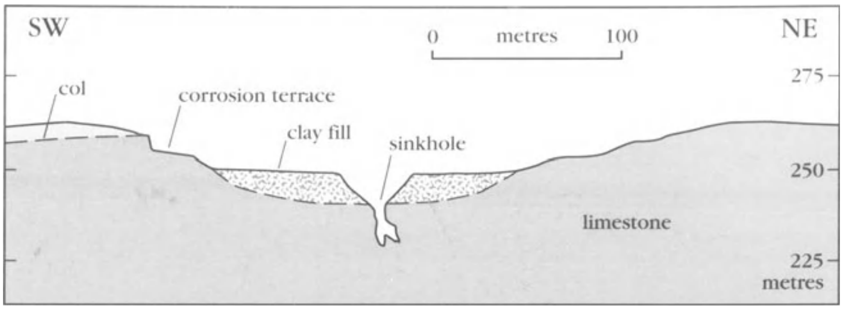


Figure 5.17 Cross-section through the depression and sinkhole of Cross Swallet (after Ford and Stanton, 1968).

cut the overflow channels. Hillwash, and perhaps windblown silt, formed the loessic silty clay deposited in the lakes, and helped seal the lake beds. In the Cross Swallet basin, the presence of the terrace indicates a stable lake surface at the col level. It is suggested that hydrostatic pressure was great enough to maintain slow talik leakage through the clay and the underlying permafrost into the limestone beneath.

The two basins combine to show all the features associated with Mendip closed depressions. These include cols leading to overflow channels, a terrace at col level etched into the limestone, old lake deposits forming flat thick clay floors, subsidence sinkholes developed in the clay fills, and impenetrable or choked caves developed below the sinkholes. Further work on the sedimentology and palynology of the loessic clay could provide important evidence on the palaeoenvironment in which the lakes were formed.

Conclusions

The site covers two of the finest major closed basins on Mendip, in an area of polygonal karst with no intervening valleys. Both basins show evidence of solutional excavation, followed by the periglacial development of lakes and overflow channels, and a return to underground drainage during interglacials.

SANDPIT HOLE AND BISHOP'S LOT

Highlights

Sandpit Hole and Bishop's Lot are two of Mendip's largest dolines. Sandpit Hole is a good example of a typical deep rocky doline, while Bishop's Lot is a broad, shallow, saucer-shaped doline.

Introduction

These two separate dolines are located about 4 km north of Wells and are developed on the flat surface of the Mendip plateau (Figure 5.1 and 5.15). They are of particular interest because they constitute two of the largest unfilled dolines on Mendip, and have distinctly contrasting morphologies. Both are developed on the Carboniferous Limestone and have been cited as evidence for the various theories which have been proposed for the formation of Mendip dolines (Stride, A.H. and Stride, R.D., 1949; Balchin and Coleman, 1959; Ford and Stanton, 1968), a resume of which is given by Smith (1975a). Both have been dug by cavers at some point, and the details of the digs at both sites are described in Barrington and Stanton (1977).

Description

Sandpit Hole is a large pit about 12 m deep and less than 50 m in diameter, with steep sides and a cliff face along one side containing several small caves. Sediment on its floor is a dolomite sand, which is largely a solutional residue left behind as a result of the weathering of granular dolomite. Below the floor, excavations by cavers show that limestone boulders continue to a depth of at least 16 m below the plateau surface. It is a fairly typical example of a Mendip doline, as yet unfilled.

Bishop's Lot is a large almost circular depression with a shallow saucer-shaped profile; it is 11 m deep and over 200 m in diameter. The margins are poorly defined and digging by Balch, around 1900, revealed a thick deposit of clay on its floor. Its morphology provides a clear contrast to that of Sandpit Hole.

Interpretation

Stride, A.H. and Stride, R.D. (1949) interpreted Sandpit Hole as being formed as a result of cavern collapse as did Coleman and Balchin (1959). More recently Ford and Stanton (1969) attributed the formation of dolines to gradual solution working down from the surface along joints, and deepened by the breakdown of the limestone at the top of the fissures. The origin of Sandpit Hole appears to be a combination of subaerial and under ground solution, undercutting and collapse of the limestone; the buried limestone boulders indicate the nature and scale of the collapse.

The nearby Whitepit closed depression, which is similar to Sandpit, has recently been excavated to reveal a cave at shallow depth. Directly below the surface depression, the cave passes through a debris pile at least 10 m across; the debris consists mainly of rounded limestone blocks, and appears to be derived largely from the rockhead zone of weathering. At Whitepit, it appears that an older open cave at a shallow depth has aided leakage of water into the limestone overlying the cave, accompanied by ravelling and partial collapse of the limestone to form the depression. It is possible that a similar mechanism can be invoked for Sandpit Hole, although more digging would be required to confirm this.

The evolution and deepening of the Bishop's Lot doline appears to have been dominated by solution rather than collapse. However, as almost nothing is known about the subsurface structure, the relative importance of subaerial solution and collapse cannot be estimated. It may represent an early form of the larger depressions, which include Brimble Pit and Cross Swallet, where premature leakage precluded any significant ponding and therefore prevented lateral expansion.

Conclusions

Two of the largest isolated dolines on Mendip have been formed by a combination of solution and collapse, and are typical of most of the depressions on the karst plateau. The two provide clearly contrasting morphologies, and represent opposite ends of the spectrum of processes and morphologies exhibited by the Mendip dolines. Sandpit is a steep-sided doline with rock walls and a floor of boulders continuing to depth, which may have been formed by collapse into an underlying cave. Bishop's Lot is a much broader, shallow depression, with a thick clay floor, developed mainly by solutional processes.

WURT PIT AND DEVIL'S PUNCH-BOWL

Highlights

Wurt Pit and Devil's Punch-Bowl are two of the most spectacular subsidence dolines in the Mendip Hills karst. They provide important evidence of the role of subsurface solution, and of leakage of water through impermeable cover rocks, in the formation of dolines on the Mendip plateau.

Introduction

These two dolines lie on the northern side of the Mendip Hills where the limestone plateau is only partly exhumed from its Mesozoic cover (Figure 5.1). Both Wurt Pit and Devil's Punch-Bowl are collapse dolines developed in the Jurassic Harptree Beds and the Triassic Mercia Mudstones, which overly the Carboniferous Limestone. In each case, rainwater is concentrated onto a series of seepage paths through the dominantly impermeable surficial rocks, into the limestone at depth, causing solution and collapse, and hence a depression. Their genesis is explained by Smith (1975a) and Barrington and Stanton (1977), and both sites are briefly described by Duff *et al.* (1985).

Description

Wurt Pit is a cup-shaped doline, 15 m deep and almost 100 m across, set in a gently sloping hillside with no associated valley features (Figure 5.18). It has a sharply defined rim and steep rocky sides. The surface rocks are the silicified limestones and mudstones of the Jurassic Harptree Beds, which are exposed on the walls of the doline. The Mercia Mudstones and Dolomitic Conglomerate are believed to underlie the site at no great depth, and they outcrop nearby. The nearest exposure of the Carboniferous Limestone is 500 m to the south-east; however, limestone is almost certainly present at depth directly beneath the doline.

Devil's Punch-Bowl is another impressive depression, over 50 m in diameter and almost

Figure 5.18 The Wurt Pit doline breaks the gently graded surface on the Harptree Beds outcrop. (Photo: A.C. Waltham.)

20 m deep. Like Wurt Pit, it is independent of the local drainage pattern, but it does have a small trench valley into it, and there is usually a small pool on its floor. Mercia Mudstones are exposed in the walls of the depression, and rotted siliceous material exposed in the trench may represent part of the Harptree beds. The Dolomitic Conglomerate and Carboniferous limestone appear in outcrops at distances of around 500 m north, west and south of the doline, and must underlie the site.

Interpretation

Wurt Pit is an excellent example of a subsidence doline developed in consolidated cover rocks. It was clearly formed by the solution of Carboniferous limestone at depth, followed by collapse and subsidence of the relatively impermeable and insoluble cover rocks (Smith, 1975a; Barrington and Stanton, 1977). The water responsible for the solution could have come from either or both of two sources; it may have been lateral flow entirely within the underlying limestone, but it was almost certainly joined by aggressive surface water leaking down through fissures in the Jurassic cover rocks. If lateral groundwater flow within the Carboniferous limestone was the dominant agent, then Wurt Pit may be more correctly described as a collapse doline.

Devil's Punch-Bowl has an origin which is broadly similar to that of Wurt Pit, except that the Mercia Mudstones in which it lies are almost completely impermeable. The ephemeral lake is a consequence of the very low surface permeability, and it drains only very slowly underground, where the morphology of any caves and fissures is unknown. Recent explorations have revealed cave systems beneath Wigmore (Jarratt, 1991; Hughes, 1991) and Attborough swallets, both of which are dolines comparable to the Devil's Punch-Bowl, located a few kilometres to the east. These have stream caves developed in the carbonates of the Carboniferous limestone and the Dolomitic Conglomerate, with active tributary passages and chambers formed wholly within the Mercia Mudstone. These mudstone caves appear to have developed as piping failures, enlarging progressively headwards, but the seepage flow which causes the piping erosion may have been initiated along calcareous horizons within the Mercia Mudstone. The same processes may be, or may have been, active beneath the Devil's Punch-Bowl doline.

Conclusions

The site covers two of Mendip's largest collapse dolines, and both are excellent examples of subsurface solution creating surface depressions, aided by leakage and piping through the surficial rocks, irrespective of surface morphology.

LAMB LEER CAVERN

Highlights

Lamb Leer Cavern is a fragment of a formerly phreatic cave system whose position, remote from the present catchments, is strikingly anomalous in the overall pattern of Mendip caves. It contains one of the largest chambers in Mendip and rare stratified aragonite flowstone. The area above the cave was the site for one of the earliest attempts to locate a cave by geophysical methods.

Introduction

Lamb Leer Cavern is located near the northern edge of the Mendip limestone plateau, 2 km south of Compton Martin (Figure 5.1). It is remote from any significant modern swallets and also from the anticlinal core of Old Red Sandstone, which provides most of the allogenic input to the present cave systems in Mendip, and lies more than 100 m above the adjacent lowlands. The cave is developed in fine-grained, chinastone facies of the upper part of the Clifton Down Limestone, dipping 15° east, and the east–west Lamb Leer Fault passes through it.

Little has been published on Lamb Leer. Descriptions of the passages are given by Barrington and Stanton (1977) and Irwin and Jarratt (1992), and its geomorphology is mentioned only briefly by Smith (1975a) and Stanton (1983).

Description

The cave is accessible through a mined shaft which intersects natural passage at a depth of 20 m. Downslope to the north, 100 m of passage passes through Beehive Chamber, with a 4 m high stalagmite boss at its centre, and continues beneath a 300 mm thick aragonitic flowstone floor to enter the east side of the Great Chamber, 20 m above its floor. The chamber is over 35 m high and 20 m in diameter. From it, a partly mined rift runs west for 60 m to the Cave of Falling Waters, where a small stream sinks in the floor and drains to Rickford Rising (Figure 5.2) (Barrington and Stanton, 1977). North from the chamber is St Valentine's Series, a complex of small phreatic tubes with some larger rifts and chambers, in places well decorated with speleothems. Extensive clastic sediment deposits and calcite and aragonite flowstones are preserved in several parts of the cave.

In 1938, L.S.Palmer undertook a resistivity survey of the area above Lamb Leer, one of the earliest attempts to locate caves by geophysical methods. He found an anomaly over the known cave and also a second anomaly suggesting another large cavity lies 130 m NNW off the Main Chamber (Barrington and Stanton, 1977). Palmer's Chamber, as it is known, remains unverified.

Interpretation

The position of Lamb Leer is anomalous among Mendip caves in its great distance from any present source of allogenic input. All of the cave passages are phreatic in origin, developed below their contemporary water table, yet the adjacent lowlands are now more than 100 m below the level of the cave. The Lamb Leer Fault may have influenced drainage routes and cave development. Large, isolated, phreatic chambers are known in various Mendip caves, but the Lamb Lear chamber is uncommonly large in relationship to its associated passages. The cave's distance from the main catchments on the Old Red Sandstone (Figure 5.1) suggests that it may represent the middle reaches of a system, formerly fed by sinks and vadose inlets much closer to the stratigraphic base of the limestone, whose upstream extension has been destroyed by surface lowering of the limestone plateau. The comparable middle reaches of the active Mendip caves of the Priddy–Wookey system (Figure 5.15) and the Charterhouse–Cheddar system remain inaccessible. Only the caves of the smaller St Dunstan's Well catchment can be explored over most of their length (Figure 5.20). Hence Lamb Leer may provide further information on this part of the anatomy of a Mendip karst drainage system.

Alternatively, Lamb Leer may have been fed by sinks developed on a formerly more extensive cover of Mesozoic rocks. Either scenario implies a considerable age for the system, perhaps extending back more than a million years to a time soon after the exhumation of the plateau from beneath the cover of Mesozoic strata. Investigation of the sediments and speleothems within the cave, including the aragonite flowstones, may confirm this, or at least establish a minimum age and sequence of development for the system. Such information would be extremely valuable in interpreting the geomorphological evolution of this area during the Pleistocene and earlier.

Conclusion

Lamb Leer Cavern is a fragment of an ancient phreatic system now isolated from present catchments as a result of surface lowering. It appears to be a relic from Tertiary drainage patterns, related to a higher plateau surface or a more extensive Mesozoic cover across the Carboniferous Limestone. The large, isolated chamber and the aragonite flowstone are two unusual features which make Lamb Leer so distinctive.

THRUPE LANE SWALLET

Highlights

Thrupe Lane Swallet is the most extensive vertical cave system in Mendip, containing the deepest vadose shaft in southern England. It provides a striking contrast to the more gently inclined passages of other Mendip caves which are controlled by bedding planes and joints, and it demonstrates the overriding major influence which faults may have on cave development.

Introduction

Thrupe Lane Swallet is a major stream sink for water draining south off the Beacon Hill inlier in the eastern Mendips (Figure 5.1). The Old Red Sandstone and Lower Limestone Shales are faulted against the Black Rock Limestone to the south by the east–west Thrupe Fault. The water resurges at St Andrews Well in Wells. Descriptions of the cave are found in Barrington and Stanton (1977), Irwin and Jarratt (1992), and Meade-King (1984), but there are no geomorphological studies to date.

Description

The cave contains just over 1400 m of passages, descending to a depth of 120 m (Figure 5.19). It is entered through an excavated shaft in one of a line of dolines which engulf two streams. All the sinking water is encountered again in the cave, where it follows a complex branching route through steeply descending rifts and inclined bedding plane passages. The cave system comprises a series of both active and abandoned, sub-parallel rifts trending close to north–south, containing vertical shafts up to 60 m deep and linked by smaller inclined passages; the lowest point is a choked rift.

Interpretation

Thrupe Lane Swallet has developed by vadose invasion and enlargement of a series of rifts previously opened by phreatic solution. It is atypical of Mendip caves due to its dominantly vertical development. The sub-parallel rifts and vertical shafts have been developed within the influence of major fractures associated with the Thrupe Fault. The bedding dips at 30° south-west, and smaller

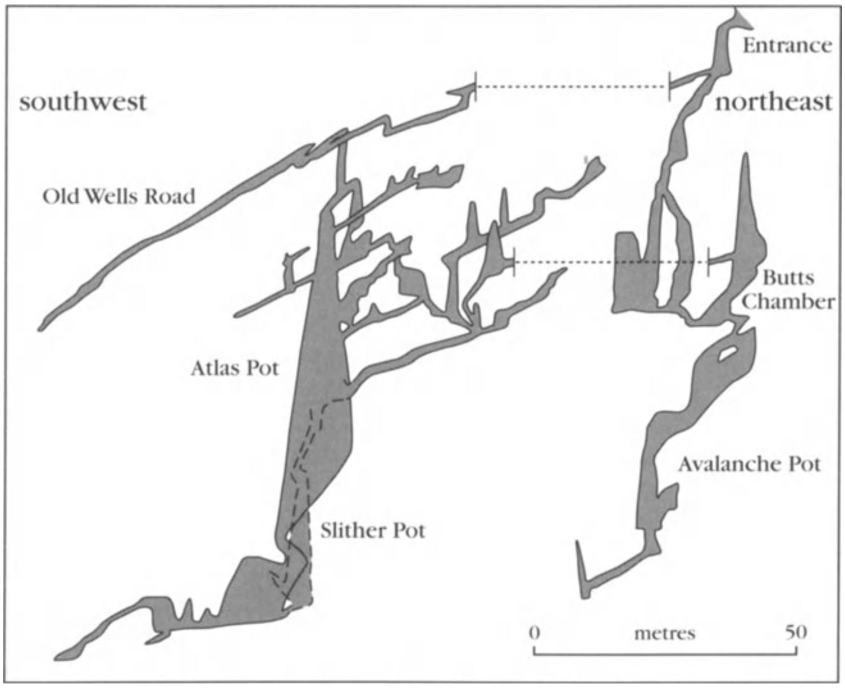


Figure 5.19 Projected profile through Thrupe Lane Swallet (from survey by Mendip Nature Research Committee).

downdip drains follow bedding-joint intersections to provide the links between the vertical rifts. At least six bedding planes have acted as inclined inception horizons, reflected in the pattern of cave development just updip of Atlas Pot (Figure 5.19). Only Reservoir Hole and Rhino Rift have comparable depth/length ratios in the Mendip karst, and Thrupe Lane Swallet is more akin to the vadose shaft systems in the Yorkshire Dales.

Conclusion

The cave is a complex vertical system of shafts and rifts developed in dipping limestone adjacent to a fault. Its vadose shafts demonstrate an unusual aspect of cave development in the dipping limestones of the Mendip Hills.

ST DUNSTAN'S WELL CATCHMENT CAVES

Highlights

The caves of the St Dunstan's Well catchment contain the most abundant, and best preserved, calcite deposits in the Mendip karst. They represent the only significant cave systems in Mendip which can be explored in almost their entirety from sink to rising. Their gently graded profiles in steeply dipping limestone contrast with the looped profiles of caves developed in more gently dipping strata on the southern flanks of the Mendips.

Introduction

The catchment covers a number of caves developed in the Carboniferous Limestone on the northern limb of the Beacon Hill pericline, all draining to St Dunstan's Well (Figure 5.1). Allogenic water drains off the Silurian basalts, Old Red Sandstone and Lower Limestone Shales of the inlier, and sinks at various points as it crosses the karst. The limestone dips north at 50–80°, though the local relief is very subdued. Consequently the caves have little vertical range, none exceeding 50 m in depth. The passages cut across the bedding into successively higher stratigraphic horizons in their course towards the resurgence of St Dunstan's Well, lying at the contact of the Carboniferous limestone with the overlying Namurian Quartzitic Sandstone. The Withybrook Fault passes through the Fairy Cave Quarry area, aligned NNW and dipping 50° west, with a small downthrow to the west; the fault has a brecciated zone, 15–20 m wide, with calcite and ferromanganese mineralization.

The cave systems of the western part of the catchment, in and around Fairy Cave Quarry, were comprehensively described by Price (1977, 1983). Passage descriptions of all of the caves are in Barrington and Stanton (1977) and Irwin and Jarratt (1992). The hydrology and water chemistry were investigated in some detail by Drew (1968, 1970), with further brief comments by Atkinson *et al.* (1973), Drew (1974) and Edwards (1994). Various aspects of the caves and their hydrology have been described and discussed in Smith (1975a).

Description

The western part of the site is centred around the now disused Fairy Cave Quarry which, during its working life, intersected the passages of two major connected cave systems (Figure 5.20) and provided the only known entrances to the caves. More than 4500 m of passages have been recorded, but 800 m of this has since been destroyed by quarrying. The remaining cave fragments opening in the quarry faces have been given separate names.

The western branch of the quarry caves is formed by Withyhill Cave and the connected Hillwithy-Hilliers-Fairy Cave System (Figure 5.20). Withyhill Cave is the largely abandoned, upstream segment of the system, with 740 m of passages. It comprises a single, gently meandering passage with two tributary elements at its upstream end, one reaching close to the sink at Withybrook Slocker. In places a phreatic half-tube is discernible in the roof above a vadose trench, but the original form of the passage is much obscured by collapse and by the profusion of calcite speleothems. These include stalactites, large stalagmite bosses, helictites, curtains, gour pools and crystal pools on a scale unmatched by any other cave on Mendip. The downstream continuation of Withyhill is the Hillwithy-Hilliers-Fairy system, with 1200 m of passages entered from Fairy Cave, an old phreatic inlet. The main part of this section is a simple meandering phreatic tube or rift, with a vadose trench discernible in places, now occupied by a misfit stream; there are a few collapse chambers.

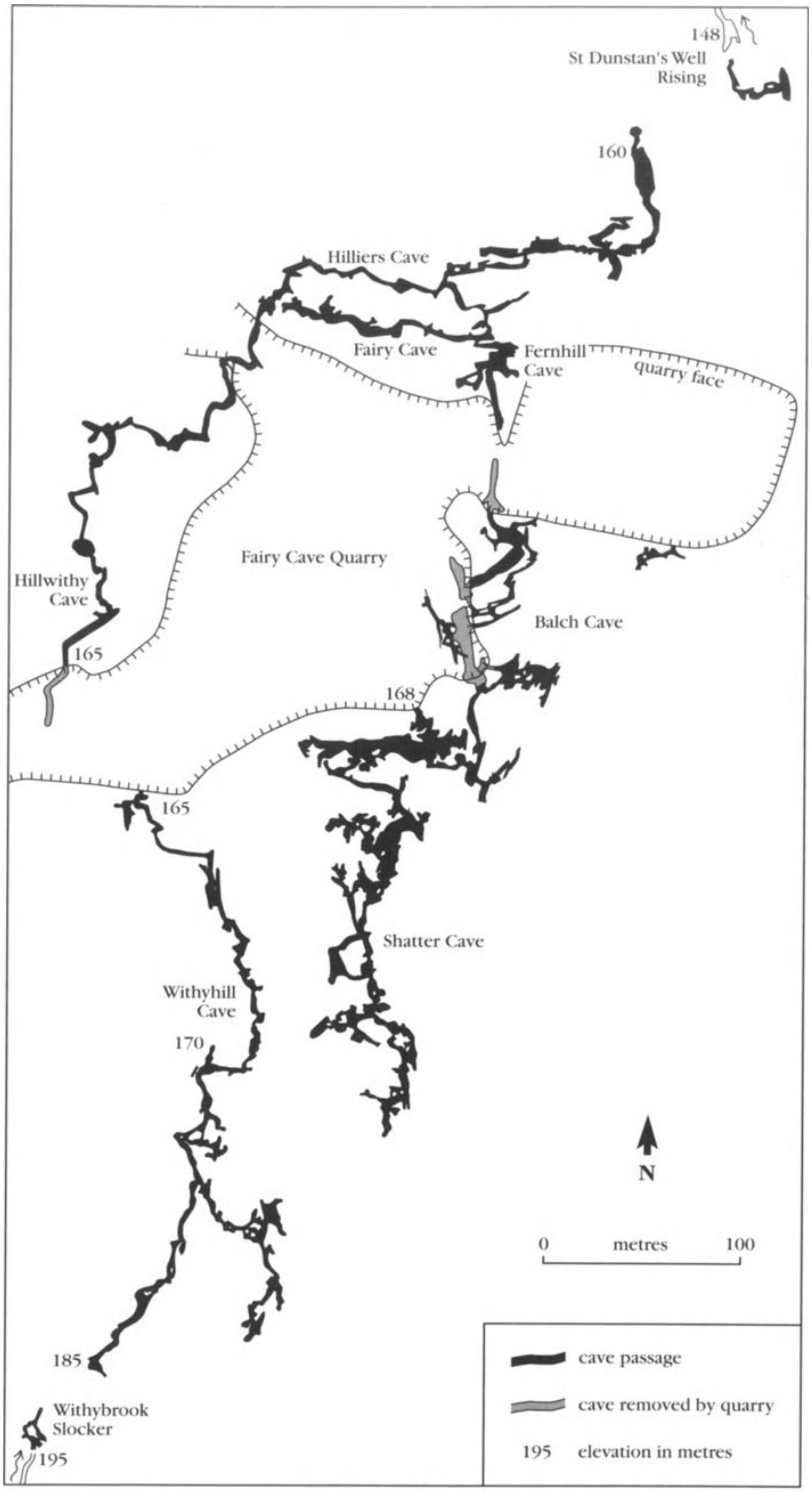


Figure 5.20 Outline map of the cave systems revealed where the Fairy Cave Quarry cut into the limestone outcrop (from survey by Cerberus Caving Club).

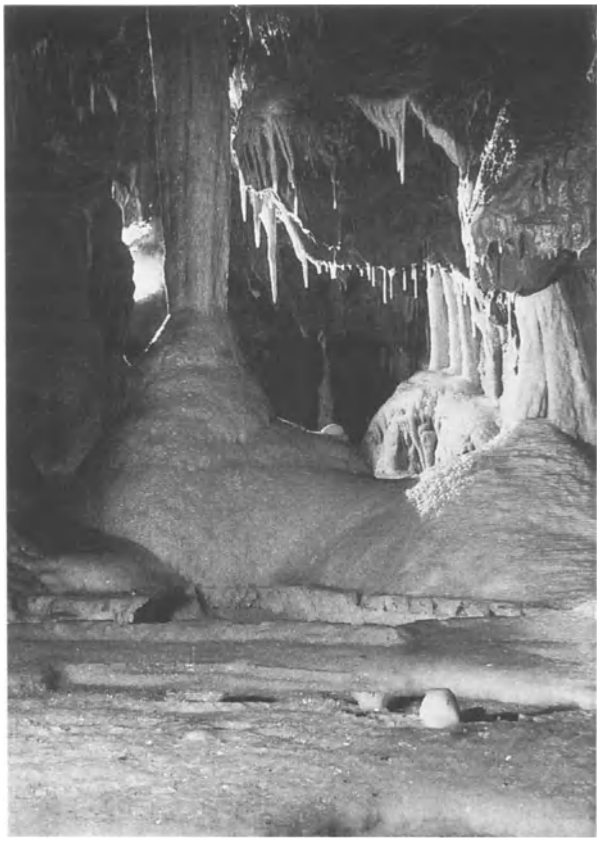


Figure 5.21 Pillar Chamber in Shatter Cave. (Photo: A.C. Waltham.)

Fairy Cave and much of Hilliers Cave are developed along the strike, and the cave is blocked in a well decorated chamber choked close to the surface, possibly at the site of a former resurgence only 70 m from the present risings.

Shatter Cave, with 1200 m of passages, is the upstream segment of the eastern branch of the main system breached by the quarry (Figure 5.20). It is mainly developed in the fractured limestone adjacent to the Withybrook Fault, and consists of a series of collapse chambers, each up to 20 m across, connected by smaller rifts and bedding plane passages. Abundant speleothems of all types, with notable crystal pools and curtains, are only marginally less impressive than those in Withyhill Cave (Figure 5.21). There is evidence of at least two phases of calcite deposition, as yet undated, with an interval of erosion or disturbance when many speleothems were broken up and re-cemented by later growth. Upstream the passage is blocked by collapse; and the downstream continuation has been largely destroyed by quarrying.

Stoke Lane Slocker is an important swallet cave in the eastern part of the catchment (Figure 5.1). It has 2200 m of mapped passages (Figure 5.22), and carries a stream which contributes 75% of the swallet input to the St Dunstan's Well risings (Drew, 1968). This stream flows to the northwest, before turning along the strike near its explored limit. Much of the streamway is a rejuvenated phreatic passage rarely more than 1 m high or wide. It follows a very gentle gradient, so that dips in the roof level have created six sections of permanently flooded passage; the end of the known cave lies 900 m east of the resurgence at St Dunstan's Well. Above the streamway, a series of small phreatic tubes and rifts link to high-level phreatic chambers, modified by collapse. Some of these chambers are decorated with an abundance and variety of speleothems, including some very large stalagmite bosses. Bone Chamber is the largest, 35 m long and 15 m wide, approaching very close to the surface; it lacks calcite deposits, and its floor is strewn with boulders and mud, containing charcoal and bones which are probably very recent (Tratman, 1975).

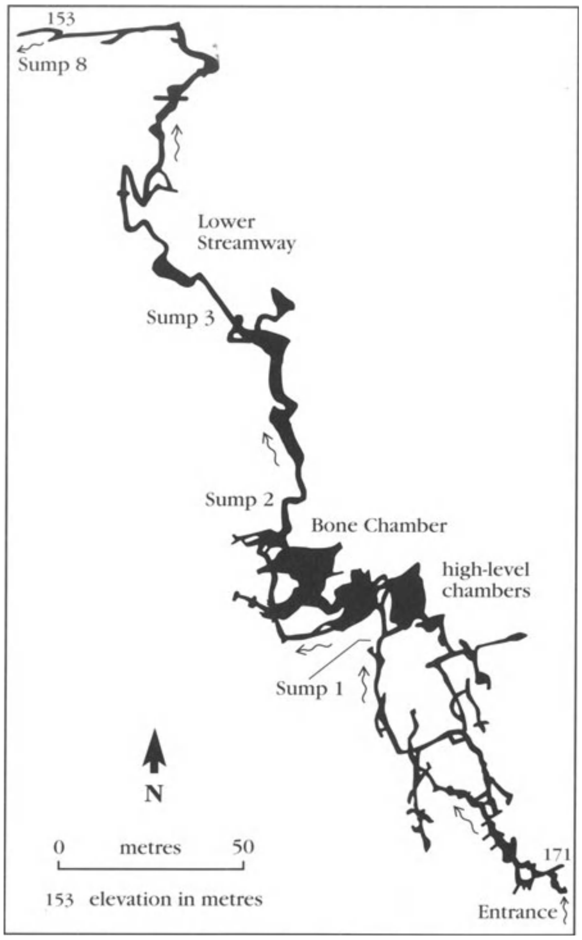


Figure 5.22 Outline map of Stoke Lane Slocker (from surveys by Wessex Cave Club and Cave Diving Group).

Interpretation

The importance of these caves lies primarily in the abundance and variety of speleothems which they contain. These are on a scale unmatched elsewhere in Mendip and equalled by only a few other sites in Britain. Investigation of these caves and comparison with other sites notably rich in calcite speleothems may well reveal information on the climatic, topographic and geological factors which influence speleothem development. The dominance of straw stalactites in Pennine caves such as Strans Gill Pot, and the prominence of much more massive stalagmites in Otter Hole at low altitude in South Wales, both stand comparison with the more mixed speleothem assemblages in these East Mendip caves. The contrasts are probably due to geographically dictated climatic and palaeoclimatic differences; more detailed and quantitative study of the speleothems and their geological environments is needed before the implications on palaeoenvironmental reconstructions can be assessed.

The profiles of these caves presents a significant contrast to those developed in the more gently dipping limestones draining the southern flanks of western Mendip. The Priddy–Wookey and Charterhouse–Cheddar systems are characterized by large phreatic loops developed down the bedding and up joints whereas, despite the much steeper dip of the limestone in the St Dunstan's Well catchment, these caves have gently graded profiles. The fracture density within the limestones has been high enough for the caves to develop on an almost graded profile, without deflection by the bedding planes into deep loops (Ford, 1971).

These caves of eastern Mendip show a sequence of development, from phreatic chambers, followed by phreatic conduits close to a graded profile, and then rejuvenation and modification by vadose erosion, with associated collapse and calcite deposition. This sequence reflects changes in karst drainage associated with landscape evolution through the Pleistocene; absolute dating of the calcite speleothems is required to recognize the time-scale involved.

Conclusion

The catchment contains the accessible fragments of three cave systems, all of which are notable for the exceptional profusion, variety and beauty of their calcite speleothems. The caves can be explored over almost their full length from sink to rising, whereas the middle reaches of most other Mendip cave systems remain inaccessible. The gently graded profiles of the passages, contrasting with other Mendip caves, are a consequence of the steep dip and close fracturing of the limestone.

Chapter 6

Karst in Wales

INTRODUCTION

Carboniferous limestones have outcrops in various parts of both North and South Wales, but most of the significant karst and cave development is located in the narrow outcrop of limestone, known as the North Crop, which fringes the northern side of the South Wales coalfield. This outcrop is over 100 km long, though seldom more than 2 km wide (Figure 6.1). Within its sinuous belt and under the adjacent grit cover, a number of limestone caves include the deepest and many of the longest systems in Britain. Though the area is not distinguished by dramatic limestone landscapes and spectacular karst landforms, it does contain a turlough and an extensive interstratal karst with some large doline fields, both features which are not well represented elsewhere in Britain.

Stratigraphically the limestones of the North Crop range through much of the Dinantian, within the Lower Carboniferous, but erosional and non-depositional breaks mean that the full sequence is not found in any one section. The full thickness of the exposed carbonates varies from 120 m to 150 m along the outcrop (Ramsbottom, 1973; George *et al.*, 1976; Wright, 1986a; Barclay *et al.*, 1988; Barclay, 1989; Lowe, 1989b). All are of shelf facies, consisting of micrites, sparites, bioclastic limestones and oolites, and some horizons are extensively dolomitized. Most are well bedded, and they contain many thin shale, mudstone, sandstone and palaeosol horizons, the latter overlying shallow zones of palaeokarstic features (Wright, 1982, 1986b).

The North Crop limestones are underlain by the Lower Dinantian Lower Limestone Shale; this rests, with only slight unconformity, on the sandstones and shales of the Devonian, which rise northwards to form the escarpments of the Brecon Beacons and Black Mountain. In some places the limestone is capped by a thin Upper Dinantian shale. Elsewhere this is cut out, and the strong, coarse-grained Basal Grit, of Namurian age, rests directly on the limestone. The permeability of this caprock has been responsible for the extensive interstratal karst which distinguishes much of the North Crop with its doline fields formed in the grit.

Along the North Crop, the limestone dips gently

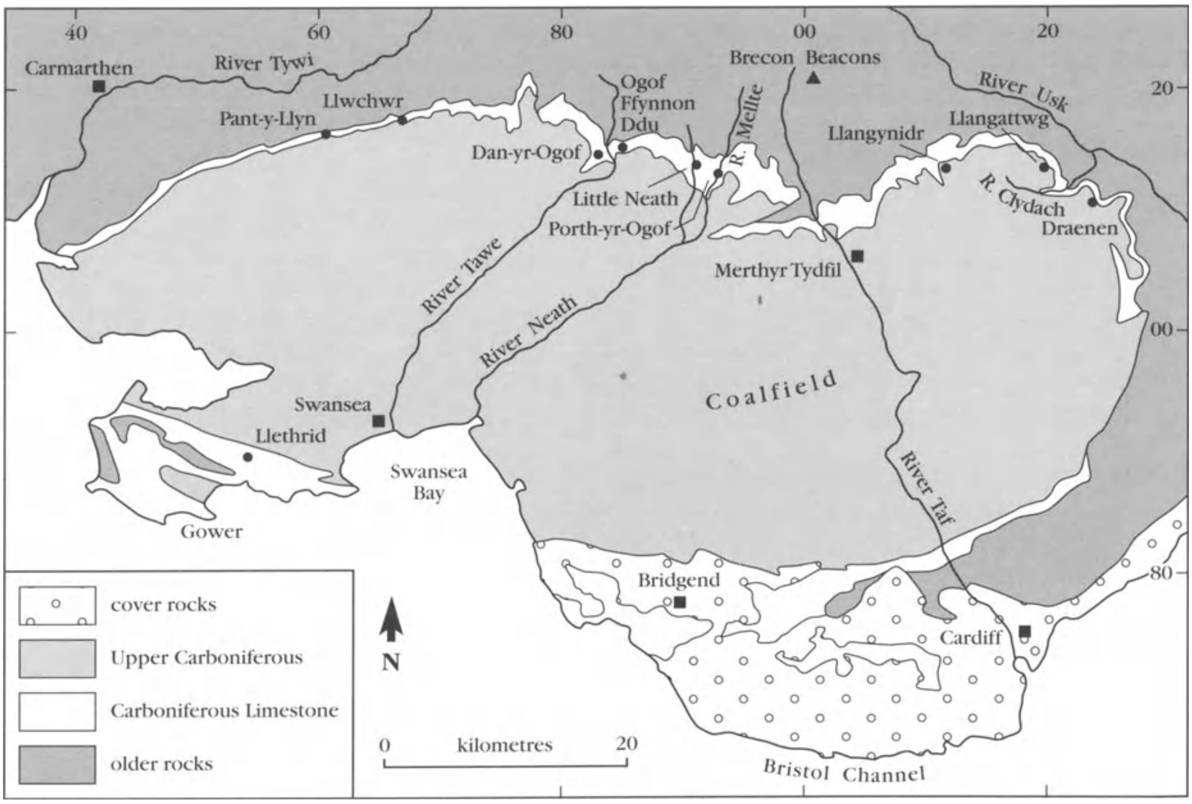


Figure 6.1 Outline map of the karst areas around the perimeter of the South Wales coalfield, with locations referred to in the text. The cover rocks in the south are Triassic and Jurassic mudstones and thin limestones.

south into the coalfield syncline. Dips are generally less than 15°. Numerous north-south faults cross the outcrop, and there are local zones of more severe disturbance, orientated SW-NE, with major faults and steep folding (Lowe, 1989b).

The karst of the North Crop

Surface karst features are not conspicuous components of the limestone landscapes of South Wales, and there is none of the spectacular landmarks which distinguish the Carboniferous limestones in much of England. This is partly because the limestone is surrounded by topographically prominent sandstones which lie stratigraphically both above and below. Much of the limestone outcrop is little more than a line of scars backed by a narrow dip slope and overlooked by the scarp face of the strong Namurian Basal Grit.

The limestone outcrops were subjected to glaciation during the Devensian, but limestone pavements are developed on only a few of the interfluves where the dip is low; much of the karst is now veneered with till, and limestone pavements are not extensive (Thomas, 1970). Ice moved south from central Wales and the Devonian sandstone mountains, deepening pre-Devensian valleys right across the narrow limestone outcrop (Bowen, 1970: Bowen and Henry, 1984; Crowther, 1989; Campbell and Bowen, 1989). The modern drainage from the north is well organized to utilize these deep, gently graded valleys; the Rivers Taff and Tawe cross the limestone entirely above ground (Figure 6.1). The Hepste, Mellte and Nedd Fechan (Little Neath) all sink into the limestone, though only the Mellte fails to use an overground flood route. There are no deep gorges in the limestone, though lines of white scars line some of the smaller dry valleys on the limestone slopes.

There are some notable doline fields on the limestone, mostly of subsidence dolines developed in the thicker mantles of glacial till. These are overshadowed by the extensive fields of large dolines which have formed in the Namurian Basal Grit where it forms a cap on the limestone dip slopes. The Grit dolines are the result of interstratal karst - where the limestone beneath has been removed by solution, followed by subsidence and collapse of the insoluble Grit cover (Thomas, 1974). The dolines occur on most of the Grit plateaus immediately south of the limestone outcrop of the North Crop, with the finest on Mynydd Llangynidr; they are unmatched anywhere else in Britain.

The caves of the North Crop

The most important feature of the North Crop drainage is the southward flow off the higher slopes of Devonian Old Red Sandstone in the north. Though the limestone outcrop is only narrow, its position across the regional trend of slope and surface drainage allows it to capture very large supplies of allogenic water. Where a valley outlet exists in the same bed, the favourable hydraulic gradients through the limestone have created conditions ideal for cave development. Compared with the Yorkshire Dales karst, the caves of South Wales are few in number, but those that do exist are notably long and deep (Ford, 1989a).

A characteristic of many North Crop caves is that allogenic water sinks into them at or near the stratigraphic base of the limestone. The immediate underground drainage is then downdip, until the contemporary water table is reached close to the level of the adjacent valley breaching the limestone outcrop. Phreatic flow then develops broadly along the strike. This situation still exists at the western end of the North Crop, where drainage, locally from both sides of the narrow limestone outcrop, is along the strike to the flooded resurgence of Llygad Llwchwr (Figure 6.1) (Ford, 1989a). At most other sites, new systems of phreatic strike drainage have developed further downdip, in response to subsequent rejuvenations. New vadose inlets, extending considerable distances down the gentle dip, have intersected the old phreatic trunk caves, creating the very extensive passage networks which give these Welsh caves their great length. The North Crop has four of the five longest cave systems in Britain (Table 1.1), and also the deepest where Ogof Ffynnon Ddu drains obliquely downdip into the Swansea Valley.

Most of the North Crop caves can be ascribed to one of three broad types, which are characterized according to the relationship between the narrow limestone outcrop and the local topography. Valley floor sites with major stream sinks include the caves of Porth-yr-Ogof and the Little Neath River where large passages are developed beneath or close to the normally dry surface valleys. Where streams sink into the limestone outcrops high on the major interfluves, caves develop down the hydraulic gradient and obliquely to the adjacent valley floor; the caves of the Swansea Valley are of this type, and the contrasting patterns of Ogof Ffynnon Ddu and Dan-yr-Ogof reflect a relationship between geological structure and valley orientation. The third cave type underlies the gently sloping Grit plateaus, and carries drainage from sinks in the marginal limestone outcrop on the higher, updip side through to risings in the lower side; the caves of Llangattwg have this pattern, where the lower edge of the gentle dipslope has been trimmed by recession of the Clydach Gorge.

The sheer size of the North Crop caves give them a special place in any review of Britain's karst. They also have an exceptional diversity of passage morphology and depositional detail, and their long Pleistocene histories are recorded in their complex passage networks.

The karst and caves in outlying areas of Wales

Apart from the North Crop, the Carboniferous Limestone forms outcrops scattered across North and South Wales. They all have their own distinctive limestone sequences, structure and karst features, but the surface landforms and the caves are more limited in scale than those of the North Crop.

The Wye Valley

Carbonates nearly 300 m thick crop out around Chepstow, and continue eastwards to the Forest of Dean, over the border into England (Chapter 7). A few small stream sinks, and some shallow dry valleys in the farmed lowland, are almost the only expressions of karst processes, though the incised meanders of the River Wye do have some cliffs of dolomite shrouded in vegetation. The one truly remarkable feature of the area is Otter Hole, a substantial cave system cut in the Lower Dinantian Lower Dolomite. This cave is unique in that both its resurgence and its only accessible entrance passage lie in the intertidal zone, but it is especially renowned for the very large calcite stalagmites in one of its chambers. These are on a scale unmatched elsewhere in Britain, and are more comparable to caves in Mediterranean environments; they probably reflect higher solution and deposition rates beneath a soil-covered karst further south and at lower altitude than other major caves in Britain's Carboniferous Limestone.

The Gower Peninsula

The limestones along the southern margin of the main South Wales coalfield syncline have been broken into two fragments by the coastal incursion of Swansea Bay. The western fragment forms the Gower Peninsula, and contains the only notable karst features. The Dinantian limestones thicken to the south and west in South Wales (Lowe, 1989b), and Gower has a sequence of pure limestone more than 400 m thick. Hercynian compression increased along the syncline towards the west; the limestone was steeply folded, and now forms a series of narrow outcrops between belts of sandstone.

Dry valleys cross the limestone outcrop where drainage is underground. The known caves are mainly small, but there are some larger old chambers; these could be remnants of more extensive pre-Devensian cave development which is also responsible for some recent collapse features in the Bishopton Valley (Ede and Bull, 1989). A number of caves open in the coastal cliffs of Gower; Bacon Hole and Minchin Hole, both on the southeast coast, are very old solution cave fragments, now most notable for their extensive sequences of Pleistocene sediments and archaeological material (Stringer, 1977; Sutcliffe, 1981; Stringer *et al.*, 1986; Bowen *et al.*, 1989).

South Glamorgan

East of Swansea Bay, the Carboniferous Limestone underlies part of the lowland of southern Glamorgan. Much of it is covered by Triassic and Jurassic mudstones or glaciofluvial sediments. The outcrops bear few signs of a karstic landscape, and there are no significant caves. Palaeokarstic fissures in these limestones contain Triassic and Jurassic sediments, and comparable, larger karst conduits contain the hematite ore deposits once worked at Llanharry (Simms, 1990). Thin, nearly horizontal limestones within the Mesozoic cover support limited karstic development around Bridgend.

Solution fissures, potholes and subsidence dolines have been recorded in the Liassic limestones, and a cave in Triassic limestone was found to have over 100 m of rifts and phreatic tubes (North, 1952). Holocene and modern tufas lie in valleys cut into the coastal cliffs of Lias limestones south of Bridgend (Campbell and Bowen, 1989).

South Dyfed

The Carboniferous Limestone thickens to over 1000 m in the south-western corner of Wales, but much of the succession is thin bedded with high proportions of intercalated shale. The limestones are tightly folded in the Hercynian compression zone, and form only narrow outcrops. Inland, the glaciated platform has little sign of karst, except for a few small caves largely choked with sediment (Davies, 1989). The high cliffs of the south coast contain numerous caves; many of these have karstic origins, and have been breached or modified by marine action. Some caves in the wave-battered cliffs contain evidence of human occupation, which must have occurred when the sea was far from its present position during the Devensian (Davies, 1989).

The Clwydian Hills

Carboniferous Limestone forms high ground on both sides of the Vale of Clwyd, at its most conspicuous on the great escarpment of Eglwyseg Mountain, just north of Llangollen (Figure 6.2). The Asbian Loggerheads Limestone is the dominant unit, only 100 m thick at Eglwyseg, but over 500 m thick further north.

East of Clwyd, Halkyn Mountain is part of a limestone belt with poorly developed surface karst, whose natural underground drainage all flowed to St Winifride's Well at Holywell. The limestone is laced with mine workings, developed to extract the rich ores of lead and zinc (Richardson, 1937; Warwick, 1968; Appleton, 1989); the main production was from about 1800 until 1958. The mines intercepted many stream caves and large phreatic chambers extending above and below sea level; Powell's Lode Cavern is 70 m long and over 30 m high and wide. Drainage adits, mined through the impermeable Namurian cover towards the east, lowered the water tables throughout the limestone. This action permanently dried up the natural resurgence at St Winifride's Well, a vauclusian rising in a faulted anticline of Brigantian black limestones and chert beds at the top of the main carbonate succession (the rising is an important religious site and its flow has been reinstated by a concealed diversion of water from another nearby drainage tunnel). The adits also drained extensive phreatic cave systems beneath the Alyn Gorge, where the middle course of the River Alyn is normally dry in summer. Halkyn Mountain has a

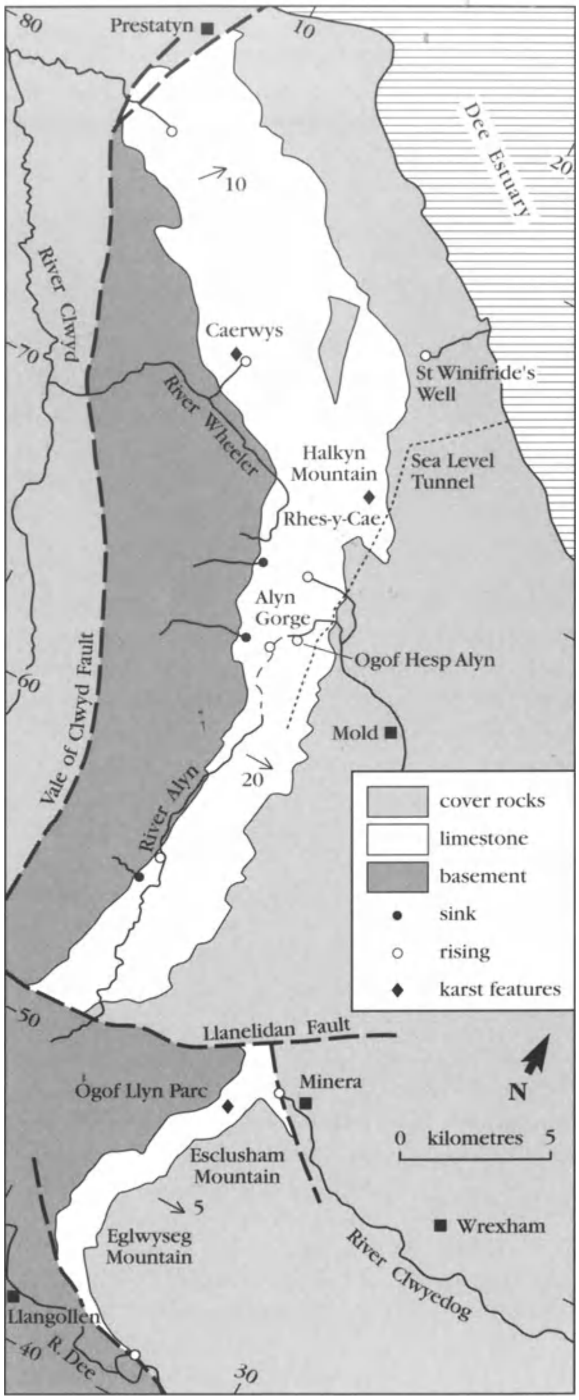


Figure 6.2 Outline map of karst features in the Carboniferous Limestone of eastern Clwyd, North Wales, with locations referred to in the text. The main rivers and risings are shown as they were before disturbance by the mine drainage. The basement is Ordovician shale; the cover rocks are Upper Carboniferous and Triassic clastics. Many of the steps on the boundaries are due to minor faults.

number of pocket deposits where subsided Tertiary sediments are preserved in solution depressions; the Rhes-y-Cae pit is the best documented (Walsh and Brown, 1971). Postglacial tufa deposits at Caerwys are the most extensive in Britain; they have been heavily quarried, but this has revealed the structure of the barrage, pool and cave deposits (Pedley, 1987).

South of the Llanelidan Fault, the splendid Minera cave system lies beneath the northern slopes of Esclusham Mountain; it also has its lower phreatic passages partially drained by the mining activity. To the south, Eglwyseg Mountain has towering limestone cliffs and some of the finest limestone pavements in Wales. The outcrop is a topographic high, so no allogenic drainage reaches it, and there are no influent caves; its autogenic waters collect in a karstic drainage system feeding sediment-choked risings in the Dee Valley.

West of Clwyd, the limestone outcrops are smaller, and the karst is poorly developed. The River Elwy crosses the limestone in a gently graded, alluviated valley. In its northern slope, Pontnewydd Cave is a truncated fragment of large cave passage which was almost full of sediments; excavation of these has revealed a record of Pleistocene environments and human occupation extending back 240 ka (Green, 1984) - it is the most northerly site in Europe with human remains of this age. The limestone continues through broken outcrops to the Great Orme, at Llandudno, where the karst drainage is so poorly integrated that adits were driven to permit mining down to sea level.

Anglesey

The upper Dinantian limestones form two outcrops in southern Anglesey, but the lowland aspect and the thick cover of glacial and glaciofluvial debris preclude significant karst development. The island's limestones are most notable for the spectacular palaeokarst exposed in the wave-cut platform and low cliffs of the east coast (Baughen and Walsh, 1980; Walkden and Davies, 1983). Cylindrical sandstone pipes, about 1 m in diameter, penetrate the limestone for up to 3 m. The sandstone fills are Dinantian; they lie in hollows which appear to be solutional features excavated in a temporarily uplifted coastal platform, though some may be moulins developed by wave action.

DAN-YR-OGOF

Highlights

Dan-yr-Ogof is an outstanding example of a cave system with contrasting geological controls on its configuration; it has fault-guided inlet passages draining to a trunk route which is predominantly bedding-controlled close to the axis of a minor syncline. The cave contains classic examples of phreatic and vadose passage morphology, some of which are now superbly decorated with calcite speleothems.

Introduction

The Dan-yr-Ogof cave system includes the truncated fragment known as Tunnel Cave and is located on the western side of the Tawe, or Swansea, Valley, north-east of Ystradgynlais (Figure 6.1). Parts of both Dan-yr-Ogof and Tunnel Cave are operated as show caves, the latter under the name of Cathedral Cave; above the two entrances, Ogof yr Esgyrn is a cave fragment which has yielded many Bronze Age and later artefacts (Mason, 1968). The caves are developed almost entirely in the Carboniferous Dowlais Limestone, of Holkerian age, which locally reaches a thickness of about 100 m; some passages extend into the Asbian limestone above. Old Red Sandstone crops out to the north, and part of the cave lies beneath the cap of Namurian Basal Grit which overlies the limestone to the south. The gentle southerly dip of the Palaeozoic succession is interrupted by the Cribarth Disturbance (Owen, 1954; Lowe, 1989b), a belt of tight folds and faulting. Immediately north of this, the cave lies within a shallow, asymmetric syncline in the limestone (Figure 6.3). Several minor faults extend north or NNE from the Cribarth Disturbance, parallel to the major joint set, and a minor joint set is orientated WNW–ESE.

Allogenic water reaches the caves mainly from the Old Red Sandstone slopes to the north, though it appears that much of the flow which formerly entered the system has been captured by the River Haffes to the north. Percolation water from the limestone outcrop contributes significantly to the underground flow, and there are numerous dolines and small stream sinks on the hillsides above the cave.

The Dan-yr-Ogof cave system has been described by Coase and Judson (1977) and Coase

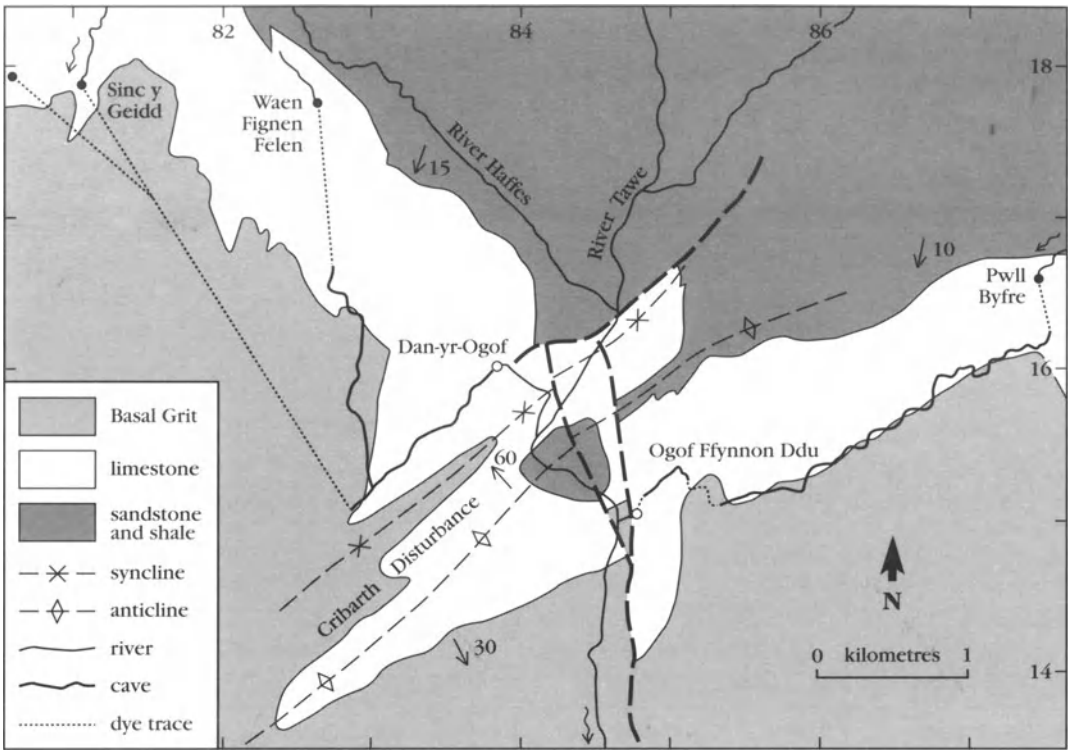


Figure 6.3 Geological map of the North Crop of the Carboniferous Limestone where it is crossed by the River Tawe in the Swansea Valley. Many small faults are omitted to improve clarity. The sandstones and shales below the limestone are mainly Devonian but include the Lower Limestone Shale from the Carboniferous. The only caves marked are the main stream passages in Dan-yr-Ogof and Ogof Ffynnon Ddu.

(1967, 1975, 1989), and the hydrology is reviewed by Gascoine (1989). Short descriptions of the cave passages are given by Stratford (1995), though there are significant more recent discoveries (Kealy, 1992; Murlis, 1992).

Description

Dan-yr-Ogof contains more than 16 km of cave passages (Figure 6.4). The south-eastern arm of the system, largely known as Dan-yr-Ogof 1 and part of Dan-yr-Ogof 2 (DYO1 and DYO2), is the trunk route of the cave, with passages at several levels draining to the north-east. The linear passages of Dan-yr-Ogof 3 (DYO3) form the western arm of the system and are inlet passages draining south, largely along a series of north-south faults.

The downstream end of DYO1 has an artificial tunnel into the show cave just above the resurgence. This intercepts a long high-level passage, which connects with the lower, active passage at various points. The main river cascades through a series of lakes, ponded by sediment banks and rock bars within the old, horizontal, phreatic tubes. Downstream it flows through the sumps of the Battle of Britain Series (Murlis, 1992); upstream it emerges from more totally flooded passages, and is only seen again in the Syphon Series of DYO1, in the magnificent phreatic borehole of Bakerloo Straight in DYO2, and in the complex series of partly flooded phreatic tubes of Mazeways. Above the lower, active level, there is an extensive, and complex, series of meandering high-level passages, locally well decorated with calcite speleothems and containing thick, clastic, sediment sequences. The show cave high level has undercut vadose passages with remnants of a phreatic tube preserved in the roof (Figure 6.5). Further west, this level is dominated by old phreatic passages, modified considerably by vadose entrenchment. Extensive collapse and choking with sediment have broken these passages into a series of isolated fragments; the chokes have been bypassed via much smaller high-level phreatic passages - including the series of small rifts which provide the route into DYO2.

A massive choke immediately north-east of Gerrard Platten Hall lies directly below the Crater, a large collapse doline on the surface. The high

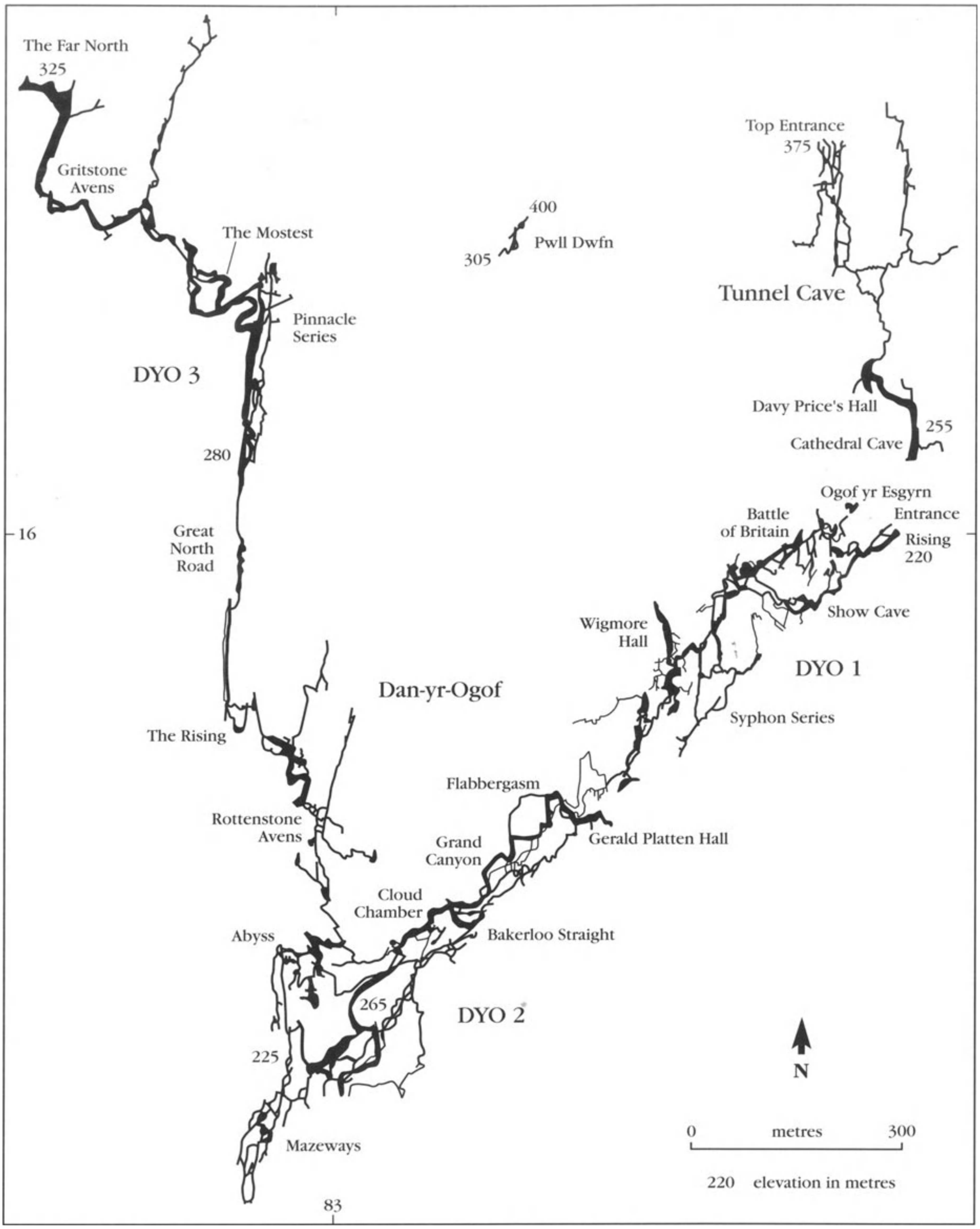


Figure 6.4 Outline map of Dan-yr-Ogof and Cathedral Cave (from survey by South Wales Caving Club).

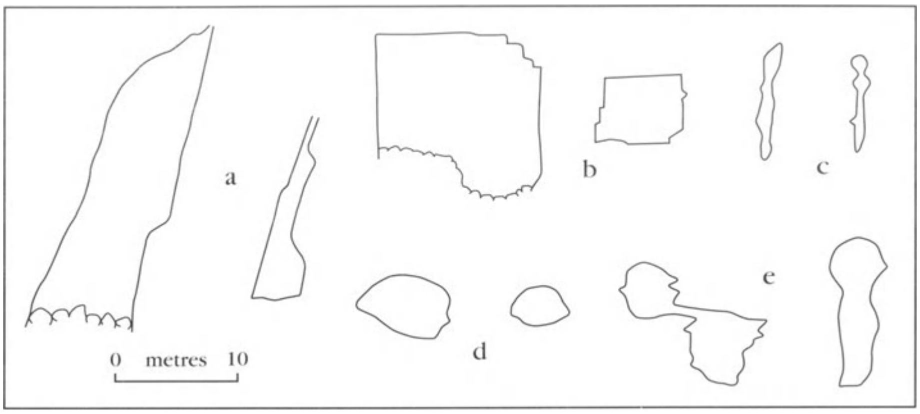


Figure 6.5 Passage cross-sections in Dan-yr-Ogof: (a) fault-guided rifts in the Great North Road; (b) collapse-modified tunnels in the Far North; (c) deep vadose canyons in Tunnel Cave and DYO2; (d) phreatic tubes in the synclinal zone of DYO2; (e) phreatic tubes with large vadose floor trenches in DYO1 and DYO2. (After Coase, 1967, and Coase and Judson, 1977.)

level continues into the Grand Canyon, which has a classic keyhole shape, with a vadose trench 2 m wide and up to 7 m deep cut in the floor of a large meandering phreatic tube (Figure 6.5). Flabbergasm Chasm is a magnificent phreatic tube up to 3 m wide forming an abandoned loop north of the Grand Canyon roof tube. It is decorated with calcite straw stalactites up to 2.5 m long and gypsum oulopholite flowers, while crystal pools, mud-cracks and drip-pits adorn the floor. Further west the main passage widens to 8 m, but thick clastic sediments reduce its present height to 2 m, before Monk Hall, Cloud Chamber and Hangar Passage form a section clear for their full heights and richly decorated with straw stalactites and other calcite speleothems (Figure 6.6). This large old passage is partially blocked by collapse at several points, and eventually ends to the west at clay

Figure 6.6 Calcite straw stalactites hang from the arched phreatic roof of Cloud Camber in Dan-yr-Ogof. (Photo: J.R. Wooldridge.)

and boulder chokes. Phreatic tubes at a lower level (Figure 6.5) include Bakerloo Straight, and are largely abandoned as the main water flows through another, lower set of flooded conduits.

From Cloud Chamber, the ponded Green Canal passage links through to more, large, dry passage at the southern end of the fault-controlled, western limb of the cave system. To the south, sections of large passage are blocked by clay and boulder chokes, and a network of smaller phreatic tubes extends through a flooded section to Mazeways Two. Dali's Delight, close to the Abyss, has irregular scalloped pillars etched into the Honeycombed Sandstone, a distinctive band of basal Asbian arenaceous limestone 1 m thick. To the north, a narrow vadose canyon is a flood route from the north which passes beneath the Rottenstone Avens, and leads upstream to junctions where high-level rifts, decorated with helictites, pass over the sumped section at The Rising.

The Great North Road is the main passage in DYO3; it is a large vadose canyon modified greatly by collapse along a series of closely spaced, steeply dipping fault planes (Figure 6.5). At Pinnacle Chamber, the passage is 10 m wide and 20 m high, with the Pinnacle Series of high-level passages developed above. Further north, a superb section of undercut, meandering vadose passage swings round from the west below a phreatic tube, 6 m high and 15 m wide. The two unite briefly upstream in a classic keyhole-shaped passage, before the phreatic tube turns north again in The Mostest, beautifully decorated with coloured flowstone, gour pools and calcite crystals around a dried-out pool. Beyond a junction with an inlet from the north, the main passage contains numerous large boulders of grit and quartz conglomerate beneath the Gritstone Avens. Large sediment banks precede a massive terminal choke in The Far North, a passage 13 m high and wide and modified by block collapse (Figure 6.5), at the end of the explored cave.

Tunnel Cave contains more than 2100 m of passages (Figure 6.4). The northern inlets of the cave are descending series of narrow vadose rifts, locally with well developed roof tubes (Figure 6.5). These unite downdip, to the south, into a vadose canyon which leads into the large passage of Davy Price's Hall, extensively modified by collapse and containing thick banks of sand, silt and mud. This chamber is now open as a show cave, under the name of Cathedral Cave, with an artificial entrance close to its tiny active outlet.

Interpretation

Both the Great North Road and Tunnel Cave have developed due to drainage almost straight down the regional dip. The linear form of the Great North Road reflects its development on a series of north–south faults and associated fractures, while Tunnel Cave follows only joints which are less extensive. The gradient of DYO3 is less than the dip, so that it climbs stratigraphically on the fault planes - reflecting initial development under phreatic conditions. Both these inlets, and a third inlet from Sinc y Geidd (Figure 6.3), drain into the south-eastern arm of the cave, developed close to the trough of the asymmetrical syncline north of the Cribarth Disturbance.

The axis of the syncline is almost level, and the nearly horizontal phreatic passages have drained towards the aquifer outlet in the Tawe Valley. The cave is not in the trough of the main fold, whose surface expression is the tongue of Grit outcrop just to the south-east (Figure 6.3). However, dips recorded in the cave clearly show the presence of a shallow synclinal flexure, repeatedly displaced by small crossing faults, the axis of which is rigorously followed by the cave (Coase and Judson, 1977). Joints have exerted a minor influence by creating a network of fissures which the main flow utilized; many passage segments are joint aligned, but the main cave nowhere strays far from the direct line to the valley resurgence. The confluence of the Great North Road faults with the syncline is the site of a sprawling complex of passages in Mazeways, which extends as development along the strike by water from Sinc y Geidd.

The series of passage levels, and the sediment and speleothem deposits which they contain, record a long and complex history which awaits an absolute chronology based on uranium-series dating of speleothems. Initially the cave system consisted of a series of small, fracture-controlled, phreatic rifts and tubes which drained downdip to the south, into the syncline of the Cribarth Disturbance. Slower flows to the north-east within the syncline trough enlarged the more complex network of fissures which were the ancestors of the many passages now forming DYO2 and DYO1. Remnant from this phase may be the small, high-level, phreatic tubes which survive as bypasses around the massive chokes in the southern part of the system. The large high-level trunk passages, mostly on DYO2, represent the main phase of cave development, since fragmented by collapse and sediment infill. The large size of the trunk passages probably indicates a much higher flow than that of the present active streamway.

Subsequent drawdown has resulted in extensive vadose modification of many of these old phreatic passages, notably on the shallow phreatic loops caused by the interplay of joint control and dip within the syncline. Downcutting of the Tawe Valley, and erosion through some of the phreatic loops within the cave, has favoured enlargement of new drainage routes at lower levels through the same fissure networks. This process has been repeated a number of times in different parts of the system, but the majority of the development has been phreatic in the synclinal trough; the active drainage route is still largely flooded. Vadose entrenchment of the phreatic tubes has been deepest close to the resurgence, in direct response to the surface lowering, and in the steeper passage gradient of Tunnel Cave.

The abrasiveness of sand sediment washed into the cave may account for the largest section of passage being that closest to the former sinks feeding into the Far North. These large passages date probably from a period prior to the capture of much of the surface drainage by the River Haffes (Figure 6.3). The choke in Gerrard Platten Hall lies 40 m below a large collapse doline, and the feature represents an early stage of the dissection and eventual destruction of the cave. Tunnel Cave appears to have been an inlet to Dan-yr-Ogof before the side of the Tawe Valley retreated far enough to remove the junction; its truncation is another aspect of cave destruction.

The stages in the development of the caves, with their consecutive sequences of passages and repeated rejuvenations, must relate to the series of ice advances and the glacial and fluvial valley deepening during the Pleistocene. Much of the original allogenic drainage has been lost to the River Haffes, in an unusual example of underground drainage being captured by a surface stream. The resurgence is now perched above the valley floor, as the water emerges from a truncated phreatic tube very close to the base of the cavernous limestone. A preliminary interpretation of events ascribes different drainage routes and cave passages to the interglacial stages of the Pleistocene (Coase, 1989), but it remains conjectural without absolute dates from the cave sediments. Environmental and chronological data have yet to be elucidated from the thick clastic and speleothem sequences in the various cave levels, to obtain a clearer picture of the evolution of both the cave system and its surrounding landscape.

Conclusion

Dan-yr-Ogof is a major cave system with large, fault-guided passages uniting in a drained phreatic trunk route along the axis of a syncline. It shows very clearly the effect of various geological controls on cave development, and it contrasts with the adjacent Ogof Ffynnon Ddu, a cave system with a conspicuously different morphology in the same limestone in a different structural situation on the other side of the Cribarth Disturbance (Figure 6.3). Rejuvenation in response to valley incision has left an extensive series of partly abandoned, high-level passages, many of which are classic examples of their type, superbly decorated with calcite speleothems.

OGOF FFYNNON DDU

Highlights

Ogof Ffynnon Ddu is the deepest cave system in Britain, and is also one of the most extensive. The complex network of large high-level passages, the exceptionally long vadose streamway, and the many inlets perched high above it, provide an unparalleled record of drainage evolution in the limestone.

Introduction

Ogof Ffynnon Ddu lies beneath the eastern slopes of the Tawe Valley (or Swansea Valley) at Penwyllt, upstream of Ystradgynlais (Figure 6.1). It is the second longest cave system in Britain, with around 50 km of explored passages, and is Britain's deepest cave with a vertical range of 308 m.

The cave is developed entirely within the Holkerian Dowlais Limestone, which is locally just under 100 m thick. This limestone has a broadly uniform dip of about 10° to the south, as it lies clear of the Cribarth Disturbance (Figure 6.3), and it forms an outcrop less than 1 km wide across the upland interfluve between the Neath and Tawe valleys. It is broken by numerous faults, mostly orientated north–south with displacements of up to 35 m, and the major joint sets trend roughly north–south and east–west. A series of gentle fold flexures have their axes aligned close to the regional dip. Allogenic water flows south from the Old Red Sandstone slopes, and the only large surface stream feeds the main sink at Pwll Byfre. Underground flow is westwards through Ogof Ffynnon Ddu to the resurgence of Ffynnon Ddu, close to the floor of the Tawe Valley. Most of the upland slopes have a thin veneer of till which obscures truncated passages known from inside the cave system.

The progressive exploration and understanding of Ogof Ffynnon Ddu has been documented primarily by Railton (1953), O'Reilly *et al.* (1969) and Smart and Christopher (1989), and the main cave passages are described by Stratford (1995). The relationship of the cave to the geology has been discussed by Glennie (1950) and Charity and Christopher (1977), while aspects of the hydrology have been investigated by O'Reilly and Bray (1974) and Bray and O'Reilly (1974).

Description

A main stream passage extends through the length of the Ogof Ffynnon Ddu cave system, but most of the 50 km of known passages constitute complex, three-dimensional networks of active and abandoned caves (Figure 6.7).

The main streamway is 5 km long, covering most of the distance between sink and rising. The upper end is in the large chamber of Smith's Armoury, where the water from the Pwll Byfre sink emerges through a choke of sandstone boulders. Most of the streamway is a magnificent, clean washed, vadose canyon, 2–5 m wide and 5–30 m high. The Marble Showers area is notably spectacular where the canyon walls, cut in dark limestone streaked by white calcite veins on small faults, are washed by inlets from the roof. The water cascades and swirls through numerous moulins and deep pools, and over ledges where dolomitic horizons have resisted solution more than the adjacent limestone. There are few waterfalls along the main streamway, for it descends 300 m largely by following the bedding obliquely downdip. At floor level, the stream follows a contorted course where meanders have enlarged as they have been entrenched. The higher levels of the canyon are commonly aligned on joints, so that they appear as straight rifts; the Traverses (Figure 6.8) are high in a spectacular straight canyon over 40 m high, now abandoned where the stream passes through a short flooded loop between the canyons of OFD3 and OFD2. From just below the Piccadilly junction, the stream route is again through a flooded loop beneath the Rawl Series, until it emerges in

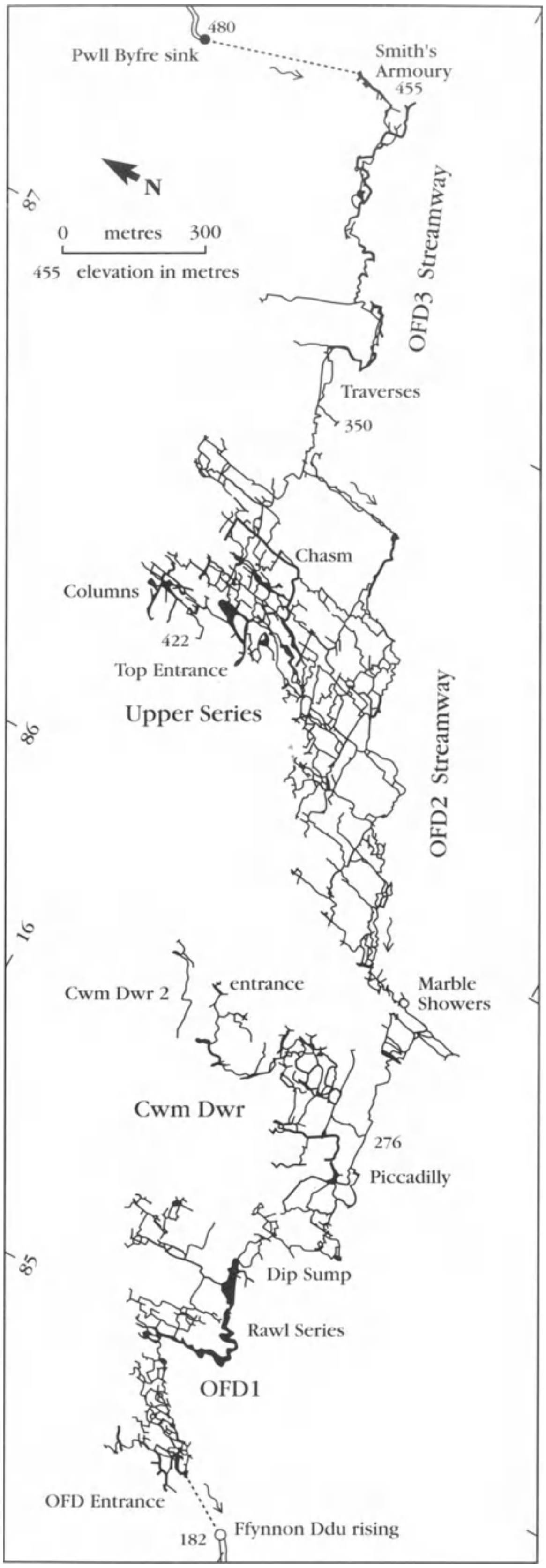


Figure 6.7 Outline map of Ogof Ffynnon Ddu (from survey by South Wales Caving Club).

Figure 6.8 The deep vadose canyon where the Traverses in OFD3 are high in the roof above the upper end of the streamway in Ogof Ffynnon Ddu. The ledges are created by lithological contrasts in the limestone beds. (Photo: J.R. Wooldridge.)

the OFD1 streamway. This drains into a sump not far from the Ffynnon Ddu resurgence pool.

Extensive active and relict passages form networks at multiple levels, almost entirely on the north, updip side of the streamway. The Upper Series of OFD2 is the most complex, with a maze of interconnected phreatic tubes and vadose canyons. Many of the caves are aligned on joints or faults, and some fractures guide three separate passages stacked vertically above each other. The largest passages are old trunk routes over 10 m wide, now broken into fragments by roof collapse. One deepens into a large vadose canyon forming the Chasm, and another has been truncated by surface lowering to create the Top Entrance. Sections of the old caves are beautifully decorated by calcite dripstone, of which the Columns are the most distinctive (Figure 6.9); a stalagmite floor over an eroded clastic fill in the Upper Series has been dated to 267 ka by uranium-series analysis (Smart and Christopher, 1989). Younger vadose canyons have been cut by invading streams right through the mazes of old passages; they drain down the dip to the main streamway. Even though these are deeply entrenched, their incision rates have not matched that of the main stream, and most are perched as roof inlets.

Downstream in OFD2, another series of large, abandoned, phreatic passages forms a high level, reaching towards a truncation at the Cwm Dwr Entrance. Cwm Dwr 2 is an isolated fragment of tributary streamway (Herbert and Langford, 1991).The relict passages from Cwm Dwr to Piccadilly also extend downstream to form a passable link over the flooded section of streamway into OFD1. These continue through the spacious passages of the Rawl Series, which are the only extensive old passages on the downdip side of the present stream route. They continue into another very complex maze which can be followed above the modern stream canyons, to emerge from the original OFD Entrance where the abandoned passage is truncated by the side of the Tawe Valley.

Pant Mawr Pot lies 1500 m east of Pwll Byfre (Alexander and Jones, 1959; Moore, 1989). In the floor of a large shakehole, a shaft drops into a single passage up to 8 m wide and 5 m high, with extensive clastic sediment fills. Though this largely abandoned passage may once have related to Ogof Ffynnon Ddu, its underfit stream now drains into the Neath Valley further east.

Interpretation

The evolution history of Ogof Ffynnon Ddu is long and complex. The narrow outcrop of dipping limestone, from the interfluve ridge to the valley floor, has dictated the overall pattern of karst drainage - obliquely down the dip to the contemporary valley floor resurgence. This pattern has survived through successive deepening of the Tawe Valley, creating consecutive series of passages superimposed into each other. The overall drainage oblique to the dip has been developed by utilization alternately of downdip and strike fractures. Passage development further down the dip has been inhibited where longer loops would have passed beneath the Grit cover into limestone less favoured by authigenic solutional enlargement of its fissure network. New downdip

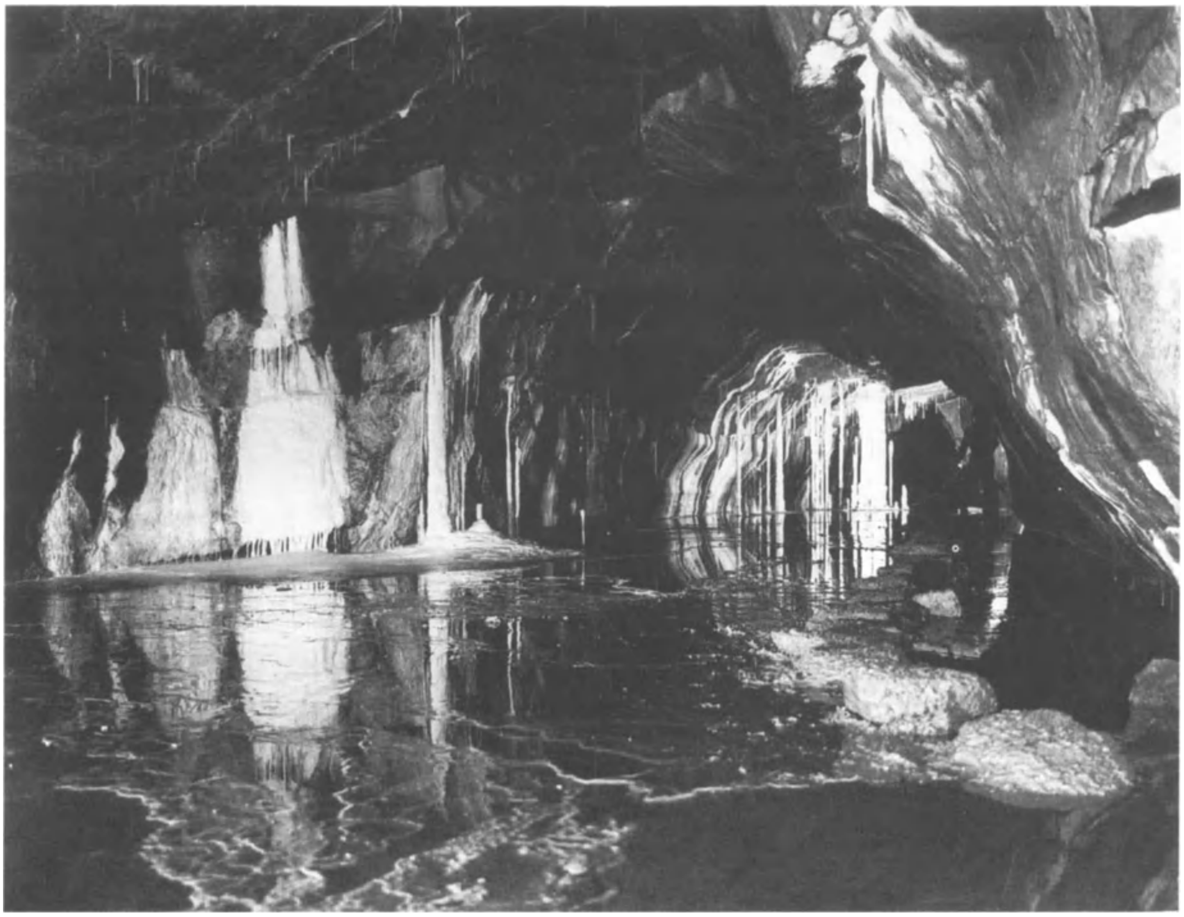


Figure 6.9 The Columns in Ogof Ffynnon Ddu - calcite stalactites and stalagmites which have grown to connection in a fossil phreatic tube. (Photo: South Wales Caving Club.)

drainage paths have therefore developed largely in response to the surface lowering and downdip shift of the outcrop, with simultaneous lowering of the valley floor. The valley deepening also permitted the development of new, lower resurgence sites, while the downstream ends of the older, higher phreatic passages, which had also discharged to the west, were progressively destroyed by surface erosion.

Solutional fissure enlargement in the well bedded and well fractured limestone of Ogof Ffynnon Ddu has led to an uncommonly large number of drainage captures in both the phreatic and vadose parts of the aquifer. This has led to the development of a complex multi-level network cave quite distinct from predominantly two-dimensional maze caves, such as Mossdale Caverns (Figure 2.48), in which passage capture has played only a minor role. The local relief on cave drainage routes created sections of steep downstream gradients; these became the sites of rapid updip vadose entrenchment of the phreatic loops, leading to the gradual elimination of phreatic segments with a corresponding increase in total length of the vadose streamways. This is an important mechanism of passage evolution in many caves in dipping limestone, and Figure 1.7 is based on detailed observations in Ogof Ffynnon Ddu. The process is well demonstrated in the main streamway, where the vadose canyons are exceptionally deep; the original passages are preserved over many crests of the old phreatic loops, and new phreatic loops are developing beneath some of the old canyons. The result is a hybrid cave, intermediate between water table and vadose drawdown caves (Ford and Ewers, 1978; Smart and Christopher, 1989).

The rectilinear pattern of so much of Ogof Ffynnon Ddu is ample evidence of the role of the tectonic fractures in the establishment of the karst drainage and the cave passages. There are further expressions of geological control in the caves. The major networks of high-level passages coincide with gentle anticlinal structures which plunge down the regional dip, while the intervening synclines house far fewer cave passages (Charity and Christopher, 1977); this may be due to either the tensional opening of fissures over the fold crests, or the earlier exposure of the limestone on the structural highs. Bedding of the limestone has also influenced passage morphology, where projecting ledges are formed by more resistant dolomite horizons, and where passages with square sections and flat roofs have been modified by collapse. In the upper part of the Dowlais Limestone, the *Composita ficoides* bed is more sparsely jointed than most; roof collapse and upward stoping frequently stop at this horizon, which forms the roof in many parts of OFD2 Upper Series. The Pwll Byfre sink is close to the base of the limestone, but the cave system climbs stratigraphically in numerous phreatic lifts, so that the past and present resurgences lie at the top of the Dowlais Limestone.

Although evidence of only the last two glaciations has been recognized in this area it is probable that the development of the cave system has been influenced by earlier ones, surface evidence of which has been entirely removed by later glaciations. The effects of these glaciations include truncation of near-surface passages, vadose entrenchment consequent on base-level lowering, and subsequent infilling of passages with glaciofluvial sediments. Solutional enlargement of the cave passages takes place more rapidly in warm interglacial environments, and some of the cave levels may correspond to resurgences at interglacial valley floor levels, though no detailed correlations have yet been made. The one dated flowstone from the OFD2 Upper Series shows that these passages were drained by the Hoxnian stage (267 ka), but these passages are among the oldest in the cave and were probably dry long before this flowstone was deposited. Vadose entrenchment in the Traverses of OFD3 totals about 75 m, and a mean entrenchment rate of 100 mm ka^{-1} may be interpreted from comparable dated sites elsewhere in Britain (Gascoyne *et al.*, 1983a). This suggests an age of about 750 000 years for the OFD3 streamway, and a history for the whole cave is likely to span more than a million years (Smart and Christopher, 1989).

The sequential cross-cutting relationship of many passages and the extensive sediment and speleothem deposits which Ogof Ffynnon Ddu contains gives this system enormous potential for elucidating the Pleistocene history of the upland area, evidence for the early part of which has been entirely removed from the surface landscape. Pant Mawr Pot may represent a fragment of large passage which was once related to the older components of the high levels in Ogof Ffynnon Ddu. This would imply that the Tawe Valley was deeply entrenched before the Neath Valley was excavated, but further speculation is inappropriate until more is known of the interfluve caves so heavily choked with sediment.

Conclusion

Ogof Ffynon Ddu is a very extensive cave system developed by drainage obliquely through a dipping bed of limestone. The prevailing southerly dip, the minor folds and the joint sets have all exerted a strong control on the configuration of the cave passages. Continued incision through a large depth range has allowed the passages to evolve within these geological constraints over a very long timespan. The morphology of the passage network provides a striking contrast with the nearby cave system of Dan-yr-Ogof, and provides many features of detail which are among the finest in Britain.

LITTLE NEATH RIVER CAVE

Highlights

The caves of the Afon Nedd Fechan provide an excellent example of the progressive underground capture of a surface river in a limestone karst. They contain passages in all stages of development, including immature vadose inlets, a large main streamway, an active phreas, partly and completely abandoned high levels, and truncated fragments beneath the active part of the surface river bed.

Introduction

The caves adjacent to the Afon Nedd Fechan, west of Ystradfellte, include the Little Neath River Cave, Bridge Cave, Pwll y Rhyd, White Lady Cave and Town Drain (Figures 6.1, 6.10). The underground drainage is developed largely in the Holkerian Dowlais Limestone, which dips south at less than 10°, and is broken by a number of dip faults. Draining south from the Old Red Sandstone slopes of the Fforest Fawr mountains, the headwaters of the Afon Nedd Fechan (Little Neath River) sink at various points across the Carboniferous Limestone outcrop. All the water from the sinks in the river bed rises from deep fissures at Pwll Du, where the top of the limestone dips beneath the sandstone of the Namurian Basal Grit.

The morphology of the Little Neath River Cave is well documented by Norton *et al.* (1967), Standing *et al.* (1971) and Mullen (1987, 1988, 1990), and all the caves are briefly described by Moore (1989) and Stratford (1995).

Description

In dry weather the entire flow of the Nedd Fechan sinks within 200 m of leaving the sandstone outcrop into impenetrable riverbed fissures. The first open sink is the Flood Entrance of Little Neath River Cave. The main sink of the river is also choked, but drains into Bridge Cave, which has a dry entrance in through a doline. The water then flows through a short sump, and east along a wide bedding plane canal into the Main Stream Passage of Little Neath River Cave (Figure 6.10). This is the trunk route through the cave system, which has nearly 9 km of mapped passages, including the small joint-controlled tributaries from the Flood Entrance.

Much of the gently graded Main Stream Passage is 10 m high and wide, but it is broken by a series of shallow phreatic loops, each up to 200 m long, which are permanently flooded. The faulted and fractured limestone walls break away in numerous large collapse blocks, and there are some extensive banks of clastic fill. Abandoned oxbows and high-level passages leave and join the streamway at various points, and some of these carry invading tributary streams. The Old World and the New World Series are both complexes of old passages, at an elevation of about 260 m, close to the cave river level at the upstream end but 40 m above at the south end of New World; they include tall avens and collapse chambers containing extensive calcite speleothems. The present limit of exploration is 600 m from the resurgence, at a point 27 m deep in Sump 8, whose water surface level is the same as that of the resurgence. This flooded passage appears to pass through a downfaulted block, 400 m wide, in which the top of the limestone is below resurgence level. The faulting is more complex than the surface outcrops shown

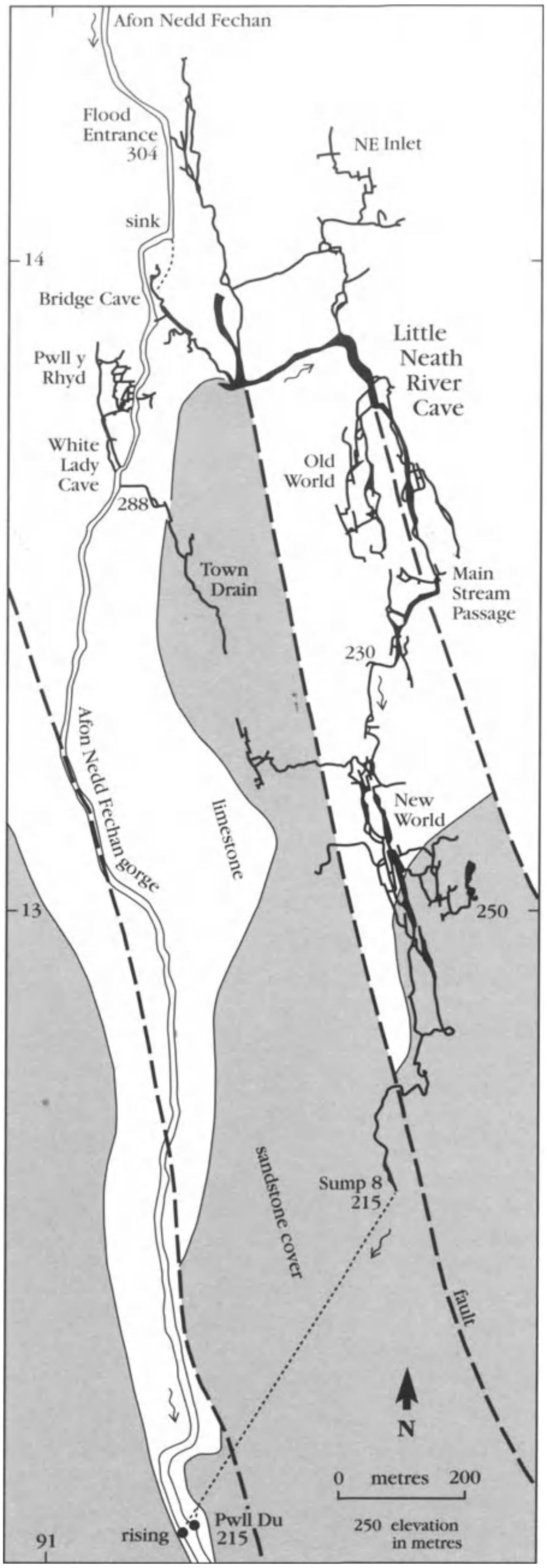


Figure 6.10 Outline map of the Little Neath River Cave, its surface geology and the adjacent caves of Pwll y Rhyd (from survey by University of Bristol Speleological Society).

on Figure 6.10, and Sump 8 may lie on or very close to the underground line of the main fault.

The bed of the Nedd Fechan continues south of the sinks into the Little Neath River Cave, and takes flood overflows as far as the open rift of Pwll y Rhyd (Figure 6.10). This is an unroofed, phreatic rift, exposed across the river bed. From its floor, a network of small phreatic rifts and tubes extends beneath the west bank, and carries flood waters to the fine elliptical tube which returns to daylight as White Lady Cave. Between Pwll y Rhyd and the White Lady flood rising, a narrow limestone gorge has breached the old phreatic caves; it is dry other than in exceptional floods, and represents the surface drainage route prior to underground capture. Town Drain, beneath the east bank, carries flood flows through immature rift passages which probably drain to the nearby inlet in Little Neath River Cave.

Interpretation

All the caves of the Nedd Fechan show strong geological control, with passage development along bedding planes, joints and a number of small faults. In plan, the main cave drains almost straight downdip, though it is deflected slightly to the east by the faults. In profile, it climbs steadily through the stratigraphy, by way of small phreatic lifts; the sinks are close to the base of the limestone, and it finally resurges from fissures which carry the flow to the top of the sequence. The lower end of the cave occupies a phreatic loop which extends beneath the downfaulted block of sandstone. The severe flooding which occurs in the lower reaches of the known cave suggests that this phreatic loop may act as a sediment trap, which restricts flow to the resurgence.

The Little Neath River Cave has a more complex morphology than many other valley floor river caves, such as Porth yr Ogof and Ogof Hen Ffynhonau, which are developed mainly at one level. Passages up to 30 m above the present streamway indicate a complex history for the cave. The main passages in Little Neath River Cave drain east of south, away from the surface river course, and it has been suggested (Moore, 1989) that early phases of the cave fed a resurgence in the Mellte Valley, 2 km to the east (Figure 6.1). It is more likely that the passage orientation is a feature of the local geology, and the resurgence has always been to the south-west (Mullen, 1990), even though the limestone outcrops are at lower level in the eastern valley. The New World Series represents the earliest underground drainage route, active until it was captured by the present streamway. These large old passages were subsequently filled by extensive sediment deposits, which have only been partly removed by the present streams.

Cave development has also been influenced by changes in the surface topography and drainage imposed by the Pleistocene glaciations. The Nedd Fechan, Mellte and Hepste originally flowed south-east as headwaters of the Cynon catchment, until they were captured by Pleistocene excavation of the Vale of Neath (North, 1962). The rejuvenated rivers entrenched into their present steep gorges, exposing the limestone at lower levels and rapidly draining previously phreatic sections of the karst aquifer. The Nedd Fechan gorge, entrenched along a fault zone west of the caves, is a youthful feature which may postdate the main downcutting of the Mellte and Hepste valleys. Its course across the limestone was probably excavated by high meltwater flows in cold stages of the late Pleistocene, creating the site of a new outlet for the cave water in subsequent warm stages. Any diversion of the cave drainage, from an earlier route to the Mellte, has not yet been dated, but the Nedd Fechan gorge is pre-Devensian as its lower reaches contain glacial till. This relatively late incision also accounts for the breaching of the caves at Pwll y Rhyd.

Conclusion

The caves of the Afon Nedd Fechan constitute an excellent example of karstic development beneath a large river valley crossing a narrow limestone outcrop. Their multiple phases of vadose and phreatic passages represent an early capture of the surface drainage, flow patterns dictated by geological constraints, and a diversion of the outlet path in response to surface rejuvenation.

PORTH-YR-OGOF

Highlights

Porth-yr-Ogof is a spectacular river cave, with the main conduit accessible from sink to resurgence. It is a fine example of underground capture of surface drainage and of the initial stages of development of a river gorge through cavern collapse.

Introduction

Porth-yr-Ogof lies in the floor of the Mellte Valley, south of Ystradfellte (Figure 6.1). Several streams drain south from the Old Red Sandstone slopes to converge near Ystradfellte, before crossing onto the lowest beds of the Carboniferous Limestone dipping south at about 4°. In dry weather, the water all flows underground into massive limestone at the impenetrable Church Sink. In flood conditions, a surface stream continues south for a further 600 m to enter the Main Entrance of Porth-yr-Ogof at the end of a short limestone gorge. The entire cave is developed near the top of the Holkerian Dowlais Limestone. After 300 m underground, the river resurges through a deep pool, 700 m upstream of a fault which crosses the valley at the end of the limestone outcrop. A shallow rocky ravine lies almost directly above the cave; it is permanently dry and its floor is breached by a collapse into the main cave passage.

The cave and its geomorphology have been described and discussed by Standing and Lloyd (1970), Lloyd (1980) and Waltham and Everett (1989), and the Mellte hydrochemistry was discussed by Groom and Williams (1965).

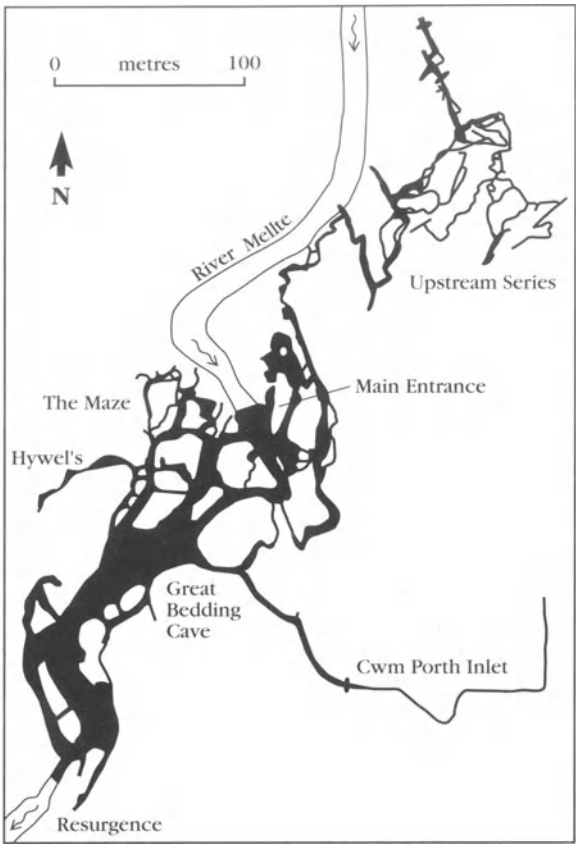


Figure 6.11 Outline map of Porth-yr-Ogof (from surveys by University of Bristol Speleological Society and Cave Diving Group). The dry valley between the sink and resurgence lies almost directly over the largest cave passages.

Description

Porth-yr-Ogof has nearly 2500 m of passages within a very small area and reached by fifteen entrances (Figure 6.11). These include the main river entrance and exit, a number of incidental joint fissures and three roof collapses towards the resurgence. The Main Entrance is 15 m wide with a shallow arched roof into a wide bedding plane chamber, with the river in a trench between wide rock shelves (Figure 6.12); it lies at the lower end of a short gorge between vertical limestone walls. Inside the cave, the Mellte flows in a wide passage with numerous oxbow loops at water level. The Great Bedding Cave is up to 30 m wide, with shallow pools and shingle banks spanned by an unbroken limestone slab beneath a gently dipping bedding plane. Downstream the passage narrows into the joint-guided resurgence rift. The baseflow of the Mellte enters Church Sink and drains into the flooded tubes and rifts which are the active part of the Upstream Series. The whole cave is distinguished by the braided form of its passages, with numerous loops extending in the bedding planes on both sides of the main drainage path. These are particularly complex on the western side, where passages on two bedding planes are connected by small shafts in the Maze. Cwm Porth Inlet is a flooded tributary on the east side, probably gathering water from small sinks along the limestone boundary (Burke, 1967). Hywel's Grotto is an old distributary on the west side, now decorated with calcite speleothems.

Interpretation

Passage morphology throughout Porth-yr-Ogof is closely controlled by the bedding planes and fractures within the massive limestone, though the effects are masked on the cave map (Figure 6.11) by the braiding on the major bedding planes. The main passages are formed on four bedding planes, each separated by limestone beds about 1 m thick. These are all exposed in the Main Entrance and first chamber, and are most conspicuous where the second from the top forms the wide roof span of the Great Bedding Cave. This wide cave sur-

Figure 6.12 The main entrance to Porth-yr-Ogof with the River Mellte flowing in between rock terraces determined by the limestone bedding. (Photo: A.C. Waltham.)

vives where the limestone beds are less fractured. The main joints are aligned NNW and NNE, controlling the rift passages around the resurgence and in the Upstream Series, and the collapse entrances at the lower end of the Great Bedding Cave. Where they are more densely packed, they have allowed more roof collapse to form the gorge upstream of the Main Entrance.

Initial development of the cave was by phreatic solution opening up fracture routes through the limestone beneath the river bed. Aided by the steep gradient of the rejuvenated Mellte (North, 1962) and the hydraulic continuity within the four dipping bedding planes, the karstic conduits were soon large enough to take the entire flow underground. A short phreatic lift at the resurgence was removed due to incision by its overflow, and the whole cave was then further enlarged in a vadose environment. Evolution has now reached a stage where the upstream end of the cave is being progressively unroofed to create the upstream gorge. Adjacent to this, the cycle is restarting where renewed underground capture is developing in the Upstream Series, which is still largely within the phreas and cannot yet take flood flows.

The gorge upstream of the Main Entrance represents the finest example in Britain of gorge incision by progressive roof collapse downstream from an influent cave entrance; unroofing of the cave is also progressing on a small scale where the old dry ravine is collapsing into the river cave beneath.

Conclusion

Porth-yr-Ogof is Britain's finest example of a completely vadose river cave in a valley floor environment. It may be compared with valley floor caves in Nidderdale, Chapel le Dale and the Alyn Valley, but its morphology is far more diverse. It shows every stage of underground capture, from early phreatic fissure enlargement through to collapse and transformation into a subaerial gorge. Within the cave, the wide roof spans in single limestone beds are particularly spectacular and are unmatched elsewhere in Britain.

MYNYDD LLANGYNIDR

Highlights

Mynydd Llangynidr contains the finest array of collapse dolines and subsidence basins seen anywhere in Britain, and clearly demonstrates the surface geomorphic effects of interstratal karst.

Introduction

Mynydd Llangynidr is located on the summit of an escarpment overlooking the Usk Valley on the northern edge of the South Wales coalfield syncline. The escarpment is formed by the Carboniferous Limestone, but is capped by

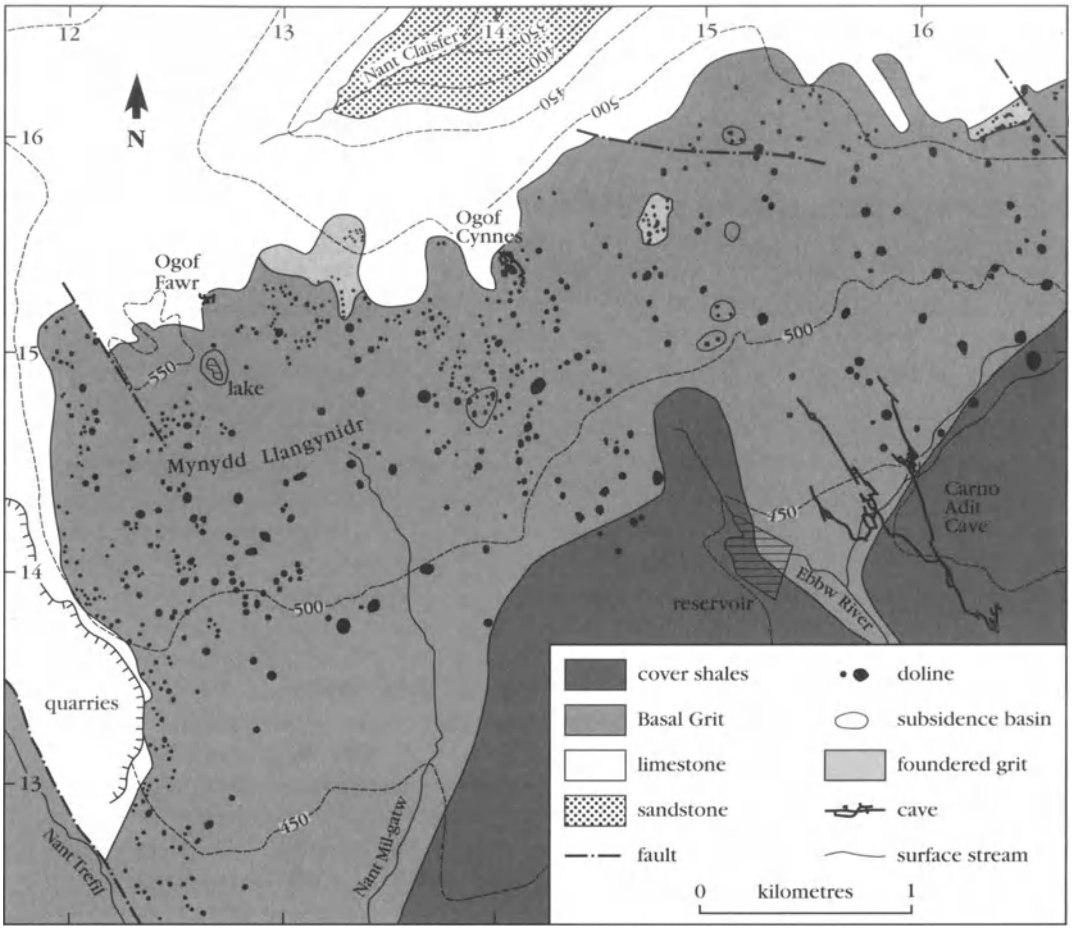


Figure 6.13 Geological map of the doline field on Mynydd Llangynidr (partly after Thomas, 1974: Ogof Carno from survey by Brynmawr Caving Club). The cover rocks are Namurian shales above the Basal Grit and Coal Measures. The sandstone beneath the limestone is Devonian. Much of the limestone outcrop is covered by soliflucted Grit blocks.

Namurian Basal Grit, and the moorland of the dip slope is pocked by a spectacular suite of collapse dolines and subsidence basins. These form a packed doline field 2 km wide near the crest of the escarpment, which is fringed by foundered masses of the Grit lying on the limestone of the scarp face (Figure 6.13). The site clearly demonstrates how subsurface, interstratal karst solution can induce collapse and therefore produce dolines and subsidence basins in the non-carbonate outcrop.

Thomas was the first to describe the geomorphic effects of the interstratal karst on the limestone and sandstone outcrops all along the northern rim of the South Wales coalfield (Thomas, 1954, 1963, 1973, 1974). The link between surface collapses and underlying cave systems was examined by Bull (1977) in the adjacent Mynydd Llangattwg, while Battiau-Queney (1980, 1986) recognized a buried palaeokarst exposed in quarries 4 km west of Llangynidr, and Smart and Christopher (1989) attribute some examples of large masses of foundered Millstone Grit to faulting.

Description

Mynydd Llangynidr is a bleak moorland rising to 550 m. To the north it is bounded by a steep scarp face overlooking the Usk Valley, and to the south an extensive dip slope extends towards Tredegar and the South Wales coalfield. Dinantian limestones, about 120 m thick and dipping 2–5° SSE, form the main escarpment. The dip slope is capped by the Basal Grit, a strong, coarse-grained sandstone forming the lowest unit of the Namurian Millstone Grit Series; the thickness of the Grit cover progressively increases from the escarpment edge to about 30 m where it gains a cover of Namurian shales.

The dip slope has a spectacular assemblage of dolines. Over 500 dolines are concentrated in an area of less than 10 km^2 (Figure 6.13). The largest of the dolines is 55 m wide and 17 m deep; an average diameter of 29 m (Thomas, 1974) is significantly greater than that of typical solutional dolines formed in limestone. Most have a roughly symmetrical, inverted conical cross-section. Thomas (1954) investigated 437 collapse dolines, across the whole of the South Wales interstratal karst belt, and found that nearly 75% of them have a depth:diameter ratio of 1:3. The remainder include both steep-sided, rocky dolines, and also broad and shallow depressions with saucer profiles. The doline sides have an average slope of about 30° on a veneer of Grit blocks and solifluction deposits; exposed rock walls form only part of the perimeters of about 30% of the dolines. The steeper funnel-like dolines dominate the Llangynidr interstratal karst, but there are also seven much larger subsidence basins. These are oval shaped, mostly no more than 5 m deep, and are up to 150 m across. One basin contains the lake of Garn Fawr, and the others have sediment floors pitted with smaller dolines. The northern edge of the escarpment, overlooking the Claisfer Valley, has an area of more than 12 ha of foundered Basal Grit on the edge of the limestone outcrop. Beyond this, collapsed and soliflucted Grit debris overlies the limestone in a zone about 200 m wide which is almost continuous along the edge of the solid outcrop of the limestone.

There are two small caves on Mynydd Llangynidr (Figure 6.13). Ogof Cynnes is a cave with 900 m of rift passages formed in the top beds of the limestone beneath the Grit cover. The entrance passage has a roof of Basal Grit, and the complex of narrow rifts are plastered with mud and blocked by Grit boulder chokes; there are no large collapse chambers in the cave. The cave passes beneath the floor of the entrance doline, and the passages end in chokes almost beneath adjacent dolines in the Grit (Glover, 1993). Ogof Fawr (Chartist Cave) is another small cave close to the upper boundary of the limestone; it also has a Grit roof at its entrance, and leads to rift passages and three spacious collapse chambers, which have no surface expression. Over 5 km of cave passages in Carno Adit Cave lie in the Dowlais Limestone, at a depth of about 100 m beneath the eastern edge of Llangynidr (Figure 6.13); they appear to be the downdip continuations of influent caves including Ogof Cynnes, but are only accessible from the drainage adit and are not directly related to the surface features (Gascoine, 1991; Bailey, 1992; Rogers, 1992).

Some of the dolines within the Basal Grit have small streams sinking within them, but the main surface drainage is to the south. Water also sinks at the Grit/limestone boundary, and underground drainage resurges at both Ffynnon Shon Sheffrey, in the Trefil valley (Figure 6.13), and Ffynnon Gisfaen in the Clydach Gorge, 6 km to the south-east (Gascoine, 1989).

Interpretation

The doline fields seen on Mynydd Llangynidr are interstratal karst landforms, where subsurface solution has induced deformation and collapse of the overlying cover rocks (Thomas, 1974). Interstratal karst occurs along much of the adjacent Lower Carboniferous outcrop, but is particularly well developed on Mynydd Llangynidr, where the dip slope gradient is very close to the regional dip, so that the cover of Basal Grit remains thin across a broad belt.

Where the Basal Grit is thinnest, along the northern margin of the interstratal karst belt, solutional erosion is at a maximum; groundwater

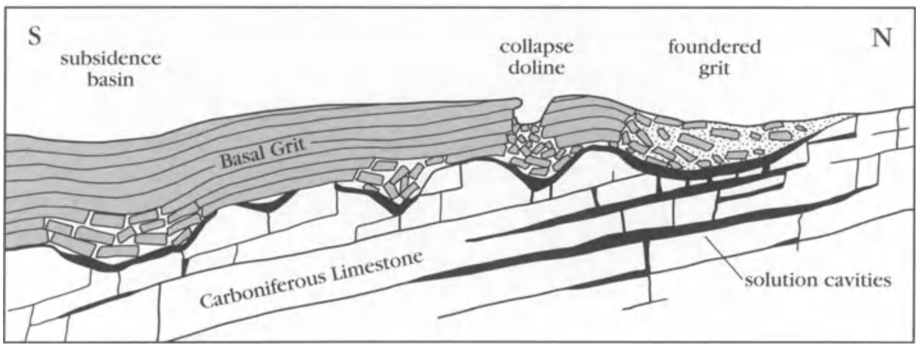


Figure 6.14 Diagrammatic cross-section through the three types of surface depression formed in the Basal Grit due to solution of the limestone beneath (after Thomas, 1974).

recharge occurs through the thin and broken cover, and the high hydraulic gradient near the scarp crest maintains underground flow. In this zone of thin cover, solutional cavitation of the limestone causes collapse of the Grit which is transmitted directly to the surface, resulting in formation of the steep-sided collapse dolines (Figure 6.14). Processes in this zone are comparable to those observed in shafts beneath the Grit close to the Mellte Valley (Burke and Bird, 1966; Burke, 1967).

Further down the dip slope, the Grit cover is thicker, and lower fracture permeability at depth reduces the solutional activity. Collapses into solutional cavities within the limestone of this zone is generally confined to the lower beds in the Grit sequence. Repeated collapse expresses itself on the surface as a shallow basin caused by sagging of the uppermost beds (Figure 6.14). Thomas (1954) estimated that in one block covering 6 ha, the total volume of the dolines was 180 000 m^3, equivalent to solutional removal of a continuous bed of limestone 3 m thick. Battiau-Queney (1980) suggested that the widespread rotting of the Basal Grit resulted from the solution of the silica under a deep regolith cover in the warmer climates of the Tertiary, and limestone solution only followed when uplift allowed circulation through the rotted zones; though the history of the solutional activity may remain open to debate, the dolines are clearly the result of cavity development within the limestone. The maze of rift passages in the top limestone beds in Ogof Cynnes, and the collapse chambers in Ogof Fawr, indicate the style of cavity development which ultimately will cause undermining and collapse of the Grit cover. Neither of these has developed to the point where collapse of the cover rocks can be seen underground as clearly as in Siambre Ddu (see below).

Along the northern edge of the escarpment, the limestone has undergone solutional erosion, and the Grit forms only a minor escarpment. There have been several phases of subsidence and collapse of the overlying Grit, which in places is so distorted that little of the original structure remains; these areas have been mapped as 'foundered Basal Grit' (Figure 6.14). However, some of these grit outliers, outside the Llangynidr karst, may be due to faulting, rather than solutional subsidence (Christopher and Smart, 1989). The age of the initial karstification may be estimated from measured solution rates and the extent of solutional lowering of the foundered Grit masses; maximum lowering has occurred outside the Llangynidr area and have implied ages of 10–15 Ma (Thomas, 1963), but these figures take no account of acidic waters from pyrite oxidation, climatic change and focusing of the drainage as the depressions evolve.

On Mynydd Llangynidr, the pattern of collapse dolines may reflect the form of the irregular unconformable interface between the Basal Grit and the underlying limestone (Thomas, 1974). Solutional activity may have been concentrated to produce major caverns along the bases of depression within the plane of unconformity; many caves, including Ogof Cynnes, are developed at the unconformity. There is no clear relationship between the collapse dolines of Llangynidr and collapse features directly underground in the few known cave systems. On the adjacent Llangattwg plateau, direct links can be traced between the surface collapses and boulder chokes in Ogof Agen Allwedd (Bull, 1977), but similar links are absent from Ogof Cynnes. On Llangynidr, the Basal Grit is very strong and forms extensive roof spans at a number of points in this cave. Once a collapse doline has formed, it acts as a focus for groundwater recharge, often by highly acidic peat bog run-off, thus perpetuating the solutional development.

Many features of the interstratal karst of Mynydd Llangynidr merit further attention. These include the effect of river rejuvenation and lowering water tables (Crowther, 1989), and examination of the nature, extent and age of some of the subsidence features, especially the foundered Grit masses north of the main outcrop.

Conclusions

Mynydd Llangynidr is Britain's finest example of interstratal karst. Doline fields within the grit outcrops are a special feature of the karst on the gently dipping escarpments of South Wales, and Llangynidr has the densest, largest and most spectacular assemblage of dolines. Collapse dolines, broad shallow subsidence basins and large masses of foundered grit all occur within a small area.

MYNYDD LLANGATTWG CAVES

Highlights

The caves beneath Mynydd Llangattwg form one of the two most extensive integrated systems in

Britain, containing exceptionally large karst conduits containing important sediment sequences. The network of ancient and active passages records a very long history of karst drainage modification in response to surface downcutting.

Introduction

The Llangattwg cave systems include Agen Allwedd, Daren Cilau and Craig a Ffynnon, all lying beneath a moorland escarpment at the eastern end of the North Crop (Figure 6.1). Mynydd Llangattwg is a peat-covered upland formed by the Basal Grit dip slope overlying the Carboniferous Limestone; it is almost a plateau, as the dip is only 2-3°. The Grit dips south beneath a cover of Namurian shales and sandstones followed by the Lower Coal Measures. Scarp faces to the north and east have steep screes and quarried crags in the limestone, overlooking the Old Red Sandstone floor of the Usk Valley. Along the southern margin of the plateau the River Clydach flows east through a steep gorge tributary to the River Usk. Input drainage to the caves is through numerous small sinks and percolation into the outcrops of both the limestone and the fractured and permeable Basal Grit. Pwll y Cwm is the largest of the multiple resurgences along the floor of the Clydach Gorge.

All the main cave passages are developed in the 50 m of Chadian oolitic limestones and dolomites within the Abercriban Oolite Group: the most cavernous unit is the Blaen Onneu Oolite (Wright, 1986a). These are overlain by the thin impermeable mudstones of the Arundian Llanelly Formation, and the Holkerian Dowlais Limestone which contains very little explored cave under Llangattwg. Palaeokarstic horizons, with associated mudstones or palaeosols, are developed at several levels in the oolites, and also at the junction of the Dowlais Limestone with the overlying Basal Grit. The limestone dips 2-3° to the southwest, with minor flexures producing local dips up to 15°. The area is traversed by a number of faults, mostly trending ENE or NNW, and there are well developed joint sets trending NNW and NNE.

A wealth of publications refer to Agen Allwedd, but the main passages of Daren Cilau have only been explored since 1984 and therefore largely await the scientific study that they warrant. The cave geomorphology has been comprehensively assessed by Smart and Gardener (1989), and the passage details and hydrology were described by Stratford (1995), Stevens (1992) and Gascoine (1989). Detailed accounts cover the succession of major new discoveries in Ogof Agen Allwedd (Leitch, 1960, 1973; Jenkins, 1963; Gardener, 1983; Tomalin, 1987; Abbot and Murgatroyd, 1988; Price, 1988), Ogof Daren Cilau (Gardener, 1984, 1985, 1986; Gardener and Westlake, 1985; Farr, 1985, 1986, 1993) and Ogof Craig a Ffynnon (Parker, 1978; Gascoine, 1979). Agen Allwedd has also been a prime site for research on the environments and paleomagnetism of clastic cave sediments (Bull, 1975, 1976a, 1978, 1981; Noel *et al.*, 1979, 1981; Noel, 1983, 1986, 1988).

Description

More than 65 km of cave passage have been explored under Mynydd Llangattwg, in three major cave systems, Ogof Agen Allwedd, Ogof Daren Cilau and Ogof Craig a Ffynnon, together with several smaller caves and isolated sections of larger systems (Figure 6.14). Morphologically these constitute a single cave system, but links have not yet been explored between the three main caves. Each has its own single dry entrance, though Daren Cilau can also be entered through the flooded passages at the resurgence.

Ogof Agen Allwedd

This is the most westerly of the three major caves, was the first to be explored, and now has more than 34 km of mapped passages (Figure 6.15). Its main stream drains from the Remembrance Series under the northern tip of Mynydd Llangattwg, along Turkey Streamway and the lower part of the Main Streamway, into Maytime Series where active phreatic loops link sections of vadose streamway. The water then flows into the lower reaches of Daren Cilau. The streamway steadily increases in size downstream; many sections are 5-10 m high and wide, and the downstream passage is a massive phreatic tunnel now partially drained. Vadose trenches are discontinuous as they are mainly incised through the crests of low phreatic loops.

The major active tributaries are two vadose streamways which drain in from the north. The first is a stream from the entrance in the northern escarpment, which collects other inlets and forms the upper part of the Main Streamway, above the junction with the larger Turkey Streamway. The second is the long narrow canyon of Southern

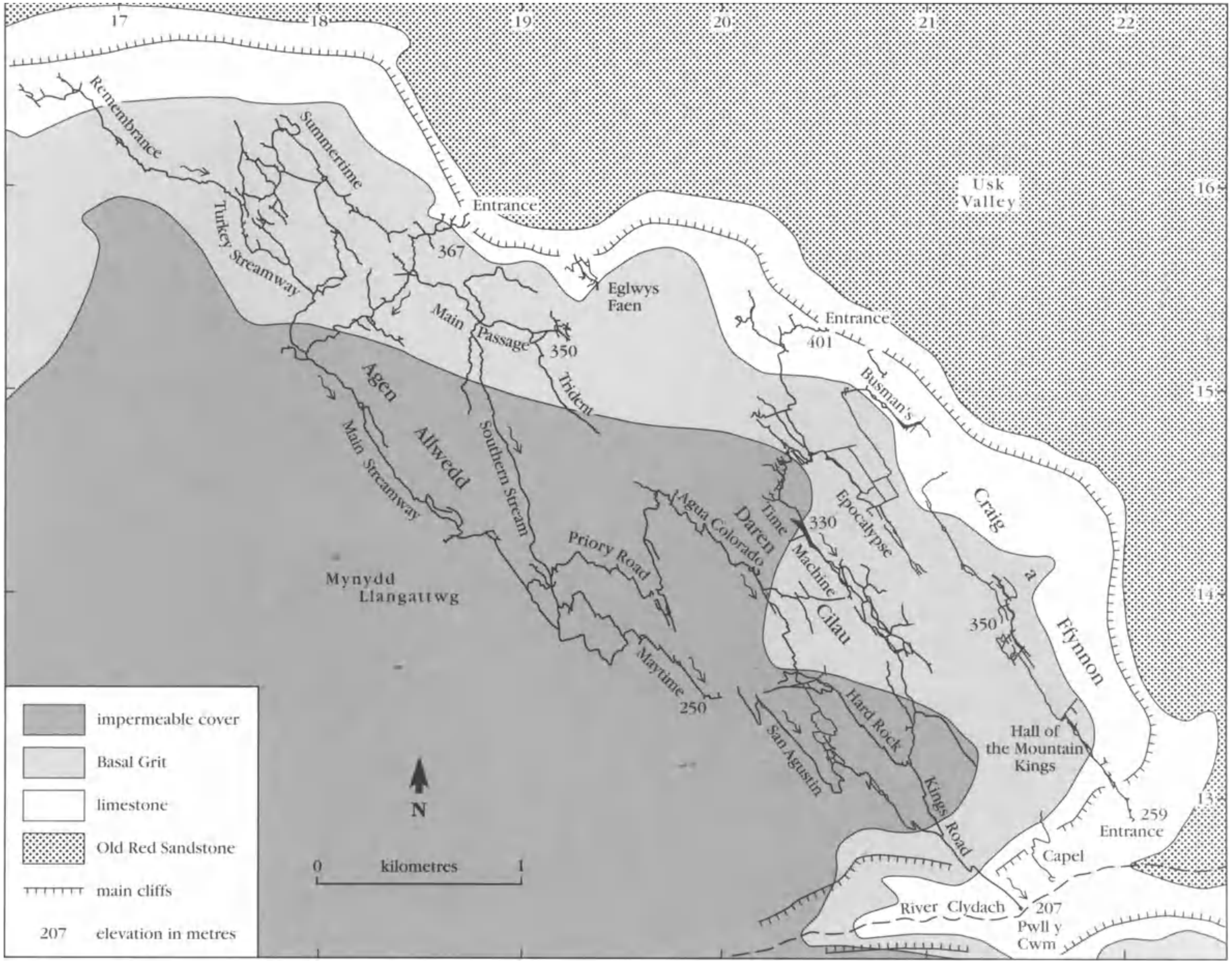


Figure 6.15 Outline map of the caves under Mynydd Llangattwg (from surveys by Chelsea Speleological Society, British Speleological Association, Cwmbran Caving Club, Cave Diving Group and others).

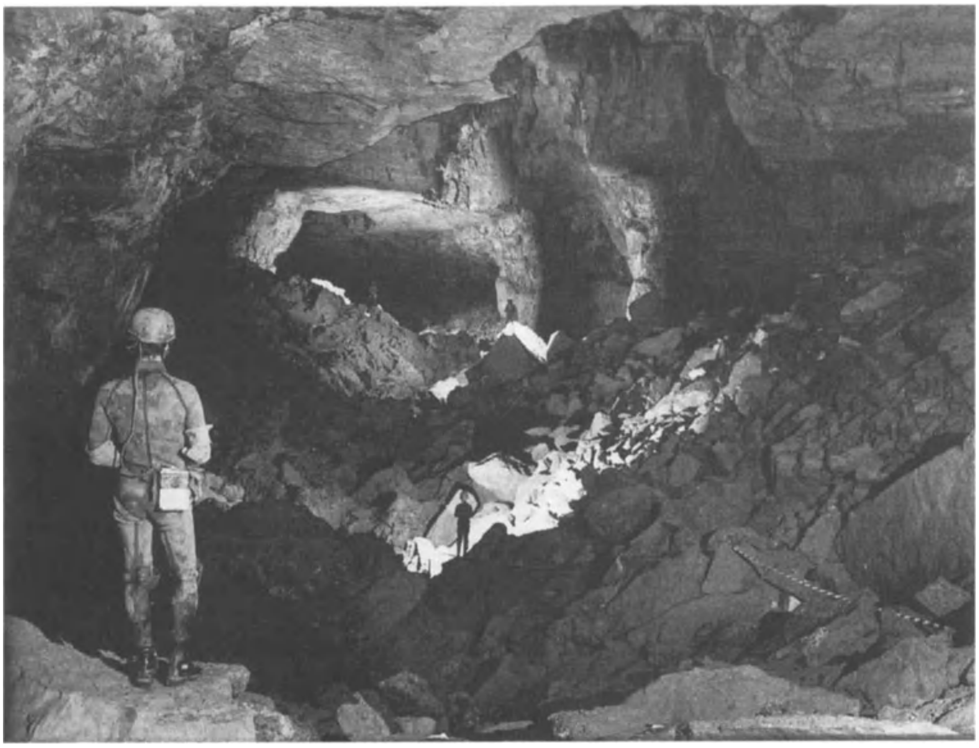


Figure 6.16 The Time Machine in Ogof Daren Cilau, the largest cave passage in Britain. (Photo: C.D. Westlake.)

Stream Passage. The active streamways provide the links between the Main Streamway and the major abandoned passages which lie further north-east. Largest of these are the old phreatic tunnels of Summertime Series and Main Passage, which are mostly 10-20 m high and wide. They are now partly blocked by enormous collapse piles, boulder chokes, and thick sequences of clastic sediment; deep profiles have been cut through the fills, notably at the south-eastern end of Main Passage. Smaller passages off these trunk routes are numerous and at various levels, and some connect with fragments of other large trunk passages; the Trident and Priory Road tunnels are blocked with sediment before they reach the passages of Daren Cilau.

Thick clastic sediments are conspicuous in much of Agen Allwedd, and completely fill some of the older passages. The fill in Main Passage includes a capping horizon of fine lacustrine silt, whose laminations are remarkably consistent along the passage (Bull, 1981). Calcite speleothems are present at only a few places in the cave, as inflows of percolation water are severely curtailed by the impermeable cover rocks over these deeper zones of the escarpment (Figure 6.15). However, there are very fine stalactites, extensively overgrown by helictites, in the high-level passage into Maytime. Mud formations with drip features are more widespread in the old passages, and some clay banks are thickly covered by bladed gypsum crystals. In a few places, gypsum crystals have caused tabular roof collapse by growing in the limestone bedding planes.

Ogof Daren Cilau

The 30 km of passages explored from the Daren Cilau entrance lie at the heart of the Llangattwg cave system, though its passages are currently isolated from the continuations in Agen Allwedd and Craig a Ffynnon by the incidental distribution of impenetrable boulder collapses and sediment chokes. The main caves of Daren Cilau consist of sections of very large old conduit linked by smaller passages, both active and relict (Figure 6.15). Sections of three streamways are encountered, and all eventually converge to enter the terminal sump, where an active phreatic loop extends to the Pwll y Cwm resurgence; this loop reaches a depth of 217 m below the highest point in the cave.

The Entrance Series has a small vadose passage into a rift and a short section of large, old passage blocked at both ends by collapse. Beyond this, a series of rifts and muddy, abandoned passages extends to a large chamber and a major junction. To the east, Epocalypse Way is a major old trunk route, up to 8 m wide, with smaller vadose passages leading off into Antler Passage and Busman's Holiday. To the west, a complex series of rifts and high-levels leads to the Time Machine. This is a massive tunnel so heavily modified by vadose undercutting and block collapse that its phreatic origins are barely recognizable (Figure 6.16); much of it is 30 m high and wide, making it the largest cave passage in Britain. The Time Machine divides at its southern end, but the main route continues into the high-level Kings Road, beyond the Hard Rock series of small rifts and larger chambers entering from the west. Kings Road continues to the lower streamway which feeds to the Terminal Sump into the flooded connection to the resurgence. From this open streamway, several flooded sections guard the major inlets from the west. Water from Agen Allwedd drains down the large vadose streamway of San Agustin, but a short flooded section has not yet been explored to make the final link. The much longer tributary inlet of Agua Colorado carries a smaller stream, and near its head provides access to high-levels which reach chokes very close to the similar Priory Road passage in Agen Allwedd.

Thick sequences of clastic sediments occur in many of the large old trunk passages of Daren Cilau. Calcite speleothems are restricted beneath the impermeable cover on the plateau, but are more common in the eastern sector. Passages around Epocalypse Way contain some very fine displays of straws, helictites, stalactites and stalagmites, beside a spectacular profusion of multi-coloured aragonite anthodites up to 40 mm long (Kendall, 1988).

Ogof Craig a Ffynnon

Lying east of Daren Cilau, nearly 9 km of caves in Ogof Craig a Ffynnon are reached through a passage truncated in the side of the entrenched Clydach Gorge (Figure 6.15). A first section of large old abandoned tunnel ends at the Hall of the Mountain Kings, but much of the continuation breaks into series of long, straight, parallel rifts, some of which link down to short sections of active streamway. The cave is broken by five areas of massive collapse; some of these contain large blocks of sandstone and lie directly below collapse dolines on the moor over 100 m above. Craig a Ffynnon now lies well clear of the impermeable cover on the limestone, and percolation drainage has deposited many excellent calcite speleothems; the older passages also contain extensive clastic sediments.

Minor caves

There are few open stream sinks on the plateau moorland. Llangattwg Swallet takes the largest flow into a collapse doline in the cap of Basal Grit. Other dolines reach down into the Dowlais Limestone, but none can be followed to depths greater than 40 m. Sediments in Pwll Gwynt were deposited with reversed magnetic polarity, indicating an age greater than 780 000 years. Of the smaller caves around the escarpment rim, Eglwys Faen is the longest with passages which start large, only to terminate in major collapses. In the Clydach Gorge, the most important caves are the Pwll y Cwm resurgence and its flood overflow through the narrow rifts of Elm Hole, and Ogof Capel, a vadose streamway well decorated with calcite dripstone (Gardener, 1988).

Interpretation

The overall drainage trend in the Llangattwg limestone is to the south-east, to resurgence positions determined by the surface topography. This direction is almost across the low SSW dip. There has, therefore, been a long history of vadose inlet drainage roughly downdip to join phreatic trunk conduits along the strike; the latter have shifted progressively downdip in response to valley downcutting and lowering of resurgence levels, leaving parallel abandoned caves north-east of the active drains. Both vadose and phreatic components of the cave system have been constrained to a zone of limestone only 50 m thick, wherein they have been subject to strong directional control by tectonic fractures. The influence of joints and faults on the passage details is evident in the rectilinear patterns in many parts of the cave system. The strong influence of two sets of joints on a strike phreatic passage is clearly seen in the modern stream route into Maytime; the cave's zigzag pattern through the joint grid includes downdip loops which are still flooded between updip segments which are now drained.

Palaeokarst and palaeosol horizons have

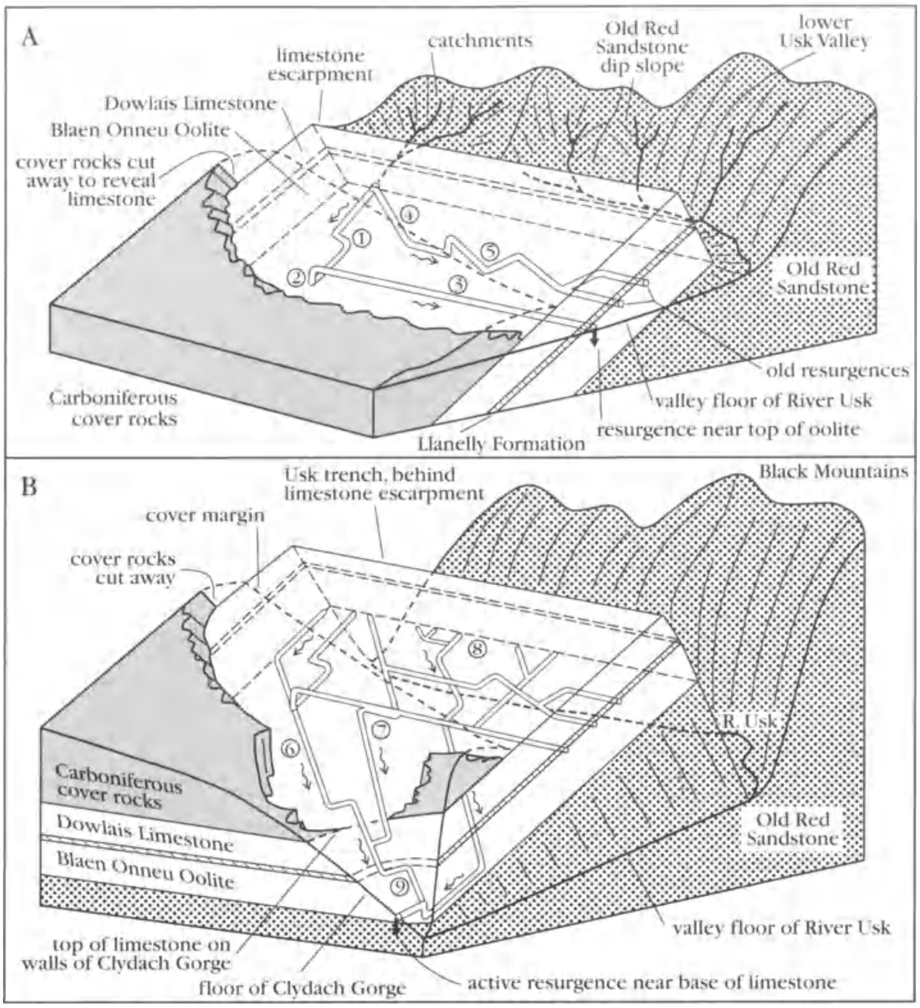


Figure 6.17 Block diagrams showing stages in the evolution of the Llangattwg caves. (A) In the early stage, the Usk headwaters drain into the cave: 1 = vadose flow downdip; 2 = phreatic lift taking water from the base to near the top of the oolites; 3 = phreatic strike flow; 4 = vadose flow on joints; 5 = phreatic conduit looping along joints. (B) In the later stage, the limestone escarpment is perched above the Usk trench: 6 = main drainage on joints with greater downdip component; 7 = older passages invaded by vadose streams; 8 = vadose inlets from limestone outcrop and from glacier melt during Pleistocene; 9 = limited phreatic development behind resurgence. (After Smart and Gardener, 1989.)

influenced the stratigraphic positions of passage development, acting as aquicludes or inception horizons and also limiting the extent of upward collapse stoping; hence many passages are roofed by these beds. Stratigraphy has also exercised broader controls. Inlets to the overlying Dowlais Limestone drain along the shales and calcite mudstones of the Llanelly Formation for considerable distances before finding routes through fractures into the cavernous limestones below. The impermeable cover, of Namurian and younger clastics, has largely excluded percolation water from the caves, and there are few calcite and aragonite speleothems, except in the uncapped passages of Craig a Ffynnon. Some very delicate speleothems appear to have been formed by evaporation of water which was intergranular seepage through the Grit. The Namurian shales are pyritiferous, and may have been the source of the sulphate which has formed so much gypsum in parts of Agen Allwedd.

Smart and Gardener (1989) interpreted the isolated sections of very large trunk passage as parts of ancient systems of essentially strike-orientated phreatic conduits. Major streams flowing off the Old Red Sandstone dip slope entered sinks into vadose passages which drained downdip, or obliquely along the joints, to feed the trunk cave drainage; this was orientated ESE towards presently unknown resurgence sites along the eastern outcrop of the limestone (Figure 6.17). The main sinking drainage was then captured by the River Usk whose valley was deepened, fluvially and glacially, along the outcrop of the softer rocks around the Devonian/Carboniferous boundary. The beheaded trunk passages were largely choked by collapse and sediment infill, while new, smaller vadose caves developed obliquely downdip under joint control, and in many places cut across the older, strike-orientated, phreatic trunks. New trunk drains at lower positions, down the dip, were originally phreatic, but resurgence lowering left them in the vadose environment except for the lower phreatic loops. They have since undergone extensive vadose modification, with considerable entrenchment at the crests of phreatic loops. The vadose passages developed in the lower beds of the limestone, and some have subsequently entrenched into the underlying shales, while the phreatic caves had lifting segments which took them and the early resurgences to the top of the oolitic limestones just beneath the Llanelly Formation (Figure 6.17).

The complexity of the Llangattwg cave systems, with the variation in size and morphology of their passages, reflects a long and complex history extending well back into the Pleistocene. The very large sizes of the old trunk caves is incompatible with the modern drainage to the limestone. The oldest passages were fed by headwaters of the River Usk, but this river was then entrenched below the plateau sinks. Glaciers breached the Old Red Sandstone escarpment to flow down the Usk Valley: they could have supplied large flows of sediment-charged, aggressively acidic meltwater to the high-level marginal sinks in the limestone. Some of the Llangattwg caves may therefore be unusual in having been active largely during the glacial stages of the Pleistocene. During the interglacials, water charged with biogenic carbon dioxide had less effect because of the very small flows reaching the elevated limestone outcrops. This concept is supported by the glacial ancestry of much of the cave sediment (Smart and Christopher, 1989; Bull, 1976a, 1980), but the sediment sequences in the main caves have not yet yielded a framework of absolute dates.

While the upper end of the cave system was being modified in response to entrenchment of the Usk, the lower end was rejuvenated as the Clydach Gorge retreated into the limestone escarpment. The ages of the earliest trunk passages are not yet known, but they appear to predate much of the glacial excavation of the Usk Valley, and probably date from the early Pleistocene. Early resurgences could have been close to the mouth of a proto-Clydach valley. The chronology of entrenchment of the Clydach Gorge is debatable. South of the Clydach, Ogof Draenen contains some old abandoned cave passages which were formed by northward drainage into the Clydach Gorge, thus indicating its considerable age (see Figure 6.18). Draenen has no passages large enough to represent the downstream continuations of the old Llangattwg trunk conduits, truncated by a younger Gorge, but these could lie downdip of the known cave system. The age of the Clydach Gorge therefore remains uncertain, but current evidence suggests that it is older than most of the caves.

Unlike other systems of comparable size and complexity, such as the Ease Gill caves and Ogof Ffynnon Ddu, the Llangattwg cave system appears to have experienced a major change in drainage pattern at some point in its development. Establishing the nature and cause of this change is fundamental to understanding the geomorphological history of this area.

Conclusion

The caves of Llangattwg form one of the most extensive systems in Britain, which developed largely beneath an impermeable cover. They record the evolution of a massive karst drainage system over a very considerable timespan, during which the surface drainage underwent major changes and captured much of the early sinking water. Subsequently these caves were modified by smaller percolation flows and possibly by episodic invasion by glacier marginal drainage. The much greater underground flows of the past formed trunk drains, one of which is now the largest single cave passage in Britain. The old caves contain exceptional clastic sediment sequences of great stratigraphic value, and some unusual speleothems of calcite, aragonite and gypsum.

OGOF DRAENEN

This is a proposed GCR site,
of which only Siambre Ddu is currently
a designated SSSI

Highlights

Ogof Draenen is a major cave system developed in Carboniferous Limestone almost entirely beneath its own allogenic catchment on the overlying Coal Measures. Large passages record an unusually clear example of flow diversion related to differential rates of incision in two adjacent valleys. Abandoned phreatic conduits contain the finest examples in Britain of solutional wall notches associated with sediment deposition or ponding. The adjacent cave of Siambre Ddu is Britain's finest example of a chamber which is an element of interstratal karst, formed by solution directly beneath a cover of Millstone Grit.

Introduction

Ogof Draenen is developed beneath the escarpment and moorland extending south-east of the Clydach Gorge (Fig 6.1). The outcrop of the limestone is confined to a narrow sinuous strip along the steep, north-facing scarp face of the escarpment and an elongate window in the Afon Lwyd valley south of Blaenavon. Most of the cave lies beneath the outcrop of the cover rocks; less than 10 m of Namurian Basal Grit is overlain by thin shales and sandstones, also of the Namurian, and a much greater thickness of Westphalian Coal Measures. The cave passages of Draenen are developed at numerous levels throughout almost the entire thickness of the Abercriban Oolite Group. These Lower Dinantian limestones are 50 m thick on Gilwern Hill, but thin to 30 m on the Blorenge, due to progressive eastward overstep of the unconformable Basal Grit cover. Other than the massive, bioclastic Gilwern Oolite at the top of the sequence, much of the carbonate is extensively dolomitized and some is thinly bedded. The dip is 4° south-west, and the dominant joint set is NNW–SSE, nearly parallel to several small faults.

There are no major sinks feeding the cave, but many of the streams draining across the escarpment have at least part of their flow captured as they cross the limestone outcrop. The largest risings proven to take flows from Ogof Draenen are at Snatchwood and Pontnewynydd, both 7 km south of Blaenavon down the Lwyd Valley and just north of the Trevithin Fault. Some cave water drains to small risings in both Cwm Llanwenarth and Cwm Dyar.

Ogof Draenen was discovered late in 1994, and the only published descriptions are those by the explorers (Bolt *et al.*, 1994, 1995) and one brief discussion of the geomorphology (Simms *et al.*, 1996). The hydrology is discussed by Gascoine (1994), and Siambre Ddu is briefly referred to in Stratford (1995), Thomas (1974) and Gascoine (1982).

Description

The entrance to Ogof Draenen is a small phreatic tube developed high in the Gilwern Oolite and truncated by the hillside above Cwm Llanwenarth. It is the only access to more than 48 km of mapped cave passages (Fig 6.18). The entrance series descends to Tea Junction, where it meets the long passages aligned almost on the strike on the western side of the system. To the south, Beyond a Choke is a major active streamway; the large phreatic passage is increasingly entrenched to the south, with a narrow, locally joint-guided, vadose canyon more than 30 m deep beneath an almost horizontal phreatic tube. Towards its southern end the phreatic tube descends to stream level in a series of loops. Strewn with boulders in the northern section, it has a fine sediment floor and deep pools towards the south. North of Tea Junction, Gilwern Passage is a major abandoned conduit with several tributary passages from complex areas of rifts and collapse; its underfit stream now drains to a rising in Cwm Dyar. The Gilwern Passage series contains some fine speleothems, including calcite rafts; thick sediment sequences contain false floors, clastics with epsilon cross-bedding, numerous coal clasts, and well preserved current ripples which indicate northward flow.

A complex series of mainly abandoned passages lies east of Beyond a Choke (Fig 6.18). Upstream Passage contains thick deposits of laminated sands and clays similar to those in Main Passage of Agen Allwedd; they drain south, but the current bedding indicates flow to the north, and the sediment fill may be from temporary glacial inputs at passage breaches, which were remote from the sinks at the head of the stream caves. A minor inlet extends north to give access to more than 2 km of abandoned high-level passages in Waterfall Series. South of the entrance series, Lamb and Fox Chamber is a large collapse feature extending into the upper part of the Abercriban Oolite; its walls are undercut by an exceptionally wide solutional notch, and have sands and gravels preserved on remnant ledges. Two vadose canyon passages drain towards the south; the eastern passage has a small phreatic tube in its roof and extends to a further junction. The Canyon is a large paragenetic canyon passage with a vadose trench incised

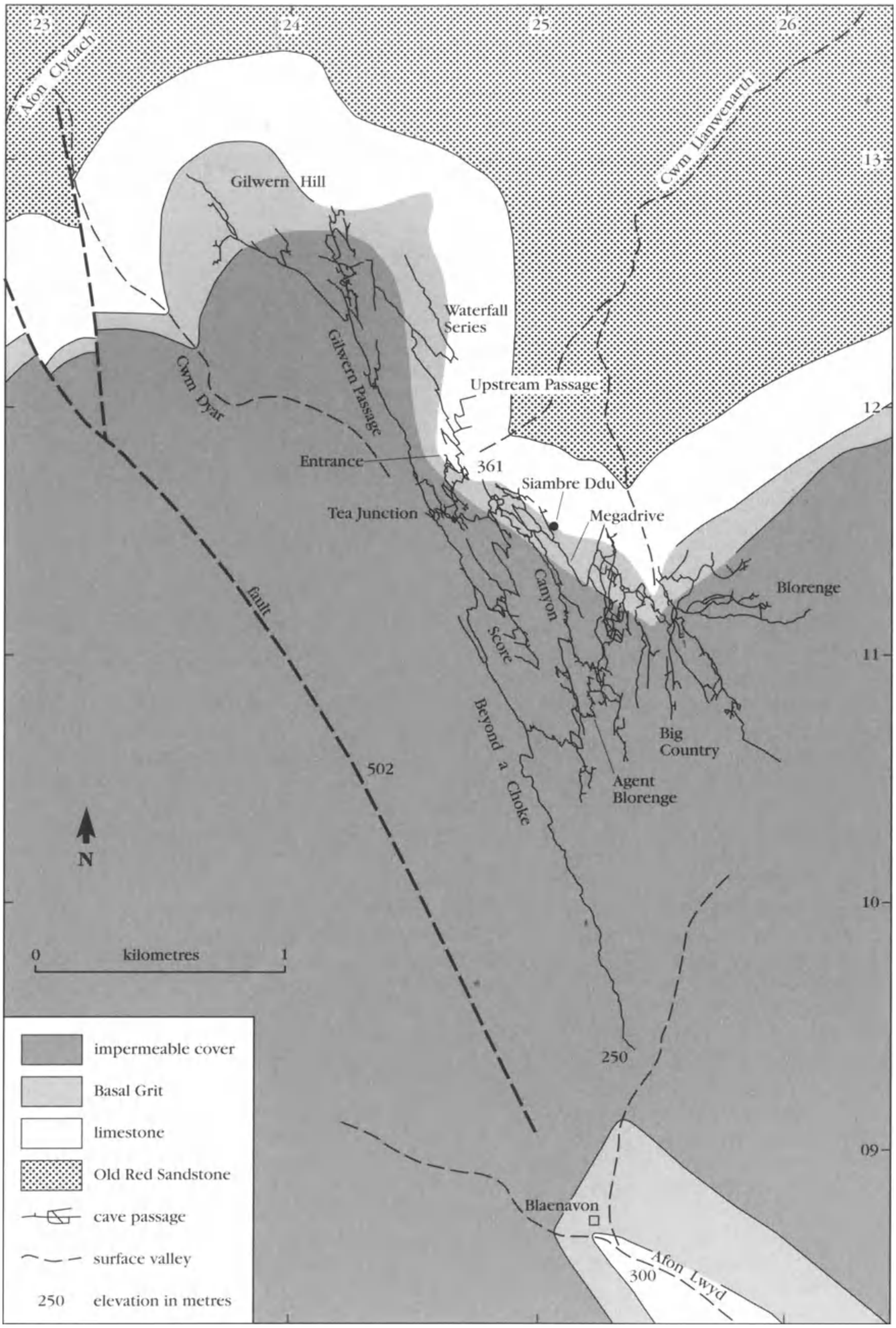


Figure 6.18 Outline map of Ogof Draenen; this is only a centreline plot of the cave, with no indication of passage widths (from survey by Chelsea Speleological Society).

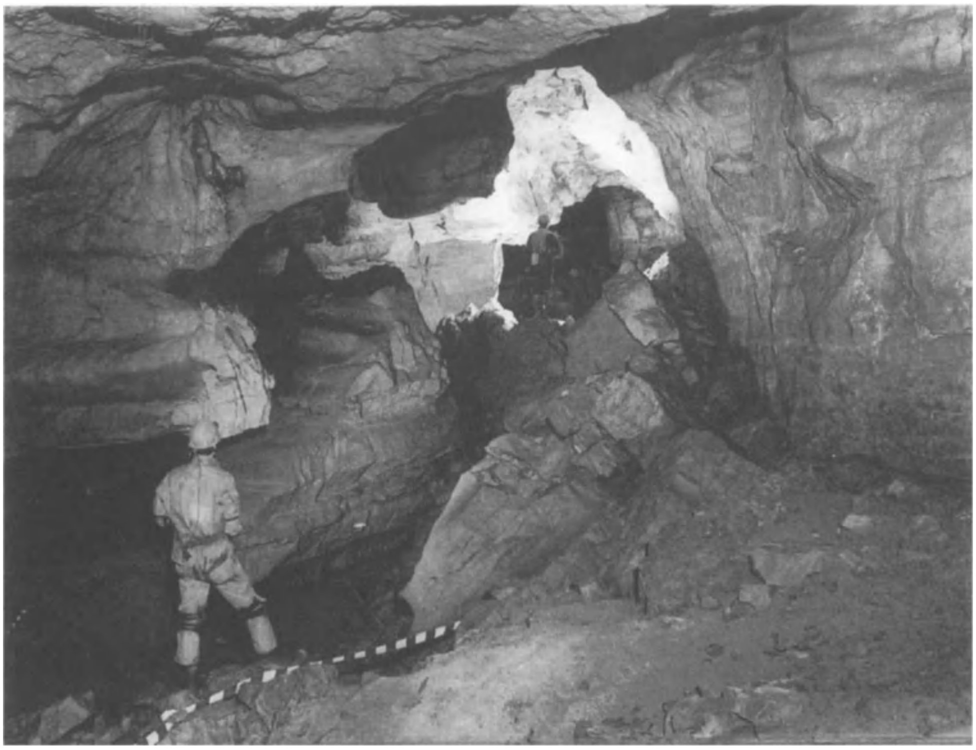


Figure 6.19 The abandoned conduit of Megadrive in Ogof Draenen, with preserved solutional notches along the wall on the left. (Photo: C.D. Westlake.)

beneath. Megadrive is a major abandoned phreatic conduit with a meandering vadose trench incised several metres into its floor; it contains extensive sediment banks, has very fine solutional notches along its walls (Figure 6.19) and extends to a large choked chamber close behind the scarp face.

A complex series of abandoned passages south of Megadrive includes St David's Hall, the largest collapsed chamber yet found in Ogof Draenen, and connect to the active streamway of Agent Blorenge. This descends steeply via a series of cascades into a trench incised more than 30 m beneath the original phreatic tube; it joins the lower part of the Beyond a Choke trunk streamway. Abandoned phreatic conduits in the Elliptic Passage Series extend east of Megadrive; these connect with passages extending beneath Blorenge to the Big Country, where tributaries unite in a passage which is up to 10 m high and 20 m wide and drains to the south.

Calcite speleothem development is very limited through most of the cave, due to the lack of percolation water beneath the impermeable cover rocks. Gypsum crystals are locally abundant on walls, fallen blocks and sediments in some abandoned passages. Helictites and anthodites are spectacularly abundant in some of the abandoned passages. Accumulations of bat guano are significant in some passages near the scarp face where old entrances are now blocked. Bradyodont fish teeth and fin spines are commonly etched in relief on passage walls in the richly bioclastic Craig-y-Caer Coral Bed at the base of the Gilwern Oolite.

Siambre Ddu lies in the scarp face east of the Draenen entrance (Figure 6.18), with a short entrance passage leading to a single chamber 25 m in diameter and 10 m high. The roof is an almost flat bedding plane in the strong sandstone of the Basal Grit, though early stages of collapse of individual beds has started to modify this into an arched profile. A thick shale is exposed in the walls beneath the Grit, and the underlying limestone is obscured by the collapse debris which litters the floor. This includes fragments of a fossil tree trunk of *Lepidodendron*, whose cast, 6 m long, survives in the roof. The chamber roof and walls, and many fallen blocks, are coated with a soft black deposit, which is high in manganese (Gascoine, 1982) and is probably a mixture of oxides and humates. Siambre Ddu lies directly above some dripping avens, floored with Grit boulders, in a side passage off Megadrive in Ogof Draenen.

Interpretation

The main passages of Ogof Draenen are largely aligned close to the strike in a thin unit of gently dipping limestone. Successive conduits have developed due to progressive downdip shift of the drainage in response to the availability of lower resurgence sites in new outcrops of the limestone exposed by surface lowering. The uniform alignment of the Gilwern and Beyond a Choke Passages also reflects the orientation of the dominant joint set parallel to a strike-slip fault. Most of the passages are developed along joints adjacent to the fault, suggesting that the latter is relatively impermeable. A minor fault is crossed in one chamber which is modified by collapse, but the lack of collapse throughout such a great length of passage adjacent to a fault is remarkable. Lithological factors have influenced the cave passage shapes. Passages in the Gilwern Oolite and other massive units have largely retained their original solutional profile, while those in more thinly bedded carbonates lower in the succession have been greatly modified by collapse.

Passages in Ogof Draenen are developed at multiple levels within the limestone, and show a complex pattern of flow. Passages in the lower part of the Abercriban Oolite were formed by flow either to north or south, or in some cases to both. In the Gilwern Oolite almost all passages, including Megadrive, which is the highest and oldest trunk conduit in the system, have scallops indicating flow to the south. The fossil passages in Waterfall Series appear to represent an upstream extension of Megadrive, but the intervening section has been removed by retreat of the escarpment face. An abundance of coarse, angular sediment derived from the Millstone Grit suggests that the main sinks lay only a short distance to the north, but have since been destroyed by scarp retreat. Deposition of this sediment at the vadose-phreatic interface was responsible for promoting lateral dissolution and creating the spectacular wall notches in this part of the cave. If Megadrive continued along the same trend as the other main drains in the system, then the original high-level resurgence may have been located in Cwn y Nant, a re-entrant in the scarp south-east of Blaenavon. Indiana Highway, with its headward extension in Waterfall Series, and the Canyon represent downdip captures of the Megadrive flow, which was probably in response to a westward shift of the resurgence, either due to scarp retreat or to lower site in the Afon Lwyd valley.

Subsequently, White Arch Passage developed as a separate drain, lower in the limestone, extending north into Gilwern Passage, where scallops and sediment bedforms indicate consistent flow to the north. It was also fed by inlets draining north along the Score, perhaps derived from sinks on Blorenge. The underfit stream in Gilwern Passage now drains to a resurgence in Cwn Dyar.

The modern main streamway of Beyond a Choke is the lowest passage in the cave and drains south from Tea Junction down a very gentle gradient. A Choke marks the capture point where the earlier northward flow was re-routed towards the south. During the initial stages of this capture, water may have continued to flow north along Gilwern Passage as well as south along the precursor of Beyond a Choke. The preservation of current ripples which are unmodified or draped by mud suggests that the final abandonment of Gilwern Passage was comparatively sudden. Water continued to drain north along the Score after the capture, and supplied small capture passages westwards towards Beyond a Choke. The main capture was relatively recent, but sufficient time has elapsed for the remaining misfit streams in the Gilwern Passage series to entrench to the level of the main trunk passage. The deep and narrow vadose canyon in Beyond a Choke, and the comparable entrenchment in Agent Blorenge, suggests rapid incision since the abandonment of the northward flow route. Much of the Beyond a Choke streamway appears to be graded to current base level; it has a very low gradient with stretches of sediment floor and local ponding, but incision continues in the steeper, boulder-strewn descent of its upper reaches.

The flow re-routing from south to north then south again allows a sequence of scarp retreat and valley incision to be constructed for the cave catchment, though no dates for the cave sediments are yet available. In the earliest stage of cave development, water sinking along the eastern outcrop of the limestone, at altitudes of more than 380 m, drained southwards almost along the strike, until it could drain down the dip to limestone outcrops at altitudes around 310 m, exposed by differential scarp retreat. This stage predated the incision of both the Afons Lwyd and Clydach.

The second stage of development was associated with more rapid scarp retreat to the north, in the region of the present Clydach Gorge. The unroofing of the limestone at an altitude of about 250 m caused a reversal of the hydraulic gradient towards a new resurgence in the proto-Clydach valley. The new cave drainage to the north, via Gilwern Passage, was offset to the west of its southbound predecessor, as it shifted both downdip and into lower parts of the Abercriban Oolite. This stage of development in Ogof Draenen may be correlated tentatively with the development of Ogof Craig a Fflynnon, which still drains to a resurgence at a similar altitude on the opposite side of the Clydach Valley. The abundance of coal debris in the Gilwern Passage sediment suggests rapid stripping of overlying Coal Measures, perhaps associated with a poorly vegetated, periglacial environment. Remnants of cross-bedded sands in Upstream Passage are also the produce of a cold environment.

In the final stage of cave development, the main flow was captured southwards along Beyond a Choke, in response to incision of the Afon Lwyd towards the 120 m level of the present main risings; these were far below rising sites in the Clydach Gorge.

The present resurgence of drainage from the Llangattwg caves may lie at Pwll y Cwm only due to the breaching of a deep phreatic conduit by the incision of the Afon Clydach (Smart and Gardner, 1989). Prior to this breach, water may have continued southwards to an unidentified lower than any passages yet discovered in Ogof Draenen. The consistent northward flow in Gilwern Passage indicates that Ogof Draenen was not formed by water draining from a former eastward extension of the Llangattwg caves prior to incision of the Clydach Valley. The resurgences from Gilwern Passage and from Ogof Craig a Ffynnon lie at close to the same altitude, and may indicate comparable ages for the two caves. The older, high-level passages in Ogof Daren Cilau and Agen Allwedd may also have drained towards this resurgence unless they developed in response to the incision of the Agon Lwyd, providing a low-level outlet more than 10 km to the south. The Megadrive trunk conduit predates the main Llangattwg caves, but the second and third phases of development on Ogof Draenen may have been contemporary with the evolution of the Llangattwg drainage system, when the Afon Clydach became a major control on the regional cave development.

Dating of the caves is not yet possible. Early Pleistocene origins are indicated by the altitude of the large fossil phreatic conduits high above the modern base level and their truncation by the modern hillside. Deep vadose entrenchment of the Beyond a Choke main streamway and its Agent Blorenge tributary suggests that the presently active passages are also of considerable age. Dating of the cave sediments will establish a chronology for the flow diversions already recognized within Ogof Draenen; this will provide evidence for the relative ages of the Clydach Gorge and the Afon Lwyd valley, and also relative rates of scarp retreat in this region of South Wales.

The abundance of coal clasts in sediments throughout Ogof Draenen suggests that allogenic recharge to the narrow limestone outcrop has been from the west, draining directly from the Coal Measures which overlie much of the cave; there is no positive evidence for input from a former Old Red Sandstone catchment to the east, comparable to that on Llangattwg. Downdip drainage into the limestone beneath the Namurian clastics has produced the interstratal karst of which Ogof Draenen is a component.

Siambre Ddu is also a feature of the interstratal karst, where a cave chamber is undergoing progressive collapse and upward stoping. It lies only 15 m below the modern surface, and will ultimately form a collapse doline in the overlying Millstone Grit. Collapse of the cap rock into the solutional chamber in the limestone has blocked and obscured the shafts into the underlying cave passage, but comparison may be made with the shafts of the Mellte Valley formed by dripwater immediately beneath the Grit (Burke and Bird, 1966; Burke, 1967). Continuing stages in the evolution of Siambre Ddu may be compared to the massive Grit chokes in Craig a Ffynnon, under Llangattwg, and the doline fields of Llangynidr. Iron and manganese deposits are well developed in the Siambre Ddu chamber, and may be comparable with the complex iron minerals formed by oxidation of pyrite and deposited with peat in shafts in the Mellte Valley (Burke, 1970).

Conclusion

Ogof Draenen is a major cave system with a downdip sequence of abandoned and active, strike-aligned conduits. These have evolved through an environment of changing hydrology where flow from the central part of the cave was first to the south, then to the north, and finally to the south again, in response to the early incision of the proto-Clydach and the later incision of the Afon Lwyd. The scale of the flow reversals is unmatched elsewhere in Britain, and their ease and rapidity were largely due to the very low gradients in major cave conduits which were developed almost along the strike. Both Ogof Draenen and Siambre Ddu are components of a system of interstratal karst, and the latter represents an early stage in the progressive development of a collapse doline in the overlying Basal Grit.

OTTER HOLE

Highlights

Otter Hole contains a profusion of calcite stalactites and massive stalagmites on a scale unmatched by any other cave in Britain, and is the only major cave in the country located entirely within dolomites.

Introduction

Otter Hole lies in the west bank of the River Wye, just north of Chepstow (Figure 1.11). It has more than 3200 m of mapped passages developed entirely in the Lower Dolomite, a 100 m thick sequence of well-bedded dolomites and dolomitic limestones in the Courceyan stage of the Carboniferous Limestone. The main cave roughly follows the strike of the carbonates towards the east; dips are less than 10° to the south, and they swing around shallow flexures which plunge down the dip. The Vicarage Fault and the major joints are orientated NNW, parallel to the plunging flexures. Allogenic water drains into sinks along the boundary of the dolomite with the underlying Lower Limestone Shales, which separate it from the Old Red Sandstone to the north and west. Drainage through the cave flows to a resurgence between high and low water levels on the west bank of the River Wye about 100 m downstream of the entrance. The cave is intertidal at its lowest point, and rises to elevations of about 40 m.

The cave has been described by Elliott *et al.* (1979) and Westlake *et al.* (1989), and its diverse troglobitic fauna is recorded by Chapman (1979).

Description

An abandoned bedding plane passage, opening in the west bank of the River Wye a few metres above high-tide level, is the only entrance to Otter Hole. This joins the active streamway at the lowest part of the system, where backflooding at each high tide creates the unique tidal sump. At low tides, this can be passed to reach the active streamway which extends upstream along a fault-guided rift, and west to where the water emerges from low bedding plane passages with several flooded sections.

Above the active streamway, a series of old high-level passages extends to the west and north-east (Figure 6.20). Crystal Ball Passage is 200 m long to a choke and its rifts are up to 10 m high; there are many gour banks and spectacularly coloured curtains, and the crystal balls are roughly spherical calcite growths developed round the ends of straw stalactites where they dip below the

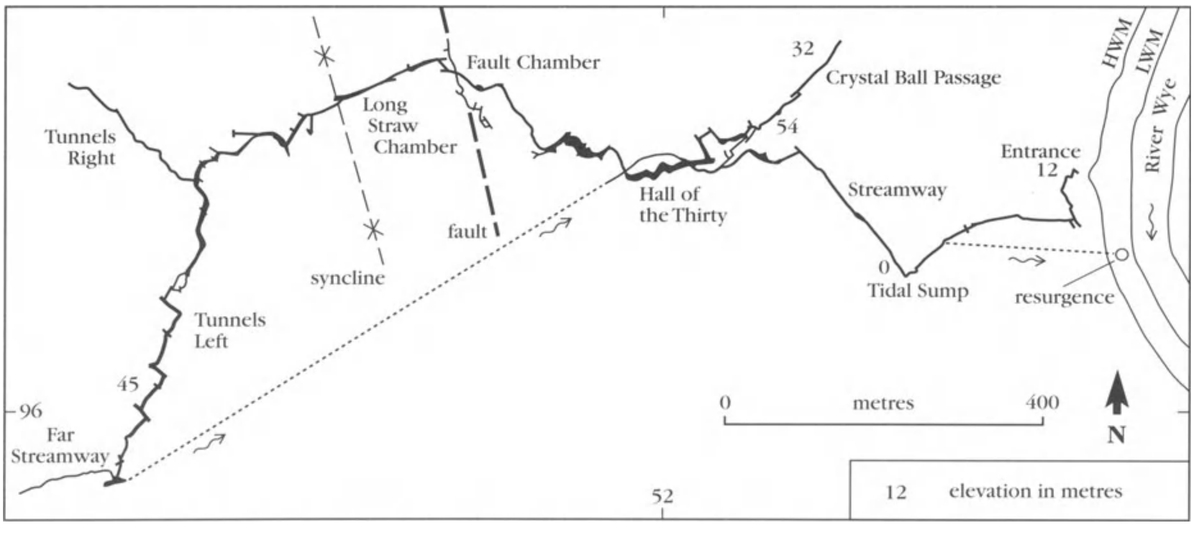


Figure 6.20 Outline map of Otter Hole (from survey by Birmingham University Speleological Society).

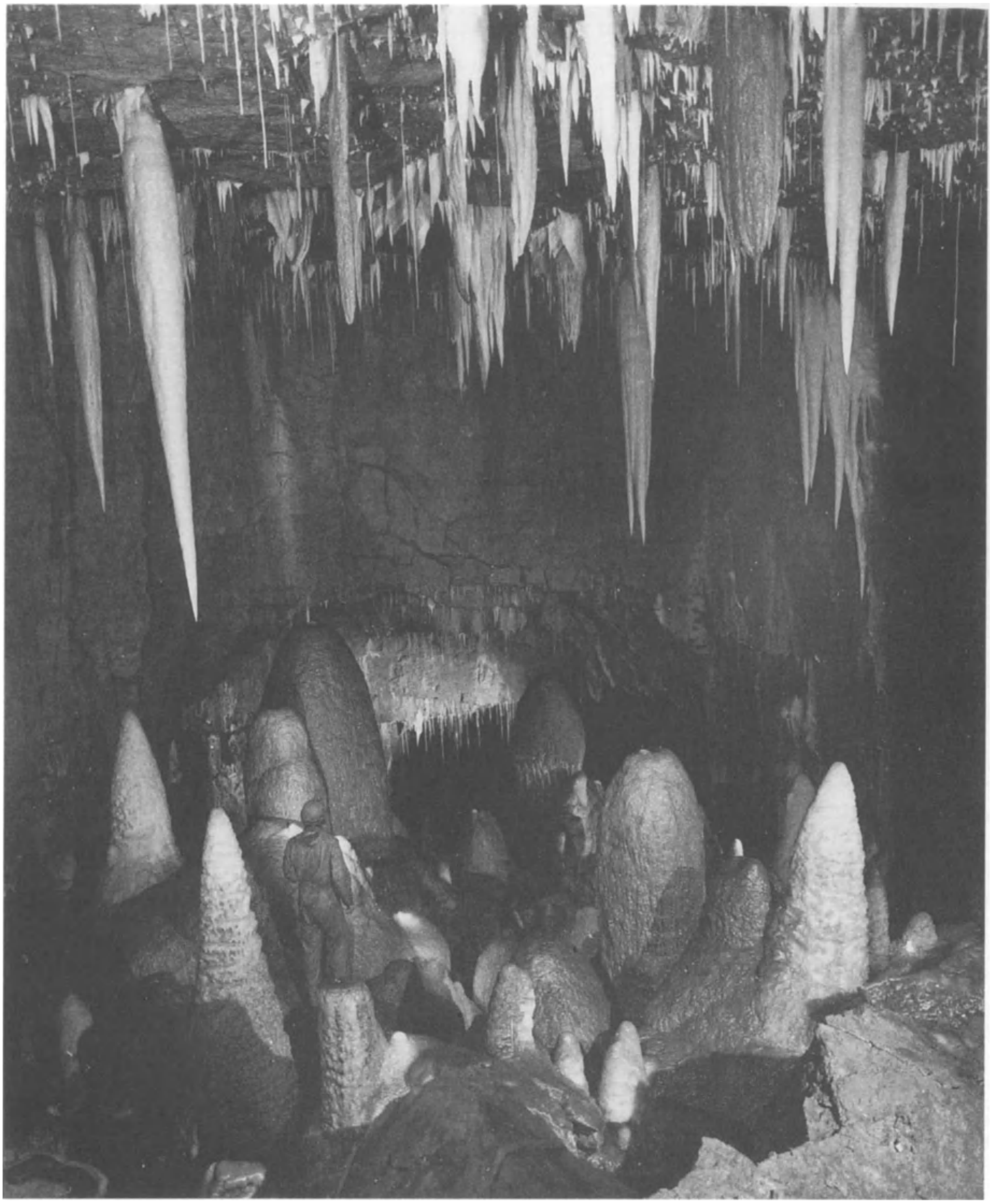


Figure 6.21 The massive stalagmites and stalactites in the Hall of the Thirty in Otter Hole. (Photo: J.R. Wooldridge.)

surface of gour pools. The main high-level passage to the west enlarges into the Hall of the Thirty. This chamber has a breakdown floor and contains a magnificent display of stalactites and stalagmite bosses on a scale unparalleled elsewhere in Britain. The most notable features are the many calcite stalagmites up to 6 m high and over a metre in diameter with cylindrical profiles and domed tops (Figure 6.21).

Beyond the Hall of the Thirty, the passage has only extensive mud formations, before it crosses the Vicarage Fault in Fault Chamber. To the west, it is again profusely decorated with gour and crystal pools, flowstone and immense numbers of straw stalactites up to 4 m long. The finest of these are in Long Straw Chamber, where the cave crosses the trough of a shallow plunging syncline. From Tunnels Junction, the larger upstream passage is Tunnels Left, a phreatic tube 4 m in diameter with many high cross-rifts; at its farthest explored limit, this is intersected by a short section of modern streamway, choked in both directions.

Interpretation

Otter Hole is developed within a thick, uniform sequence of dolomites and dolomitic limestones, lithologies which elsewhere in Britain have very limited cave development. Passage location within the carbonates is closely controlled by bedding, with the cave curving round the shallow plunging syncline while maintaining an almost constant elevation. Joints and faults guided the initial flow paths, and greatly influence the passage morphology.

The cave lies at a lower altitude than any other major system in Britain, and its intertidal entrance series provides a unique restriction on access to the passages beyond the tidally fluctuating sump. This low altitude environment, and the sheltered west coast location, may account for the scale of speleothem development in Otter Hole. The massive cylindrical stalagmites are typical of those formed by high inflows of saturated dripwater which continue to deposit calcite as they flow down the stalagmite sides. They are unlike any others in Britain, and are comparable only to those in southern European caves formed in warmer, Mediterranean climates with thick soil and vegetation cover on the carbonate outcrops. Their presence suggests a local climatic regime during parts of the Pleistocene which was significantly warmer than climates in Britain's other karst areas, mostly further north and at higher altitude. Development of the main caves in the dolomites may have been favoured by the warmer climatic conditions in earlier times, but could have been on favourable inception horizons within more fractured or porous secondary dolomites. The old high-level caves were later drained and invaded by the percolation water which formed the calcite decorations.

The sequence of development of the passages in Otter Hole correlates closely with the history of the River Wye, where four levels of terraces at altitudes of 3–60 m lie above a buried channel at least 15 m deep beneath the estuarine alluvium at Chepstow. In the earliest stage of Otter Hole, the main phreatic drainage flowed from Tunnels Left, along the main high-level cave, and out to a contemporary resurgence beyond the end of Crystal Ball Passage. The downstream end of this passage lies at about 30 m OD, just below the level of the second terrace. Further incision of the Wye, in response to the low sea level of a Pleistocene glacial stage, drained the old phreatic trunk caves into a lower route; this now forms the phreatic elements in the upper part of the active streamway, and the old phreatic passages of the entrance series, which correlate with an old resurgence at the level of the 3 m terrace. Lowering of base level during the Devensian caused vadose entrenchment in the streamway, and abandonment of the old resurgence in favour of the present one. The intertidal position today reflects a subsequent sea-level rise in the Holocene. The higher Wye terraces and the earlier stages of the cave development are not yet dated; analogy with other sites suggests a Hoxnian age for the old trunk cave and an Ipswichian timing for the main stalagmite deposition.

Conclusion

Otter Hole contains the largest stalagmites in Britain, which are comparable with speleothems in caves of the Mediterranean regions, and appear to reflect the southerly site at low altitude. The passage levels correlate with terraces on the River Wye. They are unusual in their extension down into the intertidal zone, and their development within a dolomitic sequence - both features not seen in other large caves in Britain.

PANT-Y-LLYN

Highlights

Pant-y-llyn is the only turlough in the Welsh karst, and is the only clearly defined example of a seasonal lake in Britain. It demonstrates the complex interaction between geological, geomorphological and hydrological controls on groundwater levels in a karst aquifer.

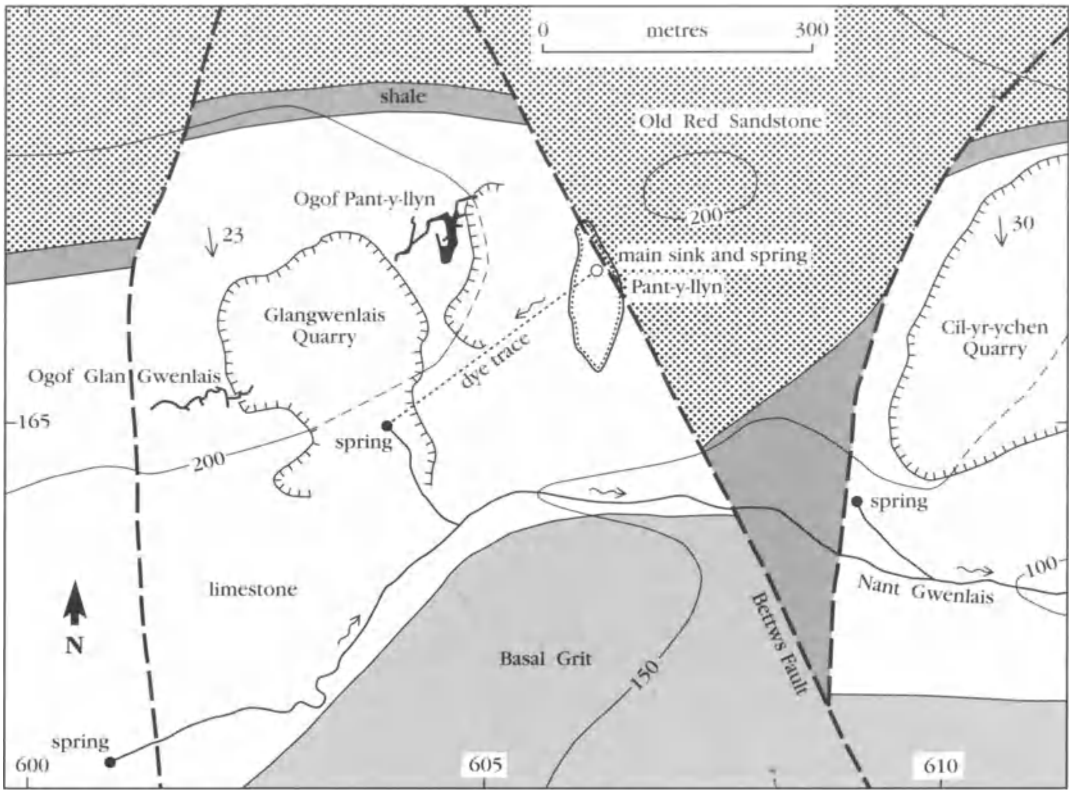


Figure 6.22 Geological map of the area around Pant-y-llyn.

Introduction

Turloughs are seasonal lakes that occur in lowland karst terranes and are best developed on the Carboniferous Limestone of western Ireland (Coxon, 1986, 1987a, b; Drew and Daly, 1993). Their water levels and intermittent appearance are related to the fluctuations of the regional water table, and they have no surface drainage, influent or effluent except for direct rainfall in the very small basin. The seasonal lake of Pant-y-llyn lies on the narrow Carboniferous Limestone outcrop near Llandybie (Figure 6.1). It slowly fills every autumn, and remains full until it drains in spring; the cyclicity is clearly related to the seasonal water table variations and is barely influenced by individual storm events. Pant-y-llyn appears to be the only turlough in Britain, as the ephemeral lakes in the chalk karst of East Anglia are in partially plugged dolines and do not have simple patterns of annual flooding.

The turlough is briefly described by Davies and Stringer (1991), Campbell *et al.* (1992) and Hardwick and Gunn (1995). The important biology of the site is described by Rundle (1993) and Blackstock *et al.* (1993), and the nearby caves are recorded by Adams and Jones (1984), Jones *et al.* (1984) and Jones (1991).

Description

The turlough is located in a closed depression within a saddle which is cut north to south through the narrow, broken limestone escarpment; it is 1500 m west of a deeper valley which carries the Afon Marlas through the limestone ridge. Quarries have removed much of the limestone both east and west of Pant-y-llyn. At its maximum extent in winter, the lake is 160 m long and up to 60 m wide, with a surface level of 160 m and a maximum depth of 3 m (Figure 6.22). The bed of the turlough is covered by a thin layer of organic debris, underlain by deposits of yellow-brown, silty clay. The lake level usually falls slowly through April, May and June, by draining through fissures in the bed; in summer the site is dry. Standing water first appears in autumn in a small pool at the northern end of the depression, where the main springs and sinks lie. It remains full over the winter; when it ices over, a small patch remains unfrozen over the spring site where warm groundwater emerges.

Pant-y-llyn lies on the northern margin of the South Wales coalfield syncline (Figure 6.1). The rocks in the area dip south at 20–40°, and include the Devonian Old Red Sandstone, the Dinantian Lower Limestone Shale, followed by about 200 m of limestone, and the Namurian Basal Grit. The turlough is underlain by the Bettws Fault, with an apparent downthrow to the west of about 200 m. This has brought the Dinantian Limestones into contact with the Devonian Sandstones; the eastern bank of the depression is cut into the sandstone, while the rest of the turlough lies on the limestone (Figure 6.22). The topography of the area consists of a low fragmented escarpment of Carboniferous Limestone, with a series of strike valleys running NE–SW, such as the Nant Gwenlais valley. The Bettws Fault zone appears to have offered a zone of weakness through the escarpment; during the Pleistocene, this was exploited either by ice moving south or by glacial meltwater (Bowen, 1965). More than 6 m of glacial till lies both north and south of the turlough, but is absent under the turlough site.

The regional drainage on the surface and in the limestone is eastwards to the lower outcrop in the Marlas valley. The Bettws Fault breaks this trend, as the limestones are not in contact across it; groundwater in the Pant-y-llyn block is therefore impounded, and its water table fluctuates between levels of 155 m in summer and 160 m in winter. Sinks and springs at higher levels drain the limestone west of a minor fault west of Ogof Glan Gwenlais (Figure 6.22). About 50 m above the level of the turlough, Ogof Pant-y-llyn has 350 m of phreatic rifts and bedding cave passages. Ogof Glan Gwenlais has 200 m of old phreatic passage aligned on the strike at an elevation of about 165 m (Figure 6.22).

Interpretation

Pant-y-llyn conforms to the three main criteria for distinguishing turloughs from other seasonal lakes (Coxon, 1986): it exhibits seasonal flooding to a depth of over 0.5 m for part of the year and a dry floor for part of the year, it is recharged via ephemeral springs or estavelles, and it empties via swallets or estavelles with no surface outlet. Dye tracing by Hardwick and Gunn (1995) indicated that the turlough was probably not fed by discrete well defined conduits, but instead receives and loses its water to the local groundwater body in its western flank. Recharge is mainly from the west, where the unbroken aquifer continues along the strike of the limestones, and outflow is to the west, as sandstone lies to the east. Tracing studies, limited by the lack of access to land around the quarries, have shown that water from the turlough reappears in fissure risings in the floor of the temporarily inactive Glangwenlais Quarry (Figure 6.22), and then flows into the Nant Gwenlais; it may also feed into the Nant Gwenlais from other unknown springs or seepages. The lower spring in the Gwenlais valley east of the Bettws Fault is fed only from the limestone around the Cil-yr-ychen Quarry.

Seasonal recharge and discharge of the turlough reflect seasonal variations in the groundwater surface. The local water table level is partly dictated by spring levels, which are where the valleys intercept cave conduits on inception horizons at various stratigraphic levels in the dipping limestone; there appears to be none in the youngest beds which have the lowest outcrop just west of the Bettws Fault. Groundwater storage is in fracture and bedding plane fissures opened by solution, but flow is probably impeded by both the immaturity of the karst and large amounts of inwashed clastic sediment. Perched water tables may lie behind rising phreatic loops and sediment chokes. A combination of these factors accounts for the turlough water table being perched 5–10 m above the Gwenlais valley. The water table fluctuation is caused by the limited capacity of conduits draining the groundwater to the risings. Excess recharge in autumn and winter raises the piezometric surface, causing the turlough to fill from fissures in the limestone; as recharge declines in late spring and early summer, the piezometric surface lowers and the turlough drains out via the same fissures in its bed. Direct drainage eastwards is impeded by the impermeable Devonian strata on the eastern side of the Bettws Fault. The flow pattern is not known in detail, but the known caves do demonstrate the importance of flow along the strike. The relationships between rainfall and turlough recharge, and the lag times involved, are still not known in detail, nor is the exact catchment area of the turlough.

Conclusions

Pant-y-llyn is a small turlough which reaches its maximum size in winter, when it measures 160 m long and 3 m deep. In late spring and early summer the lake drains through its floor. The lake appears to be the surface expression of the local water table. Dye tracing shows that the turlough drains to the Glangwenlais Quarry to the west and to the Nant Gwenlais via an unknown route. Pant-y-llyn is the only known turlough in Britain.

LLETHRID VALLEY

Highlights

The Llethrid valley system is a fine example of a complete dry valley system extending across the limestone outcrop, between a stream sink and a resurgence, each close to the limestone boundaries. Two caves include parts of the flood route of the underground drainage, and another has yielded an important Pleistocene fauna.

Introduction

The Llethrid valley is cut through the complete sequence of the Carboniferous limestones where the outcrop lies across the core of the Gower Peninsula (Figure 6.1). From the sink which swallows the water draining off the Namurian and Upper Limestone Shales, the valley is dry for 1500 m southwards to the resurgence, where older, impure limestones reach the surface. The underground route of the base flow is inaccessible, but Llethrid Swallet and Tooth Cave have passages which act as a flood overflow route.

The evolution of the valley system has been discussed by Groom (1971; and in Atkinson and Smart, 1977), though there is controversy over the local effects and extent of the Devensian glaciation (Bowen and Henry, 1984; Bowen *et al.*, 1989, Campbell and Bowen, 1989). The karst landforms of the Gower are summarized in Ede and Bull (1989), the caves are documented in Price (1984),

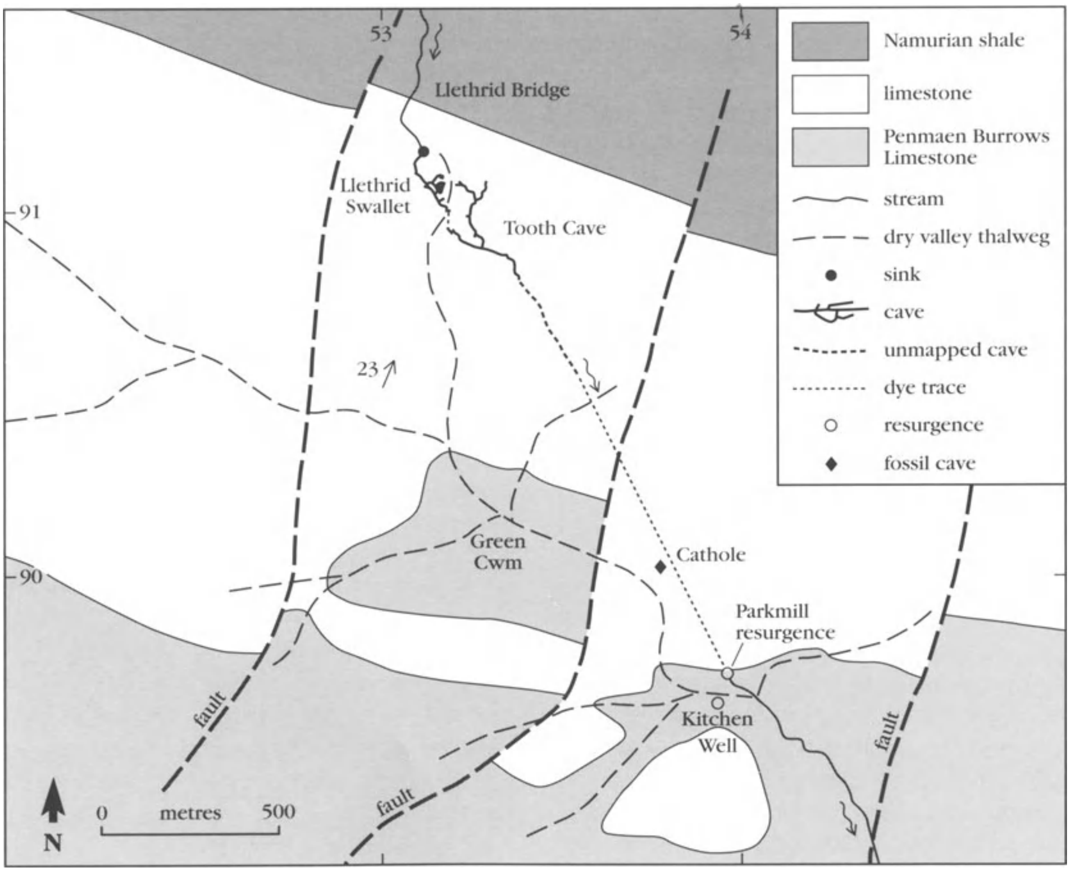


Figure 6.23 Geological map of the dry valleys and caves of Llethrid. The position of the downstream section in Tooth Cave is only approximate, as it is normally flooded and has not been mapped in detail.

Stratford (1995) and Oldham (1982), and the karst hydrology is considered in Ede (1973). Cathole is the most important of the inland caves in the Gower with respect to its Pleistocene fossils and Bronze Age human remains (Campbell, 1977; Campbell and Bowen, 1989).

Description

The valley drains against the northerly dip of the Carboniferous limestones, which is mostly at about 20° except where it steepens near Llethrid Swallet. A thin band of Oystermouth Beds, the local equivalent of the Upper Limestone Shales, separates the limestone from the Namurian shales which form the headwater catchment on Pengwern Common. Over 400 m of massive, pure limestones are exposed down the valley (Figure 6.23), before the less permeable Penmaen Burrows Limestone is reached; these are dolomitic and shale-rich, and continue south of the resurgence to a faulted contact with Devonian sandstones.

Upstream of Llethrid Bridge, a permanent stream flows on the shales. Where it passes onto the highest limestones, the water flows underground via a number of choked sinks. Downstream, the valley is dry and meanders south, as a trench cut 30 m into the limestone plateau. Its sides are steep, with crags and lengths of bare cliffs, and several dry tributary valleys drain into the main valley. Most of the dry valley has a broad grassy floor; this is widest at the Green Cwm where the main valley confluence is cut in the weaker Penmaen Burrows Limestone. A relict stream channel is present only along a short section just above the confluence.

The main sink, only active at high stage, is the entrance to Llethrid Swallet; this has over 300 m of small stream passage, which forms a flood route almost inactive at low stage, with a large sloping chamber heavily modified by bedding collapse above it (Figure 6.23). In the left bank, shortly below the sink, Tooth Cave has 300 m of low muddy passage leading into another section of the flood route about a kilometre long; this is normally dry down to an almost permanent sump, through which the downstream end is rarely accessible, and the whole passage fills in flood. Cathole has a wide entrance in a cliff 17 m above the valley floor; it is a cave remnant with two dry chambers, in which sediments have yielded a cold Devensian fauna and Creswellian artefacts (Campbell, 1977; Campbell and Bowen, 1989).

The Parkmill, or Wellhead, resurgence lies in the eastern side of the valley, at an elevation of about 15 m, nearly 30 m below the sink; it is impenetrable and now lies under a pool impounded to facilitate pumped abstraction. Llethrid Swallet provides about 20% of the water at the resurgence, with a flow-through time of 20 hours; the rest comes from percolation input and other smaller sinks. On the opposite side of the valley, Kitchen Well is a smaller spring, fed entirely by percolation water. It has a mean calcium hardness of 197 ppm in winter, and 206 ppm in summer; both these values are higher than those for the Parkmill resurgence, which is fed partly by swallet water and is frequently diluted by flood flows with low solute loads (Ede, 1973). Below the resurgences, the stream flows on the surface over the impure limestones to reach the sea at Oxwich Bay.

Interpretation

The Llethrid valley is incised into the 60 m coastal platform surface of Gower, which was probably developed in the early Pleistocene (Ede and Bull, 1989). Valley entrenchment was therefore later than this, and underground drainage was initiated to form phreatic caves, as the limestone plateau was exposed. The main chamber in Llethrid Swallet may be a remnant from this early phase. The dry valley system was almost certainly incised during periglacial periods when the underlying cave systems were choked with till or ice, allowing surface flow. Groom (1971) suggested that the main dry valleys on the Gower were incised during the penultimate interglacial, although this is unlikely unless underground drainage had not become fully integrated by this time. The northern part of the valley was probably glaciated during the Devensian (Bowen *et al.*, 1986), and meltwater from this and earlier glaciations almost certainly flowed down the valley. The meandering nature of the valley indicates its fluvial origin, while extensive infill by solifluction during periglacial episodes accounts for the smooth valley profile.

The known caves all lie in the younger part of the limestone sequence, where groundwater flow appears to have been captured on some particularly favourable cave-inception horizons (Lowe, 1989b). Beyond the explored limits of the caves, the flow route descends through the stratigraphic sequence, but rises to the surface above the dolomitic Penmaen Burrows Limestone. All the accessible passages in Llethrid Swallet and Tooth Cave pre-date the modern phase of cave formation and are only active under flood conditions. The modern drainage route is immature and unable to cope with flood discharges; it may be developing in smaller cave passages within the stratigraphically lower limestones. Dating of stalagmites in the Llethrid Valley caves has the potential to provide estimates for the timing of valley incision, which would define a minimum age for the formation of the 60 m planation surface.

Conclusions

Llethrid Valley contains some of the best karst features on the Gower. It is an excellent example of a karst dry valley with a complete spectrum of allogenic and autogenic underground drainage, and active and abandoned caves.

MINERA CAVES

Highlights

The caves of Minera form the most extensive integrated cave system yet explored in North Wales, with a major trunk conduit draining from active and inactive tributaries to a series of progressively more recent, and lower, resurgence passages.

Introduction

The Minera cave systems lie beneath the northern slopes of Esclusham Mountain, west of Minera and south of the Clywedog Valley (Figure 6.2). Ogof Llyn Parc, Ogof Dydd Byraf and Ogof Llyn Du are fragments of an integrated karst drainage system; they are now separated only by short sections of passage choked with sediment (Figure 6.24). The Carboniferous Limestone dips at 5–15° just south of east, and is truncated to the north by the Llanelidan Fault which downthrows the overlying Cefn y Fedw Sandstone of the Millstone Grit. The limestone unconformably overlies cleaved Ordovician shales which crop out to the west. Cave development within the site extends down through the Brigantian and Asbian succession of the Cefn Mawr, Loggerheads and Leete Limestones.

Water draining from the moorland in the west collects into the Aber Sychnant. This flows north across thick till and through a short limestone gorge, before turning east to become the River Clywedog, where the remains of another limestone gorge survive in the floor of the older, abandoned Minera quarries. The main allogenic input to the caves is through sinks west of and along the Aber Sychnant where it crosses the limestone west of Ogof Llyn Parc. Other sinks lie further south, near the outcrop of the Cefn y Gist Vein.

The only published accounts of the explored caves within this system are those of Appleton (1986, 1987, 1989), with a brief account of the divers' extension to Ogof Llyn Du by Whybro (1987).

Description

Swallow holes along the course of the Aber Sychnant are immature fissures largely choked with limestone blocks, gravel and peat.

Ogof Llyn Parc is entered through mined shafts and levels on the Pool Park Vein. These intercept the main cave at a depth of 130 m, from where natural cave passages extend for over 4000 m, with a vertical range of 115 m (Figure 6.24). The main streamway emerges from the Quarry, a passage up to 12 m wide and 8 m high, modified by roof collapse of the thinly bedded Leete Limestone south of the Pool Park Vein. Downstream, the Miners' River Passage is a phreatic tube 3 m high and 7 m wide following the strike of the beds. It enlarges as it is joined by three other tributaries flowing downdip from the west. The Northwest Inlet flows for almost half of its length over a floor of Ordovician shales, and forms a 3 m cascade over a minor fault with a 3 m throw. Fault Inlet follows a poorly mineralized fault with the Ordovician shales exposed on its north side. The water now sinks into impenetrable bedding planes towards the lower end of the Miners' River Passage. Prior to draining when the miners cut adits below the natural resurgence, the main stream flowed along the Northeast Passage. Both this and a high-level passage along the strike further updip lead into the Master Passage, up to 6 m wide and 11 m high, developed along the north side of a small fault. This was the main trunk drain for the whole mountain, and a number of large, ancient, phreatic passages converge onto it. These abandoned passages contain abundant calcite speleothems and clastic sediments, some in sequences of stalagmite layers interbedded with

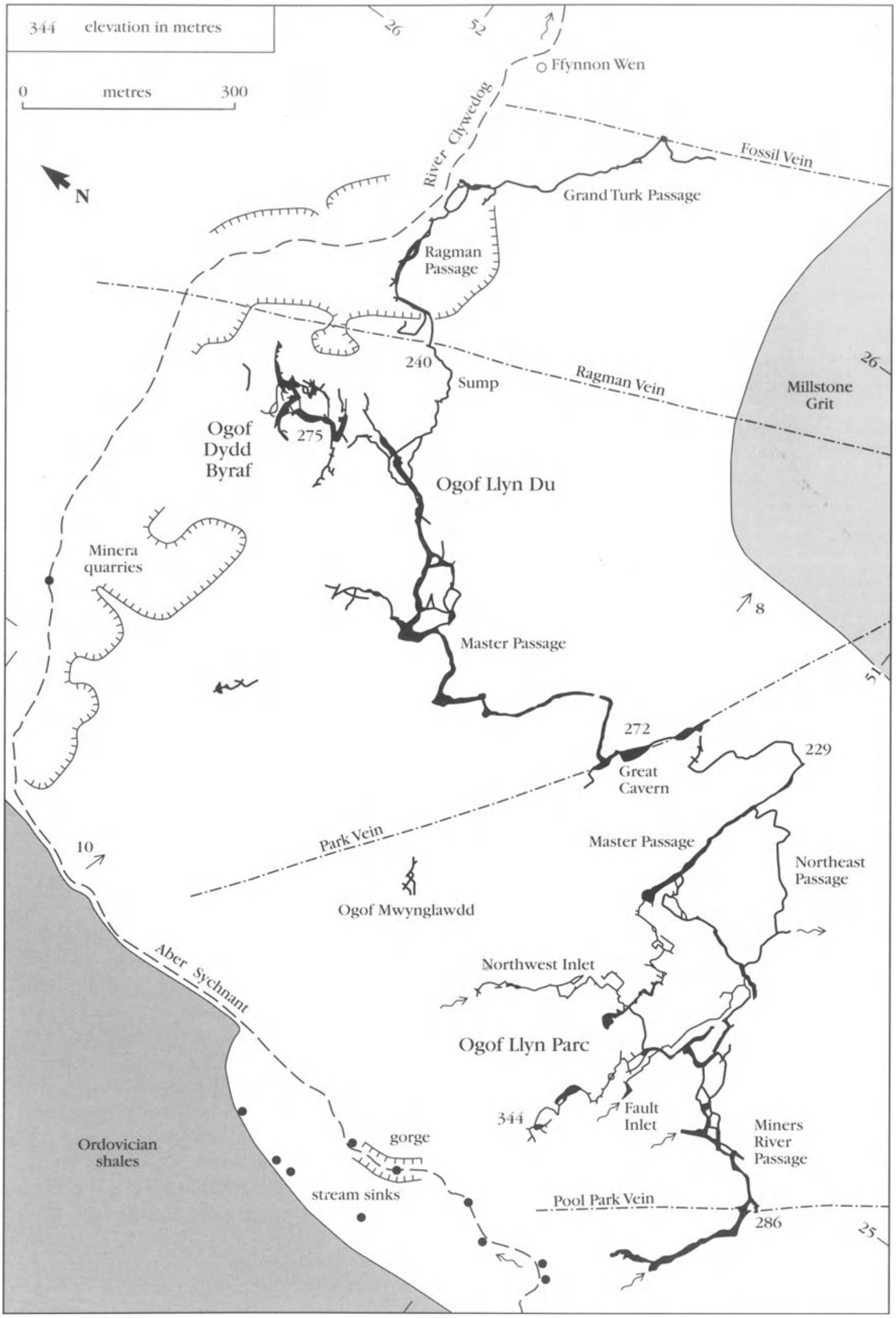


Figure 6.24 Outline map of the caves of Minera (from surveys by North Wales Caving Club). The Ffynnon Wen resurgence is no longer active as its water has been captured by the deep mine adits.

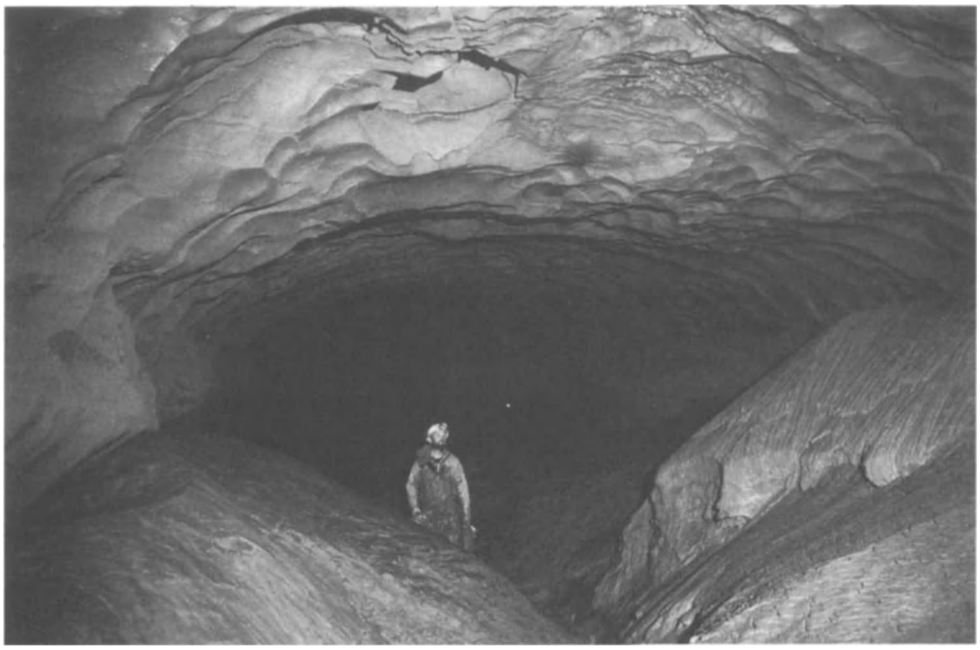


Figure 6.25 The Master Passage of the Minera caves where it leaves the Park Vein just north of the Great Cavern. This phreatic tube was beneath the water table until it was drained by the mine adits. (Photo: P.J. Appleton.)

clastics containing broken and detached speleothems. The lowest section of the Master Passage, on its easterly loop down the dip, contains thick clastic sediments exposed in a rejuvenated stream trench, before it turns north to a sand and gravel choke close to the Park Vein.

Three chambers, up to 70 m high and 20 m wide, are developed along the Park Vein, and are accessible via a separate 90 m deep mined shaft (Figure 6.24). The Great Cavern lies downstream of the sand choke at the end of Ogof Llyn Parc, and the downstream continuation is a phreatic tube 6 m in diameter choked with gravel in a downward loop (Figure 6.25).

The Master Passage of the Minera caves continues through Llyn Du Cavern (Figure 6.24). A large elliptical phreatic tube contains some flooded sections before turning north-east along the strike with a number of loops and tributaries; a large talus cone which contains sandstone blocks indicates a connection with the surface some 60 m above. The larger passage continues at a high level, to a choke close to Ogof Dydd Byraf, with a smaller inlet choked only a few metres from the northern cave system. A smaller phreatic tube lies directly beneath parts of the high level, and is linked to it by various younger vadose inlets. This then turns downdip into a phreatic loop which ends at a lift on the Ragman Vein; the whole loop, 26 m deep and 270 m long, is still flooded. Downstream from the Ragman Sump, a phreatic tube on the bedding descends from the crest of the loop, entrenched by a vadose canyon, with abundant stalactites and gour dams up to 1.5 m high. The old phreatic passage extends to the site of a former resurgence in the floor of the Clywedog valley. More recently, a lower route has taken the main drainage of the area to the Ffynnon Wen rising, prior to its capture by the miners' drainage adits. The Grand Turk Passage carried water against the dip, to the same old resurgence. It is a low phreatic tube on one bedding plane in the lower part of the Loggerheads Limestone, and the source of this water is unknown.

Access to Ogof Dydd Byraf is gained via mined passages which have intercepted a natural rift. Most of the cave is a relict phreatic system on two main levels. A lower main passage, developed in the middle of the massive Loggerheads Limestone, extends north through a series of decorated chambers, where elliptical tubes 4 m wide have been modified by roof collapse. High solution cavities lie on cross-joints, and one section opens into a three-dimensional network of passages developed on three bedding planes linked by steep joint fissures. These passages contain some very fine calcite deposits, with gours, crystal pools, cave pearls, curtains and white and green stalactites; there are also many mud formations and sequences of fine silts and muds deposited in slowly moving water. The upper level of passages are developed at the base of the more thinly bedded Cefn Mawr Limestone and have suffered more extensive collapse. They lie about 35 m above the present floor of the Clywedog Gorge. Some of the caves are choked with clastic sediment, and some contain a profusion of deep-red stalactites, stalagmites and helictites, with stalagmite crusts formed on mud flakes.

Many fragments of cave passage were intersected by the faces of the disused Minera Quarry. They represent remnants of old inlets to the Minera caves, and most have been choked with calcite and clastic sediments; stalagmites from one cave have been dated to about 56 ka.

Interpretation

The Minera caves developed in several phases as water tables dropped in direct response to the lowering of the resurgence outlets. These new resurgences were formed as a consequence of surface lowering through the Pleistocene, and more specifically through excavation of the Clywedog valley through Minera, which progressively exposed the top of the limestone at lower altitudes downdip to the east. Vadose drainage has always been essentially eastwards, downdip as far as the water table. Below this level, the phreatic drainage was northwards, roughly along the strike, to the contemporary outlet, at or near the lowest limestone outcrop. Successive routes were displaced downdip to the south-east, with greater lowering and displacement closer to the shifting resurgences at the downstream end of the system. Consequently, the old phreatic trunk routes diverge from a common source in the Miners' River Passage in Ogof Llyn Parc. Inlet passages have stayed in the same positions while their upper ends have been progressively removed by scarp retreat in their direction.

The shifting patterns of karst drainage can therefore be reconstructed and correlated with the geomorphological history of the area. The high-level passages in Llyn Parc, Llyn Du and Dydd Byraf are remnants of the same trunk route which was one of the earliest established through the limestone. Other fragments of ancient high-level routes, now almost entirely choked with sediment, were cut by the working faces of the Minera quarries. Later downdip captures include the Northeast Passage in Llyn Parc and the lower tube and Ragman Passage in Llyn Du. Youngest of all are the still immature and not yet accessible modern drainage routes into the low-level mine adits. The extensive sediment and speleothem deposits within the various parts of the system may enable a detailed chronology of the Pleistocene history of the area to be reconstructed.

Within the broad pattern of strike drainage, the morphology of the cave passages demonstrates strong control by geological features. Thick shale beds, palaeokarst horizons, the basal clastic sequence of the limestone, and the contact with the Ordovician shales are all well exposed underground, and clearly influence the passage positions and morphology. Major phreatic passages are developed on bedding planes along the strike, while at least two of the vadose inlets flow for part of their length downdip along the underlying Ordovician shales. In Ogof Llyn Du, vadose incision has occurred in the drained phreatic tube, downdip on the bedding, immediately east of the Ragman Vein; this was the downstream limb of the phreatic uploop with the lift on the vein.

Both vadose and phreatic passages show a close control by faults and major joints. Enhanced solution along mineralized faults has led to the development of large chambers, notably the series of caverns along Park Vein, though smaller examples are widespread through the system. The dip-orientated faults and veins divert the strike-orientated flow into deep phreatic loops out to the east; the most conspicuous is the loop down the Master Passage in Llyn Parc and back up through the Great Cavern on the Park Vein. Phreatic lifts also occur where the bedding planes feed drainage to the mineral veins; the 26 m deep Ragman Sump in Ogof Llyn Du is the deepest accessible lift, while the downstream end of Ogof Llyn Parc is another, now largely choked with sediment in its second downloop. The Llyn Parc lift is now active only when a sufficient hydrostatic head has been raised on the upstream side to breach the sediment plug by lifting sediment the full height of the lift. Fluidization of the sand under these flood conditions has polished the limestone walls over the lift crest.

Conclusion

The Minera caves represent one of the finest examples of completely integrated karst drainage in Britain. The system contains active and abandoned vadose inlets feeding to a sequence of phreatic strike passages which developed successively downdip in response to surface lowering and karstic rejuvenation. This pattern is repeated elsewhere in the dipping limestones of Britain; in their geometry, Yorkshire's Kingsdale and the Priddy caves drainage on Mendip differ only with respect to the magnitude of the local dips. All these cave systems have their oldest elements drained and abandoned, but only Minera has its recent phreatic passages accessible due to drainage by mined adits. Minera also has cave passages on mineral veins and along the base of the limestone, and the calcite speleothems in Ogof Dydd Byraf exhibit mineral colouring on a scale which is rare in Britain.

ALYN GORGE CAVES

Highlights

The caves of the Alyn Gorge provide excellent examples of both shallow and deep phreatic drainage systems which now lie within the vadose zone as a result of mine drainage.

Introduction

Ogof Hesp Alyn and Ogof Hen Fynhonau lie beneath the southern flank of the Alyn Gorge, upstream of Mold (Figure 6.2). These two caves represent components of the former underground feeders of the Alyn River, subsequently drained by mining activities. The Carboniferous Limestone rests uncomformably on Silurian mudstones and dips at about 15° east; the caves are formed in the Asbian Loggerheads Limestone. This is broken by two sets of faults; an east–west trending set are mostly mineralized, and a north–south trending set are generally barren but some have notably wide breccia zones. The caves are developed adjacent to the Alyn River where it flows through the Alyn Gorge on its route across the limestone outcrop.

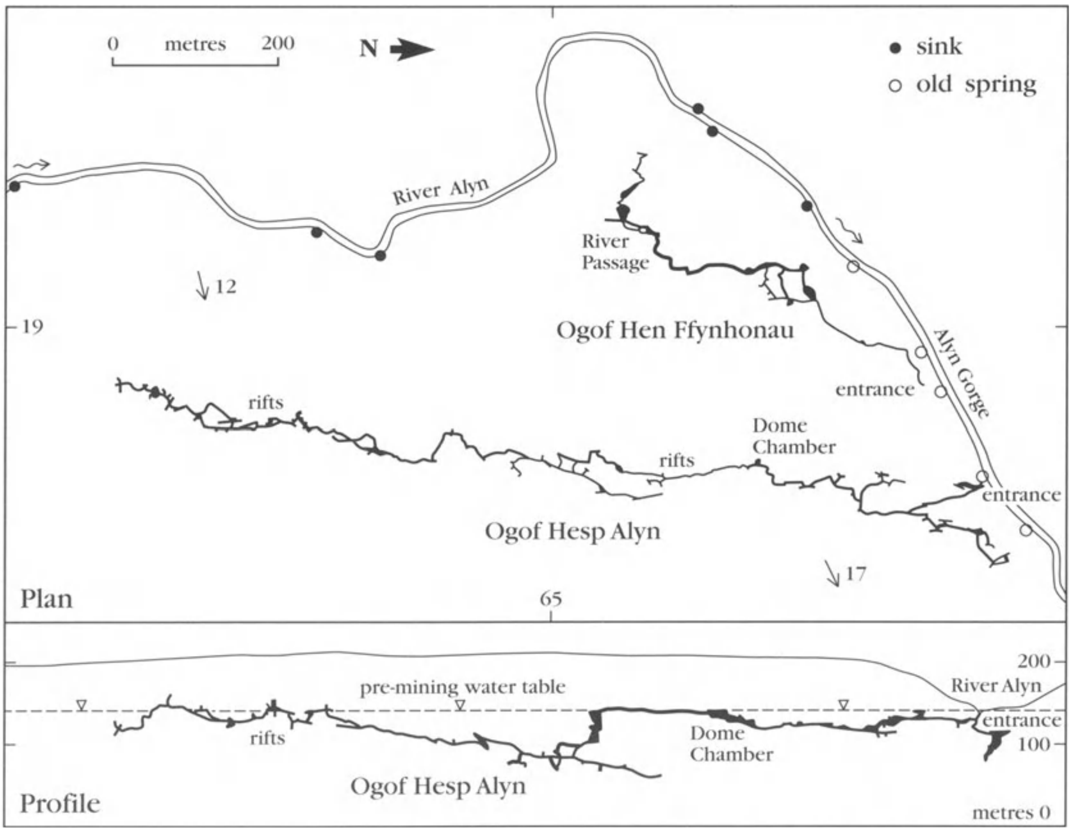


Figure 6.26 Plan and profile of the caves beneath the River Alyn (from surveys by North Wales Caving Club). The surface river loses water at the various sinkholes along its course on the limestone. Only Ogof Hesp Alyn is shown on the profile; it was almost completely flooded until deep mine adits captured its water.

The only published accounts of these caves and their hydrology are by Appleton (1974, 1984, 1989).

Description

The entrance of Ogof Hesp Alyn lies on the south bank of the Alyn Gorge. Over 2000 m of passages have been mapped in the cave, most of them along the phreatic trunk route which has a series of loops over a vertical range of 90 m (Figure 6.26). Most of the cave was beneath the water table, until it was drained in about 1901. Lead and zinc ores were mined from zones beneath the natural water table, and this required massive pumping until deep drainage adits were driven to drain the limestone almost to sea level. The new adits drained most of the cave by capturing the flow which it originally fed to risings at its northern end. The Hesp Alyn passages are a mixture of rounded phreatic boreholes, tall rifts on the main faults, and elliptical tubes on bedding planes. Slow phreatic flow has etched networks of fissures out of some fracture zones. Some of the downloops are partly choked with collapse debris and clastic sediment, or contain perched sumps, and small modern inlets have invaded some sections. The upstream end of the cave lies on a mineral vein breached by a short length of active streamway.

Ogof Hen Ffynhonau lies close to the west of Hesp Alyn, with its entrance in a mineralized fault adjacent to the main group of pre-mining springs in the Alyn Gorge. Over 800 m of passages have been mapped in the cave, and most are on bedding planes along the strike, with rifts where the main faults are crossed. They carry part of the River Alyn flow, and reach over half way to the sinkhole sources in the river bed (Figure 6.26). Large calcite decorations have been deposited in some of the passages, which were mostly above the water table prior to the mine drainage.

Interpretation

The two caves in the Alyn Gorge present a unique combination of two contrasting resurgence systems, one shallow and one deep, both now made accessible through artificial drainage of the phreas.

Ogof Hen Ffynhonau represents a relatively shallow drainage system which lay close to the local water table prior to mining; calcite dripstone accumulated in some dry parts. Sections of both phreatic and vadose passage typify the morphology of a cave close to the valley floor, which determined the levels of the local water table at the crests of its undulating profile. Much of the flow was derived locally, from sinks further south on the Alyn River, though some additional flow may perhaps have come from a deeper source. A natural rift choked with large rounded boulders lies 10 m above river level near the entrance and may represent a former resurgence.

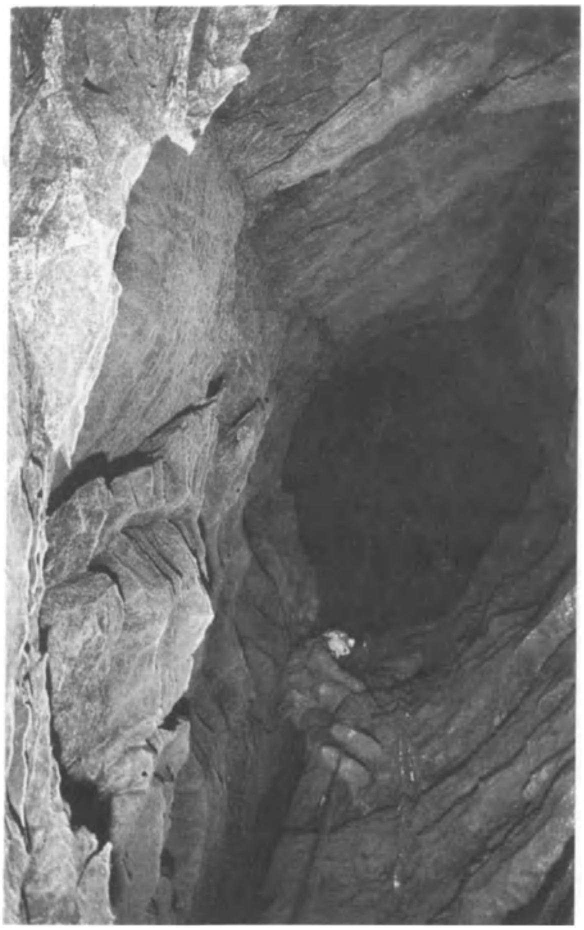


Figure 6.27 Looking up the 15 m shaft in Ogof Hesp Alyn. This phreatic lift on a major joint was active and completely submerged until mine drainage lowered the water table in 1901. (Photo: P.J.Appleton.)

Ogof Hesp Alyn has a quite different morphology, reflecting its deeper phreatic origin away from the valley. Water flowing northwards, almost along the strike following the intersections of fractures and bedding planes. This orientation took the flow to increasing depths until vertical phreatic lifts developed on veins or cross-faults (Figure 6.27); these took the water to higher bedding planes before finally escaping through springs at river level. The conduit may have drained a large area of limestone to the south; there is a dearth of known sinks nearby. Short sections of vadose trench were cut through the crests of two loops close to the pre-mining water level, but the rest of the cave developed entirely under phreatic conditions. Its drainage by mining activities has been so recent that there has not been any significant vadose modification. The cave shows clearly the effect of geological structure on phreatic development, in particular the influence of major jointing, and also shows many phreatic solution features, notably solution pockets, phreatic tubes and large phreatic lifts. The linear arrangement of passages in the northern half of the system is partly due to a coincidence of strong north–south jointing with the northwards hydraulic gradient.

Formerly the Alyn River flowed northwards beyond Cilcain and through what is now the Wheeler Valley (Figure 6.2). At some point it was diverted across the limestone outcrop, thereby excavating the Alyn Gorge (Embleton, 1964), and it is clear that the caves must largely, if not entirely, postdate this event. The river is cutting down through glacial till along parts of its route through the gorge, suggesting that its diversion was pre-Devensian. Ogof Hen Ffynhonau appears initially to have resurged some 10 m above the present river level, suggesting that the caves too pre-date the last glaciation. The sediments and stalagmite sequences in Ogof Hen Ffynhonau are a consequence of these later events, but a chronology has not yet been established.

Conclusion

The Alyn Gorge site provides a unique example of cave development in almost purely phreatic conditions from just above the water table to a depth of 80 m below resurgence level. Lowering of the water table by deep mine drainage has revealed some long phreatic caves, which have their morphology uniquely well preserved. There is almost no vadose modification or sediment infilling. The caves offer a unique insight to the anatomy of a flooded karst aquifer.

Chapter 7

Outlying karst areas in England

INTRODUCTION

England's karst landscapes are dominated by two rock types. By far the largest area of karst is provided by the Cretaceous Chalk (Figure 1.1), but this is a very distinctive karst type - almost totally lacking the bare rock outcrops and accessible cave systems which are characteristic of karst in the stronger limestones. Consequently, the Carboniferous limestones are more widely known for their karst landforms, even though they occupy a much smaller area in England. These strong limestones contain nearly all England's caves within the justifiably famous karst landscapes of the Pennine and Mendip Hills. There is an additional scatter of karst features on the same limestones in outlying outcrops, mainly in the west country.

Beyond the Chalk and the Carboniferous Limestone, karst landforms are developed on a lesser scale on a variety of carbonates and non-carbonates. Limestones range through most systems of the stratigraphic column, but their limited karst features are commonly regarded as poor relations of better developed sites on the Carboniferous and Chalk. Both gypsum and salt lie buried beneath the lowlands of England, but their surface expression in karst landscapes is virtually limited to the subsidence basins over the salt. The remaining, more obscure types of karst and pseudokarst are insignificant in Britain.

Karst on the outlying Carboniferous limestones

In the land areas each side of the Bristol Channel, the Dinantian limestones of the Lower Carboniferous form a scatter of outcrops which emerge through the cover of mainly Triassic mudstones. The largest of these forms the Mendip Hills, with the splendid caves and karst features already described in Chapter 5. Separate inliers further west form Brean Down and the islands of Steep Holm and Flat Holm, where solutional features are little more than details of the landscape (Simon *et al.*, 1961). An escarpment of the same Dinantian limestones forms the Clifton Downs, best known for their deep dissection by the Avon Gorge - a magnificent feature of superimposed drainage, inherited from the Triassic cover. The gorge has exposed many small fissures and solutional rifts in its walls, but there is minimal karstic expression on the Downs, which are high enough to receive no modern allogenic drainage. The hot springs at Bath lie over faults within Mesozoic cover rocks, but their source appears to be meteoric water which has drained through a syncline in the Carboniferous limestone deep enough to be geothermally heated (Kellaway, 1991). North of Bristol, there are facies changes within the Dinantian sequence, and karst features are very minor in the dolomites and thin limestones intercalated with clastic rocks.

North of the Bristol Channel, Dinantian limestones underlie the Forest of Dean basin. Their thickness is variable and they are partly cut out by the basal unconformity of the Westphalian, so that their main outcrop is round the western flank of the Forest and south across the Wye Valley to Chepstow (Figure 1.2). Within most outcrops on the English side of the border, the main karst features are formed in the Holkerian limestones, overlying less karstified dolomites. The main topographic features of the area are not karstic, but a series of small sinkholes feed an extensive cave system beneath a sandstone cover and draining to the Slaughter Rising in the bank of the River Wye.

Carboniferous outcrops in the Birmingham area lack any limestones as the region was a landmass in Dinantian times. Sedimentation in the Carboniferous basin north of this produced the thick sequence of carbonate, clastic and Coal Measure rocks which now form the Pennines and much of the rest of northern England. The karst and caves in the large Peak District inlier have been described in Chapter 4; the northern Pennines have even more extensive outcrops of Carboniferous limestone, which are also more varied in structure, and these are host to the many caves and karst landforms described in Chapters 2 and 3.

Outside the Pennines, the Dinantian rocks have large outcrops but very limited modern karst, though some of the Midland's inliers contain notable features of fossil karst (Simms, 1990). In northern Lancashire, the Clitheroe area lies on basinal facies of the Dinantian which are dominated by shale sequences; isolated outcrops of thinly bedded and reef limestones have almost no karstic expression. Similarly, the Dinantian sequences north of Weardale have only very thin carbonate units whose sinuous valley side outcrops are marked only by isolated scars and the shortest of underground drainage loops.

The Lake District has lost to erosion its Carboniferous cover, which originally included thick Dinantian limestones. These rocks now form an annular outcrop around much of the Lower Palaeozoic inlier, and the thicker limestone sequences remain in the south and east. Faulted blocks of limestone around the eastern arm of Morecambe Bay were scoured by Pleistocene ice, leaving some very fine pavements in addition to some small cave systems. Further extensive limestone pavements are formed on the Dinantian outcrops north of the Howgill Fells. Both these groups of sites (Figure 3.1) are described in Chapter 3 as they are so closely related to the Pennine karst. There are numerous small cave systems in the limestones all around the Lake District (Brook *et al.*, 1994), and the geomorphology of those on the southern fringe, around Morecambe Bay, was described by Ashmead (1969, 1974a).

The limestones continue west round Morecambe Bay, across the Cartmel and Furness peninsulas. On Cartmel, Humphrey Head has more limestone pavements and a few short caves. Kirkhead Cavern, a phreatic rift modified by marine erosion and now left behind a raised beach, has been excavated to reveal cryoturbated Devensian clastics overlain by a Holocene cave earth with flints and other artefacts (Gresswell, 1958; King, 1974). The Roudsea Wood Cave is a joint-controlled network formed in the shallow phreas adjacent to a Devensian lake, in the same style as the Hale Moss caves east of the Bay (Ashmead, 1974a). The Dinantian limestones of the Furness peninsula are best known for their hematite orebodies, which have been valuable sources of iron. The ore bodies included veins and flats and also the unique sops. These were shaped like buried dolines over 100 m deep and up to 300 m across, with layered fills of sand over massive hematite over limestone rubble. More than 50 sops are known, mostly in the Holkerian Park Limestone west of Dalton. They appear to represent a style of interstratal palaeokarst where late Mesozoic saline waters leached iron from the overlying Triassic sandstones, invaded karstic fissures below, created and then filled larger solutional cavities in the limestone, and then promoted collapse of the cover by further solution (Rose and Dunham, 1977). Later erosion stripped the sandstone cover, and the sops were covered only by glacial till, until the hematite was mined; their sites are now marked by massive, flooded subsidence bowls. The Quaternary karst in Furness includes the relict phreatic chambers of Stainton Cavern, the abandoned resurgence cave in Henning Valley and many other small caves (Ashmead, 1974a).

Facies changes reduce the limestone to bands less than 20 m thick interbedded with clastic rocks in the Dinantian on the northern rim of the Lake District. The limestone outcrops are extensively blanketed by glacial till, and the few known caves include The Swilly Hole in the Asbian Fifth Limestone; this has over 800 m of relict rift passages invaded by a sinking stream in the entrance zone (Brook *et al.*, 1994).

The chalk karst

The large areas of chalk karst across the south-eastern half of England are very distinctive in the styles of their landforms and their underground drainage. Both characteristics are a function of the mechanical properties of the chalk rock. It is a pure, white, weak, friable, poorly lithified, porous limestone. Its matrix is composed largely of crystals, fragments and skeletons of coccoliths; most particles are <0.002 mm across, and it is poorly recrystallized as it is largely low-magnesium calcite which was stable at low burial depths (Hancock, 1975, 1993). The rock is massive, with poorly defined bedding and few large fractures. Matrix porosity is generally >30%, but its permeability is low through the tiny pore spaces. The high bulk permeability of chalk is due to its networks of microfractures, many of which are enlarged by solution. The Chalk forms a single unit over 300 m thick in the Upper Cretaceous.

The distinctive landscape of chalk karst is a softly contoured grassy upland, often known as downland or downs. Though the chalk is weak, with a strength (UCS) of 5–30 mPa, it forms the high ground as the outcrops are surrounded by weaker clays and are little eroded by surface water. The low strength precludes scar formation, except in vertical sea cliffs, undercut by wave action faster than any surface degradation. Devensian ice covered little of the chalk, and earlier glaciations only reached to the Chilterns (Figure 1.2); glacial till masks only some of the eastern outcrop (Figure 7.1). The rounded landforms are the product of periglacial weathering. The top 10 m of most chalk outcrops have been so frost-shattered that they now form a rubble weak chalk or a thixotropic putty chalk (Higginbottom, 1966). Solifluction was widespread during the Devensian cold stages, and combe rock is a chalk head common on valley floors. Chalk outcrops were covered by woodland until the clearances between Mesolithic and

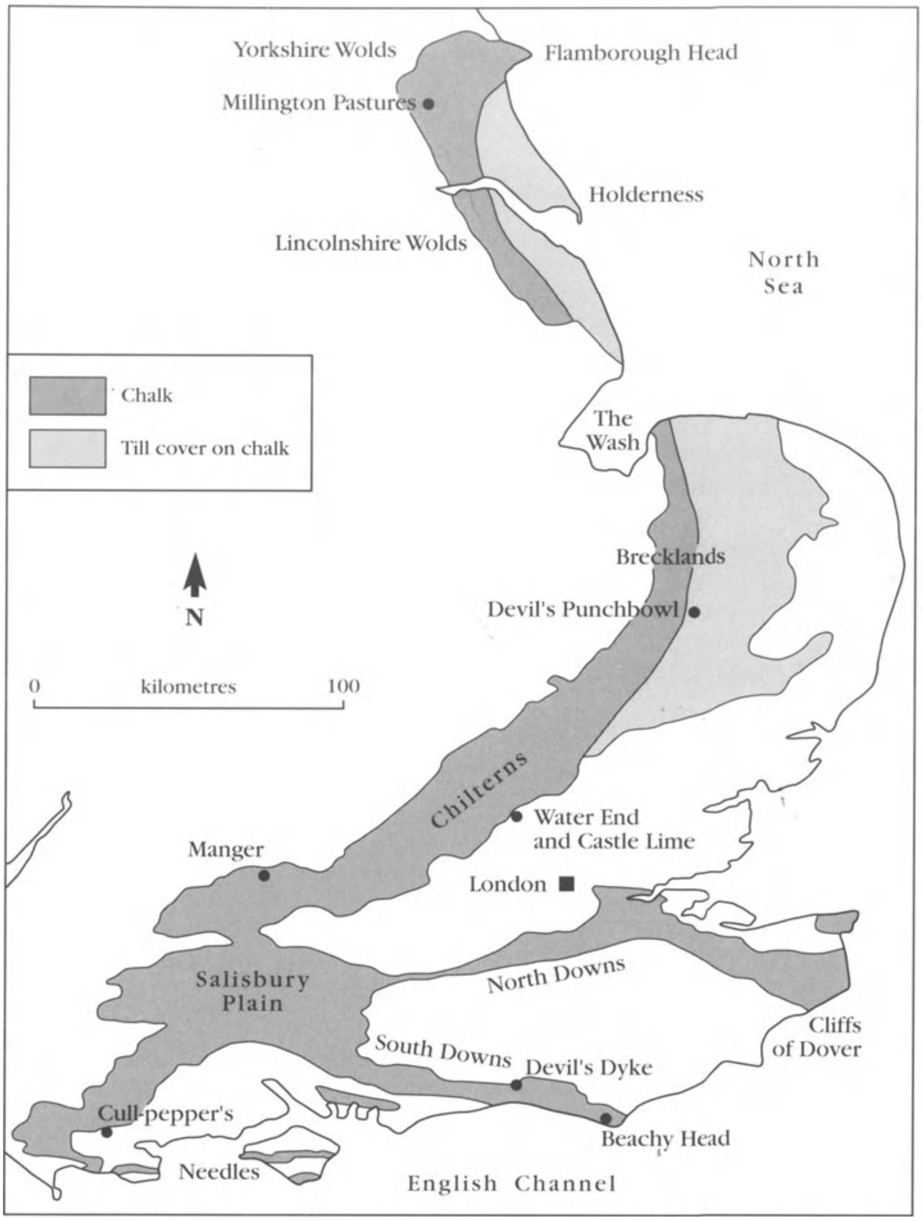


Figure 7.1 Outline map of the chalk karst of England, with locations documented in the text. Superficial deposits occur on many parts of the Chalk outcrop; only the large areas of glacial till are distinguished on this map, as they mask most topographic expression of the karst.

medieval times, and the subsequent sheep grazing has maintained the short turf of the modern downland.

Dry valleys are common on the chalk karst escarpments - largely the result of solifluction and snowmelt erosion during the Devensian. They form steep combes on the scarp faces and large dendritic systems on the dip slopes; the Manger, Devil's Dyke and Millington Pastures represent both these styles. Springs are common at their lower ends, and winter rises of the water table create many seasonal surface streams, known as bournes. Diffuse input of rainfall creates very few dolines on the main chalk karst. Along the margin of the Tertiary cover, on the very gentle dip slopes, allogenic drainage creates active sinkholes, deep subsidence dolines and the pipes which are largely clay-filled solutional fissures. The best of these features fringe, breach or underlie the feather-edge of the Tertiary cover in the London basin, but large dolines also punctuate the Quaternary cover in the East Anglian Brecklands and the Dorset Heaths.

The Chalk is the most important aquifer in

England, yielding about half of the country's pumped supplies. Its permeability is high and very variable; it has minimal flow though the matrix pores, high diffuse flow through the fracture networks, and very high flows through solutionally enlarged fissures (Lowe, 1992a; Price *et al.*, 1993; Mortimore, 1993; Price, 1994; Younger and Elliot, 1995). Most infiltration is through the matrix, and through the fractures where flow rates are higher. Secondary opening of fissures takes place where flows converge, mainly beneath valley floors; they are generally absent under interfluves (Price, 1994). Within these fissures, flow rates are much higher; they are an asset to water abstraction, but may also transmit pollution (Atkinson and Smith, 1974; Price *et al.*, 1992; Banks *et al.*, 1995). Solutional enlargement to the scale of explorable caves is relatively unusual in south-east England, but active caves beneath the Water End sinkholes and the relict cave at Beachy Head demonstrate this aspect of chalk karst hydrology (Reeve, 1979). Many more caves are known in the more indurated chalk of France, but stronger lithologies at the ends of the English outcrop, at Flamborough Head and Beer, contain only fragments of cave passage.

Karst of the minor limestones

Silurian limestones include the Aymestry and Much Wenlock Formations which form parallel escarpments in Shropshire where there are weak shales below, between and above them. Each limestone is less than 30 m thick, and though they are strong rocks, they are impure, argillaceous and well bedded. They have almost no solutional features on their outcrops, and are not exploited as aquifers.

Devonian limestones form numerous small outcrops scattered through the structurally complex clastic sequences in both north and south Devon. All these Devonian rocks are lightly metamorphosed, and some of the limestones are commercially known as marbles, though their scale of recrystallization has not been enough to destroy their fossil shell structures. Surface karst landforms are insignificant in the fluvial landscapes, even on the largest outcrops forming much of the headlands on both sides of Tor Bay, but there are many notable caves. Berry Head has many small caves formed at the marine/freshwater interface during high interglacial sea levels of the Pleistocene (Proctor, 1988; Proctor and Smart, 1991). Kent's Cavern at Torquay has about 400 m of large, old, phreatic passages forming a maze now truncated in the wall of the Ilsham Valley (Proctor and Smart, 1989). These contain thick clastic sequences interbedded with stalagmite floors, of which the lower one is over 350 000 years old (Proctor, 1995). The faunal remains are extremely important (Campbell and Sampson, 1971), including an abundance of pre-Anglian cave bear bones in the lower sediments and Devensian mammoth, rhinoceros and deer in the middle layer. Tornewton Cave lies about 10 km inland; the cave is short, but the sediments contain a rare sequence of Devensian and Wolstonian (or Anglian) cold faunas separated by an interglacial bed with hippopotamus and hundreds of hyaenas (Sutcliffe and Zeuner, 1962). Further west, the largest caves in Devon lie in the Buckfastleigh limestones, and the Kitley Caves, near to Plymouth, are small phreatic chambers with spectacular wall notches formed at past water tables. In north Devon, the limestone outcrops are even smaller, but do contain a number of small caves east of Ilfracombe, including Napps Cave with its remarkable aragonite deposits.

Permian carbonate is represented by the impure, thinly bedded dolomites of the Magnesian Limestone, whose outcrop extends from Nottingham to Middlesborough. In County Durham it is a productive aquifer with a high fissure permeability. Karst landforms include lines of sinkholes along some of the outcrop margins along the low escarpment, and small abandoned caves in some of the valleys which cross it. The shallow gorge through Creswell Crags, on the Nottinghamshire/Derbyshire border, has five caves exposed in its walls, all of which are truncated fragments of phreatic rifts and mazes. Stalagmite layers within them have been dated back as far as 300 ka (Rowe, Atkinson and Ienkinson, 1989), and the clastic sediments have yielded important Devensian and earlier animal and human remains (Jenkinson, 1984, 1989). The site is best known as the type locality of the Cresswellian culture which occupied the cave in the late Devensian interstadial. Further north, the Knaresborough Gorge contains some large active tufa screens (Burgess and Cooper, 1993), and the tectonic fissures of Farnham Cave and Smeaton Pot contain limited features of solution and calcite redeposition (Lowe, 1978; Brook *et al.*, 1988).

Jurassic limestones have extensive outcrops in the scarplands across the heart of England (Downing, 1994). The Great and Inferior Oolites form the broad escarpment of the Cotswolds

(Figure 1.2), where the dip slope is crossed by numerous shallow dry valleys, mostly floored with Devensian head. The Great Oolite is the more productive aquifer; its high fissure density makes it a diffuse flow aquifer, where the flow catchment boundaries are poorly defined due to the low component of conduit flow (Smart, 1976; Atkinson and Smart, 1977). There is no input of allogenic drainage, and sinkholes are few. Further north, the Lincolnshire Limestone replaces the Inferior Oolite, where it forms the broad plateaus in Northamptonshire and the scarp of the Lincoln Ridge. It is a major aquifer with secondary fissure flow (Downing and Williams, 1969; Rushton *et al.*, 1982), but has no known caves and only limited areas of sinkholes along the boundary of its cover rocks and drift (Hindley, 1965). North of the Humber, the main limestones are in the Coralline Oolite Formation, in the Hambleton Hills, southwest of the North Yorkshire Moors. Their Windypits are tectonic caves with only incidental solutional features (Cooper *et al.*, 1976, 1982), but some small active and dry phreatic caves are recorded (Cooper and Halliwell, 1976, Brook *et al.*, 1988); the remnant, phreatic, bedding plane passages in Kirkdale Cave are best known for their role as Pleistocene hyaena dens which have yielded large numbers of mammalian bones of cold and warm environments (Buckland, 1822; Boylan, 1981). The Isle of Portland is capped by the Portland Limestone, near the top of the Jurassic succession; this is heavily fissured, and contains both tectonic caves and many relict solutional caves truncated in the western cliffs (Ford and Hooper, 1964; MacTavish, 1975; Graham and Ryder, 1983).

Salt and gypsum karst

Rock salt, which is almost pure halite, forms thick units in the Triassic Mercia Mudstones, beneath the Cheshire Plain (Figure 1.2) and in other smaller Triassic basins. In Cheshire, the Wilkesley and Northwich Halites are each sequences over 100 m thick consisting of salt interbedded with mudstone. Nowhere do they survive at outcrop, but they lie beneath thick covers of permeable drift where groundwater solution has left residual breccias of collapsed mudstone over their buried outcrops (Evans, 1970; Earp and Taylor, 1986). The landforms of the salt karst are restricted to subsidence hollows, formed where circulation of the brine has allowed continued solution by influxes of unsaturated groundwater. The well known meres, including Rostherne, are flooded dolines which evolved through much of the Holocene, while many of the linear subsidence hollows, such as Moston Flash, have deepened considerably during the last hundred years in response to artificial brine pumping (Waltham, 1989).

Anhydrite occurs in the Permian succession of England, but in thinner units than the salt. At depths less than about 100 m, it hydrates to form gypsum, which in turn is normally completely dissolved within the weathering zone; only at a few sites near Ripon, in Yorkshire, does gypsum occur in temporary surface outcrops (James *et al.*, 1981). Karst features on the gypsum include numerous dolines with active subsidence, notably around Ripon (Cooper, 1986; Powell *et al.*, 1992; Burgess and Cooper, 1993; Patterson *et al.*, 1995), and large breccia pipes in the overlying beds which are the product of interstratal karst but have little modern surface expression (Smith, D.B., 1972; Cooper, 1988). The largest gypsum caves in Britain are at Houtsay, in the Permian gypsum of the Vale of Eden: they have about 100 m of tubular phreatic passages (Ryder and Cooper, 1993), and may indicate the extent of caves which are known only as fragments exposed in mines and quarries, elsewhere in the Permo-Triassic of England.

SLAUGHTER STREAM CAVE

This is a proposed GCR site, not yet designated as an SSSI

Highlights

An underground catchment area is contained in the plunging Worcester Syncline, where modern cave drainage converges on the axial zone and flows to the Slaughter Rising. Active and abandoned passages, within the Slaughter Stream Cave and adjacent caves, show strong stratigraphical and structural guidance of their origin and development. Palaeokarstic features, including conduits partially infilled by Triassic iron ores, and abandoned caverns containing mammalian bones, provide evidence of a long and complex history of cave development.

Introduction

The Slaughter Rising lies on the eastern bank of the River Wye, south of Symonds Yat (Figure 7.2),

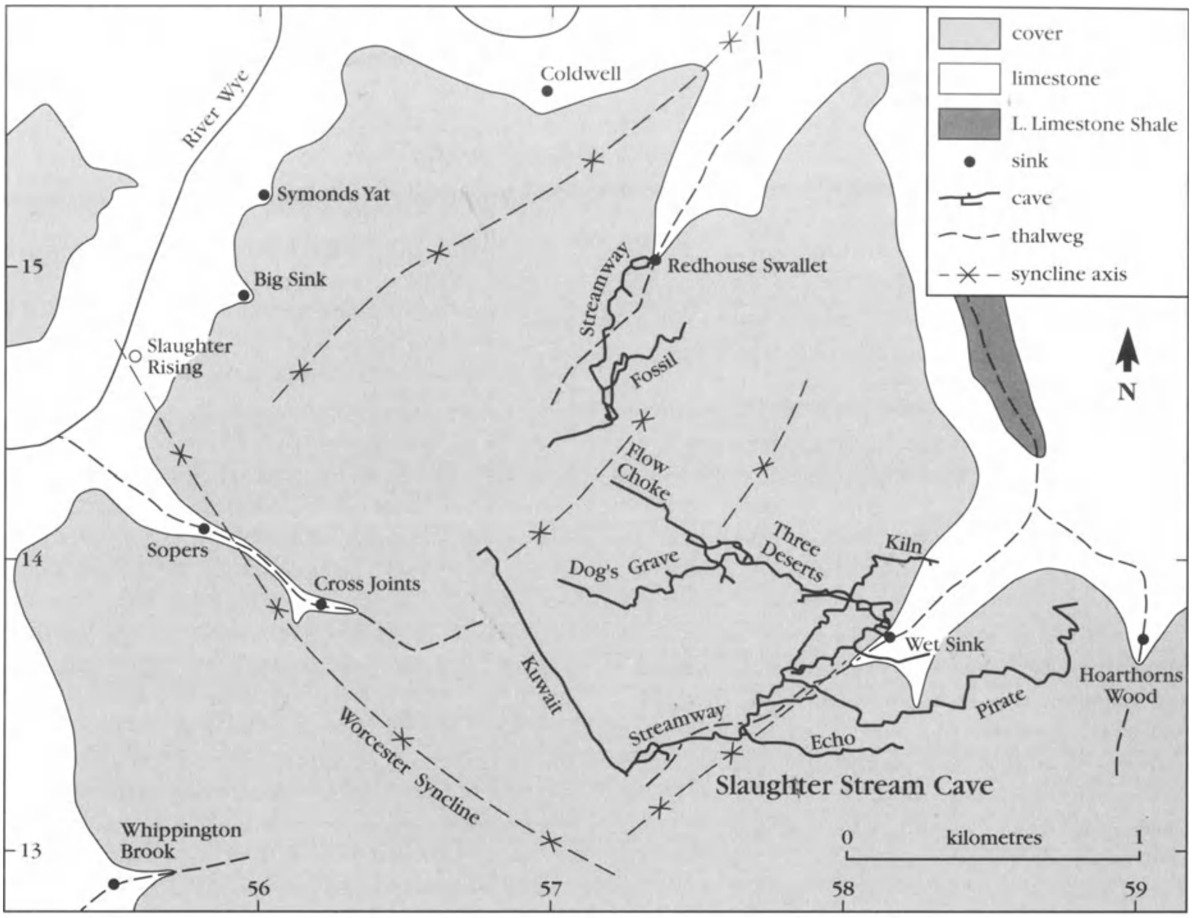


Figure 7.2 Outline map of the caves in the catchment of the Slaughter Rising. The cover rocks are the Drybrook Sandstone and the Upper Coal Measures. All the sinks marked on the map have been dye traced to the Slaughter Rising (from survey by Royal Forest of Dean Caving Club).

and gathers allogenic input from sinkholes as far apart as Coldwell Swallet, Whippington Brook Swallet and Hoarthorns Wood Swallet (Figure 7.2). It is the resurgence for nearly all the underground drainage which gathers in the Worcester Syncline, an arm to the north-west off the Forest of Dean basinal structure. Most of the feeder swallets have developed where water runs off the cap of Holkerian Drybrook Sandstone, and down through fissures in the Arundian Whitehead Limestone and the Chadian Crease Limestone, each about 25 m thick. The main cave passages are developed in the underlying Courceyan Lower Dolomite, which is about 70 m thick. Coldwell Swallet lies in the Crease Limestone as the Whitehead is overstepped by the Coal Measures. The Slaughter Stream Cave is the most extensive cave known within the catchment. Its 12 km of mapped passage are only a small part of the total network behind the Slaughter Rising; cave passages on both flanks of the Worcester Syncline include high-level remnants and the active drains, all forming parts of an ancient system which is still evolving.

The catchment of the Slaughter Rising was defined by a series of dye tests carried out by the local caving clubs (Standing, 1967; Solari, 1974; Lowe 1989a). Subsequently the cave passages were explored beneath Wet Sink (now known as Slaughter Stream Cave) and Redhouse Swallet (Clark, 1991; Taylor, 1993). Stratigraphical control of the cave geomorphology demonstrates the important role of inception horizons (Lowe, 1992b, 1993) and also the links with the palaeokarst and hematite deposits widespread in the Forest of Dean limestones (Trotter, 1942; Welch and Trotter, 1960).

Description

The only access to the Slaughter Stream Cave (Figure 7.2), most of which lies under the impermeable cover, is through immature fissures beneath Wet Sink, into a series of shafts and narrow rifts which meets the Main Stream Passage 52 m below the entrance. Downstream the passage is generally 2 m wide and up to 4 m high, but the stream is lost into flooded passages south of an overflow route through unmodified, partly sand-filled, solutional tubes. There are two inlet passages from the east; both are series of narrow canyons and rifts, of which Pirate Passage reaches over 1800 m. The stream is rejoined beyond its flooded section, and Echo Passage is another inlet with water emerging from a choked sump. The lower streamway is up to 15 m high, with walls in heavily corroded rock. A canyon develops in the floor of a wider phreatic roof passage, until the stream is again lost into flooded bedding planes. A primitive bedding passage at roof level continues above the sump, and turns north-east into Kuwait Passage. This long rift is 2 m wide, with black walls and white calcite formations, but becomes narrower after 1000 m, where a small streamway is met, and the current end of the cave is a static sump.

Upstream from the entrance series, Zurree Aven leads up into extensive upper passages. Abandoned passages lead into the Chunnel, which continues 10 m wide and 6 m high westward into a complex of rifts and chokes. Kiln Passage is an inlet with about 500 m of twisting canyon, ending near the surface excavation of Kiln Hole. The Three Deserts Series continues to the west and into Dog's Grave Passage. At low level, an abandoned streamway leads to a high rift with large crystals on its walls. The upper passage continues as a large tunnel to Helictite Rift, well decorated with helictites and dripstone of white calcite; this continues over a stream which is covered by a calcite crust. A short, choked passage leads over white calcite flakes into the Snow Garden, a high, narrow streamway with a white calcite floor. Flow Choke Passage is a series of abandoned rifts north-west from the Three Deserts into a major abandoned streamway, 10 m high, continuing in a straight line to a choke cemented by calcite flowstone.

The 12 km of mapped passages in the Slaughter Stream Cave cover a vertical range of 99 m. They include a magnificent active vadose streamway cut beneath a primitive phreatic route, as well as several, abandoned drains modified by vadose trenches. Abandoned passages contain thick beds of sand and silt, with sedimentary structures including ripple marks, and parts of Kuwait Passage and Helictite Rift contain very fine calcite formations. The cave contains an unusual number of mammalian bones, which have fallen or been washed in, or have been relocated by stream activity. The streamway contains bones of a hippopotamus, at least 125 000 years old. The Graveyard in the high levels near the Chunnel contains bones of domesticated species, and possibly human remains, and an auroch bone (2000-3000 years old) was found in a nearby passage. The skeleton of a dog in Dog's Grave Passage, and associated tracks in a nearby oxbow passage, present a mystery with regard to the animal's route into the system.

North-west of Wet Sink, a major stream is swallowed in a blind valley near Redhouse Lane, where over 2000 m of cave is now known in Redhouse Swallet (Figure 7.2). A series of narrow rifts, originally choked with sediment and boulders, drops about 30 m into the main stream passage. Downstream, this passage is of varied morphology, with deep canals, high rifts, collapse zones and abandoned loops over flooded sections; it ends in a choke over a flooded rift. The Fossil Series has a small inlet stream and several large chambers, including Bowen Chamber with its thick clastic deposits, and Missed Chamber.

Interpretation

The caves of the Slaughter catchment demonstrate the role of inception horizons, where chemical contrasts within the carbonate sequence have created specific stratigraphic horizons favourable to solution and cave development. These inception horizons are features of the host rocks, and the cave inception may be traced back to a limited scale of cavity opening soon after the limestones were formed, long before the caves were greatly enlarged by drainage beneath more modern landscapes (Lowe, 1992, 1993).

Though exploiting fissures locally, the main drain and its tributaries in the Slaughter Stream Cave are cut beneath a bedding-guided solutional conduit, segments of which survive unmodified where drainage is captured by fault-guided short-circuits. Drainage in both the Slaughter and Redhouse cave systems initially flows away from the rising, maintaining a constant horizon and following local hydraulic gradients south-westwards into the trough of the Worcester Syncline (Figure 7.2). Stratigraphical guidance was crucial to underground conduit growth, but fracture guidance was important locally, where joints facilitate minor conduit sidestepping within the same horizon. Faults may be involved in offsetting and linking zones of stratigraphical guidance, but it is unconfirmed whether the major linear rifts, including Kuwait Passage, follow fault planes.

Superimposed across the Worcester Syncline is a series of gentle, asymmetrical folds whose axes plunge towards the main fold trough (Figure 7.2). These minor folds have fundamentally influenced the direction of past and present underground drainage. The modern drainage collects in the main syncline and then escapes by upward leakage through fissures to the impenetrable Slaughter Rising. It is unknown if the main cave conduits behind the rising are deep in the Lower Dolomite, at the stratigraphic level of the Slaughter Stream Cave; alternatively, they may lie in the Crease or Whitehead Limestones downstream of phreatic lifts from the known caves.

The earliest speleogenesis in the area probably predated the downcutting of the Wye, by many millions of years. A pre-Triassic maturity is likely, and inception activity commencing during the Carboniferous is a realistic possibility. The relationships of abandoned high-level passages to the active drains and to nearby relict cave fragments are yet to be elucidated, but the abandoned high-levels, of Flow Choke and Dog's Grave Passages, may once have carried drainage away from the River Wye, eastwards into the groundwater reservoir of the Forest of Dean basin. If this was so, these drains must have been conceived before the Wye incised its valley and captured the karst drainage during the late Tertiary. Whether the trunk passages in the main caves were contemporary with the passage segments truncated in the sides of the Wye Valley is unknown. The clastic sediments could provide the evidence; these may reflect a temporary engulfment of part of the proto-Wye during incision, or may have been deposited in pre-existing tunnels cut before the incision. It has been suggested that the Wye Valley relict caves formed before iron ore emplacement in the Triassic period, and some clastic sediment may be of late Triassic age (Lowe, 1993). Large chambers in Symonds Yat and Cross Joints Swallets were formed by solution of the Crease Limestone and collapse of the overlying beds. These resemble the voids in the local iron ore mines, where high-grade ore was removed from the host bedrock (Lowe, 1989, 1993).

Existing phreatic conduits, and any less well developed parallel or tributary routes of the same age, were drained following uplift, and then carried underfit vadose streams. After uplift, phreatic flow continued at lower levels, along fissures and inception routes that were conceived and partially developed before the uplift. Development of these lower routes was by the underflow which they carried along their own favoured hydraulic gradients, which need not have mirrored the hydraulic gradient favoured by simultaneous flow at higher levels. In the Slaughter Stream Cave, deep underflow continued, probably augmented by fissure leakage from underfit streams in the overlying primary conduits. As the Wye cut through the Worcester Syncline, favourable inception horizons in the aquifer were exposed first on the northern and southern fold flanks, and later in the fold core. With the breaching of each stratigraphic horizon in the fold core, a potential spring site was exposed and offered hydraulic gradients more advantageous than those towards more distant outlets.

Whether the high-level routes were imprinted within their guiding horizons before or after the minor fold ripples formed across the limbs of the Worcester Syncline is unknown. However, after downcutting and rejuvenation, these minor synclinal troughs offered a means of turning vadose drainage away from its regional, eastward trend and towards the advantageous westward route along the Worcester Syncline. In the Slaughter Stream Cave the roofs of the Main Streamway and its tributaries are at a constant horizon, close to the axis and across the limbs of a minor syncline. Redhouse Swallet lies in another syncline, and more parallel independent drains may exist in the other synclines where the drainage direction changes beneath abandoned high-level passages.

The presence within Slaughter Stream Cave of bones, animal tracks and ancient man-made detritus indicates that other surface connections have existed in the past. Some detritus, and bones of butchered domesticated species, may have been 'tipped' into shafts or active collapses at the surface, eventually to enter the cave. Other debris, of wild and domesticated species, might have been washed in, though this would have required sinks far more open than those which were excavated to give the modern access. Other bones, and local guano deposits, indicate that bats and small rodents have occupied the caves, but how they entered the caves is unknown. Most intriguing are the tracks and skeleton found in Dog's Grave Passage, which lies far beneath the impermeable cover remote from any sinkhole sites. It is difficult to imagine how the animal gained access, either uninjured or sufficiently uninjured to move around until its eventual death.

Conclusions

The underground catchment of the Slaughter Rising is a karst of synclinally folded carbonates whose stratigraphy and structure have guided the inception and development of a long and complex cave system. Among the known cave passages and chambers, there are elements which were well developed before late Triassic times. Underground drainage directions have changed radically during this long history, largely in response to the deepening of the Wye Valley, but possibly due partly to the earlier effects of tectonic deformation. Traces of animals preserved in the caves indicate that their recent history includes further significant changes.

BUCKFASTLEIGH CAVES

Highlights

The Buckfastleigh caves are the most extensive in England which are developed in pre-Carboniferous limestones.

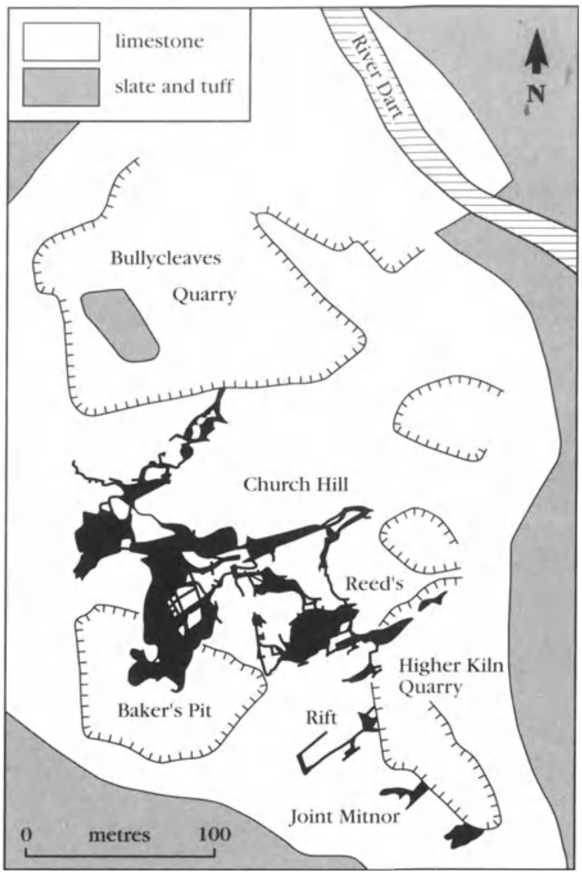


Figure 7.3 Outline map of the caves of Buckfastleigh (from surveys by Devon Speleological Society).

Introduction

A series of caves lie in separate outcrops of structurally complex, Middle Devonian limestone in the valleys of the River Dart south-east of Dartmoor in Devon (Figure 1.2). The largest of the caves are beneath Church Hill on the outskirts of Buckfastleigh, but significant other caves are at Pridhamsleigh and further west in the Dean Valley. Many of the cave passages and aspects of their geomorphology have been described by Hooper (1956, 1960), Vowler (1980) and Neill (1988).

Description

The Devonian limestones occur mainly as a series of reefs within clastic formations; they now have very dispersed outcrops, due to the reef distribution and the subsequent folding, faulting and thrusting. Regional metamorphism has left them in a slate sequence; the carbonates are known locally as marbles, but the limited recrystallization has not destroyed their many fossils. In places the limestones contain interbedded volcanic ash and are cut by small lamprophyre dykes.

The most extensive caves lie in Church Hill, Buckfastleigh, which is a small limestone outlier partly underlain by thrust planes. The system of Reed's Cave and Baker's Pit extends through most of the hill to entrances in quarries on opposite flanks (Figure 7.3). Though contained within an area of less than 4 ha, the caves have been surveyed to a total length of more than 3000 m, as they form an intricate maze on several levels - though some of the length is within complex collapse areas. Most passages are small in cross-section, except where they open out into chambers which survive between the zones of collapse. Some of these chambers contain small, well-preserved calcite and aragonite deposits. Several other small caves within the hill are not connected to the main system, but contain important calcite and clastic deposits. Joint Mitnor Cave

(Figure 7.3) contains one of Britain's richest Ipswichian mammalian bone deposits (Sutcliffe, 1960). The smaller caves in the Higher Kiln Quarry are important bat sanctuaries.

Pridhamsleigh Cavern contains more than 1000 m of passages. The relict parts of the cave include chambers up to 30 m wide connected by complex series of solution tubes and bedding plane passages. These contain considerable quantities of mud, with minor calcite and aragonite deposits.

Beneath the relict levels lies an active phreas, with several levels of development leading off a flooded shaft more than 40 m deep, in which stalagmites have been observed at a depth of 12 m. A large chamber, Pridhamsleigh II (Mulholland, 1992), lies within the flooded zone, which is hydrologically linked to the adjacent River Ashburn, with water levels in the cave fluctuating by up to 10 m in response to rainfall.

The small group of caves in the Dean Valley includes Bunker's Hole and Potter's Wood Cave. These consist mainly of phreatic chambers which have been intersected by small vadose streamways, and are notable for their range of mineral deposits, including goethite, with spectacular calcite helictites and aragonite crystals growths.

Interpretation

The caves of South Devon are largely phreatic in origin, and their morphology exhibits significant influence by faulting and vein mineralization within the limestone. There has been only limited vadose modification. The multi-level sequence of active and abandoned caves may be correlated with phases of valley incision and terrace formation in the Dart Valley, through at least the later parts of the Pleistocene. The speleothems 12 m below the present water surface in Pridamsleigh Cavern indicate that valley floor aggradation has caused a rise in the water table subsequent to the valley incision which had earlier drained most of the caves. The sediment and speleothem sequences within the caves may enable a more detailed chronology to be constructed through this interval, while the submerged speleothems may provide valuable data on sea-level fluctuations during this time.

The bone deposits of Joint Mitnor Cave contain remains of elephant, hippopotamus, lion, hyaena, deer and fox within an assemblage richer in species than any other Ipswichian deposit in Britain. The fauna is indicative of a warm environment when the bones accumulated as a pit-fall deposit, forming a debris cone beneath a shaft which was open to the surface.

Conclusion

The small and complex cave systems are the longest which are developed in the Devonian limestones. They show multi-level development related to downcutting and then aggradation in the adjacent valleys of the River Dart during the Pleistocene. One cave contains an unequalled interglacial assemblage of mammal bones, and the secondary carbonate mineralization is notable for including aragonite.

NAPPS CAVE

Highlights

This cave has only a short length of relict passages, but these contain some of the finest aragonite speleothems in Britain.

Introduction

Napps Cave was exposed at the turn of the century in a quarry into the eastern slopes of Napps Hill, west of Combe Martin on the north coast of Devon (Figure 1.2). The cave is a small remnant system of abandoned passages developed in a dipping band of limestone less than 20 m thick within the Middle Devonian Ilfracombe Beds. The cave is rarely visited as its aragonite is very fragile, and is briefly recorded only by Vowler (1980, 1981).

Description

Napps Cave consists of several elongate chambers connected by small rifts. Access to the cave is via one of these rifts where it has been intercepted by a quarry face, and about 200 m of passages have been explored. The walls of most of the rifts in the cave are decorated with anthodites - spectacular clusters of radiating aragonite crystals up to 70 mm in length (Figure 7.4).; these vary from pure white to browns and greys due to iron stain-

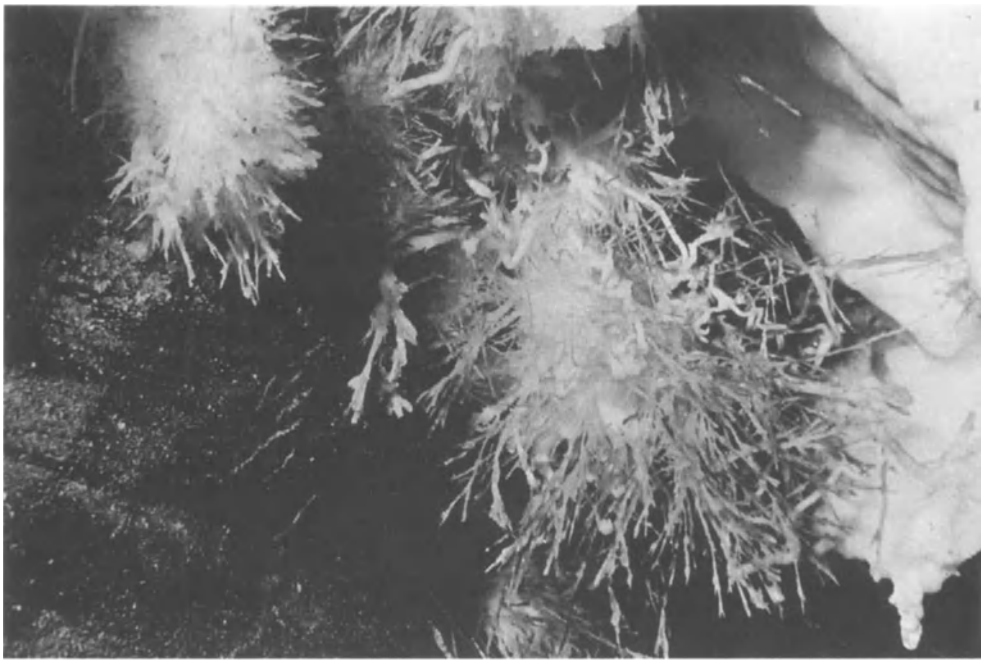


Figure 7.4 Clusters of delicate aragonite needles on the walls of a rift in Napps Cave. (Photo: F. Vowler.)

ing. Floor deposits include layered stalagmite of both aragonite and calcite, overlying thick clay, and there are various dripstone features including stalactites and very delicate helictites.

Interpretation

The morphology of Napps Cave is that of a short network of broad tubular passages connected by narrow rifts. The whole cave is a phreatic remnant, parts of which have developed along the steeply dipping boundaries between the limestone and adjacent mudstone.

Cave anthodites are generally the product of deposition by slowly moving film water which is lost to evaporation. This makes them distinct from dripstone which is deposited due to loss of carbon dioxide. Aragonite is the common mineral of anthodites, and generally precipitates in place of calcite where the carbonate is high in strontium; this situation is more common in thin limestones within clastic sequences. The precise controls on aragonite deposition and anthodite growth are unknown, but the environment of Napps Cave appears to fit the general case.

Although this cave is very small, the aragonite formations which it contains are the largest and most spectacular in Britain. The controls on their development await detailed research.

Conclusion

This isolated small cave is highly valued for the aragonite crystal formations, which are the largest and finest in Britain.

CULL-PEPPER'S DISH

Highlights

The doline fields in the chalk karst of Dorset are noteworthy for their density, and Cull-pepper's Dish is the largest, most spectacular and best developed of the individual dolines. Nearly all the dolines lie in a covered karst where Tertiary and Quaternary sands overlie the chalk karst and their relationship to datable archaeological remains suggests a very short time-scale of evolution.

Introduction

The Dorset heathlands, along the northern edge of the western Hampshire Basin, are pock-marked with an impressive array of dolines. Over 370 are recorded on Puddletown Heath, and there are more than 100 on the smaller Southover Heath. Cull-pepper's Dish is the largest and most spectacular individual doline, and occurs on the Bryants

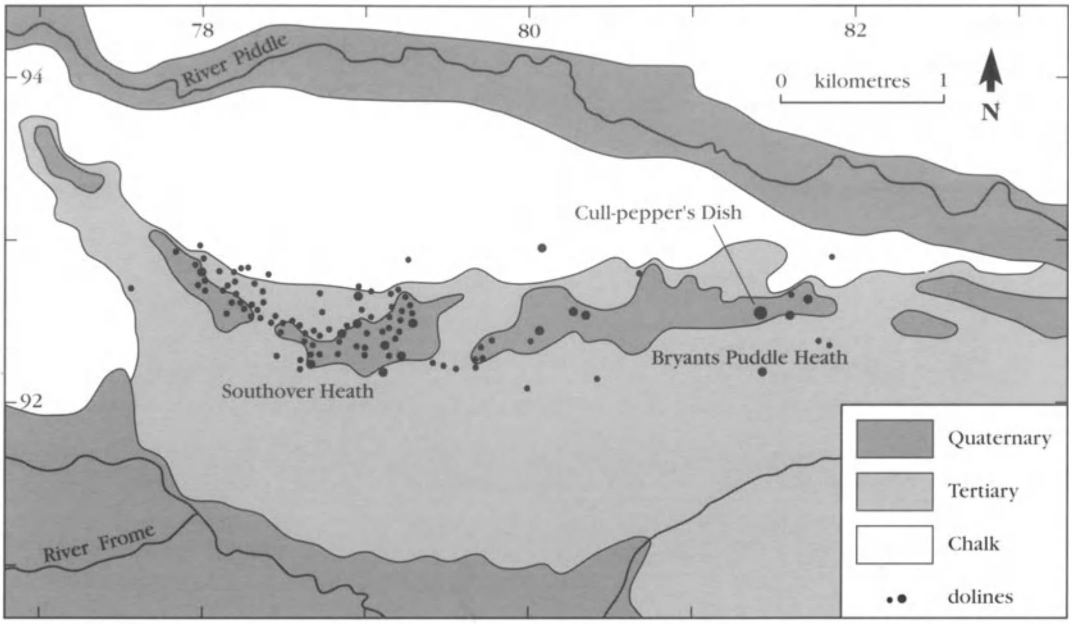


Figure 7.5 Geological map of Cull-pepper's Dish and the doline fields of the adjacent Dorset heaths on sediments overlying the chalk. The Tertiary rocks include the Reading, London and Bagshot Beds. The Quaternary rocks include plateau gravels along the central strip of heathland, and also alluvium in the valleys north and south (after Sperling *et al.*, 1977).

Puddle Heath, along with just a few other dolines of larger than average size (Figure 7.1).

The Dorset doline fields gained early mentions by Stevenson (1812), Mansel-Pleydell (1873), Fisher (1858, 1859) and Reid (1899), and the site geology is described by Wilson *et al.* (1958). The doline genesis was explained by Sperling *et al.* (1977), and again further considered by House (1991, 1992) and Goudie and Gardner (1985). These chalk karst dolines are comparable with some of those developed on the Carboniferous Limestone (Coleman and Balchin, 1959; Thomas, 1974).

Description

The doline fields of the Dorset heathlands reach densities of 99 km^{-2} on Southover Heath and 157 km^{-2} on Puddletown Heath, where the majority of the dolines are only about 10 m across. Cull-pepper's Dish, located in woodland on Bryants Puddle Heath, is the largest single doline in the area, and lies in a region of lower doline density (Figure 7.5). It is a conical depression, 21 m deep with a mean diameter of 86 m, slightly elliptical in plan, with uniformly graded sides sloping at about 30° (Sperling *et al.*, 1977). The whole depth of the visible doline is developed within the clastic sediments which survive over the chalk. These are the sands and gravels, with minor clay horizons, of the Eocene Reading Beds, overlain by a thin spread of Pleistocene plateau gravels. Beneath the Reading gravels, the Cretaceous Upper Chalk is a soft pure-white limestone over 100 m thick, which dips at less than 3° (Wilson *et al.*, 1958). Soils mask all the bedrock, except at a few exposures of unconsolidated sand. In the floor of the doline, a sinkhole is choked with soil between loose blocks of chalk; there is no evidence of any collapse structure.

Two other large dolines occur in the immediate vicinity, but Cull-pepper's Dish is the largest and least vegetated, and consequently the best exposed. They all lie in the heathlands at altitudes around 80 m, overlooking the chalk slope to the valley floor 50 m below. None of the dolines bears any relationship to the surface drainage pattern which is poorly developed on the very permeable sand.

Interpretation

The formation of the dolines was initially ascribed to the subsidence of the clastic material into pipes in the subjacent chalk, owing to the percolation of rainwater, which dissolved the chalk (Fisher, 1859). Reid (1899) further suggested that the moist climate and acidic peaty soils combined with the vertically extensive vadose zone to provide conditions favourable to doline development. It is notable that the doline fields lie along the ridge where groundwater can drain rapidly downwards to a deep water table within the chalk. The same location provides the required thickness of the sediment cover, where it thins enough to permit substantial through drainage.

Solution of the underlying chalk by highly acidic percolation water from the peaty heathlands may be locally concentrated by discontinuous clay-rich beds in the very variable Reading Beds. This creates small cavities and fissure networks in the Chalk. Collapse may occur when they reach a critical size, but is unlikely to be a significant process as there are no comparable collapse features in the exposed chalk to the north. With or without any collapse, the loose unconsolidated sands ravel and slump into the chalk voids, and are carried to depth by the vadose drainage. Most of the dolines are formed where the Reading Beds are capped by Quaternary gravels - which are even more permeable and prone to ravelling. At some sites, less concentrated recharge may cause solution and subsequent lowering at the interface between the chalk and the overlying material, producing more gradual surface lowering.

The age of the dolines is open to debate. They clearly postdate the Pleistocene gravels in which they are formed, but these deposits on the higher levels of the heaths are probably older than Ipswichian. New dolines continue to form, and small collapses have been recorded in recent years (Goudie and Gardner, 1985). Doline development within the last 4000 years has disturbed a Bronze Age burial site (House, 1992), but this is located west of Dorchester and correlations with Cull-pepper's Dish are tenuous.

Conclusions

Cull-pepper's Dish is the largest and most spectacular of the dolines developed on the covered chalk karst of Dorset. It provides an excellent example of a conical subsidence doline formed in unconsolidated sands and gravels of Tertiary and Quaternary age.

THE MANGER

Highlights

Incised into the escarpment of the Berkshire Downs, the Manger is one of the finest chalk dry valleys, or combes, in Britain. It is especially notable for the series of steep chutes which serrate its southern flank and for its well documented floor sediments and downslope alluvial fan.

Introduction

The Manger is one of many short dry valleys which etch the chalk escarpments of southern England. It is incised into the northern escarpment of the Berkshire Downs, overlooking the Vale of the White Horse (Figure 7.1). The Manger's importance lies in the sediments which mantle the floor of the valley and fan out into the Vale. These provide evidence for the geomorphic evolution of the valley, which is pertinent to the many other similar dry valleys in England's chalk downland. The debate over the origins of the chalk dry valleys has extended through many years (Smith, 1975b). The two main suppositions are that the valleys were cut by normal stream action which has since gone underground (Chandler, 1909; Fagg, 1923, 1954; Sparks and Lewis, 1957; Small, 1962, 1964), or that they were cut by runoff and solifluction processes under periglacial conditions (Reid, 1887, 1892; Bull, 1936, 1940; Kerney *et al.*, 1964; Sheail, 1971). The geomorphology of the Manger has been described by Arkell (1947), Beckinsale (1954), Paterson (1977) and Goudie and Gardner (1985).

Description

Overlooking the broad Vale of the White Horse, the chalk escarpment rises steeply for 100 m, and is scored by many short, steep, dry valleys. The Vale is developed largely on the Mesozoic Gault and Kimmeridge clays, which are separated from the chalk by a narrow band of Greensand cropping out at the foot of the scarp. The Lower and Middle Chalks form the bulk of the escarpment, with some of the Upper Chalk surviving along the crest; they dip 1-3° south.

The Manger combe is a rounded valley about 500 m long and 50 m deep, cut into the scarp-face

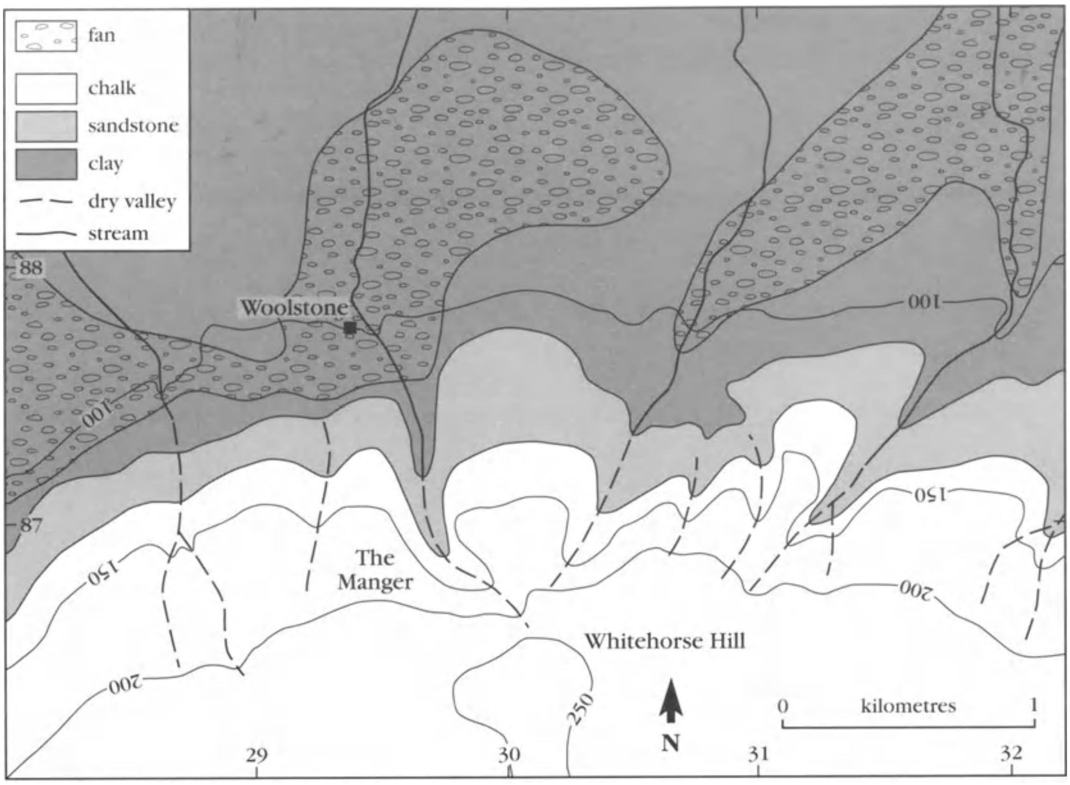


Figure 7.6 Geological map of the Berkshire Downs scarp face, with its dry valleys, or combes, including the Manger, and associated fan deposits.

below Whitehorse Hill (Figure 7.6). It is totally dry. Its floor gradient is up to 36° near the head of the combe, but this eases to less than 10° lower down. The long profile is a smooth graded slope, convex over the upper rim and concave on the lower slopes; the cross-valley profiles are generally symmetrical. On the southern slope of the combe, the left bank is corrugated by a series of ten much smaller combes or furrows (Figure 7.7). These are each about 120 m long and 40 m across, and are incised by about 5 m.

Both the geomorphology of the valley and the stratigraphy of the combe floor alluvium and fan deposits were investigated by Paterson (1977), using electrical resistivity surveys, field mapping, excavated trenches and over 800 augered boreholes. The floor of the valley is cut in the chalk until it breaks through to the underlying sandstone (Figure 7.8). Bedrock is covered by up to 5 m of white, angular, chalk rubble and silt; above this are up to 3 m of grey-brown, humic, chalk silts with occasional chalk and flint fragments, capped by a thin layer of topsoil (Paterson, 1977). These deposits can be traced the whole length of the combe, and the white chalk rubble alone extends into the gently graded fan of chalk detritus which reaches out over 2 km from the foot of the scarp face (Figure 7.6). A terrestrial molluscan fauna is preserved in a cryoturbated chalk debris in the equivalent fan below the combe to the west of the Manger (Paterson, 1971, 1976).

Interpretation

The origin of the Manger was ascribed to spring sapping when the water table was higher (Arkell, 1947), and the gullies on the south side were interpreted as old spring sites which represent former positions of the spring line. This concept is, however, incompatible with the large amount of sediment remaining within the Manger. Excavation by surface run-off and solifluction is indicated by depositional evidence from the banded gravels in the combe floor (Paterson, 1977). The Manger was carved out by an abundance of meltwater from annual snow banks on the summit of the escarpment. This was aided by seasonal solifluction on the valley sides, which led to major deposition on the valley floor, but the importance of meltwater transport is demonstrated by the large fans of well sorted chalk

Figure 7.7 The dry valley of the Manger seen from its head; the chalk of its south slope is scored by the series of furrows or small-scale combes, above the flatter valley floor veneered by solifluction debris. (Photo: A.C. Waltham.)

debris extending out onto the clay floor of the Vale.

The dry furrows, or small-scale combes, in the south slope of the main dry valley have been attributed by Paterson (1977) to the sites of former springs, which supplied water into the valley and provided the loci for intensive freeze–thaw action and other periglacial processes. Alternatively, they may be avalanche tracks (Goudie and Gardner, 1985), dating from Devensian periglacial environments. Avalanches can cause erosion of a weak bedrock such as frost-shattered chalk. The north-east aspect of the slope makes it a prime site for snowdrift accumulation in winds from the south-west; spring avalanches would scour the thawing, weakened, active layer within the chalk, and would repeatedly follow the same tracks.

The volume of sediments in the alluvial fan represents about a quarter of the volume of the Manger valley. This implies that much of the material eroded is still present in the fans, and suggests a relatively recent origin for the deposits. The stratigraphy of the deposits, the nature of their

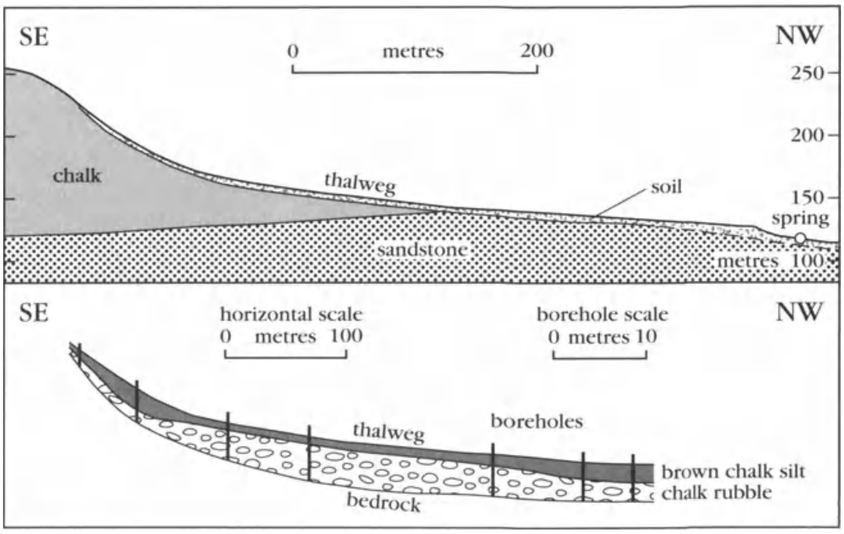


Figure 7.8 Long profiles of the floor deposits in the Manger. The upper profile is drawn to true horizontal and vertical scales. In the lower profile the soil thicknesses are increased by a factor of 8 (after Paterson, 1977).

molluscan faunas, the absence of deep leaching and chemical weathering, and their geomorphic relationship to the surrounding land, all indicate a Devensian age. Paterson (1976) assigned the molluscan assemblage to the Allerød Interstadial of the Late Devensian. He then dated the soliflucted chalky debris to the succeeding cold environment of the Loch Lomond Stadial (10.8–10.3 ka), and the brown chalk silts to hillwash following clearance and cultivation in the Holocene. Though there is no surviving evidence of pre-Devensian sediments, the scale of the Manger, compared to the volume of sediments, demonstrates that the valley was incised over several cold phases during the Pleistocene. In each of the intervening warmer stages, and in the Holocene, renewed underground drainage led to the combe becoming a temporarily inactive dry valley.

Conclusions

The Manger is a spectacular combe incised into the scarp face of the Berkshire Downs; it is a particularly fine example of this type of dry valley which is distinctive of England's chalk karst. The sediments preserved in the valley floor and in a fan extending out in the Vale to the north provide striking evidence of solifluction and periglacial excavation during the Devensian glaciation. The furrows preserved on the southern side may represent either an abandoned spring line or relict avalanche tracks. The dry combe is excellent evidence of the karstic preservation of a periglacial landform.

BEACHY HEAD CAVE

Highlights

Beachy Head Cave is the longest and best example in Britain of a phreatic conduit developed in chalk.

Introduction

Beachy Head Cave is entered from an opening close to the foot of the cliffs 400 m west of Beachy Head, East Sussex (Figure 7.1). The cave is formed in the lower beds of the Senonian Upper Chalk, which locally dips very gently to the west. It lies immediately above a tabular flint band 20 mm thick. There is no apparent relationship between passage orientation and surface topography. The cave was described by Reeve (1980), and cave development in chalk has been reviewed by Lowe (1992a).

Description

The opening to the cave is on a ledge 4 m above the base of the cliff, where marine erosion has caused cliff retreat to intersect a remnant phreatic cave passage (Figure 7.9). Most of the cave comprises a phreatic tube about a metre in diameter but there are a few phreatic domes and small avens. The cave is close to horizontal, as it follows the one bedding horizon, but it changes level to follow this horizon across some small faults.

The passage north-east from the entrance

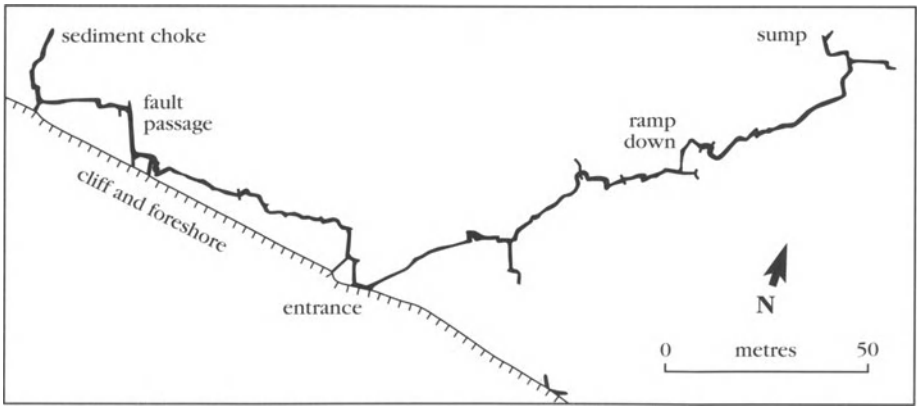


Figure 7.9 Outline map of Beachy Head Cave (from survey by Chelsea Speleological Society).

extends for about 180 m, obliquely away from the cliff face, to where a static sump has prevented further exploration. Several smaller passages branch off at various points and a small cluster of botryoidal stalactites is present near the end. The western passage extends almost parallel with the cliff face, for about 145 m to a clay choke. Daylight can be seen at several places where small branch passages open to the cliff face. One short stretch of passage is aligned on a fault, and has well-developed phreatic domes in the roof.

Interpretation

The development of Beachy Head Cave appears to have been entirely phreatic; there is no evidence of vadose modification. The thin tabular flint band which floors the passage appears to have played a fundamental role in cave development. It may have acted as an aquiclude which prevented downward migration of water, and initiating cave development immediately above; alternatively its chemical contrast may have created a favourable inception horizon. The cave is phreatic, and its altitude means that it predates at least 5 m of water-table lowering. It may be much older, and it now carries no drainage flow; it is unrelated to the present topography, but its depth of over 100 m below the South Downs surface renders this of little significance. Beachy Head Cave may represent a relict example of the type of passage that must extend below currently active sinks in chalk, such as those at Water End in Hertfordshire (Walsh and Ockenden, 1982).

Conclusion

The cave is the only one in British chalk with any significant amount of accessible passage, and is therefore an important demonstration of the existence of conduits and the role of conduit flow in the heavily exploited chalk aquifer.

DEVIL'S DYKE

Highlights

The Devil's Dyke is the largest, and perhaps the most famous, of the combes developed on the chalk karst of England's downlands. Its sheer size provides ample evidence of the effectiveness of the solifluction and other periglacial processes which operated during cold stages of the Pleistocene.

Introduction

Many short, steep-sided, dry valley, or combes, are cut into the scarp face of the Cretaceous Chalk escarpments of southern England. The Devil's Dyke, incised into the northern side of the South Downs, north of Hove (Figure 7.1), is probably the largest, most spectacular and most well known of all the chalk combes. Its unusually large size is partly due to its two stages of excavation, but it still provides an excellent example of the scale of periglacial activity in the chalk karst.

Debate over the origins of the Devil's Dyke and the many other chalk combes has been extensive. The main hypotheses have centred on either spring sapping and normal stream erosion in temperate climates (Chandler, 1909; Fagg, 1923, 1954; Sparks and Lewis, 1957; Small, 1962, 1964), or erosion under periglacial conditions by solifluction and surface run-off (Reid, 1887, 1892; Bull, 1936, 1940; Kerney *et al.*, 1964; Sheail, 1971). Specific studies of the Devil's Dyke include those by Martin (1920), Sherlock (1929), Wooldridge (1929), Bull (1936) and Small (1962, 1964).

Description

The Devil's Dyke is incised into the north-facing scarp face of the South Downs; it cuts through most of the Lower and Middle Chalk, and finishes in the Upper Greensand. The valley is just over 1000 m long, is 400 m wide at its rim and reaches a depth of over 80 m (Figure 7.10). Its head lies close to the crest of the scarp, and it descends over 130 m to debouch onto the clay vale to the north. It begins as a steeply descending, steep-sided combe with an initial gradient of about 25%, but soon develops a flatter floor. The valley cross-profiles are broadly V-shaped, but are slightly asymmetrical, with steeper slopes on the southern, downdip side. The valley walls are unusually steep, locally exceeding 30°, before flattening sharply to gentle convex profiles in their upper parts. A ribbon of soliflucted chalk debris, known as combe rock, follows the thalweg of the dry valley.

The combe heads north-east, until it swings to

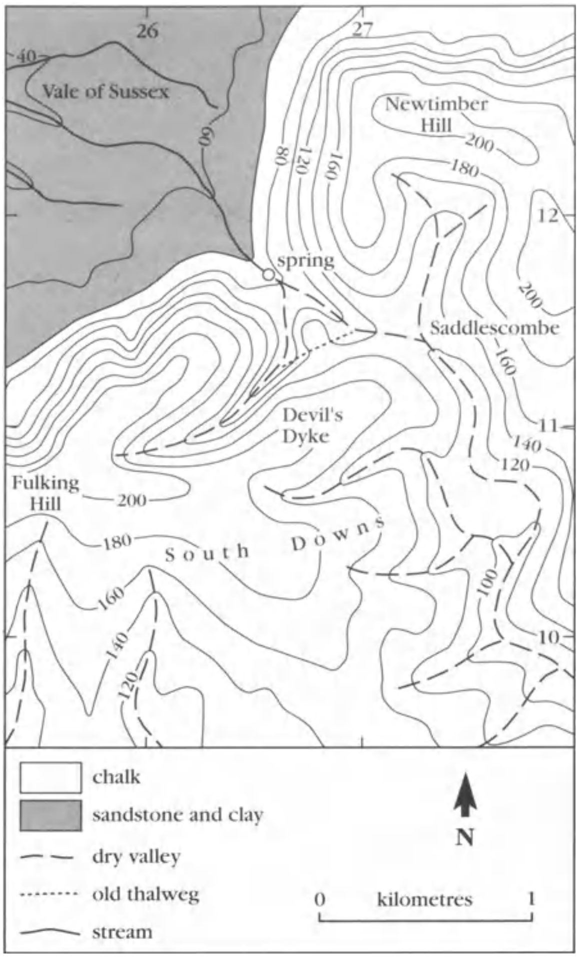


Figure 7.10 Topographic map of the Devil's Dyke and the Saddlescombe dry valleys on the South Downs.

the north, and flattens and widens considerably below a marked step. Lower down, the dry valley again steepens, and ends at the foot of the scarp above a powerful group of springs. A tributary valley on the south at its lower end is the only one to enter the combe; its head lies at a col over into a large dry valley system cut in the dip slope (Figure 7.10). Their is no trace of any related valley feature in the clay vale to the north.

Interpretation

Early claims that the Devil's Dyke was excavated by glaciers (Martin, 1920; Sherlock, 1929) were refuted by Wooldridge (1929) and Bull (1936) who proposed that it was incised by meltwater flowing from melting snow caps during periglacial periods. Fagg (1923, 1954) suggested that the chalk dry valleys were cut by normal stream action, and were then desiccated by scarp retreat when new springs caused a fall in the local water table.

Small (1962) suggested that the Devil's Dyke was originally the upper segment of Saddlescombe, which was captured by headward erosion in a scarp face valley. The sharp bend in the combe may be regarded as the elbow of capture, and the saddle on the small spur just to the east may be a remnant of the original valley floor (Figure 7.10). Following capture, the new steeper gradient caused the valley to become overdeepend. Small proposed that this was due in part to spring sapping and associated stream erosion, a mechanism which has also been put forward for the formation of several other chalk dry valleys such as the Manger in Oxfordshire (Arkell, 1947; Sparks and Lewis, 1957). He also noted the role of erosion by snowmelt run-off during periglacial episodes, and speculated that a former impermeable capping of clay-with-flints may have caused surface drainage under wetter conditions.

The chalk is very susceptible to frost shattering in its weathering profile, followed by erosion by surface run-off and solifluction processes, under periglacial conditions (Paterson, 1977). Valley excavation in the chalk thereby progressed during each cold stage of the Pleistocene, and became dormant when underground drainage resumed during each subsequent warm stage. About half the depth of the Devil's Dyke was cut when it was a high tributary of the Saddlescombe valley system on the dip slope. Probably very late in the Pleistocene, it was rejuvenated when it was captured by the scarp face valley, and periglacial processes were enhanced on the new, steeper slopes. Even though it appears to have a two-phase history, the Devil's Dyke represents an excellent example of the massive scale of periglacial activity on the chalk downlands.

Conclusions

The Devil's Dyke is the largest and most impressive of the many combes incised into Britain's chalk karst. Its origins are complex, with capture and rejuvenation creating its present overdeepened form. The dimensions of the valley have important implications for ideas on the scale and effectiveness of solifluction and meltwater run-off on the chalk downlands of England during periglacial stages of the Pleistocene.

WATER END SWALLOW HOLES

Highlights

The Water End swallow holes are some of the largest, best developed and most accessible swallow holes in the chalk karst of southern England. They admirably demonstrate the role of conduit flow within the Chalk aquifer.

Introduction

Mimmshall Brook has a catchment area of 45 km^2 draining the high ground to the south and west of Potters Bar (Figure 7.1). It drains an area of Tertiary London Clay and flows north onto the Cretaceous Chalk outcrop. For 3 km, the stream flows in a shallow valley floored by thin gravelly alluvium which rests on the chalk. At Water End, it drains underground through a series of sinkholes. Two other small streams drain into the same broad depression, and also sink into the chalk. In wet weather, the sinks cannot cope with the flow, and a lake develops; in extreme flood, this overflows into a surface course to the west to join the River Colne. The sinking water resurges between 7 and 15 km to the east at four springs along the River Lea.

The site has been described by Wooldridge and Kirkaldy (1937), Evans (1944), Waltham (1969), Ockenden (1972) and Reeve (1979). Kirkaldy (1950) included the area in his study of chalk solution in the Mimms Valley, and Walsh and Ockenden (1982) reviewed their detailed hydrological studies of the sinkhole complex.

Description

Water End lies at the edge of the London basin where the Tertiary outlier is eroded to expose the underlying chalk. Thin sandy Reading Beds are capped by London Clay, both of which are Eocene, and Pleistocene gravels form the higher ground around the valley; chalk is exposed along the floors of the main and tributary valleys (Figure 7.11). The chalk dips to the south at less than 0.5°. Glacial till is extensive to the north, and floors a shallow trough just north of Water End.

The sinkhole complex at the end of Mimmshall Brook occupies a large blind valley up to 10 m deep. The main sinks are clustered within an area of 1 ha at the end of the Mimmshall Brook surface course; this zone normally has about 15 discrete

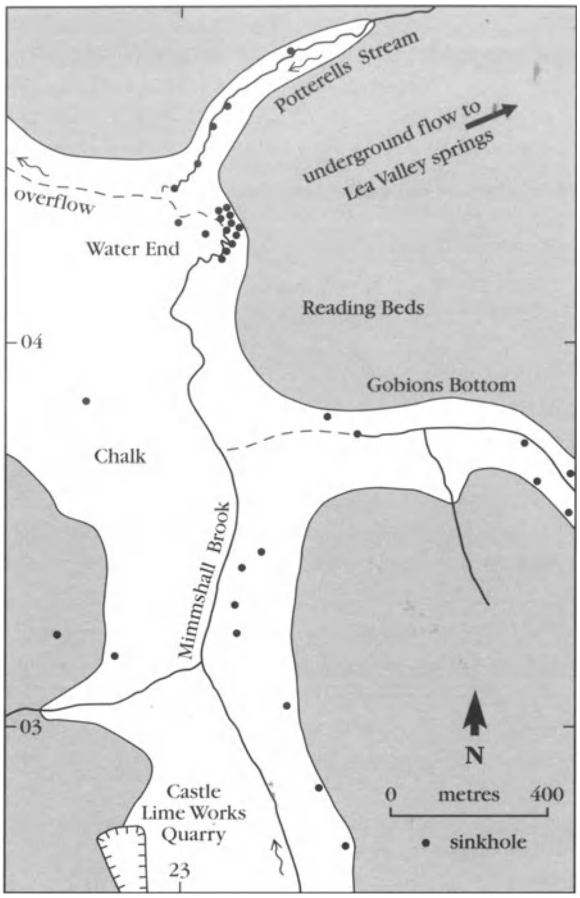


Figure 7.11 Geological map of the Mimmshall Valley with the Water End sinkholes and the Castle Lime Works Quarry.

sinks and depressions, but the number may change after a major flood (Figure 7.11). Under normal weather conditions most of the water sinks at a single point, but under progressively wetter conditions the main sink is overwhelmed and a series of other sinks in the valley are utilized. In flood, the combined discharge of the streams may exceed the capacity of the swallets; a lake then forms and expands to cover more than 2 ha before overflowing down the normally dry valley to the west, to join the River Colne (Figure 7.11). Average stream discharge is about 80 l s^{-1}, but the swallets have a capacity of about 1 m^3 s^{-1}, and this is exceeded in flood (Walsh and Ockenden, 1982). Fluorescein dye tracing proved that the water resurged at four springs spread out down 12 km of the Lea Valley (Morris and Fowler, 1937); these lie between 8 and 15 km from Water End, at elevations 20–45 m lower, and the flow-through rates averaged about 5500 m per day.

A second smaller stream, the Potterells Stream, drains the high ground to the north-east and also sinks in the northern corner of the Water End depression. Numerous sinkholes have been recorded along its lower course, which may be described as a degraded uvala, although the water normally flows across it (Wooldridge and Kirkaldy, 1937). A third very small stream sinks in Gobions Bottom, at the edge of the depression, and there are further sinkholes and dolines upstream along the floor of the Mimmshall Valley.

At the main sinks, the chalk lies just below the surface, buried under a few metres of sediment into which the streams are entrenched. Two main types of sinkhole are developed in the alluvium. Some are funnel-shaped holes with steep conical sides, in which the water drains directly into fissures within the chalk; they are often partially or totally choked with vegetation and domestic refuse. The others are shallow soakaway basins, in which the water gradually seeps through the sediment flooring the depression. Both types are essentially subsidence sinkholes, where the main surface feature is developed within the sediment overlying the chalk, but some collapse of the chalk does occur. New sinkholes appear at irregular intervals, and surface detail may be noticeably modified after a flood inundation. In some of the sinkholes, cave passages have been followed within the chalk for lengths of 10–20 m; they consist of narrow rifts with phreatic solutional enlargements and the water cascades along and down narrow floor fissures. Sediment chokes and flood risk have precluded further exploration.

Auger drilling across the sinkhole basin has revealed a layered sequence in the sediments (Walsh and Ockenden, 1982), which reach about 5 m in total thickness:

3. Dark-grey or black organic silts and clays, overlying light-brown silty clay, occasionally gravelly in part.
2. Greenish brown clay, overlying brown sands.
1. Brown silty clay with occasional chalk fragments, resting on chalk.

Interpretation

The initiation of the swallow hole complex appears to have been in the Pleistocene. The Radlett Mimms depression, in which the Mimmshall Brook is located, may have originated as part of a meander loop from a former course of the River Thames (Wooldridge and Kirkaldy, 1937). Subsequently, the underfit Mimmshall Brook was diverted by a plug of glacial till, which blocked the natural northerly continuation of the valley towards Hatfield, causing it to flow northwest to the River Colne. Since the last glaciation, a thin spread of gravels has been deposited over the valley floor.

The swallow holes appear to have formed after the gravels were deposited, as they cannot be shown to be older than the gravels (Walsh and Ockenden, 1982). Thus the sinkholes appear to be relatively recent features, formed over chalk fissures which were enlarged by solution over a very much longer period; new collapses occur at frequent intervals (Evans, 1944; Reeve, 1979; Walsh and Ockenden, 1982), usually after the basin has flooded.

The upper sediment layer (3) is interpreted as flood lake deposits, laid down during the last few decades or centuries (Walsh and Ockenden, 1982); the change from silty clay to organic clay is ascribed to geographical or vegetational changes in the drainage basin. The middle layer of sediment appears to be identical to the Reading Beds which crop out on the slopes to the east. The lowest clastic horizon (1) is a thin insoluble weathering residue of the chalk. Pollen analysis of the sediments (Walsh and Ockenden, 1982) indicates that they are postglacial. The Reading Beds (layer 2) lie under the basin at levels below the chalk on all sides, which suggests that they have been lowered by solution of the chalk on which they rest.

The nature of the cavity system under the swallow holes, and the pattern of underground drainage, remain conjectural. Dispersion of the dye from the single sink to the four widely spaced springs suggests that the water sinking in the swallow holes does not flow through a discrete cave system; the flow is probably mostly through a network of micro-fissures and a maze of larger fissures. The rapid flow rate suggests that there is a significantly large passage allowing vadose flow for at least part of the flow route, probably in the initial stages. The North Mimms Well, 500 m from Water End, has a very stable water table about 12 m below the sinkhole level, and recorded none of the dye injected at the sinks. This suggests that the chalk is honeycombed by a fissure system which allows free drainage from the sinks down to the regional water table, as seen in the short vadose sections of accessible cave. The large amount of clastic sediment which has been carried into the stream sinks during formation of the surface depressions is a further indication of the considerable size of the solutional cavities within the chalk.

Quantitative dye-tracing from the sinks under different stage conditions would provide further data on the nature of the underground drainage network beneath the swallow holes, and this would have implications for waste disposal and water management in Chalk areas.

Conclusions

The Water End swallow holes are the best and largest examples in Britain of permanent stream sinks developed in the chalk karst, and they clearly demonstrate the role of conduit flow within the Chalk aquifer.

CASTLE LIME WORKS QUARRY

Highlights

The preserved walls of this quarry provide the finest exposures in Britain through clay-filled pipes and fissures within the Chalk. The infills to these solution features are Tertiary sediments much older than the clay-with-flints commonly found on the Chalk outcrops.

Introduction

The old chalk pit of the Castle Lime Works lies in the western slope of the Mimmshall Valley (Figures 7.1 and 7.11). It is partially backfilled, but a part of the final working faces has been retained to provide a clean section through the sediment cover and into the bedrock chalk. The rockhead is pitted with solution pipes and cavities, which are infilled with clays and sands of Tertiary age. These pipes are seen to penetrate several metres down into the chalk, and some continue beneath the floor of the exposed section; though they vary greatly in profile, they are collectively known as pipes. The Tertiary rocks crop out on higher ground immediately west of the quarry, and consist of thin Thanet Sands overlain by the sands and clays of the Palaeogene Reading Beds.

The geology and hydrogeology of the Mimms Valley was discussed by Wooldridge and Kirkaldy (1937), but a more thorough examination of the site was carried out by Kirkaldy (1950). The petrography and origin of the deposits infilling the pipes were studied and discussed by Thorez *et al.* (1971).

Description

The solution pipes and cavities occur on the eastern face of the quarry, which stands nearly 5 m

Figure 7.12 A section of the preserved face in the Castle Lime Works Quarry exposing the extremely irregular upper surface of the chalk, broken by clay-filled pipes and broader depressions. (Photo: A.C. Waltham.)

high and is a preserved, permanent exposure (Figure 7.12); other faces of the quarry have been backfilled or are still active. The poorly bedded, horizontal Upper Chalk is exposed; it is irregularly fractured and has ill-defined bands of nodular flint. It is overlain by less than a metre of Quaternary reddish clay containing flints, and this supports a thin layer of topsoil.

A complex rockhead relief and infilled solution features have long been recognized during the progressive advance of the quarry faces. Though many individual features have since been lost to the quarrying, three main types of filled pipes can be recognized, all lying beneath 1–2 m overburden of stony clay soil (Thorez *et al.*, 1971):

1. Shallow basins up to 3 m deep and 30 m wide. These contain disturbed units of the flint-rich glauconitic Bullhead Bed, the basal member of the Palaeogene deposits, overlain by grey and brown pebbly sands with clay partings.
2. Steeply inclined or vertical cylindrical pipes, between 0.5 and 5.0 m wide; many were seen extending to depths greater than 12 m in parts of the quarry now destroyed. They are mostly infilled with grey and brown pebbly sand and are lined with dark-brown clay.
3. Horizontal seams of dark-brown clay and sand, up to 0.5 m thick, veining the chalk and occurring to 20 m below ground level; these include the 'sheet-pipes' described by Kirkaldy (1950).

The chalk surface along the preserved face is pitted with a series of these filled cavities; they vary considerably in shape, from shallow, rounded depressions, to steep-sided pipes up to a metre deep, and all are infilled with flint-bearing clay soils. Three larger pipes up to 3 m deep and wide are largely infilled with varieties of brown clays, reddish clays and sandy deposits derived from the overlying Tertiary rocks. Many of the pipes are lined with a dark-brown, very porous clay. The mineralogy and microstructures of these sediments were documented in detail by Thorez *et al.* (1971).

Interpretation

The pipes now exposed in the quarry formed as solutional voids or small caves in the chalk beneath the Tertiary cover rocks, causing the overlying sands and clays to slump into them. An early interpretation related the pipes to a group of fossil swallow holes, similar to the modern features at Water End (Kirkaldy, 1950). However, the quarry pipes are clearly subsurface solution features; structures within the Tertiary fills show that they have slumped and collapsed into them and were not the result of subaerial sedimentation in surface depressions.

The mineralogical composition of the infills supports this hypothesis, by showing that the pipe sediments are mainly a correct sequence of the Palaeogene cover rocks overlying the chalk (Thorez *et al.*, 1971); the clastic material does not have an inverted sequence which would evolve from erosion and redeposition. The lowest few metres of pebbly sand in the pipes relate to the Thanet Beds, which thin out beneath their cover 10 km south-east of the quarry. The higher sediments are typical of the Reading Beds which now outcrop on the hillside immediately west of the quarry. It appears that the chalk is initially dissolved beneath a cover of Thanet Sands and Reading Beds, which then slump into the underlying void. The relatively coherent Bullhead Bed at the base of the Palaeogene deposits was strong enough to support the upper sandier material until the cavities were enlarged sufficiently for collapse to occur. Disrupted Bullhead Bed material occurs up to 3 m below the original sub-Palaeogene surface.

The dark clay lining many of the pipes was deposited from suspension in the percolating groundwater, into the voids created by solution of the chalk. The clay material was derived from the overlying Palaeogene sediments, with only a small component derived from the insoluble chalk residue (Thorez *et al.*, 1971). The clays infilling the sheet pipes appear to have a similar origin. If the solution of the chalk and redeposition of the clay was concentrated near the water table, the sheet pipes may represent palaeo-water tables. It appears that solution of the chalk has been enhanced along the feather edge of the Tertiary cover rocks (Edmonds, 1983), where concentrated allogenic recharge enabled greater solutional activity to create the multiplicity of pipes and solutional basins.

The age of the solutional features is open to some debate, and may cover a considerable range. It appears that the larger pipes are a feature of early solution, possibly during the late Tertiary, and are only preserved beneath the Tertiary cover. Thanet and Bullhead material has been preserved within the pipes while it is missing from nearby undisturbed stratigraphic sequences; this indicates solutional lowering before deposition of the Reading Beds. However, Reading material is also slumped into the pipes beneath undisturbed Quaternary soils, indicating renewed solution and subsidence in the later Tertiary. Estimation of solution rates in pipes elsewhere on the Chalk suggests that they form in 5–10 ka (de Bruijn, 1983), but this does not indicate the absolute age of fossil features. The more widespread smaller pits, infilled with clay-with-flints, are probably of Quaternary age. Some of the material infilling the upper parts of the larger pipes is heterogeneous, suggesting that it was reworked by solifluction, and infilled surface depressions that overlay any collapse feature (Thorez *et al*., 1971). Chalk solution and sediment collapse associated with the pipes continues today, away from the quarry, at numerous sites where subsidence is observed.

Conclusions

The solutional features exposed along the eastern rockhead of the Castle Lime Works Quarry are some of the best preserved examples of clay-filled pipes in the chalk karst of Britain. They demonstrate a significant component of chalk karst morphology and provide evidence of the role of subsurface solution of the carbonate rock. The processes of chalk solution and subsequent subsidence of the cover rocks are analogous to those in the interstratal karst doline fields of South Wales (see Chapter 6), and also to those which preserved the Brassington Formation in the Peak District sinkholes (see Chapter 4).

DEVIL'S PUNCHBOWL

Introduction

The Devil's Punchbowl is a fine subsidence doline. It is the deepest and most spectacular of the many dolines developed on the covered chalk karst of the Norfolk Breckland, just north of Thetford.

Introduction

The sandy heathlands of the Breckland in Norfolk (Figure 7.1) are pock-marked by dozens of small dolines or depressions, often partially filled with water. These depressions and meres are developed on a thick cover of boulder clay overlying the chalk. The Devil's Punchbowl is the clearest example of these depressions. Its morphology is especially interesting as it is intermediate between the conical collapse dolines and the larger shallower subsidence basins both of which are common in the Breckland. The origins of the Devil's Punchbowl and the many other pits and depressions in Norfolk were discussed by Clarke (1903), Marr (1913), Jones and Lewis (1941), Prince (1962, 1964) and Sparks *et al.* (1972), and were reviewed by Day and Goudie (1978) and Goudie and Gardner (1985).

Description

The dolines of East Anglia are particularly numerous north and east of Thetford, in the area of heaths known as the Breckland. The chalk is overlain by up to 30 m of Anglian (and possibly 'Wolstonian') till which is mainly of sandy composition; nowhere is rock exposed at the surface. Scattered across the plateau are many dolines, with their surface forms entirely developed within the glacial drift. Most of these are steep conical depressions, up to 20 m across, but there are also about ten larger, shallower, saucer-shaped depressions up to 150 m across and covering up to 12 ha. These normally contain small lakes or meres.

The Devil's Punchbowl is one of the finest of the Breckland dolines (Figure 7.13). It has the profile of a shallow inverted cone, 6 m deep with sides sloping at up to 18°; almost perfectly circular, the doline has a surface diameter of about 150 m and its lake covers an area of about 0.6 ha. The lake level fluctuates by 2–3 m with changes in the groundwater level, in the aquifer which is contiguous between the chalk and the permeable till; the fluctuations are not simply seasonal, as the Punchbowl may be dry for many years at a time and then contain water for several years. The position of the doline is independent of the local valleys and surface streams.

Interpretation

Jones and Lewis (1941) described the circular Breckland meres as swallow holes due to the solution of the chalk and, in some cases, the collapse of the surface into underground cavities. Acidic drainage from the peaty heathlands has percolated

Figure 7.13 The Devil's Punchbowl doline with a lake on its floor in April 1982. (Photo: A.C. Waltham.)

through the glacial drift and dissolved the chalk below, causing subsequent settlement of the overlying glacial drift and subsidence on the surface. This mechanism characterizes the many forms of subsidence doline formed by ravelling and surface lowering in poorly consolidated clastic cover materials overlying cavernous carbonates. Analogies may be drawn with dolines in Dorset (Sperling *et al.*, 1977) and the widespread shakeholes in the till of the Yorkshire Dales karst.

The form of the Devil's Punchbowl appears to be intermediate between the small, steep conical collapses developed by ravelling into a single chalk fissure, and the larger shallow depressions formed by subsidence over a broader zone of rockhead solution; the Punchbowl may therefore provide the genetic link between the two styles of Breckland doline. The main group of Breckland meres, including the Devil's Punchbowl, all lie on a local dome, about 8 m high, in the water table (Day and Goudie, 1978), demonstrating the importance of downward seepage through their floors. This infiltration, and downwashing, is very slow, as the mere levels fluctuate with a time lag of some months behind the rainfall patterns. The age of the Breckland dolines is unknown, except that they postdate the glacial drift in which they are formed.

Alternative modes of origin for the many depressions in the Norfolk landscape include mineral workings, marl pits, and thaw sinks, besides the karstic landforms (Marr, 1913; Prince, 1962, 1964). No single explanation could account for the large numbers and uneven distribution, but the artificially excavated pits are recognizable by their small and irregular forms. The hypothesis of the thaw sinks has been applied to the Devil's Punchbowl and other Breckland meres (Prince, 1964; Sparks *et al.*, 1972). A pingo is formed by ice expansion within the shallow soil layers, and a soil cover may then slump off its domed surface so that a subsequent thaw leaves a depression: this cover sliding may leave marginal ramparts, but the Breckland meres have none of these features, which are recognized at other Norfolk sites.

Conclusions

The Devil's Punchbowl is the most spectacular of the many dolines developed in the covered chalk karst of the Breckland. Its origins have been strongly debated, but it is a fine example of a subsidence doline caused by subsurface solution of the chalk and subsequent settlement of the glacial drift cover. Its morphology suggests a genetic link between the large shallow basins containing meres and the smaller, steeper conical dolines, both of which are common in the Breckland.

MILLINGTON PASTURES

Highlights

The dry valley system of Millington Pastures may qualify as the finest dry valley system on the Chalk of Britain. With its deeply incised, dendritic form and well developed head deposits, it is an excellent representative of this important and ubiquitous karst landform.

Introduction

The Yorkshire Wolds form the crest of a wide chalk escarpment with a gentle dip to the east (Figure 7.1). Many dry valley networks are cut into both the scarp face and the dip slope. Millington Pastures is the finest and deepest of these, incised in the steep scarp face. It consists of a superb dendritic valley network with eight separate branches, all converging into Millington Bottom, east of Pocklington village. The thick head deposits in the valley floor are well preserved. The dry valleys of the Yorkshire Wolds have been poorly documented in comparison with the chalk valleys of southern Britain. Their origins have been discussed by Cole (1879, 1887), Mortimer (1885), Lewin (1969) and De Boer (1974), and many of the arguments put forward to explain the dry chalk valleys of southern England apply equally well.

Description

Millington Pastures has a dendritic system of eight converging dry valleys entrenched by up to 100 m

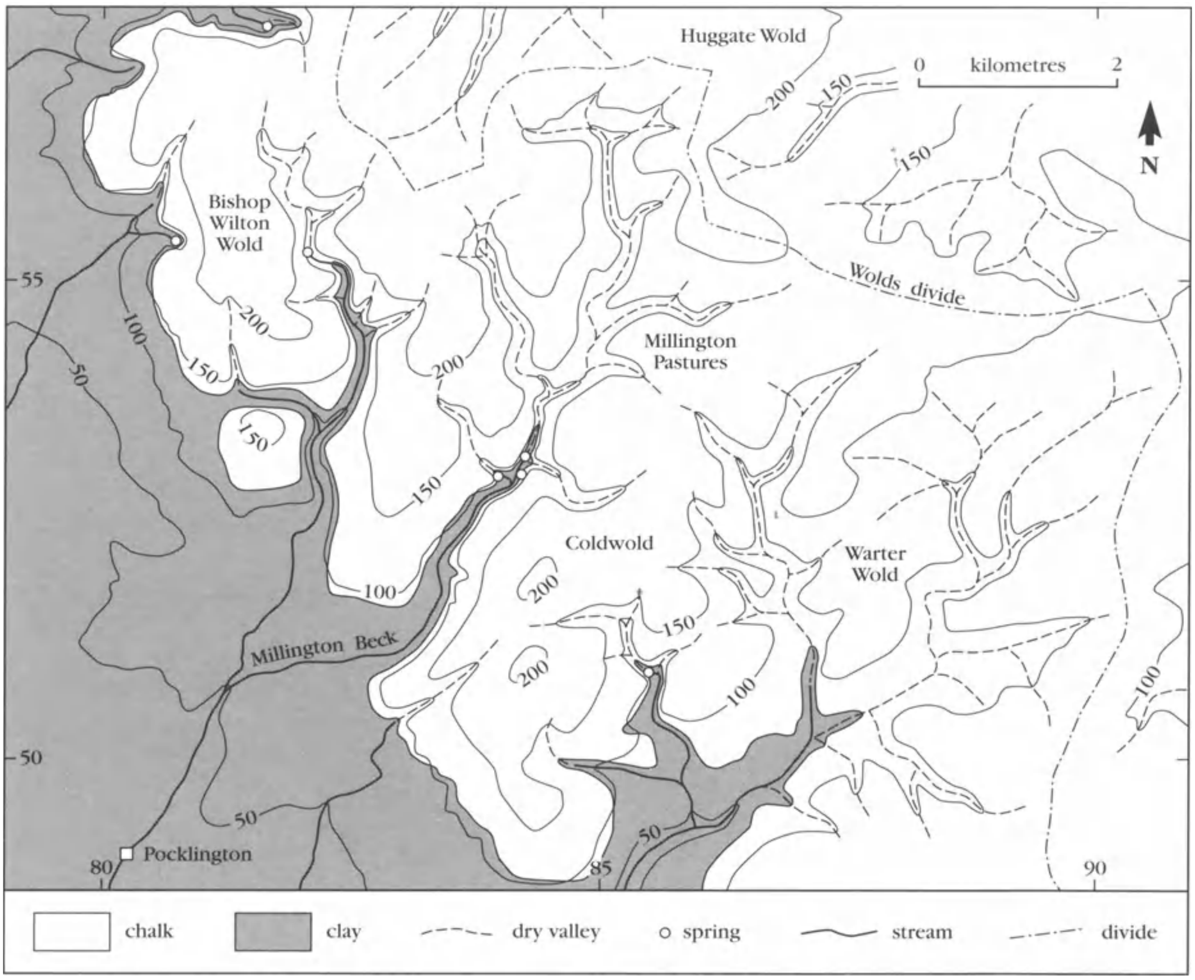


Figure 7.14 Outline map of the dry valley systems of Millington Pastures and its neighbours in the Yorkshire Wolds chalk escarpment. The chalk outcrop includes those of the impure, red Ferriby and Hunstanton Chalks, forming the lowest 25 m. Only the larger springs are marked.

Figure 7.15 One of the main dry valleys in the Millington Pastures system, with an almost flat floor of soliflucted head beneath slopes cut in weathered chalk. (Photo: A.C. Waltham.)

into the chalk wolds (Figure 7.14). They cut through the entire surviving Chalk sequence in the western flank of the Yorkshire Wolds escarpment; the dip is a few degrees to the east, but the chalk is cambered along the scarp edge and valley sides. The chalk locally lies unconformably on Liassic clays, which are exposed in the floor of Millington Bottom, where three powerful springs are fed by groundwater from the chalk to feed Millington Beck. Upstream of the springs, which are all now capped, the valleys are completely dry except after extreme rainfall events (Cole, 1887). Several sections through weathered chalk and the overlying combe rock are exposed in old quarries and pits, but there are no natural chalk outcrops.

The valley floors are gently graded, but steepen rapidly up into their headwalls. Many of the valley sides have very uniform gradients for most of their height, with a short convex section at the top. The valley cross-sections show a flattening where the thicker head deposits occupy the valley floors (Figure 7.15). The dry valley systems of both Millington Pastures and Warter Wold are among the largest of the scarp face systems. Their headward retreat has been so extensive that they have shifted the topographic divide of the Wolds escarpment towards the dip slope (Figure 7.14); the trendline of high points along the interfluves passes through Coldwold.

Interpretation

As with the many other chalk dry valleys in southern England, many different hypotheses have been proposed to explain the origin of the dry valleys of the Yorkshire Wolds. Mortimer (1885) suggested that they were formed by periods of crustal upheaval creating fractures in the crust, subsequently rounded by denudation. Cole preferred a more realistic explanation (1879, 1887) after noting the effects of a prolonged sharp frost followed by a rapid thaw. He witnessed surface flow occurring in many normally dry valleys causing extensive erosion and commented that 'nothing could more plainly show what rapid denudation of the chalk dry valleys might be carried out under glacial conditions'. However, he still maintained that chemical denudation was the primary cause for their origin. Lewin (1969) provided a comprehensive review of the dry valleys of the Yorkshire Wolds. He noted that there were at least three generations of valley which can be identified: old wide valleys which continued as wind gaps through the interfluves; younger dendritic valleys such as Millington Pastures; and the Devensian glacial meltwater channels. He reviewed the several hypotheses for their origins: by meltwater erosion of frozen ground, by subsurface solution and collapse, and by dissection resulting from scarp retreat and climatic change. He concluded that the dendritic valleys such as Millington Pastures formed by headward erosion by surface streams and were later abandoned due to climate change. These valley systems are clearly distinct from the glacial meltwater channels which cross parts of the Wolds without gathering tributaries and commonly with neither upstream nor downstream continuations.

By analogy with similar valleys in the Chalk of southern England such as the Manger (Paterson, 1977), and the Devil's Dyke (Small, 1962), it is clear that the dry valleys were excavated by a combination of solifluction and subaerial fluvial action under periglacial conditions. Ice sheets covered the Wolds in the earlier Pleistocene glaciations, but they were not covered by Devensian ice. During a long period of Devensian periglacial conditions, extensive and recurrent solifluction flows moved frost-shattered and saturated chalk debris into and down the growing valleys. Sediment removal and valley incision were further aided by snow meltwater flowing over the permafrost each summer.

The valleys subsequently became dry relict features when the climate ameliorated and all precipitation could percolate into groundwater systems. The chalk is a very permeable aquifer, with extensive diffuse flow. Beneath Millington Pastures the groundwater appears to converge on some form of conduit systems; though the springs are all close to the impermeable base of the chalk, they are in the valley sides where fissures dictate their sites, and not simply at the lowest exit points.

Conclusions

The chalk of Millington Pastures is scored by a singularly well developed system of dry valleys. It is a particularly spectacular example of a dendritic, dry valley network, more compact and deeply incised than any other in the Chalk of Britain. Slope morphologies and solifluction deposits are clearly displayed, and small old quarries expose useful sections through the weathered chalk and head deposits.

MOSTON LONG FLASH

Highlights

Moston Long Flash is a lake lying in one of a pair of well developed linear subsidence depressions above the Triassic salt beds of Cheshire. These are two of the clearest examples of this landform, which characterizes the Cheshire salt karst, and they are still deepening by active subsidence.

Introduction

The lake of Moston Long Flash lies on the Cheshire Plain 4 km south of Middlewich (Figure 1.2), at a site underlain by the Triassic salt beds. Natural subsidence features on the salt include linear depressions a few metres deep and several kilometres long, cutting across both hills and valleys, and also broader areas of ground lowering. Moston Long Flash is an example of an actively subsiding linear depression; initially its development was slow and natural, but accelerated greatly within the last 70 years as a result of brine abstraction. The salt beds of the Cheshire basin have been extracted since the Middle Ages, both by mining and by brine pumping.

The geology of the Cheshire basin salt deposits is outlined by Evans *et al.* (1968) and Earp and Taylor (1986). The subsidence features of the Cheshire basin were described by Calvert (1915), Wallwork (1956, 1960) and Waltham (1989), while the specific processes behind the subsidence at Moston Long Flash were examined by Oates (1981).

Description

Moston Long Flash is developed on over 20 m of permeable glacial till and glaciofluvial drift, of Devensian age. The underlying Triassic Mercia Mudstone sequence includes the Wilkesley Halite, which is a formation over 100 m thick consisting of alternating beds of mudstone and halite: individual beds of almost pure halite are 0.5–20 m thick, and the whole formation contains about 50% soluble salt.

The flashes of the Cheshire karst are lakes which form rapidly in depressions which subside below the water table due to subsurface salt solution. Moston Long Flash is a recently formed lake within an active linear subsidence (Figure 7.16). The whole subsidence landform extends for over 3 km, and is about 200 m wide and 3–10 m deep. Its gently curving cross-profile, remarkably uniform along its length, is asymmetric; a gently graded slope lies opposite a steeper bank which is often scored by a series of small slip scars as the

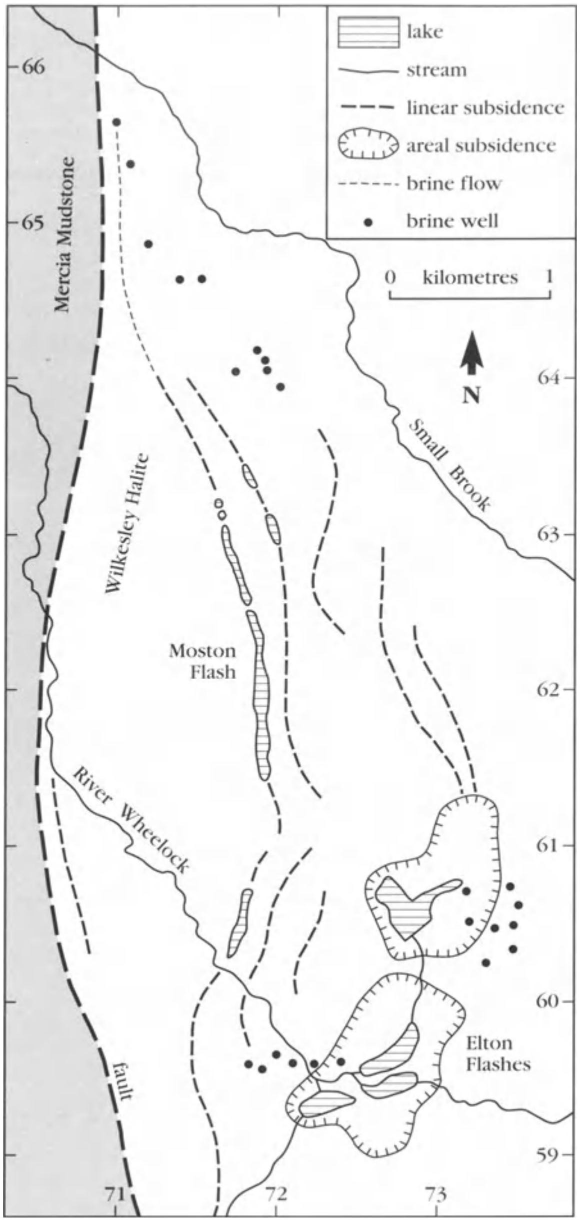


Figure 7.16 Outline map of Moston Flash and the adjacent linear and areal subsidences formed over the Wylkesley Halite. There is no solid outcrop as the entire area is covered by about 20 m of glacial till and glaciofluvial gravels. All the brine wells have now ceased pumping (after Oates, 1981, and Waltham, 1989).

depression subsides and enlarges in its direction. Some parts of the linear subsidence feature have the topographic appearance of a valley, but it is totally independent of the valleys of the area which have been formed by surface water erosion since the Devensian.

Subsidence has persisted over the last 70 years, often at rates in excess of 77 mm year^{-1} (Waltham, 1989); this was measured at a reference post on the edge of the depression, and subsidence rates were certainly higher in the centre of the flash. The lake first appeared in the 1920s, expanded first to the south and then extended to the north. Active subsidence continues to affect the adjacent farmland and farm buildings, and is clearly demonstrated by the repeated repairs to the road which crosses the flash (Figure 7.17). A second linear subsidence feature lies north-east of, and almost parallel to, the flash. It is smaller, less active and contains only a few small ponds where it crosses shallow valleys on the drift terrain. Subsidence rates in both features greatly reduced when brine abstraction stopped in 1978, but slow movement does continue today.

Interpretation

Where halite beds reach rockhead, the exposed salt is dissolved by groundwater flow at the base of the drift cover. The remaining insoluble mudstone beds collapse to create a permeable breccia zone, which may deepen to reach a thickness of over 50 m (Figure 7.18). Groundwater flows through the breccia, as a layer of saturated brine in contact with the halite, along the buried surface locally known as the wet rockhead. Solution is negligible where the impermeable halite is buried beneath impermeable mudstone, at the contact misleadingly known as the 'dry rockhead'. The brine is overlain by fresh water of lower density, and continued solution is dependent on an inflow of fresh water reaching the halite, usually after the brine has flowed out to natural brine springs or has been artificially pumped out (Calvert, 1915; Waltham, 1989).

The commonest type of subsidence feature is the linear trough of which Moston Long Flash is the prime example. These depressions are formed where solution of the underlying salt beds has been accentuated along zones of concentrated groundwater flow, locally known as brine streams, at the rockhead interface of the halite and breccia, usually 50-120 m below the surface. Slow natural subsidence does occur along these brine streams; but this is greatly accelerated where the saturated brine is artificially abstracted, so that unsaturated groundwater flows into contact with the halite. Wild brining is the process of pumping from boreholes sunk into the natural underground brine streams, and one of their effects has been that Cheshire's brine springs have all ceased to flow.

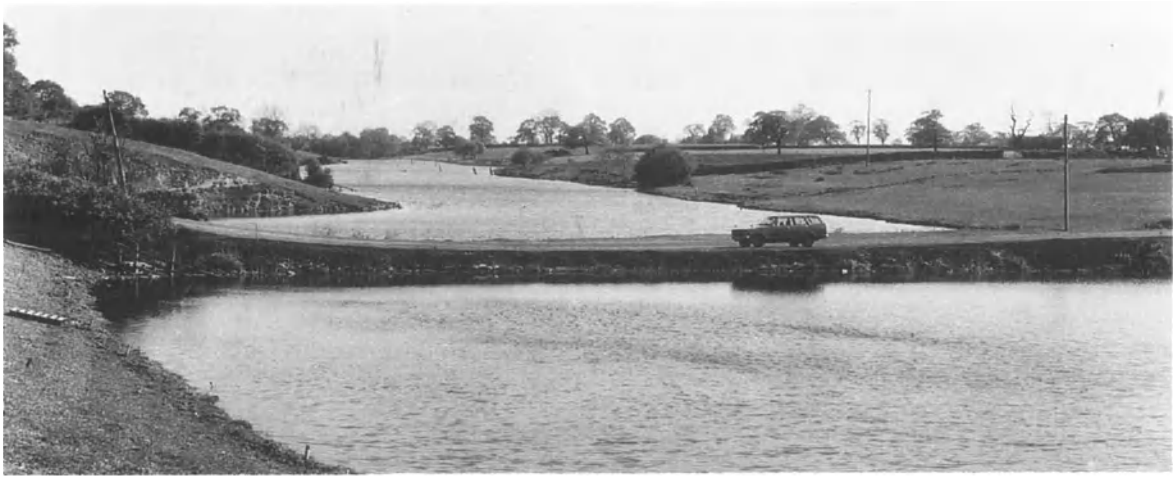


Figure 7.17 The linear subsidence of Moston Flash, looking south where the left face is steeper because it is retreating in the direction of dip as the depression enlarges. (Photo: A.C. Waltham.)

By correlation of the increasing volume of the subsiding depression with the volumes of pumped brine at nearby wells, Oates (1981) showed that the recent rapid subsidence of Moston Long Flash was due largely to brine pumping at a well 2 km to the north (Figure 7.16), and that the subsidence lagged about a year behind the pumping. The pumping draws fresh groundwater laterally through the surface drift and breccia, and solution and subsidence occur where it first meets the halite, far from the abstraction borehole. Saturated flow in a brine stream causes no subsidence. The linear subsidences are broadly aligned with the regional groundwater flow, but most appear to be located over the rockhead outcrops of individual beds of pure salt or along fracture zones. Moston Long Flash appears to follow the bedding until its brine stream turns parallel to a fault. The Elton Flashes are in areal subsidences, which are broad, less well defined depressions formed where fresh water is drawn into contact with the halite rockhead beneath surface streams and valleys (Figure 7.16).

The linear subsidence containing Moston Long Flash is almost certainly post-Devensian, formed after the salt rockhead was scoured by ice and then blanketed with drift. Subsidence has accelerated since the Middle Ages, and especially over the last 70 years, as a result of brine abstraction.

Conclusions

The active linear subsidences of Moston Long Flash and its smaller neighbour are excellent examples of the landforms developed by solution of underlying salt beds; they are characteristic of the Cheshire salt karst. Both features are clearly identifiable, and Moston is the largest active flash in the Cheshire Plain.

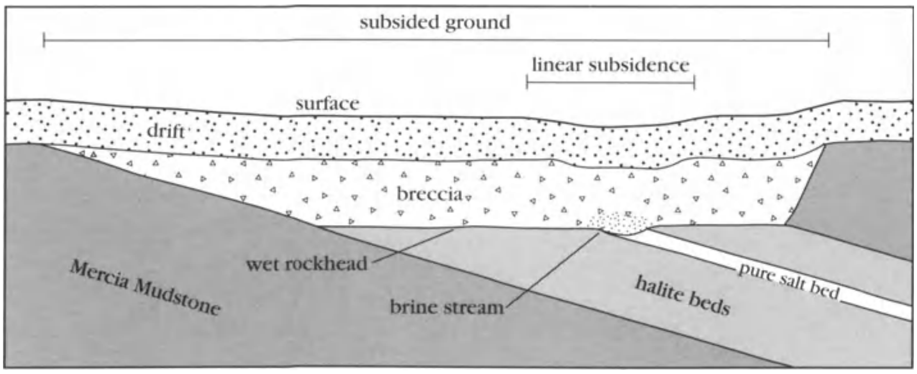


Figure 7.18 Diagrammatic section through the breccia of solutional residue at the rockhead in salt karst, with a brine stream flowing beneath an active linear subsidence like Moston Long Flash (from Waltham, 1989).

ROSTHERNE MERE

Highlights

Rostherne Mere lies in one of the finest and clearest examples of a subsidence basin formed over the Triassic salt beds of Cheshire. The meres of the Cheshire Plain are the most conspicuous feature of Britain's only area of salt karst.

Introduction

Rostherne Mere, north of Knutsford (Figure 1.2), is a lake within a roughly circular subsidence basin which was formed by solution of salt beds at depth. The buried salt is dissolved and removed where it is exposed to groundwater flow, causing regional and localized subsidence and collapse at the surface. Many depressions become lakes as they subside below the shallow water table of the plain. Lakes formed rapidly by accelerated subsidence in recent times are locally known as flashes - such as Moston Long Flash. The meres, including Rostherne, have a longer history of slow subsidence.

The subsidence features of the Cheshire salt karst have been described by Calvert (1915), Wallwork (1956), Reynolds (1979) and Waltham (1989), and the geology of the plain is outlined in Evans *et al.* (1968) and Earp and Taylor (1986). The bathymetry of Rostherne Mere was surveyed by Tattersall and Coward (1914) and redrawn by Pritchard (1961).

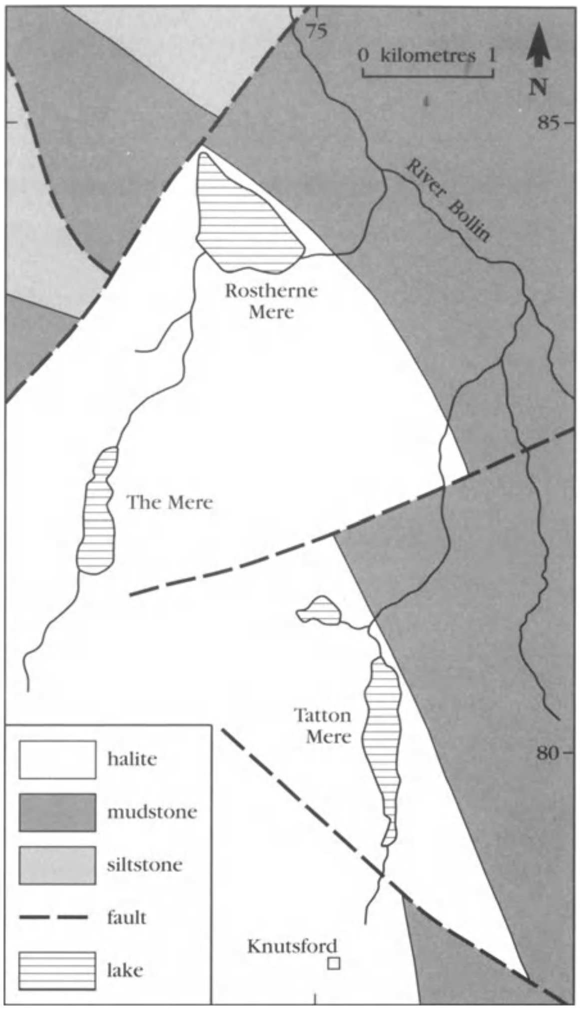


Figure 7.19 Solid geology map of the area around Rostherne Mere and the other adjacent subsidence depressions over the Northwich Halite, buried by a complete cover of glacial and glaciofluvial drift.

Description

Rostherne Mere is a lake with a surface area of 48.7 ha, and a maximum diameter of 1200 m (Figure 7.19). It lies in a shallow bowl which reaches about 35 m below the level of the surrounding terrain, and 27 m of this depth is now submerged. The sides of the depression slope gently down into the mere at about 5°, and there are no rock outcrops. The mere lies on a stream course which drains out to the east. The depression is an incidental part of the drainage system and is an isolated feature in the plain landscape of low relief, though other comparable meres occur nearby.

The mere depression is formed within the cover of Devensian glacial till and glaciofluvial sands. The drift overlies the Northwich Halite, a formation of 200 m of interbedded halite and mudstone within the Triassic Mercia Mudstone Group. On the halites the rockhead is covered by a breccia of mudstone left as a solution residue (Figure 7.18). The wet rockhead at the base of this breccia is generally over 100 m deep on the Northwich Halite (Oates, 1981). The depth to the breccia/drift interface at Rostherne Mere is unknown.

Interpretation

Both the halites and the mudstones are impermeable, and the only groundwater flow is in the permeable drift. Solution of the beds of halite therefore takes place where they meet the rockhead, causing collapse of the interbedded mudstones to create the residual breccia. The removal of the soluble salt by the groundwater flow, through both the drift and the breccia, caused the surface lowering and the formation of a subsidence basin, which subsequently flooded to form the mere. Continued solution of the rockhead halite relies on a supply of fresh water, but most of the rockhead on the halite is covered by a layer of dense, saturated brine, incapable of further solution. Both Rostherne and Tatton Meres overlie the Northwich Halite close to its buried edge, where groundwater flow off the adjacent rockhead of Mercia Mudstone first encounters the soluble salt (Figure 7.19).

The deep profile of the depressions containing Rostherne and some other meres is more than can be accounted for by differential solution and subsidence. The meres may be self-perpetuating in that, once formed in an incidental slight depression, they may gather surface water and act as supply points to the salt below. The same infiltration flow may cause some ravelling of the cover sediments into solution cavities beneath and around the deepening mere; this process was suggested as a mechanism behind some crater subsidences (Evans *et al.*, 1968).

A large part of the surface depression containing Rostherne Mere was clearly formed by subsidence which postdates the Devensian drift. Pritchard (1961) suggested that the subsidence may have been localized over an ice-excavated hollow; the known morphology and bathymetry provide no positive evidence for this, and the profiles of both the surface and the rockhead will have been substantially modified by postglacial solution.

Conclusions

Rostherne Mere occupies one of the prime examples of a subsidence basin developed over the Triassic Halites of the Cheshire Plain. Solution of the underlying salt beds, by natural groundwater circulation, has caused the surface subsidence. Its clear morphology makes it representative of these diagnostic features of the Cheshire salt karst, and it provides an excellent contrast with the linear subsidences such as Moston Long Flash.

Chapter 8

Karst in Scotland

INTRODUCTION

Karst is a very minor component of Scotland's landscape. Only at Assynt has underground drainage left significant lengths of dry valley floor and limestone solutional features on a scale where they create distinctive landforms. Elsewhere, the very small outcrops of various limestones are almost lost in landscapes dominated by erosional and depositional products of Pleistocene glaciations on a grand scale.

The Assynt karst

A narrow belt of carbonate outcrops extend north–south through the hills immediately east of Inchnadamph, at the head of Loch Assynt; the outcrops are discontinuous due to complexities of the geological structure, and are nowhere more than 3 km wide. The main karst features and caves are in the Traligill and Allt nan Uamh basins, both close to Inchnadamph (see Figure 8.2), but there are also caves in the Knockan basin 10 km to the south and on the Achmore plateau 3 km to the north (Lawson, 1988).

All the carbonates are in the Durness Group, a sequence of grey, bedded dolomites, 100 m thick, of Cambrian and Ordovician age (Johnson and Parsons, 1979). They are underlain by Cambrian quartzites. These rocks lie within a zone of major Caledonian thrusts, of which the Moine Thrust is the highest. Thrust planes have left klippen of Eocambrian sandstones forming the summit outliers of Beinn an Fhuarain and Beinn nan Cnaimhseag on top of the Durness carbonates. Curvature of the thrust planes creates a broadly basinal structure within the dolomites south-east of Inchnadamph, but bedding planes within the thrust sheets dip at various angles. Many of the caves are formed on the planes of the thrusts and faults, and groundwater flow is also constrained by a number of igneous intrusions.

Each Pleistocene glaciation covered the area with ice and largely removed earlier karstic landforms (Atkinson *et al.*, 1995). Glacial till, glaciofluvial debris and blanket peat bogs mask much of the carbonate outcrop, and periglacial weathering during the Loch Lomond Stadial created areas of frost-shattered debris and solifluction flows. A few rocky gorges expose the carbonates and there are a few high crags (Figure 8.1), but there are no extensive doline fields. There are few limestone pavements, because the complex geological structure has not allowed the glacial stripping of extensive stratimorphs.

The sinks, dry valleys and risings are the most conspicuous features of the karst. Allogenic

Figure 8.1 The limestone crags of Creag nan Uamh containing the Bone Caves, south of the Allt nan Uamh, seen from Beinn nan Cnaimhseag, with the Claonaite valley on the left. (Photo: T.J. Lawson.)

drainage is from the higher ground of Breabag and Conival in the east, and most of it sinks where it crosses onto the dolomite. Both the Traligill River and the stream of Allt nan Uamh flow entirely underground, except in very wet weather, though some streams maintain surface courses over the dolomite, mainly supported on mantles of glacial till. The associated caves carry water in postglacial stream trenches and flooded passages, and they also contain many passages which pre-date the last glaciation.

Outlying limestone areas

The Durness dolomites extend both north and south of Assynt, but with fewer karst features. Smoo Cave lies beneath a raised beach platform on the north coast; it has a large entrance chamber at the head of a tidal inlet, and an inner chamber with a stream cascading though its roof; the karstic passages and chambers were modified by wave action when sea levels were higher (Ford, 1959). Many small sinks and caves occur in the Durness dolomites just east of Loch Slapin, on the Isle of Skye; they include the stream cave of Uamh Cinn Ghlinn where the Allt nan Leac flows underground for 350 m (Ryder, 1974). The same carbonates have more small caves in the hills around Kishorn and some caves and limestone pavements in Glen Creran (Jeffreys, 1975, 1984).

Thin Jurassic limestones contain several small caves around Broadford, on Skye, and also the cave of Uamh nan Breagaire which has over 500 m of rift and bedding plane passages beneath a small limestone gorge at Applecross (Ryder, 1982). The mountain of Schiehallion, south-east of Loch Rannoch, is formed in Dalradian quartzites and phyllites with thin bands of marble; short caves have been formed where these are crossed by streams (Jeffreys, 1984). Uamh nan Uachdar is a cave with 60 m of passage in a Dalradian marble at an elevation of about 900 m in the Grampians south of Glen Spean (Young, 1992). There are no significant, recorded karst features in the thin limestone units in the Carboniferous of the Midland Valley.

Limestones are poorly represented in Scotland, and this short list of karst features is unlikely ever to become much longer. Pleistocene and Holocene sediments have been excavated from many caves and rock shelters, but these are mostly marine features preserved in fossil cliff lines behind raised beaches, and are not in limestone.

TRALIGILL VALLEY

Highlights

The Traligill Valley contains Scotland's finest karst scenery and is Britain's most recently deglaciated karst area. Its fine surface glaciokarst and its well documented underground drainage, developed in the Durness Limestones, provide a detailed record of landform development through the Pleistocene.

Introduction

The Traligill Valley karst lies just to the east of Inchnadamph, on the dolomites of the Durness Group, which crop out along a narrow and structurally complex belt. Drainage from the impermeable quartzites of Ben More Assynt in the east flows underground on reaching the dolomites, resurging several kilometres further down the valley. A spectacular variety of rock scars, pavements, blind valleys, dry valleys and dolines represents the progressive development of the karst drainage.

The geology of the Assynt karst and caves has been discussed by Ford (1959), Johnson and Parsons (1979) and Lawson (1988). The cave geomorphology is more fully described by Lawson (1983, 1986, 1988), and the later cave discoveries are recorded by Taviner (1993), Jeffreys (1994) and Mulholland (1994). The karst hydrology has been assessed through quantitative dye tracing (Newson and Atkinson, 1970; Smart *et al.*, 1986; Lawson, 1988), and stalagmite studies have yielded chronological and environmental data for the late Pleistocene (Lawson, 1982; Atkinson *et al.*, 1986; Baker *et al.*, 1993, 1995a).

Description

The Traligill River drains a section of wild moorland terrain dominated by the glaciated landforms which typify the Scottish Highlands. Karst features are not overly conspicuous, but a cave drainage system underlies the valley floor and distinguishes the morphological landforms of the site. About half of the 17 km^2 catchment is underlain by the Durness Carbonates (Figure 8.2). The upper basin extends onto quartzite outcrops and is an undulating plateau at elevations around 300 m, extensively mantled by peat and thick fluvioglacial sands. In contrast, the lower valley is narrower

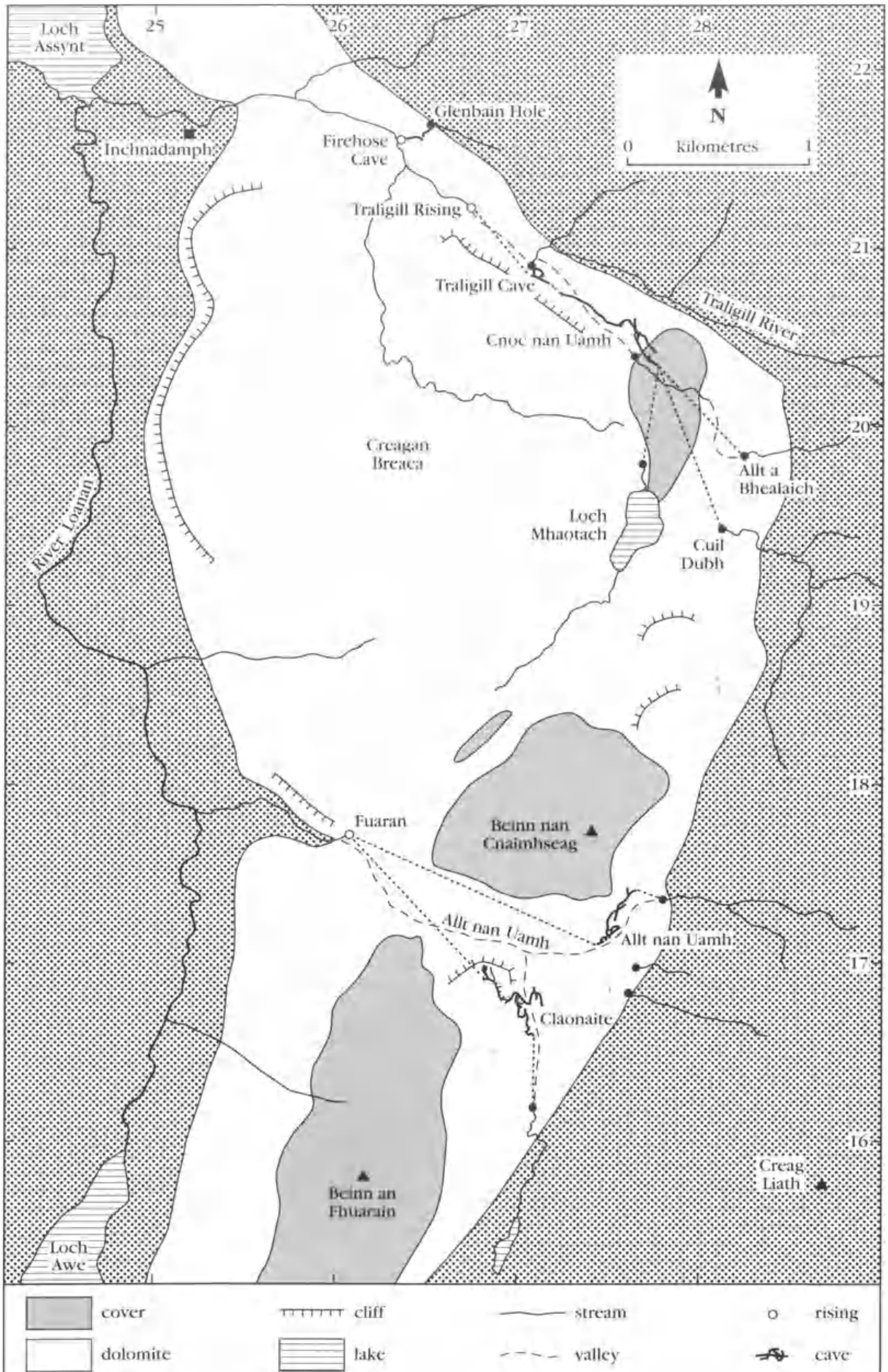


Figure 8.2 Geological map of the main karst belt in Assynt, containing the caves of the Traligill and Allt nan Uamh Valleys. The dolomites belong to the Durness Group and are underlain by the Lower Palaeozoic quartzites. The cover rocks are klippe of Cambrian quartzite and Eocambrian sandstone lying over major thrust planes.

and steeper and is dry over much of its length across the carbonate outcrop. The Traligill River flows into Loch Assynt at an elevation of 70 m.

The Cambrian–Ordovician rocks are composed of basal quartzites, overlain by the 'Pipe Rock' orthoquartzites, which are in turn overlain by shales and dolomites, and finally by the calcareous Durness Group; these are bedded dolomites about 100 m thick which are host to all the karst and caves. The simple stratigraphy is complicated by later tectonic events, and thrust planes and high-angle reverse faults occur throughout the Traligill Valley (Figure 8.3). These include the Traligill Main Thrust, on which lie some steeply inclined segments of cave passage and some asymetrical subaerial ravines along part of the carbonate margin on the north side of the lower valley. Higher thrust planes cap the dolomites, and an outlying klippe of Cambrian quartzite has the cave drainage passing beneath it.

Recurrent glaciation has sculpted all the larger features of the modern landscape, and ice striae, till, erratics and outwash are widespread. Karst features are most conspicuous in the valley floor where it is dry for 2 km, broken by closed depressions, some floored with gravel deposits. Surface streams from the orthoquartzites and peat deposits sink at the contact with the carbonates (Figure 8.2), and ten sinks are known to converge on the cave streamway in Cnoc nan Uamh (Smart *et al.*, 1986). Additional input is provided from the sinks around the Lower Traligill Cave, and all the water resurges at the Traligill Rising and other adjacent risings towards the lower end of the dolomite outcrop.

Of the many caves in the Traligill basin (Jeffreys, 1984; Lawson, 1988), the largest is the Cnoc nan Uamh system which extends for a large part of the drainage line from sink to rising. It has over 2200 m of mapped passages, covering a depth range of nearly 100 m (Figure 8.4). A complex high-level series of phreatic passages, oxbows and collapse chambers, with thick gravel deposits overlain by stalagmite, has been breached at the Uamh an Tartair entrance, and upstream the water emerges from two active phreatic sections. Downstream the cave is again breached at the Uamh an Uisge entrances, immediately above the Waterslide, where the water forms a spectacular cascade down an inclined thrust plane (Figure 8.5). Short sumps at the foot of the downdip cascades lead to a gently graded streamway almost along the strike. This has sections of beautifully decorated high-level phreatic bedding planes, and ends where the water roars into a small foaming sump close to the upstream limit in the Lower Traligill Cave. The lower caves in the valley contain steep, vadose, tributary streamways, abandoned high levels and various small chambers, but none has been explored for more than a few hundred metres. The main flow resurges at the Traligill Rising, but some water escapes to the lower risings where it is joined by the cave stream from Glenbain Hole resurging from Firehose Cave.

Figure 8.3 The Traligill River entrenched on the steeply inclined Traligill Thrust, upstream of Traligill Cave. (Photo: T.J. Lawson.)

The broad carbonate bench of Creagan Breaca has a thick peat cover, with surface drainage to the west, against the dip of the thinly bedded dolomites; the karst has many dolines but no known caves.

Interpretation

The karst landscapes of the Traligill Valley demonstrate the effects of complex geological structures

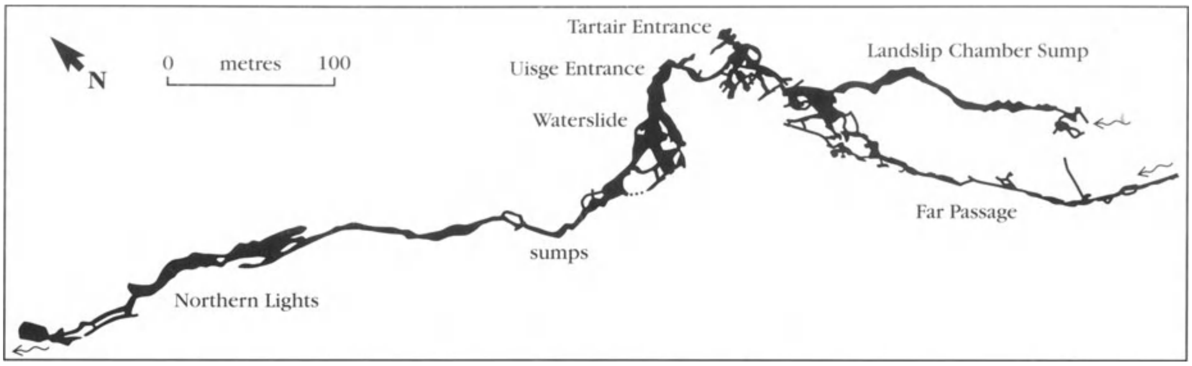


Figure 8.4 Outline map of the cave system of Cnoc nan Uamh (from survey by Grampian Speleological Group).

and multiple glaciations on the surface morphology. Several knickpoints along the valley record the surface lowering and subsequent rejuvenation of the valley after each glacial event. This has led to a progressive lowering of the local water table, causing a series of underground drainage captures. The structurally complex nature of the limestone is reflected in the rather irregular morphology of the caves, with joints, bedding planes and thrust planes all influencing cave development. The drainage below Cnoc nan Uamh follows the Traligill Main Thrust, and a series of successive downstream resurgences further down this thrust plane have captured the water (Taviner, 1993).

The modern hydrology of the caves is complex; large streams converge on a single trunk drain in Cnoc nan Uamh, and then flow into a distributary system feeding several resurgences (Smart *et al.*, 1986). Passage morphologies are varied, and abandoned high-level passages represent earlier stages of development. The multiple resurgences and many flood overflow channels may indicate the immature and constricted nature of many of the lower conduits.

Cave underflow has caused the partial desiccation of the lower Traligill valley, which is now only active as a flood route in wet weather. Higher up, the valley is almost completely dry and dolines punctuate the valley floor. This process of progressive underground capture had resulted in a complex range of landforms, including valleys with permanent surface flow and underflow, abandoned resurgences, dry channels and closed depressions.

By analogy with the caves of the Allt nan Uamh, just to the south (Figure 8.2), many of the high-level cave passages pre-date the Ipswichian interglacial, and were abandoned as the drainage occupied the modern streamways. Within these older passages, calcite stalagmites have yielded dates between 38 and 26 ka, or less than 9 ka (Atkinson *et al.*, 1986). The older dates show that conditions were warm enough to permit groundwater recharge and solutional activity between 38 and 26 ka, which suggests that there was a mid-Devensian deglaciation of the valley; subglacial water may excavate caves but normally lacks the exchangeable carbon dioxide which permits extensive calcite deposition. These caves clearly survived beneath their temporary glacial covers, but how much of the older systems were destroyed is not known. Their extensive clastic fills are interpreted as the product of flooding during some stages of the glacial events, with the main influx of coarse gravels and sands probably during deglaciation.

A Holocene stalagmite from the upper level of Cnoc nan Uamh, has luminescent banding which

Figure 8.5 The Waterslide in Cnoc nan Uamh. (Photo: A.C. Waltham.)

represents annual cycles of deposition (Baker *et al.*, 1993, 1995a); the same banding reveals a short acceleration of growth around 3150 years ago, which may record the Hekla 3 eruption in Iceland. The stalagmite and clastic sequences in the caves clearly provide evidence for each glacial modification of the valley during the Devensian cold stage, and further dating may indicate rates of base-level lowering.

Conclusions

The Traligill valley contains Scotland's finest glaciokarst landscape. Its cave drainage is well documented and provides an excellent example of a complex network dominated by structural rather than lithological controls; the thrust plane caves are particularly unusual. Some of the caves are immature, but sediments in the abandoned caves provide a unique record of underground drainage during a warmer phase of the middle Devensian. The dry valley morphology has evolved in response to sequential lowering of base levels and associated underground captures.

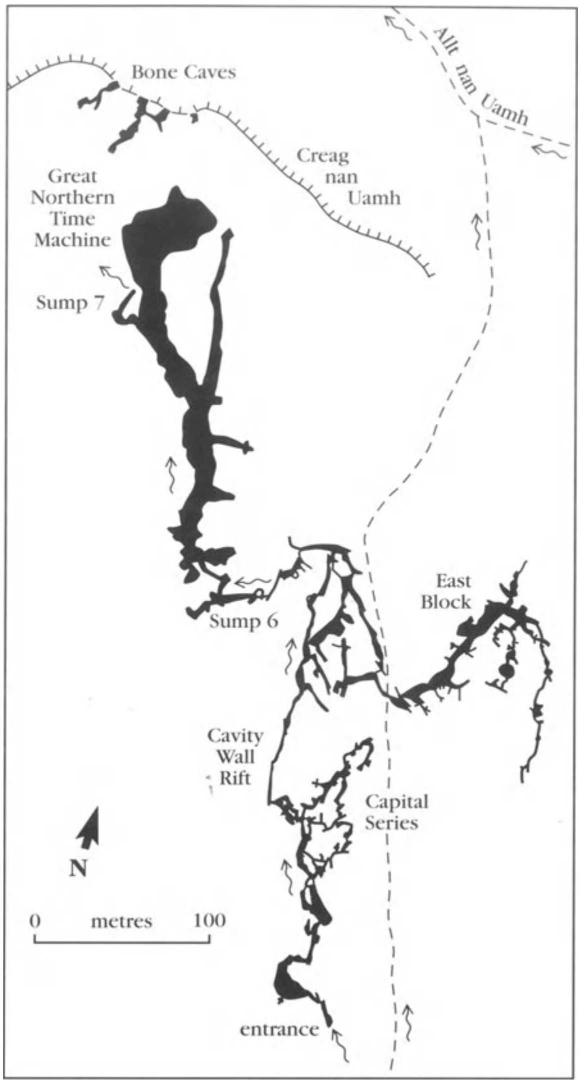


Figure 8.6 Outline map of Uamh an Claonaite; the survey beyond sump 6 is only a preliminary drawing (from surveys by Grampian Speleological Group).

ALLT NAN UAMH CAVES

Highlights

The site contains two of the most extensive cave systems in Scotland, developed in structurally complex Durness dolomites, and containing sequences of calcite and clastic sediments related to the Pleistocene glacial events.

Introduction

The caves are located in the upper basin of the Allt nan Uamh valley, south-east of Inchnadamph (Figure 8.2). Ten caves have been explored, all of which are developed in thrust sheets of Cambrian–Ordovician Durness Limestone. Allogenic streams drain from outcrops of the underlying Cambrian quartzites in the east, as well as from an overlying klippes of Torridonian sandstone both north and south of the valley. The two main cave systems transmit water from sinks in the upper part of the Allt nan Uamh valley towards a common resurgence at Fuaran Allt nan Uamh. The mapped cave passages cover only a small part of the 2000 m horizontally and 150 m vertically between the sinks and resurgence. They have been described by Ford (1959), Jeffreys (1984) and Lawson (1988).

Description

The Allt nan Uamh Stream Cave drains the northern sector of the basin. More than 1500 m of mapped passages are developed at two levels. The upper series of large, abandoned phreatic passages contains extensive clastic deposits and areas of collapse, but has few speleothems. Many of the passages are aligned on joints, and some of the steeply inclined rifts are developed on thrust planes. A lower series of narrow vadose rifts and low bedding plane passages carries the main water into a constricted sump on a thrust plane. Upstream, long flooded sections of passage alternate with spacious vadose canyons and cascades, above which lies a series of avens and abandoned high-level passages.

The southern drainage passes through the Uamh an Claonaite system, which has over 2600 m of explored passages (Figure 8.6). The modern cave stream flows along the western side of the system; much of its known length is in a vadose canyon, with several cascades, interrupted by flooded sections. Cavity Wall Rift is a fault-guided section where the stream descends gradually along two sections of solutionally enlarged thrust planes which separate two lithologically distinct limestones. The streamway turns south-west through a series of sumps, and then north-west below high-level chambers to a cascade into another sump. Off the main streamway, there are three series of abandoned passages, all at altitudes of 300-340 m. Capital Series consists of a series of largely abandoned, oxbow passages, partly filled with sediment from past periods of ponding. East Block is another high-level series of large phreatic passages, partially blocked by collapse and thick deposits of clastic sediment. The Great Northern Time Machine and other wide chambers above the far streamway are choked very close to the Bone Caves. The old passages contain suites of clastic sediments and calcite deposits which have yielded 22 dates (Atkinson *et al.*, 1995).

The remaining short caves include the Bone Caves which are truncated fragments of old passage in the cliffs of Creag nan Uamh, high above the modern valley floor; the sediments in these caves have provided an important source of Quaternary faunal remains (Lawson, 1993, 1995a). The sinks are choked, except for one shaft 7 m deep into a narrow fissure, and the resurgences are immature.

Interpretation

The Allt nan Uamh caves demonstrate close control by geological structures, including bedding planes, joints, faults and thrust planes within the tectonically complex dolomite beds.

The old high-level passages with their thick clastic sediment sequences preserve a record of the climate and landscape history of this area through much of the Pleistocene. The high-level relict passages in Allt nan Uamh Stream Cave and Uamh an Claonaite, and the truncated passages of the Bone Caves, are the remains of a series of conduits formed close beneath a palaeo-water table at about 340 m. They were abandoned when the main streamway of each cave formed at its present lower level, in response to glacial rejuvenation of the resurgence site. Flowstone from the abandoned passages in Uamh an Claonaite has been radiometrically dated to ages of 12-192 ka (Lawson, 1981, 1988; Atkinson *et al.*, 1995). These results suggest that the main phase of cave development was no later than the Hoxnian, since when the resurgence level has been lowered by about 130 m, probably by three stages of glacial erosion.

The glacial rejuvenation drained the phreatic passages and created the sites for Ipswichian deposition of the oldest stalagmites. The Devensian clastic sediments within the caves include subglacial silts, proglacial sands and gravels, and lag gravels and cobble beds left by meltwater scour during deglaciation (Atkinson *et al.*, 1995; Lawson, 1995b). There was also at least one phase of solutional activity and stalagmite deposition within the Devensian.

Conclusion

The Allt nan Uamh valley contains two of Scotland's most extensive and complex cave systems, as well as some of Britain's longest caves developed in non-Carboniferous limestone. These caves provide excellent examples of the nature of karst hydrology and geomorphology in structurally complex thrust sheets of dolomite, and their high-level passages contain a valuable sedimentary record of Pleistocene events.

References

Abbot, S. and Murgatroyd, R. (1988) Digging through to Maytime. *Descent*, **81**, 20-3.

Adams, A.E., Abdel Aziz, A.A. and Horbury, A.D. (1990) Controls on Dinantian sedimentation in South Cumbria and surrounding areas of Northwest England. *Proceedings of the Geologists' Association*, **101**, 19-30.

Adams, J. and Jones, K. (1984) Ogof Glan Gwenlais. *Isca Caving Club Journal*, **2**, 53-9.

Aitkenhead, N., Chisholm, J.I. and Stevenson, I.P. (1985) *The Geology of the Country around the Towns of Buxton, Leek and Bakewell*, British Geological Survey Memoir, Sheet 111.

Alexander, J.M. and Jones, J.C. (1959) The Survey of Pant Mawr Pot, South Wales. *Cave Research Group Publication* 9.

Allwright, P., Holmes, R., Rushforth, A. and Shepley, D. (1993) Ibbeth Peril: Upstream Downstream Passage. *Caves and Caving*, **62**, 6-8.

Andrews, J.E., Dennis, P.F. and Pedley, H.M. (1994) Stable isotope record of palaeoclimatic change in a British Holocene tufa. *The Holocene*, **4**, 349-55.

Andrews, P. (1990) *Owls, Caves and Fossils: Predation, Preservation and Accumulation of Small Mammal Bones in Caves, with an Analysis of the Pleistocene Cave Faunas from Westbury-sub-Mendip, Somerset, UK.* Natural History Museum, London.

Appleton, P. (1974) Subterranean courses of the River Alyn, including Ogof Hesp Alyn, North Wales. *Transactions of the British Cave Research Association*, **1**, 29-42.

Appleton, P. (1984) Ogof Hesp Alyn. *Caves and Caving*, **24**, 21-4.

Appleton, P. (1986) Ogof Llyn Du, North Wales - the rediscovery of the 'Llyn Du Cavern'. *Caves and Caving*, **33**, 4-6.

Appleton, P. (1987) The discovery of Ogof Llyn Parc, North Wales. *Caves and Caving*, **37**, 30-3.

Appleton, P. (1989) Limestones and caves of North Wales. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 217-54.

Arkell, W.J. (1947) *The Geology of Oxford*, Oxford University Press, Oxford.

Arthurton, R.S., Johnson, E.W. and Mundy, D.J. (1988) *Geology of the Country around Settle*, British Geological Survey Memoir, Sheet 60.

Ashmead, P. (1967) The origin and development of Ease Gill Caverns. *Transactions of the Cave Research Group of Great Britain*, **9**, 104-12.

Ashmead, P. (1969) The origin and development of caves in the Morecambe Bay area. *Transactions of the Cave Research Group of Great Britain*, **11**, 201-8.

Ashmead, P. (1974a) The caves and karst of the Morecambe Bay area. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 201-26.

Ashmead, P. (1974b) Development of the Caves of Casterton Fell. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 250-72.

Atkinson, F. (1950) The Cavern, Ireby Fell. *Cave Science*, **2** (9), 21-7.

Atkinson, F. (1963) Some notes on the formation of caverns in the Craven area of North-west Yorkshire. *Proceedings of the British Speleological Association*, **1**, 67-75.

Atkinson, T.C. (1967) The geomorphology of Longwood Swallet, Charterhouse-on-Mendip. *Proceedings of the University of Bristol Speleological Society*, **11**, 161-85.

References

Atkinson, T.C. (1968a) The earliest stages of underground drainage in limestones - a speculative discussion. *Proceedings of the British Speleological Association*, **6**, 53-70.

Atkinson, T.C. (1968b) Tracing swallet waters using *Lycopodium* spores. *Transactions of the Cave Research Group of Great Britain*, **10**, 99-106.

Atkinson, T.C. (1977) Diffuse flow and conduit flow in limestone terrain in the Mendip Hills, Somerset (Great Britain). *Journal of Hydrology*, **35**, 93-110.

Atkinson, T.C., Drew, D.P. and High, C. (1967) *Mendip Karst Hydrology Research Project, Phases One and Two*. Occasional publication of the Wessex Cave Club, **2** (1).

Atkinson, T.C., Harmon, R.S., Smart, P.L. and Waltham, A.C. (1978) Palaeoclimatic and geomorphic implications of $^{230}Th/^{234}U$ dates on speleothems from Britain. *Nature*, **272**, 24-8.

Atkinson, T.C., Lavis, J.J., Smith, D.I. and Whitaker, R.J. (1973) Experiments in tracing underground waters in limestones. *Journal of Hydrology*, **19**, 323-49.

Atkinson, T.C., Lawson, T.J. and Hebdon, N.J. (1995) Karst geomorphology. In *The Quaternary of Assynt and Coigach: a guide* (ed. T.J. Lawson), Quaternary Research Association, Cambridge, 61-86.

Atkinson, T.C., Lawson, T.J., Smart, P.L. *et al.* (1986). New data on speleothem deposition and palaeoclimate in Britain over the last forty thousand years. *Journal of Quaternary Science*, **1**, 67-72.

Atkinson, T.C. and Rowe, P.J. (1992) Applications of dating to denudation chronology and landscape evolution. In *Uranium Series Disequilibria* (2nd edn) (eds M. Ivanovich and R.S. Harmon), Clarendon Press, Oxford, 669-703.

Atkinson, T.C. and Smart, P.L. (1977) *Caves and Karst of Southern England and South Wales. Guidebook for the International Congress of Speleology*, British Cave Research Association.

Atkinson, T.C. and Smart, P.L. (1982) Fleet Street, Manor Farm Swallet, Charterhouse-on-Mendip, ST 498.556. *Proceedings of the University of Bristol Speleological Society*, **16**, 85-91.

Atkinson, T.C., Smart, P.L. and Andrews, J.N. (1984) Uranium series dating of speleothems from Mendip Caves. 1: Rhino Rift, Charterhouse-on-Mendip. *Proceedings of the University of Bristol Speleological Society*, **17**, 55-69.

Atkinson, T.C. and Smith, D.I. (1974) Rapid groundwater flow in fissures in the Chalk: an example from south Hampshire. *Quarterly Journal of Engineering Geology*, **7**, 197-205.

Bailey, C. (1992) Sweet and sour adit. *Descent*, **106**, 8-9.

Baker, A. (1993) 'Speleothem growth rates and paleoclimates.' PhD thesis, Bristol University.

Baker, A., Barnes, W.L., Edwards, R.L. *et al.* (1995a) The Hekla 3 volcanic eruption recorded in a Scottish speleothem? *Holocene*, **5**, 336-42.

Baker, A., Edwards, R.L., Richards, D.A. and Smart, P.L. (1993) Annual growth banding in a cave stalagmite. *Nature*, **364**, 518-20.

Baker, A., Smart, P.L. and Edwards, R.L. (1995b) Paleoclimate implications of mass spectrometric dating of a British flowstone. *Geology*, **23**, 309-12.

Baker, A., Smart, P.L. and Edwards, R.L. (1996) Mass spectrometric dating of flowstones from Stump Cross Caverns and Lancaster Hole, Yorkshire: paleoclimatic implications. *Journal of Quaternary Science*, **11**, 107-14.

Baker, A., Smart, P.L. and Ford, D.C. (1993) Northwest European palaeoclimate as indicated by growth frequency variations of secondary calcite deposits. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **100**, 291-301.

Baksi, A.K., Hsu, V., McWilliams, E. and Farrar, E. (1992) $^{40}Ar/^{39}Ar$ dating of the Brunhes-Matuyama geomagnetic field reversal. *Science*, **256**, 356-7.

Ball, T.K. and Jones, J.C. (1990) Speleogenesis in the Limestone Outcrop North of the South Wales Coalfield: the role of micro-organisms in the oxidation of sulphides and hydrocarbons. *Transactions of the British Cave Research Association*, **17**, 3-8.

Bamber, H.A. (1948) The Lathkill Caverns of Derbyshire. *Transactions of the British Cave Research Association*, **1**, 148-51.

Bamber, H.A. (1951) The hydrology of the Lathkill Dale area of Derbyshire. *Transactions of the British Cave Research Association*, **2**, 293-304.

Banks, D., Davies, C. and Davies, W. (1995) The chalk as a karstic aquifer: evidence from a tracer test at Stanford Dingley, Berkshire, UK. *Quarterly Journal of Engineering Geology*, **28**, S31-S38.

Barclay, W.J. (1989) *South Wales Coalfield, Part II, the Country Around Abergavenny*. British

Geological Survey Coalfield Memoir.

Barclay, W.J., Taylor, K and Thomas, L.P. (1988) *South Wales Coalfield, Part V, the Country Around Merthyr Tydfil*. British Geological Survey Coalfield Memoir.

Barrington, N. and Stanton, W.I. (1977) *Mendip: The Complete Caves and a View of the Hills*, Cheddar Valley Press, Cheddar.

Battiau-Queney, Y. (1980) 'Contribution a l'étude geomorphologique du Massif Gallois'. PhD thesis, l'Université de Bretagne Occidentale, France.

Battiau-Queney, Y. (1986) Buried paleokarstic features in South Wales: examples from Vaynor and Cwar Yr Ystrad quarries, near Merthyr Tydfil. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 551-67.

Batty, G. (1957) Hammer Pot. *Northern Pennine Club Journal*, **1**, 26-31.

Batty, G. (1967) Gingling Hole extension. *Northern Pennine Club Journal*, **3**, 1-5.

Baughen, D.J. and Walsh, P.T. (1980) Palaeokarst phenomena in the Carboniferous Limestone of Anglesey, North Wales. *Transactions of the British Cave Research Association*, **7**, 13-30.

Beck, H.M. (1984) *Gaping Gill, 150 years of exploration*, Hale, London.

Beck, J.S. (1975) The caves of the Foolow-Eyam-Stoney Middleton area, Derbyshire, and their genesis. *Transactions of the British Cave Research Association,* **2**, 1-11.

Beck, J.S. (1977) The caves of the Foolow-Eyam-Stoney Middleton area. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 361-82.

Beck, J.S. (1980) 'Aspects of the speleogenesis in the Carboniferous Limestone of North Derbyshire.' PhD thesis, University of Lancaster.

Beck, J.S. (1990) Streaks/Merlin - the saga continues. *Descent*, **97**, 10-11.

Beck, J.S. (1991) Peak Cavern: eight hours to Heaven and back. *Descent*, **103**, 8-9.

Beckinsale, R.P. (1954) Geomorphology. In *The Oxford Region* (eds A.F. Martin and R.W. Steel), Oxford University Press, Oxford.

Bennett, A.F., Bennett, M.R. and Doyle, P. (1995) Paving the way for conservation. *Geology Today*, **11**, 98-100.

Blackstock, T.H., Duigan, C.A., Stephens, D.P. and Yeo, M.J.M. (1993) Vegetation zonation and invertebrate fauna in Pant-y-llyn, an unusual seasonal lake in South Wales, UK. *Aquatic Conservation: Marine and Freshwater Ecosystems*, **3**, 253-68.

Bögli, A. (1960) Kalklosung und Karrenbildung. *Zeitschrift für Geomorphologie*, Supplementband, **2**, 4-21.

Bögli, A. (1964) Mischungskorrosion - ein Breitag zur Verkarstungsproblem. *Erdkunde*, **18**, 83-92.

Bögli, A. (1971) Corrosion by mixing of karst waters. *Transactions of the Cave Group of Great Britain*, **13**, 109-14.

Bögli, A. (1980) *Karst Hydrology and Physical Speleology*, Springer Verlag.

Bolt, P., Garman, A., Hicks, R. and Kendall, A. (1994) Ogof Draenen. *Descent*, **121**, 26-7.

Bolt, P., Garman, A., Lovett, B. and Munn, K. (1995) Twenty thousand metres under Pwll Du. *Descent,* **122**, 20-3.

Booth, T. S. (1905) Jockey Hole and Rift Pot. *Yorkshire Ramblers Club Journal*, **6**, 145-56.

Bottrell, S. and Gunn, J. (1991) Flow-switching in the Castleton karst aquifer. *Transactions of the British Cave Research Association*, **18**, 47-9.

Boulter, M.C. (1971) A palynological study of two of the Neogene plant beds in Derbyshire. *Bulletin of the British Museum of Natural History (Geology)*, **19**, 360-411.

Boulter, M.C., Ford, T.D., Ijtaba, M. and Walsh, P.T. (1971) Brassington Formation: a newly recognised Tertiary formation in the southern Pennines. *Nature*, **231**, 134-6.

Bowen, D.Q. (1965) 'Contributions to the geomorphology of central South Wales.' PhD thesis, University of Wales.

Bowen, D. Q. (1970) Southeast and central South Wales. In *The Glaciations of Wales and Adjoining Regions* (ed. C.A. Lewis), Longman, Harlow, 197-227.

Bowen, D.Q. and Henry, A. (1984) *Wales: Gower, Preseli and Fforest Fawr*, Quaternary Research Association Field Guide, Cambridge.

Bowen, D.Q., Rose, J., McCabe, A.M. and Sutherland, D.G. (1986) Correlation of Quaternary glaciations in England, Ireland, Scotland and Wales. *Quaternary Science Reviews*, **5**, 299-340.

Bowen, D.Q., Sykes, G.A., Reeves, A., Miller, G.H., Andrews, J.T., Brew, J.S. and Hare, P.E. (1989) Amino acid geochronology of raised beaches in southwest Britain. *Quaternary Science Reviews*, **4**, 279-318.

Boylan, P.J. (1981) A new revision of the Pleistocene mammalian faunas of Kirkdale

Cave, Yorkshire. *Proceedings of the Yorkshire Geological Society*, **43**, 253-80.

Bramwell, D. (1964) The excavation at Elderbush Cave, Wetton, Staffs. *North Staffordshire Journal of Field Studies*, **4**, 46-60.

Bramwell, D. (1977) Archaeology and palaeontology. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 263-91.

Bray, L.G. and O'Reilly, P.M. (1974) Preliminary oxidation studies on some waters from the Ogof Ffynnon Ddu system, Breconshire. *Transactions of the British Cave Research Association,* **1**, 75-84.

Bretz, J.H. (1942) Vadose and phreatic features of limestone caves. *Journal of Geology*, **50**, 675-811.

Brindle, D. (1949) Car Pot breakthrough. *Yorkshire Ramblers Club Journal*, **7**, 248-52.

Brindle, D. (1954) The Dowber Gill Passage in Dow Cave, Kettlewell. *Journal of Craven Pothole Club*, **1**, 263-9.

Brindle, D. (1955) The Dow Cave-Providence Pot system, Wharfedale. *Journal of Craven Pothole Club*, **2**, 4-10.

Brindle, N. (1956) New Goyden. *Journal of Craven Pothole Club*, **2**, 63-73.

Broadhurst, F.M. (1972) *Castleton: Itinerary 6*, Geologists' Association Guide, **26**, 27-30

Broadhurst, F.M. and Simpson, I.M. (1973) Bathymetry on a Carboniferous Reef. *Lethaia*, **6**, 367-81.

Brodrick, H. (1905) Notes on the geological features of Rift Pot. *Yorkshire Ramblers Club Journal*, **2** (6), 157-9.

Brook, A., Brook, D., Griffiths, J. and Long, M.H. (1991) *Northern Caves. Volume 2: Three Peaks,* Dalesman, Clapham.

Brook, D. (1969) The drainage and development of West Kingsdale. *University of Leeds Speleological Association Explorations Journal*, **1**, 70-3.

Brook, D. (1971a) A note on the Black Keld System. *Proceedings of the British Speleological Association,* **46**, 21-2.

Brook, D. (1971b) Caves and caving in Kingsdale. *Journal of the British Speleological Association,* **48**, 33-47.

Brook, D. (1974a) Cave development in Kingsdale. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 310-34.

Brook, D. (1974b) The caves of the Black Keld drainage system. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 422-33.

Brook, D. and Crabtree, H. (eds) (1969a) Kingsdale. *University of Leeds Speleological Association Explorations Journal*, **1**, 7-21.

Brook, D. and Crabtree, H. (eds) (1969b) Chapel-le-Dale. *University of Leeds Speleological Association Explorations Journal*, **1**, 23-40.

Brook, D., Davies, G.M., Long, M.H. and Ryder, P.F. (1988) *Northern Caves. Volume 1: Wharfedale and the North-East*, Dalesman, Clapham.

Brook, D., Long, M.H., Griffiths, J. and Ryder, P.F. (1994) *Northern Caves. Volume 3: Three Counties System and the North-West*, Dalesman, Clapham.

Brown, H.A., Otlet, R.L. and Sweeting, M.M. (1986) Stable isotopes - an investigation into their application in karst hydrology in the UK, with special reference to the Malham area, north Yorkshire. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 213-31.

Buckland, W. (1822) Account of an assemblage of fossil teeth and bones of elephant, hippopotamus, bear, tiger and hyaena, and sixteen other animals, discovered in a cave at Kirkdale, Yorkshire, in the year 1921. *Philosophical Transactions of the Royal Society*, **112**, 171-236.

Bull, A.J. (1936) Studies in the geomorphology of the South Downs. *Proceedings of the Geologists' Association*, **47**, 99-129.

Bull, A.J. (1940) Cold conditions and landforms in the South Downs. *Proceedings of the Geologists' Association*, **51**, 53-71.

Bull, P.A. (1975) Birdseye structures in caves. *Transactions of the British Cave Research Association*, **2**, 35-40.

Bull, P.A. (1976a) An electron microscope study of cave sediments from Agen Allwedd, Powys. *Transactions of the British Cave Research Association*, **3**, 7-14.

Bull, P.A. (1976b) Cave sediment studies in South Wales. *Studies in Speleology*, **3**, 13-24.

Bull, P.A. (1976c) Dendritic surge marks in caves. *Transactions of the British Cave Research Association*, **3**, 1-5.

Bull, P.A. (1977) Boulder chokes and doline relationships. *Proceedings of the 7th International Speleological Congress*, British Cave Research Association, Bridgwater, 93-6.

Bull, P.A. (1978) A study of stream gravels from a cave: Agen Allwedd, Powys, South Wales. *Zeitschrift für Geomorphologie*, **22**, 275-96.

Bull, P.A. (1980) Towards a reconstruction of timescales and palaeoenvironments from cave sediment studies. In *Timescales in Geomorphology* (eds R.A. Cullingford, D.A. Davidson and J. Lewin), Wiley, London, 177-87.

Bull, P.A. (1981) Some fine-grained sedimentation phenomena in caves. *Earth Surface Processes and Landforms*, **6**, 11-22.

Bull, P.A. and Carpenter, I.R. (1978) Sedimentological investigations of Goatchurch Cavern and Sidcot Swallet, Burrington Coombe, Somerset. *Proceedings of the University of Bristol Speleological Society*, **15**, 53-74.

Burek, C.V. (1977) The Pleistocene Ice Age and after. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 87-128.

Burek, C.V. (1991) Quaternary history and glacial deposits of the Peak District. In *Glacial Deposits in Great Britain and Ireland* (eds J. Ehlers, P.L. Gibbard and J. Rose), Balkema, Rotterdam, 193-201.

Burgess, I.C. and Cooper, A.H. (1993) *Geology of the Country around Harrogate*, British Geological Survey Memoir, Sheet 62.

Burke, A.R. (1967) Geomorphology and speleogenesis of vertical shafts in Carboniferous limestone at Ystradfellte, Breconshire. *Proceedings of the British Speleological Association*, **5**, 17-46.

Burke, A.R. (1970) Deposition of stalactitic and related forms of peat: genesis and bacterial oxidation. *Transactions of the Cave Research Group*, **12**, 247-58.

Burke, A.R. and Bird, P.F. (1966) A new mechanism for the formation of vertical shafts in Carboniferous limestone. *Nature*, **210**, 831-32.

Callaway, C. (1902) The zig-zag course of Cheddar Gorge. *Geological Magazine*, **9**, 67-9.

Calvert, A.F. (1915) *Salt in Cheshire,* Spon, London.

Campbell, J.B. (1977) *The Upper Palaeolithic of Britain: A Study of Man and Nature in the Late Ice Age,* Clarendon Press, Oxford (2 vols).

Campbell, J.B. and Sampson, G.G. (1971) A new analysis of Kent's Cavern, Devon, England. *University of Oregon Anthropological Papers*, **3**, 1-40.

Campbell, S. and Bowen, D.Q. (1989) *Quaternary of Wales,* Nature Conservancy Council, Peterborough.

Campbell, S., Gunn, J. and Hardwick, P. (1992) Pant-y-llyn - the first Welsh turlough? *Earth Science Conservation*, **31**, 3-7.

Carter, W.L. and Dwerryhouse A.R. (1904) The underground waters of Northwest Yorkshire; part 2, the underground waters of Ingleborough. *Proceedings of the Yorkshire Geological Society*, **15**, 248-92.

Chandler, R.H. (1909) On some dry chalk valley features. *Geological Magazine*, **6**, 538-9.

Chapman, P. (1979) The biology of Otter Hole, near Chepstow. *Transactions of the British Cave Research Association*, **6**, 159-67.

Charity, R.A.P. and Christopher, N.S.J. (1977) The stratigraphy and structure of the Ogof Ffynnon Ddu area. *Transactions of the British Cave Research Association*, **4**, 403-16.

Christopher, N.S.J. (1980) A preliminary flood pulse study of Russett Well, Derbyshire. *Transactions of the British Cave Research Association*, **7**, 1-12.

Christopher, N.S.J. (1984) Further water tracing experiments at Castleton, Derbyshire. *Transactions of the British Cave Research Association*, **11**, 86-8.

Christopher, N. and Beck, J.S. (1977) A survey of Carlswark Cavern, Stoney Middleton, Derbyshire, with geological and hydrological notes. *Transactions of the British Cave Research Association*, **4**, 361-6.

Christopher, N.S.J., Crabtree, R.W., Culshaw, S.M. *et al.* (1981) A hydrological study of the Castleton area, Derbyshire. *Transactions of the British Cave Research Association*, **8**, 189-206.

Christopher, N. and Wilcock, J. (1991) Temporal and seasonal controls on the composition of Derbyshire resurgences. *Transactions of the British Cave Research Association*, **18**, 51-4.

Clark, A. (1991) Three Deserts at Slaughter Stream Cave. *Descent*, **101**, 11.

Clark, R. (1967) A contribution to glacial studies of the Malham Tarn area. *Field Studies*, **2**, 479-91.

Clarke, W.G. (1903) The meres of Wretham Heath. *Transactions of the Norfolk and Norwich Naturalists Society*, 7.

Clayden, B.W. and Findlay, D.C. (1960) Mendip derived gravels and their relationship to combes (abstract). *Proceedings of the 3rd Conference on the Geology and Geomorphology of Southwest England*, Royal

Geological Society of Cornwall, 24.

Clayton, K.M. (1966) The origins of the landforms of the Malham area. *Field Studies*, **2**, 359-84.

Clayton, K.M. (1979) The Midlands and the Southern Pennines. In *The Geomorphology of the British Isles, Part II; Eastern and Central England* (eds A. Straw and K.M. Clayton), Methuen, London, 143-240.

Clayton, K.M. (1981) Explanatory description of the landforms of the Malham area. *Field Studies*, **5**, 389-423.

Clough, P. and Clough, S. (1981) Cliff Force Cave - the Spar Shop Series. *Journal of Craven Pothole Club*, **6**, 131-2.

Coase, A.C. (1967) Some preliminary observations on the geomorphology of the Dan yr Ogof system. *Proceedings of the British Speleological Association*, **5**, 53-67.

Coase, A.C. (1975) 'The geomorphology of the Dan-yr-Ogof caves.' PhD thesis, Leicester University.

Coase, A.C. (1989) Dan yr Ogof. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 190-207.

Coase, A.C. and Judson, D.M. (1977) Dan yr Ogof and its associated caves. *Transactions of the British Cave Research Association*, **4**, 245-344.

Coe, R.G. (1968) Birks Fell Cave extensions. *Journal of the Craven Pothole Club*, **4**, 111-20.

Cole, E.M. (1879) On the origin and formation of the Wold dales. *Proceedings of the Yorkshire Geological Society*, **7**, 128-40.

Cole, E.M. (1887) Note on dry valleys in the Chalk. *Proceedings of the Yorkshire Geological Society*, **9**, 343-6.

Coleman, A.M. and Balchin, W.G.V. (1959) The origin and development of surface depressions in the Mendip Hills. *Proceedings of the Geologists' Association*, **70**, 291-309.

Collcutt, S.N. (1985) Analysis of sediments in Gough's Cave, Cheddar, Somerset, and their bearing on the Palaeolithic archaeology. *Proceedings of the University of Bristol Speleological Society*, **17**, 129-40.

Cook, L.B. (1950) Stump Cross Caverns: past and present. *Journal of Craven Pothole Club*, **1**, 70-3.

Coope, G.R. (1977) Fossil coleopteran assemblages as sensitive indicators of climatic changes during the Devensian (last) cold stage. *Philosophical Transactions of the Royal Society of London*, **B280**, 313-40.

Cooper, A.H. (1986) Foundered strata and subsidence resulting from the dissolution of Permian gypsum in the Ripon and Bedale areas, North Yorkshire. In *The English Zechstein and Related Topics* (eds G.M. Harwood and D.B. Smith), Geological Society of London, Special Publication, **22**, 127-39.

Cooper, A.H. (1988) Subsidence resulting from the dissolution of Permian gypsum in the Ripon area: its relevance to mining and water abstraction. In *Engineering Geology of Underground Movements* (eds F.G. Bell, M.G. Culshaw, J.C. Cripps and M.A. Lovell). Geological Society Engineering Geology Special Publication, **5**, 387-90.

Cooper, R.G. and Halliwell, R.A. (1976) A relict karst feature in the Hambleton Hills, North Yorkshire. *Proceedings of the Yorkshire Geological Society*, **41**, 71-3.

Cooper, R.G., Ryder, P.F. and Solman, K.R. (1976) The North Yorkshire windypits: a review. *Transactions of the British Cave Research Association*, **3**, 77-94.

Cooper, R.G., Ryder, P.F. and Solman, K.R. (1982) The windypits in Duncombe Park, Helmslet, North Yorkshire. *Transactions of the British Cave Research Association*, **9**, 1-14.

Corbel, J. (1957) *Les karsts du Nord Ouest de l'Europe* Institute Etudes Rhodaniènes Université de Lyon, Memoirs et Documents, 12.

Cordingley, J. (1986) *The Peak Cavern system - A caver's guide,* Vitagraph Books, Manchester.

Cordingley, J. (1988) Peak Cavern: further exploration in Stemple Highway. *Descent*, **85**, 28-31.

Cordingley, J. (1989) Peak Cavern: progress in Far Sump extension. *Descent*, **86**, 28-9.

Cordingley, J. (1990) Probing the secrets behind Malham Cove. *Descent*, **93**, 32-3.

Coxon, C.E. (1986) 'A study of the hydrology and geomorphology of turloughs.' PhD thesis, Trinity College, Dublin.

Coxon, C.E. (1987a) The spatial distribution of turloughs. *Irish Geography*, **20**, 11-23.

Coxon, C.E. (1987b) An examination of the characteristics of turloughs using multivariate techniques. *Irish Geography*, **20**, 24-42.

Crabtree, R.W. (1979) Quantitative fluorometric dye tracing, Rickford and Langford resurgences, Northern Mendip. *Proceedings of the University of Bristol Speleological Society*, **15**, 129-41.

Crowther, J. (1989) Karst geomorphology of South

Wales. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 20-39.

Currant, A.P. (1987) Late Pleistocene Saiga Antelope *Saiga tatarica* on Mendip. *Proceedings of the University of Bristol Speleological Society*, **18**, 74-80.

Cvijić, J. (1893) Das Karstphanomen. *Geographische Abhandlungen herausgegeben von A. Penck*, **5** (3), 218-329.

Cvijić, J. (1918) Hydrographie souterraine et evolution morphologique du karst. *Recu Travaux d'Institute de Geographie Alpine*, **6** (40), 376-420.

Davies, B. (1984) Rift Pot. *Caves and Caving*, **25**, 2-5.

Davies, G.M. (1974a) The caves of Nidderdale. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 434-9.

Davies, G.M. (1974b) New Goyden Pot. *University of Leeds Speleological Association Review*, **13**, 22-4.

Davies, M. (1989) Recent advances in cave archaeology in southwest Wales. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 79-91.

Davies, R.H. and Stringer, R.N. (1991) Pant-y-Llyn: a Welsh turlough, or now you see it, now you don't. *Isca Caving Club Journal*, **13**, 36-9.

Davis, W.M. (1899) The geographical cycle. *Geographical Journal*, **14**, 481-504.

Davis, W.M. (1930) Origin of limestone caverns. *Geological Society of America Bulletin*, **41**, 475-628.

Dawkins, W.B. (1862) On a Hyena-den at Wookey Hole near Wells. *Quarterly Journal of the Geological Society*, **18**, 115-25.

Dawkins, W.B. (1875) The Mammalia found at Windy Knoll. *Quarterly Journal of the Geological Society*, **31**, 246-55.

Day, M.J. and Goudie, A.S. (1978) Why the Devil's Punchbowl dries up. *Geographical Magazine*, **50**, 381-5.

Daykyns, J.R., Gunn, W., Strahan, A. and Tiddeman, R.H. (1890) *The Geology of the Country around Ingleborough with Parts of Wensleydale and Wharfedale*. British Geological Survey Memoir.

De Boer, G. (1974) Physiographic evolution. In *The Geology and Mineral Resources of Yorkshire* (eds D.H. Rayner and J.E. Hemingway), Yorkshire Geological Society, Leeds.

De Bruijin, R.G.M. (1983) Some considerations on the factors that influence the formation of solution pipes in chalk rock. *Bulletin of the International Association of Engineering Geology*, **28**, 141-6.

Donovan, D.T. (1969) Geomorphology and hydrology of the central Mendips. *Proceedings of the University of Bristol Speleological Society*, **12**, 63-74.

Donovan, D.T. (1988) The Late Pleistocene sequence at Wells, Somerset. *Proceedings of the University of Bristol Speleological Society*, **18**, 241-57.

Doughty, P.S. (1968) Joint densities and their relation to lithology in the Great Scar Limestone. *Proceedings of the Yorkshire Geological Society*, **36**, 479-512.

Downing, R.A. (1994) Jurassic limestone environments. *British Cave Research Association Cave Studies Series*, **5**, 25-9.

Downing, R.A. and Williams, B.P.J. (1969) *The Groundwater Hydrology of the Lincolnshire Limestone*, Water Resources Board, Reading.

Drew, D.P. (1966) The water table concept in limestones. *Proceedings of the British Speleological Association*, **4**, 57-67.

Drew, D.P. (1968) A study of the limestone hydrology of the St Dunstan's Well and Ashwick drainage basins, eastern Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **11**, 257-72.

Drew, D.P. (1970) Limestone solution within the East Mendip area, Somerset. *Transactions of the Cave Research Group of Great Britain*, **12**, 259-70.

Drew, D.P. (1974) Quantity and rate of limestone solution on the eastern Mendip Hills. *Transactions of the British Cave Research Association*, **1**, 93-100.

Drew, D.P. (1975a) The limestone hydrology of the Mendip Hills. In *Limestones and Caves of the Mendip Hills* (ed. D.I. Smith), David and Charles, Newton Abbot, 171-213.

Drew, D.P. (1975b) The caves of Mendip. In *Limestones and Caves of the Mendip Hills* (ed. D.I. Smith), David and Charles, Newton Abbot, 214-312.

Drew, D.P. and Daly, D. (1993) *Groundwater and Karstification in Mid-Galway, South Mayo and North Clare*, Geological Survey of Ireland Report Series 93/3.

Drew, D.P., Newson, M.D. and Smith, D.L. (1968) *Mendip Karst Hydrology Research Project: Phase Three*, Wessex Cave Club Occasional Publication, **2** (2).

Dreybrodt, W. (1981a) Kinetics of the dissolution of calcite and its application to karstification. *Chemical Geology*, **31**, 245-69.

Dreybrodt, W. (1981b) Mixing corrosion in $CaCO_3$-CO_2-H_2O systems and its role in the karstification of limestone areas. *Chemical Geology*, **32**, 221-36.

Dreybrodt, W. (1988) The kinetics of calcite dissolution and its consequences to karst evolution from the initial to the mature state. *National Speleological Society Bulletin*, **49**, 31-49.

Duff, K.L., McKirdy, A.P. and Harley, M.J. (eds) (1985) *New Sites for Old: A Students' Guide to the Geology of the East Mendips,* Nature Conservancy Council, Peterborough.

Dunham, K.C., Hemingway, J.E., Versey, H.C. and Wilcockson, W.H. (1953) A guide to the geology of the district around Ingleborough. *Proceedings of the Yorkshire Geological Society*, **29**, 77-115.

Dunham, K.C. and Stubblefield, C.J. (1945) The stratigraphy, structure and mineralization of the Greenhow mining area, Yorkshire. *Quarterly Journal of the Geological Society*, **100**, 209-68.

Dunham, K.C. and Wilson, A.A. (1985) *Geology of the Northern Pennine Orefield, Volume 2, Stainmore to Craven*, British Geological Survey Memoir Economic Memoir, Sheets 40, 41 and 50.

Earp, J.R. and Taylor, B.J. (1986) *Geology of the Country around Chester and Winsford.* British Geological Survey Memoir, Sheet 109.

Ede, D. P. (1973) Aspects of karst hydrology in South Wales. *Geographical Journal*, **139**, 285-94.

Ede, D.P. and Bull, P.A. (1989) Swallets and caves of the Gower peninsula. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 211-16.

Edmonds, C.N. (1983) Towards the prediction of subsidence risk upon the chalk outcrop. *Quarterly Journal of Engineering Geology*, **16**, 261-6.

Edwards, A.J. (1994) 'Hydrogeological aspects of waste disposal in a Carboniferous Limestone aquifer.' PhD thesis, University of Bristol.

Elliott, J.V., Westlake, C.F. and Tringham, M.E. (1979) Otter Hole, near Chepstow, Wales. *Transactions of the British Cave Research Association*, **6**, 143-58.

Elliott, J.V., Westlake, C.F. and Tringham, M.E. (1989) Otter Hole, Chepstow. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 117-23.

Embleton, C. (1964) Sub-glacial drainage and supposed ice-dammed lakes in northeast Wales. *Proceedings of the Geologists' Association*, **75**, 31-8.

Evans, P. (1944) Field meeting at Water End, North Mimms, Hertfordshire. *Proceedings of the Geologists' Association*, **55**, 189.

Evans, W.B. (1970) The Triassic salt deposits of northwestern England. *Quarterly Journal of the Geological Society*, **126**, 103-23.

Evans, W.B., Wilson, A., Taylor, B.J. and Price, D. (1968) *Geology of the Country around Macclesfield, Congleton, Crewe and Middlewich*. British Geological Survey Memoir, Sheet 110.

Eyre, J. (1989) *The Ease Gill Cave System*, British Cave Research Association, Speleo-history Series, 1.

Eyre, J. and Ashmead, P. (1967) Lancaster Hole and the Ease Gill Caverns. *Transactions of the Cave Research Group of Great Britain*, **9**, 61-123.

Fagg, C.C. (1923) The recession of the Chalk escarpment and the development of Chalk valleys in the regional survey area. *Proceedings and Transactions of Croydon Natural History and Scientific Society*, **9**, 93-112.

Fagg, C.C. (1954) The coombes and embayments of the Chalk escarpment. *Proceedings and Transactions of Croydon Natural History and Scientific Society*, **12**, 117-31.

Farr, M.J. (1985) How Britain's most spectacular passage was discovered. *Descent*, **64**, 12-15.

Farr, M.J. (1986) Through Llangattwg Mountain. *Descent*, **72**, 19-26.

Farr, M.J. (1993) Busman's Holiday. *Descent*, **113**, 20-21.

Farrant, A.R. (1991) The Gough's Cave system: exploration since 1985, and a reappraisal of the geomorphology. *Proceedings of the University of Bristol Speleological Society*, **19**, 3-17.

Farrant, A.R. (1995) 'Long term Quaternary chronologies from cave deposits.' PhD thesis, Geography Department, University of Bristol.

Findlay, D.C. (1965) *The soils of the Mendip District of Somerset,* Agricultural Research Council, Memoir of the Soil Survey of Great Britain, England and Wales. Harpenden.

Fisher, O. (1858) On some natural pits on the heaths of Dorsetshire. *Geologist*, **1**, 527.

Fisher, O. (1859) On some natural pits on the heaths of Dorsetshire. *Quarterly Journal of* *the Geological Society of London*, **15**, 187-8.

Flindall, R., Swainn, J. and Hayes, A. (1981) A survey of the Masson Cave-cum-Mine complex, Matlock. *Bulletin of the Peak District Mines Historical Society*, **8**, 103-8.

Foley, I. (1930) Lost John's Cave. *Yorkshire Ramblers Club Journal*, **6**, 44-59.

Ford, D.C. (1963) 'Aspects of the geomorphology of the Mendip Hills.' PhD thesis, Oxford University.

Ford, D.C. (1964) On the geomorphic history of G.B. Cave, Charterhouse-on-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **10**, 149-88.

Ford, D.C. (1965a) Stream potholes as indicators of erosion phases in limestone caves. *Bulletin of the National Speleological Society of America*, **27**, 27-32.

Ford, D.C. (1965b) The origin of limestone caverns: a model from the central Mendip Hills, England. *Bulletin of the National Speleological Society of America*, **27**, 109-32.

Ford, D.C. (1965c) The sequence of development in Swildon's Hole. *Wessex Cave Club Journal*, **8**, 198-205.

Ford, D.C. (1966a) Calcium carbonate evolution in some central Mendip caves. *Proceedings of the University of Bristol Speleological Society*, **11**, 46-53.

Ford, D.C. (1968) Features of cavern development in central Mendip. *Transactions of the Cave Research Group of Great Britain*, **10**, 11-25.

Ford, D.C (1971) Geologic structure and a new explanation for limestone cave genesis. *Transactions of the Cave Research Group of Great Britain*, **13**, 81-94.

Ford, D.C. (1988) Characteristics of dissolutional cave systems in carbonate rocks. In *Paleokarst* (eds N.P. James and P.W. Choquette), Springer Verlag, 25-57.

Ford, D.C. and Ewers, R.O. (1978) The development of limestone caves in length and depth. *Canadian Journal of Earth Sciences*, **15**, 1783-98.

Ford, D.C. and Stanton, W.I. (1968) The geomorphology of the south-central Mendip Hills. *Proceedings Geologists' Association*, **79**, 401-27.

Ford, D.C. and Williams, P.W. (1989) *Karst Geomorphology and Hydrology*, Unwin Hyman, London.

Ford, T.D. (1956) The Speedwell Cavern. *Transactions of the Cave Research Group of Great Britain*, **2**, 97-124.

Ford, T.D. (1959) The Sutherland caves. *Transactions of the Cave Research Group of Great Britain*, **5**, 141-90.

Ford, T.D. (1964a) Fossil karst in Derbyshire. *Proceedings of the British Speleological Association*, **2**, 49-62.

Ford, T.D. (1964b) The Goyden Pot drainage system, Nidderdale, Yorkshire. *Transactions of the Cave Research Group of Great Britain*, **6**, 81-90.

Ford, T.D. (1966b) The underground drainage systems of the Castleton area, Derbyshire, and their evolution. *Transactions of the British Cave Research Association*, **5**, 369-96.

Ford, T.D. (1969) The Blue John fluorspar deposits of Treak Cliff in relation to the Boulder Bed. *Proceedings of the Yorkshire Geological Society*, **37**, 153-8.

Ford, T.D. (1975) *Ingleborough Cavern and Gaping Gill* (cave guide), Dalesman, Clapham.

Ford, T.D. (1976) The geology of caves. In *The Science of Speleology* (eds T.D. Ford and C.H.D. Cullingford), Academic Press, 11-60.

Ford, T.D. (ed.) (1977a) *Limestones and Caves of the Peak District*, Geo Books, Norwich.

Ford, T.D. (1977b) The caves of the Castleton area. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 297-346.

Ford, T.D. (1977c) The caves of Buxton and the upper Wye valley. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 383-93.

Ford, T.D. (1978) *The Speedwell Cavern* (guide book), R. and D. Harrison, Castleton.

Ford, T.D. (1980) *The Story of Treak Cliff Cavern* (guide book), Harold Harrison, Castleton.

Ford, T.D. (1984) Palaeokarsts in Britain. *Transactions of the British Cave Research Association*, **11**, 246-64.

Ford, T.D. (1986a) The evolution of the Castleton cave systems and related features, Derbyshire. *Mercian Geologist*, **10**, 91-114.

Ford, T.D. (1986b) The evolution of the Castleton cave systems, Derbyshire. *Transactions of the British Cave Research Association*, **13**, 131-48.

Ford, T.D. (1987) The origin of the Winnats Pass, Castleton, Derbyshire. *Mercian Geologist*, **10**, 241-9.

Ford, T.D. (ed.) (1989a) *Limestones and Caves of Wales*, Cambridge University Press, Cambridge.

Ford, T.D. (1989b) Tufa - the whole dam story. *Transactions of the British Cave Research Association*, **16**, 39-49.

Ford, T.D. (ed.) (1991) Peak and Speedwell Caverns-exploration and science. *Transactions of the British Cave Research Association*, **18**, 1-58.

Ford, T.D. and Beck, J.S. (1977) The caves of the Lathkill area. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 395-409.

Ford, T.D. and Burek, C.V. (1976) Anomalous limestone gorges in Derbyshire. *Mercian Geologist*, **6**, 59-66.

Ford, T.D., Burek, C.V. and Beck, J.S. (1975) The evolution of Bradwell Dale and its caves. *Transactions of the British Cave Research Association*, **2**, 133-40.

Ford, T.D., Burek, C.V. and Beck, J.S. (1977a) The caves of the Bradwell area. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 347-59.

Ford, T.D., Flindall, R. and Worley, N. (1977b) The caves and mines of the Matlock and Wirksworth area. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 409-33.

Ford, T.D., Gascoyne, M. and Beck, J.S. (1983) Speleothem dates and Pleistocene chronology in the Peak District of Derbyshire. *Transactions of the British Cave Research Association*, **10**, 103-15.

Ford, T.D. and Gunn, J. (1990) *Caves and Karst of the Peak District.* BCRA Cave Studies Series, Number 3.

Ford, T.D. and Hooper, M.J. (1964) The caves of the Isle of Portland. *Transactions of the Cave Research Group of Great Britain*, **7**, 11-35.

Ford, T.D. and King, R.J. (1969) The origin of the silica sand pockets in the Derbyshire Limestone. *Mercian Geologist*, **3**, 51-69.

Friederich, H. (1981) 'The hydrochemistry of recharge in the unsaturated zone, with special reference to the Carboniferous Limestone aquifer of the Mendip Hills.' PhD thesis, University of Bristol.

Friederich, H. and Smart, P.L. (1981) Dye tracer studies of the unsaturated zone recharge of Carboniferous Limestone aquifer of the Mendip Hills, England. *Proceedings Eighth International Congress of Speleology, Kentucky*, National Speleological Society, Huntsville, USA, 283-6.

Friederich, H. and Smart, P.L. (1982) The classification of autogenic percolation waters in karst aquifers: a study in G.B. Cave, Mendip Hills, Somerset. *Proceedings of the University of Bristol Speleological Society*, **16**, 143-60.

Gale, S.J. (1981a) 'Karst palaeoenvironments.' PhD thesis, University of Keele.

Gale, S.J. (1981b) The geomorphology of the Morecambe Bay karst and its implications for landscape chronology. *Zeitschrift für Geomorphologie*, **25**, 457-69.

Gale, S.J. (1984) Quaternary hydrological development in the Morecambe Bay karst, northwest England. *Norsk Geografiske Tidsskrift*, **38**, 185-92.

Gams, I. (1969) Some morphological characteristics of the Dinaric karst. *Geographical Journal*, **135**, 563-74.

Gams, I. (1974) *Kras*. Izdala Slovenska.

Gams, I. (1978) The polje: the problem of its definition. *Zeitschrift für Geomorphologie*, **22**, 170-81.

Gardener, C.G. (1983) Agen Allwedd, Trident and the missing link. *Caves and Caving*, **21**, 20-4.

Gardener, C.G. (1984) Daren Cilau - the route to the heart of the Llangattwg Mountain. *Descent*, **61**, 17-24.

Gardener, C.G. (1985) Daren Cilau - a new 5 km extension and the largest underground passage in Britain. *Caves and Caving*, **28**, 6-11.

Gardener, C.G. (1986) Cascading water and echoes of David Bowie from the depths of Daren Cilau. *Caves and Caving*, **33**, 20-6.

Gardener, C.G. (1988) The secret depths of the Llangattock Mountains. *Caves and Caving*, **40**, 26-31.

Gardener, C.G. and Westlake, C. (1985) Beyond the further reaches of Daren Cilau. *Caves and Caving*, **29**, 18-22.

Gardner, J.H. (1935) Origin and development of limestone caverns. *Bulletin of the Geological Society of America*, **46**, 1255-74.

Garwood, E.J. and Goodyear, E. (1924) The Lower Carboniferous succession in the northwest of England. *Quarterly Journal of the Geological Society*, **68**, 449-586.

Gascoine, W. (1979) Ogof Craig a Ffynnon. *Cambrian Caving Council Journal*, **5**, 66-9.

Gascoine, W. (1982) The formation of black deposits in some caves of Southeast Wales. *Transactions of the British Cave Research Association*, **9**, 165-75.

Gascoine, W. (1989) Hydrology of the limestone outcrop north of the coalfield. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 40-55.

Gascoine, W. (1991) Carno's secret. *Descent*, **102**, 20-2.

Gascoine, W. (1994) Water tracing at Pwll Du. *Caves and Caving*, **65**, 4-5.

Gascoyne, M. (ed.) (1973) The caves and potholes of East Kingsdale and Scales Moor - an area review. *Lancaster University Speleological Society Journal*, **1** (3), 11-58.

Gascoyne, M., Currant, A.P. and Lord, T.C. (1981) Ipswichian fauna of Victoria Cave and the marine paleoclimatic record. *Nature*, **294**, 652-4.

Gascoyne, M. and Ford, D.C. (1984) Uranium series dating of speleothems: Part 2, Results from the Yorkshire Dales and implications for cave development and Quaternary climates. *Transactions of the British Cave Research Association*, **11**, 65-85.

Gascoyne, M., Ford, D.C. and Schwarcz, H.P. (1978) Uranium-series dating and stable isotope studies of speleothems: Part 1, theory and techniques. *Transactions of the British Cave Research Association*, **5**, 91-111.

Gascoyne, M., Ford, D.C. and Schwarcz, H.P. (1983a) Rates of cave and landform development in the Yorkshire Dales from speleothem age data. *Earth Surface Processes and Landforms*, **8**, 557-68.

Gascoyne, M., Schwarcz, H.P. and Ford, D.C. (1983b) Uranium-series ages of speleothems from northwest England: correlation with Quaternary climate. *Philosophical Transactions of the Royal Society of London*, **B301**, 143-64.

Gatacre, E.V., Stanton, W.I. and Winsor, D. (1980) *Wookey Hole* (guide book), Wookey Hole Caves Ltd.

George, T.N., Johnson, G. A. J., Mitchell, M. *et al.* (1976) *A Correlation of Dinantian Rocks in the British Isles,* Geological Society of London, Special Report, **7**.

Gill, D.W. and Beck, J.S. (1991) *Caves of the Peak District,* Dalesman, Clapham.

Glasser, N. F. and Barber, G. (1995) Cave conservation plans: the role of English Nature. *Transactions of the British Cave Research Association*, **21**, 33-6.

Glennie, E. A. (1950) Further notes on Ogof Ffynnon Ddu. *Transactions of the Cave Research Group of Great Britain*, **1** (3), 1-47.

Glover, D. (1993) Ogof Cynnes. *Descent*, **112**, 30-1.

Glover, R.R. (1974) Cave development in the Gaping Gill system. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 343-84.

Goddard, F.J. (1944) GB Cave, Charterhouse-on-Mendip. *Proceedings of the University of Bristol Speleological Society*, **5**, 105-13.

Goldie, H. S. (1973) The limestone pavements of Craven. *Transactions of the Cave Research Group of Great Britain*, **15**, 175-90.

Goldie, H.S. (1976) 'Limestone pavements with special reference to Northwest England.' PhD thesis, Oxford University.

Goldie, H.S. (1981) Morphometry of the limestone pavements of Farleton Knott, Cumbria. *Transactions of the British Cave Research. Association*, **8**, 207-22.

Goldie, H.S. (1986) Human influence on landforms: the case of limestone pavements. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 515-40.

Goldie, H.S. (1993) The legal protection of limestone pavements in Great Britain. *Environmental Geology*, **21**, 160-6.

Gordon, D., Smart, P.L., Ford, D.C. *et al.* (1989) Dating of the late Pleistocene interglacial and interstadial periods in the United Kingdom from speleothem growth frequency. *Quaternary Research*, **31**, 14-26.

Gordon, J.E. and Sutherland, D.G. (1993) *Quaternary of Scotland*, Chapman and Hall, London.

Gosden, M.S. (1968) Peat deposits of Scar Close, Ingleborough, Yorkshire. *Journal of Ecology*, **56**, 345-53.

Goudie, A.S. and Gardner, R. (1985) *Discovering Landscape in England and Wales,* Allen and Unwin, London.

Graham, N. and Ryder, P.F. (1983) Sandy Hole, Isle of Portland. *Transactions of the British Cave Research Association*, **10**, 171-80.

Grandison, N. (1965) Mossdale Caverns. *Proceedings of the British Speleological Association*, **3**, 43-56.

Green, G.W. and Welch, F.B.A. (1965) *Geology of the Country around Wells and Cheddar,* British Geological Survey Memoir, Sheet 280.

Green, H.S. (1984). *Pontnewydd Cave: A Lower Palaeolithic Hominid site in Wales: the First Report*, National Museum of Wales, Cardiff.

Gresswell, R.K. (1958) The postglacial raised beach in Furness and Lyth, North Morecambe Bay. *Transactions of the Institute of British Geographers*, **25**, 79-103.

Griffiths, G.E. (1927) Juniper Gulf. *Yorkshire Ramblers Club Journal*, **5**, 209-14.

Groom, G.E. (1971) Geomorphology. In *Swansea and its Region* (ed. W.G. Balchin), British

Association Handbook, Swansea, 29-40.
Groom, G.E. and Williams, V.H. (1965) The solution of limestone in South Wales. *Geographical Journal*, **131**, 37-41.
Grund, A. (1903) Die Karsthydrographie. Studien aus Westbosnien. *Geographisches Abhandlungen herausgegeben von A. Penck*, 7 (3), 103-200.
Grund, A. (1914) Der geographische Zyklus im Karst. *Zeitschrift Geschalt Erdkunde*, 1914, 621-40.
Gunn, J. (1991) Water-tracing experiments in the Castleton Karst, 1950-1990. *Transactions of the British Cave Research Association*, **18**, 43-6.
Gunn, J. and Edmans, A. (1989) The Wye Head systems: some hydrological observations. *Caves and Caving*, **45**, 35.
Halliwell, R.A. (1974) A history of karst studies in Yorkshire. *Transactions of the British Cave Research Association*, **1**, 223-30.
Halliwell, R.A. (1979a) Gradual changes in the hydrology of the Yorkshire Dales demonstrated by tourist descriptions. *Transactions of the British Cave Research Association,* **6**, 36-40.
Halliwell, R.A. (1979b) Influence of contrasted rock types and geological structure on solutional processes in North-west Yorkshire. In *Geographical Approaches to Fluvial Processes* (ed. A.F. Pitty), Geo Books, Norwich, 51-71.
Halliwell, R.A. (1980) Karst waters of the Ingleborough area, North Yorkshire. *Proceedings of the University of Bristol Speleological Society*, **15**, 183-205.
Halliwell, R.A., Cavanagh, A.H. and Pitty, A.F. (1975) The influence of the pre-Carboniferous basement rocks on karst development in the Ingleton area of North-west Yorkshire. In *Karst Processes and Relevant Landforms* (ed. I. Gams), University of Ljubljana, 161-3.
Hancock, J.M. (1975) The petrology of the chalk. *Proceedings of the Geologists' Association*, **86**, 499-535.
Hancock, J.M. (1993) The formation and diagenesis of chalk. In *The Hydrogeology of the Chalk of Northwest Europe* (eds R.A. Downing, M. Price and G.P. Jones), Clarendon Press, Oxford.
Hanwell, J.D. and Newson, M.D. (1969) The frequency of severe storms over the Mendip Hills, Somerset. *Transactions of the Cave Research Group of Great Britain*, **11**, 209-12.
Hanwell, J.D. and Newson, M.D. (1970) *The Great Storms and Floods of July 1968 on Mendip*, Wessex Cave Club Occasional Paper, Series 2, no. 2.
Hardwick, P. and Gunn, J. (1995) Landform-groundwater interactions in the Gwenlais karst, South Wales. In *Geomorphology and groundwater* (ed. A. Brown), Wiley, Chichester.
Hartley, P. (1972) Description of Red Moss Pot. *Burnley Caving Club Review*, **1**, 31-4.
Hawkins, A.B. and Kellaway, G.A. (1971) Field meeting at Bristol with special reference to new evidence of glaciation. *Proceedings of the Geologists' Association*, **82**, 267-92.
Hennig, G.J., Grun, R. and Brunnacker, K. (1983) Speleothems, travertines and paleoclimates. *Quaternary Research*, **20**, 1-29.
Herbert, M. and Langford, H. (1991) Cwm Dwr II: its rediscovery and exploration. *Descent*, **100**, 38-40.
Heys, B. (1957) Mostly Hammer Pot. *Northern Pennine Club Journal*, **1**, 32-9.
Hicks, P.F. (1959) The Yoredale rocks of Ingleborough, Yorkshire. *Proceedings of the Yorkshire Geological Society*, **32**, 31-44.
Higginbottom, I.E. (1966) The engineering geology of chalk. *Proceedings of the Symposium on Chalk in Earthworks and Foundations* (1965), Institute of Civil Engineers, 1-13.
Hill, C.A. (1987) Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains. *New Mexico Bureau of Mines and Mineral Resources Bulletin*, **117**.
Hindley, R. (1965) Sinkholes on the Lincolnshire Limestone between Grantham and Stamford. *East Midlands Geographer*, **3**, 454-60.
Holmes, R. (1994) The hydrology of the Ibbeth Peril Cave System. *Caves and Caving*, **65**, 7.
Hooper, J.H.D. (1956) The Buckfastleigh Caverns (Devon). *Cave Science*, **4**, 96-121.
Hooper, J.H.D. (1960) The Buckfastleigh Caverns (Devon). *Cave Science*, **4**, 259-72.
House, M.R. (1991) Dorset dolines: Part 1, The higher Kingston road cutting. *Proceedings of the Dorset Natural History and Archaeological Society*, **112**, 105-8.
House, M.R. (1992) Dorset dolines: Part 2, Bronkham Hill. *Proceedings of the Dorset Natural History and Archaeological Society*, **112**, 113, 149-55.
Howarth J.H. *et al.* (1900) The underground waters of north-west Yorkshire. *Proceedings of the Yorkshire Geological and Polytechnic Society*, **14**, 1-44.

Howe, J.A. (1897) Notes on the pockets of sand and clay in the limestone of Derbyshire and Staffordshire. *Transactions and Annual Report of the North Staffordshire Field Club*, **31**, 143-9.

Hudson, R.G.S. (1933) The scenery and geology of northwest Yorkshire. *Proceedings of the Geological Association*, **44** 228-55.

Hughes, T. (1991) Wigmore Swallet. *Belfry Bulletin*, **462**, 10-13.

Hughes, T. McK. (1909) Ingleborough Part IV. The Carboniferous rocks. *Proceedings of the Yorkshire Geological Society*, **16**, 253-80.

Imbrie, J., Hayes, J.D., Martinson, D.G., McIntyre, A. *et al.* (1984) The orbital theory of Pleistocene climate: support from a revised chronology of the marine record. In *Milankovitch and Climate* (eds Burger, A.L., Imbrie, J., Hayes, J., Kukla, G., and Saltzman, B.), Klewer Academic, Amsterdam.

Ineson, P.R. and Ford, T.D. (1982) The South Pennine orefield: its genetic theories and eastward extension. *Mercian Geologist*, **8**, 285-303.

Irwin, D.J. (1991) *St Cuthbert's Swallet*, Bristol Exploration Club.

Irwin, D.J. and Jarratt, A.R. (1992) *Mendip Underground* (2nd edn), Mendip Publishing, Castle Cary.

Ixer, R.A. and Vaughan, D.J. (1993) Lead-zinc-fluorite-barite deposits of the Pennines, North Wales and the Mendips. In *Mineralisation in the British Isles* (eds Pattrick, R.A.D. and Polya, D.A.), Chapman and Hall, London, 355-411.

Jackson, J.W. and Storrs, Fox, W. (1913) The occurrence of Lynx in North Wales and Derbyshire. *Geological Magazine*, **10**, 159-62.

Jacobi, R.M. (1985) The history and literature of Pleistocene discoveries at Gough's Cave, Cheddar, Somerset. *Proceedings of the University of Bristol Speleological Society*, **17**, 102-15.

James, A.N., Cooper, A.H. and Holliday, D.W. (1981) Solution of the gypsum cliff (Permian, Middle Marl) by the River Ure at Ripon Parks, North Yorkshire. *Proceedings of the Yorkshire Geological Society*, **43**, 433-50.

Jarratt, A.R. (1991) The excavation and exploration of Wigmore Swallet. *Belfry Bulletin,* **460**, 19-23.

Jeffreys, A.L. (1975) Uamh Coire Sheilach, Argyllshire. *Bulletin of the British Cave Research Association*, **10**, 8-11.

Jeffreys, A.L. (1984) *Scotland Underground*, Oldham, Crymych.

Jeffreys, A.L. (1994) Scotland's longest cave. *Descent*, **119**, 16-17.

Jenkins, D.A. (1963) Notes on the geology of Agen Allwedd and Mynydd Llangattwg. *Proceedings of the British Speleological Association Annual Conference*, **1**, 49-65.

Jenkinson, R. (1989) The archaeological caves of Creswell Crags. *Transactions of the British Cave Research Association*, **16**, 91-4.

Jenkinson, R.D.S. (1984) *Creswell Crags: Late Pleistocene sites in the East Midlands*, British Archaeological Reports, British Series, 122.

Jennings, J.N. (1985) *Karst Geomorphology*, Blackwell, Oxford.

Johnson, M.R.W. and Parsons, I. (1979) *MacGregor and Phemister's Geological Guide to the Assynt district of Sutherland,* Edinburgh Geological Society, Edinburgh.

Johnson, R.H. (1967) Some glacial, periglacial and karstic landforms in the Sparrowpit-Doveholes area of north Derbyshire. *East Midland Geographer*, **4**, 224-38.

Jones, D.M.H. (1957) Fairy Hole Cave, Weardale. *Yorkshire Ramblers Club Journal*, **8**, 118-26.

Jones, K. (1991) Hydrology of Pant-y-Llyn. *Isca Caving Club Journal*, **13**, 40-3.

Jones, K., Bryan, G. and Adams, J. (1984) Ogof Pant-y-Llyn. *Isca Caving Club Journal*, **2**, 14-36.

Jones, O.T. and Lewis, W.V. (1941) Water levels in Fowl Mere and other Breckland meres. *Geographical Journal*, **97**, 158-66.

Judson, D.M. (1964) Mongo Gill Hole 1964 Extension. *Journal of the Craven Pothole Club*, **3**, 172-80.

Katzer, F. (1909) *Karst und Karsthydrographie*, Zur Kunde der Balkanhalbinsel, Sarajevo, **8**.

Kealy, L. (1992) Dan-yr-Ogof 1991. *Caves and Caving*, **56**, 15-18.

Kellaway, G.A. (1991) Structural and glacial control of thermal water emission in the Avon basin at Bath. In *Hot Springs of Bath* (ed. G.A. Kellaway), Bath City Council, Bath, 205-42.

Kendall, A. (1988) Aragonite in Ogof Daren Cilau. *Transactions of the British Cave Research Association*, **15**, 83-4.

Kent, P.E. (1957) Triassic relics and the 1,000 foot surface in the Southern Pennines. *East Midlands Geographer*, **1**, 3-10.

Kerney, M.P., Brown, E.H. and Chandler, T.S. (1964) The Late-glacial and Post-glacial history of the Chalk escarpment near Brook, Kent. *Philosophical Transactions of the Royal Society* **B248**, 135-204.

King, A. (1974) A review of archaeological work in the caves of northwest England. In *Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 182-200.

King, C.A.M. (1969) Trend surface analysis of central Pennine erosion surfaces. *Transactions of the Institute of British Geographers*, **47**, 47-69.

King, C.A.M. (1976) *Northern England* (The geomorphology of the British Isles), Methuen, London.

Kirkaldy, J.F. (1950) Solution of the Chalk in the Mimms Valley, Hertfordshire. *Proceedings of the Geological Association*, **61**, 219-24.

Knighton, A.D. (1975) Form adjustment in a limestone dry valley at Castleton, Derbyshire. *East Midlands Geographer*, **6**, 130-5.

Langthorne, C. (1976) The exploration and survey of the Cliff Force system. *Journal of the Moldywarp Speleological Group,* **9**, 5-10.

Latham, A.G., Schwarz, H.P., Ford, D.C. and Pearce, G.W. (1979) Palaeomagnetism of stalagmite deposits. *Nature*, **280**, 383-5.

Lawson, T.J. (1981) First Scottish date from the last interglacial. *Scottish Journal of Geology*, **17**, 301-3.

Lawson, T.J. (1982) Uranium series dating of certain Assynt speleothems: preliminary results. *Grampian Speleological Group Bulletin*, Series 2, **1** (4), 8-12.

Lawson, T.J. (1983) 'Quaternary geomorphology of the Assynt area, N.W. Scotland.' PhD thesis, University of Edinburgh.

Lawson, T.J. (1986) Loch Lomond advance glaciers in Assynt, Sutherland, and their palaeoclimatic implications. *Scottish Journal of Geology*, **22**, 289-98.

Lawson, T.J. (1988) *Caves of Assynt,* Grampian Speleological Group, Occasional Publication, **6**.

Lawson, T.J. (1993) Creag nan Uamh. In *Quaternary of Scotland* (eds J.E. Gordon and D.G. Sutherland), Chapman and Hall, London, 127-33.

Lawson, T.J. (1995a) The Creag nan Uamh caves. In *The Quaternary of Assynt and Coigach: A guide* (ed. T. J. Lawson), Quarternary Research Association, 87-103.

Lawson, T.J. (1995b) An analysis of sediments in caves in the Assynt area, NW Scotland. *Cave and Karst Science*, **22**, 23-30.

Leakey, R.D. (1947) The caverns of Mossdale Scar. *Cave Science*, **1**, 7-18.

Lehmann, H. (1936) Morphologische Studien auf Java. *Geographische Abhandler*, **3**.

Leitch, D.E. (1960) *Ogof Agen Allwedd in relation to the Mynydd Llangattwg*. Cave Research Association Publication, **10**.

Leitch, D.E. (1973) In *A Review of the Speleogenesis of Agen Allwedd*, Hereford Caving Club 21st Anniversary Publication, 7-43.

Leroi-Gourhan, A. (1985) Pollen analysis of sediment samples from Gough's Cave, Cheddar. *Proceedings of the University of Bristol Speleological Society*, **17**, 141-4.

Levitan, B.M., Audsley, A., Hawkes, C.J. *et al.* (1989) Charterhouse Warren Farm Swallet, Mendip, Somerset: exploration, geomorphology, taphonomy and archaeology. *Proceedings of the University of Bristol Speleological Society*, **18**, 171-240.

Lewin, J. (1969) *The Yorkshire Wolds, A Study in Geomorphology*, University of Hull Occasional Paper in Geography, **11**.

Lloyd, J. and King, E. (1780) An account of Eldon Hole in Derbyshire. *Philosophical Transactions of the Royal Society*, **61**, 250-65.

Lloyd, O.C. (1980) Porth yr Ogof, Ystradfellte, Powys. *Proceedings of the University of Bristol Speleological Society*, **15**, 259.

Long, M.H. (1969) Developments in Wharfedale and Langstrothdale. *Cave Diving Group Newsletter*, **117**, 6-16.

Long, M.H. (1971) Dentdale - present and future. *Journal of the British Speleological Association*, **6**, 9-17.

Long, M.H. (1974) The caves of Wharfedale. In *Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 410-21.

Long, M.H. (1992) Hydrology and flooding risks [of Sleets Gill Cave]. *Descent*, **106**, 22.

Lowe, D.J. (1978) Farnham Cave, a rift cave in the Magnesian Limestone. *Transactions of the British Cave Research Association*, **5**, 23-8.

Lowe, D.J. (1989a) Limestones and caves of the Forest of Dean. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 106-16.

Lowe, D.J. (1989b) The geology of the Carboniferous Limestone of South Wales. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 3-19.

Lowe, D. J. (1992a) Chalk caves revisited.

Transactions of the British Cave Research Association, **19**, 55-8.

Lowe, D.J. (1992b) 'The origin of limestone caverns: an inception horizon hypothesis.' PhD thesis, Manchester Metropolitan University.

Lowe, D.J. (1992c) A historical review of concepts of speleogenesis. *Transactions of the British Cave Research Association*, **19** 63-90.

Lowe, D.J. (1993).The Forest of Dean caves and karst: inception horizons and iron-ore deposits. *Transactions of the British Cave Research Association*, **20**, 31-43.

Lyon, M.K. (1974) The caves of Dentdale. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 227-34.

McConnel, R.B. (1939) The relic surfaces of the Howgill Fells. *Proceedings of the Yorkshire Geological Society*, **24**, 152-64.

Macfadyen, W.A. (1970) *Geological Highlights of the West Country,* Butterworths, London.

Macklin, M.G. (1985) Floodplain sedimentation in the Upper Axe valley, Mendip, England. *Transactions of the Institute of British Geographers*, **10**, 235-44.

MacTavish, A. (1975) New Portland cave - Blacknor Hole. *Descent*, **32**, 4-9.

Malott, C.A. (1937) Invasion theory of cavern development. *Proceedings of the Geological Society of America*, 323.

Mansel-Pleydell, J.C. (1873) A brief memoir on the geology of Dorset. *Geological Magazine*, **10**, 402-13, 438-47.

Marr, J.E. (1913) The meres of Breckland. *Proceedings of the Cambridge Philosophical Society for Mathematical and Physical Sciences,* **17**, 58-61.

Martel, E.A. (1921) *Nouveau traité des eaux souterraines*, Doin, Paris.

Martin, E.A. (1920) The glaciation of the South Downs. *Transactions of the Southeast Union of Scientific Societies*, 13-30.

Martinson, D.G., Pisias, N.J., Hayes, J.D. *et al.* (1987) Age dating and the orbital theory of the Ice Ages, development of a high resolution nought to 300,000 year chronostratigraphy. *Quaternary Research*, **27**, 1-29.

Mason, E.J. (1968) Ogof yr Esgyrn, Dan-yr-Ogof Caves, Brecknock. *Archaeologia Cambriensis*, **117**, 18-71.

Meade-King, S. (1984) Full of eastern promise - the continuing saga of Thrupe Lane Swallet. *Wessex Caving Club Journal*, **13**, 47-56.

Millward, R. and Robinson, A.H.W. (1975) *The Peak District,* Eyre Methuen, London.

Milner, A.J. (1972) *The Caves of the Alum Pot area*, University of Leeds Speleological Association.

Mitchell, W.A. (ed.) (1991) *Western Pennines: Field Guide*, Quaternary Research Association, Cambridge.

Mitchell, W.A. (1994) Drumlins in ice sheet reconstructions, with reference to the western Pennines, northern England. *Sedimentary Geology*, **91**, 313-31.

Moisley, H.A. (1955) Some karstic features in the Malham Tarn district. *Annual Report, Council for the Promotion of Field Studies*, 33-42.

Monico, P. (ed.) (1989a) Mossdale Caverns and Langcliffe Pot. *University of Leeds Speleological Association Explorations Journal*, **2**, 1-41.

Monico, P. (ed.) (1989b) Sleets Gill Cave. *University of Leeds Speleological Association Explorations Journal*, **2**, 47-52.

Monico, P. (ed.) (1989c) Penyghent Pot. *University of Leeds Speleological Association Explorations Journal*, **2**, 65-80.

Monico, P. (1992) Underwater Dentdale. *Descent*, **108**, 10.

Monico, P. (ed.) (1995) *Northern Sump Index,* Cave Diving Group.

Moore, S. (1989) The Afon Nedd Fechan caves. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 165-76.

Morris, H.M. and Fowler, C.H. (1937) The flow and bacteriology of underground water in the Lea valley. *32nd Annual Report of the Metropolitan Water Board*, London.

Mortimer, J.R. (1885) On the origin of the Chalk dales of Yorkshire. *Proceedings of the Yorkshire Geological Society*, **9**, 29-42.

Mortimore, R.N. (1993) Chalk water and engineering geology. In *The Hydrogeology of the Chalk of Northwest Europe* (eds. R.A. Downing, M. Price and G.P. Jones), Clarendon Press, Oxford.

Moseley, F. (1972) A tectonic history of northwest England. *Quarterly Journal of the Geological Society*, **128**, 561-98.

Moseley, F. (1973) Orientations and origins of joints, faults and folds in Carboniferous limestones of N.W. England. *Transactions of the. Cave Research. Group*. **15**, 99-106.

Mostaghel, M.A. and Ford, T.D. (1986) A sedimentary basin evolution for ore genesis in the South Pennine Orefield. *Mercian Geologist*, **10**, 209-24.

Mulholland, P. (1992) Prid's sonic survey. *Descent*, **107**, 29.

Mulholland, P. (1994) New discoveries in Cnoc nan Uamh. *Grampian Speleological Group Bulletin,* Third Series, **3** (2), 19-22.

Mullen, G.J. (1987) Little Neath River Cave. *Caves and Caving*, **37**, 8-11.

Mullen, G.J. (1988) The Little Neath River Cave 1971-1987. *Proceedings of the University of Bristol Speleological Society*, **18**, 314-16.

Mullen, G.J. (1990) Little Neath River Cave. *Transactions of the British Cave Research Association,* **17**, 135.

Murlis, B. (1992) Pushing Dan-yr-Ogof. *Descent*, **107**, 20-2.

Murphy, P.J. (1993) Lithological analysis of sediment samples from the Rushup Edge-Speedwell Cavern-Peak Cavern cave system, Castleton, Derbyshire. *Transactions of the British Cave Research Association,* **20**, 30.

Myers, J.O. (1948) The formation of Yorkshire caves and potholes. *Transactions of the Cave Research Group of Great Britain*, **1**, 26-9.

Mylroie, J.E. and Carew, J.L. (1987) Field evidence of the minimum time for speleogenesis. *National Speleological Society Bulletin*, **49**, 67-72.

Nash, D. (1991) Review of available literature. *Cave Science*, **18**, 41-3.

NCMRS (1980) *The Mines of Grassington Moor*. Northern Cavern and Mine Research Society, British Mining Records, **13**.

Neill, A. (1988) Joint Mitnor Cave, Buckfastleigh, Devon. *Caves and Caving*, **39**, 38-9.

Newson, M.D. (1969) Some geomorphological implications of the Mendip flood of July 1968. *Transactions of the Cave Research Group of Great Britain*, **11**, 213-14.

Newson, M.D. (1972) Rickford and Langford resurgences, Mendip Hills, Somerset: a problem in limestone hydrology. *Proceedings of the University of Bristol Speleological Society*, **13**, 105-12.

Newson, M.D. and Atkinson, T.C. (1970) Report of water tracing in the Inchnadamph area. *Grampian Speleological Group Bulletin*, **4**, 32.

Nixon, D. (1991) Shooting the White River. *Descent*, **101**, 20-2.

Nixon, D. (1992) The White River Series, Peak Cavern. *Caves and Caving*, **55**, 2-3.

Noel, M. (1983) The magnetic remanence and anisotropy of susceptibility of cave sediments from Agen Allwedd, South Wales. *Geophysical Journal of the Royal Astronomical Society*, **72**, 557-70.

Noel, M. (1986) The palaeomagnetism and magnetic fabric of cave sediments from Pwll y Gwynt, South Wales. *Physics of the Earth and Planetary Interiors*, **44**, 62-71.

Noel, M. (1987) The magnetostratigraphy of cave sediments in Masson Hill, Derbyshire. *Proceedings of the Yorkshire Geological Society*, **46**, 193-201.

Noel, M. (1988) Palaeomagnetism of cave sediments from Mynydd Llangattwg. *Transactions of the British Cave Research Association*, **15**, 3-9.

Noel, M., Homonko, P. and Bull, P.A. (1979) The palaeomagnetism of sediments from Agen Allwedd, Powys. *Transactions of the British Cave Research Association*, **9**, 134-41.

Noel, M., Retallick, W.G. and Bull, P.A. (1981) Further palaeomagnetic studies of sediments from Agen Allwedd. *Transactions of the British Cave Research Association*, **8**, 178-87.

Noel, M., Shaw, R.P. and Ford, T.D. (1984) A palaeomagnetic reversal in early Quaternary sediments in Masson Hill, Matlock, Derbyshire. *Mercian Geologist*, **9**, 235-42.

North, F.J. (1952) Some geological aspects of subsidence not due to mining. *Proceedings of the South Wales Institute of Engineers*, **68**, 127-53.

North, F.J. (1962) *The River Scenery at the Head of the Vale of Neath* (4th edn), National Museum of Wales, Cardiff.

Norton, M.G. (1966) Interim report on the Ladder Dig Series, G.B. Cave, Charterhouse-on-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **11**, 63-70.

Norton, M.G., Savage, D.A. and Standing, P.A. (1967) The Little Neath River Cave, South Wales. *Proceedings of the University of Bristol Speleological Society*, **11**, 186-200.

O'Connor, J. (1964) The geology of the area around Malham Tarn, Yorkshire. *Field Studies*, **2**, 53-82.

O'Connor, J., Williams, D.S.F. and Davies, G.M. (1974) Karst features of Malham and the Craven Fault zone. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 395-409.

O'Reilly, P.M. and Bray, L.G. (1974) A preliminary hydrological study in Ogof Ffynnon Ddu, Breconshire. *Transactions of the British Cave Research Association*, **1**, 65-75.

O'Reilly, P.M., O'Reilly, S.E. and Fairburn, C.M. (1969) *Ogof Ffynnon Ddu*, South Wales Caving Club, Penwyllt.

Oakman, C.D. (1979) Derbyshire sough hydrogeology and the artificial drainage of the Stanton syncline near Matlock, Derbyshire. *Transactions of the British Cave Research Association*, **6**, 169-94.

Oates, N.K. (1981) 'Wild brine extraction and related subsidence in the Cheshire saltfield.' PhD thesis, University of Aston.

Ockenden, A. (1972) An investigation of the swallet holes at Waterend. *Peleobates, Croydon Caving Club*, **19**, 18-21.

Oldfield, F. (1960) Late Quaternary changes in climate, vegetation and sea level in lowland Lonsdale. *Transactions of the Institute of British Geographers*, **28**, 99-117.

Oldham, T. (1982) *Caves of Gower,* Oldham, Crymych, Dyfed.

Owen, T.R. (1954) The structure of the Neath disturbance between Bryniau, Gleision and Glynneath, South Wales. *Quarterly Journal of the Geological Society of London*, **109**, 333-65.

Palmer, A.N. (1975) The origin of maze caves. *National Speleological Society Bulletin*, **37**, 56-76.

Palmer, A.N. (1984) Geomorphic interpretation of karst features. In *Groundwater as a Geomorphic Agent* (ed. R.G. LaFleur), Allen and Unwin, London, 173-209.

Palmer, A.N. (1987) Cave levels and their interpretation. *National Speleological Society Bulletin*, **49**, 50-66.

Palmer, A. N. (1991) Origin and morphology of limestone caves. *Geological Society of America Bulletin*, **103**, 1-21.

Palmer, R. (1988) Cheddar Caves '88. *Descent*, **83**, 34-6.

Parker, J. (1978) Ogof Craig a Ffynnon - 4 miles of new cave found in South Wales. *Descent*, **38**, 16-26.

Parkinson, D. (1953) The Carboniferous Limestone at Treak Cliff, Derbyshire with notes on the structure of the Castleton reef-belt. *Proceedings of the Geologists' Association*, **64**, 251-68.

Parry, J.T. (1960) The limestone pavements of northwestern England. *Canadian Geographer*, **16**, 14-21.

Patchett, A.N. (1953) Clapham Cave. *Journal of Bradford Pothole Club*, **1**, 21-31.

Paterson, K. (1971) Weichselian deposits and fossil periglacial structures in North Berkshire. *Proceedings of the Geologists' Association*, **82**, 455-68.

Paterson, K. (1977) Scarp-face dry valleys near Wantage, Oxfordshire. *Transactions of the Institute of British Geographers*, **2**, 192-204.

Patterson, D.A., Davey, J.C., Cooper, A.H., and Ferris, J.K. (1995) The investigation of dissolution subsidence incorporating microgravity geophysics at Ripon, Yorkshire. *Quarterly Journal of Engineering Geology*, **28**, 83-94.

Pedley, H.M. (1987) The Flandrian (Quaternary) Caerwys tufa, North Wales: an ancient tufa barrage deposit. *Proceedings of the Yorkshire Geological Society*, **46**, 141-52.

Pedley, H.M. (1993) Sedimentology of the Late Quaternary barrage tufas in the Wye and Lathkill valleys, north Derbyshire. *Proceedings of the Yorkshire Geological Society*, **49**, 197-206.

Pentecost, A. (1981) The tufa deposits of the Malham District. *Field Studies*, **2**, 365-87.

Pfeiffer, P. (1991). 'Geologische Einheiten eines Limestone Pavement Giebetes, Hutton Roof Complex, Cumbria, Nordwest England.' Diplomarbeit, Geographishen Instituts, Universität Tübingen.

Pigott, C.D. (1962) Soil formation and development on the Carboniferous limestone of Derbyshire. *Journal of Ecology*, **50**, 145-56.

Pigott, C.D. (1965). The structure of limestone surfaces in Derbyshire. *Geographical Journal,* **131**, 41-4.

Pilkington, J. (1789) *A View of the Present State of Derbyshire* (2 vols), Derby.

Pitty, A.F. (1968) The scale and significance of solutional loss from the limestone tract of the southern Pennines. *Proceedings of the Geologists' Association*, **79**, 153-77.

Pitty, A.F. (1969) Rates of seepage in Poole's Cavern, Derbyshire. *Proceedings of the British Speleological Association*, **7**, 7-15.

Pitty, A.F. (1971) Evidence related to the development of caverns from karst water studies in Peak Cavern, Derbyshire. *Transactions of the British Cave Research Association*, **13**, 53-5.

Pitty, A.F. (1974) Karst water studies in and around Ingleborough Cavern. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 127-39.

Pitty, A.F., Ternan, J.L., Haliwell, R.A. and Crowther, J. (1986) Karst water temperatures and the shaping of Malham Cove, Yorkshire. In

New Directions in Karst (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 281-91.

Pohl, E.R. (1955) *Vertical shafts in limestone caves*, National Speleological Society of America Occasional Paper, **2**.

Potts, J. (1977) The caves of the Dove and Manifold valleys. In *Limestones and Caves of the Peak District* (ed. T.D. Ford), Geo Books, Norwich, 445-61.

Powell, J.H., Cooper, A.H. and Benfield, A.C. (1992) *Geology of the Country around Thirsk*, British Geological Survey Memoir, Sheet 52.

Powell, R. (1954) Providence Pot - Dowber Gill. *Journal of the Craven Pothole Club*, **1**, 269-73.

Price, D. (1988) Resurrection Passage: a dry route to Maytime. *Caves and Caving*, **40**, 11-13.

Price, G. (1977) *Fairy Cave Quarry. A Study of the Caves,* Cerberus Speleological Society, Occasional Publication no. 1.

Price, G. (1983) The caves of Fairy Cave Quarry. *Studies in Speleology*, **4**, 71-6.

Price, G. (1984) Llethrid Swallet, Tooth Cave and the caves associated with the Wellhead Resurgence, Gower Peninsula. *Cerberus Speleological Society Journal*, **14**, 13-19.

Price, M. (1994) Cretaceous limestone (chalk) environments. *Cave Studies Series*, British Cave Research Association, **5**, 30-5.

Price, M., Downing, R.A. and Edmunds, W.M. (1993) The chalk as an aquifer. In *The Hydrogeology of the Chalk of Northwest Europe* (eds R.A. Downing, M. Price and G.P. Jones), Clarendon Press, Oxford, 35-58.

Price, M., Atkinson, T.C., Barker, J.A. *et al.* (1992) A tracer study of the danger posed to a chalk aquifer by contaminated highway runoff. *Proceedings of the Institute of Civil Engineers, Water Maritime and Energy Journal*, **96**, 9-18.

Priesnitz, K. (1985) Nichtbiogene kalklösung am Lough Leane und am Muckross Lake (SW-Irland). *Berliner Geographische Studien*, **16**, 55-69.

Prince, H.C. (1962) Pits and ponds in Norfolk. *Erkunde*, **16**, 10-31.

Prince, H.C. (1964) The origin of pits and depressions in Norfolk. *Geography*, **49**, 15-32.

Pritchard, T.O. (1961) *Management Plan, Rostherne Mere National Nature Reserve, Cheshire,* Nature Conservancy, Shrewsbury.

Proctor, C.J. (1988) Sea level related caves on Berry Head, South Devon. *Transactions of the British Cave Research Association*, **15**, 39-49.

Proctor, C.J. (1995) 'A British Pleistocene chronology based on uranium series and electron spin resonance dating of speleothem.' PhD thesis, University of Bristol.

Proctor, C.J. and Smart, P.L. (1989) A new survey of Kent's Cavern, Devon. *Proceedings of the University of Bristol Speleological Society*, **18**, 422-9.

Proctor, C.J. and Smart, P.L. (1991) A dated cave sediment record of Pleistocene transgressions on Berry Head, southwest England. *Journal of Quaternary Science*, **6**, 233-44.

Proudlove, G.S. (1985) Recent explorations in Peak Cavern. *Caves and Caving*, **29**, 3-12.

Quirk, D.G. (1986) Mineralisation and stress history in North Derbyshire. *Bulletin of the Peak District Mines and Historical Society*, **9**, 333-86.

Quirk, D.G. (1993) Origin of the Peak District orefield. *Bulletin of the Peak District Mines and Historical Society*, **12**, 4-15.

Railton, C.L. (1953) *The Ogof Ffynnon Ddu System*, Cave Research Group of Great Britain, Publication no. 6.

Raistrick, A. (1931) The glaciation of Wharfedale, Yorkshire. *Proceedings of the Yorkshire Geological Society*, **22**, 9-30.

Raistrick, A. (1938) Mineral deposits in the Settle-Malham district, Yorkshire. *Naturalist* (1938), 119-25.

Raistrick, A. (1954) The calamine mines, Malham, Yorkshire. *Proceedings of the University of Durham Philosophical Society*, **11**, 125-30.

Ramsbottom, W.H.C. (1973) Transgressions and regressions in the Dinantian: a new synthesis of British Dinantian stratigraphy. *Proceedings of the Yorkshire Geological Society*, **39**, 567-607.

Ramsbottom, W.H.C. (1974) Dinantian. In *The Geology and Mineral Resources of Yorkshire* (eds D.H. Rayner and J.E. Hemingway), Yorkshire Geological Society, 47-73.

Ratcliffe, A. (ed.) (1977) *A Nature Conservation Review: The Selection of Biological Sites of National Importance to Nature Conservation. Volume 2: Site Accounts*, Cambridge University Press, Cambridge.

Rauch, H.W. and White, W.B. (1970) Lithologic controls on the development of solution porosity in carbonate aquifers. *Water Resources Research*, **6**, 1175-92.

Rauch, H.W. and White, W.B. (1977) Dissolution kinetics of carbonate rocks: 1. Effects of lithology on dissolution rate. *Water Resources Research*, **13**, 381-94.

Rayner, D.H. (1953) The Lower Carboniferous rocks in the north of England - a review. *Proceedings of the Yorkshire Geological Society*, **28**, 231.

Reeve, T. (1979) *Caves and Swallets in Chalk*, Chelsea Speleological Society Records, **9**.

Reeve, T. (1980) The discovery and exploration of Beachy Head Cave; a cave system in Cretaceous Chalk. *William Pengelly Cave Studies Trust Newsletter*, **37**, 1-4.

Reid, C. (1887) On the origin of dry Chalk valleys and of Coombe rock. *Quarterly Journal of the Geological Society of London*, **43**, 364-73.

Reid, C. (1892) The Pleistocene deposits of the Sussex coast and their equivalents in other districts. *Quarterly Journal of the Geological Society of London*, **48**, 344-61.

Reid, C. (1899) *The Geology of the Country around Dorchester,* British Geological Survey Memoir, Sheet 328.

Reynolds, C.S. (1979) The limnology of the eutrophic meres of the Shropshire-Cheshire plain: a review. *Field Studies*, **5**, 93-173.

Reynolds, S.M. (1927) The Mendips. *Geography*, **14**, 187-92.

Rhoades, R. and Sinacori, N.M. (1941) Patterns of groundwater flow and solution. *Journal of Geology*, **49**, 785-94.

Richardson, D.T. (1974) Karst waters of the Alum Pot area. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 140-8.

Richardson, J.B. (1937) A revival of lead mining at Halkyn, North Wales. *Transactions of the Institute of Mining and Metallurgy*, **46**, 339-461.

Robey, J.A. (1965) The drainage of the area between the River Wye and the River Lathkill. *Proceedings of the British Speleological Association* **3**, 1-10.

Rogers, N. (1992) Open Season leads to the Precious Years. *Descent*, **105**, 20-2.

Roglić, J. (1938) Morphologie der Poljen von Kupres und Vukovsko. *Zeitschrift Geschalt Erdkunde*, **7/8**, 291-316.

Rose, J. (1987) Status of the Wolstonian glaciation in the British Quaternary. *Quaternary Newsletter*, **53**, 1-9.

Rose, L. and Vincent, P.J. (1986a) Alkalinity measurements in karst water studies. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 1-15.

Rose, L. and Vincent, P.J. (1986b) The kamenitzas of Gaitbarrows National Nature Reserve, north Lancashire, England. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 473-96.

Rose, L. and Vincent, P.J. (1986c) Some aspects of the morphometry of grikes - a mixture model approach. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 497-514.

Rose, W.C.C. and Dunham, K.C. (1977) *Geology and Hematite Deposits of South Cumbria*, Geological Survey of Great Britain Economic Memoir, Sheets 58 and 48.

Rowe, P.J., Atkinson, T.C. and Jenkinson, R.D.S. (1989a) Uranium-series dating of cave deposits at Creswell Crags Gorge, England. *Transactions of the British Cave Research Association*, **16**, 3-17.

Rowe, P.J., Austin, T. and Atkinson, T.C. (1989) The Quaternary evolution of the South Pennines. *Transactions of the British Cave Research Association*, **16**, 117-21.

Rundle, S.D. (1993) The invertebrate community of Pant-y-llyn, a Welsh turlough. *Bulletin of the British Ecology Society*, **24**, 215-21.

Rushton, K.R., Smith, E.J. and Tomlinson, L.M. (1982) An improved understanding of flow in a limestone aquifer using field evidence and mathematical models. *Journal of the Institute of Water Engineers and Scientists*, **36**, 369-87.

Ryder, P.F. (1974) The caves of the Beinn an Dubhaich area, Isle of Skye. *Transactions of the British Cave Research Association*, **1** 101-24.

Ryder, P.F. (1975) Phreatic network caves in the Swaledale area, Yorkshire. *Transactions of the British Cave Research Association*, **2**, 177-92.

Ryder, P.F. (1981) Cliff Force Cave. *Journal of Craven Pothole Club*, **6**, 132-5.

Ryder, P.F. (1982) MSG expedition to Skye and Applecross. *Grampian Speleological Group Bulletin,* Second Series, **3** (5), 30-1.

Ryder, P.F. and Cooper, A.H. (1993) A cave system in Permian gypsum at Houtsay quarry, Newbiggin, Cumbria, England. *Transactions of the British Cave Research Association*, **20**, 23-8.

Sadler, H.E. (1964) The origin of the 'Beach Beds' in the Lower Carboniferous of Castleton, Derbyshire. *Geological Magazine*, **101**, 360-72.

Salmon, L.B. (1956) Giant's Hole, Castleton, Derbyshire. *Cave Science*, **4**, 1-33.

Salmon, L.B. (1959) Giant's Hole, Castleton, Derbyshire (second report). *Cave Science*, **4**, 230-40.

Salmon, L.B. and Boldock, G. (1951a) Nettle Pot, Castleton, Derbyshire. *Cave Science*, **2**, 331-8.

Salmon, L.B. and Boldock, G. (1951b) Oxlow Cavern, Castleton, Derbyshire. *Cave Science*, **3**, 13-20.

Savage, D. (1969) The visible effects of the flood of July 10th 1968 in and around G.B. Cave, Charterhouse-on-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **12**, 123-6.

Schwarzacher, W. (1958) The stratification of the Great Scar Limestone in the Settle district of Yorkshire. *Liverpool and Manchester Geological Journal*, **2**, 124-42.

Shackleton, N.J., Berger, A. and Peltier, W.R. (1990) Alternative astronomical calibration of the Lower Pleistocene timescales based on ODP Site 677. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **81**, 251-61.

Shaw, J. (1983) *The Geology of the Area around Malham Tarn, North Yorkshire*. Occasional Publication of the Field Studies Council, **3**.

Shaw, R.P. (1983) Rediscovery of the lost Pilkington's Cavern, Castleton, Derbyshire. *Transactions of the British Cave Research Association*, **10**, 1-8.

Shaw, T.R. (1992) *History of Cave Science: The Scientific Investigation of Limestone Caves to 1900*, Oldham, Crymych.

Sheail, G.M. (1971) Coombes of the white Chalk downs. *Geographical Magazine*, **43**, 471-3.

Sherlock, R.L. (1929) The origin of the Devil's Dyke, near Brighton. *Proceedings of the Geologists' Association*, **40**, 371-2.

Simms, M.J. (1990) Triassic palaeokarst in Britain. *Transactions of the British Cave Research Association*, **17**, 93-101.

Simms, M.J. (1994) Emplacement and preservation of vertebrates in caves and fissures. *Zoological Journal of the Linnean Society*, **112**, 261-83.

Simms, M., Farrant, A. and Hunt, J. (1996) Ogof Draenen. *Caves and Caving*, **71**, 10-13.

Simon, A.M., Donovan, D.T. and Taylor, M. (1961) The stratigraphy and archaeology of the Late Glacial and Postglacial deposits at Brean Down, Somerset. *Proceedings of the University of Bristol Speleological Society*, **9**, 67-136.

Simpson, E. (1935) Notes on the formation of the Yorkshire caves and potholes. *Proceedings of the University of Bristol Speleological Society*, **4**, 224-32.

Simpson, E. (1950) Lost John's Cave. *Cave Science*, **2** (13), 206-17.

Simpson, E. (1967) Malham waters and Pikedaw Calamine Mine. *Journal of the British Speleological Association*, **41**, 24-9.

Simpson, I.M. and Broadhurst, F. (1969) A boulder bed at Treak Cliff, Derbyshire. *Proceedings of the Yorkshire Geological Society*, **37**, 141-52.

Small, R.J. (1962) A short note on the origin of Devil's Dyke near Brighton. *Proceedings of the Geologists' Association*, **73**, 187-92.

Small, R.J. (1964) The escarpment dry valleys of the Wiltshire Chalk. *Transactions of the Institute of British Geographers*, **34**, 33-52.

Smart, P.L. (1976) Catchment delimitation in karst areas by the use of quantitative tracer methods. *Proceedings of the 3rd International Symposium on Underground Water Tracing*, Bled, Yugoslavia, 291-8.

Smart, P.L. (1981) Variation of conduit flow velocities with discharge in the Longwood sinks to Cheddar Rising systems, Mendip Hills. *Proceedings of the 8th International Congress on Speleology*, **1**, 333-7.

Smart, P.L., Atkinson, T.C, Laidlaw, I.M.S. *et al.* (1986) Comparison of the results of quantitative and non-quantitative tracer results for the determination of karst conduit networks: an example from the Traligill Basin, Scotland. *Earth Surface Process and Landforms*, **11**, 249-61.

Smart, P.L and Christopher, N.S.J. (1989) Ogof Ffynnon Ddu. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 177-89.

Smart, P.L. and Frances, P.D. (1991) *Quaternary Dating Methods - A User's Guide*, Quaternary Research Association Technical guide, **4**, Cambridge.

Smart, P.L. and Gardener, C.G. (1989) The Mynydd Llangattwg cave systems. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 124-51.

Smart, P.L. and Hodge, P.G. (1979) A pulse-wave test at Charterhouse. *Journal of the Wessex Cave Club,* **15** (176), 132-6.

Smart, P.L. and Hodge, P. (1980) Determination of the character of the Longwood sinks to Cheddar resurgence conduit using an artificial pulse wave. *Transactions of the British Cave*

Research Association, **8**, 208-11.

Smart, P.L., Moody, P.D., Moody, A.A.D. and Chapman, P.R.J. (1984) Charterhouse Cave: exploration, geomorphology and fauna. *Proceedings of the University of Bristol Speleological Society*, **17**, 5-27.

Smart, P.L., Palmer, R.J., Whitaker, F.F. and Wright, V.P. (1988a) Neptunian dykes and fissure fills: an overview and account of some modern examples. In *Paleokarst* (eds N. P. James and P. W. Choquette), Springer Verlag, 149-63.

Smart, P.L., Smith, B.W., Chandra, H., Andrews, J.N. *et al.* (1988b) An intercomparison of E.S.R. and uranium series ages for Quaternary speleothem calcite. *Quaternary Science Reviews*, **7**, 411-16.

Smart, P.L. and Stanton, W.I. (1974) Manor Farm Swallet, Charterhouse-upon-Mendip: an account and geomorphology. *Proceedings of the University of Bristol Speleological Society*, **13**, 391-402.

Smart, P.L., Waltham, A.C., Yang, M. and Zhang, Y. (1986) Karst geomorphology of western Guizhou, China. *Transactions of the British Cave Research Association*, **13**, 89-103.

Smith, D.B. (1972) Foundered strata, collapse breccias and subsidence features of the English Zechstein. In *Geology of Saline Deposits* (ed. G. Richter-Bernburg), UNESCO Earth Sciences, **7**, 255-69.

Smith, D.I. and Mead, D.G. (1962) The solution of limestone, with special reference to Mendip. *Proceedings of the University of Bristol Speleological Society*, **9**, 188-211.

Smith, D.I. (1972) The solution of limestone in an Arctic environment. In *Institute of British Geographers Special Publication*, **4**, 187-200.

Smith, D.I. (ed.) (1975a) *Limestones and Caves of the Mendip Hills,* David and Charles, Newton Abbott.

Smith, D.I. (1975b) The problems of limestone dry valleys - implications of recent work in limestone hydrology. In *Processes in Physical and Human Geography* (eds R. Peel, M. Chisholm and P. Haggett), Heinemann, London, 130-47.

Smith, D.I. (1977) Limestone features and the geomorphological evolution of the Mendip Hills. In *Geological Excursions in the Bristol District* (ed. R.J.G. Savage), University of Bristol, Bristol, 65-72.

Smith, D.I. and Atkinson, T.C. (1977) Underground flow in cavernous areas with special reference to the Malham area. *Field Studies*, **4**, 597-616.

Smith, D.I. and Newson, M.D. (1974) The dynamics of solutional and mechanical erosion in limestone catchments on the Mendip Hills, Somerset. *Institute British Geographers Special Publication*, **6**, 155-67.

Smith, P.B. and Waltham, A.C. (1973) The P8 cave system. *Journal of the British Speleological Association*, **50**, 21-8.

Solari, R.A. (1974) Hydrology of the Slaughter Rising. *Cave Projects Group Newsletter,* **5**, 54-65.

Sparks, B.W. and Lewis, W.V. (1957) Escarpment dry valleys near Pegsdon, Hertfordshire. *Proceedings of the Geologists' Association*, **68**, 26-38.

Sparks, B.W., Williams, R.B.G. and Bell, F.G. (1972) Presumed ground-ice depressions in East Anglia. *Proceedings of the Royal Society,* **A327**, 329-43.

Spencer, H.E.P. and Melville, R.V. (1974) The Pleistocene mammalian fauna of Dove Holes, Derbyshire. *Bulletin of the Geological Survey of Great Britain*, **48**, 43-53.

Sperling, C.H.B., Goudie, A.S., Stoddart, D.R. and Poole, G.G. (1977) Dolines of the Dorset Chalklands and other areas in southern Britain. *Transactions of the Institute of British Geographers*, **2**, 205-23.

Standing, I.J. (1967) Speleology in Gloucestershire. *Cave Research Group Newsletter,* **108**, 4-6.

Standing, P.A. and Lloyd, O.C. (1970) Porth yr Ogof. *Proceedings of the University of Bristol Speleological Society*, **12**, 213-29.

Standing, P.A., Newson, M.D. and Wilkins, A.G (1971) Second report on the Little Neath River Cave. *Proceedings of the University of Bristol Speleological Society*, **12**, 303-25.

Stanton, W.I. (1965) The digging at the end of Gough's Cave and its bearing on the chances at Cheddar. *Journal of the Wessex Cave Club*, **8**, 277-83.

Stanton, W.I. (1972) Rhino Rift: survey notes and divers observations. *Journal of the Wessex Cave Club*, **12**, 48-50.

Stanton, W.I. (1977) A view of the hills. In *Mendip: The Complete Caves and a View of the Hills* (eds N. Barrington and W.I Stanton), Cheddar Valley Press, Cheddar.

Stanton, W.I. (1983) Digging for Mendip caves. *Studies in Speleology*, **4**, 77-83.

Stanton, W.I. (1985) Cheddar Gorge and Gough's Cave. *Proceedings of the University of Bristol Speleological Society*, **17**, 121-8.

Stanton, W.I. (1987) Waterwheel Swallet, Charterhouse-on-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **18**, 3-19.

Stanton, W.I. (1991) The habitat and origin of lead ore in Grebe Swallet Mine, Charterhouse-on-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **19**, 43-65.

Stanton, W.I. and Smart, P.L. (1981) Repeated dye traces of underground streams in the Mendip Hills, Somerset. *Proceedings of the University of Bristol Speleological Society*, **16**, 47-58.

Stenner, R.D. (1968) Water tracing in St Cuthbert's Swallet, Priddy, Somerset. *Transactions of the Cave Research Group of Great Britain*, **10**, 49-60.

Stenner, R.D. (1973) A study of the hydrology of GB Cave, Charterhouse-upon-Mendip, Somerset. *Proceedings of the University of Bristol Speleological Society*, **13**, 171-226.

Stenner, R.D. (1978) The concentration of cadmium, copper, lead and zinc in sediments from some caves and associated surface streams on Mendip, Somerset. *Transactions of the British Cave Research Association*, **5**, 113-20.

Stevens, J. (1992) *An Exploration Journal of Llangattwg Mountain*, Chelsea Speleological Society Records, **19**.

Stevenson, R. and Palmer, R. (1986) Cheddar River Cave. *Caves and Caving*, **33**, 16-18.

Stevenson, W. (1812) *General View of the Agriculture of the County of Dorset,* London.

Stratford, T. (1986) *Caves of South Wales,* Cordee, Leicester.

Stride, A.H. and Stride, R.D. (1949) The formation of the Mendip caves. *British Caver*, **19**, 6-25.

Stride, R.D. and Stride, A.H. (1946) Longwood Swallet, Charterhouse-on-Mendip. *Proceedings of the University of Bristol Speleological Society*, **5**, 183-7.

Stride, R.D. and Stride, A.H. (1949) August Hole. *Proceedings of the University of Bristol Speleological Society*, **6**, 14-22.

Stringer, C.B. (1977) Evidence of climatic change and human occupation during the last interglacial at Bacon Hole Cave, Gower. *Journal of Gower Society*, **28**, 36-44.

Stringer, C.B., Currant, A.P., Schwarz, H.P. and Collcutt, S.N. (1986) Age of Pleistocene faunas from Bacon Hole, Wales. *Nature*, **320**, 59-62.

Stuart, A.J. (1983) Pleistocene bone caves in Britain and Ireland. *Studies in Speleology*, **4**, 9-36.

Sutcliffe, A.J. (1960) Joint Mitnor Cave, Buckfastleigh. *Transactions of the Torquay Natural History Society*, **13**, 3-28.

Sutcliffe, A.J., (1981) Progress report on excavations in Minchin Hole, Gower. *Quaternary Newsletter*, **3**, 1-17.

Sutcliffe, A.J., Lord, T.C., Harmon, R.S. *et al.* (1985) Wolverine in northern England at about 83 000 years B.P.: faunal evidence for climatic change during isotope stage 5. *Quaternary Research*, **24**, 73-86.

Sutcliffe, A.J. and Zeuner, F.E. (1962) Excavations in the Torbryan Caves, Devon, 1. Tornewton Cave. *Proceedings of the Devon Archaeological Exploration Society*, **5**, 127-45.

Sutcliffe, J.R. (1974) The caves of Barbondale and the Dent Fault zone. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 235-49.

Sutcliffe, J.R. (1985) Knock Fell Caverns. *Gritstone Club Journal*, (1985), 70-4.

Sweeting, G.S. and Sweeting, M.M. (1969) Some aspects of the Carboniferous Limestone in relation to its landforms. *Mediterranée*, **7**, 201-9.

Sweeting, M.M. (1950) Erosion cycles and limestone caverns in the Ingleborough district of Yorkshire. *Geographical Journal*, **115**, 63-78.

Sweeting, M.M. (1958) The karstlands of Jamaica. *Geographical Journal*, **124**, 184-99.

Sweeting, M.M. (1966) The weathering of limestones. In *Essays in Geomorphology* (ed. G.H. Dury), Heinemann, 177-210.

Sweeting, M.M. (1972) *Karst Landforms*, Macmillan, London.

Sweeting, M.M. (1974) Karst geomorphology. In *Limestones and Caves of North-west England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 46-78.

Swinnerton, A.C. (1932) Origin of limestone caverns. *Bulletin of the Geological Society of America*, **43**, 662-93.

Tate, T. (1879) The source of the River Aire. *Proceedings of the Yorkshire Geological and Polytechnic Society*, **7**, 177-86.

Tattersall, W.M . and Coward, T.A. (1914) Faunal survey of Rostherne Mere. I. Introduction and methods. *Memoirs of the Proceedings of the Manchester Literary and Philosophical Society*, **58**, 1-21.

Taviner, R. (1993) Traligill - a reappraisal of speleological potential. *Grampian Speleological Group Bulletin Series* 3, **2** (5), 30-9.

Taylor, D.M., Griffiths, H.I., Pedley, H.M. and Prince, I. (1994) Radiocarbon dated Holocene pollen and ostracod sequences from barrage tufa-dammed fluvial systems in the White Peak, Derbyshire, UK. *The Holocene*, **4**, 356-64.

Taylor, P. (1993) Red House Swallet. *Caves and Caving*, **60**, 28.

Thistlewood, L. and Noel, M. (1991) A palaeomagnetic study of sediments from maypole Inlet, Peak Cavern. *Transactions of the British Cave Research Association*, **17**, 55-8.

Thomas, T.M. (1954) Swallow holes on the Millstone Grit and the Carboniferous Limestone of the South Wales coalfield. *Geographical Journal*, **120**, 468-75.

Thomas, T.M. (1963) Solution subsidence in south-east Carmarthenshire and south-west Breconshire. *Transactions of the Institute of British Geographers*, **33**, 45-60.

Thomas, T.M. (1970) The limestone pavements of the north crop of the South Wales coalfield. *Transactions of the Institute of British Geographers*, **50**, 87-104.

Thomas, T.M. (1973) Solution subsidence mechanisms and end products in south-east Breconshire. *Transactions of the Institute of British Geographers*, **60**, 69-86.

Thomas, T.M. (1974) The South Wales interstratal karst. *Transactions of the British Cave Research Association*, **1**, 131-52.

Thorez, J., Bullock, P., Catt, J.A. and Weir, A.H. (1971) The petrography and origin of deposits filling solution pipes in the chalk near South Mimms, Hertfordshire. *Geological Magazine*, **108**, 413-23.

Thrailkill, J. (1968) Chemical and hydrological factors in the excavation of limestone caves. *Bulletin of the Geological Society of America*, **79**, 19-46.

Tiddeman, R.H. (1872) On the evidence for the ice sheet in north Lancashire and adjacent parts of Yorkshire and Westmorland. *Proceedings of the Geologists' Association*, **28**, 471-91.

Tomalin, S. (1987) Extensions in Gothic Passage - Agen Allwedd. *Caves and Caving*, **36**, 29-32.

Towler, P.A. (1977) 'A geological survey and an investigation into the relation with the present environment of tufa formation in Lathkill Dale.' PhD thesis, University of Bristol.

Tratman, E.K. (1963) The hydrology of the Burrington area, Somerset. *Proceedings of the University of Bristol Speleological Society*, **10**, 22-57.

Tratman, E.K. (1975) The cave archaeology and palaeontology of Mendip. In *Limestones and Caves of the Mendip Hills* (ed. D.I. Smith), David and Charles, Newton Abbot, 352-91.

Tratman, E.K., Donovan, D.T. and Campbell, J.B. (1971) The Hyaena Den (Wookey Hole), Mendip Hills, Somerset. *Proceedings of the University of Bristol Speleological Society*, **12**, 245-79.

Trotter, F.M. (1929) The Tertiary uplift and resultant drainage of the Alston Block and adjacent area. *Proceedings of the Yorkshire Geological Society*, **21**, 161-80.

Trotter, F.M. (1942) *Geology of the Forest of Dean Coal and Iron-ore Field*. Geological Survey of Great Britain Memoir.

Trudgill, S.T. (1977) The making of Cheddar Gorge. *Geographical Magazine*, **50**, 196-9.

Trudgill, S.T. (1985a) *Limestone Geomorphology*, Longman, Harlow.

Trudgill, S.T. (1985b) Field observations of limestone weathering and erosion in the Malham district, North Yorkshire. *Field Studies*, **6**, 201-36.

Underhill, J.R., Gayer, R.A., Woodcock, N.H., Donnelly, R. *et al.* (1988) The Dent Fault System, northern England - reinterpreted as a major oblique-slip fault zone. *Journal of the Geological Society, London*, **145**, 303-16.

Vincent, P.J. (1981) Some observations on the so-called relict karst of the Morecambe Bay region, Northwest England. *Revue de Géologie Dynamique et de Géographie Physique*, **23**, 143-50.

Vincent, P.J. (1982) Snow patches on Farleton Fell, southeast Cumbria, *Geographical Journal*, **148**, 337-42.

Vincent, P.J. (1983) The dissolving landscape: step karren. *Geographical Magazine*, **55**, 508-10.

Vincent, P J. (1995) Limestone pavements in the British Isles: a review. *Geographical Journal*, **161**, 265-74.

Vincent, P.J. and Lee, M.P. (1981) Some observations on the loess around Morecambe Bay, North-West England. *Proceedings of the Yorkshire Geological Society*, **43**, 281-94.

Vowler, F. (1980) The caves of Devon. *Caving International Magazine*, **6/7**, 58-62.

Vowler, F. (1981) The caves of north Devon. *William Pengelly Cave Studies Trust Newsletter*, **38**, 1-7.

Walkden, G. and Davies, J. (1983) Polyphase erosion of subaerial omission surfaces in the late

Dinantian of Anglesey, North Wales. *Sedimentology*, **30**, 861-78.

Wallwork, K.L. (1956) Subsidence in the Mid-Cheshire industrial area. *Geographical Journal*, **122**, 40-53.

Wallwork, K.L. (1960) Some problems of subsidence and land use in the Mid-Cheshire industrial area. *Geographical Journal*, **126**, 191-9.

Walsh, P.T., Boulter, M.C, Ijtaba, M. and Urbani, D.M. (1972) The preservation of the Neogene Brassington Formation of the Southern Pennines and its bearing on the evolution of Upland Britain. *Journal of the Geological Society of London*, **128**, 519-59.

Walsh, P.T. and Brown, E.H. (1971) Solution subsidence outliers containing probable Tertiary sediment in northeast Wales. *Geological Journal*, **7**, 299-320.

Walsh, P.T., Collins, P., Ijtaba, M. *et al.* (1980) Palaeocurrent directions and their bearing on the origin of the Brassington Formation (Miocene-Pliocene) of the southern Pennines, Derbyshire, England. *Mercian Geologist*, **8**, 1.

Walsh, P.T. and Ockenden, A.C. (1982) Hydrogeological observations at the Water End swallow hole complex, North Mimms, Hertfordshire. *Transactions of the British Cave Research Association*, **9**, 184-94.

Waltham, A.C. (1969) Swallets and caves in Chalk. *Journal of London University Caving Clubs*, **9**, 3-5.

Waltham, A.C. (1970) Cave development in the limestone of the Ingleborough district. *Geographical Journal*, **136**, 574-85.

Waltham, A.C. (1971a) Controlling factors in the development of limestone caves. *Transactions of the Cave Research Group of Great Britain*, **13**, 73-80.

Waltham, A.C. (1971b) Shale units in the Great Scar Limestone of the southern Askrigg Block. *Proceedings of the Yorkshire Geological Society*, **38**, 285-92.

Waltham, A.C. (ed.) (1974a) *The Limestones and Caves of North-West England,* David and Charles, Newton Abbot.

Waltham, A.C. (1974b) The geology of the southern Askrigg Block. In *Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 25-45.

Waltham, A.C. (1974c) The geomorphology of the caves of North-west England. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 79-105.

Waltham, A.C. (1974d) Speleogenesis of the caves of Leck Fell. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 273-309.

Waltham, A.C. (1974e) The caves of Ribblesdale. In *The Limestones and Caves of North-West England* (ed. A.C. Waltham), David and Charles, Newton Abbot, 385-92.

Waltham, A.C. (1977a) Cave development at the base of the limestone in Yorkshire. *Proceedings of the 7th International Speleological Congress*, British Cave Research Association, Bridgwater, 421-3.

Waltham, A.C. (1977b) White Scar Cave, Ingleton. *Transactions of the British Cave Research Association*, **4**, 345-53.

Waltham, A.C. (1981) Leck Fell and the Three Counties System. *Transactions of the British Cave Research Association*, **6**, 29-32.

Waltham, A.C. (1984) *Caves, Crags and Gorges*, Constable, London.

Waltham, A.C. (1986) Valley excavation in the Yorkshire Dales karst. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 541-50.

Waltham, A.C. (1989) *Ground Subsidence*, Blackie, London.

Waltham, A.C. (1990) Geomorphic evolution of the Ingleborough karst. *Transactions of the British Cave Research Association*, **17**, 9-18.

Waltham, A.C. and Brook, D.B. (1980a) The Three Counties cave systems. *Transactions of the British cave research Association*, **7**, 121.

Waltham, A.C. and Brook, D.B. (1980b) Geomorphological observations in the limestone caves of the Gunung Mulu National Park, Sarawak. *Transactions of the British Cave Research Association*, **7**, 123-39.

Waltham, A.C., Brook, D.B., Statham, O.W. and Yeadon, T.G. (1981) Swinsto Hole, Kingsdale: a type example of cave development in the limestone of northern England. *Geographical Journal*, **147**, 350-3.

Waltham, A.C. and Davies, M. (1987) *Caves and Karst of the Yorkshire Dales,* British Cave Research Association Cave Studies Series, **1**.

Waltham, A.C. and Everett, D.G. (1989) The caves of the Mellte and Hepste valleys area. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 155-64.

Waltham, A.C. and Harmon, R.S. (1977) Chronology of cave development in the

Yorkshire Dales, England. *Proceedings of the 7th International Speleological Congress*, British Cave Research Association, Bridgwater, 423-5.

Waltham, A.C. and Hatherley, P. (1983) The caves of Leck Fell. *Transactions of the British Cave Research Association*, **10**, 245-7.

Waltham, A.C. and Tillotson, A.C. (1989) *The Geomorphology of Ingleborough,* Nature Conservancy Council, Peterborough, **1**.

Ward, S.D. and Evans, D.F. (1975) 'A botanical survey and conservation assessment of British limestone pavements. Unpublished report, Institute of Terrestrial Ecology, Volumes 1-9.

Ward, S.D. and Evans, D.F. (1976) Conservation assessment of British limestone pavements based on floristic criteria. *Biological Conservation*, **9**, 217-33.

Warwick, G.T. (1953) 'The geomorphology of the Dove-Manifold region.' PhD thesis, University of Birmingham.

Warwick, G.T. (1962) British caving regions. In *British Caving* (ed. C.H.D. Cullingford), Routledge and Kegan Paul, London, 120-217.

Warwick, G.T. (1964) Dry valleys of the Southern Pennines. *Erkunde*, **18**, 116-23.

Warwick, G.T. (1968) Some primitive features in British caves. *Proceedings of the 4th International Congress on Speleology,* Speleological Society of Yugoslavia, Ljubljana, **3**, 239-52.

Waters, R.S and Johnson, R.H. (1958) The terraces of the Derbyshire Derwent. *East Midlands Geographer*, **2**, 3-15.

Webb, S. (1995) Conservation of limestone pavement. *Transactions of the British Cave Research Association*, **21**, 97-100.

Webster, R. (1969) 'The Romano-British settlements of Westmorland: a study in cultural ecology.' PhD thesis, University of Reading.

Welch, F.B.A. and Trotter, F.M. (1960) *Geology of the Country around Monmouth and Chepstow,* Geological Survey of Great Britain Memoir, Sheets 233 and 250.

Westlake, C.D. (1967) Giant's Hole and Oxlow Cavern: some notes on the development and recent exploration of the system. *Proceedings of the British Speleological Association*, 1-12.

Westlake, C.D., Elliott, J.V. and Tringham, M.E. (1989) Otter Hole, Chepstow. In *Limestones and Caves of Wales* (ed. T.D. Ford), Cambridge University Press, Cambridge, 117-23.

Whitaker, F.F. and Smart, P.L. (1993) Circulation of saline ground water in carbonate platforms – a review and case study from the Bahamas. In *Diagenesis and Basin Development* (eds A.D. Horbury and A.G. Robinson), American Association of Petroleum Geologists Studies in Geology, **36**, 113-32.

White, W.B. (1976) Cave minerals and speleothems. In *The Science of Speleology* (eds T.D. Ford and C.H.D. Cullingford), Academic, 267-327.

White, W.B. (1988) *Geomorphology and Hydrology of Karst Terrains,* Oxford University Press, Oxford.

White, W.B. and Longyear, J. (1962) Some limitations on speleo-genetic speculation imposed by the hydraulics of groundwater flow in limestone. *Nitteny Grotto Newsletter (National Speleological Society)*, **10** (9), 155-67.

Whybro, P. (1987) Ogof Llyn Du. *Caves and Caving*, **36**, 35-6.

Williams, N.J. and Farrant, A.R. (1992) Drunkard's Hole. *Proceedings of the University of Bristol Speleological Society*, **19**, 265-72.

Williams, P.W. (1966) Limestone pavements with special reference to western Ireland. *Transactions of the Institute of British Geographers*, **40**, 155-72.

Williams, P.W. (1983) The role of the subcutaneous zone in karst hydrology. *Journal of Hydrology*, **61**, 45-67.

Williams, P.W. (1985) Subcutaneous hydrology and the development of doline and cockpit karst. *Zeitschrift für Geomorphologie*, **29**, 463-82.

Wilson, A.A. (1974) Developments in limestone geology in the Ingleton-Settle area. *Transactions of the British Cave Research Association*, **1**, 61-4.

Wilson, A.A. (1983) *Masham; geological map Sheet 51,* British Geological Survey.

Wilson, P. (1979) Surface features of quartz sand grains from the Brassington Formation. *Mercian Geologist*, **7**, 19-30.

Wilson, W., Welch, F.B.A., Robbie, J.A. and Green, G.W. (1958) *The Geology of the Country around Bridport and Yeovil,* British Geological Survey Memoir, Sheet 327.

Winwood, H.H. and Woodward, H.B. (1891) Excursion to the Mendip Hills. *Proceedings of the Geologists' Association*, **11**, 151-216.

Wooldridge, S.W. (1929) Discussion (of Sherlock, 1929). *Proceedings of the Geologists' Association*, **40**, 372.

Wooldridge, S.W. and Kirkaldy, J.F. (1937). The geology of the Mimms valley. *Proceedings of* *the Geologists' Association* **48,** 307-15.

Worley, N.E. and Nash, D.A. (1977) The geological evolution of the Jugholes Caves, Matlock, Derbyshire. *Transactions of the British Cave Research Association*, 4, 389-401.

Worthington, S.R.H. and Ford, D.C. (1995) High sulfate concentrations in limestone springs: an important factor in conduit initiation? *Environmental Geology*, **25**, 9-15.

Wright, B. (1987) The Maypole Inlet dig, Peak Cavern. *Caves and Caving*, **38**, 9-11.

Wright, V.P. (1982) The recognition and interpretation of paleokarsts: two examples from the Lower Carboniferous of South Wales. *Journal of Sedimentary Petrology*, **52**, 83-94.

Wright, V.P. (1986a) Facies sequences on a carbonate ramp: the Carboniferous Limestone of South Wales. *Sedimentology*, **33**, 221-41.

Wright, V.P. (1986b) The polyphase karstification of the Carboniferous Limestone in South Wales. In *New Directions in Karst* (eds K. Paterson and M.M. Sweeting), Geo Books, Norwich, 569-80.

Yates, H. (1934) Goyden Pot, Nidderdale. *Yorkshire Ramblers Club Journal*, **6**, 216-28.

Yeadon, G. (1975) The new Boreham Cave extensions. *Kendal Caving Club Journal*, **8**, 14-18.

Yeadon, T.G. (1985) Tate Galleries, Lost John's System, Leck Fell. *Caves and Caving*, **28**, 2-4.

Young, I. (1992) The highest cave in the British Isles. *Descent*, **103**, 32-3.

Younger, P.L. and Elliot, T. (1995) Chalk fracture system characteristics: implications for flow and solute transport. *Quarterly Journal of Engineering Geology*, **28**, S39-S50.

Zhang, Z. (1980) Karst types in China. *GeoJournal*, 4, 541-70.

Glossary

This glossary provides simple explanations of the more important technical and arcane terms used in the Introductions to the chapters and in the Highlights and Conclusions of Chapters 2 to 8. These explanations do not pretend to be scientific definitions but are intended to help the general reader. Stratigraphical terms are omitted as they are given context within the tables and figures.

Absolute dating: method of determining the date of formation of a rock or mineral; includes **radiometric dating**, **electron spin resonance**, **thermoluminescence** and **paleomagnetic** techniques.

Adit: horizontal tunnel, for access or drainage, mined into a hillside.

Allogenic: derived from another source region; in **karst**, most commonly describing drainage flows derived from surface runoff that originates on adjacent non-karstic rocks.

Aquiclude: impermeable rock capable of absorbing water but generally acting as the boundary to an **aquifer**.

Aquifer: body of rock permeable enough to transmit significant flows of groundwater.

Anticline: upfold of rock which rises in the middle, with older rock in its core.

Aragonite: natural mineral formed of calcium carbonate ($CaCO_3$); it has the same composition as the more common calcite, but differs in its atomic structure and crystal shape.

Artesian aquifer: an aquifer, confined by an overlying **aquiclude** within a **syncline** or comparable geological structure, so that its groundwater has enough hydrostatic pressure to rise to the surface through a fissure or a well.

Autogenic drainage: drainage water derived entirely from rainfall directly onto the rock outcrop.

Barrage tufa: natural dam of calcareous deposits which holds back stream water in a pool.

Bedding plane: planar feature in sedimentary rocks, created by an interruption or change within the original sediment deposition; commonly a site of cave inception and development.

Bedding plane cave: cave passage formed by dissolutional enlargement of a **bedding plane** in a limestone; in horizontal or gently dipping limestone it is therefore a very low and wide passage. The term may also refer to a cave initially formed on a bedding plane, but subsequently enlarged into a tube or entrenched into a canyon.

Bioclastic: consisting largely of broken shell debris.

Bourne: seasonal surface stream in a normally dry valley, resulting from winter rise in the water table, typically in the **chalk karst**.

Calcite: the common and stable mineral form of calcium carbonate ($CaCO_3$), which is the dominant component of limestone and secondary cave deposits due to its dissolution and reprecipitation by natural waters at normal temperatures.

Carbonic acid: weak natural acid formed by the absorption of atmospheric carbon dioxide into rainwater and soilwater, capable of the reversible reaction of dissolving or precipitating calcium carbonate.

Carboniferous: any rock formed within the eponymous period of geological time about

360–290 million years ago.

Cave: best defined as 'a natural cavity large enough to be entered by a person'. Most caves are formed by dissolution of limestone within karst landscapes, but others include glacier, sea, lava and **tectonic caves**.

Cave pearl: roughly spherical concretion of calcium carbonate, usually calcite, which grows by concentric accretion in cave pools saturated in lime and disturbed by dripping water.

Cavernous karst: limestone landscape that includes both surface karst features and also extensive cave systems.

Chalk: weak, friable and porous, poorly lithified, white limestone; Britain's chalk is a specific rock unit of late Cretaceous age.

Chalk karst: distinctive type of **karst** landscape formed on **chalk**; distinguished by systems of dry valleys, rounded hillsides with few rock scars and efficient underground drainage with few cave-sized conduits.

Chert: hard, microcrystalline silica mineral (SiO_2) that commonly occurs as nodules or bands within limestone.

Chinastone: very fine-grained calcite mudstone.

Choke: blockage in a cave passage formed either by inwashed sediment or breakdown debris from roof collapse; the latter commonly forms a boulder choke.

Clint: block of limestone within a **limestone pavement,** bounded by **grykes**.

Coccolith: calcareous skeleton of a type of microscopic marine alga, made of calcite plates, and forming a major component of **chalk**.

Collapse stoping: progressive breakdown and upward collapse of a cave ceiling, which causes the open cavity to migrate, or stope, to higher levels.

Combe: dry valley in **chalk karst**, excavated by meltwater river and modified by **solifluction** during past phases of **periglacial** conditions.

Combe rock: surface layer of periglacial **head** formed largely of chalk, mostly found on floors of valleys or **combes** in chalk karst.

Conduit: dissolutional void, generally greater than 100 mm in diameter, in limestone; larger than a fissure, and including cave passages.

Confined aquifer: aquifer containing water under pressure because it is confined by an overlying **aquiclude**.

Connate water: fossil water that has been trapped within the sediment since the time of deposition.

Cryoturbation: disturbance by the growth of ice crystals, including the expansion of frost wedges within soils in **periglacial** environments.

Dissolution: natural process of dissolving a solid; specifically in karst processes, the dissolving of carbonate rock to create a liquid solution of calcium and bicarbonate ions in water; also known as **solution**.

Doline: roughly circular, closed depression in the ground surface, with no drainage outlet except underground through a central **sinkhole**; may be regarded as the diagnostic landform of karst topography.

Dolomite: carbonate mineral very similar to **calcite**, except that it contains magnesium within its composition $CaMg(CO_3)_2$; also a sedimentary rock very similar to limestone, except that it contains more of the mineral dolomite than calcite.

Drumlin: low, rounded hill of glacial till, which was moulded into a streamlined shape by glacier ice passing over it.

Dry valley: fluvial valley, cut by a subaerial stream or river, then abandoned and left dry when underground drainage developed in a limestone **karst** region.

Electron spin resonance (ESR): a measure of the exposure of a calcite speleothem to environmental radiation, indicative of its age.

Epikarst: shallow zone immediately below the ground surface in a **karst** landscape, generally including the soil and the buried rock surface where much of the limestone dissolution occurs.

Estavelle: cave that acts as either a sinkhole or a spring in response to fluctuations in the level of the water table.

Evaporite: mineral or rock precipitated from water due to evaporation, most importantly **gypsum** and **salt**.

Exsurgence: natural spring fed entirely by percolation water, as opposed to a **resurgence** which is fed largely by sinking streams.

Fault: fracture within rock along which there has been significant displacement due to tectonic deformation.

Fissure: natural opening in rock, nominally 10–100 mm wide; therefore wider than a fracture, but smaller than a **conduit** or **cave**.

Flash: flooded depression formed by subsidence in **salt karst**.

Flint: mineral variety of **chert** which occurs as nodules and bands in **chalk**.

Flowstone: mineral deposited by flowing water on the walls or floor of a cave; mostly formed of **calcite**.

Fluviokarst: limestone landscape characterized by dendritic systems of **dry valleys**.

Fossil karst: karst landforms created by erosion in past eras of geological time, and now preserved within the rock sequence; better known as **paleokarst**.

Glaciokarst: karst landscape developed on limestone outcrops which have been scoured by glaciers; characterized by bare rock scars and limestone **pavements**.

Gour: barrier of calcite **flowstone** which dams a pool within a cave passage, and continues to grow by deposition from the overflowing water.

Greywacke: strong, old sandstone, with a significant clay content due to its turbidite origins, which has been metamorphosed to slate grade.

Gryke: fissure between **clints** within a limestone **pavement**, formed by dissolutional enlargement of a joint; Yorkshire name for **kluftkarren**.

Gypsum: white or colourless mineral of composition $CaSO_4.2H_2O$; may form massive gypsum rock as an **evaporite** deposit; or forms crystals in caves by reaction between limestone and sulphuric acid derived from the oxidation of pyrite.

Head: surface layer of soil and rock debris, developed by frost shattering and moved downslope by **solifluction** largely under periglacial conditions.

Helictite: small **stalactitic** form of calcite, with twisted and contorted form, due to deposition from seepage water supplied by capillary action independent of gravity flow.

Hydraulic gradient: slope of the water table within an **aquifer**.

Influent cave: cave with surface stream flowing into it.

Inlier: outcrop of old rocks surrounded by younger rocks, commonly formed by erosion of deep valleys.

Interstadial: short phase of warm climatic conditions within a longer cold stage of glaciation.

Interstratal karst: caves and karst features formed by underground dissolution of a rock unit covered by insoluble rocks which occur at outcrop.

Interfluve: upland between adjacent valleys.

Joint: fracture produced in rocks by tectonic deformation.

Kamenitza: dissolution basin, generally less than a metre across, formed on a limestone outcrop; self-deepening due to solution by regularly recharged rainwater.

Karren: small dissolution features formed on limestone outcrops and on limestone surfaces beneath a soil cover; dominated by channels or runnels, mostly 10–500 mm deep, which are entrenched to leave sharp or rounded intervening ridges; a German term now used throughout international literature.

Karst: distinctive terrain created by erosion of a soluble rock, where the topography and landforms are a consequence of efficient underground drainage; characterized by **caves**, **dolines**, **sinkholes** and **dry valleys**, and mainly developed on limestone.

Klippe: eroded remnant outlier of a unit of rock lying above a thrust plane.

Kluftkarren: deep open fissure formed by dissolutional enlargement of a joint or fault within limestone.

Limestone: sedimentary rock composed largely of calcium carbonate ($CaCO_3$) in the form of the mineral **calcite**, and therefore soluble in weak acids including rain and soil water; strong, well lithified limestones may stand in high vertical cliffs and can span large **cave** passages formed within them by dissolutional enlargement of fractures.

Knickpoint: break in slope of the long profile of a river, produced by rejuvenation or headward erosion and commonly marked by a waterfall or rapids.

Loess: fine-grained sediment of windblown silt and clay, largely derived from cold **periglacial** deserts.

Magnetic reversal: periodic reversal of the Earth's dipole magnetic field, which can be recognized by the orientation of ferric mineral particles in old sediments and rocks.

Maze cave: cave network of intersecting joint fissures or interconnecting and contemporaneous passage loops.

Mere: shallow lake, notably formed by flooding of a subsidence depression in the **salt karst** of Cheshire.

Metamorphism: process of change of the mineralogy and structure of rocks which are subjected to increases in pressure and temperature within the Earth's crust.

Meteoric water: water that originated from atmospheric rain or snow fall.

Micrite: fine-grained, chemically precipitated calcareous mud, which forms the matrix of many limestones.

Monocline: fold of rock which has one steeply dipping limb, with nearly horizontal rocks on both sides.

Moulin: hollow or pothole formed by scour in the bed of a stream.

Neptunian dyke: fissure or cavity infilled with younger sediment, commonly a feature of **paleokarst**.

Network cave: style of **maze cave** formed by the dissolutional enlargement of multiple intersecting joint systems.

Nunatak: mountain peak that projects above surrounding ice sheets and is subjected to intense frost action but is not scoured by glacial erosion.

Oolite: limestone made of rounded calcium carbonate grains (ooliths), produced by concretionary deposition in warm and shallow seas with high evaporation rates.

Outlier: outcrop of young rocks surrounded by older rocks, commonly left as an erosional remnant on a hill top.

Oxygen isotope stage: time subdivision of the Pleistocene based on the ratios of oxygen isotopes in marine sediments, which largely reflect changes in the ocean temperatures and world climates.

Paleokarst: karst landforms created by erosion in past eras of geological time, and now preserved within the rock sequence.

Paleomagnetism: remnant magnetism preserved in some ferric minerals as a feature of the Earth's magnetic polarity and field orientation at the time of the host rock or sediment formation, and changes in these provide a basis for stratigraphic dating.

Paleosol: fossil soil, buried and preserved by younger deposits.

Paragenesis: enlargement of a cave passage due to dissolution and progressive raising of the ceiling while erosion of the floor is inhibited by a protective layer of sediment.

Pavement: limestone **bedding plane** surface scraped clean by glaciers and subsequently carved by dissolution into **clints** and **grykes**.

Perched conduit: cave passage above the regional water table.

Pericline: fold of rock which is curved unequally in two directions, so that erosion of it creates an elliptical outcrop.

Periglacial: zone or environment peripheral to glaciers, so that it is very cold but is not covered by ice sheets; it is characterized by the frozen ground known as **permafrost**.

Permafrost: permanently frozen ground within a **periglacial** environment; may extend to more than 100 m deep, but the active layer of the top few metres thaws each summer and then refreezes in winter.

Phreas: the saturated zone of the ground, below the **water table**, where all pore spaces, fissures and caves are filled with groundwater.

Phreatic cave: cave passage or cave system developed within the **phreas**, therefore characterized by vertical loops, reverse gradients and rounded passage shapes due to dissolution of the floor, walls and ceiling.

Phreatic lift: section of a phreatic passage that carries, or carried, water upwards in a downstream direction.

Phreatic tube: tubular cave passage formed by almost equal dissolution of the walls, ceiling and floor, while full of water within the **phreas**; abandoned tubes are common and may be filled with sediment.

Phreatic zone: synonym for **phreas**.

Pipe: cylindrical or conical dissolutional cavity, which is infilled and buried by younger sediments.

Piping failure: process whereby a cave is developed headwards in unconsolidated sediment as the fine particles are first removed by seepage water, allowing the larger particles to be washed out next, and ultimately leading to collapse of the void roof.

Polje: large karstic depression with a flat floor and sharp breaks of slope to its rock walls.

Polygonal karst: mature karst terrain where **dolines** have replaced valleys as the main landform, and have partially coalesced to leave a polygonal network of **interfluves**.

Pothole: single shaft, or entrance to a cave system, or an entire cave system that is dominantly vertical; also a **moulin** in a stream bed.

Proto-cave: natural void linking a potential water input point to an output within an **aquifer**, but which is still too small to be entered by a person.

Pseudokarst: landscape with caves, dolines and underground features, therefore resembling a true **karst**, but developed on rocks and soils by processes other than dissolution of rock.

Puddled: artificially lined with impermeable clay; sealing the floors of water channels.

Pyrite: widespread, naturally occurring iron sulphide mineral, FeS_2, which may be oxidized to form corrosive sulphuric acid.

Pyroclastic: consisting of fragments of volcanic material produced by explosive eruptions.

Radiometric dating: method of **absolute dating**, or determining the date of formation of a rock, based on the known decay rates of radioactive elements, such as carbon-14 or uranium-234, which naturally occur in measurable quantities in certain rocks.

Rake: local name for a mineral vein large enough to be mined in the Derbyshire Peak District.

Rejuvenation: initiation of a new cycle of erosion, normally as a consequence of lowered base level, climatic change or tectonic uplift.

Resurgence: natural spring fed partly by sinking streams, as opposed to an **exurgence** which is fed only by percolation water.

Rift: tall, narrow and generally straight cave passage, formed by dissolutional widening of a joint or fault.

Rillenkarren: small dissolution runnels, generally 10–20 mm wide and deep, cut into sloping, exposed limestone surfaces by subaerial dissolution, leaving small intervening ridges with sharp crests.

Rinnenkarren: dissolution runnels about 200 mm wide, with rounded troughs and sharp rims, cut into sloping rock surfaces by rainwater runoff.

Rising: point where underground water rises to the ground surface; synonym for **spring**.

Rundkarren: dissolution runnels, generally 100–400 mm wide and deep, with rounded ridges between rounded troughs, formed largely by dissolution beneath a soil cover.

Runnels: small channels formed by dissolution in **karren** landforms.

Salt: white or colourless mineral of composition NaCl; may form massive beds of rocksalt as an **evaporite** deposit; also known as halite.

Salt karst: karst landscape developed over beds of buried rocksalt, with subsidence features created by rapid dissolution of the salt.

Secondary minerals: minerals formed by chemical alteration, hydration or oxidation of older minerals.

Shakehole: local name in the northern Pennines for a **subsidence doline** formed in glacial **till** overlying limestone.

Sink: point where a stream sinks and disappears below ground; the water may filter through a sediment **choke** or flow into an open **cave**.

Sinkhole: synonym for either an active stream **sink** or a **doline**.

Solifluction: downslope movement of saturated sediment or soil debris, occurring most commonly in **periglacial** environments.

Solution: synonym for **dissolution**; also the product of the dissolution process where a solid component is dissolved within a liquid.

Sop: local name for a large **doline** in the Furness region, filled with sand, limestone rubble and hematite iron ore.

Sough: local name for a drainage **adit** cut by miners to lower the water table in the limestone of the Derbyshire Peak District.

Sparite: coarsely crystalline interstitial cement of limestones, generally comprised of calcite or aragonite, formed in the transformation of the sediment into a limestone.

Speleothem: general term for all cave mineral deposits, mostly formed of calcite by precipitation from lime-saturated groundwater.

Spitzkarren: residual limestone pinnacles or blades with sharp crests.

Spring: point where underground water emerges onto the ground surface from any aquifer; the largest springs are mostly the outlets from limestone caves.

Stadial: cold phase within the Pleistocene period.

Stalactite: calcite **speleothem** hanging from a cave roof, where it grows by deposition from lime-saturated percolation water entering the cave.

Stalagmite: calcite **speleothem** growing upwards from a cave floor, where it forms by deposition from drips of water still saturated with lime when they fall from the cave roof.

Stratimorph: landform such as a plateau whose shape (or morphology) was determined by erosional stripping of rocks to expose a single strong surface on a **bedding plane** (or stratum).

Straw stalactite: fragile, thin-walled tube of calcite, produced by the continuous precipitation of calcite around the rim of water drops before they fell away; the diameter of the tube is therefore that of a drop of water, about 5 mm.

Subsidence doline: surface depression or **doline** formed entirely in the soil or sediment cover above a limestone, where the cover material has been washed down into underlying limestone fissures by percolating water.

Suffosion: process of washing soil particles down into an underlying fissure; a type of **pip-**

ing failure; also the mechanism behind the development of **subsidence dolines**, which are sometimes called suffosion dolines.

Sump: section of flooded or submerged cave passage; may be a short perched sump within a **vadose cave**, or an entire active **phreatic cave** below the regional water table.

Swallet: local name for a **sinkhole** in the karst of the Mendip Hills.

Swallow hole: synonym for **sinkhole**.

Syncline: downfold of rock which sinks in the middle, with younger rock in its core.

Tectonic: produced by large-scale Earth movements.

Tectonic cave: fissure or rift cave, or other opening, produced by ground movement, notably at the head of a landslip block.

Thalweg: line down the floor of a valley; therefore the position occupied by a river unless the valley is dry.

Thermoluminescence: measure of the exposure of a calcite **speleothem** to environmental radiation, and therefore indicative of its age.

Thixotropic: characteristic of some sediments which transform from a solid or plastic state to behave as a liquid when disturbed.

Thrust: gently dipping or horizontal **fault**.

Till: unsorted, non-stratified sediment deposited directly by glacial ice; commonly known as boulder clay or glacial till.

Toadstone: local name for heavily weathered or altered, basaltic lava or volcanic tuff, occurring within the limestone sequence of the Derbyshire Peak District.

Tor: mass of rock rising above the surrounding landscape, commonly shaped by frost shattering or exfoliation in past **periglacial** conditions.

Travertine: calcareous mineral deposited by flowing water, commonly where plants or algae promote carbonate precipitation by extracting carbon dioxide from the water; normally refers to deposits formed outside caves, thereby excluding **speleothems**.

Trittkarren: small stepped bevels cut into a gently sloping limestone surface by dissolution.

Troglobite: animal which permanently dwells underground beyond the daylight zone of a cave.

Tropical karst: extreme development of relief in a karst landscape, creating residual hills shaped as cones or towers.

Tufa: soft, porous variety of **travertine**.

Turlough: karst depression that may be dry or flooded in response to seasonal fluctuations of the **water table**.

Unconformity: boundary separating two sedimentary rock units whose bedding structures are not parallel to each other, due to a period of folding and erosion between their times of formation.

Underfit stream: stream that is clearly smaller than the ancestral stream or river which excavated the cave or valley now host to the underfit.

Uniclinal shift: sideways shift of a stream channel, notably one which cuts down obliquely and downdip to follow a weak bed from its strike outcrop, leaving a stronger underlying bed to form the updip valley side.

Uranium-series dating: the method of **absolute dating** used most widely on cave material.

Vadose canyon: cave passage with roughly parallel, vertical sides formed by continued floor entrenchment or incision by a free-flowing **vadose** stream.

Vadose cave: cave passage or cave system developed by freely flowing streams above the **water table** and with air above any water surface, therefore characterized by continuous downslope profiles and canyon passages formed by erosion only of the floor.

Vadose shaft: roughly cylindrical, vertical section of cave above the water table, which contains or contained a waterfall and was largely excavated by dissolution by spraywater.

Vadose zone: zone of rock above the **water table** with groundwater freely flowing downwards and with cavities only partially filled with water; also referred to as the unsaturated zone.

Vauclusian rising: karst rising where water flows up a flooded cave passage, within the **phreas** and under pressure, to emerge at the surface.

Wadi: valley with intermittent stream flow in a semi-arid region.

Water table: the level within a rock mass below which all voids are filled with groundwater; above it the **vadose zone** is freely draining, and below it the **phreatic zone** is totally and permanently saturated.

Wayboard: thin bed of altered and weathered volcanic ash or **toadstone**, within the limestone sequence of the Derbyshire Peak District.

Site locations

Locations of the sites are indicated by conventional National Grid References, for purposes of filing and data retrieval. SSSI boundaries are only fully defined on maps which are filed with the site documents. 4 or 6 figure references, of the whole site or the main feature, are cited as appropriate; some caves are cited only by their main or sole entrance. References linked by dashes define lines of features or the approximate extents of very large sites. References separated by commas define individual features grouped into the same site.

The Yorkshire Dales karst

Ease Gill Cave System	SD6778 — SD6681
Kingsdale caves	SD6976 — SD7079
Scales Moor	SD7075 — SD7277
Ingleborough karst	SD7173 — SD7678 — SD7970
Ingleborough caves	SD7173 — SD7478 — SD7776 — SD7571
Birkwith caves	SD8075 — SD8077
Brants Gill catchment caves	SD8172 — SD8569 — SD8274
Pikedaw Calamine Caverns	SD875640
Malham Cove and Gordale Scar	SD8964 — SD9164
High Mark	SD9168 — SD9467
Penyghent Gill	SD857734
Sleets Gill Cave	SD959692
Boreham Cave	SD926726
Strans Gill Pot	SD916788
Birks Fell caves	SD9376
Dow Cave	SD984744
Black Keld catchment area	SD995711, SE016697
Conistone Old Pasture	SD9867 — SD9966 — SD9968

Outlying karst areas of the Northern Pennines

Hutton Roof	SD5576 — SD5678
Farleton Knott	SD5479
Gait Barrows	SD482775
Hale Moss caves	SD500775
Short Gill Cavern	SD670847
Upper Dentdale caves	SD732861 — SD741864

Stump Cross Caves	SE089632 — SE092635
Nidderdale caves	SE100764 — SE105730 - SE088739
Hell Gill	SD787969
Cliff Force Cave and the Buttertubs	SD875690
The Clouds	SD7399 — NY7400
Great Asby Scar	NY6409 — NY6809
Little Asby Scar and Potts Valley	NY6809 — NY7009
Helbeck Scars	NY7916 — NY7620
God's Bridge	NY957126
Knock Fell Caverns	NY720307
Fairy Holes	NY943368

The Peak District karst

Castleton caves	SK099813 — SK149826
Winnats Pass	SK136826
Cave Dale	SK150825
Bradwell Dale	SK171792 — SK174808
Bagshaw Cavern	SK171809
Stoney Middleton caves	SK2175
Poole's Cavern	SK050725
Lathkill Dale	SK1566 — SK2165
Upper Lathkill Dale caves	SK171659, SK161646, SK144674
Green Lane Pits	SK165626
Masson Hill caves	SK2859 — SK2958
Dove Dale	SK1358 — SK1451
Manifold Valley	SK0951 — SK0956

The Mendip Hills karst

Burrington Combe	ST4758
Charterhouse caves	ST471564 — ST498556
Cheddar Gorge	ST4653 — ST4854
Cheddar caves	ST467539, ST474544
Priddy caves	ST531513 — ST549501
Wookey Hole	ST532480
Brimble Pit and Cross Swallet	ST509508, ST508500
Sandpit Hole and Bishop's Lot	ST532498, ST550495
Wurt Pit and Devil's Punch Bowl	ST559539, ST544537
Lamb Leer Cavern	ST543550
Thrupe Lane Swallet	ST603456
St Dunstan's Well catchment caves	ST657477, ST669474

Karst in Wales

Dan-yr-Ogof	SN838160
Ogof Ffynnon Ddu	SN848152 — SN873166
Little Neath River Cave	SN912142
Porth-yr-Ogof	SN928124
Mynydd Llangynidr	SO1214 — SO1515
Mynydd Llangattwg caves	SO1616 — SO2112 — SO2015

Ogof Draenen	SO2312 — SO2509 — SO2611
Otter Hole	ST526963
Pant-y-llyn	SN606166
Llethrid Valley	SS5390
Minera caves	SJ2550 — SJ2551
Alyn Gorge caves	SJ191653

Outlying karst areas in England

Slaughter Stream Cave	SO582137
Buckfastleigh caves	SX743666, SX748678, SX735652
Napps Cave	SS565475
Cull-pepper's Dish	SY814925
The Manger	SU298868
Beachy Head Cave	TQ579953
Devil's Dyke	TQ2611
Water End swallow holes	TL231043
Castle Lime Works Quarry	TL228026
Devil's Punchbowl	TL878892
Millington Pastures	SE8443 — SE8555
Moston Long Flash	SJ719620
Rostherne Mere	SJ745842

Karst in Scotland

Traligill Valley	NC2621 — NC2819
Allt nan Uamh caves	NC275172, NC271166

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