<u>Karst hydrogeology of the southern catchment of the River Wye, Derbyshire</u>

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Abstract.

A conceptual model of the regional hydrogeology of the White Peak, considered fundamental to the understanding of the local (Wye) catchment has been presented. Specific to the local catchment, an investigation of the karst hydrogeology has been carried out in the context of its geological setting using results from: tracer experiments, chemical analyses of spring water, and hydrograph analyses; alongside detailed consideration of speleogenetic processes and terrain evaluation. Derived from these studies, a conceptual model has been developed, which represents the catchment hydrogeology in a number of hydrogeological units. Their attribution reflects the lithological differences and material responses to both stress and mineralization that have exerted significant influence on speleogenetic processes in the catchment. The units exhibit different recharge, through-flow and resurgence characteristics. Speleogenetic processes in some of the bedrock units support the inception horizon hypothesis. Flow paths typically pass through more than one hydrogeological unit.

Lead-zinc-fluorite-baryte mineralization is associated with the dominant hydrogeological unit on the eastern side of the catchment. The mineral deposits were subject to several phases of exploitation facilitated by dewatering via drainage adits (soughs). Records pertaining to the soughs have been used to contribute to an understanding of the changes in groundwater levels as a consequence of mineral exploitation. A case study focused on Lathkill Dale has been used to test the catchment model and further explore human impacts on the hydrogeology.

The major contribution of this work is in furthering the understanding of the hydrogeology and speleogenetic processes operating in the catchment. This is supplemented by additional contributions to the understanding of the distribution of superficial deposits within the catchment. Speculation regarding mineralizing processes; geomorphology; functioning of karst aquifers; seasonality of the groundwater chemistry; climate change, and the engineering properties of the bedrock may encourage further research in these areas.

Acknowledgements:

The compilation of this thesis is merely a step in what has been a long path in achieving a level of understanding, albeit undoubtedly naïve, of the karst hydrogeology of the White Peak. A number of groups and individuals have offered help, advice and support at different stages of the study. Two of these individuals, Professor John Gunn (University of Huddersfield) and Dr. David Lowe (British Geological Survey), have made themselves available and worked tirelessly to provide guidance throughout. Further to this, my thanks extend beyond the research, to the lessons in rigour and perseverance required in the completion of this work. The University of Huddersfield is acknowledged for the financial support in the form of a bursary over a period of three years. As are the fieldwork contributions that were made to the Limestone Research Group (University of Huddersfield) by English Nature. Dr C Hunt and Dr T Ford are thanked for taking part in the examination process.

Placement students working in the Limestone Research Group, including Claire Mathews, Paul Taylor, Chris Hustler and Helen Lamb are thanked for their assistance with fieldwork and the laboratory work that they have carried out. Particular thanks are also extended to Mr Steve Newsome, who offered endless enthusiasm for, and invaluable assistance in, the field. Thanks also to: John Taylor and his colleague, leaders of a memorable trip into Lathkill Head Cave; Mr Jon Humble (Natural England) and Mr David Webb, who carried out flow monitoring in Bateman's Shaft; Mr Paul Chandler (Peak District Mines Historical Society) excursion leader to Yatestoop Sough, and the Black Marble Mine, Bakewell; Mr and Mrs Blackwell who allowed access to springs at Dirtlow Farm, Sheldon; High Peak District Council Park Keepers who facilitated access to Bakewell Recreation Ground thermal spring, and the quarry managers at Hindlow and Topley Pike who were kind enough to allow site visits and discuss aspects of quarry management and the hydrogeology of the catchment.

The generosity of: Mr Tim Colman (British Geological Survey) in sharing knowledge with respect of mineralization and providing references; Dr Alan Dykes in reading and commenting on early copies of chapters of the thesis; Mr Steve Green (High Peak Borough Council) in providing access to meteorological records; Mr Ben Le Bas (Natural England) and Dr Steve Worthington in conducting constructive reviews of chapters 11 and 8 respectively, and also for supplying papers, is much appreciated. Thanks are also extended to my former employers and colleagues at RSA Geotechnics Limited and to my current employer and colleagues, at the British Geological Survey (Natural Environment Research Council).

Dr A.E. Adams, University of Manchester is acknowledged for instilling an early interest in carbonate sedimentology that has been beneficial to this research. Similarly, without the lessons in how to apply the tools of the geologist to engineering geology and hydrogeology, learnt under the guidance of Dr M.H. de Freitas, it is unlikely that this thesis would have been submitted. This is also the appropriate forum to extend my thanks to, and acknowledge that, without the support of Mr E.A. Barber it is unlikely that I would have been recognised as an appropriate candidate for further education.

To my family and also my friends, as well as thanks for their support, I would like to apologise for my neglect of them over the last six years.

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Chapter 1: Introduction.

1.1 Background.

Limestone aquifers are perhaps the most challenging of the major aquifers in terms of characterisation and modelling. This is primarily attributable to the triple porosity of the medium, which is clearly defined by Worthington and Smart (2004), as matrix; fracture (dissolutionally enlarged fractures of <0.01 m), and channel or conduit porosities. It is exceptional, perhaps in the context of heavily fractured limestone, that a limestone aquifer could in any way be considered as isotropic.

Building on previous work carried out by the Limestone Research Group (University of Huddersfield), for example Gunn (1998), the aim of this thesis was to analyse the nature and extent of human impacts on the hydrology of the southern catchment of the River Wye, Derbyshire through the development of a pre-Roman conceptual model of its hydrogeology. As with all aquifers the key to developing a representative conceptual model is deriving an understanding of the storage and transmissivity of the aquifer. In the context of a karst aquifer this requires an understanding of the speleogenetic processes that guide the development of permeability. The area of investigation, as defined in the title of the thesis, is the southern catchment of the River Wye. This area stretches from Buxton in the west, to the confluence of the River Wye with the River Derwent at Rowsley in the east. The catchment includes the catchments of the Rivers Wye, Lathkill and Bradford and extends as far south as the southern boundary of the catchment of the River Bradford (Figure 1.1). However, reflecting the geographical distribution of available source information, some areas within the catchment have formed the focus for more detailed research and some aspects of the research extend over a broader area.

Situated in the area referred to as the White Peak, the research area is underlain by Carboniferous shelf limestones, largely deposited on faulted basement rocks. During subsidence and burial the limestone was subject to Mississippi Valley-type mineralization. The forces of erosion have broadly removed later sediments, which were of limited thickness. It would appear that the current topography has been carved into a former erosional surface, primarily by the Anglian glaciations, as the area was largely free of ice during the Devensian. The current topography comprises upland areas of 200 to 400 m above Ordnance Datum (OD), with the valleys of the Wye, Lathkill and Bradford dropping from approximately 300 m OD in the west to approximately 98 m OD at the confluence of the River Wye with the River Derwent (Figure 1.1). Exploitation of the lead-zinc mineralization occurred in a number of phases, probably commencing during the Roman period and locally continuing until the early 1900s; the phasing being influenced by the price of lead and by technological improvements in methods of dewatering the mine workings. By the seventeenth century most of the lead had been worked down to the "water table" and dewatering became necessary. Initial dewatering was achieved using rag and chain pumps, or similar, and the directing of groundwater into natural cavities (underground swallow holes). Sough driving, i.e. the excavation of tunnels or adits from the valleys to the mine workings, as a means of dewatering the workings, appears to date to the 1640s. The earliest soughs were local excavations, then in order to achieve greater depths of dewatering the soughs were driven from increasingly distal, but lower valleys.

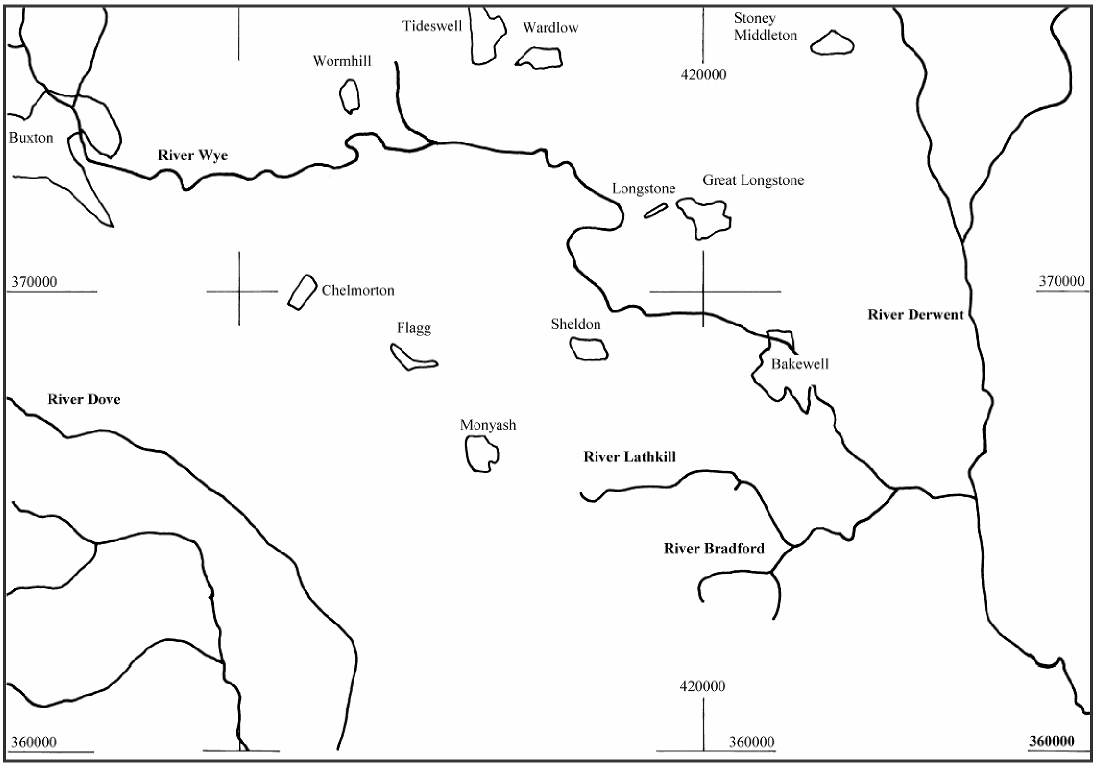


Figure 1.1: Research area defined in the text (section 1.1).

It is generally accepted that groundwater levels in the Carboniferous Limestone aquifer of the Peak District have fallen in response to the impacts of the human exploitation of the mineral reserves of the limestone (Christopher et al., 1977, Bamber, 1951). However, although previous work in the area has covered: geology (e.g. Adams, 1980; Aitkenhead et al., 1985; Biggins, 1969; Berry 1984; Butcher and Ford, 1977; Cope, 1973; Dunham, 1973; Ford, 1999; Gutteridge, 1983; Ineson and Dagger, 1985; Schofield, 1982 ; Smith et al., 1985; Walkden, 1970), mineralization (Carruthers and Strahan, 1923; Ineson and Ford, 1982; Quirk, 1986; Worley, 1978) speleogenesis (Beck, 1980; Ford, 1989), geochemistry (Bertenshaw, 1981; Burek, 1978; Christopher, 1981; Edmunds, 1971), hydrogeology (Bamber, 1951; Downing et al., 1970; Raffety et al., 1953) and sough hydrology (Oakman, 1979; Rieuwerts, 1981); no clear conceptual model has been developed for the hydrogeology. In order to consider the human impacts the development of a conceptual model was necessarily the first step. A number of conceptual models of karst hydrology have been developed, for example Motyka (1998), Smart and Hobbs (1986) and White (1969), but none were directly applicable to this research area. In developing a conceptual model a secondary aim of the research was to develop the understanding of speleogenetic processes that have and continue to operate in the research area. Sweeting (1972, p. 260) observed "*Hence many of the disputes about the relationship of the caves to former erosion levels and to river terraces etc., arise from the fact that fluviokarst landforms are developed differently and have had a different history from true karst forms*". This is an interesting statement in that it implies that in the absence of glacio-fluvial influences, the hydrology resulting from karst processes could be modelled. Yet, the starting point for interpretation of the Peak District by most previous authors (Beck, 1980; Burek, 1978) has been to try to relate terraces and cave levels to base levels imposed by a chronology of events. Strata that exhibit relatively low angles of dip underlie the major portion of the research area; therefore this was a good area to test the inception-horizon hypothesis of Lowe (1992). The development of a conceptual model and investigation of speleogenetic processes required the identification of the key hydrogeological features and the hydrogeological history that characterise the research area.

To optimise any local or intermediate catchment-based hydrogeological study it is necessary to set the catchment in the context of the regional catchment, as implied by Tóth (1963) and nowhere is this more important than in the context of a limestone catchment. Furthermore, literature reviews (Chapter 5) have identified that work carried out in other areas of the White Peak has implications for the understanding of hydrogeology of the northern catchment of the River Wye and therefore it is inevitable that much of the work presented in the thesis extends beyond the southern catchment of the River Wye.

The conceptual model developed in this thesis has been derived from an assessment of a large number of resources. It was that opinion of this author that to report this assessment in the traditional form with a single literature review situated towards the front of the thesis would result in an unbalanced document with the reader being required to carry information from one chapter to another. An outline of the structure of the thesis follows in section 1.2, to assist the user. A glossary of terms used in the thesis has been included as Appendix 1.1.

1.2 Thesis structure.

As a hydrogeological study researched by a geologist it is inevitable that the starting point should be the development of an understanding of the geological attributes with the potential to influence the hydrogeology. Thus, Chapter 2 presents a description of the types and variability of limestone of the White Peak. The types of limestone are closely linked to their depositional environments and are influenced by the occurrence of extrusive igneous rocks, which are also described in Chapter 2. The depositional environment and structure are set in the context of the tectonic setting, as a means of assisting in the interpretation of the influence of the structure on the development of groundwater flow-paths. Chapter 2 also considers the Post-Dinantian solid geology, the influence of the Quaternary and introduces aspects of the geomorphology that are evident in the White Peak and reflect hydrological processes.

The extremes of permeability (cave to matrix) that are represented in a mature karst aquifer necessitate very different methods of investigation and inevitably indicators of the way in which a karst aquifer functions come from a number of sources. Accordingly, the research that forms the substance of chapters 3 to 8 is derived from a multi-facetted approach, as summarised in Figure 1.2 (designed to assist the reader to navigate through the thesis). There is an extensive literature associated with each of these subject areas and the investigation of the literature has formed an integral part of this research. One of the ancillary aims of the thesis was to reinterpret earlier work in the light of more recent understanding of karst hydrology. Accordingly, the findings of the literature reviews are incorporated within the context of the appropriate chapter.

Some of the early chapters draw particularly heavily on the findings of literature searches. For example, Chapter 3, which considers human exploitation of the resources of the limestone, in particular the mining of lead and zinc and the associated construction of soughs, relies on the work of Ford and Rieuwerts (2000), Rieuwerts (1980, 1981 and 2000) and Oakman (1979). This reflects the difficulty for an inexperienced caver to access the mine workings and the time requirements associated with archival research. Despite these constraints, this author has included additional information from components of personal archive research. Chapter 3 also reconsiders aspects of the mineralization processes and the extensive chert deposits (with the potential to act as aquitards). Present and historic groundwater exploitation from the limestone aquifer is described in the context of resource exploitation.

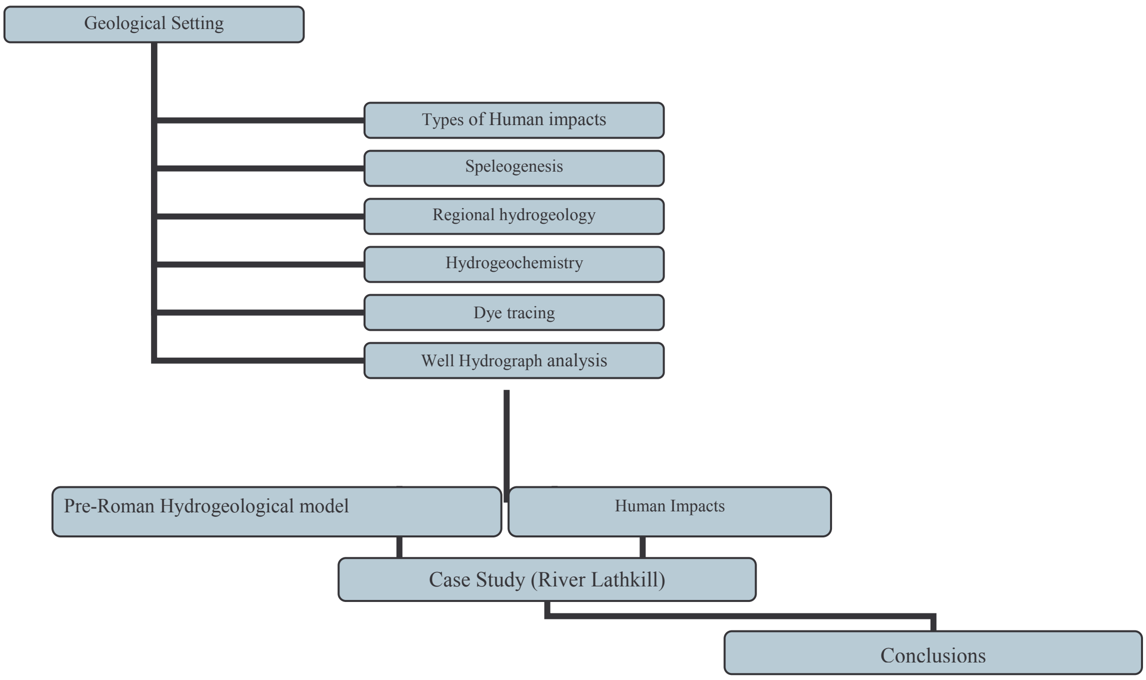


Figure 1.2: Thesis structure.

Archaeological interest (Bramwell, 1977) has shown the intrigue that caves (conduits of dimensions sufficient for human access) have held for man since prehistoric times. This interest continues to the present day and the information provided from the exploration of caves is crucial with respect to understanding the factors that guide speleogenesis. For this investigation the main source of information regarding the caves of the area has been Beck (1980) and Beck and Gill (1991).

Accordingly, Chapter 4 also includes a large component of literature review, but also presents new work with respect to potential speleogenetic processes operating in the White Peak and crucially considers recharge, storage and flow-paths in the karst aquifer. Other chapters, such as Chapter 5, develop current research themes, for instance Gunn et al. (2006) and relies unashamedly upon lateral thought, giving consideration to the significance of the thermal springs, the length of flow-paths, the distribution of springs, caves and the regional variation in the chemistry of the thermal springs. This leads on to a consideration of the potential for underflow (groundwater flow-paths extending beyond the area of the limestone outcrop).

Chapter 6 reinterprets the work of Christopher (1981) and Edmunds (1971), in the light of the current understanding of the hydrogeology of the White Peak. It also uses the work of Smith (2000) to present classifications for the springs, which are based on (interpreted) geological formation at source. It is considered by this author that this work provides supportive evidence for the speleogenetic processes postulated in Chapter 4.

Tracing experiments form a key technique in the investigation of karst aquifers. The main source of information regarding dye-tracing experiments has been the work of the Limestone Research Group, University of Huddersfield. In addition to more recent work carried out by Banks and Gunn (2003), Chapter 7 reinterprets the work of earlier experiments carried out by: Gunn (1990 and 1998); Gunn and Hardwick (1999); Gunn, Hardwick and Lowe (1999); Hardwick (1996a, 1996b and 1996c); Hardwick and Gunn (1994 and 1995), Hardwick and Hyland (1991a and 1991b) in the light of the current understanding of the hydrogeology of the White Peak. For reference purposes, the available results of Limestone Research Group (University of Huddersfield) experiments across the entirety of the White Peak have been interpreted.

The seasonality of an aquifer and its response to individual storm events gives an indication of the storativity and transmissivity of the aquifer. This type of information has been considered both in terms of the hydrogeochemistry of the aquifer (Chapter 6) and the groundwater recession curves, as indicated by river discharge (Chapters 10 and 11) and borehole hydrographs (Chapter 8). The latter gives an indication of the contribution of matrix/ fracture components of the aquifer and provides valuable supplementary detail to the results of tracing experiments (Chapter 7). The analysis presented in Chapter 8 also considers the interpretation of the form of the seasonal recession curve in the context of the karst aquifer. This is understood to be one of the first analyses of this type that has been carried out in a karst aquifer.

The results of the investigatory work that are summarised in chapters 3 to 7 have enabled this author to group the limestone formations as hydrogeological domains and to develop a conceptual model for groundwater flow in the White Peak. This, together with an interpretation of the landscape evolution, has enabled the author to present a pre-Roman conceptual model for the hydrogeology of the research area (Chapter 9) and also to assess the human impacts on the model (Chapter 10). The results of flow monitoring of the River Lathkill have provided additional information that renders the River Lathkill a suitable case study to test some of the ideas presented in the conceptual model and in addition to provide further, supplementary considerations. The flow monitoring was carried out by this author, with other members of the Limestone Research Group, on consecutive years during the course of the research for this thesis, in the context of commissioned research being carried out by the Limestone Research Group for English Nature.

It is not surprising that during the course of the preparation of the thesis, a maze of potential research avenues have opened up. These avenues have been waymarked within the conclusions of the thesis (Chapter 12).

Chapter 2: Geological and geomorphological considerations.

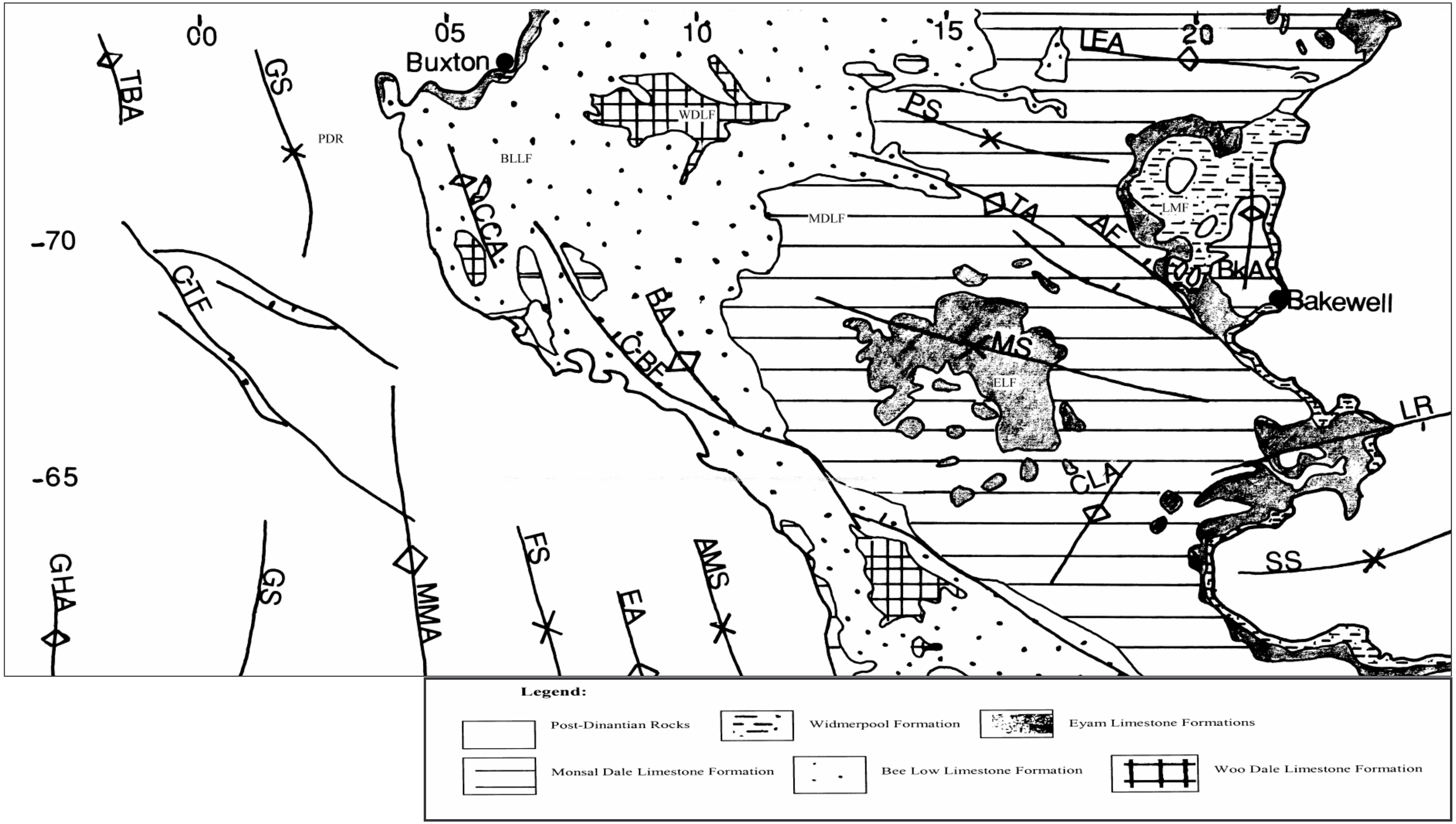
2.1 Introduction.

Klimchouk and Ford (2000b) present a case for working towards a complete understanding of the geological history as a means to interpreting the hydrogeology and speleogenesis of karst terrains. Inherent within the detailed analysis of the geology of a specific area is the requirement to understand the regional context. This is the approach that has been adopted in this analysis of the geological influences on the hydrogeology and speleogenesis in the Wye valley.

The 1: 50 000 Series British Geological Survey Sheet 111 *Buxton*, Solid and Drift edition, shows the area of the White Peak to be underlain by Lower Carboniferous (Dinantian) shelf limestones (BGS, 1978). Broadly the strata form an inlier within Namurian and Westphalian (Silesian) strata and young to the east. Interbedded with the shelf limestones are a number of penecontemporaneous lavas. Reefs surround the shelf limestones and there are further developments of reefs, as exemplified by those to the east of Monyash (SK 150666). The area is crossed by a number of faults and mineral veins, which trend predominantly northwest to southeast and west to east. There are terrace-like areas of glacial till, at about 220 to 240 m OD in the Wye Valley, to the south of Bakewell. Figure 2.1 shows the main geological features of the area.

Stratigraphical and lithological studies of the Derbyshire Dome date back over 200 years; important studies of the last thirty years being those of Aitkenhead et al. (1985); George et al. (1976), Strank (1985) with respect to the stratigraphy; and Aitkenhead et al. (1985); Berry (1984); Gutteridge (1991b); Schofield and Adams (1985), Walkden (1987) with respect to the lithology. Furthermore, authors such as Beck (1980); Oakman (1979); Worley (1978), Quirk (1993) have provided geological descriptions within the context of research on the speleogenesis, mineralization and mining in the area. Since the preparation of these works, understanding of both the concepts of speleogenesis and interpretation of the geology has evolved and the following is largely an attempt to integrate these concepts as a framework for subsequent chapters covering aspects of speleogenesis and hydrogeology.

At an early stage of the research this author identified that significant faults influence the hydrogeology, particularly in forming zones of groundwater resurgence and in defining hydrogeological boundaries. It was also clear that there are a number of other geological influences on hydrogeology, particularly in the context of inception horizons (Lowe, 1992), which indicated a need to pursue geological detail, for example: boundaries and anomalies in structure, lithology, diagenesis and chemistry. Accordingly, considerable attention has been given to the detail of the literature, for example the descriptions of caves, in particular the excellent, detailed descriptions presented by Beck



Key: Structures - AF Arrock Fault; AMS Archford Moor Syncline; BA Brierlow Anticline; BkA Bakewell Anticline; C-BF Cronkston-Bonsall Fault; CCA Countess Cliff Anticline; CLA Calling Low Anticline; C-TF Cut-Thorn Fault; FS Fernyford Syncline; GS Goyt Syncline; GHA Gun Hill Anticline; LEA Longstone Edge Anticline; LR Long Rake; MMA Mixon Morridge Anticline; MS Monyash Syncline; PS Priestcliffe Syncline; SS Stanton Syncline; TA Taddington Anticline; TBA Todd Brook Anticline. **Formations –** BLLF Bee Low Limestone Formation; ELF Eyam Limestone Formation; LMF Widmerpool Formation; MDLF Monsal Dale Limestone Formation; WDLF Woo Dale Limestone Formation.
PDR Post-Dinantian Rocks.

Figure 2.1: Principal Structures and Formations of the research area.

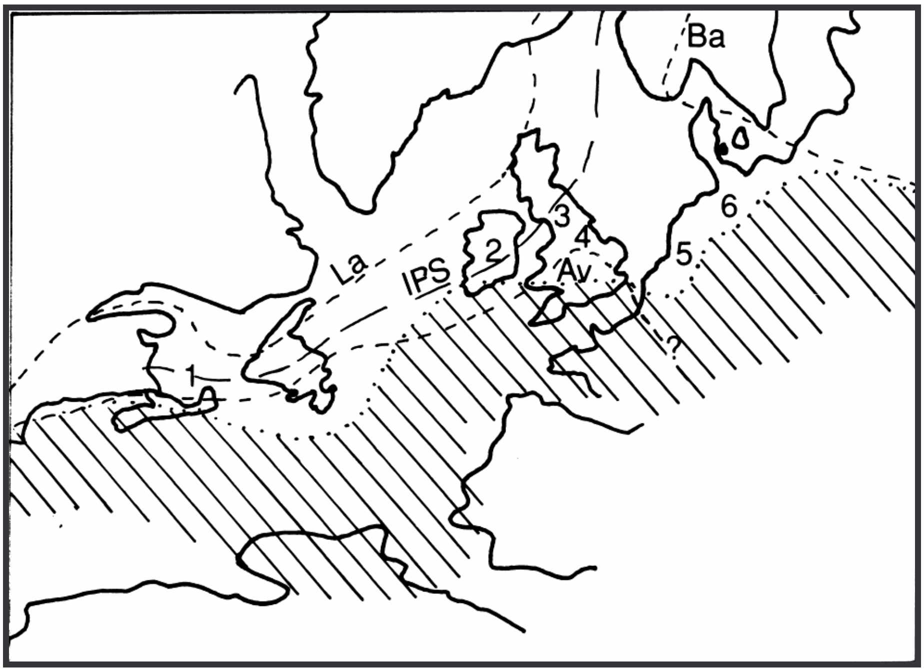
(1980), which suggest a form of bedding guidance in speleogenesis, particularly in the area of the River Lathkill. From these beginnings the following geological considerations have been pursued: tectonic setting, structural setting, sedimentary styles, reef facies, clay wayboards, diagenesis, post-Dinantian solid geology, Pleistocene and Quaternary sedimentation and weathering.

2.2 Tectonic setting.

The enduring guidance of the basement on the subsequent geological history and on the hydrogeology of the area is perhaps more important than has previously been recognised, for example Gunn (1992). The following comprises a brief summary of the tectonic history. More detailed reviews are presented by Bluck et al. (1992); Fraser and Gawthorpe (2003); Gutteridge (1989); Haszeldine (1989); Leeder (1987 and 1988), Smith and Smith (1989).

Bluck et al. (1992) have interpreted the foundations of the British Isles as a series of suspect terranes, or areas of similar basements defined by prominent fault systems derived from a Proterozoic Supercontinent (Piper, 1982). The setting of the area of the Derbyshire Dome, on the northern edge of Avalonia (the Midlands Microcraton, Pharaoh et al., 1987) is shown on Figure 2.2. The Midlands Microcraton, which underlies much of central England, comprises late-Proterozoic rocks. To the north of the Iapetus suture the deformed basement is thought to have formed in mid-Proterozoic times as an integral part of the Laurentian Shield (Thorpe et al., 1984). The area between appears to have been controlled by the evolution of a subduction zone comparable with that of the more extensively documented Welsh Basin (Pharaoh et al. 1987) and associated with the closure of the Iapetus Ocean. The deflexion of the resulting Caledonian orogeny, around the Avalonian Terrane imposed the dominant northwest to southeast-trending structural grain, the Charnian trend (Rogers, 1983; Guion et al., 2000). At this time Britain occupied an equatorial position and as the Peak District formed low-lying land there is no sedimentary record of the period. However, to the south of the Cronkston-Bonsall Fault, the Caldon Low Borehole record identifies that a significant thickness of Devonian sediments accumulated in the North Staffordshire area. During the Carboniferous, shelf-seas became established on the northern edge of the Rheno-Hercynian Ocean.

Closure of the Rheic Ocean (or proto-Tethys) took place towards the end of the Devonian, as Gondwanaland moved north, forming a northerly dipping subduction zone beneath the Midlands Craton, to meet Laurentia in the formation of Pangea. As the Variscan deformation front moved north the Derbyshire Dome was subject to north to south-trending extensional stresses. This was associated with a period of back-arc basin formation and subsidence, followed by thermal subsidence and foreland basin development during the Silesian (Leeder, 1988) referred to as post-rift subsidence by Smith et al., 2005). Leeder (1988) postulated that in this setting localised melting of the asthenosphere gave rise to the widespread basaltic vulcanicity. The evidence suggests that north-south extension was accommodated in the pre-existing structural grain. Alternative interpretations of the tensional tectonic setting have been presented: for example, Haszeldine (1989) preferred the concept of reactivation of



Key:

Plate names:
Av = Avalonia; La = Laurentia; Ba = Baltica
Plate collisions Late Silurian to Early Devonian

IPS = Iapetus suture

Hatched area denotes the Variscan foreland during the Late Devonian to Tournasian Sedimentary basins forming in front of this include:

1	North Appalachians
2	Central Ireland
3	Craven
4	East Midland (Derbyshire Dome)
5	Dutch
6	Northwest German

Figure 2.2: Tectonic setting of the Derbyshire Dome (adapted from Smith and Smith, 1989).

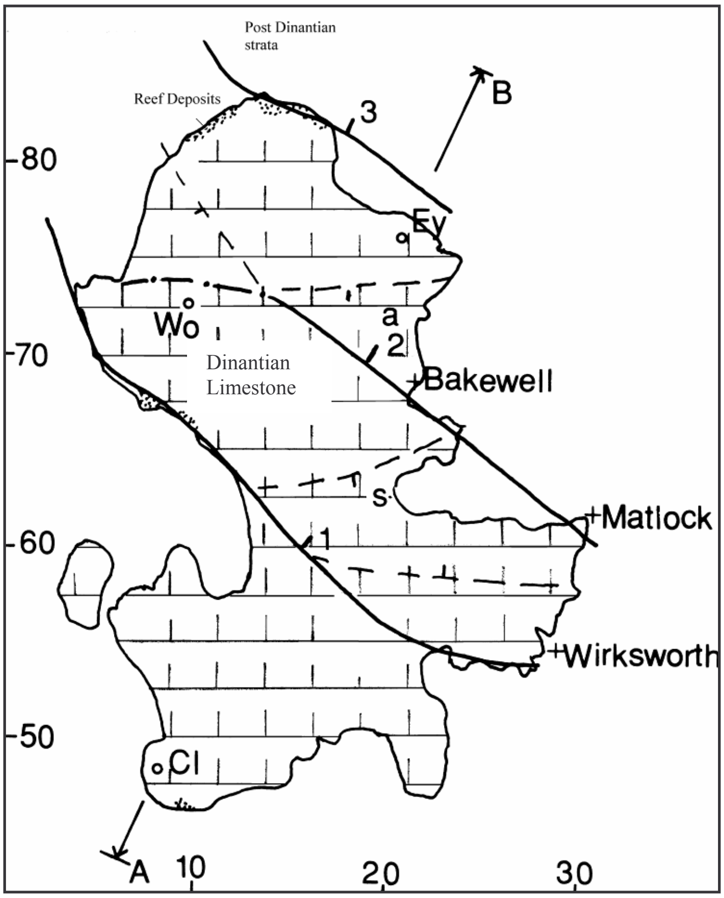
north-south lineaments, associated with the commencement of opening of the Atlantic Ocean; and the concept of dextral shearing from Maritime America across to the Irish and British area, associated with the inter-plate forces between the Laurentian and Baltic plates (Smith and Smith, 1989). It is also important to note, that although the absence of Carboniferous ocean floor in the present day oceans makes relative pole and plate positioning difficult, it is widely accepted that Gondwana extended southwards to the South Pole and therefore was subject to glaciation at this time. During the later Carboniferous regional east to west compressive stresses developed as the Variscan effect moved north, which resulted in north to south-trending folds and faults. On the western side of the study area there is a north-south structural grain, following the Pennine Axis. The structural grain is associated with the Malvern Line, and a north to south-trending extensional fault, which lies on the western side of the Goyt Trough, the southerly extension being referred to as the Hawkins Line (Fraser and Gawthorpe, 2003).

Smith and Smith (1989) and Quirk (1993) suggest that the formation of the North Sea, during the Permian and Mesozoic, was associated with easterly tilting. Palaeogene uplift, thought to be associated with igneous underplating (Holdsworth et al. 2000) is attributed to the opening of the Atlantic. Later, Neogene basin movements are attributed to the Alpine orogeny. However, the impacts of these movements on the Peak District were less significant.

2.3 Structural setting.

Comparison of the findings of the Woo Dale Borehole (Cope, 1973) and the Eyam Borehole (Dunham, 1973 and Strank, 1985) demonstrates that the Dinantian sediments were laid down on an irregular basement. This author's interpretation is shown on Figure 2.3. Assuming geometric constraints on fault kinematics, ancient zones of tension can be compared with current tensional environments (Jackson and McKenzie, 1983). Modern examples of extensional tectonics include the North Sea (Gibbs, 1984 and Guion et al., 2000), the Aegean Sea (Jackson and McKenzie, 1983 and Leeder 1988) and the Gulf Coast Plain of the U.S.A. (Johnson, 1982). It is this line of enquiry that has dominated structural research of the area of the Derbyshire Dome over the last thirty years. Of particular note is the concept that faults or basin development initiated in an extensional setting can become zones of inversion. This has significance both in terms of sedimentation and controls on weathering (considered in Chapter 9) and therefore also on the pattern of surface water drainage of the area. That some of the faults were active during sedimentation is implicit in the use of the term 'growth-faults' (Aitkenhead et al., 2002).

Evolving from the tilt block model (Grayson and Oldham, 1987; Miller and Grayson, 1982), the current model for the Derbyshire Dome is one of a carbonate platform formed around a basement high and composed of three half-grabens. Support for this model comes from a number of geophysical studies: gravity data (Cornwell and Walker, 1989; Lee, 1986; Maroof, 1976); seismic refraction (Rogers, 1983), seismic reflection (Maguire, 1987; Smith, and Smith, 1989).



Key:

Cl	Caldon Low Borehole
Ey	Eyam Borehole
Wo	Woo Dale Borehole
EG	Edale Gulf
a	Ashford Basin
s	Stanton Basin

1. Cronkston-Bonsall Fault (Cornwell and Walker, 1989; Gawthorpe et al., 1989)
2. Bakewell Fault (Smith et al., 1985)
3. Edale Fault (Fraser and Gawthorpe, 2003)

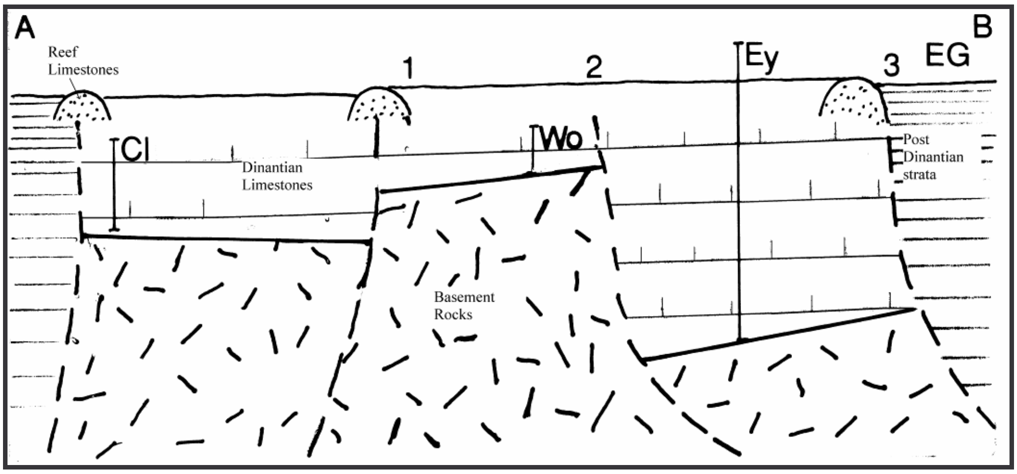


Figure 2.3: Schematic basement profile (adapted from Smith et al., 1985).

Gawthorpe et al. (1989) and Gutteridge (1989) have defined the northern part of the high as the footwall of the east to west-trending, northerly downthrowing Edale Fault and the southern margin as being associated with the Cinderhill Fault (Whitcombe and Maguire, 1981). The easterly margin is defined by the easterly downthrowing Ashford Fault, but at the western margin there is no distinct structure. The Derbyshire Dome is divided into three blocks by the east-west Cronkston-Bonsall Fault and the Bakewell Fault (Smith et al., 1985) a sub-surface fault associated with the Bakewell to Taddington Anticline. Uplift of the northern area appears to have occurred in the Holkerian, possibly due to footwall uplift, resulting in an unconformity between the Woo Dale and the Bee Low Limestone Formations, which is increasingly marked to the south (Schofield and Adams, 1985). An intra-shelf basin, the Ashford Basin, developed at the beginning of the Brigantian, as a consequence of reactivation of the Bakewell Fault. Late in the Brigantian, as a fault terrace developed, the basin spread southwards as an eastward-dipping carbonate ramp developed as a result of footwall collapse of the southern fault block. Similar fault guidance of sedimentation has been observed by Gawthorpe (1987) for the Bowland Basin. It was probably at this time that the faults that comprise northnorthwest to southsoutheasterly-trending mineral veins in the northern part of the Monyash Syncline were reactivated.

The Bakewell Fault to the south, and the Longstone Anticline to the north, form the boundaries to the Ashford Basin. To the east the Bakewell Anticline (Aitkenhead et al., 1985) comprises another growth fault (not named by Aitkenhead et al., 1985), defining the eastern edge of the Ashford intra-shelf basin (Gutteridge, 1987 and 1989). Reversal of slip on the Arrock Fault (Aitkenhead et al., 1985), associated with the southeastern section of the Bakewell Fault (as defined on Figure 2.3) during the late Brigantian caused the development of minor north to south-trending anticlines at the eastern margin of the Derbyshire Dome. Gutteridge (1987) has postulated a second, slightly later, intra-shelf basin, the Stanton Basin, in a very similar setting with the Cronkston-Bonsall Fault to the south, Long Vein to the north and the Millclose Anticline controlling the eastern boundary. Evidence for the intra-shelf basin development necessarily comes from detailed sedimentological studies (Gawthorpe et al. 1989), some of the earliest being on the Station Quarry Beds (Butcher and Ford, 1973; Cope 1933; Gutteridge 1987 and 1989, Walkden, 1977). Other detailed studies of laminated units indicative of slumping associated with active basin development include the work of Adams and Cossey (1978); Gutteridge (1989), Padzierski (1982). Gutteridge (1989) also carried out studies of facies associations.

A number of faults, which have been enduringly active, are suspected to form significant hydrogeological boundaries. Furthermore, it is possible to speculate that the Ashford Fault, which defines the eastern boundary of the Edale tilt block, may extend farther south to join with the fault associated with Bubble Springs (SK 20516611) in Lathkilldale. Boundaries to the west are less easily defined. Biggins (1969), in his interpretation of aeromagnetic variations, suggested that the western margin of the outcrop of the Dinantian Limestone comprises a tilt axis. Later work appears to confirm this (Rogers, 1983). Both Rogers (1983) and Guion et al. (2000) suggest that the axis may reflect a continuation of the Malvernoid trend (section 2.2). It is also interesting to note that both of the half graben boundaries that have been identified appear to hinge from the area of Buxton. Furthermore, that the Cronkston-Bonsall Fault is evident in gravity data (Cornwell and Walker, 1989) suggests that it is a basement fault; whereas the Bakewell Fault is evident from seismic data (Fraser and Gawthorpe, 2003; Smith and Smith, 1989), suggesting that it is a higher-level fault, extending to the basement.

East to west-trending faults can also be identified in the area. These are likely to be compressional faults, associated with the northerly migration of the Variscan front, as Africa docked against Europe. Some north to south-trending faults and dextral shear along pre-existing faults, as well as folding (Butcher and Ford, 1977), is likely to have resulted from the tectonic adjustments associated with the juxtaposition of Gondwanaland, the Baltic and Laurentian plates. Later north- to south-trending fault-sets can be attributed to tension associated with the opening of the Atlantic, although much of the tension is likely to have been accommodated in pre-existing fault sets. Further consideration is given to the faulting in Chapter 9.

2.4 Sedimentary styles.

Sedimentary styles varied across the carbonate platform. On the northern, fault-controlled, margin slopes of 30 to 40° were established (Broadhurst and Simpson, 1973). On the southern boundary marginal slopes of 5 to 10° have been determined (Guion et al., 2000). Sedimentation in this area comprised fine-grained sediments with slump deposits. Similar sequences were deposited within the intra-shelf basins, e.g. Rosewood Marble (Adams and Cossey, 1978), with transgressive periods being marked by coral beds (Guion et al., 2000). Carbonate sand bodies were established at the platform margins and shelf carbonates comprised shallowing-upward sequences, as described further below. The basinal facies exhibit greater variability, e.g. the development of bands of chert and this has been shown to influence speleogenetic processes. Table 2.1 provides a summary of the geological formations of the carbonate platform and further consideration of the lithologies is presented below.

Table 2.1: Formations of the Derbyshire Dome.

Series	Stage	Formation	Coral Brachiopod Zone
	Brigantian	Longstone Mudstone Formation Eyam Limestone Formation Monsal Dale Limestone Formation	D2
	Asbian	Bee Low Limestone Formation*	D1
	Holkerian	Woo Dale Limestone Formation	S2
	Arundian		C2 S1
Visean	Chadian		γ C1
	Ivorian	Anhydrite	K2
Tournasian	Hastarian		

* The base of the Bee Low Limestone lies marginally above the Asbian – Holkerian boundary

The Woo Dale Limestone Formation comprises a sequence of open shelf limestones, overlain by finer-grained peritidal deposits, which migrated from north to south across the shelf and are overlain by restricted lagoon (packstones and grainstones) deposits (Schofield and Adams, 1985). Limestones of the Asbian and Brigantian stages have been shown (Walkden, 1987) to be cyclic, with each bed showing evidence of progressive shallowing (regressive sequences). The Monsal Dale Limestone Formation is divided into dark and pale facies. The dark facies largely comprise biosparites and biomicrites and range in colour from mottled shades of grey to true dark grey (Cox et al., 1977). The pale facies are dominated by massively-bedded biomicrosparites (Cox et al., 1977). Marker horizons are common throughout (Cox et al., 1977). Typically a cycle of the pale facies of the Monsal Dale Limestone Formation comprise basal, coarse, bioclastic grainstone/packstone grading to bioturbated, bioclastic, packstone/ wackestone with coral bands, grading to fine-grained, sorted bioclast peloid grainstones, locally capped by wackestones and fenestrae (Gutteridge, 1989). Walkden (1974) has shown that the upper parts of cycles show evidence of emergence, including calcrete textures and paleokarstic surfaces, which he estimates took in the order of 30 000 to 100 000 years to form. The surfaces are mammillated, with hummocky or potholed morphologies, they cut fossils, suggesting that they are erosional, and have pits perpendicular to bedding (rather than to vertical), suggesting that they are contemporaneous with deposition and are unrelated to structural features (Walkden, 1987). Paleokarstic surfaces have also been found at the unconformable boundaries between formations. The lower parts of the Monsal Dale Limestone cycles comprise skeletal carbonate sands indicative of a sub tidal environment. The depositional environment has been interpreted by Aitkenhead et al. (1985) as one of a shallow carbonate shelf subject to interplay between eustatic and tectonic controls on sedimentation times. Walkden (1987) noted a change in lithotypes between early Asbian and late Brigantian The change took the form of an increase in the thickness of the cycles, attributed to either progressively faster rates of subsidence of the platform, or to an increase in the size and duration of the cyclic oscillations. The evidence favours eustatic changes, of a type that could have been brought about by the fluctuations in the ice mass covering Gondwanaland (see section 2.2), superimposed upon the tectonic control. Walkden's work (1987) also shows that the evidence for the cyclicity is even found at the scale of the limestone cements and cathodoluminescence revealed zonation in the meteoric cements attributable to the cyclicity.

In a study of sedimentary styles in the Bowland Basin, it has been noted by Gawthorpe et al. (1989) that periods of tectonic activity are associated with an increased clastic input. The situation is similar in the Derbyshire Dome, where Walkden (1987) observed that there are subtle changes in the sedimentary cycles through the Asbian and Brigantian. The development of paleosols occurred during the later Asbian and early Brigantian, with an increasing clastic input towards the end of the Brigantian. This was associated with the formation of mudstones and reduced exposure of the limestones during the regressions. During tectonically quieter periods, such as the Asbian, the carbonate shelf was separated from areas of more basinal sedimentation by marginal apron reefs and the Earl Sterndale and Brassington reef belts formed part of a continuous basin margin.

Well-developed bedding is generally considered to be characteristic of limestones, e.g. Harrison (1981). The bedding varies considerably from thin to massive, e.g. the contrast between the thinly bedded Miller's Dale Limestone Member overlying more massively bedded Chee Tor Limestone Member, as exposed in the Monsal Dale railway cutting (SK 171725). However, Simpson (1985), drawing from his work on the Lower Carboniferous Limestone of the Gower, South Wales, suggested that care is required in the interpretation of bedding planes. In thin section apparent bedding planes appear as concentrations of stylolites, or pressure dissolution seams. In some limestones this is associated with clear sedimentological evidence of bedding, but in other layered limestones, for example those studied by Simpson (1985) that are relatively homogeneous and lack lithological changes or sedimentary structures, such evidence is lacking. Where depositional layering exists it is considered by this author that this will have a strong influence on the siting of stylolites and pressure dissolution seams.

2.5 Reef facies.

Reef limestones occur in both the shelf and off-shelf facies and in a range of stratigraphic horizons. Waulsortian-type mud mounds are restricted to the Tournaisian to Chadian (Miller and Grayson, 1982). The Waulsortian mud mounds have been studied extensively (e.g. Bridges and Chapman, 1988; Bridges, Gutteridge and Pickard, 1995; Miller and Grayson, 1982) and have been shown to have a characteristic structure, with extensive *stromatactis* and fissuring. The Waulsortian reefs are not exposed within the area of investigation, although they are suspected to occur at depth (Miller and Grayson, 1982) and they are significant because they appear to mark block boundaries, e.g. associated with the Edale Fault and the association of the Earl Sterndale reefs with the Derby lineament (southern margin of the magnetic basement beneath the Derbyshire Dome, Cronkston-Bonsall Fault), pointing to a potential relationship between the lineament and the Carboniferous shelf (Cornwell and Walker, 1989).

The Asbian to Brigantian reefs that do occur in the area are most commonly referred to, e.g. Gutteridge (1983) as knoll reefs, flat reefs and apron reefs, “build-ups”, or “lime mud build-ups”. The different types of reef have common lithological features comprising poorly bedded pale- to mid-grey micritic limestones with dips reflecting the original slopes of the reef flanks at progressive growth stages. The detailed structure of the reefs appears to have varied. Reefs fringing the Woo Dale Limestone were formed almost entirely of micritic mud; others had a micritic core surrounded by crinoidal flanks Gutteridge, 1983). Aitkenhead et al. (1985) note that the junction between the core and flank, or between core and inter-reef limestones, is generally sharp and for mapping purposes is taken as the boundary of the core of the knoll reefs. However, in a few reefs in the Eyam Limestone, the flank beds have also been included in the reef (Aitkenhead et al., 1985). The knoll reefs, or mud-mounds, of the Eyam Limestone have been described by Biggins (1969) and Gutteridge (1983, 1995). Reef development is associated with bioclastic carbonate sands deposited at the shallow part of an intra-platform ramp. Laterally accreted mounds occur to the west (indicative of shallower conditions) and vertically accreted mounds to the east. Typically they comprise a fine-grained structureless core with a bedded fringe passing into inter-reef facies dominated by the presence of crinoid fragments, indicative of a crinoid “thicket” around the mounds (Gutteridge, 1995) and with pockets of brachiopods, placing the reefs in the crinoid-brachiopod-fenestrate bryozoan build-ups (Bridges et al., 1995).

Erosional surfaces have been identified within the build-ups. Biggins (1969) in a study of High Tor, Matlock, observed cyclothemic facies variations related to windward organic progradation and regression "*in response to earth movements*". He also described a plane of disconformity associated with regression, invariably associated with a thin layer of tuff, swelling clays and haematite (clay wayboard [section 2.6], indicative of a period of emergence). In the stratigraphically lower of the mud mounds of the Eyam Limestone, caliche crust has been found to cap at least two of the knolls (Adams, 1980; Gutteridge, 1995). Vadose cement is associated, also indicative of emergence. The upper surfaces of the stratigraphically higher mounds of the same formation appear to have been subject to high-energy erosion and scouring, although they are not thought to have been subaerially exposed. Identification of paleokarstic surfaces, indicative of a significant period of erosion, at the top of carbonate mud mounds (reef deposits) in the area of Lathkill Dale has enabled Gutteridge (1991a) to suggest that the mud mounds in Lathkill Dale actually lie within the Monsal Dale Limestone Formation, rather than within the overlying Eyam Limestone Formation.

The core of the mud mounds is characterised by syn-depositional cavities, including cavities resembling *stromatactis*, but less interconnected, with rims of marine cement, suggesting their origin may be one of shelter voids, or it could relate to shearing of the sediment prior to diagenesis (Gutteridge, 1995). Additionally, the mound cores are cut by fractures that show evidence of formation at varying stages of lithification indicative of surface microbial binding stabilising the mud mound, below which fissuring occurred within the less consolidated mud (Gutteridge, 1995). In the apron-reefs that fringe much of the shelf margin in the Bee Low Limestone Formation there is a discontinuous wall-like core of micritic algal biolithite, passing into a transitional back-reef on the shelf side and into an outward-dipping fore reef on the off-shelf side (Gutteridge, 1995).

2.6 Extrusive igneous rocks and clay wayboards.

The British Geological Survey (1978) Sheet 111 recognises the Lower Miller's Dale Lava and the Ravensdale Tuff Members within the Bee Low Limestone Formation. Within the overlying Monsal Dale Limestone Formation the following have also been recognised within the area of this research: Upper Miller's Dale Lava, Lees Bottom Lava, Shacklow Lava, Litton Tuff, Lathkill Lodge Lava, Bradford Dale Lava, Upper Alport Lava, Lower Alport Lava and the Conksbury Bridge Lava. More recently the Lathkill Lodge Lava and the Conksbury Bridge Lava have been recognised as members of the Fallgate Volcanic Formation (Waters et al., 2006). These have been interpreted (Aitkenhead et al., 1985) as penecontemporaneous lavas, or tuffs. The lavas are predominantly subaerial deposits of vesicular olivine basalt (Aitkenhead et al., 1985), characteristic of continental rifting. Submarine deposits have been identified in boreholes and are associated with thicker volcanic sequences (Aitkenhead et al., 1985). Correlation between the lavas is problematical and cannot be relied upon (Ford, 1977). Furthermore, a number of mining terms have been applied to the lavas including 'toadstone', 'blackstone' and 'channel', which can make interpretation of mining records difficult.

Walters and Ineson (1981) have presented detail with respect to the geographical distribution of the lavas. Earlier volcanic activity is represented within the sequence encountered at the base of the Woo Dale Borehole (Cope 1973).

There are dolerite sills, e.g. at Buxton (Waterswallows) and dykes, e.g. in Great Rocks Dale, near Peak Forest (BGS, 1978). A vent neck and associated dykes have been identified at Calton Hill, above Taddington and others have been identified (BGS, 1978) in Peter's Dale, Monks Dale and near Bakewell (SK 211670 and SK 207684 respectively). A probable vent was intersected in lead mines at Hucklow (Walters and Ineson, 1981). Kirton (1984) presented a table correlating volcanic events of the East and West Midlands and Derbyshire with the South Midlands, the chronology is presented as Table 2.2, below. It is significant in demonstrating that volcanic activity was on-going during mineralization, which forms the subject of Chapter 3 of this thesis.

Table 2.2: Chronology of Carboniferous Vulcanicity (adapted from Kirton, 1984).

Location	Event	Age (Ma)
Monks Dale Vent	Vent intrusion	339
Calton Hill Vent	Vent intrusion	339
	Lower Miller's Dale Lava	336
	Upper Miller's Dale Lava	334
	K-bentonite Clays	329-327
Waterswallows	Hypabyssal Intrusion *	321-312
Calton Hill Vent	Vent intrusion and lava *	301

* Radiometric age; represents a minimum age, because of potential hydrothermal interference (Walters and Ineson, 1981).

Walters and Ineson (1981) have suggested that the centre of activity lay in an area to the east of the limestone outcrop. Evidence for this comes from the distribution of the lavas and from geophysics, for example Biggins (1969), cites aero-magnetic evidence. Aitkenhead et al. (1985) described the Alport volcanic centre where, in contrast with the rest of the Dome, volcanic rocks (the Fallgate Formation) predominate over limestones. There is no sign of the volcanic centre at the surface, as the overlying limestones conceal it. Smith et al. (1985) suggested that the broad distribution of Dinantian volcanism, in a northwest to southeast-trending swathe, extending across the eastern margin of the outcrop, corresponds with the position and trend of the Bakewell Fault. However, whilst the regional geophysics do show a northwest to southeast aligned magnetic basement and the Grantham Lineament appears to be closely associated with basement intrusions, this does not correlate exactly with the Bakewell Fault. This author has also noted alignment of the Bakewell vents with the Ashford Fault and the Edensor Anticline, thus associated with the eastern margin of the shelf.

In addition to lavas, the limestones are commonly interbedded with potassium bentonite clay horizons, interpreted as layers of weathered volcanic ash, and known as clay wayboards, resting on paleokarstic surfaces (Walkden, 1974). The maximum depth of hollows in the paleokarstic surfaces is generally in the order of 2 m, increasing to about 4 m in the Monsal Dale Limestone Formation. The depth probably reflects the depth to the water table at the time of formation and the increased maximum depth in the Monsal Dale Limestone is in keeping with the change in the magnitude of the cyclicity and the sedimentary style (see section 2.4). The clay wayboards are less well developed within the intra-shelf basin facies, which suggests that the basins were not subject to subaerial exposure. In the intra-shelf basinal setting the volcanic dust was commonly mixed with terrigenous inputs to the basin and from this environment shales are interbedded with the limestone (Walkden, 1970). Berry (1984) identified thirty to forty clay wayboards within the Asbian limestones of North Derbyshire. The absence of extrusive igneous rocks and less frequent wayboards from stages earlier than the Asbian is further confirmation of the change in the tectonic setting from this time.

Weathering of the lavas results in clay formation, which can be significant in terms of the hydrogeology. Trail (1939) observed that the lavas were weathered to clay on both the upper and lower surfaces, particularly where they had been in contact with mineralizing fluids.

2.7 Diagenesis.

Key diagenetic processes occurring within limestones have been identified as: microbial micritisation, dissolution, cementation, neomorphism and compaction (Tucker and Wright, 1990). All of these processes have been identified in the course of microscopic studies of the Derbyshire limestones (Berry, 1984; Biggins, 1969; Gutteridge, 1983; Hollis and Walkden, 1996; Padzierski, 1982; Schofield, 1983; Walkden, 1970, 1974, 1977, 1987). The majority of these processes are associated with a reduction in porosity and the low matrix porosity of the limestones has been found to have a significant influence on the hydrogeology. Physical compaction is less efficient at reducing porosity, because it leaves wider pore-throats (Wright, 2002).

Cementation is a process that can occur at all stages of diagenesis (Wright, 2002). Walkden and Berry (1984) and Walkden and Williams (1991) identified five stages of cementing. The stages were designated: Pre-zone 1 and zones 1 to 4. They recorded a history of shallow, meteoric, vadose cementation to meteoric phreatic cementation, followed ultimately by vein and stylolite infill cementation (zone 4) associated with mineralization. Petrological analyses of the cementation has identified zoning within the Zone 2 cement, that corresponds with the cyclicity and emergence attributed to the waxing and waning of the Gondwanan ice sheet (section 2.4). Zone 3 cement has been shown to be the most volumetrically significant cement (70%+ of the pore space), requiring external sources of water and carbonate. It has been postulated that emplacement was during the early Silesian by meteoric water sourced from a karst landscape on the uplifted eastern edge of the Derbyshire-East Midland shelf (see also section 2.8). Light $\delta^{18}O$ values have been determined that could either reflect burial temperatures, an unusually high local heat flow, or highly fractionated hinterland-derived meteoric water (Walkden and Williams, 1991). The requirement for externally derived carbonate and the meteoric source of water suggest a period of significant karst development towards the end of the Carboniferous. Climatic studies suggest that the region was subject to monsoonal conditions and evidence for significant weathering is seen at Leek, where Permo-Triassic deposits overlie heavily weathered and faulted Carboniferous strata (Aitkenhead et al., 1985).

Dissolutional diagenesis takes a variety of forms in the limestones of the Derbyshire Dome. The formation of paleokarstic surfaces (section 2.4), comprises one form of dissolutional diagenesis. Grain to grain dissolution is an efficient process in cementing limestone, commonly associated with stylolite development. Stylolites have been found to play a role in the hydrogeology of the limestones and therefore are described further below.

Berry (1984) observed six styles of stylolite in the Asbian limestones of North Derbyshire and found that they occur along original discontinuities, commonly being associated with clay wayboards and containing black insoluble residues with greenish grey clay associated with some column peaks and troughs, indicating that they are modified former emergent surfaces. They predominantly fall within the non-sutured seam dissolution style of stylolites (Wanless, 1979). Grain to grain sutured seam dissolution was found to be rare within the Asbian limestones (Berry, 1984). Berry (1984) also observed that microstylolites were most common within the darker, finer-grained limestones, which contain a high proportion of non-soluble material and that smooth microstylolites only formed within the darker facies. They were found to be predominantly parallel to bedding, even in dipping sequences of rock. The stylolites are off-set by major faults indicating that they pre-date the Hercynian orogeny. They show no increase in frequency with depth.

There are two other aspects associated with diagenesis that have been shown to have an influence on the hydrogeology. One is the process of dolomitization, which takes a number of differing forms within the context of the Derbyshire Dome. Considerations with respect to dolomitization within the research area have been included as Appendix 2.1. The other, possibly related, aspect is the mineralization of the limestones. Human exploitation of the lead zinc mineralization is pertinent to the hydrogeology (Chapter 3). Aspects of burial dissolution are considered in Chapter 4.

2.8 Post-Dinantian solid geology.

The monsoonal climatic conditions and associated increases in rainfall of the late Dinantian influenced the southerly progradation of deltaic sediments in the Namurian, comprising cyclic fining upward sequences, with more significant cycles being referred to as mesothems (Ramsbottom, 1979). Westphalian sediments were generally finer, being deposited in a delta plain/ estuarine environment. The majority of these sediments must have been eroded from the Derbyshire Dome, but it is suspected that they were formerly in the order of 2 km thick (Quirk, 1993). Remnant evidence of mid Permian to late Triassic, Jurassic, late Cretaceous, or Tertiary sedimentation is scanty. Aitkenhead et al. (1985) pointed to the presence of neptunian dykes of red sandstone and siltstone of presumed Triassic age, for example occurring in the Dove valley (SK 14545673), as evidence of the former Triassic cover. Burek (1991) suggested that this is the product of chemical weathering during a period of prolonged warm weather. At outcrop, in the Leek outlier, where the Triassic cover fills a valley, it can be seen that the Triassic strata were deposited on folded and faulted Carboniferous rocks (Aitkenhead et al., 1985).

Ford, (1977) argued that from the Cretaceous onwards Britain lay on the edge of the downwarping North Sea Basin, the evolution of which was influenced by Arctic to North Atlantic rifting (Cope et al., 1992). Periodic uplift associated with this enabled firstly the stripping of the Cretaceous chalk (of which there is no evidence remaining), Triassic and Silesian strata and then the deposition and weathering of the Brassington Formation (Walsh et al., 1972), which is a Tertiary (Miocene to Pliocene) terrestrial deposit. Evidence of the Brassington Formation comes from its exposure to the south of the subject area of this study, within a large number of solution hollows of various dimensions, extending to depths in the order of 160 m. These hollows occur in a northwest to southeast belt in the order of 20 km long and 4 km wide. The dissolution features are largely defined by the distribution of dolomitised limestones. It is thought that the sediment was derived from the receding Triassic margin and laid down by a braided river system (Ford, 1977).

2.9 Quaternary sedimentation and erosion.

Whilst there are extensive deposits of loess, soliflucted head deposits and screes, there is a paucity of glacial deposits. Aitkenhead et al. (1985) suggest that the paucity of glacial deposits in the Peak District can be attributed to the relative absence of ice during the most recent (Devensian) glaciation. The main evidence for this lies in the laterally extensive occurrence of loess deposits over the Derbyshire Dome. Outcrops of glacial till do occur at a number of interfluve locations, in particular associated with the Wye / Derwent and Lathkill / Derwent confluences. However, these small relicts are attributed to an earlier ice advance. Erratics found on the limestone plateau (330 to 360 m OD) are indicative of Anglian ice having crossed the region. Head deposits suspected to be of periglacial origin and attributed to the Devensian occur throughout. Spreads of glacial sand and gravel derived from various sources occur throughout the Peak District. Quaternary deposits are also preserved within the fissures, conduits and caves of the limestone (Shaw, 1984). Tufa deposits occur and these have, at least in part, been dated to the Quaternary (Aitkenhead et al., 1985). The depositional processes associated with tufa deposits are subject to on-going research. For whilst colder conditions would permit water to absorb more carbon-dioxide than warmer conditions, the amount of plant growth and therefore production of carbon-dioxide is considerably less during the colder conditions. Furthermore some tufa deposition continues to the present day (Chapters 9 and 11).

The earliest recorded evidence of Pleistocene activity in the Peak District occurs in the high-level caves. Speleothem development in Elder Bush Cave, in the Manifold Valley, has been dated (Rowe et al., 1989) to the Pastonian Stage (Oxygen Isotope Stage 63, 1.77 Ma). In Water Icicle Close Cavern above the Lathkill valley speleothem development has been ascribed (Ford et al., 1983) to at least the Hoxnian (Oxygen Isotope Stage 11, 0.4 Ma, or earlier) and in Old Jant Mine, Masson, Matlock glacial sediments show a palaeomagnetic reversal, argued to have occurred at the Bruhnes/ Matuyama event of 0.77 Ma (Noel et al., 1984). Bramwell (1977) described the faunas associated with Victoria Quarry, Dove Holes. These were subsequently described by Burek (1991) as "*rare warm temperate faunas normally assigned to the Lower Pleistocene*" [Villefranchian Stage, Bramwell (1977, p.276)]. During the Hoxnian (Oxygen Isotope Stage 11, or older), relatively warm, moist conditions were conducive to the formation of cement in screes that developed in periglacial conditions and tufas. Associated with this, speleothem development was active in the high level caves at Castleton (Ford et al., 1983). The Ipswichian Stage (Oxygen Isotope Stage 5) was associated with soil formation and in-situ weathering forming extensive deposits of insoluble residue (Burek, 1978; Pigott, 1962). The periglacial conditions that prevailed through the Devensian are associated with the development of scree slopes, dolomite tors, solifluction (and head deposits), soil creep, cryoturbation and ice wedges. The late-Devensian interstadial was associated with the triggering of mass movement of more cohesive strata, including weathered limestone and aeolian deposits. These deposits were subject to cryoturbation during the following cold spell and have been mapped as 'silty drift' (Pigott, 1962).

Burek (1991) assigned pockets of till that occur in the interfluve locations, to the Wolstonian Glaciation (Oxygen Isotope Stage 8). More recently, evidence for the Wolstonian (European Saalian) in Britain has been doubted (Bowen, 1991), particularly as the type location has been assigned to the Anglian (Woodcock, 2000). However there remain arguments in favour of the existence of the Wolstonian Glaciation, in particular the existence of the Welton Till in Lincolnshire, which has been shown to lie between the Hoxnian and the Ipswichian interglacials. Nevertheless, Aitkenhead et al. (2002) have ascribed the glacial till deposits of the research area to an Anglian Glaciation (Oxygen Isotope Stage 12). Burek (1977, 1978 and 1991) described the glaciation as being characterised by extensive till deposition in Stoney Middleton Dale and the valleys of the Wye and Lathkill. Till deposits associated with this glaciation have been found at 60 to 248 m OD. Clast orientation indicates easterly flow and amalgamation with Derwent ice resulting in the placement of a lodgement till seen at Shining Bank Quarry (SK 229652) and Raper Pit (SK 217653). It is considered that the distribution of the till should not be considered in isolation as an indicator of ice movement, for it has been noted by this author that the till appears to be better preserved on the clay rich strata, in particular on the Widmerpool Formation and Namurian mudstones (Figure 2.4). The ice is considered to have been Lake District ice that moved south and came into contact with Irish Sea and Welsh ice. This guided the ice along the Goyt Valley to pass through a narrow col at Dove Holes (Johnson, 1967), along Peak Dale and along the Wye (where it over-rode the ridge at Monsal Head), Dove and Manifold valleys.

The valleys are characterised by a number of terraces (Clayton, 1953; Johnson (1969); Walsh et al. (1972), Waters and Johnson (1958). The terraces are erosional features, thought to be associated with eustatic change in base level (relative change of land or sea level). These have been attributed to knick-point recession following the Pleistocene sea-level changes (Warwick, 1976). Clayton (1953) identified the highest terrace as being of Mio-Pliocene age, using the international stratigraphy (section 2.10). Additionally, there are terrace deposits and Aitkenhead et al. (2002, p. 97) have suggested, "*deposition and vertical and lateral incision leading to terrace formation were primarily controlled by changes of climate, augmented by sustained eustatic uplift. Rivers aggraded coarse sediment more* *rapidly during the periglacial periods when the run-off and bed load increased significantly and when the lower reaches were rapidly adjusting to lowering sea level. These sediments were then incised and terraced during the early part of the next interglacial cycle, with regional uplift facilitating these processes.*" Some of the terraces have been linked to the retreat of the Devensian ice that lay to the west of the area (Aitkenhead et al., 1985). Aitkenhead et al. (op. cit.) observed that the glacial sand and gravel deposits extend through the same height range as the till, although Burek (1991) suggested that they overlie the till. They comprise clean sands with gravel of northwesterly origin. River Terrace Deposits are limited to the area of Bakewell (Aitkenhead et al., 1985), in the order of 1 to 2 m above the level of the flood plain. These deposits, predominantly aggraded during flood events, date to the Early Flandrian (Aitkenhead et al., 2002).

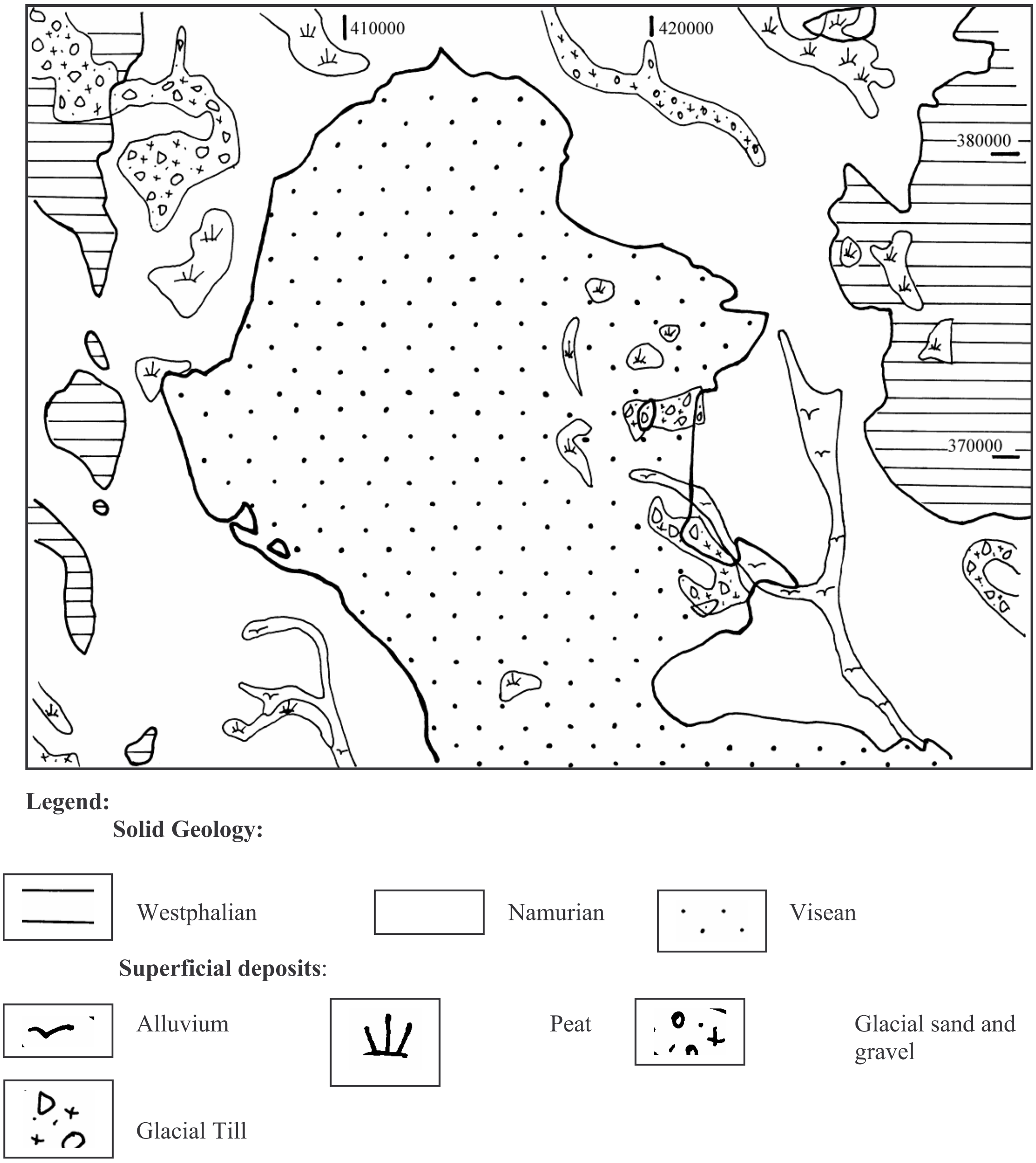


Figure 2.4: Distribution of Glacial Till

Interpretation of the glacial history of the Derbyshire Dome is difficult, because the area has been subject to repeated early glaciations. In considering the results of flow monitoring in the River Lathkill this author has speculated about the possibility of there being buried valleys, because the dales would form a potential focus for subglacial/meltwater channels (section 11.5). A buried valley of one glaciation would be likely to form a focus for valley deepening, as a consequence of exposure resulting from the removal of the channel fill during the subsequent glaciation. The British Geological Survey Mineral Assessment Reports contain a number of boreholes logs. The locations of the boreholes predominantly lie above the valley sides, probably largely reflecting the greater ease of access to these positions, resulting in a dearth of information with respect to the valley bottoms. However, research carried out by Pedley (1993) to investigate tufa deposits has confirmed the presence of buried valleys at SK 216650 and SK 170710 in the Lathkill and Wye respectively. With respect to the Wye Valley, the gorge widens downstream of Cressbrook Dale.

2.10 Geomorphological evolution of the 'Derbyshire Dome'.

The geomorphological evolution of the Peak District has attracted surprisingly little attention. Douglas (1987) stressed the importance of the tectonic history of west-central Britain in influencing geomorphology and suggested that Dinantian landforms now form important exhumed and buried landscape features. Significant examples include the exhumed karst exhibited by knoll reefs, which Warwick (1962) suggested are more susceptible to the development of karst, whereas Ford and Burek (1976) have shown that they have guided parts of the river Bradford and Bradwell gorge. The buried karst associated with the clay wayboards had a possible influence on conduit inception (sections 2.6 and Appendix 2.2). As noted with the glacial geology, interpretation of earlier phases of karst development is difficult where they have been removed, or over printed by more recent processes.

A number of geomorphological processes active during the Quaternary require consideration. These include: glacial and periglacial processes (section 2.9), glacio-fluvial and fluvial processes, and dissolution. Gunn (1985b) points out that it is the predominance of the dissolutional processes that characterise karst systems. Indeed, the importance of dissolution warrants separate review, as presented within the consideration of speleogenesis (Chapter 4) and hydrogeochemistry (Chapter 6). The surface landforms that are associated with dissolutional activity include closed depressions and karren. Below ground the landforms comprise dissolutionally widened fractures, conduits and caves (Chapter 4). The superficial cover of the Peak District, albeit poorly developed, obscures many of the surface landforms. Furthermore the naturally occurring features have been masked by human impacts, such as the extensive mining and quarrying. Implicit in the existence of large areas of limestone plateau that are devoid of surface drainage is the existence of well developed underground drainage systems, which form the subject of a large part of this thesis. Surface expressions include solution dolines, suffosion dolines, buried dolines and subsidence dolines. The identification of buried dolines filled with Head deposits suggests that some dolines pre-date the Devensian glaciation. Most authors note that karren is only poorly developed (e.g. Gunn, 1985a), which suggests that fracturing provides surface porosity.

Fluvial and glacio-fluvial processes are significant in the consideration of the geomorphology of the study area. Extensive upland planation has been chronologically correlated to the Mio-Pliocene (Clayton, 1953). Below this level polycyclic valleys have been cut. Johnson (1957, tabulated by Burek, 1977) identified eight terraces, including the present day terrace, in the River Wye and these have been correlated with the terraces that Clayton (1953) has identified in the Trent catchment. The terraces have proved to be useful in correlation. Burek (1977) suggested that they are likely to correspond to the on-set of interglacial periods and Beck (1980) related periods of speleogenesis to some of the terraces. Both of these hypotheses are logical. The position of the White Peak rendered it susceptible to invasion by glacial meltwater from pre-Late Devensian Pennine and Trans-Pennine Ice (Aitkenhead et al., 2002) and possibly even some meltwater from the Dimlington Stadial. More recent work has questioned both the level (Walsh et al., 1972 and Walsh et al., 1999) and the timing of planation, for example Bowen (1999) has pointed out the poor reliability of dating based on the characteristic vegetation present in the interglacial deposits used to constrain the dating of glacial deposits. Bowen (1999) also noted the ongoing identification of additional stages in the classification of the Quaternary. In the context of the Peak District there is a potential for the identification of more stages between the Bruhnes Matuyama boundary and the Anglian (Oxygen Isotope Stage 12) stage. Ongoing research with respect to the underplating associated with the opening of the Atlantic Ocean and initiation of the Iceland plume (e.g. Tiley et al., 2004) has shown how denudation can be calculated as a function of underplating. Tiley et al. (2004) indicate that the White Peak falls within an area of maximum denudation (possibly up to 3 km). Clearly this sheds uncertainty on the assumptions made by earlier authors and is considered further in Chapter 9.

Some attention has been given to the evolution of the river systems of the Derbyshire Dome. The findings of some of these studies have been summarised below for reference in subsequent chapters. Linton (1956) interpolated pre-Pleistocene drainage at a level of approximately 490 m OD. He postulated a strong east-west trend to the drainage of the area, with the catchment of the rivers Lathkill and Bradford draining northwards into the river Wye, which extended farther east towards the River Derwent.

Johnson (1957) developed a model to show how the Gratton Dale Stream, a tributary of the River Bradford (which follows the edge of the Stanton Basin) that evolved on the Namurian strata in the area of the Stanton Basin (section 2.3) cut back to capture the stream in Long Dale. Using the data on the terraces Johnson (1957) also suggested that the course of the upper Lathkill was established in the Early Pleistocene and that it was captured by the lower Lathkill (by headward erosion to the northwest, across a Namurian cover). Prior to capture by the lower Lathkill, the upper Lathkill would have flowed eastwards to the Derwent. At the time that river capture took place it is considered (Johnson, 1957) that river base level was probably controlled by the level of the interface between the Namurian sandstones and mudstones and that it would have been in the order of 100 m above the current day river level, flowing on the Namurian strata.

Examination of Ordnance Survey maps indicates that the Ashford intra-shelf basin has exercised some guidance on both the pre-Pleistocene (Straw, 1968) and the current alignment of the River Wye, albeit that this may just reflect the basinal control on facies distribution. Johnson (1957) favoured this view; in his interpretation, strike-streams that developed on the Namurian mudstones were subsequently superimposed on the limestone, thereby giving rise to the impressive meander observed at Monsal Head. The identification of the broad Pilsley Terrace, approximately coincident with the Ashford Basin, at a level of approximately 180 to 195 m OD, (above the Hathersage Terrace, which forms a surface at approximately 158 to 177 m OD and is overlain by glacial till (Waters and Johnson, 1958)), led Straw (1968) to suggest that there has been glacial diversion of the River Wye. However, Johnson (1969) suggested that the pre-glacial drainage may have comprised a Wye and a Lower Wye trending northwest to southeast, having had sufficient time after superimposition to develop concordance with the structure of the limestone. The Lower Wye inherited the Monsal Dale meander from the Namurian cover and captured the Wye headwaters by headward erosion.

Ford and Burek (1976) have demonstrated how reef limestones in the gorge of the River Derwent have prevented the uniclinal shift of streams evolved on the Namurian mudstones and caused them to become entrenched within the reef systems. Gorges of Bradwell and Bradford Dales are anomalous in the same way. Warwick (1976) opined that spring-head retreat is a more potent factor than cavern collapse in the formation of new valleys, and that this is followed by uplift, which results in limestone incision, the development of vadose drainage and hanging tributaries. Clearly other processes have influenced gorge development and should also be considered. Such processes might include passages worn by water imprisoned between two parallel mineral veins (Warwick, 1976) and cambering, whereby the stress relief with which this is associated will result in the loosening of blocks, which in the right structural setting may facilitate gorge development.

The occurrence of dry valleys is characteristic of the limestone. They exhibit the same dendritic pattern as is found on other rocks and can be seen to extend on to the Namurian strata. This provides further evidence to suggest that they have been inherited from a drainage network established on an impermeable cover, which has been interpreted as largely mudstones (Johnson, 1969; Warwick, 1964). It should also be noted that prior to karstification the limestone is also largely impermeable and also that further development of the dry valleys is likely to have occurred during periglacial conditions. Associated with the occurrence of dry valleys is the presence of blind valleys, where streams coming off overlying strata sink into the limestones.

2.11 Themes for surface and underground landform development.

A number of implications for surface and underground landform development were interpreted from the geological assessment of the research area. These have provided direction for the research and analysis presented in the following chapters. Some have been shown to be of only minor influence in the evolution of the karst hydrogeology. The main themes have been tabulated below.

Table 2.3: Themes for surface and underground development research.

Theme:
Structural and geological evidence points to the Cronkston-Bonsall Fault being a basement fault. The other main fault in the area of investigation is the Bakewell Fault, which is particularly evident in seismic data (Smith and Smith, 1989; Smith et al., 1985; Fraser and Gawthorpe, 2003) and is suspected to extend to the basement. Both faults appear to hinge from the western side of the dome, suggesting to this author a series of partially formed half grabens, or step faulting formed during an increasingly tensional environment with the effect migrating towards the east and culminating in the intrashelf basin formation noted by Gutteridge (1987, 1989). Clearly there is a potential for this to influence the regional hydrogeology of the Derbyshire Dome (Chapter 5).
In the plateau areas open tensional faults are likely to form a target for surface water (Chapters 4, 6 and 9).
The paleokarstic surfaces and associated clay wayboards would appear to be prime locations for conduit inception. Implicit in this is a form of bedding guidance (Chapters 4 and 9).
Ford (1977) has observed that a product of the alteration of iron minerals in the lavas is pyrite, which is found scattered as small cubes in the contact layers of both toadstone and limestone, thereby providing a source for sulphuric acid generation and consequently a potential for inception
Generally it has been accepted that the lavas, which exhibit low vertical permeability, can form significant aquitards within the limestones. However, it should be noted that jointing and fissuring of the lavas may provide flow paths for groundwater movement through the lava. Furthermore, where the lavas are exposed e.g. in Millers Dale, they appear to have been subject to extensive weathering (exfoliation), which indicates that stress relief has provided greater aerial exposure for weathering and implicit in this is the potential for groundwater movement through the lavas, where sufficient permeability contrast occurs. Sedimentological studies have shown that in some of the limestones the occlusion by cementation has resulted in very low permeabilities, suggesting that relatively there is a potential for clay wayboards (with higher horizontal permeability) to act as aquifers. A further, more complex, speleogenetic process is postulated in Chapter 4.
In the Woo Dale Limestone Formation the bedding is reported to be laterally discontinuous and fissuring less well developed, accordingly different inception mechanisms have been sought (Chapter 4). Dolomitization and dedolomitization appear to be significant in establishing the potential flow paths in these deposits (Appendix 2.1). Consideration also needs to be given to the lithological variation in the Woo Dale Limestone Formation, for these deposits contain micritic facies and in the Yorkshire karst it has been observed that micrite is considerably more soluble than sparite (Sweeting, 1979).
It is conceivable that the accumulations of insoluble and organic matter associated with the formation of stylolites may form a focus for ground water movement. In support of this Schofield (1982) has shown that some mineralizing fluids associated with the Woo Dale Limestone Formation passed along stylolites (Chapter 9).
Warwick (1976) has suggested that the reef facies are more susceptible to karst. He worked extensively in the Dove and Manifold valleys, where marginal reef facies are encountered. Ford (1989) has observed that the upper beds of the reef limestones in the Castleton area were eroded following uplift at the end of the Carboniferous and he noted the occurrence of fissures, or deep grykes. Ford (1989) also noted the development of small phreatic tube cave systems beneath the erosion surface. However, consideration should also be given to the possibility that these fissures could be tension gaps. The karst development observed in the Castleton area is not necessarily reflected in the shelf knoll reefs of this study area. Furthermore it may be argued that the predominance of swallow-holes around the perimeter of the outcrop of the limestones is attributable to the aggressive waters derived from the surrounding strata and in some localities to the presence of tensional faults guiding the points of recharge. Speleogenesis associated with the reef limestones in the study area appears to be dominated by paleokarst. For example, Worley (1978) observed mineralization in hydrothermal karst within a reef deposit in Magpie Mine. Perhaps equally significantly it has also been noted that the dense structure of the reefs does appear to be significant in guiding water movement (Ford and Burek, 1976).
An understanding of mineralization may contribute to understanding the hydrogeology of the limestone (Chapter 3).
The former distribution of the Namurian strata capping the limestone is likely to have been significant in karst development, because these strata would exhibit a lower permeability and give rise to more acidic groundwater, thereby increasing the potential to exert some guidance on the distribution of doline or sink formation (Christopher et al., 1977 and Chapter 9).
The evidence suggests that erosive glaciers invaded the area during the Early Anglian, although much of the cover rock is likely to have been removed before this time. It is likely that high rates of dissolution were associated with periods when glacial meltwaters invaded the area, attributable to both increased volumes of water and elevated concentrations of carbon dioxide associated with renewed growth of vegetation.
It is possible that the melting of valley glaciers could be associated with valley cambering (Chapters 3 and 9).
Burek (1977) observed that whilst tufa deposition is attributed to interstadials, tufa formation in the current interstadial appears to be very limited. This is considered further (Chapters 6 and 10).

Theme:

The significance of the river captures described in section 2.10 is considered in Chapter 9. However, it should be noted that it is suspected that the river captures took place at levels considerably elevated above current base levels. Beck (1980) has already investigated the relationship between cave development and the development of the river terraces.
The knick point and gorge associated with Lover's Leap (SK 07157274) is considered worthy of further investigation, because it appears to transect the dominant structural trend, it reveals significant dissolution along sub-vertical fissure surfaces and it extends from approximately 275 m OD, virtually to the current base level associated with the River Wye (Chapter 9).

Chapter 3: Mineralization and human exploitation of the resources of the limestone.

3.1 Introduction.

The aims of this chapter are first to describe the mineralization of the Derbyshire Dome and then to present factual data on how the limestone resources have been exploited. Boden (1960, p. 53) described how "*Half of the population of England now lives within a 65-mile radius in the industrial conurbations. Yet the population density of limestone areas is and always has been low. In contrast to the marked effect of the industrial growth on landscape of the coalfields, North Derbyshire has changed little, yet it has contributed much to the development surrounding it*." The situation today remains similar, it is reported that 16 million people live within one hour's travel of the National Park, which has a low density and ageing population (Peak District National Park Authority, 1998). The contribution Boden (1960) referred to was primarily that of the limestone itself, which has been and still is utilised in construction, industry and agriculture. Further to this, the limestone has offered additional resources including potable water, lead, zinc, chert and fluorspar. The exploitation of each of these resources has had a differing impact on the hydrogeology of the area, as described below and analysed in Chapter 10.

3.2 Mineralization.

Although a detailed study of mineralization is beyond the remit of this thesis the form and genesis of the mineralization is considered significant to the understanding of the hydrogeology for three reasons, namely:

i) it is clear that the mineralized zones potentially form conduits for groundwater movement. Ford and Worley (1977a) observed that with uplift, the mineralized faults form a partially integrated system of conduits;

ii) the mineralization can be seen as a "frozen" view of fluid movement through the limestone, albeit of hydrothermal fluids moving at the time of mineralization,

iii) methods of mineral exploitation have influenced the hydrogeology.

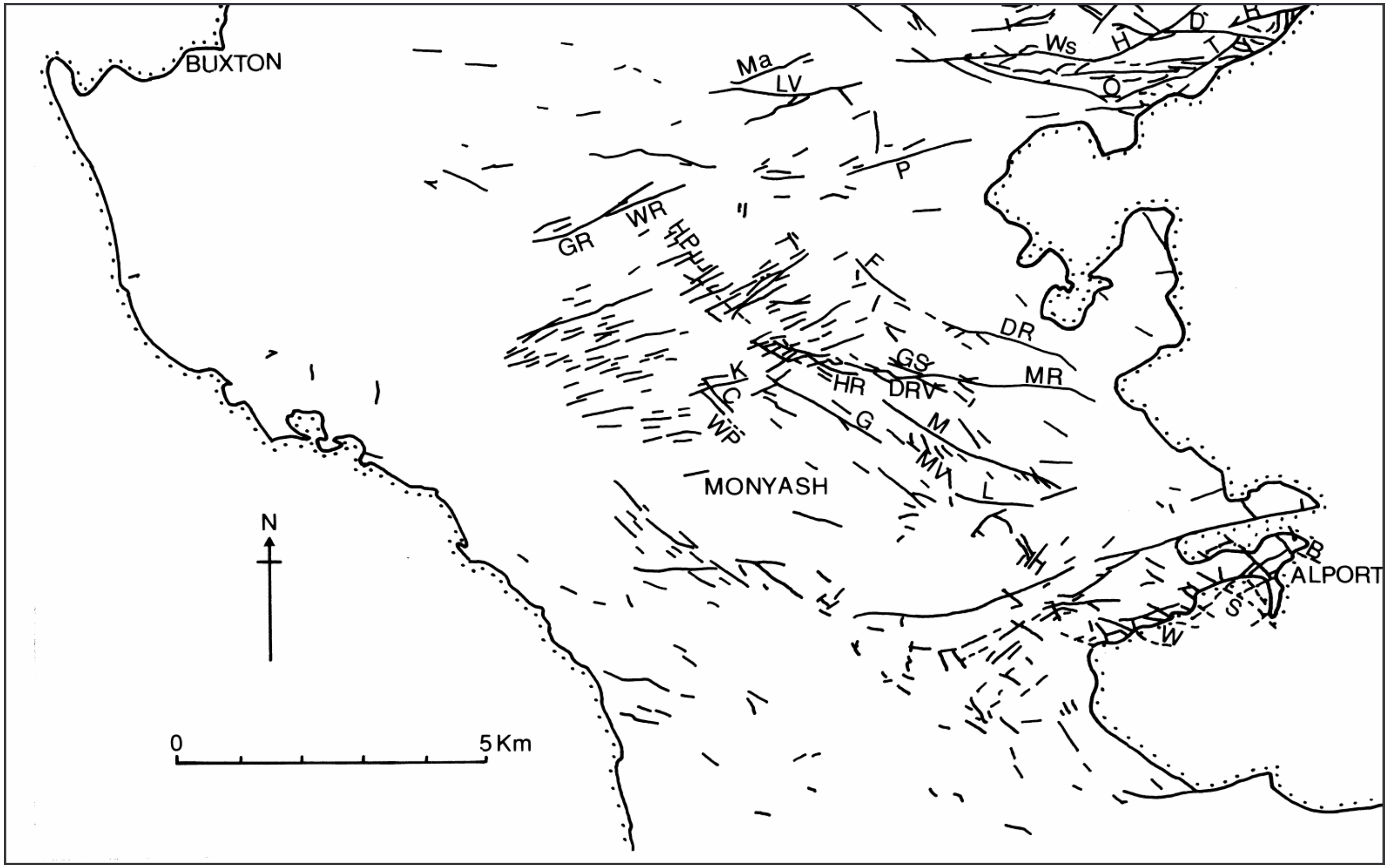
Mineralization is closely associated with the northeastern side of the Derbyshire Dome and with dominant fault trends. Broadly, the mineralized veins trend eastnortheast to westsouthwest and northnorthwest to southsoutheast. Economic mineralization dies out with increasing depth and within this study area the mineralization is generally not shown within the outcrop of the Woo Dale Limestone Formation. Nichol et al. (1970) suggested that the richest ore is developed immediately below the contact with the Namurian strata and beneath the lava horizons in the limestone and that the ore can be traced for many metres along strike, but rarely extends more than about 30 m vertically.

Figure 3.1 shows dominant mineral veins that have been worked in the research area. The mineral deposits comprise fluorite, barite, calcite, sphalerite and galena. They have been classified in terms of their form as rakes, scrins, flats and pipes. Detailed descriptions of the forms of deposit have been presented by Ford and Rieuwerts (2000), Worley and Ford (1977a) and Worley (1978). Quirk (1993), presented a detailed classification of ore bodies, relating the form to the mechanism of mineralization. The ore deposits have also been shown to be zoned with fluorite forming the dominant gangue mineral in the veins in the zone along the easternmost one to two kilometres of the limestone outcrop (Mostaghel, 1983); barite the dominant gangue mineral in the adjacent one to two kilometres, and calcite to the west of this. The zonal changes can be followed along single rakes, e.g. along Long Rake at Youlgreave. However, there are anomalies in the pattern of zoning, as described by Firman and Bagshaw (1974) and attributed to structural guidance and to multiple phases of mineralization (Firman and Bagshaw (1974), an observation that has also been made by this author.

It is generally accepted that the process of mineral deposition was similar to that invoked for the hydrothermal, or metasomatic, Mississippi Valley Ore deposits, having crystallized from fluids of the order of 130˚ C at a depth of ~2 km, remote from the influences of contemporary igneous activity (Lowe, 2000b). A number of theories have been put forward with respect to the genesis of the deposits. This author favours the sequence that has been presented below. The arguments are considered further and a broader range of mineralization processes and element sources is considered in Appendix 3.1.

Following Namurian and Westphalian sedimentation and burial associated with high local heat flow, with temperatures of approximately 80 to 125˚ C, an early phase of fluoride and hydrocarbons was mobilised from the adjacent troughs (or Gulfs). Deeper burial mobilised a second phase of hydrocarbons, associated with the release of iron and manganese-rich mineralising fluids. Subsequently all of the mineralising elements reached a maximum activity and this was associated with an increase in the salinity of the mineralizing fluids. Temperatures during this stage were beginning to cool and reached a maximum of approximately 120˚ C. This is the only phase of mineralization that was associated with significant barium precipitation. Much of the evidence for this come from studies of the mineral vein cements (Hollis and Walkden, 1996). In the area of the dolomitised Woo Dale Limestone it would seem likely that the mineralizing fluids described above post-date an early phase of dolomitization associated with the maximum depth of burial and compression of the Carboniferous Limestone (Appendix 2.1).

In support of the sequence presented above, the following has been noted: very acidic solutions would be required for the transport of lead and reduced sulphide in the same solution (Anderson, 1975); as there is very little extensive evidence of karstification indicative of acidic mineralizing fluids, separate sources of sulphate (carbonate groundwater) and metal (buried mudstones) are postulated. Downing (1967) has shown that sulphate-rich waters are likely to have occurred at depth. Descriptions of mineralization (Worley, 1978 and 1976; Worley and Ford, 1977a and b, Quirk, 1993) and of paleokarst (Ford, 1989) do suggest dissolution during the early stage of lead mineralization and during subsequent stages of mineralization. The presence of hydrocarbons in association with the mineralization has been noted by many authors including Ewbank et al. (1995); Ford and Worley (1977); Ineson and Ford



Key: B Bowers Rake; C Crimbo Vein; D Deep Rake; DR Dirtlow Rake; DRV Dirty Redsoil Vein; F Fieldgrove Vein; G Greensward Rake; GR Grove Rake; GS Great Stones Vein; H High Rake; HP Hubbadale Pipe; HR Hard Rake; K Kiln Hill Rake; L Lathkill Vein; LV Lees Vein; M Mandale Rake; MR Magshaw Mine; Ma Maury Rake; MV Mycross Vein; O Ox Pasture Vein; P Putwell Hill Vein; R Red Rake; S Sutton Vein; T Tor Vein; W Wenley Hill Vein; WR Wham Rake; WP Whalf Pipe; Ws Watersaw Rake

Figure 3.1: Map extract to show mineral vein locations (adapted from Aitkenhead et al., 1985).

(1982), Schofield (1982). The mudstones within the Edale Gulf were noted as being the most likely source for the hydrocarbons by Ewbank et al. (1995). They are also considered to be the likely source for the metals by a number of authors including Kendrick et al. (2002); Quirk (1993); Schofield (1982); Hollis and Walkden (1996). Colman et al. (1989) postulated the concept of seismic pumping as a means of introducing the mineralizing fluids into the limestone and this is supported by this author's observation of the slickenside and striated form of many of the mineralized veins; Ford and Worley (1977a) described how movement on a fault has resulted in convexities and concavities being moved opposite to one another; Worley (1978, p. 206; p. 239) observed: "*replacement took place in a pulsatory fashion*" and "*paragenetic sequences cannot be matched across the orefield indicating the plumbing system was not always open and some parts became sealed off, whilst others remained open in response to faulting*"; and Worley (1978) observed that there was much disaggregation of mineral deposits during deposition, partly attributed to gravity, but partly attributed to phasing of mineralization. Sulphate reduction could be attributable to either bacterial or thermal reduction (Machel, 2001).

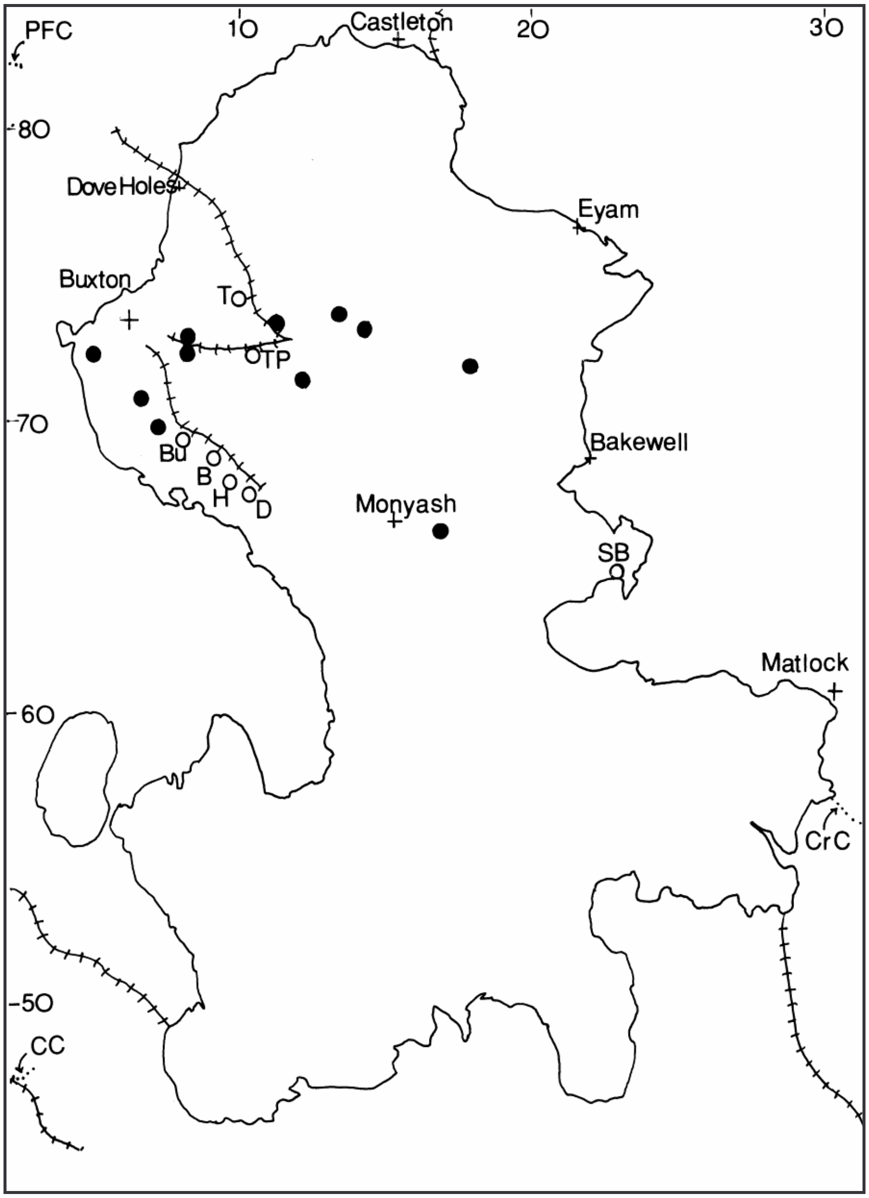
3.3 Chert Deposits.

Chert bands are common throughout the area, particularly in association with the Monsal Dale Limestone Formation and it is generally accepted that the source of the silica is siliceous sponge spicules, as observed by Jessop (1931). However, the presence of chert mines within the area of Bakewell is indicative of a significant source of silica specific to the area, which is associated with the north to south-trending Bakewell Anticline. The close association of the chert with faulting (Brown, 1976) suggests to this author that some of the chert actually forms a mineral deposit, with a source to the east. In support of this, if the Gulfs (Edale, Widmerpool and North Staffordshire) are accepted as the source for the mineralizing fluids there would have been a potential for associated silica mobilization. The mobilization of silica in early mineralizing fluids would also account for the intergrowth of fluorite and quartz that was observed by Worley (1978). Firman and Bagshaw (1974) also described some of the chert as a product of mineralization. The findings of limited experimental work on samples of the chert (Appendix 3.2) demonstrate that some of the chert is associated with mineralization. Another potential source of silicification is the silica released from clay wayboards during pressure solution (Appendix 2.1). The presence of the extensive bands of silica may be significant, hydrogeologically, because of their potential to act as aquitards and also because chert may play a role in speleogenetic processes more specifically in limiting dissolution (Appendix 2.1). A more extensive description of chert deposits and other potential formational processes (e.g. biogenic or associated with clay wayboards) are presented in Appendix 3.2.

3.4 Quarrying.

Quarrying for limestone (Figure 3.2) has a long history; Leach (1996) suggested that in the Peak District its use dates at least to the Bronze Age, when it was used in the construction of barrows and stone circles.

Lime mortar was found in the walls of the Roman baths at Buxton (Jackson, 1950). Initially the limestone was won for building, but then from the late 1500s, agrarian improvements and the early commencement of industrialisation led to a rapidly increasing demand for lime. The demand was so great that medieval exploitation of lime from the commons had to be controlled (Leach, 1996).



Key:
Open circles represent active quarries:
B Brierlow Bu Buxton D Dowlow H Hindlow SB Shining Bank T Tunstead TP Topley Pike
Closed circles represent closed quarries
Hatched line represents freight line railway
Dotted lines represent canals: CC Cauldon CrC Cromford PFC Peak Forest

Figure 3.2: Map to show current areas of limestone quarrying in the Wye catchment (adapted from Harrison et al., 1985).

Development for trade purposes was associated with kiln construction close to the toll roads, the main centres being focused on Buxton and Stoney Middleton. More recently, the considerable effort associated with the working of limestone for building is seen as prohibitively expensive and its current use is largely confined to road construction and railway ballast. The proportion of limestone used in aggregate increased from about 20% of limestone production in 1954 to more than 50% in 1971 (Gunn et al., 1985). Continued increase in the proportion of limestone used in aggregate is indicated by statistics presented by the East Midlands Aggregates Working Party (1992 and 1996). The statistics for 1989 indicate that 69% of limestone sales in Derbyshire were for aggregate, but by 1996 the proportion had increased to 92% of limestone sales. Other specialist uses for the high quality lime from Hillhead and Tunstead quarries, include agricultural applications, cleaning emissions (flue-gas desulphurisation), paper finishing, the water industry, steel industry, the chemical industry (including toothpaste, washing powder, dye stuffs and the sugar beet industry) and construction industries (including cements, aircrete blocks and soil stabilisation). Historically lime was also used in tanning (Bunting, 2006). Limestone that is still quarried as building stone predominantly comes from the Wirksworth area (Leach, 1996). However even within this research area there are localised sources in occasional use, e.g. the Once a Week Quarry, near Monyash.

Transport of the product, be it lime or limestone, has influenced the development of the White Peak. Early transport was by packhorse trails and then by canal (the Peak Forest Canal [Figure 3.2] opened in 1796), with tramway connections to Dove Holes (1797) and to the Cromford Canal (1822), but with the coming of the railways (in the mid- to late-1800s) rail transport became increasingly important. Jackson (1950) described how the opening of the Rowsley to Manchester railway line (1864) provided a focus for the opening of quarries including: Great Rocks Lime and Stone Company in 1864, Buxton Central Lime and Stone Company at Blackwell Mill in 1885, East Buxton Lime Company in Miller's Dale in 1880, and Miller's Dale and Oldham Lime Company at Miller's Dale in 1878. A Branch line from Blackwell Mill to Buxton followed and was opened in 1865, giving rise to J. Slater at Ashwood Dale in 1865 and New Buxton Lime Company at Cowdale in 1898 (Jackson, 1950). Jackson (1950) identified that the High Peak Railway track was re-laid in the 1870s, in standard gauge, and this encouraged Richard Briggs and Sons of Clitheroe to open a new works at Dowlow in 1874. Boden (1960) has suggested that by 1900 most quarries had realigned themselves to one or other of the railways. In 1994, in the interests of minimising the impact of road haulage, investment in the rail equipment ensured that up to half a million tonnes of stone a year could be transported from Tunstead Quarry to Hindlow for processing (Anonymous, 1994) and in 1999 RMC Aggregates partnered with Buxton Lime Industries (owned by Minorco) to form Buxton Rail to transport materials from both Tunstead and Dove Holes to the second runway contract at Manchester Airport (Anonymous, 1999). More recently road transport has become increasingly important as the product applications have become more specialised. Further detail with respect to the history of quarrying is presented by Ineson and Dagger (1985), Leach (1996), Jackson (1950) and Jackson (1964). The main areas of quarrying activity within the subject area of this thesis have been shown on Figure 3.2.

To date no planning permission has been granted for working below the ‘water-table’, however at Topley Pike and also at Tunstead quarrying operations do follow the seasonal fluctuation in the ‘water-table’. It follows also, that some of the quarries form windows into the ‘water-table’. More significantly though, the opening of quarries has a direct impact on storage, particularly storage of the epikarst and deep fissures that have been removed by the quarrying activities. Furthermore, stress relief induced by quarrying (stress relief and blasting) opens an artificial zone of storage deeper in the aquifer and where connections are made with conduits this is likely to result in more rapid spring response to rainfall events. Quarrying also introduces fines into the groundwater system (Hunt, C.O. personal communication, 2007). Processing of the quarried limestone requires water and consequently many of the quarries have licensed ground water abstractions. A list of the licensed abstractions from the Carboniferous Limestones, which dates to 1997, included the following for this research area:

- 600 m^3/day from SK 08716909, licensed to RMC Industrial Minerals Limited (Brierlow)
- 68.20 m^3/day from SK 16906090, licensed to D.S.F. Refractories Limited (Deep Dale)
- 19 m^3/day from SK 10137244, licensed to Tarmac Roadstone North West Limited (Topley Pike)
- 90.90 m^3/day from SK 07906953, licensed to Tarmac Roadstone North West Limited (Hillhead)
- 109 m^3/day from SK 08177262, licensed to Croxton and Garry Limited (Ashwood Dale Quarry)
- 6319 m^3/day from SK 09867248, licensed to Buxton Lime Industries Limited (Topley Pike)
- 23 m^3/day from SK 09846803, licensed to Buxton Lime Industries Limited (Brierlow)

For reference purposes, a more recent list of abstractions from the White Peak, as published by the Environment Agency, has been included as Appendix 3.7. Details of the abstraction values can be obtained by application to the Environment Agency.

Gunn et al. (1985) calculated the volumes of material that have been quarried. These figures have been updated by this author (Appendix 3.8), using statistics presented in the United Kingdom Minerals Yearbooks (1988 to 2002, published by the British Geological Survey), and are presented in Figure 3.3. Clearly the impact of this in terms of water storage requires consideration in terms of the storativity of each the limestone formations and at least an estimation of the contribution of each geological formation to the annual figures.

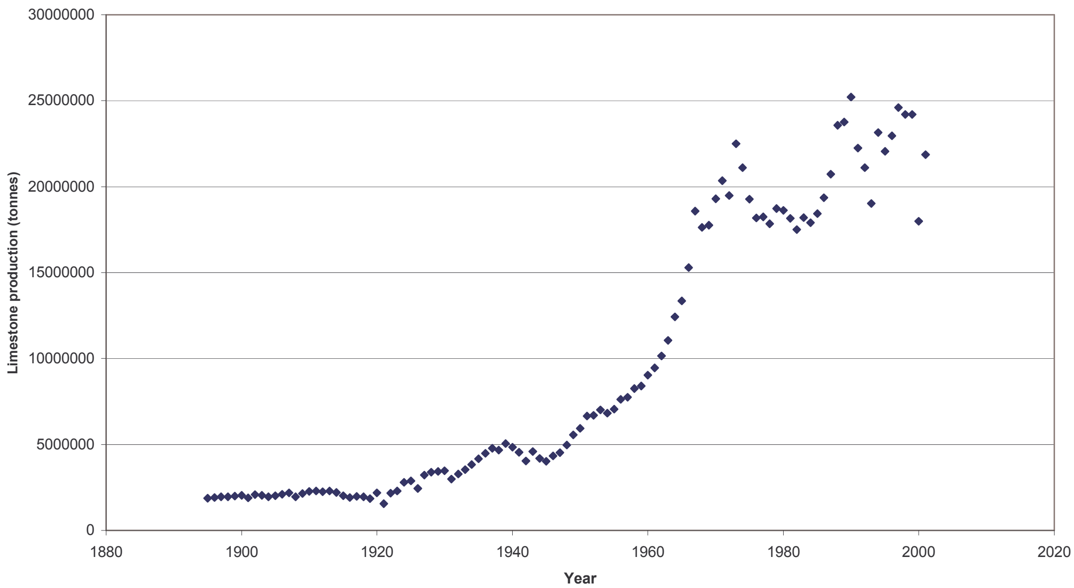


Figure 3.3: White Peak limestone production 1985 to 2001.

3.5 Potable water.

3.5.1 Springs.

Water supply must have remained an issue throughout history, particularly for settlements on the limestone, but surprisingly there are few references to the problems of potable water supply, possibly because records were lost during the various reorganisations in the local authorities and water supply companies and possibly also as a consequence of changes to the administration of potable drinking water supply. This author has carried out searches in a number of places including the records offices at Matlock and Sheffield, the Internet based 'Access to Archives', Derbyshire Dales District Council, Bakewell Town Council, Youlgreave Water Company and Severn Trent Water Authority. A summary of these findings is presented as Appendix 3.3.

The findings indicate that the majority of the springs in the limestone of the Peak District are no longer used for potable supply. The spring supplies were vulnerable to drought and were particularly prone to contamination as a consequence of poor agricultural practice, mining activities and poor waste management. This highlights another of the key areas of potential human impact, which is one of quality. Of particular use to this thesis is the list of wells and springs presented by Stephens (1929), which, together with additional data from Waters (1949) forms the subject of Table 3.1, below. The names of the springs and wells presented are those used by Stephens (1929). Many are now known by different names. Springs with current abstraction licences are listed in Appendix 3.8.

Table 3.1: Limestone springs and wells in the study area, as detailed by Stephens (1929) and edited by this author.

Spring or Well	Parish supplied	Estimated NGR (SK)	Notes	Geology
Aldwark (Duckett Well of Farey, 1811)	Aldwark	228573	A few springs in the village, at 305 to 320 m OD. There was also a supply at Lidgate Farm, which did not fail in the drought of 1921	Junction of Monsal Dale Limestone with Lower Matlock Lava
Alport	Alport	215639	Supplied village and neighbouring farms	Fault in Eyam Limestone
Bleakley Plantation	Youlgreave	21616298	At 213 m OD	Ashover Grit over mudstone
Brook Head	Tideswell, Litton	14367731	Tideswell Water Committee. Yield reduced to 20 000 gallons/day during the 1921 drought. 346 m OD. There was a shortage of water in 1940 (Waters, 1949)	Miller's Dale Limestone over Lower Miller's Dale Lava
Calton Hill Quarry	Blackwell	12107125	Originally supplied by Five Wells, spring at 350 to 366 m OD supplied the village by gravity	Dolerite, although Stephens suggests Carboniferous Limestone
Camphill Farm	Great Hucklow	18207853	Gravitational supply from 378 m OD	Namurian Shale Grit over mudstone
Chelmorton	Chelmorton	11527033	Bakewell Rural District Council Supply. Spring also called Illy Willy Water, at 382 m OD. Later supply to Chelmorton, installed 1926, pumped from old mine shaft. By 1949 Waters (1949) described the supply as a covered reservoir of 1140050 litres capacity and top water level of 369 m OD, which was constructed in 1938. The minimum yield of the spring was recorded as 22800 litres/24-hours.	Monsal Dale Limestone over Upper Miller's Dale Lava
Cressbrook	Cressbrook	16657300	Stephens (1929) states that Cressbrook "receives a supply from a spring at the riverside to the west of the house". This author considers this to be the spring at the grid reference presented.	Base of Monsal Dale Limestone
Five Wells	Taddington	12647110	Bakewell Rural District Council, at approximately 396 m OD. Minimum gauging April 1924, 0.8 litres/sec. Supplied Taddington, Priestcliffe and the Waterloo Hotel. Taddington also relied on collected rainfall. Also previously supplied part of Blackwell. Originally provided by the owner of the Waterloo Hotel	Monsal Dale Limestone over Upper Miller's Dale Lava
Flagg Hall	Flagg	13766905	Supply installed 1926. Disused mine shaft, ground level 322 m OD, water 46 m below. Water chlorinated by 1949 (Waters, 1949), water lifted by pump to a reservoir. Not shown on Ordnance Survey sheet. In 1935 the mains were extended to Monyash	Source likely to be boundary between Bee Low Limestone and overlying Monsal Dale Limestone

Spring or Well	Parish supplied	Estimated NGR (SK)	Notes	Geology
Grangemill	Grangemill	24095775	Supply part from roadside spring at 91 m OD and part from stream, was piped to several premises in Grangemill	BGS (Sheet 111) shows fault in Bee Low Limestone, Stephens (1929) suggests junction of Carboniferous Limestone with Toadstone
Ivonbrook	Ivonbrook Grange	24005877	At 274 m OD, supplied the farm	Boundary of Shothouse Spring Tuff and Monsal Dale Limestone, Lower Matlock Lava below, possibly on fault. Stephens (1929) suggests junction of Carboniferous Limestone with Toadstone
Kingsterndale (now referred to as Ashwood Dale Resurgence)	Kingsterndale	08917222	Stephens (1929) wrongly gives the direction N rather than NE of the Lodge. Strong spring raised by ram to a tank near the Cottage, Kingsterndale. Supplied the Hall, Vicarage and other houses. 248 m OD. Waters (1949) described the supply as two springs issuing at 248 and 253 m OD, from the Carboniferous Limestone, near Pig Tor.	Fault in Woo Dale Limestone
Little Hucklow (Naylor [1983] lists Silver Well under Little Hucklow)	Little Hucklow	15477920	Largely relied on collected rainfall, but Stephens (1929) noted "*has also a few springs*". The only evidence of this on the 1:25 000 Series Ordnance Survey Sheet is Blakes Pool at 366 m OD, at the NGR quoted.	Monsal Dale Limestone
Middleton	Middleton	19996327	Spring near Mill Pond, also supplies Lomberdale House in Youlgreave.	Boundary between Eyam Limestone and underlying Monsal Dale Limestone
Middleton-by-Wirksworth	Middleton-by-Wirksworth	27355658	Gravitational supply to reservoir to supply Middleton.	Monsal Dale Limestone
Monsal Dale	Brushfield	16997110	Supplied by ram from spring in Monsal Dale.	Rises on a fault, Monsal Dale Limestone, immediately beneath the Shacklow Wood Lava
Otter Hole	Buxton	04607330	Stephens (1929) suggested this taps water from fissures in the limestone, it is actually a large spring.	Boundary between Monsal Dale Limestone and overlying Namurian mudstones, with Kinderscout Grit above
Peak Forest	Peak Forest	e.g. 12827941 (Adam Well) and 12977958 (Cop Well)	Relied on shallow wells, liable to pollution. At the time Stephens (1929) prepared his report there was a scheme proposed to bring water from springs along Rushup Edge, to the north. Waters (1949) reported supply from nine springs issuing between 421 and 463 m OD, from the Shale Grit at Rushup Edge.	The wells are likely to be perched on the dolerite
Not named (but must be the spring currently referred to as Cowdale or Rockhead Spring)	Rockhead	08667229	Lifted by ram, supplied Rockhead hamlet. 253 m OD	Faulted contact between Bee Low Limestone and Woo Dale Limestone
River Lathkill	Over Haddon	20806606	Supplied by ram (remnants still evident)	Monsal Dale Limestone over Conksbury Bridge Lava

Spring or Well	Parish supplied	Estimated NGR (SK)	Notes	Geology
Shacklow Wood	Sheldon	17556976	36480 litres/day provided by the Rural District Council, remainder of parish dependent on rainfall. Water raised by water wheel with direct drive to a triple ram pump, to an open reservoir of capacity 8000 gallons and top water level 335 m OD.	Contact between Monsal Dale Limestone and Shacklow Wood Lava
Shothouse Spring	Ivonbrook Grange	24205892	Supplied the Grange, 274 m OD	Boundary of Shothouse Spring Tuff and Monsal Dale Limestone. Stephens (1929) suggests junction of Carboniferous Limestone with Toadstone
(Not named)	Smerrill Grange	19906192		Supplied by roof water only
Tissington Spires	Tissington	14555219	Dovedale. Raised by ram to supply Tissington Hall. 152 m OD	
Tissington: Hall Well	Tissington	17425235	Minimum yield 39400 litres/day. 235 m OD.	
Turner Lodge	Buxton	06207528	Small reservoir, north of the town, taps an underground spring	Kinderscout Grit
Wormhill Moor	Wormhill	10737621	Supplied by Chapel-en-le-Frith Rural District Council. Pumping station, ground water intercepted 1.5 m below ground level and pumped to a reservoir on Bole Hill. Average yield, 228000 litres/day; minimum yield 137900 litres/day. 366 m OD	Miller's Dale Limestone over Lower Miller's Dale Lava
Wormhill Moor	Hargatewall	10677597	Raised by wind-pump. 375 m OD	Junction: Miller's Dale Limestone overlain by Upper Miller's Dale Lava, suspected northeast to southwest- trending fault
Wormhill Springs (West)	Wormhill	12317352	213 m OD	Northwest to southeast-trending fault in Chee Tor Limestone
Wye Dale	Pictor Hall and Woolow in Green Fairfield Parish	09477242 09277249	248 m OD 249 m OD	Fault at boundary between Woo Dale Limestone and Woo Dale Dolomite; Boundary between Woo Dale Limestone and Woo Dale Dolomite

3.5.2 Boreholes.

The use of boreholes for potable supply to the area has not been extensive, largely because of the difficulty in identifying reliable underground resources of adequate yield and also because of the costs associated with deep well construction in the limestone. The following have been identified from a 1997 listing of ground water licences for abstraction from the Carboniferous Limestone:

- 2182 m^3/day from SK 05007150, licensed to Severn Trent Water (Stanley Moor)
- 2000 m^3/day from SK 14207332, licensed to Severn Trent Water (Monks Dale)

Neither of these supplies was being utilised in 1997 and this continues to be the case. However they are considered significant because they identify the location of significant reserves of water. In the case of

Monks Dale the borehole targeted ground water that is confined beneath the Upper and Lower Millers Dale Lava. The Ladmanlow (Stanley Moor) Borehole proved Bee Low Limestone to 73.6 m and continued to 177 m in limestone, but the borehole is situated close to the western edge of the limestone and at least some of the groundwater is likely to emanate from the Namurian, Kinderscoutian strata, with storage in a dominant fault zone. In addition to the public water supply a private water supply was licensed:
- 9 m^3/day from SK 11247277, licensed to K.M. and A.M. Drew (Great Rocks Dale)

Licences have also been issued for abstractions for agricultural use. The following were included in the 1997 listing:
- 16 m^3/day from SK 14207360, licensed to Mr W H Wilkinson (Monks Dale)
- 1 m^3/day from SK 19266854, licensed to W.B. and A.G. Blackshaw (Dirtlow Farm)
- 15 m^3/day from SK 10697596, licensed to Mrs D.M. Drewry (Tunstead)
- 3 m^3/day from SK 10757618, licensed to Mr F.H. Mosley (Wormhill)
- 20 m^3/day from SK 11457191, licensed to Mr R. W. Ardern (Calton Farm)

Industrial process water abstraction licensing included:
- 480 m^3/day from SK 07247193, licensed to Lion Norwestern Foods Limited (Staden). Now transferred to Hockenhull Enterprises (Antigua) Limited, this licence permits i) powder blending, ii) bottled drink manufacture and iii) compensation water (compensation for water abstraction from Rockhead Spring, formerly known as Cowdale Spring).

Wells with current abstraction licences are listed in Appendix 3.8.

3.6 Mining.

There are extensive records pertaining to the history of lead mining in Derbyshire. Primary sources include the Derbyshire Records Office at Matlock, Sheffield Record Office, Chatsworth Estate, and the local studies library Matlock. Further sources include the works of Carruthers and Strahan (1923), Ford and Rieuwerts (2000), Kirkham (1968), Rieuwerts (1980, 1981 and 2000), Stokes (1996) and also the museum and records of the Peak District Mines Historical Society. Records with respect to specific areas and mines are also particularly useful sources and these include: descriptions of the caves and veins of Castleton, the Matlock Mines, the Stanton Syncline and Wirksworth areas (Ford, 2000, 2001, 2003 and 2005, Oakman, 1980), of Millclose Mine (Traill, 1939), of the Magpie Mines (Butcher, 1975; Ford, 1980; Willies 1974 and 1980; Willies et al., 1980, Worley, 1976), of mines in the Taddington area (Barnatt and Heathcote, 2003; Kirkham, 1964; Shaw, 1980, Worley et al, 1978) and of chert mines in the area of Bakewell (Brown, 1976, Critchley et al., 1975). Some understanding of the history, customs and terminology is a prerequisite to interpreting and understanding the primary sources. Mining in the Peak District dates to Roman times and possibly earlier. Evidence for this comes from inscribed Roman lead ingots (pigs) and finds in Early Bronze age barrows (Ford and Rieuwerts, 2000). At the time of Domesday there were seven plumbariae in Derbyshire, including one at Ashford and one at Bakewell. They were suspected to comprise communal, combined ore-washing and smelting sites (Ford and Rieuwerts, 2000). Carruthers and Strahan (1923) suggested that the Derbyshire lead ores were among the first to be worked in Britain. The demand for lead grew, particularly during the Middle Ages, when it was used for roofing and water systems. Specific booms were associated with increases in the economic value of lead and occurred in the mid- 1700s and the mid- 1800s after which there was general decline in the industry.

The main mining areas are detailed in Appendix 3.4. Minerals that were mined include galena, sphalerite, cerussite (White Ore), blende, calamine, fluorite, barite and calcite. A more comprehensive list of minerals can be found in Ford et al. (1993). The ore occurs in rakes, scrins, pockets, caverns, pipes and flats and also replacement ore bodies (Appendix 1.1). Many of the pipes were found by following leaders from the veins. Ford and Rieuwerts (2000) observed that pipes and flats are commonly associated with wayboards. In Mill Close Mine (SK 261624) 'wing deposits' (triangular form of flat deposit) were associated with the underside of clay wayboards (Traill, 1939). The ore itself is usually concentrated between layers of calc-spar, barites, or fluorspar (gangue minerals) parallel to the walls of the cavity containing the mineral, although locally galena is found as isolated crystals scattered through the 'vein stuff'. The form of the minerals and mineralization is covered in detail by Ford et al. (1993), Ford and Worley (1977) and Quirk (1986). Carruthers and Strahan (1923) reported that generally the lead deposits diminished with depth and although workings at Millclose extended close to sea level the reserves at that depth were uneconomic for exploitation. It has been suggested that fluorite rose up the sequence farther, although Traill (1939) observed that pyrite and barite were also more plentiful at higher levels. This suggests that whilst density separation must have occurred, the nature of the mineralization was guided by element supply (Appendix 3.1 and section 3.2).

Although limited evidence of Roman lead mining has been found (Ford and Rieuwerts, 2000; Kirkham, 1968) it is considered that early working would predominantly have been from the surface. A particularly good example of the near surface workings can be seen beside the footpath between Chelmorton and the Waterloo Inn (SK 13157135), albeit that these workings are more likely to date to the sixteenth century, as they were subject to relatively early attempts at dewatering (Appendix 3.4). Towards the latter part of the eighteenth century the shallower parts of many of the veins had been all but exhausted (Carruthers and Strahan, 1923). Deeper mining was carried out via shafts, e.g. in Lathkill Dale in the area of Haddon Grove. The earliest form of mining via shafts comprised the formation of bell pits, examples of which can be found in the area of Elton Moor (Rieuwerts, personal communication, 2005). However, it should be noted that Ford and Rieuwerts (2000) point out that many features with the appearance of bell pits (the ring of arisings at ground level) were merely exploratory pits excavated to examine the vein at surface and not true bell pits. These features are more readily identifiable on aerial photographs and make it easy to identify the line of mineral veins. The results of preliminary aerial photograph studies carried out at the Cambridge Aerial Photograph museum (J. K. St Joseph collection) have been appended (Appendix 3.5) to demonstrate the benefit of aerial photographs in the identification of otherwise unmarked mineral veins.

The administration of the lead mines is unique, with its own set of laws and customs (Appendix 3.6). With such a long history of mining there has clearly been significant scope for technological development (Ford and Rieuwerts, 2000 and Rieuwerts, 1980). A brief summary is presented below because it assists in appreciating the increasing impact of mining with time. Until the middle of the seventeenth century over-night 'firing' was the method used to loosen the ore. Blasting was introduced between 1636 and 1645 and mechanised drilling was introduced in Magpie mine (SK 172682) in 1874. With respect to haulage, hand carting gave way to the horse-gin, whiskets gave way to corves, or sledges; wooden rails were introduced in the eighteenth century, iron rails in the 1820s and steam was introduced in the late 1700s, with the occasional use of boats in later soughs. Improvements in ventilation were brought about by the introduction of cupolas and falling water. Technological aspects of ore dressing and smelting were also improved.

Hillocking, the retrieval of gangue minerals from former waste heaps, was carried out from about 1720, attributable in part to the development of wet sieving techniques (jigging), and has continued to a greater or lesser degree to the present. As technology improved and poorer grade ore could be utilised, organised stoping was carried out in preference to merely following veins. Improvements in technology led to changes in the scale of operation and encouraged the eighteenth and nineteenth-century expansion of many previously worked mines. Ford and Rieuwerts (2000, p. 46) stated that "*The second half of the eighteenth century was distinguished by large ore strikes and some mines returning immense profits, although the smaller mines exploited by the working shareholders continued to be very much in evidence*". Ford and Rieuwerts (2000) and Willies (1976) observed that by the nineteenth century larger interests were acquired, as Consolidated Titles, rather than individual meers along specific veins, with workings being extended to depths in the order of 250 m. It should also be noted, however, that adoption of new technology was patchy and was generally only available to larger concerns. By 1860 there was a rapid decline in employment in lead mining (Ford and Rieuwerts, 2000). The exception to this was Mill Close Mine, which reopened in 1859 and remained open until 1939. Additionally, workings for fluorspar have continued to the present day.

Drainage was a very significant factor in the progress of lead mining and is more significant to the understanding of the hydrogeology, so it is considered in more detail. By the seventeenth century most of the lead had been worked down to the 'water-table' and dewatering became necessary. At first this took the form of rag and chain pumps, or similar, and the directing of groundwater into natural cavities

(Rieuwerts, 1980). Sough driving commenced at about this time, i.e. the excavation of tunnels or adits from the valleys to the mine workings, as a means of dewatering the workings. Rieuwerts (1981) identified three stages in the draining of the Alport mining field (Appendix 3.4). The first phase (pre 1700) comprised the driving of individual vein soughs, the second the construction of cross-cut levels to dewater a group of mines (this included Alport and Grime Soughs [Appendix 3.4]) and finally the construction of Hillcarr Sough. Six of the earliest soughs to have been constructed within the study area were: Hardyhead Sough (1653), Maury Sough (1653), Wheels Rake Old Sough (1660), a sough at Youlgreave (1669), Stanton Sough (1695) and a sough in the Alport area referred to as Tools Sough (Rieuwerts, 1980, Appendix 3.4). These soughs were all driven to individual mines. The earliest soughs followed mineral veins, at least in part to overcome surveying difficulties. Later the formation of cross-cut veins and even main soughs (including Hillcarr sough) followed the Namurian mudstone/ limestone boundary wherever possible, because of the relative ease of excavation in the mudstones (Rieuwerts, 1980). To a certain extent the routes of soughs were influenced by landowners, because the excavation of soughs did not have the same right to free excavation that was offered by mining rights. Water wheels and occasionally even windmills became important and then, during the eighteenth century, steam engines were introduced to the mines, enabling the lifting of water to the pre-existing drainage levels. Kirkham (1952) pointed out that it was essential that the soughs were maintained, silt removal being a routine requirement, in order to minimise the work that was being asked of the pumps.

As noted above, the gradual development of the Alport mining field exemplifies the evolution of technological improvements in drainage. Willies (1976) suggested that soughs prior to Hillcarr (Appendix 3.7) had drained the Alport mining field to the level of the Rivers Lathkill and Bradford. Hillcarr Sough reached Guy Vein (SK 220636) in 1787 and by 1796 the area was drained to about 22 m below this. By 1801 consideration was being given to the installation of water pressure engines to further lower the level of dewatering (Willies, 1976). More specifically, engines were employed in Shining Sough (SK 228648) on Sutton Vein (an extension of Stoney Lee Vein). In 1842, despite problems of ground water associated with the clay wayboards (DRO D504B/L246/23) hydraulic engines were installed in Guy Vein, but the volume of water overcame several attempts of dewatering. Furthermore, it was established that there was a paucity of ore with depth. During this period the best finds were on Prospect Vein (a southerly extension of Break Vein), but this was 18 m above Hillcarr Sough (Willies, 1976). As Ford (1982) has described, drainage proposals comprised the area of mining that most required plans and sections, because the soughs commonly passed land without mineral vein and therefore it was necessary to obtain wayleaves.

Much of the existing information regarding the location of the soughs comes from detailed examination of historical records by Rieuwerts (1980, 1981) and Ford and Rieuwerts (2000). This author has carried out some additional searches and has found that the records are widespread, with variable coverage of individual areas, largely because, as with many records, they are collections from a number of sources and therefore they reflect the interest of the source. The examination of early historic records requires care because of the change in the use of terminology, for example in the interpretation of the word sough. It would appear that the term was sometimes applied to early mines, although this may reflect locations where the sough was driven along the bottom level of the vein, e.g. Smallpenny Sough, Lathkill Dale (Ford and Rieuwerts, 2000). Kirkham (1952) referred to a sough vein as a worked out rake vein utilised for dewatering along its floor. Further problems have been described by Ford (1982), including that of the use of north arrows, which for magnetic north need a date for adjustment for accuracy. Geological observations, which began to appear on plans and sections from about the middle of the eighteenth century (Ford, 1982) can be confusing, in particular the use of the term blackstone, which has been applied to both dark facies of the limestone and to the toadstone (lava). With respect to measurement, the mining unit for measurement along a vein was a meer, but this differed between the liberties (Appendix 1.1). Stokes (1996) listed the following: in the Wapentake of Wirksworth and the manor of Ashford the meer is 29 yards (26.5 m), in the Manors of Peak Forest, Litton and Tideswell 32 yards (29.3 m) and in the manor of Youlgreave 28 yards (25.6 m).

Clearly the principal soughs have the major influence on the current hydrogeology, but it is the minor soughs that provide the better evidence for former ground water levels. Appendix 3.4 includes a summary of the published detail with respect to the soughs of the area. A summary of soughs recorded as discharging water is presented as Table 3.2.

Table 3.2: Summary of soughs recorded as discharging water.

Sough	National Grid Reference	Construction period	Tail Altitude m (OD)	Receiving water	Area Drained	Current status
Maury Sough	SK 15097311	1774	<305	Wye	Maury Vein	Perennial discharge
Hardyhead Sough	SK 12977120	1642/1653	370	Wye via Magpie?	Groove Rake	Perennial discharge
Magpie Sough	SK 17926957	1873-1881	143	Wye	Magpie Mine and Townhead Vein in particular	Perennial discharge
Wheal Sough	SK 160695	pre 1753-1767	250	Wye via Magpie Sough	Hubberdale Mine	Intermittent flow in Deep Dale
Haredale Sough	SK 21606880	1777	124	Wye	Mockshaw Rake	Water issues under corn mill, probably in culvert.
Lathkilldale Sough	SK 20206612	1743-1773	158	Lathkill	Lathkilldale Vein	Intermittent, with perennial flow at Bateman's Shaft (SK 194658)
Mandale Sough	SK 197661	1820	168	Lathkill	Lathkilldale Vein	Intermittent
Gank Hole	SK 18976582	1880s	176	Lathkill	Gank Hole Vein	Intermittent

Sough	National Grid Reference	Construction period	Tail Altitude m (OD)	Receiving water	Area Drained	Current status
Black (Oxclose) Sough	SK 24176580	1743, extended in the 1830s	105	Wye	Cross cut level connecting a number of veins see Appendix 3.4	Intermittent
Smallpenny Sough	SK 18006578		190	Lathkill	Smallpenny Vein	Intermittent
Rowsley Level	SK 25116562	1669	102	Wye	Vein Sough	Long bolt, intermittent seepage
Rainstor Sough	SK 23876545	1720	110	Lathkill	Vein Sough	Collapsed bolt, intermittent seepage
Bowers Rake Sough	SK 23346509	Pre 1718	125	Lathkill	Vein Sough	
Alport Sough	SK 22786487	1706, extended 1749	120	Lathkill	Cross cut level connecting a number of veins see Appendix 3.4	Submerged
Shining Sough	SK 22876482	1756	120	Lathkill	Cross cut level connecting a number of veins see Appendix 3.4	
Harthill Sough	SK 231646	1706-8		Lathkill/Ivy Bar Brook	Cross cut level connecting a number of veins see Appendix 3.4	
Hartle Calf Croft Sough (Priesthill Sough)	SK 230646	1742, 1750	130	Ivy Brook	Vein Sough	
Wheels Rake Old Sough	SK 22756448	1660	c.175	Bradford	Vein Sough	
Leewall Sough	SK 21836435	<1718	130	Bradford	Vein Sough	
Nick Sough	SK 20356390	not known	155	Bradford	Vein Sough	
Dale Sough	SK 199635	1748	170-175	Bradford	Vein Sough	Issues, adjacent to adit
Gratton Dale Sough	SK 20806080		250	Bradford	Vein Sough	
Hillcarr Sough (comprises a number of levels as described in Appendix 3.4)	SK 25845372	1766-1880s	96	Derwent	Alport Mines	Partially blocked, but still perennial discharge

Other materials have been mined (Adams and Cossey, 1978; Ford, 1999, and Ford and Rieuwerts, 2000). These include 'black marble' a dark, fine-grained limestone that was mined at Ashford (where 'Rosewood Marble' was also mined), Birds Eye Marble (from Nettler Dale), and crinoidal limestone from Ricklow Quarry (SK 165661). Although referred to as marbles they are not metacarbonates, but limestones that are attractive when polished. They were worked above the water table. Although they are of local importance they are not considered by this author to have a significant impact on the hydrogeology. Chert was mined from the mid eighteenth century, for use in the Potteries (Appendix 3.1).

Chapter 4: Speleogenesis: the origin of karst aquifers.

4.1 Introduction.

It is beyond the remit of this thesis to consider speleogenesis in detail, but a hydrogeologic assessment of the Wye catchment is impossible without reference to speleogenesis, for as Klimchouk and Ford (2000a, p. 47) stated "*Speleogenesis can be viewed as the creation and evolution of organized permeability structures in a rock that have been developed as the result of dissolutional enlargement of an earlier porosity.*" Whilst there is an extremely extensive literature covering many aspects of speleogenesis, the following discussion has, by necessity, been restricted to areas that this author has found to be particularly relevant to the research area. These areas have been identified from: i) consideration of the hydrogeology in terms of inputs, storage, transfer mechanisms and outputs; ii) the geological and geomorphological detail combined with field studies (Chapters 2, 5 and 11); iii) the results of the examination of geochemical data (Chapter 6); and iv) the results of dye tracing tests (Chapter 7).

Much of the work in this chapter would not have been possible were it not for the very thorough descriptions of the caves of north Derbyshire presented by Beck (1980), supplemented by numerous cave surveys prepared by members of local caving groups. The location of the majority of these caves is shown on Figure 4.1 (caves listed in key, pages 48-50). The aim of this chapter is to draw out those aspects of speleogenesis that can be used to develop a conceptual hydrogeological model for the area of investigation. The science of speleogenesis has its own vocabulary; accordingly many of the definitions used in this section are included in the glossary (Appendix 1.1). The chapter is subdivided between an overview of speleogenetic processes, the regional karst setting and conceptual models (sections 4.2 to 4.5); and hydrogeological considerations in terms of inputs, storage, transfer mechanisms and outputs (sections 4.6 to 4.9); followed by a brief consideration of the application of flow nets.

4.2 Conceptual models of karst hydrogeology.

Theoretical models of karst hydrogeology, at all scales, focus on one or more of the following: input, storage, transfer mechanisms and output in karst systems, commonly with little attention being given to boundary conditions. Historic conceptual models of limestone hydrogeology have been described by numerous authors including: Fetter (2001); Ford, D.C. (1971, 1998); Lowe (1992); Milanovic (1981); Motyka (1998), and White (1969 and 2000). Motyka (1998) presented a particularly useful summary table, which has been reproduced as Table 4.1. The brief descriptions of conceptual models presented below are those that have specific components that have been referred to in the preparation of the thesis; this summary is not to be considered as a comprehensive literature review of conceptual models of karst hydrogeology.

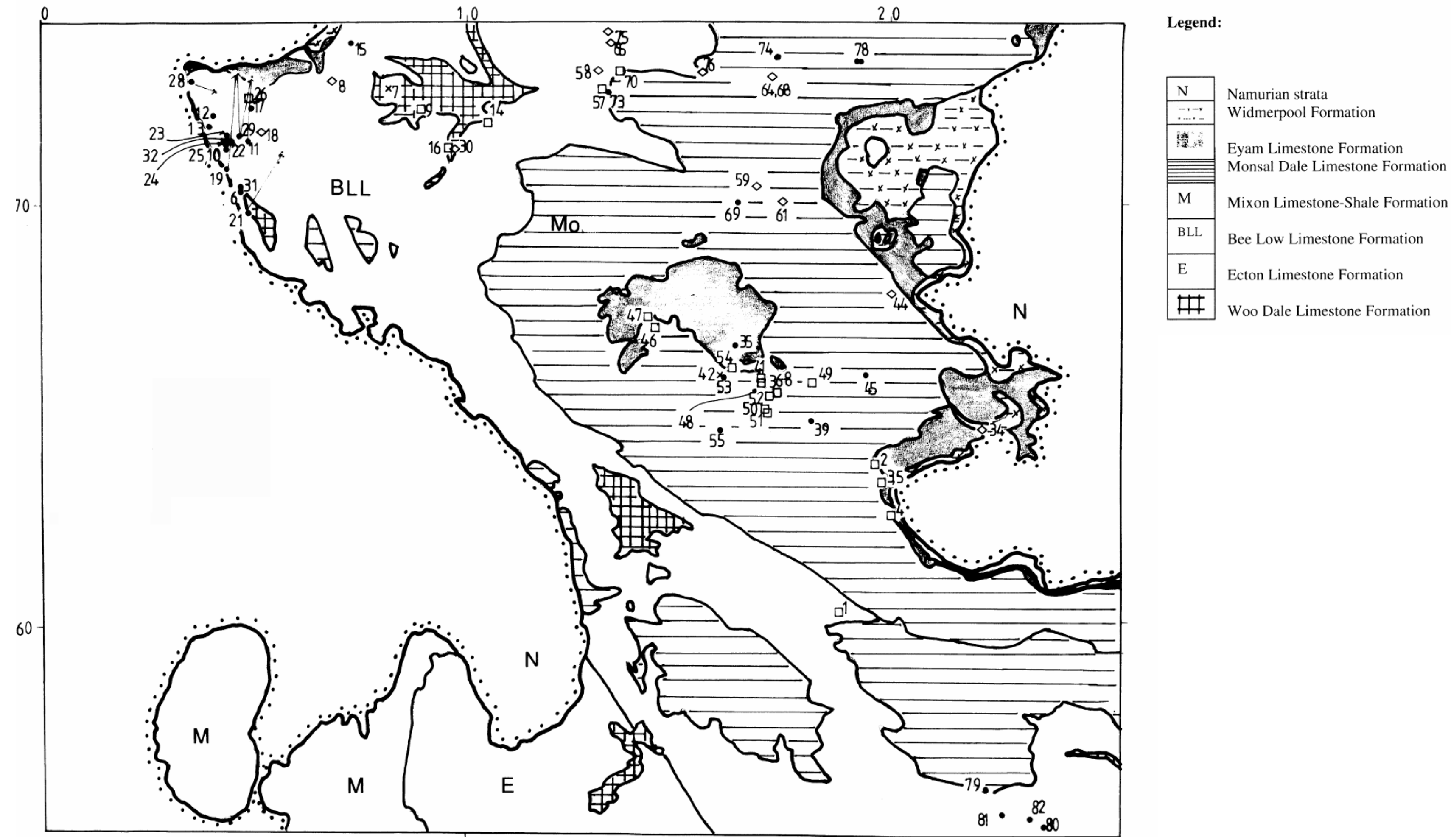


Figure 4.1: Cave locations.
<u>Key:</u> ● Sink (used here to cover both current and relict sinks; ͳ Bedding plane cave; ◊ Rift cave; x unclassified cave.
Many caves fall in more than one category (overleaf)

Key to Figure 4.1.

	Cave Name:	National Grid Reference (SK):	Altitude (m AOD):	Details (Orientation, length, depth, references):	Type:
	River Bradford:				
1	Dolomite Cave	188602	270	5 m long. Branch of Long Dale, near dolomite/limestone boundary. Outer and inner chamber	
2	Lomberdale Cave	196638	235	Entrance on south side of dale.	Bedding
3	Longshot Caves (Rushden Caves)	198633	176	61 m long trends south. Old outlet, water rising at Well Head in main valley. Lower passage parallel to dale side	Bedding
4	Lowfields Cave (Badger Hole)	201625	180	9 m long. East bank of Rowlow Brook. Former outlet for sinks thought to be higher up.	Bedding
5	Owlet Hole	198633	183	3 m long. In cliff face. Closes to a vadose slot. Archaeological excavations (Bramwell and Wood, 1947)	
	Buxton Catchment				
6	Anthony Hill Shakeholes	047703	375	Two small sinks south of Stanley Moor Reservoir (Figure 5.3)	
7	April Cave	082727	299	Cunning Dale, 3 m of passage in Woo Dale Dolomite	
8	Ashwood Dale Cave	06887291	300	North side of dale, immediately to the east of the sewage treatment works	Rift
9	Ashwood Dale Resurgence	08957222	246	Tight passage for 6m	Bedding
10	Axe Hole	044713	375	60 m long north of Stanley Moor Reservoir (Figure 5.3). Traced to Brook Bottom and to Otter Hole and Wye Head via Poole’s Cavern	
11	Borehole Swallet	049715	348	Sink (see also Figure 5.3). Traced to Wye Head via Poole's Cavern	
12	Can Holes 1	041721	366	Sink (Figure 5.3)	
13	Can Holes 2	040718	366	Sink (Figure 5.3)	
14	Churn Holes	10547186	264	60 m long at the head of Marl Dale (Drakeley, 1981 and Turner, 1899)	Bedding
15	Cunningdale Swallet	073738	305	Associated with faults and alluvium	
16	Deep Dale Cave	09267129	292	21 m long bedding plane cave, with rift in wood above (Smith, 1956)	Bedding
17	Green Lane Pot	050726	305	21 m deep, 24m long, part of Poole's Cavern to Wye Head system	
18	Grinlow Rock Shelter	052717	345	3 m long. In cliff face. Closes to a vadose slot. Archaeological excavations (Bramwell and Wood, 1947)	Rift
19	Jake's Hole	04457080	380	24 m long, 11m deep. Sink near south wall of Stanley Moor Reservoir (see also Figure 5.3)	
20	Jake's Hole (lower)	04457080		Small stream sinks and connects with Otter Hole and Wye Head, via Poole's cavern	
21	Leap Edge Swallet (Dale Head Swallet)	04906975	389	South of Stanley Moor, active swallet traced to Brook Bottom (see also Figure 5.3)	
22	Nail Pot	045715	375	15 m long, 15m deep, infilled	
23	Old Bill's Swallet	044716	390	3 m long. Small stream sinks in wet weather (Figure 5.3)	
24	Perseverance Pot	044714	381	37 m long. 27 m deep. North of Stanley Moor (Downhill, 1964)	
25	Plunge Hole	044713	372	9 m long. 15 m deep. North of Stanley Moor Reservoir Traced to Brook Bottom, Otter Hole and Wye Head via Poole's Cavern (Figure 5.3)	
26	Poole's Cavern	050725	335	244 m long. Stream from Stanley Moor swallets	Bedding
27	Resurgence Swallet	04607330	366	37 m long resurges at Otter Hole (Anon, 1963) (Figure 5.3)	
28	Shay Lodge Sinks	035729	396	46 m long, three sinks, water rises at Dog Holes (Figure 5.3)	
29	Stanley Moor Cave	047716	372	21 m long, large doline, Stanley Moor	
30	Thirst House Cave	09707124	271	58 m long. Entrance 4.5 m wide, 6 m long. (Ward, 1985)	Rift
31	Turncliff Swallet	04727034	366	3 m long traced to Brook Bottom, Otter Hole and Wye Head via Poole's Cavern (Figure 5.3)	
32	Virgin Pot	044715	381	14 m deep, doline north of Stanley Moor Reservoir	

	Cave Name:	National Grid Reference (SK):	Altitude (m AOD):	Details (Orientation, length, depth, references):	Type:
33	Waterswallows Cave	079749	335	Stream sink, traced to bed of River Wye	Bedding and Rift
	River Lathkill:				
34	Alport tufa caves	221646	137	6 m long. Small rock shelters in tufa, collapse in 1963 revealed moonmilk in cavities	
35	Boulder Pot	164665	267	Doline	Rift
36	Cales Dale Cave (lower)	174654	206	1055 m long part of Lathkill system, resurgence cave	Bedding
37	Cales Dale Cave (upper)	17306544	232	60 m long, part of Lathkill system, resurgence cave	Bedding
38	Cales Dale New Cave	17366538	204	20 m long, part of Lathkill System	Bedding
39	Calling Low Holes	181648	312	Dolines, one to partially collapsed cavern	
40	Cascade Cavern (Rumbling Hole)	158664	250	41 m long, 18m deep, access via mine shaft	Vadose shaft
41	Critchlow Cave	17106593	204	Part of the Lathkill system	Bedding
42	Engine Close Mine	16056596	251	46 m long, 9 m deep, access via shaft, small natural solution cavities	
43	Freezland Mine	16086594	256	6 m deep shaft intersects phreatic tube	Bedding
44	Green Cowden Quarry Cave	200678	260	Quarry revealed cave	Rift
45	Haddon Hole	194659	167	sink in bed of River Lathkill, lost	Rift
46	Hillocks Mine	145672	285	pipe deposits and bedding plane cave at 51 m depth	Bedding and pipe
47	Knotlow Cavern	14386739	290	Mined and natural cavities associated with pipe deposits	Bedding and pipe
48	Lathkill Head Cave	17076588	206	Lathkill system bedding cave, but upper access via rifting	Bedding
49	Lathkill Resurgence Cave	18136568	183	61 m long, bedding cave below waterfall	Bedding
50	Lynx Cave	17236509	232	9 m long, 45 m above valley floor (possibly valley intersection of southeasterly flow associated with an inception horizon)	Bedding
51	One Ash Cave	172651	234	30 m long, west side of dale, 9m above valley floor (possibly valley intersection of southeasterly flow associated with an inception horizon)	Bedding
52	One Ash Shelter	17306556	232	Rock shelter, low blocked phreatic tube, suggests rift access to inception horizon related cave to this author	Bedding
53	Raven Mine	16096590	268	58 m deep, partly natural shaft, natural passage near base, possibly sink beside reef deposit providing access to inception horizon related passage	Bedding
54	Ricklow Cave	16366607	225	131 m long, 16 m deep, part of Lathkill system	Bedding
55	Water Icicle Close Cavern	16106460	325	213 m long, 32 m deep, this author opines that this is likely to be a relict vadose tube associated with faulting and the feather edge of the reef deposit	Rift
	River Wye				
56	Bee Low Pot	092793	403	Behind Bee Low Quarry, heavily fluted shaft, doline	
57	Blackwell Dale Cave	133728	251	bedding related, opening to valley on a ledge, possibly southeasterly flow path intercepted by Blackwell Dale	Bedding
58	Chee Dale Rift Cave	132732	201	48 m long, 12 m deep, 3.6 m above river level, associated with fault	Rift
59	Demons Dale Cave	1689 7045	198	Large rock shelter, 1.5 m high, 5 m long, large volume resurges in flood	
60	Grindlow Cavern	17257713	240	30 m long, 12 m deep, a shaft to a bedding plane cave	Bedding
61	Hob's House Cave	176712	240	Back of landslide on north side of Fin Cop, narrow descending fissure	
62	Lamb Pots	100795 to 104795	285-300	7 open natural shafts, 9-18 m deep.	
63	Lingard's Cave	17067506	274	18m long, west side of Cressbrook Dale, 46 m above valley floor	

	Cave Name:	National Grid Reference (SK):	Altitude (m AOD):	Details (Orientation, length, depth, references):	Type:
64	Lumb Hole	17257313	192	Cressbrook Resurgence, 24 m long, east side of gorge section of Cressbrook Dale, discharging large stream above lava	Rift
65	Monks Dale Cave	134739	270	9m deep west side of dale 61 m above valley floor, sink to rift	Rift
66	Monks Hole	134750	262	8 m long, 30 m above valley floor, short blind pot, east side of valley	Rift
67	Neptune Mine (Ney Green Mine)	175745	267	348 m long, 28 m deep. East side of Cressbrook Dale adit reaches backfilled rift	Rift
68	Old Cressbrook Cave	17237312	201	15 m long, directly opposite Lumb hole, large entrance, diminishes in size after 9 m, this author suggests that this must once have flowed east to Lumb Hole	Rift
69	Old Woman's Cave	165708	210	13 m long, 6 m deep, 30 m above road, south side of Taddington Dale above crags, chamber with small passages leading off, likely to be fault related former doline	
70	Percy's Resurgence Cave (Millers Dale 4 Spring)	138732	175	Above road, south side of bridge at Millers Dale, small issue, used to supply Dale Hotel with water, appears to rise from beneath lava	
71	Pips Cave	133783	244	8 m long, east side of dale above road, blocked with calcite, suspected by this author to have been fed by doline beneath Head deposits	
72	Pittle Hole	133783	375	14 m deep, immediately west of road connecting Pittle Mere and Little Hucklow, foot of dry valley, shaft to rift, see notes above	Rift
73	Porn Pot	133728	251	Directly above Blackwell Dale Cave 4 m fissure	Rift
74	Ravencliffe Cave	17397356	305	High on the east side of Cressbrook Dale. Single large chamber	
75	Shelob's Lair	134741	274	6 m long, 8 m deep. High on the west side of Monks Dale and north of Monks Dale Cave. Small entrance to the top of a rift. Stream heard below	Rift
76	Tideswell Dale Cave	15557319	213	30 m long, 183 m upstream from the junction of Tideswell Dale and Millers Dale on the east side of the footpath. Bedding plane cave enlarged by mining	Bedding
77	Wardlow Crawl Cave	178756	247	3 m long, just above dale floor level, 125 m down the dale from Wardlow Mires Swallet on northwest side. Blocked by stalagmite	Bedding
78	Watersaw Swallets	19257337 19287338	335	4 m deep, shallow valley north of the opencast workings on Watersaw Rake. Fed by water from the peat cover. Lost	
	Bradbourne Brook				
79	Hoe Grange Quarry Cave	223560	335	6 m long, quarried away (Arnold-Bemrose, 1904), relict sink to bedding plane	
80	Manystones Quarry Cave	237551	320	Solution caves	
81	Rains Cave	226553	330	9 m long chamber, sloping with crawls off chamber (Ward, 1889). Interpreted as a relict sink	
82	Water Low Cavern	233553		Chamber 9 m below ground, 3 m wide and 9 m high, passage at north end	

Table 4.1: Examples of definitions of karstic hydrographic zones (from Motyka, 1998).

<table>
<tr><th>Cvijic (1918); Castany (1963)</th><th>Sokolov (1967)</th><th>Mangin (1975)</th><th>Jakucs (1977)</th><th>Muller (1981)</th><th>Bonacci (1987)</th><th colspan="2">Ford, Williams (1989)</th></tr>
<tr><td rowspan="4">Zone of aeration</td><td rowspan="3">Zone of aeration</td><td rowspan="2">Epikarst</td><td rowspan="2">Zone of infiltration</td><td rowspan="2">Epikarst</td><td rowspan="3">Vadose zone</td><td rowspan="3">Unsaturated (vadose) zone</td><td>Soil</td></tr>
<tr><td>Subcutaneous (epikarstic) zone</td></tr>
<tr><td rowspan="2">Zone of infiltration</td><td rowspan="2">Zone of downward filtration</td><td>Unsaturated (vadose) zone</td><td>Free draining percolation zone</td></tr>
<tr><td>Zone of seasonal fluctuation of water level</td><td>Intermittently saturated zone</td><td>Occasionally vadose and phreatic zone</td><td colspan="2">Intermittently saturated zone</td></tr>
<tr><td rowspan="3">Zone of saturation</td><td rowspan="2">Zone of full saturation</td><td rowspan="3">Eukarst</td><td rowspan="2">Lenticular zone with dissolution by pressure and mixing corrosion</td><td rowspan="3">Saturated (phreatic) zone</td><td rowspan="3">Phreatic zone</td><td rowspan="3">Saturated (phreatic) zone</td><td>Shallow phreatic zone</td></tr>
<tr><td>Deep phreatic (bathyphreatic) zone</td></tr>
<tr><td>Zone of deep circulation</td><td>Zone of the inactive deep karst</td><td>Stagnant phreatic zone</td></tr>
</table>

Table 4.2: Terminology adopted by White (1969) for carbonate aquifer systems in regions of low to moderate relief.

Flow type	Hydrological control	Associated cave type
I Diffuse Flow	Gross Lithology; shaley limestones; crystalline dolomites; high primary porosity	Caves rare, small. Have irregular patterns
II Free Flow	Thick, Massive Soluble Rocks	Integrated conduit cave systems
A Perched	Karst system underlain by impervious rocks near or above base level	Cave system perched – often free air surface
1. Open	Soluble rocks extended upward to level surface	Sinkhole inputs; heavy sediment load; short channel morphology caves
2. Capped	Aquifer overlain by impervious rock	Vertical shaft inputs; lateral flow under capping beds; long integrated caves
B Deep	Karst system extends to considerable depth below base level	Flow is through submerged conduits
1. Open	Soluble rocks extend to land surface	Short tubular abandoned caves likely to be sediment choked
2. Capped	Aquifer overlain by impervious rocks	Long, integrated conduits under caprock. Active level of system inundated
III Confined Flow	Structural and Stratigraphic Controls	
A Artesian	Impervious beds which force flows below regional base level	Inclined 3-D network caves
B Sandwich	Thin beds of soluble rock between impervious beds	Horizontal 2-d network caves

During the 1970s and 1980s conceptual models were largely aimed at characterising karst aquifers on the basis of specific components of the system, e.g. Shuster and White (1971) suggested that transmission processes controlled spring behaviour i.e. conduit vs. diffuse transmission. Drake and Harmon (1973) classified waters on the basis of saturation index with respect to calcite and the equilibrium carbon dioxide partial pressure (open or closed systems). White (1969 and 1977) classified carbonate terrains by relating flow type to hydrogeological environment. The summary table from White's 1969 paper has been reproduced as Table 4.2.

In more recent years these ideas have been developed, for example Palmer (2000; and Figure 4.2). Similarly, the terminology has evolved and although White (2002) referred to diffuse infiltration others such as Worthington and Smart (2004) prefer the term 'dispersed'. There has also been more recent investigation of the form of conduits and their relationship with fractures (Worthington and Smart, 2004), which has been described in section 4.5. Palmer's (2000) conceptual model can readily be applied to the area considered by this thesis and therefore a copy of the model has been presented as Figure 4.2. The concept of phreatic caves has been extended with the introduction of the terms epiphreatic and bathyphreatic (Ford and Williams, 1989).

Gunn (1985 a,b) considered free-flow aquifers as systems, defined by recharge, transfer mechanisms and resurgence, rather than developing a classification based on a single component of the system. Recharge was viewed as being of autogenic (dispersed or concentrated) or allogenic (dispersed or concentrated), which is generally reflected in the chemical aggressiveness of the water. Similarly, Smart and Hobbs (1986) considered aquifers as systems and each of the components: recharge, flow and storage were defined in terms of a continuum with end members, which they suggested could be plotted on three orthogonal axes. With respect to recharge, maximum discharge was seen as the characterizing parameters and flow was seen to vary between conduits and diffuse end members. This model was developed from the perspective of groundwater supply and protection. It incorporated the concept of dynamic storage, reflecting seasonal variations in groundwater storage. White (1993), in a consideration of karst processes, developed a terminology that described a conceptual model of karst drainage. The model was based on the assumption that conduits act as master drains and receive infeeders from sinkhole drains, with upstream infeeders where sinking streams connect. White (1993) suggested that the conduits commonly lie close to base level, but undulate both above and below it. This is similar to the concept of stream order in river basin analysis. The storage volume lying between the water-table and local base level constitutes the dynamic storage and it is this component that maintains spring flow during drought periods. The concepts of continuum of end members and the "ordering" of fissures are incorporated in the conceptual models of Worthington (1999) and Worthington and Smart (2004), which described the karst aquifer as a triple porosity (or permeability) medium, with matrix; fracture (dissolutionally enlarged fractures of <0.01m), and channel, or conduit porosities.

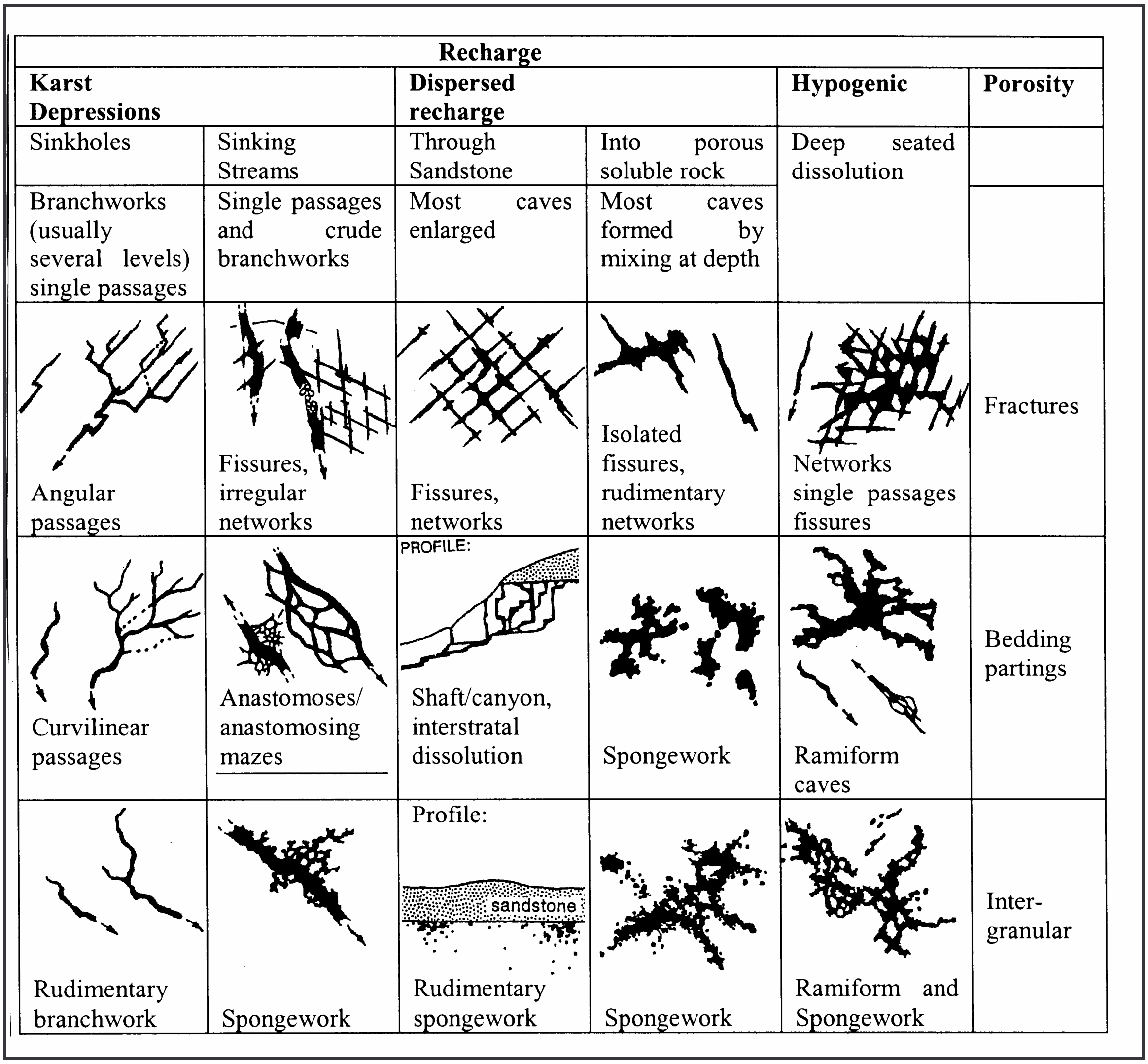


Figure 4.2: Classification of cave plan forms (adapted from Palmer, 2000).

Klimchouk (2003) considered confined (artesian) systems and presented a conceptual model based on the concept of transverse speleogenesis. He proposed that flow paths form across soluble beds (transverse speleogenesis) that are sandwiched between insoluble fissured formations, with horizontal flow along and vertical flow through the soluble horizon. The amount and direction of hydraulic communication across dividing beds depends on the relationship of the heads of adjacent aquifers, which in turn, are influenced by surface topography, with an overall gradual vertical and horizontal transition between net recharge and discharge (Chapter 5). He suggested that there is a specific hydrogeologic mechanism that suppresses the positive flow-dissolution feedback (increased dissolution achieved by increased discharge, particularly on achieving breakthrough, section 4.4) accounting for the development of more homogeneous and more pervasive channeling and maze patterns. This results from the decrease in the vertical hydraulic gradient achieved by breakthrough in the confined setting. Klimchouk (2003) considered that groundwater aggressiveness can be achieved by: groundwater mixing, groundwater cooling, sulphate reduction and dedolomitization. Rates of vertical water exchange depend on: permeability, thickness, continuity, number of dividing beds and tectonic regime (the process being favoured in zones of uplift). Klimchouk (2003) observed that the ratio of the cave length to the field area and the cave porosity are an order of magnitude higher in confined, compared with unconfined settings. In the context of the White Peak, this may be applicable to the processes involved in the dissolution of carbonates at the base of the limestone.

Whilst generalized conceptual models are important in guiding thought processes with respect to limestone aquifers; it is apparent from the vast array of models and the application of different models to different aspects of karst hydrogeology, that there is not a single a model that adequately characterizes all karst aquifers. Furthermore, the conceptual model will vary according to its proposed end-use. Ford (1998, p. 91) suggests that "*General Systems Theory is broad and non-specific, however it invites researchers to organize their investigations of interesting phenomena in a systematic manner but does not push them towards conclusions that will be confined within some limited range of concepts,..*" It is also true to say that it is only with a sufficiently detailed and perceptive conceptual model that any attempt of mathematical modelling can be carried out. Accordingly, it is important to realize that the derivation of a site specific conceptual model is an essential tool in the analysis of a karst aquifer. It is this aim that forms the focus of Chapter 9 of this thesis and the following sections contribute to the detail of the model.

4.3 Regional setting and ages of karstification.

At first sight the regional speleogenetic setting may seem irrelevant to the hydrogeology of the catchment of the River Wye. However, water tracing (Chapter 7) has proved a connection that is at odds with the current hydrogeological setting, i.e. the connection between Illy Willy Water at Chelmorton (SK 11537033) and Ashwood Dale Rising (SK 08957223). Accordingly, it is postulated that the associated flow paths correspond with superceded aspects of the regional hydrogeology (Chapter 9) and this section considers the karst setting in the context of current karst terminology.

Gunn (1992) described the current setting of the White Peak as ponded karst; the limestone is surrounded by non carbonate beds. This certainly describes the surface exposure of the limestone and is probably an appropriate model for the interpretation of the unconfined aquifers; however consideration should also be given to the easterly and southeasterly continuation of the limestone beneath a cover of younger strata. The geological structure described in Chapter 2 indicates that the regional karst context is one of a disrupted basin, i.e. layered structure prevails, but it is disrupted by block and cross-cutting structures, which result from the stress history of the region. The current exokarst setting can be classified as open, unconfined subsurface per descensum karst, with a thin mantle (Klimchouk and Ford, 2000a). The geological description presents evidence for syngenetic, paleokarstic surfaces, associated with penecontemporaneous lavas providing a potential for interstratal karst gestation. However, intrastratal karst also appears to be significant in the unconfined meteoric zone. Additionally, the pattern of mineralization within the Peak District indicates the occurrence of cross-cutting expulsion water circulation, with some evidence of per ascensum hydrothermal karst development (see below), associated with a depth of burial of 2 to 3 km. The presence of a former Silesian cover to the Peak District suggests that the karst has evolved through a sequence of subjacent, entrenched and exposed karst. Accordingly, the system was once confined and consideration needs to be given to those aspects of karst, which have been inherited from earlier stages. The situation is further complicated, because, as Palmer (1991 and 2000), Worthington (1991), Worthington and Ford (1995) and Klimchouk and Ford (2000a) identified, there is an increasing awareness that dissolutional features originate at deep-seated horizons, without apparent relationship to the land surface.

Klimchouk and Ford (2000b) emphasised the evolving processes associated with basinal evolution through eogenesis, mesogenesis and telogenesis, but balanced this with the concept of inheritance. In the context of the White Peak, the majority of the evidence relating to groundwater circulation during eogenesis is related to cementation. The potential for deep-seated flow path development is considered in section 4.4. The expulsion of depositional (connate) waters appears to have been linked with the development of regional flow paths, which are independent of the current hydrogeology (Chapter 5). The process of mineralization appears to be associated with the dewatering and with hydrocarbon evolution and metalliferous release and expulsion from adjacent basins, namely the Edale, Widmerpool and North Staffordshire Gulfs, during mesogenesis (Hollis and Walkden, 1996; section 2.7). Klimchouk and Ford (2000b) suggest that mesogenesis is generally associated with more uniform speleogenesis. In the context of the White Peak, speleogenesis during mesogenesis probably comprised three phases: basinal dewatering, migration of hydrocarbons and mineralization with contemporaneous and subsequent cementation.

It has been suggested by Klimchouk and Ford (2000b) that the highest degree of heterogeneity of porosity develops towards the end of mesogenesis, as the carbonates emerge to the subsurface and surface. This reflects a number of processes, including stress relief (particularly associated with the opening of bedding planes); changes in stress associated with changing tectonic regimes; circulation by meteoric water and denudation. This can also be said of the karst of the White Peak. However, although there is heterogeneity of porosity, it is not random, for as Palmer (1991, 2000) has described, the location of caves is determined by the distribution of recharge and discharge areas. Furthermore, in considering the catchments of the River Lathkill and the River Wye it should be noted that there are differences in the geologic setting. In Lathkill Dale the hydraulic gradient coincides with the dip of the beds, whereas along sections of the Wye valley the hydraulic gradient is along the strike of the limestone. Klimchouk and Ford (2000b) argued that with time the importance of fabric diminishes in favour of fissure network porosity, which can be accounted for by the increasing impact of stress relief as a result of uplift.

Implicit in the above description is a Carboniferous age for the hydrothermal per ascensum paleokarst. Ironically, timing with respect to the more accessible per descensum karst is not so clear. Palmer (2002), having assessed the kinematics of the laminar flow associated with inception and breakthrough (section 4.4), presented a nomogram, which relates breakthrough times to the initial opening aperture, temperature and the ratio of the hydraulic gradient to the flow path length (Figure 4.3). This author has considered the implications for the inception horizon from which Lathkill Head Cave forms an effluent, overflow cave. A level survey carried out by this author (Appendix 11.1), indicates a gradient of approximately 1:90. Taking an assumed length of 5.5 km for an initial opening of 0.1 cm, a breakthrough time of 10^5 years is obtained from Palmer's nomogram. Reducing the initial opening to 0.01 cm increases this time to the order of 10 million years. The amount of initial opening would reflect the state of stress of the rock mass, with a greater potential for inception following uplift. White (2002, p. 94) states that the "*evolution from fractures to conduits is discontinuous*", this is because of the thresholds reached and consequential increase in dissolution rates associated with reaching the critical width of 0.01 m. Once this threshold is reached conduit enlargement to the size of typical cave passages can be achieve in a time scale measurable in thousands of years (White, 2002).

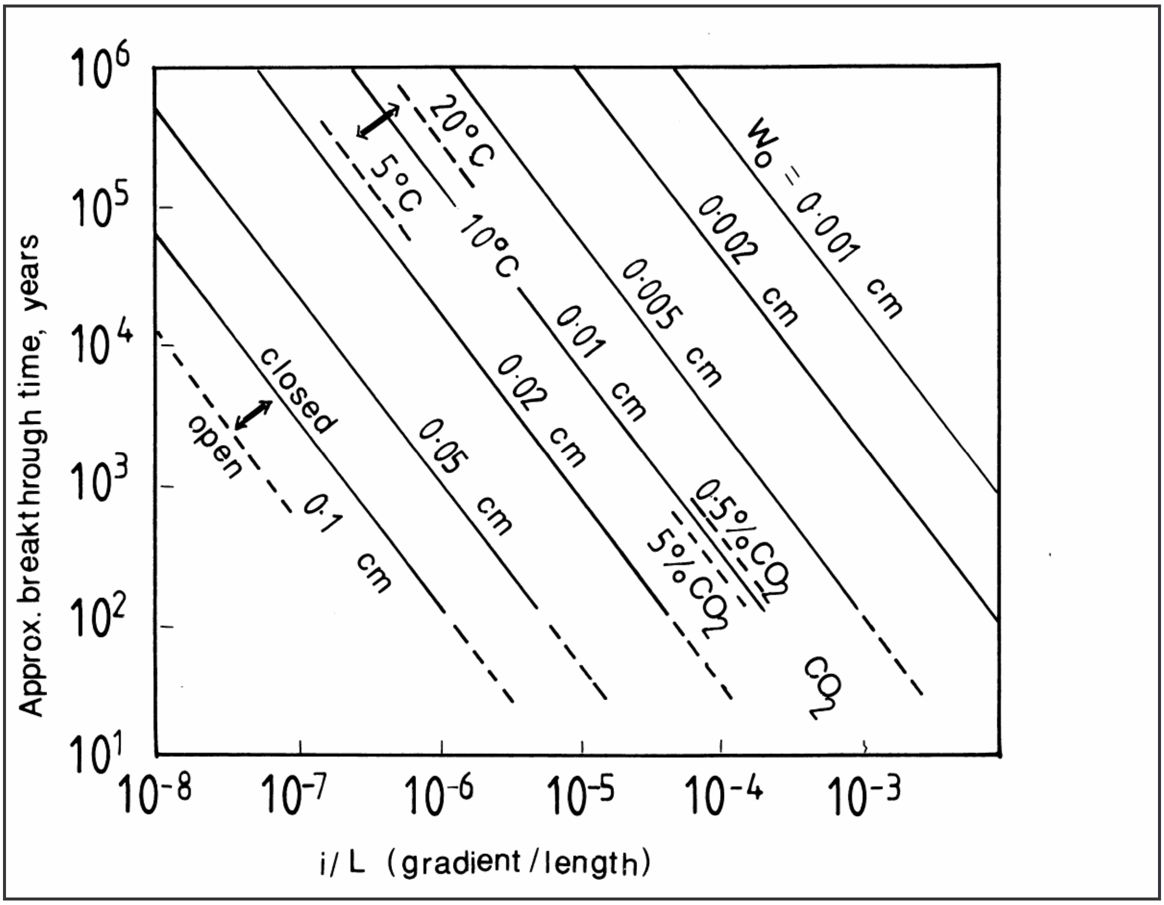


Figure 4.3: Breakthrough times (adapted from Palmer, 2000).

These findings lead to two alternative possibilities reflecting the estimate of the initial opening. One is to conclude that the process of valley cambering, which is evident in Lathkilldale and could be attributed to unloading of ice following an early Anglian stadial, facilitated gestation. The alternative is that gestation dates at least to Miocene uplift (in the order of 25 Ma) and possibly to Palaeocene (60 Ma) uplift, associated with the opening of the North Atlantic when climatic conditions became more temperate following the dry climatic conditions of the late Cretaceous. The latter is preferred by this author. Firstly, because not all of the caves systems described by Beck (1980) are concordant with current valley settings, for example those associated with Blackwell Dale (Figure 4.1) and secondly, in consideration of the available data regarding the dating of speleothems (section 2.9).

Isotope dating of speleothems provides minimum ages for the last significantly erosive phase of speleogenesis and although it does not place a time on inception it indicates the pre-existence of conduits. The oldest dates in the White Peak have been attributed to the Pastonian and were recorded by Rowe et al. (1989) (section 2.9). There is further evidence for pre-Pleistocene speleogenetic activity. For example, Ford (1977) points to the fact that the dissolution hollows containing sediments of the Brassington Formation may provide an indication of pre-Tertiary base levels of the area, although no resurgence caves have been found to confirm this and the actual level has not been defined, e.g. Walsh et al. (1999). Interestingly, remnants of Namurian mudstones were found in the dissolution hollows that were associated with the Brassington Formation (Walsh et al., 1972 and Ford, 1977). This indicates that although later sediments (Permian to mid Cretaceous) may have been eroded, possibly the mudstone cover was probably maintained until the phases of uplift associated with the Alpine orogeny (Eocene epoch, 60 – 35 Ma).

4.4 Dissolutional processes and speleogenetic development.

Understanding the dissolutional processes operating in the karst environment is essential in the identification of potential speleogenetic processes; accordingly this section explores these processes. In the unconfined setting, dissolutional processes are dominated by carbonic acid reactions fed by carbon dioxide derived from a range of biogenic processes (Atkinson, 1977b). Palmer (1991, 2002) described unconfined speleogenetic processes in limestone, identifying that initial openings are slowly enlarged by groundwater, which is close to saturation with respect to calcium carbonate, and that the routes conducting the highest discharge are subject to greater dissolution. Consequently, conduit development requires a substantial hydraulic gradient and conduit initiation is commonly attributed to uplift. For a given partial pressure of carbon dioxide, the initial rate of dissolution decreases in an approximately linear manner with increasing calcium carbonate content, but at 60-90% saturation the dissolution rate decreases rapidly (the kinetic threshold), enabling deeper penetration and delaying the final stage of saturation with respect to calcite. Thus, slow uniform dissolution is required for the gestation of long passages. Favoured routes achieve breakthrough in the time spans indicated on Figure 4.3.

The solubility of dolomite is less than that of calcite, but research on the dissolution kinetics of dolomite (Herman and White, 1985), especially with respect to the effects of lithology and fluid velocity, indicates that dissolution of dolomite proceeds in a similar manner to that of calcium carbonate. The experimental work of Herman and White (1985) showed that fine grained samples were prone to higher dissolution rates than coarser grained ones and preferential dissolution occurred at the outer edges of rock grains and energetically favoured sites within individual crystals. At low levels of saturation the rate of dissolution was less affected by the Reynolds number (Herman and White,

1985) and sedimentary and hydrothermal dolomites dissolved at comparable rates. Observations made by Bakalowicz et al. (1987) demonstrated that in settings where calcite and dolomite are both present, dolomite dissolution will continue as calcite saturation is reached.

Karst aquifers are usually described in terms of a triple permeability model (White 1988 and 2002) comprising matrix and fracture permeability, connected by channels (Worthington, 1999 and section 8.3), or conduits. Flow in pre-dissolutional openings is laminar. As dissolution proceeds and openings reach the hydrodynamic threshold (generally considered to be 10 mm, Fetter, 2000), turbulent flow develops and the rate of carbonate rock dissolution increases rapidly as circulating water becomes less saturated. This opening width also coincides with the kinetic threshold and the onset of clastic sediment transport (White, 2002). In practice there must be a continuum of fissure to conduit and cave sizes and although Smart and Worthington (2004) point out that there is little capillary movement of water upwards through the unsaturated zone to sustain surface evaporation it is the opinion of this author that this probably reflects the shape of fissure (tending to increase in size upwards as a consequence of per descensum dissolution), rather than the absence of capillary water.

The time required to reach the maximum rate of wall retreat (generally ~ 0.01 to 0.1 cm/year) varies inversely with flow distance and temperature and directly with initial fracture width, discharge, gradient and concentration of carbon dioxide (Palmer, 1991). The rate of dissolution increases with discharge until a maximum is achieved. During periods of flooding, flow rates are higher and water is generally more aggressive, with a lower pH and lower concentration of calcium. Furthermore, additional flow paths will be activated, both in the vadose and the phreatic zones, facilitating rapid dissolution and increasing the efficiency of the system. As the relationship between calcite solubility and temperature is an inverse one, cooling of rising thermal water can increase its aggressiveness. Similarly, vadose water may be cooled as it approaches phreatic water, thereby making it increasingly aggressive. Undersaturation with respect to calcite can also result when two solutions, each saturated with respect to calcite, but in equilibrium with vapour phases at different values of carbon dioxide, are mixed, an effect referred to as mixing corrosion (James, 2004). Thrailkill (1968) observed that rates of carbon dioxide out-gassing can exceed rates of calcium carbonate precipitation, resulting in super saturation, with a potential to mask any mixing effects. The form of cave networks reflects the hydrological processes (Palmer, 1991, 2000 and 2002). This is attributed to flow path competition and to the influence of the low head of main passages in capturing recharge from the surrounding area, although this only occurs in the phreatic zone. In the vadose zone there are fewer tendencies to convergence as passages maintain hydraulic independence in targeting the phreatic zone. In the final stages of conduit development the rate of enlargement decreases as the conduit becomes air-filled and loses aggressiveness due to the loss of carbon dioxide. Condensation corrosion is possible in this environment (James, 2004).

Newson (1971) emphasized the importance of abrasion and sediment transport as an erosive process in cave streams. Studies of scallops (Gale, 1984 and Lauritzen et al., 1985) suggested that abrasion is most effective during periods of flood (generally corresponding with the upper 5% of flow). The thin nature of the soil cover over much of the White Peak (Burek, 1977) would suggest that surface sediment sources are limited, although there is a plentiful supply of largely aeolian sediment that was washed into the karst system during the Pleistocene, with a potential for mobilization during periods of flooding. In practice it is more likely that the sediment acts as a form of armoring, thereby slowing speleogenetic processes.

Even in the unconfined setting, carbonic acid is not always the source of aggressiveness in carbonate dissolution. In Level Crevice Cave, Dubuque, Iowa, Morehouse (1968) identified sulphuric acid, produced as a result of the oxidation of pyrite and marcasite, as the cave forming solution in the Ordovician Galena Dolomite. The depth to which oxygenated groundwater circulated limited the depth of the process. Lowe (2000b) attributed the inception of Notts Pot Cave and Short Drop Cave in the Yorkshire Dales, England to the effects of the pyrite rich Notts Pot Coal Seam sequence. The occurrence of limonite to considerable depths, for instance in the Bee Low Limestone in a borehole in Bee Low Quarry (SK 08547904) to a depth of 100 m (Harrison, 1981), indicates the occurrence of oxygenating conditions as a result of water circulation, either now or at some time in the past with a consequential potential for pyrite oxidation to considerable depth. Other sources of pyrite were identified in Chapter 2. Worley (1978) noted corrosion associated with oxidation of marcasite (Appendix 3.4).

Egemeier (1981) described a process termed “replacement-solution” in thermal caves in Wyoming. This involves the replacement of limestone by gypsum crystallizing from a thin film derived from aqueous oxidation of gaseous hydrogen sulphide originating from thermal springs, associated with regional flow paths. A hydrocarbon source has been established for the hydrogen sulphide (Hill 1994 and Jagnow et al., 2000). Subsequently, dissolution of the gypsum occurs as it reaches unstable proportions and collapses from the cave roof, to be removed by the underlying streams in solution. This achieves upward development of the caves. Associated with the thermal springs are metals deposited in hydrogen-sulphide rich muds. Such processes may have been active during the development of paleokarst (section 4.7), associated with the lead mineralization of the Peak District.

In the context of confined flow, later Palaeozoic, Mesozoic and Cenozoic strata once capped the limestone of the White Peak. Thicknesses in the order of 3 km were suggested by Ford and Gunn (1992), although Walkden and Williams (1991) and Hollis and Walkden (1996) suggested that areas of limestone were probably exposed by the mid-Triassic. Prior to the erosion of these strata, the processes of transverse speleogenesis (Klimchouk, 2002) may have been active. Furthermore, the geological setting suggests the potential for evaporite dissolution at depth. Evaporite dissolution rates are commonly higher than those of calcite and the switch to high order kinetics occurs when solution is extremely close to saturation (Klimchouk, 2002). Accordingly, conduit formation in gypsum requires steep gradients, wider initial openings, or shorter flow paths to achieve breakthrough. Klimchouk (2002) suggests that this will result in the formation of either maze form caves, or discrete conduits.

Under pressure the plasticity of gypsum can cause it to deform and even flow. Where localized occurrences of evaporites have been proved towards the base of the Carboniferous Limestone in the Peak District (as at Eyam, Dunham, 1973), steep hydraulic gradients are suspected to be associated with the rising and falling limb of the regional flow paths (Chapter 5). The chemical interaction between carbonates and sulphates can greatly increase the solubility of dolomite, gypsum and anhydrite (Palmer, 1995). This process is discussed further below.

Palmer (1995) suggested that a common origin for deep-seated porosity is the bacterial or thermal reduction of sulphates in anoxic zones by organic carbon compounds. Such processes have been considered in the analysis of the mineralization of the Peak District (Chapter 3). It is considered that active dissolution of deep sulphates is also ongoing (Chapters 5 and 6). In closed systems, replacement of gypsum or anhydrite by calcite can produce up to 50% and 20% porosity, respectively, but these percentages are rarely achieved because exact mole-for-mole replacement is rare. Bischoff et al. (1994) described a process whereby dissolution of gypsum drives dedolomitization. Palmer (1995) also noted the importance of the common ion effect in controlling the relative solubility of calcite and dolomite. Dissolved gypsum or anhydrite diminishes the solubility of both limestone and dolomite, but the effect on limestone is greater. As a result, dolomite becomes far more soluble than calcite or aragonite, and selective dissolution of dolomite can occur. Calcite is precipitated by mole for mole exchange with the dolomite as calcium is released by the dissolution of gypsum. The process maintains low bicarbonate and pH, causing dolomite to continue to dissolve. It is reported to increase the solubility of dolomite up to five times (Plummer and Back, 1980) and the solubility of gypsum 1.56 times. This was also reported as a process in deep-seated flow paths, associated with elevated calcium, magnesium and sulphate concentrations, near Oak Ridge, Tennessee (Saunders and Toran, 1994). The process has also been described by Back et al. (1983) in South Dakota and Wyoming. Palmer (2000) suggested that the presence of salinity can drive this process further. The geology of the Wye catchment would be conducive to this type of reaction at depth. Schofield (1982) observed an apparent catalytic effect of gypsum on dedolomitization. In considering the potential for this type of karstification of the syn-depositional dolomite of the Woo Dale Limestones it is interesting to note that Goldstrand and Shevenell (1994) identified a depth limit to the process, which was 60 m in the Copper Ridge Dolomite (cherty dolostone) and Maynardville Limestone (supratidal and subtidal dolostone and limestone), Oak Ridge, Tennessee. Below this depth groundwater movement was dominated by fracture flow. Within the context of the White Peak such a process (section 6.4) may have been initiated by sedimentary dewatering.

The influence of cold phases of the Pleistocene on the development of conduits must also have been significant, particularly in the context of vadose development, both because of groundwater temperature and also the volume of water and associated fluid mechanics. As well as the increased rate of dissolution associated with the volume of water, meltwater would temporarily have enhanced mechanical erosion in influent caves. Where conduits are characterised by a low gradient the bed load accumulates until such time that the velocity is large enough to transport it. Hence, many streams sink through accumulations of boulders, which are remnant from periods of significantly higher discharge. Enlargement of the mouths of cavities can, in part, be attributed to periglacial weathering (Warwick, 1976). Even during current climatic conditions frost action is still likely to be active. However it should also be noted that a major source of sediment in conduits results from solifluction and there has been translocation of loess.

4.5 Geological guidance.

Structural guidance of conduit inception comes from the faults and mineral veins, jointing and tectonic displacements (Klimchouk and Ford, 2000b and Lowe and Gunn, 1997), which are not necessarily obvious. Warwick (1976, p. 83) observed that "*.. where the joint pattern is coarse, flow tends to be concentrated on a smaller number of channels and so enlargement into caves is commoner*" and Goldstrand and Shevenell (1994) pointed to a limiting depth of karstification reflecting the depth to which fissures were open, as a consequence of the depth to which stress relief is effective. Faults influence the hydrology by causing discontinuity and in some cases providing a physical opening. Some mineral veins may act as hydrologic barriers whilst others act as conduits for karst waters (Waltham, 1971). In the Northern Pennines, Warwick (1962) identified that thicker mineral veins were normally less porous than limestone and phreatic tubes commonly formed on the up-dip side of some of the veins in the mines. Fetter (2001) suggested that as greater volumes of groundwater movement, and therefore dissolution, take place at the intersection of joints, or joints with bedding planes consideration needs to be given to the guiding potential of the strike of intersecting joint sets, attention should be given to the importance of the investigation of fracture traces in understanding the hydrogeology of limestone terrains, as suggested by Latmann (1958).

The relative importance of bedding and structure in conduit development formed a focus for research in the 1970s (Ford, 1971; Ford and Ewers, 1978; Waltham, 1971). Waltham (1971) suggested that bedding is a stronger guiding factor in the evolution of vadose conduits, whereas jointing is a stronger guiding factor in phreatic conduit development. Ford and Ewers (1978) argued that bedding planes should be the more important guide as they are continuous to the boundaries of the limestone mass. However, their model (which comprises: bathyphreatic caves; deep phreatic caves; mixture of phreatic and water-table levelled components, and ideal water-table caves) also reflects fissure guidance, which increases with time.

Ford (1971) investigated the significance of bedding dip and joint frequency in the development of caves, categorizing them as: vadose, phreatic, water-table and true artesian, concluding that the types of cave that will develop depend on the frequency of groundwater fed fissures and the geometric relation of the fissures to bedding. He suggested that the higher the hydraulic conductivity (intensity of fissuring) the more likely the water-table condition is achieved and as a consequence upper tiers of caves have deeper phreatic loops than lower ones. Ford (1971) suggested that caves were most likely to develop in steeply dipping rocks with lower hydraulic conductivity, as a result of bedding plane guidance. He suggested that where the ratio of penetrable bedding plane length to joint is large, linear anastomotic caves develop in gently dipping beds and dip tubes with ascending/descending chimneys following joints in steeply dipping rocks. Worthington (1991), from a study of international cave statistics established the empirical relationship: $R = 1/(\tan\theta \sin\theta)$, where R is the ratio of bedding planes to joints used by conduits and θ is the angle subtended by the strike and the flow direction, with values of less than 1 occurring in rare karst with high dips, where flow is not strike adjusted. In a similar way, Palmer (1991 and 2000) has developed a conceptual model that relates cave morphology to geological formational influences; thus, bedding plane- guided passages are sinuous and curvilinear, dissolutionally enlarged joints and high angle faults form lensoid, fissure-like passages. In the context of the White Peak, Christopher et al. (1977) suggested that flow in phreatic conduits is generally along the strike of preferred bedding planes, particularly in areas of shallow dip, whereas flow in the vadose zones tends to follow dip and in the absence of other factors the plan-form of phreatic networks reflects the local fold structures. However, the former suggestion does not appear to have been confirmed by this study. For example, groundwater follows bedding in the Monsal Dale Limestone into the phreatic zone e.g. Bubble Springs (SK 20406612). Evidence from the Lathkill cave system indicates that the latter suggestion reflects the structural guidance of point recharge (Chapter 9). Worthington (1991 and 2001) established an empirical relationship between the dip of bedding, flow path length and the maximum depth of the flow path, which is utilized in Chapter 5.

Waltham (1971) and Ford and Ewers (1978) suggested that bedding planes can display features indicative of preferential selection for cave formation, e.g. partings of mudstone. Lowe (2000a) has taken this further and has demonstrated that the earliest (inception) phase of cave development may begin during diagenesis, reflecting the occurrence of localized, relative impurities within carbonate sequences. Such impurities may comprise atypical chemical, or physical properties, which may provide the pre-existing network of openings connecting recharge and discharge areas that is required for subsurface flow and the onset of speleogenesis (Lowe 1992, 2000b). Further to the inception horizons associated with the clay wayboards and the unconformities, this author has tried to identify additional potential speleogenetic processes within the strata that underlie the area of investigation, as summarized below. Lowe and Gunn (1997) suggest that inception and gestation is guided by the regional hydraulic gradient, with a number of inception horizons being active at a given time.

Geological guidance in the White Peak is mainly achieved through permeability contrasts and can be considered in terms of both lithology and structure-related guidance. Some of the contrasts were described in Chapter 2, in particular, the presence of the lavas, tuffs and clay wayboards, which are associated with the paleokarstic surfaces (Walkden, 1970, 1974 and 1977). Clay wayboards and weathered lavas can act as aquitards to both ascending and descending fluids. However, they transmit groundwater laterally and this has been seen to be significant where fissures do not penetrate a given wayboard. Lavas that are more competent are susceptible to fracturing and therefore have been found to be permeable. It might also be anticipated that dissolutional weathering associated with the paleokarstic surfaces would have increased the permeability of the exposed surface, thereby potentially forming inception horizons, particularly in the zone of unconformities, although such secondary permeability formed at this stage is likely to have been occluded by subsequent cementation. It has been speculated that laterally extensive stylolites may act as aquitards because of the decrease in permeability resulting from their formation (Wanless, 1979). However, Schofield (1982) observed that some stylolites formed groundwater flow paths. This may be a reflection of their association with dolomite (Appendix 4.1). Mineral veins in the area of investigation have been found to comprise significant storage areas (section 10.2) and are commonly, as a result of inheritance from the covered karst setting, the focus for valley formation. The formations of the area have been the subject of research, primarily in the context of investigations for oil, which has shed some light on their sedimentology, e.g. Schofield (1982) and Gutteridge (1983). Detail that is considered potentially relevant to the context of speleogenesis is described, together with the findings of petrological studies (Appendix 4.1). This thesis has identified a number of differing settings for inception including: stylolites in the Woo Dale Limestone Formation and towards the base of the Chee Tor Limestone Member; shell beds in Lathkill Dale, in the Monsal Dale Limestone Formation; and microstylolites-related dedolomite associated with clay wayboards in the Miller's Dale Limestone Member of the Bee Low Limestone Formation and mudstones in the Eyam Limestone Formation (Appendix 4.1). In the Chee Tor Limestone Member of the Bee Low Limestone Formation speleogenesis has created a network karst at surface, which is particularly evident in the sides of Lovers Leap gorge (SK 08507225).

4.6 Inputs in the research area.

4.6.1 Concentrated allogenic recharge.

Gunn (1992, p. 31) wrote that "*in the upper Wye basin immature and largely impenetrable stream-sinks on Stanley Moor flow through Poole's Cavern, a large but essentially relict effluent cave of phreatic origin, to both Wye Head and Otter Hole risings. At both risings the water emerges through impenetrable fissures but underground velocities are rapid and this is clearly conduit flow as opposed to a fissure flow system. However, lower down the Wye there are no allogenic inputs and recharge is predominantly diffuse autogenic with some concentration by closed depressions. There are a large number of springs none of which have any penetrable cave passage. This together with the general lack of variation in chemistry suggests that they are fed by fissure flow systems.*" Clearly the allogenic recharge largely comprises surface water run-off from the Namurian mudstones that largely enclose the limestones. There appears to be a close association between the occurrence of concentrated allogenic recharge and the position of faults and ramp reef deposits (Figure 5.3). Some of the allogenic recharge resurges after a few kilometers, but some forms underflow.

4.6.2 Dispersed allogenic recharge.

The context of the regional hydrogeology (Chapter 5) indicates that consideration should be given to the contribution of allogenic groundwater from the Goyt Syncline at a range of depths. Conceptually this might be perceived as dispersed allogenic recharge, or if via conduits, as further concentrated allogenic recharge.

4.6.3 Concentrated autogenic recharge.

Autogenic recharge, which is derived from the area of the limestone, can also occur as either concentrated recharge via sinks (dolines), or dispersed recharge. Dolines take a number of forms, normally within the categories: solution, collapse, buried and subsidence dolines (Williams, 2004). It has been observed that there is a close association between the location of dolines and the presence of dominant fissures, including faults and mineral veins. Dolines in the White Peak are particularly associated with specific settings, such as: valley sides; faults, or at the edge of till or head deposits. In the area of the Southern White Peak, Al-Sabti (1977) grouped doline settings into the following: 58.44% population on hills (includes valley side setting); 26.7% over plateaux; 14.86% valleys. He found that in the valley setting 42% of dolines were associated with valley sides but the majority were associated with hills (features rising above plateau level), which suggests to this author a probable association with run-off from volcanic horizons, for example Chelmorton (SK 115703), where perched springs, associated with the volcanic horizons, sink again once they reach the limestone (but see also the discussion below). Dolines are not necessarily easily identified in the White Peak. One reason for this is the similarity of the surface expression of dolines with that of shafts remnant from historic mining (e.g. in the area of Blackwell, SK 125717) and another is the practice of utilizing dolines for waste, as carried out by some farmers, or else for the locating of dewponds. These difficulties are also evident in the examination of aerial photographs. This author found a predominance of identifiable dissolutional features associated with the valley sides and also with dominant faults. For example, this author was able to follow the northwesterly extension of the faults crossing Deep Dale (west), simply by following the dolines. Furthermore, dolines are clearly visible in the Bee Low Limestone Formation on either side of the A515 in the vicinity of SK 076703. Indeed, it would seem that the dolines are more visible in the Bee Low Limestone Formation than they are in the Monsal Dale Limestone, or the Eyam Limestone Formations.

The form of the caves provides further evidence of concentrated autogenic recharge. Palmer (2000, p. 77) stated that "*Specific cave patterns, i.e. branchwork vs. maze patterns, are controlled by the nature of the groundwater recharge*". The implication of this is that the prevailing branchwork form to caves in the Wye catchment is indicative of a predominance of point recharge to cave systems. Furthermore, it should be noted that the majority of the explored cave systems within the Wye catchment are caves within the Monsal Dale Limestone Formation. This provides supporting evidence for the occurrence of dolines in the Monsal Dale Limestone Formation. Over much of the eastern side of the research area the Monsal Dale Limestone Formation is exposed at ground surface. Being younger than the Bee Low and Woo Dale Limestone formations, the Monsal Dale Limestone Formation potentially represents the area of most recent "unroofing" of the Namurian mudstones. Therefore, the guidance for the point recharge targeting pre-existing, structurally-guided phreatic tubes developed on inception horizons could be the gradually retreating edge of the Silesian cover rocks as suggested by Beck (1980).

However, it has been observed by this author that in some areas, most notably in Lathkilldale (Fern Dale and Lathkill Head) concentrated recharge is focused on mineral veins, which stand as bluffs behind the solution dolines. This leads this author to conclude that the mineral veins are acting in the same way as the faults and fissure sets that form the focus for doline development in the Bee Low Limestones, but the mineral veins respond in a different way. This may be because they are less soluble as suggested by Waltham (1971), or because they are so open as a result of the mineralization processes that further dissolution is not required to achieve the required capacity. The groundwater storage associated with mineral veins has also been noted by Ford (2000). At some locations, the groundwater storage associated with mineral veins may have provided additional loading on strata dipping in towards the valleys, thereby facilitating sliding failure (section 11.4).

During the course of fieldwork and the supervision of the excavation of a pipeline trench (Banks, 2002) it has been observed by this author that where the surface of the limestone is exposed beneath the cover it is generally found to be smooth and rounded, undulating slightly (a subdued form of poorly developed rundkarren), as might be expected of a surface that has been subject to chemical erosion beneath a soil cover. On some surfaces cockling has been observed, notably in Lathkill Dale and Arbor Low, apparently related to the growth of lichen, moss or other forms of plant growth. Microrillenkarren are rarely developed. One conclusion from this is that fissuring is required to convey water beneath the surface, but as field observations indicate that the dominant fissure spacing in the Monsal Dale Limestone and the Eyam Limestone Formations is greater than in the Bee Low Limestone Formation, particularly in the Chee Tor Limestone Member, it is likely that larger dolines form in these strata, with dissolution being focused both down fissures and along bedding between fissure sets. Examination of the 1: 50 000 Series geological sheet reveals a predominance of superficial head deposits in pockets to the east of the region, in fact largely in the areas associated with mineralization. This and the fact that Burek (1978) observed that the upper reaches of many dry valleys are commonly shallow depressions (e.g. Calling Low), together with the geochemistry of the springs that is described in Chapter 6; leads this author to conclude that dolines in the Monsal Dale Limestone and the Eyam Limestones formations are largely to be found within and beneath the pockets of head deposit, as seen in the vicinity of Calling Low (SK 187647). This is interesting because it suggests that they predate the Devensian. Unfortunately, not all of the pockets of head deposits are shown on the geological sheet. Examination of the geological map does indicate a close association between faults (or extensions of faults) and the occurrence of head deposits. In the Monsal Dale Limestone Formation the dolines appear to fall within regional subsidence basins, suspected by this author to incorporate inception horizon-related storage, which influences spring chemistry (Chapter 6).

In conclusion, the evidence suggests that the geology guides the form of the dolines. Smaller, more numerous solution dolines, possibly more recently developed, with a deeper surface expression are associated with the more heavily fissured Bee Low Limestone Formation and perhaps more specifically the Chee Tor Limestone Member, whereas larger, less common, buried dolines are associated with the Monsal Dale Limestone. However, localized occurrences of smaller, fault-related solution dolines that also occur in the Monsal Dale Limestone e.g, associated with the feather edge of the Conksbury Bridge Lava, below Over Haddon at SK 20656625 and also associated with mineral veins, should be noted.

4.6.4 Dispersed autogenic recharge.

The extent of dispersed autogenic recharge is difficult to ascertain. Gunn (personal communication, 2005) has suggested that the best evidence for dispersed recharge is the extensive area that has no concentrated recharge. However, as stated above, dolines are commonly difficult to locate. Furthermore, the catchment area of the associated subsidence basins has not been determined and where they intercept inception horizons and clay wayboards there is a potential for significant catchment areas. The occurrence of dispersed autogenic recharge via matrix and fracture porosity is likely to be associated with storage and greater retention times, contributing to the baseflow of spring discharge.

4.7 Storage and paleokarst.

Broadly, the major storage of unconfined groundwater corresponds with the synclinal setting, whereas that of confined groundwater corresponds with anticlines. The detail is considerably more complex. Williams (1983) described the problems of identifying the location of storage within the karst aquifer. Smart and Hobbs (1986) presented a method of defining storage in terms of end members ranging from saturated to unsaturated storage. Unsaturated storage comprises that which is partially air filled for much of the year and includes epikarst storage (the underlying part of the unsaturated zone being termed the transmission zone). In the saturated zone they defined two subdivisions: that of dynamic storage (storage above spring level in unconfined aquifers) and that of perennial storage, which occurs below spring level in unconfined aquifers and throughout confined aquifers.

Conceptually, dissolutional enlargement of dolines and fissures forms part of a self-sustaining system of vadose speleogenetic development, which is required to provide an outlet for water and sediment being taken through the 'system' (Klimchouk, 2000). Wider fissures encourage more through-flow of greater volumes of aggressive (under-saturated with respect to calcium carbonate) water. Soil residues can provide permeability barriers in the epikarst, e.g. blocking fissures, but may also form significant areas of storage. Williams (1983) suggested that the amount and depth of dissolution varies with: rainfall, time, lithology, soil composition and thickness, partial pressure of CO_2 and whether the system is open or closed.

It is generally considered that the epikarst serves a function in storage (Gunn, 1985b, Williams, 1983) and concentration of flow into the vadose zone via the fissure network, which takes a form comparable with that of increasing stream order. The epikarst forms as a result of stress relief, weathering and dissolution, but appears to be less well developed in the Peak District than in some other limestone areas of Britain and Ireland such as the Yorkshire Dales and Burren (Gunn, personal communication, 2004). Unlike the Peak District these areas were ice covered in the Devensian. However, work by

Friederich and Smart (1982) suggests that the epikarst is well developed in the Mendips, which were also ice-free. Accordingly, it is possible that the poor development of the epikarst may reflect the combination of lower temperatures during the Pleistocene and the capping of loess deposits that occur over the plateaus of the White Peak, or alternatively greater permeability in the unweathered limestone, possibly attributable to the presence of the mineral veins and the clay wayboards, or structural differences and lower densities of fissuring in specific formations of the White Peak.

Klimchouk (2000) suggested that, at the base of the epikarst, flow is focused along a few major fissures and preferential enlargement occurs from the base of the epikarst downwards, which ultimately results in the development of hidden shafts. In zones of high storage this process is pronounced due to the piston effect of high input from the epikarst, resulting in high discharge from the base. Klimchouk (2000) defined a percolation threshold at the base of the epikarstic zone that causes flow to concentrate while passing from the epikarstic zone to the vadose zone. As the number of discharge points increases retention decreases. However, in the absence of a well-developed epikarst these processes must be transferred to the vadose zone and in zones of low epikarst storage the hydraulic response of shafts is smoothed.

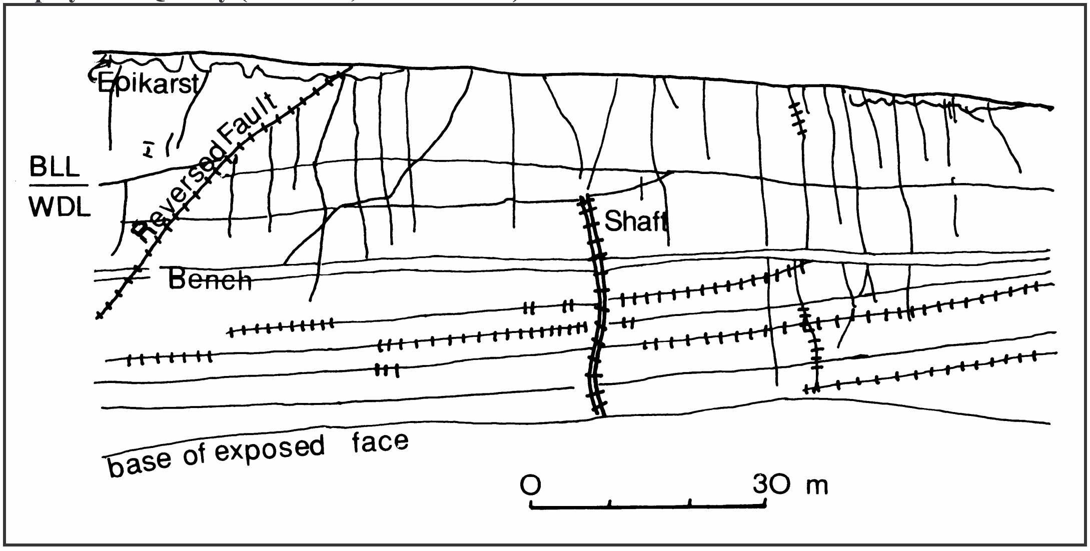
In order to consider these ideas further Topley Pike Quarry and Hindlow Quarry were visited with the aim of examining epikarst profiles (Figure 4.4). In Hindlow Quarry fissures in the Bee Low Limestone Formation were filled with superficial deposits and residual limestone and the fissuring continued throughout the sequence, virtually uninterrupted by bedding, with the exception of one dominant bed near the base of the exposed sequence. Dominant fissures occurred at approximately 30 m intervals and the epikarst was barely developed. At Topley Pike Quarry there was evidence of some superficial material filling fissures, but the volume was considerably less than that observed in Hindlow Quarry. Indeed, this is the reason that quarrying ceased at Hindlow Quarry and Buxton Lime concentrated on quarrying the Woo Dale Limestone at Tunstead in the 1970s. At Topley Pike Quarry the regular fissuring of the Bee Low Limestone Formation was evident, commonly continuing some distance into the Woo Dale Limestone Formation, but the fissures were not as open to the same depths observed in Hindlow Quarry (depths in excess of 20 m). At Topley Pike Quarry sediment was also evident in some bedding horizons, but some care is needed in the interpretation, because much of the bedding-related staining is a consequence of oxidation as a result of groundwater movement along bedding planes and also dedolomitization.

Observations presented as Figure 4.4, appear to confirm the significance of percolation thresholds both in concentrating flow to the vadose zone (Klimchouk, 2000) and in terms of epikarst development. Percolation thresholds have been found to operate on a number of scales, e.g. associated with major unconformities, as at Topley Pike, or associated with clay wayboards. Within the vadose zone this author has observed that where clay wayboards are visible, the limestone above and below clay wayboards shows differing fracture patterns and that only some, dominant, fissures actually pass through the wayboards. Klimchouk (2000) suggests that the initial contrast in permeability between the top layer of an exposed rock and the bulk mass at depth is due to the combined effect of stress relief and weathering, with the degree of fissuring being related to the rate of uplift. It would appear that this is likely to be the situation in the White Peak of Derbyshire. Additionally, formational differences have been noted by this author (Chapter 9), which suggests that material response should also be considered. As fissure frequency decreases with depth, flow must be guided towards a smaller number of more dominant fissures.

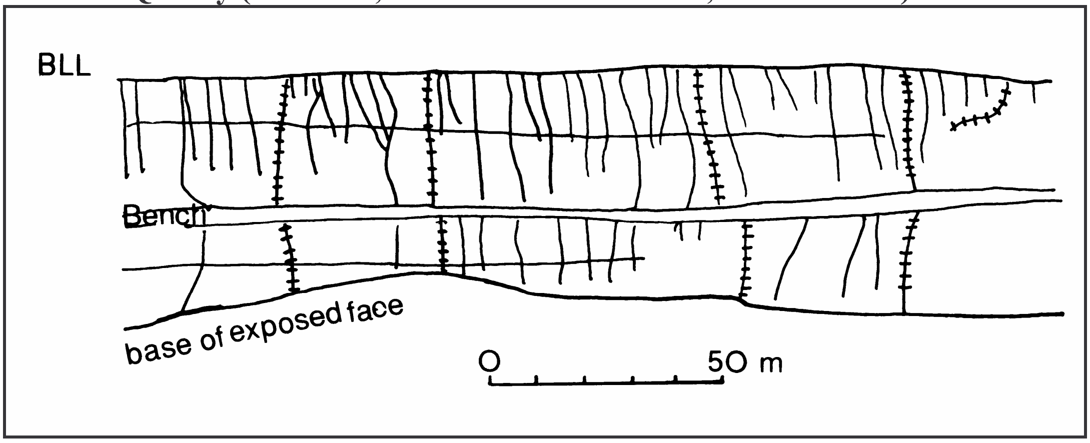
Following periods of wet weather it is clear that water stored along zones of clay wayboards (or mudstones) that are associated with bedding planes and stylolites and can be considered as inception horizons, or subsidiary fissures or conduits (Smart, 1999). The presence of stored water is also implicated by the preference of moss and tree root growth along these zones. Interestingly however, preferential dissolution along the wayboards appears to be very limited (millimeters, or centimeters at the most). Caves of branchwork form (bedding plane caves of Beck, 1980) occur at the boundary between the Bee Low Limestone Formation and the underlying Woo Dale Limestone Formation (examples being Thirst House Cave, Deep Dale Cave, Ashwood Dale Rising and the caves in Blackwell Dale). The caves comprise structurally-guided phreatic tubes. The work of Palmer (2000) suggests that the branchwork form is indicative of point recharge. However, bearing in mind the phreatic origin of these caves it is likely that early dissolution within the dominant fissures was by upward flow of water. Field observations made by this author indicate that point recharge occurs via dominant fissures (Figure 4.4), thus it is likely that dominant fissures have subsequently been targeted by per descensum flow paths.

Extensive development of subvertical fissuring in the Bee Low Limestone Formation is apparent along the A6 through Ashwood Dale (SK 067729). Significant fissure enlargement in the Bee Low Limestone Formation on either side of Lover's Leap (SK 07107268) is also apparent. Ineson and Dagger (1985) observed 'castellation' in the Bee Low Limestone Formation in the vicinity of Tunstead Quarry. This can be likened to a less mature form of clint and gryke development, which is associated with dispersed drainage. Interpolation of the extent of Namurian sandstone cap rock has convinced this author that the maze-like development is not attributable to the presence of a former porous cover. Instead it appears to be attributable to the more massive nature of the Bee Low Limestones and their susceptibility to fissuring. Palmer (2000) has observed that network mazes are limited to predominantly fractured aquifers and this would seem to describe the prevailing situation in the Bee Low Limestone, albeit that the karst is relatively immature.

Topley Pike Quarry (west face, SK 10177222).



Hindlow Quarry (west face, in Bee Low Limestones, SK 09306775).



Key: BLL Bee Low Limestone Formation; WDL Woo Dale Limestone Formation; Hatching indicative of iron-staining associated with dominant fissures and bedding planes.

Figure 4.4: Epikarst revealed in quarry profiles.

Motyka (1998) presented a conceptual model representing types of porous space in carbonate rocks. It comprised a modification to the broadly accepted triple porosity model (White, 1988), incorporating an additional porosity type, defined as "filled forms", which comprise re-cemented, fractured zones. With a little modification this model lends itself as a conceptual model of storage in the Wye catchment. In this area six porosity types have been identified: soil and head; faults and mineral veins; rock matrix; fissures; bedding plane/stylolites, and conduits. In considering matrix porosity some division needs to be made between carbonate shelf and reef limestones (Palmer, 1991 and Schnoebelen and Krothe, 1999). Examination of mining records indicates that mineralized veins form a significant focus for groundwater transfer and as such have the potential to act as groundwater divides.

Perhaps with the exception of the reef limestones, with which paleokarst has been associated; matrix porosity in this research area is extremely low, as indicated by petrological studies carried out by Hollis and Walkden (1996) and Schofield (1982). Although surface water is rarely observed, the storage capacity of the soil is exceeded in some valley locations following storm events. For example, this author has observed surface water in Lathkill Dale upstream of the usual source of the river during periods of heavy rainfall. This indicates that the storage of the superficial deposits (soil and a mixture of loess and limestone residue) can be exceeded in valley settings.

With respect to paleokarst, Ford (1995) differentiates between descending meteoric waters (per descensum) and ascending waters and gases (per ascensum). He shows that per ascensum waters exhibit a close relation with paleokarst, such that in many instances the presence of paleokarst is essential to the existence of modern per ascensum karst. The location of such features is independent of surface processes. By contrast per descensum karst water results from the hydrological systems of the cover and therefore it is only by chance that they are superimposed on paleokarst features. However, it has been shown (Chapter 9) that in the context of the White Peak paleokarst is dominated by per ascensum hydrothermal karst, which provides a focus for per descensum groundwater flow, because the inherited drainage pattern has been guided by the same dominant faults that provided the target for mineralization.

Smart and Hobbs (1986) suggested that storage is best measured in terms of the ratio of the storage volume to the annual recharge. Although this is ideal, it is not possible without detailed water balance calculations and more generally research has been directed at specific components of discharge. For instance, Atkinson (1977a) looked at the storage exhibited by recession curves for streams in the Mendip Hills Somerset, where he attributed 60 to 80 % of spring discharge to flow via conduits and assessed the storativity in the diffuse flow zone to be 0.92 %. This author has considered discharge components of the River Wye in Chapter 10.

4.8 Flow paths.

Whilst Palmer (1991) and Waltham (1971) have observed that phreatic passages originate along routes of greatest hydraulic efficiency and epigenic caves develop synchronously with the surrounding landscape, this is not necessarily true for hypogenic (deep-seated) caves (Palmer, 1991). In the Peak District evidence for hypogene flow paths is evident in the presence of thermal springs (Worthington and Ford 1995b, Chapter 5). The thermal springs have been shown to have a high sulphate content, increased thermal flux towards the regional base level, flow paths circulating to depths of at least 600 m and residence times of at least 15 to 20 years (Edmunds, 1971). The importance of deep-seated flow paths (hypogenic caves) in terms of understanding the regional hydrogeology is described in Chapter 5. To a large extent one can only surmise the nature of karst at depth. The evidence presented by Palmer (1991) suggests that thermal springs are probably associated with ramifying cave forms; however this has not been proven. Worthington (1991) has suggested that decreasing viscosity with depth facilitates increased flow rates along some preferred, deep, flow paths, thereby allowing them to enlarge more rapidly than those at shallow depth and that potentially this could be associated with the development of cave tiers. The interesting conundrum with deep flow paths is whether or not they develop from the surface point of recharge. It is the opinion of this author that the types of dissolutional processes described by Bischoff et al. (1994 and section 4.4), whereby gypsum dissolution drives dedolomitization, could have been activated by per ascensum basinal dewatering, probably driven by the depth of burial that was required for dewatering to occur. Dewatering would have been rendered more efficient by the associated decrease in viscosity of the groundwater.

Per ascensum flow paths are implicated in the mineralization of the White Peak. Ford (1989) in his description of the hydrothermal paleokarst of the Peak District described three host settings for hydrothermal karst associated with mineralization: a) fault fractures and joints b) favourable lithologies of limestone to yield replacement ore deposits (pipes) and some flats c) pre-existing cavities (possibly including flats). Ford (1989) suggested that large bodies of hot water formed the hydrothermal karst, possibly associated with an early pre-mineralization aggressive phase and probably reached the surface as springs that may have been responsible for pipe vein caverns. These flow paths probably developed soon after the regional flow paths developed. They would appear to be associated with the maximum depth of burial of the limestone and the dewatering of the adjacent Widmerpool Gulf and the Edale Gulf. They appear to post date the dolomitization in the Wye Valley (Schofield, 1982).

Evidence for seismic pumping (Colman et al., 1989) associated with mineralization comes from the slickensided form on the mineralized faults. It is likely that the flow paths rose to the surface, as suggested by Ford (1989), citing the Golconda Mine, near Brassington, as one such egress. This author opines that Thirst House Cave (SK 097713) could be another. The mineralization flow paths are discordant with the current hydrogeology and with that perceived for the Triassic and the Pleistocene. Accordingly, although the mineralized veins form significant zones of groundwater storage, there is no reason to suspect that they form significant flow paths.

4.9 Outputs.

The majority of interest with respect to outputs has been within the fields of spring hydrographs and spring hydrogeochemistry (Chapter 6). Outputs can also be considered in terms of per ascensum and per descensum karst. With respect to per ascensum karst, Palmer (2000) suggested that caves formed by upward movement of groundwater are focused on areas where the rising groundwater is able to escape. Ford (1989) also recognized this with respect to hydrothermal paleokarst. Typically, effluent per ascensum caves would be described by Palmer (1991) as ramiform caves e.g. Carlsbad Cavern (Egemeier, 1981). However, the Buxton thermal springs ascend from depth along a fracture (Barker et al., 2000). Palmer (1991, p. 19) observed, "*Inactive hypogenic caves are most difficult to distinguish in a humid climate. Invading surface water tends to overwhelm the deep-seated processes, or to modify the caves so that their pre-existing hypogenic features are masked*".

With respect to per descensum karst, Warwick (1962, p. 88) observed that "*Former effluent caves are found in the valley sides, often just below former valley floors, though these rarely penetrate for any great distance and were often used as dwelling places by man and beast during the Pleistocene period.*" Based on personal observations it is considered more likely that valley floor lowering intersected the caves. The Lathkill Dale Cave system seems typical of others in that it has formed on an inception horizon, below the valley floor, its location being attributable to the valley side location of the recharge. This also supports the potential for tiered flow path development. Springs along the Wye Valley appear to be closely related to faults and are suspected to be predominantly channel and fissure fed.

On the basis of hydrogeochemical data (Chapter 6) it is possible to differentiate between underflow and overflow springs. In Lathkill Dale (Chapter 11) bedding planes form the main lateral routes along which water moves from sinks to rising. Groundwater rises where structural guidance overcomes lithological guidance; such structural guidance is particularly associated with fault planes and dominant fissures. The structure-guided flow paths are dispersed, as seen in particular at Bubble Springs and in Ashwood Dale. The more dispersed guidance could reflect the degree of fissuring associated with the fault zone, slickensiding, mineralization, or the fault displacement attributable to tectonic movements that occurred during mineralization. For instance, a considerable amount of calcite veining has been observed in the limestone from which the Ashwood Dale Springs (SK 08957223) emerge. That suspected underflow and overflow springs can be identified indicates that a number of different flow paths target the same faults, or dominant fissures. This supports the concept of tiered flow, which is described further below.

4.10 Consideration of flow net application.

Thrailkill (1968) and Worthington (1991) consider that flow paths within limestone can be modelled using flow nets (the Darcy and the Hagen-Poiseuille flow nets respectively). In particular Palmer's (1991) observation that most caves are well adjusted to present patterns of recharge, the concept of stacked karstic groundwater basins (White, 1993 and Worthington, 1991) and the ideas of Klimchouk and Ford (2000a) that karst represents an hydrogeological cycle characterised by progressively expanding meteoric groundwater circulation support this concept. Furthermore, the observations of researchers investigating the sedimentology of the Carboniferous Limestone (Hollis and Walkden, 1996; Schofield, 1982, and Walkden and Williams, 1991), indicate that cement zones 1 and 2 (identified by cathodoluminescence) were precipitated in the meteoric phreatic environment, zone 3 cement was precipitated during shallow burial and is thought to have occluded most of the remnant palaeoporosity. During the early stage of burial basinal dewatering would have been active, through what must have been much more permeable limestone, albeit in a confined setting, probably with dominant flow paths developing on the edges of the tectonic system (Palmer, 2000). Consequently groundwater flow paths, initiated by basinal dewatering, lend themselves to modelling with flow nets (Chapter 5).

On the local scale, the concepts of: per ascensum groundwater from inception horizons; of unconfined water becoming confined, because of lithological guidance; and of the low hydraulic resistance of conduits make boundary conditions difficult to determine and introduce an anisotropy, which renders the superimposition of local basins on the regional flow net more impractical. These difficulties have been described in some detail by Worthington (1991).

Chapter 5: Regional hydrogeological setting.

5.1 Introduction.

Tóth (1963) has shown that groundwater can be considered in terms of local, intermediate and regional flow systems. In the context of this study, these may be likened to the drainage to the rivers Lathkill and Bradford; Wye; Derwent and Trent, respectively. It is the intermediate and regional flow systems that are the main focus of this section. This section also includes an introduction to the main areas of research in the local context that forms the subject of subsequent chapters. In terms of regional hydrogeology, the White Peak lies immediately to the east of the North Sea to Irish Sea water parting. Downing et al. (1987) divided England and Wales into four groundwater provinces, namely: Eastern Province, Hampshire Province, Severn Province and North-west Province. The area of the White Peak falls towards the western edge of the Eastern Province (Figure 5.1) and to the north of the Exe to Tees line, the boundary between Upland and Lowland Britain. This imposes a regional west to east flow trend that is reflected in the surface water drainage of the area. Wilson and Luheshi (1987) quote Gale (personal communication) as interpreting a potentiometric gradient of ~1 m/km from borehole stem pressure data for the Dinantian, which can be compared with the very steep, intermediate hydraulic gradient of ~1m/ 100 m across the White Peak, as shown by Downing et al. (1970). However, by virtue of its triple porosity and in particular the inception horizon-guided dissolutional component of the permeability, the permeability of the aquifer is anisotropic and at a local scale flow does not always travel perpendicular to the current potentiometric surfaces. Downing et al. (1987) pointed out that, regional and local rates of groundwater flow can vary by an order of magnitude, reflecting different aquifer properties.

Tóth (1963) demonstrated that where there is little or no topography, local flow paths will not develop and flow will be dominated by regional flow patterns. Worthington (1991) used the Hagen Poiseuille equation ($Q=\Pi\ pgr^4S/8n$, where Q is discharge, p fluid density, n fluid viscosity, r hydraulic radius, g gravitational acceleration and S the hydraulic gradient) to show that the effects of the thermal gradient on fluid viscosity and density render deeper, long, flow paths more efficient than shallow flow paths in catchments longer than approximately 3 km. However, Thrailkill (1968) suggested that in practice, flow path depth typically reaches in the order of 1% of the catchment length. Worthington (1991) suggests that this may be due to tightening of fissures, or hydrochemical conditions. Nevertheless, there is substantial evidence that deep-seated flow paths do occur within the limestones of the Derbyshire Dome. Consideration of the deeper flow paths forms a significant part of this chapter.

In this assessment of the regional hydrogeology, the limestone is set in its structural context with consideration being given to: the effect of relief on the hydraulic gradient; recharge in its geological context, and confirmation of the direction of the regional hydraulic gradient. More specifically, evidence for the regional hydraulic gradient has been drawn from: the rivers as flow paths; the distribution of springs; variations in groundwater chemistry; heat flow; the presence of thermal springs;

and the orientation of "bedding plane" caves. From this and an analysis of the regional base-level it has been possible to use the physical dimensions of the limestone to generate a conceptual model for the regional hydrogeology.

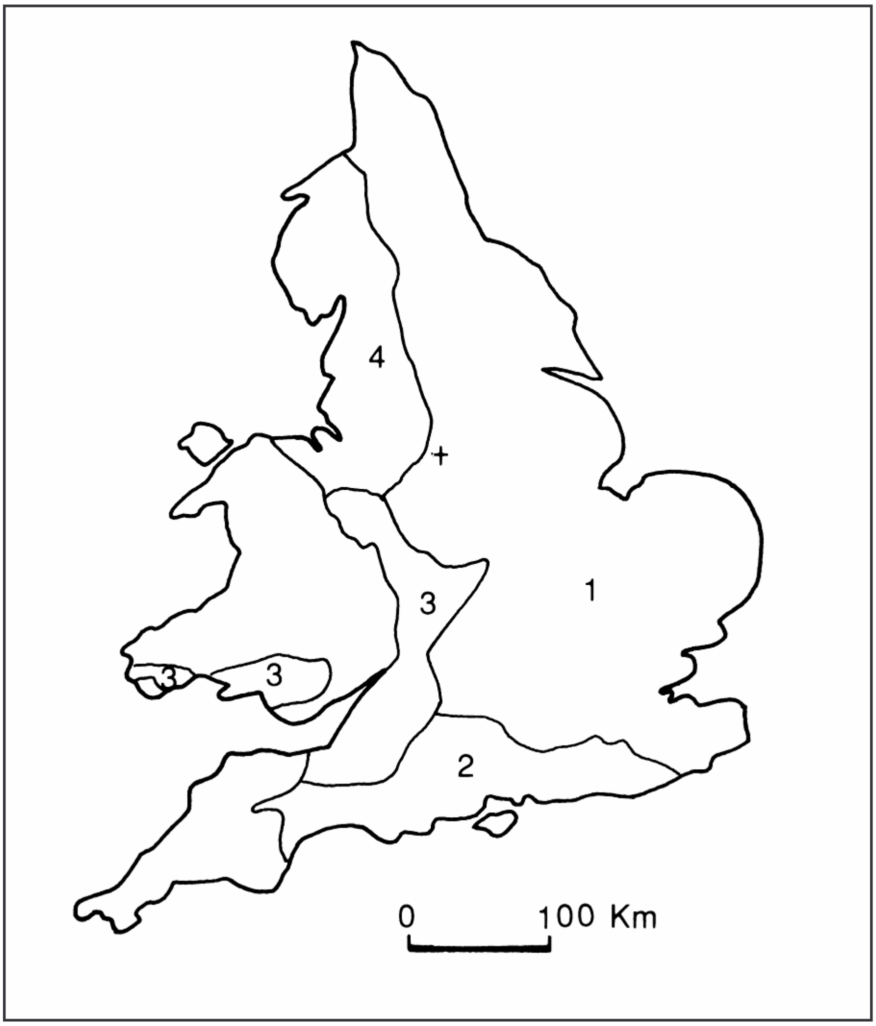
5.2 Relief.

White (1969) differentiated between regions of moderate relief, where hydraulic gradients are sufficiently low that cave systems are in equilibrium with local base-level, and alpine areas of extreme relief where floods and snow melt waters follow flow paths perched considerably above local base-level. The plateau surface of the Peak District is at an elevation of approximately 300 to 400 m OD, the Wye Head Spring, the source of the River Wye is at an elevation of 308 m OD and the confluence of the River Wye with the River Derwent, at Rowsley, is at an elevation of approximately 98 m OD. Although topography is locally steep, the region should be considered as one of moderate relief in terms of the classification of White (1969), with any perching of groundwater being attributable to geology rather than relief. Notwithstanding this, there have been localised exceptions following periods of extreme weather conditions, most notably when water has flowed along dry valleys and elevated paleokarst has been reactivated following periods of prolonged cold weather, as has been observed in the upper Lathkill valley.

In the situation where cave systems are in equilibrium with local base-level it is appropriate to make reference to the water-table. In an unconfined aquifer the water-table is the level to which groundwater rises. In a confined aquifer the water level may rise to a level above the top of the aquifer (artesian conditions), or may fall below it. Confined groundwater conditions have been encountered at some levels in this study of the limestone. For confined aquifers a map of the hydraulic head is referred to as the potentiometric surface and essentially this is what has been used to estimate the intermediate hydraulic gradient. However, care is required in interpretation for, as Freeze and Cherry (1979, p. 49) point out, "*The concept of a potentiometric surface is only rigorously valid for horizontal flow in horizontal aquifers. The condition of horizontal flow is met only in aquifers with hydraulic conductivities that are much higher than those in the associated confining beds*". Otherwise the key issue is that it is a 2-dimensional surface representing 3-dimensional flow regimes.

5.3 Structural setting.

The structural setting of the White Peak (section 2.3) comprises a faulted basement forming a structural high, against which a southeasterly dipping carbonate ramp formed. Basement faults comprise growth faults that are evident at surface as either faults or anticlines. This author opines that the distribution of thermal springs provides further evidence for the location of basement faulting. Others have considered the relationship of thermal springs with faults and Edmunds et al. (1971, p. 142) stated, "*in general, the regional geological structure is not a dominant factor in controlling the sites of mineral waters, but it may play a decisive role in the siting of thermal springs*". Worthington (1991) established an association of some of the thermal springs with fault zones and Barker et al. (2000) described the close relationship of the Buxton thermal spring with faulting. Examination of the geological setting local to each of the thermal springs within this study area (Table 5.1, below; temperatures presented in Table 5.5) does not always demonstrate a close relationship between thermal springs and fault zones. However, this probably reflects masking of fault zones by later sedimentation and it has been noted by this author that each of the thermal springs is associated with anticlinal features that are likely to be associated with growth faults. The distribution of thermal springs is shown on Figure 5.2.



Key to Groundwater Provinces:

1	Eastern Province
2	Hampshire Province
3	Severn Province
4	North-west Province

Non-designated areas are underlain by Pre-Carboniferous rocks

\+ Represents location of the White Peak of Derbyshire

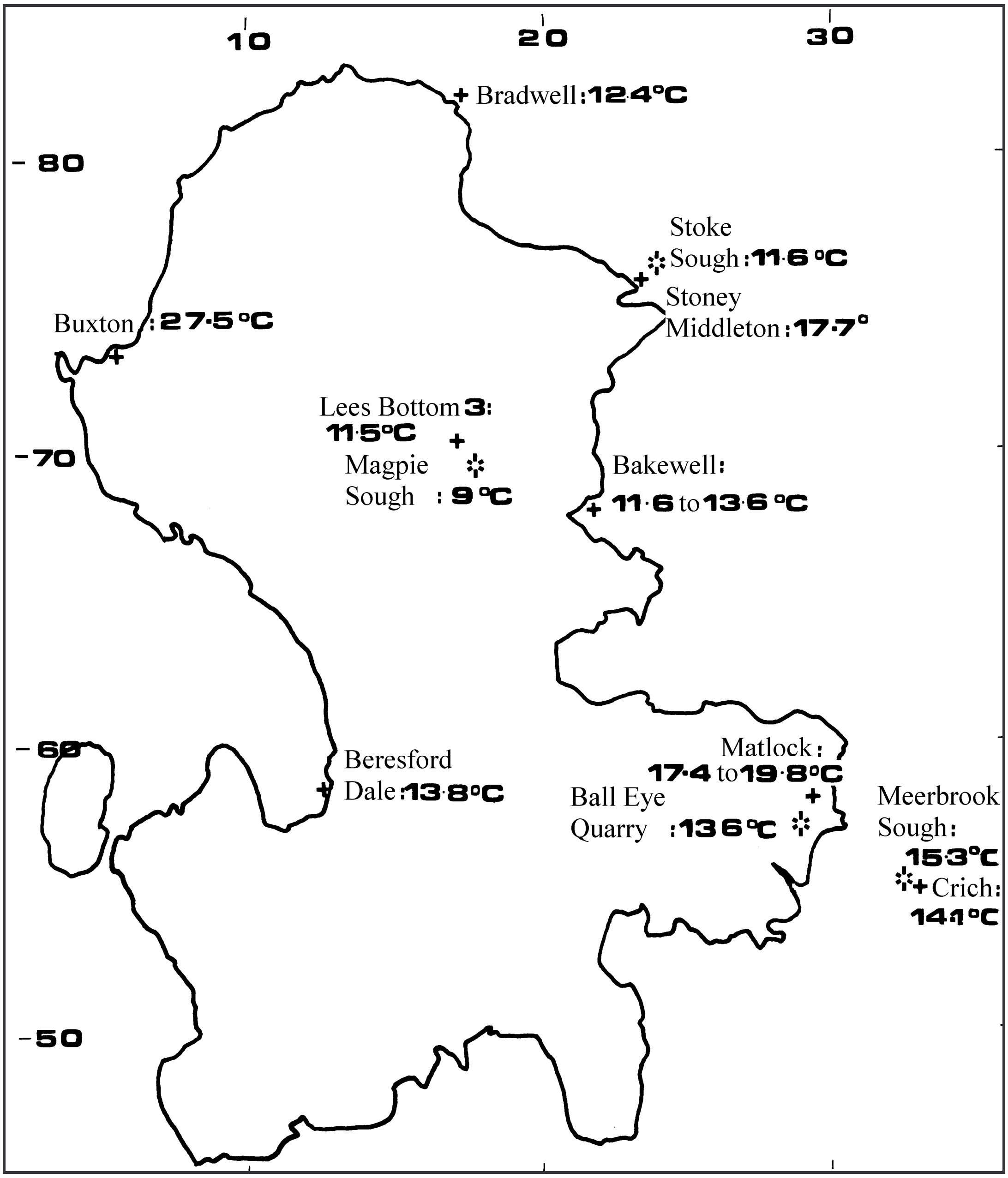
Figure 5.1: Regional hydrogeological setting (adapted from Downing et al., 1987).

Table 5.1: Local geological settings of each thermal spring/ sough.

Location	National Grid Reference (SK)	Level (m OD)	Geological Setting (interpreted from BGS 1: 50 000 geology maps)
Bradwell	17398114	175	Apron reef inlier in Namurian mudstones, possibly associated with the western end of the Hope Valley Anticline.
Stoke Sough	24007640	143	Namurian strata at surface, association with underflow from the Woo Dale Limestone Formation.
Stoney Middleton	2320 7560	145	Southern end of knoll reefs in Eyam Limestone Formation, beds dip to the east at 10 to 20˚.
St. Anne's Well, Buxton	05707350	298	Alluvium appears to cap Monsal Dale Limestone Formation, at the western end of the Bee Low Anticline; possible that western end of the associated fault extends to this point.
Lees Bottom 3 (Lower Dimindale)	17157059	147	Monsal Dale Limestone Formation, no fault shown but possible association with a northerly extension to the Arrock Fault, located to the north of the Taddington Anticline. Beds dip c. 5˚ to the south, probably fed by underflow from the Woo Dale Limestone Formation.
Magpie Sough	17896956	143	Sough draining the Taddington Anticline, boil up with elevated temperature, suspected to be underflow from the Woo Dale Limestone Formation.
Bakewell British Legion	21806860	118	Immediately to the east of the boundary between the Eyam Limestone Formation and the overlying Namurian strata. No fault shown, but association with the eastern side of the Bakewell Anticline.
Bakewell, Recreation Ground	22006810	120	As above, probably faulted at depth.
Beresford Dale	12805860	210	Reefs on the western side of the limestone outcrop, likely association with faulting associated with the Dove Dale Anticline.
Matlock East Bank Rising	294 05820	86	Alluvium over Monsal Dale Limestone Formation. Not immediately on a fault, association with mineralized zone at eastern end of Cronkston-Bonsall fault zone and the Matlock Anticline.
Matlock, New Bath Hospital	29305790	122	Tufa, otherwise as above.
Matlock, Fountain Bath	29405840	90	Mineralized fault, associated with the eastern end of the Cronkston-Bonsall Fault.
Matlock Ball Eye Quarry	28905730		Immediately to the south of the southeasterly extension of the Cronkston-Bonsall Fault.
Meerbrook Sough	32705520	80	Sough along Gang Vein, possibly a southeasterly extension of the Cronkston-Bonsall Fault zone, suspected underflow in the Woo Dale Limestone Formation.
Ridgeway Sough, Crich	332549 (Edmunds, 1971)		Inlier of Carboniferous Limestone in the Derwent Valley.

Glover (1831, p. 22), quoting Farey (1811, p. 505) included an additional potential thermal spring in the listing of springs, under Middleton by Wirksworth, "*W (Boota) on 3rd toadstone; and NNW (Wood) hot, formerly, and an open bath in Bonsal Dale*".

Thirteen groundwater systems largely focused on river or dry valley systems have been identified within the White Peak (Beck and Gill, 1991; Christopher et al., 1977; Downing et al., 1970; Edmunds, 1971), see also Chapter 7 for description in terms of recharge, transmission and discharge. These systems correspond to the local and intermediate systems of Tóth (1963).



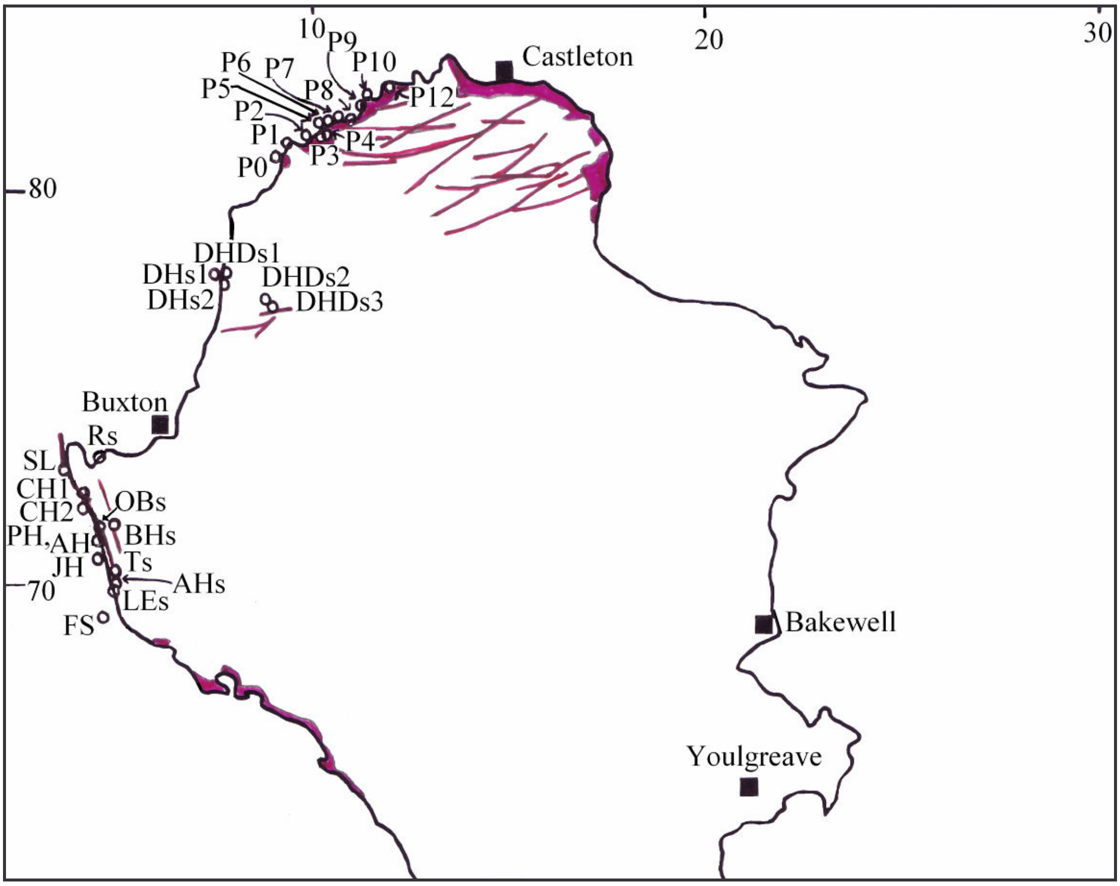
Key:

All temperatures in ˚C
+ Thermal Spring (temperature exceeds average bedrock temperature (Field, 1999))
* Warm water encountered

Figure 5.2: Location of the thermal springs (spring setting presented in Table 5.1).

5.4 Recharge in its geological setting.

Recharge falls in four categories: allogenic dispersed, allogenic concentrated, autogenic dispersed and autogenic concentrated (Chapter 4). The distribution of known stream sinks (Christopher et al., 1977, Figure 5.3) indicates that concentrated allogenic recharge appears to be particularly dominant around the periphery of the White Peak and most notably in the northwest, northeast and southwest. Concentrated autogenic recharge does occur and a database including details of known sinks was prepared by Gunn (1998). Dispersed autogenic recharge is thought to dominate over much of the remainder of the area, although consideration should also be given to the number of solution dolines that have been infilled by farmers (Gunn and Lowe, personal communication, 2004) and to the head filled hollows that this author considers likely to be buried dolines (section 4.6.3 and section 11.3). The close relationship between structure and the distribution of dolines suggests to this author that it is, at least in part, inherited from the former cover.



Legend: Apron reefs shaded; lines = faults or mineral veins
Key:
AH Axe Hole; AHs Anthony Hill Swallet; BHs Borehole swallet; CH1 Can Hole 1; CH2 Can Hole 2; DH Dove Holes; DHD Dove Holes Dale; FS Fairthorn Sink; JH Jake's swallet; LEs Leap Edge Swallet; OBs Old Bill's Swallet; PH Plunge Hole; Rs Resurgence swallet; Ts Tuncliff Swallet

Figure 5.3: Stream sinks of the Wye and Peakshole Sub basins (Adapted from Christopher et al. (1977).

Early Namurian deposits were dominated by mudstones, with subordinate interbedded, thin quartz sandstones and siltstones. In later Namurian and Westphalian times a delta complex deposited the massive coarse-grained sandstones of the Millstone Grit Group, the lowest of which is the Kinderscout Grit. The delta complex is understood to have migrated progradationally southwards and westwards into the region, in particular filling the Edale Gulf and the Widmerpool Gulf, but eventually also covering the Midlands Craton. This is reflected in the apparent rise to the north of the Kinderscout Grit

(British Geological Survey, 1978). Additionally, sedimentation on the eastern side of the Derbyshire Dome is more conformable than that to the west. The rise in the surface of the base of the Kinderscout Grit is such that it is likely to have been in direct contact with the limestone in the northern part of the Derbyshire Dome. The westerly progradation of the strata appears to have been associated with a thickening of the basal part of the sequence to the east; hence a greater thickness of shaley, clastic rocks occurs in the east. The Namurian sediments were laid down on a folded, faulted and heavily weathered limestone surface. The evidence suggests that locally the surface of the limestone would have risen above the early phases of Namurian deposition in the area, thereby giving rise to further, localised contact between the gritstones and the limestones. Where the limestone was formerly capped by the gritstones it is considered likely that diffuse recharge is better developed. Subsequent faulting resulted in the juxtaposition of sandstones and mudstones. As surface water courses cut down through the cover rocks the faulted contacts between the coarse-grained sandstones and the mudstones became the focus for surface water courses, albeit that they have become slightly displaced as they cut down below the contact. Evidence (Figure 9.6) for this comes from an examination of the surface watercourses immediately beyond the surface exposure of the limestone, a particular example being the River Manifold and its tributaries. Furthermore, Warwick (1964) showed that the network of dry valleys on the limestone exhibits a similar pattern to the drainage of the surrounding Silesian strata. Accordingly, it is reasonable to assume that there was similar drainage guidance on the former cover of the White Peak.

5.5 Defining the regional hydraulic gradient.

5.5.1 An outline of the indicators and the problems of interpretation.

Evidence for the regional flow direction comes from a number of sources including: the distribution of surface water courses; the results of hydrogeochemical investigations; the distribution of springs and stream sinks; flow vectors indicated by cave passages; flow vectors indicated by the results of dye tracing tests; and studies of heat flow. Each of these factors is discussed below, with the exception of the results of dye tracing tests, which form the subject of Chapter 7.

Mineralization is closely associated with basinal dewatering (Hollis and Walkden, 1996; Chapter 3), thus it is logical that an initial hypothesis of this thesis was that as mineralizing fluids were likely to represent one of the earliest phases of fluid flow in the limestone, the distribution of minerals offers the potential to assist in the identification of flow paths and the hydraulic gradients in the limestone. The examination of potential speleogenetic processes has confirmed this to be the case; in particular the lavas and clay wayboards have guided mineralizing fluids. The occurrence of mineral veins indicates that some faults are likely to form important flow paths. However, the concept of seismic pumping (Chapter 3) suggests that it is unlikely that the distribution of the minerals gives an indication of likely hydraulic gradients.

5.5.2 Rivers as flow paths.

As rivers form the base-level for the local and intermediate systems they provide evidence of the hydraulic gradient of these systems. The evolutionary history of the river systems gives an indication of the changes in the hydraulic gradient of the area, which are likely to reflect the regional hydraulic gradient. This history (section 2.10) indicates a trend of west to east drainage (River Wye) evolving from a more southeasterly direction, perhaps reflecting early concordance with the dominant structure, in particular with the basement structure (subsection 5.5.6). For example, the current course of the River Lathkill reflects the southeasterly hydraulic gradient that has been implied above for this part of the Dome. However, the early Pleistocene course of the upper part of the River Lathkill (Johnson, 1957) is suspected to have been to the east. Fault-guidance is again likely, because the river follows Lathkill Dale mineral vein. Interestingly the river is not displaced from the mineral vein in the same way that rivers are on the adjacent Namurian surface. This local discordance with the regional hydraulic gradient and centralisation of the river over the Lathkill mineral vein can be attributed to the trapping of the river by mud mounds in the limestone, as suggested by Ford and Burek (1976). Thus, the upper Lathkill does not appear to reflect the regional hydraulic gradient, but the overall course of the river does. These concepts are developed further in chapters 9 and 11.

5.5.3 Distribution of springs.

The distribution of springs (Figure 10.4) presents a more complex picture. The first thing to note is that upstream of Wormhill Springs (SK 12467342) the springs are fewer in number, but are of a larger discharge. Physically the springs appear to follow the regional hydraulic gradient of the local and intermediate systems, with outputs occurring at gradually lower levels to the east and southeast. However, it is interesting to note that some of the springs emerge from under beds of lava, suggesting that they are confined to varying degrees and indicating that locally the hydrogeology is complicated by the potential for confined as well as unconfined flow.

There are difficulties associated with defining individual spring catchments for a number of reasons, including the following:

-the existence of paleokarst systems;

-preferential speleogenetic development of specific horizons, which can take groundwater across surface water catchment divides;

-speleogenetic development along specific structures, or lithological features, which may also take groundwater across surface water catchment divides;

-inception at a number of levels, with vadose flow targeting different levels at different stages of groundwater level. Thus groundwater from the same sink may take water in differing directions depending on the stage, and

-divergent flow occurs at low stage, as hydrogeological groundwater divides subside.

Evidence that different flow paths are activated at differing stages also comes from the springs that operate as 'underflow' and 'overflow' springs, with the overflow component of the resurgence only being activated during periods of 'high' discharge, e.g. Ashford Dale Spring.

The hydrogeological significance of the lava beds has been noted by authors including Downing et al., (1970) and Edmunds (1971). The lavas generally act as aquitards, particularly where they are weathered (the weathering commonly occurring as a consequence of mineralization). Consequently they can either perch or confine groundwater. Perched (overflow) springs include Chelmorton (SK 103694), Illy Willy Water (also at Chelmorton, SK 113703) and Five Wells at Taddington (SK 126711). In some areas, for example where the rock is free of the influence of mineralization, it would seem that rock fissure flow predominates. The speleogenetic influence of the clay wayboards was noted in Chapter 4.

5.5.4 Regional variation in groundwater chemistry.

The results of a hydrogeochemical study, carried out by Downing (1967), present further evidence of the regional hydraulic gradient. This study established regional variations in groundwater chemistry in an easterly direction away from the Derbyshire Dome. Calcium bicarbonate water was shown to give way to calcium sulphate and then chloride groundwaters. The brines are considered to be the result of diagenesis and the expelling of formation waters, in which flow is driven by the hydraulic head in the Derbyshire Dome, driving water from the Carboniferous Limestone upwards into the Silesian strata (Downing, 1967). The boundary between the calcium sulphate waters and the chloride waters lies marginally to the west of Eakring, a point of rising groundwater (subsection 5.5.5), indicating the source of the sulphate is associated with the Dinantian Limestone. On the local scale groundwater chemistry can be used to assist in the definition of spring catchments. Edmunds (1971) attempted to relate groundwater spring chemistry to geology and Christopher (1981) used spring chemistry to: differentiate between allogenic and autogenic springs; differentiate between open and closed systems; identify thermal springs, and also to classify geological influence on groundwater chemistry (Chapter 6).

Gunn et al. (2006) used isotopic data on groundwater sulphate, inorganic carbon and strontium to establish sulphate sources. Sulphate in 'Matlock' type waters was interpreted as having been derived from buried evaporites. The 'Matlock' chemistry was shown to be significantly different to that of the thermal water at Buxton, which is considered to be derived from Namurian sandstones (Gunn et al., 2006, section 5.5.6).

5.5.5 Heat flow.

Further evidence for the regional flow pattern postulated in this thesis is provided by Downing et al. (1987), who found that pressure data from drill-stem tests in the Millstone Grit Group, identified a 'pressure ridge' associated with an area of less saline water at the easterly limit of the Millstone Grit Group. Downing et al. (1987) also noted that the rise of the groundwater from the Carboniferous

Limestone coincides with an area of a regional heat flow anomaly in the area of a faulted, north to south-trending, anticline at Eakring. In the UK the overall heat flow pattern comprises regional anomalies superimposed on a fairly uniform background of about 52 mW/m^2, although values of 60 to 65 mW/m^2 are common to the east of England. At Eakring heat flows of 114 to 120 mW/m^2 have been reported by Gale and Holliday (1984). It is suggested (Downing et al., 1987) that the heat anomaly can be attributed to the rise of thermal water, resulting from geothermal heating of deep groundwater, which the groundwater chemistry suggests to be derived from the west.

Support for the hypothesis of the origin of the thermal anomaly at Eakring resulting from groundwater movement from the limestone also comes from the results of finite element, numerical modelling of heat and fluid flow carried out by Wilson and Luheshi (1987), albeit that such modelling is inevitably limited by the paucity of information with respect to the bulk permeabilities of the strata within the model. Furthermore, although the model was run for a range of permeabilities within the Dinantian Limestones, it does not appear to have been run with a permeability contrast at the base of the limestone i.e. to represent karstification. It is significant that the model shows that upward flow path generation is sensitive to slight variations in the permeability contrasts, which is relevant to the conceptual model postulated in Chapter 9.

Further evidence that the heat influx at Eakring could be a consequence of heat transfer from elsewhere comes from an investigation of the geothermal potential of the UK carried out during the late 1970s to early 1980s as part of a Europe-wide research and demonstration project in geothermal energy. All of the available data have been summarised, together with the findings of additional investigations in a variety of forms and reference to this can be found in the literature (Burley et al., 1984; Downing and Gray, 1984; 1986; Downing et al., 1987; Gale and Holliday, 1984; Richardson and Oxburgh, 1978, Rollin, 1984). Rollin (1984) reported a geothermal gradient of 26.4° C/km for onshore UK. However, the situation in the White Peak appears to be below average. Indeed, it would appear that there is a heat transfer in an easterly direction from the White Peak towards Eakring.

In considering the geothermal gradient, it is necessary to assess the heat source and the conductivity of the overlying materials. Richardson and Oxburgh (1978) suggest that the heat production (averaging 2.22 μW/m^3) is less fractionated in the upper part of the crust underlying England and Wales than in axial zones of orogenic belts characterised by high-grade metamorphism and magmatism. Average heat flow rates are 58.58 mW/m^2. In the area of the White Peak heat flow values of 17 mW/m^2 have been determined (Richardson and Oxburgh, 1978), with 35 mW/m^2 in the Cheshire Basin, indicating that average geothermal gradients cannot be used in the assessment of flow depths in the limestone. Data specific to the White Peak appear to be restricted to measurements from the Eyam Borehole (Richardson and Oxburgh, 1978), which indicated a very low heat flow, with a negative gradient below about 470 m, taken to indicate downward water flow at these depths. At 622 m depth the equilibrium temperature measurement in the borehole was 11.3° C, representing a temperature gradient of 5.1° C/km (including the negative below 470 m). Mathematical modelling carried out by Rollin

(1984), using the more reliable of the heat flow data suggests that the heat flow of the White Peak is actually in the order of 50 mW/m^2.

It would seem that deep-seated flow paths transfer heat from the basement rocks beneath the limestone to the regions outside the outcrop of the limestone. The findings derived from the Eyam Borehole are very significant because they also suggest the potential for a significant volume of meteoric water targeting underflow routes on the eastern side of the White Peak.

5.5.6 Thermal springs.

Evidence for the occurrence of deep flow paths can be interpreted from the occurrence of thermal springs within the region (Barker et al., 2000; Christopher, 1981; Downing et al., 1970; Edmunds, 1971; Stephens, 1929). Thermal springs, characteristic of many limestone basins, are attributable to the presence of conduits, which enable water to be conveyed from considerable depths to the surface with minimal mixing with surface water and minimal conductive loss, thereby ensuring that the elevated temperature is maintained. The locations of such springs in the Peak District and their associated temperatures are presented on Figure 5.2. Thermal waters have also been encountered in boreholes to the east of the area shown on Figure 5.2. For example in No. 1 oil well, Eckington, water from the limestone at 2898 feet (883 m) had a temperature of 48.8˚ C (Downing et al., 1970). The results of the geothermal investigations point to the heat source for the thermal springs being the basement rocks.

A radiogenic source for strontium in the Buxton thermals is suggested by Gunn et al. (2006) on the basis of a higher $^{87}Sr/^{86}Sr$ ratio in the Buxton thermals. If, as Barker et al. (2000) suggested, the heat source for the thermal springs is fluid contact with the basement then the temperature of the thermal springs could be used to contribute further to the understanding of the form of the basement. Interestingly, the distribution of thermal springs (Figure 5.2) supports the structural interpretation presented in section 2.3. It is suspected that this reflects the focusing of recharge on dominant faults and the way in which faults act as aquitards to horizontal flow paths, thereby guiding resurgence. It would appear that it is the form of the fault blocks that guides the regional hydraulic gradient and this could account, in part, for the significant volume of water associated with the Stanton Syncline (Oakman, 1979).

If the top of the basement is uneven, as postulated above, there exists the potential for evaporites to have formed in the resultant inter-block basins in a similar manner to that observed on the Franco-Belgian border (Beckelynck et al., 1984). Indeed, anhydrite has been encountered in the Eyam Borehole (Dunham, 1973) and Hathern No.1 Hydrocarbon Borehole (SK 51582416, Carney et al., 2001). Gypsum karstification requires constant recharge to ensure groundwater mixing and prevent density stratification, which ultimately would preclude further dissolution (Klimchouk, 2002). Gunn et al. (2006) have inferred the occurrence of gypsum dissolution from the groundwater chemistry of the Matlock risings. In terms of the regional flow paths developed at the base of the limestone it seems most plausible that dissolution results from evolving processes, during mesogenesis groundwater movement is directed from areas of greatest subsidence to margins of the basins (Klimchouk and Ford, 2000a) where rocks are at shallower depth (expulsion water). During telogenesis hydrostatic head drives meteoric groundwater circulation. This suggests that early speleogenetic processes may have been initiated both by dewatering of the limestone and by basinal groundwater being driven out of the Goyt Syncline driving speleogenetic processes described in Chapter 4.

In a simplistic way it could be assumed that the depth from which thermal springs emanate can be calculated from the geothermal gradient. This assumes that the cooling of the resurgent water is limited, there is no mixing with meteoric water and necessitates the adoption of a geothermal gradient. A global value of 25˚ C/km is suggested by Freeze and Cherry (1979). Richardson and Oxburgh (1978), state that heat flow is obtained from the product of thermal conductivity and vertical temperature gradient. They present a mean thermal conductivity for the Carboniferous Limestone that is 3.473 W/mK. A heat flow of 50 mW/m^2 (as indicated on contour maps presented by Rollin, 1984) indicates a temperature gradient of 14.4˚ C/km. This value is close to the 15˚ C/km suggested by Barker et al. (2000). Assuming a background, surface, water temperature of 8.0˚ C for the White Peak [rounded up from the 7.7˚ C determined by Edmunds (1971)], this would indicate a flow depth of 1300 m for St. Anne's Well, Buxton, which is 27.5˚ C. This depth exceeds the depth of the base of the Dinantian Limestones, as determined in the Woo Dale Borehole (SK 40993726) and therefore either indicates that the basement is deeper at this location, or suggests a source other than the White Peak for the thermal water. Based on the interpretation of the basement geology presented in Chapter 2 (Figure 2.3), this author prefers the latter interpretation.

At first sight the presence of the high temperature rising at Buxton, and also at Beresford, on the western edge of the outcrop of the limestone appears anomalous on the basis of this description of the regional hydraulic gradient, as well as on the basis of the depth calculated from the temperature, as described above. However, the rising could be accounted for in at least three ways: it could reflect a local difference in the depth to the basement rocks; it could be local recharge water sinking at the boundary between the Upper and Lower Carboniferous deposits; or the deep-seated flow paths may incorporate warm water derived from the west. Further evidence for a westerly source for the groundwater comes from comparison of the depth calculated for the source of the spring water at Buxton with the strata encountered in the Woo Dale Borehole, which indicates that basement rocks were encountered at 273.60 m (Cope, 1973). Examination of the 1: 50 000 Series British Geological Survey Sheets 99 '*Chapel en le Frith*' and 111 '*Buxton*' suggests to this author that a likely source is that of southerly moving water in granular beds in the Silesian strata, more specifically the Kinderscout Grit which is faulted and nearly vertically bedded, on the western side of the Goyt Syncline and returns equally steeply on the eastern side of the syncline. It has been argued in Chapter 4 that recharge is associated with faults and therefore it is equally plausible that the recharge occurs via a number of the sandstone bodies in contact with the significant fault zone on the western side of the Goyt Syncline.

In an assessment of the validity of Hagen-Poiseuille flow net conditions, Worthington (1991) established that for a range of karst settings there is an empirical relationship between flow depth, the length of the flow path and the dip of the strata, which he defined as:

$$D_m = 0.11(L_x \ \sin\theta)^{0.81}$$

A modified form of the equation was presented by Worthington (2001):

$$D_m = 0.18(L_x \ \sin\theta)^{0.79}$$

Where: D_m=mean depth of flow; L_X = Flow path length and θ = angle of dip of bedding.

By changing the subject of the latter equation, estimates of the flow lengths associated with the depths that have been calculated from the temperature data, have been made for the thermal springs on the eastern side of the White Peak (Table 5.2). In these calculations, which give the theoretical minimum flow lengths, it has been assumed that the bedding is vertical, reflecting the apparent structural controls on the occurrence of the thermal springs.

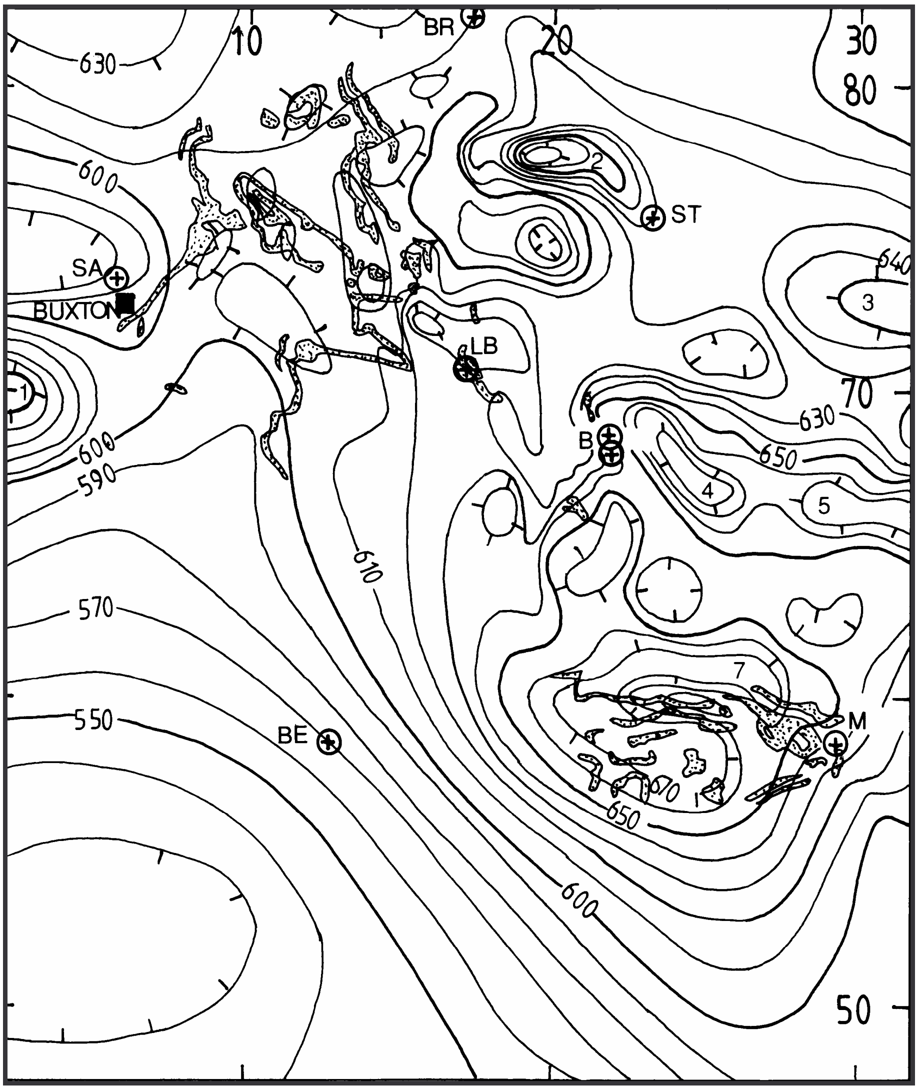
Table 5.2: Flow path lengths for thermal springs, calculated after Worthington (2001).

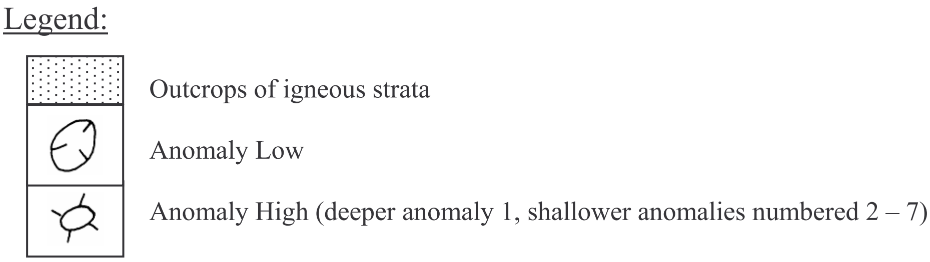
Spring Location	Temperature ˚C	Depth of flow path below water-table (m), calculated from geothermal gradient 15˚ C/km	Calculated flow path length (km)
Buxton	27.5	1300	76.6
Bradwell	12.4	325	13.2
Stoney Middleton	17.7	680	33.7
Lees Bottom 3	11.5	265	10.2
Bakewell	13.3	385	16.4
Matlock	19.8	820	42.8
Crich	14.1	440	19.4

The calculated flow path lengths suggest a potential for a significant contribution from outside the limestone catchment. Whilst it is the preferred interpretation of this author that the groundwater rising on the western side of the limestone, including the Buxton, St Anne's, thermal spring is derived from faulting associated with the Todd Anticline, it seems unlikely that this is the source of the springs on the eastern side of the catchment, where the temperature is lower.

Barker et al. (2000) pointed to a geothermal heat source, with 'pyroclastics' shown on their conceptual drawing. Furthermore, Gunn et al. (2006) suggest a source of radiogenic strontium associated with the Buxton thermal. The location of the springs has been plotted on the aeromagnetic map of the Buxton district (Aitkenhead et al., 1985). This map (Figure 5.4) shows a strong positive anomaly to the south west of Buxton and because this anomaly shows little change in signature as a ground determined anomaly compared with the aeromagnetic anomaly it is considered (Aitkenhead et al., 1985) that this is a magnetic anomaly at considerable depth (1.5 to 2.0 km below Ordnance Datum) and it is thought to lie in the pre-Carboniferous Basin. It is probably this source that is implicated by Barker et al. (2000).

The presence of this material provides a potential mechanism for the Carboniferous hinge effect along this boundary of the limestone noted by Biggins (1969).





Key:
Aeromagnetic contours at 10 nT intervals
BR Bradwell Thermal Spring; SA St Anne's Thermal Spring; ST Stoney Middleton Thermal/Spring; LB Lees Bottom 3 Thermal Spring; B Bakewell Thermal Springs (Recreation Ground and British Legion); BE Beresford Thermal Springs; M Matlock Thermal Springs (New Bath Hotel and East Bank Rising)

Figure 5.4: Location of thermal springs superimposed on the map of aeromagnetic anomalies (after Aitkenhead et al., 1985).

Shallower, positive, anomalies are shown (numbered 2 to 7), with depths thought to be in the order of several hundred metres. These anomalies have been related to thick accumulations of igneous rocks, probably the sources of the lavas that are interbedded with the limestones. The series of unmarked positive anomalies extending from the north through to the southeast are thought to correspond with deeper magnetic bodies present in the Precambrian basement. The absence of any associated Bouger anomaly suggests that the anomalies do not have any significant density contrast with the other materials in the basement. This evidence suggests variability in the availability of basement geothermal energy.

Groundwater/ bedrock reactions can be used as an indicator of flow paths, where the geology is known in detail. Edmunds (1971) examined the chemistry of the thermal springs and suggested that groups of waters from the same thermal centre have similar compositions, but that there is considerable variation between the centres. Thus, he differentiated between the thermal waters of Buxton, Bakewell, Matlock and Bradwell. Gunn et al. (2006) demonstrated that the allogenic, Buxton thermal had a higher Mg/Ca ratio and lower sulphate content than the other White Peak thermal waters, which they interpreted to be broadly similar on the basis of geochemistry and isotope data (Chapter 6). Accordingly, it was considered by this author that if the postulated fault-block distribution (Figure 2.3) does exist, there is a possibility that this would also be reflected in the groundwater chemistry. A search of the literature has revealed the following analyses:

Table 5.3: Chemistries of thermal spring and thermal sough waters (all units, except pH and Conductivity (µS/cm) in mg/l).

Thermal Rising	Data Source	pH	Condu-ctivity	Total Hardn-ess	HCO_3	Ca	Mg	Na	K	Sr	SO_4	Cl	NO_3
Bradwell	Christopher (1981), after Edmunds (1971)	7.2		580	178	162	42	240	5.2	5.4	326	420	0.2
Stoke Sough	Gunn et al. (2006)	6.54	1052		193	112	25.1	45	3.4	1.41	188.3	77	2.5
Stoney Middleton	Gunn et al. (2006)	7.56	909		242	92	28.2	60	1.4	0.71	93.5	90.2	1.6
Buxton	Gunn et al. (2006)	7.06	509		241	59	20.2	24.2	1.1	0.71	12.2	38.8	<0.05
Lees Bottom 3 (Lower Dimindale)	Gunn et al. (2006)	7.89	631		239	104	6.1	13.1	1.2	0.2	37.2	31.2	17.7
Magpie Sough	Gunn et al. (2006)	7.8	568		220	91	9.1	7.7	0.8	0.45	32.6	47.2	11.7
Bakewell British Legion	Christopher (1981), after Edmunds (1971)	7.37		473	178	161	17	19	1.5	6.67	262	23	3.1
Bakewell Recreation Ground	Christopher (1981), after Edmunds (1971)	7.44		563	163	186	24	19	0.9	12.3	384	23	0.1
Beresford Dale	Christopher (1981), after Edmunds (1971)	7.22		281	244	99	8	8	1.6	0.1	37	12	2.1
Matlock East Bank Rising	Gunn et al. (2006)	7.11	819		246	95	26.6	22.9	1.1	1.71	133.9	47.8	3.1
Matlock New Bath Hotel	Gunn et al. (2006)	6.48	850		238	86	24.6	21.3	0.6	0.71	134.7	49.3	2.8
Matlock Fountain Bath	Christopher (1981), after Edmunds (1971)	7.77		511	229	103	38	29	0.8		192	52	3.1
Matlock Petrifying Well	Christopher (1981)	7.53		399	224	116	26.5	27.5	0.9		140	61	1.8
Ball Eye	Gunn et al. (2006)	7.55		351	210	97	26	19	2		101	39	1.8

Thermal Rising	Data Source	pH	Conductivity	Total Hardness	HCO_3	Ca	Mg	Na	K	Sr	SO_4	Cl	NO_3
Quarry													
Meerbrook Sough	Gunn et al. (2006)	7.98	608		212	86	22.1	12.8	1.1	0.5	59.8	24.7	10.2
Crich, Ridgeway Sough	Christopher (1981), after Edmunds (1971)	7.43		446	192	155	14	25	1.8	1.7	230	47	0.37

Edmunds (1971, p. 14) suggested that the high SO_4^{2-} in the Bakewell waters may be "*closely related to mineralized veins, which possibly provided easy access to the surface for deeply circulating meteoric waters*". An alternative explanation is that the higher sulphates reflect the depositional environment of the early Dinantian intra-block basin areas described above, for instance Gunn et al. (2006) suggest an anhydrite source to account for the very high sulphate encountered in the Matlock thermal springs. Anhydrite dissolution during burial in the Carboniferous is also likely to have been the source of the sulphur incorporated in the mineral deposits and made available by thermal reduction (Machel, 2001) during burial. Thus, the basement structure in influencing the distribution of the inter-block basins has guided the distribution of the mineral deposits, for example anhydrite was not encountered in the Woo Dale Borehole.

Given the extensive nature of the chert deposits in the Bakewell area (Appendix 3.2) this author was interested to see whether or not the Bakewell Risings contained elevated concentrations of silica. On 6 May 2004 a groundwater sample was obtained from Bakewell Recreation Ground. The analytical results have been presented as Appendix 5.1. The Bakewell Thermals are also interesting in that they are characterised by particularly high strontium to calcium ratio in comparison with the other thermal springs (Edmunds, 1971); attributable to particularly elevated strontium concentrations. Evidence from Table A3.2/1 (Appendix 3.2) also indicates elevated strontium concentrations associated with the Pretoria Chert Mine chert. The chemistry (low bicarbonate and high sulphate) points to a source of celestine ($SrSO_4$). It is considered unlikely that this is a hydrothermal mineral associated with the mineralization of the White Peak, because Edmunds (1971) has shown that strontium concentrations are lower in the adjacent Namurian strata (considered to be the source for mineralization, Chapter 3). It is more likely that it is derived from buried evaporites. However, as there is a complete solid solution series between barytes and celestine it is possible that the celestine is associated with barytes. Furthermore, the calcium concentrations are relatively high (161-186 mg/l, Table 5.3) and magnesium concentrations are moderate (17 to 24 mg/l, Table 5.3) suggesting the possibility of dolomite dissolution. Dolomite is commonly the host for celestine (Deer, Howie and Zussman, 1980). In support of the buried evaporite hypothesis, the chemistry, with relatively low bicarbonate (163-175mg/l) could also be indicative of a process comparable with that described by Bischoff et al. (1994), sub-section 4.4, whereby evaporite dissolution drives dedolomitization.

Edmunds (1971) suggested that the elevated sodium and chloride concentrations in thermal water sourced from Bradwell, possibly reflect the remnants of connate water. As for the Bakewell Thermals, an alternative hypothesis is that the groundwater chemistries are indicative of the occurrence of different suites of evaporite minerals in basinal areas formed by regional movement of the fault blocks. To illustrate this, the following "inter-block" basinal evaporite mineral assemblages could be inferred from the reported groundwater chemistries (Table 5.3):

Table 5.4: Inferred Evaporites.

Thermal Centre	Inferred Evaporite	Molecular notation
Bradwell	halite, sylvite, anhydrite, kieserite	$NaCl$; KCl; $CaSO_4$; $MgSO_4$
Stoney Middleton	halite	$NaCl$
Buxton	None	
Lees Bottom/Magpie Sough	halite	$NaCl$
Bakewell	celestine,anhydrite	$SrSO_4$; $CaSO_4$
Beresford Dale	None	
Matlock	anhydrite, kieserite	$CaSO_4$; $MgSO_4$
Ball Eye Quarry/Meerbrook Sough/ Crich	halite	$NaCl$

Potassium salts are highly soluble and generally the last to form during the evaporation of sea-water. They are most commonly encountered in sabkha type environments, which are the type of environment that has been associated with the deposits of anhydrite encountered in the Eyam Borehole (Dunham, 1973). If the elevated concentrations of potassium determined in the Bradwell groundwater (Table 5.3) were derived from salts it could be inferred that the basin occupied a higher topographic level. Furthermore the inferred sequence of evaporites from Matlock north to Bradwell is typical of the sequence that might be encountered in a barred marine basin (Tucker, 1991), with anhydrite being precipitated in the deeper part of the basin, halite farther up the sequence and potash salts at the top. Meerbrook Sough and Crich lie to the south of the Cronkston-Bonsall Fault and appear to have a different chemistry to the groundwater at Matlock.

Other models of the thermal waters could be considered, for instance, using silica geothermometery Pentecost (1999) suggested a maximum temperature of 35˚ C for the Matlock springs at depth. He attributed the eightfold increase in the concentration of strontium, fourfold increase in potassium, six fold in magnesium and twofold in sodium and sulphate in the thermal waters relative to the cold springs, to a combination of residence time, pressure solution, and contact with the Millstone Grit Group, and/or basalt. Accordingly he suggested a maximum circulation depth in the order of 500 m, with groundwater being derived from either the Millstone Grit Group, or the limestones to the west of Matlock, becoming confined below the Lower Lava and ascending at the lava outcrops. For water to become confined in this way an up hydraulic gradient (northwesterly) source would be considered more likely by this author. The geothermal gradient associated with this hypothesis would be 38˚ C/km and clearly this would contradict the concept of thermal flux from the limestone.

5.5.7 Caves as indicators of flow vectors.

Drawn largely from the work of Beck and Gill (1991) the location of 'bedding plane' caves within the study area have been plotted and presented as Figure 4.1. The inferred directions of flow (Appendix 4.2, determined from the data presented by Beck and Gill (1991)) provide additional evidence for a generally easterly hydraulic gradient in the northern part of the area, with a more southeasterly trend, towards the Cronkston-Bonsall Fault, in the south. However, there are exceptions, which should be accommodated in any hydrogeological modelling (Chapter 9).

5.6 Defining the regional base level.

Having defined the regional hydraulic gradient it is logical to look to the east and southeast for the regional base-level determined by the fault blocks under investigation. The easterly and southeasterly hydraulic gradient and the drainage of the river Derwent, and also the River Dove, into the River Trent (Downing et al., 1970) point to the River Trent as the regional base-level. During the early development of the Midlands drainage system the River Trent captured streams flowing east through the Jurassic escarpment at Lincoln (and Ancaster) and diverted them to the Humber (Downing et al., 1970, Howard and Knight, 2004), probably as a consequence of glacial impedence caused by ice occupying the Vale of Belvoir. Additionally, Aitkenhead et al. (2005) and Howard and Knight (2004) noted the existence of west to east oriented tunnel valleys filled with till and possibly pre-dating the Anglian glaciation in the area of the confluence of the rivers Derwent and the Trent, further suggesting early development of a southeasterly hydraulic gradient. The easterly watershed of the River Trent is very low, being formed of Lias Clays. Furthermore, the overlying Upper Carboniferous strata, where mudstones are present above the Millstone Grit Group, confine the groundwater in the limestone in an easterly direction, such that any recharge to the River Trent from the limestone must be via faults, allowing mixing with groundwater in the sandstones, with resultant masking of the chemistry of the limestone groundwater. Stem pressures at Eakring (in the order of 37 km to the eastnortheast of Matlock) indicate that local recharge to the River Trent is potentially possible, although examination of the geology sheet indicates that faulting is apparently less pronounced in the northern part of the Lower Trent River Basin than in the south, and to the east of the River Trent the easterly dip of the strata is much less disturbed. It should also be noted that many faults in the Triassic strata have not been shown by the British Geological Survey, particularly in areas of undifferentiated Keuper Marl (Lowe, personal communication, 2004). Accordingly, for modelling purposes the adoption of the River Trent as the regional base level is appropriate. However, the structural evidence suggests the likelihood of further, albeit minimal, underflow to the southeast.

It is considered unlikely by this author that groundwater from the fault blocks being investigated in this thesis target the up-faulted inliers of Carboniferous Limestone in the area of Ticknall and Breedon (immediately to the north of inliers of pre-Carboniferous Charnian Supergroup basement rocks), because they are situated to the southeast of the study area, albeit that these strata are known to be karstified (Carney et al., 2001).

5.7 Towards a quantification of underflow.

The temperature and discharge data presented in Table 5.5 have been used to calculate the thermal output from the springs (Gunn et al., 2006 and Worthington, 1991). Worthington (1991) calculated that 96 % of the thermal flux of the White Peak emanates at Matlock.

Table 5.5: Temperature and discharge of the thermal springs (from Edmunds, 1971).

Thermal centre and source	Temperature ˚C	Discharge (l/sec)
1. Buxton (St. Anne's Well)	27.5	10.6
2. Bradwell Spring	12.4	0.7
3. Stoney Middleton Spring	17.7	1.3
Stoke Sough	11.6	1.3*
4. Bakewell British Legion (spring)	11.6	9.3
Bakewell Recreation Ground (spring)	13.3	0.2
5. Matlock Fountain Bath	13.3	0.2
New Bath Hotel	19.8	6.3
East Bank of Derwent	17.4	0.5
Ball Eye Quarry	13.6	1.5 (pumped)
Meerbrook Sough	15.3	790
6. Crich (Ridgeway Sough)	14.1	4.0
7. Beresford Dale Spring	13.8	1.7
8. Lower Dimindale Spring	11.5	4.2

* thought to be in error by Gunn, who has noted higher discharges (personal communication, 2005).
Note also, Edmunds (1971) did not include thermal water in Magpie Sough with the thermal springs.

Further consideration of the thermal flux at Matlock, in the light of the fault-block interpretation of the structure (Figure 5.2) indicates that there are a number of other factors to consider:

i) It would appear that the limestone has a lower than 'average' geothermal gradient, as a consequence of heat flux, which is known to convey heat towards Eakring.

ii) There may be more than one heat source for the basement rocks.

iii) If the fault block interpretation of this chapter is correct, Meerbrook Sough and Ridgeway Sough drain a different block to that drained by the Matlock springs. In this case the thermal output of the southern block would be significantly greater than that of the Woo Dale block (as represented by the Matlock group of springs).

iv) "Scarp" slopes in the basement blocks are inferred as being significant in directing groundwater flow up zones of discontinuity. This is not seen as contradicting the flow mechanism hypothesised for the Buxton spring by Barker et al. (2000).

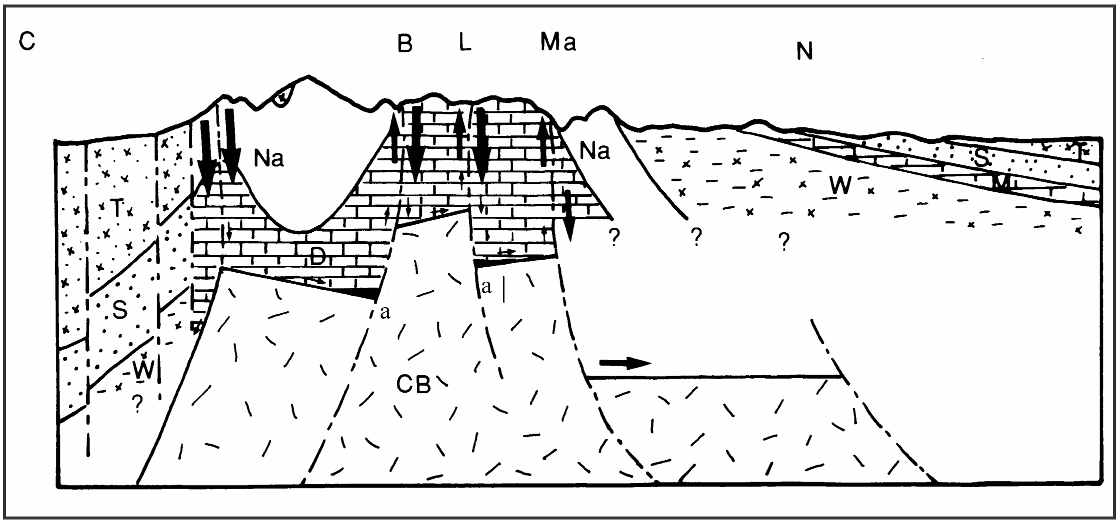
Hypothesised flow paths (Figure 5.5) indicate the occurrence of underflow. However, it is considered by this author that the relatively high permeability associated with the flow paths feeding the thermal springs captures a significant proportion of the potential underflow, thereby minimising the actual underflow from the region.

5.8 Application of a flow net to the regional hydrogeology.

Tóth (1963) demonstrated that the physical dimensions and topography of a flow system influence flow net development. Thrailkill (1968) showed that the pattern of flow in a limestone aquifer is similar under Darcy, laminar (Hagen-Poiseuille), or turbulent (Darcy-Weisbach) flow conditions, providing that the aquifer is wide relative to its depth. The dimensions of the Peak District flow system are in the order of at least 60 km wide (west to east) by 270 m (Woo Dale Borehole) to 1800 m (Eyam Borehole) deep. Based on this ratio only local flow paths would be expected to develop in an isotropic aquifer (Tóth 1963). Thus flow net development of the aquifer is unlikely. However, Worthington (1991) has suggested that the thermal springs of the Peak District can be attributed to a deep flow net, although an argument that he presents in support of this is that the springs are not artesian and the evidence provided in this thesis would not entirely support this. Furthermore, as flow from the limestone is suspected to reach Eakring, the width of the aquifer can be taken as 90 km.

A conceptual model showing the suspected flow paths is presented as Figure 5.5. The significance of sulphate in deep-seated flow paths has been described in sub-section 5.5.6. Accordingly inter-block basinal salts are shown on the conceptual model of the regional hydrogeology (Figure 5.5). The exact form of the speleogenetic processes involved in the development of this dominant zone of flow remains unanswered. The geological setting suggests that it is very plausible that the flow paths started to form as very early drainage paths for the release of pore pressure induced by sedimentary loading, thus effectively comprising inception horizons (Lowe, 1992). Alternatively flow paths may have developed at a later stage, as a consequence of gypsum driven dedolomitization. Flow net development of the type envisaged by Worthington (1991) would require karstification to have commenced at the stage at which infiltrating surface water breached the cover of Upper Carboniferous, Permo-Triassic and Cretaceous deposits, although it should be noted that initial "breaching" was probably via dominant faults long before actual cover removal occurred. At the intermediate basin scale, the geological guidance of dissolution is such that flow net application would be a misleading representation of flow in the aquifer. For example, the dominance of inception horizon-related flow and observations at Millclose Mine (Traill, 1939, p. 887) that "*South and west of the mine the limestone surface reaches an elevation 900 feet or so above the level of the River Derwent, which controls the groundwater level. This circumstance, along with the alternation of permeable and impermeable layers, and the presence, near the river, of open fissures, gives the conditions necessary for artesian circulation. This is put forward as a possible explanation of the extreme depth of the oxidised zone, and may even explain the original cavernization.*" Perhaps on the local scale however, for example the scale of an individual spring outlet the flow net is again applicable, because it is modelling unconfined conditions and fissures contribute more evenly to the permeability, which is partly attributable to the effects of stress relief (Thrailkill, 1968).

Brassington (2007) has presented an alternative conceptual model. This is further described in section 12.2



Key: Localities: C Cheshire Basin; B Buxton; L Lees Bottom; Ma Matlock; N Nottingham
Lithologies: T Triassic Mudstones; S Sherwood Sandstone; M Magnesian Limestone; W Westphalian; Na Namurian; CB Caledonian Basement; a indicates possible location of inter-block anhydrite deposits. Vertical exaggeration x 20.
Arrows indicate groundwater flow paths

Figure 5.5: Conceptual model of the regional hydrogeology (adapted from the Hydrogeological Map of England and Wales).

Chapter 6: Hydrogeochemistry and the analysis of springs.

6.1 Introduction.

Research carried out by a number of karst specialists has demonstrated that groundwater can be characterized by the degree of equilibrium between the water and the wall rock (Drake and Harmon, 1973; Richardson, 1968; Shuster and White, 1971, Smith and Atkinson, 1976). Various authors have used groundwater chemistry to assess the nature of springs, for instance the degree of equilibrium between the groundwater and the wall rock has been taken as an indicator of the type of flow feeding the spring (e.g. Atkinson, 1977a; Shuster and White, 1971; Worthington, 1991, Worthington and Ford, 1995). Shuster and White (1971) classified diffuse and conduit flow waters, Bertenshaw (1981) vadose and phreatic, or open and closed systems and Worthington (1991) underflow and overflow systems. Kehew (2001, p. 16) suggested, "*By knowing the state of equilibrium between the water and minerals within the aquifer, we can predict the type of reactions that are occurring or would be likely to occur.*" It is in this context that this thesis aimed to reassess the chemistry of the karst waters of the area of investigation. The sources of data used in the assessment are described in section 6.2.

Models based solely on spring water chemistry do need to be treated with caution, because as Newson (1971) identified and Christopher (1981) confirmed, the majority of dissolution appears to occur at rockhead i.e. within the epikarst. Gunn (1986, p. 382) reported that net erosion budgets have shown that "*a high proportion of solution (50 to 80 %) occurs within several metres of the surface in the soil (if present) and subcutaneous zone..*". Where epikarst dissolution is the dominant process, groundwater chemistry will be more significantly influenced by the geology of any covering of superficial material and the epikarst, which could mask any variation in groundwater chemistry resulting from dissolution along the remainder of the length of the flow paths. Newson (1971) also observed that in the Mendips the mechanical processes of erosion (abrasion) are dominant when streams are in flood. Notwithstanding these observations, the aim of this chapter is to try to identify hydrogeochemical markers to assist in the identification of flow paths. One reason for persisting with the theme of geological characterization of groundwater chemistry is the observation that many dolines are associated with faults (Chapter 5) and where the cover materials are thin and storage is low, such dolines offer the potential to guide a source of rapid autogenic recharge, facilitating dissolution deeper in the aquifer.

A number of factors influence geochemistry, including: flow-through time (water-soil contact or water-rock contact time); atmospheric conditions, including temperature; thickness and type of topsoil; thickness of epikarst and bedrock geology, thus making classification difficult. That these parameters can be listed indicates that classification should be possible, albeit that it may be complex. Generally the aim is to derive a classification based on the minimum number of parameters possible.

Accordingly, the classification schemes that have been established to date have been derived from: individual parameters (Hem, 1970), ratios of parameters (Downing, 1967, Hem, 1970, and Vervier, 1990), flow-through times based on hardness and P_{CO_2} (Pitty, 1966), seasonal variation of parameters (Shuster and White, 1971) and variation with discharge (Jacobson and Langmuir, 1974). Attempts have been made to apply a number of these methods of classification to the spring waters of the White Peak and the findings are presented in sections 6.3.2 and 6.5. The key aim of this work is to identify whether the geochemistry of the springs can be used to identify the contributory flow paths within the aquifer.

6.2 Previous work and sources of information with respect to the chemistry of the springs of the Wye catchment.

A regional study of the chemistry of spring sediments was reported by Nichol et al. (1970). This was a meticulous piece of research; however it is of limited use to this investigation because of the dearth of data from the limestone shelf, owing to the absence of surface water courses within the outcrop of the limestone. Accordingly, it would appear that the mineralization of the limestone is not properly represented in the data. The findings provide a useful reference for the interpretation of background metal concentrations (Appendix 6.1). Similar results were obtained by Salzmann (2002) working on sediment samples in Lathkill Dale (Appendix 6.1).

A range of chemical parameters including: pH, total hardness, ammoniacal nitrogen, anions and cations (including: strontium, lithium, rubidium, caesium, aluminium, chromium, copper, iron, molybdenum, manganese, lead, nickel, titanium, zinc, bicarbonate, sulphate, chloride, fluoride, bromide, iodide, nitrate, nitrogen dioxide, hydrogen phosphate, selenium oxide, arsenic oxide and borate) were determined by Edmunds (1971) for a selection of spring, sough and borehole waters. Edmunds (1971) classified groundwaters on the basis of the geology at the point of egress (section 6.5), but saw no reason to try to subdivide the limestone groundwater on the basis of geological formation. Mine drainage waters showed enrichment with lead, selenium-oxide, nickel, iron, strontium, zinc, borate and fluoride, whilst thermal waters showed enrichment in all constituents except nitrate relative to the limestone background (Edmunds, 1971). However, Bertenshaw (1981) established that fluoride was the best hydrogeochemical indicator of mineralization.

In the initial chapters of his work on the groundwaters of north Derbyshire, Christopher (1981) examined the temporal variability of twenty sites in north Derbyshire and northeast Cheshire. Ten of the Derbyshire sites fell within this research area. He concluded that the dominant control on temporal variability of resurgences was rainfall during the preceding 48 hours, but there were also other influences, including: the degree of contact with mudstone, dolomite and lava; any pollution; and the concept of open versus closed systems. The terms open and closed system are defined by Appelo and Postma (2005), Gunn (1986) and Smith and Atkinson (1976). The open system is one in which gas, water and rock are all in contact with one another such that carbon dioxide is available to replace that used up in the reaction of limestone and carbonic acid. The closed system is one in which gas and water come into equilibrium, but a replacement supply of carbon dioxide is not continuously available as the reaction of limestone and carbonic acid takes place. As Christopher (1981) acknowledges, there is a gradation between fully open and fully closed systems. In the context of trying to identify flow paths in the limestone it is worth noting that deep, regional flow paths comprise closed systems (section 6.4). This is important because the identification of a "closed system" could be indicative of a more significant contribution from rising groundwater. For this reason this author considers it important to interpret the geochemistry in the context of the geological setting, for example tensional fault zones offer the means of connection between differing groundwater contributions. Furthermore, as noted in Chapter 4, there are other karst processes operating within soluble rock aquifers.

Christopher (1981) initially classified the sites that he investigated in five groups, namely: surface streams; resurgences fed by surface streams; intermediate types; thermals; and percolation resurgences (open or closed). These categories demonstrate an aqueous evolution reflected in increasing total hardness, dissolved calcium concentration and reduced variances of many ionic species. This largely relates to the H_2O-CO_2-CaO system approaching equilibrium (Jacobson and Langmuir, 1974; Shuster and White, 1971, White, 1969).

Christopher (1981) found that intermediate category springs (intermediate between the surface streams, allogenic resurgences and percolation fed autogenic resurgences) were characterized by median total hardness in the range 240 to 280 mg/l, >90 mg/l Ca, temperature variance of 3.4 % or less (compared with >8 % in surface waters) and greater than 80 % saturation with respect to calcite. Deep Dale Main Resurgence (SK 097714) and Wormhill Springs (SK 123735) fall within the intermediate type category. These springs emanate at, or close to, the boundary between the Woo Dale Limestone and the overlying Bee Low Limestone formations (Appendix 6.2) and are associated with major faults (Chapter 2). This author opines that they receive a considerable input of groundwater from the Woo Dale Limestone Formation (section 6.5) and that they represent mature karst systems, as indicated by the results of dye tracing (Chapter 7).

Open percolation resurgences identified by Christopher (1981) included: Magpie Sough (SK 180696), Lumb Hole (SK 173733) and Moss Well (SK 178723). They appear to be associated with mineral veins. The hydrogeochemistry of open percolation resurgences was shown to have many similarities with the intermediate sites, e.g. with respect to temperature, total hardness, Ca, Mg, Na and K, with variance in the same ranges, except with respect to K, which is slightly higher. The main difference was with the saturation index with respect to calcite (SIc) and P_{CO_2}. All of the open percolation sites were supersaturated with respect to calcite, with high P_{CO_2}.

Closed percolation-fed resurgences identified by Christopher (1981) included: Cheedale Bridge Resurgence (SK 128735); Lower Cheedale (SK 130734); Brindley's Well (SK 124742); Ashwood Dale Resurgence (SK 089722), and Cowdale (more recently referred to as Rockhead Spring at SK 084722). These resurgences have catchments that are predominantly within the Bee Low Limestone Formation.

They exhibited high total hardness (median total hardness in the range 271 to 308 mg/l), high calcium contents (99 to 119 mg/l) and low P_{CO_2}. With the exception of Lower Cheedale and Brindley's Well there was a low variability in these parameters. Christopher (1981) attributed the variability of the chemistry of the Lower Cheedale and Brindley's Well resurgences to lower summer flows, producing almost stagnant resurgence pools that warm and evaporate. Furthermore, with small catchments and rapid flow-through they did not reach equilibrium with the less reactive mineral phases and therefore exhibited lower concentrations of magnesium despite being on the lava (Christopher, 1981).

From the distribution of results, Christopher (1981) observed that the lowest temperatures corresponded with high altitude allogenic sinks, adjacent to the Namurian mudstones, on the northwestern edge of the Dinantian outcrop. Elsewhere average temperatures, with the exception of the thermal springs, were broadly in the range 9.6 to 11.2° C. There was a general broad band of low magnesium water in the northwest of the Dinantian outcrop, with higher concentrations to the southeast and centered on the areas of thermal water. Christopher's (1981) results indicated an area of high hardness associated with northern outcrops of lava and areas towards the centre of the dome, remote from the margin of the Namurian strata. Christopher (1981) observed that virtually all calcium fluctuations could be related to flood dilution rather than to temperature or P_{CO_2}. Work carried out by this author broadly appears to substantiate that of Christopher (1981); has developed some of the ideas; identified additional groundwater types (sections 6.4 and 6.5) and has emphasized the contribution of underflow components.

Smith (2000) collected water samples from a number of springs in the Wye catchment on an approximately quarterly basis during 1998, commencing February 1998. The results of the analyses were published by Smith et al. (2001). The limitations of the results should be recognized. The monitoring was only carried out on a quarterly basis, for biological rather than hydrogeological purposes. Accordingly, at the time of monitoring, no consideration was given to groundwater levels. The mean value of ionic charge balance was 5.75 %, which is higher than the 5 % value that Freeze and Cherry (1979) recommend as a reasonable limit for accepting an analysis as valid. Ionic charge balance values as high as 13.62 % have been determined by SOLMINEQ.GW, for the data, with 23 of the values exceeding 7 %, the limit adopted by Christopher (1981). These data have not been rejected because there appear to be consistencies to the ionic imbalances that have been determined and rather than being solely date-related the imbalances show some correlation with catchment formation. Furthermore, it has been noticed that in some cases the pattern of charge errors extends to analyses carried out by others within the Limestone Research Group. In particular it is noted, that where negative imbalances have been observed, these are usually associated with either the Monsal Dale Limestone, or with the Woo Dale Limestone and also with Litton Spring. This indicates that the analyses might not cover an extensive enough range of cations and anions, particularly where there is mineralization of the limestones.

Terjesen et al. (1961) demonstrated experimentally that the presence of low concentrations of metals could act as an inhibitor to the formation of calcium bicarbonate. They attributed this to an increase in the surface concentration of carbonate ions. This property of inhibition is increasingly effective as equilibrium is reached and results in an apparent equilibrium for calcium, which is considerably, lower than it would be without the presence of the metals. A number of metal ions were tested as inhibitors and the ions were placed in an order of decreasing effectiveness as inhibitors, namely lead, lanthanum, yttrium, scandium, cadmium, copper, gold, zinc, germanium and manganese. The analyses presented in Appendix 6.3 indicate that there are low concentrations of metals present in solution in the groundwater and therefore this property may help to account for negative imbalances. Contamination can also inhibit solubility; Berner and Morse (1974) found that trace concentrations of phosphate depressed calcite solubility rates by up to three orders of magnitude. A further reason for ionic imbalance is one of incomplete analysis.

Christopher (1981) also discussed the problem of ionic balance and pointed out that there were some high ionic balance values in the Edmunds (1971) data, which he attributed to the presence of road salt. Clearly this does need to be borne in mind when considering ionic balance, particularly as many of the springs are in close proximity to the A6 road. However, elevated chloride concentrations do not always correspond with the roadside locations and although cattle salt licks may provide another source of salt, the approach taken by this author has been to accept the data, unless there is strong reason or evidence to suspect contamination.

The analyses published by Smith et al. (2001) have been tabulated (Table 6.2), with springs grouped according to the source catchment geology interpreted by this author. Lees Bottom 3 Spring is considered to consist of a significant proportion of thermal water and thus the dominant chemical markers are likely to be derived within the deeper portion of the flow path rather than in the unsaturated zone.

Table 6.1 Springs sampled for analysis by Smith (2000).

Spring	National Grid Reference	Geology at Egress
Cowdale (Rockhead) Spring	SK 08667229	Woo Dale Limestone
Kidtor Spring	SK 08697219	Woo Dale Limestone
Ashwood Dale Resurgence	SK 08957222	Woo Dale Limestone
Woolow	SK 09477242	Woo Dale Limestone
Topley Pike	SK 10047251	Woo Dale Limestone
Wormhill Springs (West)	SK 12217351	Bee Low Limestone
Wormhill Springs (East)	SK 12467342	Bee Low Limestone
Cheedale (West)	SK 12737343	Bee Low Limestone
Cheedale (East)	SK 13007340	Bee Low Limestone
Litton	SK 16527306	Bee Low Limestone
White Cliff Spring	SK 18217179	Monsal Dale Limestone
Lees Bottom 1	SK 17127074	Monsal Dale Limestone
Lees Bottom 2	SK 17147073	Monsal Dale Limestone
Lees Bottom 3	SK 17137059	Monsal Dale Limestone
Lees Bottom 5	SK 17187035	Monsal Dale Limestone
Great Shacklow 1	SK 17686987	Monsal Dale Limestone over Great Shacklow Lava
Great Shacklow 2	SK 17897073	Monsal Dale Limestone

6.3 Potential influences on karst waters of the 'Derbyshire Dome'.

6.3.1 Inputs and soil chemistry.

Appelo and Postma (2005) suggested that over the oceans the composition of rainwater can be likened to diluted seawater with chloride concentrations of the order of 10 to 15 mg/l. The salt originates from aerosols derived from the oceans. As air masses move over land dust and gases modify the composition of rainwater, for example Appelo and Postma (2005) identified a number of additional chloride sources in industrialized areas including various combustion processes and evaporation from seawater used as coolants. Raper (1989) monitored variations in precipitation chemistry as part of an investigation of acid deposition in the Derbyshire High Peak District. The results of Raper's (1989) work indicate that recharge chemistry is influenced both by the precipitation chemistry and by dry deposition (the difference between the composition of rainwater collected by wet-only gauges and of rainwater collected by bulk samplers, therefore comprising particulate aerosols and adsorbed atmospheric gases). However, in considering the results it should be noted that Ridder et al., (1984, referenced by Appelo and Postma, 2005) found that bulk samplers overestimate the wet input of the major ions by a factor of about 1.3. Nevertheless, Raper (1989) identified that the proportional cationic composition of precipitation (Table 6.3) differed from that of the groundwater. The difference is primarily attributed to dissolution of calcium carbonate and perhaps not surprisingly, groundwater was found to exhibit a proportionately greater composition of calcium.

Raper identified close temporal and spatial association of calcium and non-marine sulphate. This was attributed to the reaction of regional gaseous sulphur dioxide and calcium derived from quarrying. In the bulk deposition non-marine sulphate, nitrate and ammonium were at a maximum during the summer months, whereas sulphur dioxide and oxides of nitrogen were at a maximum during the winter. Sodium and chloride were also at a maximum during the winter months. The majority of Raper's (1989) monitoring sites lay to the north and northeast of Buxton, largely to the north of this author's research area. Deposition of non-marine sulphate, nitrate, ammonium, sodium, chloride and hydrogen were found to increase in a northerly direction, whereas calcium concentrations were higher in an area centred on and to the east of Buxton. The high concentrations of calcium close to the Wye Valley are thought to reflect the locations of the mineral extraction areas. The associated deposition of calcium sulphate suggests a tree canopy influence focused on the Wye valley. Tree canopies intercept bulk deposition. During the summer months evapotranspiration results in a loss of rainwater and an accumulation of salt deposition within the canopy. Subsequently the autumn dry deposition is flushed into the ground with the rain (Appelo and Postma, 2005). Edmunds (1971) has argued that there is a similar accumulation of salts in the epikarst as a consequence of evapotranspirational losses. It is also relevant to note that trees can also be important in concentrating chloride in pore-waters, for example Price et al. (1993) noted that chloride concentrations approaching 1000 mg/l have been found in pore-waters beneath beech trees, the chloride being concentrated by the use of more pure water.

Table 6.2: Summary of spring chemistry (published by Smith et al., 2001) along the Wye Valley, grouped according to the interpreted dominant source catchment geology (identified by formation).

Spring	SIc	SId	Electrolytic Conductivity	Total Dissolved Solids	Mg	Ca	Na	K	Cl	SO_4	Total Hardness as $CaCO_3$	CO_2 (bars)	Dissolved Oxygen (mg/l)
Lees Bottom 3	0.42 0.96	1.5 2.5	360 584	408 440	17.2 19.7	80.2 101	11.4 15.9	0.5 0.8	21.8 26.5	37.3 61.3	274 330	0.0013 0.0074	4.9 7.3
Woo Dale Limestone catchment predominates:													
Kidtor	-0.17 0.76	-0.33 1.54	764 821	615 571	5.5 7.0	109.4 127	20.8 29.3	2.3 4.5	40.1 50.6	24.2 46.8	263 343	0.0054 0.0562	2.5 5.5
Topley Pike	-0.62 0.71	-0.99 1.51	514 641	404 456	6.1 11.0	84.4 103	10.8 22.3	1.4 2.8	22.0 39.9	26.8 63.1	240 282	0.002 0.0406	5.7 14.0
Woolow	-0.33 0.77	-0.33 1.58	622 720	420 446	5.4 9.8	94.1 104	14.2 19.8	0.5 0.9	22.3 56.7	23.5 42.1	260 282	0.0025 0.0347	8.8 14.6
Means for group	-0.37 0.75	-0.55 1.54	633 727	480 491	5.7 9.3	96.0 111.3	15.2 23.8	1.4 2.7	28.1 49.1	26.8 50.7	254 302	0.0030 0.0438	5.7 11.37
Woo Dale Limestone catchment predominates; with a proportion of allogenic recharge:													
Wormhill West	-0.56 0.33	-0.85 0.84	595 622	428 451	6.3 7.6	71.7 101.5	16.2 25.4	2.4 4.9	35.1 43.1	47.8 65.2	206 279	0.0058 0.035	8.0 10.5
Wormhill East	-0.06 0.33	0.03 0.84	589 687	431 451	6.4 7.5	80.9 101	10.5 22.0	3.2 4.3	33.2 41.8	47.8 66.3	230 279	0.0076 0.0136	7.8 12.5
Means for group	-0.31 0.33	-0.41 0.84	592 654	429 451	6.3 7.6	76.3 101.2	13.3 23.7	2.8 4.6	34.1 42.4	47.8 55.7	218 279	0.0067 0.0243	7.9 11.5
Bee Low Limestone catchment predominates:													
Litton	-0.39 0.73	-0.81 0.02	379 586	354 411	1.3 5.8	71.2 100	8.4 14.9	0.6 1.0	16.0 38.0	20.7 42.8	191 257	0.0121 0.024	11.4 12.9
Cowdale	-0.03 0.68	0.14 1.70	506 561	390 445	7.2 9.6	89 99.9	9.4 10.4	0.6 1.5	9.5 16.2	25.0 57.4	262 281	0.0086 0.0178	9.7 18.0
Ashwood Dale	-0.42 0.48	-0.67 1.09	498 611	412 458	5.0 7.6	94.7 110.5	10.6 15.2	0.7 1.0	24.6 52.3	31.0 35.0	257 307	0.0053 0.0436	7.5 13.7
Cheedale West	-0.12 0.06	-0.09 0.26	449 646	379 451	6.2 6.5	82.8 97	12.0 17.7	1.6 3.3	23.0 31.9	38.5 55.7	265 270	0.0100 0.0183	11.8 12.9
Cheedale East	-0.23 0.33	-0.41 0.84	367 619	372 464	5.9 7.0	80.4 112	8.8 20.3	0.1 1.8	17.0 32.2	33.3 47.5	230 304	0.0076 0.02965	10.5 14.6
Means for group	-0.24 0.46	-0.21 0.78	440 605	381 446	5.1 7.3	83.6 103.9	9.8 15.7	0.7 1.7	18.0 34.1	29.7 47.7	241 284	0.0097 0.026	10.2 14.4
Monsal Dale Limestone catchment predominates:													
White Cliff	-0.66 0.796	-1.38 1.87	403 546	388 431	4.0 7.4	82.2 95	11.1 14.8	0.3 1.5	26.1 33.8	29.3 41.7	236 254	0.0016 0.06	9.8 13.9
Lees Bottom 1	-0.64 0.99	-1.24 2.17	591 628	420 457	4.4 7.9	85.9 110	12.3 17.5	0.6 1.4	26.4 32.5	30.6 50.6	233 307	0.0013 0.0604	9.2 11.1
Lees Bottom 2 upr	0.02 0.98	0.30 2.26	437 609	405 437	4.6 9.9	83.3 105	12.5 16.3	0.8 2.0	26.5 31.4	30.2 55.5	229 307	0.0012 0.0121	9.2 9.9
Lees Bottom 5	0.16 1.04	1.50 2.50	429 561	399 455	3.9 10.4	70.4 103	12.0 17.7	1.1 2.0	17.8 22.4	28.8 49.6	200 300	0.0012 0.0099	7.5 8.9
Great Shacklow 2	-0.04 0.51	0.09 1.13	386 580	387 420	5.9 7.6	82.5 108	9.6 11.8	0.3 1.3	18.3 25.6	27.7 49.6	237 297	0.0046 0.0142	9.3 10.3
Great Shacklow 1	-0.11 0.52	-0.15 1.28	542 578	376 440	6.1 7.2	85.8 107.6	10.0 11.5	1.0 1.4	17.8 26.5	27.7 49.9	244 297	0.0037 0.0199	8.4 11.3
Means for group	-0.21 0.81	-0.15 1.87	465 584	396 440	4.8 8.4	81.7 104.8	11.2 14.9	0.7 1.6	22.1 28.7	29.0 49.5	230 294	0.0021 0.0441	8.9 10.9

Data presented as ranges of concentrations (mg/l), except Sic, Sid and electrolytic conductivity (µS/cm).

Clearly another process that influences the spring chemistry, by virtue of its importance in terms of carbonate dissolution, is the contribution of carbon dioxide. Evidence presented by the National Oceanic and Atmospheric Administration (NOAA), supports the concept of rising atmospheric carbon dioxide concentrations. Mean carbon dioxide determined at a number of surface sites including the Mauna Loa Observatory (Hawaii) and globally averaged, indicate that the concentration of atmospheric carbon dioxide for 2006 is close to 380 ppm (http://cdiac.ornl.gov/trends/co2/sio-mlo.htm). Examination of the calculated P_{CO_2} values presented in Tables 6.2 and 6.4 reveals that the majority of the springs support a higher concentration of carbon dioxide than this. The concentrations presented in Tables 6.2 and 6.4 were calculated using the computer programme SOLMINEQ.GW. The method is further described in Appendix 6.4.

Table 6.3: Monthly concentration of ions (mg/l) at Harpur Hill, 1987 -1988 (calculated from original deposition data presented by Raper, 1989).

Period	H	SO_4-S	Cl	NO_3-N	Na	Mg	Ca	NH_4-N	Rainfall (mm)
May	0.02	1.17	2.17	0.66	0.97	0.14	1.07	0.72	50.30
June	0.0.06	1.61	1.67	0.62	0.23	0.04	0.82	0.71	119.00
July	<0.19	1.54	2.31	0.77	0.38	<0.19	2.11	0.77	5.20
August	0.08	1.90	1.96	0.63	0.49	0.082	0.99	0.71	73.50
September	0.04	1.18	3.01	0.41	0.79	0.12	0.67	0.54	76.10
October	0.04	1.06	3.98	0.38	1.56	0.23	0.91	0.52	109.90
November	0.001	0.12	0.36	0.04	0.16	0.02	0.15	0.05	800.80
December	<0.02	1.65	2.42	0.27	0.81	0.11	2.08	0.41	55.70
January	0.02	0.56	3.64	0.21	1.42	0.18	0.13	0.34	130.20
February	0.06	0.99	6.86	0.37	3.97	0.44	0.85	0.39	56.60
March	0.01	0.99	2.98	0.96	0.13	0.38	0.31	0.0052	129.10
April	<0.03	2.59	2.68	0.86	0.73	0.11	2.46	0.62	37.00

P_{CO_2} determinations calculated from the Edmunds (1971) data (Table 6.4) were between 0.005 and 2.8 %, the majority falling within the range 0.1 to 2.8 %. It is generally accepted (Pitty, 1966; Atkinson, 1977b) that soil is the main source of carbon dioxide in groundwater, where it is generated by biological activity and infiltrates with groundwater. The elevation of these values is likely to reflect agricultural practice and the application of manure to pasture land. Smith and Atkinson (1976, p. 377) found: "*In general, carbon dioxide concentrations increase with depth and it is the carbon dioxide concentration at the base of the soil profile which determines how much limestone is dissolved.*" This can be attributed to its density, which is greater than air. It is normal to consider the generation of carbon dioxide as a near surface process, reaching a maximum towards the base of the root zone. However, in the context of the Peak District it has been observed that sediment fills fissures to a considerable depth and therefore there is a potential for some carbon dioxide production to extend to greater depth than might normally be expected.

Deeper sources of carbon dioxide, associated with bacterial or microbial decay of organic matter in the joints and fissures of the unsaturated zone were identified in the Mendip Hills (Atkinson, 1977b, p. 118), where this was termed "*ground air CO_2*". Atkinson (1977b) suggested that both aerobic and anaerobic bacterial processes generate ground gas. Where fissures are open there is a potential for oxygenated water to come into contact with the deeper soil. Another source of carbon dioxide that has not been considered previously in the context of the Peak District is that which could be derived from naturally occurring hydrocarbons in the Eyam, Monsal Dale and Woo Dale Limestone. More specifically, this author has observed the presence of relict hydrocarbons, understood to have been associated with the mineralization of the area, associated with the inception horizons in the Eyam Limestone (Chapter 3). These are suspected to form a potential carbon dioxide reservoir. Ewbank et al. (1996) determined that these hydrocarbons comprise a bitumen – oil suite that shows good correlation with a source of mudstone in the Edale Gulf. Ewbank et al. (1993, p. 584) stated "*The bitumen suite exhibits a range of biodegredative effects, extending from partial to complete removal of n-alkanes, through development of an unresolved complex mixture of hydrocarbons, to the removal of acyclic isopenoids.*" This suggests that, at least locally, there is further potential for biodegradation with resultant carbon dioxide generation. Schofield and Adams (1985) observed that their proposed Topley Pike Member of the Woo Dale Limestone contains discontinuous seams of carbonaceous material up to 20 mm thick, particularly in the uppermost 10 m of the Woo Dale Limestone of Tunstead Quarry, which they have interpreted as the residues left after large-scale pressure solution of the limestone and which potentially form another carbon dioxide reservoir. Similar sources have been found to be important elsewhere, for example Plummer and Back (1980), in their mass balance assessment of a Floridan aquifer, established an additional source of carbon dioxide, which they suspected to be derived from oxidation of lignite accompanying sulphate reduction within the aquifer. The identification of additional sources of carbon dioxide indicates another potential mechanism for groundwater mixing (section 6.3.3).

Notwithstanding the potential sources for carbon dioxide that have been postulated above, some of the P_{CO_2} concentrations that were determined from Smith's (2000) data (Table 6.2) exceeded the concentrations that are considered to be typical for UK soils (<3.5 %, Gunn personal communication, 2005). It is considered most likely that this reflects inaccuracy in the field determination of the pH. The variation in the production of carbon dioxide is influenced by: temperature, soil organic content, soil moisture content, soil density, types of vegetation and soil thickness. Accordingly, seasonality of carbon dioxide production is the norm, reflecting the increased activity of microorganisms with increases in temperature such that carbon dioxide concentrations reach a maximum in late summer (Kehew, 2001). Where this is not evident in the data there may be additional grounds for suspecting measurement error. In this context, Figure 6.1 indicates that many of the P_{CO_2} determinations for February 1998 appear to be anomalous.

It is also worthy of note that some of the P_{CO_2} determinations from the Edmunds (1971) data appear to be anomalously low. The procedures reported by Edmunds (1971) indicate that there is less likelihood of error in the determination of pH than with the Smith (2000) data. Furthermore, this author sampled two of the springs in November 2004, when high pH values were again recorded (Appendix 6.5). Accordingly, consideration has been given to the possible geochemical processes influencing the pH at Dirtlow Farm 1, Dirtlow Farm 2 and Miller's Dale 1. The simplest explanation is that these spring waters represent groundwater from systems that are closed to carbon dioxide. A closed system can arise where active dissolution occurs below the water table, rather than in the unsaturated zone, reflecting the lower rate of diffusion of carbon dioxide in groundwater compared with the more rapid exchange of carbon dioxide in the gas phase of the unsaturated zone (Appelo and Postma, 2005). The springs identified as Dirtlow Farm 1 (SK 188686), Dirtlow Farm 2 (SK 193685) and Miller's Dale 1 (SK 137731) are all springs that emanate from the Monsal Dale Limestone in areas associated with chert deposits and silicification, which, depending on the flow path availability, reduces the availability of calcium carbonate available for dissolution in the unsaturated zone. Furthermore the presence of the low permeability chert deposits and also the clay wayboards associated with the Monsal Dale Limestone perches groundwater, thereby reducing carbon dioxide diffusion. A third contributory factor could be the storage of groundwater in partially saturated clay wayboards.

Table 6.4: P_{CO_2} concentrations (bars) determined from Edmunds (1971) data using SOLMINEQ. GW (to convert to percentage x 100).

Spring:	Calculated P_{CO_2}	Spring:	Calculated P_{CO_2}
Dirtlow Farm – 1	0.00005	Litton Mill – 1	0.00723
Dirtlow Farm – 2	0.00061	Deep Dale	0.00842
Miller's Dale – 1	0.00108	Bakewell British Legion	0.00848
Ravensdale Cottages	0.00126	Lathkill Head Cave	0.00872
Taddington High Well	0.00127	Calesdale	0.0088
Nether Low	0.00226	Priestcliffe	0.00996
Stoke Sough	0.00295	Yatestoop Sough	0.01024
Hillcarr Sough	0.00506	Pictor	0.01074
Great Shacklow	0.00523	Monks Dale	0.01124
Magpie Sough	0.00523	Lower Dimindale	0.01157
Mandale Sough	0.00523	Bradwell Thermal	0.01208
Litton Mill – 2	0.00612	Woo Dale	0.0125
Miller's Dale – 2	0.00632	Staden Farm	0.01512
Bakewell Recreation Ground	0.00651	Waterloo Inn Sough	0.02189
Carter's Mill	0.00675	Chelmorton	0.02803

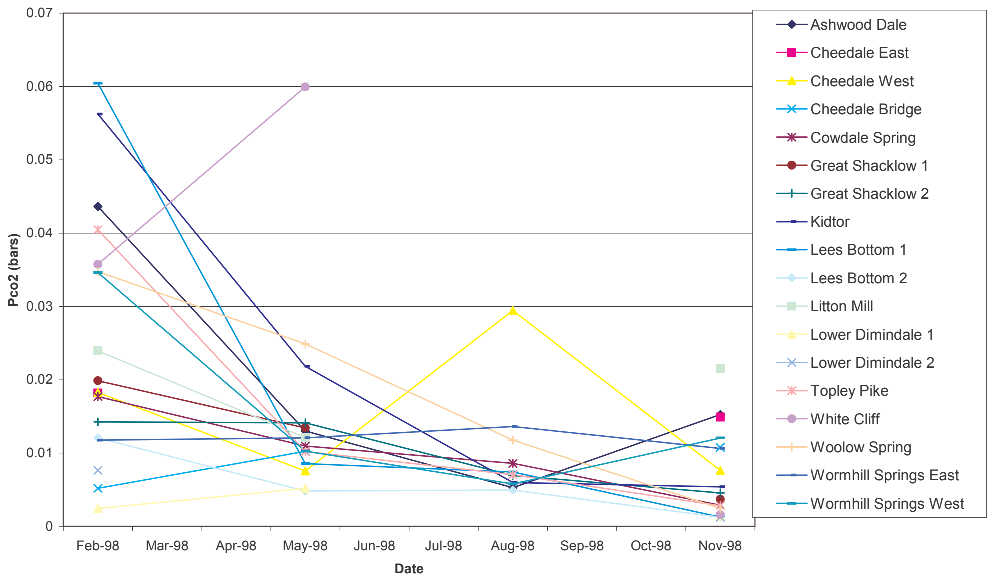


Figure 6.1: P_{CO_2} concentration against time (data from Smith, 2000).

An alternative explanation for the high pH values (and consequentially low P_{CO_2}) relates to the presence of hydrocarbons in association with chert deposits and localized mineralization associated with the Monsal Dale Limestone (Chapter 3), for it has also been observed by this author that the springs with the elevated pH determinations also contain relatively elevated concentrations of silica (Appendix 6.5). In the presence of certain organic acids quartz solubility is enhanced (Bennett et al., 1988). Exposed quartz surfaces, like most silicate minerals in aqueous environments, are characterized by hydroxyl ions (Kehew, 2001). Bennett et al. (1988) suggested that bonds between organic acid ligands and surficial hydroxyl polarize and weaken the interior Si-O bonds such that entire acid-silica complexes are released to solution. It was shown by Cozzarelli et al. (1990) that organic acid compounds are formed by bacteria from alkylbenzene precursors in anoxic environments associated with the downstream tail of hydrocarbon plumes. Given the presence of hydrocarbons associated with the mineralization of the Peak District it would seem that there is a potential for similar processes to occur in the Monsal Dale Limestone. Work carried out by Ewbank et al., (1993) indicates a paucity of aromatic hydrocarbons, but a range of aliphatic and NSO (resin) compounds. Although the latter formed the focus for the work of Ewbank et al. (1993) in the establishment of source rocks for the hydrocarbons, gas chromatograms were presented for the bitumens and they do show the presence of variable amounts of lighter hydrocarbons, which were found to be dominated by n-alkanes/alkenes (Ewbank et al., 1993). Furthermore, the production of organic acids associated with other hydrocarbons, including gasoline and creosote has also been described by Cozzarelli et al. (1994). Thus it would seem that there is a potential for hydrocarbon related dissolution of silica with an associated rise in the pH of the spring water.

Clearly, as much of the dissolution associated with spring water chemistry occurs close to ground surface, the nature of the cover materials is particularly important. However, springs are commonly fed by multiple sources and therefore the chemistry is also influenced by the relative discharge of the various flow paths. Combining of flow paths can also be significant in providing waters of different chemistries that become aggressive upon mixing (section 6.3.3). Burek (1978) confirmed that the soils of Derbyshire are either residual deposits from limestone dissolution, or aeolian sediment derived from the surrounding Silesian strata. She showed that the clay fraction of the soils contains illite, chlorite and feldspar fragments, which would be likely to result in water enriched with potassium, magnesium, sodium, calcium and silica. Enrichment of the superficial deposits by potassium is particularly evident in Table 4.3 of Burek (1978), modified and presented here as Table 6.5. Mudstones are enriched in potassium, due to ion exchange processes during sedimentation and the potassium is retained in the breakdown fraction of the soil as a result of adsorption onto other clay minerals. Burek (1978) established that there is a relatively even distribution of potassium over the region, except for a localized high (7 %) in a temporary exposure in Lees Bottom (SK 171705). Interestingly, this appears to be evident in the chemistry of the spring designated Lees Bottom 5 (section 6.5). It is clear from this, that variation in the potassium concentration in spring waters is likely to be influenced by the degree of contact of the groundwater with the superficial deposits.

Allogenic water, i.e. run-off from the bordering Namurian strata, provides a source of potentially aggressive water feeding into ponors around the edge of the Dinantian outcrop (Figure 5.3). Christopher (1981) characterised these waters as being undersaturated with respect to calcite and dolomite and having low total hardness, with high P_{CO_2}. Raper (1989) identified magnesium as the dominant base cation in the surface water catchments that he investigated. However, Christopher (1981) identified that the contribution of magnesium, which is often interpreted as an indicator of dolomitic bedrock (Shuster and White, 1971) generally increased in a southeasterly direction across the outcrop of the limestone. Thus it increases down hydraulic gradient, suggesting that the bedrock is a significant contributory factor. Clearly, magnesium requires careful consideration in the context of the spring chemistries.

Table 6.5: Comparison of bedrock geochemistry with superficial sediment data (after Burek, 1978).

Unit:	EY	MDL1	MDL2	MDL3	BLL	WDL	ODS	MG	DL	CM	T	SD	M1	M2
Element /Oxide														
CaO	53.2	54.3	53.3	54.4	56	55.4	8.6	0.1	31.6	-	-	7		11
SiO_2	1.7	1.1	2.45	0.62	0.15	0.16	49	97.9	1.34	-	-	63.87		58.7
Al_2O_3	0.53	0.09	0.26	0.08	0.01	0.03	14.1	0.57	0.06	-	-	13.27		9.8
MgO	0.47	0.38	0.64	0.27	0.21	0.26	8.3	0.06	20.17	-	-	1.1		1.3
K_2O	0.07	0.02	0.05	0.01	0	0.01	0.44	0.16	0.03	-	-	1.57		1.5
Fe_2O_3	462	240	581	333	199	146				-	-			
Total Fe							10.6	0.45	0.34	-	-	4.77		2.6
Cu	16	12	15	12	6	9	66	20	24	55	10	68	110	162
Zn	16	19	27	20	17	12		100	209		2200	534	120	823
Pb	31	6	7	4	2	15			15	50	95	484	21	39
Mn	257	114	114	137	136	86	1500	600	856	600-8500	2100	1724	900	832
TiO_2	-	-	220	-	-	-	1.74	0.36	-	0.3-0.6	0.3-0.6	0.26	0.45	0.26
Ba	-	-	5	-	-	-	80	10	-	-	-	1422	739	1473
Co	5	-	-	-	5	5	35	20	-	97	25	28	30	75
Cr	-	-	-	-	-	-	310	60	-	97	50	146	100	162
Ga	2	-	-	-	-	-	18	10	-	-	6	18	23	14
Li	5	-	5	-	5	5	25	-	8	-	38	61	62	45
Mo	-	-	-	-	-	-	-	2	-	2	2	41	11	127
Ni	5	-	8	-	5	5	230	45	-	130-200	25	130	100	304
Sn	-	-	-	-	-	-	-	5	-	9	8	1	5	0
Sr	-	-	550	-	610	610	220	10	100	-	20	184	300	362
V	-	-	-	-	-	-	170	85	-	85-200	40	195	270	945
Zr	-	-	10	-	-	-	100	500	-	-	-	433	-	76

Data from Burek, 1978 all major oxides except Fe_2O_3 reported as %. Trace elements mg/kg.

Key:

EL Eyam Limestone; MDL1 Monsal Dale Limestone (upper pale facies); MDL2 Monsal Dale Limestone (dark facies); MDL3 Monsal Dale Limestone (lower pale facies); BLL Bee Low Limestone; WDL Woo Dale Limestone; ODS Olivine dolerite sill; MG Millstone Grit Group; DL Dolomitic limestone; CM Coal Measures; T Triassic, Sherwood Sandstone Group; SD Superficial deposits (mean of 436 samples); M1 Namurian mudstone; M2 Namurian mudstone.

6.3.2 Spring discharge.

"What distinguishes karst springs is that they are output points from a dendritic network of conduits" (Smart and Worthington, 2004b, p. 699). The dendritic form is such that tiers of conduits feed many springs and a single spring may discharge via a number of distributaries, for example Ashwood Dale Resurgence (Appendix 6.2). In the case of tiers of conduits the lowermost members are termed underflow springs and the higher springs overflow springs (Smart and Worthington, 2004b). High level, overflow springs are generally characterised by more variable discharge and chemistry than underflow springs. For clarity, it should also be noted (Gunn, personal communication, 2006) that distributary conduits occur at points of divergent flow, for example Illy Willy Water (subsection 7.2.4). Over much of the catchment there is a significant seasonal recession in groundwater levels, as determined in boreholes and reported in Chapter 8. Therefore, it might be anticipated that some springs will cease to flow during periods of low groundwater conditions, as described in the context of potable water supplies in Chapters 3 and 10. The permanence of flow at other locations is likely to be

indicative of distal ‘bank storage’, or dispersed recharge reaching the spring via slower flow paths than those utilized in the immediate response to rainfall events. However, discharge is also influenced by the depth of flow paths, for instance Smart (1983), studied in the order of eighty karst springs in a vertical range of over 150 m along the Castleguard Valley, Alberta and concluded that the springs appeared to be organized into a vertical hierarchy with steady sustained flow from the lowest ‘underflow’ springs and short lived, highly variable flow from higher level springs, with uppermost systems having been abandoned. Worthington (1991) suggested that the simplest method of differentiating spring types is in the ratio of maximum to minimum annual discharge. Therefore, it might be anticipated that the springs could be divided between the underflow and overflow categories relatively easily, as the ephemeral springs (e.g. Deep Dale, Lathkill Head Cave and Cales Dale Upper and Lower Springs) would automatically be classified as overflow springs. However the concept is more tenuous, as such an argument can only be applied to unconfined springs and within the area of this investigation it is considered that some of the springs are fed by confined groundwater. Nevertheless, there are grounds for separating the overflow springs that are perched by the Upper and Lower Miller’s Dale Lava in the Monsal Dale and Miller’s Dale Limestone respectively. Many of the perched springs are ephemeral, although some, such as Illy Willy Water (SK 11507034), have a large enough catchment to provide a continuous supply of water (Appendix 6.2).

Consideration of the chemistry of the discharge forms the main focus of this chapter and attention has been given to a number of ways of classifying the discharge chemistries. In particular, the approach of Drake and Harmon (1973), which differentiated between phreatic-fed diffuse springs and vadose conduit-fed springs of carbonate waters from the Pennsylvania area on the basis of SIc and P_{CO_2} has been identified as a useful starting point, although the findings of a similar analysis carried out on geochemical data for the springs of the Wye catchment proved to be very different (section 6.5).

6.3.3 Flow paths within the aquifer.

Permeability in karst aquifers is achieved via a dendritic network of conduits fed by fractures exhibiting wide ranges in dissolutional effects. The potential for geological guidance of flow paths was considered in Chapter 2. Where there is geological guidance there is also a potential for a geochemical influence on the karst groundwater. In particular it is considered that the following could influence the groundwater chemistry: the presence of lavas, mineralization and dolomitization (Christopher, 1981 and section 6.2), the presence of chert (section 6.3.1) and of clay wayboards. The majority of clay wayboards are characterised by the occurrence of kaolinite with anatase (titanium oxide) and by a relative absence of illite/smectite layering (Walkden, 1970). In the submerged, basinal environment the volcanic dust was mixed with terrigenous inputs and the resultant wayboard shales (Walkden, 1970; Chapter 2) are characterised by an absence of anatase, a strong quartz peak and variable illite/smectite content. Walkden (1970) suggested that the illite is an alteration product of potassium-rich bentonite clays. The potassium content of weathered lava (clay) is greater than that of the fresh lava (Walkden, 1970), indicating that potassium has been introduced, for example by autometomorphism of the clay during cooling and/or hydrothermal alteration of the clay. These differences suggest that clay wayboards could release potassium, or anatase to groundwater, but are unlikely to provide any other distinctive tracer. Mudstone precursors to the shales have the potential to release potassium, aluminum, calcium and sodium. Clearly this is potentially difficult to distinguish from the influence of the superficial deposits (Table 6.5). Christopher (1977) suggested that, because potassium is readily adsorbed by clay minerals as a result of cation exchange, the potassium/sodium ratio could potentially be used as an indicator of age, as the ratio will decrease with time. This suggests that during low groundwater conditions scavenging of potassium from groundwater could occur. Examination of Table 6.5 indicates that the influence of the chemistry of the superficial deposits has the potential to mask such an effect.

Stability relationships for H_2O, CaO and SiO_2; H_2O, Na_2O and SiO_2; and H_2O, K_2O and SiO_2 determined by Christopher (1981) identified stability with respect to kaolinite ($Al_4Si_4O_{10}(OH)_8$), rather than with illite or montmorillonite, which would be the other clay minerals to be derived from weathering of the clay wayboards. In accordance with this, Edmunds (1971) identified a considerable proportion of kaolinite on filter residues of water samples taken from lava-affected sites. Similar observations were made by Christopher (1981) of the suspended matter in Magpie Sough. This tends to suggest a predominance of dissolution close to ground surface. The thermal waters were found to be saturated with respect to quartz, but undersaturated with respect to hydrosilic acid (H_4SiO_4) (Christopher, 1980).

Flow paths are guided by hydraulic controls such as hydraulic head and degree of confinement and these influences will also affect the geochemistry of the springs, largely as a consequence of residence time. Another significant influence on geochemistry is the degree of interaction between the flow paths. More specific observations in this context include the following: Appelo and Postma (2005) noted the importance of calcite dissolution caused by the mixing of groundwaters with differing carbon dioxide concentrations; Gunn (2003) observed that the significance of divergent flow in karst aquifers, as identified from dye tracing tests (Chapter 7) is often overlooked, and Thrailkill (1968) suggested that in response to the downstream decrease in head karst groundwater will flow approximately parallel to the stream bed beneath its bed and banks. The stress relief associated with valley formation encourages such movement. Others have made similar observations and it is considered likely that there are similar zones adjacent to the River Wye and reference to such bodies of water is made in mining records, e.g. Millclose Mine (Kirkham, 1968). Such zones potentially would facilitate mixing and diffusion and in the context of the White Peak appear to be associated with dominant inception horizons. However, the variation in spring chemistries indicates that the connecting fractures are of considerably lower permeability than the conduits, such that the conduit water retains a characteristic chemistry.

Worthington (1991) and Worthington and Ford (1995b) working in Alberta and also considering the springs of the Peak District, England, identified that the water derived from thermal springs comprised two components, a low-sulphate component and a sulphate-rich component, which was between 5 and

31˚ C warmer, subsequently attributed to evaporite dissolution (Gunn et al., 2006, section 6.4, p. 113) on the basis of the sulphur isotope ratios. The higher temperature, sulphate-rich waters were attributed to deep flow paths and the low sulphate water to higher flow paths discharging to the same spring. In accordance with this, the lower flow component was described as underflow and the upper flow component as overflow. Higher contents of syn-depositional evaporites have been postulated for the Woo Dale Limestone, because of their depositional environment (Schofield and Adams, 1985). The conceptual model of regional flow paths presented in Chapters 5 and 9 is one attributed to the presence of evaporites in interblock basins that are prone to speleogenesis. The increased sulphate content with depth of flow path appears to be common to many limestone aquifers. This could be seen as an indicator of downward development of flow paths; however there could be a chemical or thermodynamic explanation for the observation, which may facilitate upward development of karst initiated by basinal dewatering (Chapter 4). This author has calculated solubility products for calcite, anhydrite and gypsum for temperatures of 10, 25 and 40˚ C. The results are tabulated below.

Table 6.6: Solubility products calculated for calcite, anhydrite and gypsum.

Temperature (˚C)		10	25	40
Temperature (˚K)		283.15	298.15	313.15
Mineral:	**Solubility Product (K_{sp})*:**			
Calcite		$10^{-8.35}$	$10^{-8.46}$	$10^{-8.56}$
Anhydrite		$10^{-4.29}$	$10^{-4.44}$	$10^{-4.59}$
Gypsum		$10^{-4.61}$	$10^{-4.62}$	$10^{-4.63}$

* Calculated from: $\ln K = \frac{\Delta H_R - T\Delta S_R}{-RT}$

Where: H_R is enthalpy, S_R is entropy (taken from Kehew, 2001) and T is temperature ˚K
R is the universal gas constant 8.3143×10^{-3}

Reflecting the exothermic nature of the reaction, all of the minerals exhibit lower solubility with increased temperature in the range presented above. Anhydrite is consistently more soluble and gypsum is relatively more soluble with increased temperature. Worthington (1991) points out that there is likely to be reduced viscosity with depth, but this alone does not explain the sulphate enrichment. Other considerations include the tightening of fissures with depth and the effect of pressure on the solubility product of the minerals, about which there is little published, but would be represented by a modification to R in the equation presented above. Pyrite derived sulphate is considered in the context of the local flow paths described in section 6.5.

Clearly the thermal springs can be classified as underflow springs; however it is considered that some of the intermediate flow paths that emanate from springs could also be classified as underflow springs and it would appear (from the work of Worthington (1991) and Worthington and Ford (1995b)) that these should be identifiable from ranking the sulphate content (section 6.5). To try to rank the full range of flow paths from underflow to overflow this author has investigated a number of other chemical parameters. Temperature is considered to be the most reliable indicator of flow path depth (Chapters 5 and 10). Correlation data presented in Table 6.7 indicates that strontium is a good indicator of underflow springs. Table 6.7 indicates very strong correlation between strontium and sulphate (indicating that the strontium could be occurring as celestine, associated with the syndepositional gypsum). Correlation of strontium with lithium is also evident in Table 6.7. Although no single parameter can be taken as an indicator of underflow springs, the parameters that are considered to be the most likely indicators of longer residence time at depth are strontium, sulphate, lithium and magnesium. The absence of a single marker is attributed to the fact that each spring comprises groundwater derived from a number of differing flow paths.

Accordingly, it is concluded that detailed spring by spring analysis, which considers geological setting and the results of dye tracing (Chapter 7) as well as the groundwater chemistry is required in the assessment of intermediate spring flow paths (section 6.5).

Table 6.7: Correlation of cations, anions, SIc, SId and Pco_2 with temperature and each other (32 locations, data from Edmunds, 1971).

	Temp	Ca	Mg	Sr	Na	K	Li	Zn	HCO_3	SO_4	Cl	F	PCO_2	Sic	Sid
Total H															
Temp	1														
Ca	0.55	1													
Mg	0.57	0.39	1												
Sr	0.72	0.66	0.58	1											
Na	-0.14	0.25	0.58	0.13	1										
K	0.46	0.29	0.48	0.24	0.33	1									
Li	0.67	0.65	0.76	0.8	0.38	0.54	1								
Zn	-0.17	0.1	-0.15	-0.06	-0.006	0.62	-0.05	1							
HCO_3	0.26	0.48	-0.03	-0.16	-0.14	0.16	-0.1	0.24	1						
SO_4	0.67	0.68	0.67	0.95	0.27	0.38	0.92	-0.04	-0.19	1					
Cl	-0.19	0.22	0.54	0.08	0.99	0.29	0.31	-0.003	-0.14	0.22	1				
F	0.15	0.33	0.63	0.44	0.69	0.42	0.5	0.7	-0.21	0.53	0.67	1			
PCO_2	0.08	0.26	-0.11	-0.06	-0.15	0.21	0.02	0.14	0.45	-0.03	-0.16	-0.33	1		
Sic	-0.13	-0.02	-0.07	-0.1	0.14	-0.26	-0.19	0.007	0.001	-0.14	0.16	0.21	-0.73	1	
Sid	0.13	-0.03	0.42	0.12	0.33	-0.08	0.08	-0.16	-0.09	0.09	0.33	0.44	-0.74	0.79	1

6.3.4 Seasonality.

Edmunds (1971) pointed to the need to consider seasonality and argued that maximum evapotranspiration associated with the summer months concentrates constituents entering the ground to as much as twice the concentration measured in the atmospheric precipitation. In addition, Edmunds (1971) suggested that the overall concentration of dissolved solids derived from atmospheric precipitation may be greater in infiltrating waters in areas of lower rainfall, in the southeastern part of the limestone recharge area. Arguably the available data are not refined enough to show such regional variation.

Shuster and White (1971) used seasonality as an indicator of flow type. More specifically they related the variability of total hardness (expressed as a percentage coefficient of variation i.e. standard deviation/mean) to the type of resurgence. Shuster and White (1971) considered that springs with a variability of greater than 10 % represented conduit flow and those with less than 5 % diffuse, or percolation flow (dispersed flow). Shuster and White (1971) found that variation in carbon dioxide pressures were more related to source areas. Seasonality was also considered by Christopher (1981), who suggested that different defining coefficients of variation might be applicable in other catchments. Deriving comparable (with Shuster and White, 1971) coefficients of variation with respect to total hardness, Jacobson and Langmuir (1974) concluded that discharge was a more important influence on water chemistry than season; particularly for dispersed recharge type springs.

Christopher (1981) corrected his data for random error and looked at the standard error of the mean relative to an absolute standard (taken as Cowdale [Rockhead] Spring). The data matrix that was generated was analysed using cluster analysis. The classifications that resulted did not concur with those derived from the analysis of the cation and anion data, with only four classes resulting (surface, allogenic mixed, intermediate and percolation) (Appendix 6.6). Christopher (1981) ranked the variability, with the following result: temperature variability > potassium > total hardness > bicarbonate > sodium > calcium > magnesium > sulphate > nitrate > silica > chloride. Christopher (1981, section 4.6) suggested that this "*illustrates the non-equilibrium conditions over the range of the results towards calcium carbonate and highlights the anomalous position of potassium*". The findings appear to reflect the fact that within the context of the White Peak the superficial deposits and epikarst significantly influence the spring discharge waters. This is considered further in the spring-by-spring analysis (Appendix 6.2). The deeper groundwater contributions are characterised by higher strontium, lithium, sulphate and magnesium. Dispersed distal recharge (fracture/matrix), if isolated, is characterised by a greater degree of equilibrium with the host geology than proximal and rapid throughputs. Deeper groundwater and dispersed recharge waters are subject to dilution by proximal recharge. Accordingly, seasonality is more likely to reflect flow-through time (as indicated by the saturation index with respect to calcite), relative dilution of the deeper groundwater contribution (e.g. reduced sulphate), and fluctuations in the anions and cations derived from, or concentrated in the superficial deposits, in particular potassium, sodium, chloride, nitrate and also sulphate.

6.4 Regional considerations.

Downing (1967) identified regional variations in the groundwater chemistry to the east of the Derbyshire Dome. Close to outcrop the dominant groundwater type comprises the calcium bicarbonate type, with high sulphate contents being attributed to pyrite, or to small amounts of sulphate in the limestone and the anions being balanced by alkaline earth cations. To the east, down dip, the concentration of sodium, potassium, calcium, magnesium and chloride all increase and the groundwater type grades from calcium bicarbonate to calcium sulphate and then chloride. Downing (1967) suggested that the increased sulphate content of the calcium sulphate groundwaters, appears to be associated with an increase in the concentration of calcium (which would be compatible with gypsum, or anhydrite dissolution). The chloride waters are associated with an increase in the sodium content of the groundwater. Probably reflecting the regional groundwater chemistry, broad increases in the calcium, chloride and sulphate concentration of the springs appear to occur at greater stratigraphical depth. This variation is also of interest in the context of the mineralization in the area. If it is accepted that the hydrothermal mineralization was derived from connate water, largely driven out of the adjacent Silesian strata (Chapter 3), then a strong association might be anticipated between mineralization (including the dolomitization of the Woo Dale Limestone) and more saline water, as shown by fluid inclusion studies reported by Ewbank et al. (1993). It has been observed that higher lead concentrations are associated with more chloride-rich groundwater derived from the Woo Dale Limestone (Appendix 6.3). Saunders and Toran (1994) have suggested that upward diffusion of sodium and chloride derived from brines can occur. There seems no reason why such processes should not be considered in the context of the Derbyshire Dome.

Table 6.8: Springs and soughs that have been classified as thermal (data from Edmunds, 1971).

Spring	National Grid Reference	Published Temperature (˚C)	Spring	National Grid Reference	Published Temperature (˚C)
Bakewell British Legion	SK 218686	11.6	Matlock Fountain Bath	SK 294584	19.7
Bakewell Recreation Ground	SK 220681	13.3	Matlock New Bath Hotel	SK 293579	19.8
Ball Eye Quarry Borehole	SK 289573	13.6	Lees Bottom 3 (Lower Dimindale)	SK 180696	14.3
Beresford Spring	SK 128586	13.8	Meerbrook Sough 1	SK 327552	17.0
Bradwell Spring	SK 174820	12.4	Meerbrook Sough 2		17.0
Buxton (St Anne's Well)	SK 057735	27.5	Ridgeway Sough (Crich)	SK 332549	14.1
Magpie Sough	SK 180696	14.3	Stoke Sough	SK 240764	11.6
Matlock East Bank Rising	SK 294582	17.4	Stoney Middleton	SK 232756	17.7

The significance of the thermal springs in the context of regional flow paths was described in Chapter 5. A thermal spring is defined as one where the temperature of the water is above the average temperature of the superficial rock (EPA, 1999). A mean air temperature of 7.6˚ C for Buxton for the period 1976 to 1980 was calculated from daily values provided by Peak District Borough Council. The average temperature of the superficial rock has not been determined.

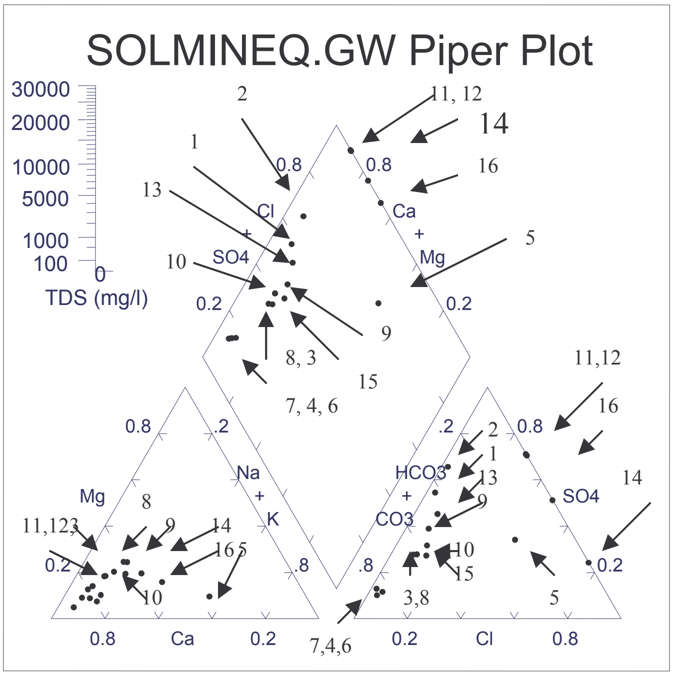
Based on determinations of tritium, Edmunds (1971) suggested that the thermal waters were at least 15 to 20 years old. Pentecost (1999) has observed that comparison of historic analyses of the thermal spring waters indicates that essentially there has been no significant change in their chemistry during the history of their study, although this author has observed that more recent analyses such as those presented by Gunn et al. (2006) do indicate that there have been increases in nitrate concentrations, which are probably attributable to changes in farming practices and are likely to represent contributions from the vadose zone. For example, the nitrate determinations for Matlock East Bank Rising and Matlock New Bath Hotel were 0.8 and 0.4 mg/l respectively (Edmunds, 1971) and 3.1 and 2.8 mg/l respectively (Gunn et al., 2006). However, there has been no measurable increase in the nitrate content determined for St Anne's Well, Buxton.

Edmunds (1971) found that, relative to other waters, the thermal waters were enriched with most constituents except nitrate, but with regional differences in trace element chemistry from different thermal centres. In particular Edmunds (1971) stated that Buxton thermal waters have similar hydrogeochemistry to the waters from non-mineralized areas of the limestone, but with relatively enriched manganese and arsenic concentrations, suggesting to Edmunds (1971) derivation of at least some of the water from the Millstone Grit Group. More recent research by Gunn et al. (2006) supports this suggestion. The concept of different chemistries associated with different thermal centres provides support for the concept of differing flow paths (Chapter 5). The enrichment can be attributed to a number of factors including the increase in solubility with depth due to the rise in hydrostatic pressure (section 6.3.3 and Thrailkill, 1968) and increased ionization. To some extent the evolution in the groundwater chemistry is masked because each spring is fed by a number of differing flow paths.

The cluster analyses carried out by Christopher (1981) also point to groups of thermal springs. The groups identified by Christopher (1981) include: the Matlock/Buxton Group, the Bakewell Group, with the Bradwell and Stoney Middleton thermals being separate sites of distinct composition. This is similar to Edmunds (1971), although he also classified Meerbrook Sough, Lees Bottom 3 (Lower Dimindale) and Beresford as thermals. Gunn et al. (2006) subdivided the thermal springs into those of the Buxton type and those of the Matlock type. However, their investigation did not include the Bakewell thermal waters. The Buxton type is thought to be derived from deep sandstone aquifers in the Silesian deposits to the west of the limestone outcrop and those of the Matlock type are seen to be derived from within the limestone. They exhibit elevated sulphate concentrations, which the results of isotopic analysis suggest to be derived from interaction with buried evaporites in the Lower Carboniferous succession (Gunn et al., 2006). This author has used the groundwater analytical package SOLMINEQ.GW to plot the analyses presented by Edmunds (1971) on a Piper plot (Figure 6.2). Table 6.9 shows that, using the convention of Freeze and Cherry (1979) the springs fall within six clusters that are apparently independent of the geology at egress.

Pentecost (1999) noted that the Matlock Springs contained nitrate but no nitrite, that dissolved oxygen was found in all waters and that the concentration of carbon dioxide was equivalent to atmospheric equilibrium partial pressures typical of soil atmospheres. This author has used the SOLMINEQ.GW package to determine the partial pressures of carbon dioxide that are indicated by the published analyses for the thermal springs (Table 6.10). For springs with elevated nitrate concentrations it is likely that higher concentrations of Pco_2 indicate mixing with water derived from higher flow paths, with the P_{CO_2} being carried down from the topsoil-rooting zone. This suggests that the higher the P_{CO_2} the less reliable is temperature as an indicator of flow path depth. Higher concentrations of carbon dioxide are associated with the Matlock springs, Lees Bottom 3, Beresford Spring and with Bradwell Spring, indicating that these springs receive a greater contribution of groundwater from higher-level flow paths. However, it is interesting to note that St Anne's Well, Buxton shows the highest calculated concentration of carbon dioxide. This corresponds with lower ionic strength and with <0.05 mg/l of nitrate, suggesting the presence of a deeper source of carbon dioxide (subsection 6.3.1).

This author has also given consideration to whether or not the chemistry of the thermal springs provides any evidence for the process of gypsum-driven dedolomitization (Bischoff et al., 1994 and Chapter 4), which would be another process to explain the isotopic evidence for sulphate being derived from buried evaporites (Gunn et al., 2006).



Key: 1 Bakewell British Legion; 2 Bakewell Recreation Ground; 3 Ball Eye Quarry; 4 Beresford Spring; 5 Bradwell Thermal; 6 Lees Bottom 3; 7 Magpie Sough; 8 Matlock East Bank Rising; 9 Matlock Fountain Bath; 10 Matlock New Bath Hotel; 11 Meerbrook Sough - 1; 12 Meerbrook Sough – 2; 13 Ridgeway Sough; 14 St Anne's, Buxton; 15 Stoke Sough; 16 Stoney Middleton.

Figure 6.2: SOLMINEQ. GW Piper plot for the thermal waters (data from Edmunds, 1971).

Table 6.9: Classification of thermal springs (data from Edmunds, 1971).

Classification of thermal springs based on groundwater type:	Representative Springs	Geology at egress (see also Table 5.1 and Figure 5.2)
Calcium bicarbonate	Matlock East Bank Rising	Alluvium over Monsal Dale Limestone
	Matlock New Bath Hotel	Alluvium over Monsal Dale Limestone
	Matlock Fountain	Mineralized fault in Monsal Dale Limestone
	Ball Eye Quarry	Southeasterly extension of the Cronkston-Bonsall Fault in the dolomitized Monsal Dale Limestone Formation
	Stoke Sough	Silesian strata
	Lees Bottom 3	Monsal Dale Limestone
	Beresford Spring	Apron reef in Bee Low Limestone Formation.
	Magpie Sough	Monsal Dale Limestone
Calcium sulphate/Calcium bicarbonate	Ridgeway Sough, Crich	Woo Dale Limestone
Calcium chloride	St Anne's Well, Buxton	Alluvium over Monsal Dale Limestone
Calcium sulphate/ Calcium chloride	Stoney Middleton	Southern end of knoll reefs in Eyam Limestone
Calcium sulphate	Meerbrook Sough 1	Woo Dale Limestone
	Meerbrook Sough 2	Woo Dale Limestone
	Bakewell British Legion	Eyam Limestone/Silesian
	Bakewell Recreation Ground	Eyam Limestone/Silesian
Sodium chloride	Bradwell Spring	Apron reef inlier in Millstone Grit Series

Table 6.10 Concentrations of carbon dioxide determined for the thermal waters (data from Edmunds, 1971).

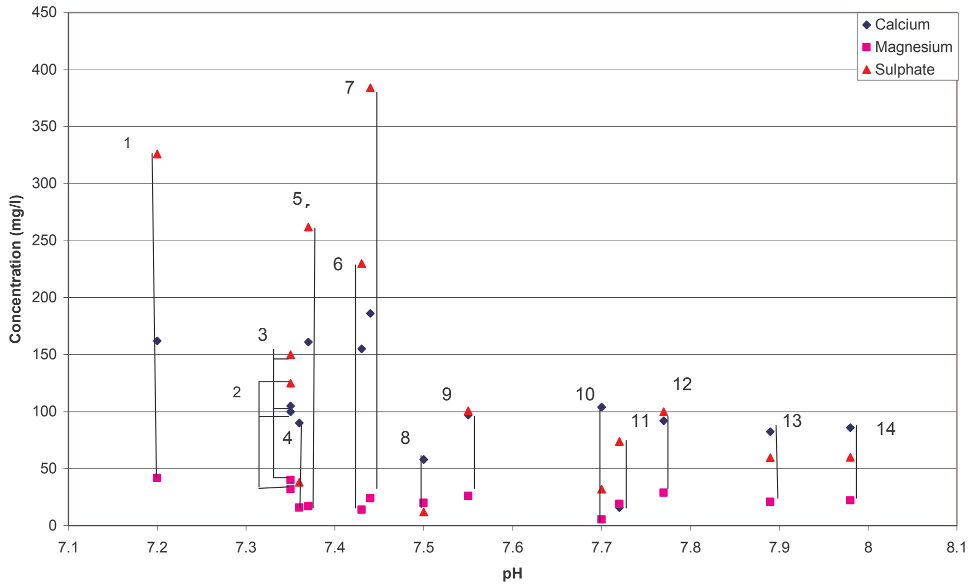
Spring/ Groundwater	Partial pressure of carbon dioxide (bars)	Carbon dioxide (%)	Spring/ Groundwater	Partial pressure of carbon dioxide (bars)	Carbon dioxide (%)
Bakewell British Legion	0.0077	0.77	Matlock Fountain Bath	0.0043	0.43
Bakewell Recreation Ground	0.0060	0.60	Matlock New Bath Hotel	0.0107	1.07
Ball Eye Quarry Borehole	0.0063	0.63	Matlock East Bank Rising	0.0110	1.10
Beresford Spring	0.0170	1.70	Meerbrook Sough 1	0.0020	0.20
Bradwell Spring	0.0118	1.18	Meerbrook Sough 2	0.0029	0.29
Buxton (St Anne's Well)	0.0206	2.06	Ridgeway Sough (Crich)	0.0074	0.73
Lees Bottom 3	0.0106	1.06	Stoke Sough	0.0026	0.26
Magpie Sough	0.0049	0.49	Stoney Middleton	0.0059	0.59

Saunders and Toran (1994) identified three lines of evidence for gypsum-driven dissolution, namely textural evidence, decreasing pH with increasing sulphate content of groundwater and increase in the magnesium content of groundwater with increasing sulphate content. Textural evidence requires rock sampling, but Figure 6.3, which is a plot of calcium, magnesium and sulphate against pH, indicates that, potentially, this reaction may be associated with the Matlock springs and Bradwell Spring, where sulphate and magnesium contents are higher and pH lower than in the other thermal springs. The elevated calcium and sulphate concentrations determined in the Bakewell springs and Ridgeway Sough would suggest gypsum or anhydrite dissolution, or an association with mineralization.

It is also interesting to note that the chemistry of Stoney Middleton Spring, with marginally elevated manganese and arsenic, as determined by Edmunds (1971), suggests that it receives a contribution from Namurian strata (as at St Anne's Well, Buxton), probably locally from the adjacent outliers of Namurian strata, or possibly from more distal stream sinks associated with the boundary of the Dinantian outcrop. The elevated silica content of the Bakewell Recreation Ground Spring was considered in Chapter 5.

6.5 Classification of non-thermal spring waters.

In order to characterise flow paths within the aquifer it was considered necessary to first identify the best method of classification for the springs. Clearly, if a classification could be related directly to geological formation then this would provide the simplest means of interpretation. However, this has proved difficult for two principal reasons, firstly because of the mixing of groundwater from different sources and secondly because the major component of dissolution occurs within the vadose zone. Edmunds's (1971) classification was based on the geology at resurgence and he suggested that four main rock groups have a potential influence on groundwater chemistry, namely: pure limestone, basic igneous rocks, lead-zinc-barite-fluorite mineralization and Millstone Grit. Worthington (1991) suggested that the simplest method of differentiating spring types is the ratio of maximum to minimum annual discharge. Such an argument can only be applied to unconfined springs and within the area of this investigation it is considered that some of the springs, including a number of the Lees Bottom springs, are fed by groundwater that is confined.



Key: 1.Bradwell Thermal; 2 Matlock East Bank Rising; 3 Matlock New Bath Hotel; 4 Lees Bottom 3; 5 Bakewell British Legion; 6 Ridgeway Sough; 7 Bakewell Recreation Ground; 8 St Anne's Well, Buxton; 9 Ball Eye Quarry; 10 Magpie Sough; 11 Stoke Sough; 12 Stoney Middleton; 13 and 14 Meerbrook Sough 1 and 2.

Figure 6.3: Plot of calcium, magnesium and sulphate against pH for the thermal springs (data from Edmunds, 1971).

Following his earlier classification, described in section 6.2, Christopher (1981) presented a more refined classification of the waters using statistical analysis (Appendix 6.6). It divided the waters into four strong hierarchical groups: general limestone, grit/shale, dolomitic and thermal. At higher levels of clustering the groups were subdivided further. The general limestone group was subdivided into two dominant clusters: one was characterized by high potassium and the other by low magnesium and low sodium. Whilst this may be a mathematically more robust classification it only provides a broad indication of dominant flow paths in the aquifer.

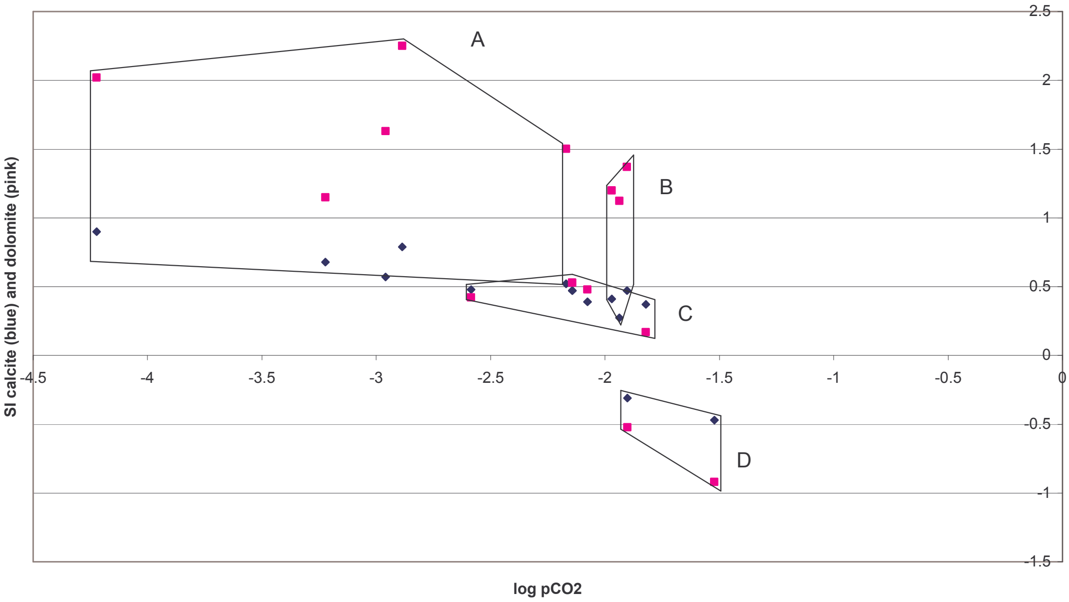
It has already been stressed that one of the problems with finding a means of classifying the springs on the basis of geological formation is the fact that the majority of springs are fed by water infiltrating more than one stratum. Furthermore, where water is stored there is also a potential for groundwater mixing and interaction with the bedrock in the storage zone. This author attempted to derive a classification of the Edmunds (1971) data based on the work of Drake and Harmon (1973). Spring chemistry data (Edmund, 1971) pertinent to the area of this research, were entered into the computer program SOLMINEQ.GW. Based on the input of temperature, pH and concentrations of the cations and anions the SOLMINEQ.GW program converts concentrations to milliequivalent values and calculates: the ionic ratio, the theoretical CO_2 pressure (P_{CO_2}) and saturation indices of the water with respect to a number of minerals, including dolomite (SId) and calcite (SIc). The method of calculation of these parameters is presented in Appendix 6.4. The geology at the point of egress was established from published geological data and field observations. Table 6.11 presents the springs that were selected for classification.

Table 6.11: Springs associated with specific formations (interpreted from the data of Edmunds, 1971).

Formation	**Representative Springs, with Limestone Research Group names in brackets and National Grid References** (six figure references presented by Edmunds, 1971)
Woo Dale Limestone	Lower Dimindale (Lees Bottom 3): SK17147060; Woo Dale (Woolow): SK 09477242; Pictor: SK 09277251
Bee Low Limestone	Nether Low: SK 111692; Litton Mill – 2 (Litton Mill): SK 161729; Deep Dale (Deep Dale Main Resurgence): SK 097714; Staden Farm: SK 074 723
Monsal Dale Limestone	Dirtlow Farm 1: SK 188686; Dirtlow Farm 2: SK 193685; Millers Dale 1 (Miller's Dale 4): SK 137731; Millers Dale 2 (Miller's Dale 5) SK 146734;Taddington High Well: SK 144708
Namurian	Underhill Well: SK 088664; Underhill Farm: SK 093661

Having considered the potential significance of dolomitization in speleogenetic processes (Chapter 4), this author was particularly interested in the saturation index with respect to dolomite. It was found that a broad classification could be derived based on the calculated saturation indexes of calcite and dolomite (SIc and SId) plotted against the calculated partial pressure of carbon dioxide (Figure 6.4). Of particular interest are the low values of calculated log P_{CO_2} that are achieved in the Monsal Dale Limestone Formation. One might anticipate that low concentrations of carbon dioxide would be associated with closed systems associated with deeper flow paths, or even underflow springs. However, in this case a greater degree of closure appears to be associated with the strata encountered at higher stratigraphical levels. These observations support the early findings of Christopher (1981), falling within the closed percolation fed resurgences group (section 6.2). Generally the low values observed in the Monsal Dale Limestone Formation are attributed by this author to the predominance of dispersed flow and active dissolution beneath the zone of carbon dioxide replenishment, possibly reflecting the dominance of flow along clay wayboards and inception horizons (Chapter 4 and considered further in Chapters 8, 9 and 11). The lowest values that were determined have already been described and have been attributed to additional processes (section 6.3). It has also been noted that the saturation indices with respect to dolomite that are achieved in the Woo Dale Limestone Formation are considerably higher than those achieved in the Bee Low Limestone Formation and the open flow of the springs in the Namurian deposits is, as would be expected, associated with under-saturation with respect to both calcite and dolomite.

Very high levels of saturation with respect to both calcium and dolomite are achieved in the Monsal Dale Limestone Formation. Supersaturation with respect to calcite has been observed elsewhere, although this may in part be attributable to the selection of solubility equilibria in the calculation of the saturation index (Langmuir, 1971, Appendix 6.4). Furthermore, the distribution of tufa, as indicated both by examination of geological sheets and by field observations is largely associated with the outcrop of the Monsal Dale Limestone, which also supports the concept of super-saturation being achieved in the Monsal Dale Limestone. Additionally, Smith's (2000) results indicate lower dissolved oxygen contents in the springs associated with the Monsal Dale Limestone Formation. Many of the springs in the Monsal Dale Limestone Formation are overflow springs and are unlikely to receive any contribution from rising thermal water.



Key

A	Monsal Dale Limestone (inception horizon-guided flow, with poorer developed fissuring)
B	Woo Dale Limestone (stylolites and fissure-dominated flow)
C	Bee Low Limestone (fissure-dominated flow)
D	Surface streams on the Silesian mudstones

Figure 6.4: Plot of saturation index (calcite blue and dolomite pink) against log Pco_2 of selected springs (data from Edmunds, 1971).

Figure 6.4 indicates that there are lower values of SIc associated with the Bee Low Limestone Formation. This is also supported by the Smith (2000) data (Table 6.2). As clearly described in the literature (Palmer, 2002 and Thrailkill, 1968), low values of saturation can be brought about in a number of ways, for example mixing saturated waters of differing P_{CO_2}, limestone resistance to dissolution, or by rapid through-flow times that do not provide sufficient time for rock/groundwater equilibration. Whilst it is plausible that the P_{CO_2} determinations are indicative of the form of differing flow paths in the limestones, the concentrations of cations and anions are considered more likely to reflect the sources of water and because the majority of dissolution occurs within the vadose zone, the geology at the predominant source of the groundwater. Accordingly, in the Bee Low Limestone Formation it is considered likely that the low values of saturation with respect to calcite and dolomite, are brought about by the high permeability of the vadose zone (as seen in soakage and dye tracing tests, Chapter 7) resulting from the closely-spaced fissuring observed in the Bee Low Limestone Formation with consequential reduction in dissolution at rock head; by sediment infill of fissures which minimises surface area contact of the limestone with groundwater; and also by the massiveness, purity and high sparite content, resulting in a lower solubility of the Bee Low Limestone Formation. The purity is attributed to the condensed nature of the recession component of the depositional cycles in the Bee Low Limestone and relative absence of detrital material (Walkden, 1987). The range between the mean minimum and mean maximum concentration of P_{CO_2} is lower in the Bee Low Limestone (Table 6.2) and the minimum values are generally higher than in the other geological formations. This supports the concept of there being a source of carbon dioxide that extends into the aquifer and is thought to be the aerobic and anaerobic biogenic degradation of the organic component of the cover deposits that overlie and fill fissures that penetrate the Bee Low Limestone more deeply than other formations.

The Woo Dale Limestone is also very pure (Harrison and Adlam, 1985), yet high saturation indices with respect to both calcite and dolomite are achieved in the Woo Dale Limestone (Figure 6.4, Table 6.2). However, P_{CO_2} concentrations do not get as low as those of the Monsal Dale Limestone, suggesting a more open system (the intermediate springs of Christopher's (1981) early classification). Where the Woo Dale Limestone Formation is exposed in a cutting in Ashwood Dale, it is moderately thinly bedded (beds apparently as thin as 10 cm) and although the bedding is somewhat uneven it has clearly been subject to stress relief and is open. There is a close association between faults and springs in the Woo Dale Limestone. The quarry manager at Topley Pike (Mr S Jones, personal communication, 2004) has described how surface water and groundwater accumulating on the quarry floor appears to drain to an approximately east to west trending, partially mineralized, fault on the northern perimeter of the quarry. It has been speculated (Chapter 3) that dominant fissures adopt the function of dolines in this setting. The findings presented in Figure 6.4 for the Woo Dale Limestone indicate that a higher saturation with respect to dolomite is achieved than in the Bee Low Limestone, this may reflect the contribution of deeper groundwater that has been in contact with dolomitized areas of the Woo Dale Limestone Formation.

In order to reflect the significance of the influence of the epikarst on groundwater chemistry, Table 6.2 presents Smith's (2000) data in groups selected on the basis of: i) suspected dominant source geology (influence of dissolution in the vadose zone) and ii) the likely source of groundwater. Thus, it might be considered a conceptual classification. Seasonality of the data is apparent, however flaws in the data have been identified (section 6.3.1) and with only four data points per spring, statistical verification of the data would not be valid. A spring by spring hydrogeochemical description is presented as Appendix 6.2. In the consideration of the data presented in Table 6.2 it has been noted that although there is some overlap, the parameters that show the greatest degree of variation between the groups are: SIc and SId, electrolytic conductivity, total hardness, magnesium, sodium, potassium, chloride and sulphate. Total hardness determinations were generally higher than average determinations for limestone in temperate regions (210 mg/l, Smith and Atkinson, 1976). It was only in the samples obtained from Litton Spring that there were any determinations below this value. The lower total hardness determined for the Bee Low Limestone Formation is in keeping with the lower values of SIc and SId and also with the lower electrolytic conductivity determinations. The highest total hardness concentrations were determined for Lees Bottom 3 Spring, which is not surprising given the likely depth and length of the flow paths associated with this spring. Particularly high electrolytic conductivity values have been determined for Kidtor Spring. In this case, the evidence of elevated nitrate concentrations suggests that this is, at least in part, attributable to contamination. Generally, however, it would appear that the high total dissolved solids content determined in the Woo Dale Limestone is associated with higher ionization, associated with higher concentrations of sodium and chloride, which may be associated with the mineralization in the Woo Dale Limestone Formation.

The potential subdivision of the springs with recharge from the Monsal Dale Limestone Formation, between overflow springs and those that receive a contribution of underflow (Lees Bottom 1 Spring), was suggested above in the light of the findings of Christopher's (1981) work. The results presented in Table 6.2 appear to support this hypothesis. It should also be noted that the influence of the lava appears to be evident in the chemistries of Great Shacklow 1 and 2 springs, which are associated with the Shacklow Wood Lava Member of the Monsal Dale Limestone (Table 6.2).

With respect to magnesium, most noticeable is the high concentration determined in Lees Bottom 3 (Table 6.2), attributed to contact with dolomitized areas of the Woo Dale Limestone Formation. Lees Bottom 3 has been classified by this author as a thermal spring, with an elevated temperature (11.5˚ C). The relatively elevated concentrations of magnesium determined in Topley Pike, Woolow and Wormhill springs are also attributed to contributions from the dolomitized Woo Dale Limestone. Shuster and White (1971) found that the calcium/magnesium ratio of groundwaters derived from calcium carbonate rocks is significantly higher (in the order of 3 to 8 times higher) than that from dolomitic limestones. Dolomitization of the Woo Dale Limestone, which occurs at depth, is of limited extent, albeit larger than the surface area exposed in the Wye Valley. Concentrations of magnesium in the Bee Low Limestone, in Litton Spring in particular, are noticeably lower. Although, to the south of this research area high magnesium determinations are associated with the dolomitization of the Miller's Dale Limestone adjacent to the Cronkston-Bonsall Fault. Marginally lower concentrations were determined in the Monsal Dale Limestone, which may partly be attributable to the processes of dolomitization and dedolomitization (Chapter 4), a process which must have resulted in the release of magnesium originally derived from the clay wayboards during the process of dedolomitization. This author has observed dolomitization associated with faulting in Ashwood Dale. This is evident at Ashwood Dale Spring, where the limestone is more brownish grey and heavily veined with calcite and it is also evident in limestone exposed in Cowdale, again associated with calcite veining (see also Appendix 4.1). As Ashwood Dale Resurgence is situated on the same fault as Cowdale [Rockhead] Spring, it would seem likely that the apparently elevated magnesium concentrations (for the Bee Low Limestone) is at least in part, attributable to dolomitization associated with faults in the Bee Low Limestone Formation.

Of particular note in the context of the chemistry of the Wormhill Springs are the elevated sulphate concentrations, some of which may be derived from the dissolution of calcium sulphate in the Woo

Dale Limestone Formation. The range of sulphate concentrations in the different spring groups overlap, but it is suggested that there are two different sources of sulphate. In the Monsal Dale and the Bee Low limestones sulphate is largely derived from the superficial deposits and is also associated with the volcanic deposits, whereas that in the Woo Dale Limestone is attributed to dissolution of calcium sulphate, other evaporites and to the oxidation of ore minerals. Sulphate derived from the superficial deposits has been attributed to fertilizers, natural organic sulphates from organisms and plant material, oxidation of pyrite and other sulphide minerals (Webber, 1995). Webber (1995) examined the isotopic composition of sulphur derived from a number of springs, which were generally to the north and southeast of the focus of this research and suggested that groundwaters that were light in the isotope $\gamma^{34}S$ had received sulphur from one of the following: oxidation of pyrite, addition of percolation water with lighter isotopic signature, or oxidation of sulphide ore minerals. Interestingly, the Wormhill Springs were among the springs monitored by Webber (1995) and they were found to be isotopically light. This author opines that there are two sources for the isotopically light signature, one is the considerable thickness of superficial deposits, observed in the valley above Wormhill Springs (Appendix 6.2), which hosts pyrite, the oxidation of which provides one source and the other is allogenic recharge from the Namurian strata. By contrast, sulphate in thermal springs, generally in significantly higher concentrations, had a higher isotopic signature attributed to dissolution of evaporites. Generally, as with the other chemical considerations, groundwater mixing makes it difficult to interpret the isotopic signature. This work has been further developed by Gunn et al. (2006). Cases of more mineralized infiltrating water than stored karstic water have been described in the literature for other areas, e.g. the Foussoubie system in France (Vervier, 1990). In the context of the White Peak, it was noted in section 6.3.1 that Raper (1989) measured higher bulk deposition rates for calcium sulphate during the winter months. Smith's (2000) results indicated higher concentrations of sulphate in both of the Wormhill Springs in February. Furthermore, these springs lie beneath a tree canopy, with a potential for calcium sulphate deposition both at the point of recharge and at the point of monitoring. It would seem feasible that the presence of remnant hydrocarbons associated with the stylolites and inception horizons described in the Monsal Dale Limestone is another source of sulphate in the Monsal Dale Limestone.

Concentrations of sodium and chloride were at a maximum in the Woo Dale Limestone Formation sourced springs. The global mean concentration of chloride in rainwater is 3.8 mg/l and of sodium is 1 mg/l (Hitchon et al., 1999). Accordingly, the values determined indicate marginal elevation (particularly if compared with the values determined for the thermal springs (section 6.4). Christopher (1981) described the problems associated with contamination by road salt. These data do not support this theory, there is little evidence of seasonal effects and where there is seasonal evidence it indicates higher concentrations during the summer months, perhaps supporting the concentrating effect of evaporation detailed by Edmunds (1971). Furthermore, the springs where the highest concentrations have been determined (Wormhill Springs) and the associated predicted flow paths are farther from roads than other springs, which are adjacent to the A6 and have lower chloride concentrations. This author opines that the elevated chloride concentrations (Kidtor, Wormhill and Lees Bottom Springs) are associated with recharge via buried doline systems (Appendix 6.2) that are filled with superficial deposits, within which chloride is concentrated by evapotranspiration. The concentrating of chloride by beech trees was noted in section 6.3.1; it is also worth noting that there is an extensive area of former estate land in King Sterndale, above Kidtor Spring, which supports a significant number of beech trees. The mineralization of the limestone was associated with sodium chloride brines (Ewebank et al., 1993), indicating a potential source associated with mineralized veins in the Woo Dale Limestone, which more specifically may influence the chemistry of the Wormhill Springs. Another potential source of chloride is an agricultural one and this is suspected where the chloride determinations correlate closely with nitrate.

The spring by spring descriptions (Appendix 6.2) were derived, in part, from correlation of the results with borehole and rainfall data. The mean discharge values that were calculated by Smith et al. (2001), are based on only four measurements and are presented for indicative purposes only. Analytical concerns with respect to pH and ionic balance were described in sections 6.3.1 and 6.2 respectively. The evidence of Appendix 6.2 suggests that in the Monsal Dale Limestone Formation there is a broad tendency for a reduction in the concentration of potassium with increased discharge. If the principle of closed, inception horizon-dominated flow is accepted then this can be seen as further evidence that potassium is largely derived from the superficial deposits (Table 6.5). Further evidence for this comes from the observation that the lowest concentration of potassium was determined in Lees Bottom 3 Spring, an underflow spring (section 6.4). Potassium concentrations appear to be marginally elevated in the Wormhill Springs; this author has observed extensive deposits of superficial materials in the suspected catchment for Wormhill Springs (Appendix 6.2).

Christopher (1981) concluded that much dissolution occurred in the epikarst. The evidence presented above, which suggests that sulphate and potassium can be derived from the superficial materials, supports these conclusions and also the conceptual classification presented in Table 6.2. It would appear that the closely-spaced fissuring in the Chee Tor Limestone Member, where it is exposed, facilitates rapid groundwater movement to depth. In the Woo Dale Limestone groundwater moves along dominant fissures, which are tighter than in the Chee Tor Limestone, thereby reducing the rate of flow (Table 9.5) and encouraging greater dissolution. Dolines formed in the Monsal Dale Limestone Formation appear to feed inception horizon related channels, which are also fed by subvertical fissure flow and along which active dissolution occurs.

6.6 Trace elements.

Edmunds (1971) made special reference to trace constituents. Although lead (up to 16 μg/l) was detected in most of the mine-drainage waters, but not in the non-mineralized areas, there was little systematic pattern to the distribution of other metals. Of the trace metals, nickel was broadly occurring, with a median concentration of 1.8 μg/l and was found in both mineralized and non-mineralized areas, thus it was speculated that the limestone itself may be the source of the nickel

(Edmunds, 1971). Zinc was determined in all water samples. The median was 54 μg/l. Zinc concentrations from the soughs and mineralized areas were generally higher than those from the limestone; however, zinc contamination from artificial sources meant that little reliance could be placed on the occurrence of any anomalies. The Limestone Research Group determined metal concentrations of spring water analysed during the Illy Willy Water dye tracing experiment (Chapter 7) and these results form the subject of Appendix 6.3.

6.7 Conclusions drawn from the analysis of selected springs.

Much of this work supports and builds upon the conclusions of the initial work carried out by Christopher (1981). Observations with respect to the hydrogeology have been used to carry out further assessment of the hydrogeochemistry. From this, the following findings are considered useful in the hydrogeological interpretation of the Wye catchment.

i) Overflow springs can only be categorically defined as such where they are ephemeral springs in the Monsal Dale Limestone.

ii) A significant number of springs rise on faults and it is considered that many discharge a proportion of confined groundwater (baseflow or underflow). Of these springs, some are intermittent and have been termed ‘seasonal underflow springs’ by this author. Lees Bottom Spring 5 (characterised by higher potassium concentrations), Great Shacklow Springs 1 and 2, White Cliff Spring and Litton Spring have all been classified as overflow springs, in the assessment of the data presented by Smith (2000), (Appendix 6.2).

iii) Whilst temperature is the best indicator of groundwater contributions from depth, strontium has been shown to be the best chemical indicator of deep flow paths. Elevated concentrations of sulphate, lithium and magnesium are also associated with deep flow paths.

iv) Fluoride is the best indicator of groundwater contact with mineralization (Bertenshaw, 1981).

v) Seasonally low P_{CO_2} determinations and high saturation indices with respect to calcite and dolomite are associated with the Monsal Dale Limestone Formation. Although the susceptibility of the springs to seasonality suggests that dissolution largely occurs in the epikarst it is considered that the low P_{CO_2} determinations and high saturation indices with respect to calcite and dolomite are characteristic of closed flow paths in the Monsal Dale Limestone in areas overlain by a limited thickness of superficial deposits. The evidence of closed flow paths support the concept of perched groundwater in the Monsal Dale Limestone Formation.

vi) Extremely low P_{CO_2} determinations are associated with the presence of chert bands in the Monsal Dale Limestone.

vii) Springs in the vicinity of Bakewell support relatively elevated concentrations of silica. Silica solubility may be increased by the presence of hydrocarbons associated with zones of microstylolites above clay wayboards in the Monsal Dale Limestone.

viii) Springs with catchments in the Bee Low Limestone appear to be characterised by lower concentrations of total dissolved solids. Many of the springs in the Bee Low Limestone Formation were identified by Christopher (1981) as closed percolation resurgences. It is suspected by this author that this reflects the significance of inception horizon related storage in the Miller's Dale Limestone Member of the Bee Low Limestone Formation and at the boundary of the Miller's Dale Limestone Member with the underlying Woo Dale Limestone Formation.

ix) Within this area of investigation springs classified as 'open percolation resurgences' by Christopher, 1981) have actually been found to be soughs. They are characterized by supersaturation with respect to calcite, with high P_{CO_2}. It is suspected by this author that, in addition to collecting water from open, dispersed, groundwater resurgences they form zones of groundwater mixing, also collecting water from: closed, dispersed groundwater associated with inception horizon-related storage; clay wayboard storage and underflow rising on faults.

x) Evapotranspiration appears to be significant in concentrating chloride in the epikarst and superficial deposits, It has been speculated that this may be influenced by vegetation, more specifically by beech trees. It is also likely that the concentrations of salts, as indicated by seasonality in the concentrations of sodium and chloride, occur towards the top of the sequence of superficial deposits, where the effects of atmospheric variation are greatest. By contrast, it is suspected that potassium is derived from deeper in the profile of the superficial deposits, where a longer soil contact time is required for leaching, thus fluctuations in the concentration of potassium do not mirror those of chloride.

xi) Although there are concerns regarding the quality of Kidtor Spring, it is worth noting that this is probably the only overflow spring that has been monitored in the Woo Dale Limestone Formation. Further monitoring of this spring may provide useful data regarding groundwater-rock interaction in this formation.

xii) There is a more stable chemistry and higher total hardness associated with the Woo Dale Limestone Formation catchment and this is likely to reflect the lesser extent of cover over the Woo Dale Limestone.

xiii) Storage in near surface exposures of the Woo Dale Limestone Formation is likely to be associated with voids formed within the bedding that results from stress relief.

xiv) It would appear to this author that underflow springs are associated with faults with a trend that is perpendicular to that of the groundwater flow path.

It is considered that the analysis of the hydrogeochemistry would benefit from temporal monitoring of temperature, strontium, lithium, sulphate and magnesium in order to try and assess the contribution from the deeper flow paths. It is also noted that there has been no chemical characterization of the groundwater in the Namurian sandstones and this information may be useful in identifying zones of groundwater recharge from the Namurian sandstones to the Dinantian limestones.

As a point to carry forward for the conceptual hydrogeological model, it is interesting to consider the distribution of overflow, baseflow and underflow springs. Cowdale (Rockhead) Spring has been classified as an underflow spring, with suspected recharge (and therefore a regional hydraulic gradient) from the southwest. Similarly, Lees Bottom 1 has been classified as an underflow spring, indicating an easterly regional hydraulic gradient at this location. The overflow springs are associated with the Monsal Dale Limestone Formation and with the area between Wormhill Springs and Lees Bottom 1 Spring, defining a stretch of the River Wye that is considered to be seasonally perched. As baseflow springs target the local groundwater system (effectively the nearest river, deepened by meltwater and when they formed, by valley glaciers) the associated flow paths are not necessarily indicative of the regional hydraulic gradient. It is also possible that the development of baseflow springs is closely related to glacial episodes (glacial meltwater), whereas the flow paths identified by underflow springs correspond more closely with the regional hydrogeology, which appears to be influenced by the distribution of the faulted basement blocks (Chapter 5).

Chapter 7: Water tracing experiments.

7.1 Introduction.

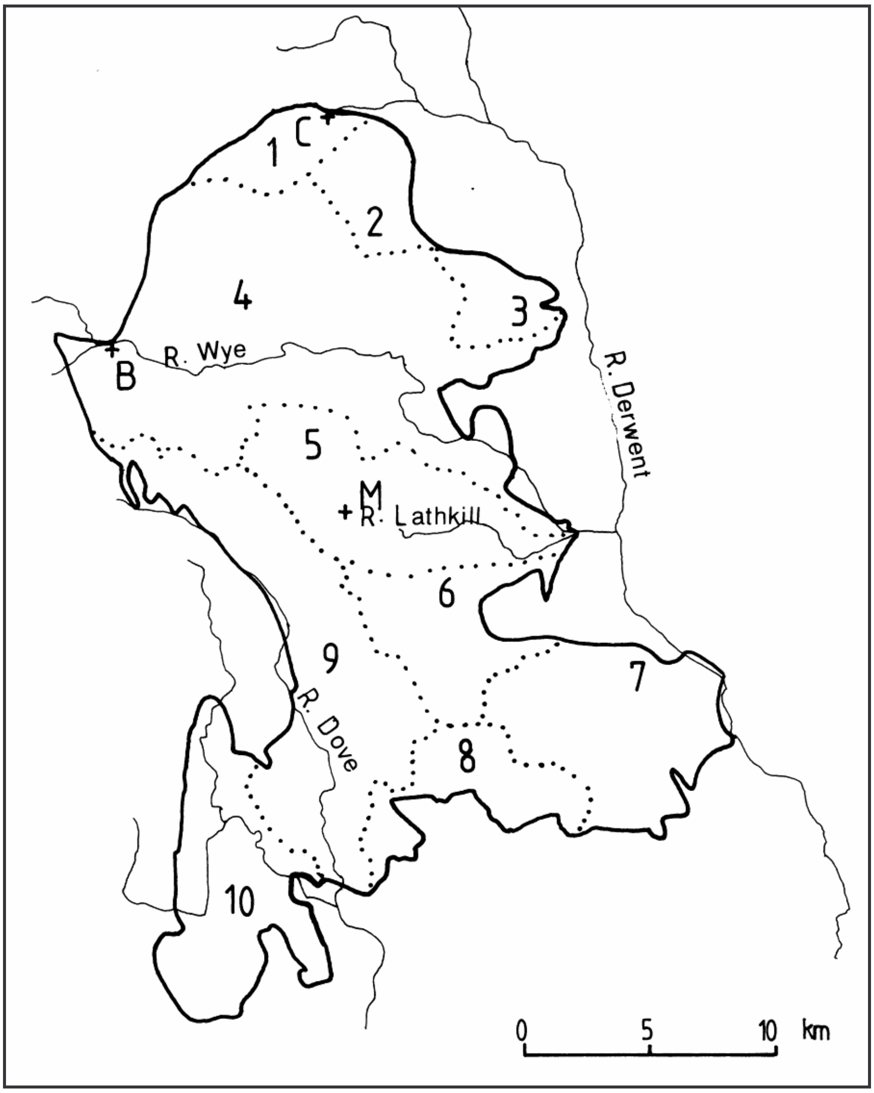
Groundwater basins in karst rarely coincide exactly with surface water basins. White (1993, p. 476) stated "*Recharge near the basin boundaries flows into all adjacent basins, showing that the concept of a ground-water basin divide, although sound on a regional scale, is somewhat fuzzy if examined in close detail*". Water tracing, the technique of injecting a marker into groundwater and monitoring the output time and location, has proved a very useful tool in helping to define groundwater basin catchments. Fluorescent dyes are the most commonly used tracer, but others, such as spores, salt or polystyrene; have also been injected. Since its formation, the Limestone Research Group, University of Huddersfield, has been commissioned to carry out a number of water-tracing experiments to provide evidence of flow vectors, likely flow paths and the range of permeability within the White Peak. This author has only had direct involvement with two of the experiments described in this thesis. This is partly because the number of experiments that can be conducted is restricted on technical grounds by the rate at which injected dye is removed from the system and partly because of the costs and time involved in effective water-tracing. In addition to the water-tracing experiments with which the author has been directly involved, the results of past water-tracing experiments carried out within the subject area have been made available and have been reinterpreted within the context of the interpretation of the geology and karst terminology and processes described in this thesis. In particular, the reader's attention is drawn to the use of the terms matrix, fracture and channel flow (chapters 1, 4, 8 and Appendix 1.1). In some cases this represents a reinterpretation of the data, but any reinterpretation is solely the work of this author and it should not be judged against earlier interpretations made on the basis of the site-specific information that was available at the time of the preparation of the experiment report.

Procedures for conducting water-tracing experiments are well described in a number of standard texts, including Brassington (1988) and White (1993). Methods employed in the various types of tracing experiments are broadly similar. The Limestone Research Group prefers fluorescent dyes, because in the concentrations used they are perceived to be non-toxic. Evaluation of the different tracers is beyond the remit of this thesis, but useful summaries can be found in Smart and Laidlaw (1977), Trudgill (1987) and Kasnavia et al. (1999). Most commonly used are fluorescein (CI 45350 Acid Yellow 73) and rhodamine WT (CI Acid Red 388). Wherever possible the Limestone Research Group injects two (occasionally three) different tracer dyes at different locations on the same day. This provides economies of scale when sampling twenty to forty potential receptor locations, which might be several kilometres apart. Prior to the commencement of the test the appropriate sampling protocol is planned and samples are obtained to determine background concentrations of the dyes. Sampling generally takes two forms, namely discrete water samples and fluocapteurs (Appendix 7.3).

Upon return to the laboratory the discrete water samples are kept cool and out of sunlight and fluocapteurs are washed and dried. An elutant is used to extract the fluorescent dye. Both the discrete water samples and the samples of elutant from the fluocapteurs are analysed using luminescence spectrofluorimetry. The analytical method is sensitive to the separation between excitation and emission. An Hitachi F4500 luminescence spectrofluorimeter, with 20 nm separation is used by the Limestone Research Group. A plot of intensity (fluorescent units) against excitation wavelength is produced. The presence of fluorescein dye in water is indicated by a clear peak at 493 nm and of rhodamine WT in water by a peak at 560 nm. Unfortunately there are a number of difficulties relating to the interpretation of the results. In particular a number of other organic substances fluoresce at similar wavelengths, and fluorescent products are present in some pollutants such as in antifreeze products, and the aromatic hydrocarbons in coal charcoal fluoresce over a range of wavelengths. These problems have been described by Smart and Simpson (2001, 2002) and Smart and Karunaratne (2001); see also Appendix 7.3. Accordingly, although the baseline monitoring carried out immediately prior to a test should assist in the interpretation, by subtraction of background fluorescence, the interpretation of minor peaks, where dye may have been detected in very low concentrations, can be problematical and commonly is reliant on additional information such as the reproducibility of the results. Having identified the sites at which positive connections have been made it is possible to use the first arrival time and the distance between input and output, to calculate minimum velocities of tracer movement.

The duration over which dye continues to emerge at a given output point provides additional information on the behaviour of the aquifer and, ideally, tests would continue until all of the injected dye has emerged. It has been found that dye can continue to emerge from some springs for several months. Furthermore, seasonal fluctuations in groundwater level can be identified in borehole groundwater records (Chapter 8) and it is common to find that where water-tracing spans the period of seasonal recession the dye may appear at an output point, be stored as groundwater levels fall and then re-emerge with the seasonal recovery in groundwater level. As Martin and Dean (1999) point out "*As elevations of the water surfaces change, water may use different conduits as flow paths, or conduits may become completely filled. Consequently, dye trace studies would have to be conducted frequently and at various stages of water level in order to accurately determine the residence time of water in the subsurface, as well as the relationship of residence time to river stage or water-table elevation*". Within this study area further interpretational difficulty is encountered where springs have been given different names by different authors.

Beck and Gill (1991) presented a map of catchment areas in the White Peak, based largely on the surface rivers, combined with the speleological and geological knowledge of the authors and with some reference to water tracing work. Gunn (1998) adopted the same boundaries in the preparation of a map of the groundwater sub-basins and these have also been used as main headings in this chapter, with subheadings corresponding to individual experiments. The groundwater sub-basin boundaries have been modified by this author in the light of the results of the experiments presented below and in the light of the conceptual model (Chapter 9; Figure 7.1). Detail with respect to the individual experiments can be found in Appendix 7.1.



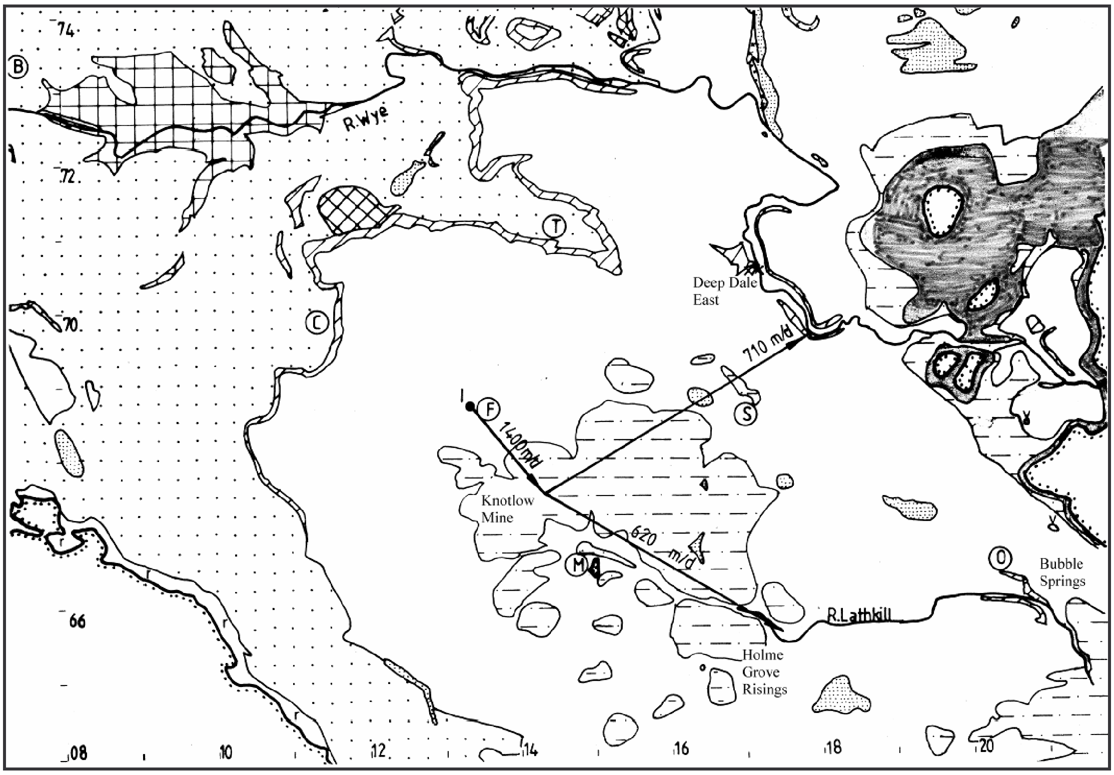
Key: Sub-basins: 1: Castleton; 2: Bradwell; 3: Stoney Middleton; 4: Wye; 5: Lathkill; 6: Bradford; 7: Matlock and Wirksworth; 8: Parwich and Bradbourne; 9: Dove; 10: Hamps.
Reference locations: B Buxton; C Castleton; M Monyash

Figure 7.1: Groundwater sub-basins of the White Peak (after Beck and Gill, 1991).

7.2 The Lathkill Sub-basin.

7.2.1. Investigation of a soakaway at Flagg, Derbyshire (Appendix 7.1, Hardwick and Gunn, 1995).

On 23 May 1995 rhodamine WT was injected into the Monsal Dale Limestone Formation via a soakaway at Flagg (SK 13366866). Dye was recovered from: Knotlow Mine, Magpie Sough, Holme Grove Risings and Bubble Springs (Figure 7.2). Dye was not recovered from Deep Dale (East).



Key: I Injection point. C Chelmorton; F Flagg; M Monyash; S Sheldon; T Taddington.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. x monitoring location. r = reef deposits; v volcanic vent

Legend: Head

Widmerpool Formation

Eyam Limestone Formation

Monsal Dale Limestone Formation

Bee Low Limestone Formation

Woo Dale Limestone Formation

Interbedded lavas

Dolerite

Figure 7.2: Results of water-tracing: soakaway in Flagg (Appendix 7.1).

The flow rate, calculated to be in the order of 1400 m/day to Knotlow Mine, is suspected to indicate channel-guided flow. It is also suspected that the connections to Holme Grove Risings, Bubble Springs and Magpie Sough are via Knotlow Mine. The mean flow rate to Holme Grove Risings has been calculated to be 740 m/day. If routed via Knotlow mine a minimum flow rate of 620 m/day (Knotlow to Holme Grove Risings) has been calculated and is likely to be indicative of saturated zone fracture flow to Holme Grove Risings. The calculated flow rate of 1060 m/day (1050 m/day if via Knotlow) to Bubble Springs could be indicative of channel flow, but is more likely to represent surface water flow from Holme Grove Risings to Bubble Springs. There was no proven connection with the Deep Dale (East, more recently referred to as the Lees Bottom) springs, or with the River Wye upstream of Deep Dale. The Monsal Dale Limestone dips approximately 2˚ to the southsoutheast (BGS Sheet 111). Immediately to the north and south (in the order of 150 m) of the injection point the Monsal Dale Limestone is crossed by a number of northeast to southwest-trending mineralized faults. The dominant inception horizons are probably marginally above the top of the Upper Miller's Dale Lava and the boundary between the Bee Low Limestone Formation and the overlying Monsal Dale Limestone Formation. The absence of a connection with Deep Dale (East) indicates that the connection with Magpie Mine is via Knotlow Mine, rather than Hubberdale Mine. The minimum flow rate from Knotlow Mine to Magpie Sough has been calculated as 720 m/day (with a mean value of 670 m/day from Flagg to Magpie Sough). The calculated values are thought to be indicative of saturated zone fracture flow from Knotlow to Magpie Sough.

7.2.2 Investigation of a soakaway at Flagg Sewage Treatment Works (Appendix 7.1, Hardwick, 1996a).

On 7 December 1995 rhodamine WT was injected into the Monsal Dale Limestone Formation via a soakaway in Flagg STW (SK 13446868). Dye was recovered from Knotlow Mine and Bubble Springs. Holme Grove Risings were not monitored. Unlike the test reported above, there was no proven connection with Magpie Sough. The absence of a connection with Magpie Sough suggests to this author that groundwater levels had fallen below the level of the connection identified in 7.2.1. Evidence for the lower groundwater conditions comes from examination of borehole data (Chapter 8), e.g. the groundwater level in the Bull I' Th' Thorne Borehole (SK 128665) was 243.86 m OD and falling on 23 May 1995 and was 225.33 m OD and rising on 19 December, 1995. It should also be noted that the flow rate determined between the injection point and Knotlow Mine (480 m/day) could be a significant underestimate, as the monitoring was carried out after three days and in the test carried out in the soakaway (see section 7.2.1) the dye reached Knotlow Mine after 1 day. However, the significantly lower rate to Bubble Springs (a mean velocity of 250 m/day, or 300 m/day from Knotlow Mine to Bubble Springs) does appear to indicate a change in the hydrological conditions and is also probably attributable to a proportionately greater contribution of confined water flowing beneath the Upper Miller's Dale Lava. Furthermore, it should be noted that the actual flow rate could have been slower, these values were calculated from an assumed 21-day recovery, as the fluocapteurs were replaced after 20 and 30 days and the dye was recovered on the 30 day visit. Consideration of the rainfall records over the duration of the test offers an alternative hypothesis. It is possible that dye was stored in the karst system as a consequence of the low groundwater conditions and that it was "flushed" through in response to the significant rainfall event on 21 December 1995 (Figure 7.3, p. 131). As Holme Grove Risings was not monitored it is plausible that the dye recovered at Bubble Springs had actually travelled downstream from Holme Grove Risings.

7.2.3 Tracing effluent discharges from Chelmorton and Taddington Sewage Treatment Works, Derbyshire (Appendix 7.1, Hardwick, 1996b).

On 26 March 1996, rhodamine WT was injected into one of the sewage treatment lagoons situated over the Upper Miller's Dale Lava at Taddington Sewage Treatment Works (SK 15157100). Dye was recovered at: Knotlow Mine and Holme Grove Risings (460 m/day from Knotlow Mine, Figure 7.4). The rate to Knotlow Mine is not known due to the monitoring frequency (dye could have reached Knotlow Mine anywhere between three and seven days after injection, giving a minimum flow rate of 500 m/day). This test was conducted during a period of relatively high groundwater conditions (on the

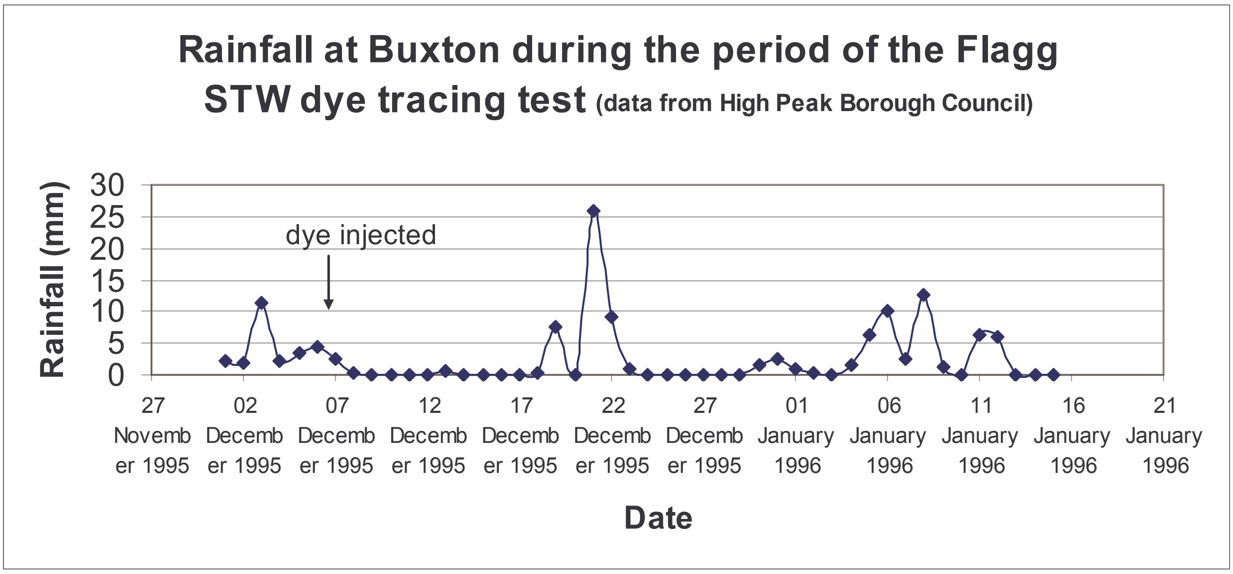
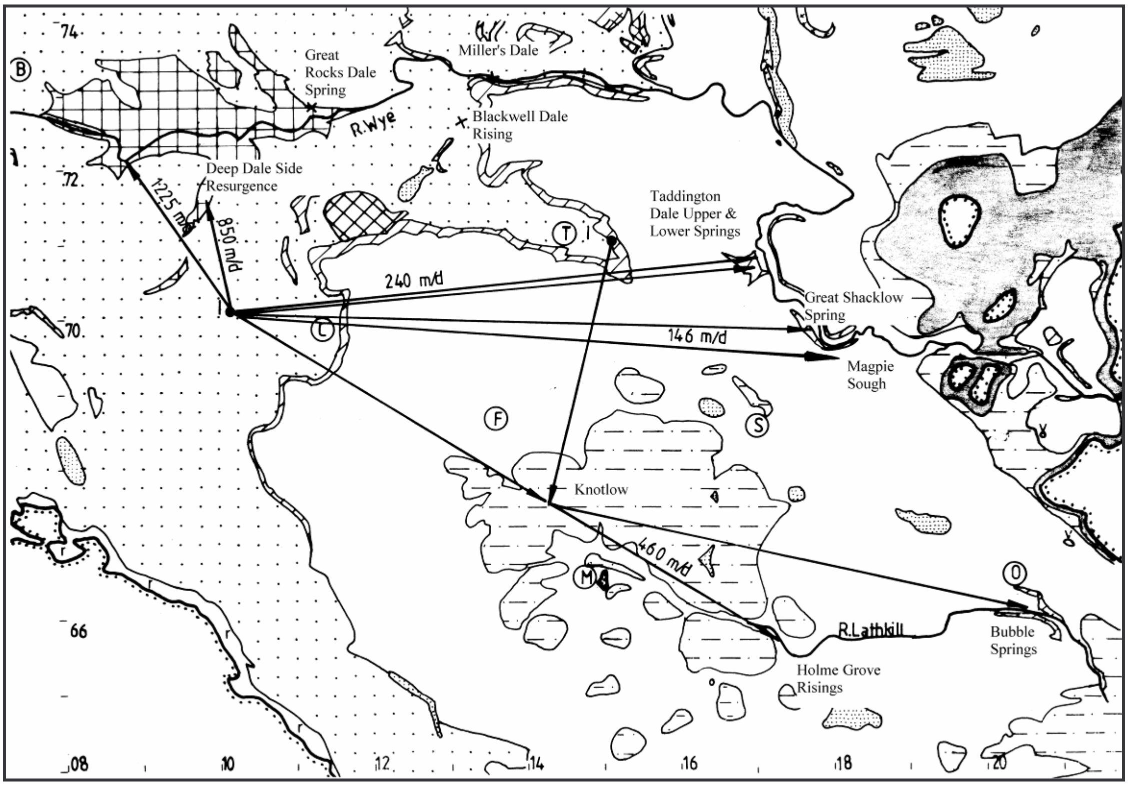


Figure 7.3: Rainfall at Buxton December 1995 and January 1996.

18 March 1996 the groundwater level was 250 m OD in the Bull I' Th' Thorne Borehole), which suggests to this author that during periods of high groundwater levels, faults associated with the Taddington Anticline act as a localised groundwater divide. It would seem most likely that groundwater reached Knotlow Mine via Hubberdale Mine and that the flow path to Holme Grove Risings is also via Knotlow Mine.

On the same date (26 March 1996) fluorescein was injected into the Bee Low Limestone Formation (Chee Tor Limestone Member) via the soakaway at Chelmorton Sewage Treatment Works (SK 10157015). It is suspected that dye injected at this location rapidly entered the main body of groundwater via dissolutionally enlarged fractures in the Chee Tor Limestone Member. Dye was recovered at: the River Wye upstream of Deep Dale (west); Deep Dale (west); the River Wye at Miller's Dale (likely to be dye from upstream); Taddington Dale 1 Spring (more recently referred to as Lees Bottom 2 (lower) by the Limestone Research Group); Great Shacklow Wood Spring (more recently referred to as Shacklow Wood 2 by the Limestone Research Group); Magpie Sough; Bubble Springs; Holme Grove Risings; and Knotlow Mine (Figure 7.4). The results of more recent tracing (subsection 7.2.4) have confirmed that the dye recovered from upstream of Deep Dale (west) is likely to have arrived via Ashwood Dale Rising, possibly from the underflow springs (that have been observed at river level, immediately to the north of the A6), which continue to flow, even after flow at the main rising ceases. Unfortunately the level of the receiving groundwater body (at the Sewage Treatment Works) was not recorded (and may not have been known) at the time of the water-tracing test. However, evidence from a section drawn during the Illy Willy Water experiment (subsection



Key: I Injection point. C Chelmorton; F Flagg; M Monyash; S Sheldon; T Taddington.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent

Legend: Head; Widmerpool Formation; Eyam Limestone Formation; Monsal Dale Limestone Formation; Bee Low Limestone Formation; Woo Dale Limestone Formation; Interbedded lavas; Dolerite

Figure 7.4: The results of water-tracing: Chelmorton and Taddington Sewage Treatment Works (Appendix 7.1).

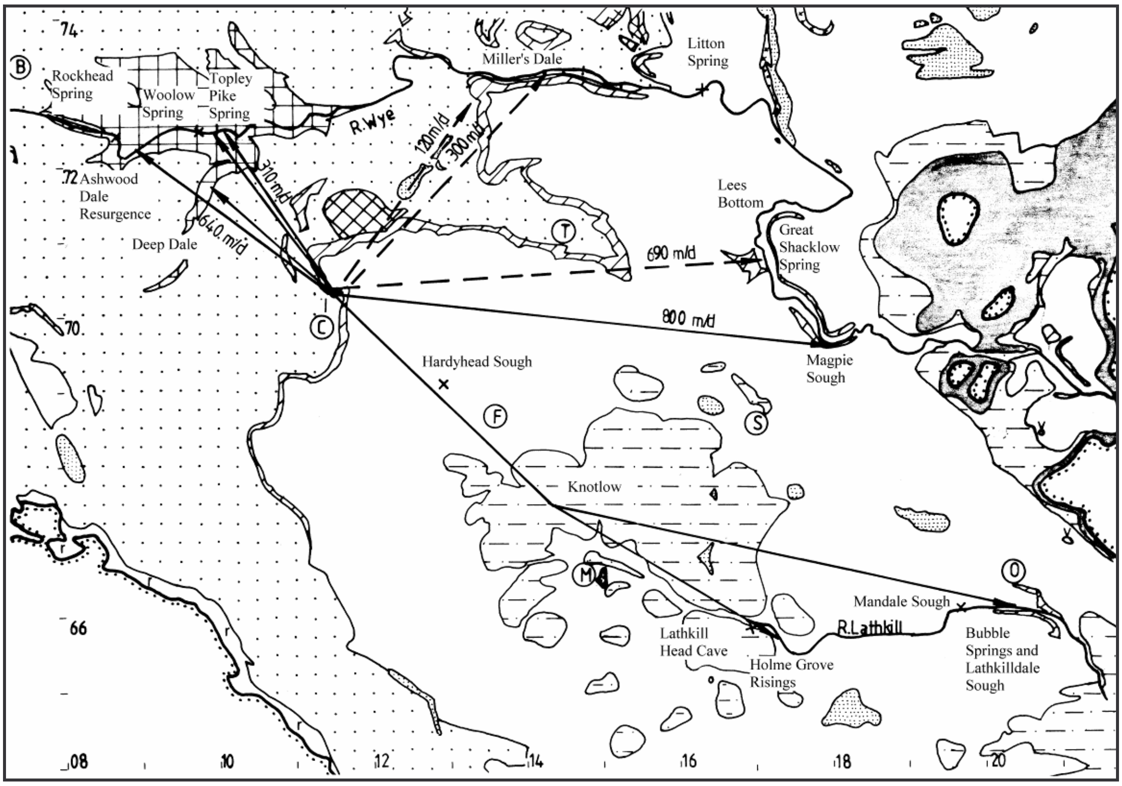
7.2.4) and the level of the Bull I' Th' Thorne Borehole indicate a groundwater level in the order of 250 to 253 m OD in the Bee Low Limestone Formation at Chelmorton. The receiving springs were at elevations of between 250 m OD (Ashwood Dale Spring) and 143 m OD (Magpie Sough). Interestingly, groundwater rises from the Monsal Dale Limestone Formation at Lees Bottom 2 Lower, Great Shacklow, Bubble Springs and Holme Grove Risings. This confirms that these springs are not just fed by inception horizon-related fractures, but they form baseflow springs rising on faults (Chapter 6). Flow rates for the fluorescein have been calculated to be in the order of 1225 m/day to Ashwood Dale, indicating channel flow, which is likely to be focused on the inception horizon associated with the Woo Dale Limestone Formation (Figure 7.6).

The connection between Chelmorton and Knotlow Mine is likely to have been deeper than that between Taddington and Knotlow. The base of the Upper Miller's Dale Lava Member is at a level of approximately 220 m OD where it feathers out to the northwest of Knotlow Mine (Figure 7.6), suggesting that groundwater flowing beneath the Upper Miller's Dale Lava would be increasingly confined to the southeast. However, the monitoring was insufficiently close to enable comparisons to be made. Both traces suggest flow rates in the order of 450 m/day between Knotlow and Holme Grove Risings. The arrival of fluorescein at Bubble Springs ahead of its arrival at Holme Grove Risings suggests the presence of an independent route to Bubble Springs, although it is suspected that this route is also via Knotlow.

7.2.4 **Illy Willy Water experiment** (Appendix 7.1, Banks and Gunn, 2003).

The Illy Willy Water is situated at the northern end of Chelmorton village (SK 11537033). It comprises a short stream that rises from water perched where the Monsal Dale Limestone is underlain by the Upper Miller's Dale Lava and sinks after a few metres into the Bee Low Limestone Formation (Miller's Dale Limestone Member) at the western end of a dominant northeast to southwest-trending mineral vein. The spring is perennial and was once the sole water supply for the village (Roberts, 1964, Spencer and Robey, 1973). Fluorescein was injected into the sink to investigate the routes taken by the sinking stream. To capitalise on the labour investment an additional water-tracing test was commissioned and rhodamine WT was injected into the area where water discharging from Hardyhead Sough (SK 30007132) sinks into the Miller's Dale Limestone Member. The tests were commenced on 24 April, 2002. Being near the boundary of the Lathkill catchment, divergent flow was hypothesised and therefore a large number of resurgences were selected for monitoring. The tests were carried out during a period of falling groundwater levels. Monitoring continued until 20 August, 2002. There was no obvious recovery of rhodamine WT, which suggests that either recharge in this area follows a deep underflow drainage route and does not reappear in the Wye catchment (Banks and Gunn, 2002), or that flow is dispersed. Beswick (2002) described how historically, during periods of high discharge, the overflow water from Five Wells (SK 12557110) ran down the fields and into drinking troughs in the lay-by at the bottom of Waterloo Hill (SK 12757151). The water was piped under the road and continued to flow across the fields to supply a trough in a lay-by opposite the entrance to Spring Hill House in Priestcliffe Ditch (SK 12907190) and any surplus water would flow down the dale to the River Wye. This suggests that epikarst is very poorly developed over the Miller's Dale Limestone Member, whereas it is considered to be better developed in the Chee Tor Limestone Member of the Bee Low Limestone Formation (subsection 7.2.3).

Most of the fluorescein that was injected targeted Ashwood Dale Resurgence and also Deep Dale (west) Rising, situated to the northwest of the injection point. There was rapid flow of a lesser proportion of dye to Knotlow Mine and on to Lathkill Head Cave, Holme Grove Risings, Bubble Springs and Mandale Sough. This author established that groundwater levels in the mineral vein, towards its eastern end, were below the depth reached by a 175 m-dip meter (< 245 m OD).



Key: I Injection point. C Chelmorton; F Flagg; M Monyash; S Sheldon; T Taddington.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent

Legend: Head — Widmerpool Formation

Eyam Limestone Formation — Monsal Dale Limestone Formation

Bee Low Limestone Formation — Woo Dale Limestone Formation

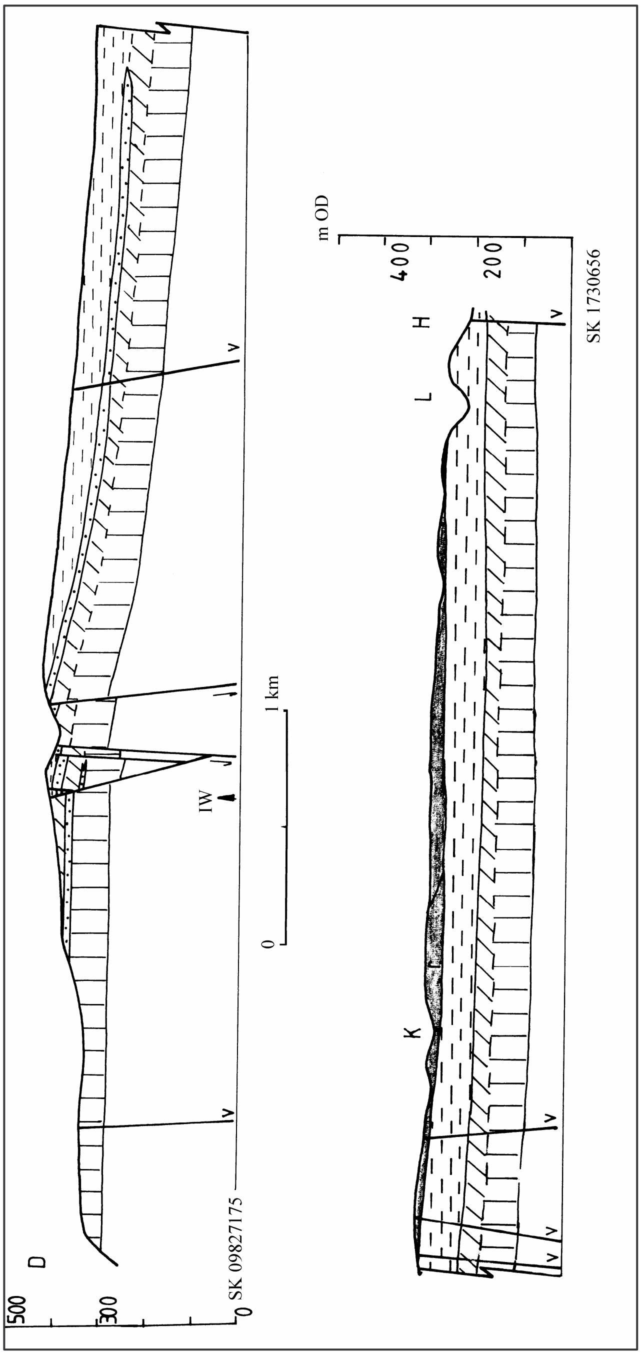
Interbedded lavas — Dolerite

Figure 7.5: Water-tracing from Illy Willy Water, Chelmorton (Appendix 7.1).

The groundwater level determined for the Environment Agency, in the Bull I' Th' Thorne Borehole on 14 April 2002 was 253.4 m OD. This, the level of Ashwood Dale Spring and evidence from a geological cross-section (Figure 7.6, p. 136) constructed by this author indicate that the groundwater level at the injection point was likely to have been at approximately 250 to 255 m OD (similar to the level estimated for the Chelmorton Sewage Treatment Works trace (subsection 7.2.3). This evidence points to groundwater flow in channels associated with inception in the Woo Dale Limestone Formation and the boundary of the Woo Dale Limestone Formation with the overlying Bee Low Limestone Formation (Figure 7.6). With respect to the fluorescein detected at Lathkill Head Cave, it is considered most likely that this travelled via Knotlow Mine. The dye reached Knotlow Mine after 5 to 10 days, whereas it reached Lathkill Head Cave after 10 to 17 days. As with the tracing carried out at Flagg, Taddington and Chelmorton Sewage Treatment Works, it is also suspected that the flow to Holme Grove Risings, Mandale Sough and Bubble Springs was also via Knotlow Mine, with flow rates to Knotlow exceeding those beyond Knotlow. However, the time interval between monitoring visits was not short enough to discriminate the difference in flow rates. The evidence from Figure 7.6 (p. 136) suggests that if channels at the boundary between the Woo Dale Limestone and the overlying Chee Tor Limestone Member guide flow paths into the Monyash Syncline, then groundwater must rise to reach Knotlow Mine. The rise of groundwater could be guided by the northeast to southwest-trending mineral vein that is situated immediately to the northeast of Knotlow Mine. Further evidence to support this comes from the fact that dye reaches Holme Grove Risings, before reaching Lathkill Head Cave, which suggests differing flow paths.

Trace amounts of fluorescein, with minimum flow rates in the order of 117 to 313 m/day targeted Woolow Spring, Topley Pike Spring, Blackwell Dale Spring and Miller's Dale Spring 8. Woolow Spring and Topley Pike Spring lie marginally below the Chee Tor Limestone/Woo Dale Limestone boundary. The boundary between the Chee Tor Limestone and the Woo Dale Limestone is thought to be close to the base of Blackwell Dale at the location at which the spring is monitored, although it may be considerably deeper to the north of the fault at SK 13327297, which is suspected by this author to approximate to the fault block boundary, (Chapter 5). Blackwell Dale and Miller's Dale lie to the northeast of Chelmorton and it is suspected that the dye at these locations results from fracture flow associated with inception horizon-related fracture flow at the formational boundary between the Bee Low Limestone and the underlying Woo Dale Limestone formations and also within the Woo Dale Limestone Formation. Faster flow rates were determined to the east and the southeast (down hydraulic gradient). The strata young to the east towards Magpie Sough and Shacklow Spring 2, with flow rising on significant faults (the Taddington Anticline fault to Magpie Mine and the southern end of a northnorthwest to southsoutheast-trending fault at Brushfield), again confirming that these are baseflow springs (Chapter 6).

The long duration over which dye continued to emerge (seventeen weeks) indicates that significant storage of groundwater occurs. Furthermore, during a period of low groundwater levels (9 to 26 July, 2002) the dye did not emerge from the main orifice of Ashwood Dale Resurgence, but it continued to emerge once groundwater levels were restored following a rainfall event (Figure 10.2).



Key: D Deep Dale; IW Illy Willy Water; K Knotlow; L Lathkill Head Cave; H Holme Grove Risings. v Mineral vein; r reef deposit

Legend: Eyam Limestone Formation

Monsal Dale Limestone Formation

Lower and Upper Miller's Dale Lava

Bee Low Limestone Formation: Miller's Dale Limestone Member

Chee Tor Limestone Member

Woo Dale Limestone Formation

Figure 7.6: Geological cross-section from Deep Dale to Holme Grove Risings.

7.2.5 Conclusions drawn from water-tracing carried out in the Lathkill Catchment.

Appendix 7.2 comprises a table that provides a brief summary of a number of tracing experiments in the Lathkill catchment, including a number of traces reported by Wheeldon (1992). Clearly, the calculated flow rates can only be as accurate as the monitoring frequency, which varies between tests, primarily for strategic reasons. Furthermore the flow rates are representative of the prevailing conditions during the period of the test and it is known that the hydrological setting is variable. Nevertheless, the results of these experiments indicate the following:

i) Water-tracing in the Monsal Dale Limestone Formation

Groundwater in the northwestern end of the catchment is perched by the Upper Miller's Dale Lava. Flow paths and rates in the Monsal Dale Limestone appear to be affected by groundwater levels. The findings of water-tracing from Taddington suggest that during low groundwater conditions some of the connections with Magpie Mine become ineffective (as indicated by water-tracing from Flagg STW in December 1996). During high groundwater conditions the elevated head associated with storage in faults associated with the Taddington Anticline acts as a groundwater divide. Accordingly, Magpie Sough captures a greater proportion of the water that would have been in the River Lathkill during periods of intermediate groundwater conditions. Flow rates determined from the water-tracing experiments indicate lower flow rates to Bubble Springs during periods of higher groundwater levels. This may be a response to changes in the piezometric surface, as described below.

ii) Water-tracing in the Bee Low Limestone Formation

The cross-section (Figure 7.6, p. 136) indicates that during periods of low groundwater conditions, groundwater levels can fall below the level of the base of the Upper Miller's Dale Lava, whereas in high groundwater conditions piezometric levels are higher and groundwater in the Miller's Dale Limestone is increasingly confined to the east (towards the feather-edge of the lava). The axis of the Monyash Syncline follows the dominant northwest to southeast structural grain and appears to guide the channel development that is implicated by the tracing carried out in the Miller's Dale Limestone Member. This would appear to support the hypothesis of inception horizon-related fracture development in the Monsal Dale Limestone Formation and in the Miller's Dale Limestone Member of the Bee Low Limestone Formation. It has been established that many of the flow paths follow routes that pass through Knotlow Mine. This mine occupies a position towards the centre of the syncline, at the featheredge of the Upper Monsal Dale Lava, thus it can be likened to a funnel receiving perched groundwater, particularly during low groundwater conditions and providing access to groundwater at depth. Evidence from the Illy Willy Water water-tracing test suggests that groundwater also rises into Knotlow Mine. It is suspected that this occurs where groundwater meets the mineral vein immediately to the west of Knotlow Mine. Appendix 7.2 provides further evidence of lower flow rates to Knotlow Mine during periods of low groundwater conditions (rates fall from approximately 1400 m/day to 470 m/day). Similar observations have been made for flow rates to Bubble Springs, which are considerably lower during periods of low groundwater conditions (ranging from 250 to 1560 m/day). This suggests that different flow paths are in operation, because, for example, groundwater rising into Knotlow is under a lower head.

It is suspected that flow rates of the order of 500 to 1550 m/day (Appendix 7.2) are indicative of mature channels associated with inception horizons in the Monsal Dale Limestone Formation, the Miller's Dale Limestone Member, at the boundary between the Miller's Dale Limestone Member and the Chee Tor Limestone Member and at the boundary between the Chee Tor Limestone Member and the underlying Woo Dale Limestone Formation. Flow rates to Magpie Sough are generally considerable lower than this, with flow rates of 210 to 270 m/day appearing to be typical, with a single value of

810 m/day having been recorded. It is considered likely by this author that confined water enters Magpie Sough at the location of the Taddington Anticline Fault, which is down-thrown to the north and coincides with the interpolated, feather-edge, of the Upper Miller's Dale Lava. At depth, below the level of Magpie Mine, Woo Dale Limestone is brought into contact with the Chee Tor Limestone Member, which would encourage southeasterly flowing water in the Woo Dale Limestone to rise up the fault, and this is the likely source of the warm water (Christopher, 1981; section 10.2 of this thesis) that has been observed to enter Magpie Sough. It is plausible that the flow rates are indicative of stylolites-related fracture flow in the Woo Dale Limestone Formation (Chapter 9).

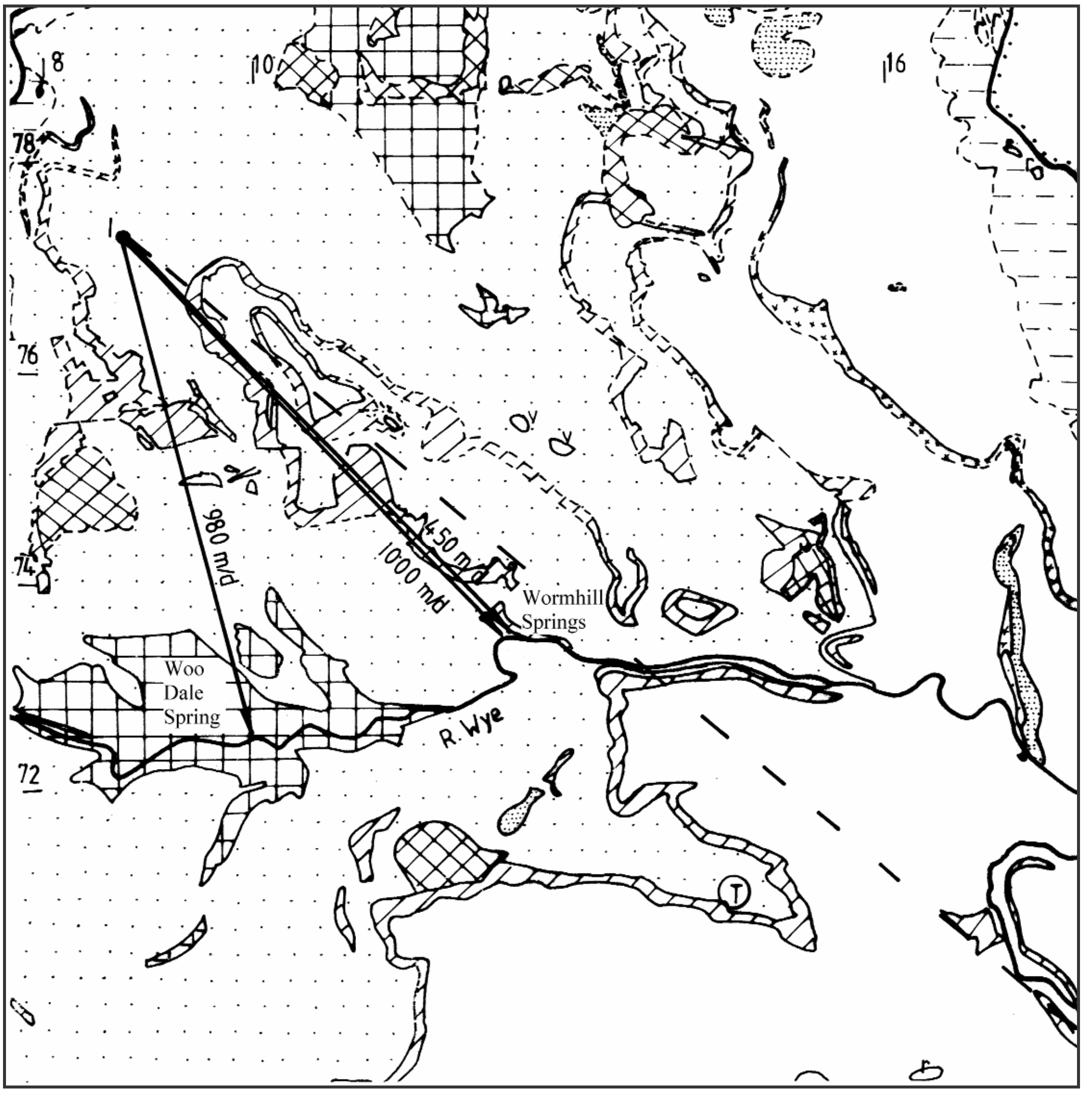
7.3 The Wye Sub-basin.

7.3.1 Documented background.

Upstream tributaries of the River Wye rise on the Namurian strata of Axe Edge Moor. Downstream, the flow is augmented by several springs, the largest being: Dog Holes (SK 04107270), Otter Hole (SK 04607330), Golf Ball Resurgence (SK 04707330) and the Wye Head Resurgence (SK 04997304). All of these springs discharge a mixture of autogenic and allogenic recharge from small streams that rise on the Namurian rocks and sink close to the boundary with the limestone. These sinks and springs are described by Beck and Gill (1991), Christopher et al. (1977) and Gunn (1998). As the River Wye flows east it receives further contributions from a number of springs and ephemeral stream flows. Initial water tracing experiments in the upper Wye were undertaken in the course of undergraduate project-work, designed to elucidate flow paths (Edmans, 1988; Gilman, 1985). Gilman's (1985) experiments identified apparently phreatic flow between Shay Lodge and Dog Holes Resurgence (SK 041727), where it is suspected by this author that groundwater rises on a northerly extension of a northnorthwest-trending fault that is downthrown to the east. Connections were also proven between: the sinks known as Anthony Hill Swallet (SK 04707030), Plunge Hole (SK 04407130), Axe Hole (SK 04407130) and Jakes Hole (SK 04457080) and resurgences at Otter Hole and Wye Head; between the sinks known as Leap Edge Swallet (SK 04906975) and Anthony Hill Swallet and the resurgence at Brook Bottom (where the angular stream plan reflects the guiding influence of the fissuring in the Chee Tor Limestone Member on the stream's down-cutting, exposing the Woo Dale Limestone Formation at the base); and between Borehole Swallet (SK 04907150) and Wye Head. An overflow route connects Poole's Cavern, which has formed in the Miller's Dale Limestone Member of the Bee Low Limestone Formation, with Wye Head. Plunge Hole and Axe Hole have also been found to connect with Brook Bottom during high groundwater conditions (Gunn, personal communication, 2007). Further downstream a series of experiments were commissioned to assist with developing the understanding of the effects of the past and future quarrying activity upon the hydrogeology. It is the opinion of this author that, with the exception of Poole's Cavern, which is likely to be associated with ancient river systems (Chapter 9), these are relatively recent (Pleistocene) connections with the River Wye.

7.3.2 **Drainage waters from RMC Roadstone Limited Dove Holes Quarry, Buxton** (Appendix 7.1, Hardwick, 1996c).

On 7 August 1996 rhodamine WT was injected into a sink hole into flowing water, at the western end of a northnorthwest to southsoutheast-trending fault in the Chee Tor Limestone Member, in the roadside adjacent to Dale Head Road (SK 088773), Dove Holes Derbyshire. It is suspected by this author that the groundwater flow was in the Woo Dale Limestone Formation, which is in the order of 70 m below ground level (230 m OD). Fifteen stream and river sites falling within the Wye and Goyt catchments were monitored (Appendix 7.1). Dye was recovered from some of the sites within the Wye catchment, namely: the River Wye downstream of Woo Dale, Wormhill Springs (East and West) and the River Wye at Taddington Dale. The test was carried out during a period of low groundwater conditions and the Great Rocks Dale springs were observed to be dry for the duration of the test. The flow rates to Woo Dale and Wormhill Springs (Main), in the order of 1000 m/day (Figure 7.7, p. 138), are comparable with those from Cunningdale Shack to Wormhill Springs East (subsection 7.3.3), but two to three times higher than the flow rates suspected to be associated with the Woo Dale Limestone Formation in the Lathkill sub-basin. As the springs egress from the Woo Dale Limestone, it is likely that the flow path falls within the Woo Dale Limestone Formation. The lower flow rate to Wormhill Springs East could be indicative of overflow from Wormhill Springs Main rising. The test was carried out during periods of low groundwater levels and therefore it has not been possible to look at the effect of higher groundwater levels.



Key: I Injection point. T Taddington
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Broken line suspected connection with Magpie Sough.

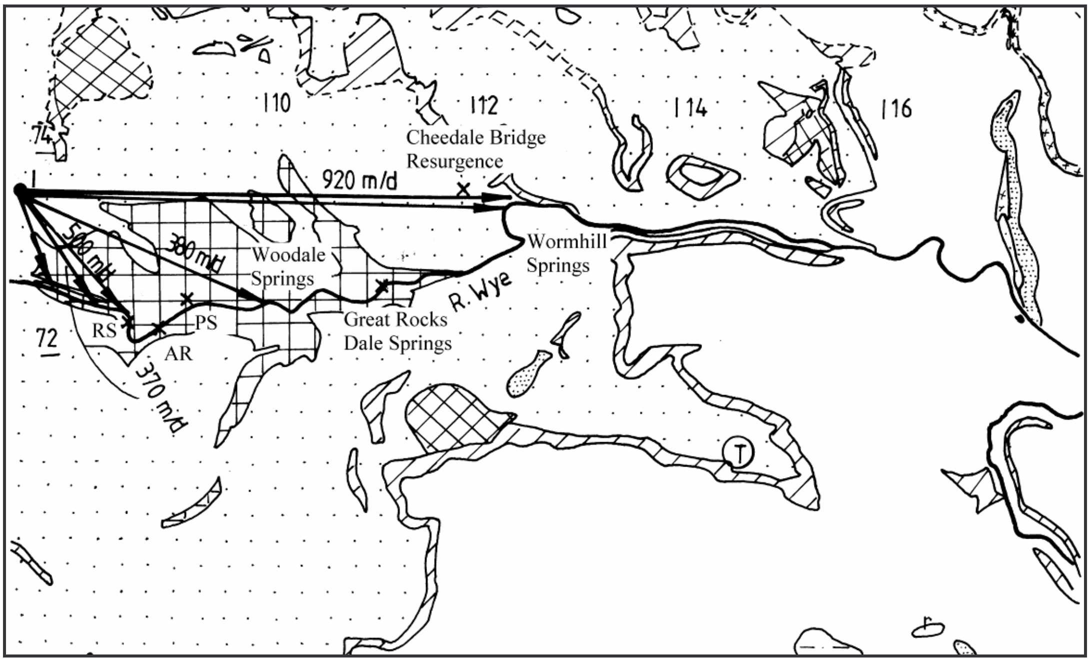
Legend: Head; Widmerpool Formation; Eyam Limestone Formation; Monsal Dale Limestone Formation; Bee Low Limestone Formation; Woo Dale Limestone Formation; Interbedded lavas; Dolerite

Figure 7.7: Proven connections with the RMC Roadstone Limited Dove Holes Quarry, Buxton (Appendix 7.1).

7.3.3 Water tracing from Cunning Dale Shack near Buxton, Derbyshire (Appendix 7.1, Hardwick and Hyland, 1991a).

On 29 August 1991 fluorescein was injected into the easterly dipping Chee Tor Limestone Member via Cunning Dale Shack (SK 07577359). Dye was recovered from the River Wye Weir, the River Wye between Cunning Dale and Pig Tor, Woo Dale Spring (No.2), Wormhill Spring (East), and the River Wye upstream of Wormhill Springs (Figure 7.8). Dye was also recovered from the Great Rocks Dale springs, where flow was intermittent; accordingly flow rates were not calculated.

The shack (doline) lies on an east to west-trending fault, which is downthrown to the north, adjacent to the point that it is met by a northwest to southeast-trending fault, which is downthrown to the southwest. This brings Miller's Dale Limestone in the downthrown block to the northwest of the doline, into contact with the Chee Tor Limestone Member to the south and the Lower Miller's Dale Lava to the east. As a consequence it is plausible that the doline formed by a combination of run-off from the Lower Miller's Dale Lava and inception horizon-guided flow in the Miller's Dale Limestone Formation, coming into contact with the Chee Tor Limestone Member. It is suspected by this author that the east to west-trending fault, the Woo Dale Anticline growth fault, is associated with the northern limit of the Woo Dale Fault block (Chapter 5). The exposure of the Woo Dale Limestone Formation to the south indicates that at this point beds dip to the southeast. In the vicinity of the doline the limestone is overlain by alluvium, which occupies part of the bed of Cunning Dale.



Key: I Injection point; T Taddington; AR Ashwood Dale Resurgence; PS Pictor Spring; RS Rockhead Spring. Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone. Lines with arrowheads indicate connections proven by water-tracing. X Monitoring location.

Legend: Head; Widmerpool Formation; Eyam Limestone Formation; Monsal Dale Limestone Formation; Bee Low Limestone Formation; Woo Dale Limestone Formation; Interbedded lavas; Dolerite

Figure 7.8: Results of water-tracing: Cunning Dale Shack (Appendix 7.1).

Additional work (Hardwick and Hyland, 1991a and 1991b) revealed that the greatest concentration of dye targeted the River Wye in close proximity to the Buxton Sewage Treatment Works outfall at SK 08007253, which appears to be at a faulted contact between the Woo Dale Limestone Formation and the Chee Tor Limestone Member that is downthrown to the south. This suggests the possibility of storage associated with the east to west fault immediately to the north the River Wye in the vicinity of Buxton Sewage Treatment Works. It was observed that there was a general increase in the dye recovery following a significant rainfall event on 8 September 1991, which further implies a significant component of storage (possibly associated with fissuring in the Chee Tor Limestone Member, Chapter 9) and also reinforces the rapid flow-through times. The dye recovery from Woo Dale Spring (No. 2) was also large.

Ground level at the dye injection point is approximately 300 m OD and the River Wye at Buxton Sewage Treatment works is at about 250 m OD. Minimum flow rates are shown on Figure 7.8. These rates can be subdivided into flow vectors with a rate of 920 m/day to the east and 300 to 500 m/day to the south and southeast. The implications are:

i) An almost vertical flow path to the water table, with phreatic fracture-related flow to the River Wye; or possibly vadose flow down Cunning Dale. The former is considered more likely by this author, because of the dominance of sub-vertical fissuring in the Chee Tor Limestone Member.

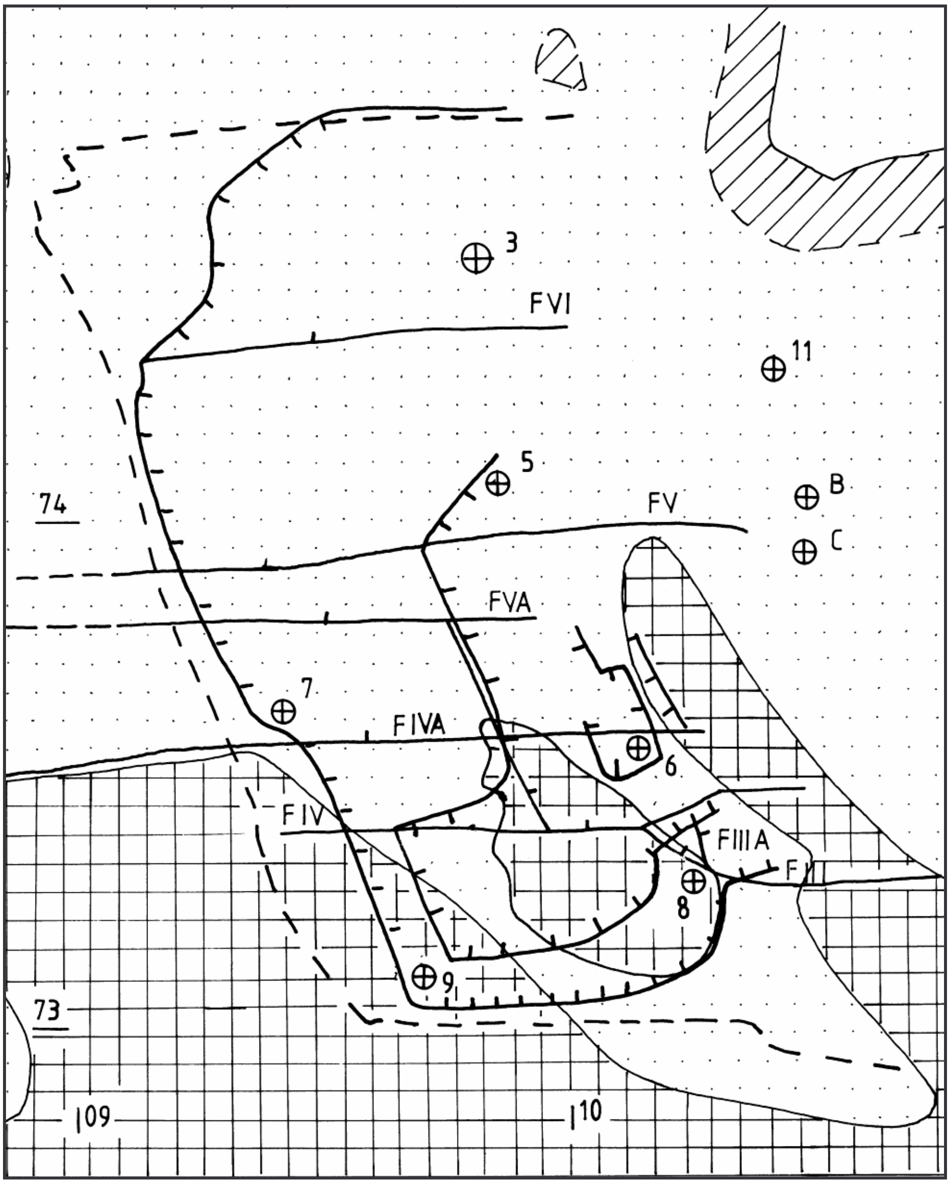
ii) Easterly flow towards Wormhill Springs, with flow rates of 920 m/day likely to be indicative of channel, flow associated with inception horizons at the boundary between the Chee Tor Limestone Member and the underlying Woo Dale Limestone Formation.

iii) Another flow vector associated with the eastsoutheasterly fault zone and recharge to fracture flow in the Woo Dale Limestone Formation, directed down Woo Dale. However, it should be noted that there was some concern about possible contamination of this monitoring point by dye moving down the River Wye.

7.3.4. **Tunstead Quarry** (Appendix 7.1, Hardwick and Gunn, 1994).

On 22 November 1993 fluorescein and rhodamine WT were injected into the Woo Dale Limestone Formation via two boreholes within the upper lift of the quarry, namely boreholes 3 and 7 respectively (Figures 7.9 to 7.11, pp. 143, 145 and 146). Dye was recovered from: boreholes in lower lifts in the quarry, boreholes in Old Moor, the confluence of Woo Dale with the River Wye, Wormhill Springs and Chee Dale Bridge Rising. Monitoring was carried out during the period 22 November to 23 December, 1993. The geology of Tunstead Quarry, which has been excavated into the Woo Dale Limestone Formation, has been described by Ineson and Dagger (1985). The dip of the beds is in the order of 2° to the southeast (BGS Sheet 111). This is coincident with the structural grain of the region. Clearly, imposed upon this there is significant east to west faulting, which is coincident with the regional hydraulic gradient (Figure 7.9, p. 143).

The results of the rhodamine-trace from Borehole 7 (Figure 7.10, p. 145) indicate that the most rapid movement of rhodamine WT was to the east towards Wormhill Springs East (WHE). Velocities in the order of 700 m/day were recorded in this direction. The top of the Woo Dale Limestone Formation is exposed in the River Wye immediately upstream of the monitoring point WAW (SK 12287330), dye



Key:
Circles with cross and number/letter = borehole positions; Faults shown as lines with dash to indicate downthrow. Faults referenced F with a numerical designation (Ineson and Dagger, 1985). Dashed line planning boundary in 1985.

Legend:

Bee Low Limestone Formation

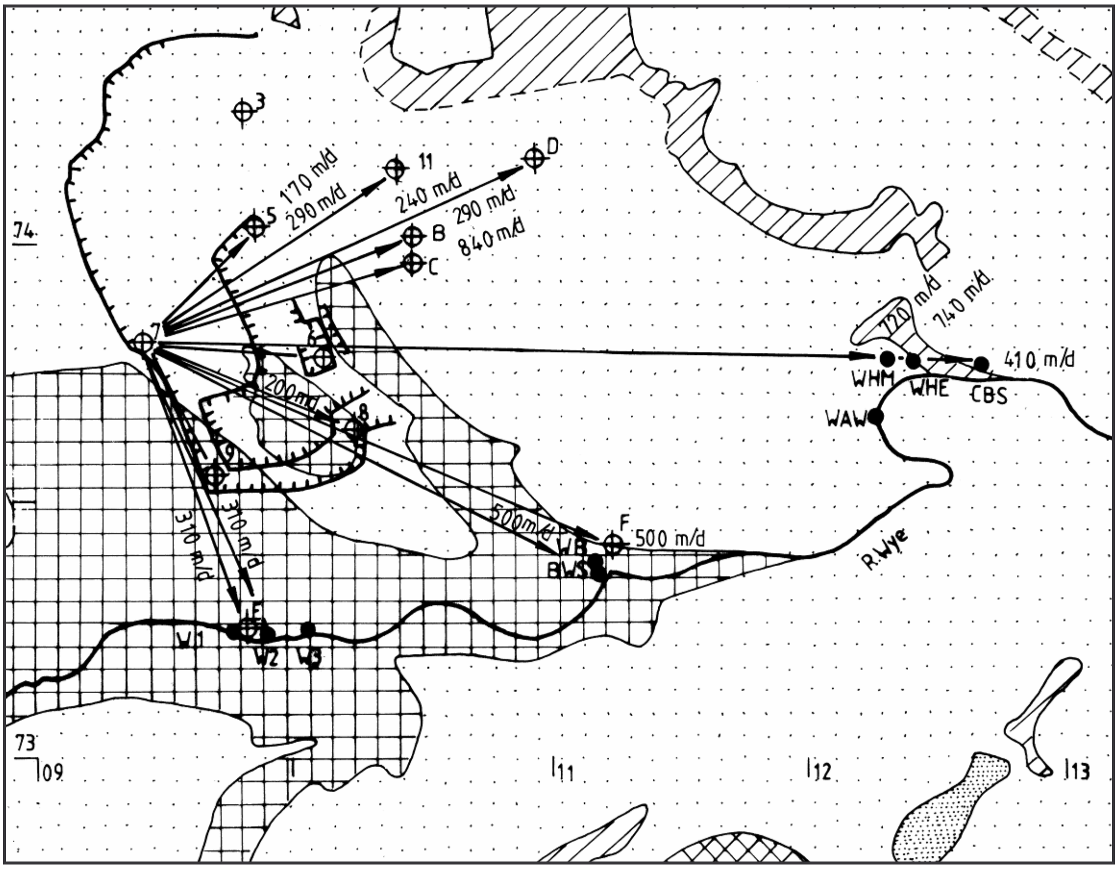
Woo Dale Limestone Formation

Interbedded lavas

Figure 7.9: Faulting in Tunstead Quarry (after Ineson and Dagger, 1985).

was not recovered from this location, which suggests that flow paths are predominantly in the Woo Dale Limestone Formation. It is considered that groundwater in the Woo Dale Limestone Formation rises via channels on a partially mineralized fault that trends northwest to southeast immediately to the southwest of the springs. The more rapid recovery of dye from Wormhill Springs East suggests that the groundwater rises into the fault zone and to channels that feed the Chee Dale Bridge Rising (CBS). Thus CBS can be considered to be an overflow spring. The slower and shorter duration of recovery at WMS could be attributable to the proportionately greater contribution to this spring from the large catchment area to the north of the spring. Rapid flow to Blackwell Spring (BWS) and Borehole F at the confluence of Great Rocks Dale and the River Wye (500 m/day) is likely to occur as fracture flow in the Woo Dale Limestone Formation. There was also rapid flow of rhodamine WT towards the southwestern corner of the quarry (to the southsoutheast of the injection well). The velocities in this direction were found to be in the order of 310 m/day. Structurally related fracture-guided fracture flow is considered the most likely guiding influence on the flow vector. The water-tracing test identified slower, fracture flow to the other boreholes within the quarry, confirming the divergent flow from the fault zone (Fault IV, Figure 7.9), which is indicative of dye injection close to the potentiometric surface. It was observed by Hardwick and Gunn (1994) that a portion of the rhodamine moved along a discrete horizon at 268 m OD within the Woo Dale Limestone. The test was carried out during the winter months and fluctuations in groundwater levels were noted at the time of the test. More specifically, Borehole 9 was dry during the period 2 to 8 December 1993.

The findings of the fluorescein-trace injected into Borehole 3 (Figure 7.11, p. 146) were confirmatory of the findings of the trace from Borehole 7; although there were apparent variations in the rate of flow and there was no recovery from Wormhill Main Spring (WHM) or from Chee Dale Bridge Spring (CBS). Velocities towards the confluence of Woo Dale with the River Wye were more rapid (up to 720 m/day), than from Borehole 7, suggesting the possibility that as well as the east to west orientated channels that target Wormhill Springs; there are southerly-orientated channels that target Woo Dale. Lower apparent velocities in the main quarry and towards the confluence of Great Rocks Dale with the River Wye, probably reflect the observation that Borehole 7 was situated on, or very close to, the dominant fault FIVA (Ineson and Dagger, 1985; Figure 7.9, p. 143), which guided groundwater towards dominant joint or fracture sets.



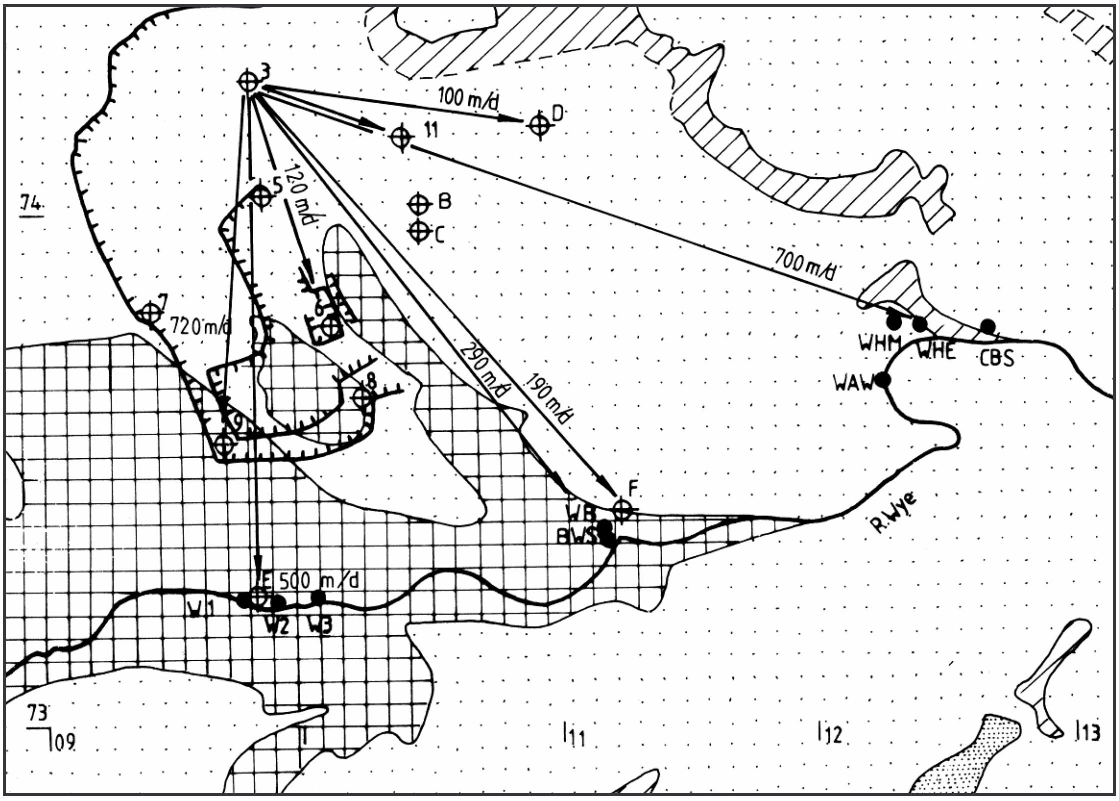
Key:
Circles with cross and number/letter = borehole positions; lines with arrowheads indicate connections proven by water-tracing.
BWS Blackwell Spring; CBS Chee Dale Bridge Rising; W1 River Wye upstream of Woo Dale Springs; W2 Woo Dale Spring 2; W3 Woo Dale Spring 3; WAW River Wye upstream of Wormhill Springs; WB Woo Dale Borehole water; WHE Wormhill Spring (East), WMS Wormhill Main Spring.
Faults recorded on Figure 7.9 (p. 143).

Legend:

Key: I Injection point. C Chelmorton; F Flagg; M Monyash; S Sheldon; T Taddington.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent

Legend: Head — Bee Low Limestone Formation — Woo Dale Limestone Formation — Interbedded lavas — Dolerite

Figure 7.10: Results of water-tracing from Borehole 7, Tunstead Quarry (Appendix 7.1).



Key:
Circles with cross and number/letter = borehole positions; lines with arrowheads indicate connections proven by water-tracing.
BWS Blackwell Spring; CBS Chee Dale Bridge Rising; W1 River Wye upstream of Woo Dale Springs; W2 Woo Dale Spring 2; W3 Woo Dale Spring 3; WAW River Wye upstream of Wormhill Springs; WB Woo Dale Borehole water; WHE Wormhill Spring (East), WMS Wormhill Main Spring.
Faults recorded on Figure 7.9 (p. 143).

Legend: Head | Bee Low Limestone Formation | Woo Dale Limestone Formation | Interbedded lavas

Figure 7.11: Results of water-tracing from Borehole 3, Tunstead Quarry (Appendix 7.1).

7.3.5 **Landfill Site, Taddington, Derbyshire** (Appendix 7.1, Gunn and Hardwick, 1999).

Calton Hill Landfill site occupies part of a former quarry in a dolerite intrusion associated with the volcanic vent at SK 11957125. On 22 June 1999 fluorescein dye was injected into the Miller's Dale Limestone Member via a sink hole at SK 11857169 (Figure 7.12, p. 148), and rhodamine WT was injected into a soakaway at SK 12437173. After a period of twenty days and no dye having been recovered from the monitoring sites a further injection of fluorescein was made. The locations into which the dye was injected lie within the Miller's Dale Limestone Member. SK 12437173 falls on the line of an east to west-trending rake, although the line of the rake is not shown at surface in this area it would seem unlikely that it is not connected at depth. The association of dolines with rakes was noted in Chapter 4. It is also interesting to note that the BGS (1: 50 000 scale, sheet 111) indicates the occurrence of extensive Head deposits in this area, which are also commonly associated with dolines

(Chapter 4). Furthermore, the injection points are situated close to the crest of an anticline, a structural setting commonly associated with more intense fissuring.

Although extensive monitoring was carried out (Appendix 7.1), there were no convincing dye recoveries in the period up to 22 September 1999. The Limestone Research Group (Gunn and Hardwick, 1999) presented a number of possible reasons for the lack of dye recovery, which are summarised below:

i) The tracer remained in the groundwater body beneath the injection sites, or moved too slowly towards the monitoring sites to be detected during the monitoring period. Hardwick and Gunn (1999) observed that work elsewhere in the Peak District suggests that fracture flow can be as low as 26 m/day. At these velocities dye could have taken 77 to 115 days to reach sites 2 to 3 km from the point of injection, a timescale that exceeded that for the monitoring period.

ii) The tracer was held in temporary storage within the limestone aquifer:

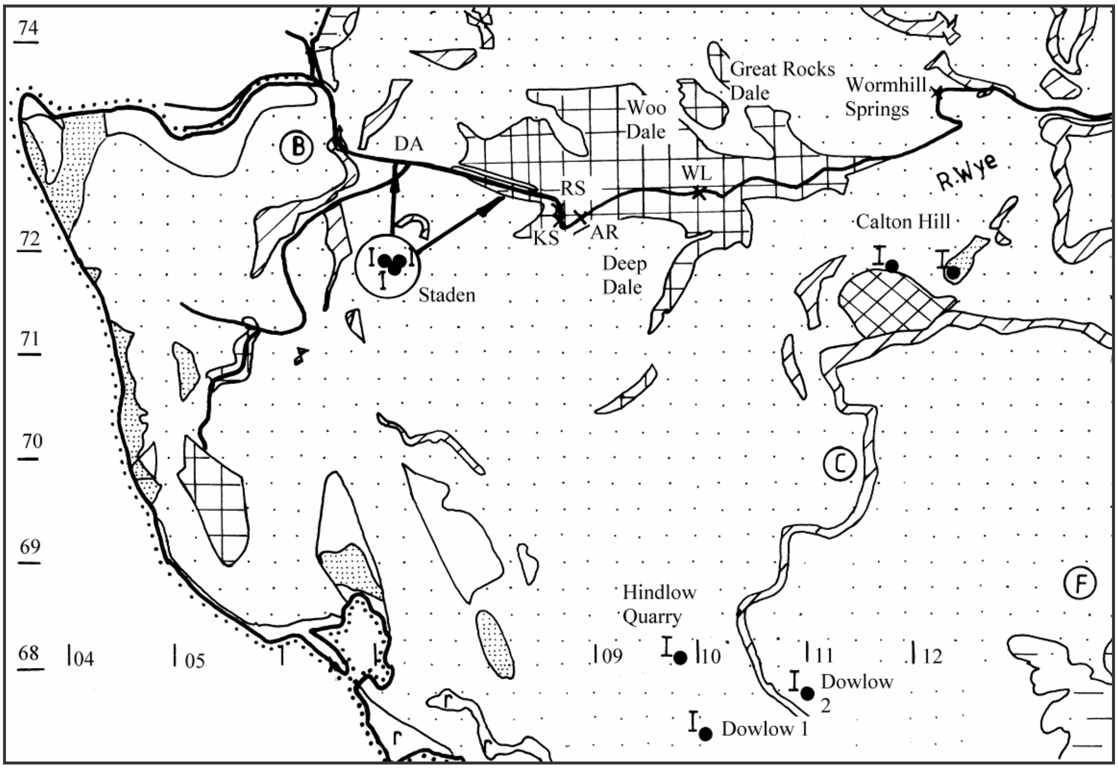
22 June to 17 September 1999 was a period of falling groundwater levels, as indicated by the drying up of many of the springs. Recovery beginning during the middle of September (evidence for which comes from the groundwater levels recorded in the Highcliffe Farm Borehole (Chapter 8)), thus dye could have reached sites during October/ November 1999.

iii) Tracer travelled to one or more of the monitored sites in concentrations diluted below the limit of detection in an extremely large volume of groundwater.

iv) Tracer was transported deep into the aquifer.

v) Tracer was adsorbed onto siliceous sediments and clays and, in the case of rhodamine WT, onto the materials used to construct the soakaway.

This author considered another possibility, namely that storage could takes place within the mineral vein. It is quite likely that the vein was worked to at least 150 m depth (approximately to spring level). Therefore, taking a 4 km-long rake of 2 m width, with 25% void, the total dye and flush input being approximately 13500 litres the rise in water level would be less than 1 cm. It has also been noted by this author that Brushfield Spring does not appear to have been monitored, however it is understood (Gunn personal communication 2003) that this is because Brushfield Spring was dry throughout the monitoring period, a period of low groundwater levels. The chemistry of this spring, with elevated chloride and nitrate, does indicate that it could receive contaminated water, for which the most obvious source would be a contribution from the area of Calton Hill landfill site. It is possible that the link with Brushfield Spring would be via the mineral vein. Nevertheless, this hypothesis has been rejected because the River Wye was being monitored at this point and there was no evidence of any dye emerging into the River Wye. Accordingly, the interpretation favoured by this author is that vertical flow paths are dominant in the Chee Tor Limestone Member of the Bee Low Limestone Formation, and the tracer has been transported to the boundary between the Woo Dale Limestone and the overlying Bee Low Limestone formations, or into the Woo Dale Limestone (Limestone Research Group point iv). Based on the current hydrogeological setting it would seem most likely that this groundwater takes a southeasterly route, towards the Taddington Anticline and Magpie Sough. The Bee Low Limestone/ Woo Dale Limestone formational boundary lies below the river level and the groundwater, as indicated by the findings of water-tracing carried out in the Lathkill catchment becomes increasingly confined beneath the Lower Miller's Dale Lava (Figure 7.6, p. 136).



Key: I Injection point. B Buxton; C Chelmorton; F Flagg; DA Devonshire Arms; KS Kidtor Spring; RS Rockhead Spring; Woolow Spring. Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone. Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent.

Legend:

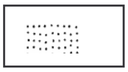
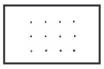
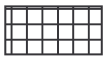
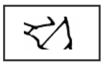
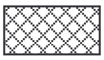
Head		Head Monsal Dale Limestone Formation	
Bee Low Limestone Formation		Woo Dale Limestone Formation	
Interbedded lavas		Dolerite	

Figure 7.12: Location of water-tracing tests at Calton Hill, Staden, Hindlow Quarry and Dowlow Quarry (Appendix 7.1).

7.3.6 The watershed between the River Dove and River Wye to the southeast of Buxton, Derbyshire (Appendix 7.1, Gunn et al., 1999).

On 12 February 1999 tracers were injected into three boreholes: Hindlow Borehole (rhodamine WT), Dowlow Borehole 1 (fluorescein) and Dowlow Borehole 2 (Tinopal ABP). Monitoring of the sites continued until 15 November 1999. During this period there was no convincing tracer recovery from any of the monitoring sites (Gunn et al., 1999). It was calculated (Gunn et al., 1999) that most of the dye injected into the Hindlow Borehole and Dowlow Borehole 1 entered the aquifer. However, some dye remained within the borehole and it was found that the dye concentrations did not decrease throughout the monitoring period. Groundwater levels were not determined during the monitoring period. The constancy of the tracer concentrations indicates that groundwater levels either remained constant, or fell during the monitoring period. It can be concluded that there was no, or very little, additional groundwater targeting the boreholes i.e. the boreholes were hydrologically isolated within the Bee Low Limestone Formation. The concentration of tracer in Dowlow Borehole 2 was not determined, because of the narrow diameter of the dip tube. However, groundwater hydrographs obtained from the Environment Agency (Chapter 8) show seasonal fluctuations in groundwater levels throughout the limestone. Although discharge is not measured, the seasonality indicates at least minimal flow of groundwater. Structural guidance would take groundwater in a southsoutheasterly direction. This author opines that there is groundwater storage associated with the Cronkston-Bonsall Fault (Chapter 9) and therefore would suggest that groundwater flow is more likely to be marginally higher during low groundwater conditions.

7.3.7 **Staden** (Appendix 7.1, Appendix 7.3).

On 25 August 2002 the Limestone Research Group injected fluorescein into the Chee Tor Limestone Member via an excavation for a soakaway at the Rockhead Spring water bottling plant (SK 07187194). Dye was recovered from: the River Wye at Cowdale (700 m/day) and Topley Pike (1450 m/day), but not from Cowdale (Rockhead) Spring (Figure 7.12, p. 148). The recovery at Topley Pike is suspected to be via fracture/channel discharge further upstream in the River Wye. It is considered most likely that this occurs at the Bee Low Limestone/ Woo Dale Limestone boundary (as indicated by the groundwater level in the Staden Borehole (Figure 8.12)). Monitoring continued until 31 October 2002 and detection of dye continued in the River Wye at Rockhead throughout this period. Dye was not detected in the River Wye upstream of Lover’s Leap (SK 0717275). However it should be noted that this test was carried out during a period of low groundwater levels (the groundwater level in the Bull I’ Th’ Thorne Borehole (Chapter 8) was at approximately 234.4 m OD). This author walked along the stretch of the River Wye between Cowdale (Rockhead) Spring and Buxton Sewage Treatment Works. The river is concrete lined upstream of Cunning Dale. An operative at the Buxton Sewage Treatment Works, Alec Mamczur (personal communication, 1 May, 2003), who had been living in the area for about 50 years, suggested that this is to prevent loss of water from the bed of the river and it may be covering an old mill wheel pit. He also noted that it was only from the preceding 9 to 10 years that there had been completely dry periods in Lover’s Leap. Prior to that there was always “at least a trickle of water to be seen in the valley”. A colleague (Chris Hustler) identified a spring in the River Wye at the Devonshire Arms Public House, in the order of 1100 m to the northeast of the injection point. A fluocapteur was positioned at this location and significant peaks were recorded on 16 and 30 October 2002.

In order to confirm these findings a second dye trace was carried out by the Limestone Research Group, which commenced on 24 March 2003. 5 litres of 40% fluorescein dye was added to one trial pit (situated at SK 07087191) and 6 litres of rhodamine WT was added to another trial pit (situated at SK 07137190). Fluorescein dye was recovered from: the River Wye upstream of Lover’s Leap, the River Wye upstream of the Devonshire Arms, the River Wye at the Devonshire Arms, the spring at the

Devonshire Arms and the River Wye upstream of Cowdale (Rockhead) Spring. Traces of fluorescein dye were also suspected to be recovered from Lover's Leap, Cowdale (Rockhead) Spring and Woolow Spring. The latter were all recorded on 27 March, 2003. Small quantities of rhodamine were recovered from the River Wye upstream of the Devonshire Arms, at the Devonshire Arms, upstream of Cowdale (Rockhead) Spring and at Cowdale Spring. It was suspected that rhodamine was also recovered from the Devonshire Arms Spring, and possibly Lover's Leap. However, it should be noted that considerable difficulty was experienced differentiating peaks during this test (Appendix 7.3). The occurrence of dye in the River Wye upstream of Lover's Leap during this experiment may be attributable to the higher groundwater conditions (the groundwater level in the Bull I' Th' Thorne Borehole (Chapter 8) was at approximately 258.8 m OD, but falling). The findings, like those of the Illy Willy Water trace, demonstrate the difficulties that the Limestone Research Group has experienced with the successful use of this dye.

Consideration of these results (Figure 7.12, p. 148) and examination of British Geological Survey Sheet 111 suggest to this author that the dye that reaches the River Wye upstream of Cowdale (Rockhead) Spring rises on a continuation of the fault upon which the Cowdale (Rockhead) and Ashwood Dale springs rise. The time taken for dye recovery suggests fracture flow, possibly guided by the boundary between the Woo Dale Limestone and the overlying Bee Low Limestone, or associated with another inception horizon near to the top of the Woo Dale Limestone Formation. The recovery of dye from upstream of Lover's Leap indicates northwesterly, fracture-guided, flow vectors, which are likely to be vadose flow paths. Groundwater monitoring carried out during the monitoring of the water-tracing test established that water levels in the borehole at Rockhead Spring water bottling plant appear to be independent of the discharge at Cowdale (Rockhead) Spring, which suggests that faults locally act as groundwater divides.

7.3.8 Conclusions drawn from water tracing experiments carried out in the Wye Sub-basin.

The lithological guidance of the hydrogeology described in the Lathkill catchment appears equally significant in parts of the Wye catchment. The evidence from the tracing experiments in this area highlights the potential for significant conduit development in the Woo Dale Limestone and at the boundary between the Woo Dale Limestone Formation and the overlying Bee Low Limestone Formation. It is also important to note that the findings appear to confirm what might be suspected from visual examination of the geology map. This is that even though the springs at the Devonshire Arms, Cowdale [Rockhead] and Ashwood Dale resurgences appear to lie on the same fault, they also appear to be fed by independent conduits probably at lower stratigraphic horizons in the Woo Dale Limestone Formation in an easterly direction. At the western end of the catchment, in the area of Dove Holes, flow rates to Wormhill Springs suggest the possibility of a mature karst system, possible even of cave proportions, parallel to and to the north of the river at a level that is likely to be below 250 m OD. Edmans (1988) described vadose (overflow) paths linking sinks on the western edge outcrop of the limestone and to the west of Buxton. It is suspected by this author that these are attributable to inception horizon-guided conduit development in the northerly dipping Miller's Dale Limestone

Formation, with which Poole's Cavern is also associated. The likely processes associated with inception horizon development in the Miller's Dale Limestone Member are described in Chapter 4. Evidence provided by Edmans (1988) indicates that there is deeper phreatic flow between these sinks and Wye Head Resurgence and Otter Hole. There is also evidence to suggest that groundwater rising from Shay Lodge at Dog Holes is confined (Gilman, 1985).

Results from the various tracing experiments appear to confirm that conduit development is more limited in the Chee Tor Limestone Member (Chapter 6). Permeability in the Chee Tor Limestone results from the intensity of fissuring (Chapter 6, Ineson and Dagger, 1985). It is likely that surface recharge is focused and that further internal focusing (Pruess, 1999), particularly on the boundary between the Chee Tor Limestone Member and the underlying Woo Dale Limestone Formation, occurs at depth and similarly, that groundwater inputs to the Chee Tor Limestone are guided by vertical flow paths, whereas horizontal flow predominates in the Miller's Dale Limestone Formation and in the Woo Dale Limestone Formation.

The dominance of structurally-guided (Chapter 9) fracture flow, which encourages vertical flow paths and relative paucity of conduit development in the Chee Tor Limestone Member, has been noted. This is attributed to its massiveness and its relative purity, which increases in a westerly direction (Sadler, 1964) and reflects the relative absence of the shallow-water facies of the cyclic regressions (Appendix 4). Clearly, the fracture pattern reflects the local structure, with fractures tending to be more open in anticlinal settings, reflecting the extension within the upper surface of each bed. In the area of Tunstead, Ineson and Dagger (1985) describe a number of east-west tensional faults (Figure 7.9, p. 143). These lie towards the northern edge of the Woo Dale Fault block (Chapter 4). Wormhill Springs Main Rising is associated with the eastern extent of this fault block. The variation in the dip of the bedding, at the point at which the northwest to southeast-trending fault tends to an east to west trend at its northern end, provides further evidence of basement influence. It is considered most likely by this author that Wormhill Springs East Rising is encouraged by the paucity of conduit development in the Chee Tor Limestone, which is downthrown to the east, bringing conduits or inception horizons in the Woo Dale Limestone against the Chee Tor Limestone, such that the groundwater rises up the northwest to southeast jointing. The top of the lowest borehole used in the Tunstead Quarry tracer experiment was 260 m OD. Although groundwater levels were not monitored during the test, it can be surmised that flow in the Woo Dale Limestone was below this level.

By contrast the Woo Dale Limestone shows great variation in purity. Ineson and Dagger (1985) describe a number of features, in the form of lithological variations that could be associated with preferred zones of conduit development in the upper beds of the Woo Dale Limestone. These include: the occurrence of cavities, which they attribute to the shrinkage that accompanies dolomitization; the occurrence of three or four shell beds, characterised by bands of *Daviesiella* sp. with *D. llangollensis* in the higher bands; and alternations of dolomitized limestone and pure limestone, which has been shown to occur from a depth of 60 m below the Chee Tor Limestone in the Tunstead area. Schofield and

Adams (1985) noted the occurrence of coal in the Woo Dale Limestone, another potential focus for inception (Chapter 2).

At the western end of the catchment to the south of the river, northnortheasterly and easterly vectors have been established by water-tracing. Surface watercourses in this area appear to show strong structural guidance (particularly fault and fracture-related guidance), which is discordant with the hydraulic gradient indicative of recent (Pleistocene) development. Interestingly, the explored passages of Poole's Cavern have a predominantly northerly trend, with easterly development. This is discussed further in Chapter 9. However, in the northwestern edge of the Wye catchment it has been reported by Downing et al. (1970) that groundwater entering the Dove Holes railway tunnel drainage water is drained to the northwest via the Mersey Basin. Further evidence for this comes from water-tracing carried out in February, 1985 (Gunn, personal communication 2004). This is effectively a local reversal of the hydraulic gradient. The observations with respect to groundwater levels falling in the area of Lover's Leap suggest that there may be a lowering of the piezometric surface to the west of the limestone. Farther to the east, northwesterly and northeasterly flow vectors have been identified from the Illy Willy Water and Staden tracing experiments. This is attributed to the dispersed nature of the fracture-related flow that has developed on inception horizons. It may also reflect a localised reversal of the local catchment-scale hydraulic gradient, attributable to the deepening of the River Wye during the Anglian glaciation (Chapter 9).

It is interesting to note that the dye loss associated with the A515 quarries (7.3.6) and also with Calton Hill (subsection 7.3.5) potentially indicates the significance of the faults and associated anticlines (in this case Long Dale, the Cronkston-Bonsall Fault and the Taddington Anticline, the Taddington Fault) as significant zones of groundwater storage. Further evidence for this comes from the work of Oakman (1979), which identified that there was significant fault storage associated with the Cronkston-Bonsall Fault.

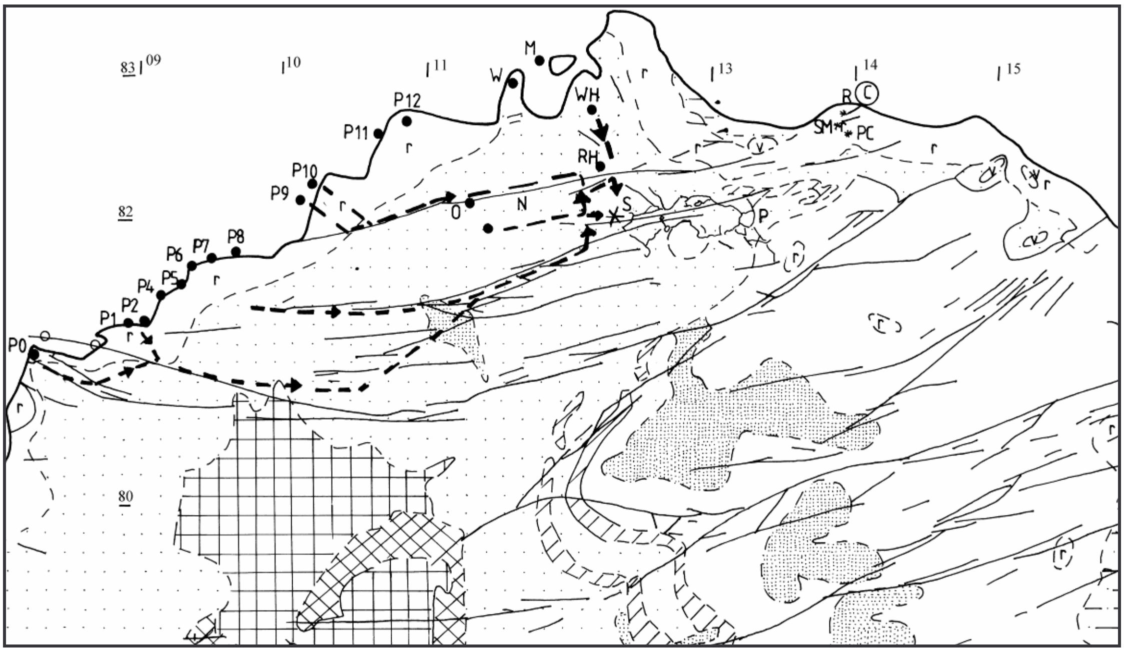
7.4 Peripheral sub-basins.

7.4.1 The Peakshole Water Sub-basin.

Gunn (1998) has summarised the findings of work carried out in the northern part of the Peakshole Water sub-basin (Figures 7.13 and 7.14, adapted from Gunn, 1998). Most of the flow in this sub-basin is captured by three springs associated with Peak Cavern, namely: Peak Cavern Rising, Russet Well and Slop Moll Rising (Figure 7.13), in the marginal reef facies of the limestone. Many of the sinks on the western edge of the limestone, where the Namurian strata rest unconformably against the Dinantian limestone, are connected. P0 and P1 flow to Coalpit Hole Rake, to the north, P2 to P8 converge on New Rake, and P9 to P12 flow to Faucet Rake. All of the sinks drain to two major inlet sumps in Speedwell Cavern, namely Main Rising and Whirlpool Rising (Gunn, 1998). Ford and Rieuwerts (2000) observed that Hollandtwine Mine was once nearly 190 m deep with drainage via a natural "swallow", resurging in Peak Cavern (Gunn, personal communication 2006 advised that this is via

Lake Passage). Gunn (1998) also noted that there are five other caves that collect dispersed flow and drain to Speedwell Cavern, these are Nettle Pot, Rowter Hole, Winnats Head Cave, Blue John Cavern and Treak Cliff Cavern. Gunn (1998) reported that tracing experiments carried out in the area of Eldon Hole (SK 11618093) also identified connections with Speedwell.

Whereas the presence of alluvium in Perry Dale and Dam Dale might indicate at least surface movement of water in a southerly direction, it is clear from the description above, that the groundwater flow is to the east. It is likely that the surface drainage marks the location of a significant glacial meltwater drainage channel. The evidence suggests that in this part of the White Peak the mineralized veins, in particular the western end of Dirtlow Rake (Coalpit Hole Rake), act as aquitards to groundwater, restricting any southerly flow. Examination of the topography reveals that with an elevation of approximately 140 m OD the River Noe is more likely to form the local base-level, thereby reinforcing the easterly hydraulic gradient of the Peak-Speedwell system. It would seem that the presence of the mineral veins, supported by human impacts (artificial drainage of the mine workings) has facilitated underground capture of southerly drainage on the western side of the White Peak. Further evidence to support this hypothesis comes from the investigatory work carried out by Professor Shotton (DRO D3040 L/W1/44 92-94). This established a groundwater level of approximately 242 m OD in Hazard Mine and 274 m OD in Coalpit Hole No. 10 (at the western end of Dirtlow Rake), whereas the level of Russett Well and Peak Cavern Resurgence are approximately 187 and 188 m OD respectively. In a letter to Mr Thompson of the Derwent Valley Water Board, dated 21 November, 1951 Professor Shotton wrote "*... The Coalpithole level holds true even when only a million gallons a day is coming from the Peakshole Water. If we had a water table controlling this outflow in any way resembling the normal situated ground of the chalk or the Bunter, we should have to assume that the outflow was a function of the gradient of the water table. Yet we know that Peakshole can increase its yield sixty-fold without a big change in the level at Coalpithole. There is no fluctuation of water table commensurate with the oscillation in outflow.*" Thus it would seem that the discharge is largely controlled by the permeability offered by the Peak-Speedwell system. Ford (2000) has also commented on the significance of vein cavities as a focus for groundwater. Gunn (1998) has described how the hydrological regime of the Peak-Speedwell system changes in response to the groundwater conditions, noting that under low to moderate flow conditions Peak Cavern Rising discharges only autogenic recharge, but that during high flow conditions water backs up in Speedwell such that water enters higher level conduits in Peak cavern resulting in an allogenic component to the discharge at Peak Cavern Rising.



Key: C Castleton; M Mam Tor Swallet; N Nettle Pot; O Oxlow Cavern; P and S Peak-Speedwell Cavern system; RH Rowter Hole; W Windy Knoll Cave.
Asterix marks resurgences: PC Peak Cavern Rising, R Russett Well, SM Slop Moll. Shaded circles stream sinks. Mineral veins shown.
Heavy line marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent.

Legend: Head — Eyam Limestone Formation

Monsal Dale Limestone Formation — Bee Low Limestone Formation

Woo Dale Limestone Formation — Interbedded lavas

Dolerite

Figure 7.13: Results of groundwater tracing in the Peakshole Water sub-basin (after Gunn, 1998).

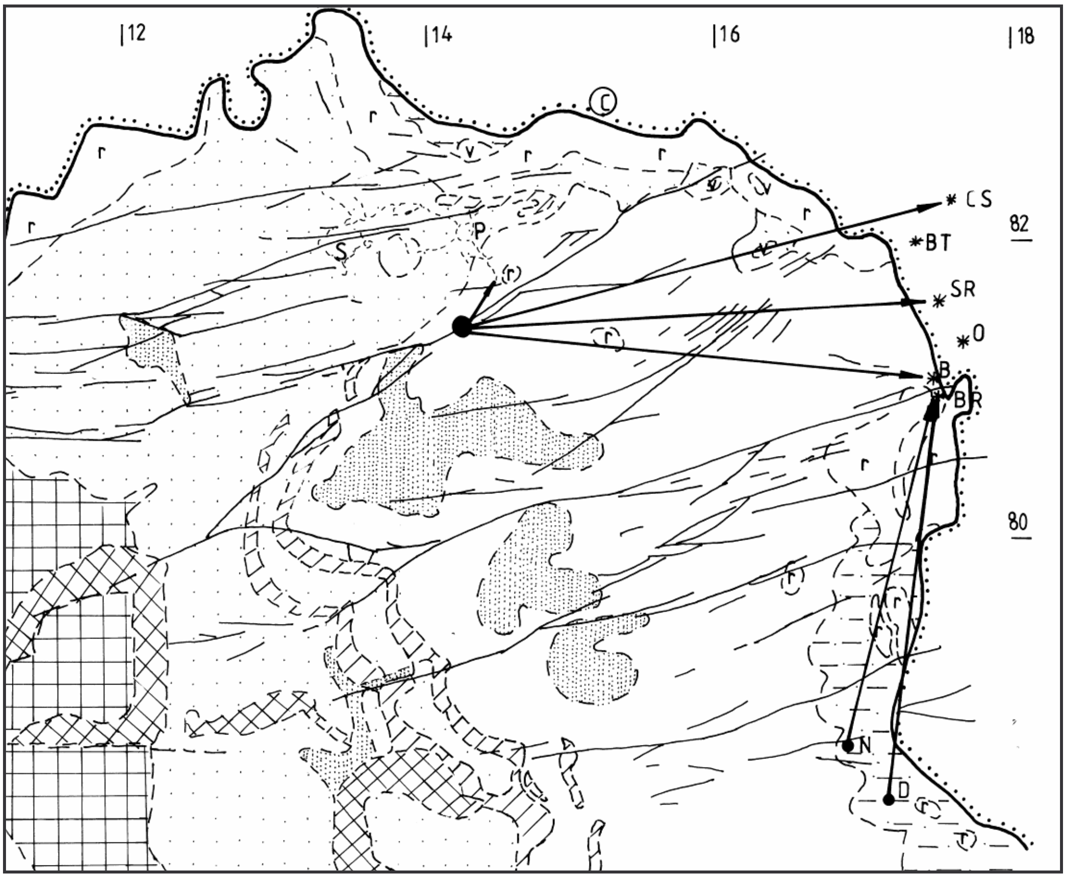
Although the differentiation of the Chee Tor Limestone Member and the Miller's Dale Limestone Member is not always made by the BGS (Sheet 111) it is very interesting to note that the Peak-Speedwell cave system (indicated approximately on Figure 7.13) is most likely to fall within the Miller's Dale Limestone Member. This system drains to the northeast, which suggests that at the time of formation there was a lower base-level than that of the River Derwent in a northeasterly direction. With respect to the depth of flow, Beck (1980, p. 87) observed that *"... the western reaches of the Peak/Speedwell cave system may penetrate the S_2 beds* (Woo Dale Limestone Formation*), as may the deepest parts of the Blue John Cavern, the Giant's Hole/Oxlow Caverns Complex and Nettle Pot*." Observations made in Chapter 4 of preferential zones of karstification in the Woo Dale Limestone Formation suggest that this will be the case.

Whilst considering the geology in this area, which lies outside the main focus of the thesis, it is also interesting to note the presence of mineralized veins in the Chee Tor Limestone Member (absent in this Member farther south), which suggests to this author that the former shale cover was only tens of metres above the existing ground surface at these locations (Chapter 9) during mineralization. Furthermore, Ford (1977) has recorded the presence of neptunian dykes in the reef limestones of the area, therefore it follows that fractures in the Chee Tor Limestone Member would have been more open in this zone of uplift, thus facilitating early integration of surface water recharge with the inception horizon-related karst.

Further evidence that Dirtlow Rake forms a groundwater divide lies to the south. Tracer investigations of Dirtlow Rake (SK 14258133 and 13478125) have proven connections both with the Peak-Speedwell system and also with soughs discharging into Bradwell Brook and Springwell Resurgence (Gunn 2000b and Figure 7.14). Thus it would appear that some easterly flow occurs via the Miller's Dale Limestone and the Monsal Dale Limestone and is released via the soughs that drain to Bradwell Brook, a northerly flowing tributary of the southeasterly flowing River Noe, a tributary of the River Derwent. However, tracer experiments have not proven any connection with either Bagshawe Resurgence (SK 17408110), or with the thermal spring at Bradwell.

7.4.2 The Bradwell Brook Sub-basin.

At the upstream source of Bradwell Brook is the Bagshawe Resurgence (approximately 183 m OD), which is fed by water from Bagshawe Cavern. The results of water-tracing tests (Figure 7.14) used to define this sub-basin have been reported by Gunn (1998) and Whitehead (2000). The Bradwell Brook sub-basin is, at least in part, perched in the Monsal Dale Limestone Formation, above the Derwent and the Wye sub-basin, as indicated by Downing et al. (1970) and Professor Shotton (DRO D3040L/W1/44/94). In a letter (22.9.52) to Mr Thompson of the North Derbyshire Water Board, Professor Shotton reported the groundwater levels in the Hucklow boreholes (1 to 7) to be in the order of 261 m OD to 204 m OD. By contrast, the presence of warm water in Moorwood Sough and Stoke Sough is indicative of deeper, probably confined water rising from beneath the volcanic deposits and the Chee Tor Limestone Member. The natural outlet for the perched water is Bagshawe Risings. The situation appears to be very similar to that of Lathkill Head Cave in that the cave system appears to be formed largely within the Monsal Dale Limestone Formation. However, its northerly flow vector, which is closely associated with the position of the former Namurian cover, appears to confirm the observations made with respect to the Peakshole Water sub-basin, that there would appear to have been a lower base-level than the Derwent lying to the north of the limestone outcrop at the time of inception.



Key: Shaded circle Injection point. C Castleton D Duce Hole N Nether Water Swallet P Peak Cavern system S Speedwell system. Asterix marks resurgences: B Brookside Sough Tail; BR Bagshawe Resurgence; BT Bradwell Thermal Spring; CS Stream from Krondstadt Sough; O The Ooms Resurgence; SR Springwell Resurgence.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent. Mineral veins shown.

Legend: Head — Eyam Limestone Formation

Monsal Dale Limestone Formation Formation — Bee Low Limestone

Woo Dale Limestone Formation — Interbedded lavas

Dolerite

Figure 7.14: Peakshole Water and Bradwell Brook sub-basins, showing water-tracing within the Bradwell Brook sub-basin (after Gunn, 1998).

Gunn (1998) described the poor definition of the boundary between the Bradwell Brook sub-basin and the Wye sub-basin to the west. Beck (1980) suggested that the boundary follows the Litton Tuff and, where that peters out, the Miller's Dale Lava. Certainly for the perched water this author concurs. However, flow vectors in the groundwater beneath the Chee Tor Limestone Member are likely to differ, with a southeasterly or easterly vector being considered most likely, as indicated by the location of the thermal springs and the thermal water encountered in Stoke Sough (SK 239769). The mineral veins that in part appear to provide storage for the lower body of groundwater complicate the situation. The path from Duce Hole (SK 18697746) to Bagshawe (Figure 7.14) was reported to be via Milldam Mine (SK 175813, Hucklow Edge Vein) and dye is reported to have continued to emerge over a prolonged period of time (Whitehead, 2000), which was interpreted as being indicative of storage within a considerable body of water (Whitehead, 2000). This author agrees that there is likely to be significant storage associated with the mineral veins, but has also observed that flow rates are also influenced by the formation bearing the predominant flow path. For example, lower flow rates appear to be associated with the Monsal Dale Limestone and Miller's Dale Limestone and Whitehead (2000) from tracings injected at Cartledge Farm (SK 179778) determined flow rates of 19 to 20 m/hour to Bagshawe Resurgence (SK 174809) and the Ooms (Bradwell Brook, SK 176815). Consideration of these results indicates the possibility that stored groundwater associated with mineral veins (and or fault zones) feeds inception horizon-related channels.

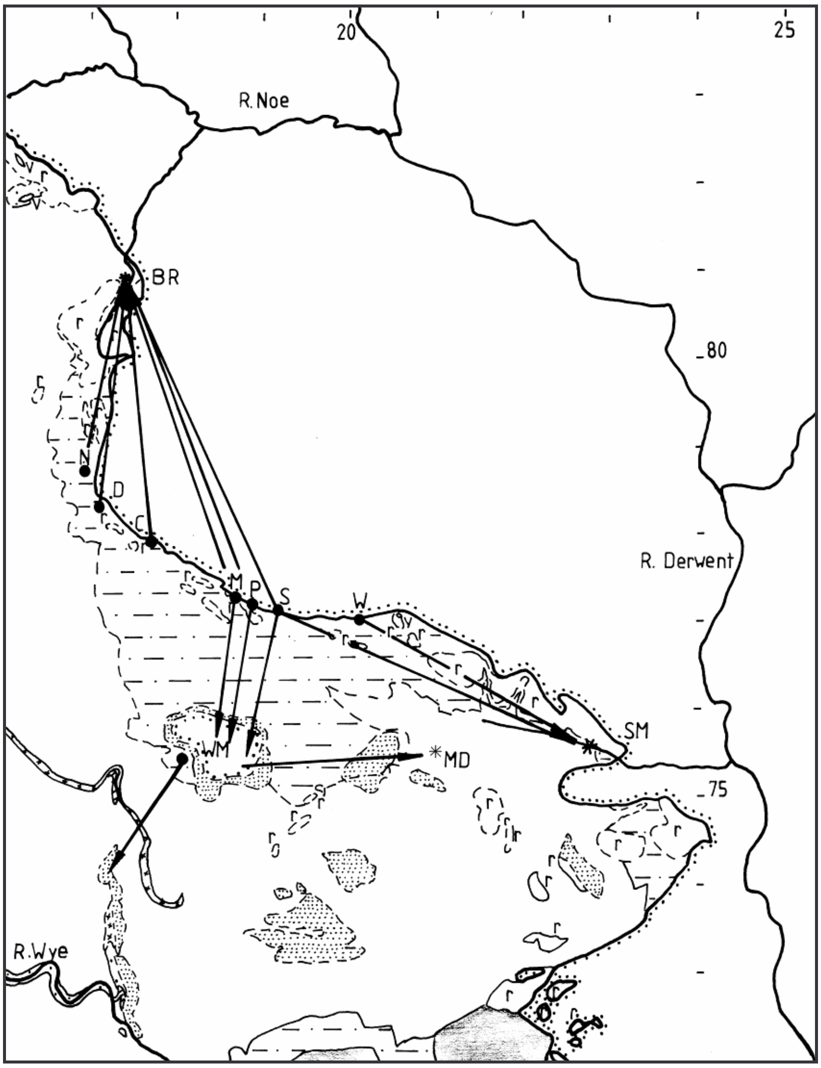
7.4.3 The Stoney Middleton Sub-basin.

The groundwater divide between the Bradwell and Stoney Middleton sub-basins would appear to lie beneath Stanley Moor (to the west of Foolow), an area that has a relatively low density of mineral veins (with less potential for drainage). Work carried out by Whitehead (2000) has confirmed connections between Waterfall Swallet (SK 199770) and Wardlow Mires (SK 178757), Cressbrook Resurgence (SK 174728) and Moorwood Sough (SK 231754), and a possible connection with Watergrove Sough (SK 211758). Further to this, in a letter to Professor Shotton dated 3 March 1952, Mr Thompson (DRO D3040L/W1/44/92-94) reported that Edwin Maltby, the son of a former Glebe Mines Manager, had described how prior to the working of Glebe Mine, water disappearing at Waterfall Swallet followed a natural underground river passing near the Glebe Mine and emerging into Middleton Dale about "*150 yards below the junction of the road leading to Eyam*" (Carlswark Cavern, Gunn personal communication, 2006) and that the underground stream, which was encountered in Glebe Mine, was diverted to Moorwood Sough. The Carlswark Cavern complex has been described by Beck (1973). It comprises a network of phreatic channels at the base of the silicified Lower Shell Bed in the Monsal Dale Limestone Formation. At a lower stratigraphical level, but still in the pale facies of the Monsal Dale Limestone Formation, the Lower Complex is associated with rifting, with a long deep sump at the base of the rift.

Whitehead (2000) also identified a weak connection between Piece End Swallet (SK 189773) and Cressbrook Resurgence as well as connections between Piece End Swallet and: Watergrove Sough, Moorwood Sough, Stoney Middleton Thermal (SK 232756) and The Ooms (SK 176815). However there was no clear evidence of a connection between Piece End Swallet and Wardlow Mires (Figure 7.15, p. 157). The in-situ groundwater quality data presented by Whitehead (2000), more specifically the electrolytic conductivities and temperatures, indicate that there is a component of deeper groundwater entering Moorwood Sough and that the groundwater at Cressbrook Resurgence, which is at the junction of the Lower Miller's Dale Lava and the Bee Low Limestone Formation, at least in part, rises from the Woo Dale Limestone Formation. The connection with the Ooms suggests that Piece End Swallet lies on the groundwater divide between the Bradwell and Stoney Middleton sub-basins. Whitehead (2000) recorded temperatures consistently in the order of 9.7 to 9.8° C for the Ooms. This suggests to this author that there is a component of thermal groundwater associated with this spring. Thus it is likely that the flow path between Piece End Swallet and Bradwell is deep-seated. Furthermore, it is suspected that there is a northnortheast-trending fault, thrown down to the east (which has not been mapped), with which Bradwell Brook is associated. The former thermal baths at Navio (SK 181827) would also fall on the postulated alignment of the fault, as would the Bradwell thermal rising (SK 17398114). The sink at Wardlow Mires (SK 178757) is an estavelle, overflowing to Cressbrook Dale during periods of high groundwater conditions. It is interesting to note that tracer tests that have been reported by Whitehead (2000) indicate mixing of groundwater from the Linen Dale swallets (focused on Waterfall Swallet SK 19887705) with the thermal water rising at Stoney Middleton. Clearly this is likely to impact on the temperature of the rising.

Gunn (1990) carried out water tracing in the area of Darlton Quarry (SK 217757). Here upper and lower aquifers separated by the Litton Tuff, or its associated wayboard, have been proved. Fluorescein was injected into the lower aquifer and rhodamine into the upper aquifer, via boreholes. Upper aquifer connection with Hawkenedge Well (SK 21567582), at the time of tracing thought to be a natural spring on a fault in the northeasterly dipping Monsal Dale Limestone, to the west of Dale Brook was proved. Gunn (1990) noted that Hawkenedge Well is known to discharge large concentrations of suspended sediment on occasions. Whilst this has been attributed to sediment associated with Victory Shaft, the occurrence of a large area of head in the vicinity of Burnt Heath (SK 203753) is likely to be associated with dolines, thereby providing an additional source for the sediment. However, it should also be noted that Kirkham (1968) has referred to Hawkenedge Well as Oakenedge Sough. Flow vectors in the lower aquifer were found to target Dirty Rake Victory Level, Moorwood Sough and also Dale Brook (but not via Hawkenedge Well). This suggests an easterly hydraulic gradient for the confined groundwater, its flow being impeded either as a result of faulting, or mineralization associated with the Dirty Rake. It would also appear that the valley-forming processes associated with Dale Brook are associated with an easterly extension of the west to east-trending mineralized fault that is shown on the British Geological Survey Sheet 99 *Chapel en le Frith*. Furthermore, the evidence suggests to this author that the hypothesised fault is associated with rising underflow. Calculated velocities of flow in the upper aquifer were as high as 350 m/hour, whereas those in the lower aquifer were in the order of 10 m/hour.

A number of interesting points arise from these results, when considered in the geological context. The setting of Dale Brook is comparable with that of the River Lathkill. From the description presented by Beck (1973), comparisons can be made with Lathkill Head Cave in terms of inception horizon-related conduit development. Furthermore, northerly migration of the brook appears to have been restricted by the reef deposits immediately to the north of the brook. Watersaw Rake appears to form the northern limit of the Ashford Basin (Gutteridge, 1987; Chapter 2). Examination of the distribution of the strata and the absence of the Litton Tuff to the south of the Watersaw Rake also appear to indicate a further intra-block basin centred on the area of Wardlow Mires. Support for this comes from the observation that the Eyam Limestone Formation is shown to be horizontally bedded, although the dip of the Monsal



Key: Shaded circle Injection point. C Cartledge Farm Swallet D Duce Hole M Mrs Smythes Swallet N Nether Water Swallet P Piece End Swallet S Swevic House Swallet W Waterfall Swallet. Asterix marks resurgences: BR Bagshawe Resurgence; MD Middleton Dale; SM Stoney Middleton; WM Wardlow Mires Inlier. Stoke Sough not shown.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent.

Legend: Head — Widmerpool Formation

Eyam Limestone Formation — Monsal Dale Limestone Formation

Bee Low Limestone Formation — Woo Dale Limestone Formation

Interbedded lavas

Figure 7.15: Results of water-tracing: Stoney Middleton Sub-basin (Appendix 7.1).

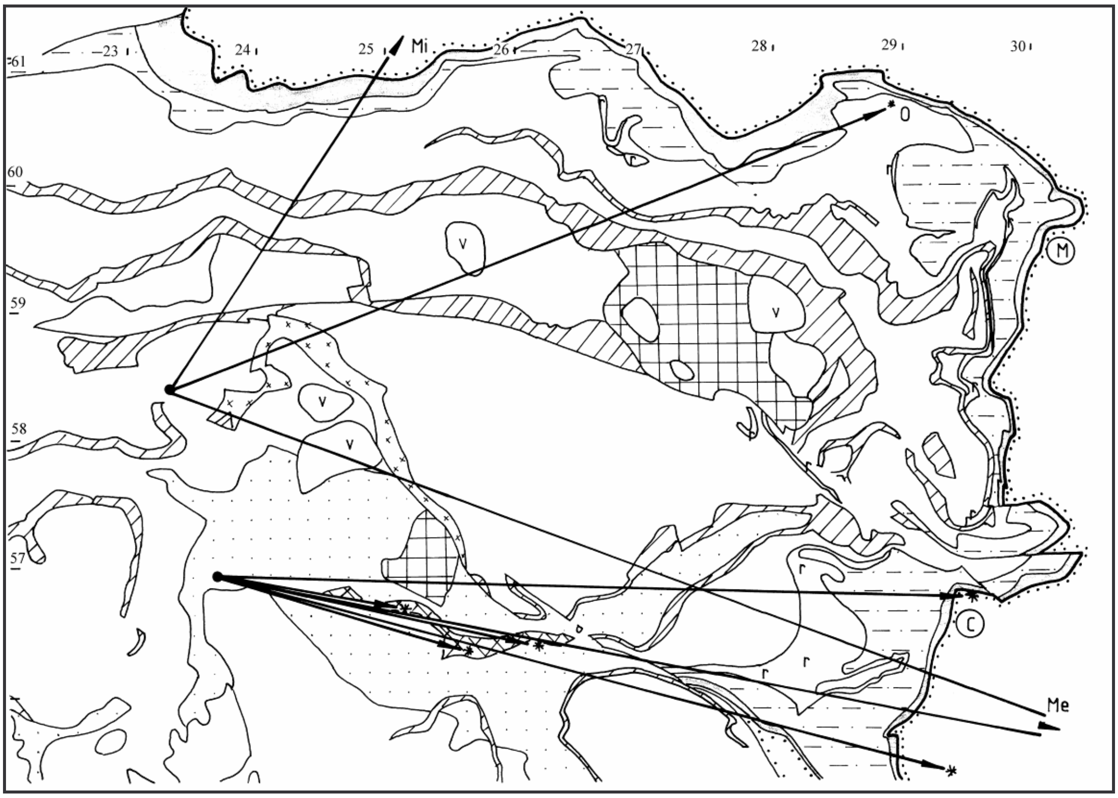
Dale Limestone Formation is more variable, but broadly appears to be to the north. If a basinal structure is assumed, it is likely that the Monsal Dale Limestone Formation to the north of Wardlow Mires dips to the south, thus facilitating inception horizon-related channel flow in the Monsal Dale

Limestone. It would seem most likely that the easterly directed flow is guided by the mineral veins. The northern boundary of this basin is likely to be associated with Dirtlow Rake (sub-section 7.4.2). Connections between Cartledge Farm Swallet, Mrs Smythes Swallet and Piece End Swallet appear to represent strike-oriented flow. Hucklow Vein and Watersaw Rake are likely to form significant zones of groundwater storage (comparable with Dirtlow Rake to the north, sub-section 7.4.2), thereby contributing to the response of the Wardlow Mires estavelle. The significance of the Hucklow Vein in terms of storage is supported by the requirement for the construction of Stoke Sough, which was driven west from the Derwent Valley in approximately 1720; likewise the importance of Watersaw Rake is evident in the considerable difficulties experienced in attempting to dewater this rake (Rieuwerts, 1987).

7.4.4 The Derwent South (Matlock) Sub-basin.

Water tracing experiments carried out in the Derwent South sub-basin (Hardwick and Gunn, 1999 and Hyland et al., 1994) appear to provide further support for the hypotheses presented for the Wye and Lathkill sub-basins. However, it is noted that the Hopedale Limestone (southerly equivalent of the Bee Low Limestone Formation) is not subdivided in the same way and the influence of the Chee Tor Limestone Member cannot be claimed to prevail. Dyes introduced into boreholes 1 (SK 23875706), 5 (SK 23575693) and 7 (SK 23505683) in Brassington Quarry, in February 1999 (Hardwick and Gunn, 1999), were injected into the Bee Low Limestone Formation. Flow vectors to the east and southeast were identified (Figure 7.16). Proximal risings (SK 25115669, SK 25625638 and SK 26155641), were from the Woo Dale Limestone, more specifically a porcellanous limestone band that was formerly referred to as the Griffe Grange Member (BGS Sheet 112). The more distal risings (SK 29515678 and SK 29345690) appear to be associated with northwest to southeast faults, downthrown to the south. These faults are associated with the Cronkston-Bonsall Fault. Traces of dye were detected in Meerbrook Sough (SK 32795520), which indicates that some of the confined water in the Woo Dale Limestone rises into the mineralized faults, in particular Gang Vein is suspected. Water-tracing in Meerbrook Sough (Hardwick et al., 1996) established peak tracer velocities (lithium chloride) of 0.20 to 0.27 m/s (~17280 to 23300 m/day) in the sough. Lower velocities were attributed to delays resulting from flooded shafts and tributary routes, with secondary peaks attributed to sough blockages such as winze gates, or roof falls.

The results for Brassington Quarry (SK235570) can be contrasted with those for Ivonbrook Quarry (SK 23325842), which has been excavated into the Monsal Dale Limestone Formation. From this quarry the dominant flow vector was found to be towards Meerbrook Sough (Hyland et al, 1994). Traces of dye were also determined at Oxclose Sough (SK 28906061) and Millclose Sough (SK 26446256). Comparison of this experiment with the Brassington Quarry experiment indicates two differing groundwater bodies (with higher groundwater in the Monsal Dale Limestone Formation), which are locally connected by the faults, in particular the mineralized faults. The results of the experiments (Figure 7.16) also support the concept of a greater degree of divergence in the Monsal Dale Limestone Formation than in the underlying formations.



Key: Shaded circle Injection point. C Cronkston; M Matlock; Me Meerbrook Sough; Mi Millclose Sough; O Oxclose Sough.
Heavy line with parallel dots marks edge of the outcrop of the Dinantian Limestone.
Lines with arrowheads indicate connections proven by water-tracing. r = reef deposits; v volcanic vent.
An arrowhead removed from stem indicates direction only (not distance).

Legend: Eyam Limestone Formation — Monsal Dale Limestone Formation

Bee Low Limestone Formation — Woo Dale Limestone Formation

Interbedded lavas

Dolerite — Volcanic tuff

Figure 7.16: Results of water-tracing: Derwent South (Matlock) Sub-basin (Appendix 7.1).

7.5 Conclusions.

Water-tracing has confirmed the presence of many fast flow routes (with flow rates in excess of 500 to 800 m/day) within which turbulent flow must predominate. Typically these routes are identified when tracer is injected into stream sinks, but they have also been identified by injection into boreholes (for instance at Tunstead Quarry). Generally, the fast flow is concurrent with, and followed by, a prolonged period of what has been interpreted as fracture-fed dispersed flow, indicative of storage in the aquifer and commonly detected at a larger number of springs, possibly associated with complex underflow paths, but more generally interpreted as divergent dispersed flow, associated with inception horizon-related fractures. Implicit in this is that there is a significant component of storage. Where tracer is injected into the epikarst its movement is guided by fracture flow at a range of stages of dissolutional enlargement. Where there are channels, or conduits, in close enough proximity to influence flow paths at least some of the flow will target the conduit. Beyond the influence of conduits, the fracture flow will be guided by the prevailing hydraulic gradient.

As with groundwater chemistry (Chapter 6), the fact that flow paths pass between different formations makes it difficult to relate flow rates to specific formations. For example, there is a lack of information regarding flow rates in the Monsal Dale Limestone Formation and the Miller's Dale Limestone Member, as it is only where these aquifers are perched (for example at Taddington) that they are isolated. Further difficulties arise because many of the tracing experiments are designed to establish flow path connections, rather than to establish flow rates, and this is reflected in the monitoring frequency. Wherever possible this author has speculated about the dominant flow paths in each of the experiments and some broad conclusions have been drawn from this speculation. There is a substantial pool of evidence for channel/ conduit development at the boundary between the Bee Low Limestone Formation and the underlying Woo Dale Limestone Formation, e.g. the Cunning Dale experiments, and in the Woo Dale Limestone Formation, where inception is suspected to be related to stylolites, for example the Tunstead Quarry experiments. Evidence from the Staden water-tracing experiments appears to indicate that springs at the Devonshire Arms, Cowdale and Ashwood Dale Resurgences appear to rise on the same fault from increasingly lower stratigraphical level in the Woo Dale Limestone Formation respectively. It is suspected that at depth the Chee Tor Limestone Member acts as a local aquitard. Some support for this hypothesis comes from the hydraulic isolation of the boreholes in the limestone quarries adjacent to the A515. However, closer to the surface, reflecting the response to stress relief, the opening of fissures in the Chee Tor Limestone Member results in higher rates of fracture flow as indicated by the high flow rates associated with the open excavations in the area of Staden. Water-tracing tests carried out in the Bee Low Limestone Formation, for example the Illy Willy Water trace; indicate rapid vertical movement of water to the Woo Dale Limestone. In the context of flow rates it is interesting to note that White (1969) observed that "*Artesian karst aquifers are not different in principle from other artesian aquifers except that the size of the solutional openings offers very little friction and thus very little head loss. However, the flow velocities in artesian karst aquifers seem to be slower than those in the free flow aquifers, possibly because of longer flow paths, and possibly because of other rate limiting factors such as discharge into overlying rocks of lower permeability*." In the context of the Wye catchment and also particularly evident in the Peak-Speedwell system, it would seem that one of the main rate limiting factors is the capacity of the dendritic channels that feed the resurgences. Another limiting factor is the low gradient of the inception horizon-related fractures. A number of resurgences are associated with faults and the importance of faults as groundwater divides, for example Dirtlow Rake and the Arrack Fault (Taddington Anticline), is also evident in the results of these experiments. None of the tracer tests from which dye was recovered were taken to the point of zero return of the tracer and therefore the tracer tests provide only limited information on the nature of the fracture/matrix flow, thus making the information derived from borehole hydrographs (Chapter 8) all the more important in terms of developing an understanding of this component of the flow.

Broadly the hydraulic gradient is to the east and southeast, although localised exceptions have been encountered, in particular in the northern part of the White Peak and also to the north of Chelmorton, where mature karst in the Woo Dale Limestone targets Ashwood Dale and at Staden where mature karst suspected to be at the boundary between the Woo Dale Limestone and the overlying Bee Low Limestone formations targets the River Wye at the Devonshire Arms. It would seem most likely that these localised reversals of gradient are attributable to glacial deepening of the River Wye. Groundwater appears to be concentrated in two zones, one perched above the Lower Miller's Dale Lava and the other concentrated in the Woo Dale Limestone. Connection between the aquifers occurs in the mineralized veins and in the estavelle at Wardlow Mires. It would appear that much of the groundwater in the Woo Dale Limestone only rises to the River Wye in geologically favourable situations, in particular at the location of faults, as described above, and that some of the groundwater in the Woo Dale Limestone moves out of the region as underflow (Chapter 5).

Chapter 8: Groundwater parameters derived from monitoring wells.

8.1 Introduction.

The constraints on using groundwater-levels derived from boreholes to draw conclusions regarding the hydrogeological characteristics of karstified limestone aquifers are well known. Smart and Worthington (2004a) and Worthington and Smart (2004) point out that given the low volume of limestone that is occupied by conduit networks, the probability of a borehole intersecting a major, cave sized conduit is only about 1 – 2%. Indeed, the likelihood of encountering a specific porosity element decreases as the dimensions of the element increase. Accordingly Worthington and Ford (1995a) argue that borehole techniques, which extend the test zone into the aquifer, such as the use of tracers, packer tests or pumping tests are required to confirm the existence of the network of interconnected channels. With respect to this study, the results of dye-tracing tests (Chapter 7) have been used to confirm the existence of interconnected channels comprising dissolutionally enlarged fractures and conduits (including caves). An additional source of data, comprising time-series related groundwater-levels for a number of monitoring wells, was made available by the Environment Agency. The groundwater-levels have contributed to the potentiometric surface generated in Chapter 9. Whilst, as suggested above, this may not be the preferred method of investigation; it was considered that the completeness of the dataset (the frequency of monitoring was approximately weekly during the period from well installation until 1983, when the monitoring frequency was reduced (Table 8.1)) justified further examination. Consequently, the groundwater-level fluctuations form the subject of the major part of this chapter. Further justification for the use of the dataset is the concept that, unlike rock strength, which can be demonstrated to increase with a reduction in sample size, but similarly reflecting the dominance of "matrix" behaviour, and for the reasons presented above, hydraulic conductivity generally decreases with a reduction in the volume of sample analysed, therefore the starting hypothesis was that the borehole dataset, which provide access to a small volume of rock, relative to the volume of the aquifer as a whole, offer a potential to characterise the hydraulic conductivity of the "matrix". Conceptually the results of this monitoring are also interesting because they reflect the hydrological properties of the zone of dynamic storage (Chapter 5 and Smart and Hobbs, 1986).

Water-level fluctuations in groundwater monitoring wells fall within four basic types: (1) changes in groundwater storage, (2) response to fluctuations of atmospheric pressure in contact with the water surface in wells, (3) deformation of aquifers and (4) disturbances within the well (Davis and De Wiest, 1966). Changes in storage account for most large fluctuations of water-levels and this forms the main focus of the following analysis of the results of the monitoring of the groundwater wells. Smart (1999, p. 155) modelled the effect of a number of karst features on observation well records, including: subsidiary conduits (conduits linking the well to primary conduits), overflows, overflows to reservoirs and surface recharge; and concluded that water-levels have a complex relationship to water flux in subsidiary conduits and suggested that the water-level effects are "*subtle and ambiguous*". Smart (1999, p. 155) also concluded that periodic monitoring of water levels "*will not allow characterization* *of the subsidiary conduits and sustains the continued use of inappropriate groundwater models in karst*", suggesting that more diagnostic information could be obtained from spring discharge measurements. Notwithstanding these conclusions, and in the absence of any spring discharge measurements, it is considered that (armed with an understanding of the regional geology, speleogenetic setting and the benefit of borehole logs) some verification of the conceptual model can be derived from analysis of the groundwater monitoring levels.

In addition to the data from the Environment Agency, groundwater monitoring records provided by: Hockenhull Enterprises Limited, The Limestone Research Group, University of Huddersfield and North Derbyshire Water Board (data held by the Limestone Research Group) have been considered. Climatic data used in the assessment of the hydrographs were obtained from the British Atmospheric Data Centre for the years 1977 to 1982. Temperature data were obtained for Buxton, and rainfall data for: Ashford, Bakewell, Buxton, Great Hucklow School, Monyash Vicarage, Peak Forest (Conies Farm) and Stanley Moor. Additional meteorological data for Buxton were obtained from High Peak Borough Council. It should also be noted that with respect to the time series data associated with the Environment Agency wells (Table 8.1) a consequence of the longevity of the dataset is that most of the drillers' records dating to the period of well construction have been lost. Furthermore, in some cases well construction details are ambiguous and detailed geological records (sought from both the Environment Agency and the British Geological Survey) were only available for some of the wells.

Table 8.1: Physical details of the Environment Agency groundwater monitoring wells.

Well	National Grid Reference	Availability of data	Ground Level (m AOD)	Groundwater-level (m AOD)		
				Maximum	Minimum	Range
Bee Low	SK 08567901	12.10.77 to 1998	354.40	337.74	286.07	51.67
Bull I' Th' Thorne	SK 12786655	11.4.69 to 1998	356.38	269.59	224.04	45.55
Dale Head Farm	SK S20776897	30.4.76 to 1998	154.55	132.17	129.12	3.05
Highcliffe Farm	SK 08197135	15.12.76 to 1996	344.96	284.64	249.00	35.64
Hucklow South	SK 17777762	11.4.69 to 1998	301.82	280.56	248.00	32.56
Nutseats Quarry	SK 2370 6582	11.3.76 to 1998	113.10	112.79	109.14	3.65
Oddo House	SK 21816072	11.4.69 to 1998	280.55	256.33	250.5	5.83
Peak Forest	SK 12017880	21.8.73 to 1998	335.28	333.87	324.29	9.58
Victory Quarry	SK 07507691	15.3.77 to 1998	352.16	345.25	330.26	14.99

8.2 Groundwater-levels.

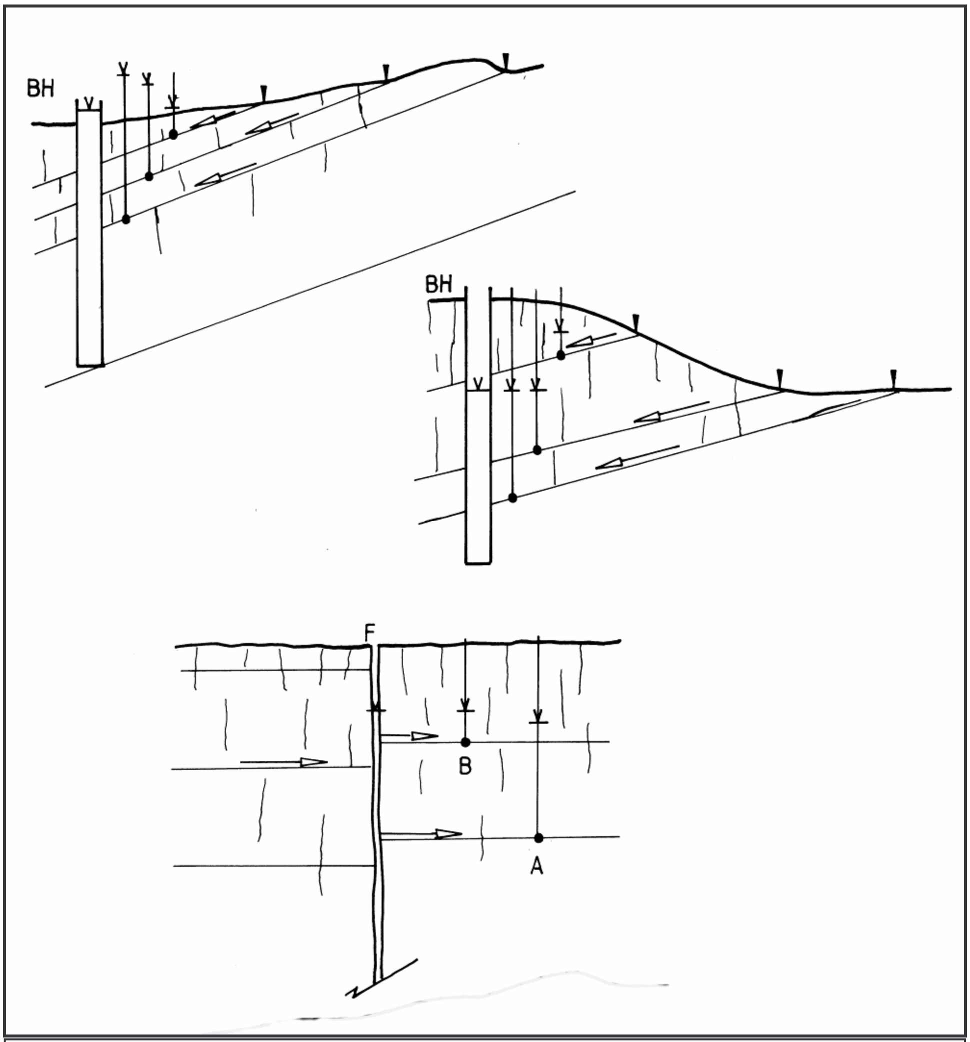
Historically, there has been considerable debate about the existence of a water table in limestone. This debate has been more than adequately summarised by Lowe (1992) and Worthington (1991). For the purposes of this dissertation, analysis has been carried out in terms of the potentiometric surface of the groundwater, which concurs with the work of Edmunds (1971) and the resultant groundwater contour map that he prepared. Indeed, groundwater-levels determined in the wells are in keeping with the potentiometric surface prepared by Edmunds (1971), which suggests that the contour map was prepared from these data. Clearly however, whilst this is relevant on a regional scale, it is potentially misleading on the local scale, because groundwater at different levels may be under differing heads, which are masked by the penetration of more than one level by a borehole, in which vertical flow can occur. In the context of this thesis, which presents evidence of inception horizon-related guidance of groundwater flow, there are a number of ways in which the differing heads may develop (Figure 8.1).

To add to the understanding of the groundwater-levels in the Peak District a groundwater-level map combining spring data with the potentiometric surface contours has been prepared (Figures 8.2 and 12.1). It is clear from this that there are areas of 'perched' groundwater associated with the interbedded lavas. The interbedded lavas also have the effect of confining groundwater resulting in springs emanating under pressure from beneath the lava e.g. in the area of Lees Bottom, where the springs are also associated with extensive tufa precipitation (Chapter 6). However, it should be noted that the location of Taddington High Well (SK 14407080) indicates that additional scenarios occur, for at this location the spring discharges from the Upper Miller's Dale Lava and at this location it would appear that there is considerable storage associated with fissuring in the lava.

Broadly, the potentiometric surface reflects the easterly and southeasterly hydraulic gradient that has been identified in the discussion of the regional hydrogeological setting (Chapter 5). However, there are areas of perched water that are associated with dynamic storage in the Monsal Dale Limestone and there are also mounds of water associated with recharge from the adjacent Namurian strata, e.g. the groundwater mound to the south of Buxton.

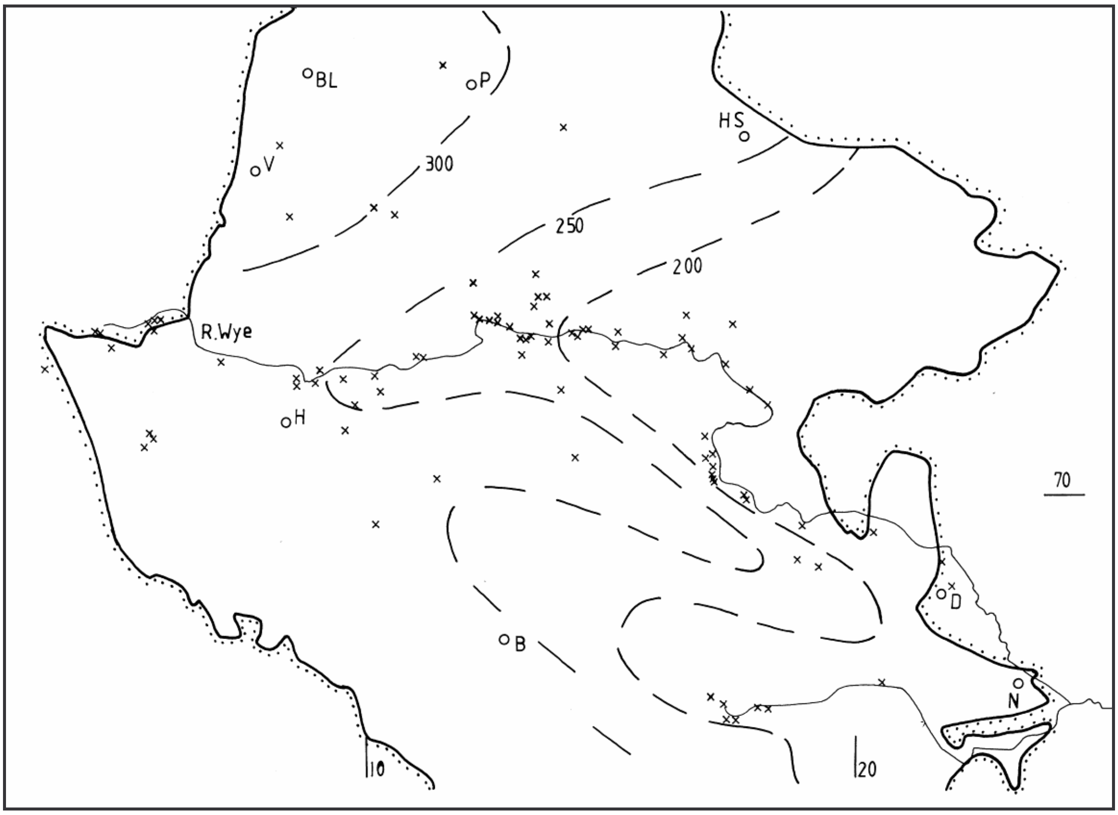
8.3 Aquifer characterisation.

In the interpretation of groundwater-levels it is first necessary to characterise the aquifer. The current conceptual model of karst hydrogeology is one of a triple porosity medium as described by Worthington and Smart (2004, p. 401): "*Most pre-Cenozoic carbonate aquifers have well-developed bedding planes and joints, and these intersecting fractures result in fracture porosity and hydraulic conductivity in these aquifers. The matrix blocks between the fractures yield a second type of porosity and hydraulic conductivity, and the linear, interconnecting conduits form a third order of porosity and hydraulic conductivity*." Clearly the geologist needs to understand that the term matrix flow in this context (Worthington et al., 2000), does not relate to the intergranular fine-grained matrix that surrounds clasts in a sedimentary rock and is know to have been occluded by a number of phases of cementation in the limestone of the research area; instead it primarily refers to any interconnected voids (largely fine factures) in the rock mass between the fractures forming the fracture porosity. The interconnecting porosity within each of the porosity types is referred to as channel porosity (Worthington and Smart, 2004), or conduit porosity. Conduits form the principal pathways for groundwater movement and they are fed by fractures and matrix flow. Thus the hierarchy is comparable with that of stream ordering. Worthington and Smart (2004) suggest that >95% of the porosity is rock matrix porosity, but because pores are usually small and poorly connected matrix hydraulic conductivity is low. Fracture hydraulic conductivity is usually three to four orders greater than that of the matrix hydraulic conductivity, reflecting the interconnection of the fractures with typical apertures of 0.01 to 0.1 mm (Worthington and Smart, 2004). The hydraulic conductivity of dissolutionally enlarged fractures is a further one to two orders of magnitude greater.



Key: BH Borehole, solid circles piezometers, v standing water level, open arrows indicate groundwater flow paths and occluded triangles surface water recharge. Note: slightly lower head in A than B reflects the influence of seepage forces.

Figure 8.1: Piezometric head and groundwater-levels in boreholes.



Key: B Bull I' Th' Thorne Borehole; BL Bee Low Borehole; D Dale Head Farm Borehole; H Highcliffe Farm Borehole; HS Hucklow South Borehole; N Nutseats Quarry Borehole; P Peak Forest Borehole; V Victory Quarry Borehole. Spring names presented in Chapter 10 (Figure 10.4)
Legend: x Location of spring (locations from Gunn (1998)); O borehole Location. Groundwater contours shown at 50 m intervals.

Figure 8.2: Groundwater level contours map to show borehole and spring positions.

Worthington and Smart (2004, p. 401) state that "… *in most carbonates flow into boreholes occurs at just a few locations, and these locations are fractures that have been greatly enlarged by dissolution*." In the absence of drillers' records and geophysical logs (e.g. temperature, flow velocity or conductivity) the depths of the inflows of groundwater to the wells is not known and therefore one of the aims of the analysis was to try and establish the likely levels of groundwater inflow to each of the boreholes.

It is clear from assessment of the hydrogeochemistry (Chapter 6) that there is a variable degree of confinement within the limestone. Brassington (1988, p. 81) states that "*changes in atmosphere pressure produce significant fluctuations in wells in confined aquifers*", this indicates that one means of assessing the degree of confinement in the aquifers penetrated by the boreholes could potentially be in the determination of the barometric efficiency of the wells (the ratio of water-level changes in the well to the inverse of water-level changes in a water barometer):

$\alpha = -\Delta W / \Delta B$, where W is the water surface elevation in the well and B the barometric pressure

Although barometric efficiency may be a significant parameter it has not to this author's knowledge been used to characterise karst aquifers, although barometric responses have been measured, for example Larocque, et al., (1998). Springs may also be subject to barometric influence, particularly where they are deeper and approach vertical flow paths. It is also considered by this author that barometric effects may influence groundwater movement through the aquifer, as there is empirical evidence that groundwater moves in pulses (Gunn, personal communication, 2003; Bottrell, personal communication, 2003).

Davis and De Wiest (1966) suggest that water-level fluctuations caused by pressure on the water surface in wells are generally of two types: short term fluctuation brought about by gusts of wind passing over the mouth of the well and fluctuations due to changes in barometric pressure. Rasmussen and Crawford (1997) identified short and long term barometric effects, albeit on a different scale, which provide diagnostic information regarding the confinement of the aquifer. An instantaneous response was shown to be indicative of confined conditions and a delayed response due to borehole storage or skin effects with delayed total head response in unconfined aquifers due to the air diffusivity coefficient within the unsaturated zone. Hare and Morse (1997) demonstrated that in some circumstances, e.g. inside landfill containment systems, wells in unconfined aquifers can have barometric efficiencies greater than those typically observed for wells in confined aquifers. It is plausible that this may be likened to the containment imposed by an "impermeable" fault zone and therefore careful consideration is required in the assessment of the nature of confinement.

Wells in the limestone usually comprise a steel casing to support the bore through more weathered or fractured ground, including the epikarst, and beneath this the well comprises an unsupported 'open hole'. Accordingly, it might be considered that the boreholes are unlikely to show a barometric response, because the response zone is not discrete and therefore none of the wells are sealed within a specific confined aquifer and vertical flow can occur within the well. However, it is considered by this author that discharge from the unsaturated zone could occur in response to falls in barometric pressure causing a reduction in tensional storage in fine fractures in the limestone. This form of barometric response does not appear to have been reported previously and is a vadose (unsaturated) zone response. Water movement in the vadose zone is essentially vertical, with lateral movement along channels, ultimately feeding to dissolutionally enlarged fissures and then conduits. It is clear from the constancy of supply to perennial springs that there must be considerable storage in the vadose zone. Storage is likely to take two forms, one being dissolutionally enlarged cavities with restricted outflows (comparable with macro-pore storage in soil) and given the range of porosity sizes in the limestone the other is likely to be capillary water. Smart and Worthington (2004) noted that there is little capillary movement of water upwards through the unsaturated zone. It is suspected by this author that this reflects the predominance of porosity in fissures that widen upwards, thereby precluding capillary rise. The presence of capillary water is evident in the drip waters observed in caves, e.g. Friederich and Smart (1982). If the matric potential of the capillary water is susceptible to changes in barometric pressure, there is a significant potential for the release of stored capillary water during periods of rapid falls in barometric pressure, thereby inducing an unsaturated zone response to barometric pressure, which is more likely to be evident in discharges to springs collecting water from a large geographic area, than groundwater-levels in the boreholes. The occurrence of clay wayboards is also likely to be significant in terms of groundwater storage. The clays will exhibit high matric potential, which will be overcome by the removal of infiltrating recharge water feeding the "quick flow".

8.4 Characterisation of the recession curve portion of well hydrographs.

Regular seasonal recession characteristics can be seen in the hydrographs (Appendix 8.1). In unconfined conditions, with predominantly vertical drainage to channels, this is likely to reflect the drainage of units of differing hydraulic conductivity in the limestone. The regularity of the seasonal recession characteristics have been observed by a number of workers including Headworth (1971), Moore (1992), Powers and Shevenell (2000) and Shevenell (1996). The form of the recession curve observed in springs forms the basis for the derivation of basic hydrologic information about karst aquifers (Padilla et al., 1994).

It might be argued (as for example by Ben Fretwell, personal communication in an Unsaturated Zone Model Users Group, May 2006) that the stepped parabolic form of the recession curve in karst aquifers is typical of a fracture-flow aquifer and simply reflects the contribution of a continuum of fracture sizes. This concept is explored in the analysis of spring discharge curves (Padilla et al., 1994 and section 8.7). Nevertheless, it should also be noted that the karst system is more heterogeneous than a fractured rock aquifer, because there are a number of additional factors to consider, including the following:

- karst systems comprise continuums of both laminar and turbulent flow and there is a greater range of forms of porosity and its connectivity to achieve hydraulic conductivity than in fracture systems;
- the hydraulic conductivity of the karst system evolves from a broader range of influences, it is not simply a product of the stress fields that have influenced the aquifer;
- because the karst system achieves its hydraulic conductivity by dissolution, its hydraulic conductivity evolves and reflects the state of maturity of the karstification process; and
- preferential flow paths (channels) are an integral feature of karst systems, which must lessen the significance of proximal and distal flow path considerations, by shortening flow path lengths.

Further to these observations and a consideration of the results of dye-tracing experiments (Chapter 7) it can be seen that dissolutional activity is not only important in creating hydraulic conductivity, but is also important in generating storage. Thus it can be argued that the relatively steep hydraulic gradient across the White Peak (Chapter 5), which contrasts with the very low hydraulic gradient of an individual inception horizon-related fracture, or conduit, is a facet of both the storage and the limiting dimensions of the matrix, fracture and channel flow paths that impose storage by limiting outflow rates. The evidence provided in Chapters 3 and 4 indicates that karst processes operating in the White Peak are strongly related to bedding (with bedding have a significant guiding influence on the development of inception horizons). This is significant in considering both storage and flow path processes. Sump formation is associated with more steeply dipping conduits, which reflects the funnelling of water and therefore focused dissolutional activity. In the situation where flow paths are sub-horizontal, it is inevitable that during low to normal groundwater conditions dissolutional activity will be focused more evenly on the floor and sides of the channel, with the potential for downward dissolutional activity. During peak flow conditions, in situations where laminar flow prevails, frictional forces limit flow rates around the perimeter of the channel, focusing higher rates of flow towards the middle of the channel and thereby protecting the storage zone that develops beneath the elevation of the flow path outlet, or spring resurgence level and maintaining storage. Clearly the surface area of a sump is likely to be considerably less than that of a channel, therefore it follows that a sump-fed resurgence will respond much more quickly to storm events than a system in which channels form the major component of storage. Thus it can be argued that the responsiveness of the aquifer is a function of its form as well as the state of maturity of karstification.

The epikarst forms an integral part of the karst system. The range of porosity types associated with the epikarst has been described in Chapter 4. Shafts provide a rapid response to storm events, thereby contributing to the higher hydraulic conductivity response of the recession curve. It is suspected that the localised occurrences of Head deposits that are particularly associated with the Monsal Dale Limestone Formation and the Miller's Dale Limestone Member form zones of storage that provide a significant contribution to base-flow.

The elevation to which seasonal groundwater recession falls within a borehole is influenced by a number of factors including: the amount and intensity of rainfall events in a given year, the extent of evapotranspiration, aquifer elevation, hydraulic conductivity and storage of the aquifer and the levels of horizontal inflows (fracture or channels connections) to the borehole. If, as hypothesised in the conceptual model of karst hydrogeology, vertical drainage to channels is occurring, this would be reflected in a stepped recession curve as each unit is consecutively drained. Furthermore a more parabolic form to the recession curve would be more likely to indicate greater matrix hydraulic conductivity. Where channels are encountered at depth, this should be reflected in a reduction in the rate of recession (i.e. reflecting the lateral inflow, increased hydraulic conductivity and storage associated with these zones). Furthermore, repetitions of the elevation defining the base of a recession could be indicative of a significant channel that limits the depth to which the groundwater-level will fall in the borehole. Equally, this could be indicative of an impermeable layer at depth.

Storm events are also characterized by recession curves, which are superimposed on, and take a similar form to, the seasonal recession curve. Following the method of Rorabaugh (1960), Powers and Shevenell (2000) have suggested that the form of karst well hydrograph recessions, following specific storm events, can be approximated by three straight line segments on a semi logarithmic plot. This has been likened to the recession curve of a surface stream, with each segment representing a different type of storage. The steeper portion of the curve is thought to represent the drainage of conduits, the middle portion fractures and the lower part the matrix, up hydraulic gradient of the monitoring point. Such analysis requires closely spaced monitoring intervals following specific storm events. Closer examination of the published curves suggests that the straight line portioning is not entirely representative and that perhaps the recession curves could be more reliably interpreted by division into a number of differing recession curves. Given the heterogeneous nature of karst it seems inappropriate to assume that discharge, which is driven by hydraulic gradient and depends on the hydraulic conductivity, which is known to vary, can be represented by the head determined in a nearby borehole, unless that borehole is in very close proximity to, and in the flow path of, the discharge point.

It was noted by Headworth (1971) that the recession portions of hydrographs are characterised by a specific form and that the form of the storm recession curve generally approximates to part of the seasonal recession curve monitored within a specific well. This suggests a response to a characteristic range of storage units within the zone of dynamic storage in the limestone. Headworth (1971) suggested that in the situation where unconfined conditions exist, the rate of fall of the water table must be influenced by lithological variation and the drainage of different lithological units, hence changing porosity ranges. Thus the recession curve needs to be related to groundwater-level. In accordance with these observations Headworth (1971) developed methods to interpret natural groundwater-level fluctuations in the Chalk of Hampshire. He used the seasonal recession component of the hydrograph to evaluate the rate of recession, a recession constant, the coefficient of storage, the rate of apparent percolation and an estimation of the co-efficient of transmissibility. Although developed for the Chalk, the methods are considered applicable to the interpretation of limestone hydrographs, because as noted in section 8.3 the wells are most likely to represent the matrix/fracture hydraulic conductivity of the aquifers. Furthermore, these methods are considered to be relevant because they essentially comprise a statistical method of analysis of seasonal recession curves monitored over numerous consecutive years and because they directly relate the rate of change of the gradient of the recession curve to groundwater level, they overcome the problem of variations in the monitoring frequency. The method (Headworth, 1971 and Appendix 8.2) was used to generate a master recession curve for each borehole (Figures 8.3 to 8.11). The stepped recession curves that result have been observed in karst settings by other authors, including Headworth (1971), Powers and Shevenell (2000) and Smart (1999). It would seem plausible that the base of the steps indicates the location of inflows to the borehole (channels, section 8.3). Given the nature of the geology it is also plausible that these points also identify the locations of inception horizons. Such an explanation does not contradict the interpretation of multi-level conduits (Smart, 1999). The differing form of individual steps of the recession curve indicates zones of differing hydraulic conductivity (section 3.8).

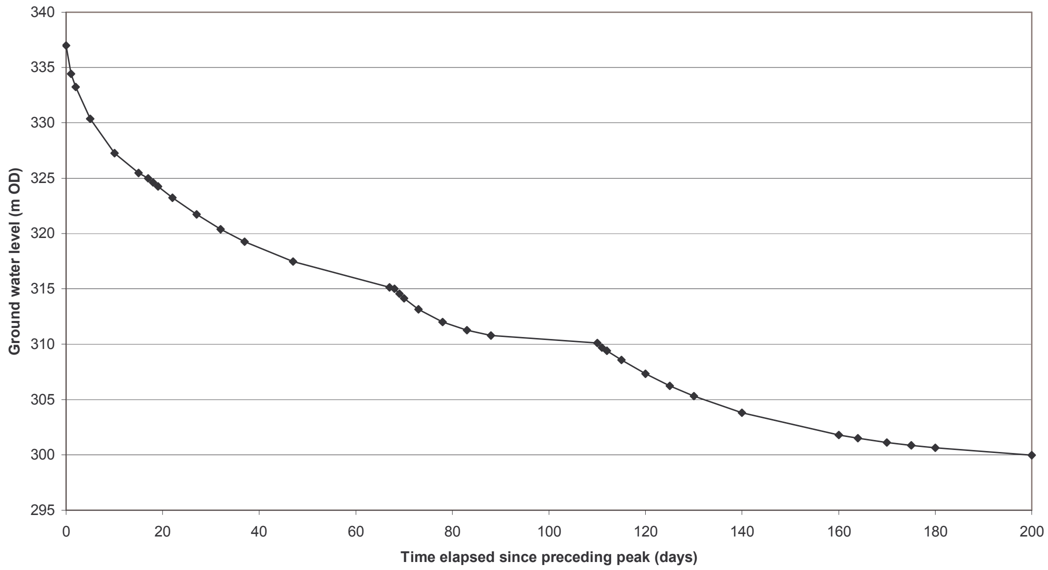


Figure 8.3: Master recession curve: Bee Low Borehole.

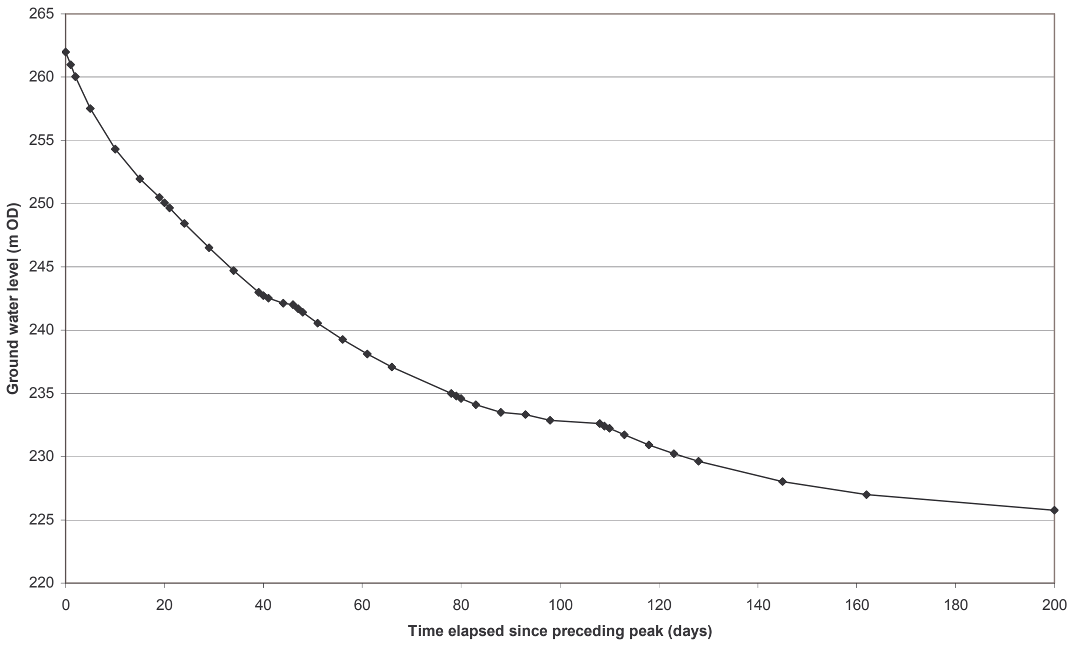


Figure 8.4: Master recession curve: Bull I' Th' Thorne Borehole.

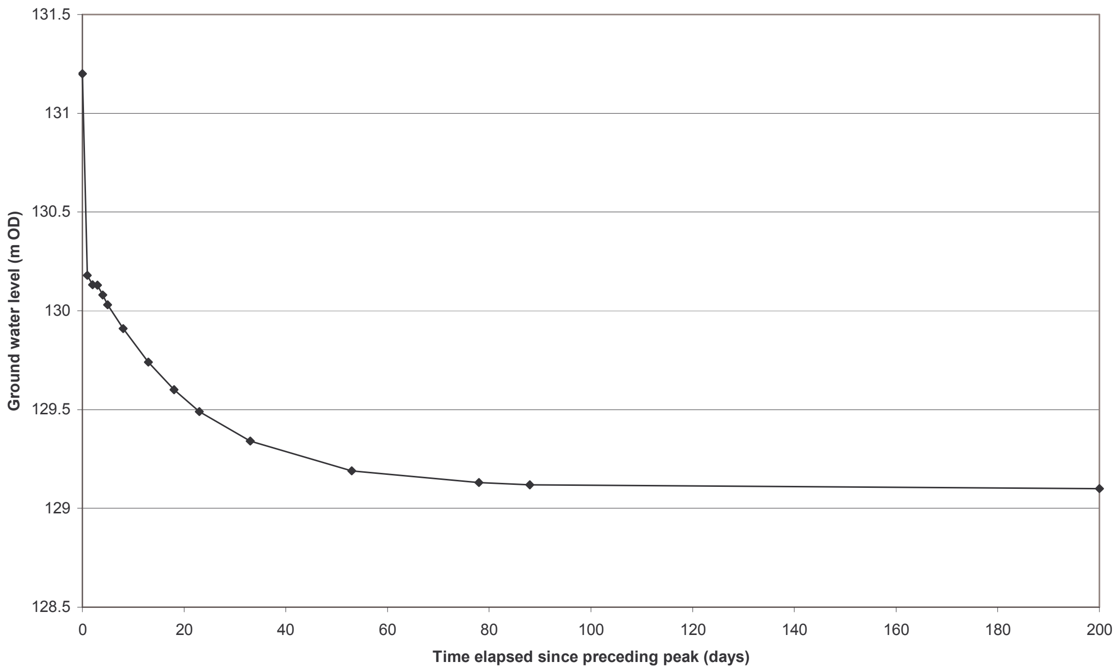


Figure 8.5: Master recession curve: Dale Head Farm Borehole.

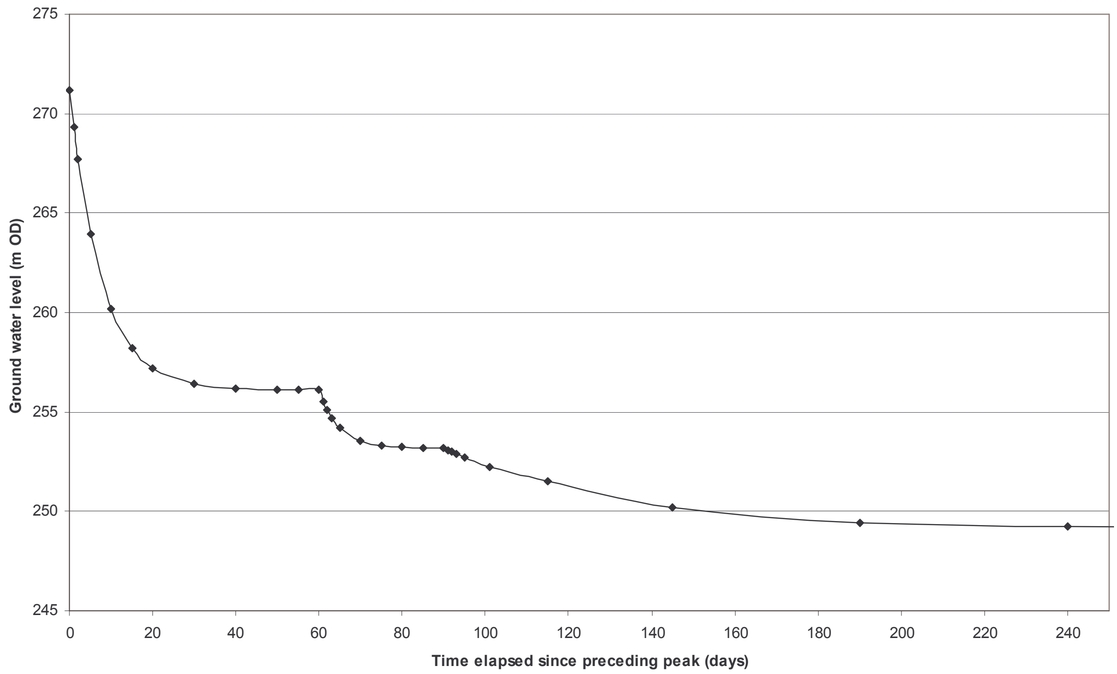


Figure 8.6: Master recession curve: Highcliffe Farm Borehole.

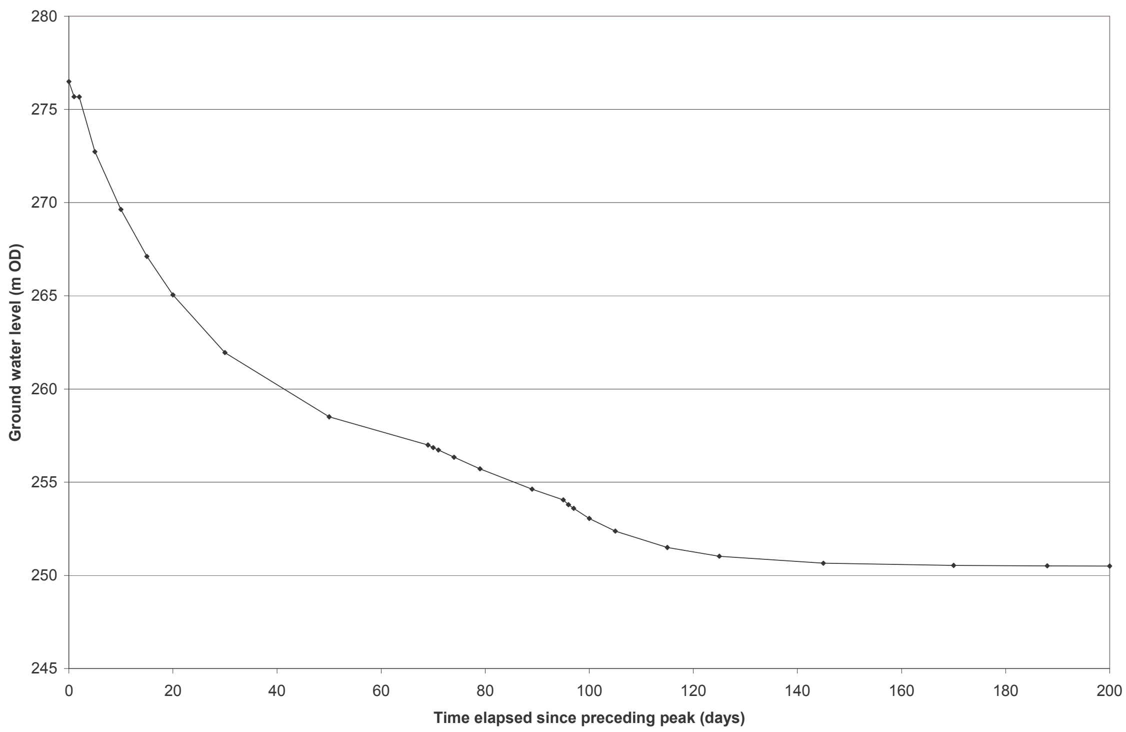


Figure 8.7: Master recession curve: Hucklow South Borehole.

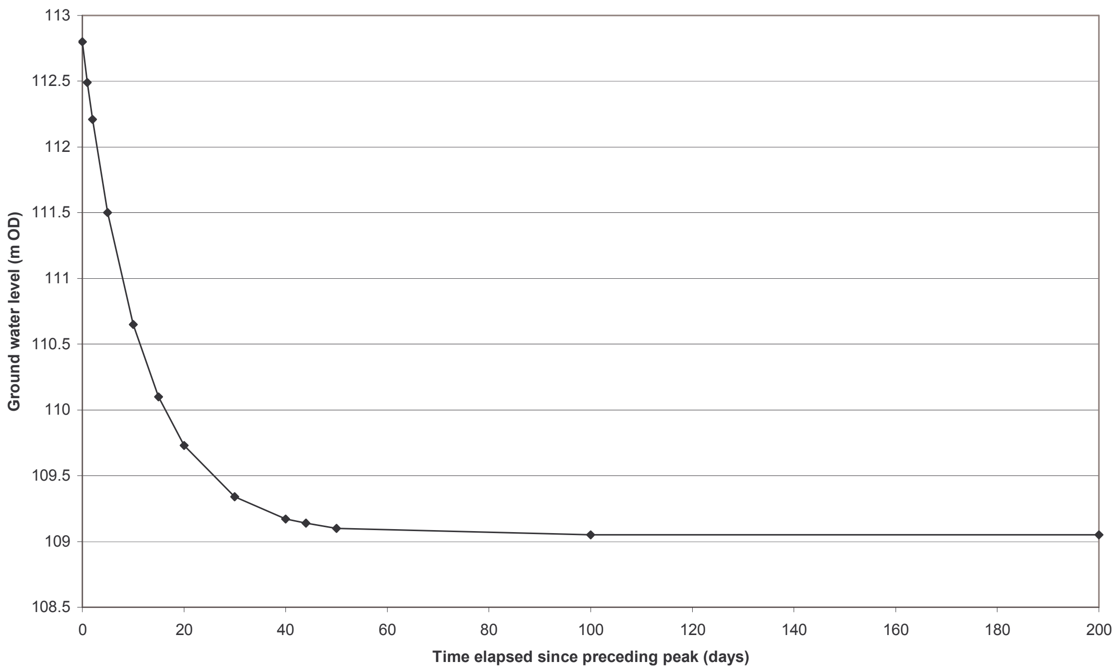


Figure 8.8: Master recession curve: Nutseats Quarry Borehole.

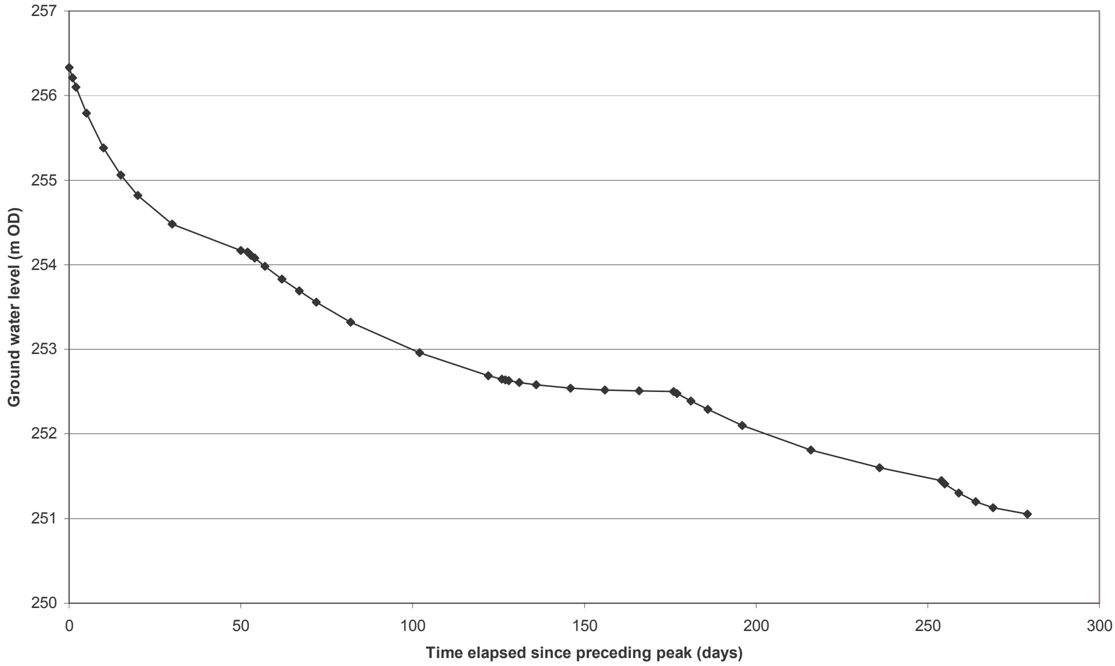


Figure 8.9: Master recession curve: Oddo House Borehole.

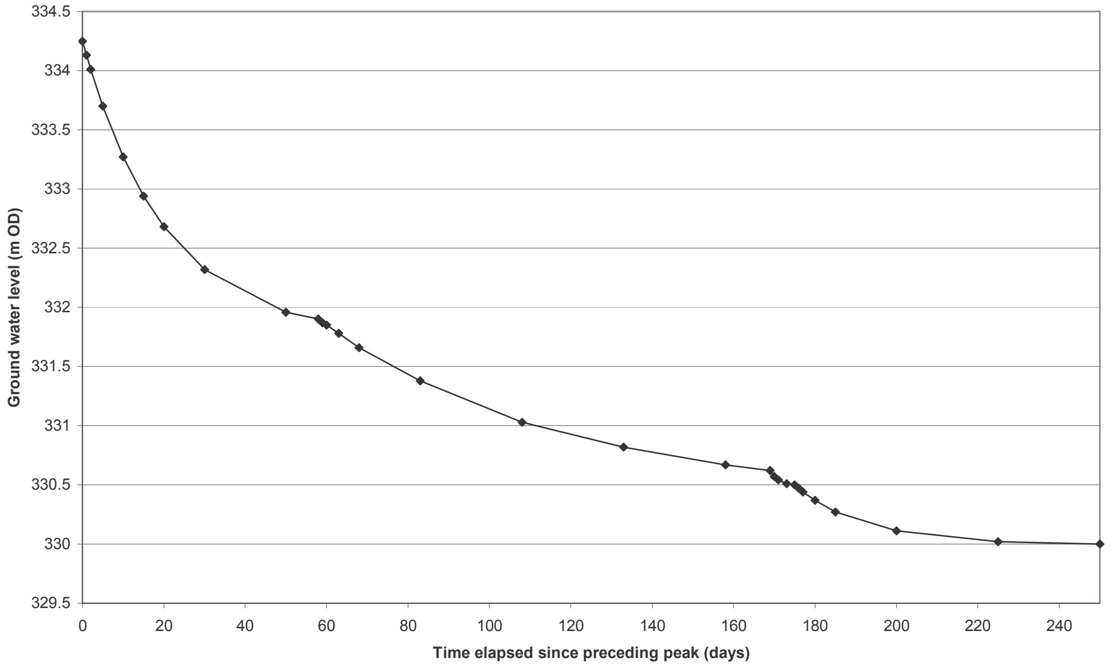


Figure 8.10: Master recession curve: Peak Forest Borehole.

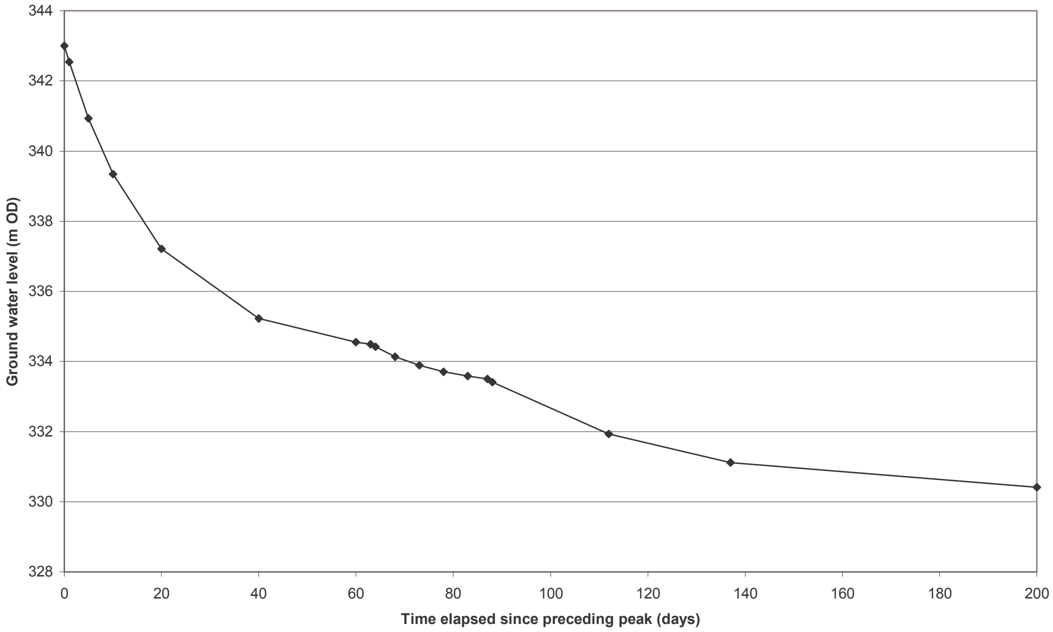


Figure 8.11: Master recession curve: Victory Quarry Borehole.

8.5 Analysis of the master recession curves.

8.5.1 Bee Low Borehole (SK 08567901; ground level 354.4 m OD):

The grid reference for this hydrograph corresponds with borehole SK07NE/44 (Harrison, 1981). The strata penetrated by the borehole are thought to comprise the Chee Tor Limestone Member. The overlying Miller's Dale Limestone Member can be seen capping the Chee Tor Limestone in Bee Low Quarry (SK 092791). The limestones exposed in the Quarry exhibit predominantly eastnortheast to westsouthwest-trending fissuring and south easterly-dipping bedding. The intense fissuring can be seen in Plate 8.1 below, although it should be noted that quarrying utilised blasting techniques, which contributes to the openness of the fissuring. Below ground the fissures close with depth and this is reflected in the shallower gradient of the first step of the master recession curve (Figure 8.3), compared with that derived for the Bull I' Th' Thorne Borehole.

It is suspected that the hydraulic gradient is to the southeast and that the high elevation of the standing water level (286.07 to 355.27 m OD) reflects: a dome of recharge from the adjacent Namurian strata; groundwater that is perched by the Miller's Dale Limestone, and where the Chee Tor Limestone is exposed, groundwater storage in the fissuring. The borehole lies outside the area of influence of the Dove Holes Tunnel (Chapter 3). The master recession curve (Figure 8.3) comprises three distinct steps. It is likely the first represents the presence of gradually tightening fissuring to a level of approximately 325 m OD, with a hydraulic conductivity in the order of 2.75 x 10^{-5} m/day, which seems to be typical of the uppermost level of the zone of dynamic storage. The indistinct step in the recession curve coincides with the base of a shell bed and a possible fracture into the borehole. Beneath this the biosparites and biopelsparites exhibit a lower hydraulic conductivity (8.67 x 10^{-6} m/day) and appear to drain to another channel, associated with clay coated stylolites at 39.4 m depth (315 m OD). Immediately below this level the ground water level falls away more quickly through a zone of stylolites with a hydraulic conductivity of approximately 3.34 x 10^{-5} m/day. This zone is underlain by another less permeable zone biopelsparite and pelsparite (8.9 x 10^{-6} m/day).

Plate 8.1: Bee Low Quarry October 2004.

8.5.2 Bull I' Th' Thorne Borehole (SK 12786655; ground level 356.38 m OD):

The records for this borehole were not published. A copy of the log was obtained from the British Geological Survey. However, groundwater levels recorded in the hydrograph fall below the recorded depth of the borehole in the borehole log (SK 16 NW/5). Correspondence with the British Geological Survey (David Lowe, personal communication February 2004) has not shed any further light on this apparent anomaly, although it would appear that there is a possibility that the borehole was deepened in accordance with instruction from Midland Region Environment Agency staff. The master recession curve (Figure 8.4) comprises six indistinct steps. It is likely that the first represents the presence of gradually tightening fissuring to a level of approximately 250.5 m OD with two minor changes in hydraulic conductivity in the "*light grey, mealy and ?oolitic limestones*". Pale blue shale encountered at 258.2 m OD was interpreted by the geologist as the lateral continuation of the Upper Miller's Dale Lava, thereby indicating the water table is in the Bee Low Limestone Formation (Miller's Dale Limestone Member). Inception horizons have been identified in the Miller's Dale Limestone (Chapter 4), however it is likely that they are not so well developed at this location, which falls on the edge of the Ashford Syncline an area that is associated with basinal limestones and an absence of clay wayboards. Ground water levels were in the range 224.04 to 269.59 m OD, with occasional values reaching levels of up to 269 m OD, indicating that this lies close to the Lathkill sub-basin ground water divide, associated with the Cronkston-Bonsall Fault, with an eastsoutheast-trending hydraulic gradient (Figure 8.2).

8.5.3 Dale Head Farm Borehole (SK 20776897; ground level 154.55 m OD):

This grid reference corresponds with Borehole SK 26 NW 12 (Bridge and Gozzard, 1981, p. 27), however this borehole is referred to as Field House and it is not located at the grid reference quoted by the Environment Agency. The Environment Agency recorded the location as SK 20816801. This author could not locate the borehole in the field at the Environment Agency grid reference; furthermore the ground level at this grid reference does not correspond. The borehole, which is 100 m deep, penetrated the Widmerpool Formation, Eyam Limestone Formation, Monsal Dale Limestone Formation and the Conksbury Bridge Lava. The master recession curve (Figure 8.5) comprises only two steps. The first exhibits a hydraulic conductivity of 3.42×10^{-5} m/day, which appears to be typical of the uppermost level of the zone of dynamic storage. Beneath this there is a single curve with an apparent hydraulic conductivity of 1.51×10^{-6} m/day. Water levels ranged between 125.34 and 132.17 m OD. These levels fall within the Monsal Dale Limestone Formation and immediately below the lowest level, at 115.53 m OD the borehole records the occurrence of two gaps in the core either side of a chert band of 39 and 23 cm width. Accordingly, it is suspected that the ground water levels recorded in this borehole comprise fracture flow, which could be confined and is likely to be associated with an inception horizon. Clearly therefore, the hydraulic conductivity calculated for this zone is unlikely to be correct. At this location the beds dip towards the north, therefore it is suspected that the borehole has intercepted fracture flow feeding to the River Wye. Interestingly, the level of the River Wye at Bakewell is approximately 140 m OD. Thus, this supports the findings of the analysis of the Monks Dale Borehole pumping test (section 8.7) that the River Wye is 'perched' for at least part of the year in the reaches between Monks Dale and Bakewell, where the river bed is underlain by a series of lava beds.

8.5.4 Highcliffe Farm Borehole (SK 08197135; ground level 344.96 m OD):

This reference appears to correspond with Borehole SK 07 SE 51 (Harrison, 1981, p. 88). The borehole, which is 100 m deep, penetrated the Chee Tor Limestone Member and continued into the Woo Dale Limestone Formation at 307.58 m OD. The master recession curve (Figure 8.6) comprises three distinct steps. The first, which ranges from 284.64 m OD down to 256.10 m OD, exhibits a hydraulic conductivity of 1.53×10^{-5} m/day, marginally lower than that determined for the Bee Low Borehole. The base of this zone falls on the boundary between "*Biomicrite, buff-grey, very fine arenite, frequent quartz euhedra, common black clay partings, limonite staining and weathered dolomite*" above; and "*Biopelsparite, buff grey, fine arenite, well sorted*" (Harrison, 1981, p. 89) below.

It is possible that there is a fracture entering the borehole at this boundary, evidence for which comes in part from the observation that the dolomite is weathered and from the occurrence of limonite. Dolomite was found to be present in other, higher, units, in the borehole, but was not described as weathered in those locations. The hydraulic conductivity of the subsequent step has been calculated as 7.65×10^{-5} m/day. The borehole log does not provide any evidence for the increase in hydraulic conductivity. It is possible that it is attributable to a greater degree of fracturing, possibly attributable to enhanced brittleness in this zone; alternatively it may be due to the presence of stylolites. However, the occurrence of stylolites does not appear to have been recorded on the borehole logs in this area (Harrison, 1981). The lowest step extends to 249.2 m OD, with a marginally lower, but still relatively high hydraulic conductivity of 1.05×10^{-5} m/day. At 249.2 m OD, the borehole log records "*Biomicrite, dark grey, very fine arenite, well sorted; common black clay streaks and partings*" (Harrison, 1981, p. 90), which suggests that the borehole intersected channel flow associated with the clay parings. It is suspected that this borehole falls within the catchment of Ashwood Dale Resurgence.

8.5.5 The Hucklow South Borehole (SK 17777762; ground level 301.82 m OD):

This is one of the boreholes that were drilled during investigatory work undertaken by the North Derbyshire Water Board (Raffety et al., 1953). A copy of the borehole record (SK 17 NE/4) was obtained from the British Geological Survey. The hydrograph for the borehole (Appendix 8.1) shows a maximum standing water level of 280.56 m OD and a minimum of 248 m OD. The borehole was 123.60 m deep and penetrated the Eyam Limestone Formation and the underlying Monsal Dale Limestone Formation, with its base in the Upper Miller's Dale Lava. At this location the limestones dip in an easterly direction. Edmunds (1971) showed that ground water in the area of Stanley Moor is perched. The findings of this borehole have confirmed the ground water levels and indicate that the 'perched' condition of the water is probably attributable to the presence of basalt lava, dipping to the north east towards the rim of reef limestones sandwiched between the Eyam Limestone and the Namurian strata to the north east. The master recession curve comprises three relatively indistinct steps. The base of the first step, at 257 m OD, coincides with the boundary between the Eyam Limestone and the underlying Monsal Dale Limestone. This suggests that the borehole has intersected fracture flow at this level and there is a strong possibility that this is associated with an inception horizon. The identification of this layer appears to confirm the suitability of this method of interpretation. The basal unit of the Eyam Limestone Formation at this location comprises dark limestone, black chert and black shale. The hydraulic conductivity indicated by the upper part of the master recession curve is 8.63×10^{-5} m OD. It should also be noted that on a few occasions the highest standing water level exceeded that shown on the master recession curve (Figure 8.7), with groundwater levels as high as 280.56 m OD being recorded. The middle part of the master recession curve appears to drain at a uniform rate down to a level of 254 m OD, with a relatively high hydraulic conductivity, determined to be 5.16×10^{-4} m/day. The lowest part of the master recession curve also exhibits relatively high hydraulic conductivity; a value of 4.75×10^{-4} m/day has been calculated. The bases of the lower steps of the master recession curve both fall within a unit described as "*Light grey limestone,* *averaging 23% chert*" (British Geological Survey borehole record SK 17 NE/4). Thus there is no further means of interpreting how the ground water moves through this unit, although it might be speculated that the chert could be influential in guiding ground water movement, particularly if it occurs as bands.

8.5.6 Nutseats Quarry Borehole (SK 23706582; ground level 113.10 m OD):

The hydrograph associated with this borehole (Appendix 8.1) indicates that ground water levels fall within a narrow range, extending between 112.79 and 108.66 m OD. The borehole is borehole SK 26 NW 14 (Bridge and Gozzard, 1981). Beneath a veneer of made ground the borehole penetrated the Eyam Limestone Formation, Monsal Dale Limestone Formation, Conksbury Bridge Lava and was completed in the Lathkill Lodge Lava at a depth of 100.05 m. The master recession curve takes the form of a smooth recession curve. In the absence of drillers' records for the borehole it is difficult to assess at what levels ground water enters the borehole. The presence of a substantial gap (0.59 m) in the recovered core at 110 m OD which falls within the zone of dynamic storage indicated by the hydrograph and is not reflected in the master recession curve suggests that the ground water levels recorded in the borehole comprise confined ground water. It is suspected that the borehole has intersected flow associated with fissuring, probably associated with inception in the Monsal Dale Limestone Formation. The borehole was situated immediately to the north of Long Rake, a suspected zone of ground water storage. As it would appear that this represents confined flow conditions, the hydraulic conductivity derived from the recession curve (1.84 x 10^{-6} m/day) is very unlikely to be representative of the true hydraulic conductivity of the strata. The borehole was drilled in the order of 250 m to the west of the confluence of the River Bradford with the River Wye. The level of the River Wye at this location is approximately 110 m OD. Therefore it is likely that the ground water is in hydraulic continuity with the River Wye.

8.5.7 Oddo House Farm Borehole (SK 21816072; ground level 280.55 m OD):

This borehole, which was drilled to 45.72 m depth in the dolomitized Monsal Dale Limestone Formation, has permanent pumping equipment installed within it. The hydrograph (Appendix 8.1) does not show any evidence of abstraction between October 1976 and 1991. The borehole record available for this borehole is limited to the driller's record (supplied by the British Geological Survey, record SK 26 SW/15), which indicates the occurrence of karstified, possibly dedolomitized, limestone with drift to 258.30 m OD (22.25 m depth), capping Monsal Dale Limestone, comprising "*hard limestone with occasional chert bands approximately 0.30m thick, with many sandy clay beds*". The hydrograph indicates that ground water levels range between 250.22 and 256.33 m OD. During drilling ground water was struck at 243.36 m OD and rose to 244.28 m OD. These levels are lower than those indicated by the well hydrograph. It has been noted that the borehole was drilled in July 1969. The findings indicate that the ground water was encountered in the Monsal Dale Limestone Formation, possibly perched on shales and clay contemporaneous with the Lower Matlock Lava. The presence of

Coast Rake in the order of 150 m to the north of the borehole provides a potential storage zone associated with this groundwater. The significance of Coast Rake in terms of groundwater storage is implicit in the need for the construction of Wrathe Sough (SK208608, Rieuwerts, 1987). In the absence of detailed logs it has not been possible to correlate the steps in the master recession curve (Figure 8.9) with specific geological features. The master recession curve comprises five steps. High permeabilities have been calculated (in the range of 1.1 to 3.93 x 10^{-3}m/day).

8.5.8 Peak Forest Borehole (SK 12017880; ground level 342.04 m OD):

This borehole penetrated ash (volcanic clay) and dolerite. The only record that was obtained for this borehole was the driller's record, supplied by the British Geological Survey (Borehole SK 17/NW/8). During drilling, ground water was encountered at a level of 289.56 m OD, in a zone of more weathered dolerite towards the base of the borehole, and the borehole was cased with plain casing between ground level and 287.56 m OD. The standing water levels monitored by the Environment Agency range between 324.29 and 333.87 m OD. The water levels indicate confined ground water conditions, because the water level has risen into the cased section of the borehole. It seems likely that ground water is confined by the fresher dolerite. Interestingly however, the master recession curve (Figure 8.10) is stepped. The dip of bedding in the limestones is to the east, coincident with the hydraulic gradient. Accordingly the source of this ground water is likely to lie to the west, with inception horizon guided flow becoming increasingly confined down hydraulic gradient, beneath the lava, in an easterly direction. The stepped form of the master recession curve may be a response to drainage in the limestone beyond the outcrop of the lava. The hydrograph for this borehole (Figure A8.1.8) shows lower recession levels in 1975 and 1976. It is suspected that the borehole was used as an abstraction well during this period.

8.5.9 Victory Quarry Borehole (SK 07507691; ground level 355.2 m OD):

Like the Hucklow South Borehole, this is one of the boreholes that were carried out during investigatory work undertaken by the Derwent Valley Water Board (Raffety et al., 1953). The hydrograph (Appendix 8.1) indicates that ground water levels range between 330.26 and 345.25 m OD. It is possible that this is Peak House Borehole (Harrison, 1981, p. 57), however the grid reference of this borehole is given as SK 07517677, in the order of 14 m to the south of the Environment Agency record. Furthermore the depth of the published record was 130.10 m, whilst that given on the drillers' records for the Environment Agency location (supplied by the British Geological Survey) only extends to 106.68 m depth. However, it is possible that the borehole was deepened, because correspondence with respect to the North Derbyshire Water Board investigation, seen by this author and held in the Derbyshire Records office does make reference to changes to borehole depths during the course of the investigation. It seems likely therefore, that the borehole penetrated the Monsal Dale Limestone Formation, Miller's Dale Limestone and the Dove Holes Tuff, the Lower Miller's Dale Lava (between 91.09 and 116.8 m depth) and entered the Chee Tor Limestone at 116.80 m depth.

The driller's record indicates that during drilling, ground water was struck at depths of 18.59 m, 86.87 m and 92.96 m. The borehole log does not provide any evidence for the occurrence of fractures, or channels at 18.59 m and 86.87 m depth. At 29.20 m depth the borehole record indicates the occurrence of silicified bioclasts and patchy disseminated black clay at the base of a biosparite. The fracture at 92.67 m depth is likely to be associated with the unit at 88.72 to 88.9 m depth, described as "*Clay, ochreous*", or the underlying unit, which was recorded by the driller as "Sand", albeit at a different level, but immediately overlying the Lower Miller's Dale Lava. The rest water level upon completion of the borehole was recorded as 14.63 m depth (340.57 m OD), which indicates that confined ground water conditions were encountered, unfortunately the strike rise level associated with each strike was not recorded. At this location the dip of the strata is to the west, dipping beneath the Namurian strata that cap the limestones to the west. It is suspected that the source of the ground water is, at least in part, recharge from the Namurian strata. The master recession curve (Figure 8.11) comprises three indistinct steps; in this situation of confined ground water conditions it is not realistic to use the master recession curve to assess the hydraulic conductivity of the strata in the borehole. Furthermore, the master recession curve cannot be interpreted on the basis of the findings of the borehole log, because the zone of influence extends beyond the outcrop of the strata encountered in the borehole.

8.6 Characterisation of the rising limb portion of well hydrographs.

Smart (1999) suggested that flow velocities directed away from primary conduits are generally greater than those returning. Similarly, the recession curves show a much more rapid recovery than decline, indicative of hysteresis, as observed in studies of unsaturated flow in more isotropic lithologies and possibly attributable, at least in part, to the profile of the voids that form the porosity in the unsaturated zone

Constantz (1982) has shown that unsaturated flow is very sensitive to changes in viscosity, as demonstrated by the relationship of unsaturated hydraulic conductivity to intrinsic hydraulic conductivity of the soil:

$$K(\theta) = \frac{k_r(\theta)\, k p_w g}{\mu_w}$$

Where: $K(\theta)$ = unsaturated hydraulic conductivity

$k_r(\theta)$ = relative conductivity (ranging from 0 to 1.0) that is the ratio of the unsaturated hydraulic conductivity at a given moisture content (θ) to the saturated hydraulic conductivity

k = intrinsic hydraulic conductivity

p_w = density of water at a given temperature

g = acceleration due to gravity

μ_w = dynamic viscosity of water at a given soil temperature

A change in temperature from 2 to 25° C can cause unsaturated hydraulic conductivity to increase by an order of magnitude. The lower temperatures associated with autumn recharge and the associated increase in viscosity and decrease in hydraulic conductivity may account for why; following significant autumn storm events; the recession curve does not regain the same base level that is achieved earlier in the season and furthermore may also help to explain the sudden very rapid recovery of the recession curve. It is also interesting to note that Davies and Jones (1999) modelled advective heat flux via concentrated rapid recharge and conductive heat flux transfer by diffusive flow to demonstrate that heat carried through dissolutionally enlarged fractures can penetrate the heterothermic (Luetscher and Jeannin, 2004) vadose zone of the aquifer more rapidly, thereby lowering the bulk aquifer heat capacity. However, during periods of maximum evapotranspiration the advective component of the heat flux is diminished and conductive heat flux is thought to dominate throughout the growing season.

Analysis of the rising portions of the Bee Low Borehole hydrograph was attempted, but it was found that the form of the recovery curve could not be established because of the low frequency of monitoring. Furthermore, by definition the rising limb occurs during the period of heavier rainfall and the occurrence of more storm events also masks steps in the rising limb of the curve. Notwithstanding this, visual examination of specific curves does suggest that there is correlation between some of the breaks in the rising limb and the level of channels identified from the recession portions of the hydrograph. Furthermore, it is notable that the rise in groundwater-levels is coincident with periods of lower temperature. Statistically this is hard to isolate, because of the close association of rainfall with cold fronts and therefore lower temperatures.

8.7 Pumping test analyses.

Relatively few full-scale pumping tests have been carried out in the White Peak. There are a number of reasons why this is so. Given the depth to the water table in the plateau areas, and the requirement for rotary drilling to penetrate the limestone, the costs involved are prohibitive. Furthermore, such a test would require extreme care in planning, because of the triple porosity nature of karst, as exemplified by the findings of the water-tracing experiments described in Chapter 7 and the recession curve analysis of section 8.4. The problems associated with predicting localised flow vectors make the positioning of useful observation wells extremely difficult. Small-scale pumping tests appear to be carried out routinely to obtain specific local information within a single borehole, particularly where it is scheduled as a groundwater supply borehole, but there is little published information and very little data available for further analyses. Two sets of data have been made available for this research and these are described in turn below. In assessing the results it should be noted that, in general, pumping tests provide hydrological information that relates to the behaviour of the saturated zone and therefore this is not directly comparable with data obtained from the zone of dynamic storage.

In the autumn of 1970 a borehole at SK 14187332 (Monks Dale), with an estimated ground level of 175 m OD, was deepened, acidised and pump tested on 6 October 1971, for abstraction assessment purposes on behalf of the North Derbyshire Water Board. The borehole penetrated the basal part of the Miller's Dale Limestone Member; the Lower Miller's Dale Lava, Chee Tor Limestone Member and extended into the Woo Dale Limestone Formation and its lower dolomitised facies. During drilling, records were made of the most significant water strikes, but the associated rest water-levels have not been recorded. The inflows that were recorded were in the Chee Tor Limestone Member (at levels of 98.8 and 79.9 m OD) and in dolomitised Woo Dale Limestone (-39 m OD). It is suspected that the second strike was associated with a band of grey shale at 71.6 m OD and that the first was associated with a sand-filled channel. The strike in the Woo Dale Limestone Formation was associated with yellow brown dolomitic limestones, described on the log as *"softer beds e.g. Dolomitic shale"*. In a letter to the Nature Conservancy (copy held by the Limestone Research Group), dated 23 August 1971, C.H. Crombie (Engineer and Manager North Derbyshire Water Board) noted, "*There were certainly other points of ingress which could not be readily identified and there were clear indications of additional water below the 702 feet* [214 m] *mark. It is absolutely certain that all the water is coming from beneath the toadstone and since the yield of the hole was only 6, 000 gallons* [27360 litres] *per hour when the drilled depth was 460 feet* [140 m] *it is safe to assume that the majority of the water is coming from a depth in excess of this*". The latter part of this statement indicates that any confinement of the groundwater is not solely attributable to the lava.

In support of this observation, the draw-down curve, plotted as log draw-down against log time and as draw-down against log time (Appendix 8.4) is indicative of unconfined conditions, with delayed yield (British Standard BS 6316:1983) i.e. there is no evidence of groundwater confinement beneath the Lower Miller's Dale Lava, although the records clearly indicate the presence of sub-artesian water. It is suspected by this author that this is attributable to the occurrence of inception horizon related conduits in the Miller's Dale Limestone Member. The discharge rate was 26 litres/sec. Prior to and during the course of the pumping test, flow monitoring was carried out at the following locations: Monks Dale Farm Spring (SK 142736), Brewery Spring (SK 137731) and Ravensdale Spring (SK 171737). In addition, groundwater-level monitoring was carried out in Peters Dale Borehole (SK 132752) and Topley Pike Borehole (SK 101724). During the monitoring period the observation points do not appear to have shown any response to the pumping (Appendix 8.4).

An easterly hydraulic gradient is hypothesised for this part of the White Peak and the following should also be noted: Ravensdale Spring is fed by water "perched" above the Ravensdale Tuff; Brewery Spring and Monks Dale Farm Spring are considered by this author to be fed by water "perched" above the Upper Miller's Dale Lava; and the groundwater-level in the Topley Pike Borehole also appears to be "perched". Hence, the lack of response of the observation points to the pumping is not entirely unexpected. Further consideration the form of the hydrographs plotted for these monitoring points (Appendix 8.4) and correlation with the rainfall at Monyash Vicarage (BADC) has identified a number of interesting points:

i) Hydrographs for Monks Dale Farm Spring and Brewery Spring are of a similar form and show correlation with the rainfall 2 days prior to monitoring (Pearson moment correlation coefficients, determined using Microsoft Excel, of 0.42 and 0.49 respectively, indicating that up to 25% of the water level change can be accounted for by rainfall in the preceding two days). This is not surprising as they are both high level springs and have small, localised catchments, with a shallow depth of recharge, discharging water collecting above the Upper Miller's Dale Lava.

ii) The hydrograph for Ravensdale Cottages Spring takes a different form to the other springs. It shows no correlation with the rainfall 2-days prior to monitoring (Pearson moment correlation coefficient, determined using Microsoft Excel, 0.12) and marginally greater, but still not significant correlation with the rainfall 10 days prior to monitoring (correlation coefficient 0.19, accounting for only 4% of the change). Furthermore, the curve shows a greater tendency towards a seasonal recession, which suggests a significantly greater baseflow contribution, indicative of a larger catchment area. It is suspected by this author that this spring is related to an inception horizon in the Monsal Dale Limestone and therefore the form of the recession curve would be expected to take a similar form to that of the River Lathkill (Chapter 11).

iii) The form of the Topley Pike Borehole hydrograph is comparable with that of the Ravensdale Cottages Spring, although it has a greater degree of correlation with the rainfall 2-days prior to monitoring than Ravensdale Cottages Spring (Pearson moment correlation coefficient determined using Microsoft Excel, 0.30, indicating that potentially rainfall accounts for 9% of the change in the groundwater-levels). The Topley Pike Borehole is very close to an east to west-trending fault and the apparent response to rainfall could reflect surface water recharge to faults that has been observed in the area of Topley Pike Quarry (Chapter 4).

iv) Ground level at the Peters Dale Borehole is 242 m OD. The form of the hydrograph suggests a seasonally artesian groundwater level, with the groundwater recession falling below the base of the borehole.

The last stage of the pumping test comprised the monitoring of the rising head. This author has determined a hydraulic conductivity of 5.41 x 10^{-3} m/day by analysing the results as 'a rising head test' (Somerville, 1986).

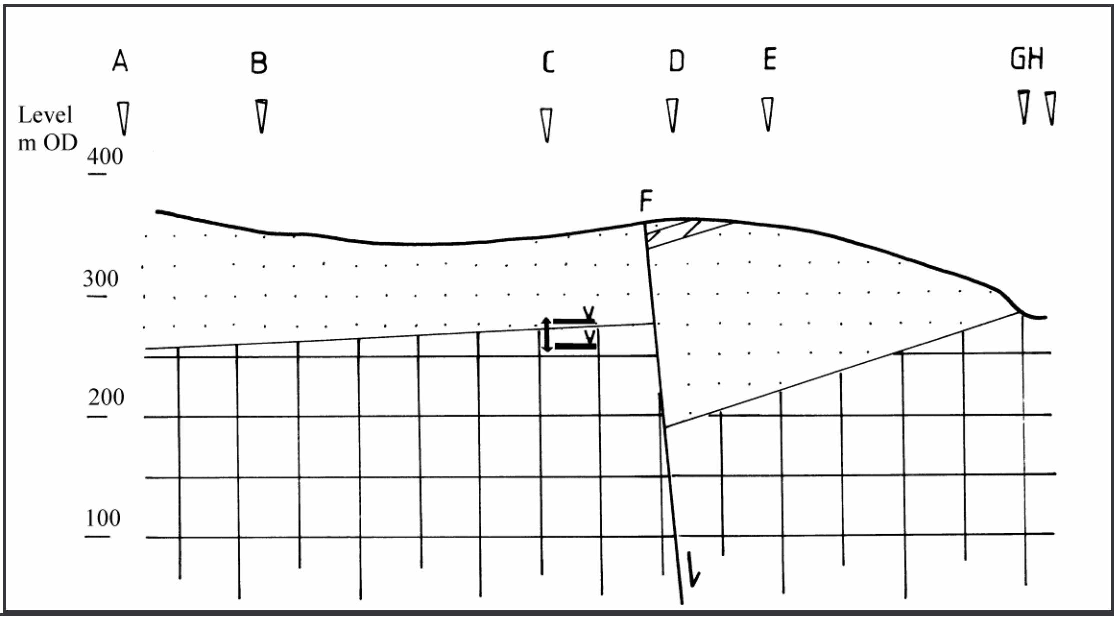
Interestingly, it would appear that for periods during the summer months the River Wye is 'perched', with surface water above the Lower Miller's Dale Lava at a higher head than the groundwater encountered in the borehole. The borehole was drilled at a level very close to river level and at the time of the pumping test a standing water-level of 2 m below ground level was recorded in the borehole (173 m OD) and the recovery level was 166.3 m OD. Furthermore, the evidence from the boreholes (Appendix 8.1) indicates that seasonal recession of groundwater continued until November 1971. However, artesian conditions are suspected to prevail during periods of high groundwater-levels.

The Staden Borehole (SK 072719) at the Rockhead Mineral Water factory was bored through the Bee Low Limestone Formation and into the underlying Woo Dale Limestone Formation to provide factory processing water. Groundwater-levels in this borehole ranged between 295.96 and 265.90 m OD, during the period September 2001 to March 2003. It is understood that explosives were used in the bottom of the borehole to increase its yield. A pumping test was carried out in this borehole in September 1989 and again in April 2003. The discharge rates were 5.3 litres/sec and 4.8 litres/sec respectively. Groundwater-levels were drawn down by 1.71 m and 1.72 m respectively. However it should be noted that the starting level for the commissioning test (1989) had probably not fully equilibrated. The plot of log draw down against log time for the 2003 test (Appendix 8.4, indicates an unconfined aquifer with a delayed yield (BS 6316:1983). Analysis of the rising head portion of both tests indicates permeabilities of 1.45 x 10^{-1} m/day (1989) and 1.69 x 10^{-1} m/day (2003).

The form of the Staden Borehole hydrograph (Appendix 8.5) takes a similar form to that of the Topley Pike observation well in the Monks Dale Borehole pumping test. It does not appear to show the pronounced seasonal recession common to the boreholes monitored by the Environment Agency. Cross-sections through this well indicate that groundwater-levels span the boundary between the Chee Tor Limestone Member and the underlying Woo Dale Limestone Formation (Figure 8.12). The form of the hydrograph and the groundwater-levels, which appear 'perched' on Edmunds' (1971) potentiometric surface, suggest that the borehole is fed by stored groundwater. Work carried out by Gunn (personal communication, 2004) has identified a correlation coefficient of 0.74 between the groundwater level in the Staden Borehole and the rainfall recorded at Buxton (this indicates that 28-day antecedent rainfall influences 55% of the change in groundwater-levels). It is the interpretation of this author that the response to rainfall reflects a proximal response that can be accounted for by the storage offered by the characteristic dissolutionally enlarged, subvertical fissuring in the Chee Tor Limestone Member (Chapter 5) and the connections that have been made with this storage by blasting in the base of the hole. It is considered likely that the main inflow to the borehole occurs at the boundary between the Chee Tor Limestone Member and the underlying Woo Dale Limestone Formation (Figure 8.12).

8.8 Baseflow contribution.

Evidence from boreholes and from exposures of limestone within quarry faces suggests that the limestone of the region can be divided into a number of hydrogeological zones. With increasing depth they comprise an unsaturated (vadose) zone, which includes the superficial cover and the epikarst (a zone of karstified limestone); a zone of fluctuating piezometric level (zone of dynamic storage, Smart and Hobbs, 1986) and the "saturated" zone. Passing through the vadose zone there are a number of dominant fractures, which are suspected to act as recharge channels (Chapter 5), accepting recharge both from dolines and from inception horizons within the vadose zone. It is evident from the records of the Monks Dale Borehole that flow within the saturated zone is also along a number of discrete channels.



Key: A SK 06977138; B Location of A515; C Location of Staden Borehole v-v indicates range of groundwater-levels; D Location of Staden Lane; E Location of unnamed lane; G River Wye; H SK 07557271. Vertical exaggeration X 2.

Legend:

Bee Low Limestone Formation

Lower Miller's Dale Lava

Woo Dale Limestone Formation

Figure 8.12: Cross-section Staden to River Wye.

Throughout the limestone there is a very wide range of pore sizes, which must be taken into account in the analysis of spring discharge data (Atkinson, 1977a; Ineson and Downing, 1964). The initial rapid fall recorded in the recession curves, following storm events, is perceived as quick drainage of larger pores, or quick flow (Atkinson, 1977a). In granular strata this is followed by drainage of smaller pores, in combination with arrival of more distal water. In karst environments flow in the unsaturated zone can be compared with soil unsaturated flow and the extremely rapid quick flow response can be seen as a separate domain. Baseflow is less likely to include a significant contribution of distal "quick flow", but instead comprises two domains comparable with macro-pore and matrix flow in the unsaturated zone of soil (Beven and Germann, 1981). Clearly there will be differences in the development of the epikarst in the plateau setting (location of most boreholes) compared with the valley setting (the location of most springs). Furthermore a borehole only intercepts a very small proportion of the aquifer. Nevertheless, it is considered that the boreholes should give a good indication of baseflow conditions (fracture/matrix flow) in the aquifer.

In the course of their work on the karst of Oak Ridge, Tennessee, Powers and Shevenell (2000) carried out field trials to determine whether water-levels in a well penetrating a conduit were directly proportional to flow at a nearby spring that discharges water from the same conduit. Accordingly, the analytical methods that were used were applicable to spring recession analyses. The detail associated with the channels draining the boreholes that have been investigated by this author is unknown. For comparative purposes some of the techniques used in spring analyses have been applied to the data from the boreholes.

Two of the techniques that have been applied were described by Padilla et al. (1994). In their analysis of springs Padilla et al. (1994) suggested that the discharge of a spring, during recession of groundwater in an unconfined aquifer, can be characterised in two ways, one as a multi-component drainage system with components of infiltration and baseflow (drainage via fracture/ matrix flow) being readily identified and the other as continuous drainage of a single reservoir.

Although the recession curves (borehole and spring) are comparable in form it is considered that the 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. Nevertheless, it would appear that there is an upper zone within which rapid drainage occurs. This zone of “quick” flow is most likely to represent drainage of larger pores and the capillary fringe. Beneath this, the hydrograph changes appear to be more subtle with a gradual slowing of the rate of recession, underlain by zones of lesser storage, represented by more rapid recession and therefore of lower hydraulic conductivity (Figure 8.12). If this interpretation is correct a greater depth of recession curve (height of the borehole water-level above base level) would be achieved in the limestones with lower matrix hydraulic conductivity.

It has been suggested by Padilla et al. (1994) that one of the best functions to describe the continuous function of a spring discharge recession hydrograph draining a single reservoir is Coutagne’s formula (1968):

$Q = CV^n$, where, Q is the flow of the spring, C is a constant; V is the stored volume available for draining via the spring and n is an exponent ranging between 0 and 2.

When n = 0 the aquifer empties at a constant rate, when n = 1, the discharge corresponds approximately to the equation for baseflow presented above and when n = 2 the solution represents an aquifer which discharges under a layered regime. This would appear to be more directly comparable with the situation monitored in the boreholes. Padilla et al. (1994) showed that:

$\alpha_{t} = \alpha_{0}/1+(n-1)\alpha_{0}t$, where: t is time, α is the recession constant at time 0 or time t as indicated by the subscript. The n value is derived from the ratio of adjacent α values, which have been derived for each step of the derived recession curves (Table 8.2).

It was found that n values between steps were very variable; however there was some consistency to the results when the ratio was determined from the uppermost and the lowest of the steps of the recession curve. It can be seen from Table 8.2 that the n values appear to fall in two groups: n=1.0 (the majority of the boreholes, and n=1.2 (Highcliffe Farm Borehole). Values of 1.00 are indicative of a constant rate of emptying of the reservoir (which with baseflow in seasonal recession is what would be anticipated). The value of 1.2 appears indicative of the greater significance of fractures (Appendix 8.3). Significantly higher values were determined for Nutseats Quarry and Dale Head Farm boreholes. It is interesting to note that these are boreholes with the lowest range in head and occupy valley bottom settings. The form of the combined recession curve for each of the boreholes shows a best fit with a third order polynomial trend line (for example Figure 8.13). This is representative of seasonal overflow (Worthington, 1991). In his assessment of recession curves Worthington (1991) was able to identify specific forms of recession curves in terms of overflow, underflow of fullflow springs. The form of the recession curves is characterised by the α value (Worthington 1991, Chapter 4). Consequently, Worthington (personal communication, 2006) suggested that this interpretation could be applied to the work carried out by Padilla et al. (1994). In Worthington's (1991) approach to the analysis, n values can exceed 2, more specifically, where a spring is an overflow spring receiving groundwater from a number of sources. Nevertheless, this interpretation cannot be directly applied to Nutseats Quarry and Dale Head Farm Borehole. It is also noticeable that these were the recession curves that were not stepped (falling within the gaining or low storage underflow spring of Worthington, 1991). The higher n values in this setting reflect a more continuous supply of groundwater associated with the valley setting.

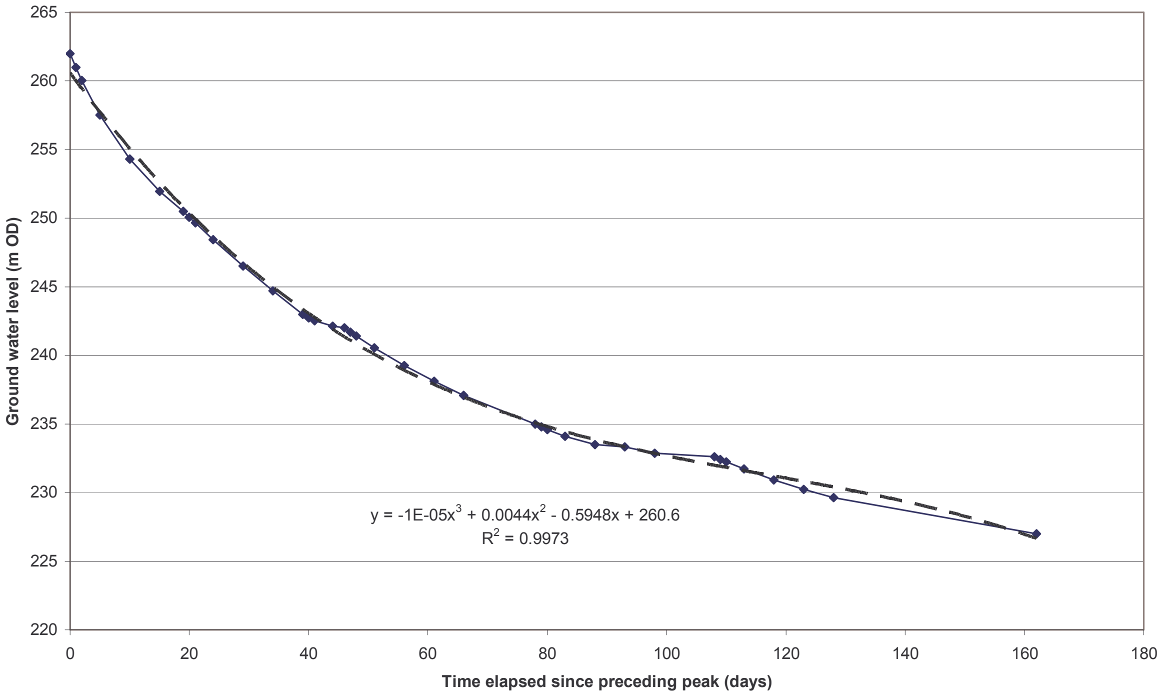


Figure 8.13: Master recession curve for Bull I' Th' Thorne Borehole, with best-fit third order curve.

8.9 Determination of hydrogeological parameters.

The heterogeneous nature of karst aquifers renders the determination of hydrogeological parameters extremely difficult. Furthermore, various authors have adopted different approaches to their determination (Atkinson, 1977, Headworth, 1971, Powers and Shevenell, 2000 and Shevenell, 1996), accordingly terms such as transmissivity and storativity in karst have tended to become more loosely defined. Clearly, in order to allow comparisons to be made between aquifers some consistency in calculation and terminology is called for. Notwithstanding such observations it is also considered that even where terminology is used strictly in accordance with the conventional definitions, such terms are not particularly well applied in karst, e.g. inherent within the concept of transmissivity is the assumption of predominantly horizontal flow, which is not applicable to large areas of karst. Furthermore the saturated thickness of a karst aquifer is particularly difficult to determine.

It is clear from examination of the hydrographs (Appendix 8.1) that the greatest range in well water levels (up to 51.67 m) is associated with the plateau setting (Bee Low, Bull I' Th' Thorne, Highcliffe Farm and Hucklow South boreholes). The range in well water levels for boreholes in the valley settings is considerably less, in the order of 3 m (Dale Head Farm and Nutseats Quarry boreholes). There was also a lower range of the groundwater-levels in the Oddo House Borehole, which is likely to reflect the higher hydraulic conductivity of the dolomitized areas of the Monsal Dale Limestone Formation and the fact that the groundwater, encountered in the Monsal Dale Limestone Formation, is suspected to be perched on shales and clay contemporaneous with the Lower Matlock Lava (Appendix 8.3). Similarly, the relatively low range in water levels associated with the Victory Quarry Borehole (14.99 m), is likely to reflect perching by the Lower Miller's Dale Lava. A low range of groundwater-levels (9.58 m) was determined for the Peak Forest Borehole. This borehole encountered confined groundwater conditions, thus the groundwater-levels are controlled by hydraulic conditions to the west of the borehole.

Headworth (1971) noted that many boreholes, like springs, show relatively rapid responses to rainfall events. Such a response is largely attributable to displacement of stored water, possibly in response to rapid groundwater infiltration comparable with the process described by kinematic wave approximation in soil (Germann and Beven, 1985) and observed for springs (Chapter 5). Accordingly, the time lag between a given rainfall event and the borehole response can be described as the rate of "apparent percolation" (Headworth, 1971). Values of apparent percolation are influenced by a number of factors, including vadose zone depth and hydraulic conductivity, the moisture content of the vadose zone prior to a rainfall event, which is influenced by the depth of the capillary fringe, and the continuity of fissure surfaces, which provide rapid routes for the flow of percolating atmospheric water. The results suggest that the rate of apparent percolation is independent of the range of water fluctuation in a given well. The high apparent rate of percolation associated with the Highcliffe Farm Borehole in the lower hydrogeological unit identified in Table 8.2, is interesting and it has been interpreted by this author as indicating a fault-guided recharge response to the borehole.

Values of matrix/fracture hydraulic conductivity have been crudely determined by interpreting the derived recession hydrographs as falling head tests in the boreholes (Somerville, 1986). The results (Table 8.2) indicate permeabilities in the order of: 10^{-3} m/day in dolomitised Monsal Dale Limestone (Oddo House Borehole), which is comparable with the results calculated from the Monks Dale Borehole pumping test (section 8.7); 10^{-4} to 10^{-5} m/day in the Monsal Dale Limestone (where the groundwater is perched and sub-horizontal fractures marginally better developed, as indicated by the lower range of the groundwater-levels) and Woo Dale Limestone; and 10^{-5} to 10^{-6} m/day in the Chee Tor Limestone Member (Bee Low Borehole). Soakage tests carried out in trial pits excavated (under the supervision of the Limestone Research Group) near the Staden Borehole indicate plateau epikarst hydraulic conductivity in the Chee Tor Limestone to range between 8 and 41 m/day. When compared with the findings of Worthington and Ford (1995a) the matrix hydraulic conductivities of the limestone, which contrast with the channel conductivities established by water tracing (Chapter 7), are relatively high for limestones of Palaeozoic age. The values are comparable with the hydraulic conductivity of clay. This may be more than coincidental, as it is clear that the clay wayboards form a significant zone of groundwater storage and transmission.

Headworth (1971) showed that storativity (Sy in Table 8.2) could be calculated from inflexions in the recession curve that can be attributed to specific rainfall events such that storativity can be calculated from the ratio of effective rainfall to the change in groundwater-level plus the recession portion of the recession curve. Atkinson (1977a) used a similar approach, but made allowance for soil moisture storage, accordingly for the purposes of this study the storativity, which corresponds closely with specific yield in the case of an unconfined aquifer, has been calculated from:

$$\text{Specific yield} = \frac{\text{Rainfall} - \text{estimated evapotranspiration} - \text{soil deficit}}{\text{Change in water-level} + \text{groundwater recession}}$$

Freeze and Cherry (1979, p. 61) suggest, "*The specific yields of unconfined aquifers are much higher than the storativities of confined aquifers. The usual range of S_y is 0.01 to 0.30. The higher values reflect the fact that releases from storage in unconfined aquifers represent an actual dewatering of the soil pores, whereas releases from storage in confined aquifers represent only the secondary effects of water expansion and aquifer compaction caused by changes in fluid pressure*." Fetter (2001, p. 102) suggests that values of storativity of 0.005, or less, are indicative of confined conditions. Values in the range 0.016 to 0.53 were determined using the above method; higher values generally coincided with the higher values of hydraulic conductivity and with valley settings. Interestingly, the value of specific yield that has been determined for the Peak Forest Borehole is 0.029 (Table 8.2). Evidence from the driller's record (section 8.5) suggests that groundwater in the Peak Forest Borehole is confined. The relatively high specific yield and the stepped nature of the master recession curve suggest that is only partially confined, reflecting more distal influences on the groundwater-levels. Values of specific yield using the Headworth method of calculation are relatively high. It is suspected that this is a reflection of the hysteresis described in section 8.5. It could also reflect the release of capillary water in response to barometric change associated with a given rainfall event.

Table 8.2: Hydrogeological parameters determined from the boreholes.

Borehole	Base of Hydrogeological Unit (m AOD)	α	Hydraulic conductivity (m/day)	Apparent percolation (m/day)	Sy	N	Levels of repeat recession bases (m AOD)
Bee Low	325.00 * 315.00 * 310.00 * 301.50	0.109 0.030 0.093 0.028	2.75×10^{-5} 8.67×10^{-6} 3.34×10^{-5} 8.9×10^{-6}	0.68 1.82 0.80	0.04 0.02 0.02	1.08	310.05
Bull I' Th' Thorne	250.50 243.00 * 242.00 235.00 232.60 * 227.00	0.064 0.013 0.251 0.023 0.085 0.025	Borehole construction details not known	7.75 1.62	0.02 0.05	1.04	234.38
Dale Head Farm	130.13 129.20	3.131 0.048	Confined water suspected	8.35	0.35 0.53	Not applic-able	129.63
Highcliffe Farm	256.10 253.20 * 249.20	0.131 0.217 0.028	1.53×10^{-5} 7.65×10^{-5} 1.05×10^{-5}	2.50 11.80	0.04 0.04	1.21	251.57 250.78 249.16
Hucklow South	257.00 254.00 *(253.28) 250.50	0.040 0.015 0.063	8.63×10^{-5} 5.16×10^{-4} 4.75×10^{-4}	1.92 5.75	0.03 0.02	0.98	254.96 256.79 256.03 252.86
Nutseats Quarry	109.14 *(109.60)	0.085	1.84×10^{-6}	0.29	0.34	Not applic-able	109.56 109.53
Oddo House	254.15 252.65 * 252.50 251.45 * 251.00	0.052 0.016 0.068 0.015 0.082	2.95×10^{-3} 1.10×10^{-3} 1.21×10^{-3} 3.93×10^{-3}	1.15	0.40	0.97	251.46 251.27 251.00
Peak Forest	331.90 * 330.62 330.50 *(330.80) 330.00 *(330.20)	0.050 0.017 0.546 0.062	2.70×10^{-5} 1.08×10^{-5} 2.70×10^{-4} 2.90×10^{-5}	0.946	0.03	0.99	330.54 330.41 330.18 330.00 329.56
Victory Quarry	334.50 * 333.50 * 330.00	0.054 0.067	Borehole construction details not known	1.97	0.03	0.99	336.49 335.13 334.53 333.40

* Confirmed by visual examination of combined recession curves (1977 – 1978), with level indicated by combined recession curve in brackets, if different to the master curve.

As Worthington (personal communication, 2006) has suggested, the calculated values of specific yield form a bimodal distribution. Eleven of the fifteen calculated values were in the range 0.02 to 0.05, which falls within the anticipated range of values for Palaeozoic carbonates. The remaining four values (in three boreholes) were in the range 0.34 to 0.53, which exceed anticipated values. The three boreholes with the highest values of specific yield were also the boreholes that show the smallest range in groundwater-levels (Dale Head Farm Borehole, Nutseats Quarry Borehole and Oddo House

Borehole, Table 8.1), located in valley settings. Accordingly, although the values appear to be elevated, they appear to provide evidence to confirm the greater storage associated with valley settings. In likening the form of the borehole recession curve to a spring hydrograph, Powers and Shevenell (2000) calculate the average non-conduit transmissivity of an unconfined aquifer from a baseflow recession curve from a spring using:

$$\text{Log}\,(Q_3/Q_4)=(T/S)(t_4-t_3)(1.071/L^2)$$

Where: L is the distance from discharge to groundwater divide; S is the storage coefficient (equal to Sy), as calculated by Atkinson (1977a, after Rorabaugh 1960, 1964); Q_3 and Q_4 are the consecutive water-levels in the borehole at times 3 and 4, indicative of the period over which baseflow was thought to be received by the borehole, i.e. the lowest part of the tripartite recession curve observed by Powers and Shevenell (2000).

Powers and Shevenell (2000) argue that this is representative of the matrix conditions in the aquifer, as confirmed by comparison of the results with field data (slug and pumping tests). Consideration of the master recession curves derived from the boreholes leads one to conclude that this method could not be applied to the White Peak boreholes, because the number of steps in the recession curve is variable. Furthermore, although the technique may be applicable to a borehole positioned in very close proximity to a spring, where the water table is more likely to be better defined and there is a greater likelihood of intersecting flow paths to the spring, it is the opinion of this author that fracture/matrix transmissivity in a triple porosity medium should not be likened to that of an isotropic medium and is likely to decrease with depth (reflecting a reduction in stress relief due to unloading by uplift and weathering) and therefore transmissivity calculated in terms of upper levels of the borehole and area of the aquifer between the borehole and the river are unlikely to be indicative of the more conventional hydrogeological definition of transmissivity (hydraulic conductivity x saturated thickness).

In his analysis of a chalk aquifer, Headworth (1971) used the more conventional method of assessing transmissivity in terms of saturated thickness. As Freeze and Cherry (1979) and Headworth (1971) note, the problem with this method lies in the assessment of the saturated thickness. Such an assessment would be made easier by the examination of accurately maintained drillers' records of exactly where groundwater ingress to the borehole takes place. In the absence of such records an assessment of transmissivity in the limestone aquifers is considered inappropriate and transmissivity has not been determined for these boreholes.

8.10 Relationship of recession curves and hydrogeological parameters to geology.

The findings of the analysis of the hydrographs have been summarised in Table 8.2 (p. 193). Changes in the gradient of the master recession curve for each of the boreholes appear to reflect a change of the hydrogeological conditions in the borehole. As Smart (1999) has modelled the changes in the hydrogeological condition can be brought about by a number of different karst features. Of particular interest to this dissertation is the identification of potential flow paths or channels, which could occur as conduits, sub-conduits (Smart, 1999), possibly related to inception horizons. The presence of conduits should be identifiable from the borehole log. Furthermore if an active conduit was encountered a low range to the groundwater-levels would be expected.

Theoretically it should be possible to make an assessment of the type of conduit by comparing the borehole log with the points of change in gradient of the recession curve. One problem with this analysis is that it relies on accurate correlation of the levels relative to Ordnance Datum. Much of the dataset that was provided by the Environment Agency and the British Geological Survey dates to the 1970s and some difficulty was experienced correlating the data (in particular borehole locations and ground levels). Furthermore drillers' records for some of the boreholes indicate partially confined groundwater, with groundwater-levels rising to a level above the strike level. Accordingly, the observations presented in section 8.5 represent the author's own "best correlation" of borehole logs and geology with the hydrographs.

With respect to establishing the geological settings that are favoured for inception, the analysis is dependent on correlation of ground levels and although this has proved difficult, the findings do indicate that fracture development appears to focus on: (i) breaks in the stratigraphic column, where inception horizons are associated with the boundary between the Eyam Limestone and the underlying Monsal Dale Limestone, as observed in the field (Chapters 10 and 11); (ii) stylolites, which are associated with the occlusion of porosity in the limestone, but appear to focus groundwater movement along them, as also observed by Schofield (1982); (iii) well sorted coarse arenites associated with inception, it is not immediately apparent from the boreholes, whether the porosity relates to the granulometric properties (good sorting), or the mineralogical composition i.e. it may be attributable to secondary porosity derived from dissolution of the crinoid stems (originally precipitated as metastable aragonite, Wright, 2002), or to aquitard development beneath, for instance due to silicification.

Fractures appear to be best developed in the Woo Dale Limestone (Highcliffe Farm Borehole), well developed in the Monsal Dale Limestone and less well developed in the Chee Tor Limestone. Results of water-tracing experiments demonstrated the significance of fracture/conduit flow in the Woo Dale Limestone (Chapter 7) and it was considered that this was largely attributable to flow along stylolites related inception horizons. The results of the falling head analyses indicate higher permeabilities in the Monsal Dale Limestone than in the Woo Dale Limestone and the Chee Tor Limestone.

8.11 Conclusions from Chapter 8.

The borehole data have proved useful in providing hydraulic parameters for the formations, which contribute to the conceptual model for the hydrogeology of the Wye catchment (Chapters 9 and 10) by contributing to the characterisation of matrix/fracture flow. Borehole hydrographs show a similar form of recession to that of the springs. The change in the form of individual steps of the recession curves indicates zones of changing hydraulic conductivity. Narrow zones of higher matrix hydraulic conductivity have been found to characterise the boreholes in the Monsal Dale Limestone Formation in particular and have been interpreted as representing the contribution of paleokarstic surfaces and clay wayboards. The association of fracture flow with specific horizons, such as the boundary between the Eyam Limestone and the underlying Monsal Dale Limestone (as observed in the Hucklow South Borehole, section 8.5) confirm field observations and provide confirmatory evidence that in addition to being the focus for conduit development, inception horizons have a laterally more extensive role in that they also provide a focus for fracture flow. The data from the boreholes have provided confirmatory evidence of the occurrence of confined flow, imposed by the low matrix hydraulic conductivity of the limestone. Whilst it would seem that the method of analysis is useful in identifying zones of fracture flow in the boreholes, analysis of the findings has not resulted in furthering the understanding of the characteristics that favour inception horizon development. The analysis of the borehole recession curves and the Staden pumping test have provided supportive information with respect to the higher storage and hydraulic conductivity of the Chee Tor Limestone Member where it is exposed at surface and fissures have opened in response to stress relief. This contrasts with its characterisation as an aquitard at depth.

Chapter 9: Towards a conceptual, pre-Roman hydrogeological model.

9.1 Introduction.

Analysis of the findings of Chapter 2 and chapters 4 to 8 has contributed to the development of a conceptual model, presented as a map of hydrogeological units (Figure 9.19) and a conceptual block diagram (Figure 9.20) for the hydrogeology of the subject area. Some of the key concepts, from each of the chapters, which support the ensuing model, are presented below (Table 9.1):

Table 9.1: Derivation of the conceptual model from the preceding research.

Chapter	Selected concepts
2: Geological Setting	i) The regional setting comprises three half grabens, for which there is geophysical evidence, formed around a basement high. Intra-shelf basins developed to the east. ii) The occurrence of lavas and clay wayboards with associated paleokarstic surfaces has been shown to have a significant impact on the hydrogeology and they may guide the development of proto-conduits along inception horizons. iii) Differing styles of sediment characterise the Woo Dale Limestone, the Bee Low Limestone and the Monsal Dale Limestone formations. Apron reef facies formed within the Bee Low Limestone around the northern and western margins of the fault blocks and micritic mud mounds formed within the shelf facies. iv) During diagenesis cementation was extensive, stylolites provide evidence of compaction and there was extensive dolomitization of the limestone, followed by mineralization. Mineralization targeted paleokarst associated with reef deposits and faults, in particular the strike slip faults of the Monsal Dale Limestone, which acted as a trap for mineralizing fluids. v) Some stylolites formed flow paths for mineralizing fluids and could subsequently have guided inception. vi) The paleokarst associated with mineralization is likely to form zones of groundwater storage. vii) The most recent stress history of the Derbyshire Dome has been one of uplift, imposing a tensional environment. viii) The former distribution of Silesian strata capping the limestone is likely to have influenced karst development. ix) There is strong structural guidance to the surface drainage.
4: Speleogenesis: the origin of karst aquifers	i) Present day per descensum dissolutional processes are dominated by carbonic acid reactions, with slow uniform dissolution being required for the gestation of long passages. Groundwater targets zones of higher permeability and these forms the focus of speleogenetic processes. Glaciations facilitated sediment abrasion. The inception of per descensum karst is likely to date to at least to the Mio-Pliocene and possibly earlier, i.e. at least to the time when groundwater began to breach the Silesian cover. ii) Other potential triggers of carbonate dissolution include sulphuric acid derived from pyrite and marcasite. iii) The regional karst context is one of a disrupted basin. The thermal springs indicate the presence of deep karst inception, considered by this author to have commenced as per ascensum karst as a consequence of basinal dewatering during diagenesis. The karst processes can only be speculated upon, they may have taken the form of "replacement solution" (Egemeier, 1981), or dedolomitization by gypsum (Bischoff et al., 1994). iv) Mineralization can be seen as a form of transverse karst, following dolomitization and the expulsion of relatively aggressive pre-mineralization fluids, evidence for which takes the form of paleokarst. Mineralization flow paths are discordant with the current hydrogeology and therefore whilst forming significant zones of storage, (e.g. Townhead vein, Magpie Mine) and vertical flow, they are not likely to form significant lateral flow paths. v) Dolines are associated with the featheredge of the Silesian strata, particularly in the northwest of the area and guide groundwater that targets the headwaters of the River Wye. vi) Mineral veins commonly appear to be a focus for valley development e.g. Lathkilldale Vein, Mandale Vein and Pasture Rake. The branchwork form of most caves suggests point recharge. There is significant evidence of structural guidance to inputs, the location of dolines commonly being related to faults or dominant fissures. The coincidence of upper Lathkill Dale and the Lathkill Cave system with the axis of the syncline provides further evidence for the importance of bedding related inception in speleogenesis. vii) The relative absence of caves in the Chee Tor Limestone Member, with the exception of a significant inception horizon towards the base and at the boundary of the Bee Low Limestone with the Woo Dale Limestone, has been noted. However the Miller's Dale Limestone Member is different, see viii, below. Quarry profiles indicate a greater degree of fissuring in the Bee Low Limestone than in either the Woo Dale Limestone or the Monsal Dale Limestone, with a significant degree of sediment infill to the fissures. Only specific dominant fissures and faults fully penetrate clay wayboards. The dominant fissures have a relatively uniform spacing and are suspected by this

Chapter	Selected concepts
	author to act as chimneys. viii) Inception horizons that have developed in the Miller's Dale Limestone, the Monsal Dale Limestone and the Eyam Limestone can be attributed to dedolomitization of diagenetic dolomite associated with paleokarstic surfaces and stylolites. ix) Both per ascensum and per descensum karst target paleokarst, which is attributed to structural guidance.
5: Regional hydrogeological setting	i) The Derbyshire Dome is situated immediately to the east of the North Sea/ Irish Sea parting, on the western side of the Eastern Ground Water Province, which imposes a regional hydraulic gradient to the east. The hydraulic gradient is reflected in the easterly flow of the surface rivers and the base level of springs, albeit that some springs comprise an outlet for confined water. ii) The topography of the Peak District can be considered as one of moderate relief in terms of the classification of White (1969). Thus conduit systems are in equilibrium with the local base level, with any perching being attributable to geology, rather than to relief. iii) Thermal springs provide evidence of deep-seated flow paths. To the east the Dinantian limestone dips beneath younger strata. The regional hydrogeochemistry provides evidence of underflow beyond the Derbyshire Dome; with bicarbonate waters giving way to calcium sulphate and chloride in an easterly direction. Further evidence comes from heat flow models and also from the drill stem pressures recorded in deep boreholes, more specifically at Eakring in boreholes drilled into the Limestone in a faulted anticline. Accordingly, it would appear that, using the terminology of Tóth (1963) the regional base level is that of the River Trent, the intermediate base level is that of the River Derwent and the local base level is that of the River Wye. iv) Deep boreholes have provided evidence of the presence of evaporites at the base of the limestone. Sulphur isotopes support the interpretation of high sulphur concentrations in thermal springs being derived from evaporites. This provides evidence for the concept of deep-seated flow paths and also suggests possible processes for karstification. v) Thermal springs are found on both the western and eastern sides of the Derbyshire Dome. The temperature of the thermal springs has been used as an indicator of flow depth. The variation in temperature of the springs points to a southeasterly hydraulic gradient. Worthington's (1991) empirical relationship between length of flow path, depth and the dip of the strata suggests flow paths with origins extending significant distances to the west of the western outcrop of the limestone, thus providing additional evidence for the difference in the groundwater chemistry between Buxton (to the west) and Matlock (to the east). This suggests that groundwater flow paths pass from the Goyt Syncline into the Limestone, which would provide a source of aggressive groundwater input at depth. vi) With respect to local recharge it is clear that there is allogenic recharge near the current margin of Silesian strata (via dolines and sinking streams). There may also be dispersed or concentrated allogenic recharge at depth, as implicated in v) above. Autogenic recharge is both concentrated via dolines and dispersed. Many dolines are associated with faults and dominant fissures, which are apparently less well developed away from the most recent Silesian boundary margins. In the Monsal Dale Limestone Formation dolines appear to be associated with solution hollows.
6: Hydrogeochemistry	i) By grouping the groundwater chemistry of the springs according to the putative recharge formation (see Table 6.2) and giving due consideration to the fact that flow paths pass through more than one formation certain trends have been related to formations, namely: - The total dissolved solids calculated by SOLMINEQ. GW was found to be higher in the Woo Dale Limestone than in the Bee Low, or the Monsal Dale limestones. - The chloride content tended to be higher in the Woo Dale Limestone than in the Bee Low and the Monsal Dale limestones. It has been suggested by this author that this could be attributable to formational fluids associated with mineralization, but alternative hypotheses have also been presented. - Total hardness and calcium concentrations were generally found to be higher in the Woo Dale Limestone and were generally notably lower in the Bee Low Limestone, which implies active dissolution in the Woo Dale Limestone and rapid through-flow, or low solubility in the Bee Low Limestone Formation. - Magnesium concentrations were notably higher in Lees Bottom 3 Spring and also, albeit to a lesser degree, in Topley Pike Spring, which has been attributed to flow through dolomitized areas of the Woo Dale Limestone Formation. - The saturation index with respect to calcite was generally found to be lower in the Bee Low Limestone, this could reflect more rapid through flow times, a greater number of input points, lower solubility of the Bee Low Limestone, or "armouring" by the sediment fill encountered in the extensive fissuring in the Bee Low Limestones. Particularly high saturation indices with respect to dolomite and calcite were achieved in some of the springs in the Monsal Dale Limestone. It is possible that the high saturation levels achieved in the Monsal Dale Limestone reflect the occurrence of closed flow paths and also of hydrocarbons associated with early stages of mineralization and associated with chert dissolution. Clearly, these observations need to be considered within the context that each spring is likely to be fed by more than one source. ii) It was found that the fluoride content is the best indicator of mineralized flow paths. iii) Groundwater chemistry appears to vary with flow conditions, as exemplified by: - Potassium and sulphate concentrations commonly appear to increase with discharge; therefore a source within the superficial deposits has been postulated. - Evidence has also been presented for deeper sources of sulphate in Lees Bottom 3 Spring and in the thermal springs. - It was found that particularly low concentrations of carbon dioxide were determined during the period of groundwater recession in the Monsal Dale Limestone. Long closed flow paths associated

Chapter	Selected concepts
	with inception horizons achieve high levels of saturation. The broad range of values is likely to reflect the change from vadose dominated to phreatic dominated inputs to the springs. iv) It would appear that the strong structural guidance to inputs has facilitated downward development of flow paths, such that vadose and phreatic flows target the same springs within a given catchment. v) That many of the thermal springs exhibit chemical enrichment supports the hypothesis of groundwater chemistry evolving to achieve equilibrium with rock chemistry.
7: Water Tracing as a means of determining flow vectors	i) The apparent flow rate of a tracer gives an indication of groundwater conditions, with confined flow rates generally being lower than unconfined flow rates. ii) Different flow paths operate at different groundwater stages. iii) During high groundwater conditions groundwater stored within faults associated with the Taddington Anticline forms a groundwater divide, which recedes as groundwater levels fall. iv) Dispersed flow is more evident where flow paths at the boundary between the Woo Dale Limestone and the overlying Chee Tor Limestone Member have been interpreted, as evident in tracing from Chelmorton Sewage Treatment Works, with further evidence from the Illy Willy Water trace. In the former dye trace, positive connections with springs at a range of levels and also in higher formations, rather than simply targeting Magpie Sough (the lowest positive connection), suggests that confined water rises at a number of locations including: Knotlow, Great Shacklow, Lees Bottom 2 (lower) and Bubble Springs. v) Evidence for groundwater recession isolating portions of conduits comes from the recovery of dye from the Illy Willy Water trace following a period of non-recovery during the peak of the groundwater recession, sixteen weeks after dye injection. vi) Apparent flow rates to Bubble Springs were lower in high groundwater conditions than in the low groundwater conditions, suggesting an increasing degree of confinement of groundwater during high groundwater conditions. vii) The results of dye-tracing appear to confirm that conduit development is rare in the Chee Tor Limestone Member, possibly owing to its high purity and massiveness. The low permeability of the matrix of the Chee Tor Limestone Member is such that when it is faulted against water bearing strata it acts as a barrier to groundwater flow causing it to move vertically within fault zones. Where the Chee Tor Limestone Member is exposed, or lies close to surface and has experienced stress relief, solutional enlargement of fissures provides the rock permeability, as indicated by dye-tracing in the trial pits at Staden (section 7.3.7). viii) High flow rates from Cunningdale Shack, Dove Holes and Tunstead Quarry to Wormhill Springs suggest the presence of a mature karst system, possibly associated with the dominant east to west faulting of the Ashwood Dale Anticline. ix) Dye injected in the area of Calton Hill and to the north of Hardyhead Sough tail was not recovered. This limestone comprises the Miller's Dale Limestone. It would appear that there is a dominant inception horizon associated with this stratum (Chapter 4), which takes groundwater southeast beneath the Monyash Syncline. x) The results of dye-tracing tests carried out in the northern part of the Derbyshire Dome indicate that the mineral veins exhibit low permeability, with the potential to form groundwater divides, acting as zones of groundwater storage, either within, or backed up against the mineral vein. xi) The dye-tracing tests have helped to define further the position of groundwater divides, thus facilitating modification of the map of principal hydrogeological features (Downing et al., 1970), see Figure 9.18.
8: Groundwater parameters derived from monitoring wells	i) Historic groundwater level data obtained from the Environment Agency were plotted. It was found that the borehole hydrographs show a similar form of recession curve to that of springs and reflected in river discharge, as in Lathkill Dale. The hydrographs were used to look at the hydraulic properties of the matrix/fracture flow of each formation. ii) It was considered that where the recession base had been repeated on a number of years the base could be indicative of a zone of higher permeability such as a conduit, and it was found that a number of bases were repeated. iii) For each borehole a master recession curve was derived. Stepped curve forms suggest zones of differing permeability. More uniform curves indicate a greater degree of isotropy to the permeability. iv) For each portion of the curve the hydraulic conductivity was calculated. The hydraulic conductivities were low, ranging between 8.67×10^{-6} and 3.93×10^{-3} m/day. It was found that within any borehole the maximum variation in permeability was one order of magnitude. Attempts were made to correlate these zones with lithological descriptions, but because of the age of the data and missing logs this was commonly impossible. Typically the hydraulic conductivity of the Bee Low Limestone was found to be in the order of 8.67×10^{-6} and 3.34×10^{-5} m/day; of the Monsal Dale Limestone 4.75×10^{-4} to 5.16×10^{-4}m/day, and of the Eyam Limestone was 8.63×10^{-5}m/day. The hydraulic conductivity of the dolomitized Monsal Dale Limestone (Oddo House Borehole) was in the order of 1.1×10^{-3} to 3.93×10^{-3} m/day. Higher hydraulic conductivities appear to be attributable to the presence of inception horizon related flow paths and possibly also to the presence of clay wayboards. v) In the boreholes that penetrated the Monsal Dale Limestone (Hucklow South and Oddo House boreholes in particular) and in the borehole in the Woo Dale Limestone (Peak Forest Borehole) narrower zones (0.12 m to 3.0 m) of higher matrix hydraulic conductivity were determined from the combined recession curves. It is thought likely that these correspond with the presence of clay wayboards and associated paleokarstic surfaces, which formed at the top of fining-upward sequences, thus the zone of higher hydraulic conductivity comprises the wayboard and the basal (coarser) component of the overlying unit. vi) The relatively high hydraulic conductivity associated with the dolomitized Monsal Dale

Chapter	Selected concepts
	Limestone encountered in the Oddo House Borehole is comparable with that determined for the dolomitized Woo Dale Limestone in the Monks Dale Borehole. vii) Pumping tests carried out at Staden indicate that at this location the Bee Low Limestone (Chee Tor Limestone Member) overlying the Woo Dale Limestone forms an unconfined aquifer with a delayed yield and a hydraulic conductivity in the order of 0.145 to 0.169 m/day. The relatively high matrix permeability indicates a contribution from fracture flow, suggesting that the borehole is fed by the solutionally enlarged fractures in the Chee Tor Limestone. viii) Data from the Monks Dale Borehole pumping test indicate that for part of the year the River Wye is perched above groundwater in the underlying limestone aquifer

9.2 A detailed consideration of the stress history and history of fluid movement in the study area.

It is the opinion of this author that an understanding of the stress history is a key to understanding the history of fluid movement and therefore the hydrogeology of the study area. This author's interpretation of this history has been summarised in Table 9.2, below:

Table 9.2: Summary of the chronological stress history and history of fluid movement in the Visean Limestone Series of the White Peak.

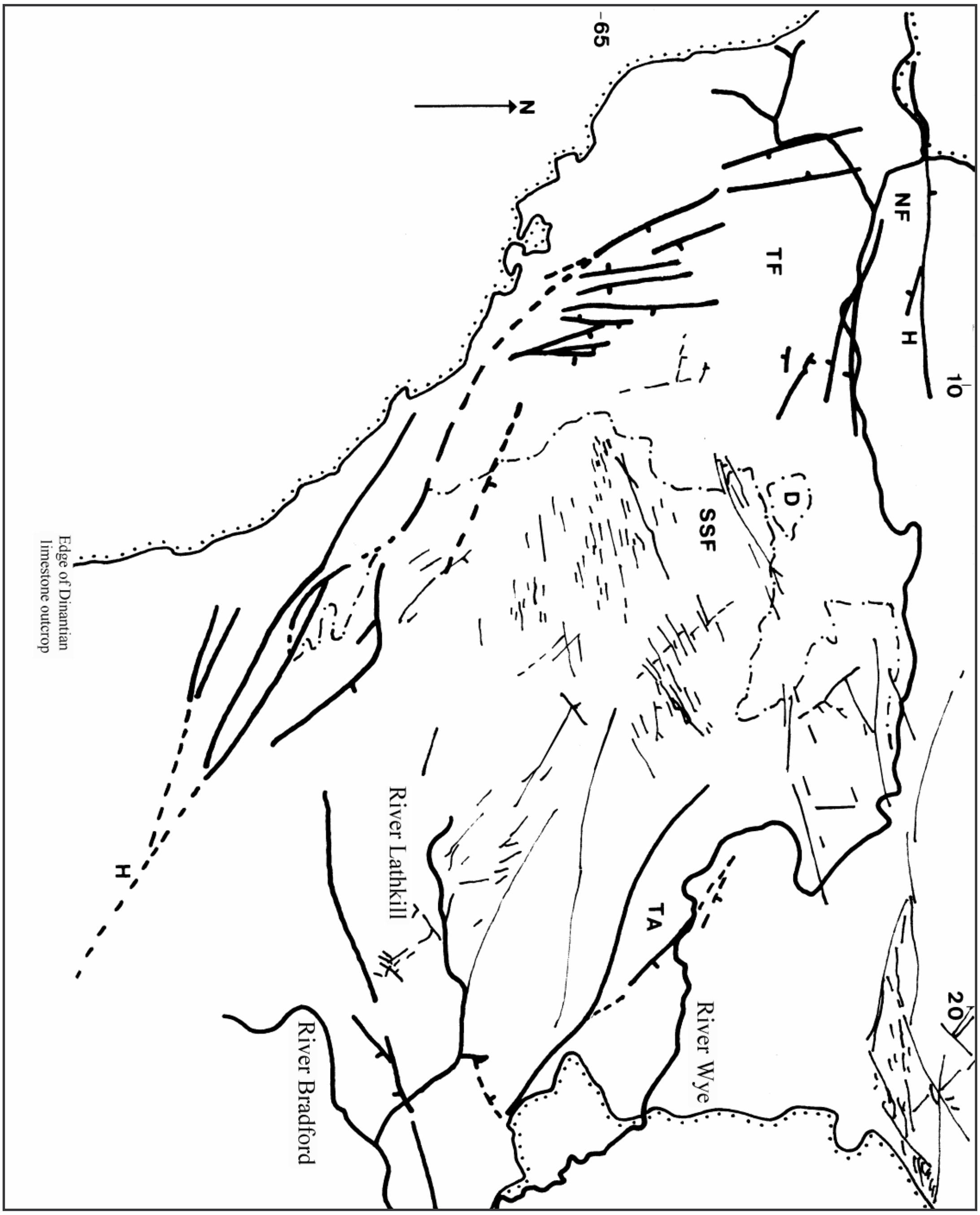
Approximate Age	Event
2-0 Ma	Valley development associated with glaciations, with associated loading and unloading of the underlying strata. Ice melt associated with doline development, extensive dissolution of fault zones and valley formation. Cambering and valley bulging.
60-2 Ma	Most of the British Isles formed a land area throughout the Tertiary. Associated with relative falls in base level (Warwick, 1962), this was largely a period of erosion and stress relief. Climatically it was a period of cooling. Terrestrial Pliocene deposits were laid down at Brassington. Walsh et al. (1972) suggested that they were deposited on a surface that would now be at 450 m OD, subsequent work is more ambiguous (Walsh et al. (1999). Walsh et al. (1972) suggested that material eroded from the Hercynian massifs has only amounted to a few tens of metres since the Pliocene.
75 Ma	During the Cretaceous the Derbyshire Platform remained emergent until the late-Cretaceous. Evidence for late-Cretaceous sedimentation is scant and is derived from an interpretation of the regional occurrence of sedimentation (Cope et al., 1992).
138 Ma	Marked uplift in the late-Jurassic exposed the Derbyshire platform, as part of the Anglo-Welsh landmass, which is suspected to have been of low relief (Cope et al., 1992).
155 Ma	Widespread marine inundation occurred in the late-Jurassic, associated with a pre-existing tensional setting associated with graben evolution. Initially shales were deposited, giving way to limestone and then again to shale as seas deepened.
	The tensional setting was linked to rifting in the North Atlantic, and continued through the Jurassic. Horst and graben movement was such that the Derbyshire shelf was exposed as part of the Pennine High in mid-Jurassic times (179 Ma). The climate was humid and warm. During the early-Jurassic predominantly marine deposition occurred, giving rise to shales (Cope et al., 1992).
235-207 Ma	The bare, desert landscape of the Permian, described below gave way, in the mid-Triassic, when the area lay to the south of a receding hilly region, to a sabkha type of environment. Later the sabkha environment gave way to fluvial/estuarine deposits, probably including red shales (Cope et al., 1992) and then to rapid marine inundation and the shales and limestones that were laid down in the late-Triassic. The tectonic setting remained largely tensional. The climate was monsoonal.
260 – 251 Ma	Throughout most of the Permian the North Derbyshire block comprised bare rugged hills with screes. The end of Permian had reduced the landscape to a rock desert (Cope et al., 1992). The climate was largely one of desert conditions. The tectonic setting was largely tensional, thus this was a period of significant stress relief.
	Variscan orogeny and inversion occurred as a result of north to south crustal extension. This was associated with doming above an upwelling of the asthenosphere and tilting about a northwest to southeast axis, with greater uplift to the east (Quirk, 1993). Tensional conditions prevailed, associated with uplift.
300 Ma	Maximum depth of burial (in the order of 2.5 km). Mineralizing elements reached a maximum concentration in Zone 4C cements, as homogenization temperatures reached 117 to 200° C. This was the only zone with significant barium concentrations. Hydrocarbon migration had ceased (Hollis and Walkden, 1996). Northeast to southwest joints developed as mineralization occurred. Zone 4D, post mineralization cement.
	Mineralization associated with Zone 4B cement and increasing salinity to fluid inclusions (Hollis and Walkden, 1996).
	Deep burial as a result of sagging due to post-rift cooling (Quirk, 1996) and as evident in the precipitation

Approximate Age	Event
	of Zone 4 cement (Hollis and Walkden, 1996). Zone 4A cement associated with hydrocarbons and an increase in Fe and Mn.
	Crustal stretching is thought to have given way to compression (Leeder, 1988). Faulting suggests initial eastnortheast to westsouthwest compression, giving rise to northnorthwest to southsoutheast and northnortheast to southsouthwest faulting (section 9.3), followed by north to south compression giving rise to the east to west trending faults that define the Wye valley to the east of Buxton and indicative of an anticlockwise shift in the stress field (Quirk, 1993). Dolomitization of the Woo Dale Limestones in Woo Dale, fluids rising up the half graben, from the southeast. Dolomitization occurred in more than one phase with some fracturing, possibly hydro fracturing in between, the second phase showing an increase in Fe and Mn (Schofield, 1982; Appendix 2.1 of this thesis).
320 Ma	Pressure dissolution and stylolite formation, associated with localised dolomitization, as the limestones became buried beneath in the order of 2 km of Silesian sediments.
320 Ma	Shallow cement: Zone 3 (volumetrically the most significant cement) associated with the uplift to the east (Walkden and Williams, 1991) and therefore elevated temperatures resulting from burial and high local heat flow. An increase in fluoride concentrations occurred between Zone 3A and Zone 3B cementation, associated with the first phase of hydrocarbon emplacement, at temperatures in the order of 82 to 125˚ C (Hollis and Walkden , 1996).
320 – 302 Ma	Namurian and Westphalian sedimentation. These sediments were laid down in cyclic sequences indicative of fluvio-deltaic conditions alternating with marine conditions, during increasingly humid climatic conditions, with sediment being derived from the northeast as a result of uplift. Marine incursions became increasingly rare in the Westphalian.
335-320 Ma	Phreatic meteoric cementation, with bright cathodoluminescence zones that correlate over several kilometres and relate to periods of slower cementation, possibly relating to waxing and waning of the Gondwanan ice sheet (Walkden and Berry, 1984 and Walkden and Williams, 1991).
	Emergence, calcrete formation and then deposition of the Longstone Mudstone.
	Emergence and then deposition of the Eyam Limestone Formation associated with the development of deeper water facies, possibly due to the final wasting of the Gondwanan ice sheet, with two intra-shelf basins forming: reactivation of the Ashford and the Stanton basins (see below) with the Long Rake Fault forming along the northern edge of the Stanton Basin in response to ongoing crustal stretching. Local meteoric vadose cementation.
328 Ma	Uplift prior to deposition of the Monsal Dale Limestone. Evidence from the fault pattern suggests that this was associated with north-south compression and the development of the east-west faulting of the Woo Dale Limestone. Downwarping in the area of the Ashford Basin, with tensional faults developing along its northern edge and periodic uplift during the deposition of the Monsal Dale Limestone. Bioturbation indicates periods of exposure, as does the presence of clay wayboards in the paler facies. Deposition of lava deposits. Local meteoric vadose cementation. The occurrence of mammilated karstic surfaces (Walkden, 1970) is perhaps the earliest evidence of non-depositional water within the limestone.
	Deposition of the Bee Low Limestone, initially very fine-grained sediments. Deposition as fining-upward sequences, with a depositional dip to the east and southeast, with emergence, karstification and deposition of volcanic ash deposits. Deposition of the Lower Miller's Dale Lava. Local meteoric vadose cementation. Onset of increased tectonic activity associated with crustal stretching in the area of the Midlands and the Peak District, a response to the closure of the proto-Tethys, juxtaposition of Africa against Europe and the onset of the Hercynian orogeny. Onset of the formation of the Widmerpool Gulf, the Edale Gulf and the Gainsborough Trough half-grabens (Cope et al., 1992). The direction of extension was largely northeast to southwest.
345 Ma	Deposition of the Woo Dale Limestone, initially as shallow-water, open-shelf deposits with early meteoric water cements, giving way to shoal deposits and then lagoonal facies, in a near-equatorial latitude.

In rock mechanics it is conventional to define the stress field associated with failure in terms of three perpendicular components, or axes (Farmer, 1983), namely σ_1 (the principal stress), σ_2 and σ_3 (the minor stress component). Failure occurs along the strike of σ_2. In considering how the stress history has impacted on the hydrogeology it is first worth considering how the strata respond to both loading and unloading. In the simplest case, given the oblate spherical form of the earth, sediments at the base of a sedimentary basin will be subject to vertical loading imposed by the accumulating sediment pile above and lateral pressures resulting from the increasingly confining nature of the basin with depth. The response to the stresses may be folding, faulting or a combination of both. Pure limestone can be described as a brittle material in that the ratio of the deformation modulus, E, to the compressive strength, σ_{cf} is high and the tensile strength of the rock is in the order of 6 to 10 times lower than the compressive strength (Farmer, 1983), therefore it is more prone to faulting than a material such as clay. This is reflected in the fissure pattern that is seen in the limestones, whereby, only some "dominant fissures" actually pass through the clay wayboards, whilst other fissures appear to end at the clay wayboards (Chapter 5). Within the area of this investigation the dominant compressional faults are the east west faults in Ashwood Dale, as indicated on Figure 9.1.

Figure 9.1: The main fault sets and their interpretation.

Key:
H Location of postulated half graben development associated with the Bakewell Fault and the Cronkston-Bonsall Fault
TA Taddington Anticline associated with growth faulting
TF Tension faulting evident in Bee Low Limestone (more brittle)
NF Normal faulting
SSF Strike slip faulting associated with the Monsal Dale Limestone, reflected east northeast to west southwest compressional faults and northwest to southeast tensional faults.
D dolerite intrusion at Calton Hill
Edge of the Dinantian Limestone outcrop
Area underlain by Monsal Dale Limestone



With uplift the compressional stresses revert to the tensional stresses, which have predominated since uplift at the end of the Carboniferous. Tensional faults predominate in the Bee Low Limestone (particularly evident in the Chee Tor Limestone Member), as indicated on Figure 9.1 where they are seen as northnorthwest to southsoutheast faults (at some locations occurring with a conjugate fault).

In the Monsal Dale Limestone clay wayboards and weathered lavas have facilitated shearing at the boundary between the volcanic strata and the limestone, thereby rotating the stress fields in an anticlockwise direction and resulting in strike-slip faulting in the Monsal Dale Limestones. Thus, compressional faults are represented by the eastnortheast to westsouthwest trending faults, possibly also driven by hydro-fracturing during mineralization. Similarly, tensional faults, which are less well developed in the Monsal Dale Limestone, are represented by the northwest to southeast trending faults. Gowan's report on Mawstone Mine, dated 1930, observed "*that the rakes which run in a northeast by southwest direction are usually poor in ore content, whereas the northwesters run at right angles and are usually ore bearing*". This would appear to provide further support for the compressional nature of the northeasterly mineral veins (DRO D359Z/239(9)). However, it should also be noted that the mineral veins may also have been guided by pre-existing fault sets resulting from the intrusion at Calton Hill, which pre-dates the mineralization, but this influence is not considered to have had great significance because the same fault orientations are maintained farther north in the absence of the volcanic intrusion. The evidence suggests that the maximum tensional forces were along a northeast to southwest axis and that they probably commenced with uplift during the late Carboniferous, thereby causing northwest to southeast strike faults to become targets for pipe deposits. Evidence in support of the later development of the pipe deposits comes from the paragenetic sequence identified by Worley (1978) for Hubbadale Mine, where barite was identified as a late-stage mineral. This also concurs with the association of late stage 4C cement with barite (Hollis and Walkden, 1996).

The brittleness of the Chee Tor Limestone Member of the Bee Low Limestone has given rise to its characteristically intense jointing and the permeability associated with the jointing has been increased by dissolution along the joints (or fractures). It has already been noted (Chapter 3) that in the area of the limestone quarries, to the south of the A515, the jointing has been subject to considerable sediment infill. Beds of lava demonstrate a material response similar to that of the Chee Tor Limestone Member. By contrast, the material response of the Miller's Dale Limestone Member is more comparable with that of the Monsal Dale Limestone Formation and therefore it would seem that hydrogeologically the Miller's Dale Limestone can be grouped with the Monsal Dale Limestone (Table 9.5). This influence is also evident in the form of the valley of the River Wye, for instance it has been noted by this author that the valley is narrower where the river passes through the Chee Tor Limestone.

Mineralized faults act as zones of storage for groundwater, with little evidence to suggest significant lateral flow along the faults, as indicated by the results of dye-tracing (Chapter 7). However, where the faults result in the juxtaposition of a permeable zone, such as a conduit, against a less permeable zone, particularly associated with the Chee Tor Limestone Member, the fault guides the flow path either upwards or downwards to another more permeable zone. Thus the faults are zones of significant vertical flow. This effect is probably influential in directing deep groundwater to the thermal rising at Buxton.

It is the opinion of this author that the dolomitization associated with the Wye valley is inextricably linked with the stress history. The dolomitized strata are exposed in a zone between the dominant east-west faulting that broadly defines the Wye valley. The petrography of the dolomitization has been described by Schofield (1982), with evidence supporting two phases of dolomitization with iron and manganese being associated with the later phase, which was separated from the first phase by a period of leaching and collapse. Schofield (1982) also observed that the second phase of dolomitization tended to move to a higher stratigraphical level than the first, yet the Bee Low Limestone Formation largely escaped dolomitization. Schofield (1982) noted a paucity of stylolites in the dolomitized Woo Dale Limestones. It has been suggested above that the east-west faults in the Wye valley are likely to have been reactivated at the peak of the compressional forces during burial. Furthermore the dolomitization described by Schofield (1982) has similarities with that of non-seam solution of Wanless (1979), although, as Schofield (1982) observed, an open system has been implicated; and it is known that the magnesium resistance to hydration, which is the main resistance to dolomitization, is overcome with pressure and increased temperature (Tucker and Wright, 1990). Dolomitization also favours reducing conditions, which are indicated by the presence of iron and manganese. Conceptually, it is difficult to separate the pressure dissolution from fluid migration and metasomatism, which would also have lowered the compressive strength of the strata; and from the neomorphism that was also observed by Schofield (1982). Schofield (1982) suggested that the source of the magnesium-rich fluids was Namurian shales, with the iron and manganese being released from the basement rocks.

As identified in the Monks Dale Borehole, the matrix of the dolomitized Woo Dale Limestone generally exhibits greater hydraulic conductivity than the non dolomitized limestone, which suggests that it should be considered a separate hydrogeological unit. The extent of pervasive dolomitization, as indicated in published borehole logs (Harrison, 1981), has been plotted by this author and was used in the preparation of the map of hydrogeological units (Figure 9.19).

9.3 The influence of cyclic sedimentation.

Syn-depositional changes in lithology throughout the shelf limestones generally take the form of fining upward sequences (Gutteridge 1989; Walkden, 1987; section 2.4 of this thesis). Higher primary permeabilities in the limestone appear to have been associated with the coarser-grained part of the sequence. Therefore, mesogenetic basinal dewatering is likely to have been initiated along the basal beds of depositional cycles. In the Monsal Dale Limestone in particular, vertical groundwater movement is likely to have been restricted by the presence of clay wayboards (occurring at the top of each of the upward-fining sequences). This suggests that the earliest flow paths that were associated with basinal dewatering are likely to have formed immediately above clay wayboards. It would appear that these zones became the target for later phases of fluid movement, in particular during mineralization.

9.4 The importance of clay wayboards and stylolites.

At first sight the importance of clay wayboards is self-evident. Clay is of low vertical permeability and therefore acts as an aquitard, impeding vertical groundwater flow. In practice the situation is considerably more complex and the clay wayboards themselves are worthy of further consideration. Walkden (1970 and 1974) broadly described the clays as potassium bentonite clays. However using x-ray diffraction techniques he determined that the clay wayboards fell within three groups, namely:

A. Longcliffe Clay 1: illite and kaolinite (77% illite, 10% kaolinite with 13% smectite), with anatase (titanium oxide).

B. Prospect Clay 2: illite and accessory anatase.

C. Monsal Clay 2: illite with peak displacement indicative of mixed layering with smectite (probably olivine derived).

The dominance of illite in these clays suggests that their engineering properties are comparable with that of marine clay, for example London Clay. To gain further information a sample of clay wayboard associated with an inception horizon in the Miller's Dale Limestone in Miller's Dale (SK 14057330) was collected and sent for analysis to determine the Atterberg limits (Figure 9.2). The clay was found to be of high plasticity and medium shrinkage potential. This indicates that the clay has a potential to shrink and swell with changes in moisture content. Thus the clay wayboards themselves form zones of groundwater storage, fed by a network of fissures in the limestone in the zone of dynamic storage (Smart and Hobbs, 1986). Potentially the clay wayboards provide two zones of groundwater storage and lateral groundwater movement. Firstly there is the zone between the clay and the limestone, from which the clay initially shrinks. Secondly, there is the water bound by the clay, which is expelled as the effective stress imposed by the limestone increases during the seasonal groundwater recession. Groundwater is replenished by subsequent rainfall events, the clay swelling as it restores its natural moisture content (close to the plastic limit). This concept may go some way to explain why some of the borehole seasonal recession curves indicate that the permeability of the zone of dynamic storage (Smart and Hobbs, 1986) is comparable with that of clay. The clay is also likely to form a significant component of matrix storage in both the Monsal Dale and Eyam Limestone formations.

The presence of the clay wayboards has been shown to be important in another aspect of the hydrogeology and that is with respect to the development of inception horizons, thereby influencing channel and conduit flow. The occurrence of zones of dedolomite associated with non-sutured seam dissolution (Wanless, 1979) was described in chapter 4 of this thesis. The association of dolomitization and micro stylolites with clay wayboards has been observed by others including Berry (1984) and Schofield (1982).

The platy material evident in the microstylolites is thought to comprise thin seams of the clay, which act as glide surfaces along which stress was released by slippage. Also, in each case the large rhombs of dolomite are dedolomitized. The calcite resulting from dedolomitization appears to be more susceptible to dissolution, as described by Evamy (1967), thereby giving rise to the inception horizon.

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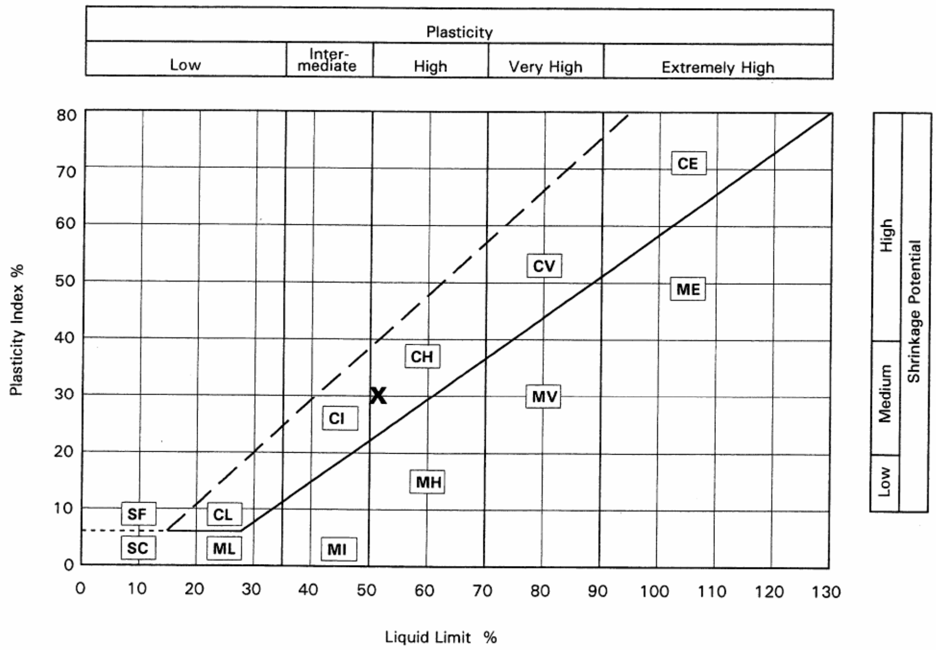
SUMMARY OF PLASTICITY INDEX AND LIQUID LIMIT USING PLASTICITY CHART



METHOD OF PREPARATION BS 1377:PART 1:1990: 7.3, 7.4
BS 1377:PART 2:1990: 3.2, 4.2, 5.2

METHOD OF TEST BS 1377:PART 2:1990: 3.2, 4, 5.3, 5.4

REMARKS Unless otherwise reported one-point cone penetrometer method used.
Plasticity chart taken from BS5930: Clause 41, Figure 31
Shrinkage potential taken from NHBC Standards - Chapter 4.2 Clause 4.2 - D4, Table 1

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Figure 9.2: Atterberg Limit Test Report.

The upper boundary of the inception horizon in the example examined from the Eyam Limestone Formation was a zone of silicification, suspected to be attributable to upward migration of silica solutions expelled from the 'shale' as a consequence of pressure dissolution. In the context of the Eyam Limestone the basinal 'shale' is suspected to have comprised grainstones with a volcanic dust component and it is the dissolution of the volcanic dust that provided the source for the silica. In the Miller's Dale Limestone the clay wayboards were better developed as a consequence of a greater degree of subaerial exposure and primarily comprise volcanic ash. Field evidence, e.g. section 11.3 suggests that the inception horizons are laterally extensive and that they form the focus for conduit development

9.5 Impact of mineralization.

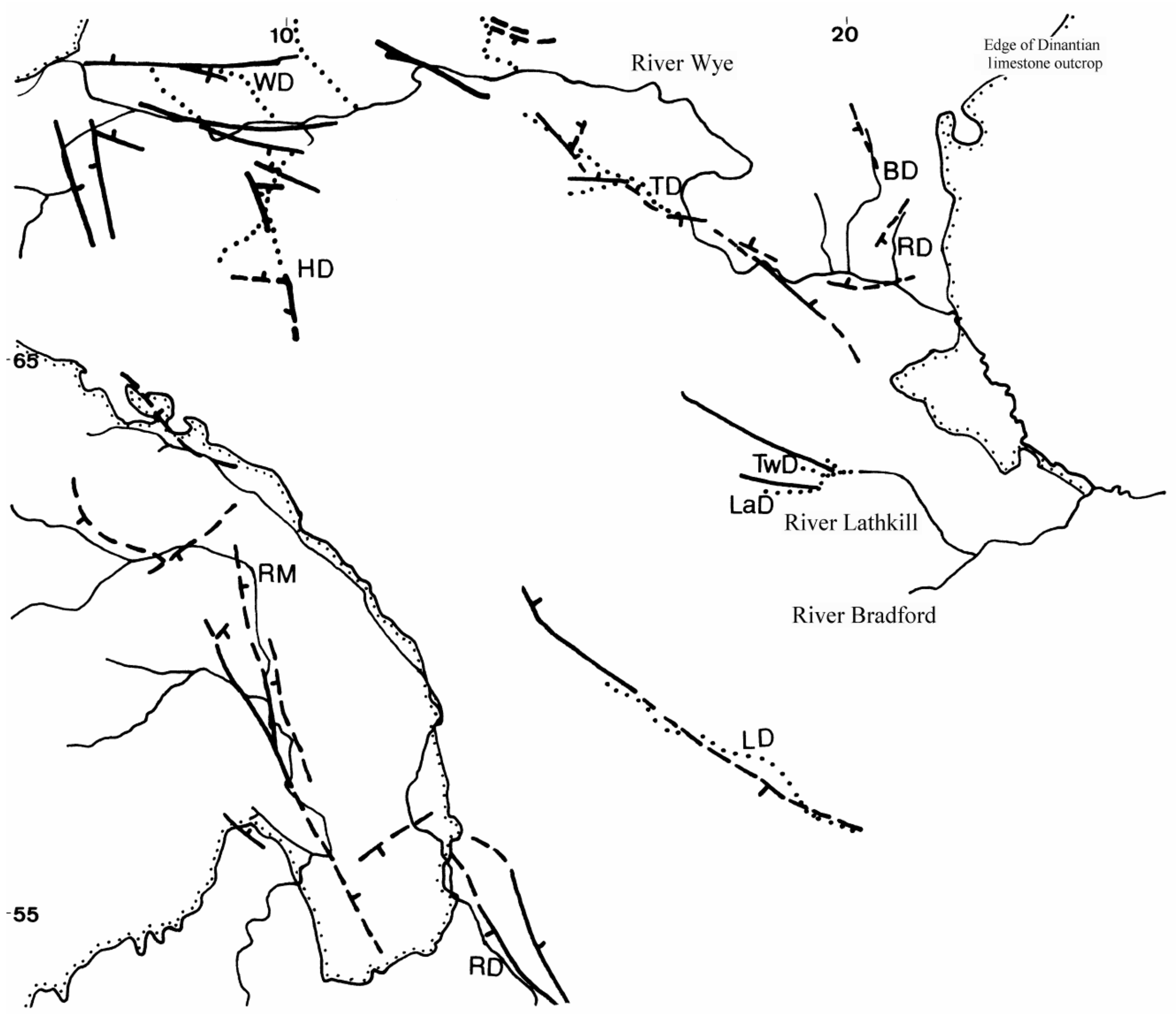
Valley formation is commonly focused on mineral veins, e.g. Mandale Rake and Lathkill Dale. During the mining of Millclose Mine zones of groundwater seepage were investigated for mineral, because a close association of groundwater with zones of mineralization was identified (Bertenshaw, 1981; Traill, 1939). Waltham (1971) observed that mineral veins in the Dinantian limestones of the North Pennines are less soluble. Similarly, Christopher et al. (1977) and Warwick (1962) observed that mineral veins were generally less permeable and although in some cases the presence of a mineral vein had no influence on flow paths, where a vein or scrin exceeded 1 m or so in thickness it was normally less porous and therefore resistant to groundwater flow, with a potential to divert flow and even to trap groundwater between parallel veins. Ford (2000 and 2003) noted the significance of vein cavities as a focus for groundwater flow. It has been noted (Chapter 4) that many dolines are associated with faults and upon further examination it is actually the non-mineralized sections of faults that form the focus for valley formation, rather than the mineral veins. It is also worth noting that Culshaw (personal communication, 2006) opines that many apparent dolines on fault lines that have been subject to lateral displacement are a facet of the faulting process, the voids resulting from the lateral movement and the juxtaposition of protrusions on the faulted surface. This observation would not appear to reflect the situation with the faults of the White Peak, which have not been subject to significant lateral displacement. Furthermore, voids present at the time of mineralization would have been subject to mineralization. Valley development has been guided by the processes associated with the breaching of the Silesian cover (section 9.6). That the vein calcite is of lower solubility is implicit in the number of valleys that end at mineral veins, as indicated in Figure 9.3. The association with valley formation is one of tension gap loading on outward dipping bedding surfaces. Loading on the mineral vein may occur both close to ground surface and at depth as consequence of the predominance of horizontal flow, failure occurring where there is a void to accommodate the slipped mass.

Figure 9.3: To illustrate the association of faults with valley forms.

Key:

…….. Dry Valley
Heavy solid lines faults, with fleck to mark downthrown side

BD Back Dale
HD Horseshoe Dale
LaD Lathkill Dale
LD Long Dale
RD Rowdale
RDo River Dove
RM River Manifold
TD Taddington Dale
TwD Twin Dale
WD Woo Dale



Little attention has been given to the phases of silicification that have occurred, some of which appear to be closely associated with mineralization (Chapter 3). Locally the silicified layers are important as they act as aquitards to groundwater movement. Furthermore, examination of thin sections taken from across an inception horizon in the Miller's Dale Limestone (Chapter 4), established that a silicified layer formed the upper boundary to the inception horizon, suggesting to this author that the source could be silica released from the clay wayboards, as a by-product of the generation of kaolinite from precursor smectite and illite clays under pressure. The formation of kaolinite takes up hydrogen from formational waters, thereby increasing pH, which is associated with the release of silica. The evidence suggests that the silica migrated away from the wayboard, concentrating and precipitating in a zone of lower pH, which in some instances is associated with mineralizing fluids (Appendix 4.1). Interestingly, silicification is more extensive in the Eyam Limestone Formation, which suggests more extensive dissolution of silica from terrestrial sediments washed into the Stanton Basin, or from sediment derived from radiolarian blooms, or that there is an additional source of silica associated with mineralization processes.

Mineralization has exerted an influence on geochemistry, although this is not as marked as might be anticipated (Chapter 6). Worley (1978) identified specific corrosion processes associated with the mineralization, including the oxidation of pyrite and of marcasite, which result in dissolution of the limestone. Thus in some mineral veins speleogenetic processes increase the storage potential of the mineral veins, although it should be noted that this was a relatively localised occurrence. Elevated chloride contents were determined for spring water taken from the Woo Dale Limestone Formation and it has been speculated that this may be attributable to connate mineralizing fluids (Chapter 6). Silica dissolution in the presence of hydrocarbons appears to be essential to understanding the low partial pressures of carbon dioxide that have been determined in the Monsal Dale Limestone (Chapter 6).

9.6 The processes of cover erosion and surface watercourse evolution.

It has been speculated, e.g. (Hollis and Walkden, 1996) that the depth of burial during mineralization was in the order of 2.5 km. Some of this depth is attributable to downwarping (Leeder, 1988), rather than simply to cover thickness. It would seem that the thickness of the Silesian cover overlying the limestones was in the order of 1.5 km (Aitkenhead et al., 2002; Ewbank et al., 1995). Throughout the Silesian Subperiod and the Permian Period the strata would have been subject to exposure to an arid climate in which the area was reduced to something of a rock desert (Cope et al., 1992). A considerable reduction in the thickness of the Silesian strata is likely to have occurred during this time period, continuing into the Triassic Period, albeit there may have been brief periods of sediment accumulation, in particular at the end of the Triassic, during the late Jurassic and late in the Cretaceous periods. However, Walkden and Williams (1991) opined that there is no evidence that the Derbyshire Platform was reburied during the Mesozoic.

Simms (2004) provided empirical evidence to show that the climatic conditions that prevailed in Britain during the Tertiary are such that continuous dissolution and consequential lowering of limestone should have left siliclastic rocks as ridges surrounding the limestone. Walsh et al. (1972), based on palaeontological analysis of the shales in dissolution hollows supporting the Brassington Formation (Pocket Deposits), calculated that 21 m of shale capped the limestone at the time at which the dissolution hollows developed (Neogene). The overlying Brassington Formation was in the order of 43 m thick and the thickness of dissolved, underlying limestone in the order of 100 to 150 m. This evidence supports the concept of active erosion and dissolution that is postulated by Simms (2004), however the exhumed landscape was not modified to the extent that Simms (2004) implies. Examination of long sections passing between the Silesian and Dinantian strata indicate that in the Derbyshire limestones the extensive lowlands that might be anticipated from prolonged exposure are not evident, indeed immediately to the east of the limestone the converse is true. This can be attributed to a number of factors, including: more recent "unroofing" of the limestone; isostatic readjustment to the "unroofing" of the limestone; or the impact of more recent glaciations (section 9.7).

The strong influence of the structure on hydrology does in itself suggest relatively recent unroofing of the limestone. The structural influence described by Al Sabti (1977) who recorded an association of dolines with fault zones, as noted by this author (Chapters 4 and 11). Johnson (1969, pp. 422-423) stated ".. *the drainage system on the Derbyshire limestone plateau appears to have been inherited from a pattern that evolved upon a cover of Namurian shale now removed and, as the limestone was exposed some of these streams have become re-aligned over parts of their courses and are adjusted to joint-set directions in the limestone*." In support of such an analysis visual examination of the geology sheets for the area reveals the close association of faults with river forms, both in areas underlain by Dinantian and Silesian strata (Figure 9.3). Where this is observed on the Silesian strata, sandstone is juxtaposed against shale on the downthrown side of faults. The association of watercourses with the downthrown side of faults is a common association. This response is not so clear in the Chee Tor Limestone Member, which may be a feature of the mapping, because there is an apparent termination of faults once they meet the Chee Tor Limestone and suspected by this author to be attributable to masking by the intensity of fissuring in this unit. Thus it would appear that the form of Deep Dale (west) has been influenced more by the edge of the mineralized zone than by the location of faults. However, at the southern end of the dale, rising into the Miller's Dale Limestone Member of the Bee Low Limestone Formation, Horseshoe Dale would appear to be guided by faulting. That the influence of faulting is evident in areas both underlain by Dinantian and Silesian strata is relevant in that it suggests that it is likely that Silesian strata overlying the Dinantian strata also supported fault-guided drainage and therefore rather than becoming re-aligned to joint sets, the drainage patterns were inherited from above.

One of the reasons for drainage being fault-guided in the Silesian strata is the juxtaposition of sandstone against shale as a consequence of faulting. It would seem likely to this author that as the surface of the Silesian strata was lowered, rivers would become losing streams, particularly where they are associated with fault zones, leaking into the limestone, targeting fissures and driving gestation of inception horizons at depth. Contemporaneous with this, there would be a potential for lateral migration of the surface watercourses as sandstone margins retreated. The strong structural guidance of inputs would impose structural influence on groundwater catchment divisions. It has also been observed that the more mature tributaries of the River Wye e.g. Great Rocks Dale and Woo Dale show a curvilinear form that is guided by faulting. It is considered by this author that, in part, this is attributable to the less resistant nature of the Woo Dale Limestone Formation, which allowed the watercourses to adjust more closely to the hydraulic gradient than did the Bee Low Limestone or, more specifically, the Chee Tor Limestone Member.

With respect to isostatic readjustment, there are several fragments of evidence that suggest that valley bulging has occurred in Lathkill Dale. This is evident in the stretch immediately upstream of Lathkill Head Cave and includes: minor tectonic movement on bedding planes, differential slip on bedding planes and evidence from the dip of the strata, which is generally to the southeast, but locally dips to the north and south, away from the valley sides (Chapter 11). Clearly, this is not of the scale that would influence differential weathering rates between the Silesian and Dinantian strata; however it is an indicator of flexural rigidity (Simms, 2004) and suggests a potential for some isostatic readjustment, which would be most likely to have shown its maximum expansion on faults.

Clayton (1953) identified an extensive Mio-Pliocene peneplain at 305m AOD and argued the case for a high-level composite surface of Neogene age, stating that "*it does seem likely that it is the Pennine counterpart of the summit surface of south-east England, the Mio-Pliocene peneplain*" (p. 35) and describing the lower surfaces (p. 33) suggested "*The complete stairway must have been formed by an interrupted fall of base-level with still-stands, if only of limited duration, every thirty to fifty feet*". Subsequently, Walsh et al. (1972), based on evidence of current sea levels, Lower Pleistocene deposits of southeast England and the sub Neogene surface of the southern Pennines suggested that Upland Britain has undergone regular uplift by about 1m per 15,250 years, since the end of the Alpine Orogeny. This suggests that the rate of uplift exceeds the rate of surface lowering. Further support for the latter hypothesis comes from Walsh et al. (1999), which provides an overview of denudation surfaces and suggests that during the Miocene much of the western half of the British Isles was part of an earlier North European Plain, with little variation in altitude, although detail with respect to how the Peak District fits into this model is not clearly presented. More recently, Tiley et al. (2004) have reported on flexural models that explore the relationship between magmatic underplating associated with the opening of the Atlantic and the Iceland Plume. This work suggests that up to 3 km of denudation occurred in the West Midlands during the Palaeogene, with evidence for subsequent smaller phases of denudation during the Oligocene-Miocene and the Pliocene. The latter studies are ongoing at Leicester and Cambridge universities, but they would seem to suggest a sequence that falls between the opposing view points of Clayton (1953) and Walsh et al. (1972).

Although some vadose downcutting of caves has been described (e.g. Beck, 1980) downcutting is not a distinguishing feature of the karst of the White Peak. In the context of the amount of uplift hypothesised this can be seen as further evidence in support of the inception horizon hypothesis, with a number of inception horizons being active concurrently. Alternatively, the poor development of vadose incision might suggest more recent cave incision; arguments against this were presented in section 4.3.

The influence of the Pleistocene forms the subject of section 9.7, but some points merit discussion here. It is considered that Devensian ice did not reach the White Peak (Burek, 1978; Clarke et al., 2004), its limit being in the area of Chapel-en-le-Frith (Aitkenhead et al., 2002). It is argued that this is why evidence of the extensive fluvial processes associated with the last deglaciation, cannot be found. Instead processes associated with periglacial conditions predominated, when weathering of the limestone would have been slower and the sandstones and mudstones would have been particularly susceptible to the effects of freeze-thaw and mass movement. Relict evidence of this type of weathering is evident, e.g. Peters Stone, Cressbrook Dale (SK 17387525), but it is likely to have occurred on a far greater scale in the shales and sandstones, thereby reducing the topographical differential between the strata and contributing to the formation of the shale vales (Johnson, 1967). Furthermore, the widespread cover of loess, which is attributed to the Devensian glaciation, was not subject to an invasion of glacial melt-water and has choked much of the surface karst (Beck, 1980). However, consideration of the maximum limit of the Devensian ice (Aitkenhead et al., 2002 and Figure 9.4) suggests that the area of Dove Holes is likely to have been inundated by glacial meltwater, probably draining via Great Rocks Dale and Monks Dale and this author has provided supporting evidence of glacial meltwater processes in valley thalweg plots (Figures 9.5 to 9.10).

The process of surface watercourse evolution is closely associated with cover erosion. Linton (1956) examined pre-Pleistocene drainage at a level of approximately 490 m OD and identified a strong west-east trend to the drainage of the area, with the catchment of the rivers Lathkill and Bradford draining north to the River Wye. With the possible exception of the easterly directed drainage between Tunstead Quarry and Wormhill Springs (section 7.3.4) this author has not found any evidence to support these observations, although it should be noted that the level of drainage that was being considered was in the order of 150 m above the current ground levels.

It would appear that superimposed on the regional hydrogeological setting, the structural and lithological guidance of the limestone resulted in the evolution of southeasterly directed watercourses on this surface, initially as a number of individual water courses and then, as the cover was breeched more extensively, river captures occurred and some rivers sank underground. In part this may be attributable to the fault patterns described above. Subsequent glaciations have reinforced the southeasterly vector to drainage. Johnson (1957) in an examination of drainage patterns showed how

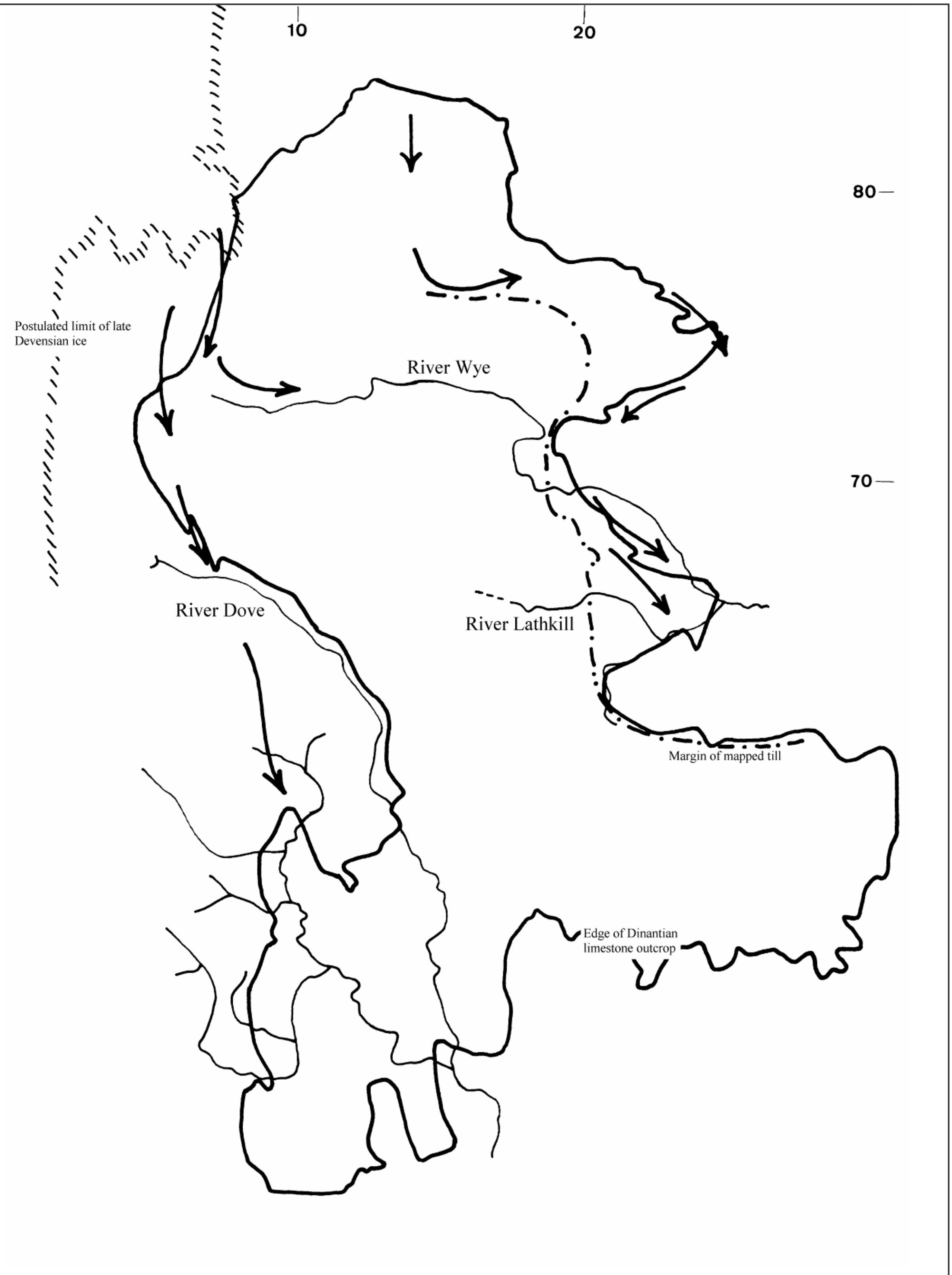


Figure 9.4: Postulated movement of Pennine Ice during Anglian I stage.
Postulated limit of late Devensian ice from Aitkenhead et al., 2002. Arrows indicate postulated direction of ice movement, evidence referenced in Table 9.3.

the remnant outcrop of the Silesian shales has influenced drainage patterns, in particular enabling the River Bradford to capture part of the Griff Grange river system. Beck (1980) related the tiers of cave complexes in the area of Stoney Middleton to the receding shale margin. In the context of inception horizon development this form of approach needs to be interpreted with care, because it is clear that a number of inception horizons are in operation at any given time and it would appear that base level lowering is reflected by the abandonment, rather than activation, of sequentially lower inception horizons.

Before going on to consider the process of cover erosion in detail it is first necessary to consider the form of the pre-Silesian surface. For instance, Simms (2004) presented a conceptual model for inverted topography, whereby unroofing of anticlinal structures occurs first, thereby facilitating early dissolutional processes in the limestone and causing anticlinal features to be left as low ground and synclinal forms as high ground. It would appear that in the Peak District the pre-Silesian surface has been adjusted by faulting to a greater degree than folding. It follows that outcrops of the Woo Dale Limestone provide an indication of where uplifted surfaces are most likely to have been encountered, namely in Woo Dale, Long Dale and Dam Dale. As the areas first exposed by unroofing and then dissolutional deepening these areas would inevitably have formed the focus for surface water. Figures 9.11 to 9.13, which have been prepared to show the conjectural surfaces, illustrate this. The amount of limestone removal implicit in these figures falls within the range calculated by Walsh et al. (1972) and the ranges quoted in Chapter 4.

Kirkham (1952, p. 8) observed, "*In very many mines the lead miner has broken into caverns, and judging by the numbers which are reported, I think that probably there are a number of large caverns at least 2-300 ft (60 to 90 m) below the surface*". Accepting that many of these caverns are likely to relate to paleokarst associated with mineralization, but bearing in mind the concept of tiered flow paths (Worthington, 1991), this author has examined geological maps to see whether or not the geological and speleological evidence supports the concepts presented above and the following observations have been noted in support of the river evolution postulated in Figures 9.14 and 9.15:-

i) The existence of an Upper and Lower Wye (Johnson, 1969) would help to account for the absence of tributary valleys to the River Wye on its south side, in the stretch between Churn Hole (SK 10507200) and Monsal Head (SK 18207160). The evidence suggests that the Lower Wye may have captured the Upper Wye by headward erosion of tributaries in the vicinity of Chee Tor (Figure 9.15), thereby causing the channel from Chee Tor, along the line of Watersaw Rake, to be abandoned. Similarly, the Upper Wye would have captured the headwaters of the River Lathkill by southerly migration and headward erosion in the Woo Dale Limestone.

ii) The evidence presented above suggests that fissure-guided incision would be anticipated in the more resistant Chee Tor Limestone Member, as observed at Chee Tor, where the form of the incised meander could be associated with fault-guided river capture processes in the vicinity of Wormhill Springs.

Incision is likely to be associated with base-level lowering. This, together with the occurrence of the deep valley at Taddington and the exposure of the darker facies of the Monsal Dale Limestone juxtaposed against the Eyam Limestone and Longstone Mudstone formations, points to a potential location for the Upper Wye. The river alignment indicated on Figure 9.15 would account for the position of the caves that have been explored in the Miller's Dale Limestone in Blackwell Dale. The geological evidence would suggest that Deep Dale (east) was an important tributary to the Lower Wye.

iii) The tightness of the meander in Ashwood Dale (at the location of Ashwood Dale Resurgence) and the occurrence of a minor outcrop of Bee Low Limestone, suggest the possibility of the incised location of river capture. Furthermore, dye recovery times from Illy Willy Water to Ashwood Dale Resurgence and the occurrence of caves in Deep Dale suggest to this author that the River Lathkill may once have had a significantly longer route, with its headwaters close to Ashwood Dale Resurgence. Evidence for this route has been lost in the heavily jointed Chee Tor Limestone Member. Data presented below indicate that the formation of Deep Dale (West) must post-date this route.

iv) It would seem plausible that Long Dale, which has been captured by the River Bradford (Johnson, 1957) extended in a northwesterly direction and formed the headwaters of the meandering river system associated with the deposition of the Brassington Formation (Walsh et al., 1972). Based on the evidence of southeasterly flow established by dye-tracing (section 7.3.1), it is even possible that Poole's Cavern was once associated with the headwaters to this system. This would also help to explain the greater concentration of sediment encountered in the Bee Low Limestone Formation at Hindlow Quarry than Tunstead Quarry and is in keeping with the observation from examination of aerial photographs of a greater number of dissolutional features along the valley sides of Long Dale. This evidence suggests that the rivers that formed on the Silesian strata predominantly flowed southeastwards.

v) Rapid downcutting of the River Wye, associated with deglaciation, resulted in localised reversal of hydraulic gradient. This is particularly evident in the area forming the headwaters of the river Wye, where the local hydraulic gradient is now to the north and also resulted in nick point migration up the valleys, locally impeded where the Lower Miller's Dale Lava caps the Bee Low Limestone (see Figure 9.5).

vi) The former route of the Derwent around Bakewell as postulated by Straw (1968) and accepted by Johnson (1969) corresponds very closely with the position of the Ashford Basin, hence the remnant outcrop of Eyam Limestone Formation and Widmerpool Formation, indicative of recent removal of the Silesian shales from this area and also a thicker sequence of the dark facies of the Monsal Dale Limestone.

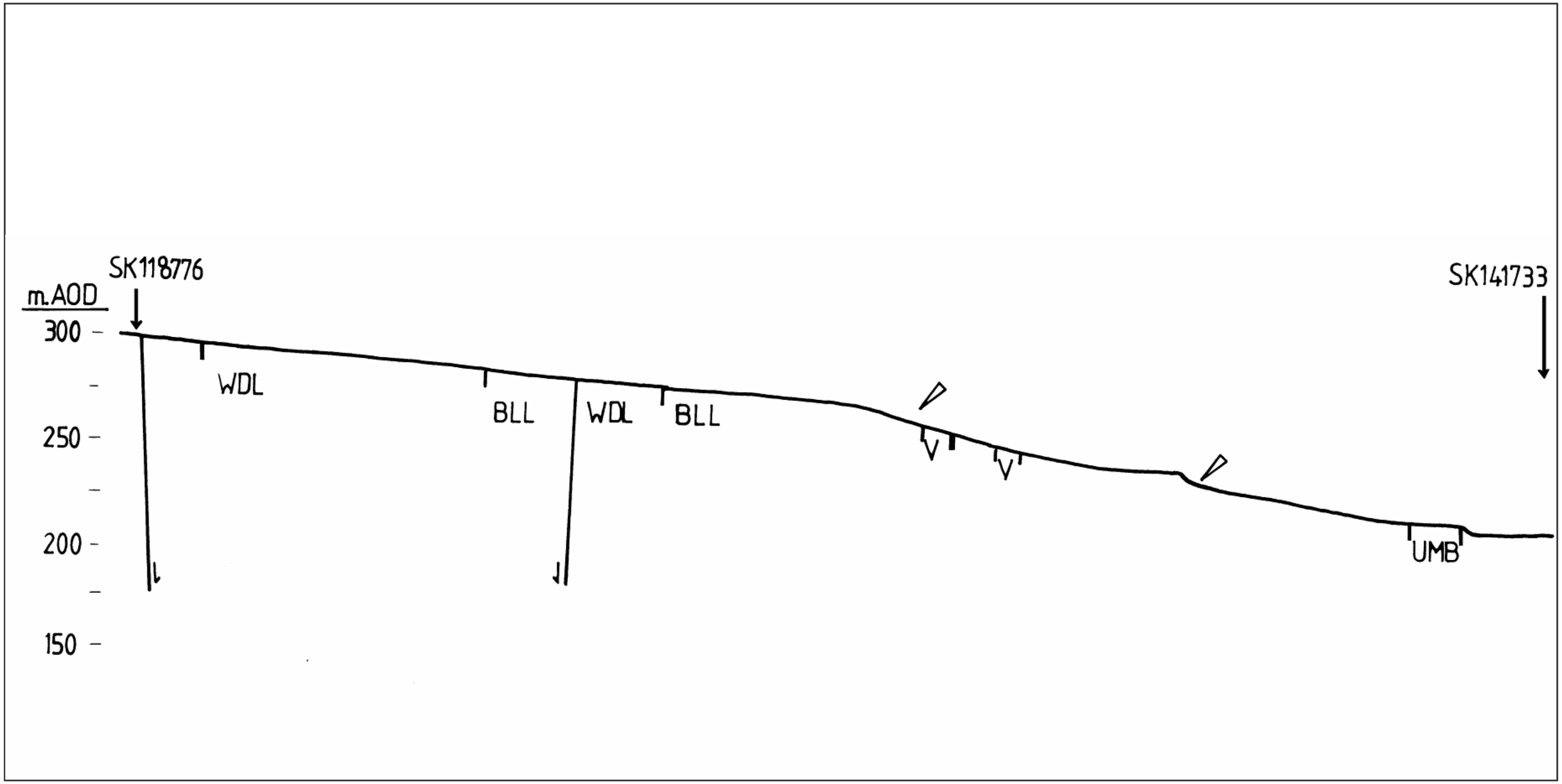


Figure 9.5: Thalweg Plot : Monks Dale.
Key: Open arrow head : Nick point that cannot be directly correlated with recognized changes in lithology, or with structure. WDL Woo Dale Limestone; BLL Bee Low Limestone; UMB Upper Miller's Dale Lava; V volcanic intrusion. Faults shown with mark to indicate down thrown side.
Scale: Vertical exaggeration x 10.
Observations: Convexo-concave form is indicative of mature valley. Nick point associated with the Upper Miller's Dale Lava suggests that it is resistant to dissolution. Upper nick point indicates that Monks Dale could have captured the Upper Wye at a minimum level of 250 m AOD.

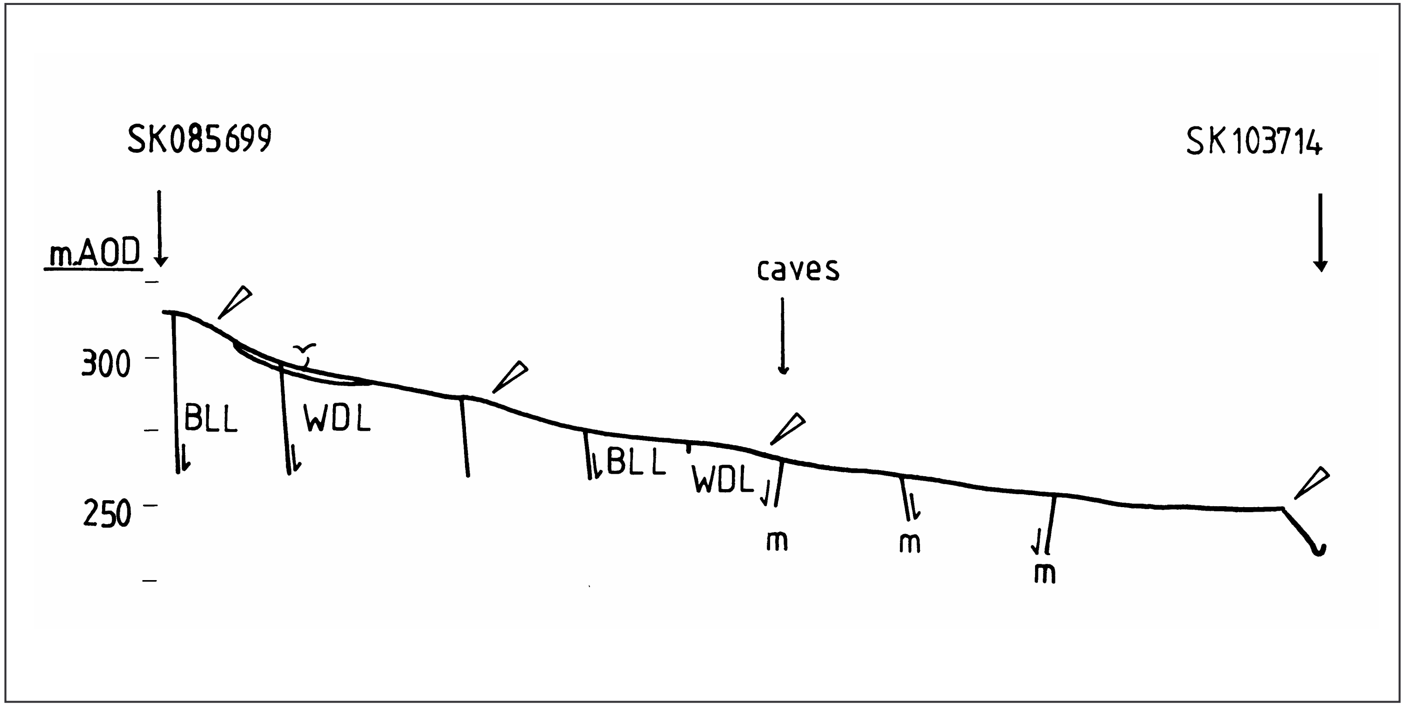


Figure 9.6: Thalweg Plot : Deep Dale (West).
Key: Open arrow head: Nick point that cannot be directly correlated with recognized changes in lithology, or with structure. ᵛ Alluvium; WDL Woo Dale Limestone; BLL Bee Low Limestone; m mineralized fault. Faults shown with mark to indicate down thrown side.
Scale: Vertical exaggeration x 10.
Observations: Convex form is indicative of a more youthful valley than Monks Dale. Note the presence of the nick point immediately adjacent to the River Wye on the south side of the river, the absence of which to the north would appear to support the hypothesis of glacial meltwater erosion of valleys to the north of the River Wye. The second highest nick point coincides with the junction of Horseshoe Dale and and Back Dale, which appears to confirm the hypothesis of cutting back to capture a water course at this location.

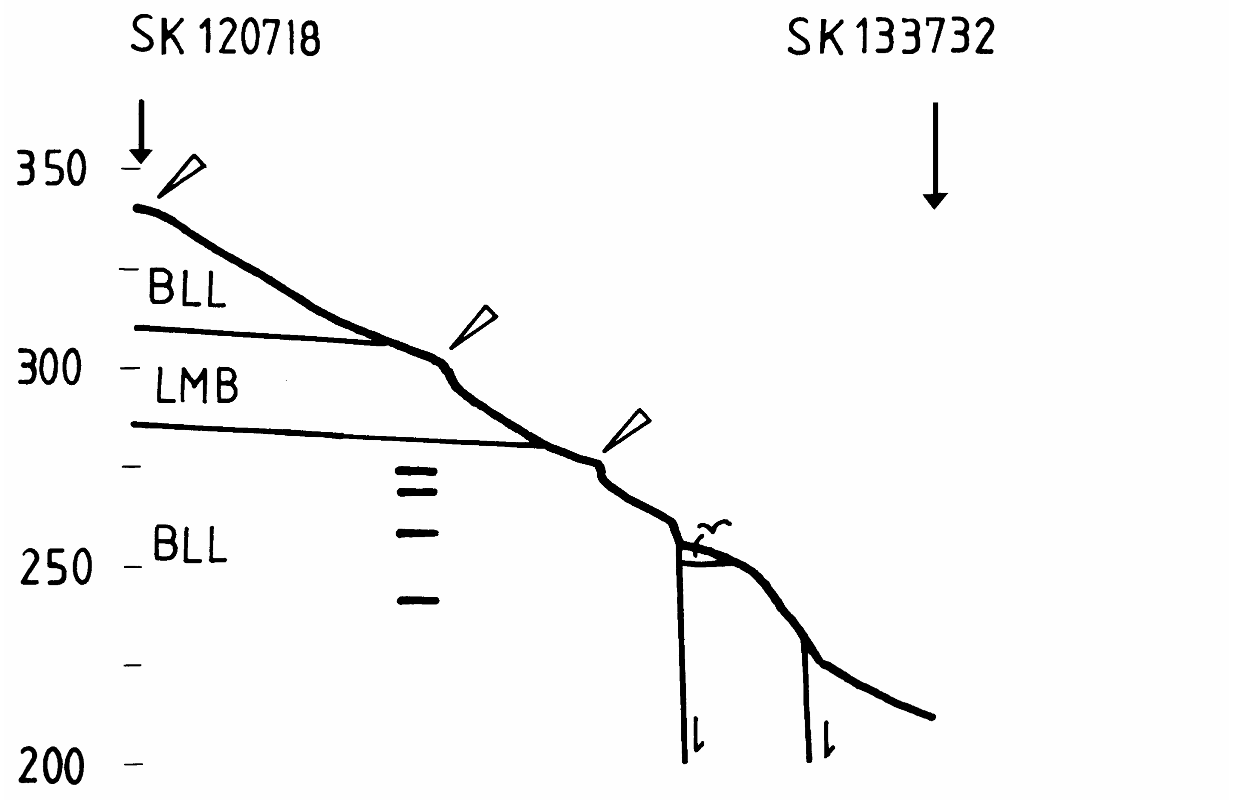


Figure 9.7: Thalweg Plot : Blackwell Dale.
Key: Open arrow head: Nick point that cannot be directly correlated with recognized changes in lithology, or with structure. ᵞ Alluivium; BLL Bee Low Limestone; LMB Lower Miller's Dale Basalt. Faults shown with mark to indicate down thrown side. Horizontal lines correspond with outlet levels of Beck (1980).
Scale: Vertical exaggeration x 10.
Observations: Convex form is indicative of a more youthful valley than Monks Dale. It is clear the faulting has influeneced the development of the thalweg. The steep gradient and youthful form indicate that the dale is a more recent feature, cut close to the river. Beck's outlet levels would appear to correspond with the nick point at 275 m AOD. The evidence suggests that flow paths extend to a minimum of 35m below the local base level.

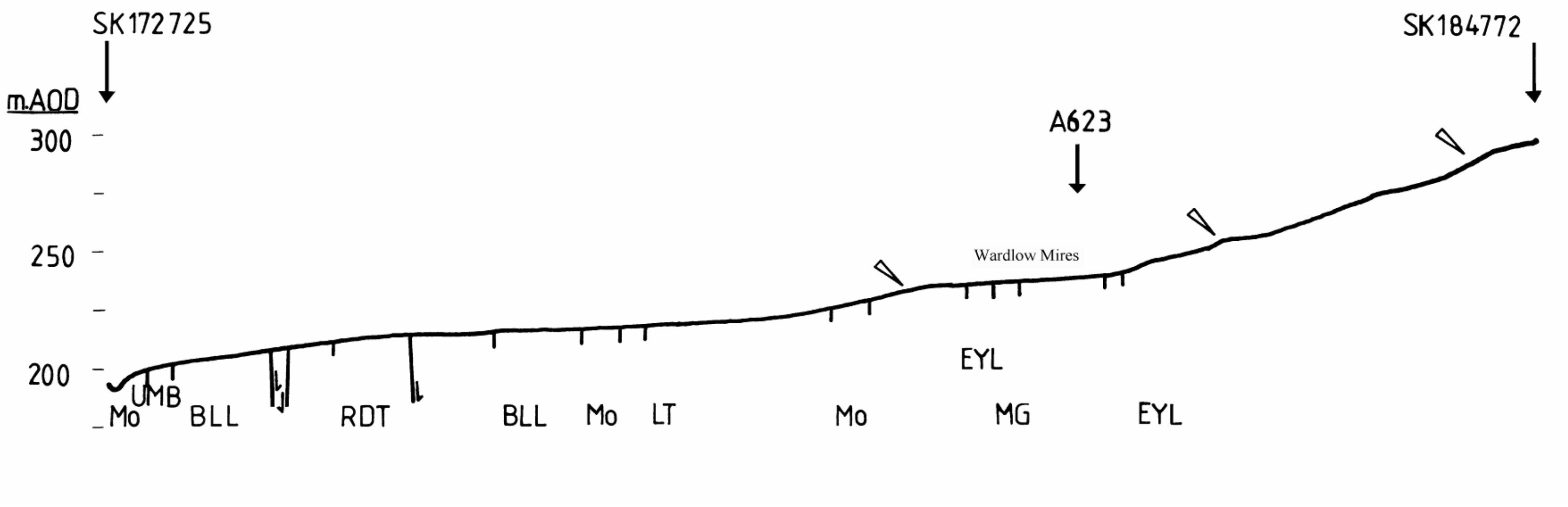


Figure 9.8: Thalweg Plot : Cressbrook Dale.
Key: Open arrow head: Nick point that cannot be directly correlated with recognized changes in lithology, or with structure. BLL Bee Low Limestone; Mo Monsal Dale Limestone; EYL Eyam Limetone; UMB Upper Miller's Dale Lava; RDT Ravensdale Tuff; LT Litton Tuff; MG Millstone Grit Group. Faults shown with mark to indicate down thrown side.
Scale: Vertical exaggeration x 10.
Observations: Concavo-convex form is indicative of a mature lower valley (suspected glacial spillway). Nick point associated with the Upper Miller's Dale Lava Formation suggests that it is resistant to dissolution. Upper nick point indicates that Monks Dale could have captured the Upper Wye at a minimum level of 250 m AOD.

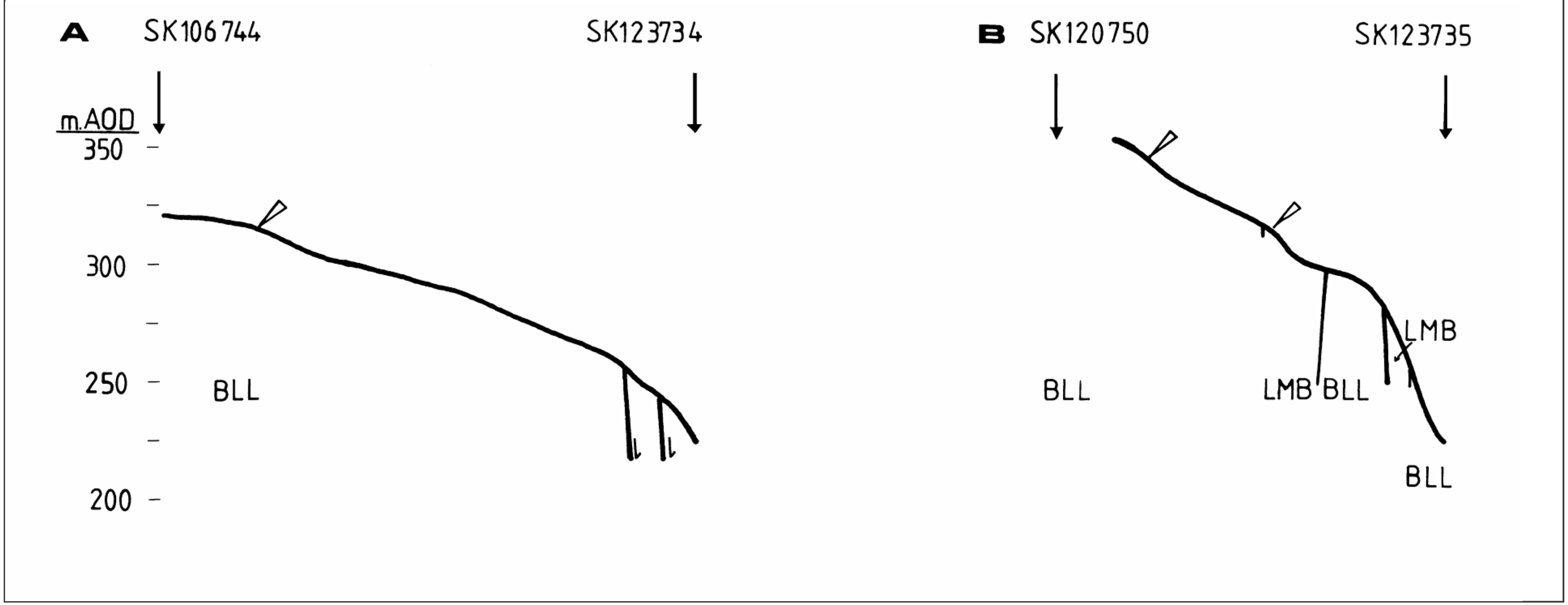


Figure 9.9: Thalweg Plots : A: Flag Dale and B: Wormhill Dale.
Key: Open arrow head: Nick point that cannot be directly correlated with recognized changes in lithology, or with structure. BLL Bee Low Limestone; LMB Lower Miller's Dale Lava.
Scale: Vertical. Exaggeration x 10.
Observations: Convex form of Flag Dale (A) indicates mature valley. Upper concave surface associated with Wormhill Dale (B) indicates more recent formation. However, the absence of a nick point at a lower level suggests that the dale acted as a meltwater spillway during the Devensian glaciation.

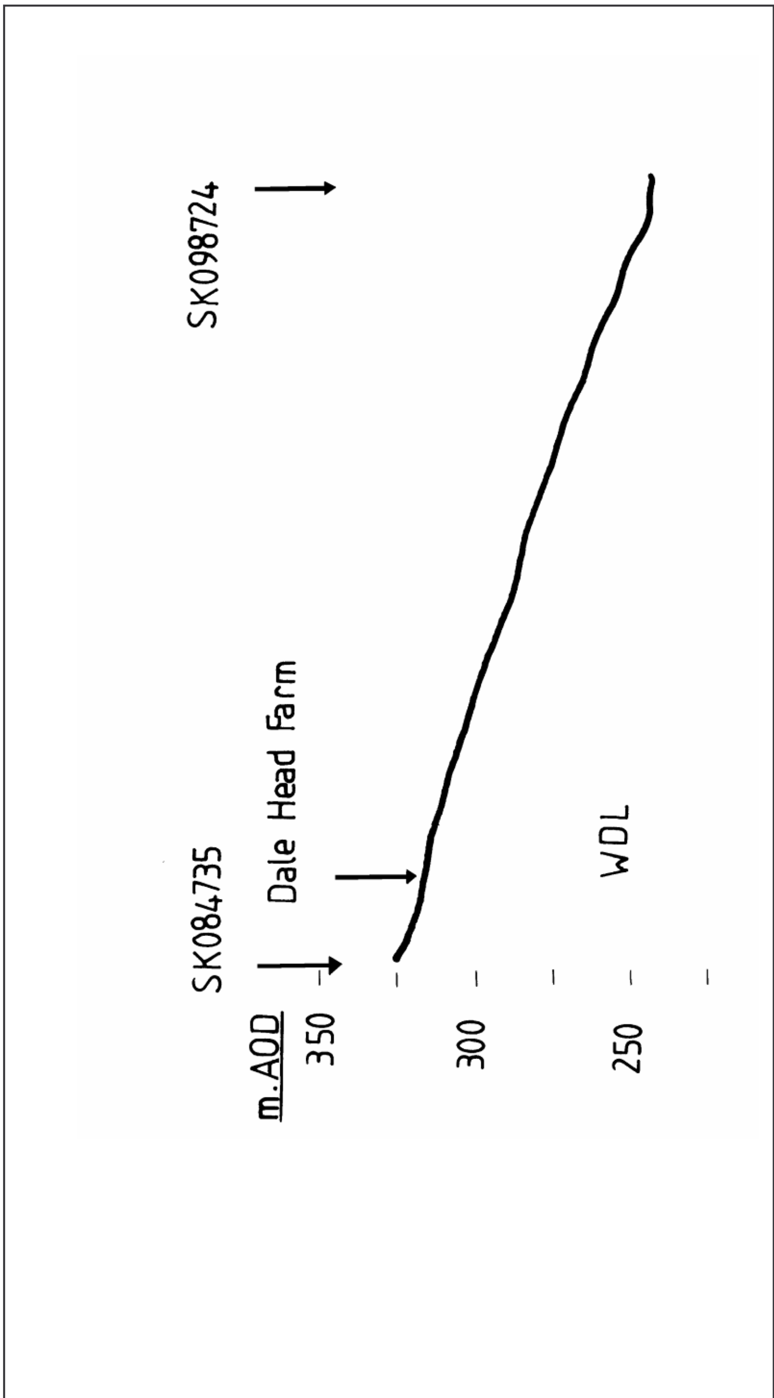


Figure 9.10: Thalweg Plot: Woo Dale.
Key WDL Woo Dale Limestone
Scale: Vertical exaggeration x 10.
Observations: Absence of nick points suggests either a single phase of development, or an erosive stage affecting the entire thalweg. The relatively shallow gradient suggests the latter, thus indicating that Woo Dale was also an outlet for glacial meltwater.

vii) It has been observed by this author that many of the pipe deposits are associated with tensional faults. Worley (1978) observed that many pipe veins have been shown to be drainage routes through which glacial meltwater has flowed and subsequent incision of the valleys has left many of the pipes in a vadose condition.

viii) The occurrence of glacial till over much of the area of the Ashford Basin provides evidence for the glacial diversion postulated by Straw (1968). Evidence of Devensian ice to the north of Dove Holes, suggests that subsequently Great Rocks Dale and Monks Dale were reactivated by glacial meltwater. Further to this, it is the opinion of this author that the relatively straight, non-structurally influenced section of lower Cressbrook Dale suggests that it may have deepened as a glacial spillway for Anglian meltwater released from ice flowing below Longstone Ridge (Figure 9.4).

Burek (1978) has suggested that these were valley glaciers and that much of the remaining area was not glaciated. However it would have been subject to considerable permafrost, hence the occurrence of the periglacial features (Chapter 2, Burek, 1978) and the broadening of shale vales (Johnson, 1967). The permafrost is also likely to have had a significant impact on flow path development. There are a number of reasons why deep phreatic flow path gestation could have occurred in response to the occurrence of permafrost. These include: the confining effect of the permafrost, the lower freezing point of mineralized water and the increased viscosity of colder water. Furthermore, the thermal conductivity of the Carboniferous Limestone has been shown to be 8.3 mcal/cm s k, whereas that of Coal Measures mudstone is 3.57 mcal/cm s k (Richardson and Oxburgh, 1978). These figures suggest that permafrost in the Carboniferous limestone would penetrate more than twice the depth of that in the Coal Measures mudstone. However, the depth of penetration of permafrost is influenced by other factors, e.g. the permafrost table is depressed by the presence of significant bodies of water and local features, such as thermal springs provide enough heat to form a zone of unfrozen ground through the permafrost (Williams, 1970), but equally the effect of the thermal springs is to conduct geothermal energy out of the region, providing a form of insulation.

Buried channel forms have been identified in Lathkill Dale (Pedley, 1993) and in the River Wye. It would appear that the current surface and near-surface geomorphological setting largely dates at least to the Anglian and therefore the majority of active local flow paths are likely to reflect the southeasterly topography that was further imposed at this time. Ford et al. (1975) have argued that it was only following the Anglian glaciation that underground routes became consistent with a hydraulic gradient to the north (towards Hope valley), although evidence from dye-tracing experiments has led this author to suggest that earlier, inception horizon-related flow is consistent with a northerly hydraulic gradient. Prior to the Anglian glaciations, the erosion surface described above and formerly attributed to the Mio-Pliocene by Clayton (1953) has been interpreted as a level, but sloping surface and the hydraulic gradient was also to the east and the southeast (Clayton, 1953). The local flow paths (Tóth, 1963) that were formed during the Pleistocene can largely be attributed to the significant volumes of meltwater. Periods of glacier wasting will have increased the rate of conduit growth by increasing volumes of water available for dissolution associated with unloading of the limestone and interbedded clay wayboards (Chapter 4).

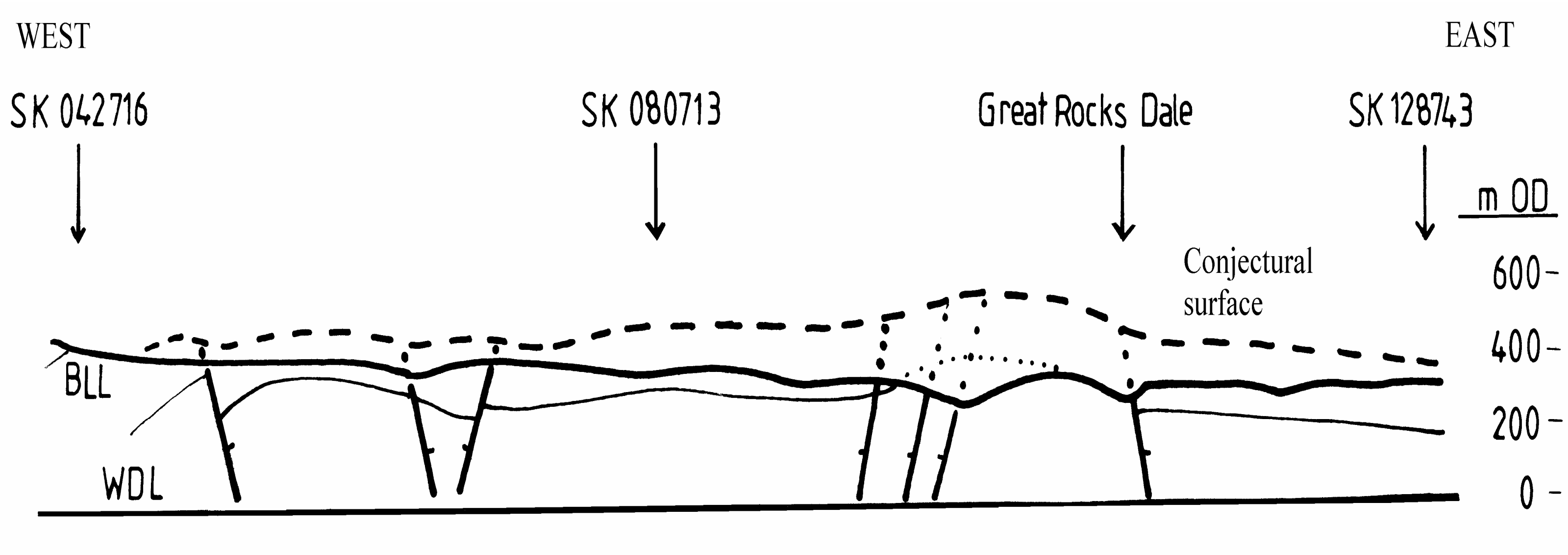


Figure 9.11: Wye valley conjectural surface.
Conjectural Pre-Silesian surface, reflecting an inverted topography (Simms, 2004) derived from an hypothesised dissolutional loss of 225 m in valley locations (former topographic highs) and 50m in plateau areas.
Key: Heavy solid line existing ground surface dashed line conjectural surface, dotted lines hypothesised geology in former cover. BLL Bee Low Limestone; WDL Woo Dale Limestone.
Scale : Vertical exaggeration x 2.

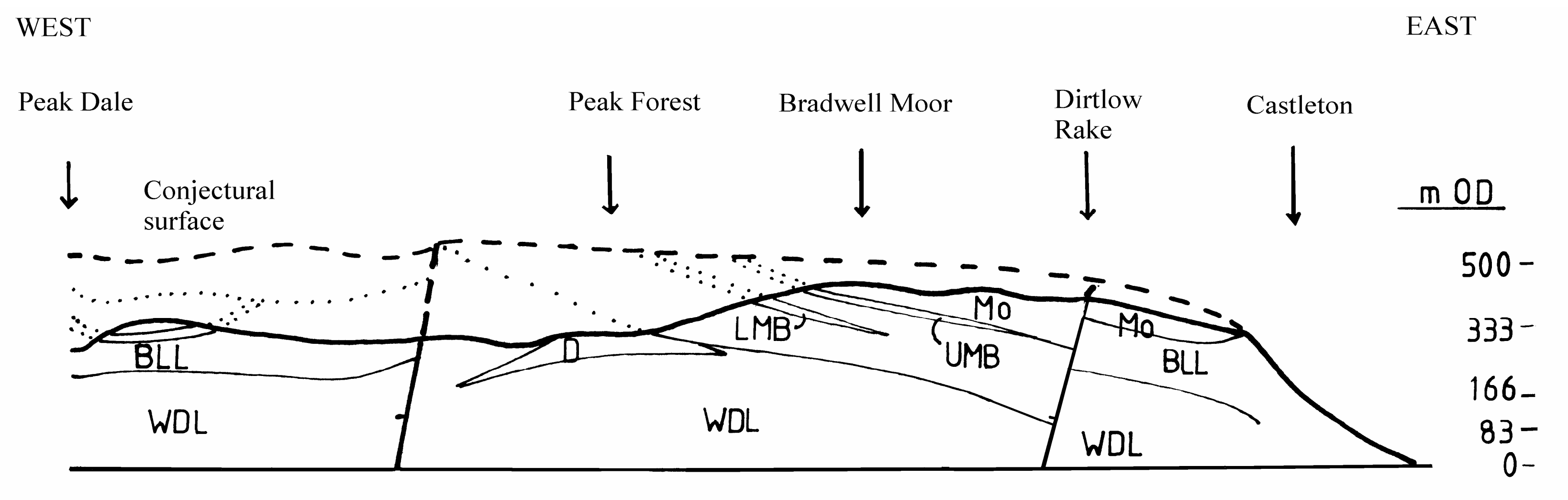


Figure 9.12: Dam Dale valley conjectural surface.
Conjectural Pre-Silesian surface, reflecting an inverted topography (Simms, 2004) derived from an hypothesised dissolutional loss of 200 m in valley locations (former topographic highs) and 50m in plateau areas. It is hypothesised that valley formation was guided by the Peak Dale Fault which juxtaposed Silesian sandstone to the southwest against shale to the northeast. Subsequently the river migrated in an easterly direction due to recharge by groundwater perched by the dolerite sill.
Key: Heavy solid line existing ground surface; dashed line conjectural surface, dotted lines hypothesised geology in former cover. BLL Bee Low Limestone; WDL Woo Dale Limestone; Mo Monsal Dale Limestone;; LMB Lower Miller's Dale Lava; Upper Miller's Dale Lava; and D dolerite.
Vertical exaggeration x 3.

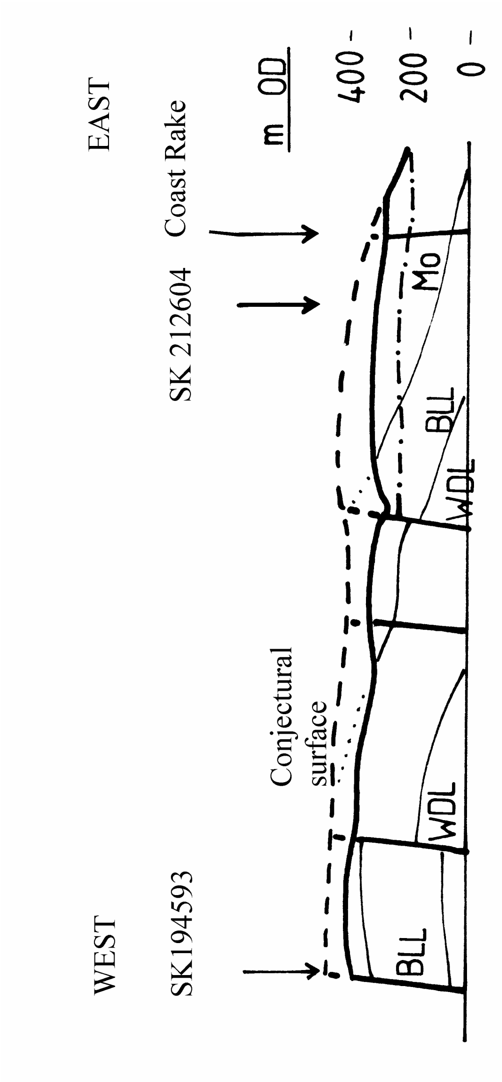


Figure 9.13: Long Dale valley conjectural surface.
Conjectural Pre-Silesian surface, reflecting an inverted topography (Simms, 2004) derived from an hypothesised dissolutional loss of 170 m in valley locations (former topographic highs) and 75m in plateau areas. It is hypothesised that valley formation was guided by the Cronkston-Bonsall Fault, which juxtaposed Silesian sandstone to the southwest against shale to the northeast.. Dolomitized limestone forms the focus for dissolution.
Key: Solid line existing ground surface; dashed line conjectural surface, dotted lines hypothesised geology in former cover; alternate dash and dot represents lower boundary of dolomitization associated with the Monsal Dale Beds adjacent to the Crronkston-Bonsall Falut. BLL Bee Low Limestone; WDL Woo Dale Limestone.
Vertical exaggeration x 2.

Burek (1978) summarised correlations between the river terraces that have been identified in the rivers Derwent and Wye; and Beck (1980) attempted to relate cave levels to the terraces with varying degrees of success. It is the opinion of this author that the poor correlation of the cave levels with the river terraces can largely be attributed to the influence of speleogenetic inception, with a number of inception horizons being active at a given time. Furthermore, as noted by C O Hunt (personal communication, 2007), given the degree of uplift that has been postulated it is very unlikely that valley form would correlate neatly with stages of cave development. The convex form of the thalweg plots of valleys considered to have carried Devensian meltwater supports this observation. For each catchment (Stoney Middleton, Bradwell, Castleton, Lathkill Dale, Blackwell Dale and Buxton) Beck (1980) established up to five development stages with estimated outlet levels for each stage. Each developmental stage was ascribed to a geological stage, thereby providing a means of assessing the fall in base level between the Mid-Pleistocene and the Present. On this basis the Bradwell, Castleton and Lathkill catchment areas

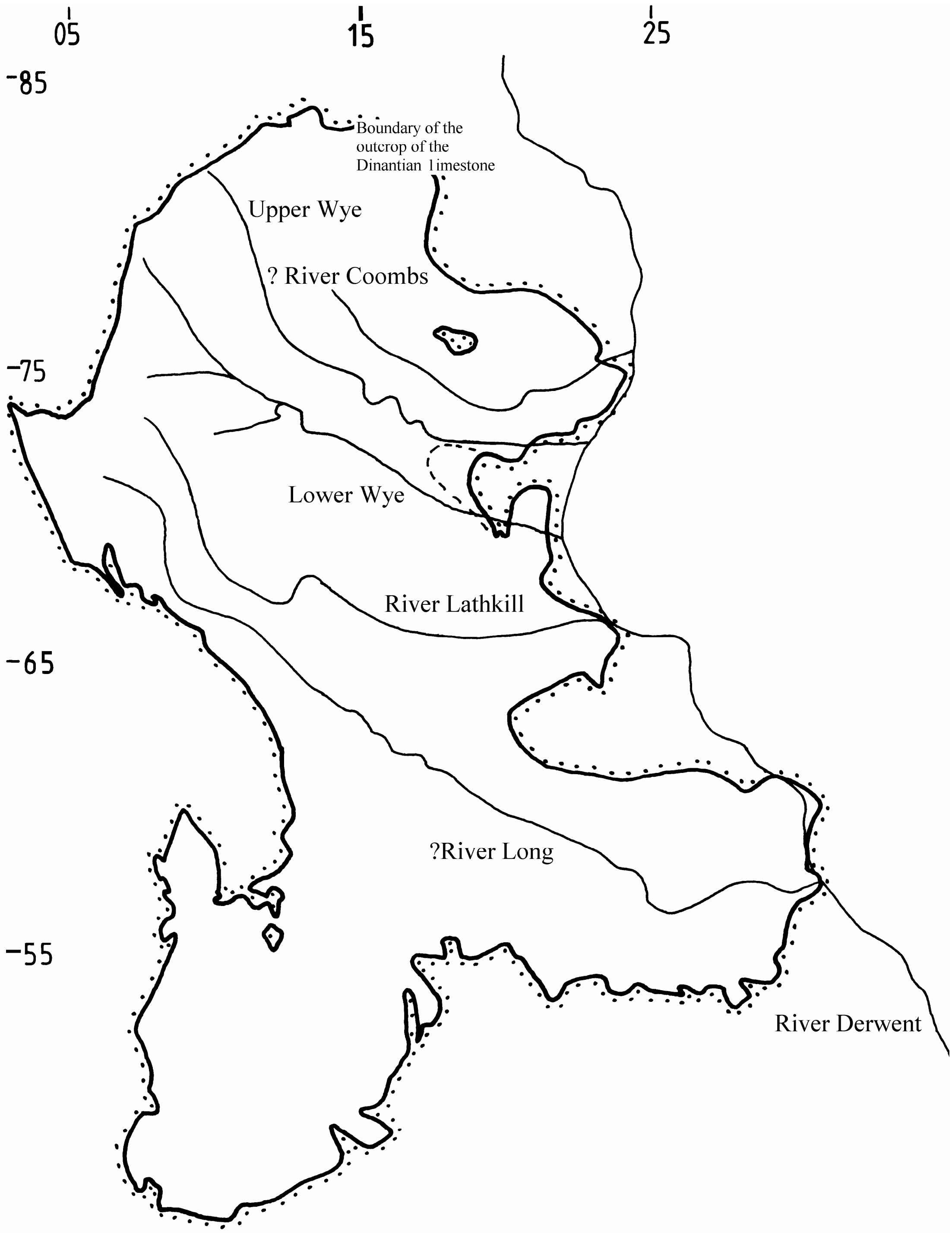


Figure 9.14: Postulated position of Upper and Lower Wye.

Key: Dashed line represents possible former extent of the Namurian shales between the Upper and Lower Wye and guiding the connection at Fin Cop (Johnson, 1957).

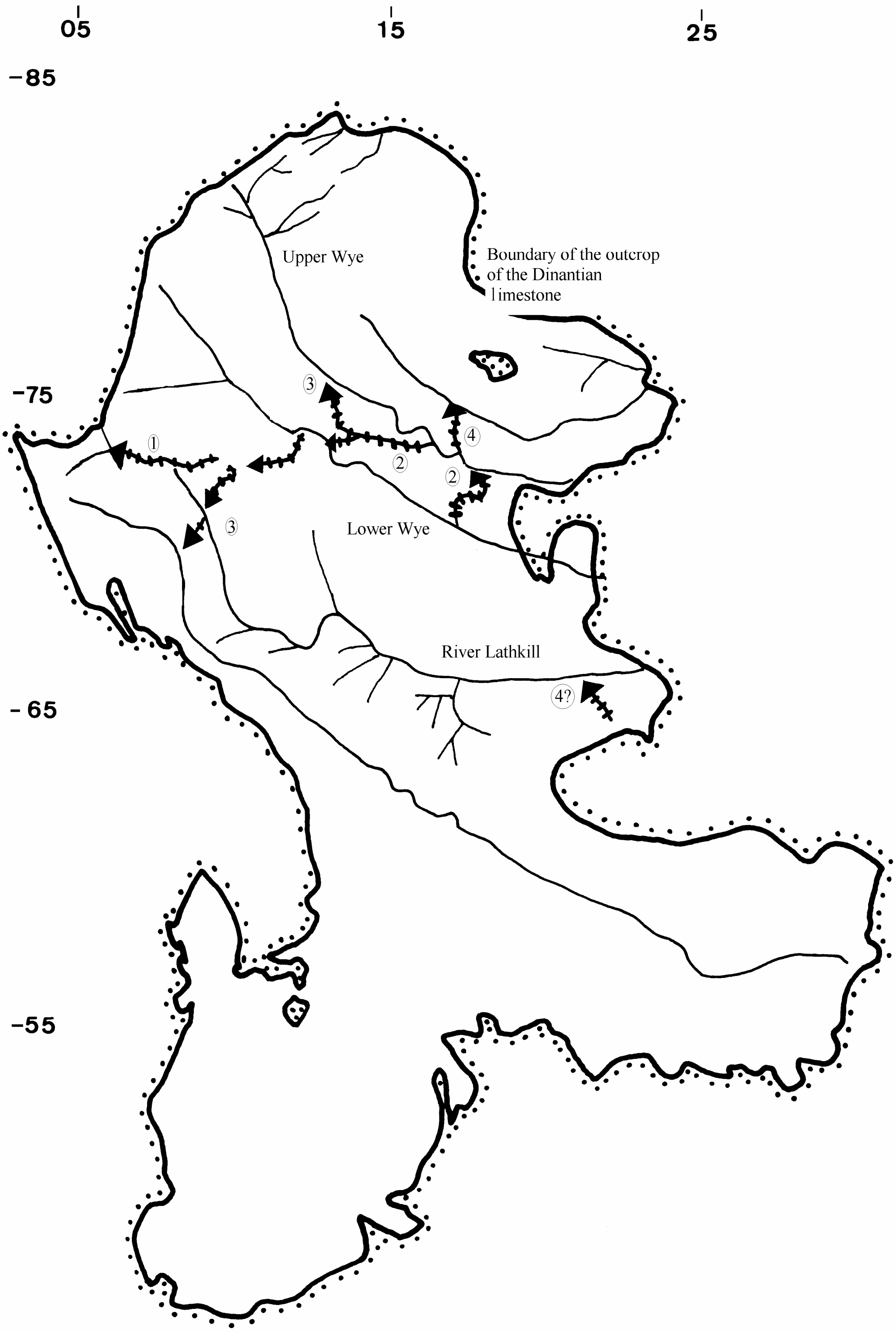


Figure 9.15: Postulated river evolution.

Key: Hatched stretches of river, with arrows indicate postulated headward erosion and river capture. Phases of headward erosion likely to be a response to periods of lowered base level. Headward erosion out of the current Wye Valley is thought to have post dated the connection of the Upper and Lower Wye. See Table 9.4 for chronology of valley evolution.

were subject to a fall in base level of 90 m, 117 m and 85 m respectively, yet in Blackwell Dale the fall in base level was a mere 33 m. Inception in Castleton is likely to be in the Woo Dale Limestone and that in Bradwell and Lathkill Dale in the Monsal Dale Limestone. Inception processes have been identified in both of these formations and therefore conceptually it is likely that lower inception horizons become increasingly active with falling base levels. Furthermore, it is likely that the development of the conduits at each stage occurs rapidly in response to the increased volumes of cold, carbon dioxide rich water associated with ice melting. This is more difficult in Blackwell Dale, where the caves all fall within the Miller's Dale Limestone, for which inception processes have been identified, but which are underlain by the Lower Miller's Dale Lava, in a synclinal setting.

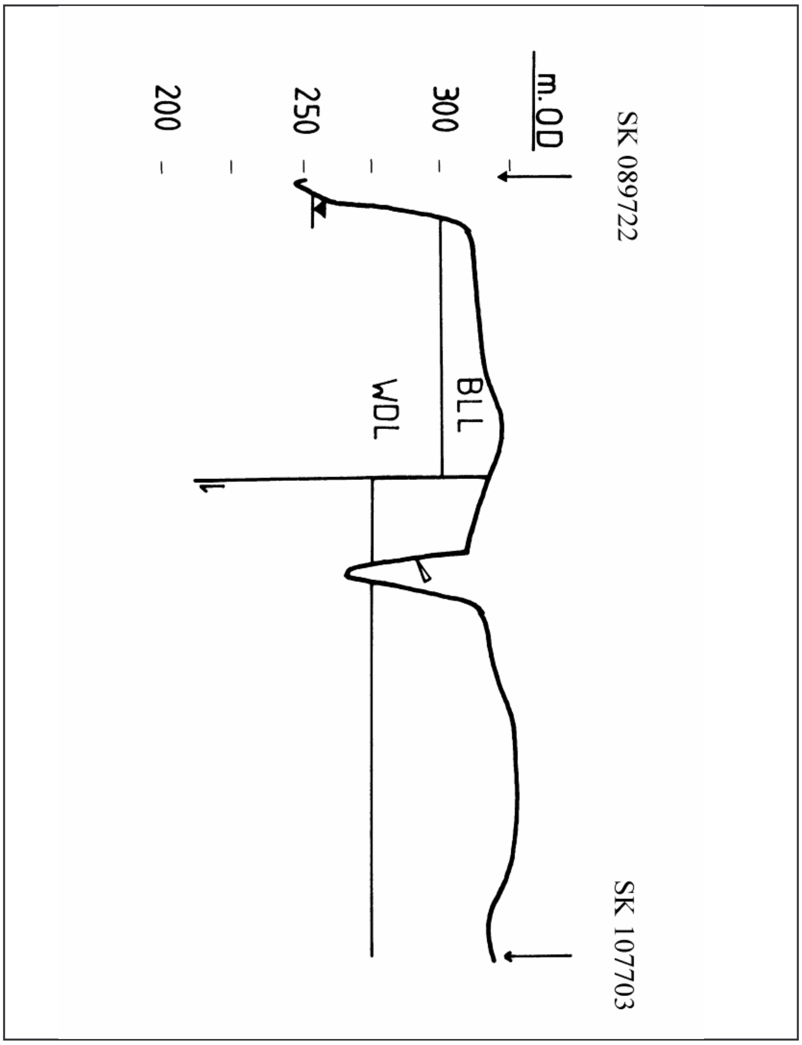


Figure 9.16: Cross section Deep Dale (West) to Ashford Dale Spring.
Scale: Vertical. Exaggeration x 10.
Key: Closed arrow head: Location of Ashford Spring; Open arrow head: Location of Deep Dale Cave; BLL Bee Low Limestone; WDL Woo Dale Limestone.
Observations: Deep Dale Cave suspected doline-fed and at boundary between the Miller's Dale Limestone and the Chee Tor Limestone Members of the Bee Low Limestone Formation. Thirst House Cave associated with dominant stylolite in Woo Dale Limestone (in the order of 9m below the top of the Woo Dale Limestone), marginally downsteam of Deep Dale Cave.

Beneath this, speleogenetic processes appear to be more restricted in the Chee Tor Limestone Member and it seems plausible that the current springs at the base of Blackwell Dale are fed by dissolutionally enlarged fissures, thus lower cave levels are apparently absent. An additional complication is seen in Lathkill Dale, where more than one inception horizon can be active at a given time, and thus the fall in base level is likely to have been more gradual than is apparent from the cave levels.

This author has plotted the thalweg of a number of the valleys (Figures 9.5 to 9.10). The plots suggest the occurrence of up to five phases of accelerated surface dissolution, or erosion. Furthermore, they appear to support the pattern of valley formation indicated in Figures 9.14 and 9.15, with convex surfaces indicating a more mature landscape, as in Monks Dale, above approximately 260 m OD. Using these plots together with the river evolution postulated in Figures 9.14 and 9.15 and the following observations it has been possible to generate a chronology of valley evolution (Table 9.4):

i) Where the Woo Dale Limestone Formation is exposed in the Wye Valley, to the east of Buxton, the northern boundary comprises the Bakewell Fault. There is a well-established conduit system associated with this fault (Chapter 7). Beck (1980) observed an absence of high level caves from the area of the Wormhill Springs i.e. in the Bee Low Limestones and used this to support the argument that the Wormhill Springs and the risings at the confluence of Great Rocks Dale and Chee Dale have been the main outlets of the underground drainage system for a long period.

ii) The Lower Wye must have been at a minimum level of 380 m OD when the headwaters of the Lathkill were captured.

iii) The Upper Wye must have captured the Lower Wye at minimum level of 340 m OD.

iv) Devensian meltwater appears to have masked any evidence of earlier periods of accelerated downcutting on the thalweg plots for Monks Dale, Woo Dale and Cressbrook Dale. The evidence for the impact of Devensian meltwater is absent from thalweg plots of valleys to the south of the River Wye. Furthermore, the relative absence of this water appears to have left nick points, such as Deep Dale (West).

v) The geology and structure also appear to influence the form of the thalwegs.

vi) It was noted in Appendix 4.1 that deposition of the Lower Miller's Dale Lava was associated with recrystallization and virtually entire cementation of the immediately underlying strata. Thus a tight interlocking mosaic of calcite crystals had been formed prior to zone 1 cementation (Berry, 1984). It was speculated that this has rendered the upper layer of the Chee Tor Limestone Member would be more resistant to dissolution. Evidence from the thalweg plot for Monks Dale and Blackwell Dale appear to support this.

9.7 Impact of the Pleistocene.

Conceptually it might be postulated that the effect of glaciations is for meltwater to enhance connections between the epikarst and the karst aquifer, as exemplified in the karst of Ireland (Williams, 1983). In the karst of the Peak District it would appear that whilst being equally significant, this process is not so obvious. This is attributable to the location of the Peak District, to the south of the limit of the most recent advance of ice, i.e. that of the Devensian, specifically of the Dimlington Stadial (Oxygen Isotope Stage 2) (Aitkenhead et al., 2002), evidence for which comes from the remnant covering of loess over much of the region. The pockets of glacial till that do exist were formerly considered to belong mostly to the Wolstonian Stadials (Oxygen Isotope Stages 8-6), but have more recently (Aitkenhead et al., 2002) been suggested to be from an Anglian glaciation (Oxygen Isotope Stage 12). Clearly this is subject to the reservations with respect to dating expressed by Bowen (1999) and described in section 2.10 (p. 25). The deposits are found at levels in the order of 165 to 215 m OD in the vicinity of Bakewell, which corresponds with the Hathersage Terrace of Johnson (1957).

The relative absence of the till from the limestone compared with the surrounding Namurian strata is worthy of comment. Even where till is encountered on the limestone it tends to be in locations where it is underlain by the Eyam Limestone Formation or the overlying Widmerpool Formation. These tend to be more shaley deposits. This distribution suggests a process relationship. For example, Evans et al. (2006) have considered subglacial processes and the coupling of ice with soft sediments to form sub-glacial till. By contrast, where the underlying strata comprise low permeability bedrock, uncoupling can occur as a consequence of elevated pore water pressures. Clearly, in this situation underlying hollows or cavities could become the focus for the deposition of sediment associated with the glacier. This type of process could also have been significant in the localised removal of the strata overlying the limestone and the deposition of sediment in the fissures, dolines and conduits in the limestone.

Burek (1978), Jowett and Charlsworth (1929) and Straw and Lewis (1962) used the distribution of erratics over the region to provide evidence of an earlier more extensive ice advance, which is likely to correspond with the ice advance associated with Oxygen Isotope Stage 16. It has been speculated (Burek, 1978) that the general absence of large-scale glacial erosion forms associated with the earlier advance indicates cold ice, with ice sheets frozen to their beds. The deglaciation of this advance would have been associated with sub-glacial drainage and the development of sub-glacial drainage channels. Accordingly, the deglaciation could have been a period associated with rapid downcutting of the River Wye and its tributaries.

The earliest recorded evidence of Pleistocene activity in the Peak District occurs in the high level caves such as Elder Bush Cave in the Manifold Valley (Rowe et al. 1989). Burek (1991) has also recognised "rare warm temperate faunas normally assigned to the Lower Pleistocene [Villefranchian stage, Bramwell (1977, p. 276)] in Victoria Quarry, Dove Holes. Clearly however, the inverted form of the topography is such that much of the evidence for the earlier glaciations has been removed by later phases of erosion. The evidence presented below indicates that there is likely to be more remnant evidence in valleys to the south of the Wye than those to the north.

Examination of the thalweg profiles indicates that towards the middle of the Wye catchment (e.g. Deep Dale (west), Blackwell Dale, Flag Dale and Wormhill) there is a distinct erosion surface with a base just below 300 m OD. There is further evidence that this was a significant event, because an erosion surface has also been identified at this level at a number of other locations including: west of Wheston Hall (SK 12507635); above Chee Dale (SK 11907315); Kid Tor (SK 09057165); above Millers Dale

(SK 14307395); above the settlement at Chee Tor (SK 12387305); above Fern Dale (SK 16206560); Burfoot (SK 16277273); above Tideswell Dale (SK 15607428) and above Ravencliffe Cave (SK 17607330). This is the Topley Pike Stage (Burek, 1978; Johnson, 1957).

It is interesting to speculate on how the fluvio-karst valleys formed. If the argument for Beestonian (Oxygen Isotope 16), or earlier glacial ice capping the region is accepted, then ice wasting is likely to have occurred at a level above 300 m OD, but below 396 m OD (Burek, 1977). In this situation it is plausible that melt water, both at the feather edge of the ice (in the order of 300 m OD) and that guided by the pre-existing valleys, became focused on dominant joints or faults in the erosion surface, as observed by Sweeting (1972) in the karst of Yorkshire. Rapid dissolution and the formation of flow paths connecting with the base level imposed by the River Derwent, which may have been lowered by permafrost, as considered below, in conjunction with mass movement triggered by solifluction, particularly associated with clay wayboards and by water loading of mineral veins would have contributed to valley evolution. Subsequent scree removal during the subsequent Anglian glaciation would have further contributed to valley development and would have been followed by another phase of dissolution by glacial meltwater

Based on field observations, the examination of Ordnance Survey maps and the evidence presented by Aitkenhead et al. (2002), Burek (1978), Johnson (1957) and Ford and Burek (1977), this author has presented an interpretation of ice movement during the later Anglian advance (Oxygen Isotope Stage 12, or 16, as suggested above), see Figure 9.4. More specifically the evidence to support the figure has been tabulated in Table 9.3, below.

Table 9.3: Evidence in support of Figure 9.4.

Evidence	**Source**
Ice breached the col at Dove Holes	Johnson (1957)
Northwest-southeast striations at Shining Bank Quarry (near Youlgreave) and west to east striations around Eyam and Stoney Middleton, together with till at Stoney Middleton	Aitkenhead et al. (2002), Straw and Lewis (1962)
Distribution of till	BGS Sheets 111 and 99
Pockets of till in Monsal Dale	Burek and Ford (1977)
The smooth U-shaped forms in Monsal Dale, Bradwell, and the Lower Lathkill	Field and topographic map
The paucity of dry valleys in the area to the west of Little Hucklow	Warwick (1964)
The over-steepened, more continuous form of Cressbrook Dale, which appears to have captured the former headwaters feeding into Litton Dale, relative to the other valleys on the north side of the River Wye. This may have formed a meltwater spillway that was also later utilised by Devensian meltwater	Field and topographic map
The direction of Pennine Ice movement indicated in the Regional Geology Guide	Aitkenhead et al. (2002)

Table 9.4: Chronology of valley evolution.

Stage [1]	River Lathkill	Lower Wye	Upper Wye
Holocene (Oxygen Isotope 1)	Speleothem growth		Speleothem growth
Late Devensian Interstadial (Oxygen Isotope 2)		Glacial meltwater targets Great Rocks Dale	Glacial meltwater guided by existing valleys targets Monks Dale and Cressbrook Dale results in convex thalweg
Late Devensian Dimlington Stadial (Oxygen Isotope Stage 2)	Ice margin immediately outside Dinantian limestone outcrop. Close proximity at Dove Holes. Screes develop.	Ice margin immediately outside Dinantian limestone outcrop. Close proximity at Dove Holes. Loess accumulates, screes develop.	Ice margin immediately outside Dinantian limestone outcrop. Close proximity at Dove Holes. Loess accumulates, screes develop
(Oxygen Isotope Stages 8 to 3)	Speleothem growth and cave deposits associated with interstadials. Development of the Lathkill Barrier.	Speleothem growth and cave deposits associated with interstadials.	Speleothem growth and cave deposits associated with interstadials.
Hoxnian (Oxygen Isotope Stage 9)	Speleothem growth e.g. Water Icicle Cave		
Anglian II (Oxygen Isotope Stage 10)			
	Blackwell Dale cuts down taking waters from the River Lathkill	Wormhill Springs exposed and Woo Dale cuts back to a tributary of Flag Dale	Cressbrook Dale reactivated as a spillway
'Swanscombian' (Oxygen Isotope 11)		Further deepening of the River Wye	Further deepening of the River Wye
Anglian I (Oxygen Isotope Stage 12)	Till deposition	Till deposition	Till deposition
	More headward capture by the Churn Hole system. Upper Lathkill captured by Lower Lathkill.	Headward erosion of Kirk Dale, Deep Dale (east) towards the mineral veins that form the groundwater divide with Lathkill Dale	Headward erosion of Cressbrook Dale towards the watercourse associated with Stoney Middleton
Cromerian	Headward erosion of Deep Dale (west) captures waters associated with the river on the Bonsall Fault	Great Rocks Dale captures a tributary to Flag Dale	Monks Dale forms by headward erosion of a tributary to capture groundwater in Peters Dale
Beestonian (Oxygen Isotope Stage 16), or earlier Stages	Formation of the River Wye and localised reversal of hydraulic gradient	Formation of the River Wye, connection with the Upper Wye along Millers Dale, Flag Dale abandoned. Localised reversal of hydraulic gradient. Deposition of 'Older Till'.	Formation of the River Wye and localised reversal of hydraulic gradient. Headward erosion of the Upper Wye along Monsal Dale. Ice blocks the downstream end of the Upper Wye and the meander around the remnant shale block overlying Monsal Head is formed (Johnson, 1957). Deposition of 'Older Till'
Early Pleistocene		Cave deposits found in Victory Quarry confirm that 'unroofing' of the Dinantian limestone had occurred.	
Mio-Pliocene	Capture of headwaters by a tributary of the Lower Wye	Tributary captures headwater of the River Lathkill (current location of Ashwood Dale Resurgence) at a minimum level of 380m AOD.	Tributary captures the Lower Wye headwaters at minimum level of 320m AOD

1. Oxygen Isotope Stages not proven, but tentatively stated to provide dating sequence.

9.8 An outline of the pre-Roman conceptual hydrogeological model.

From an analysis of the discussion above, the area can be divided into a number of hydrogeological units. The units correspond closely with geological members, but differ in detail, as described in Table 9.5, below:

Table 9.5: Description of hydrogeological units.

Hydrogeological Unit.	Description.
Unit 1:	Essentially the dolomitized areas of the Woo Dale Limestone Formation, which forms a zone of relatively high matrix/ fracture hydraulic conductivity in the Woo Dale Limestone, with conduit development and therefore a focus for springs in Ashwood Dale. The hydraulic conductivity recorded for strata encountered in the Monks Dale Borehole was 5.4 x 10^{-3} m/day.
Unit 2:	The remainder of the Woo Dale Limestone has lower fracture/ matrix hydraulic conductivity than the dolomitized areas of the Woo Dale Limestone (hydraulic conductivity in the range 1.05 x 10^{-5} to 2.7 x 10^{-4} m/day). Mature conduit development is evident, for example in the rapid movement of tracer-dye from Tunstead Quarry to Wormhill Springs. Mature conduit development appears to be associated with stylolites. The depositional environment of the Woo Dale Limestone, in particular the Vincent House Member and Topley Pike Member (Schofield and Adams, 1985) was one in which evaporites were deposited and therefore although largely replaced, groundwater might be expected to have a higher sulphate content. Groundwater storage associated with faults has been noted, e.g. at Topley Pike Quarry (Chapter 4).
Unit 3:	In the Chee Tor Limestone Member the intensity of vertical jointing, which has largely developed in tension, with dissolutional enlargement of fissures at outcrop, gives it a very high matrix hydraulic conductivity in the order of 8 to 41 m/day at, or close to outcrop. However, at depth where the fissures are tighter, the fracture/matrix hydraulic conductivity was found to be very low, ~ 8.67 x 10^{-6} m/day. Conduit development is largely absent from this unit. Where faulted against another unit, the Chee Tor Limestone Member acts as an aquitard, forcing groundwater either down or up to continue its flow path, as exemplified in Deep Dale (west), see Figure 9. 16.
Unit 4:	The Miller's Dale Limestone and Monsal Dale Limestone have a matrix/ fracture hydraulic conductivity, in the order of 5 x 10^{-4} m/day. Conduit development on inception horizons is well established in this unit (Chapter 4). Groundwater storage is largely associated with faults and also with clay wayboards, as described above. The distribution of springs suggests a larger number of lower discharge springs than associated with Unit 2.
Unit 5:	Dolomitized Miller's Dale Limestone and Monsal Dale Limestone exhibit a matrix/fracture hydraulic conductivity in the order of 2.5 x10^{-3} m/day, which is attributed to the hydraulic conductivity generated in dedolomitized limestone.
Unit 6:	Volcanic horizons: some horizons appear to be aquitards causing confinement of groundwater, at depth fissuring generates a low matrix hydraulic conductivity possibly comparable with that of the Chee Tor Limestone. Conduit development is not encountered in this unit.

The distribution of the hydrogeological units has been presented on Figure 9.19.

A number of other points require consideration in the development of a pre-Roman hydrogeological model, including:

i) The structural setting has encouraged tiers of groundwater flow paths focusing karst development and storage on the valleys. Evidence for this comes from remnants of high-level dolines being identified in close proximity to springs and the variation in the groundwater chemistry of the springs.

ii) Dominant faults have evolved as zones of groundwater storage, particularly where they are associated with anticlines (e.g. the Taddington and Grin Low anticlines) and form a target for confined water and thereby some also form groundwater divides, which can recede if groundwater pressures fall. Further evidence for the storage associated with faults comes from the apparent barometric response. A plot of the discharge at Ashford (data provided by the Environment Agency) and effective rainfall (calculated by this author from meteorological data for Buxton) has been presented as Figure 9.17.

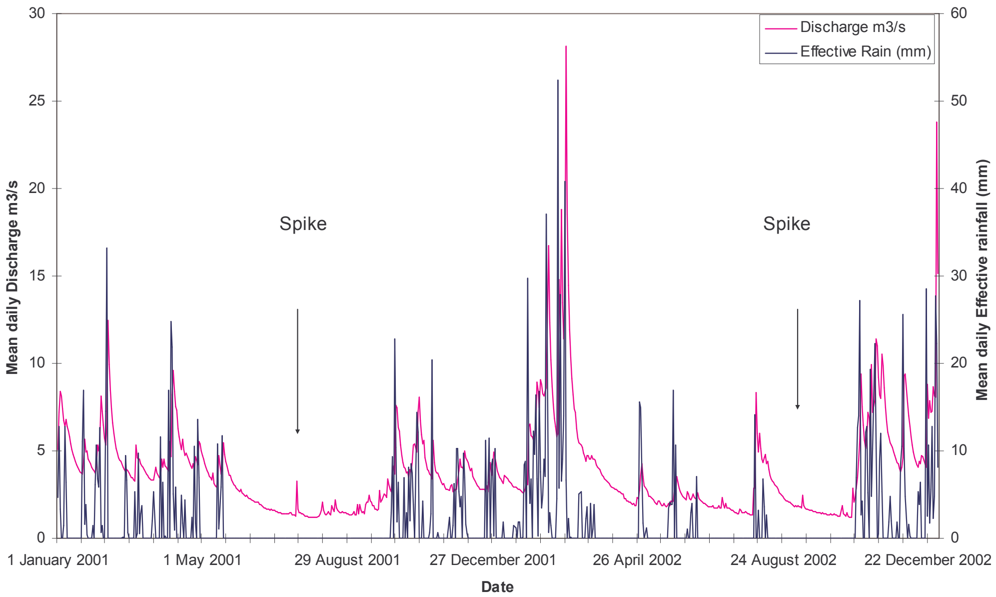


Figure 9.17: Discharge at Ashford (SK 08407250) and effective rain at Buxton 2001 to 2002 (effective rainfall calculated from rainfall data provided by High Peak Borough Council, Discharge data supplied by the Environment Agency).

Examination of the recession portions of the curves reveal that there are rises in discharges that are apparently unrelated to effective rainfall. Each of these peaks, with the exception of the peak of 10 September 2002 (which could be correlated with rainfall on the preceding day) corresponds with a significant fall in barometric pressure. It is considered by this author that the barometric effect is attributable to vertical groundwater movement on faults. Like a well, a fault that is open to the atmosphere and connected with a confined aquifer, will be subject to two opposing consequences of barometric change. In the case of increasing pressure the aquifer is loaded resulting in small rises in groundwater level as a result of transient flow (Lambe and Whitman, 1979), however this is outweighed by the response of the load on the water column, thereby resulting in a net fall in the groundwater level, thus when the pressure falls the water level rises and this response is contributed to by the release of any capillary water. The maximum increase in discharge associated with the barometric effect is 2 m^3/s (on 18 July, 2001), which was associated with a fall in barometric pressure of 6 millibars (an unloading of approximately 0.6 kN/m^2). A small response apparently occurred on 4 September 2002, when a fall in pressure of 9.1 mb resulted in a small increase in discharge (0.04 m^3/s), but this occurred higher in the recession curve and therefore was less marked.

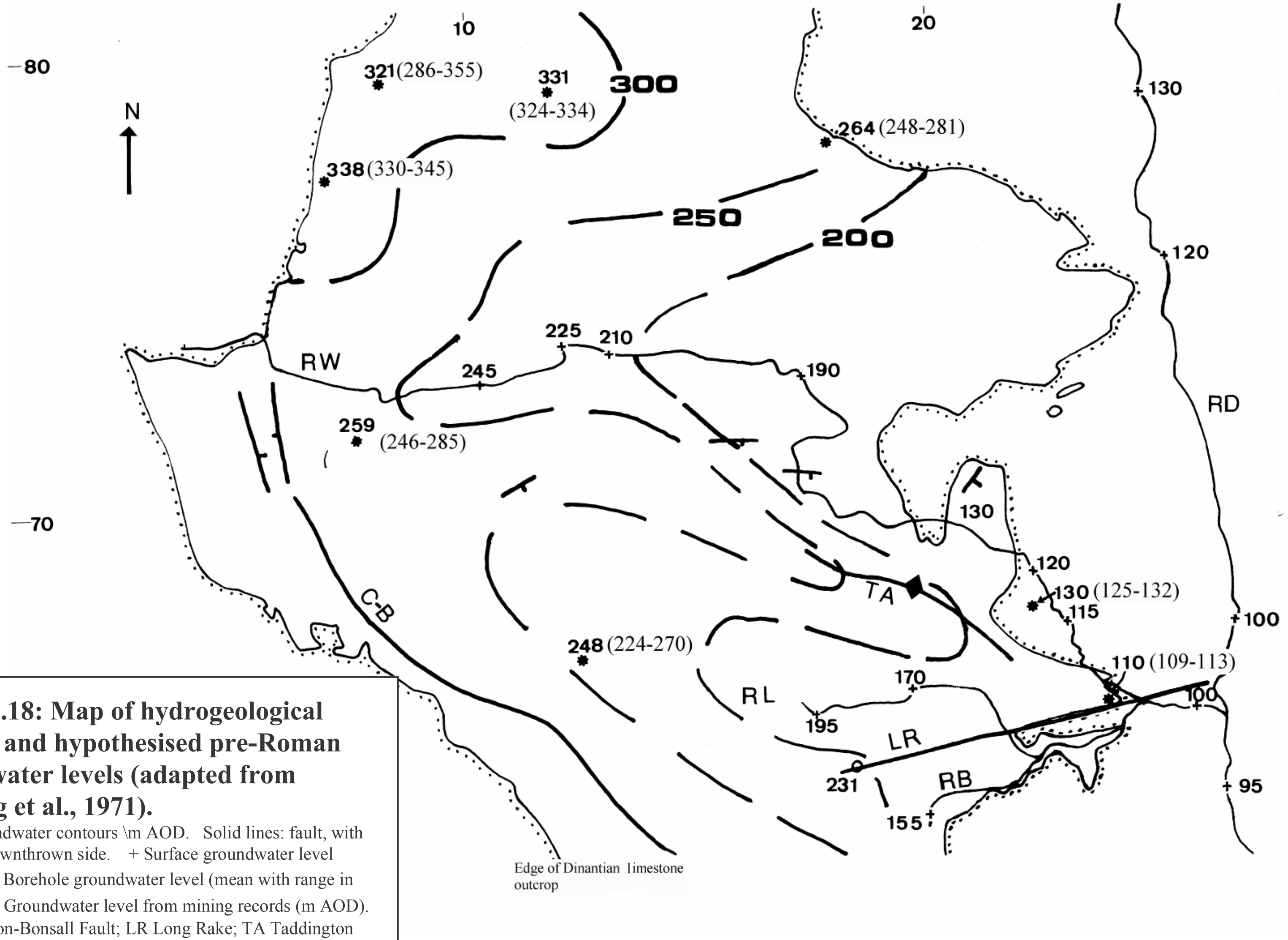


Figure 9.18: Map of hydrogeological features and hypothesised pre-Roman groundwater levels (adapted from Downing et al., 1971).

Key: Groundwater contours \m AOD. Solid lines: fault, with marker on downthrown side. + Surface groundwater level (m AOD) * Borehole groundwater level (mean with range in brackets. ◦ Groundwater level from mining records (m AOD). C-B Cronkston-Bonsall Fault; LR Long Rake; TA Taddington Anticline; RB River Bradford; RD River Derwent; RL River Lathkill; RW River Wye;

iii) Mineralized faults form vertical barriers to groundwater movement and zones of groundwater storage. As with the faults, it is considered that the associated storage occurs both within the fault zone and behind the fault or mineral vein, i.e. where groundwater storage is imposed on the inception horizons, or conduits.

iv) Closed systems associated with the inception horizon development in the Monsal Dale Limestone facilitate groundwater confinement beneath beds of lava.

v) The Chee Tor Limestone Member is heavily fissured with less conduit development.

vi) The concept of underflow and the regional hydrogeological setting i.e. the existence of underflow from proximal and distal locations and also underflow to distal locations.

vii) Domed water, as identified by Downing et al. (1970) comprises groundwater perched by lavas and stored in the Monsal Dale Limestone Formation.

viii) Many of the dye-tracing tests show evidence of fast flow followed by longer term dispersed flow, indicative of storage of dye.

ix) Because transmissivity in the limestones is low, seasonal groundwater level fluctuations are high, as indicated by the groundwater levels recorded in the boreholes. This impacts on the hydrology because water levels fall below the base of the lavas, to the extent that for part of the year the stretch of the Wye immediately downstream of Monks Dale is perched.

x) The occurrence of pyrite in clay wayboards in the Chee Tor Limestone Member suggests that groundwater has not moved through these horizons.

xi) There are a large number of high-level springs associated with groundwater bodies perched above volcanic rocks, e.g. Priestcliffe and Chelmorton.

These points have been combined in the block diagram, presented as Figure 9.20: the pre-Roman hydrogeological model.

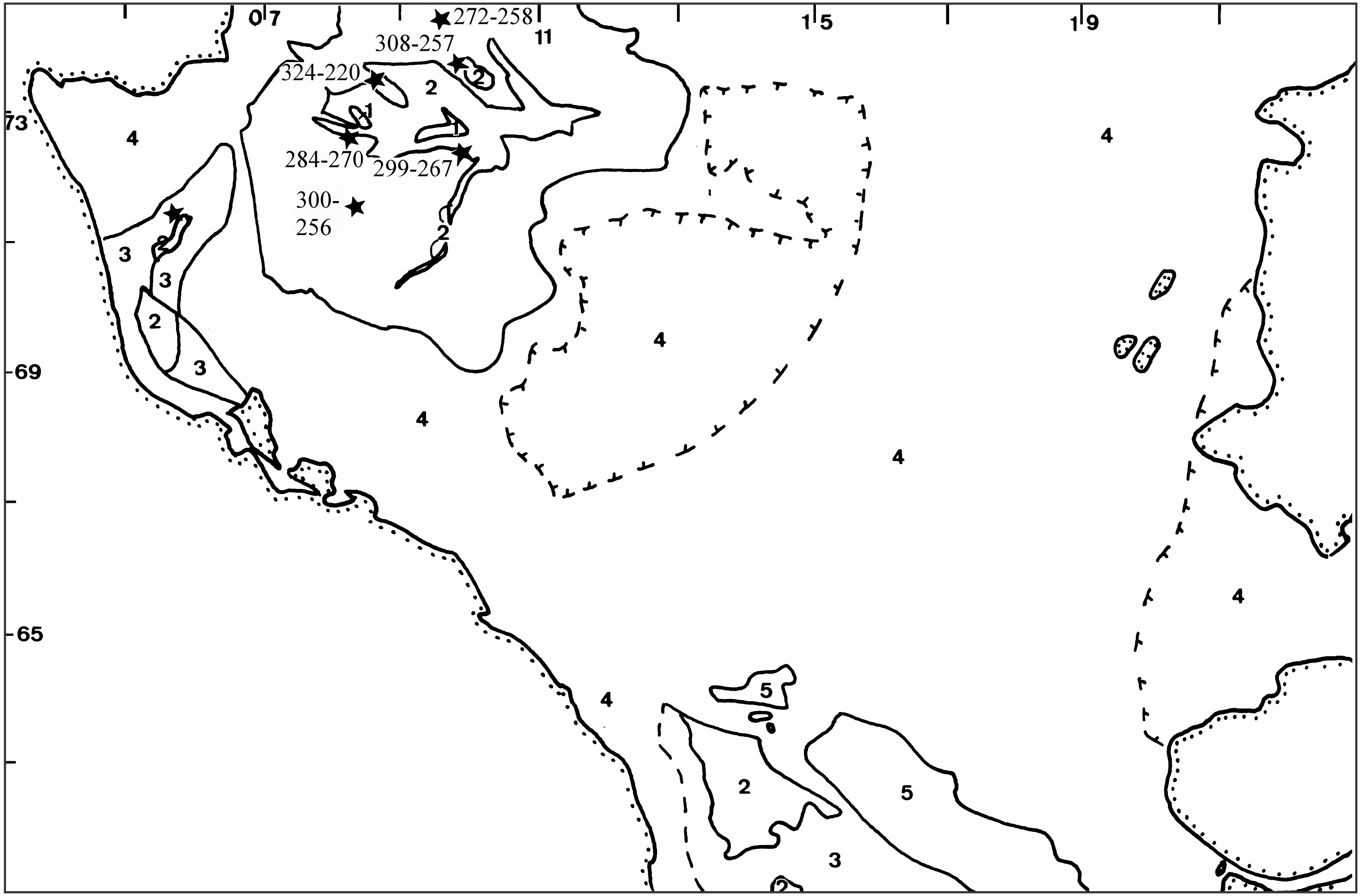
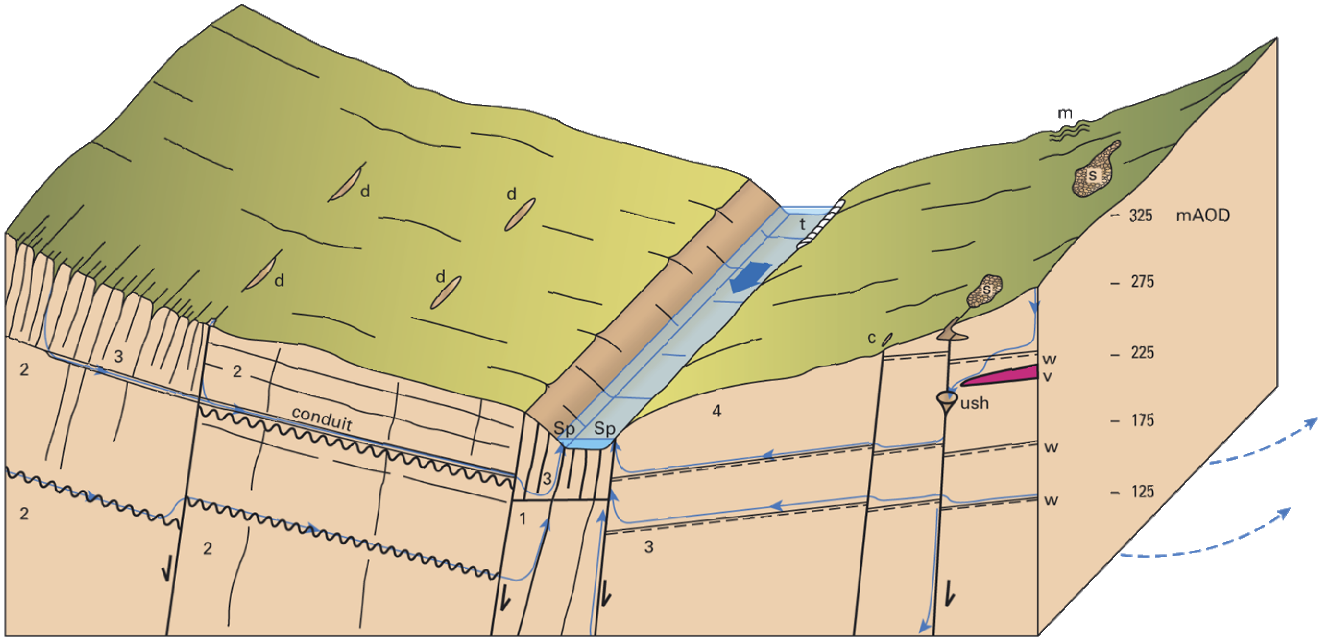


Figure 9.19: Map to show the distribution of hydrogeological units (1-5, see Table 9.5 for description).

Key: Dashed line with fleck pointing in to: Groundwater perched by underlying lavas; Line with parallel dots: Edge of the outcrop of the Dinantian limestone; Dashed line: Uncertain boundary; Unit 2 present at depth zone indicated (m AOD)



ush	Underground swallow hole	1 - 4	Hydrogeological units 1-4
t	Tufa	– – –	Clay wayboard, note conduit development at boundary between units 2 and 3
sp	Spring		Stylolite
d	Doline		Flow path
c	Cave		Fault displacement arrow on downthrow side
s	Sand pocket/ doline		Underflow
m	Mass movement down dip of mineral vein		
w	Wayboard		
v	Lava		

Figure: 9.20: Conceptual model of the limestone hydrogeology in the Wye catchment.

(Digitised by Paul Lappage).

Chapter 10: Human impacts on the hydrogeology.

10.1 Introduction.

This thesis considers human impacts under the following sub-headings: lowering of groundwater levels and alteration of hydraulic gradient; modifications to permeability; effects on storage and transmissivity; modification of groundwater divides; underflow captured by Magpie Sough; impact on groundwater quality; and climatic change. It has not been possible to give detailed consideration to all of the human impacts that have influenced the Limestone of the Peak District, e.g. no consideration has been given to atmospheric pollution as a mechanism for increasing rates of limestone dissolution, and very little consideration has been given to changes in agricultural practices or phases of deforestation, known to cause soil erosion, which Gvozdetsky (1988) suggested is the most significant of man's impacts on the karst. Although there have been phases of deforestation in the White Peak, in particular during the late Mesolithic (6090 to 5740 BP) and late Neolithic to Early Bronze Age (4100 BP), as described by Taylor et al. (1994) and also associated with the subsequent mining activities (see also Merton, 1970), the most significant human impacts appear to be associated with exploitation of the resources of the limestone (Chapter 3). Mining activities were particularly focused on the Monsal Dale Limestone and Miller's Dale Limestone formations. This is because it would appear that the strike-slip nature of the faulting, in response to seismic pumping (Colman et al., 1989) served to effectively trap the mineralizing fluids. The effects of the earliest phases of mining were largely restricted to the Monsal Dale Limestone, but as the economic pressure to extract ore from deeper levels grew, the effects of dewatering were extended to greater depth, thus into the Bee Low Limestone and to some extent the Woo Dale Limestone formations.

The human impact assessment follows the standard approach to categorising aquifers in terms of: degree of confinement, porosity, permeability, transmissivity and storage potential. However, it should be noted that the properties of limestone are more complex than those of other aquifers. This complexity is brought about by the way in which indurated limestone achieves its permeability, i.e. by dissolution. Dissolutional processes target specific, inherently vulnerable areas of the limestone (Chapter 4). The evidence that has been presented suggests that in hydrogeological unit 4 this is primarily related to bedding, in unit 3 to jointing and in unit 2 to stylolites (Table 9.5). Accordingly the mass properties of the limestone aquifer are characterised more by the secondary (channel) porosity than by matrix flow, although matrix flow in the context of the clay wayboards (unit 4) should not be ignored. The secondary (channel) porosity increases over time, as dissolution is focused on inception horizons, which may be connected vertically along joints and faults. In hydrogeological unit 3 the dissolutional enlargement of jointing appears to be limited to near surface, or surface exposure of the limestone. At depth this unit acts as an aquitard, punctured only by faults and dominant fissures. Conceptually this can be seen as a series of partially confined aquifers, particularly in the Monsal Dale Limestone Formation. Conduits (tertiary porosity) form by preferential enlargement on specific inception horizons. They provide a focal point for a number of channels and the main outlet points for drainage from the aquifer (section 10.3).

10.2 Lowering of groundwater levels and alteration of the hydraulic gradient.

Driving soughs into valley sides as a technique to drain the mine workings gradually progressed, over time, to deeper levels, extending over increasing distances to reach the deeper valleys. This resulted in a gradual lowering of groundwater levels, for instance Willies (1976, p. 146) states "*Soughs prior to Hillcarr had drained the area more or less to the level of the Lathkill-Bradford near Hawley's bridge (a level of approximately 146 m OD), but except for ventilation these were made more or less redundant, and by about 1796, in the years since 1787 when Hillcarr reached Guy vein, the whole area was drained by about 72 feet (*22 m*) below this*". Ford and Rieuwerts (2000, p. 54) stated that "*the driving of Hillcarr Sough in the latter half of the eighteenth century gave a new lease to the mines south of the River Lathkill, near Alport, but the veins rapidly became exhausted down to the level of the sough by the beginning of the nineteenth century*". In order to look at the way in which human activities have impacted on groundwater levels this author has generated a table of groundwater levels projected from the sough tail levels (Table 10.1). These values have been determined as a range. The lower value has been determined by assuming that the earliest workings would have been taken down to approximately 0.5 m below the standing water level at the head of the sough and assuming a sough gradient of 1.9 m/km (Oakman, 1979). The upper value was determined from the maximum local hydraulic gradient calculated to be 1m/50 m (interpreted from Downing et al., 1970). It is the earliest soughs, about which there are the fewest detailed records, which have the potential to provide the best indication of pre-mining groundwater levels in the Monsal Dale Limestone. These values indicate a lowering of the groundwater table by approximately 2 to 7 m. However, bearing in mind that the level of the River Derwent at the point at which it leaves the limestone catchment is approximately 75 m OD, and because mature karst aquifers cannot sustain steep hydraulic gradients, it is considered by this author that these groundwater levels are indicative of the level of perched bodies of groundwater. In support of this it has already been noted that an intermediate hydraulic gradient of 1 m/100 m was indicated by Downing et al. (1970), compared with the regional hydraulic gradient of 1 m/1 km reported by Wilson and Luheshi (1987). Many of the soughs to which reference has been made in this table were soughs of the lower reaches of the River Lathkill and the River Bradford, which are extensively underlain by volcanic strata of the Fallgate Formation, which have the potential to act as aquitards.

Other early soughs included Cross Flatts Sough (associated with Dale Sough); Dale Sough; Foxhole Sough (SK 233651), Haddon New Sough (not located); The Sugh (not traced); Mire Close Sough (not located); and Old Earles Level (Haddon Fields Deep Level), however there is insufficient information for inclusion in Table 10.1. Reference to these soughs is made by Rieuwerts (1987). It is not known whether a sough referred to as Will Sough at SK 209639 (known only from records) was constructed (Rieuwerts, 1987). Further information with respect to the soughs can also be found in Appendix 3.4.

Rieuwerts (1980) has described how sough technology improved with time, thus enabling deeper levels of dewatering to be achieved. From as early as the sixteenth century the greater depth was achieved in part by down-hole pumping. Initially this comprised hand pumping, which gave way to horse-powered pumps, followed by engines. The engines were initially water powered and then steam powered and utilised pre-existing soughs as pump-ways. Because the mining was largely carried out along mineralized fault planes (albeit that many are primarily strike-slip faults) it was being carried out in the zones of maximum vertical groundwater movement (Chapter 9). Significantly, Naylor (1983) described how when the Magpie Mine soughers broke through into mine workings on 18 August 1881 a head of 33 m was released through a pilot hole with a force of 47 lb per square inch, or 3.22 bar. Naylor's source was Willies et al. (1980), but it is not accurately reported, for it was on cutting Townhead Vein that a significant reduction in head was achieved. Nevertheless, the implication remains the same and it is from Townhead Vein that most of the flow comes from today (Ford personal communication, 2007). This author has presented the relevant detail with respect to groundwater conditions as Figure 10.1.

Table 10.1: Pre-mining groundwater levels projected from local, vein sough tails (sources: Oakman (1979) and Rieuwerts (1987)).

Sough Name:	National Grid Reference Tail:	Altitude of the tail (m OD):	Current status:	Length of sough (m):	Interpolated pre-sough ground water level (m OD):
Black Shale Pits Sough	SK 2168 6375	143		206	144-147
Blythe Sough	SK 2310 6441	130	run in	not known	
Bowers Rake Sough	SK2334 6509	125	run in	not known	
Dale Sough	SK 199 635	170-175	open bolt	not known	
Dale Vein Sough	SK 223 649	124	run in	350?	126-131
Hartle Calf Croft Sough	SK 230 646	130	run in	not known	
Leewall Sough	SK 2183 6435	130	run in	305	132-136
Nick Sough	SK 2035 6390	155	tail goite remnant	not known	
Rainstor Sough	SK 2387 6545	110	run in, but marshy	275	112-115
Rowsley Level	SK 2511 6562	102	run in	not known	
Sellers Sough	SK 236 651	110	Not located	550 estimated	112-121
Stoney Lee Sough	SK 2311 6475	120	run in		
Timperley Sough	SK 1991 6353	170-175	run in	not known	
Wheels Rake Old Sough	SK 2275 6490	120	run in	700 estimated	122-134
Hardyhead Sough	SK 1290 7120	371	Still discharges water	275	372-376

In zones of significant vertical groundwater movement the extent of groundwater lowering is more difficult to determine. The conceptual model (Chapter 9) suggests that some of the movement is underflow, possibly even from basement level, with a potential for the head to be derived from outside the outcrop of the limestone. In this situation some of the variation in head may be attributable to human activities outside the area of the outcrop of the limestone, which is considered to be beyond the remit of this thesis. Furthermore, it is clear that the head distribution across the limestone is influenced by a number of other factors, not least of which, is the potential for head loss attributable to seepage forces associated with the lower permeability flow paths. Evidence from Lathkill Dale (Chapter 11), where a dominant inception horizon has been traced, suggests to this author that the head that was encountered in the Townhead Vein is also attributable to the way in which groundwater follows dissolutional channels formed on inception horizons parallel with the dip of the limestone to depths that achieve significant levels of confinement, particularly beneath the less jointed lavas. Furthermore, it would seem reasonable to suggest that prior to its construction some of the water captured by Magpie Sough (underflow) resurged in the Bakewell thermal springs, as speculated by Ford (1980).

Evidence for the significance of the impacts of sough construction from within the area of the limestone outcrop is clear. The catchment of the River Wye upstream of Magpie Sough, as determined from the Flood Estimation Handbook (NERC, 1999) CD-ROM is 152.2 km^2. The mean discharge for the periods of baseflow (17 May to 5 October, 2001 and 19 March to 21 October, 2002) have been determined, from data supplied for the River Wye at Ashford, by the Environment Agency (Figure 10.2), as 2190 litres/sec. The discharge from Magpie Sough (SK 17926957) determined by Edmunds (1971) and Christopher (1981) approximated to 400 litres/sec, with little seasonal variation. This indicates that during low groundwater conditions up to 20 % of the baseflow of the River Wye at Ashford is derived from Magpie Sough, with a mean of approximately 12 % for the year as a whole. The results of low flow surveys carried out by the Environment Agency on a number of occasions during the period 1989 to 1995 were presented in the Hydrometric Report for the Environment Agency Midlands Region (Dixon, 1996). Data relevant to the area of this research have been presented in Appendix 10.1 of this thesis. These results indicate low discharge values for Magpie Sough to range between 325 to 434 litres/sec, with the sough discharge comprising up to 40 % of the discharge of the River Wye immediately downstream of the sough.

Clearly this has impacted on the potentiometric surface of the limestone. However, not all of the discharge from Magpie Sough should be considered additional discharge to the River Wye, because it is reported (Willies, 1980) that springs at Sheldon (Netler Dale, Figure 10.1) dried up in response to the cutting of Magpie Sough, albeit that this author has observed (18 October, 2002) that there is still a minimal discharge from the springs. Nevertheless it is considered that the excavation of Magpie Sough has increased the discharge to the River Wye upstream of Bakewell and Ford (1980) has observed that 95 % of the flow of Magpie Sough rises from Townhead Vein. It is not certain what proportion of this would have contributed underflow to the River Derwent, or beyond.

Consideration has been given to the calculation of the contribution of underflow to the River Wye. The effective rainfall figures that have been calculated, from daily rainfall figures for Buxton, for the years 2001 and 2002 (Chapter 9) have been plotted with the discharge of the River Wye at Ashford (Figure 10.2). The mean annual effective rainfall determined by this author for the years 2001 to 2003 is 680 mm. Calculated from a catchment area of 152.2 km^2 this would generate a mean discharge of

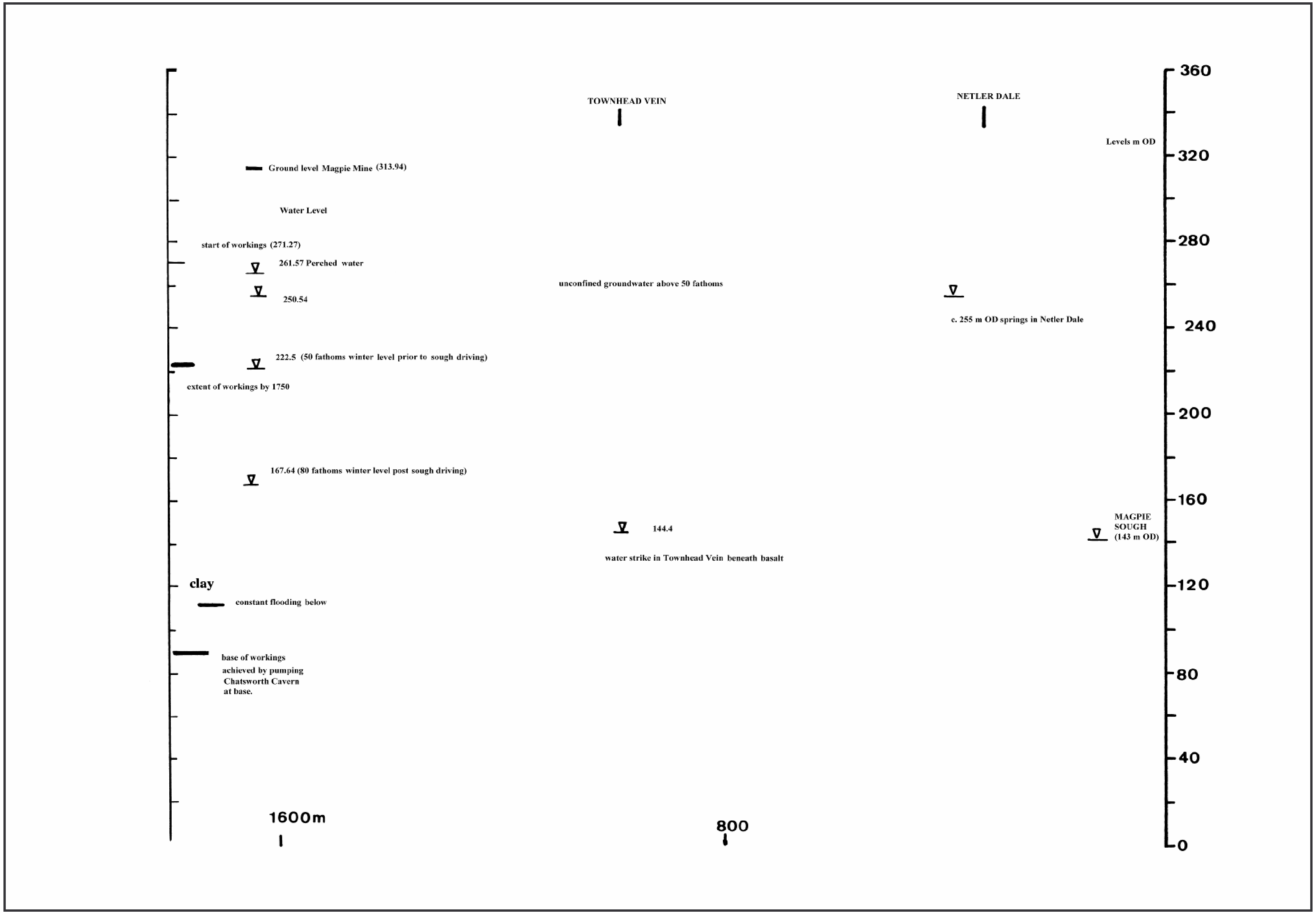


Figure 10.1: Groundwater conditions indicated by Magpie Sough (Shaft sinking commenced 1823).

3280 litres/sec at the location of Magpie Sough. The mean discharge determined for 2001 and 2002 from daily data collected by the Environment Agency at Ashford (immediately downstream of Magpie Sough and for which the Environment Agency have calculated a catchment area of 154 km^2) is 3870 litres/sec (a specific discharge of 25.1 litres/sec/km^2). Subtracting the discharge from Magpie Sough (400 litres/sec) indicated the measured discharge at Ashford to be 3470 litres/sec. Comparison of this value with the calculated value at Magpie Sough (3470-3280 litres/sec) suggests a mean underflow contribution in the order of 190 litres/sec (a specific discharge of 1.23 litres/sec/km^2). Clearly this value is open to some debate, but it might be argued that if anything it is lower than the actual value, because the topographic catchment determined by the FEH Handbook (1999) CD-ROM method includes the area of Chelmorton, which dye-tracing has shown to fall partially within the Lathkill catchment. Furthermore, it is considered by this author that at least some of the flow captured by Magpie Sough is in itself underflow. Christopher (1981) carried out groundwater sampling in Magpie Sough and found that water rising up the main shaft from workings below the sough (boil-ups) in Townhead Vein had an elevated temperature on the east side of the sough (9.7˚ C) compared with the west side (8.9˚ C), which suggests more than one source; furthermore he found that water in the Blende Vein and Main Shaft was 8.4 and 8.2˚ C respectively, whereas the water emerging from the sough tail had a temperature of 9.05˚ C. A temperature of 9.87˚ C was recorded by this author for the sough tail water on 30 April 2002. More widely also, there is considerable evidence of warm water, indicative of underflow having been captured in other soughs, e.g. Stoke Sough (SK 24007640) and Meerbrook Sough (SK33205490). Glover (1831, p. 22) describing thermal springs noted that *"... Middleton, near Wirksworth, had formerly a spring of this description, which was cut off some years since by the driving of a sough to remove the water from some lead mines in the neighbourhood*."

It has been noted that the values for underflow presented above exceed those calculated by Downing et al. (1970), namely that direct infiltration to the limestone in the Wye catchment has been calculated as 57.8 million gallons per day [mgd] (3050 litres/sec), with recharge from the Millstone Grit amounting to 1.0 mgd (53 litres/sec). The figures calculated for the Lathkill catchment were 903 litres/sec direct infiltration to the limestone and 185 litres/sec from the Millstone Grit. However, the calculations presented by Downing et al. (1970) were based on surface water catchment water balance calculations and did not consider underflow reaching or leaving the catchment.

Not surprisingly examination of the map of the principal soughs (Figure 10.3) indicates that the majority are concentrated to the east of the limestone, i.e. reflecting construction from progressively lower valley levels: local valleys, the Wye and finally the Derwent. It should also be noted that a number of soughs were driven to underground swallow holes, thereby lowering bodies of perched groundwater. This activity was particularly notable in the area of Great Hucklow, a number of which have been identified on Figure 10.3. Nevertheless, the inevitable dominance of soughs draining to the east (Magpie, Hillcarr and Meerbrook in particular) has further enhanced the southeasterly hydraulic gradient etched by glaciations, thereby increasing the hydraulic gradient of the zones of perched groundwater, with a consequent potential for increased rates of dissolution. However, the measurement of this response lies beyond the work covered by this thesis.

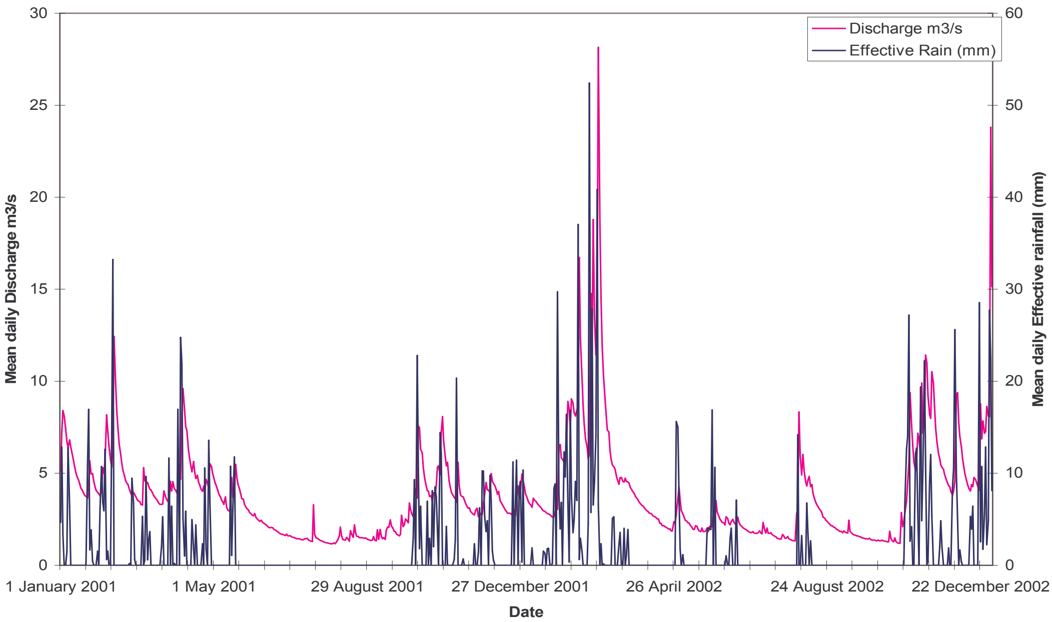


Figure 10.2: River Wye, mean daily discharge at Ashford and calculated daily effective rain at Buxton, 2001 and 2002.

10.3 Modifications to permeability.

It is interesting to compare the table of mean spring discharge with that of mean sough discharges (Table 10.2). The highest discharges are associated with the soughs, in particular Magpie Sough and formerly Hillcarr Sough, which effectively act as large conduits. Like the springs, some of the soughs, notably Lathkill Dale Sough and Mandale Sough are seasonal. It is noteworthy that there are no significant springs between Magpie Sough and Bakewell, which suggests that Magpie Sough has captured groundwater that may have fed springs further to the southeast. For instance it is known that the construction of Magpie Sough resulted in the 'drying up' of springs at Sheldon, as reported by Willies (1980). However, the groundwater feeding the springs was perched groundwater associated with clay wayboards and the underlying Great Shacklow Lava (which was drained by Magpie Sough). Therefore, an alternative explanation, and one which is preferred by this author, is that the paucity of springs actually reflects the increasingly perched nature of the River Wye to the east, as indicated by the findings of the Monks Dale Borehole (Chapter 8). Furthermore, it has been reported that in 1864 river level was artificially raised at Litton. Ford (personal communication, 2007) advised that the water level at Litton Mill is dammed by a weir of approximately 2m height, as is the water at Cressbrook

Mill. It is the opinion of this author that the hydrogeology of this area has also been influenced by extensive silica precipitation associated with precursor, or early mineralization (Chapter 3).

Table 10.2: Sough and spring discharges.

Sough:	Date:	Discharge (l/s):	Source of Data:	National Grid Reference:
Magpie Sough	Oct 73 -Nov 78	409	Christopher, 1981	SK 179 696
	1967	396	Naylor, 1983	
	1929	422-502	Naylor, 1983	
	26.10.67	400	Edmunds, 1971	
Stoke Sough	25.3.69	1.3 *	Edmunds, 1971	SK 240 766
Moorwood Sough				SK 232 754
Watergrove Sough				SK 210 759
Hillcarr Sough		370	Stephens, 1929	SK 257 637
	1929	370-422	Naylor, 1983	
	1967	224	Naylor, 1983	
		250	Edmunds, 1971	
		65	this thesis	
Basrobin Sough	1979	4.6l/s, seasonally dry	Oakman, 1979	SK 2620 6096
Yatestoop Sough				SK 264 626
Meerbrook Sough		739	Stephens, 1929	SK 327 552
		43.40 to Severn Trent	Ford and Rieuwerts, 2000	
	14.6.68	790	Edmunds, 1971	
Maury Sough	26.3.69	20	Edmunds, 1971	
Mandale Sough	25.10.67	17, seasonally dry	Edmunds, 1971	
Lathkill Dale Sough	5.4.03	24, seasonally dry	this thesis	SK 2050 6612
Basrobin Sough	1968/9	4	Edmunds, 1971	
Waterloo Inn Sough	17.10.67	4	Edmunds, 1971	
Brightside Sough	25.3.69	67	Edmunds, 1971	SK 242 745
Snitterton Sough	26.4.69	4	Edmunds, 1971	SK 282 608
Oxclose Sough	26.4.69	17	Edmunds, 1971	SK 289 606
Cromford Sough	24.3.69	24	Edmunds, 1971	SK 296 569
Odin Sough	27.3.69	7	Edmunds, 1971	SK 150 832

Spring/Outlet:	Date	Discharge (l/s)	Source of data	National Grid Reference:
Barmoor	27.3.69	2.5	Edmunds, 1971	SK 085 797
Cowdale (Rockhead) Spring	Mean, 1998	15.8	Smith et al., 2001	SK 0866 7229
Cowdale (Rockhead Spring	Nov 77 to Nov 79	8	Christopher, 1981	
Kidtor Spring	Mean, 1998	4.9	Smith et al., 2001	SK 0868 7219
Ashwood Dale Resurgence	Mean, 1998	29.3	Smith et al., 2001	SK 0895 7223
Ashwood Dale Resurgence		40	Edmunds, 1971	
Ashwood Dale Resurgence		53	Stephens, 1929	
Bole Hill Spring	25.3.69	7	Edmunds, 1971	SK 102 753
Mill Cottage Dam	25.3.69	12	Edmunds, 1971	SK 116 787
Brook Head	25.3.69	10	Edmunds, 1971	SK 141 775
Great Hucklow Stream	25.3.69	3	Edmunds, 1971	SK 179 778
Grindlow	25.3.69	0.2	Edmunds, 1971	SK 180 772

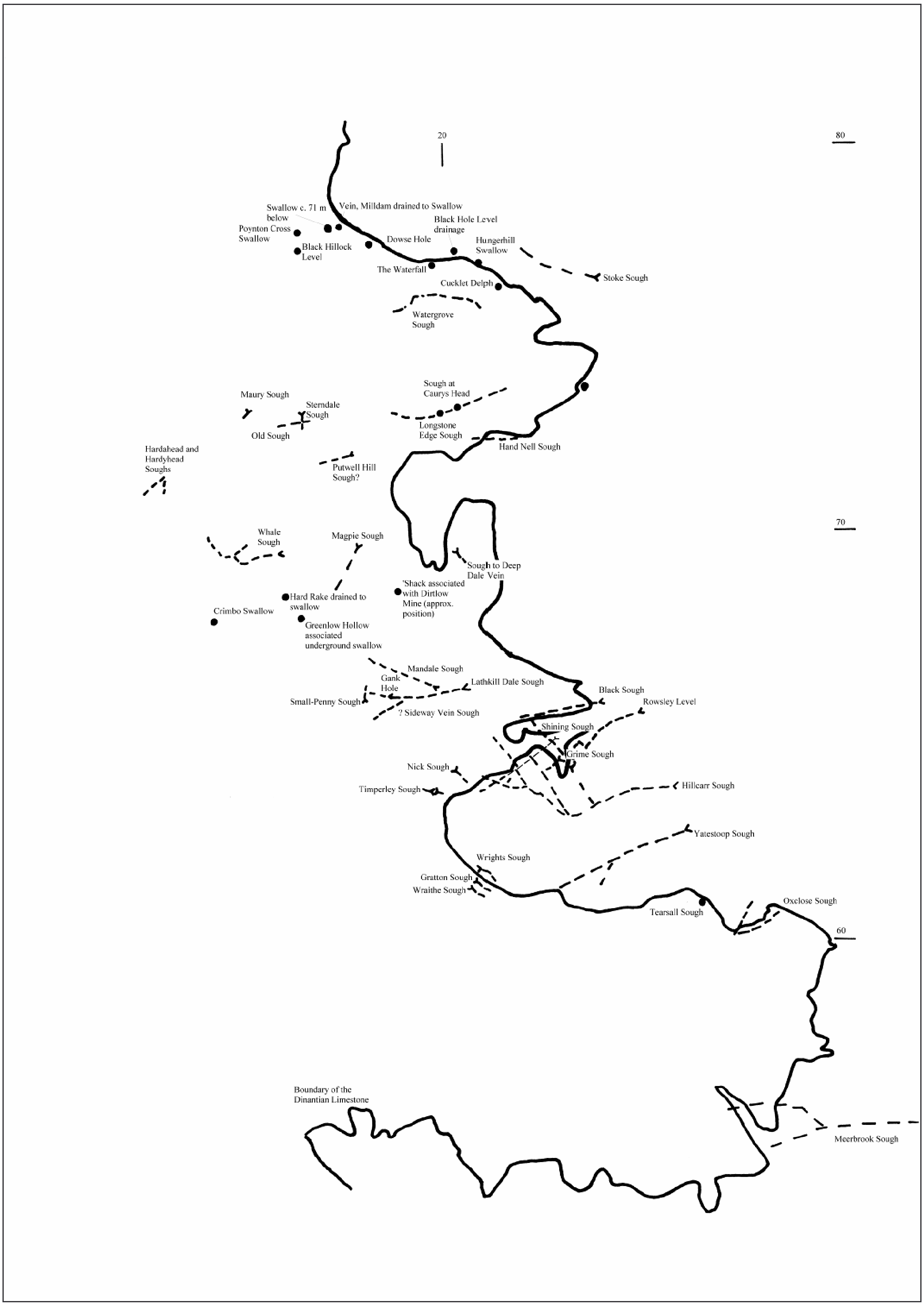
Spring/Outlet:	Date	Discharge (l/s)	Source of data	National Grid Reference:
Waterfall Swallow	25.3.69	13	Edmunds, 1971	SK 198 771
Pictor Spring (north side of Wye)	26.3.69	40	Edmunds, 1971	SK 088 722
Pictor Spring		42	Christopher et al., 1977	
Woo Dale	26.3.69	1.1	Edmunds, 1971	SK 094 725
Woolow	Mean, 1998	1.7	Smith et al., 2001	SK 0947 7242
Topley Pike	Mean, 1998	8.2	Smith et al., 2001	SK 1000 7248
Wormhill Moor	23.10.67	3	Edmunds, 1971	SK 1070 7570
Wormhill Springs West	Mean, 1998	197.6	Smith et al., 2001	SK 1231 7352
Wormhill Springs East	Mean, 1998	182.8	Smith et al., 2001	SK 1237 7352
Wormhill Springs		528	Christopher et al., 1977	
Wormhill Springs	Oct 73 to Nov 78	307	Christopher, 1981	
Monks Dale Spring	10.6.68	1.7	Edmunds, 1971	SK 137 740
Millers Dale 1	26.3.69	1.3	Edmunds, 1971	SK 137 731
Litton Mill	Mean, 1998	9.4	Smith et al., 2001	SK 1610 7295
Litton Mill 2	26.3.69	0.5	Edmunds, 1971	SK 161 729
Litton Mill 1	26.3.69	2.5	Edmunds, 1971	SK 165 732
Cheedale West	Mean, 1998	4.3	Smith et al., 2001	SK 1268 7348
Cheedale East	Mean, 1998	35.9	Smith et al., 2001	SK 1237 7346
Cheedale Bridge Resurgence	Mean, 1998	2.2	Smith et al., 2001	SK 1278 7347
White Cliff Spring	Mean, 1998	4.2	Smith et al., 2001	SK 1812 7186
Chelmorton (Illy Willy)	18.10.67	5	Edmunds, 1971	SK 1153 7033
Nether Low	23.4.69	0.1	Edmunds, 1971	SK 111 692
Taddington High Well	26.3.69	0.1	Edmunds, 1971	SK 1440 7080
Lees Bottom 1	Mean, 1998	15.7	Smith et al., 2001	SK 1704 7080
Lees Bottom 2	Mean, 1998	26.6	Smith et al., 2001	SK 1703 7071
Lees Bottom 3	Mean, 1998	3.2	Smith et al., 2001	SK 1716 7058
Lower Dimindale	26.10.67	4.2	Edmunds, 1971	
Lees Bottom 4	Mean, 1998	4.5	Smith et al., 2001	SK 1717 7044
Great Shacklow 1	Mean, 1998	2.2	Smith et al., 2001	SK 1773 9992
Great Shacklow 2	Mean, 1998	20.2	Smith et al., 2001	SK 1775 6986
Great Shacklow	26.10.67	6	Edmunds, 1971	
Brindley's Well	Nov 77 to Nov 79	2	Christopher, 1981	SK 1230 7430
Lumb Hole, (Cressbrook Dale)	Oct 73 to Nov 79	24	Christopher, 1981	SK 1735 7270
Lumb Hole, (Cressbrook Dale)		53	Stephens, 1929	
St Anne's Well, Buxton	Dec 77 to Nov 79	7	Christopher, 1981	SK 0570 7350
Buxton	1.11.67	10.6	Edmunds, 1971	
Deep Dale-Topley Pike Outfall	Jan 74 to Nov 77	100	Christopher, 1981	SK 1028 7246
Deep Dale Resurgence		25	Edmunds, 1971	SK 0970 7130
Moss Well	22.4.69	1.3	Edmunds, 1971	SK 178 721
Hay Dale	22.4.69	0.9	Edmunds, 1971	SK 178 721

Spring/Outlet:	Date	Discharge (l/s)	Source of data	National Grid Reference:
Ravensdale Cottages	22.4.69	0.3	Edmunds, 1971	SK 172 737
Wye Head		26.4	Christopher et al., 1977	SK 0499 7304
Tunstead Well	23.4.69	2.5	Edmunds, 1971	SK 109 748
Crowhill Lane	22.4.69	0.3	Edmunds, 1971	SK 204 692
Dirtlow Farm 1	22.4.69	0.1	Edmunds, 1971	SK 188686
Dirtlow Farm 2	22.4.69	0.1	Edmunds, 1971	SK 193 685
Sheldon		0.06	Stephens, 1929	SK 1755 6976
Bakewell British Legion	25.3.69	9.3, or 0.3	Edmunds, 1971	SK 2180 6860
Bakewell Recreation Ground	25.3.69	0.2	Edmunds, 1971	SK 2200 6810

***** suspected to be in error (Gunn personal communication 2005)

Of the springs, it is those that are associated with dominant faults that exhibit the highest discharges; in particular: Wormhill Springs East and West, Ashwood Dale, Pictor, Cheedale East and Lees Bottom 2. This is important in that the location of the faults gives a good indication of direction of the hydraulic gradient at these locations (approximately perpendicular to the fault). Where the springs exhibit elevated temperatures and capture underflow this is indicative of regional flow paths. There is an increase in the number of springs, albeit that they are of lower discharge, in an easterly direction along the River Wye (Figure 10.4). This is in keeping with the conceptual model (Chapter 9). The large number of lower discharge springs is associated with hydrogeological unit 4 and a significant number of the springs associated with this unit have been classified as overflow springs (Chapter 6). The data provide further evidence of the significance of groundwater storage associated with faults (within and 'backed up' against the fault), because the higher discharge springs are associated with fault zones (Figure 10.4).

Soughs effectively act as large, or very large, conduits, with the higher discharge being attributable to cross sectional area, hydraulic gradient (usually in the order of 1.9 m/km, Oakman, 1979) and increased permeability as a result of the reduced influence of channel form on discharge. The increase in the permeability indicates that effective rainfall is conveyed more rapidly to the surface water courses, either directly via soughs discharging to the River Wye, such as Magpie Sough, or Maury Sough (Figure 10.3), or in the case of higher level soughs, indirectly via the Rivers Lathkill and Bradford. Other high level soughs such as Hardyhead Sough and the soughs discharging to vadose shafts, locally increase the concentrated recharge associated with a section of a given flow path, for in the case of high level soughs like Hardyhead Sough, groundwater emerging from the sough sinks again. There is a potential for the latter situation to be associated with breaching of groundwater divides (section 10.5).



Key: Dashed line indicates line of sough.

Figure 10.3: Principal soughs of the research area (data from Downing et al., 1970 and Rieuwerts, 1987).

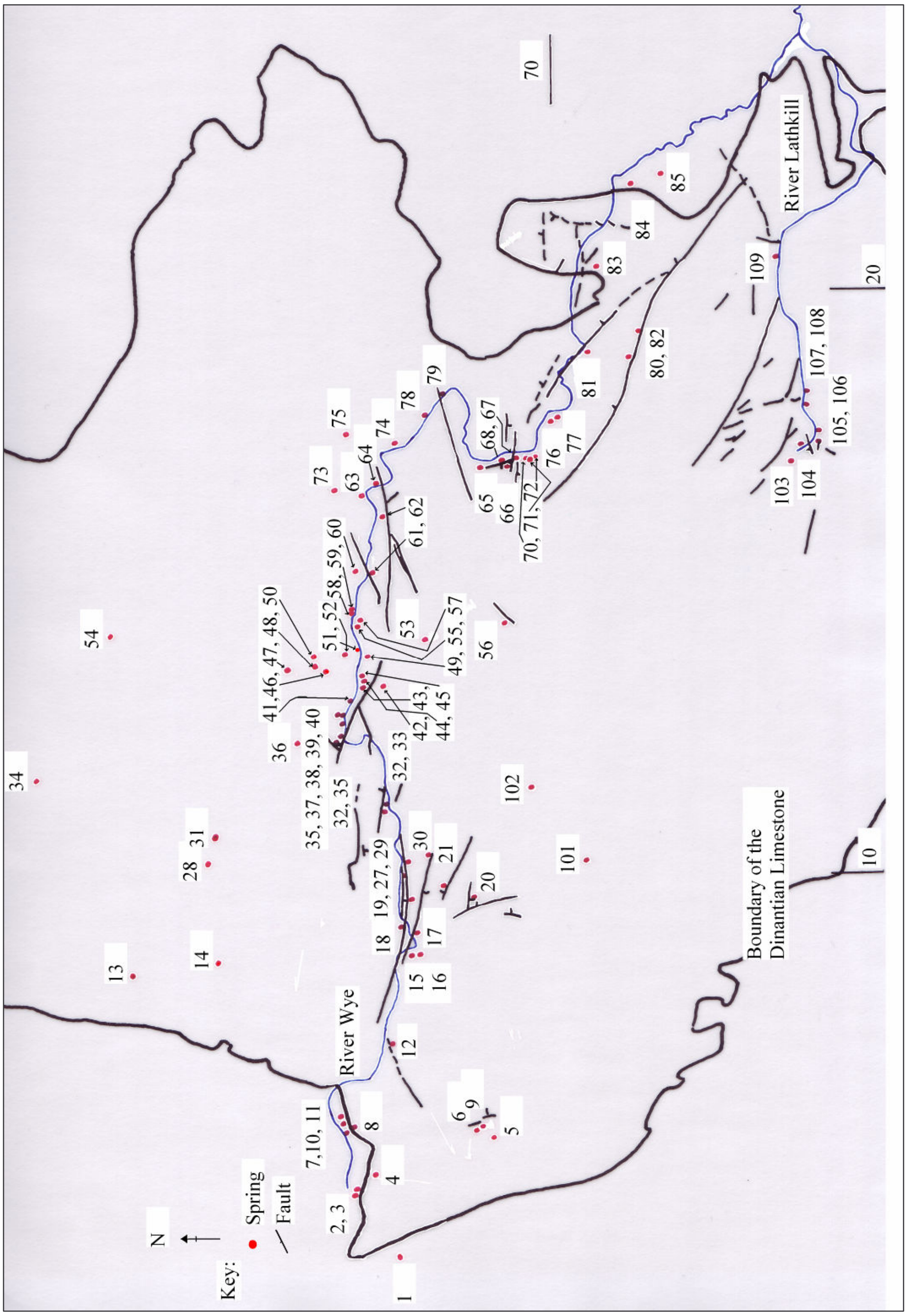


Figure 10.4 Association of springs with faults (red dots represent spring locations, for key see following pages).

Key to Springs

Map Ref.	Spring	Map Ref.	Spring
1	Dog Holes Resurgence	51	Percy's Resurgence Cave
2	Otter Hole	52	Monks Dale 1
3	Golf Ball Resurgence	53	Priestcliffe Spring 1
4	Wye Head Resurgence	54	Brook Head
5	Brook Bottom Resurgence 1		Miller's Dale 9 (control from River)
6	Brook Bottom Resurgence 3	55	Miller's Dale 8
7	Buxton Thermal	56	Taddington High Well
8	Buxton Natural Bath Springs	57	Miller's Dale 7
9	Brook Bottom Resurgence 2	58	Miller's Dale 6
10	Bingham Well (Buxton Baths)	59	Miller's Dale 5
11	Buxton Baths (Overflow)	60	Millers Dale
12	Lovers' Leap	61	Priestcliffe Spring 2
13	Dove Holes Tunnel Springs	62	Litton Mill
14	Barmoor	63	Litton Spring
15	Cowdale	64	Spring and water wheel
16	Kidtor	65	Brushfield Spring
17	Ashwood Dale Resurgence	66	Lees Bottom 2 (upper)
18	Pictor Spring	67	Lees Bottom 2 (lower)
19	Woolow	68	Lees Bottom 1
	Thirst House Cave	69	Lees Bottom 3 , or Lower Dimindale 2
20	Deepdale Resurgence	70	Lees Bottom 4
21	Deepdale Side Resurgence	71	Lees Bottom 5
22	Woo Dale 1	72	Lees Bottom 6
23	Woo Dale 2	73	Ravensdale Cottages
24	Woo Dale 3	74	Cressbrook Mill
25	Woodale 2?	75	Peter's Stone Rising
26	Woodale 1?	76	Great Shacklow 1
27	Topley Pike	77	Great Shacklow 2
28	Bole Hill Spring	78	Hay Dale Spring
29	Topley Pike Outfall	79	White Cliff Spring
30	Deep Dale Lower	80	Dirtlow Farm 1
31	Wormhill Moor	81	Rookery Bridge Spring
32	Black Rock Cottages Spring	82	Dirtlow Farm 2
33	Great Rocks Dale	83	Crowhill Lane
34	Mill Cottage (Dam Dale)	84	Bakewell British Legion
35	Wormhill Springs West	85	Bakewell Recreation Ground
36	Brindley's Well	103	Lathkill Head Cave
37	Wormhill Springs East	104	Holme Grove Risings
38	Cheedale 1 (West)	105	Cales Dale Lower
39	Cheedale 2 (East)	106	Pudding Springs
40	Cheedale Bridge Resurgence	107	Carters Mill Resurgence
41	Chee Dale Lower	108	Inception Horizon-related springs
42	Blackwell Dale Rising	109	Bubble Springs
43	Miller's Dale 1		
44	Miller's Dale 2		
45	Miller's Dale 3		
46	Monks Dale 3		
47	Monks Dale 4		
48	Monks Dale 2		
49	Miller's Dale 4		
50	Monks Dale		

Although the most significant impact on permeability has been the construction of the soughs, it should also be remembered that during the later phases of mining (early 1800s onwards) pumping was carried out and a number of soughs were utilised as pump-way channels. Whilst groundwater levels have largely been restored in these zones, it is considered likely that one of the effects of dewatering would have been to draw through some of the sedimentary fill (loess, residual limestone and glacial deposits), thereby increasing connectivity with the epikarst and further increasing the rate of recharge to soughs. In support of this it is noted that Willies (1980, p. 19) observed of Magpie Mine "*whenever there was heavy rain, or in winter, when the pumps failed to cope, the mine flooded. This in itself was a nuisance, but worse, the water brought down sand and mud into the levels which by 1844 took almost three months to clear,..*". Easier digging was achieved adjacent to the shale margin, therefore it was not unusual for soughs to follow the shale margin e.g. Hillcarr Sough (SK 25846372). More recently these soughs have been found to be more prone to collapse, for example water has backed up by at least 18 m in Hillcarr Sough and is discharging via Shining Sough (Ford personal communication, 2007). Collapse in shale must inevitably result in an upward migration of greater permeability in the cover soils.

Increases in the permeability of the limestone attributable to the soughs are largely focused along the river valleys, with a consequential increase to the 'flashiness' of the river hydrograph, particularly given that the soughs have largely been constructed in the Monsal Dale Limestone (hydrogeological unit 4 of Chapter 9), a unit in which the spring discharges are generally lower and one of the effects of sough construction is to reduce the storage associated with this unit. Furthermore, shortened flow paths divert recharge from other units, reducing the recharge to the main aquifer. Aspects of storage and transmissivity are considered further in section 10.4.

10.4 The effect on storage and transmissivity.

The form of the seasonal recession curves of the rivers must reflect the storage and transmissivity of the limestone. Ineson and Downing (1964) suggested that a river recession curve can be considered in terms of surface run-off, interflow and groundwater discharge (including bank storage). These terms are not directly applicable to a karst aquifer, yet visual examination of the recession components of the spring hydrographs of the Rivers Wye and Lathkill (Figure 10.5) suggests that they are tripartite. It is apparent that there is an immediate response to recharge events, which can be accounted for by the piston effect in karst aquifers (displacement of stored water), followed by two other responses, which must be attributable to the response of the differing components within the aquifer. However, it should be born in mind that the actual water contribution for a given 'event' may take days, weeks (Chapter 8 and Christopher, 1981), or even months and years to reach the point of discharge. Accordingly, it is considered by this author that comparable terms to those of Ineson and Downing (1964), but indicative of a very different aquifer response are required. In examining the literature it is clear that this issue has been addressed in a number of karst models. Indeed, the literature introduces such terms as: baseflow (Atkinson, 1977a) the contribution derived from slower percolation; flow from the zone of dynamic storage (Smart and Hobbs, 1986 and Chapter 4); phreatic and lower unsaturated zone recharge (Trcek and Krothe, 2002), and quick flow (Padilla et al., 1994). The phreatic recharge comprises two components within the context of the White Peak. One is defined by regional flow (referred to in this thesis as underflow) and the other as baseflow (Tóth, 1963). In terms of the phreatic flow to the local flow paths a number of factors must influential, for as water reaches channels from a number of sources (including by-pass flow from the epikarst and superficial deposits) there will be varying degrees of 'rate limiting' (section 7.5), forcing a form of bank storage in the zone of dynamic storage (Smart and Hobbs, 1986), which can only be accommodated within fissures of varying degrees of openness (due to both tectonic setting and dissolutional activity) and thereby limiting the recharge to the channels from the more distal catchment and also recharging the capillary fringe (the finest of the fissures associated with the channels). Worthington et al. (2000) consider storage in terms of matrix, fracture, channel (conduit) storage and permeability (Chapter 8). In reality there is likely to be a broad range in fracture sizes, extending from sub-visible to fracture dimensions. The situation is made more complex when one considers the storage of the epikarst and superficial deposits, including the superficial deposits that fill many of the lower order branches of conduit, which the interpretation of the geochemistry suggests to be important. Furthermore, the terms used by Worthington et al. (2000) are not directly comparable with the baseflow and dynamic storage, because they relate to zones of storage, rather than contributions to yield.

Logic suggests that the response from conduits is likely to be the most clearly defined and recognisable in the recession curve, largely due to the dominance of conduit transmission in terms of volumetric transmission of water (Atkinson, 1977a), yet even this contribution is partially masked by the on-set of discharge from other components of the zone of dynamic discharge (Smart and Hobbs, 1986). Accordingly, in the consideration of the recession curves the terms conduit-dominated flow, fracture-channel flow and baseflow (represented on the recession curve as quick, medium and low discharge conditions) are preferred by this author. It should also be acknowledged that there is considerable overlap between these responses. From the peak of the recession groundwater will discharge both from conduits, with a contribution from fissures, during the next, overlapping, portion of the curve groundwater derived from the zone of dynamic storage will largely be drained, with a contribution from the epikarst and superficial deposits. Because of the relatively low dip of the beds and the considerable influence of bedding-related features on the formation of the channels (conduits); the storage capacity of the clay wayboards; and the confined nature of some of the resurgences a minimal head increase is required for resurgence to occur and it is suspected that there is considerable remnant storage in the channels as well as in the unsaturated zone (including epikarst and superficial deposits, with an increasingly distal component). Drainage of the latter, together with a component of phreatic recharge, forms the baseflow contribution.

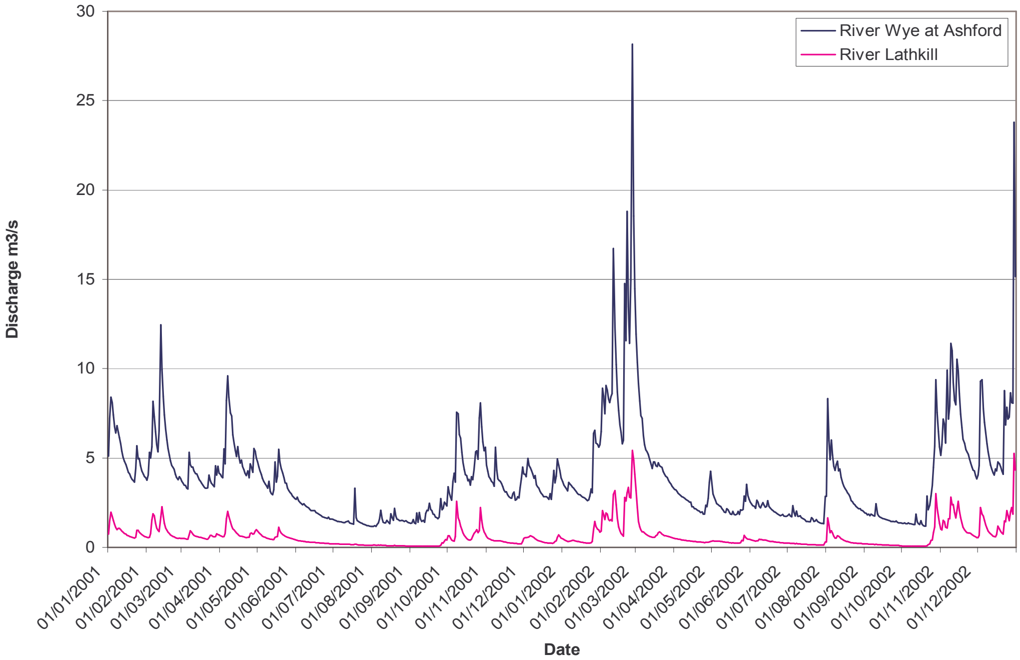


Figure 10.5: Discharge of the Rivers Wye and Lathkill 2001 and 2002 (source Environment Agency, Lathkill data adjusted to the rating curve derived by Gunn, 2002).

Trcek and Krothe (2002) successfully applied three and four component storm hydrograph separation techniques, based on natural tracers and on similar observations with respect to the form of recession curves, for karst aquifers in Indiana and Slovenia. This work also suggests that it should be possible to divide the recession curve into differing discharge components and thus to consider human impacts on storage and transmissivity. Clearly a flashy response with rapid return to baseflow would be indicative of an aquifer of high transmissivity with low storage and therefore the effect of human impacts if they are significant would be the same as those of greater maturity of the karst aquifer, i.e. to increase the 'flashiness' of the hydrograph. It is evident from Figure 10.5 that it is not just catchment size (the topographic catchment upstream of Ashford has been determined as 154 km^2 and that of the Lathkill as 33.48 km^2), which determines the form of the recession curve.

The steep gradient of the rising limb of the groundwater levels monitored in boreholes was noted in Chapter 8 of this thesis. This is also apparent in the river discharges and appears to be more marked in the limestone of the Peak District than in other areas. This contribution comprises the piston effect from rapid recharge to conduits. The rapidity of the response is likely to reflect both the maturity of the karst system and the low hydraulic gradient associated with many of the flow paths, thus it takes minimal recharge to raise the conduit water level sufficiently to generate discharge. Interestingly, the work carried out by Trcek and Krothe (2002) supports this, with phreatic flow forming the primary recession component of the storm recession curves, peaking hours ahead of water containing an epikarstic component. In the opinion of this author it is also in keeping with a contribution influenced by a barometric response, facilitating rapid release of capillary groundwater via dominant fissures and tensional fault zones. Essentially this is an immediate response of the phreatic flow contribution. It is implicit in the flood pulse study carried out by Christopher et al. (1981), when a rise in discharge levels was observed at Russett Well simultaneously with the rainfall events of 27 February 1979 and 6 October 1980.

With respect to the identification of the specific components of the recession curve, there are two commonly used approaches to the determination of baseflow from seasonal recession curves. One is the Meyboom method (Meyboom, 1961), whereby baseflow is derived from the straight line obtained by joining the base of the minor fluctuations on the recession curve represented by the plot of the log of the discharge values against time. Consideration of Figure 10.6 indicates that in the case of the limestone catchment the base of the log-recession curve does not form a straight line (the tripartite form of the recession has been described above), apparently confirming that the relative contributions of the components of the baseflow change with time. The second method, the recession curve displacement method (Rorabaugh, 1964) can be used to calculate the baseflow increase from the upward shift of the recession curve following a significant recharge event, again plotted as log discharge against time. The technique for defining the form of the baseflow curve is to define the duration of the run-off component for the event ($D = A^{0.2}$), where D is the duration in days and A is the basin area in square miles (Fetter, 2001) and to define the commencement of the straight line portion of the curve from $0.2144t_1$, where t_1 is the time taken for the baseflow to recede by 90 % on the straight line portion of the curve.

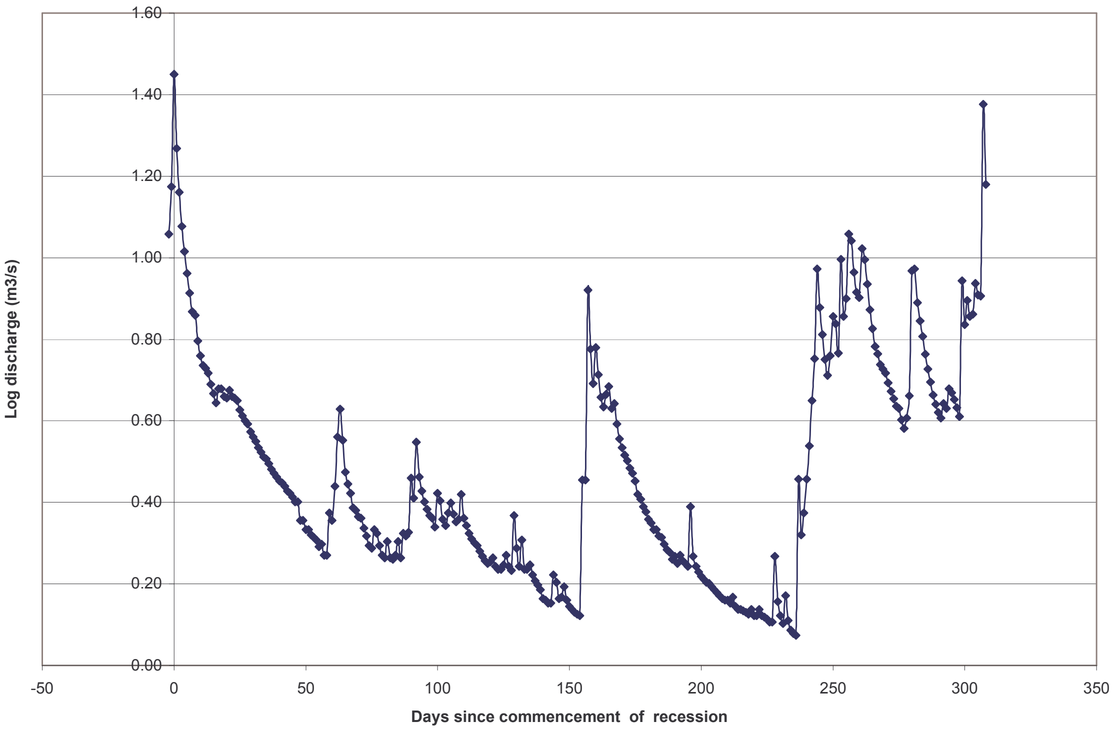


Figure 10.6: River Wye at Ashford: recession and recovery 2002 to 2003 (log discharge against time).

The components of the combined recession curve for the River Wye (Figure 10.7) were defined mathematically from the 2001 and 2002 recession curves (Figure 10.5), using the EXCEL package to determine best-fit curves (Appendix 10.3), in order to generate a theoretical combined recession curve (Figure 10.7). The low discharge component was determined (Appendix 10.3) for the periods 25 May to 27 July 2001 and 2 August to 18 October 2002. This author has observed that the medium discharge component is characterised by a straight line (see also the recession curves of the River Lathkill on Figure 10.5 and also of the River Derwent on Figure 10.9). Accordingly the gradient of the straight-line portion of the recession curve between 26 March and 25 April, 2002 and the conduit-dominated flow was determined from the preceding straight line portion (26 February to 7 March, 2002). The curves and trend lines are presented in Appendix 10.3.

The method of Rorabaugh (1964) has been used to calculate the baseflow contribution. The curve has been further subdivided to assess the contributions from conduit-dominated flow (including surface run-off) and fissure to channel recharge. The results indicate that 61 % of the discharge is derived from baseflow, 34 % from the fissure to channel flow and 5 % conduit recharge. In practice however, the idealised recession curve is never achieved because of intervening recharge events.

Similar analyses have been carried out for the River Lathkill, using Environment Agency data, modified by Professor J Gunn in accordance with the rating curve that he has developed over several years (Gunn, 2002). Comparisons of both sets of values have been made with the data presented by Atkinson (1977a) for Cheddar Spring and the findings have been presented in Table 10.3.

Table 10.3: Comparative recession data.

Catchment	Catchment Area (km^2)	Recession Period	Peak Disharge (m^3/s)	Recession Gradients			Base Discharge (m^3/s) & ($m^3/s/km^2$)
				Conduit recharge	Fissure-channel flow	Base flow (ln)	
River Wye at Ashford	154	February to July 2002	28.161	-0.8966	-0.0654	-0.018t	1.324 (0.0086)
River Lathkill at SK 19806610	33.48	February to July 2002	6.242	-0.8259	-0.0111	-0.018t	0.115 (0.0034)
Cheddar Spring (Atkinson, 1977)	39.44	February to August 1970	2.76	-0.6172*	-0.0698 *	-0.027t	0.1 (0.0025)

* calculated by this author from the straight line portion of the curves presented by Atkinson (1977a)

The gradients for base flows are steeper for Cheddar Spring, thus implying more mature phreatic conditions with lower storage. The marginally steeper gradient in the zone of fissure to channel flow also suggests the Cheddar catchment to be a more mature karst system. However, the lower gradient of the conduit recharge for the Cheddar system suggests bank storage within the conduit system, which again is likely to reflect the maturity of the conduit development, probably at least in part reflecting the greater number of siphons associated with the more steeply dipping strata. The steepness of the higher flow rate portion of the curve for the Wye at Ashford may be attributable to the impact of sough construction. Additionally, it may in part, comprise an immediate barometric response, combined with the piston effect of the displacement of water, largely stored in fissures and channels, for example associated with the clay wayboards, together with a matrix contribution, resurging via faults.

Using another method of assessment, storage has been considered in terms of the discharge of the River Wye at Ashford during two periods of baseflow: 17 May to 5 October, 2001 and 19 March to 21 October, 2002. The mean effective rainfall (0.4 mm) equates to a recharge of 2 litres/sec, whilst the mean discharge recorded at Ashford for the same period was 2190 litres/sec. If 400 litres/sec of this is derived from Magpie Sough and 190 litres/sec is derived from underflow in the order of 1598 litres/sec (73 %) of the discharge must be derived from storage in the limestone (a combination of base and fissure to channel flow).

It has been more conventional to consider recession curves in karst as comprising two components, termed quick and baseflow (Atkinson, 1977a). As described in Chapter 8 of this thesis, Padilla et al. (1994) described two methods for investigating the relative importance of baseflow and quick flow in karst springs. To allow comparison with data obtained from borehole hydrographs, these methods have been extended to the recession portion of the hydrograph for the River Wye at Ashford for the period August to October 2002 and the following parameters have been derived (Appendix 10.3):

Mangin's Analysis – $\eta = 0.083$; $\varepsilon = 0.023$; ti = 12; $q_o = 5.89\ m^3/s$; and $q_ob = 3.55\ m^3/s$

Coutagne Analysis – $\alpha_0 = 0.0156$; n = 1.50

The results of the Coutagne analysis suggest that the aquifer feeding the River Wye lies somewhere between a thick aquifer that drains at a constant discharge and an aquifer that discharges under a layered regime. The α value is generally lower and the n value higher than values determined for the boreholes (Chapter 8). This is attributable to the larger contribution of conduit flow in the catchment scale analysis. The differences justify the differing methods of analysis of the recession curves. The term ε of Mangin's analysis indicates the concavity of the quick flow, the value increasing with the degree of concavity. In the unconfined laminar flow conditions of granular aquifers, the concave form of the recession curve would be indicative of the contribution from proximal recharge (Ineson and Downing, 1964). In the Wye catchment the lesser form of the concavity reflects rate limiting factors (section 7.5) and the moderating influence of the fissure to channel flow contribution; the contribution of storage from the epikarst, and superficial deposits; and also the storage function of the clay wayboards, which have an influence in moderating the quick flow by slower circulation.

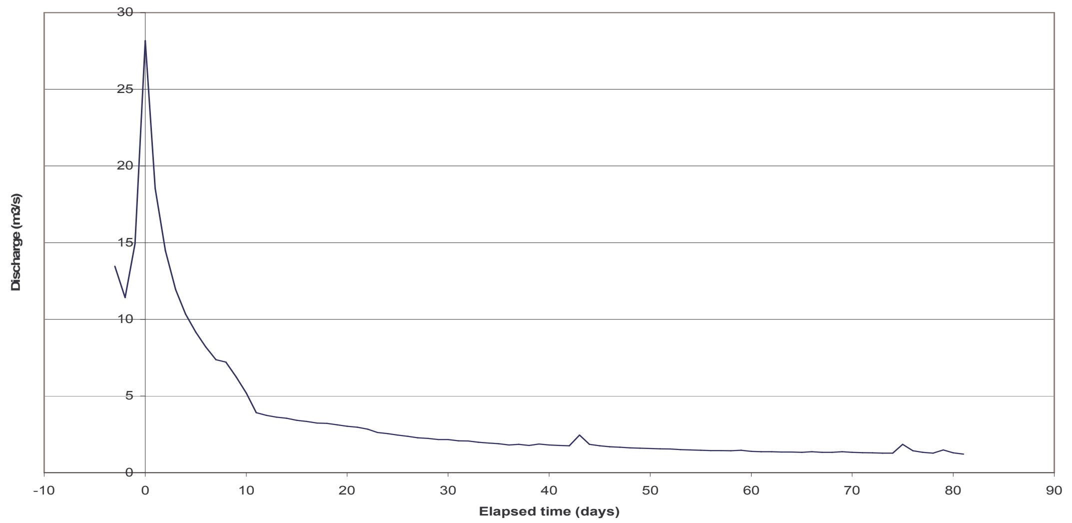


Figure 10.7: Theoretical combined recession curve for the River Wye at Ashford.

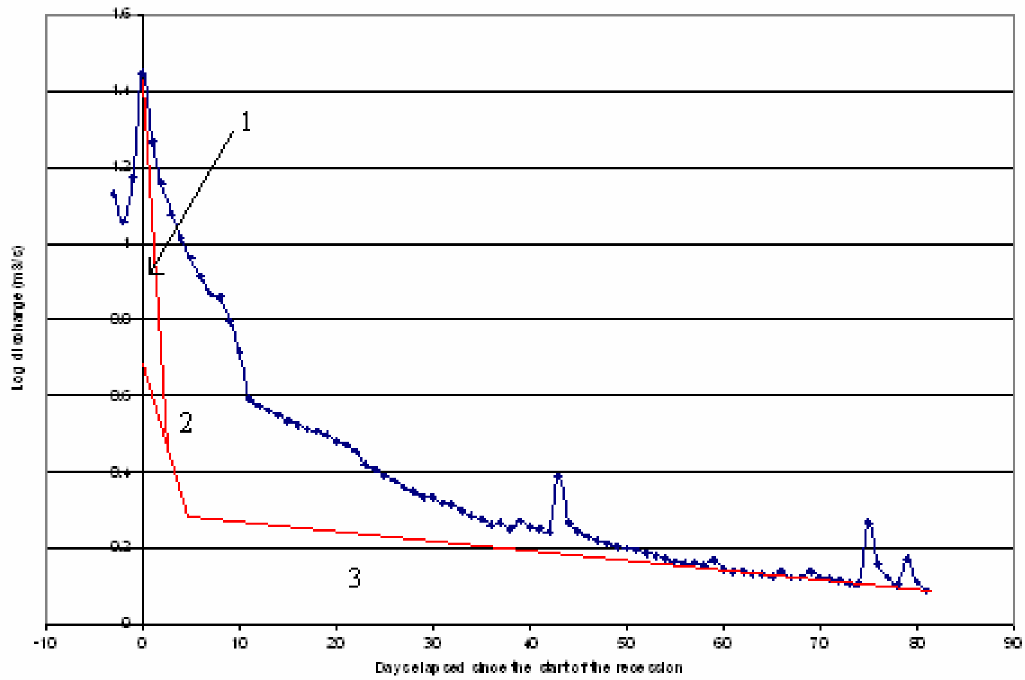


Figure 10.8: Theoretical combined recession curve plotted as log discharge against days elapsed since the start of the recession and annotated to show the contributions of conduit-dominated (1), fissure to channel (2) and base flow (3)

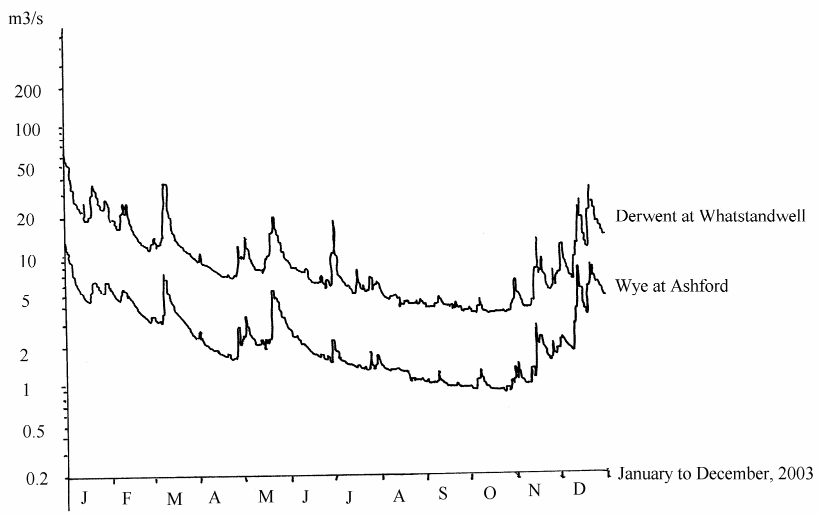


Figure 10.9: River hydrographs for the Wye and Derwent, 2003 (source Environment Agency web site).

This is in keeping with the observations that have been made above regarding transmissivity and possibly reflects the storage function of the clay wayboards and stylolites (Chapter 9). The conceptual model for this limestone terrain also incorporates a barometric influence and siphoning, which contribute to an increase in the concavity of the form of the recession curve.

Without data on the pre-mining hydrogeology it is difficult to quantify, or even qualitatively assess human impacts on the recession curve. However, the analyses suggest that the influences are only locally significant. As suggested above, it is anticipated that the human impacts of mining (primarily sough construction) would be represented by a form similar to that of a mature karst aquifer with well developed drainage via a dendritic conduit network, which is more effective in draining the fissure to channel flow and thus results in a more concave form to the recession curve, with a far greater proportion of the discharge being derived from quick flow. Indeed, it is the significant potential for the storage associated with the fault/ mineral vein zone areas that made the dewatering of the mines so difficult during mining, hence the need to construct so many individual vein soughs in the area of Alport.

A regional reduction in head was only achieved by the larger soughs, such as Magpie Sough and Hillcarr Sough, which targeted groundwater divides (the Taddington Anticline and Long Rake). The localised lowering of the water table was associated with an inevitable reduction in storage in the zone of water table lowering, but off-set against with this; there have been localised increases in specific yield, resulting from the additional storage that is offered by the void created in the mineral veins. Locally the storage function of the epikarst has been reduced as the permeability of the limestone has been increased with high level soughs diverting groundwater via vadose shafts from storage in the epikarst, to base level, or else to streams. Whilst there is substantial evidence to suggest that there were extensive phreatic cavities associated with many of the mineral veins (Ford, 1989) it is inevitable that the exploitation of mineral has resulted in additional storage, predominantly in the unsaturated zone, but in those areas that were formerly dewatered by pumping, also in the saturated zone. Locally the consequence of an increase in storage is a reduction in head, albeit that this is difficult to quantify.

For the area of the White Peak as a whole, Ford and Rieuwerts (2000, p. 10) report that between 3 and 6 million tons (2.7 to 5.5 million tonnes) of lead ore concentrates have been raised since mining began and that zinc ore production for the same period was in the order of 0.25 to 0.5 million tons (0.23 to 0.45 million tonnes). Although up to 90 % of this was originally considered waste, much of it has since been reworked for gangue minerals. Many of the workings were backfilled with arisings. If it is assumed that the workings were backfilled with arisings placed with sufficient compaction to achieve a porosity of 40 %, the resulting additional storage would amount to in the order of between 65,000 and 133,000 m^3 (calculated assuming a mean density of 45 kg/m^3 for the extracted concentrates). Much of the additional storage would be above the local water table. Realistically it is unlikely that the postulated degree of compaction could be achieved, as it is known that many of the workings were below the water table and were carried out by stoping and were abandoned as they stood, thus leaving a far greater porosity.

By contrast, quarrying, which has largely affected the epikarst and the unsaturated zone, is likely to have resulted in a loss of storage associated with in of the order of 1030 million tonnes (38,100 million m^3 of limestone between 1895 and 2001 and quarrying activities are on-going (Chapter 3). In addition, Ford and Rieuwerts (2000, p. 12) report that around 80,000 tons of fluorite (72,575 tonnes, or in the order of 2 million m^3) have been produced annually in the recent past, mainly from two sites operated by Laporte Industries plc on Longstone Edge and at Hucklow and that baryte has yielded 40,000 tons (36,280 tonnes, or 0.8 million m^3) per year in recent years. The fluorite and barite exploitation is largely carried out from surface from series of trenches, which are backfilled upon completion.

The quarrying has largely been carried out above the water table and, as described by Hobbs and Gunn (1998), such exploitation is likely to result in localised impacts, additional to the changes in storage such as:

- loss of topsoil, superficial deposits and epikarst storage resulting in direct infiltration to lower levels in the unsaturated zone, for example surface water reaching the quarry floor at Topley Pike Quarry targets a specific fault which is exposed in the quarry floor;
- reduced availability of carbon dioxide for dissolutional processes as a consequence of the removal of topsoil;

- loss of conduits in the unsaturated zone, e.g. Alsop's Cave, Castleton (Gunn personal communication, 2001);
- once a void has been opened it will potentially become a target for vadose flow, which may result in sediment being washed out of conduits connected to the quarry, or in accelerated evolution of dolines, and
- blasting generates zones of higher permeability, furthermore, zones of higher permeability are formed around the quarry as a response to stress relief, due to the relief of both vertical (overburden) and horizontal forces.

Further to the loss of epikarst storage, quarrying results in a number of definitive anthropogenic geomorphological features (Gunn and Gagen, 1989).

As noted in Chapter 3, the human impacts are clearly linked to the geology. For instance, the lead-zinc mining activities primarily correspond with the outcrop of the Monsal Dale Limestone Formation. These are the areas that are dominated by perched groundwater conditions and the implications are considered in section 10.5. Limestone production (Figure 3.3) is largely focused on the Bee Low Limestone and the Woo Dale Limestone formations. In the case of the Woo Dale Limestone Formation it is clear from the conditions at Topley Pike that the open excavation and proximity to the 'water table' encourages direct recharge to fault zones. Not only does this impact on the hydrology, reducing epikarst storage, but it also has the potential to impact on water quality, for example Christopher (1981) noted that turbidity at Wormhill Springs was related to quarrying activity. Quarrying of the Chee Tor Limestone exerts an even greater impact on the epikarst/ vadose zone storage. It is clear from the descriptions of the Chee Tor Limestone Member that the fissuring that has opened as a consequence of stress relief offers the potential for significant vadose zone storage. Evidence for this comes from the Staden Borehole (section 8.6), which appears to derive groundwater from an inception horizon corresponding with the boundary between the Bee Low Limestone Formation and the underlying Woo Dale Limestone Formation, which, at this location, would appear to be fed in part by storage from the Chee Tor Limestone Member, partly as a consequence of artificial development of connections by blasting.

Where mining has been carried out in zones adjacent to rivers, as in Bradford Dale and in Lathkill Dale in particular, but also along the River Wye in the areas of Miller's Dale and Lees Bottom, it might be anticipated that the presence of mine workings would provide an additional contribution to the zone of dynamic storage, which can be likened to seasonal bank storage. However, there is no evidence for this in the discharge data. Usually the recovery from the seasonal groundwater recession is very rapid, but during the period 5 August 2001 to 6 October 2001 the rising limb for the River Ashford was more gradual and unusually, took place over a period of weeks. Interestingly the calculated effective rainfall for the period, up to 4 October 2001 was 0, but there was precipitation during this period and the evidence suggests that some of this rainfall was effective (indicating an overestimation of the soil moisture deficit). The response of the River Lathkill for this period was very different, with no recovery occurring until 28 September 2001. Whilst this could potentially provide evidence of bank storage, it is considered more likely that it reflects the fact that the base of the groundwater recession, in the area of the River Lathkill, falls below the bed level of the River Lathkill. In considering this data it is also worth noting that the recession bases of the Lathkill data are flat, reflecting the perched nature of the bed level above summer groundwater baseflow. Furthermore, during a visit to undertake a geophysical survey of Lathkill Dale Gunn and Dykes (2000) witnessed the rapid recovery of the flow in the River Lathkill as a wetting front with a mean velocity of 0.042 m/s (Gunn, 2000a), with a source at Holme Grove Risings (196.5 m OD).

Magpie Sough requires separate consideration. There has been limited monitoring of the discharge from Magpie Sough. The values that have been obtained (Table 10.4) would suggest that the discharge appears to have remained relatively constant since the time at which it was cut (apart from periods of mine dewatering and of collapse, e.g. 1962 collapse followed by an 'explosive' release of pressure in 1966 and controlled unblocking of another collapse in 1974, [Willies, 1980]), with some seasonal variation, as indicated by the low flow data presented by Dixon (1996). This suggests that a large storage zone has been tapped, which seems unaffected by the construction of the sough. The constant rate of discharge indicates a constant head, with no evidence to suggest a long term lowering. The groundwater chemistry of the discharge from the sough suggests predominantly autogenic recharge. Therefore, it seems most likely that the head is sustained by a combination of a small component of underflow, as indicated by the groundwater contours presented on Figure 9.3, which indicates that there is a ridge in the groundwater contours associated with the Taddington Anticline and a larger component of deeply circulating surface water recharge.

In a situation where the head in an aquifer is reduced above a confining layer, the excess pore pressure below the confining layer is dissipated by transient flow. This is a process which is central to the evolution of karst aquifers where conduits become the target for groundwater flow. Furthermore the stress relief associated with the creation of an opening must also increase the permeability associated with that opening. Therefore, where the soughs form large openings and impact significantly on the local 'water table' there is a potential for them to form the focus for confined water. Thus, if the limestone by virtue of its low matrix permeability, confines the underflow, the sough becomes a focus for underflow. As described in Chapter 9 it is considered by this author that dominant tensional faults are zones of maximum vertical groundwater movement and it is suspected that prior to human impacts the high head in these zones was maintained by underflow. Further evidence in support of Magpie Sough having captured underflow comes from strontium isotope evidence presented by Gunn et al. (2006). The underflow that has been captured by the soughs must have reduced the component of underflow continuing down hydraulic gradient to the River Derwent and possibly even outside the area of the limestone outcrop. The consequence of additional groundwater mixing in these zones is likely to be an increase in rates of phreatic dissolution, thereby further increasing their storage potential. The underflow that is captured by Magpie Sough is considered further in section 10.6, below.

Table 10.4: Discharge data Magpie Sough (SK 179696).

Date:	Discharge (litres/sec):	Source of Data:
1929	422-502	Naylor, 1983
1967	396	Naylor, 1983
26.10.67	400	Edmunds, 1971
Oct 73 -Nov 78	409	Christopher, 1981
27.9.89	343	Dixon, 1996
17.10.89	352	Dixon, 1996
26.7.90	345	Dixon, 1996
18.9.91	434	Dixon, 1996

10.5 The modification of groundwater divides.

It is well documented that groundwater divides do not always coincide with surface water divides (Christopher et al., 1977). In the context of the White Peak this can be attributed to the way in which dominant tensional faults form the target for deep, confined, groundwater circulation and therefore form the groundwater divides between groundwater catchments. The guiding head is suspected by this author to be that of the Todd Brook Anticline and similar anticlines to the south. It is also suspected by this author that flow paths within groundwater divides vary between one hydrogeological unit and another. In particular the significant guidance of bedding-related inception in Unit 4 (Monsal Dale Limestone and Miller's Dale Limestone formations) is likely to guide groundwater in a more easterly direction than the fissure-related guidance of the Chee Tor Limestone Member. It would seem very likely that this variation has been enhanced by epikarst development during the Pleistocene. The constructional endeavours of the miners were also of a scale sufficient to alter, or even breach groundwater divides.

The conceptual model presented in Chapter 9 indicates that major faults and mineral veins form dominant zones of vertical groundwater movement. More specifically, it has been observed that if a fault is downthrown in the direction of the hydraulic gradient this is commonly the focus for upward groundwater movement, for example, the Arrock Fault of the Taddington Anticline (Figure 2.1) and Long Rake in Lathkill Dale. This is also seen in the area to the east of Chelmorton. Hardyhead Sough (Figure 10.3) has been found to discharge water throughout the year at a level of approximately 370m OD. The sough follows the line of a southwesterly-trending fault immediately to the north of Groove Rake, which is downthrown to the southeast. By contrast, Illy Willy Water rises from above the down-faulted bed of Upper Miller's Dale Lava, immediately to the north of Groove Rake towards its western end and then immediately sinks along the line of Groove Rake and, as described in Chapter 7, it is suspected that the sinking groundwater reaches a level of approximately 255 m OD in very close proximity to this fault.

Downing et al. (1970, p. 59), describing the discrepancy between surface and groundwater divides, suggested "*that the divide between the Wye and Lathkill appears to be south of the surface water divide between Great Low and Bole Hill (Bakewell).*" This is attributable to the influence of Magpie Sough.

Of the major soughs in the area of this research it is only Magpie Sough which actually fully breaches a groundwater divide. It is reported (Willies, 1980) that the pre-existing, high groundwater level in Magpie Mine was decreased by in the order of 50 m (to approximately 168 m OD) as a consequence of Magpie Sough breaching Townhead Vein. This would suggest that the Taddington Anticline was once a permanent groundwater divide, which has become a seasonal divide as a response to the cutting of Magpie Sough.

As reported by Raffety et al. (1953), significant modification to a groundwater divide has occurred at Dove Holes, where the construction of a railway tunnel with a gradient to the west intersected springs with southeasterly flow paths. This was described in a letter from RWS Thompson to Professor Shotton, dated 31 October 1950 (DRO D3040 L/W 1/44 92-94) "*The transition from limestone to shale occurs 1200 yards (1097 m) north of the portal of the Dove Holes tunnel, where the rail level is 279.5 m OD and the tunnel gradient is 1: 90.5 with a total length of 2983 yards (2728 m). There are a number of issues between the boundary and the Dove Holes end of the tunnel. The largest quantity was through fissures into a small transverse heading near the point of junction. There is a large culvert under the 6' way in this tunnel and the water which flows out of it at the northerly end flows down an open channel at the side of the open cutting. There is a place where a venturi tube gauge could be made quite easily. At times very large volumes of water flow down this culvert and even down the interior of the tunnel itself.*" These springs were seen by Ford (personal communication, 2007) in the early 1950s. The southeasterly flow of groundwater was also described by Hudson (1989), noting that prior to tunnel construction the presence of a swallow hole and disappearing brook, flowing underground to the southeast, at the southern end of the proposed tunnel alignment alerted the Contractor to the potential construction difficulties and associated costs, which resulted in the Midland Railway carrying out their own tunnel construction. Attempts were made to divert the stream and prevent it flowing underground, a channel being cut from the vicinity of the swallow hole towards Great Rocks Dale, the water sank in the order of 800 m to the south of the tunnel. It is reported that in the order of 6 months the water reappeared at the same point, having exceeded the storage capacity of the limestone and resumed its course along the artificial channel before sinking along a fissure near the station. This has also been confirmed by the dye-tracing carried out by the Limestone Research Group (Hardwick, 1996c and Chapter 7). Downing et al. (1970) reported that the Dove Holes railway tunnel discharges in the order of 2400 litres/sec, however this included water derived from the Millstone Grit sandstones.

10.6 Underflow captured by Magpie Sough.

Worthington (1991) has shown that there is an increase in thermal output at lower elevation among the eastern group of thermal springs that fringe the outcrop of the limestone and cites this as evidence for regional flow to the southeast. This author has tabulated the temperature and discharge values for each of the thermal outputs, recorded by Edmunds (1971) and Worthington (1991), added Bubble Springs to the list and recalculated the thermal flux for each output [thermal output (kW) = 4.2 x temperature difference between spring and background x discharge] and then by changing the subject of the formula has used the thermal flux to calculate the underflow component of each of the outputs (Table 10.5). In order to assess the underflow component it has been necessary to make assumptions about the source temperature of the thermal water. The source temperatures have been calculated by adopting a geothermal gradient for the limestone of 15° C/km and taking a depth to basement assessed from Gutteridge (1987) and Fraser and Gawthorpe (2003). Thus the assumption is that the thermal water is derived from basement level, clearly however, the actual depth of the flow paths is unknown, although the heat source is considered to be the basement rocks.

Table 10.5: Underflow component calculated for a number of sources of thermal water.

Source	Temperature (° C)	Discharge (l/s)	Heat Flux (J/s)	Deep Source temperature (°C)	Underflow (litres/sec)
Matlock Fountain Bath	19.7	11.8	580	40	6.8
Matlock East Bank Rising	17.4	0.5	20	40	0.2
Matlock New Bath Hotel	19.8	6.3	320	40	3.8
Ridgeway Sough	14.1	4	110	40	1
Meerbrook Sough	15.3	895	28570	40	275.4
Bath Sough	19.7	76	3830	40	44.9
Ball Eye Quarry Borehole	13.6	1.5	35	40	0.3
Lees Bottom 3	11.5	4.2	65	30	0.8
Magpie Sough	9	400	2185	40	16.8
Bradwell Spring	12.4	0.7	15	42	0.1
Stoney Middleton	17.7	1.3	55	42	0.5
Stoke Sough	11.6	1.3	20	42	0.3
Bakewell British Legion	11.6	9.3	150	30	0.4
Bakewell Recreation Ground	13.3	0.2	5	30	0.1
Bubble Springs	8.2	1000	2100	25	29.8

Using the same method the underflow discharge calculated for Buxton was 22 litres/sec. In this case the source temperature presented by Barker et al. (2000) was used. It is interesting to compare these data with the results of calculations presented in subsection 10.1 of this chapter which suggests an underflow of 190 litres/sec upstream of Ashford. The underflow from Lees Bottom 3, Buxton and Magpie Sough calculated above is 40 litres/sec. This suggests that there is further 150 litres/sec being discharged upstream of Ashford. It is likely that a large contribution is made via fault zones into the bed of the River Wye. Furthermore, although spring temperature can be influenced by a number of factors including aspect, vegetation cover and time of day, the variation in temperature of the springs, as recorded on 29 to 30 April 2002 and presented as Table 10.6, below indicate that other springs may also have an underflow contribution, more specifically, Cowdale (Rockhead) Spring and Holme Grove Risings (elevated temperatures recorded early in the day) and possibly others.

The data also indicate that Hillcarr Sough may capture a small component of underflow. It is reported (DRO D359Z/239(9) that when the Mawstone Mining Company cut into Timperley Vein (in the 1880s), a northeast to southwest orientated vein, at a level of about 104 to 107 m OD, it trapped a flow of water of 1500 gallons per minute (114 litres/sec) which continued to flow constantly in 1930. In addition, Kirkham (1960-1) described the groundwater problems associated with the lava horizons in Guys Engine Shaft (SK 21916382). It would appear that groundwater, increasingly confined by the

Falgate Formation in a southeasterly direction, targeted dominant faults and jointing in the Falgate Formation and it is this that caused the considerable problems dewatering this area of the mine field. Rieuwerts (1981) reported that Hillcarr Sough encountered twelve springs in a length of approximately 60 m, close to the point at which the later Thornhill Sough branches out of the main sough at SK 23956319 (Appendix 3.4 and Figure A3.4.9). The northern six brought immediate relief to the Stoney Lee Mines (SK 23126364) and Kirkham (1960-1) reported that longer term lowering of groundwater level occurred to the southeast; the lodge of the engine at Plackett Mine, Winster (SK 238611) was drained dry, and there was considerable lowering of water at Yatestoop Mine, Winster (SK 24356165), respectively a 2400 m and 1600 m distant. This suggests that in this instance the Sough encountered southeasterly dipping inception horizon-related flow paths.

Table 10.6: Temperature data for springs within the study area.

Site	Temperature (˚C)	Date and time
Rockhead (Cowdale) Spring	8.8	29-04-02/1112
Ashwood Dale	8.3	29-04-02/1130
Woolow Spring	8.1	29-04-02/1148
Topley Pike Spring	8.1	29-04-02/1206
Topley Pike Outfall	8.2	29-04-02/1258
Deep Dale (upper)	7.7	29-04-02/1327
Deep Dale (lower)	8.4	29-04-02/1337
Hardyhead Sough	7.8	30-04-02/1148
Blackwell Dale	8.5	29-04-02/1416
Maury Sough	8.3	29-04-02/1509
Miller's Dale 4	8.8	29-04-02/1443
Miller's Dale 7	8.0	29-04-02/1624
Miller's Dale 8	8.0	29-04-02/1634
Litton Spring	8.4	29-04-02/1534
Litton Mill	8.1	29-04-02/1545
Lees Bottom 1	9.8	30-04-02/1331
Lees Bottom 2 (upper)	10.7	30-04-02/1338
Lees Bottom 2 (lower)	9.9	30-04-02/1344
Lees Bottom 3	11.9	30-04-02/1353
Lees Bottom 4	Dry	30-04-02
Lees Bottom 5	9.0	30-04-02/1500
Lees Bottom 6	9.0	30-04-02/1500
Great Shacklow 1	9.5	30-04-02/1226
Great Shacklow 2	8.9	30-04-02/1240
Magpie Sough	9.9	30-04-02/1254
Holme Grove Risings	9.0	30-04-02/1040
Mandale Sough	9.6	29-04-02/1738
Lathkill Dale Sough	8.7	29-04-02/1801
Bubble Springs	9.3	29-04-02/1804
Hillcarr Sough	9.2	30-04-02/0930
River Wye	9.3	30-04-02/1600
Illy Willy Water	8.4	30-04-02/1600

In this context, although the sough construction has the potential to induce transient flow, and indeed Magpie and Meerbrook Soughs have captured underflow, it is not clear whether the underflow that reaches the soughs is actually attributable to transient flow. If it is, then the maximum increase in underflow upstream of Ashford attributable to human impacts is < 9 %.

10.7 Impact on groundwater quality.

It is commonly argued that an understanding of karst aquifers is essential because of the way in which very rapid conduit flow, which is a characteristic component of karst aquifers can transport water borne contaminants very rapidly, over long distances, with minimal attenuation (Hobbs and Gunn (1998), Smart and Hobbs (1986) and Trcek and Krothe (2002)). This problem can be exacerbated by quarrying, as the removal of overburden, including topsoil, removes what there may be in the way of attenuation and facilitates surface water infiltration to lower levels in the unsaturated zone, thus, with the potential to convey point source (such as hydrocarbons) and dispersed contamination (e.g. fertilizers, herbicides and pesticides) directly to the limestone aquifer. It might be argued that perhaps there is a greater potential for attenuation in the karst of the Peak District, where the clays associated with the clay wayboards must exhibit a higher cation exchange potential, than that of other karst regions. In Derbyshire point sources of pollution primarily comprise wastes associated with agriculture and landfill. However, there has been a long-standing problem of sewage disposal, as reported to the local Government Board on the sanitary circumstances and administration of the Bakewell Rural District (Newholme, 1910). Although, effective management of sewage disposal has improved over the years, remnant problems have been identified, for example a flow path between Knotlow Mine and Flagg Sewage Treatment Works was established following reports of 'bad air' in Knotlow Cavern and Hillocks mine. Subsequent remedial action was taken (Gunn, personal communication, 2005). However, the problems of 'bad air' continue at Knotlow and intermittent discharges of silage type effluents to the epikarst (dolines are a particularly common target for such activities), or to mine workings that are suspected and research is on-going (Gunn, 1999). Agricultural practices such as these are difficult to police and can have a significant impact on water quality. One of the water bailiffs has reported (personal communication, 2002) similar discharges of farm effluent to Blackwell Dale, which appears to have contaminated an area of tufa deposition towards the bottom of the dale.

Clearly, the response of mine workings, or soughs is similar to that of conduits. Indeed it can be argued that flow rates associated with soughs are likely to be higher, for a number of reasons including: greater hydraulic gradient, size, straighter flow path, and smoother conduit surfaces. Therefore, there is even less potential for attenuation of point source pollution where it is associated with the mine workings.

With respect to landfill operations on the limestone, Taddington (Calton Hill) landfill site is located at approximately NGR SK 117571575 at a level of 380 m to 405 m above Ordnance Datum. The site was originally operated as a dolerite quarry until the 1960s. Landfilling with waste commenced in the late 1960s, the site being the main household waste disposal site for north Derbyshire until tipping was suspended in 1989. During the period of closure, a landfill at Victory Quarry, Dove Holes was completed. The Taddington site was reopened in 1995 and was engineered. Tipping continued until December 2003. It is currently being restored. Concerns with respect to leachate generation have been addressed by installing a full leachate collection system (High Peak Planning Application NP/DDD/0803/432, MIN.3910, 15.08.03, 1188 7127/JM), leachate being taken off site by tanker. Inert disposal sites within the area include Arbor Low, near Youlgreave, where a former calcite mine is being infilled with works waste.

Another aspect of karst geomorphology that is potentially affected by groundwater quality and human impacts is tufa. It has been observed by this author that tufa deposits are associated with Hydrogeological Unit 4 (the Monsal Dale Limestone and the Miller's Dale Limestone). It would appear that in these limestones high levels of saturation with respect to calcium are achieved, which is thought to be related to the development of inception horizons (Chapter 4). However, current rates of tufa deposition are considerably reduced. Lorah and Herman (1988) and Viles and Pentecost (1999) have demonstrated that the distribution of tufa is also related to zones of more rapid water movement and to favourable sites for organic degassing. Pedley (1990 and 1993) has focused more on the biological importance of aspect, particularly with respect to the Wye and Lathkill Dale barrage tufas. There has been some discussion in the literature about the decline in the rates of tufa deposition since the Holocene (Goudie et al., 1993). This has been attributed to climatic change, tufa deposition being associated with periods of warmer, wetter weather and possibly also to human impacts, for instance Pedley (1993) has noted that tufa deposition can be impeded by increases in phosphates in the groundwater. At many localities, notably along the Wye Valley it would appear that deposition has ceased. This is considered further in section 10.8.

Detailed consideration of the impact of the mineralization and its exploitation has not been covered within the context of this thesis. Lead concentrations in the range 0 to 61 μg/litre were determined for spring samples obtained during the Illy Willy Water dye-tracing work (Appendix 6.3). Zinc concentrations were found to be in the range 1 to 78 μg/litre in the same samples. These concentrations indicate that there has been a measurable impact. However, Edmunds (1971) suggested that elevated lead concentrations did not correlate directly with areas of known workings and that although the zinc concentrations showed better correlation with areas of known workings, they were prone to contamination by weathering of metal artefacts. This is another area that calls for further research.

Although there are local problems and a potential for significant pollution, by virtue of the nature of karst aquifers, human impacts on limestone groundwater quality in the White Peak are minimal by comparison with other parts of the UK, particularly for the chalk aquifers. This is attributable to the low density of population and dearth of industry in the area.

10.8 Climatic Change.

Consideration of human impacts would seem incomplete without reference to current concerns regarding 'global warming'. Viles (2003) presented a simplified conceptual model of the impact of climate change on karst geomorphology in the UK and Ireland, making reference to five case studies, including the Yorkshire Dales. Climatic predictions for the Yorkshire Dales for the 2020s and 2050s were obtained from models run at the University of Oxford, which indicate a likely rise in mean annual temperature of between 0.5 and 2° C and an increase in annual mean effective rainfall of between 16 and 41 mm, with summer effective rainfall declining by 3 to 29 mm and winter effective rainfall increasing by 13 to 34 mm. The model predicts an increase in mean annual dissolution rate of between 2 and 5 %, which Viles (2003, p. 65) suggested "*should have only minor impacts on dissolution rates and tufa production*".

In the context of the Peak District, assuming the climatic change to be of a similar order of magnitude, the impacts may be more significant. It is generally considered (Prof. Colin Thorne, University of Nottingham, personal communication, 2004) that rainfall events will become increasingly 'stormy'. Even in limestone terrain this will result in an increase in surface run-off during rainfall events and the quick flow component of the hydrograph will be exaggerated, thereby increasing the loading on the Rivers Wye, Lathkill and Bradford, with the associated increased potential for flooding downstream. The Environment Agency has been responding to such forecasts by preparing plans for flood protection works along the River Wye (Bakewell, Ashford and Matlock Bath). The rapid response of the limestone to rainfall events is reflected in the design of protective, rather than storage flood protection measures. Thus the improvements are focused on improvements to flood walls and embankments. Flood storage improvement schemes rely on the retention of portions of flood water to control loading rates on the river. In the context of the White Peak this could only be achieved by controlling inputs from specific catchments, for example the River Lathkill catchment.

Associated with the increase in winter precipitation and its intensity, there would be a potential for increased dissolution, both as a consequence of allogenic and autogenic recharge, particularly at surface, possibly associated with minor rejuvenation of the epikarst. The increased temperatures have the potential to increase the duration of growing seasons in the White Peak and therefore increase potential production of carbon dioxide. Such an effect could be enhanced where the increasing 'flashiness' of events, together with the lower effective rainfall and therefore potential for desiccation, results in localised erosion of soil, residue and loess. Such a response could affect groundwater quality with an increase in localised turbidity resulting from storm events. A reduction in effective summer rainfall would also further concentrate ions in the unsaturated zone, causing greater seasonality of groundwater chemistry and would be likely to exaggerate the low flows recorded in the Rivers Wye, Bradford and Lathkill. However, off-set against this, Raper (1998) noted a tentative relationship between the bulk deposition of sulphate and limestone weathering in allogenic catchments. Sulphate concentrations determined in groundwater by Raper (1998) were lower than those determined by

Christopher (1980), suggesting to Raper (1998) that there had been a reduction in the mean sulphate concentration in groundwater that may be attributable to a reduction in sulphur dioxide emissions and therefore lower dissolution rates. The result of increased discharge occurring as a consequence of more intense storm events on the Silesian strata, which are dominated by interbedded sandstones and shales would be likely to result in increased surface run-off, possibly to the detriment of the recharge to the limestone. Thus the underflow component of the limestone groundwater is also likely to be subject to minor changes. Off-set against the potential loss of underflow as a consequence of reduced recharge to the Goyt Syncline, there is the potential for the head difference between the Goyt Syncline and the White Peak to increase as natural lowering of the groundwater levels in the limestone takes place.

Climatic change could also influence rates of tufa deposition, for example Goudie et al. (1993) described a number of factors that may have contributed to the decline of tufa deposition rates, which in Derbyshire primarily occurred approximately 4000 years ago. For Derbyshire the date of 4000 BP coincides with Neolithic to Early Bronze Age deforestation (Taylor et al., 1994) in Lathkilldale. This may have influenced the barrage type tufa deposition in one or more of the following ways, many of which have been described by Goudie et al. (1993):

- Deforestation, in upland areas, in the vicinity of areas of groundwater recharge may have reduced the availability of carbon dioxide directly
- indirectly deforestation may have induced soil erosion, thereby reducing the availability of carbon dioxide
- deforestation in the vicinity of the depositing resurgence may have altered the shade consequentially causing stress to mosses or cyanobacteria, which utilise carbon dioxide and precipitate calcium carbonate
- deforestation in the resurgence area allowed grazing, which must have influenced groundwater chemistry by causing eutrophication, and increases in turbidity, thereby inhibiting the bacteria and mosses responsible for tufa accumulation
- increases in turbidity would also have resulted from the increase in surface run-off resulting from deforestation, furthermore increases in surface water run-off would increase the potential for dissolution of tufa deposits during periods of high discharge.
- deforestation would also have reduced the vegetation available which assists in the physical damming of streams.

This evidence suggests that climatic change of the scale forecast for the next 50 years is unlikely to impact significantly on the rates of tufa deposition, unless it is in conjunction with reforestation, which could be driven by the changes in agricultural funding and attitudes to rural land management. However, the evidence of this author suggests that tufa deposition is also associated with the gestation of inception horizons (described in Chapter 6, section 6.5 of this thesis) and with increased seasonality (decreased summer effective rainfall and increased winter effective rainfall) the potential for an increase in tufa deposition could result.

Collison et al. (2000) considered the impact of climate change on landslide frequency at a non-karst site (in the Lower Greensand) in Kent, with a record of known landslides. They predicted that climatic change is likely to result in a lower frequency of landslides. However, this is attributable to the relative increase in evapotranspiration in the southeast of England. Viles (2003) modelled an increase in winter effective rainfall for the Yorkshire Dales; therefore it would seem likely the risk of landslides in the Peak District, where clay wayboards are interbedded with outward dipping limestones, will increase, as observed by this author in Lathkill Dale in May 2003 and in Deep Dale during the same year.

CHAPTER 11: Lathkill Dale: a case study.

11.1 Introduction.

The River Lathkill, with its tributary the River Bradford, forms a tributary of the River Wye, in Derbyshire (Figures 11.1 – 11.3). The River Lathkill is an autogenic river; its source and entire length of flow lie almost entirely within the Carboniferous Limestone. However, it has been observed by Gunn (2000a), that there are a number of small sinks associated with a shale outlier, upon which the village of Monyash is largely situated. Bubble Springs (SK 20516611) forms the upstream limit of perennial flow in the River Lathkill. The River Bradford is perennial downstream of Well Head Spring (SK 200634). The confluence of the River Bradford with the River Lathkill, at SK 22056449, in the village of Alport is associated with extensive tufa deposits (Towler, 1977).

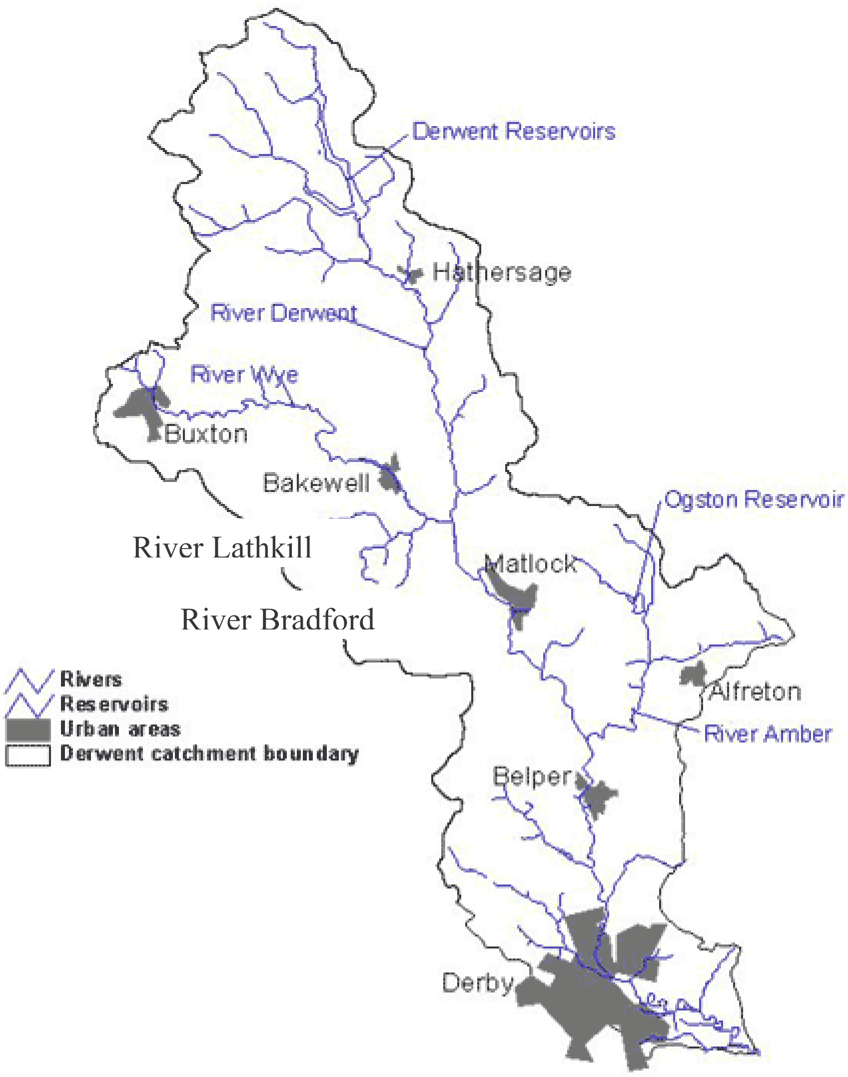


Figure 11.1: Location of the River Lathkill in the Wye catchment, Derbyshire. (Environment Agency, extract from CAM web-page)

The River Lathkill forms the axial drainage to the Monyash Syncline. The drainage of the syncline can be divided into two areas recognised by geomorphologists as the “upper” and “lower” Lathkill. This should not be confused with the Natural England definition of the Upper Lathkill Site of Special Scientific Interest (Ben le Bas, personal communication, February 2007). In the geomorphologists’ descriptions (e.g. Johnson, 1957) the course of the upper Lathkill was established in the early Pleistocene and it was captured by the lower Lathkill by headward erosion to the northwest, across the shale cover. It is considered that, prior to capture by the lower Lathkill; the upper Lathkill would have flowed eastwards to the River Derwent. Whilst adopting the upper/lower Lathkill division (Figures 11.2 and 11.3), this author notes that the commencement of the speleogenetic processes is likely to predate the Pleistocene (section 4.3). Upper Lathkill Dale (Figure 11.2) is seen as the easterly directed portion of the dale that lies upstream of Bubble Springs (SK 20516611). The valley associated with Meadow Place Grange (SK 201657) is a tributary of the lower Lathkill, which extends southeast from the Bubble Springs to the confluence of the River Lathkill with the River Bradford. This is a convenient division in terms of orientation, geology, hydrogeology and geomorphology.

There are a number of springs to which reference has been made in this chapter and these are shown on Figure 11.4. In particular, the reader’s attention is drawn to this author’s interpretation of the location of Pudding Springs (SK 17486547). This case study draws together a number of aspects described and discussed elsewhere in the thesis and both supports and further develops some of the concepts presented in chapters 9 and 10. The structure of the chapter broadly follows the general structure of the thesis, firstly setting the geological context and then going on to examine speleogenesis. Dye-tracing tests relating to Lathkill Dale have been described in Chapter 7. The addition of a section on the geomorphology, which is certainly not exhaustive, provides a link between the geological and hydrogeological aspects discussed in section 11.5 and the brief discussion of the hydrogeochemistry (section 11.6) provides further evidence to support the hypotheses presented in section 11.5.

11.2 Geological Setting.

The surface catchment of the River Lathkill is largely coincident with the Monyash Syncline. The geology is presented on Figures 7.2, 7.4 and 7.5. The visible sequence comprises the Bee Low Limestone Formation overlain by the Monsal Dale Limestone Formation, capped by the Eyam Limestone Formation, Widmerpool Formation and locally (in Monyash), by Namurian shales. In the western part of the catchment the Bee Low Limestone is separated from the overlying Monsal Dale Limestone by the Upper Miller’s Dale Lava Member of the Bee Low Limestone Formation. Towards the eastern end of the syncline, in the area of Over Haddon, the basal beds of the Monsal Dale Limestone are interbedded with the Conksbury Bridge Lava Member and the underlying Lathkill Lodge Lava Member (Figure 11.5).

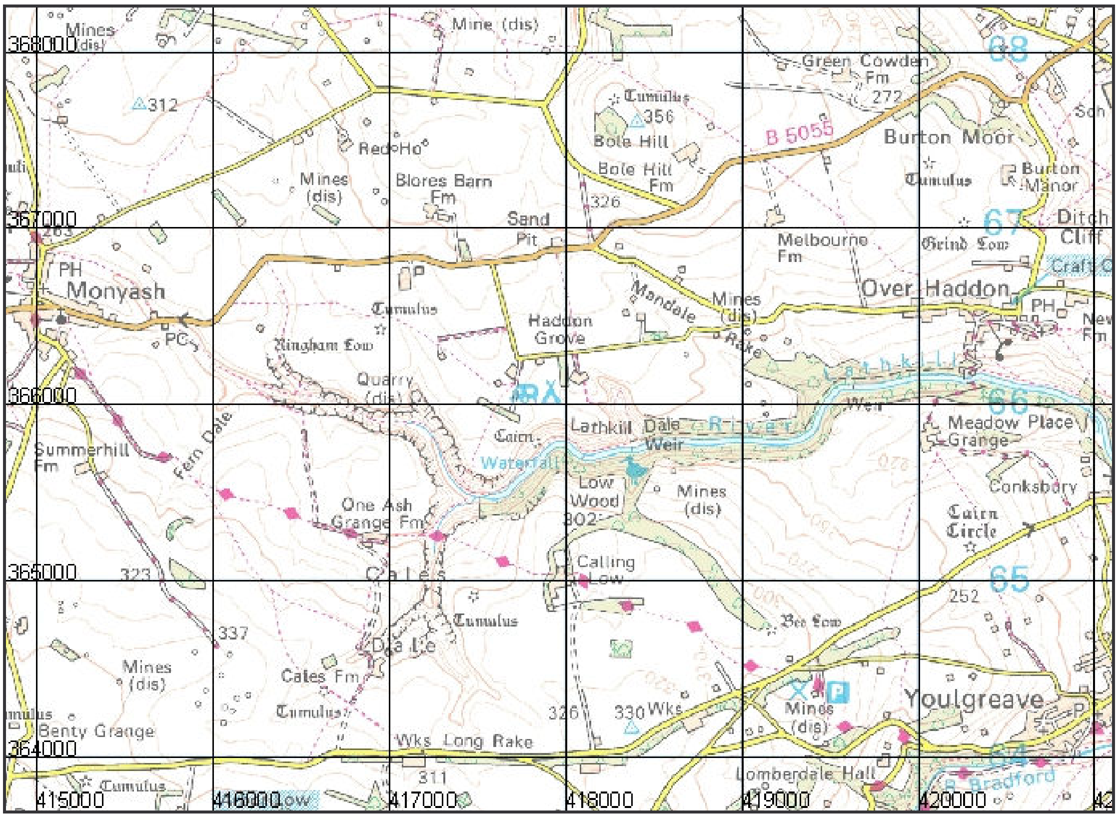


Figure 11.2: Extract of 1: 50 000 Scale Ordnance Survey map to show the location of the Upper Lathkill. (Copied in accordance with Ordnance Survey Licence 100046678).

The stratigraphical succession has been described in some detail by Aitkenhead et al. (1985), Shirley (1957) and Worley (1978). Although it is not immediately evident from examination of the 1: 50 000 Series British Geological Survey Sheet 111, the lower beds of the Monsal Dale Limestone give way to the Fallgate Volcanic Formation to the southeast, and in the vicinity of Alport (SK 221646) over half the sequence consists of volcanic rocks, which taper out southwards.

Limestones of the Asbian and Brigantian stages are shallow carbonate shelf deposits (Chapter 2), with cyclic sedimentation taking the form of progressive shallowing (regressive sequences). Walkden (1974) has shown that the upper parts of beds show evidence of emergence, including calcrete textures and paleokarstic surfaces, whereas the lower parts comprise skeletal carbonate sands indicative of a subtidal environment. Bedding in the Bee Low Limestone Formation tends to be more massive than that in the overlying Monsal Dale Limestone Formation. Stylolites are developed throughout. Reef limestones occur towards the top of the Monsal Dale Limestone (Gutteridge, 1991a) and are exposed towards the western end of upper Lathkill Dale. Good exposures can be examined in the area of Ricklow Quarry (SK 16556612). The reefs and the Eyam Limestone Formation that are exposed in Lathkill Dale were described in considerable detail by Gutteridge (1983, 1990, 1991a and 1995). There has been less extensive examination of the Monsal Dale Limestone Formation. The Monsal Dale Limestone is characterised by greater variability than the Bee Low Limestone Formation and is subdivided into pale and dark facies. The dark facies are suspected to derive their colour from the

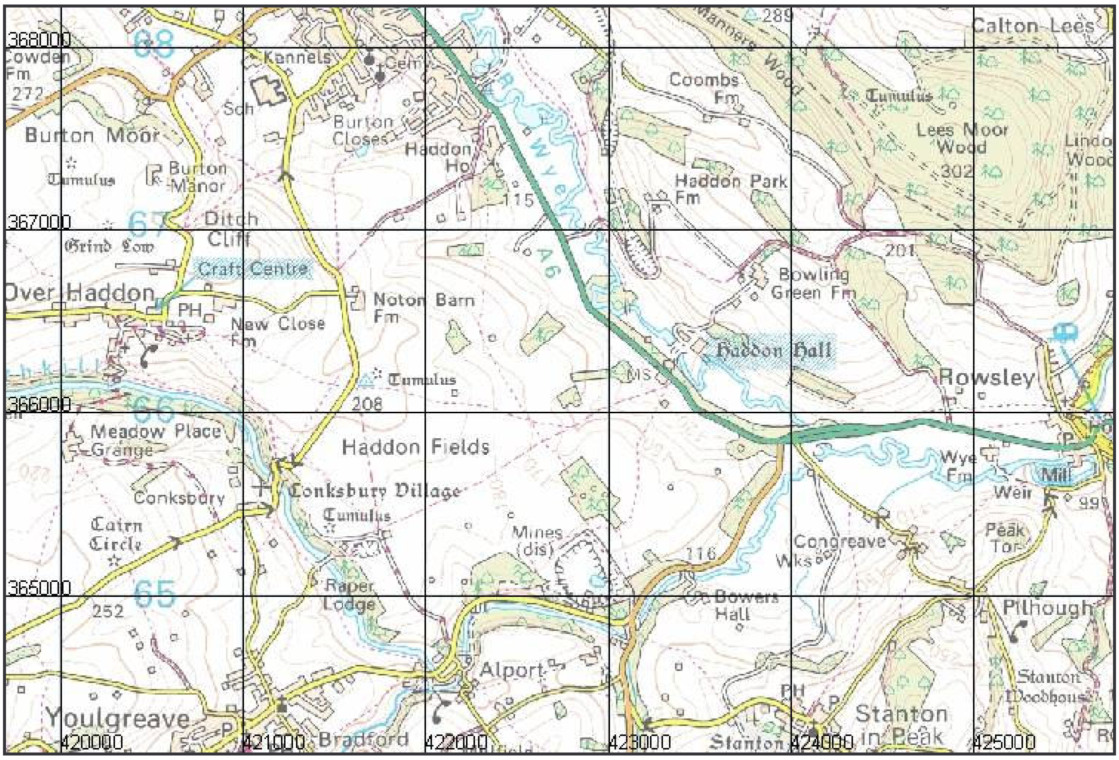
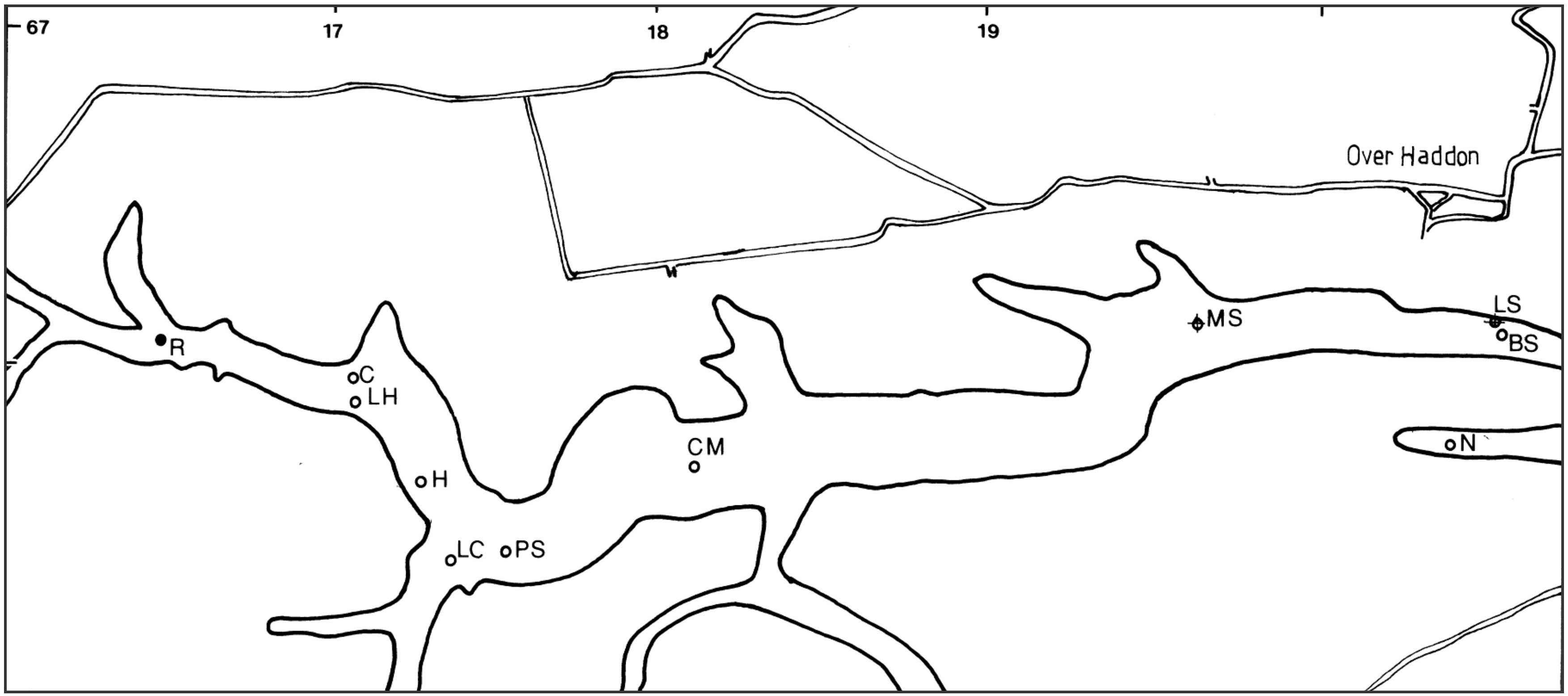


Figure 11.3: Extract of 1: 50 000 Scale Ordnance Survey map to show the location of the Lower Lathkill. (Copied in accordance with Ordnance Survey Licence 100046678).

contact with hydrocarbons, they also contain a higher percentage of clay, sulphide and P_2O_5 (Bridge and Gozzard, 1981).

Further evidence in support of the importance of hydrocarbon rich fluids is the relative purity of the Monsal Dale Limestone in the zone between the Lathkill Lodge Lava and the overlying Conksbury Bridge Lava (Bridge and Gozzard, 1981). In the area of SK 18356575 chert is interbedded with the Monsal Dale Limestone, characteristically the chert beds are in the order of 10 - 15 cm in thickness. The dark facies (black marble) was once mined from the area of SK 19006595. Chert paragenesis is considered in Appendix 3.2. Stylolites are also clearly evident in this area. The key characteristics of the Bee Low Limestone Formation have been described in Chapters 3, 7 and 9. To summarise, clay wayboards are particularly well developed in the upper part of the Bee Low Limestones (Miller's Dale Limestone Member) and the massiveness and brittleness of the underlying Chee Tor Limestone Member renders it more susceptibility to fracturing, thus where it is exposed, stress relief results in characteristic subvertical jointing. However at depth, where potential joints are closed, the Chee Tor Limestone Member acts as an aquitard.



Key:
BS Bubble Springs; CM Carter's Mill Spring; C Critchlow Cave; H Holme Grove Risings; LH Lathkill Head Cave; LC Lower Cales Dale; N Not previously named, referred to as Greaves Hollow Spring in this thesis; PS Pudding Springs; Closed circle cave, without spring: R Ricklow Cave.
Circle with cross sough discharge: LS Lathkilldale Sough; MS Mandale Sough.

Figure 11.4: Springs in Lathkill Dale.

Mineral veins in upper Lathkill Dale do not seem to be so strongly related to normal faulting as they are in other areas of the Wye catchment; instead the fissures appear to be strike-slip faults (with minor displacement). In support of this it is noted that Rieuwerts (2000) in his descriptions of Mandale and Lathkill mines described slickensides, which are indicative of horizontal movement. It has been argued (Chapter 9) that the orientation of the faults reflects a rotation in the stress field attributable to the juxtaposition of less competent volcanic deposits (clay wayboards) and more competent limestones. This fault pattern is also evident in the Monsal Dale Limestone to the north of the River Wye. In the area of upper Lathkill Dale it is possible that the fault pattern has also been influenced by the intrusion of dolerite at Calton Hill. The anticlinal setting of the mineral veins that is seen elsewhere in the region is not mirrored in upper Lathkill Dale and it would seem most likely that this is because the volcanic strata have served as a trap to mineralization.

11.3 Speleogenesis.

Chapter 4 described speleogenesis as the mode of development of limestone aquifers. This section follows a similar format and considers the aquifers in terms of inputs, storage, transfer mechanisms and outputs. The explored, active caves that form the principal flow paths (lines of transfer) within Lathkill Dale, including Cascade Cavern, Ricklow Cave, Lathkill Head Cave, Critchlow Cavern, Cales Dale Upper and Cales Dale Lower caves, have been described by Beck (1980) and Beck and Gill (1991). Subsequent exploratory work has been carried out and some of the findings have been reported by Gunn and Beck (2002) and indicate in the order of 3 km of cave in the Lathkill system.

The caves take a branchwork form (Figure 4.2). Palmer (2000, p. 78) suggested, "*cave origin is enhanced where surface runoff from large catchments is concentrated in a few small areas of ground water recharge. Caves are thus most abundant in soluble rocks that lie beneath perched stream valleys*…" and (p. 90) "*branchwork caves are especially well adjusted to the geomorphic history of the valleys into which they drain. During fluvial entrenchment, water in base-level cave streams tends to divert to successively lower levels*". Certainly the close correspondence of the Lathkill cave system with the valley is interesting (see also section 11.4). Examination of the 1: 50 000 Series British Geological Survey Sheet 111 'Buxton' Solid and Drift edition shows that the Upper Miller's Dale Lava crops out in the order of 4 km to the west of Monyash. Its outcrop forms an approximately semi-circular feature, which is locally coincident with, but locally extends to the west of, the topographic watershed of the river. The remnant of shale in the vicinity of Monyash, as well as the form of the river network and the associated dry valleys suggests that the valley has been inherited from the now eroded shale cover. This, together with the form and orientation of the caves, suggests a significant structural influence on the cave morphology. It was suggested in section 4.6.3 and by Al Sabti (1977) that dolines receiving autogenic recharge are associated with specific settings, particularly with valley sides and dominant faults. It has also been noted that the dolines are more visible in the Chee Tor Limestone Member, e.g. the lines of dolines observed along the line of the faults at King Sterndale (SK 092715) and the doline at Nether Low (SK 10266937), than they are in the Monsal Dale

Limestone, or the Eyam Limestone. The wider dominant-fissure spacing in the Monsal Dale Limestone and the Eyam Limestone appears to be associated with larger dolines, with dissolution being focused both down fissures and along bedding between fissure sets. It was also noted that dolines in the Monsal Dale and Eyam limestones are largely to be found within and beneath the pockets of Head deposits, as seen in the vicinity of Calling Low (SK 187647), with supportive evidence from aerial photograph studies (Appendix 3.5).

Similar pockets of Head deposits are evident to the west of Sheldon at SK 164688; to the west of Bole Hill at SK 175677; to the southeast of Bole Hill at SK 193763; to the east of Arbor Low at SK 168635; and Ricklow Dale at SK 163668. In support of the latter observation it has been noted that Farey (1811, p. 489) made reference to a "*swallow hole, Monyash E, Ricklow Dale*". Examination of the geology map indicates that in some locations there is a close association between faults (or extensions of faults) and the occurrence of Head deposits. It is the opinion of this author (section 9.3.6) that the solution hollows that have developed in the Monsal Dale Limestone are associated with inception horizons, which facilitate the regional subsidence component. It would appear that dolines form focal points for the solution hollows. Localized occurrences of smaller, fault-related solution dolines do also occur in the Monsal Dale Limestone e.g. in Fern Dale (SK 15706560) and below Over Haddon at SK 20656625, where they are associated with the feather edge of the Conksbury Bridge Lava.

It is significant that Farey (1811, p. 475) in describing the Bradford and Lathkill rivers observed that "*The Lathkill Excavation pursues the 1st Lime to the south side of Over-Haddon, where it descends the Series on to the 1st Toadstone, cuts through it, and lays bear a patch of the 2nd Lime; but which not being excavated so deep as the Toadstone below, occasions a sudden fall in the River, whose course, after crossing the great Bakewell Fault, again ascends on to the 1st Lime, in which it is deeply cut up to Monyash, where…*" This seems to provide further evidence that the flow path of the River Lathkill is strongly influenced by the presence of inception horizons and serves as further evidence that the faults form zones of groundwater storage, which discharge via inception horizons.

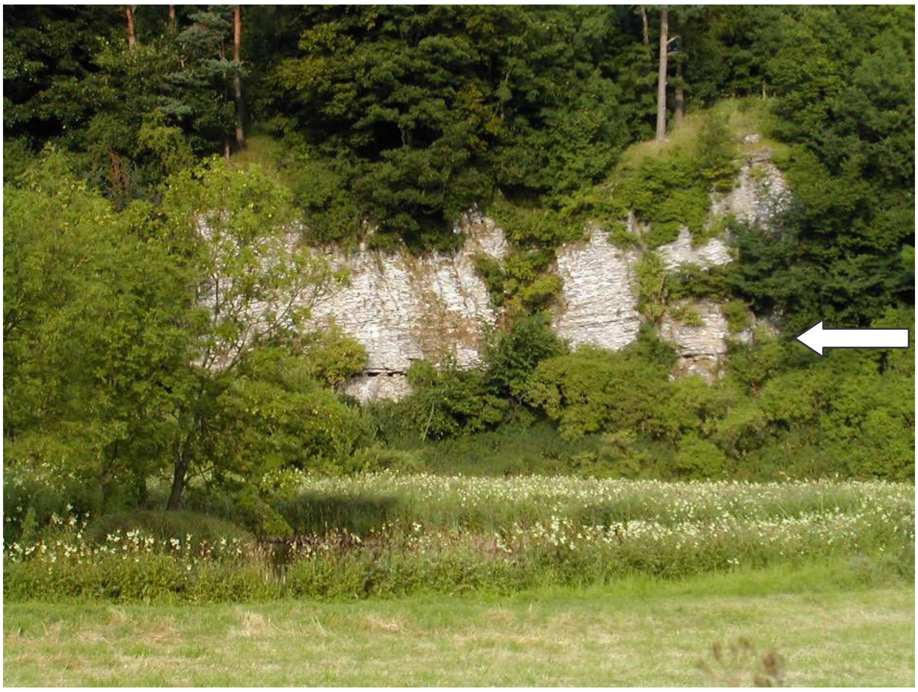
The strong influence of inception horizons in the Eyam Limestone is evident in Plate 11.1 (see also Plate A4.1.1 and Appendix 4.1). They are also evident in the lowering of cave base level reported by Beck (1980), which corresponds to switches in the dominant inception horizon. Supporting evidence for this is also provided by the association of inception horizons with tufa deposits (section 11.4). Levelling (Appendix 11.1) and field observations made by this author have identified a number of significant inception settings within Lathkill Dale. One comprises the base of the shell beds within the Monsal Dale Limestone, which appears to correspond with the current level of the River Lathkill. A second corresponds with a zone within the bedded facies of the Eyam Limestone, comprising a pale limestone towards the top of the Formation (Plate 11.2). Both of these horizons are associated with the presence of silica. This suggests that previous fluid phases targeted them because they were zones of higher permeability. In the case of the chert nodules, stylolites can be seen to pass through them, suggesting to this author that an early phase of silica was mobilised very soon after deposition of the limestone, which implies early development of inception horizons.

Careful examination of British Geological Survey logs of boreholes and sections (Aitkenhead et al., 1985) and comparison with survey levels (Appendix 11.1) provides evidence of an inception horizon within the Lathkill Cave system. It comprises the shelly basal beds of the bedded calcarenite that rest on a clay wayboard that has been interpreted by Aitkenhead et al. (1985) to be a continuation of the Conksbury Bridge Lava (approximately 50 m below the Eyam Limestones in the area of Haddon Grove and 15 m below the Lathkill Shell Bed [Aitkenhead et al., 1985]). However, the interpretation of Aitkenhead et al. 1985 is thrown into question in section 11.8. The horizon lies at, or immediately below, the boundary between the dark and the pale facies of the Monsal Dale Limestone. At first sight it would appear that the inception horizon in Lathkill Head Cave is the main carrier of ground water. However evidence presented in section 11.7 suggests that there are deeper inception horizons. Similar horizons have been identified by Ford et al. (1975) in Bradwell Cavern. The close association of the Lathkill Cave system with the synclinal structure suggests that conduit development is influenced by the structure. This is likely to reflect the fact that the base of the syncline comprises a zone of groundwater mixing (the focus for water from each side of the syncline) and also as the focus for inception horizon related flow, a zone of higher discharge.

Two previously unrecorded, inception horizon related springs (SK 2042565775 and SK 2077565787) at levels of approximately 200 and 173 m OD respectively (Figure 11.4), have been identified in the valley to the east of Meadow Place Grange (Greaves Hollow). The latter spring lays less than 5 m below the Conksbury Bridge Lava and is associated with a shell bed. The upper horizon, also associated with a shell bed, is evident in the limestone sections to the north and south of Conksbury Bridge. The occurrence of a pond at One Ash Grange (SK 1700065187) at approximately 253 m OD suggests another inception horizon related spring towards the top of the Monsal Dale Limestone.

Water Icicle Close Cavern (Beck, 1980) is situated in an area of extensive Head deposits that cap Low Moor Plantation, to the south of Calling Low (SK 16106460). The bedding related form of this cave suggests association with a high-level inception horizon in the Monsal Dale Limestone. There are mine workings immediately to the south of the shaft access to the cave and the cave is thought to have been discovered via the mine workings. Beck (1980) suggested that if the line of the northern passage of the cave is projected 300 m to the north the tube would intersect a shallow dry valley heading towards One Ash Grange and Cales Dale, which suggests association with a former inception horizon.

Examination of inception horizons in Lathkill Dale has identified an association with relatively more permeable lithologies such as the shell bed and zones of dedolomitization. The evidence from sampling the inception horizons (Appendix 4.1) suggests that impedence of ground water flow may be just as significant in the development of inception horizons. Indeed in the area of Raper Lodge the limestone immediately beneath the inception horizon was silicified to such an extent that it was barely penetrable with the diamond bit of the Limestone Research Group's rock drill. Yet, the continuity of inception horizons that is evident in Lathkill Dale (Plate 11.1) suggests either a bedding related surface that is less affected by tectonic fissuring, by implication a less brittle material such as the clay wayboards, or alternatively a material that 'heals' across fissures. In this context it is possible that this is achieved by stylolites. As described (section 2.4), non-sutured seam solution (microstylolites, microstylolite swarms and clay seams) is common in the Monsal Dale Limestone Formation. The non-sutured seam solution appears to be particularly associated with clay wayboards (Berry, 1984) and also with the darker facies of the Monsal Dale Limestone Formation, which contain more non-calcareous material. Furthermore, this author has observed that microstylolites are common within the Monsal Dale Limestone exposed in Lathkill Dale and they are even seen to pass through chert nodules. Stylolites have also been observed in samples taken from the inception horizon observed in the vicinity of Coalpit Rake. The pressure associated with the development of stylolites facilitates the movement of silica (an otherwise brittle material). The association of stylolites with wayboards and shales (section 9.5) and the zones of dolomitization (section 2.4) provides a further process for inception, i.e. dedolomitization within the oxygenating environment (Appendix 4.1). The hydraulic constraint imposed by the silicified layers has the potential to impose rate limitation on the recession curve (section 10.4).



Inception
Horizon

Plate 11.1: Inception horizon development near Raper Lodge, Lathkill Dale.

11.4 Geomorphology in Lathkill Dale.

Smith and Atkinson (1976, p. 395) suggest that "*The salient features of the karst landscape are twofold namely the dry valley system and the enclosed depression*." Broadly this is true, but within the context of Lathkill Dale there are also a number of other geomorphological features that provide additional material to assist the understanding of the evolution of the hydrology and hydrogeology. This discussion has been restricted to the observations of the author on aspects that are thought to be either previously unpublished, or directly relevant to the hydrogeological interpretation of the dale. It should be noted that the observations are not exhaustive and there remains considerable scope for developing the understanding of the dale's geomorphological evolution.

Perhaps one of the most noteworthy geomorphological features of Lathkill Dale is the apparent influence of mineral veins on valley form (Figure 11.5). The nature of the association takes a number of forms. Some of the valleys are centred on mineral veins, including parts of Lathkill Dale Vein. Others, where they occur parallel to the strike of bedding, form the head of valleys, for example at the head of Fern Dale and Lathkill Dale. In other locations, where the vein is perpendicular to the strike of bedding and parallel to the valley, the vein is found on the up-dip side of the valley, e.g. Mandale Rake. It is the opinion of this author that the coincidence of the valleys with the mineral veins provides supportive evidence both for the concept of inheritance of the visible drainage pattern from the Namurian cover (Chapter 10) and for the concept of groundwater storage in mineral veins, which might exacerbate potential instability.

The former association is useful in the analysis of the geology, particularly in providing evidence for the existence of an unmarked (on the British Geological Survey 1: 50 000 Series Sheet 111 and also the 1: 25 000 Series Sheet SK 16) fault in the area of Cales Dale. Where veins coincide with the valleys an asymmetrical form results, thus Figure 11.7 provides further support for the evidence of the fault in Cales Dale. Cales Dale is a tributary to the river, situated at a natural meander. The close association of meanders with faulting has also been observed in the River Wye (Chapter 4) and provides further supportive evidence for the inheritance of surface drainage from the former Namurian cover. The British Geological Survey 1: 50 000 Series, Sheet 111 does show a short length of mineral vein (in the order of 100 m) at SK 17196528, which was the site of a slope failure, on the public footpath, following heavy rain in August 2002. Being in the area at the time, this author took the opportunity to inspect the failure and take some structural measurements (Table 11.1) to contribute to the debate regarding the influence of faulting on valley form. Evidence for this being the location of a fault is discussed further in section 11.5

Slope failure processes characterise areas where limestone that is underlain by weathered volcanic lavas, or significant clay wayboards, and dips into the valley at an angle that exceeds the shearing angle of the partially saturated clay. This type of slide has been observed elsewhere in the White Peak, e.g. Peter's Stone in Cressbrook Dale (Ford, 1977). Such failures can also be seen in lower Lathkill Dale at SK 208657.

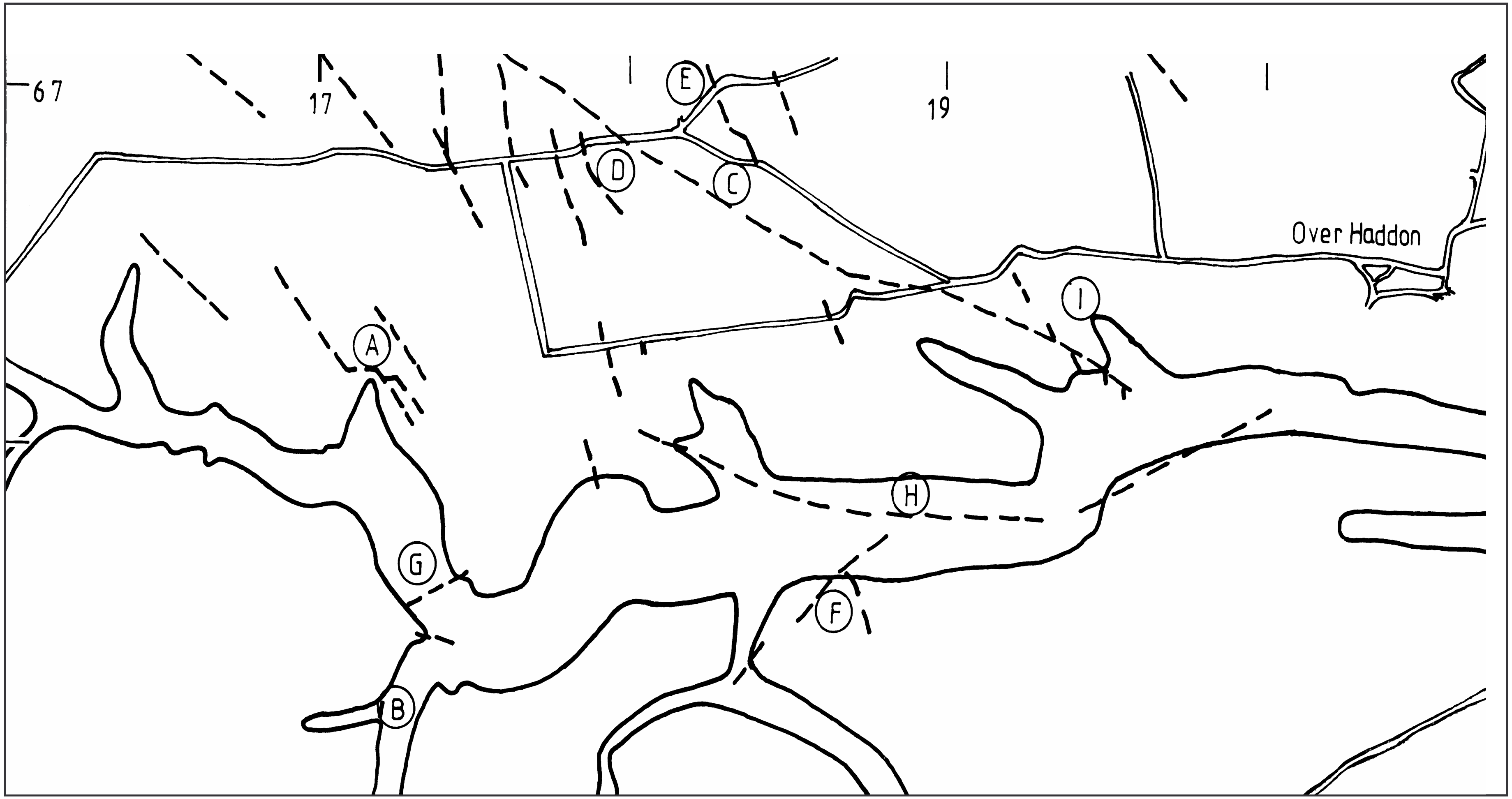
The occurrence of tufa at a number of locations in Lathkill Dale has been noted by other researchers including: Burek (1977 and 1978), Pedley (1990, 1993), Pedley et al. (2000) and Towler (1977). Pedley (1990, p. 143) defined cool freshwater tufa as "*highly porous or "spongy" freshwater carbonate rich in microphytic and macrophytic growths, leaves and woody tissue*". Pedley (1990) went on to present a number of environmental models derived from extensive fieldwork carried out on Quaternary and Recent deposits in northwestern Europe and the Mediterranean. The tufas of lower Lathkill Dale fall within the *barrage model*. They are associated with a step, of approximately 5 m, in the valley thalweg (Pedley et al., 2000).

The barrage model, where tufa is deposited rapidly at points of valley constriction, creating a downstream barrier and causing flooding of upstream barriers to form a large single lake, is typically represented by tufa at Raper Lodge (SK 21496491) and Alport (SK 22106467), as described by Pedley (1993) and Pedley et al. (2000). It has been observed by this author that the top of the tufa at Raper Lodge lies at approximately the same level as a relict inception horizon revealed in the opposite bank (Plate 11.1) and that the top of the Raper Lodge tufa lies at approximately the same level as the top of the barrier tufa at Alport (150 m OD). The geological setting of this inception horizon (close to the boundary between the Monsal Dale Limestone and the overlying Eyam Limestone) can also be identified in upper Lathkill Dale.

Extensive "sheets" of tufa have been observed in the lower Lathkill, between SK 21366527 and SK 21456520 and between SK 21466518 and SK 21566492. The tufa sheets appear to extend across the river and were identified by Pedley et al. (2000) as the location of additional barrage locations. It is suspected by this author that these are associated with another inception horizon, which is down thrown to the south at Coalpit Bridge. The setting of these deposits is similar to the association of a bed of tufa with the shell bed related inception horizon, on the south bank of the upper Lathkill, below the waterfall at SK 18126570. The association of tufa with inception horizons is an entirely logical association, which may be of use in the identification of dominant inception horizons elsewhere in the region. The consequential correlation of tufa with the ready supply of calcium carbonate from the inception horizon does not contradict Pedley's models (1993), although it does suggest a hydrogeological setting in which the dimensions of the tufa systems are influenced by the spacing of the inception horizons.

Table 11.1 Compass Clinometer Readings taken on failed slope in Cales Dale on 7 August 2002.

	Strike (˚)	Dip (˚)	Observations
Mineral Vein	018	Vertical	Surface undulation c. 3cm
Bedding	160	3 E	Bedding could have been disturbed by slope failure
Bedding	160	5 E	Bedding surfaces undulate by at least 2 cm
Joint	016	89 E	
Joint	022	87 E	
Joint	125	90	
Joint	122	83 N	
Joint	122	89 N	
Joint	140	87 N	
Joint	017	77 W	
Joint	011	89 W	



Key: A mineral vein clearly visible at the head of the valley; B Unmarked fault in Cales Dale; C Mandale Rake can be traced as a valley at surface; D, E, F Mineral veins that can be traced as valleys at surface; G Holmes Grove guides confined resurgence; H Lathkill Dale Vein, the focus of the valley along the majority of the length of the vein; I Pasture Rake can be traced as a valley at surface. Legend: Dashed lines represent mineral veins.

Figure 11.5: Relationship of mineral veins to valleys.

Plate 11.2: Tufa deposits at Alport.

At Bubble Springs the situation appears to be more complex. The bed form appears to change from a graded slope, which has the appearance of bedding planes with associated pools, eventually giving way, over a 1 m drop, to a silty, flat-bottomed area. At first sight this seems similar to the setting of the Coalpit Barrier site (Pedley 1993) at the confluence of the tufa depositing water with the relict lake, however, on closer examination it is different, because the tufa is not evident at the lower level and there are also springs in the lower bed, which have dark aureoles associated with them (Plate 11.3). These could either be micritic tufa, or fine sediment that is discharged from the conduits. This author suspects the latter. Towler (1977) described the tufa from the pool areas as being similar to that at Raper Lodge, in that it was compact, but it was generally more crumbly and she suggested that this could be attributable to river erosion. These points are considered further in section 11.5.

Tufa accumulation rates in the order of 2.5 mm/ year have been calculated from preliminary dating of cores taken from the barrier lake deposits (Andrews et al., 1994) and the oldest sediment has been dated to the Late Devensian although Andrews (personal communication, 2002) suggests that the Alport barrier has been shown to date to the Ipswichian. Oxygen isotopes have been used to suggest that air temperatures were in the order of 2˚ C higher during the period of tufa deposition. Burek (1977, 1978 and 1991) described glacial aspects of Lathkill Dale, finding that evidence from the early Anglian glaciations appears to have been masked by later glaciations. Evidence of Devensian ice is absent from the area. Burek (1977) suggested that lower Lathkill Dale was subject to glaciation during the Wolstonian, but more recent work, including Aitkenhead et al. (2002) has pushed this back to the Anglian. Pedley et al. (2000, p. 732) provided supporting evidence of glacial influence, in that they suspected that "*a basal lithoclast breccia in clay matrix*" is soliflucted till [Hunt (personal communication, 2007) opines that this is more likely to be Head]. The dating of the Lathkill barrier does not conflict with this. It is very plausible that the connection of the upper and lower Lathkill results from glacial activity during the Anglian, as the lower Lathkill lies parallel to the apparent western edge of the Till. Furthermore, Pedley (1993) has established that there is in the order of 11 m (up to 16 m was suggested in a personal communication, 2003) of channel fill in the area of the former lake associated with the tufa development. The possibility of buried channel features is of course significant to the hydrogeology, particularly as Taylor et al. (1994) suggest that the tufa deposits can be permeable.

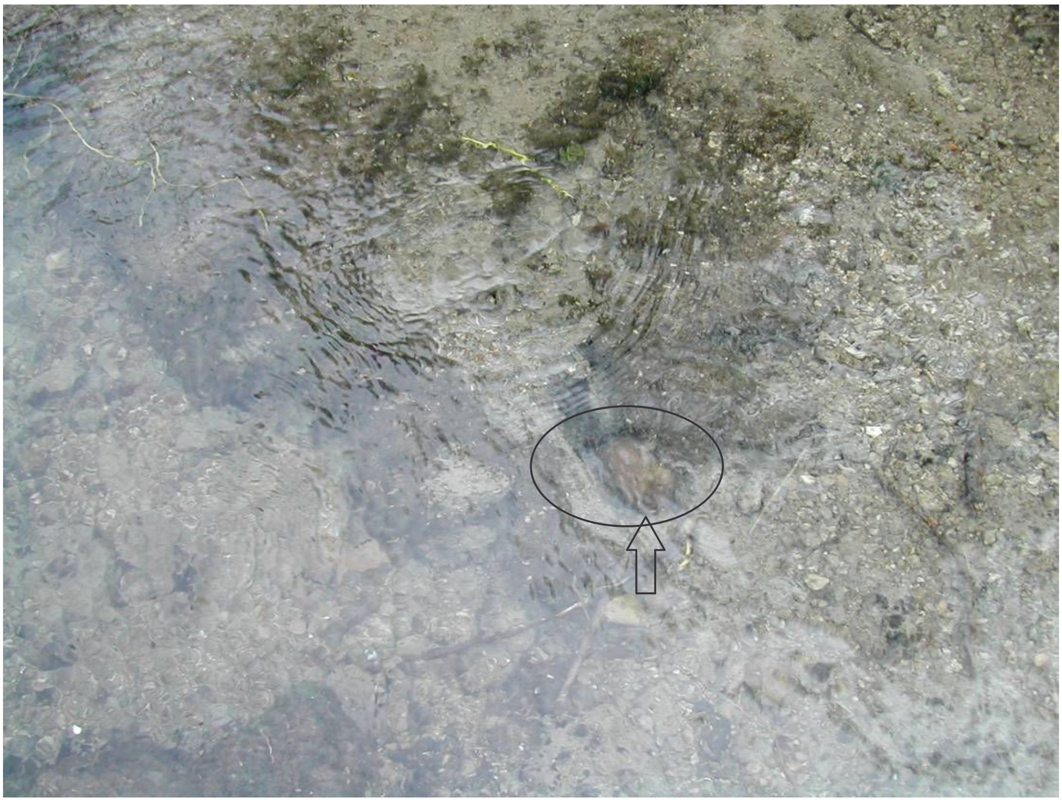


Plate 11.3: One of the springs (with dark aureole and marked with elipse) at Bubble Springs.

Straw (1968) noted that the headwaters of the upper Lathkill, formed by Cales Dale and Fern Dale, appear to form a "*hammer head*" pattern, reflecting the way that the former shale cover has influenced valley formation. Similar observations were made of the River Bradford, where Gratton Dale cut back to capture Long Dale (Johnson, 1957). However, it should be noted that this does not preclude subsurface conduit development. Indeed the association of recharge with faults or mineral veins would also guide surface erosion, particularly during periods of ice wasting. The influence of the knoll reefs in guiding the surface water drainage system (Ford and Burek, 1976) is evident in the course of the River Lathkill, although it should also be noted that the current level of the river lies below that of the base of the reefs.

Valley settings lend themselves to the development of caves as a consequence of the tensional environment produced by valley cambering (section 4.3). It is clear that cambering has occurred in Lathkill Dale. In the vicinity of Lathkill Head Cave the bedding on either side of the valley sides dips away from the axis of the valley, furthermore juxtaposed beds have shifted relative to one another (Plate 11.4, SK 16856953). The number of avens within the cave system also implies a tensional setting.

← Gaps between beds resulting from cambering

Plate 11.4: Cambering in Lathkill Dale.

One of the access points to Lathkill Head Cave (Garden Path at SK 16526594) has been created artificially (Gunn personal communication, 2006) via what is colloquially referred to as a "rift", comprising large volumes of open space resulting from stress relief of the rock. Interestingly, there is no evidence to suggest that the rift forms a major conduit for vadose flow, i.e. it is not associated with doline development and fissure surfaces are relatively fresh, even where they are coated by speleothem (Plate 11.5). Where the rift cave is natural, it appears to result from the stress relief associated with the cambering process. Sasowsky and White (1994) identified stress relief as a cave forming process and found that the tendency is for conduits to form on the down dip side of the valley. In upper Lathkill Dale the valley is coincident with a syncline, which is reflected in the cave development on both sides of the valley. It would appear that rifting has occurred along pre-existing northwest to southeast fissures and the conjugate fissure set, with which the Cales Dale Caves are associated. There is no surface expression of the cambering observed in the Lathkill setting, probably because the overlying Eyam Limestone Formation is less competent and therefore less susceptible to fracture. Further evidence for valley cambering as a process comes from the fact that in this setting the rift caves are not associated with extensive mineralization. Had they been open at the time of mineralization they would have been a focus for mineral deposits, thereby confirming that the fissuring post dates mineralization. By contrast, the bedding related component of the cave system comprises a phreatic system (Beck, 1980), which clearly pre-dates the cambering and has been interpreted by this author as inception horizon- related conduit development. To date the Lathkill Head cave system has only been explored in upper Lathkill Dale, the downstream end of exploration being at SK 1811665698.

Plate 11.5: Speleothem deposits on rift surfaces in Lathkill Head Cave.

Beck (1980, p. 215) observed "*at least three distinct cave levels in the area, all of which are related directly to early local base levels controlled by the elevation of the River Lathkill*" and suggested that the levels can be related to different stages of down cutting by the river. Implicit in this is that the base level for the system is the River Lathkill, yet a number of factors indicate that the situation is more complex, because the intermediate and regional base levels lie below the level of the River Lathkill. Such factors include the following: i) seasonally, groundwater levels fall beneath the base of the River Lathkill; ii) Beck's (1980) descriptions suggest minimal vadose down cutting associated with the bedding related caves; iii) dye-tracing test results (Chapter 7) indicate that in the region of the headwaters of the Lathkill basin (Chelmorton and Taddington) there are connections with the River Wye the base level of which is at a lower level; iv) it has been argued in Chapter 4 that speleogenetic processes associated with the development of the inception horizons are likely to pre-date the Pleistocene; and v) in Chapters 5 and 10 the concept of underflow has been described. Consequently, this author prefers the concept of flow on a number of structurally guided, active inception horizons connected by dominant fissures and exposed by valley incision. Clearly when a valley cuts through an inception horizon, springs associated with the newly formed valley will become a focus for vadose flow. The exposed spring will be subject to weathering and frost action, which encourages the geomorphological development of benches associated with the valley base at the time of exposure, particularly if the underlying bed is more resistant (e.g. silicified). Thus it is considered that the benches can be used as evidence of phases of Pleistocene down cutting (Chapter 10).

11.5 Hydrology and hydrogeology.

The surface expression of the upstream end of the River Lathkill is ephemeral and although flow is maintained farther upstream for part of the year it is only truly perennial downstream of the point known as Bubble Springs (SK 20516611). Wilson (1990, p. 151) suggests that "*an influent stream is one where base-flow is negative; that is the stream feeds the ground water instead of receiving from it* .." and (p. 152) "*an effluent stream on the other hand is fed by the ground water and acts as a drain for bordering aquifers*". On the basis of this definition the River Lathkill should be considered as an effluent stream that flows underground for part of the year, because the main contribution to the flow of the River Lathkill is ground water. Applying current karst terminology (EPA, 1999, p. 107), a losing stream is "*a stream or reach of a stream in which water flows from the stream bed into the ground. In karst terrains, losing streams may slowly sink into fractures or completely disappear down a ponor*" and thus the River Lathkill could be classed as a losing effluent stream. However, Gunn (personal communication, 2006) suggests that the upper Lathkill should be considered as an overflow-spring fed river, because it has point inputs and outputs and observed that the river not easy to classify, because it has been affected by sough construction.

It is apparent, both to the observer and from the results of the flow measurements described below (Appendix 11.2), that ground water largely enters the upper Lathkill via the southern bank. As described in section 11.3, it would appear that it is largely fed by an inception horizon associated with a shell bed. This is particularly noticeable in Lathkill Head Cave, at Cales Dale (SK 17466550) and associated with tufa (section 10.4) at the site of a waterfall (SK 18116570). A similar situation is observed in lower Lathkill Dale. On 22 February 2003 discharge was measured immediately downstream of Lathkill Head Cave (6 litres/sec); immediately downstream of Holme Grove Risings (206 litres/sec), at SK 17386556; and at the downstream end of Cales Dale (212 litres/sec, above the junction with the River Lathkill). The altitudes of these points are: 207 m OD, 199 m OD and 204 m OD respectively. The level of the associated inception horizon has been plotted (Figures 11.6 to 11.8). From the cross sections, it is clear that the shell bed in question cannot be sitting on a clay wayboard associated with the Conksbury Bridge Lava, as the level of this shell bed appears to lie beneath both the Conksbury Bridge Lava and the Lathkill Lodge Lava. It should also be noted that, because of the elevated temperature (Table 10.6) and structural position, the contribution from Holme Grove Risings is suspected to include a significant underflow contribution.

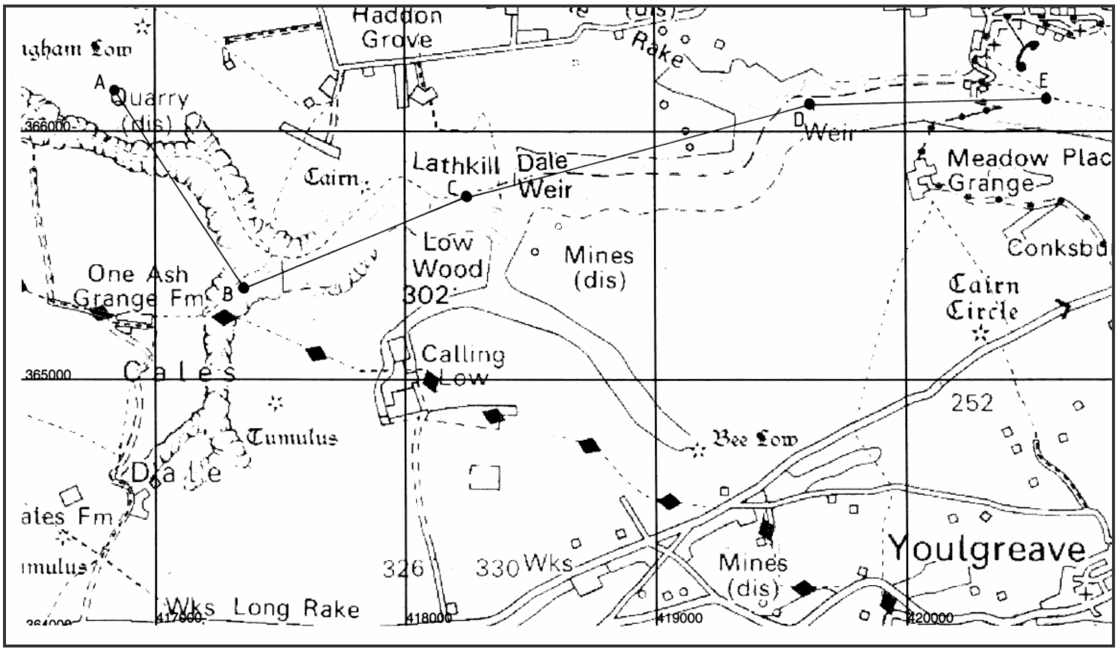


Figure 11.6: Ordnance Survey extract annotated to show line of sections A – B, B – C, C – D and D – E.
(Copied in accordance with Ordnance Survey Licence 100046678)

On the north bank of the River Lathkill the situation appears to be different, probably partly as a consequence of the greater number of dominant northwest to southeast orientated mineral veins. The experience of this author suggests that there is likely to be considerable storage associated with the mineral veins. During high groundwater conditions there is visible evidence of a contribution to the river that is associated with mineral veins, particularly to the east of Holme Grove Risings. Nevertheless, it seems anomalous that there is no visible contribution from the shell bed on the northern side of the Dale. This could be accounted for in a number of ways, for example it might be a lithological reason, such that inception has not occurred on the northern side of the dale. Palmer (2000, p. 79) observed that "*phreatic conduits preferentially form where the ground water flow is greatest, and this is strongly controlled by the widths of presolutional fissures*" and in a similar vein Klimchouk and Ford (2000b, p. 63) noted the significance of the "*heterogeneity of prespeleogenetic* *porosity of emergence*". Alternatively, it is possible, although there is little visible supporting evidence, that the strata to the north have been marginally downthrown along the Lathkill Vein, or that mined mineral veins capture groundwater that would otherwise have reached the River Lathkill (section 11.7).

At Bubble Springs the ground water contribution is from the bed of the river (Plate 11.3) via approximately twenty orifices, in the order of 5 to 10 cm in diameter. The orifices that comprise Bubble Springs are quite interesting in that their hydrological function, geological position and their karstic drainage (Smart and Worthington, 2004b) can be constrained quite closely, as described below. On 5 April 2003 three measurements of flow from the largest orifice were made, using the Limestone Research Group electromagnetic flow meter, with the sensor inverted to monitor discharge from the orifice. Velocities ranged between 0.74 and 0.79 m/second. Measurements were also made in the smaller orifices, where velocities were much lower (0.035 to 0.065 m/second). Bubble Springs rise in the Monsal Dale Limestone and stratigraphically they can be traced to the zone beneath the Lathkill Lodge Lava. Based on evidence from upstream, this suggests the presence of an inception horizon beneath this zone. Flow monitoring carried out on 5 April 2003 (Table 11.5) identified that on this occasion the discharge of the River Lathkill increased by approximately two thirds at Bubble Springs, which indicates that the source is larger than the topographic catchment inferred to contribute to the stratigraphically defined inception horizon. This is particularly evident during periods of low groundwater conditions. Accordingly, the springs appear to represent the focus of resurgence of underflow in the Monsal Dale Limestone. What is not immediately apparent is why the water should rise at this point. It is the opinion of this author that it results from the presence of a north to south-trending fault, which is downthrown to the east, such that the lower beds of the Monsal Dale Limestone are brought into contact with the down-thrown Conksbury Bridge lava and the underlying Lathkill Lodge Lava. The lavas thicken and coalesce to the east. That the position of springs is not coincident with the mapped location of the fault probably reflects stepped faulting, within a fault zone, for immediately downstream of Bubble Springs the bedding can be seen to dip moderately steeply to the east (Plate 11.6). Alternatively, there may be a zone of storage in the limestone between the Lathkill Lodge Lava and the Conksbury Bridge Lava, such that ground water rises up the fault and then apparently moves up the hydraulic gradient, beneath the Conksbury Bridge Lava, to emerge from the springs. Oakman's (1979) descriptions of significant ground water difficulties in the zone between the two lavas in the Alport mining field indicate that the latter is a plausible explanation (Figure 11.8).

The concept of the upper Lathkill being captured by the lower Lathkill was described in section 2.10. The form of the river valley has been modified both by glacial processes and by humans for a variety of uses, in particular by the lead miners, but also by farmers, and for corn milling. In order to provide more accurate information regarding the hydraulic gradient and the length of flow paths in the Lathkill catchment, this author carried out a topographical survey of the upper Lathkill from B5055 at Monyash to Bubble Springs (Appendix 11.1). The survey indicates a flow path with a gradient of 1:70, or 0.82˚, which falls within the range of typical phreatic, cave passage gradients (Worthington, 1991). Whereas

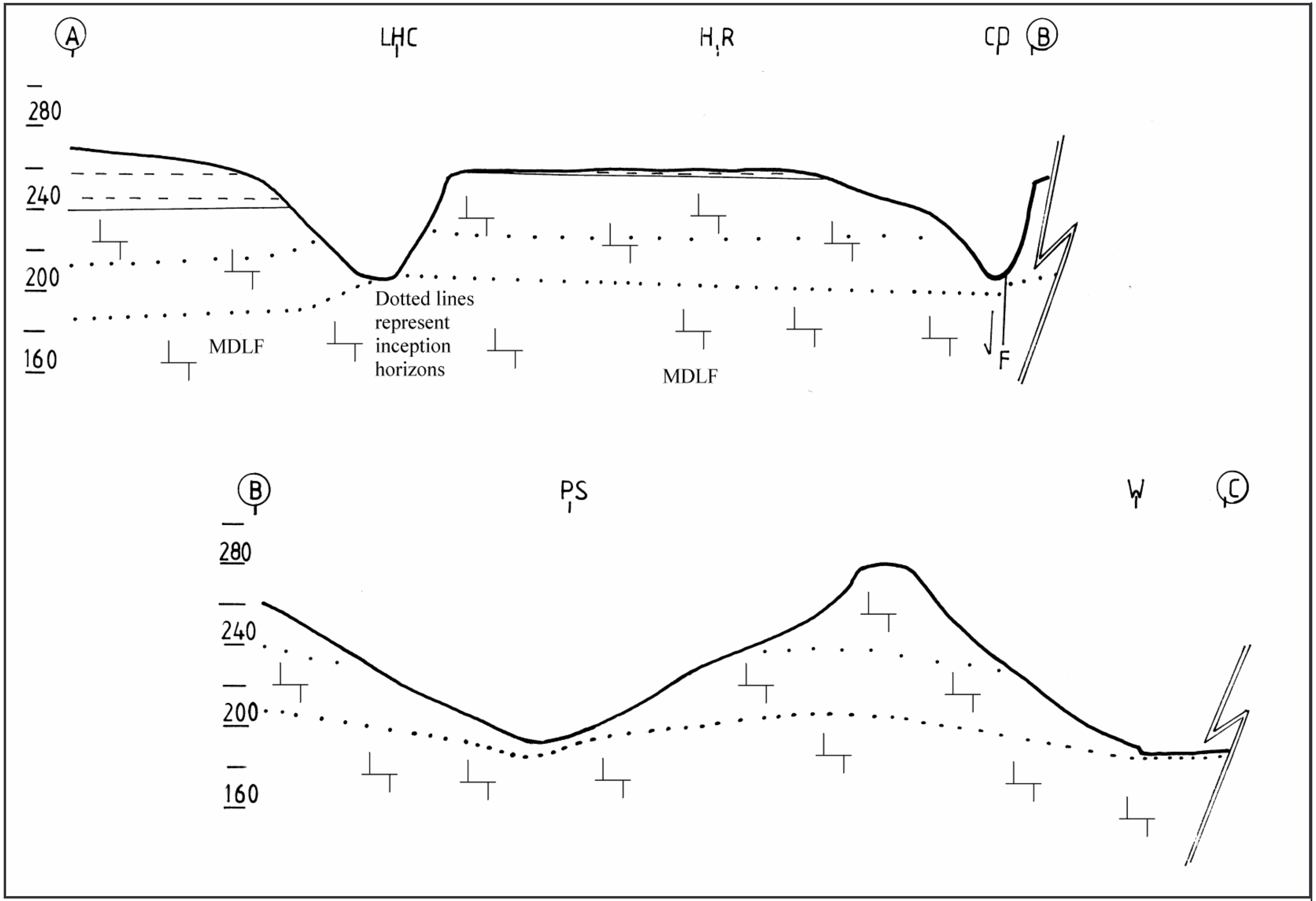


Figure 11.7: Sections A (SK 16856617) - B (SK 17376536) and B – C (SK 18256573) to show level of inception horizon.
Vertical exaggeration x 1.7 (approx) ; Levels m AOD; Line of sections shown on Figure 11.6 and continued on Figure 11.8.
Legend: dashed lines Eyam Limestone Formation; MDLF Monsal Dale Formation. Dotted line= line of inception horizon
Key: CD Cales Dale Cave; HR Holme Grove Risings (behind); LHC Lathkill Head Cave; PS Pudding Springs; W Waterfall.

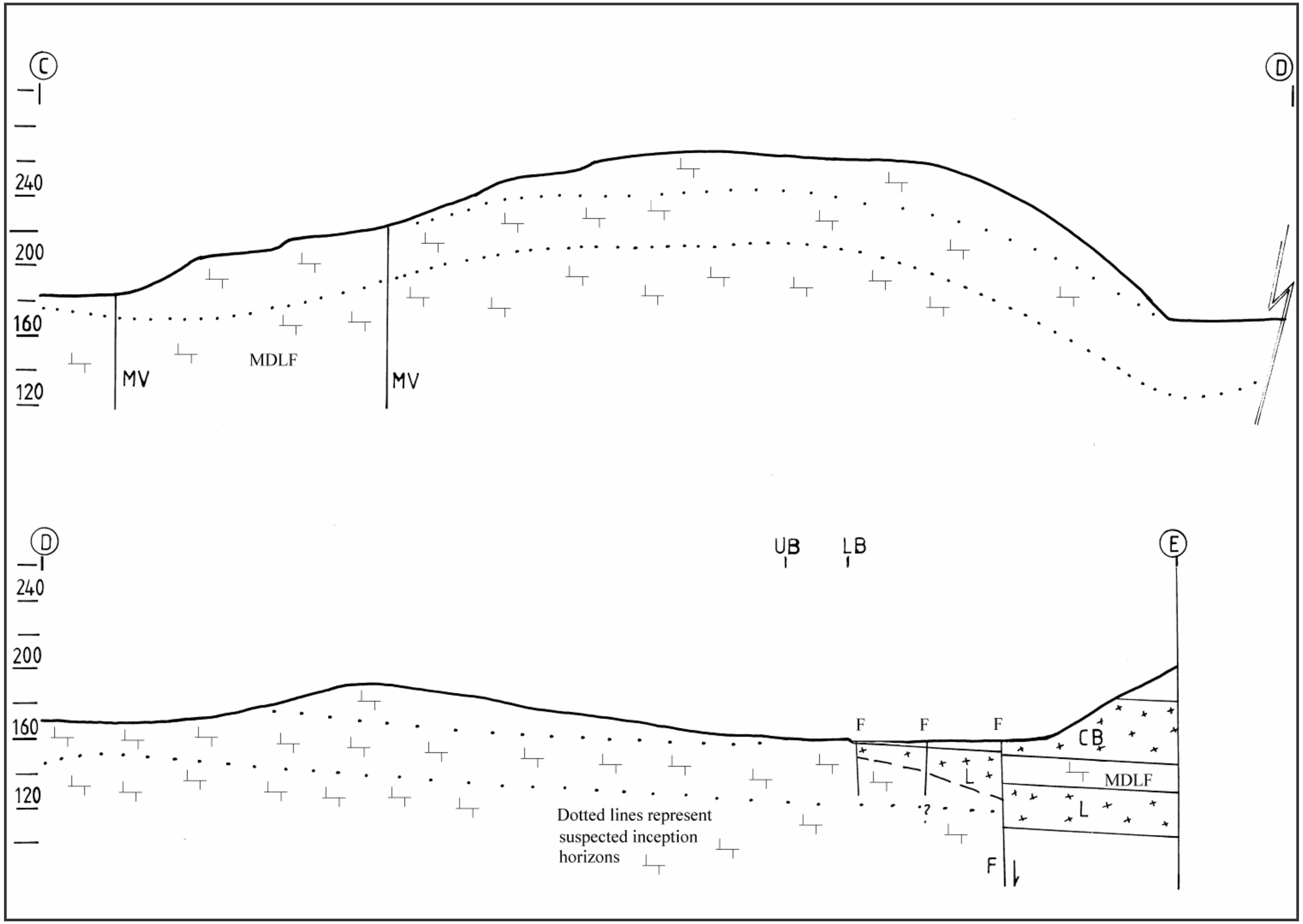


Figure 11.8: Sections C (SK 18256573) - D (SK 19626610) and D – E (SK 20576612) to show level of inception horizon.
Vertical exaggeration x1.7 (approx); Levels m AOD; Line of sections shown on Figure 11.6, continues from Figure 11.7.
Legend: XX Lava Beds; MDLF Monsal Dale Formation. Dotted line= line of inception horizon
Key: CB Conksbury Bridge Basalt; L Lathkill Lodge Lava; F Fault; MV Mineral Vein; UB Upper level of Bubble Springs; LB lower level of Bubble Springs

the close relationship of dry valleys with the mineral veins and the similarity of the network of the dry valleys with the drainage on the Namurian strata indicates inheritance from above, the significance of the inception horizon as guidance to the cave system and in providing the base flow to the river suggest that it is very likely that the river (or glacially related processes) has cut down into a karst system. However, it is interesting to note that the profile (Appendix 11.1) indicates the presence of a tufa barrage immediately upstream of Bubble Springs and therefore erosion of the tufa, as postulated by Towler (1977) is plausible.

The implication of the capture of flow at Bubble Springs is that to the east of Bubble Springs there are likely to be relict inception horizons that guide flow in an easterly direction towards the River Wye. Consideration of potential flow path directions, which are likely to be guided by the dip of the bedding, indicates that some of this flow may be captured by springs in the vicinity of Haddon Hall, for example the springs associated with Wigger Dale (SK 222668) and Fawnsdale Plantation (SK 226664). Unfortunately, this author has not sampled these springs, because they were originally thought to lie outside the area of interest as they emanate from Namurian strata.

Downstream of Bubble Springs, in the lower Lathkill, the gently folded lavas dip beneath the bed of the river to form a syncline and higher beds of the Monsal Dale Limestone form the bed of the river. Immediately to the north of Long Rake the Conksbury Bridge Lava returns to the surface. There are two faults in the vicinity of Long Rake; the throw on the northernmost fault is slight and the Conksbury Bridge Lava forms the bed of the river between the two faults. To the south of Long Rake the Conksbury Bridge Lava is downthrown beneath the bed of the river. Thus the inception horizon that is associated with the tufa formation in the vicinity of Coalpit Bridge (immediately up and downstream of the bridge) is associated with a zone immediately above the Conksbury Bridge Lava, which is particularly fossiliferous in a zone of approximately 12 m (Aitkenhead et al., 1981).

For three consecutive years (2001 to 2003) the summer recession of the upper Lathkill was monitored at six points downstream of Cowgate Pool (Figure 11.10) by the Limestone Research Group (including this author). Discharge was measured at each point at approximately fortnightly intervals between June and September. The discharge was determined using the velocity-area method (Shaw, 1994). An ADS Sensa RC2 electromagnetic flow meter was used to measure the velocity. Inevitably there are problems associated with this type of monitoring, for instance in a deep channel some flow can be occurring within the channel fill, or as considered by Downing et al. (1970, p. 35) "*Measurements of flow in a river with this sort of regime may be misleading because a considerable proportion of the total flow may be underflow in the jointed limestone*". There may also be problems associated with the accuracy of the determination using the method. In order to achieve consistency, on each occasion monitoring was carried out at the pre-determined monitoring point (marked by posts in the river bank) and at each monitoring point, where the stage allowed it, the velocity was determined at the same distances across the channel. On 17 July 2001 repeat measurements were carried out on the same day and the results indicated good reproducibility. Again to maintain consistency, the discharge presented was derived from the mean of the values calculated using the mid-section and mean-section formulae.

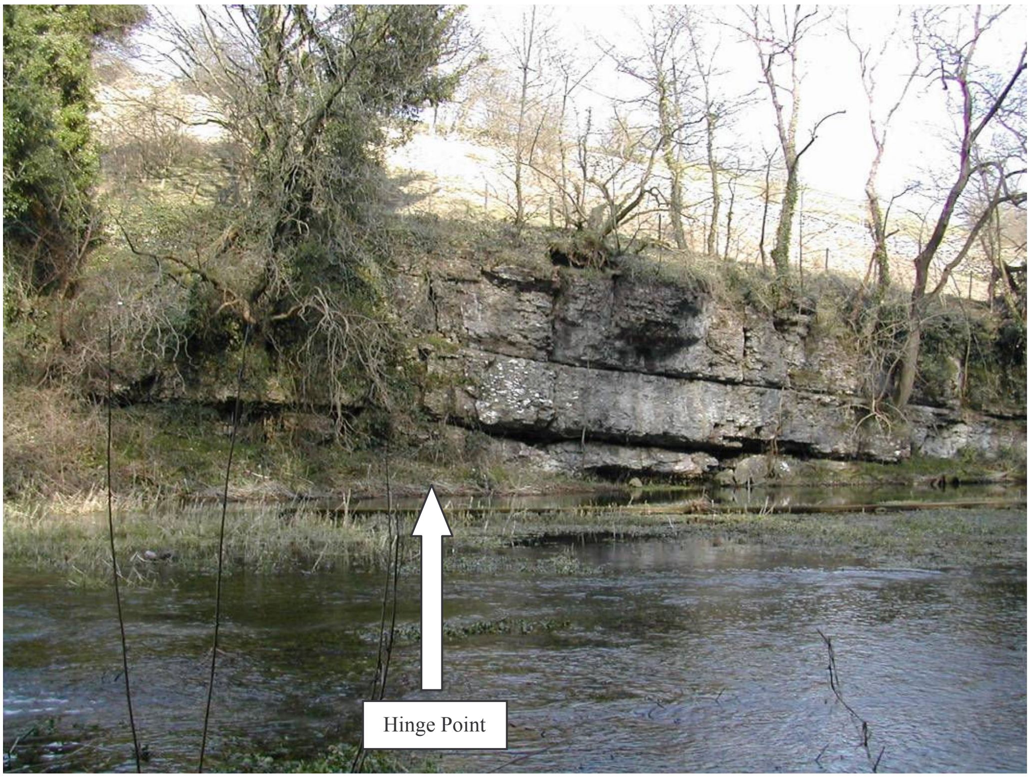


Plate 11.6: Fault zone at Bubble Springs.

The flow monitoring work was reported to English Nature (Natural England) annually by the Limestone Research Group (Gunn, 2002 and Gunn, 2003). The reporting was aimed at improving the rating curve that has been developed by Gunn (2002 and 2003) and establishing the key points of loss of flow to the river, with the ultimate aim of assisting English Nature to design works to provide sufficient flow to maintain the favourable conservation status of the river. One of the main foci for the research is to try and establish the extent of loss of discharge to mine workings that underlie reaches of the river (section 11.7). For this thesis recession curves have been prepared for each of the monitoring points, for each year, the aim being to compare the river recession curves with ground water recession curves. It was the opinion of this author that as the river is primarily fed by ground water the recession curves should be comparable. In the course of this work other interesting points have also come to light.

Discharge was generally found to be higher in 2002 and very low in 2003 (Figure 11.9). This corresponds to higher rainfall figures, as measured at Buxton and provided by High Peak Borough Council (613.2 mm during the period 1 January to 30 June 2001, 746.4 mm in the same period of 2002 and 533.7 mm for the same period in 2003). In chapter 10 it was demonstrated that the baseflow, the lower portion of the recession curve, takes an exponential form. Examination of Figure 11.9 confirms that for the measurements determined 50 m downstream of Cowgate Pool this is best defined in the 2003 data. For the 2003 data the best fit curve was $89.9e^{-0.0154t}$ ($R^2 = 0.74$), which is marginally lower (shallower) than the value ($496.4e^{-0.0248t}$, $R^2 = 0.96$) calculated from the Environment Agency data for the period 2001-2002 (Chapter 10). The curves take a similar form to that determined for the Bull I' Th' Thorne Borehole ($33.2e^{-0.0128t} + 217.3$, derived for the zone 251 to 246 m OD from groundwater levels, rather than discharge, Chapter 7). In 2001 and 2003 the form of the recession curves in the two monitoring points farthest downstream (points 1 and 2) did not exhibit an exponential basal flow and is significantly different to that observed at the four upstream locations, points 3 to 6 (Appendix 11.2). This is largely attributable to loss of groundwater to mine workings (section 11.7). Field observations support this hypothesis, for instance a discharge of 27 litres/sec was calculated to be lost to the river bed, presumably to mine workings below, at SK 18866578, on 27 August 2003, a time when the two downstream monitoring points were found to be dry. Additionally, it is the opinion of this author that the dominant inception horizon lies below the bed of the river at the easternmost of the flow monitoring points (Figure 11.10) and therefore the recession curve is effectively truncated (Figure A11.3.6).

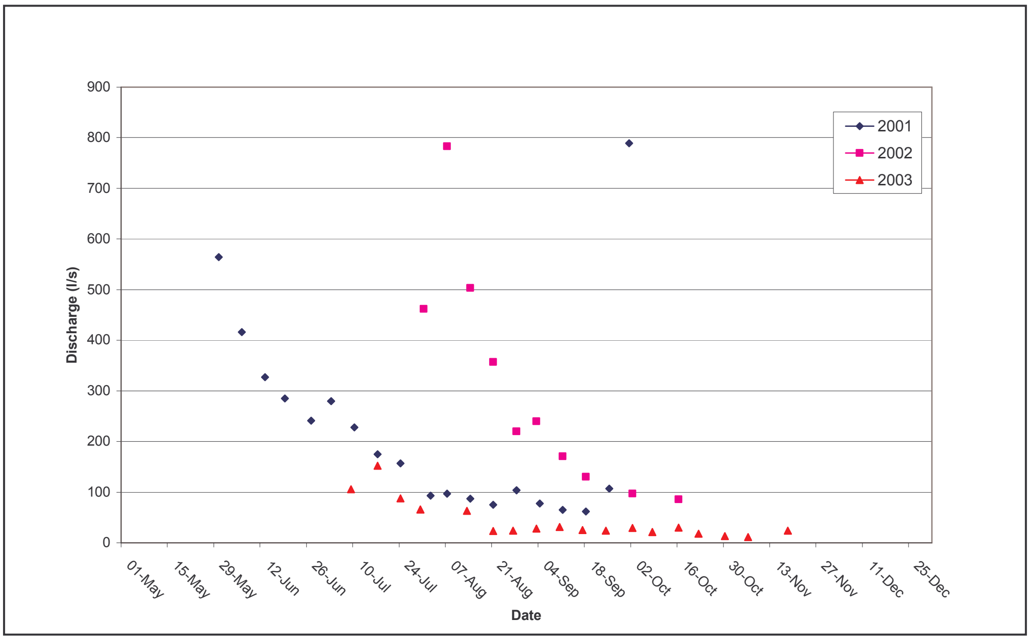
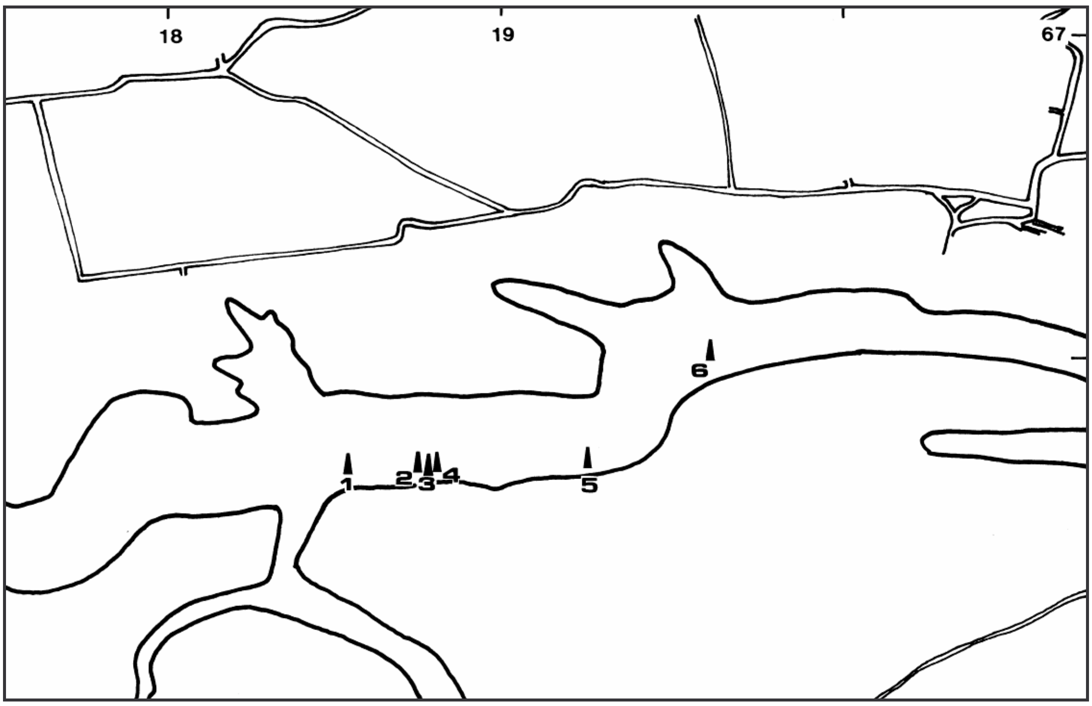


Figure 11.9: Groundwater recession River Lathkill 50m downstream of Cowgate Pool.

It has long been observed that the seasonal recovery of the river takes place over a period of a few hours. Given the findings presented above it would seem plausible that this results largely from significant joint/fissure storage in the limestone. Individually the storage capacity within any single joint is not large, but the combined displacement of water from a large number of joints and fissures ensures that groundwater rises quickly after a significant rainfall event. As the summer ends, evapotranspiration reduces (owing to lower temperatures and shorter sunlight hours), so there is more effective rainfall for a given rainfall event. Once the soil moisture deficit has been replenished water levels in the fissures rapidly rise above the level of the inception horizon(s). When a fall in barometric pressure occurs there is a potential for groundwater to rise on faults. The significance of fracture flow was described in Chapter 9 and is supported by the observation of Oakman (1979, p. 184) that "*the predominant hydrological carrier horizon was the joints themselves*".

The discharge of the River Lathkill increases considerably at the Bubble Springs resurgence (Appendix 11.4 and section 11.8). The results of periodic flow monitoring downstream of Bubble Springs (Appendix 11.4) indicate that there are further losses from the river downstream of Bubble Springs, in particular between Bubble Springs and Raper Lodge, along a stretch in which there is a dearth of mine workings. Pedley (1993) and Pedley et al. (2000) describe the use of geophysics (resistivity) and findings that demonstrate that the Wye and the Lathkill valleys incorporate deeply incised inner gorges, now partly buried, in places beneath 11 to 14 m of sediments. The earliest infills at Raper Lodge and Fin Wood are stiff brown clays, with angular Carboniferous Limestone clasts, suspected to be soliflucted till. More recent deposits include a range of sediments related to the tufa barriers, including: micrite, micritic tufa, clasts of tufa and peat (Pedley, 1993). Thus it is clear that downstream of Bubble Springs a considerable volume of flow may be occurring within the channel fill, or within joints in the limestone, as suggested above. Further evidence in support of this hypothesis comes from the observations of the Haddon Estates gamekeeper, Warren Slaney, described below.

Farey (1811, p. 505) and Glover (1831, p. 22) in lists of notable springs make reference to "*Over-Haddon South (Well), on Clay Wayboard; and SW, large at great Bakewell Fault, in Robinstye Mine*". It is suspected by this author that the latter is the spring that was harnessed to supply the village of Over Haddon. The remnants of the pipework associated with the supply can be found at SK 20786603, situated on the north-south trending fault that is downthrown to the east. It is considered by this author that this fault is the Great Bakewell Fault of Farey (1811). The springs that provided the supply comprised groundwater from inception horizons that lie between the Lathkill Lodge and Conksbury Bridge Lavas.



Flow monitoring points 1-6 (50m, 250m, 280m, 300m, 720m and 1270m downstream of Cowgate Pool respectively, at National Grid References: SK 1857365789, 1875765782, 1880865779, 1881765779, 1918365782 and 1961466098 respectively)

Figure 11.10: Location of flow monitoring points in Lathkill Dale.

The Haddon Estates gamekeeper reported (personal communication, November 2003) that "*I do not believe that there is any loss of water through the fish ponds [fishponds immediately upstream of Coalpit Bridge]. Water coming into the biggest and deepest pond doesn't leave via the designated outlet but through sink holes in the pond bottom. It then follows fissures under the path and part of the river bed and issues up where the big weir meets the pool. Drain tracings have confirmed this*." This indicates that the groundwater sinks on the fault to the north of Coalpit Bridge (Long Rake) and that some of the water rises on the fault to the south of Coalpit Bridge. This observation is significant both in terms of confirming the significance of the faults in terms of groundwater storage and also, because the fault to the south of Coalpit Bridge is not laterally extensive, in helping to explain the considerable difficulties associated with dewatering the Alport mines (Chapter 3), which would appear to receive inception horizon guided groundwater. However, it should also be noted, as suggested by Ben le Bas (personal communication, February 2007), that the loss could simply be due to a failure of the pond lining.

11.6 Hydrogeochemistry.

Christopher (1981) presented median concentrations of a range of elements for which routine analyses were carried out on fortnightly samples taken from a number of springs in the Peak District. The analyses included a number from the Lathkill catchment, which have been tabulated below:

Table 11.2: Analytical data presented by Christopher (1981).

Spring	pH	Ca	Mg	Na	K	Total Hardness	HCO_3	Cl	SO_4	NO_3	Saturation Index	pCO_2
Chelmorton (SK 115703)	7.0	116	2.0	4.0	0.3	330	242	11	32	8.9	-0.18	1.53
Lathkill Head Cave (SK 171659)	7.5	107	1.5	7.3	1.3	336	207	20	31	17.5	-0.17	2.2
Cales Dale (SK 173654)	7.5	109	2.9	10.6	1.2	336	228	30	26	7	0.32	2.26
Carter's Mill (SK 182656)	7.6	88	12	4.5	0.5	289	232	12	20	8	-0.05	2.03
Mandale Sough (SK 197661)	7.7	103	5.5	6.5	0.8	301	228	14	31	9.3	0.56	2.25
Bubble Springs (SK 210659)		97	6.4	8	0.9		218	24	21	6.8	0.29	2.26
Well Head (SK 200634)		99	8.9	5.4	1		227	18	30	15	-0.02	1.94
Gratton Dale Sough (SK 208608)	7.77	84	29.8	3.8	0.6	332	262	16	33	21.7	0.47	2.35
Elton (SK 218614)	7.68	14	7	7	1.5	62	48.4	22	56	6.2	-0.98	2.9
Friden Brickworks Borehole (SK 169608)	7.2	106	15	5	1.5	326	259	13	41	15.5	0.09	1.71

The data appear to provide supporting evidence for the conclusions that were drawn in Chapter 6. The low concentration of calcium, low total hardness and bicarbonate determined at Elton reflect the geological setting, issuing from the Namurian strata, with high sulphate concentrations derived from the pyrite. The relatively elevated chloride concentration determined in the Cales Dale Spring can be attributed to rising groundwater and the association with mineralization (although Christopher, 1981) attributes the elevated chloride to a contribution from road salt). Whilst the relatively elevated concentration of sulphate determined in waters from Friden Brickworks could be attributable to the extensive superficial deposits that cap this area, however, this borehole penetrates the Woo Dale Limestone and its chemistry is typical of the of the Woo Dale Limestone Group (Table 6.2) and this author considers that it is likely to be derived, at least in part, from the Woo Dale Limestone. The slightly elevated concentration of sulphate determined at Chelmorton is attributed to contact with the Upper Millers Dale Lava. Similarly, the relatively elevated concentration of magnesium determined in the groundwater from Gratton Dale Sough, Friden Brickworks Borehole and Carter's Mill Spring is attributed to dolomitization in the case of the former location and leaching from Head deposits in the latter two locations. Elevated concentrations of nitrate are usually associated with anthropogenic inputs; therefore it would appear that Lathkill Head Cave, Well Head Cave, Gratton Dale Sough and Friden Brickworks Borehole were subject to contamination.

With respect to metal concentrations, Salzman (2002) working on sediment samples in Lathkill Dale determined lead concentrations in the very large range of 4 - 28 448 mg/kg, copper concentrations in the range 3 to 102 mg/kg and zinc concentrations of 160 to 1600 mg/kg, with higher concentrations corresponding with areas of mine working and processing. Metal concentrations in groundwater samples were found to be relatively low, with concentrations of lead in the order of 0.03 to 0.08 mg/l, copper in the order of 0 to 0.01 mg/l and zinc in the order of 0.01 to 0.03 mg/l (Appendix 6.1).

As a consequence of trying to interpret the results of dye-tracing experiments (Chapter 7) this author became concerned that the position of fluocapteurs placed at Bubble Springs might not be truly representative if the sampling location was some distance from the spring output. Furthermore, Lathkill Dale Sough tail lies immediately to the north of Bubble Springs and it has been noted that during low groundwater conditions flow at Lathkill Dale Sough tail ceases, whilst Bubble Springs continue to resurge. Furthermore, groundwater continues to flow through the sough at Bateman's Shaft at SK 194658) during low groundwater conditions, which indicates that there is a loss from the sough to Bubble Springs. Thus the area of Bubble Springs was investigated by this author. Groundwater was found to resurge at two levels, the upper level comprising a series of spring fed pools, or hollows (in the order of 0.80 m deep) and the lower level comprised upwelling of water in the river bed, below the zone of pools. It was not known whether each spring discharge has the same groundwater chemistry. Two in-situ monitoring transects (Tables 11.3 and 11.4) were carried out across the upper level springs and in-situ monitoring of the water emanating from an orifice at the lower level was carried out (Table 11.5). pH and specific conductivity determinations were made with an environmental sonde and should be treated with caution, as pH determinations greater than 9 are considered unlikely and the specific conductivity determinations do not correspond with the laboratory derived values. However, it is the relative change in pH and the temperatures above, within and at the base of the pools (or hollows) associated with the springs that are of interest.

The results indicate differences between the spring temperatures and chemistry with those of the upstream river water chemistry. If the river temperature represents the average background temperature it would appear that the water emanating from Bubble Springs has a slightly elevated temperature with a lower pH. The spring water appears to have been evident in the lower pools of the dam (Transect 1), but not in the higher ones. It may be that the activation of orifices migrates upstream as the head increases during higher ground water conditions. There is at least 1 m between the springs in the bed of the water course and those associated with the pools, this also suggests that the springs comprise rising, confined water. It should also be noted that the change in pH within the pool environment could either be attributable to tufa deposition, as described by Towler (1977), or to mixing with the background discharge of the river.

Table 11.3: Results of in-situ water quality monitoring carried out at Bubble Springs on 22 March 2003. Transect 1 at NGR SK 20497 66127.

Distance from the north bank (m)	Observations	Temperature (˚ C)	pH	Specific Conductivity (uS/cm)
	(Water starts at 1.20 m from bank)			
1.60		7.48	8.85	645
2.50		7.48	9.12	645
4.20	At top of hollow	7.55	9.14	645
4.20	At base of hollow	8.25	8.59	654
5.00	Above hollow	7.50	9.04	646
5.00	At top of hollow	7.94	8.66	645
5.00	At base of hollow	8.26	8.21	654
6.60	At base of hollow	7.55	8.96	645
7.50		7.55	8.99	645
8.00		7.56	8.84	645
9.80	Dry			
12.00	Wet			
12.50		7.50	8.95	644
12.90	Dry			
13.60	Wet			
14.60		7.45	8.98	644
16.60	Dry			

The field observations made at Bubble Springs are particularly important in that they support the conceptual model (Chapter 9). More specifically the following can be seen at this location:

- The significance of faulting in imposing storage within the inception horizons behind the fault
- Groundwater mixing on fault zones. It would appear that the tufa deposits are associated with higher-level inception horizons in the vadose zone, with lower inception horizons discharging via the orifices below the tufa. Groundwater rising from depth appears to have a lower pH and is potentially more aggressive with a potential for cooling as it rises.

Table 11.4 : Results of in-situ water quality monitoring carried out at Bubble Springs on 22 March 2003. Transect 2 at NGR SK 20511 66115.

Distance from the north bank (m)	Observations (water starts 1.1 m from bank)	Temperature (˚ C)	pH	Specific Conductivity (uS/cm)
1.80		7.79	8.21	646
3.30		7.74	8.65	646
4.60	Above hollow	7.74	8.75	645
4.60	At top of hollow	7.74	8.79	645
4.60	At base of hollow	7.74	8.81	645
6.20	Above hollow	7.75	8.84	645
6.20	At the top of the hollow	7.74	8.84	644
6.20	At base of hollow	7.75	8.84	645
7.50		7.75	8.85	645
8.50		7.76	8.84	644
12.90		7.64	8.86	635
15.3		7.57	8.87	642
17.60		7.48	8.83	636
17.70	Dry			

Table 11.5 : Results of in-situ water quality monitoring carried out at Bubble Springs on 22 March 2003. Spring at NGR SK 20511 66105.

Spring at lowest level of Bubble Springs	Observations	Temperature (˚ C)	pH	Specific Conductivity (uS/cm)
	Dark grey deposit around orifice	8.20	7.3	651

Further groundwater sampling and analysis is required to confirm that water chemistry at these locations.

11.7 Human impacts.

Human modification to the River Lathkill has been considerable and has taken a number of forms. Water has been impounded for milling, e.g. Carters Mill and Over-Haddon Mill (HM, Figure 11.11); modifications have been made to improve the habitat for fish: leats were constructed to activate water-wheel drainage to the mines and from the 1700s a number of soughs, in particular Mandale and Lathkilldale soughs, were constructed (Rieuwerts, 2000). Extensive research into the mine workings of upper Lathkill Dale has been carried out by Rieuwerts (2000). Figure 11.11 shows the extent of known mine workings in upper Lathkill Dale (descriptions in Appendix 3.4). However, this work is not exhaustive and preliminary searches of aerial photographs (Appendix 3.5) confirmed that the workings are more extensive. This is also the finding of a comprehensive inventory and maps to show the extent of regionally and nationally important lead mining sites in the Peak District Orefield was published by Barnatt and Penny (2004), which followed an aerial photograph study carried out as part of the Lead Rakes Project funded by English Heritage (Barnatt, 2000). Mining in Lathkill Dale dates back at least as far as the thirteenth century. As elsewhere, improvements in technology meant that the depth of mining could gradually be increased, for example Mandale Rake, between Winchester's Shaft (SK 1832366630) and Haddon Grove (SK 1789866929) is known to have been initially worked opencast to at least 6 to 8 m depth, but by 1815 some of the workings had been extended to 91.5 m depth (Rieuwerts, 2000), a level of approximately 185 m OD.

The impact of human activities on the flow regime of the River Lathkill has intrigued hydrologists for a number of years (for example, Bamber, 1951; Beck, 1980; Emmett, 1984; and Oakman, 1979). It has been suggested that historically the surface water flow may have extended further up Bagshawe Dale. Although the geomorphology of Bagshawe Dale suggests that it is a continuation of Lathkill Dale, the enclosure map shows that boundaries to fields that cross Bagshawe Dale date to at least 1776. This area lies on the edge of Monyash and there does not appear to be any reference to water in the field names, which suggests the absence of surface flow at that time and it is considered by this author that continuous surface flow in Bagshawe Dale is likely to pre-date documentary evidence. Essentially it has been argued that the mining activities have caused a gradual drying of the river. However there has been little quantification of this effect and map searches have proved ambiguous. Bamber (1951) spoke to people who could recall: i) when water flowing along the entire length of the river from Lathkill Head Cave throughout the year (dated to about 1880); and ii) evidence of a strong flow at or near Carters Mill as late as 1940. Furthermore Bamber (1951, p. 294) noted that "*as recently as 1929 the river flowed strongly at all times from a point just a little higher than Calling Low Dale*" [likely to be Pudding Springs]. The latter observation is contradicted by a copy of a letter dated 30 September 1929 from Mr G E Nuttall (fishing tenant) to Mr Pearce at the Estate Office, Melbourne. This letter (located by Ben le Bas and Philip Bowler in the Estate Office Records at Melbourne) described drought conditions in the year 1929 and described that the reach above Carter's Pool dries up every summer. However it confirms point i) above and supports the hypothesis presented below that attributes the fall in groundwater levels in about 1880 to the gradual drying of the River Lathkill. Bamber (1951) and Nuttall (op. cit.) observed that in times of drought the river is dry down to Over Haddon, with the exception of very weak flows near Cales Dale and Calling Low Dale. This could equally describe the situation today. Indeed, this was the situation observed over the summer of 2001, albeit that this was not a particularly dry summer. Furthermore, an inscription in Lathkill Head Cave dating to 1871 suggests that at times it was dry enough to explore even then.

Perhaps the most convincing evidence in support of a fall in the groundwater level in the headwaters of the River Lathkill was the loss of supply to the springs that once met the water requirements for the villages of Monyash and Flagg (Appendix 3.3). There are a number of reasons that could account for the regional fall in groundwater levels, which include: changes in effective rainfall; speleogenesis causing down cutting; lead mining activity taking groundwater to lower levels and under drainage by soughs (including Hubberdale Mine [SK 14076981, Appendix 3.4] and Magpie Sough). Bamber (1951) and Edmunds (1971) suggested the possibility that Magpie Sough was capturing some of the flow of the River Lathkill. Worley and Ford (1977a, p. 161) suggested: "*The dry course with intermittent flow of the River Lathkill is probably due in part to the fall in water-table caused by the Lathkilldale Sough*". Others, such as Beck (1980) and Oakman (1979) have suggested that flow loss to

Hillcarr Sough is more significant. Documentary evidence with respect to potable water supplies (Appendix 3.3) suggests that the 'drying up' of the springs was a gradual process (section 3.6), which has been correlated with sough construction (Table 11.6). This provides supporting evidence of the significance of the impact of the construction of Magpie Sough on the catchment (Chapter 10, Bamber, 1951).

Bamber (1951), drawing on information collected by the North Midlands Group of the British Speleological Association presented groundwater levels for the period 1946 to 1950, to serve as a base-line for future studies. The levels recorded for various points along the River Lathkill and in soughs are comparable with those recorded during the survey carried out by this author (Appendix 11.1). The level recorded at Knotlow Mine is comparable with the level of the groundwater contour recorded by Downing et al. (1970). Thus there does not appear to have been any significant regional lowering of groundwater levels since the 1950s, indicating a maximum post sough construction time lag of in the order of 45 to 50 years. The springs in upper Lathkill Dale have shown a minimal response to the recent blockage of Hillcarr Sough. Yet, there was an immediate response to the driving of Magpie Sough in Townhead Vein (Chapter 10), providing additional evidence for underflow maintaining the head in dominant faults. Willies (1980) described a collapse in Magpie Sough in 1962, followed by an 'explosion' as water broke through the blockage in 1966. Willies (1980) also noted that there was another blockage that was cleared by the Peak District Mines Historical Society in 1974. However, to date this author has not located any documentary evidence describing the impact of the blockages on the hydrogeology.

Whilst this author agrees with Beck (1980) and Oakman (1979) that the groundwater collected by Hillcarr Sough is largely fed by the inception horizon that feeds the upper Lathkill, preliminary discharge calculations (in need of further measurement) indicate that the groundwater resurging at Bubble Springs broadly corresponds with the upstream recharge potential, as indicated from the topographic catchment area, suggesting that bedding guides the inception horizon related flow to the axis of the syncline and that the fault at Bubble Springs acts as a groundwater divide. This indicates that Hillcarr Sough does not capture water from the area of the upper Lathkill, although clearly the downstream reduction in head resulting from the construction of Hillcarr Sough must be distributed across the catchment, albeit that it may not be measurable. For example, the results of the Limestone Research Group flow monitoring do not appear to show any significant impact as a result of the present blockage in Hillcarr Sough. However, it is clear from the results of the flow monitoring carried out in the lower Lathkill, that downstream of Bubble Springs and probably more specifically downstream of Conksbury Bridge, there is further loss to the bed of the river (section 11.5). Oakman (1979) observed that the problem of water from the rivers Bradford and (lower) Lathkill between Youlgreave and east of Alport sinking into Hillcarr Sough was solved in the late C18th and early C19th by clay puddling and he suggests that the annual drying up of the River Lathkill may in part be due to this leakage. Christopher et al. (1977) and Emmett (1984) also suggested that part of the bed of the watercourse was lined with puddled clay in order to minimise the loss of surface water to the mine workings. However, this is not the experience of Gunn and Dykes (2000) in upper Lathkill, although it is feasible that puddled clay was used in lower Lathkill to minimise groundwater ingress to the Alport mining field. Rieuwerts (2001) noted that Danger Level reached Coalpit Bridge, which suggests that there is a potential drainage route to Hillcarr Sough via Danger Level (Figure A3.4.8).

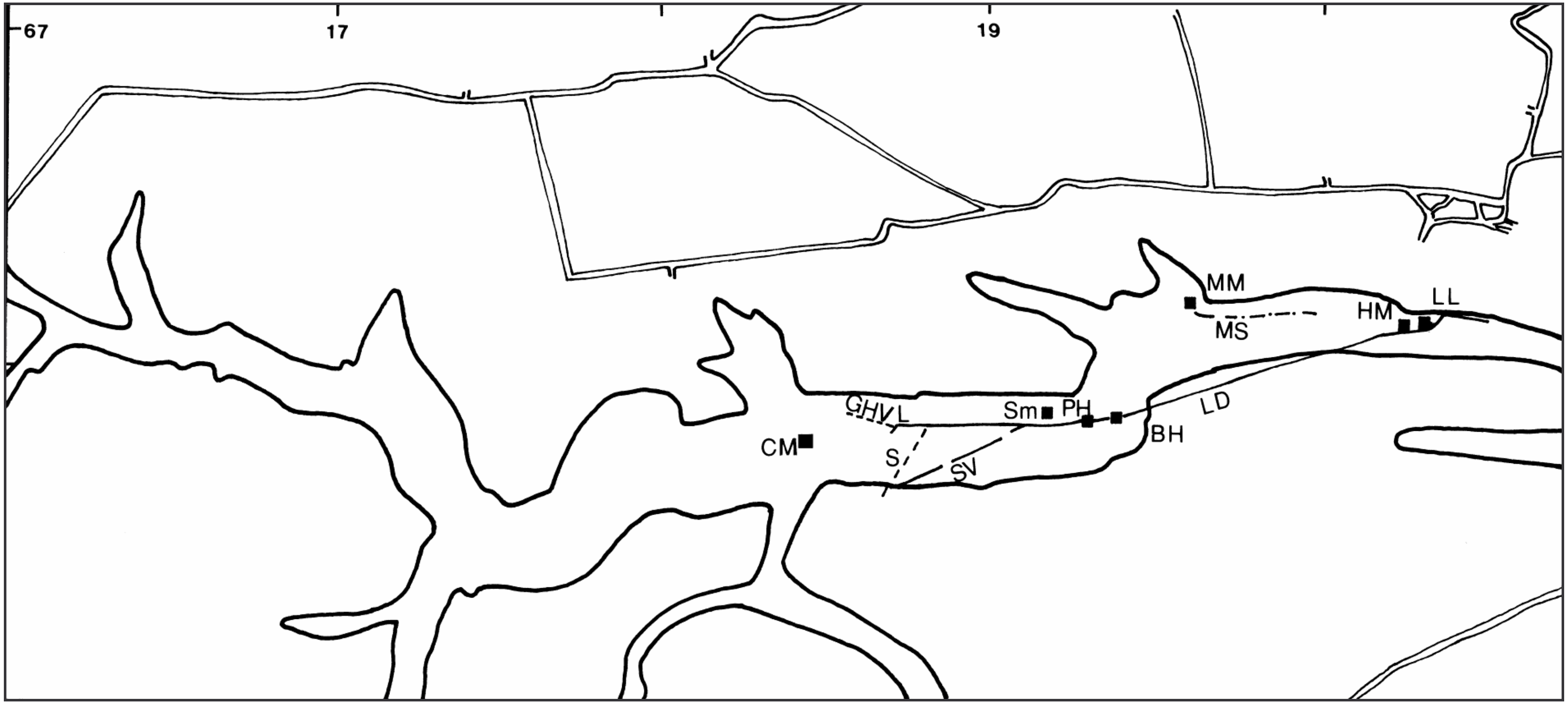
With respect to Lathkilldale Sough, it would seem that the effectiveness of the sough was dependent on the capture of inception horizon related flow. Rieuwerts (2000, p. 67) describes how "*local information relates how a great lake of water was met in Lathkill Dale Mine*". The date, or depth at which this occurred is not presented, but it does suggest that the inception horizon was captured. Further evidence comes from flow monitoring carried out by David Webb (Peak Mines Historical Society) and Jon Humble (English Heritage) beneath Bateman's Shaft on 24 September, 2003, when a discharge of 44 litres/sec was determined in Lathkilldale Sough, immediately upstream of Bateman's shaft, on an occasion when the River Lathkill was dry immediately upstream of this point as far as approximately SK 1892965800. Downstream of this point the discharge in the river increased from 4 litres/sec at monitoring point 3 to 15, 20 and 24 litres/sec at monitoring points 4 to 6 respectively (Figure 11.10). These results would appear to confirm that the dominant inception horizon lies beneath the bed of the river at this location. Clearly there is a loss of flow from the surface, this may in part be a natural occurrence, but it has undoubtedly been exacerbated by the mine workings. It has been observed that the discharge from the hypothesised location of Lathkilldale Sough Tail recedes very early in the seasonal groundwater recession; such that it cannot discharge groundwater from the mine workings during the seasonal recession. Further evidence of this comes from the flow monitoring carried out on 5 April, 2003 (Table 11.7) described below.

The flow monitoring was carried out using the procedures described in section 11.5. Positions of the monitoring locations have been marked on Figure 11.10. If all the flow loss that is implicit in the results of the monitoring was occurring to the mine workings the sum of the discharge at Lathkilldale Sough and the monitoring point upstream of Lathkill Lodge should approximate to the discharge 1270 m down stream of Cowgate Pool. That it does not, further suggests that at some point downstream of Bateman's Shaft the sough water is 'captured' by the natural drainage. Similarly, Mandale Sough dries up before the mine does.

There are three possible reasons that groundwater captured by Lathkill Dale mine does not emerge at Lathkill Sough tail. Firstly it may be attributable to the brick walls that were constructed in the sough both upstream and downstream of Bateman's Shaft. Although it is reported that these walls have fallen into disrepair (Roberts, 1998) English Nature commissioned a survey (TL Excavations, 2006) of Powder House Shaft and this established that the standing water level in the Powder House was in the order of 7.5 m higher than in the sough at Bateman's Shaft, which suggests that the intervening wall and sluice, suspected to be near the bottom of the Powder House Shaft is still holding back water. Secondly, it is possible that a component of 'self-sealing' has occurred in the sough. Rieuwerts (2000, p. 32) quoted the following extracts of a description of human remains, from the *Philosophical Transactions of the Royal Society,* Volume 43, 1744-45, pp. 266-267 (a letter dated May 1744): "*at Lathkill Dale … as the Workmen were driving a Sough or Drain to a Lead Mine, about 9 yards deep* *and 40 fathoms from the beginning of the Sough …. The place where these things was found is on every side surrounded with a rocky petrified substance of Terra Lapidae, by the miners called Tuft, so hard (as they say) to strike fire against their tools. The substance lay above the Bones …. A yard and a half thick and on either side … there being a soft coarse clay or Marl interspersed thick with little petrified Balls of the same kind of substance as the Tuft*." Although Rieuwerts opined that the material described is more likely to be lava, it seems more plausible to this author that the material is actually tufa, because this location lies very close to Bubble Springs and that it 'strikes fire against their tools' probably indicates silicification. Association with the shell bed and the occurrence of marl could be indicative of wayboard material.

Table 11.6: Correlation of sough construction with observations regarding potable water supply.

Date	Sough construction (Appendix 3.4)/Mining detail	Level of sough tail (m OD)	Observation regarding potable water supply (Appendix 3.3)
1642/53	Hardyhead Sough		
Prior to 1735	Whale Sough	250	
1743	Black Sough, Alport	102	
1743	Lathkilldale Sough	158	
1747	Crotie Sough		
1766 – 87	Hillcarr Sough	96	
1780			Drought many springs dried up (Glover, 1831)
1797	Mandale Sough	167	
1851			Mr Sutherland writes to Mr Wyatt requesting use of mine water to supply Flagg (Appendix 11.5)
1873 – 1881	Magpie Sough		
1910			Farrar observed that water for Flagg was obtained utilising a windlass at the bottom of a lead mine shaft
1914			Munro observed that: *Flagg and Monyash supply by standpipe. Monyash ample supply by springs. Flagg supply from old mine water well, inadequate.*
1914			*Shacklow Wood Springs provided an unlimited supply,* but Willies (1980) describes that the springs dried up as a consequence of the driving of Magpie Sough.
1921			Drought, many springs dried up
1926			New pump installed, Flagg (45.70 m to water)
1935			Monyash supplied from Flagg
1940			Carter's Mill Spring (SK 182 657) a strong flow until c. 1940 (Bamber, 1951)
1949			New pump installed, Flagg; extended to 54.90 m to reach water (Bamber, 1951)



Key: BH Bateman's House; CM Carter's Mill; GHV Gank Hole Vein; HM Haddon Mill; L Entrance to Lathkill Level 1884; LD Lathkilldale Vein and Level; LL Lathkill Lodge; MM Mandale Mine; MS Mandale Sough Tail, PH Powder House; SV Sideway Vein; S Sough; Sm Smithy

Figure 11.11: Principal mine workings in Lathkill Dale.

The third possibility relates to the geological structure. Rieuwerts (2000) described how bellies or boxes of ore, which tended to be the location of the great ore strikes (1820 and 1823) in Mandale Rake, were associated with 'Blackstone', which is a dark coloured, bituminous limestone, i.e. the Lower Lathkill Beds (Shirley, 1957). These beds immediately underlie the inception horizon described in section 11.3. Rieuwerts (2000) identified that these beds are encountered in Bateman's Shaft. Rieuwerts (2000) also noted that the 52 feet (15.85 m) diameter water wheel raised water at 4000 gallons/minute (304 litres/sec) from 120 feet (36 m) beneath Lathkilldale Sough. This corresponds with a level of approximately 140 m OD), marginally below the level of Magpie Sough. The closest of the Limestone Research Group monitoring points to the site of the former wheel is monitoring point 5. The maximum surface water discharge determined at monitoring point 5 in 2001 was in the order of 450 litres/sec (October 2001), with a mean discharge of 74 litres/sec (determined from the results of periodic monitoring during the seasonal recession in 2001). This provides further supportive evidence of the problems associated with dewatering the mine workings at depth and of the significance of natural flow paths beneath the level of the River Lathkill.

Table 11.7: Flow monitoring carried out on 5 April, 2003.

Date	Monitoring Location	National Grid Reference:	Discharge (litres/sec)
5 April, 2003	50 m downstream of Cowgate Pool	SK 1857365789	284
	250 m downstream of Cowgate Pool	SK 1875765782	285
	720 m downstream of Cowgate Pool	SK 1918365782	150
	1270 m downstream of Cowgate Pool	SK 1961466098	153
	Upstream of Lathkill Lodge	SK 20270366130	42
	Bubble Springs	SK 2051166105	1095
	Lathkill Dale Sough	SK 2050566124	24

Two levels are thought to exist in Mandale Sough, the lower one with a connection to Lathkilldale Sough [Tune, 1969] post dates Lathkilldale Sough (Table 11.6). This provides further evidence of a need to dewater a second groundwater regime. Although Lathkilldale Sough captured the inception horizon related flow from the south and the axial drainage from the west, it would appear that to the north the inception horizon related flow is captured by the northwest to southeast-trending mineral veins.

It should also be noted that Bamber (1951) observed that there had been a weakening of the discharge from the springs at Carters Mill. It is the opinion of this author that this could be a consequence of dewatering associated with calcite mining on Long Rake.

11.8 Underflow in upper Lathkill Dale.

Developing the themes of Chapter 10 it is appropriate to consider the extent of underflow leaving the Lathkill sub-basin. However, flow monitoring at Bubble Springs, using the techniques that are available to the Limestone Research Group, is not practical on the grounds of Health and Safety during high groundwater conditions. Using the measurement determined on 5 April 2003, the following crude assessment of underflow has been made.

Water emerging at Bubble Springs had a temperature of 8.20˚ C, on 22 March 2003 (Table 11.5), which taking a geothermal gradient of 15˚ C/km and a background water temperature of 7.7˚ C, indicates groundwater resurgence from a depth of approximately 30 m. The topographic catchment of the River Lathkill upstream of Bubble Springs has been calculated as 37.85 km^2 (using the FEH CD-ROM) and the effective rainfall for March 2003 as 56.9 mm (calculated from meteorological data for Buxton supplied to the Limestone Research Group by High Peak Borough Council). Taking Bubble Springs as the main resurgence point for the upstream catchment (although it is known from dye-tracing that a component of the recharge at Flagg and Chelmorton flows to Magpie Sough and Ashwood Dale Resurgence and visual observations have identified that untraced water also leaves the catchment via Hardyhead Sough) a discharge of 804 litres/sec would be anticipated. The measured discharge on 5 April 2003 was 1095 litres/sec. Taken in isolation this value suggests that either the catchment of Bubble Springs is larger than calculated, or that rather than underflow leaving the Lathkill sub-basin, Bubble Springs captures an underflow component. The evidence of the groundwater chemistry (section 11.6) did not identify evidence of a deeper groundwater contribution, furthermore it is not realistic to rely on the calculation presented above, because it the time lag associated with the response at Bubble Springs has not been, and cannot be, established without further discharge monitoring at this location.

Gunn (personal communication, 2006) has taken a different approach. By applying his stage discharge curve to the Environment Agency stage readings at the logger site at monitoring point 2 (Figure 11.10) he was able to calculate the volume passing the logger site in March 2003 (1644005 m^3), which using the same value of effective rainfall (56.9 mm) indicates a catchment area of 28.89 km^2. The actual catchment area (determined using the FEH CD-ROM) is 34.34 km^2. This suggests that there is a net loss (16%) upstream of the logger site. This has been interpreted as the net loss from the catchment to Magpie Sough. Whilst this is clearly a more reliable method of calculation than the calculation presented above, it does not account for inception horizon related flow that occurs beneath the river. For instance it is interesting to note that a 16% loss was recorded between monitoring points 1 and 2 (Figure 11.10) on 22 September, 2003, with a reduction from 24 to 20 litres/sec, on the occasion when the discharge in Lathilldale Sough immediately upstream of Bateman's Shaft was determined (section 11.7). Although there is no evidence to substantiate such a loss on 5 April (Table 11.7) it is interesting to note that there was no significant increase in discharge. Furthermore, during 2003 increases in discharge between monitoring points 1 and 2 were only recorded on four of twenty-one visits (Appendix 11.4). It is the opinion of this author that the loss to Magpie Sough is less than is assumed in the above calculations and that further monitoring is required at Bubble Springs in order to assess this.

In addition to the potential loss to Magpie Sough, this author also opines that there is a potential for additional underflow loss to the River Wye via easterly, inception horizon-guided flow paths, associated with the former course of the upper Lathkill (section 11.5). Further exploration of this idea could be carried out by looking at the geochemistry of the springs in the vicinity of Haddon Hall and flow and temperature monitoring in the River Wye.

Although it has generally been found (Palmer, 1987) that vadose ground water flow in limestone has a tendency to follow the dip of the limestone and phreatic flow follows the strike it is interesting to note that in Lathkill Dale, as in much of the Wye catchment, the phreatic flow, also appears to follow the dip of the limestone. Indeed, by virtue of inception horizon-related guidance, the bedding appears to impose a quasi-artesian trapping effect (Ford and Ewers, 1978). Furthermore, it is clear from the descriptions of inception horizon-guided flow above, that much of the groundwater of the upper Lathkill catchment is conveyed to the mining field of Alport and this goes some way to explain the significant groundwater problems that had to be overcome in the exploitation of the mineral in this part of the catchment, ultimately relieved by the construction of Hillcarr Sough (Chapter 3).

11.9 Conclusions regarding the hydrogeology of Lathkill Dale.

Observations made in Lathkill Dale appear to support the conceptual model presented in Chapter 9. Evidence for inception horizon-related flow paths in the Monsal Dale Limestone Formation has been presented, together with the evidence of flow paths associated with inception horizons at a range of depths. This is in keeping with the concept of tiered flow paths (Worthington, 1991). The synclinal setting guides the inception horizon-related flow to the axis of the syncline; hence the river is also associated with the axis of the syncline. The predominantly branchwork form (apparently localised speleogenesis) of the caves is in keeping with the description presented in section 4.6.

Evidence for the association of tufa deposits with specific inception horizons in the Monsal Dale Limestone Formation and the Miller's Dale Limestone Member is exemplified by the occurrence of tufa associated with closed flow paths in the Monsal Dale Limestone at a number of locations in Lathkill Dale.

Interpretations of dye-tracing test results (section 7.2.5) support the concept that the synclinal structure encourages the focusing of water perched in the northwestern part of the catchment by the Upper Miller's Dale Lava in the area of Knotlow Mine. Evidence from the Illy Willy Water dye-tracing experiment suggests inception horizon-related flow paths in the Miller's Dale Limestone Member of the Bee Low Limestone Formation are important in the northwestern part of the catchment and that groundwater probably rises into the Monsal Dale Limestone Formation via the mineral veins, in particular at Knotlow Mine.

The significance of groundwater storage associated with some fault zones and mineral veins is evident both at Bubble Springs and from the documentary evidence of significant volume of groundwater encountered in Townhead Vein during the cutting of Magpie Sough (Willies, 1980). The upward flow of groundwater in Knotlow (section 7.2.5) appears to support the hypothesis of the low permeability and resultant groundwater barrier associated with the mineral veins. Evidence for groundwater mixing (albeit limited and in need of further sampling and analysis) focused on the fault zone at Bubble Springs provides supporting evidence for the concept of significant groundwater mixing in fault zones.

Successful completion of Magpie Sough resulted in a lowering of the head associated with upper Lathkill Dale and reduced the perching effect of the Monsal Dale Limestone Formation that is seen in the Monsal Dale Limestone farther to the north (Chapter 9). The reduction in head was in the order of 50 to 60 m at Magpie Mine (Chapter 10). The reduction in head in the headwaters of the upper Lathkill is likely to have been significant. However the reduction in head would not have been evenly distributed through the limestone aquifer because it is partially maintained by the storage associated with the faults and mineral veins and by the low permeability and storage afforded by the Monsal Dale Limestone. Hence although the springs in Netler Dale were reduced to the extent that they could no longer offer a potable supply, they still flow intermittently. Accordingly, the actual reduction in head is hard to quantify and further searches of records relating to potable water supply could be beneficial in constraining this more closely. The construction of Mandale and Lathkilldale soughs has also resulted in increased permeability and reduced storage in the Monsal Dale Limestone Formation. Descriptions of Lathkilldale Sough indicate that because the soughs intercept deeper inception horizons the mine workings have increased the degree of connection between the inception horizons, thereby encouraging loss of surface water to deeper inception horizons, such that a greater proportion of flow occurs as baseflow.

Groundwater problems associated with mining in lower Lathkill Dale were even greater than those in upper Lathkill Dale. In part this is attributable to the perching effect of the Fallgate Formation, which is increasingly extensive in a southeasterly direction. Groundwater levels in the lower Lathkill were significantly reduced by the construction of Hillcarr Sough. Supporting evidence for this comes from the knowledge that at the time of the preparation of this thesis the sough was known to be blocked and groundwater levels were known to be rising. This author is not aware of any measurable impact of this on the upper Lathkill, which reflects that fact that the river is effectively an overflow spring, although it receives a baseflow contribution.

Chapter 12: Conclusions and potential future research directions.

12.1 Conclusions.

Research carried out in the course of preparing this thesis has further demonstrated the complexity of limestone aquifers. The conceptual model of triple porosity, as defined by Worthington and Smart (2004), the inception horizon hypotheis (Lowe, 1992) and a detailed assessment of the geology underpin the conceptual modelling that forms the main achievement of the thesis (Chapter 9). Whilst it is appropriate to apply the triple porosity model to the limestone of the White Peak, it could equally be argued that the limestones of the White Peak actually demonstrate quadruple porosity, because it is clear from the findings of Chapters 5, 6, and 7 that mineral veins and faults are very important in the functioning of the aquifer. The fault and mineral vein storage has been shown to be particularly important in influencing head distribution.

The conceptual model that has presented stems from an investigation of the karst hydrology in its structural and geological setting. It has been derived from the results of: tracing experiments, chemical analysis of spring water, and hydrograph analyses. Hydrogeological units defined in the model reflect formational influences on the hydrology. The influences are evident in terms of recharge, through-flow and resurgence.

During the course of the preparation of this thesis the author has attended various meetings where discussion regarding the special properties of karst aquifers and the relative contribution of fracture porosity continues. Response to stress history is one of the potential differences between a karst aquifer and a fissured rock aquifer. Unlike other hard rock aquifers, limestone is not dependent upon stress relief alone to facilitate the development of flow paths, instead flow path are largely formed by dissolution; indeed the significance of deep flow paths has been focused upon in Chapter 5. Notwithstanding this, the stress history has been shown to be important in the evolution of the limestone aquifer of the White Peak. Indeed, the stress history has been shown to have influenced: the evolution of fault blocks (Chapter 2), the regional hydrogeology (Chapter 5), the development of inception horizons (Chapter 4) and the karst processes associated with the Chee Tor Limestone Member (Chapter 9). In addition to considering the potential for limestone dissolution, the significance of geological formational differences and the existence of dissolutional constraints, for example localised silicification, have been identified and described in the thesis.

Characteristic of the limestone of the White Peak are relatively shallow angles of stratal dip. In this setting inception horizon (Lowe, 1992) focused conduit development has been demonstrated to be significant in the limestone of the Wye catchment. The identification of inception horizon-related conduit flow was fundamental to the development of the conceptual model (Chapter 9) and is a key to understanding the geological formational differences. The geological formational differences that have been identified have enabled the definition of the hydrogeological units that form an integral part of the conceptual model (Chapter 9).

Debates regarding the concept of the 'water-table' in karst aquifers, for example Lowe (1992) and Worthington (1991) have been shown to be of particular relevance in the context of the White Peak. The White Peak appears to show a particularly steep hydraulic gradient (1/100 m) when compared with the regional gradient (1/1000 m). Perching of groundwater in the Monsal Dale Limestone Formation is such that exploratory boreholes encounter a number of groundwater strikes, some of which are confined, and it is difficult to establish the true groundwater level. The range of seasonal recession in the limestone indicates that stretches of the River Wye are perched for at least part of the year. In particular the area of Monks Dale (SK142733) has been identified in this context. It has also been suggested that this helps to explain the reduced number of springs in the stretch of the river between White Cliff Spring and Lees Bottom. It has previously been observed that the interbedded lavas are largely responsible for the perching of groundwater (Downing et al., 1970). In the context of this thesis it has also been noted that extensive silicification appears to limit dissolution and this impedes drainage, i.e. imposing hydraulic constraints and thereby forcing groundwater storage and elevated groundwater levels.

In addition to the conceptual model of the hydrogeology and the assessment of the geology a number of previously unpublished ideas have been presented regarding the geology. In particular, further consideration has been given to the variety of depositional environments of the chert. The close association of chert with the Monsal Dale Limestone and Eyam Limestone formations, with potential formational processes linked to both mineralization and to silica release from clay wayboards have been hypothesised. In this context the significance of the chert deposits in the area of Bakewell has been considered and a potential source of silica, derived from an hypothesised Devonian basin (based on the Bouger anomaly low, Aitkenhead et al., 1985) has been identified. A second contribution to the understanding of the geology of the White Peak has been the identification of the close association of tufa deposits with the outcrop of strata associated with Unit 4 of the conceptual model (Miller's Dale Limestone Member, Monsal Dale Limestone Formation and Eyam Limestone Formation). Consideration has also been given to the possibility of the existence of another intra-shelf basin centred on the area of Wardlow Mires (Chapter 7). Observations regarding the distribution of glacial till and its apparent absence from the Carboniferous Limestone have been made in Chapter 9.

Examination of inception horizon-related conduit development during the course of this research has identified potential speleogenetic associations and processes (Chapter 9). Thin sections prepared from samples taken across an inception horizon in the Eyam Limestone Formation and another in the Miller's Dale Limestone Member of the Bee Low Limestone Formation identified the significance of micro-stylolites, dolomitization and dedolomitization associated with the occurrence of clay wayboards in inception horizon formation. Speculation regarding the importance of inception horizons in the solution hollows associated with pockets of Head deposits has been presented in Chapter 4. It is suspected that dolines are associated with the solution hollows. Literature reviews and field evidence have led this author to speculate about the significance of stylolites in guiding inception horizon-guided conduit development in the Woo Dale Limestone Formation. The results from dye-tracing tests and the distribution of known caves and springs have shown how the Chee Tor Member is significant as an aquitard at depth, but exhibits high permeability as a consequence of dissolution along stress relief fractures at, or close to surface. Geological understanding has also been used to try and understand the distribution of recharge and discharge points from the aquifer (Chapters 4 and 7).

Research carried out in the context of understanding the regional hydrogeology (Tóth, 1963) has further developed the work of Gunn et al. (2006) and the conceptual model for the regional hydrogeology presented as Figure 5.5 is a significant contribution to understanding the distribution of the thermal springs. The chapter also speculates about the chemical influence of salts, formerly deposited in inter-block basins within the basement rocks, on the groundwater chemistry and the speleogenetic processes associated with deep flow paths in karst aquifers. Relatively elevated silica concentrations were determined in the Bakewell thermal springs. This author opines that inception of the regional flow paths relates to dewatering of sediments in the Carboniferous basins.

Groundwater chemistry was considered in Chapter 6 and the findings have contributed to the conceptual model by further defining overflow, baseflow and underflow paths. This has been achieved by building upon Christopher's (1981) classification. Strontium and temperature have been shown to be the best indicator of underflow. Fluoride concentrations are the best indicator of groundwater contact with mineralization. As the main influence on groundwater chemistry is the dissolution that occurs in the unsaturated zone, some conclusions have been drawn regarding the nature of the recharge contributions associated with differing formations (section 6.7). Chemistry of the spring waters varies seasonally, but the variation in chemistry varies between springs. The variations have been used to interpret the dominant influences on groundwater recharge. The chemistry of some springs suggests that evapotranspiration has a significant influence on the groundwater chemistry. Sources of sulphate have also been considered, particularly in the context of the regional flow paths, but also in the context of mineralization and the head deposits (sections 5.5.4 and 5.5.6). Potential, deep sources of carbon dioxide have been speculated upon (section 6.3.1).

An overview of previous dye-tracing experiments, in conjunction with the evidence of the traces undertaken during the course of this thesis, has enable this author to point to the greater significance of divergent flow in hydrogeological unit 4, which demonstrates the laterally extensive nature of conduits in this unit. It has been speculated that this reflects the lateral continuity of the inception horizons, the formational process being linked with the distribution of the clay wayboards. The results of the dye tracing also focus on the rate limiting factors imposed, at least in part by silicification and described above.

Although a considerable amount of effort was put into the derivation of combined recession curves for a number of hydrographs for boreholes in the White Peak (Chapter 8), the findings confirm the conclusions of Smart (1999) and Worthington (personal communication, 2007) that discontinuous temporal monitoring of non-screened wells is of limited value in the understanding of limestone aquifers. However, such measurements could be useful in assessing any long-term trends in the data. Nevertheless, some conclusions have been drawn regarding permeability and the hydrogeologic parameters have been presented in Table 8.2. The low range in groundwater levels in valley, compared with plateau, settings indicates greater storage associated with the valley setting.

Evidence from mining records and the association of springs with fault locations demonstrates the importance of faults as zones of groundwater storage. Observations made at Bubble Springs, Lathkill Dale (Chapter 11) indicate that the storage occurs within and behind the fault. Confined water, as indicated by elevated spring temperatures, has been found to rise on faults, or dominant mineral veins. This observation also appears to be true of the thermal springs. This evidence has been used to substantiate the hypothesis that the faults and dominant mineral veins form zones of significant vertical groundwater flow paths.

Geomorphological evolution of the Derbyshire Dome has been used to try to establish reasons for apparently anomalous karst systems, such as the proven connection between Illy Willy Water and Ashwood Dale Resurgence (Chapter 7). This has been investigated using the concepts presented by Simms (2004) and builds on earlier work on the geomorphology (including: Burek 1977 and 1978; Clayton, 1953; Johnson, 1957; Ford and Burek, 1976, Warwick, 1962). It has been included in the development of the conceptual model (Chapter 9).

Human impacts include the direct lowering of groundwater levels, by the early construction of soughs (section 10.2) and the subsequent indirect reduction in head resulting from the construction of deeper soughs. In particular, attention has been focused on the impact of Magpie Sough. These impacts have further increased the southeasterly hydraulic gradient. Consideration has also been given to the modifications in permeability that have resulted from sough construction, both directly as a consequence of the construction of the additional ‘conduits’ and indirectly as a consequence of the removal of superficial deposits from the epikarst. The effects of quarrying include a reduction in the extent of the epikarst, resulting in more direct recharge; further influences have been listed in section 10.4. Analysis of spring recession curves and comparisons with the recession curves presented for Cheddar Spring suggests that the human impacts may have increased the conduit recharge component of the recession curve. It has also been shown that Magpie Sough penetrated a groundwater divide. Groundwater quality issues in the area formed the subject of section 10.7. Clearly, increased recharge rates resulting from removal of epikarst introduce contaminants to groundwater aquifers more quickly and with less attenuation. These impacts can also be set in the geological context. For example limestone quarrying in this research area is largely focused upon the Bee Low Limestone and Woo Dale Limestone formations.

The use of Lathkill Dale as a case study has provided additional evidence to support the conceptual model presented in Chapter 9. Further to this, the research has made a contribution that will contribute to the planning of future management of the hydrology of the dale (Ben Le Bas, personal communication, 2007). In terms of applications for this work, Figure 12.1 comprises a plot to show the potentiometric surface, springs, groundwater levels in boreholes and boundaries of the hydrogeological units defined in Chapter 9, together with representative results of dye-tracing experiments. It is hoped that this will provide a useful starting point for any future groundwater studies in the research area.

12.2 Potential directions for future research.

Whilst possibly beyond the remit of the thesis, this author considered that it would be beneficial to highlight potential research areas identified by the research that forms the subject of the thesis. Accordingly, the potential research directions that have been identified for each of the aspects of work that have been addressed have been listed below, with key words highlighted in bold for quick reference:

-Silicon or strontium isotope analyses might provide a tool to differentiate between **chert** from postulated differing depositional environments. Chert fluid inclusion studies could help to constrain formational temperatures.

-This author would be particularly interested to carry out more archival research regarding historic supplies of **potable water**. This is an aspect that was addressed in a relatively cursory manner in Chapter 3, but it is considered that collating more information on quality and quantity of the water historically abstracted and also, if available on river flows, offers a substantial potential in terms of understanding human impacts on the hydrogeology.

-One of the key points in the hydrogeological model that has been presented is the identification of the significane of the Chee Tor Limestone Member, acting as an aquitard at depth and a significant aquifer close to be surface. Additional work with respect to the **engineering geological properties** of the Chee Tor Limestone would help to constrain these properties, for example with respect to shear strength and solubility.

-The potential **speleogenetic processes** that have been identified open up a considerable number of potential research directions. In particular: with respect to furthering the understanding of the occurrence of dolomite and dedolomite; sampling and optical examination of potential inception horizons operating in the Woo Dale Limestone Formation.

-Consideration could be given to the dating of **speleothems** in Lathkill Head Cave to constrain the postulated rifting associated with cambering.

-With the exception of an undergraduate research project (Lamb, 2006), literature reviews indicate that there has been little work on the influence of **sediment load** on conduit development in the context of the White Peak.

-The distribution of the fault blocks is considered to be particularly important in understanding the regional hydrogeology of the research area, therefore any research contributing to this understanding could potentially be a valuable contribution, for example using advanced techniques of **remote sensing** or **geophysics.**

- This author would be interested to carry out some preliminary investigation of **thermal imaging** in the area of the White Peak, with a view to assessing the value of this technique for the identification of thermal waters.

-The results of the limited seasonal monitoring that has been carried out indicate that there is a potential for further **seasonal flow monitoring and chemical analysis** to contribute to the understanding of the transmission and storage processes associated with the limestone of the White Peak.

-Previous hydrogeochemical studies have focused on the limestone; it is considered that there could be some benefit from looking at allogenic recharge in more detail, in particular to determine whether there are any markers that would **differentiate allogenic/autogenic** recharge.
-This author has pointed to pockets of superficial deposits associated with hydrogeological unit 4 as zones of potential groundwater storage and therefore suggests that it would be interesting to look at **distribution of moisture** in the superficial deposits in the subsidence hollows associated with the Monsal Dale Limestone and also in the clay wayboards and the epikarst of other formations.
-Examination of Figure 12.1 suggests that **dye-tracing** in the area of **Peak Forest** would contribute to filling a gap in the testing that has been reported upon. It was suggested in Appendix 7.3 that there is considerable scope for further, rigorous experimental work with fluocapteurs, dyes and elutants.
-This author is interested to try and constrain the extent of **barometric effects** on spring and river discharge.
-It is clear from the descriptions of resource exploitation from the White Peak could be calculated on a formational basis, thereby enabling the addition of **human impacts** to the conceptual model of the hydrogeology (Chapter 9).
-Additional archival studies could be used to assess whether the driving of Magpie Sough had a measured impact on the discharge of the **Bakewell Springs** and similarly the impact of the de-watering of **Long Rake** spar mines on the discharge of the River Lathkill.
-The case study (Lathkill Dale) suggests that further monitoring is required to determine the actual losses from the Lathkill catchment to Magpie Sough. This could take the form of discharge monitoring of the River Lathkill, Bubble Springs and Magpie Sough. Further monitoring and determination of the differences in the groundwater chemistry of the upper and lower springs at **Bubble Springs** flow has the potential to offer significant understanding of the speleogenetic processes and hydrogeological function associated with the faulting in the area of Bubble Springs.
- A **regional groundwater survey of temperature and strontium** would provide further evidence to help to constrain the extent of underflow.

In concluding this work it would be churlish to ignore the recent publication by Brassington (2007), which uses some the data referenced by this author to come to very different conclusions regarding the regional hydrogeological setting of the White Peak. Brassington's (2007) hypothesis considers thermal flow systems derived from convection cells established within the limestone, during Late Carboniferous to Early Permian times, whereas the conceptual model presented in this thesis is set in a regional context and considers recharge from outside the White Peak. Resolution of this aspect of the hydrogeology is important in terms of defining source catchment protections zones (Gunn, 2007). Accordingly, in the future, this author would be keen to extend the hydrogeological investigation to the area of the Todd Anticline, immediately to the west of the White Peak.

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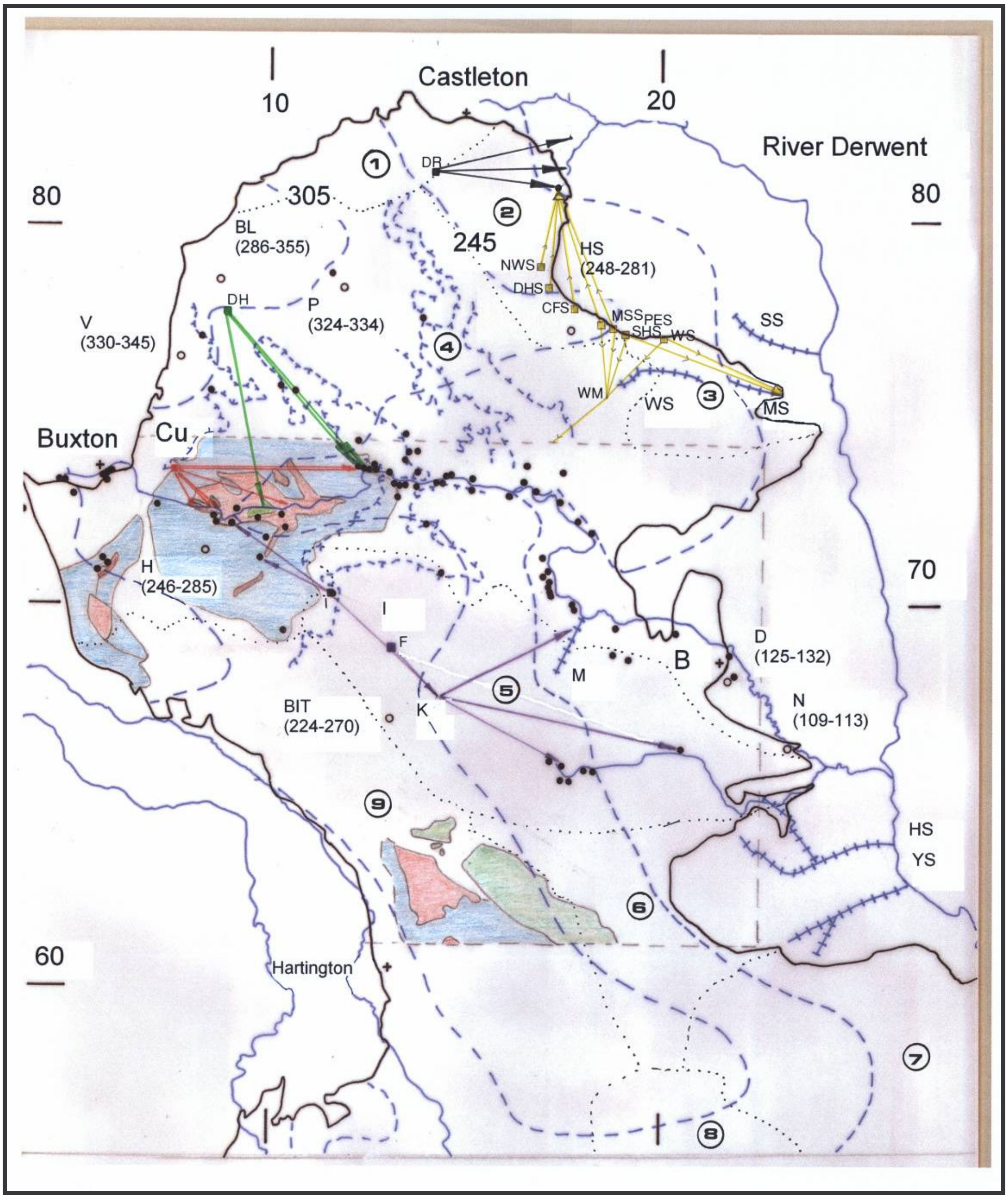

Figure 12.1: Summary map for the catchment (adapted from Downing et al., 1970). Key open circles Boreholes: BL Bee Low; P Peak Forest; HS Hucklow South; V Victory Quarry; N Nutseats Quarry; D Dale Farm; BIT Bull I' Th' Thorne; H Hinchcliffe Farm. Range of groundwater levels in brackets. Closed circles springs. Hatched lines soughs: SS Stoke; MS Moorwood; H Hillcarr; M Magpie; YS Yatestoop Sough. Dashed blued groundwater contours with levels (m OD). Coloured squares dye injection points: CFS Cartledge Farm Swallet; Cu Cunningdale; DHS Duce Hole Swallet; DR Dirtlow Rake; F Flagg; I Illy Willy Water; MSS Mrs Smythe's Swallet; NWS Nether Water Swallet; PES Piece End Swallet; SHS Swevic House Swallet; WM Wardlow Mires (not a dye injection point); WS Waterfall Swallet. Black dotted lines groundwater subbasins numbered 1-8, as in Figure 7.1. Crosses settlements, named except B Bakewell. Brown dashed line denotes limit of hydrogeological unit definition: green unit 1; red unit 2; blue unit3; unit 4 blank; unit 5 green. Dashed line with hatch denotes areas of perched groundwater.

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