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John Gunn and *Ian Fairchild*, convenors

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Groundwater in Carboniferous carbonates

Field excursion to the Derbyshire "White Peak" District
26th June 2015

John Gunn
with contributions from Andy Farrant, Paul Hardwick and Steve Worthington

Contents

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Cover photographs

A common perception of the Peak District karst hydrogeology is that it is dominated by sinking streams that feed caves which in turn discharge substantial flows of water from large springs such as Peak Cavern Rising (left). However, over much of the area where limestones crop out there is only dispersed autogenic recharge through a soil cover and surprisingly little evidence of karstic drainage, particularly at depth. During the excavation of stone from Ballidon Quarry in the southern White Peak no caves were encountered and the 100m of limestone exposure (right) shows only an epikarst below which there are no visible conduits and nothing to suggest concentrated drainage. (photographs by John Gunn)

1. INTRODUCTION

Carbonate rocks of Carboniferous (largely Viséan) age crop out widely in England and Wales but many of the outcrops are scattered and broken (Figure 1). The Peak District is one of six contiguous areas, the others being the Yorkshire Dales, the Northern Pennines, the Mendip Hills, South Wales (including the Forest of Dean) and North Wales (Gunn et al., 1998). These areas have commonly been regarded by hydrogeologists as having karstic drainage, whereas geologically more recent carbonates such as those deposited in Jurassic and Cretaceous times are commonly regarded as not being karst or, at best, 'weakly karstic'. In particular, there has been a tendency to equate karst drainage with surface landforms and to use 'caves as a measure of karst'. Hence, those carbonates with visually impressive surface karst landforms (dolines, blind and dry valleys) where there are well-developed and extensive cave systems are commonly assumed to have karstic hydrogeology and those without these landforms are assumed to be 'non-karstic'. One weakness of this simplistic approach is that caves in carbonates are simply conduits that have been enlarged by dissolution to a point where humans can explore them. Water flowing though a conduit that is, say, 0.2m in diameter and hence un-enterable will not behave any differently if after some distance the conduit attains a diameter of 0.4m and can be visited by humans. There is also an increased awareness that the marked spatial heterogeneity in carbonates at small scale (commonly expressed in terms of dual or tertiary porosity and permeability) is also present at the field scale. Hence, even the most karstic of regions, where groundwater moves through conduits and caves at velocities in the hundreds of metres per hour, is likely to also have areas where there are few conduits and where groundwater velocities may be as little as a few metres per day. The Derbyshire "White Peak" district provides an excellent location in which to discuss these concepts and their importance to management or water and mineral resources.

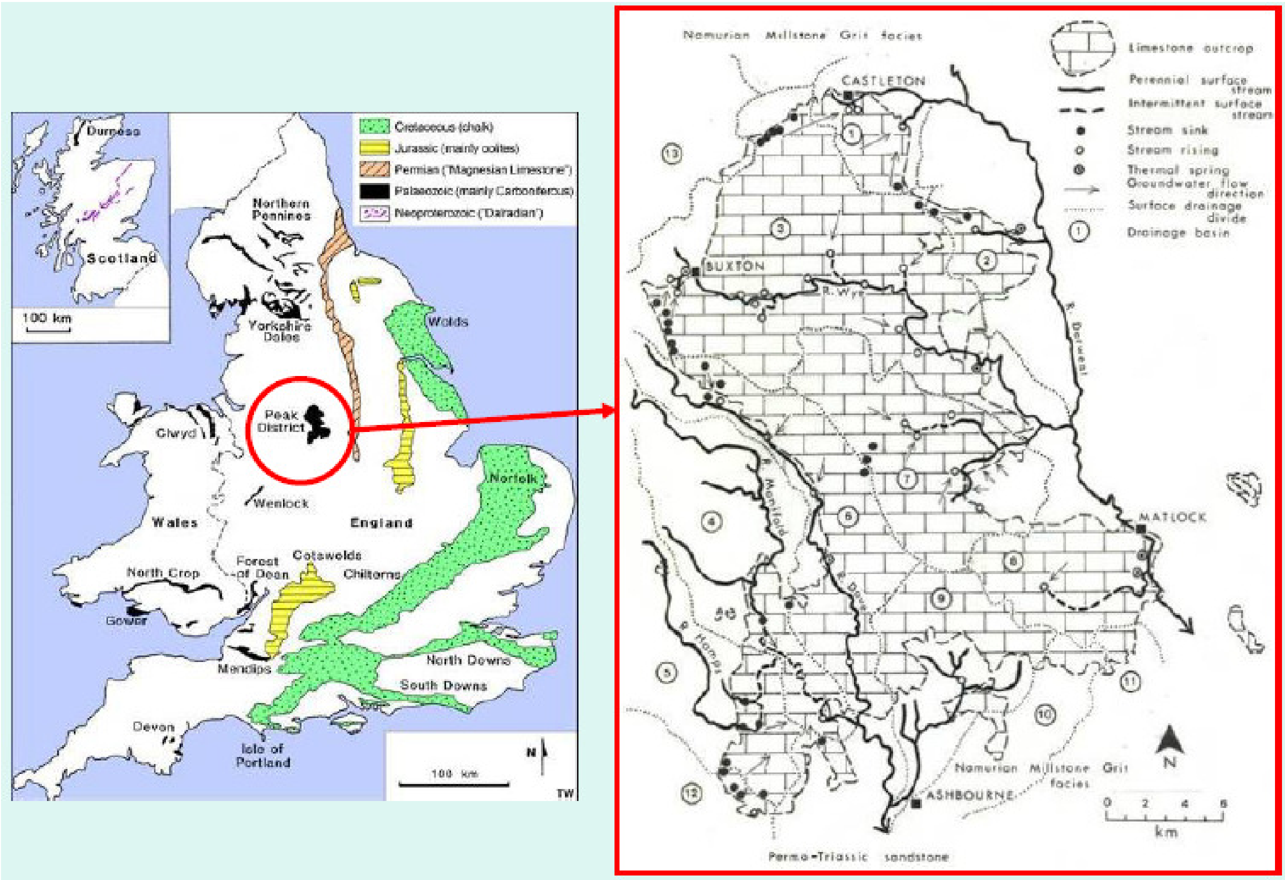

Figure 1. British carbonate rock outcrops (left) and the White Peak (right)

2. THE "WHITE PEAK" DISTRICT

The Peak District is situated in the centre of England at the southern end of the Pennine range of hills that form the main watershed of England, separating Irish Sea drainage to the west from North Sea drainage to the east. It has two distinct parts, the "Dark Peak" and the "White Peak", both comprised of Carboniferous age rocks. The Dark Peak, underlain by Namurian age sandstones and mudstones of the Millstone Grit Group, is substantially to the north but also along both west and east flanks; the "White Peak" to the south is some 40km from north to south (where it abuts Triassic sandstones) and up to 20km west to east, and largely comprised of limestone. It rises to an altitude of some 450m above sea level, and is drained by the River Derwent and its tributaries (most notably the River Wye) to the east, and by the rivers Dove and Manifold to the west. All of these rivers are tributaries of the River Trent, which flows around the area in an anticlockwise sweep.

Most of the White Peak lies within Britain's first National Park, established in 1951, although some areas were excluded on account of their high intensity of quarrying, which still continues. The towns of Buxton and Matlock, together with most of the lower Derwent Valley south of Rowsley, are also just outside the Park boundary.

3. GEOLOGICAL BACKGROUND

The bedrock geology in the White Peak is dominated by a thick succession of limestone beds of Viséan (early- to mid-Carboniferous) age that belong to the Peak Limestone Group (Figure 2; Table 1). Some 500m of beds are exposed, and deep boreholes have proved the existence of at least another 1300m in places, so that there is no impermeable basement at outcrop. The limestones were deposited in three main depositional environments: lagoonal, reefal (with mud-mound ramps and fore-reef slopes) and basinal. Subsequent folding has produced dips of up to 20° in lagoonal facies rocks, and much steeper dips in reefal rocks. Bedding planes are rare in the reef limestones but common elsewhere, with individual beds ranging in thickness from 0.5m to over 5m. Both the geomorphology and the hydrogeology reflect the varied character of the limestones. In particular, marginal reefs at Castleton (to be visited on the excursion), Upper Dovedale and around Matlock and Wirksworth form striking hills bordering the lagoonal limestones of the central 'plateau' of the White Peak.

During the extended period of limestone deposition, volcanic activity related to several active “centres” produced periodic pulses of both lava emplacement and eruption of airborne volcanic debris, some of which was deposited and now forms consolidated tuff interbeds in the immediate areas of the centres and “downwind” of them. Over much of the northern White Peak, two significant lava beds (the Lower and Upper Miller’s Dale Lava members) which are believed to derive from a centre near Tunstead (Aitkenhead *et al.,* 1985), subdivide the limestone succession (Table 1).

Table 1: Generalized lithostratigraphy (not to scale) in the central and northern White Peak.

<table>
<tr><th>Group</th><th>Formation</th><th>Member (northern White Peak)</th><th>Notes and local differences</th></tr>
<tr><td rowspan="7">Peak Limestone Group</td><td>Eyam Limestone Formation</td><td></td><td>Its base is locally unconformable and the unit may be absent, with the Monsal Dale Limestone Formation overlain by clastic rocks of the Longstone Mudstone (at the base of the Widmerpool Formation)</td></tr>
<tr><td rowspan="2">Monsal Dale Limestone Formation</td><td></td><td>In the east of the White Peak the Monsal Dale Limestone Formation is underlain by the Fallgate Volcanic Formation. Its base is currently not proven but it is likely that older limestone formations occur below.</td></tr>
<tr><td>Upper Miller’s Dale Lava Member</td><td rowspan="4">Members distinguished only locally where the presence of the named volcanic units allows subdivision as shown</td></tr>
<tr><td rowspan="3">Bee Low Limestone Formation</td><td>Miller’s Dale Limestone Member</td></tr>
<tr><td>Lower Miller’s Dale Lava Member</td></tr>
<tr><td>Chee Tor Limestone Member</td></tr>
<tr><td>Woo Dale Limestone Formation</td><td></td><td></td></tr>
</table>

In addition to these major units, various more localized volcanic rock beds are found scattered geographically and stratigraphically across and through the Peak District limestone succession. Thinner clay-rich beds, known locally as (clay) wayboards, are widespread and most likely result from fall-out of airborne volcanic dust generated during contemporaneous regional volcanic activity. Such fall-out almost certainly continued at background level throughout early Carboniferous times but the related airborne sediment only dominated within the succession (rather than simply being a low level impurity component) when limestone deposition periodically slowed down or ceased due to temporary cyclic uplift and/or fall of sea level.

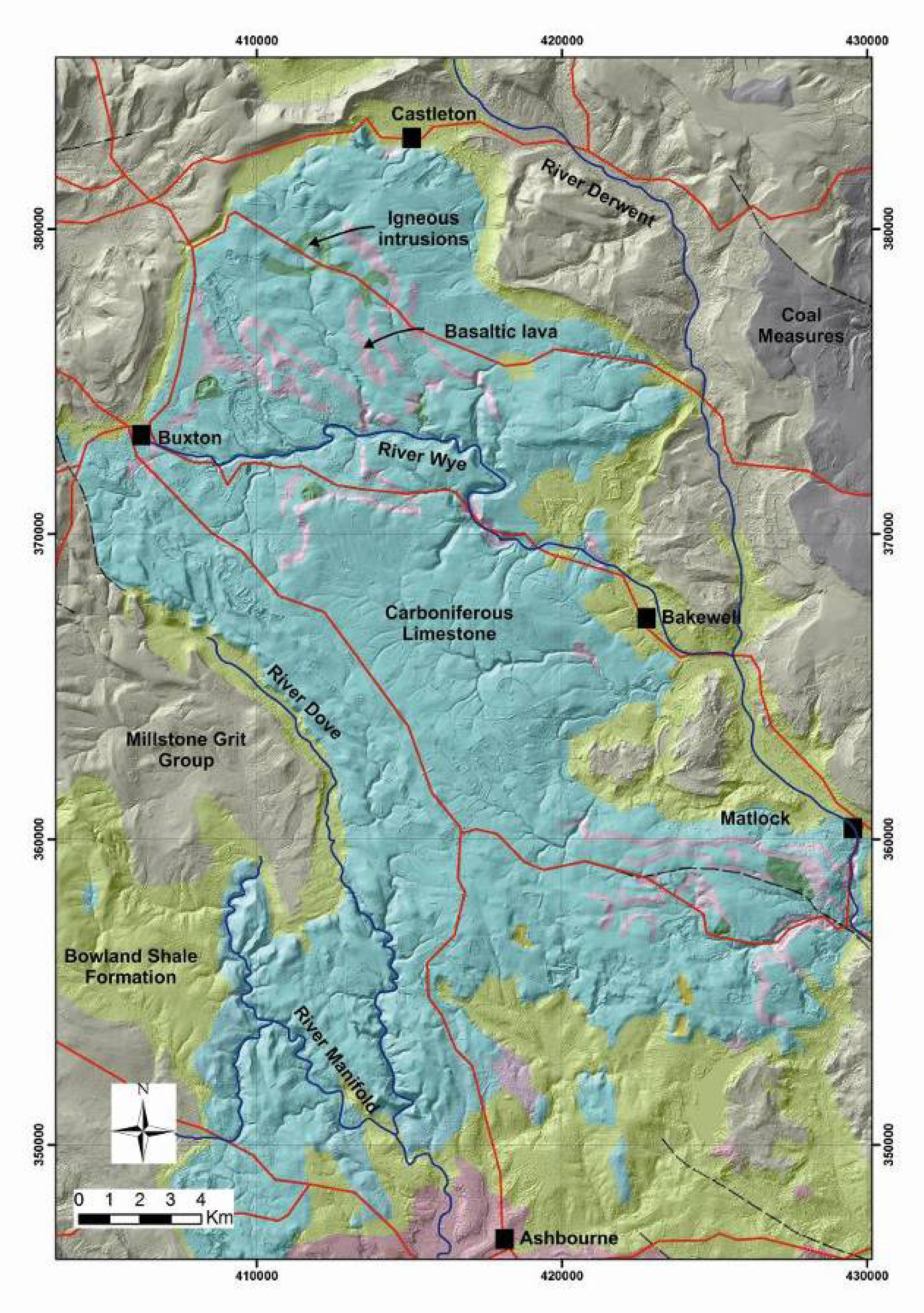

Figure 2: Geological context for the excursion. (Based upon 1:625,000 scale DigMap bedrock geological maps, with the permission of the British Geological Survey. NEXTMapTM Britain elevation data from Intermapp Technologies)

The clay wayboards and the thicker volcanic rock horizons exert at least local influences upon tertiary permeability (conduit) development in the limestone sequence. This might be because under some conditions they can act as aquitards, or it might also be because they exhibit atypical chemical and physical behaviour. Equally, these beds and horizons, which function as one type of Inception Horizon (Lowe and Gunn, 1997; Banks et al., 2009), must also affect underground water movement and storage. However, aspects of the imposed regional and local geological structure also play an important part in directing and constraining these influences and processes. The lavas and wayboards have also influenced the underground movement of mineralizing fluids. Locally-preserved palaeokarstic features indicate at least one period of sub-aerial conditions during the Carboniferous (Ford, 1984) and following burial beneath later sediments, some palaeokarstic voids were exploited as migration routes for hydrothermal fluids, becoming infilled by mineral deposits.

At the end of the Early Carboniferous episode of limestone sedimentation, there were widespread earth movements resulting in folding of the limestone beds. With fold axes being mainly west–east in the eastern part of the White Peak, beds dip either north or south, and subterranean drainage tends to flow either down dip or along the strike. Some of these folds plunge eastwards so that drainage routes within them can be both down-dip and down-plunge. The combination of facies changes in the limestone beds and the pattern of fold axes and jointing have contributed to the complexity of some cave patterns. Renewed folding took place in late Carboniferous times, generally accentuating the previous folds and also affecting the thick cover strata of deltaic sands and muds, now forming the Millstone Grit and Coal Measures groups. About 2km of these clastic sediments once lay on top of the limestone, but following later upwarping along the Pennine axis erosion has stripped off the cover and thus exhumed the limestone. The main features of the remnants of this cover are gritstone escarpments ("edges") that provide a natural boundary around the White Peak.

Stresses associated with the folding opened numerous fractures, which then became repositories for a hydrothermal mineral suite of galena, fluorite, baryte and calcite deposited from hot fluids rising from adjacent sedimentary basins. Mineralization occurred mainly in late Carboniferous times. Mineralizing fluids rose at temperatures between 70 and 120ºC at a time when the entire area was still covered by the Millstone Grit and Coal Measures. Mineralisation is a key feature of the Peak District karst, which is traversed by hundreds of veins. All known veins were mined for lead between Roman times and the mid 20th Century and their surface traces are now marked by lines of old workings and waste heaps. In the 20th century there was renewed mining, and re-working of waste heaps, for fluorspar and baryte and this continues on a much-diminished scale. One form of fluorspar, Blue John, which is of high value as an ornamental stone, occurs only within palaeokarstic voids on Treak Cliff near Castleton in the north of the White Peak.

Some 15km^2 of the limestones in the southern part of the White Peak have been affected by dolomitization, which destroyed fossils, increased porosity and reduced the rock solubility. The origin of the magnesium and the cause of the process are controversial, but dolomitization is now widely regarded as an early phase of the mineralization process. Locally the contact between dolomite and unaltered limestone has been the target of special forms of mineralization and the site of cave development, particularly in the Matlock and Brassington areas.

Following the erosional stripping of most of the Millstone Grit (and younger) cover, the later seas of the Permian Period may have transgressed the southern Pennines, but no deposits of Permian age survive. Rocks deposited during the succeeding Triassic Period are represented by sandstones, pebbly sandstones and conglomerates, the remnants of which abut the limestone in a small area near Ashbourne. The Brassington Formation, which was deposited in late Tertiary times (Miocene–Pliocene) consists of spreads of clays, sands and a few gravel layers derived from the Triassic formations. Relics of the Brassington Formation, which once covered the greater part of the southern White Peak, have subsided into collapse-structures in the limestone, to form so-called Pocket Deposits (Walsh et al., 1972).

Glaciers scoured the Peak District several times during the Pleistocene although there is still some question over the number of advances. It is generally accepted that the most recent complete ice cover was around 450,000 years ago during the Anglian Glaciation although there may have been valley glaciers during a subsequent cold period that deposited small patches of "Wolstonian" till. As the White Peak was ice free during the Devensian (the last glaciation), glacial effects and deposits are less distinct than in more recently glaciated areas such as the Yorkshire Dales. Instead there are deposits that formed under periglacial conditions: structureless masses of unsorted debris (head), fine sediment (loess) blown in by wind from the extensive glacial outwash deposits in adjacent areas, and widespread but localized scree deposits some of which have subsequently been cemented by carbonate precipitates.

4. SURFACE DRAINAGE AND GEOMORPHOLOGY

The majority of the White Peak is devoid of surface water-courses and only two allogenic-fed rivers, the Dove and the Wye, maintain perennial flow across the limestone. The River Wye, which will be viewed on the Excursion, has its headwaters on Namurian Millstone Grit Group strata to the north and south of Buxton but is fed by limestone groundwater to the east. For an unknown distance downstream of its confluence with Monk's Dale [SK 142 733] the river is underlain by the Lower Miller's Dale Lava and perching on the lava may maintain surface flow during periods of groundwater recession. Despite the absence of present day surface drainage, the most notable feature of the White Peak is that it is dissected by a dendritic system of valleys ('dales') that are dry for all, or most, of the year (Figure 3). Thus, the Peak District is an internationally important example of a relict fluviokarst. The valleys are thought to have been inherited by the limestone after originating on an impermeable clastic cover. Their desiccation largely reflects gradual karstification of drainage, aided by regional lowering of groundwater elevations in response to downcutting by major rivers. However, not all the desiccation was natural, as dewatering via mine soughs affects some rivers, most notably the middle reaches of the River Lathkill, which dry up every summer. Although the entire karst was overrun by ice during at least one glacial advance, Quaternary glaciations are not thought to have played a significant part in the incision of major river valleys. However, the smaller dry valleys would have been reactivated when the ground was frozen under periglacial conditions during the Pleistocene cold periods. Solution dolines, suffosion dolines that occur where the limestone is overlain by thick superficial loess and other drift deposits, and a few small collapse dolines pit the limestone surface, but they are mere details within the dominantly fluviokarstic landscape. Additionally, the courses of most lead veins are marked by depressions that were either modified or created by lead miners.

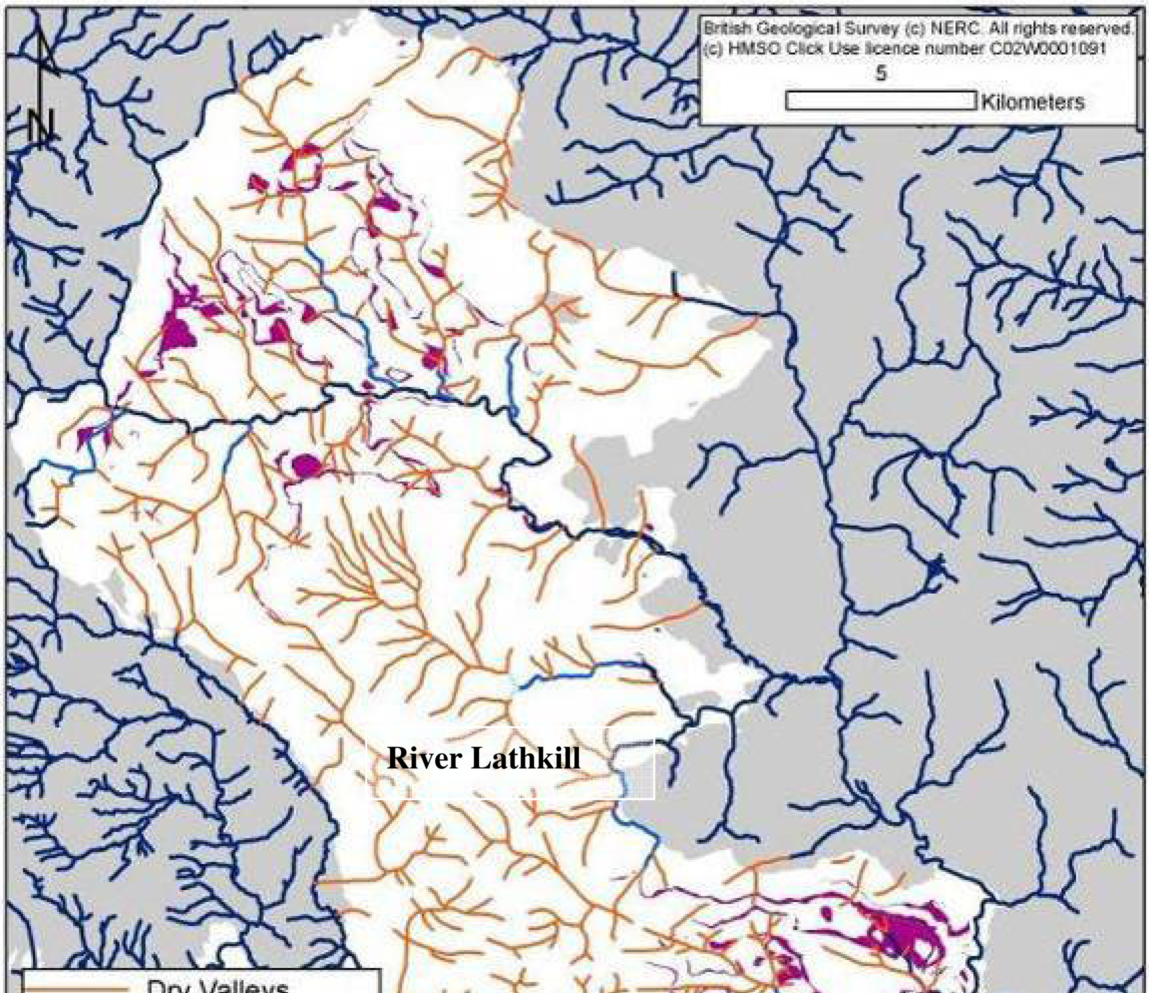

Figure 3: The Peak District fluviokarst valley network (after Warwick, 1964)
[white area = limestone outcrop; purple = volcanic rocks; remainder = younger cover rocks

Limestone pavements are rare in the Peak District; those that 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 marking the outcrops of the stronger limestone beds and especially the massive reef units. There are extensive screes, largely formed due to Pleistocene frost action, and some rocky tors, particularly in areas where dolomitised limestones crop out. Most of the surface between the valleys lies under a veneer of soil, largely formed on periglacial loess deposits. Superficial material on the steeper valley sides is commonly soliflucted, and soil creep has initiated the formation of extensive areas of terracettes. Only isolated patches of an initially widespread Anglian glacial till have survived the erosional regime that has prevailed in the area since the ice cover decayed.

Tufa deposits are a notable feature of the White Peak. The oldest, in Lathkill Dale, date back to the last interglacial but most are of Holocene age. They include both cold water and thermal water precipitates, the latter being primarily in the Matlock Bath area, which will be visited at the end of the Excursion (STOP 10). There are also tufa deposits in the Buxton area that have formed in the last hundred years where hyperalkaline groundwaters emerge and scavenge carbon dioxide. The source of the alkalinity is quicklime waste left behind after lime burning.

Quarrying of limestone has been important for over 500 years and the White Peak is one of the main suppliers of both aggregates and particularly high-purity limestones in Britain. Quarrying has impacted on the natural landforms and more limestone was removed by quarrying during the 20th century than by natural dissolutional processes acting over the whole of the Holocene (Gunn, 1993).

5. THE HYDROGEOLOGY OF THE PEAK DISTRICT LIMESTONES

The present eroded ‘dome’ structure of the Peak District was once covered by younger clastic rocks, and at that time there would have been underground drainage under confined conditions within the carbonate sequence. It is known that: (1) conduit development began in the carbonates before they were buried by younger sediments; (2) underground fluid movement into the Dome was active long before the present topography developed, with mineralising fluids entering from basins to the west and east; and (3) underflow drainage of “fresh” water migrates out from the Dome towards the southeast where it is displacing saline formational waters in Carboniferous sequences. There is no reason to suppose that such routes did not exist before the Dome was unroofed, carrying confined underground drainage from adjacent areas into the buried limestones of the Dome, through/across and (presumably) out of the Dome, according to prevailing regional hydraulic conditions. Thus, it is likely that incipient (or more active) underground drains were developed beneath the Dome prior to it achieving its present topographic expression. It is possible that some such routes (guided only by stratigraphy) were imposed, at least incipiently, before the current Dome structure itself was produced by intra-Carboniferous and end-Palaeozoic tectonism (as reflected by evidence of pre-Namurian karstification). The area may have been “unroofed” at the time of ore emplacement, but this is by no means certain. However, at that time fluids could freely enter the Dome by deeply buried routes. Ore-body morphologies confirm that significant karst cavities related both to the stratigraphy and to fracture structures, pre-dated mineralisation. In the Castleton area (Shaw, 2015; Worley, 2015) and in the Matlock area (Ford, 1984), lead miners entered caves with palaeokarst features that intersect and follow some of the mineral ore bodies.

After a long period of slow water circulation through the limestone, largely under confined conditions, the Dome was probably unroofed to its present extent during the Palaeogene/Neogene. As discussed above, Tertiary sediments were deposited in pre-existing dolines and it seems likely that there was also active groundwater circulation through conduits. The present circulation system, although guided by earlier flow paths, essentially dates from that time. Initially there would have been significant allogenic recharge from stream-sinks formed where water ran off from less permeable Namurian rocks. Evidence is provided by significant clastic sediment deposits in isolated sections of now relict cave passages several kilometres from any sources of modern allogenic recharge and many tens of metres above the present zone of active groundwater circulation. At Matlock, the Masson Hill caves contain sediment sequences that include fluvioglacial material more than 730,000 years old (Noel, 1987; Noel et al., 1984), offering fragmentary evidence that covers a long timespan of karst evolution. A second equally long record of Pleistocene events is provided by the relict caves and sediment fills exposed by limestone quarrying in Eldon Hill [see discussion at STOP 3].

The Namurian mudstones are more susceptible to erosion than the limestones and they were steadily stripped back through the late Tertiary and the Quaternary such that most of the limestone outcrop now stands topographically higher than the surrounding outcrops of younger rocks. Hence, there are only limited areas mainly around the northern periphery of the limestone outcrop where escarpments of Namurian Millstone Grit Group strata are close enough to the limestones and high enough for surface streams to cross the younger rocks and sink into the limestone. The best example is Rushup Edge [see discussion later in this guide] from where thirteen streams drain mainly via sinks in marginal reef limestones into the Peak-Speedwell Cave System.

Dispersed recharge through the cover of soil and superficial deposits dominates across most of the limestone outcrop and consequently, there are relatively few open stream-sink caves in the Peak District. 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 also few in number.

Over much of the White Peak, the underground drainage pattern is less well known than in the other British Carboniferous limestone regions, partly because water tracing experiments cannot easily be undertaken in an area largely devoid of surface drainage and partly because most of the tracing experiments that have been undertaken away from the margins have failed. The failed experiments involved tracer injection into either quarry boreholes (wells) or into small streams that start as small perched springs where volcanic rocks crop out, flow over the surface for a short distance, and sink back into the limestone. Some of the early experiments involving quarry boreholes may have used insufficient tracer as a consequence of a desire not to produce visual contamination of a prominent fishing river and a public water supply abstraction. However, Gunn & Worthington (2014) provide detailed analysis of a more recent experiment in which the amount of tracer injected was predicted to produce concentrations above the limit of detection if the dye had emerged at the monitored springs during the 50-day sampling period. One possibility is that (a) that there is not an efficient channel flow path linking the injection well with conduits that drain to the springs and (b) that groundwater velocities outside of the conduit network are very slow, less than 2m/h [See also discussion at STOP 2]. Alternatively, there could have been substantial dye losses such as adsorption onto clastic sediments or diffusion into the matrix of the rock, or it is possible that the dye emerged at springs further downstream on the Wye than monitored locations. On the same day that tracer was injected into the borehole, a different tracer was injected about 600m away at the point where a stream fed by a spring perched on basalt sinks. That tracer emerged from the springs in 37 hours and concentrations peaked after 57 hours, respectively indicating velocities of 68 and 44 m/h. These are typical of conduit flow and it was initially assumed that all such sinking streams provide concentrated recharge that enters conduits and moves relatively rapidly to springs. However, subsequent experiments have shown that this is not the case and it is now hypothesised that across parts of the area both concentrated and dispersed recharge may feed deep drainage that feeds thermal springs or drains out of the White Peak at depth (Gunn et al., 2006).

The Carboniferous Limestone in the Peak District is traversed by hundreds of mineral veins, all of which have been mined in the past for lead and zinc, and some of which are currently exploited for fluorspar. Mining has a long history, and it is clear that drainage was always a problem and became increasingly so as time progressed. Although pumps were employed in many mines, the solution adopted across the area was to construct drainage adits, called ‘soughs’, to dewater the mineral field. The soughs were designed to lower the natural water levels and drain the phreatic zone. Distant soughs are known to have intersected active underground drainage systems, effectively lowering “water tables”, at least locally. It is likely that some such “water tables” were actually stratigraphically isolated drainage systems related to individual guiding beds within the limestone sequence. Such stratigraphically discrete systems may have been regionally extensive or local, depending upon whether fractures and folds acted as barriers to drainage or as enhanced drainage routes. As each progressively lower conduit system was drained following the driving of deeper and deeper soughs, the pre-existing underground drainage pattern would be forced to evolve. Former springs would be left without underground headwaters and drainage sinking into pre-existing voids would potentially divert to the man-made risings [sough tails]. This interpretation is supported by historical accounts, which indicate that drying-up of springs was quite common following the driving of soughs. However, deep aspects of the ancient imposed underground drainage system might potentially have continued to function, and as some of the links used by the diverted drainage would be immature, there would be scope for ponding, back-up and overflow during times of higher than average recharge.

6. CAVES

There are 210 caves with a total passage length of around 50km in the White Peak, the ratio of both caves and explored passage length per unit area of limestone outcrop being markedly less than in the other Carboniferous limestone regions in Britain (Gunn et al., 1998). However, caves of a great variety of morphological forms and evolutionary histories are to be found in the Peak District. The oldest are palaeokarstic caves associated with a mid-Carboniferous phase of uplift and karstification (Ford, 1984). At about 300M years these are among the oldest known cave remnants in Britain. During the mineralization phase the pipe-veins were effectively hydrothermal caves, initially enlarged due to dissolution by acidic waters during the Carboniferous Period and subsequently partly filled or lined with minerals. Good examples are to be seen in Treak Cliff Cavern near Castleton and at Masson Hill near Matlock, some of these have been modified by the effects of later meteoric drainage.

Much younger are the currently active stream caves, particularly those of the Castleton-Bradwell and Eyam–Stoney Middleton areas. The first stage of their development, their inception, probably occurred at depth while the limestones were still completely covered, but the formation of integrated conduit networks did not occur until the early phases of exposure of limestone by stripping off the cover of shales and sandstones.
Some of the early conduits intersected previously isolated vein cavities, such as the Bottomless Pit in Speedwell Cavern, the Oxlow Caverns and Leviathan and Titan in Far Peak Cavern. Water circulation would initially have been very slow, with a rapid increase when the proto-conduits were intersected at lower levels by incising valleys on the edge of the newly exposed limestone massif. Initially, slow drainage dissolved out walls, floor and roof at the same rate, giving rise to phreatic tube caves. Then water levels fell and allogenic streams entered these tubes, eroding their floors and carving canyon-like passages with classic 'keyhole' cross-sections.

7. ITINERARY

From Birmingham we will follow the A38 north past Lichfield, joining the A515 and continuing north through Ashbourne to enter the White Peak (Figures 2 and 4). We then follow the same road towards Buxton stopping at Bull I'Th'Thorn pub, Hurdlow for refreshments and to discuss the interpretation of monitoring data from boreholes in carbonate rocks.

Figure 4: Excursion route from Birmingham to Hurdlow (red line) and from Matlock to Birmingham (blue line). Black rectangle is excursion area; see Figure 2. [Base map is OS open-source; copyright Ordnance Survey, 2015]

STOP 1: THE BULL I'TH'THORN BOREHOLE, HURDLOW

The Bull I'Th'Thorn borehole is one of a small number of long-term limestone groundwater monitoring sites in the White Peak. The borehole was drilled in 1933 and logged to a depth of 385 feet (117.35m). However, the Environment Agency database gives the depth as 230m so it would appear that it was deepened at a later, unknown, date. The datum is given as 356.38m aOD (above Ordnance Datum, the UK standard for elevation measurements). In common with other monitoring boreholes in the White Peak limestone it is an open hole. The Monsal Dale Limestone crops out at surface and the borehole log records a 'pale blue shale' from 324 to 328 feet below ground level (257.6 to 256.4m aOD) that may be a lateral continuation of the Upper Miller's Dale Lava in which case the lowest part of the hole is in the Miller's Dale Limestone (Table 1).

Water levels at the borehole have been measured since April 1969, initially at approximately weekly intervals and since the early 1980s about once a month although there was a gap between May 2008 and April 2011. Figure 5 shows the water surface elevation in the borehole from 1969 to 2014. The elevation ranges from 224.0 to 269.6m aOD and averages 248.1m aOD.

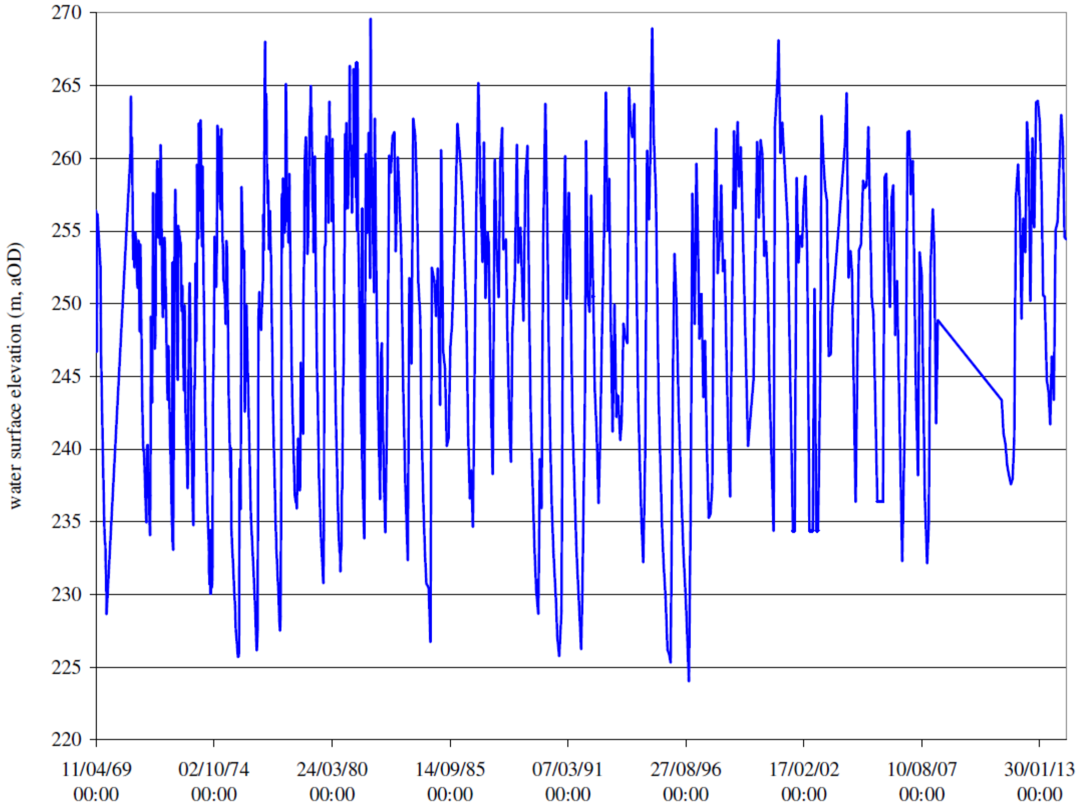

Figure 5: Water surface elevation in the Bull I'Th' Thorn borehole 1969-2014

The early, shorter interval, data (Figure 6) show that the water surface varies through the year but even at this sampling interval it is difficult to be certain how rapidly it responds to recharge. This is made more difficult by the longer interval between more recent data where there is no apparent relationship between rainfall and groundwater elevation (Figure 7).

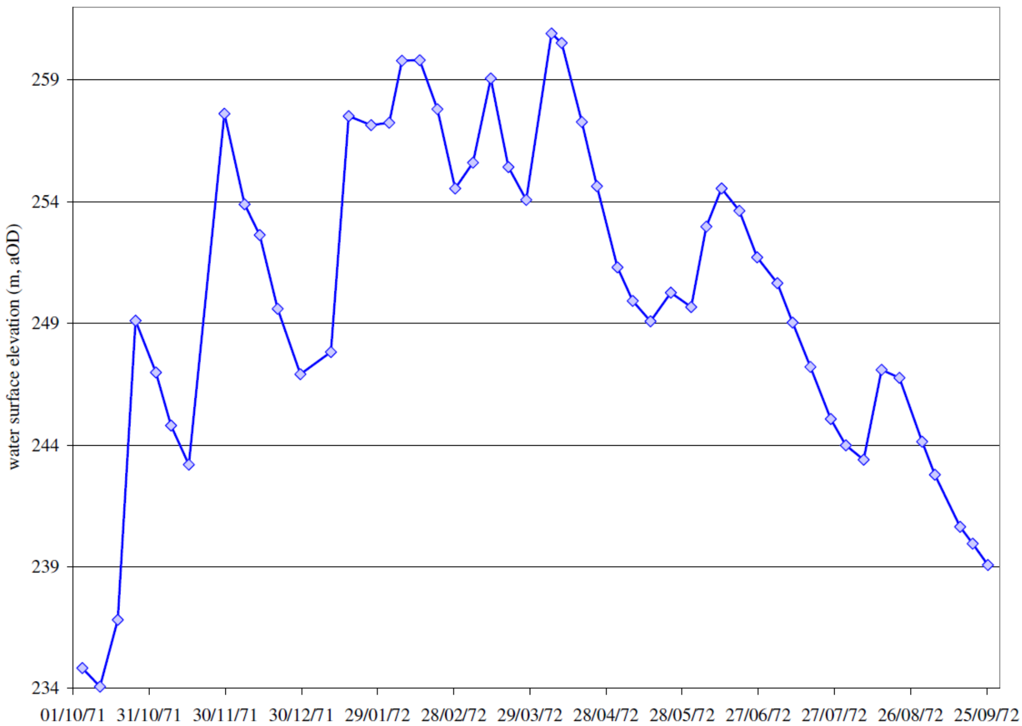

Figure 6: Water surface elevation in the Bull I'Th' Thorn borehole, 1971-1972 hydrometric year

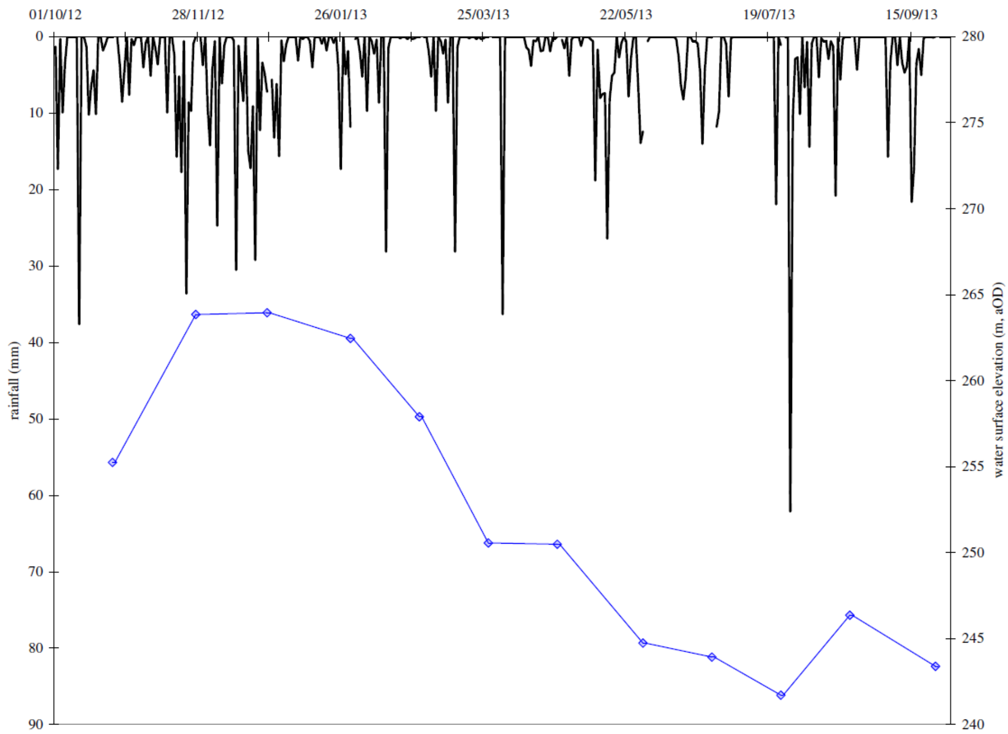

Figure 7: Water surface elevation in the Bull I'Th' Thorn borehole and daily rainfall at Buxton 2012 - 2013 hydrometric year

A key question is what the water elevation data actually tell us about the hydrogeology of the Carboniferous limestone in this part of the White Peak. The large annual variation in water level reflects the low porosity of the limestone, which is about 1%, with conduit porosity being about 0.03-0.05% (Atkinson, 1977; Worthington and Ford, 2009; Gunn and Bottrell, 2013). The borehole is almost exactly on the topographic divide between the Dove and Lathkill catchments but the lowest measured water elevation (224m) is some 10 m lower than the River Dove, 3km to the southwest, so groundwater flow in this area is most likely to the east to springs on the River Lathkill, which offers a steeper hydraulic gradient than flow to the River Wye, 6km to the north. The closest spring in Lathkill Dale is 4.3km to the southeast and is an overflow spring at an elevation of 207m. In low-flow conditions it is dry, and over 1500m of conduits have been explored (Lathkill Head Cave). The Holme Grove Risings, the lowest of which is 350m down the valley at an elevation of 196m aOD, are perennial most years but cease to flow during severe drought conditions. This occurred at the end of April 2011 when the elevation in the borehole was 243.4m giving a low-flow hydraulic gradient from the well to the spring of 0.0103. The perennial source of the Lathkill is at Bubble Springs, a further 3.1km down-valley at an elevation of 157.3m aOD. The straight lie distance from the borehole to Bubble Springs is 7.7km and the lowest elevation in the borehole in 2011 was 237.6m aOD, giving a virtually identical hydraulic gradient of 0.0102. These low values reflect the high permeability of the limestone. The hydrogeology of the area has been complicated by mining and some groundwater that previously flowed to the Lathkill has been captured by a drainage adit (Magpie Sough, visited later in the excursion if time permits) about 4.1km from the Holme Grove Risings at an elevation of 145m aOD. Following a hydrological study (Gunn, unpublished) it was recommended that a control structure be constructed at the south tail to raise groundwater elevations and restore baseflow in the River Lathkill. A group has been established to obtain funding to progress this proposal.

In boreholes in carbonates, almost all the flow usually comes from just a few solutionally-enlarged fractures (Worthington, 2015). This is almost certainly the case at Bull i'th' Thorn and although there is no downhole information at this site, Banks (2007, p. 190) has suggested that it may be possible to make deductions based on recession curve analysis: "*borehole recession curves comprise baseflow and the form of the recession is indicative of changes in aquifer storage and transmissivity with depth, attributable in part to the tightening of fissures as the depth of overburden increases with depth"*. She constructed a master recession curve (Figure 8) on which she identifies six indistinct steps thought to correspond to changes in hydraulic conditions. The recession curve is similar to that in other boreholes in karst, with an approximately exponential recession reflecting the progressive drainage of lower-permeability stores, and with drainage of the epikarst probably contributing significantly (Gunn, 1986).

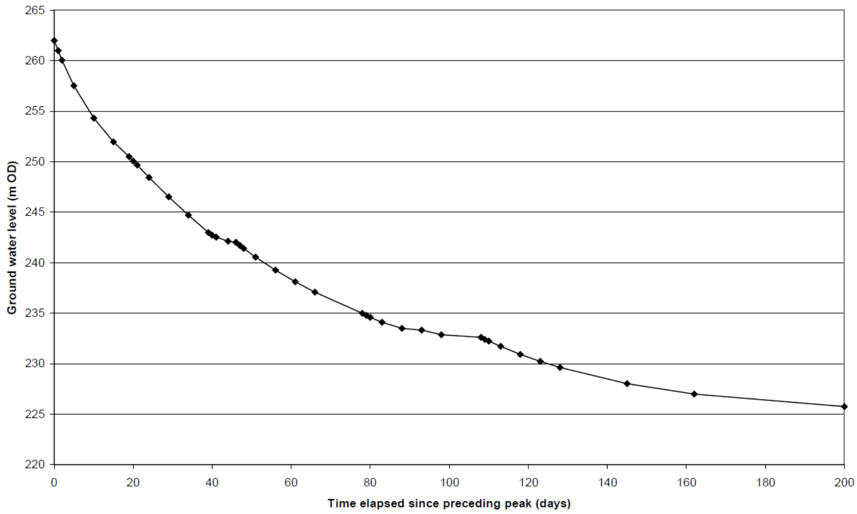

Figure 8: Master recession curve for the Bull I' Th' Thorne Borehole (Figure 8.4 in Banks, 2007)

From Hurdlow we continue northwards along the A515 (Figure 9), passing a series of large limestone quarries (Figure 10). At Brierlow Bar we turn right onto the A 5270 and follow this road to the A6. After turning left onto the A6 we will view the allogenic valley of the River Wye and Tunstead Quarry at **STOP 2**.

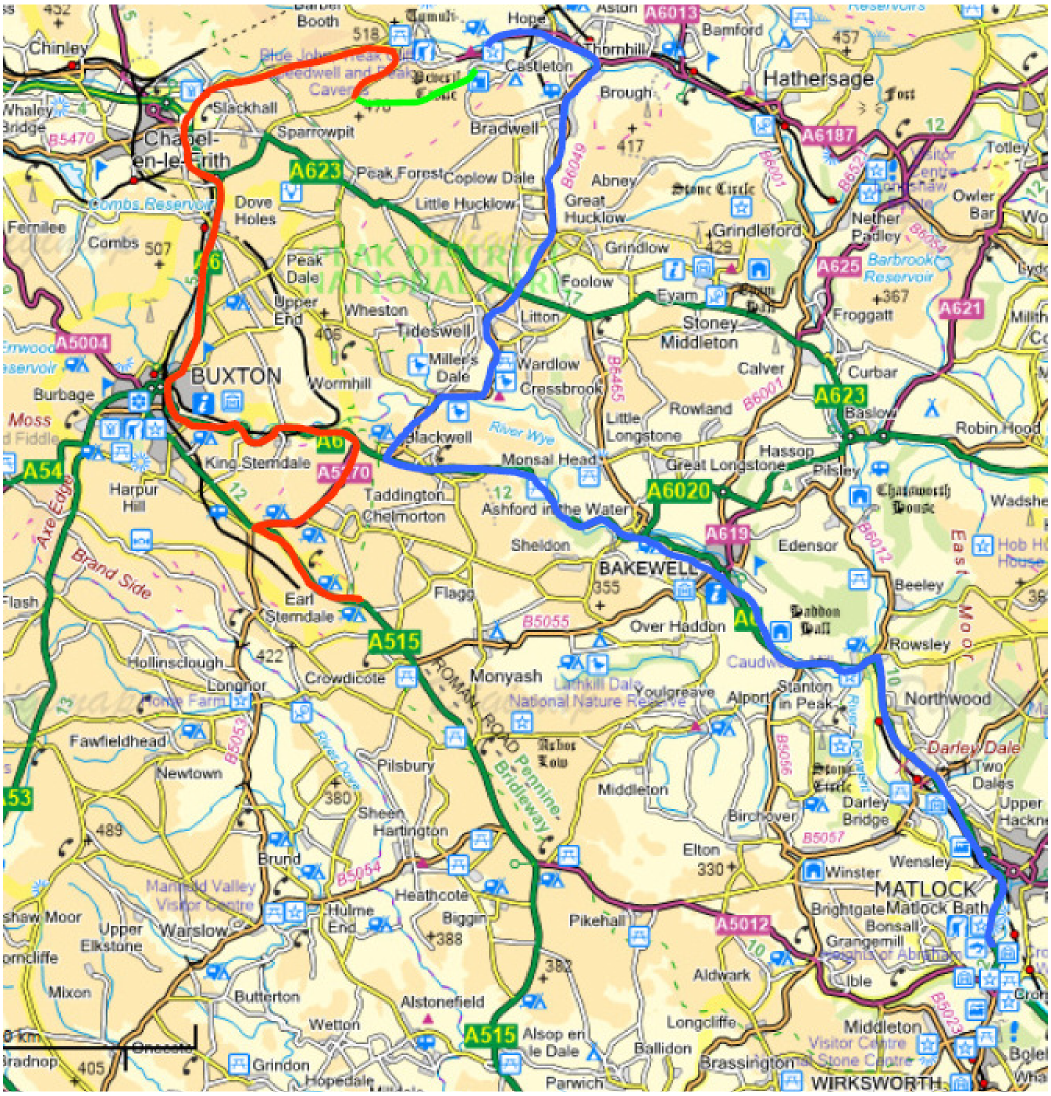

Figure 9. Red line shows route of coach from Hurdlow (Stop 1) to Eldon Hill Quarry (Stop 3)
Green line shows route of walk from Eldon Hill Quarry to Castleton
Blue line shows route of Coach from Castleton to Matlock
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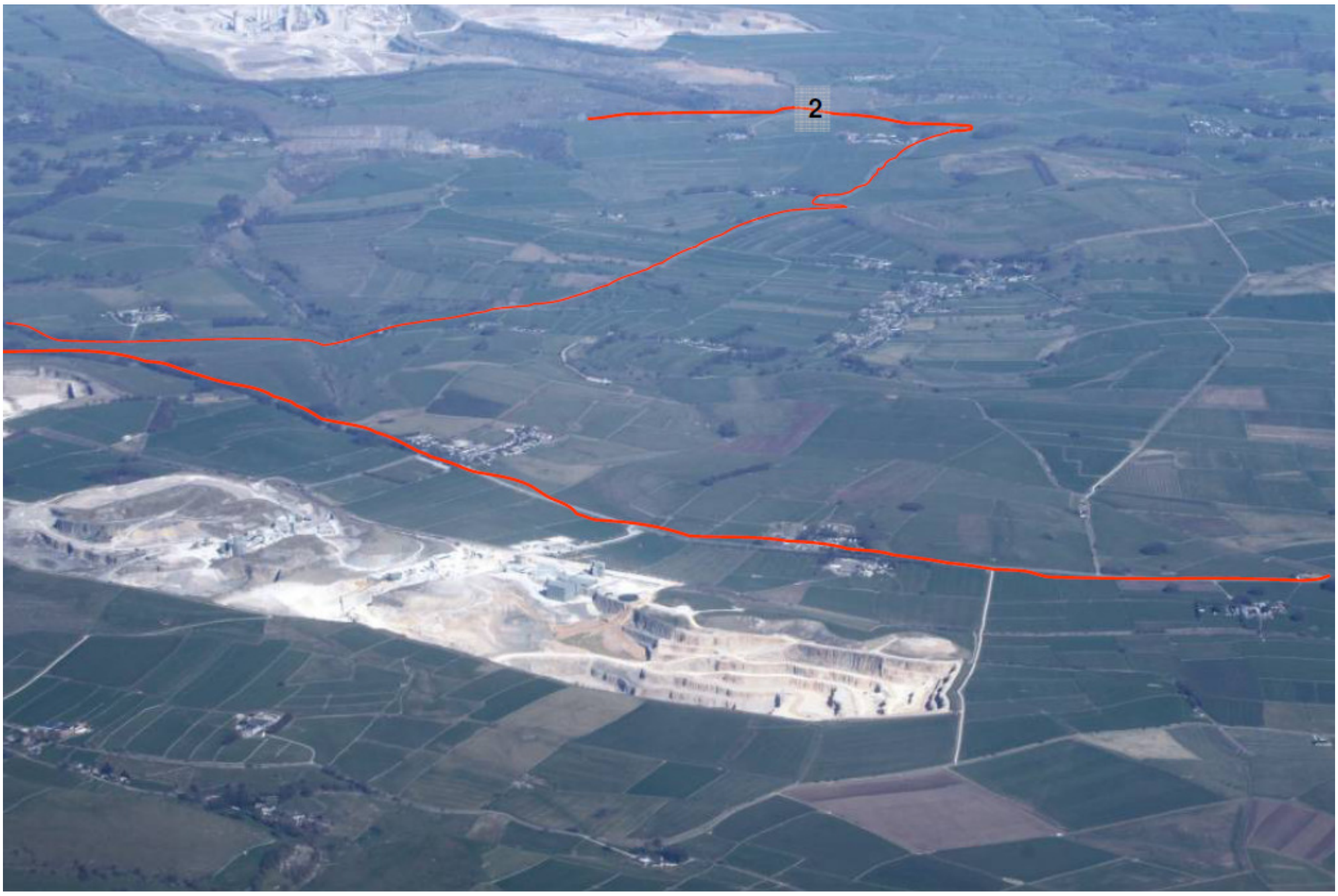

Figure 10: Red line shows route of the excursion west from Stop 1 (Hurdlow is to the right of the photo) past the Dowlow and Hindlow Quarries to Brierlow Bar (to the left of the photo). Then northeast to Stop 2 which overlooks Tunstead Quarry part of which is visible at the top of the photo. [Photo by John Gunn]

STOP 2: THE RIVER WYE AND TUNSTEAD QUARRY

The River Wye has its headwaters on Namurian Millstone Grit Group strata to the north and south of Buxton but is groundwater-fed to the east. At this stop we look down into the deeply incised valley along the reach known locally as Wye Dale and north-northwest up the tributary dry valley of Great Rocks Dale. The characteristic daleside landform sequence of bare and vegetated screes, rock headwalls and rock buttresses (Figure 11) is well illustrated together with human impacts on those landforms in the form of quarrying. Tunstead Quarry (Figure 12) is one of the largest in Britain, removing up to 5Mt of limestone per annum.

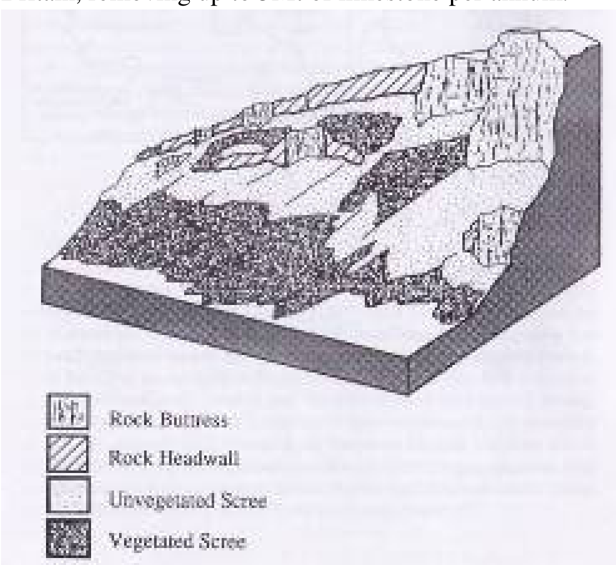

Figure 11: Characteristic White Peak daleside landform sequence (from Gunn et al., 1992)

Figure 12: View of Tunstead Quarry (to the left) and the linked Old Moor Quarry. The two quarries are separated by Great Rocks Dale. The red star shows the approximate location of Stop 2 and the location to the immediate northwest of the star is the former Central Quarry, now reclaimed by infilling with quarry waste products. Ashwood Dale Quarry which the coach will drive past is in the middle foreground. [Photo by John Gunn]

Quarries such as Tunstead are very instructive in providing a view of the vadose zone at a larger scale than is possible in caves. There are several kilometres of limestone exposed on benches that extend to almost 100m below the original land surface. The uppermost benches expose the epikarst and provide cross-sections through dolines (Figure 13) and dissolutionally enlarged joints (Figures 14 and 15A). However, there is little evidence for the development of large conduits on lower benches (Figure 15B) and these benches remain dry despite the water surface elevation in boreholes on the periphery of the void being commonly metres, and even tens of metres, above the elevation of the floor. This shows how randomly-drilled boreholes (and especially shallow ones) may fail to intersect the main drainage pathways in the Carboniferous Limestone.

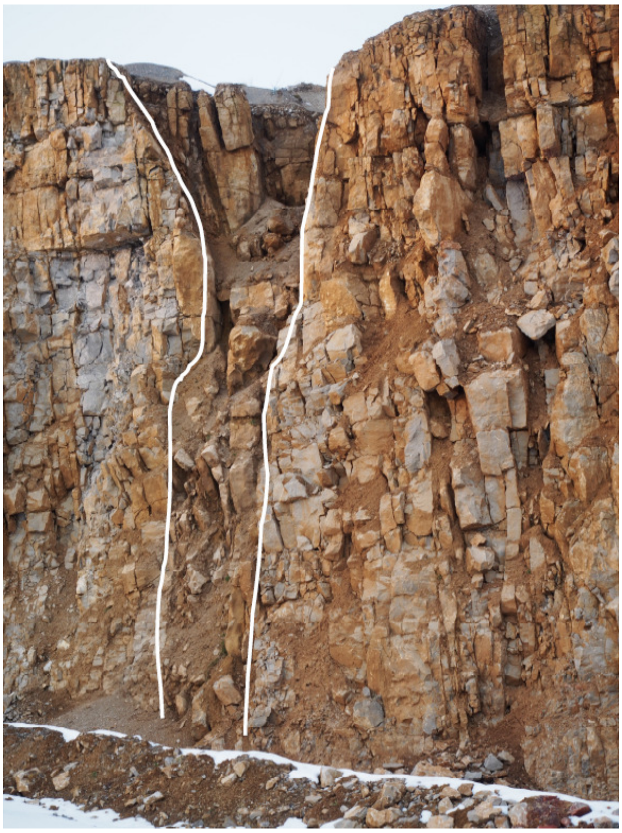

Figure 13: Cross-section through a doline (from which the fill has subsequently slumped) and underlying joint in a White Peak quarry face. The top of the face is at the pre-quarrying land surface. (Photo by John Gunn)

Figure 14: Laterally extensive dissolution feature that formed on a prominent inception horizon about 20m below the original land surface and was intersected during quarrying (Photos by John Gunn).

Figure 15

A [<u>Left</u>] shows a prominent joint that had been enlarged by dissolution and filled with sediment that slumped out following exposure during quarrying. The top of the quarry face is close to the original land surface.

<u>B [Right]</u> shows four quarry benches, each about 20m high. The quarry intersected a doline from which sediment slumped out forming the fan on the top bench. The conduit beneath the doline is clearly visible in the second face but has markedly reduced in size by the third face and is totally absent on the lowest face. It is considered likely that the vertical conduit drained to a laterally extensive conduit system probably associated with the prominent bedding plane near the base of the second face. Below that level there was little dissolution and only very slow groundwater circulation. (Photos by John Gunn)

Continuing west from STOP 2, the A6 runs alongside the River Wye through Ashwood Dale to Buxton passing the Topley Pike and Ashwood Dale limestone quarries and two contrasting south-bank springs. The first, Ashwood Dale Rising, is intermittent, drying up in summer but with a strong discharge in winter. Dye tracing has shown that it is partly fed by water that sinks in Chelmorton (Banks et al., 2009). In contrast, the Rockhead (aka Cowdale) spring, a few hundred metres up the valley, is perennial, has a low flow variability, consistent chemistry and proven absence of bacterial contamination, and has been recognised as a natural mineral water suitable for bottling without any treatment.

From Buxton the A6 trends north, passing the Buxton Mineral Water bottling plant on the right. Water from the Buxton thermal springs, at 27ºC the second warmest in Britain after Bath, is pumped some 3km to the bottling factory. The springs (Figure 16) have been shown to discharge water that has a distinctive chemistry and is elevated in ^{87}Sr above local limestone groundwaters indicating that it has followed a deep (c. 1500m) flow-path through locally confined, sandstone aquifers to the west before rising via a high permeability pathway in the limestone (Figure 17; Gunn et al., 2006).

Figure 16: A small proportion of the water from the Buxton thermal springs is discharged via The Lion's Mouth (St Anne's Well) but the bulk is pumped for bottling or used in a thermal swimming pool (Photos by John Gunn)

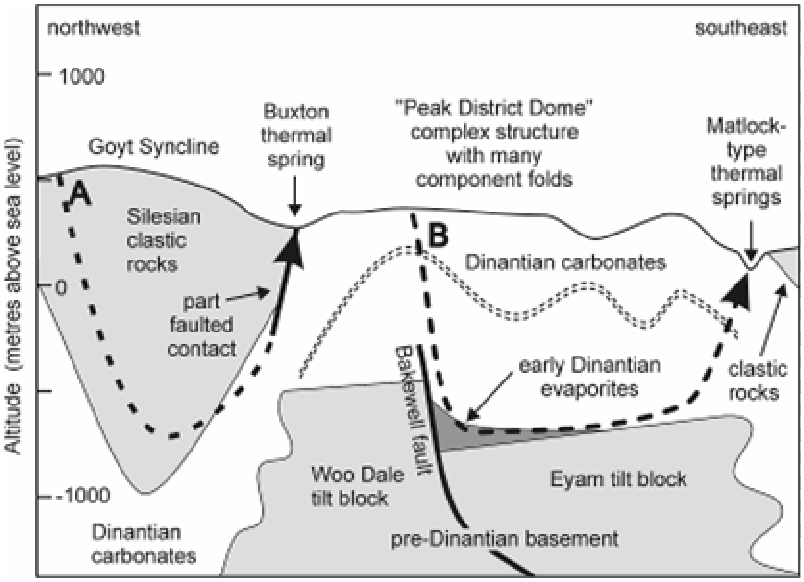

Figure 17: Composite conceptual model of the major geological components and deep thermal groundwater flow between the Goyt syncline in the west and the Derwent valley in the east showing the two main thermal spring types. All topographical and geological relationships are greatly simplified. The nature and position of the Bakewell Fault and related basement blocks (Smith et al. 1985) are speculative. Vertical exaggeration about ×20 (From Gunn et al., 2006)

In Dove Holes village the A6 crosses above a railway tunnel that was driven in the 1860s. The tunnel intersected strong springs subsequently shown by water tracing to be derived from streams that flow from Namurian strata and sink in the village. This water would originally have flowed to springs on the banks of the River Wye and would eventually have been discharged into the North Sea. However, the railway engineers diverted the water out of the northern tunnel portal and it now finds its way to the Irish Sea.

Near the village of Chapel en le Frith we leave the A6 and take a minor road that trends northeast to enter the Castleton drainage basin, the most studied part of the White Peak and the most important caving area with over 25km of explored conduit. To the north the limestones dip steeply beneath the Namurian sandstones and mudstones of Rushup Edge, the lower slopes of which are overlain by Quaternary solifluction deposits. A series of streams with a combined catchment of c.5 km^2 flow over these deposits and sink at 15 discrete points in marginal reef limestones (Figures 18 - 20). Eight of the stream-sinks are associated with accessible influent caves (Figures 18 & 21) including the 240m deep Giants - Oxlow cave system.

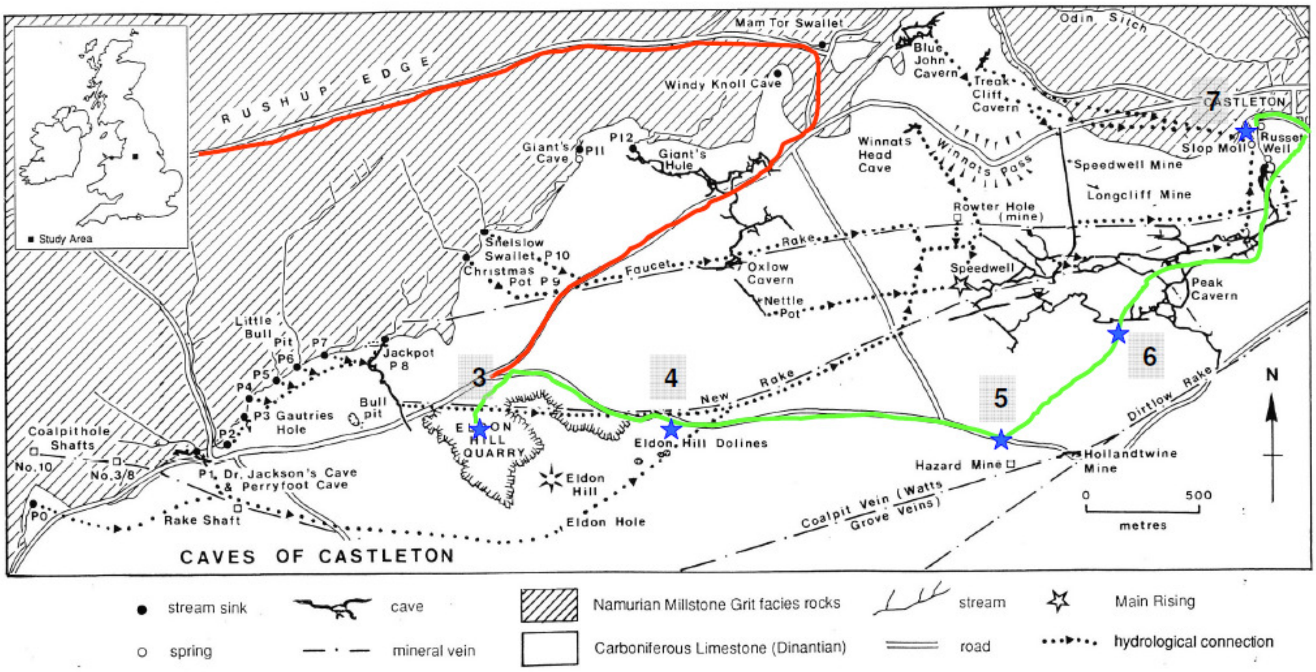

Figure 18: The Castleton karst. Red line shows route to be followed by coach along Rushup Edge to Stop 3. Green line shows footpath from Stop 3 to Stop 7 (after Gunn, 1991).

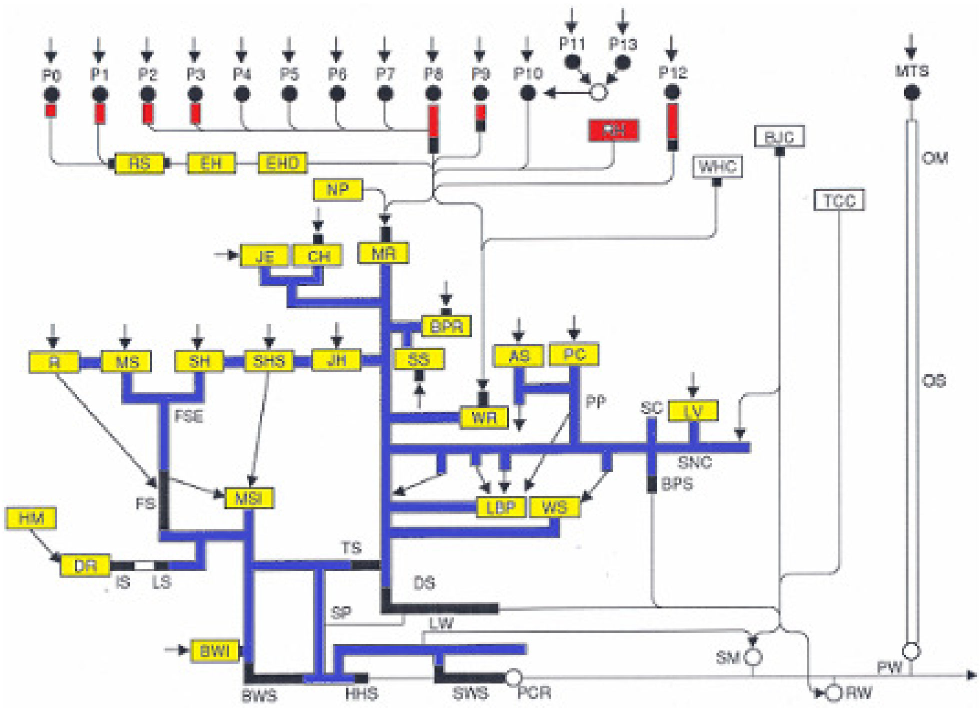

Figure 19: Hydrological links in the Castleton groundwater system.
Black circles are point inputs from allogenic stream-sinks. Red boxes show accessible sections of influent cave. Black boxes are sections of phreatic cave passage explored by divers. Blue boxes are sections of vadose cave passage. Thin lines show linkages proved by water tracing (after Gunn, 1991; key below)

Figure 2. Summary of drainage routes within the Castleton karst, based on water tracing experiments (Gunn, 1991). Key: AS, Assault Course; BPS, Bottomless Pit Sump; BRP, Bathing Pool Rising; BJC, Blue John Cavern; BWI, Buxton Water Inlet; BWS, Buxton Water Sump; CCI, Cliff Cavern Inlet; CH, Cliff Hanger; DR, Dooms Retreat; DS, Downstream Sump; EH, Eldon Hole; FS, Far Sump; FSE, FS Extension; HHS, Halfway House Sump; HM, Hollandtwine Mine; IS, Ink Sump; JE, Joint Evert; LBP, Long By-Pass; LV, Longcliffe Vein; LW, Lumbago Walk; MR, Main Rising; MS, Minor Sump; MSI, Main Stream Inlet; MTS, Mam Tor Swallet; NP, Nettle Pot; OM, Odin Mine; OS, Odin Sough; PO to P13, Rushup Edge Swallets; PCR, Peak Cavern Resurgence; PP, Pilkington's Passage; PW, Peakshole Water; R, Rasp Stream; RH, Rowter Hole; RS, Rake Shaft; RW, Russett Well; SC, Speedwell Cavern; SH, Stemple Highway; SHS, SH Sink; SNC, Speedwell Near Canal; SM, Slop Moll; SP, Speedwell Pot; SS, Secret Sump; SWS, Swine Hole Sump; TCC, Treak Cliff Cavern; TS, Treasury Sump; WHC, Winnats Head Cavern; WR, Whirlpool Rising; WS, Window Sump

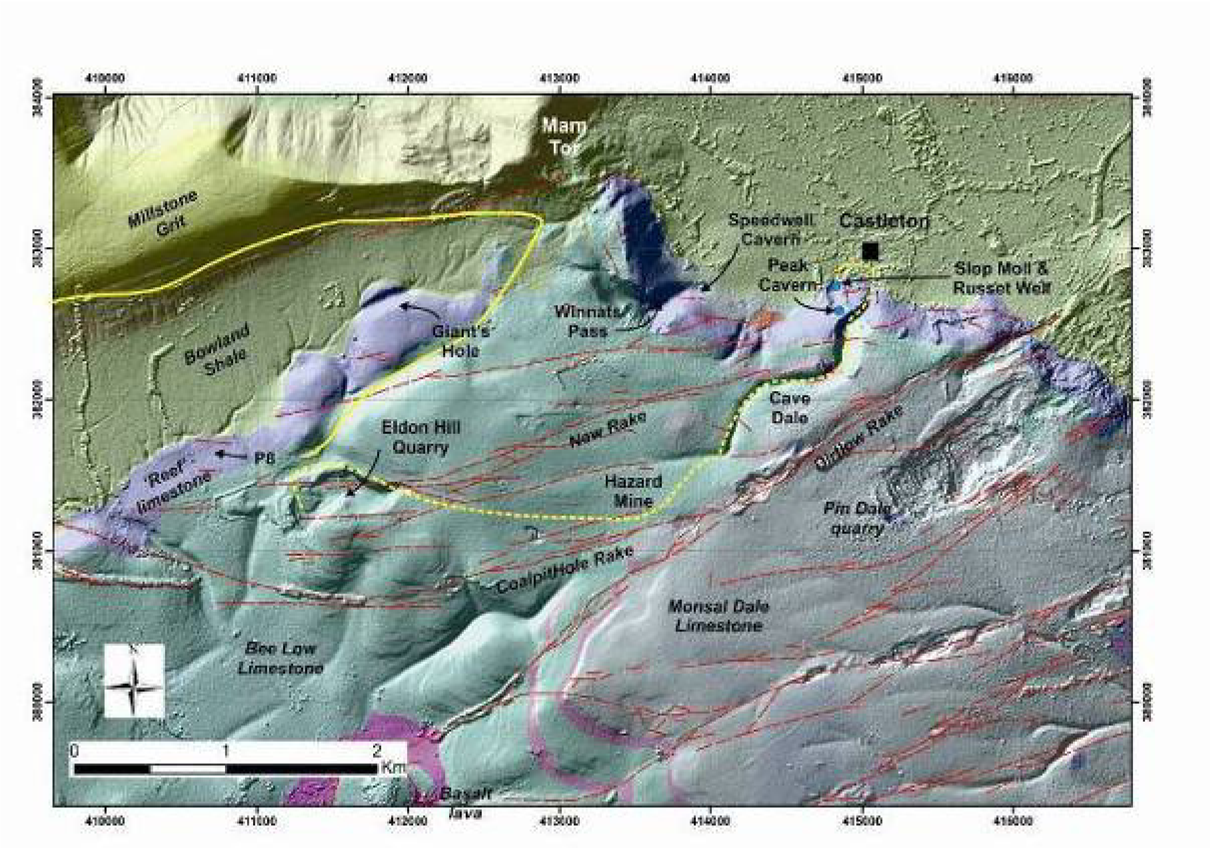

Figure 20: Geology and topography of the Castleton area. Solid yellow line is route followed by coach. Dashed yellow line is route of walk. (Based upon 1:50,000 scale DigMap bedrock geological maps, with the permission of the British Geological Survey. NEXTMapTM Britain elevation data from Intermap Technologies.)

Figure 21: The P1 stream sink and entrance to Perryfoot and Dr Jackson's caves (Photos by John Gunn)

The stream-sinks and influent caves drain to flooded conduits (sumps) which have been explored by divers, the deepest being the East Canal sump in Giant's Hole (-25m). Repeated tracing experiments (Gunn, 1991 and unpublished) have shown that water from the sinks enters Speedwell Cavern through two flooded conduits, Main Rising (MR) and Whirlpool Rising (WR), Figure 22). There is some question as to the vertical accuracy of the cave surveys that were completed under difficult conditions, but the water surface elevations at the inlet sumps appear only marginally higher than those of the outlet sumps although WR is shown as being about 8.3m higher than MR (Figure 23).

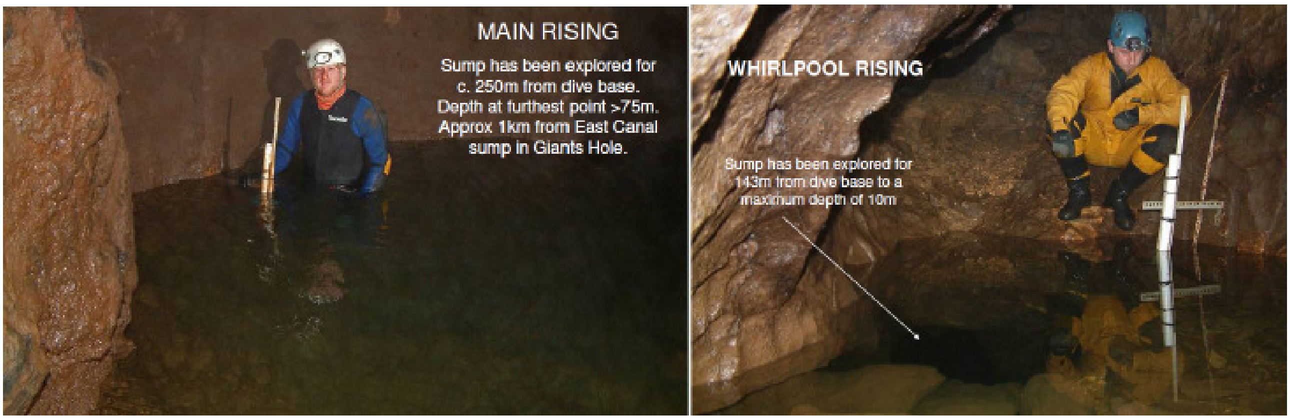

Figure 22: The two main inlets to Speedwell Cavern (Photos by Nigel Ball)

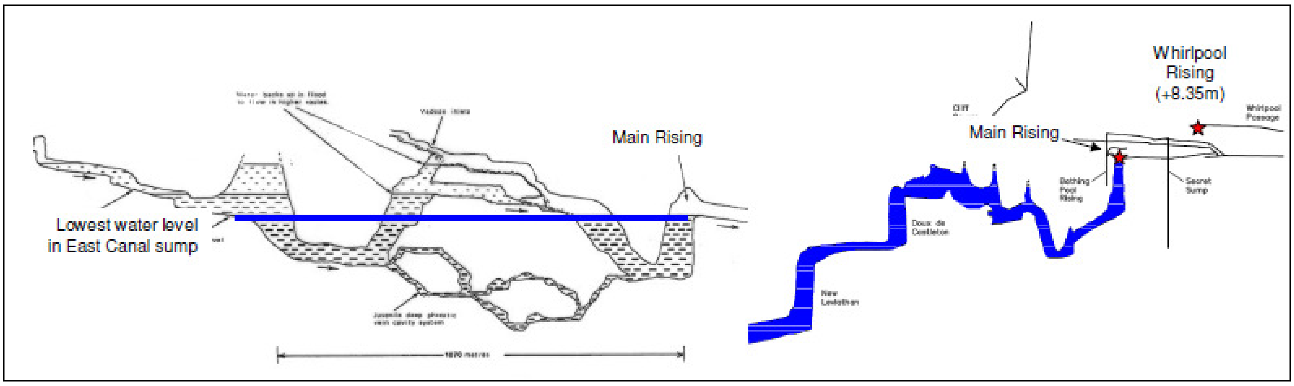

Figure 23: Left shows a conceptual model of the conduit system between the Giant's Hole East Canal inlet sump and Speedwell Cavern Main Rising compiled by Beck (1980) before exploration by divers. Right is an unpublished survey of the main Rising sump by John Cordingley. The horizontal distance to the limit of exploration is about 200m from the dive base and the depth at that point was over 75m below dive base

The MR and WR streams flow through open conduit and combine at ‘the Whirlpool’, which is about 290m downstream of each rising. There is a further 810m of open streamway (best conceptualised as a surface stream with a roof) after which a flooded conduit leads after about 550m (straight line) to the Slop Moll (SM) and Russet Well (RW) springs.

In addition to the allogenic catchment there is an autogenic catchment of about 8.4 km^2 in which the limestones are overlain by soils of loessic origin. Part of the recharge onto this area drains to MR and WR and to percolation-fed inlets in Speedwell Cavern and the remainder drains to Peak Cavern, the majority entering via two flooded conduits, Ink Sump (IS) and Far Sump (FS). FS has been dived and found to connect with Speedwell Cavern although the main water inlet is downstream of the connection point. The water discharged by the sumps flows through open cave passage and two sumps one of which has been completely explored by divers and one partly explored. The second sump discharges into the Peak Cavern gorge and the emergence is commonly known as Peak Cavern Rising (PCR).

The Peak-Speedwell Cave System is characterised by large and mature conduits developed on several levels (Ford, 2008). Over 16km of active and relict cave passages have been surveyed, including over a kilometre of permanently water-filled cave passage explored by divers. In addition to percolation inlets the system has around 50 inlet streams, 20 of which flow from explored sumps. The present day (active) Speedwell conduit has developed on a lower inception horizon to that on which the present day Peak conduit has formed and at one point the Peak Cavern stream is directly above the stream that flows through the Speedwell conduit but is separated by several metres of rock. Such crossing of underground streams at different levels is relatively common in karst areas but impossible to model using porous media assumptions. Although the waters do not mix at the crossing point the lower part of the Speedwell conduit is unable to transport all the flow under conditions of high recharge and the water backs up and enters the Peak conduit at a number of discrete points, eventually emerging from PCR. Hence, during low to medium flow conditions PCR only discharges water from the autogenic catchment but during higher flows it also discharges water from the allogenic catchment. This is discussed further at **STOP 7**. Figure 19 provides a summary of the hydrological linkages in the Castleton karst based on exploration and water tracing. Although complicated the diagram is a simplification of an even more complicated reality that provides the ‘conditioning’ for the hydrological ‘signal’ recorded by monitoring the three output springs. In many cases the system behind a spring is unknown and this example shows how difficult it can be to draw meaningful conclusions purely on the basis of output analysis.

As shown on Figure 18, the road runs to the north of Windy Knoll Cave, a short length of ancient relict cave, and to the south of Mam Tor Swallet, a small sink that has been captured by a lead mine drainage level, Odin Sough. After the swallet the road heads south and then, at the upper entrance to the Winnats Pass (Stop 8), southwest, crossing over part of the Giant's Hole - Oxlow Caverns system before reaching the now closed Eldon Hill Quarry [**STOP 3**] from where we walk to Castleton.

STOP 3: ELDON HILL QUARRY

Eldon Hill Quarry began operation in the early 1950s and after two unsuccessful planning applications for extensions finally closed in 1999. This was two years after expiry of the Planning Permission to extract stone but the company had exploited a loop-hole that allowed stone that had already been released by blasting to be removed from the site. During the period of operation over 90 fragments of cave passage that formed part of a complex network of partially or totally sediment-filled caves were intersected and in a number of cases totally destroyed. The longest section of passage known to have been destroyed was in Alsop's Cave which was exposed by quarrying in December 1994 (Figure 24).

**Figure 24:
The entrance to Alsops Cave in December 1994 shortly after it was intersected by the quarry**

A total of 208m of passage was explored and surveyed (Figure 25 left) and at the time this was the highest elevation relict phreatic passage (up to 4m diameter but largely sediment filled) in the White Peak. Despite containing ancient sediments and fine speleothem (Gunn & Beck, 2002) most of the passage was removed during the final frenzy of quarrying prior to expiry of the permission to blast. However, in 2002 cavers located an entrance that was clearly a continuation of the "Escape Route" in Alsop's Cave (Figure 25 left) and this was named Sidetrack Cave (Figure 25 right). Comparison of the left and right surveys on Figures 25 shows the extent of cave passage removed by quarrying, but almost 480m of cave passage can still be explored in Sidewalk Cave. In 2010, following digging in the southwest corner of the Quarry, Convenience Cave was entered and 130m of passage was explored (Figure 26).

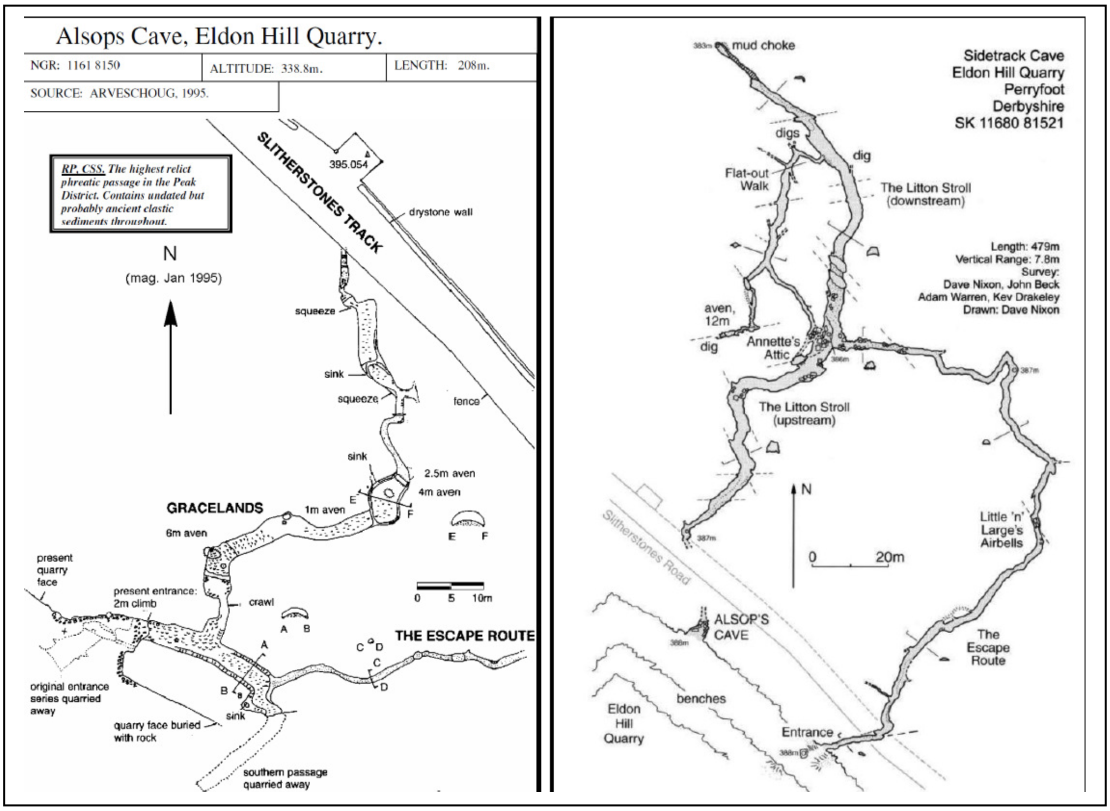

**Figure 25:
Left is a survey of Alsop's Cave in 1995 (from Gunn & Beck, 2002).
Right is a survey of Sidetrack Cave from**
www.PeakDistrictCaving.info

Figure 26: Entrance to Convenience Cave (top left and right) and typical passage showing extent of sediment fill (photos by John Gunn)

The sections now exposed in the quarry can be divided into two groups, one located in the western end of the quarry at 360-380 m aOD, and another group along the northern side of the quarry at 380-400 m aOD developed on the line of the New Rake mineral vein (including Alsop's Cave and Sidetrack Cave). The caves exposed in the quarry represent a truncated part of a relict conduit system between stream sinks draining the Namurian mudstones and sandstones to the north and springs in the Castleton area.

Many of the caves in Eldon Hill Quarry display characteristic features suggesting they were formed by paragenesis (Farrant and Smart, 2011) during repeated phases of sediment input. These caves, some of which are sinuous narrow canyons up to 12 m high, typically have smooth phreatic walls, sometimes with occasional phreatic pendants and large elongate concave grooves discordant with the bedding (Figure 27) and are nearly all totally choked with sediment. The sediment fills typically consist of an alternating sequence of coarse poorly sorted, chaotic, sandstone- or chert-rich gravels, interbedded with coarse sands and silts, locally with well developed cut and fill structures. Many sections are capped by finer grained silts and clay. The sandstone clasts are generally large (up to 30cm in diameter), sub rounded, and usually composed of immature arkosic micaceous sandstone derived from Rushup Edge, with minor amounts of shale, quartz and chert. Analysis of the sediments suggests the coarse gravels were emplaced during cold periglacial periods under 'sliding bed' conditions, similar to gravel deposits in eskers. The lack of any exotic clasts, coupled with the limited size of the cave passages, indicates the palaeocatchment area did not extend much beyond the present area and also suggests there was no erratic rich glacial till within the catchment area which could have been reworked.

Uranium series dating of flowstones in some of the sections indicated the speleothems were beyond the range of the dating technique. $^{234}U/^{238}U$ isotope ratios which approach unity suggest they may have formed a considerable time before 350ka. Palaeomagnetic samples were taken from fine grained sediments and speleothems from several of the sections in the quarry. It was found that in several cases reversed polarity samples overlay normal samples which in turn were overlain by reversed samples. This evidence, in conjunction with the Uranium isotope evidence suggests the normally magnetized sediments were deposited during the Jaramillo normal event, 1.06 to 0.9 million years ago. Excluding the caves formed in the Carboniferous the conduits in Eldon Hill Quarry are amongst the earliest in the Castleton area and are probably of a similar age to those recently described in Speedwell Cavern (Shaw, 2015; Sheldon & Wolstenholme, 2015; Worley, 2015).

Figure 27. Section through a paragenetic passage exposed by quarrying, Eldon Hill Quarry. The former passage is marked by brown mud-stained dissolutionally etched limestone. Above and to the left of the figure are several conspicuous paragenetic solution ramps dipping steeply down to the quarry floor on the left of the image. The prominent bedding plane at head height has been picked out by dissolution and displays several examples of paragenetic pendants. Remnants of the sediment fill form the scree slopes upon which the figure (approximately 1.9 m tall) is standing. The passage exits the quarry through a small 1 m diameter tube to the left of the image, which was excavated after quarrying had ceased to reveal Convenience Cave. (Photo by M J Simms).

STOP 4: ELDON HILL DOLINES

As noted above, there are surprisingly few large solution dolines in the White Peak and these three examples on the slopes of Eldon Hill are amongst the largest. They have formed along a prominent mineral vein and played a part in blocking the expansion of Eldon Hill Quarry. In 1984 the quarry operators applied for planning permission to extend the quarry to a point abutting the deepest doline (Doline 1; Figure 28). The area is within the Castleton SSSI (Site of Special Scientific Interest) and the Nature Conservancy Council commissioned an investigation of the doline geomorphology and hydrogeology. A 3.5m deep pit was excavated by hand at the base of Doline 1, the first 0.5m being recent material washed in from a motorcycle scrambling track and the remainder waterworn limestone clasts up to 0.4m long-axis in a clay-silt matrix, probably in part a periglacial loess. A solution containing 2000g of sulpho-rhodamine B was washed into the pit using 18000L water from a bowser which entered over a 1-hour period without any ponding. The trace was undertaken under very low flow conditions and the dye emerged from Russet Well and Slop Moll in Castleton **(STOP 7)** after 11 days, a straight-line velocity of 290m/day. The dye flowed through Speedwell Cavern (Figure 18) and the excavation and tracing experiment were taken as evidence that quarrying in the vicinity of the dolines would be likely to impact on underground drainage in the SSSI and that there was a potential for contaminants to enter. This evidence contributed to the rejection of the planning application, and of a subsequent application, and to the eventual closure of the quarry.

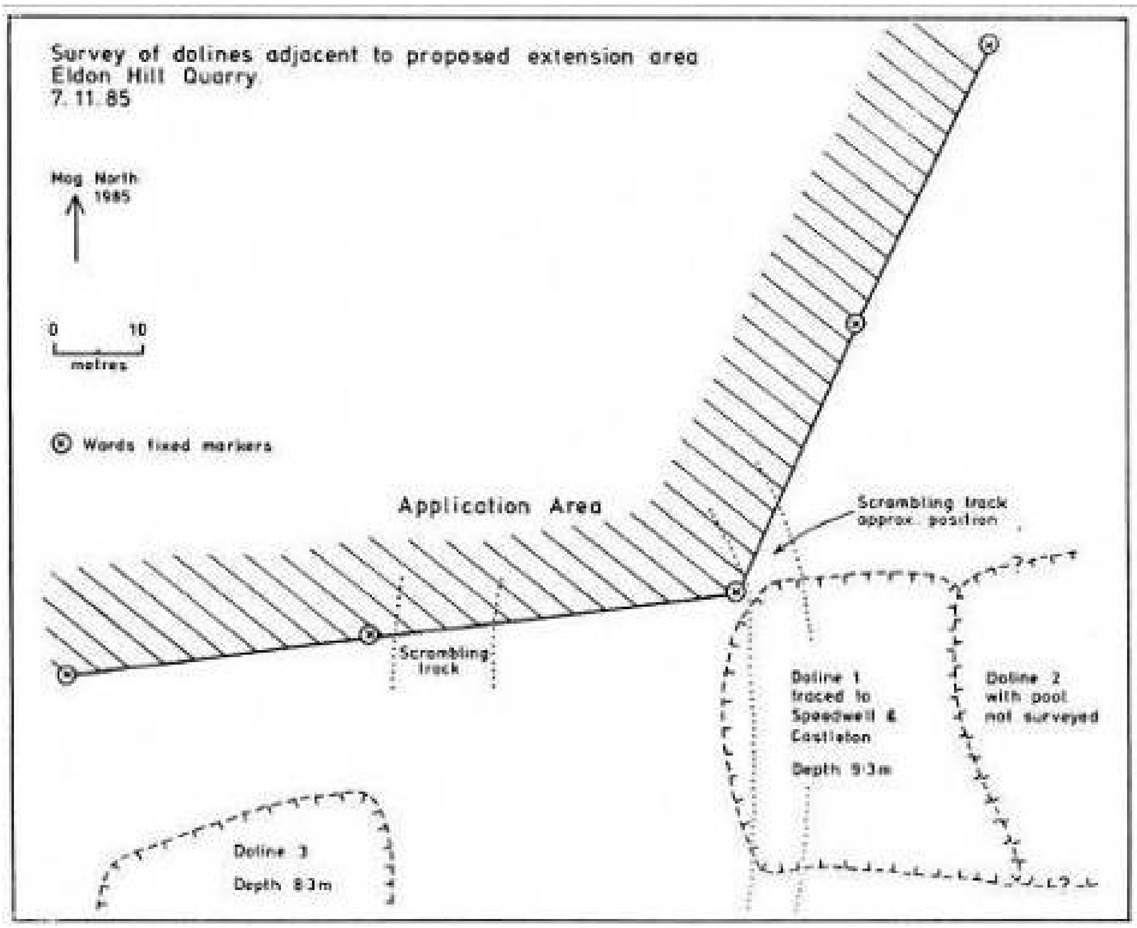

Figure 28: Eldon Hill dolines (unpublished survey by John Gunn)

STOP 5: DIRTLOW RAKE AND DOLINES

In 1987 a tracing experiment by the Limestone Research Group showed that groundwater sinking at the base of Dirtlow Pit, a now-closed fluorspar and limestone open pit quarry, drained in two directions, east to Bradwell and northeast to Peak Cavern (Figure 29). In 1999, cavers reported an orange-red staining in the water emerging from Ink Sump into Lake Passage, a major tributary in Peak Cavern (**STOP 7c**; Figure 18). The water from the sump flows through the Peak Cavern system, and the main streamway was stained orange-red for approximately 1000m downstream of Ink Sump. The area around the emerging water in Lake Passage was covered in a gelatinous biofilm that sloughed off when disturbed. Cave divers confirmed that the material was present in two sumps downstream, Buxton Water Sump and the Peak Cavern Rising sump. Although the latter is the source of the Peakshole Water no visible evidence of the pollution was detectable in the watercourse outside of the cave but the organic pollution had a serious impact on benthic cave invertebrates, resulting in the elimination of most taxa from impacted sites (Wood et al., 2008). Detailed examination of the autogenic contributory area suggested that the most likely source of the pollution was material' being stored on unimproved pasture to the west of the Dirtlow Pit (Figure 29). This material was a mixture of pulp from the production of paper and organic rich peat sludge from a water treatment works. Runoff from the material was similar in colour to that in Peak Cavern and was observed to sink in an area of closed depressions near the head of Cave Dale.

In March 2000, the Limestone Research Group was commissioned by the Environment Agency to undertake a tracer investigation to determine the destination of groundwaters draining from the closed depressions. The first depression, immediately to the west of a north-south trending dry stone wall, was observed to have become blocked such that it contained standing water but a clear flow-line showed how the water overflowed into three other depressions and it was decided to inject the tracer into the final one (Figure 30). To ensure an unequivocal result two tracer dyes were injected simultaneously: 1200g of sodium fluorescein (CI 45350 Acid Yellow 73; 2000mL of solution @ 60% weight/volume) and 500g of Rhodamine WT (CI Acid Red 388; 2500mL of solution @ 20% weight/volume). The depression was pre-flushed with c. 3650L of water from a bowser which was absorbed without any ponding. Water from the bowser was then allowed to flow into the depression and the rhodamine tracer was injected into this water as a slug at 15.30 followed by the sodium fluorescein at 15.33. A further c. 10900L of water from the bowser was released into the depression to ensure the tracers were well flushed in.

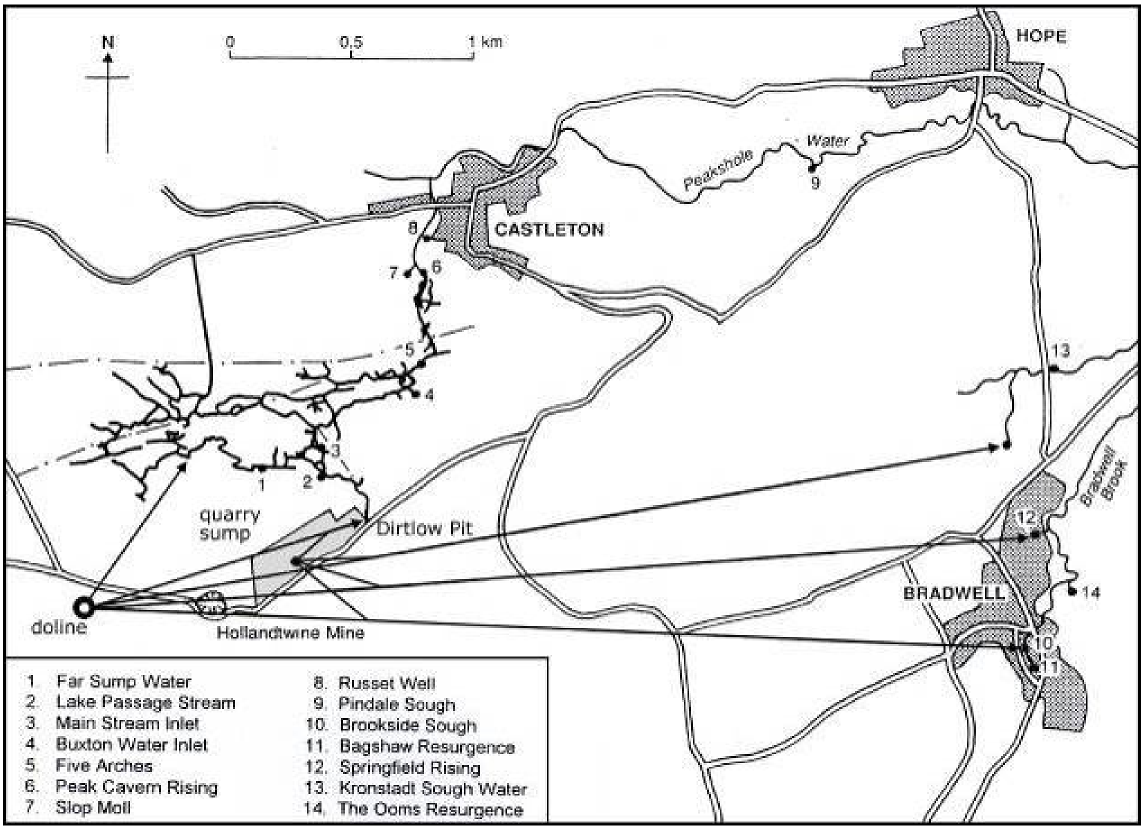

Figure 29: Tracer injection points (Dirtlow Pit and Doline), monitoring points and flow-lines

Figure 30: Tracer injection into Dirtlow Doline

Monitoring was undertaken at 14 surface and underground sites (Figure 29) using granular activated charcoal (GAC) fluocapteurs, and spot water sampling. In addition, an automatic water sampler was deployed on the Peakshole Water downstream of the Peak Cavern Rising where samples were collected at hourly intervals for 96 hours following tracer injection; at 2-hourly intervals from 98-336 hours following injection and then at 4-hourly intervals until 498 hours after tracer injection. The tracer moved in two main directions, eastwards towards Bradwell (sites 10, 12 and 13 on Figure 29) and northwards towards Peak Cavern (sites 1 and 2 on Figure 29; site 3 is downstream of site 1). The travel time to Lake Passage (site 2) was 7-10 days giving a minimum velocity of 155-221 m/day; the travel time to site 3, and by inference to upstream Far Sump, is similar. The travel time to the Bradwell sites was also 7-10 days but the distance is further giving minimum velocities of 387-589 m/day. Most of the pathway between Lake Passage and the Peak Cavern Rising is in open vadose passage but there are two sections of flooded passage one of which has been fully explored by divers and a second partially explored. The minimum tracer velocity was 643 m/day.

Two of the Bradwell recovery points are soughs (lead mine drainage adits) and the third is a watercourse that may be fed in part by a sough. Hence, it is possible that water which originally targeted Peak Cavern via Dirtlow Rake and Ink Sump has been partially captured by soughs driven west from the Bradwell area. However, short sections of cave passage interested by lead mines and quarries between Dirtlow and Bradwell suggest the drainage was divergent before the soughs were driven. Irrespective of the cause, it is clear that the Castleton and Bradwell springs, which are ~3 km apart, have part of their autogenic contributory areas in common.

STOP 6: CAVE DALE

From the area of dolines we walk northeast down Cave Dale, a typical example of a White Peak dry valley. Initially excavated in lagoonal limestones, the valley steepens markedly as it cuts through the reef belt. The proto-valley is thought to have been formed by surface drainage when the Namurian strata now seen on Rushup Edge extended further to the east. The stream cut down into the limestone but lost flow as new swallets opened to the west. Further enlargement took place during the Quaternary cold periods with both meltwater run-off and erosion when the ground remained frozen under periglacial conditions.

The somewhat unlikely entrance to Titan, the deepest natural shaft in Britain at 141m, is close to the summit of Hurd Low on the north flank of Cave Dale. Titan is a large, near vertical, solution cavity similar to others in the Castleton area that developed in association with large mineral veins (Ford, 2000). The mineralising fluids followed, and enlarged conduits that had formed in Carboniferous times and the mineralized areas then formed the inception loci for later conduit development that has overprinted the mineralization (Worley, 2015). Slow dissolution under deep phreatic conditions produced larger cavities whose large-scale morphology has been strongly influenced by the primary fractures. The cavities have no relationship to the present patterns of groundwater movement but one, Ford's Cavern (named in honour of Dr Trevor Ford, the doyen of Peak District limestone geology and speleogenesis), was intersected by at least one later conduit. Ford's Cavern is part-filled by sandy sediments that were derived from erosion of the Namurian strata exposed to the north and northwest and were by transported by fluvial processes (Shaw, 2015).

Titan was initially entered from the bottom in Speedwell Cavern and was explored upwards, initially through an extensive boulder choke and then by an impressive feat of climbing. The roof of the chamber is presently about 30m below the surface (see http://www.daveclucas.com/cms/index.php/caving-trips/9-news/74-titan-laser-scan) and the location beneath the highest point of the hill is fortuitous in that had Titan been formed a few tens of metres away in any direction it would have already been breached by surface lowering. During exploration of Titan digging in a passage fed by a percolation stream led to surface subsidence and it was decided to attempt to excavate a shaft from this point. Following long period of sustained technical excavation the 45m deep shaft was connected to Titan allowing cavers to make a through trip to Peak Cavern via Speedwell Cavern.

Continuing down valley Cave Dale runs across several of the passages in the Peak-Speedwell cave system (Figure 18) but these are all below the Cave Dale Lava and this part of the cave is notable for an absence of diffuse percolation and hence of speleothem. Instead virtually all of the vadose waters entering the cave do so as point inputs where shafts have breached the basalt. The lava crops out about 700m from the bottom of the dale where it shows a crudely hexagonal columnar form and is associated with perched springs. The water flows down the valley for about 100m before sinking on a mineral vein and entering Peak Cavern at Roger Rains House. In the lower part of the dale there are a number of cavities that were enlarged by lead miners. Peveril Castle is visible high on the left side and the dale then narrows markedly before emerging into Castleton village.

STOP 7: CASTLETON SPRINGS AND PEAK CAVERN

Castleton Village is a major visitor centre that takes its name from Peveril Castle, constructed in 1086. To the south and west of Castleton the steep north-facing limestone hillsides are the outer slopes of a complex of marginal reef limestones behind which there is a massif of lagoonal limestones. Drainage is via the Peakshole Water, which flows eastwards, joining in turn the Rivers Noe, Derwent, Trent and Humber before discharging into the North Sea. The Peakshole Water is partly fed by streams that rise on Namurian sandstones and shales but the majority of the flow is from three limestone springs: Russet Well (RW) which has been dived to a depth of 27m, Slop Moll (SM) which is impenetrable and Peak Cavern Rising (PCR) which has been dived to emerge inside Peak Cavern. RW and SM are on opposite sides of the Peakshole Water and PCR emerges just outside the Peak Cavern entrance beneath the eastern cliff (Figure 31 and cover photo). The relationships between the different discharge points are complex and vary with discharge. Under low to medium discharge conditions all of the discharge from PCR is derived from autogenic recharge and the majority of the flow from RW and SM is derived from allogenic recharge. RW and SM have a very similar chemistry and the majority of the water they discharge comes from the downstream sump in Speedwell Cavern. However, autogenic recharge onto an area to the northwest of Castleton, including flow through Blue John Cavern and Treak Cliff Cavern, flows only to RW whilst autogenic recharge that leaves Peak Cavern via the Lumbago Walk sump flows only to SM. Upstream in Peak Cavern the Speedwell Pot sump, thought to be above the conduit from Speedwell cavern, has been traced to both RW and SM so there must be a bifurcation in the conduit between this point and the input form the Lumbago Walk sump (Figure 31).

Under the low to medium discharge conditions the main streamway in Peak Cavern emerges from Buxton Water sump, flows for about 150m along a large (up to 7m diameter) passage to the Halfway House. Here the main passage trends steeply upwards (the Devil's Staircase) but the stream flows east into a much smaller passage that ends in a sump. Water enters from the southeast (Styx Inlet Sump) downstream of which the sump becomes too constricted for exploration. The water is seen again in the Swine Hole Sump and although no progress can be made upstream the sump has been dived downstream to emerge at PCR.

Following periods of intense or prolonged rainfall the discharge down the Speedwell Cavern streamway exceeds the capacity of the flooded conduit that connects to RW and SM and water begins to back-up, eventually overflowing into Peak Cavern at a number of discrete points, including Speedwell Pot and the Lumbago Walk sump where there is a reversal of flow. When this happens PCR starts to discharge both autogenic waters and an increasing proportion of allogenic water. If the input of water continues to increase the capacity of the Halfway House sump is exceeded and water begins to rise up the Devil's Staircase, eventually overflowing and continuing down through the Show Cave and into the Swine Hole passage. When the capacity of the Swine Hole sump is exceeded water begins to back up in the Swine Hole Passage and eventually overflowing out of the normally dry entrance to the cavern.

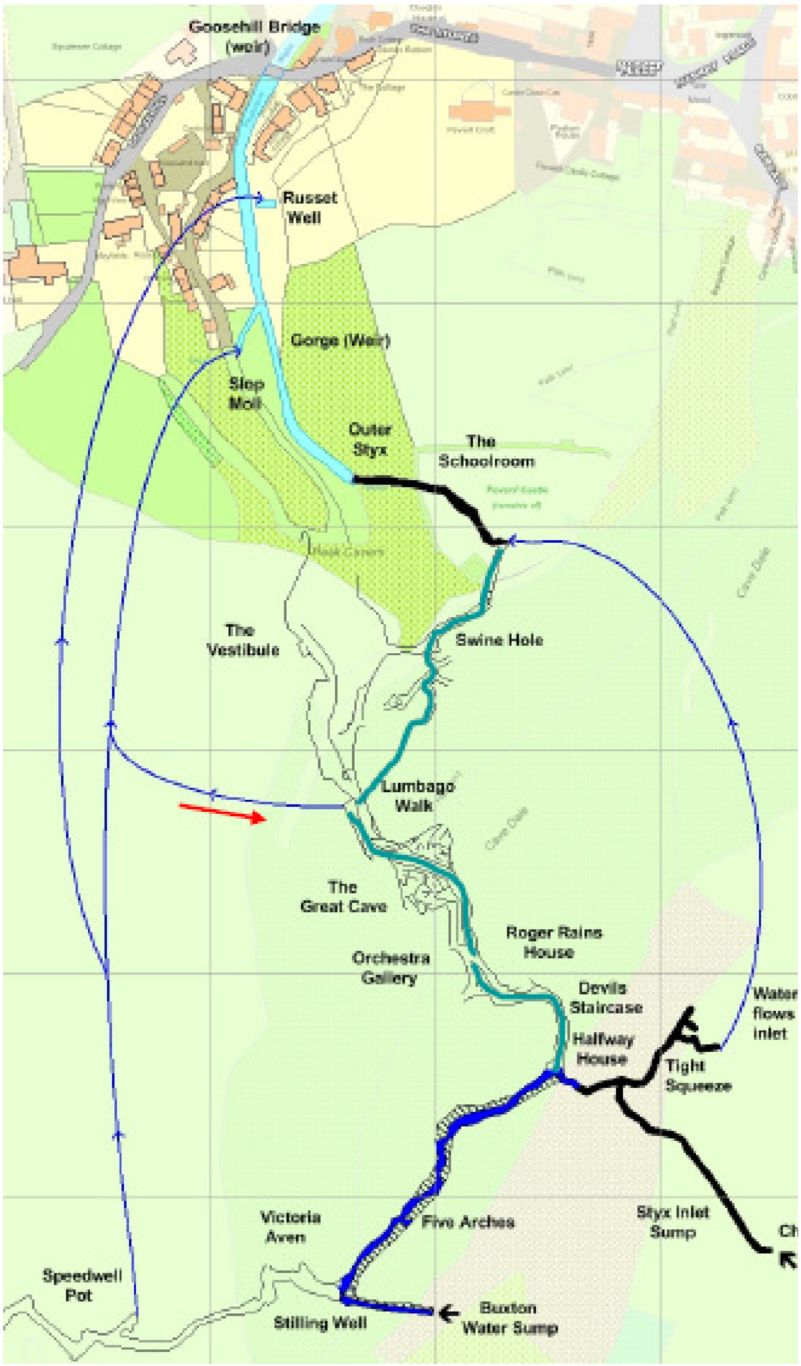

Figure 31: The outer passages of Peak Cavern and the springs feeding the Peakshole Water.
In this area the location of the conduit between Speedwell cavern downstream sump and the springs is unknown

The Peak Cavern 'internal plumbing' shows how complex the conduit networks feeding to springs can be and provides a warning to hydrogeologists that considerable caution is needed when analysing the discharge from impenetrable springs. In an early study of the Castleton karst a weir was installed downstream of Goosehill Bridge (for location see top of Figure 31) and a water level recorder was installed. This site measures the total flow from the Castleton karst (RW, Sm and PCR) and although the records from the pen and ink recorder were not always easy to interpret it was clear that flow was non-linear, exhibiting both (irregular) periodicity and anomalous troughs and peaks. Bottrell & Gunn (1991) suggested that the anomalous events were probably a consequence of flow switching between MR and WR, the two main inlet sumps in Speedwell Cavern. The left panel of Figure 32 shows three anomalous events recorded in 1987 and the top right panel shows a possible model to explain them. The water level at WR has never been observed to drop more than 3m below the lip which is shown on surveys to be 8.35m above the lip at MR. As repeated water tracing experiments (Gunn, 1991 and unpublished) have shown hydrological connectivity between MR and WR some form of permeable sediment barrier upstream of MR is required to maintain the head difference between the two points of emergence. Changes in the permeability of the barrier would affect the head difference between WR and MR and this is modelled in the lower right hand panel of Figure 32. Case A models cases A and B in the left hand panel and it is hypothesised that from MR decreases rapidly due to movement of sediment decreasing the hydraulic conductivity of the barrier. Assuming constant input of water to the system the head behind the barrier will increase (as it does during storm events in other conduits discussed above) and this will produce a small increase the amount of water transmitted through the barrier. After a period of hours further sediment movement in the barrier results in a marked increase in hydraulic conductivity and flow at MR is restored rapidly to a peak after which it declines. It was assumed that the increase in head was not sufficient to result in an increase in flow from WR, and on the basis of the discharge measurements at Goosehill Bridge it appeared that the volume of water 'lost' in the trough was greater then the volume of water 'gained' in the following peak, suggesting that the rise in head caused water to enter smaller conduits and channels from which it was released relatively slowly. Case B in the right panel models Case C in the left panel. In this case it was assumed that following a drop in hydraulic conductivity the head behind the barrier rose until the excess water overflowed into and was discharged from WR.

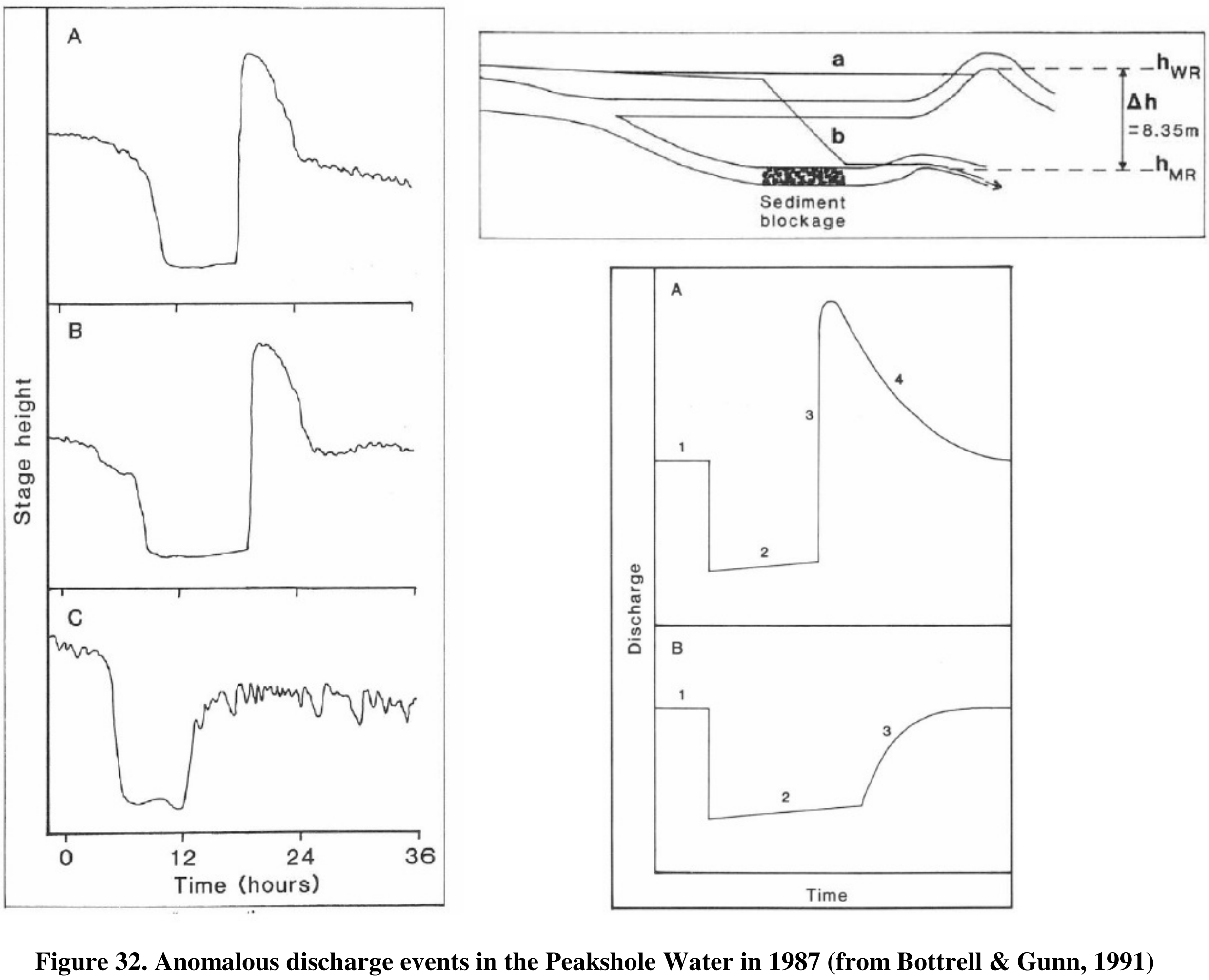

Figure 32. Anomalous discharge events in the Peakshole Water in 1987 (from Bottrell & Gunn, 1991)
Left panel shows events on (A) 4-5 March; (B) 28-29 July; and (C) 19-20 December. Top right panel is a possible explanation of the events and bottom right shows hydrographs based on this model (see text for further detail)

In February 2012 a party of cavers in Speedwell Cavern described in dramatic terms how they had encountered a flood pulse during a period when there had been no rain and water levels had (initially) been relatively low. This prompted a study of water depth at MR and WR using DIVER loggers sampling initially at 1-minute and then at 2-minute intervals. The results were spectacular and the study expanded to include collection of data from PCR, RW, SM and the combined flow of the Peakshole Water (PW) at Goosehill Bridge. Additional temperature loggers were installed to monitor the progress of flood peaks down the Speedwell cavern allogenic streamway. Gunn (2014) has described how some of the temperature data have been used but the bulk of the data remain to be analysed. Figure 33 shows the discharge at PW following an intense summer rainfall event that occurred after a long dry period. The discharge data are hourly averages from a 5-minute data series and the rainfall data are 1-hour totals. There is a remarkably short delay between the onset of rainfall and the discharge response in Castleton, although the rain gauge is located at Sparrowpit at the extreme west of the catchment and it is possible that the storm moved from east to west such that the rain reached the allogenic stream catchments before arriving at the rain gauge. The rapid response is typical of systems where there are flooded conduits through which flood pulses are transmitted instantaneously and the water will also have moved rapidly down the steep sections of vadose streamway between the flooded conduits. The recession limb shows two anomalous events (the second of which is almost identical to the event described by Bottrell & Gunn (1991)) and a period of irregular flow. Although the water depth is measured at RW and SM this cannot easily be converted into discharge but on Figure 34 the 5-minute discharge from PCR (measured at a flat-crested weir) has been subtracted from the 5-minute PW discharge to obtain the combined flow of RW and SM. The depth data from RW and SM confirm that the anomalous discharge events originate at those two springs. It is likely that the flood peak at PCR is at least in part due to overflow of allogenic water from the Speedwell conduit; hence the very rapid recession. Figure 35A shows how the water depth varied at MR and WR over a 24-hour period that encompassed the first anomalous event and Figure 35B shows how the depth varied over a 4-hour period immediately before that event. Both graphs display 1-minute interval data and further illustrate the remarkable complexity of the system.

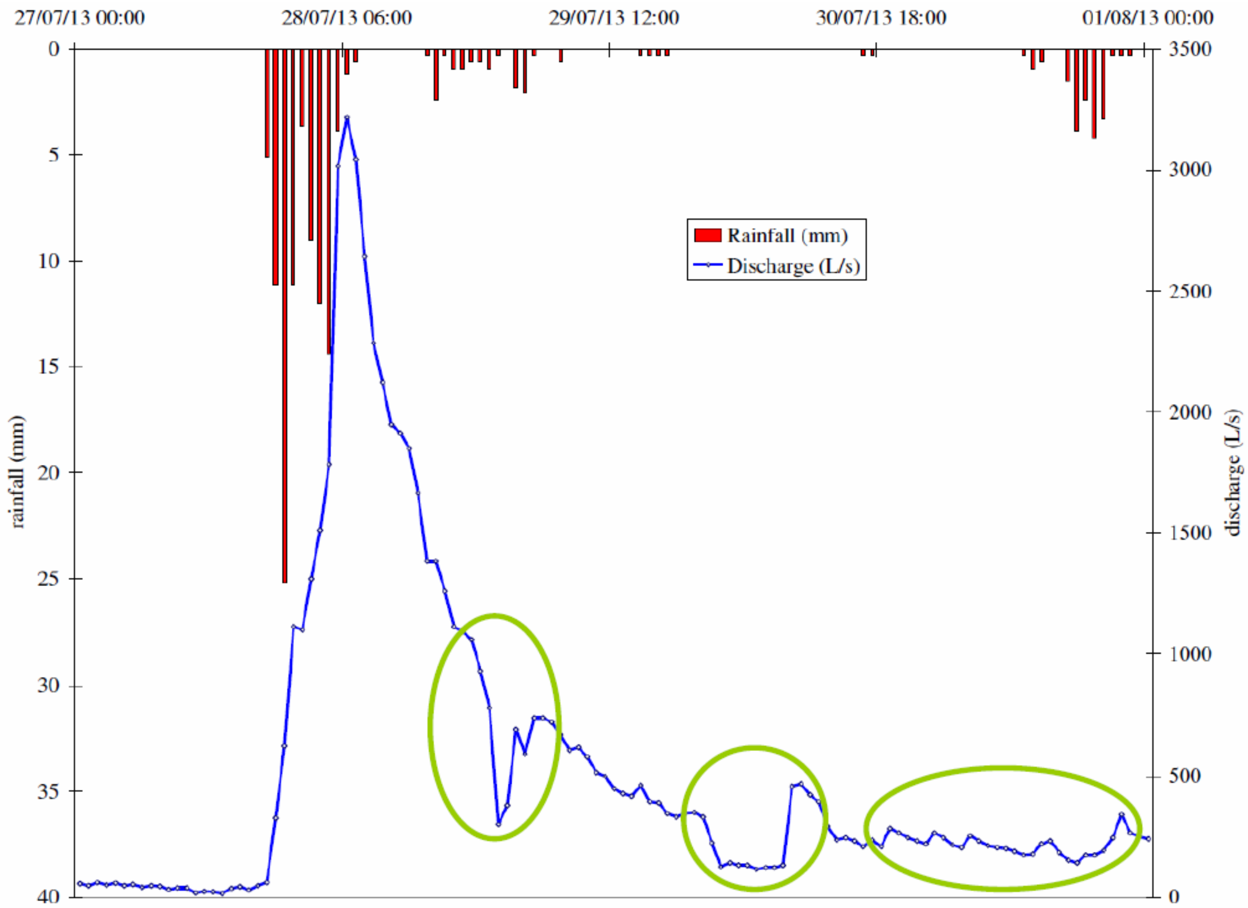

Figure 33: Hourly discharge of the Peakshole Water at Goosehill Bridge (averaged from 5-minute data) and hourly total rainfall at Sparrowpit. Two anomalous events and a period of irregular discharge are highlighted.

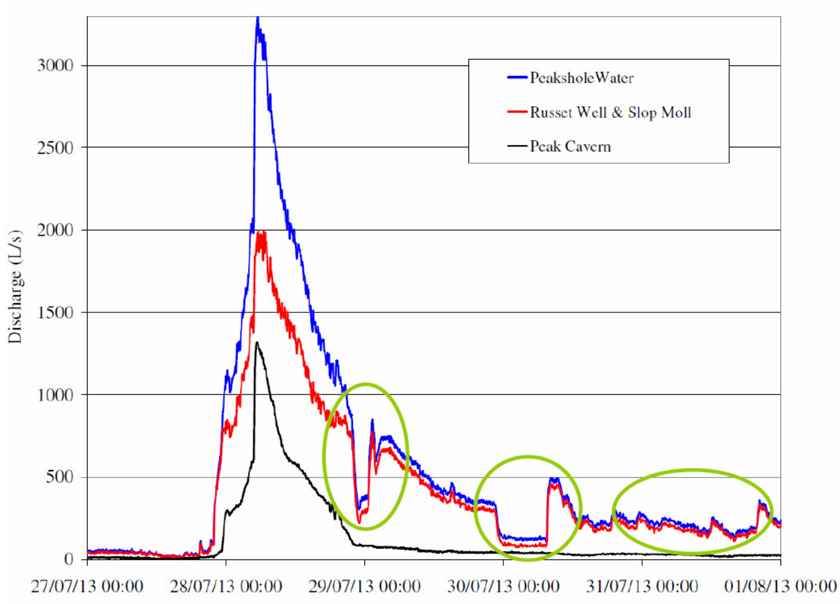

Figure 34: Discharge of the Peakshole Water and tributary springs (5-minute data RW and SM combined flow obtained as (PW - PCR)

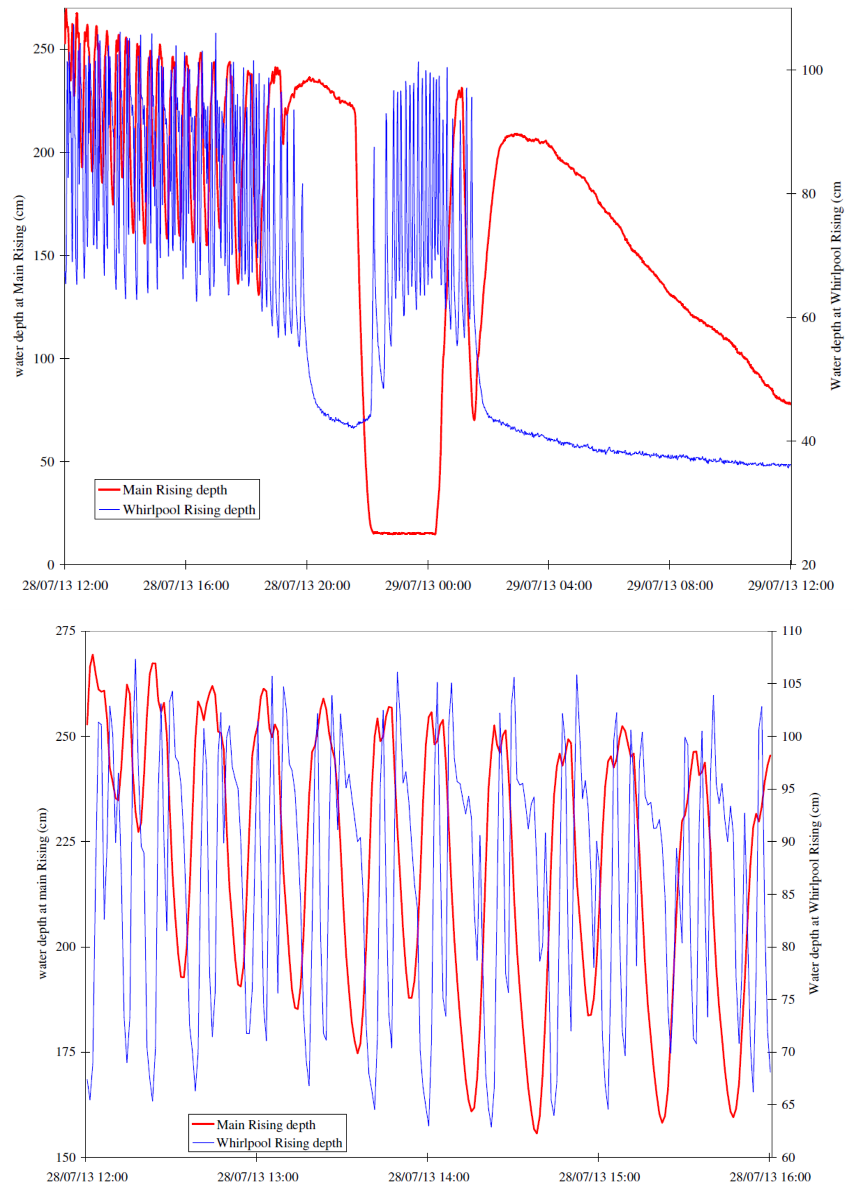

Figure 35: Water depth at Main Rising and Whirlpool Rising (1-minute data). (A) over 24 hour period and (B) over a 4 hour period

Additional insights can be obtained by adding water temperature data to the depth data. For example, Figure 36 shows a 5-hour delay between the rise in water depth and the main rise in water temperature, which reflects the time taken for the (relatively) warm recharge to travel through the flooded conduit. The short very small rise that precedes the main rise may represent rapid autogenic recharge. The minimum length of flooded conduit (that is the distance between MR and the closest input sump) is about 1200m, so the minimum water velocity through the flooded zone is 240m/h. Gunn (2014) provides additional detail but clearly there is a great deal more analysis needed before we can start to really understand this complex hydrological system.

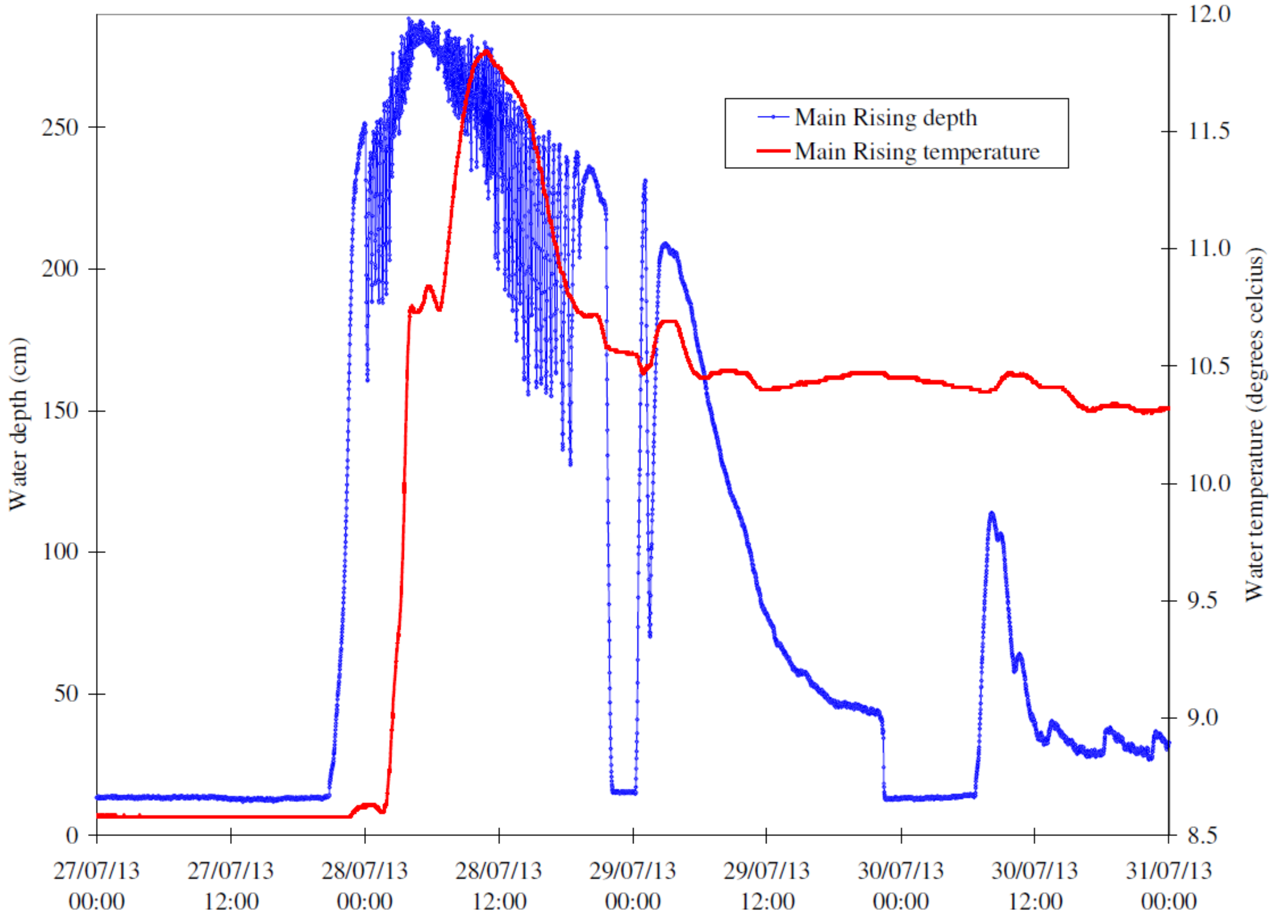

Figure 36: Water depth and temperature at Main Rising, Speedwell Cavern (1-minute data)

Peak Cavern, known for centuries as the "Devil's Arse in the Peak", is one of the four Castleton Show Caves and is notable for having the largest natural cave entrance in Britain. The entrance chamber, "The Vestibule" was once the site of a rope-walk where rope was made by hand and this tradition was revived in the 1990s. The rope-makers had cottages in the Vestibule and the roof is still partly blackened from their chimney smoke. Avens in the roof have massive tufaceous speleothems. After the Vestibule, the cave soon closes down to Lumbago Walk and the Inner Styx Passage which flood to the roof in very wet weather. Beyond, the cave opens up again into the Great Cave, where features in the roof indicate upward solution when it was in the phreatic zone. Sediment slopes rise on both sides and have yielded animal bones from a fissure, now choked, in the floor of Cave Dale. An easy walk leads into Roger Rain's House, with its shower falling from a mineral vein rising into Cave Dale where a streamlet sinks. The tourist route ends at the top of The Devil's Staircase, which descends to meet the stream at the Halfway House (see discussion above).

STOP 8: WINNATS PASS

The Pass itself has generated considerable speculation over its origin. It is a spectacular relict gorge; a dry valley barely 1km long but over 130 metres deep, cut through the belt of marginal reef limestones. Analysis has shown that the Pass was initially a shallow channel cut through the reef belt in mid-Carboniferous times. Later filled with shales of the Millstone Grit Series, it was re-excavated by meltwater running off a snow-and-ice field near Windy Knoll in the later parts of the Pleistocene. In spite of popular stories that it was formed by cavern collapse no evidence has been found to support this claim.

STOP 9: MAGPIE SOUGH TAIL

From Castleton we follow the A6187 east through Hope and turn right (south) onto the B6049. The road passes through Bradwell and ascends a dry valley onto the plateau. After crossing the A623 and passing through Tideswell we descend another dry valley to join the River Wye at Miller's Dale. The road then follows the dry valley of Blackwell Dale and ascends to join the A6. We follow this road south down the Taddington Dale dry valley to rejoin the River Wye which we follow to Bakewell. If time permits we will park and make a short walk to the tail (exit) of Magpie Sough, the last lead mine drainage level to be constructed in the White Peak (completed 1881). The sough drains into the River Wye, supplying up to a third of base flow during times of low discharge. As noted above, the sough acts as an underflow spring and has captured some of the flow from the River Lathkill (Figure 37), changing it from a perennial to an intermittent river over the reach above Bubble Springs.

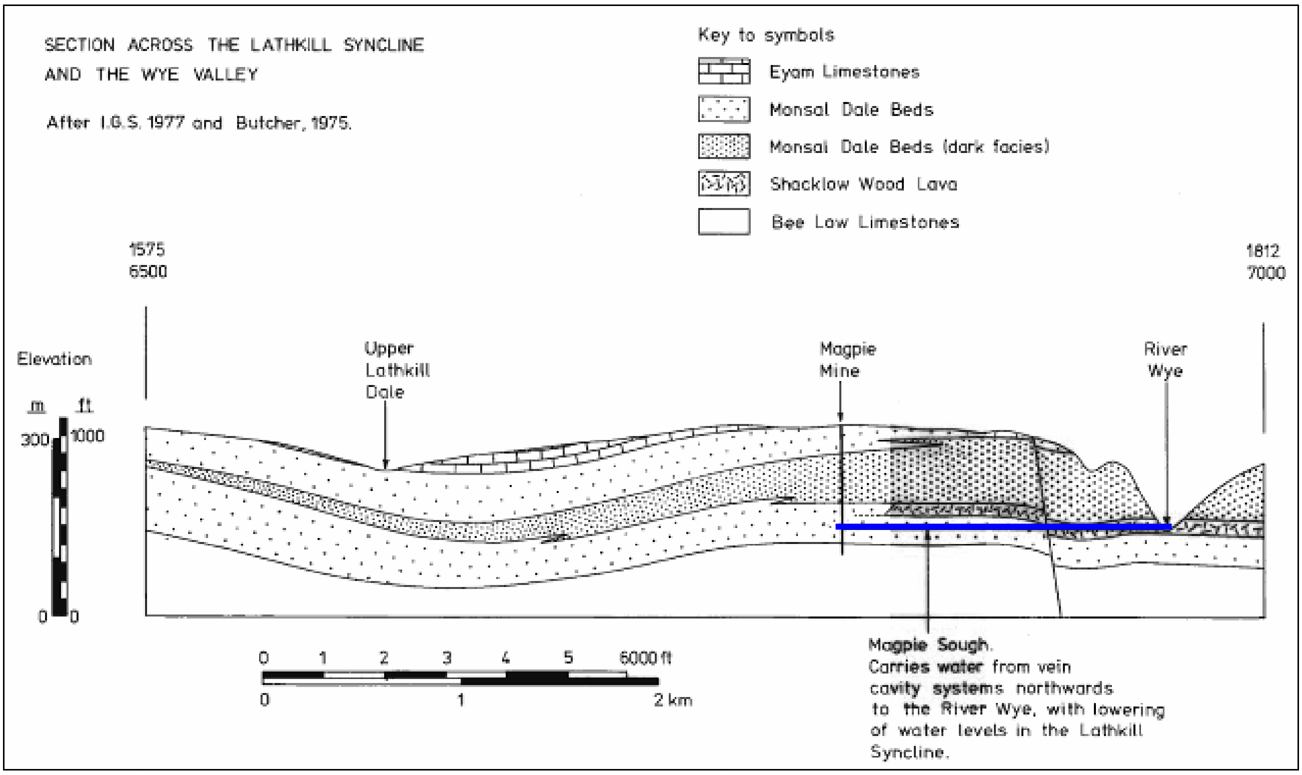

Figure 37: Geological cross-section across the Lathkill syncline and Wye Valley showing the line of Magpie Sough (from Beck, 1980).

After Bakewell the A6 continues past Haddon Hall and the confluence of the River Wye with the River Derwent to Matlock and Matlock Bath (Figure 4), a small spa town lying in the Derwent Gorge The spa developed round a series of warm carbonate springs below which are the finest examples of geothermally warmed tufa systems in Britain. These springs developed a large, coalesced spring-line tufa deposit on the western bank of the Derwent, on which much of the old town is built. Although much reduced and heavily modified by drainage engineering works, the spring-tufas are still forming. The two most active systems, known as the “New Bath” and “Old Bath” springs, form a series of steep-fronted deposits overlying a Holocene tufa deposit in Derwent Park and along the western bank of the Derwent. In terms of its size, rate of carbonate precipitation and raised temperature the Matlock Bath tufa is the only system of its kind in Britain (Rogerson et al., 2015).

In total, five thermal springs are known to have been active at Matlock Bath during the early Holocene (Pentecost, 1999). Three were still active at the end of the 20th Century, in addition to the “Petrifying Well” spring which receives water rising into Wragg Sough and is not a natural spring (Pentecost, 1999). The precipitation rate at the Petrifying Well is about 0.5mm of tufa/year, and although not a natural spring this is probably representative of maximum depositional rates through the system. “Whittakers Spring”, to the north of the town, was reported as being active by Pentecost, 1999, but no evidence of it could be found in December 2014. The two remaining springs, Old Bath and New Bath, both emerge above the Holocene tufa terrace south of the main part of Matlock Bath town. Waters emerge at 18 to 20^{o}C (Pentecost, 1999), and have done so at least since 1772 when they are reported by Thomas Percival (their first mention in the scientific literature) as emerging at 66 and 68^{o} Fahrenheit (Percival, 1772).

The thermal springs were considered to have a meteoric source on the basis of their trace elemental (Edmunds, 1971) and isotopic (Pentecost, 1999) characteristics, despite being geothermally warmed. Gunn et al. (2006) also argued for a meteoric origin and noted that the Matlock thermal springs have elevated sulphate concentrations most likely derived from interaction with deeply buried (~800m below surface) evaporites within the Carboniferous Limestone sequence (Figure 17). There is no chemical evidence for flow below the limestone and the Sr isotopic data do not indicate any significant interaction between the Matlock thermal waters and lithologies outside the Lower Carboniferous carbonate/evaporite sequence.

STOP 10: OLD BATH SPRING (from Rogerson et al., 2015)

Old Bath spring emerges from a natural conduit above a basalt bed (Upper Matlock Lava) in the Monsal Dale Limestone Formation of the Peak Limestone Group. The spring is channelled into a covered well that commemorates the now-demolished Royal Hotel which it used to feed. At this point, the spring is not producing tufa as it has not yet degassed CO_2 sufficiently to achieve supersaturation (Pentecost, 1999). The water is piped below a car park, re-emerging in a small tufa-stone quarry cut into the Holocene terrace ~200m downslope. Within the quarry, two bryophyte-covered and steep-fronted tufa cascades flow down into a small, circular garden. The semi-aquatic mosses (*Cratoneuron* sp.), liverworts and grasses provide an undulose and sponge-like surface, through which the lime-rich spring waters pass. Some of this precipitates as low-Mg calcite around the moss stalks and forms tufa skins, which rapidly bury the living plant tissues. It is considered that the precipitation is strongly promoted by plant respiration and bacterial metabolic activity (Pedley, 1994; Pedley et al., 2009; Rogerson et al., 2008), as well as by continued physico-chemical degassing of CO_2. Beneath the living veneer a typical open textured phytoherm tufa fabric is developed which is added to by further inorganic precipitation of calcite. The growing lobe of tufa has developed surface channelled rivulets with a distributary network, encouraging a lateral fan-shaped development of the tufa body with time. Under natural conditions the final product would be a lenticular longitudinal and transverse section with its widest point distal to the spring. However, management of this site has resulted in truncation of the main channel and construction of a ditch around the edge of the path with consequent highly altered flow in the distal parts of the system.

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