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Hydraulic and geological factors influencing conduit flow depth

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Abstract

There has much been speculation as to whether cave formation should occur at, above, or below the water table, but a satisfactory explanation has been lacking until recently. The last 50 years has seen extensive mapping of caves both above and, more recently, below the water table. It is now becoming apparent that there are systematic differences in depth of flow between different areas and that conduit flow to depths >100m below the water table is not uncommon. Such deep flow is facilitated by the lower viscosity of geothermally heated water at depth. Analysis of data from caves shows that depth of flow is primarily a function of flow path length, stratal dip and fracture anisotropy. This explains why conduits form at shallow depths in platform settings such as in Kentucky, at moderate depths (10–100m) in folded strata such as in England and in the Appalachian Mountains, and at depths of several hundred metres in exceptional settings where there are very long flow paths.

Keywords: Speleogenesis, conduit flow, deep flow

Introduction

The nature and extent of cave development with respect to the water table has received much attention, especially in the first half of the Twentieth Century (Lowe, 2000a). Early theories were largely speculative because few caves above the water table and none below the water table had been accurately mapped. For instance, in 1950 there were only seven caves with mapped lengths greater than 10km. In the last 50 years, however, this number has grown rapidly, reaching 125 in 1977 and 350 in 2000 (Chabert, 1977; Madelaine, 2001). The growth of mapped cave passages below the water table is even more marked. In 1950 underwater exploration in caves had barely commenced, but by 2000 there were 14 caves explored to depths of at least 150m below the water table (Farr, 2000). This impressive dataset of mapped caves provides an invaluable empirical database for ideas on cave development.

Conduit development results from the interplay between hydraulic, chemical and geological factors. Early calculations on conduit enlargement were disappointing as they showed that infiltrating water would quickly become saturated with respect to calcite, that no further dissolution would then take place, and thus caves could not be formed (Weyl, 1958). Caves clearly do exist, and so some alternative explanations were proposed, such as the presence of sulphides (Howard, 1964) or the occurrence of mixing corrosion (Bögli, 1964). At the time it thus appeared that caves might only form in some special situations and so be rare. However, this concept was overturned in the 1970s when lab experiments showed that the dissolution rate of calcite and of dolomite is non-linear and that there is a large reduction in the rate as chemical equilibrium is approached (Berner and Morse, 1974; Plummer and Wigley, 1976; Herman and White, 1985). These results were in turn incorporated into numerical models, which showed that caves form in all unconfined carbonate aquifers (Dreybrodt, 1990; Palmer, 1991). Furthermore, this process occurs not only where there is sinking stream recharge but also where there is percolation recharge (Dreybrodt, 1996).

Early studies on the depth of cave development with respect to the water table focused largely on hydraulic factors. For instance, Davis (1930) noted that some flow lines from recharge areas to discharge areas describe arcs deep below the water table during the initial pre-karstification stage of flow through an aquifer, and he suggested that cave development could occur along such deep flow paths. Swinnerton (1932) suggested that the greatest flow and thus the greatest dissolution would occur along the shortest flow path, that shallow flow paths are the shortest, and consequently that conduit formation close to the water table is favoured. Thrailkill (1968) analysed flow through shallow and deep flow paths and his findings supported Swinnerton (1932). The concept of cave development close to the water table attracted widespread support and the occurrence of deep phreatic flow presented a challenge because there was no known process that explained it. Ford and Ewers (1978) proposed that deep flow only occurs where there are a limited number of open fractures and that it is solely the presence of these open fractures that facilitates conduit development deep below the water table. However, Worthington (2001) carried out a comprehensive analysis of the equation for laminar flow through fractures and showed that the effect of geothermal heating results in increased flow for deep flow paths. Consequently, the presence of random large open fractures at depth is not required to explain deep flow.

The most widely quoted model in recent years has been that of Ford and Ewers (1978), which appeared to provide a comprehensive explanation of the occurrence of deep phreatic, shallow phreatic and water table caves. However, there has been a general failure to find examples that fit the model, and Ford (2000) stated that the model *"...does not attempt to predict what will be the effective fissure frequency and aperture in any particular topographic or geologic setting"*. This raises the spectre that cave development might occur at unpredictable depths below the water table, being related only to randomly located open fractures.

This pessimistic view of cave formation appears to be contradicted by the growing evidence from cave exploration and mapping that there are systematic differences between different areas in the depth of flow. For instance, cave diving in the Mendips, the Peak District and in the Yorkshire Dales has shown that these three areas in England have sumps that descend to as much as several tens of metres below the water table (Farr, 2000). Geomorphological studies in English caves have demonstrated that some fossil conduits had similar depths of flow (Ford, 1968; Waltham, 1974). In contrast, extensive cave exploration in Kentucky indicates that few conduits were formed more than a few metres below the water table (Palmer, 1987), whereas in some mountainous areas there are examples of conduits formed at depths >100m below the water table (Waltham and Brook, 1980; Smart, 1984; Farr, 2000, Jeannin *et al.,* 2000; Yonge, 2001).

The systematic differences between England, Kentucky and mountainous areas suggest that depth of flow is likely to be a function of one or more parameters that vary between these three settings. Worthington (2001) found that depth of flow was a function of stratal dip and flow path length, but only a limited number of parameters was considered in that study. The discussion below considers a larger number of parameters that might influence the depth of conduit flow, including hydraulic factors, fracture permeability, fracture geometry, spacing of bedding planes, and topography.

Hydraulic factors

Flow along fractures in the early stages of karstification is in the laminar regime, and can be described by the Hagen-Poiseuille equation (White, 1988, p.162). This describes head loss (h) through a circular pipe as

$$h = 8\ \eta\ v\ L\ /\ (\rho\ g\ r^2) \qquad (1)$$

where η is dynamic viscosity, ρ is the density of water, v is velocity, L is conduit length, r is conduit radius, and g is the acceleration due to gravity.

Thrailkill (1968) calculated the differences between shallow and deep flow paths in horizontally-bedded strata by using Equation 1. His results showed that shallow flow is slightly greater than deep flow, implying that conduits should develop close to the water table. However, Thrailkill used the simplifying assumption of constant temperature and constant viscosity in his calculations.

The average global geothermal gradient is about 25ºC/km, so that water following a deep flow path in a limestone aquifer will be slightly warmer than water following a shallower flow path. The density and dynamic viscosity of water both vary as a function of temperature, and can be combined to give the kinematic viscosity, which decreases by 50% between 10ºC and 40ºC. This results in increased flow in fractures at depth, increased dissolution, and an environment for preferential formation of conduits deep below the water table.

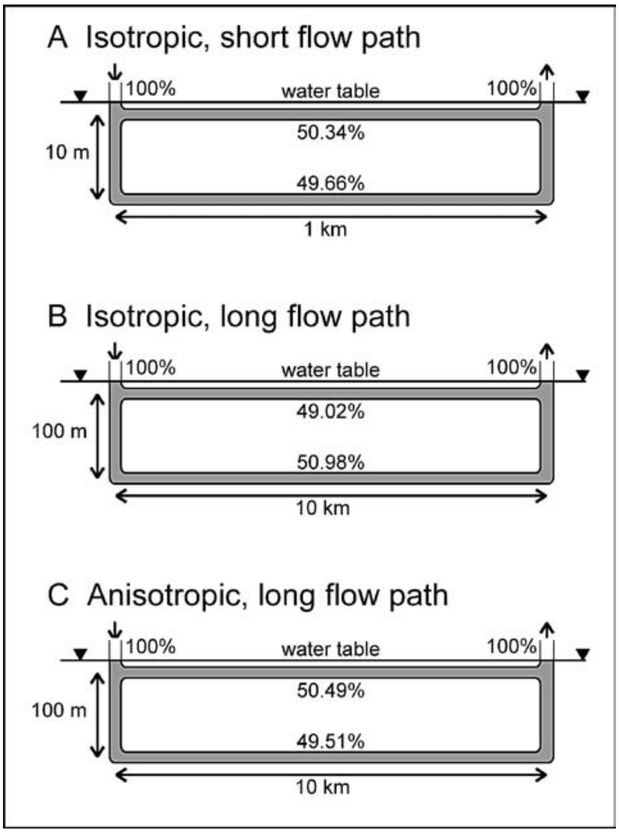


Fig. 1. Relative flow in shallow and deep conduits in flat-bedded limestone where the length/depth ratio is 100. A) Greater flow in the shallow conduit where the flow path is 1km long due to the shorter flow path in the shallow conduit. B) Greater flow in the deep conduit where the flow path is 10km long, due to the lesser viscosity at depth. C) Greater flow in the shallower conduit where initial apertures in the horizontal fractures are twice the apertures in the vertical fractures.

Fig. 1 shows a comparison of flow through shallow and deep flow paths, where the temperature at the water table is 20ºC and the geothermal gradients is 25ºC/km. A single input and single output are modelled, with flow through a shallow and a deep conduit. This is similar to the model of Thrailkill (1968), except that all terms in the Hagen Poiseuille equation are considered here. Three cases are examined in Fig. 1 and in all cases the deep flow path is 2% longer than the shallow flow path. Figures 1a and 1b are for isotropic conditions, where the apertures of the horizontal and vertical pipes are the same.

For a short (1km) flow path the lower viscosity at depth does not compensate for the longer flow path, so flow through the shallow pipe is greater (Figure 1a). However, the opposite is true for the longer (10km) flow path in Fig. 1b, with more flow passing through the deeper conduit. Worthington (2001) calculated that a distance of 3300m is the critical flow path length threshold for more efficient flow path at depth, assuming a geothermal gradient of 25ºC/km and that the strata are flat-lying. Fig.1c shows results for anisotropic conditions, where the initial apertures of the horizontal pipes are twice that of the vertical pipes. The anisotropy results in greater flow through the shallow pipe.

The hydraulic analysis shows that deep conduit development should be favoured in many settings, implying that the base of a limestone aquifer is the optimum location for conduits. In fact, such a location for conduits is rare (excepts for conduits formed in the vadose zone), implying that there must be other pertinent factors that favour shallow flow or inhibit deep flow. One possibility is anisotropy (Fig. 1c), and the evidence for this is discussed next.

Fracture permeability

Evidence of fracture permeability from cave maps and flow path analysis

Conduits developed in older (e.g. Palaeozoic) limestones are almost all oriented along bedding planes, joints, or faults, or along intersections of two of these fracture planes. Jameson (1985) found that the initial fracture guidance for a passage could best be inferred in caves where the fractures are prominent but widely spaced, bedding is massive, roof collapse is minimal, sedimentation is limited, and passage size is not so great that traces of the initial guiding fractures have been eroded away. Few caves fit all these conditions and so it is commonly difficult to identify the initial guiding fractures. Furthermore, few detailed studies of fracture guidance in caves have been made. One such study is by Jameson (1985), who found that all but 2% of the passage length in part of Friars Hole System, West Virginia, was formed along fractures.

Matrix permeability is higher in younger limestones and so fracture-guided conduits may be less important in some of these rocks. Nevertheless, in the limestones of Eocene to Miocene age at Mulu there is widespread passage development along bedding planes, joints and faults (Waltham and Brook, 1980). Furthermore, many cave passages in early Cainozoic limestones in Tonga and late Cainozoic limestones in the Bahamas are oriented along near-vertical fractures (Lowe and Gunn, 1986; Palmer, 1986).

The alignment of cave passages along bedding planes, joints and faults gives an indication of the relative permeability of these features. Mammoth Cave, Friars Hole System and Castleguard Cave provide examples of fracturing (Figures 2–5), and these caves have been widely cited in the literature (Gunn, 2004; Culver and White, 2005).

Mammoth Cave, Kentucky, is the world's longest cave and has figured prominently in discussions of cave formation. It is located in a low-dip mid-continent setting where there is little faulting or folding. Joints in the cave are sparse, individual joints have very limited vertical and horizontal extents, and few passages have clear joint orientation (Deike, 1989; Palmer, 1989a,b). In this situation, almost all passages follow bedding planes and in plan view have meandering paths (Fig. 2). Swinnerton Avenue provides a typical example. All of the approximately 1000m of passage is aligned along bedding planes except for a 3m section where flow rose on a joint (Fig. 3). Scuba diving in the Mammoth Cave area has shown that most active phreatic passages are located within a few metres of the water table. Likewise, studies of fossil passages indicate that most were within a few metres of the water table when they were active (Palmer, 1987, 1989a,b). The scarcity of clear passage guidance on either joints or faults and the frequent large width/height ratios of phreatic passages at Mammoth Cave indicate that the initial horizontal hydraulic conductivity (K_h) along bedding planes was much greater than the initial vertical hydraulic conductivity (K_z) along joints or faults.

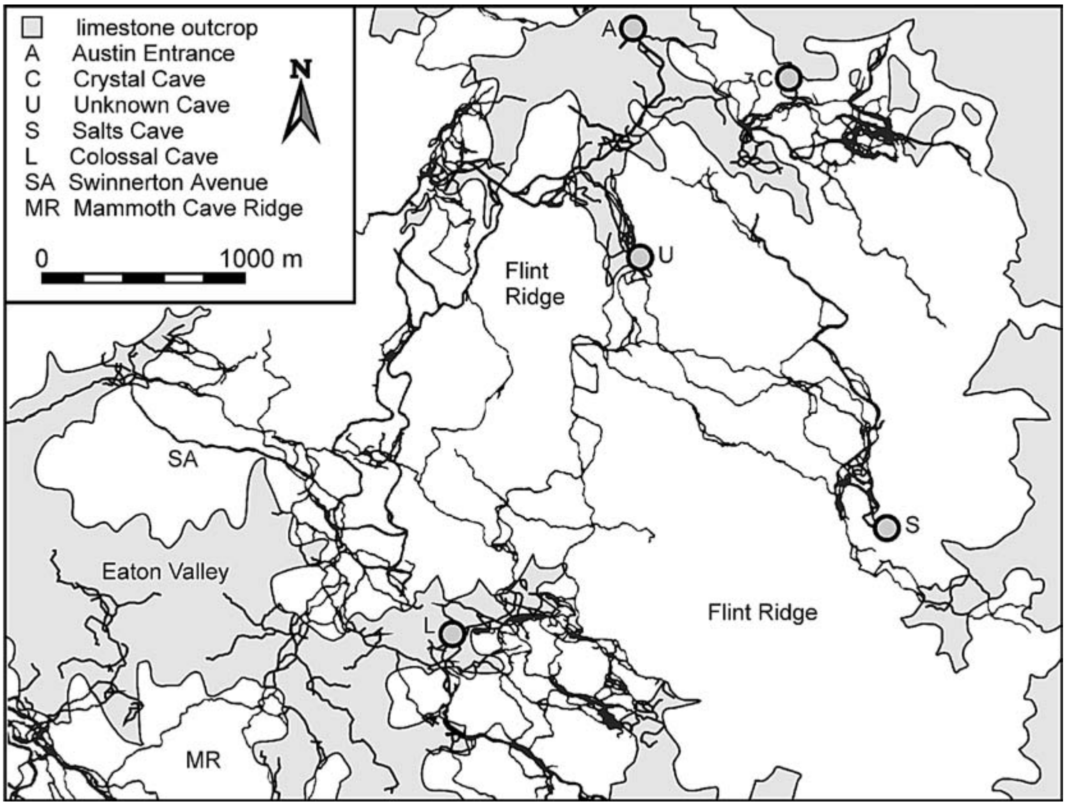


Fig. 2. Relationship of cave passages in part of the Mammoth Cave System to the surface outcrops of limestone along valleys (after cave map by Cave Research Foundation and geology map by U.S. Geological Survey).

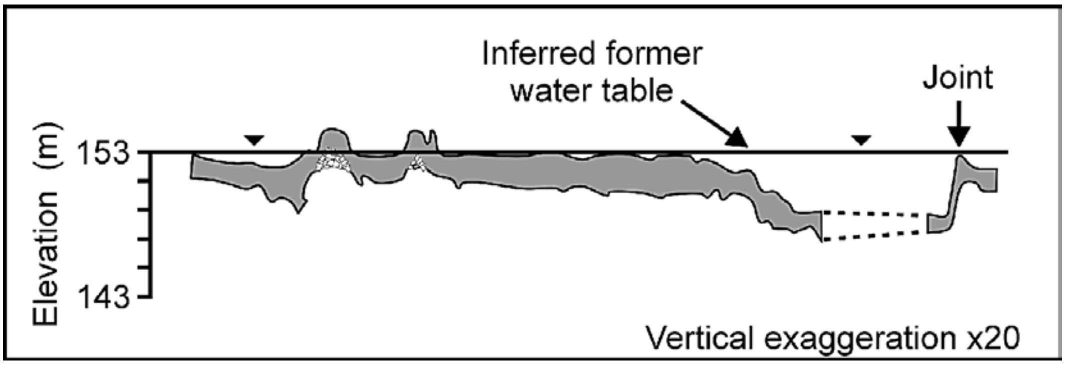


Fig. 3. Profile of part of Swinnerton Avenue in the Mammoth Cave System (after Palmer, 1989a).

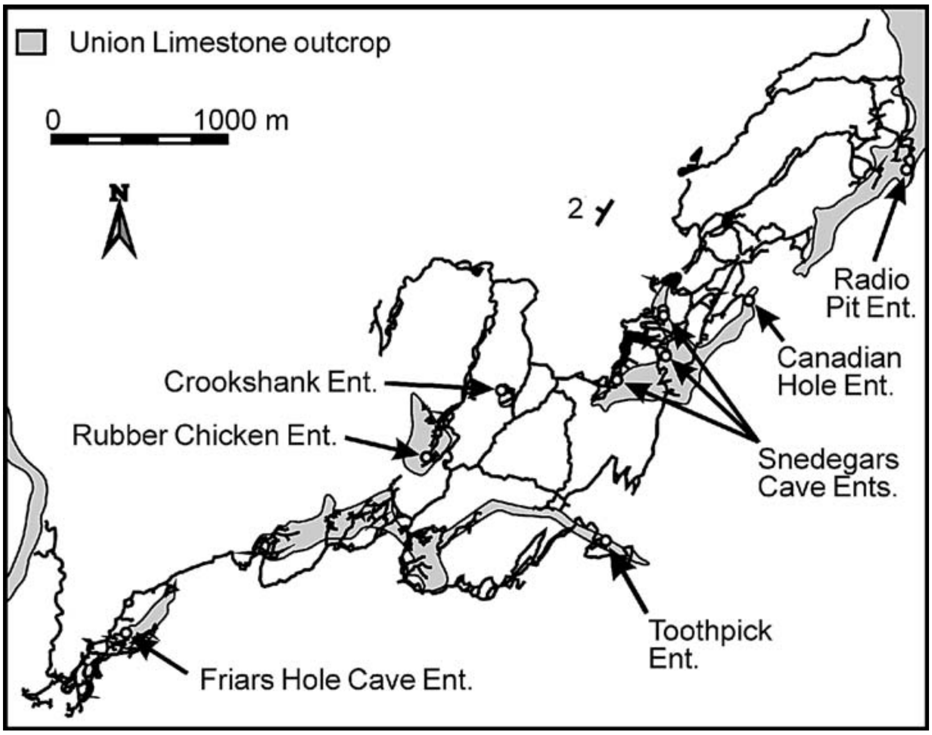


Fig. 4. Map of Friars Hole System, showing limestone inliers where recharge to the cave occurs (based on map compiled by D. Medville).

Friars Hole System, West Virginia, provides a second example (Fig. 4). The cave is the longest in the Appalachian Mountains. It lies close to the boundary between the Appalachian Plateau and Valley and Ridge Provinces, and the structure shows features of both geological provinces. The entrances to the cave lie along a valley, Friars Hole, where a series of limestone inliers have been exposed. Sinking streams typically descend downdip to the northwest through the vadose zone in the upper 50m of limestone, and then along almost horizontal base-level passages along the strike to the southwest. Many of these base-level passages are formed at the intersection of low-angle thrust faults and the contact between the Union Limestone and the underlying Pickaway Limestone (Worthington, 1984; Worthington and Medville, 2005). The detailed study by Jameson (1985) of 429m of passage in Snedegars Cave showed that 37% of the passage was oriented along bedding planes, 30% on joints, 20% on bed/joint intersections, 7% on faults, 4% on fault/joint intersections, and the remaining 2% had no guiding fracture. From a geomorphological study, Worthington (2001) inferred that the mean depth of flow in fossil conduits was 17m below the water table. The roughly equal proportions of bedding plane guidance and joint or fault guidance for passages at Friars Hole System (Jameson, 1985), as well as the elongation of phreatic passage cross-sections along either bedding planes or on faults or joints, indicates that the initial pre-dissolution apertures f bedding planes were similar to those of faults and major joints. Consequently, the initial vertical and horizontal fracture permeabilities were of similar magnitude ($K_h = K_z$).

The third example is Castleguard Cave, which is the longest cave in Canada and is located in the Rocky Mountains. The cave's topographical and geological situations are described by Ford *et al.* (2000). Fig. 5 shows the plan and profile of the cave. The main passage extends updip for 8921m from its entrance at 2010m above sea level to a termination where it is blocked by glacier ice that has intruded from the overlying Columbia Icefield. The plan shows that the main passage is oriented along a series of northwest-trending joints. However, the profile shows that the cave was almost all developed on three bedding planes (Fig. 5, profile). The lowest of these is followed by the main passage for 7045m. The two pitches are developed on joints or faults, and the remainder of the main passage is developed along bedding planes (40% of the total length) or bed-joint intersections (60%). The longest distance that a passage follows a single bed-joint intersection is 650m. There has been minor movement, possibly of just a few centimetres, on both joint and bedding plane fractures (Ford *et al.,* 2000). The geomorphology of the cave suggests that it was enlarged up to a cross-section of about 3m^2 while it was below the water table, indicating that the present entrance was more than 370m below the water table at that time. Passage cross-sections and fracture guidance suggest that $K_h = K_z$.

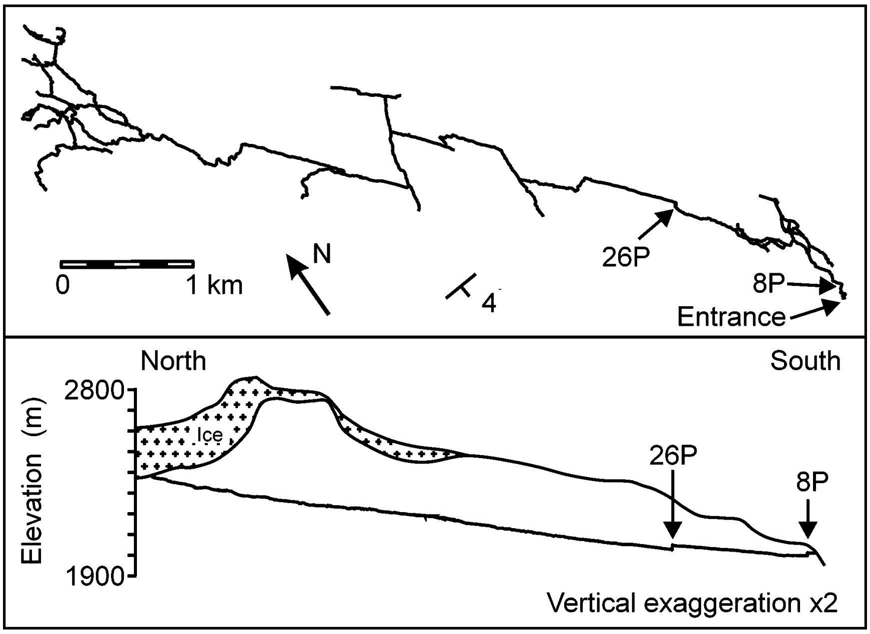


Fig. 5. Plan (top) and profile projected along the strike (bottom) of Castleguard Cave. The profile shows only the main passage (based on survey compiled by S. Worthington).

In Britain there have been many studies that have described the major guiding fractures in caves, such as in South Wales (e.g. Coase and Judson, 1977), the Forest of Dean (e.g. Elliott *et al.,* 1979), the Mendip Hills (e.g. Drew, 1975), Derbyshire (e.g. Ford, 1977) and Yorkshire (e.g. Waltham, 1974). There are many cave passages oriented along joints or faults close to sections of cave where faults and joints have much less importance than bedding planes in guiding cave passages. Leck Fell provides a notable example. The stream in Short Drop Cave flows down the axis of a plunging syncline and is perched 100m above the water table, indicating narrow joints apertures in this area. However, just 200m to the north the deep shafts of Death's Head Hole, Rumbling Hole and Big Meanie are aligned on a single vertical fault (Waltham, 1974; Waltham and Hatherley, 1983), and this demonstrates the local-scale variability in K_h / K_z . None of the above studies were carried out to the level of detail of Jameson (1985), and so there are no statistics on the relative importance of the various types of fracture or fracture intersections in providing guiding pathways for passages.

Bedding planes are used extensively in all the caves described above. However, joint and fault guidance is extremely variable between caves. In most cases bedding plane and joint or fault apertures appear to be similar, giving an isotropic fracture network ($K_h = K_z$). Less commonly the fracture network may be anisotropic, with $K_h > K_z$ or $K_h < K_z$. Mammoth Cave is an example of the former, where $K_h > K_z$, and this inhibits deep flow. Death's Head Hole is an example of the latter, with $K_h < K_z$, and this promotes deep flow.

Stress-release fracturing

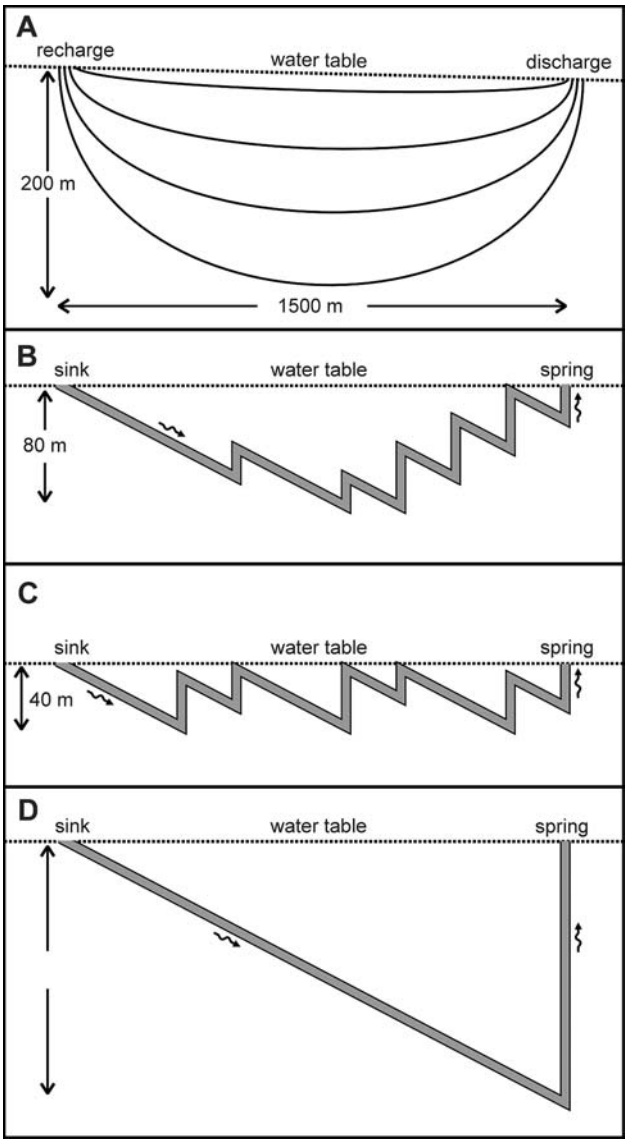
Sasowsky and White (1994) found that many long caves in Tennessee closely paralleled the sides of deep valleys, and suggested that this is due to stress-release fracturing. However, the reasons are unclear why stress-release fracturing should be important in this setting and of lesser importance in other caves in North America such as Mammoth Cave, Friars Hole System, and Castleguard Cave. Much of the passage of Mammoth Cave is found close to the centre of ridges capped by clastic rocks rather than at the flanks of these ridges where stress-release fracturing should be more important (Fig. 2). Similarly, at Friars Hole System and at Castleguard Cave there is no tendency for passages to be located where the rock overburden in minimised (Figures 4 and 5).

Stress-release fracturing is more important close to the surface and decreases sharply with depth. Therefore it will promote shallow conduit development in those cases where it is important.

Fracture geometry

Fracture geometry can have a significant influence on the depth of conduits below the water table. Two common situations are discussed below, with the flow direction either aligned in the same direction as the stratal dip or normal to the dip. In both cases steeper dips facilitate flow to greater depths below the water table (Ford, 1971). Conversely, deep flow is likely to be inhibited in low-dip strata unless open fractures facilitate vertical flow, as in Yorkshire, as noted above.

Fig. 6 depicts an aquifer where flow from a recharge area to a discharge area is in the same direction as the stratal dip. Figure 6a shows a simple initial flow field before the start of dissolution. A single source of recharge is shown, such as occurs where a sinking stream develops. Initially, there are a series of flow lines from the area of recharge to the area of discharge. The following three figures (6b,c,d) show three possibilities for where conduit enlargement might occur. Fig. 6b shows a cave with six phreatic loops and a flow path that follows a generally curving pattern below the water table. Figure 6c also shows six phreatic loops, but the conduit is in general closer to the water table than in 6b. Figure 6d shows a single phreatic loop. All three flow paths have identical lengths, with the conduit following a bedding plane for a horizontal distance of 1500m and following one or more joints vertically for a total of 180m.



Given the equal length of the flow paths in Figures 6b, 6c and 6d, it might be considered that there are equal probabilities of a conduit forming along any one of them. However, there are reasons why each one of these flow paths might be favoured. The flow path in Fig. 6b might be favoured if the early conduit dissolution was along a flow path similar to that in a porous medium and if there is a balance between the two competing factors of decreasing fracture aperture with depth (which promotes shallow flow) and decreasing viscosity with depth (which promotes deep flow). The flow path in Fig. 6c might be favoured if factors such as greater fracture apertures at shallow depth or mixing corrosion between conduit water and percolation were the most important factors. The flow path in Fig. 6d is favoured if the fracture network is isotropic as in Figure 1b.

Fig. 7 shows the equivalent situation where flow is along the strike. Fig. 7a shows flow lines along a single bedding plane from a recharge area to a discharge area and Figures 7b, 7c, and 7d show three possibilities of where a conduit might form. It is assumed that there are two major perpendicular joint sets and that conduits form at the intersection of joints with the bedding plane. The flow paths in Figures 7b, 7c and 7d have identical lengths. The reasons why one of these three might be favoured for conduit development are the same as with the downdip case (Fig. 6).

Fig. 8 shows a simplified rectilinear flow system that is similar to Figure 1 but has flow along a dipping bedding plane rather than a vertical fracture. The length of the flow path P is

$$P = L + 2D/(\sin \alpha) \qquad (2)$$

where L is the horizontal distance of the flow path, D is the depth of the conduit and α is the stratal dip in degrees. Shallow dips have increasingly long flow paths and thus there is a lower probability that the more efficient flow at depth (q.v. Equation 1) will be associated with low dips. If the depth of flow is inversely proportional to the increasing length of the flow path then the depth of flow will be proportional to the sine of the dip.

Fig. 9 shows two caves with flow paths that are similar to Figures 6b and 7b, respectively. The flow path between Grotte Annette and Trou de Glaz is primarily downdip (Fig. 9a), and is one of the major

Fig.6 (left). Potential flow paths where the flow path from a sink to a spring is downdip. A) Initial flow field; B) – D) Possible flow paths. The three flow paths in B–D have identical lengths.

flow paths in this 50km-long cave (Lismonde, 1997; Audra, 2004). It was used as the basis for Fig. 6, and the flow pattern is clearly very similar to Fig. 6b. As discussed earlier, this suggests that the pattern of preferential solution was influenced principally by the initial flow field in combination with decreasing fracture apertures with depth. However, the geology also plays an important role. The section of cave shown is developed in 190m of Cretaceous limestones. Lismonde and Marchand (1997) identified just five bedding planes within this 190m section that are preferentially used for conduit development, and most of the cave shown in Figure 8a is aligned along one of these bedding planes (Fig. 10). A relatively minor fraction of the passages has vertical flow along joints or faults (Fig. 11).

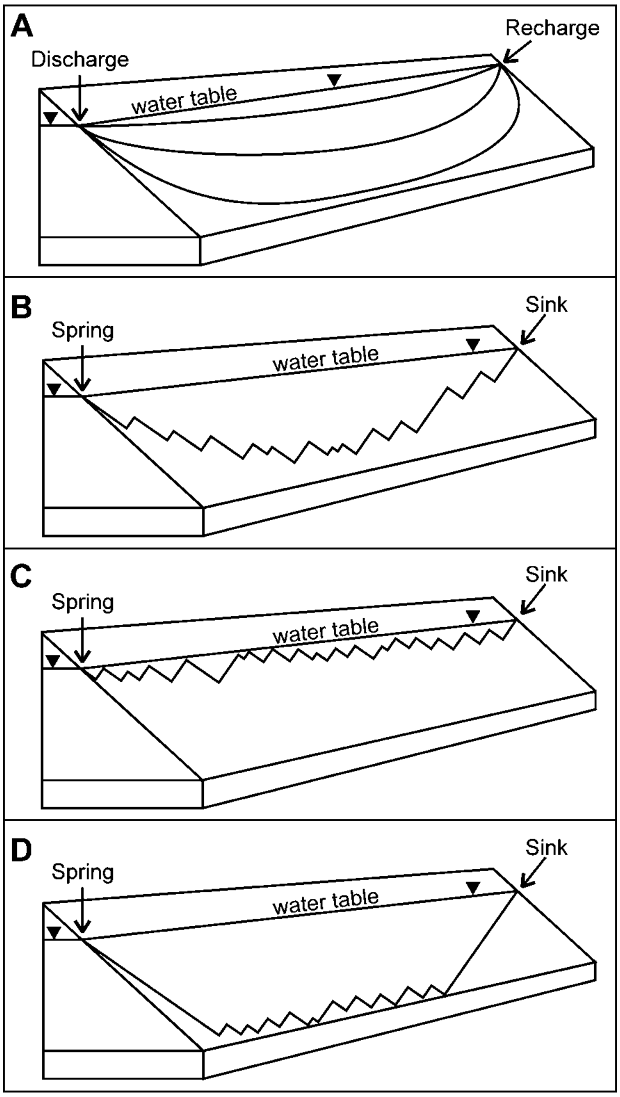


Fig. 7. Potential flow paths where the flow path from a sink to a spring is along the strike and where conduits form at the intersection of joints and a single bedding plane. A) Initial flow field; B) – D) Possible flow paths. The three flow paths in B–D have identical lengths. The dipping bedding plane is assumed to extend to much greater depths than is shown.

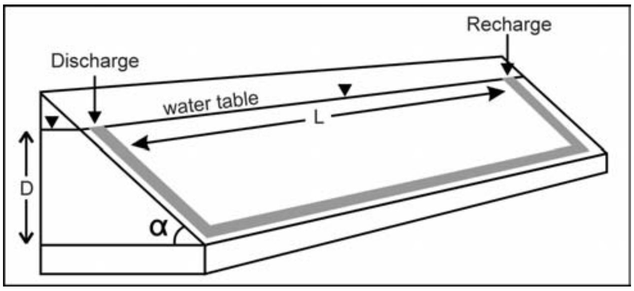


Fig. 8. A simple rectilinear flow path on a dipping bedding plane.

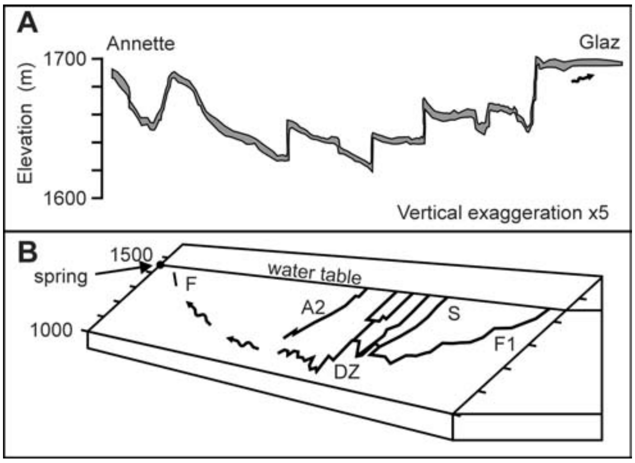


Fig. 9. Major conduit flow paths that exhibit a single curving loop below a former water table. A) Profile of downdip flow in Dent de Crolles, France (adapted from Lismonde, 1997); B) Perspective view of 12km flow path on strike at Siebenhengste (S), Fitzlischacht (F) and the caves A2 and F1, Switzerland (adapted from Jeannin *et al.,* 2000). Note: DZ = Deep Zone.

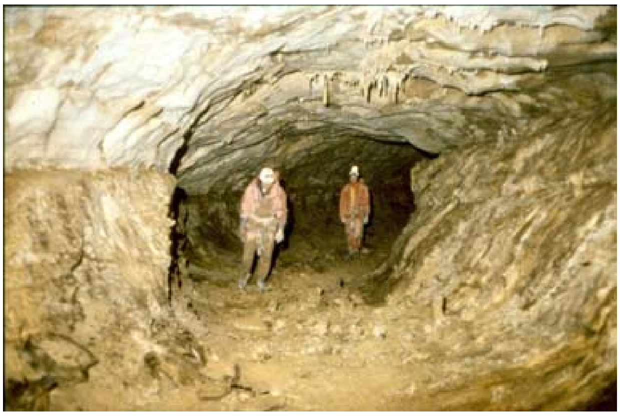


Fig. 10. Bedding plane guided phreatic tube in Trou de Glaz. The passage was formed at least 50m below the water table.

Fig. 11. Joint-guided rift in Trou de Glaz that was formed by upward flow at least 60m below the water table.

Figure 9b shows the caves of the Siebenhengste area in Switzerland that developed when the water table was at 1440m asl. Most of the passages are developed at the base of the Schrattenkalk Formation, a pure limestone that is 150–200m thick and that overlies a marly limestone. The limestones dip to the southeast at about 13º, though the dip steepens near the several faults that are present. Flow was on the strike to the southwest. There is a major strike-oriented passage that descends from F1 to the Deep Zone, which is up to 415m below the former water table, and then rises to Fitzlischacht, which is close to the former spring (Jeannin *et al.,* 2000). Four passages in Siebenhengste Cave descend downdip to depths of 200–300m below the former water table where they join the main strike-aligned conduit. The nearby cave A2 has a similar pathway.

The two flow paths shown in Fig. 9 are close to the ideal situation of a single loop below the water table, as shown in Figures 6b and 7b. However, these two examples may not be representative, and further examples are given below.

Six cave profiles are shown in Fig. 12. These examples were chosen because they all have substantial maximum depths of flow below the water table. This makes comparison to the examples in Figures 6 and 7 simpler than for caves developed closer to the water table. In explaining the section of cave shown in Fig. 12a, Waltham and Brook (1980) wrote:

> *"If the horizontality of the trunk passage was due to the influence of the original water table, there seems no reason why the cave should have developed 50 metres below water level" (Waltham and Brook, 1980, p.131).*

Waltham and Brook made similar comments on Benarat Walk (Fig. 12b), and concluded:

> *"it is clear that the caves of Mulu must contribute data to any thoughts on cave genesis..... The great horizontal or sub-horizontal phreatic tubes mostly have an uncertain relationship to their contemporary water table, and raise questions on the distinction between shallow and deep phreatic cave development" (Waltham and Brook, 1980, p.138).*

The same occurrence of horizontal passages deep below the water table is seen in Nettlebed Cave (Figures 12c and 13) and Yorkshire Pot (Fig. 12d), both of which, like the Mulu examples, have flow along the strike in steeply dipping carbonates. At Nettlebed, flow in the Oubliette was along the strike of steeply-dipping fractures in marble (Fig. 13). At the Meltdown, the flow rose up at least 70m in an almost circular phreatic tube (Fig. 12c). In Yorkshire Pot, both the Roller Coaster and its tributary, Alberta Avenue, were formed along the strike of a steeply-dipping bedding plane in limestone.

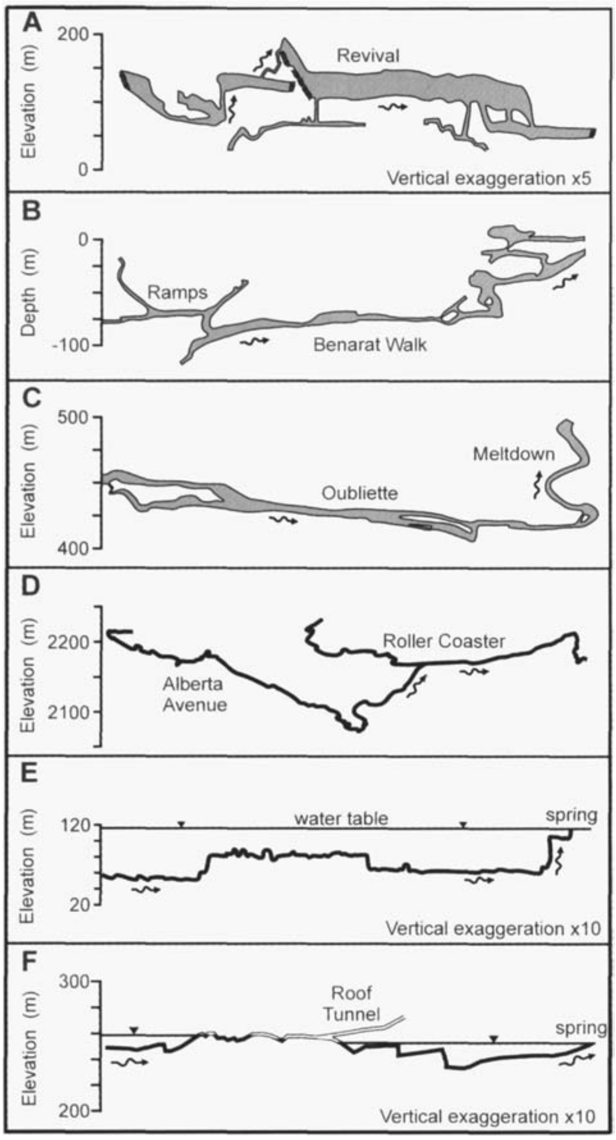


Fig.12. Profiles of conduits formed well below the water table. A) Clearwater Cave, Malaysia; B) Tiger Foot Cave, Malaysia; C) Nettlebed Cave, New Zealand; D) Yorkshire Pot, Canada; E) Doux de Coly, France; F) West Kingsdale System, England. (A and B after Waltham and Brook, 1980; C after Pugsley, 1979; D after Worthington, 1991; E after EKPP, 2004; F adapted from Brook and Brook, 1976 and Monico, 1995).

The situations at Doux de Coly and West Kingsdale are somewhat different because the conduits have formed along a number of bedding planes. Flow at Doux de Coly is along the strike of the limestone, which has a dip of 6º (Fig. 12e). Flow below the water table in the West Kingsdale System is updip (Fig. 12f). The conduit may have been formed up to 60m below the water table (Waltham, 1974) or at only 20m below the water table (Brook, 1974). Deep flow along horizontal passages 60m or more below the water table in Yorkshire has also been described for Duke Street in Ireby Fell Cavern (Waltham, 1974) and for Sleets Gill Cave (Waltham *et al.,* 1997).

Fig.13. Large phreatic tube in Nettlebed Cave, New Zealand. The flow from this phreatic passage ascended more than 80m at the Meltdown (photo by C. Pugsley).

The *"uncertain relationship to the contemporary water table"* noted by Waltham and Brook (1980, p.138) for the Mulu phreatic tubes also applies to the other caves in Fig. 12. These phreatic tubes most closely resemble the situation in Figures 6b and 7b, though not as well as Dent de Crolles and Siebenhengste (Fig. 9). This suggests that the initial flow field has had a powerful influence on subsequent cave formation in all these examples.

Data on phreatic looping below the water table may also be assessed by measuring the ratio of the depth of the crests to the depth of the bases of phreatic loops (i.e. D_c/D_b in Fig. 14). For instance, the mean depth below the water table of loop crests and loop bases is 35m and 58m, respectively for the conduit in Figure 14a, and so the crest/base ratio (D_c/D_b) is 0.60. For the similarly curving flow paths in Figures 6b, 7b, 9a and 9b the crest/base ratios are 0.52, 0.82, 0.45 and 0.77, respectively. On the other hand, the crest/base ratios are much smaller where the loop crests are close to the water table: 0.18 in Fig. 14b, 0.21 in Fig. 6c, 0.36 in Fig. 7c, and 0.28 in Fig. 9 of Ford and Ewers (1978).

Loop crest/base ratios for 12 sumps and for 8 fossil passages are given in Tables 1 and 2, respectively. The mean ratio of these twenty examples is 0.48. This high value is similar to the examples in Figures 9 and 12, and suggests that there is a tendency for the depth of phreatic conduits to be strongly influenced by hydraulic factors that cause loop crests to be well below the water table.

TABLE 1 Depth of the crests and bases of phreatic loops in 12 sumps

Cave	Length (m)	Maximum depth (m)	Loop crest mean depth (m)	Loop base mean depth (m)	Loop crest/base ratio	Number of loop crests and bases	Reference
Doux de Coly, France	5675	63	29	42	0.69	13	Figure 12e
Grotte de la Mescla, France	1510	95	30	59	0.51	18	Tardy, 1990a
Rinquelle, Germany	930	23	10.8	21.2	0.51	21	Farr, 2000
Grotte de Paques, France	810	50	4.8	19.6	0.24	11	Tardy, 1990b
Source du Lison, France	802	29	7.7	16.2	0.48	14	Isler, 1981
Joint Hole, England	700	19	3	12	0.21	11	Monico, 1995
Fontaine de Saint George, France	1520	76	5	15	0.33	15	Farr, 2000
Fontaine de Ressel, France	1865	81	19	37	0.56	12	Farr, 2000
Chaudanne Spring, Switzerland	608	140	43	59	0.69	17	Farr, 2000
Bätterich Spring, Switzerland	374	79	16	53	0.42	6	Farr, 2000
Blautopf, Germany	1250	24	10	17	0.53	25	Farr, 2000
Source de Bestouan	2955	29	5.6	18	0.31	14	Douchet, 1992

TABLE 2 Depth of the crests and bases of phreatic loops for 8 former conduit flow paths.

Cave	Length (m)	Maximum depth (m)	Loop crest mean depth (m)	Loop base mean depth (m)	Loop crest/base ratio	Number of loop crests and bases	Reference
Roller Coaster, Yorkshire Pot, Canada	522	58	33	42	0.78	12	Figure 12d
Grotte Annette - Trou de Glaz, France	1600	79	27	60	0.45	18	Figure 9a
Ogof Hesp Alyn, Wales	1100	59	6	28	0.21	12	Appleton, 1989
Réseau de Foussoubie, France	9000	130	68	77	0.89	80	Le Roux, 1989
Wookey Hole, England	1300	95	21	37	0.65	34	Farr, 2000
Rats Nest Cave, Canada	900	136	57	82	0.69	16	Yonge, 2001
Grand Circle - Bowling Alley, Cueva del Agua, Spain	1400	206	72	106	0.78	20	Smart, 1984
Big Rift - Hole in the Wall, Cueva del Agua, Spain	1200	104	65	83	0.77	27	Smart, 1984

TABLE 3 Structural and flow characteristics for 20 conduit flow paths (partly after Worthington, 2001)

No.	Cave	Geological age	Flow path length (m)	Stratal dip (degrees)	Range of elevation (m)	Mean flow depth (m)
1	Jordtulla, Norway	Palaeozoic	520	25	30	9
2	Otter Hole, Wales	Palaeozoic	3500	6	140	10
3	Friars Hole System, West Virginia	Palaeozoic	11000	2.2	200	17
4	Ogof Ffynnon Ddu, Wales	Palaeozoic	3100	12	350	35
5	Doux de Coly, France	Mesozoic	13700	6	200	43
6	Hölloch, Switzerland	Mesozoic	11000	16	1670	100
7	Nettlebed, New Zealand	Palaeozoic	7000	50	1200	120
8	Lubang Benarat, Malaysia	Cainozoic	7000	45	1600	150
9	Siebenhengste, Switzerland	Mesozoic	12000	13	1630	230
10	Mammoth Cave, Kentucky	Palaeozoic	2200	0.6	155	2
11	Horseshoe Bay Cave, Wisconsin	Palaeozoic	3000	2	30	6
12	West Kingsdale System, England	Palaeozoic	2700	3	130	35
13	Rio Encantado, Puerto Rico	Cainozoic	9500	4	260	15
14	Ogof Hesp Alyn, Wales	Palaeozoic	2000	14	80	24
15	Guanyan, China	Palaeozoic	6500	7	130	25
16	Swildons-Wookey, England	Palaeozoic	3500	15	176	40
17	Trou de Glaz, France	Mesozoic	1500	17	730	67
18	Peña Colorada, Mexico	Mesozoic	12000	40	1760	120
19	Nelfastla de Nieva, Mexico	Mesozoic	7400	70	1000	240
20	Edwards Aquifer, Texas	Mesozoic	180000	1.3	160	600

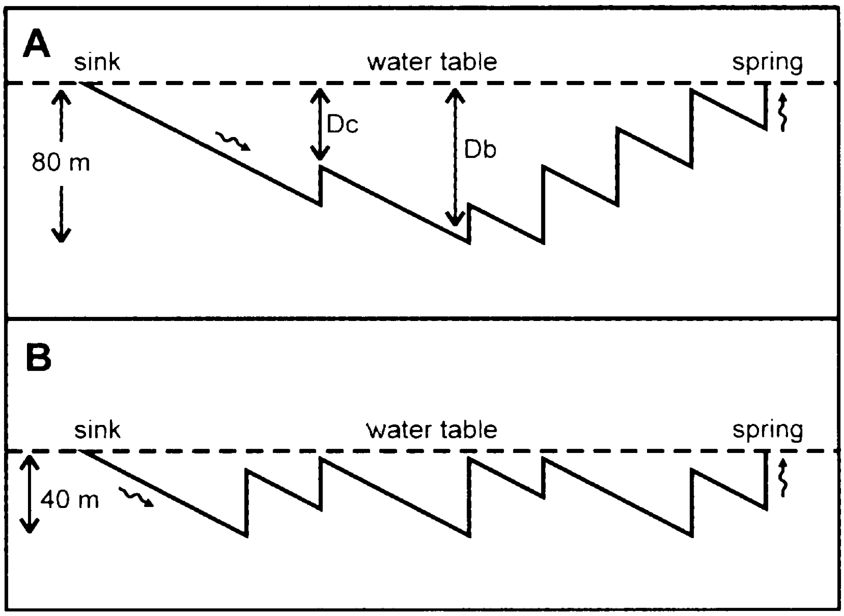


Fig.14. Phreatic flow paths with (A) a single loop deep below the water table and (B) with loop crests just below the water table.

Spacing of conduit-guiding bedding planes

Frequently, cave development occurs preferentially on a limited number of bedding planes, as noted above for the Friars Hole System, Castleguard Cave and the Dent de Crolles System. Lowe (2000b) described the preferential cave development in the Yorkshire Dales along just a few bedding planes, which he called inception horizons, and he discussed possible reasons why just a few bedding planes in a sequence are favoured for cave development.

It is likely that the presence of a limited number of inception horizons in an aquifer has only a marginal influence on the depth of flow. For instance, there are five major cave-guiding bedding planes in the 190m section of limestone in which the predominantly downdip Grotte Annette to Trou de Glaz passage is formed (Fig. 9a). If there were fewer or more major cave-guiding bedding planes in this section then the magnitude of the phreatic lifts would change, but the cave would still be able to follow a looping flow path that descends some 80m below the water table.

The number of cave-guiding bedding planes is also likely to have little influence on the depth of flow where flow is on the strike, such as at Friars Hole System (Fig. 4), Siebenhengste (Fig. 9b), or Mulu (Fig. 12a, 12b). In each case many bedding planes in the dipping limestone intersect the water table and cave passages are able to follow a single favourable bedding plane for substantial distances. On a particular bedding plane, the conduit is formed at a hydraulically favourable location. This may be just below the water table (e.g. Cleaveland Avenue, Mammoth Cave: Palmer, 1981), 10m below the water table (e.g. Friars Hole System: Fig. 4), 100m below the water table (e.g. Mulu: Figure 12a,b) or hundreds of metres below the water table (e.g. Siebenhengste: Fig. 9b).

Topography

Caves in mountainous areas tend to have deep flow paths, and several examples were noted above. These areas also tend to have steep stratal dips and large vertical ranges between the highest recharge points and spring elevations. The reasons why steep dips are associated with deep flow have been explained earlier. But the extent to which flow paths deep below the water table might be associated with mountainous topography is less clear. This is because the increase in fracture apertures in the early stages of karstification results in a large drop in hydraulic gradients and a vadose zone that may be more than 2000m thick, as at Krubera Cave in Georgia (Klimchouk *et al.,* 2004). If flow depth below the water table is a function of pre-karstification hydraulic gradients, then mountainous topography could be an important factor in promoting deep flow. On the other hand, depth of flow may be related more to the low and stable hydraulic gradients that exist following the early stage of karstification. Numerical modelling has demonstrated the rapid decline in hydraulic gradients in the early stages of karstification and so the steep early gradients may have little effect on later conduit development (Gabrovšek and Dreybrodt, 2001; Kaufmann, 2002). These ideas are tested using empirical data in the next section.

Discussion and conclusions

Discussion above of the factors influencing depth of flow suggests that deep flow is associated with longer flow paths, steeper dips, the presence of open joints and faults, and possibly with a large range in elevation in an aquifer. The four variables, depth of flow, flow path length, stratal dip and the range in elevation, can reasonably be estimated for a number of aquifers (Table 3). Worthington (2001) listed the first 19 examples shown in Table 3 and the remaining example is the aquifer associated with Comal Spring in Texas, which is the largest spring in the southwest USA. A recent numerical model of the aquifer feeding this spring includes a 140km-long conduit that is up to 1200m below the potentiometric surface (Lindgren *et al.,* 2004).

Whereas it is straightforward in many cases to evaluate flow path length, stratal dip and the range in elevation for a karst drainage basin, there is no simple way to evaluate fracture anisotropy. Conduits are developed on joints or at bed/joint intersections in most structural settings, with platform carbonates such as at Mammoth Cave being a notable exception. The latter also have low dips, and so stratal dip may act as a proxy measure of fracture anisotropy. Fig. 15 shows the relationship between depth of flow and flow path length, stratal dip and the range in elevation for the examples in Table 3. Regression of mean depth of flow against the different parameters in Fig. 15 gives:

$$D = 0.053\ L^{0.77} \quad (3)$$

$$D = 120\ \theta^{0.56} \quad (4)$$

$$D = 0.60\ E^{0.74} \quad (5)$$

$$D = 0.061\ L^{0.91}\ \theta^{0.72} \quad (6)$$

$$D = 0.016\ L^{054}\ E^{0.56} \quad (7)$$

$$D = 1.6\, E^{0.64}\ \theta^{0.72} \qquad (8)$$

$$D = 0.047\, L^{0.85}\ \theta^{0.64}\, E^{0.11} \qquad (9)$$

where D is the mean conduit depth in metres below the water table, L is the flow path length in metres, θ is the dimensionless stratal dip (equal to the sine of the dip in degrees), and E is the elevation difference in metres between the lowest spring and the highest recharge point to the aquifer. The correlation coefficients of the above seven equations are 0.64, 0.50, 0.68, 0.90, 0.80, 0.70 and 0.90, respectively. Equation 6, which uses length and dip, gives the best correlation. The addition of elevation as a dependent variable (Equation 9) does not improve the correlation. The correlation using length and dip can also be expressed as

$$D = 0.18\, (L\ \theta)^{0.81} \qquad (10)$$

This equation has a lower correlation (0.79) than Equation 6 but is simpler to express graphically with a best-fit line added (Fig. 15d). These results show that the two parameters of stratal dip and flow path length explain most of the variability in conduit flow depth.

The relationships that are apparent in Fig. 15 can be further tested by comparing pairs of caves from one area where the dip or flow path length varies between the two caves. The caves of the Alyn Gorge in Wales provide one such pair. Ogof Hen Ffynhonau and Ogof Hesp Alyn are adjacent caves with parallel flow routes. The latter lies further from the adjacent valley floor and has a deeper phreatic origin (Waltham *et al.,* 1997). The deeper flow in Ogof Hesp Alyn is explained by the longer flow path (Equations 6 and 10).

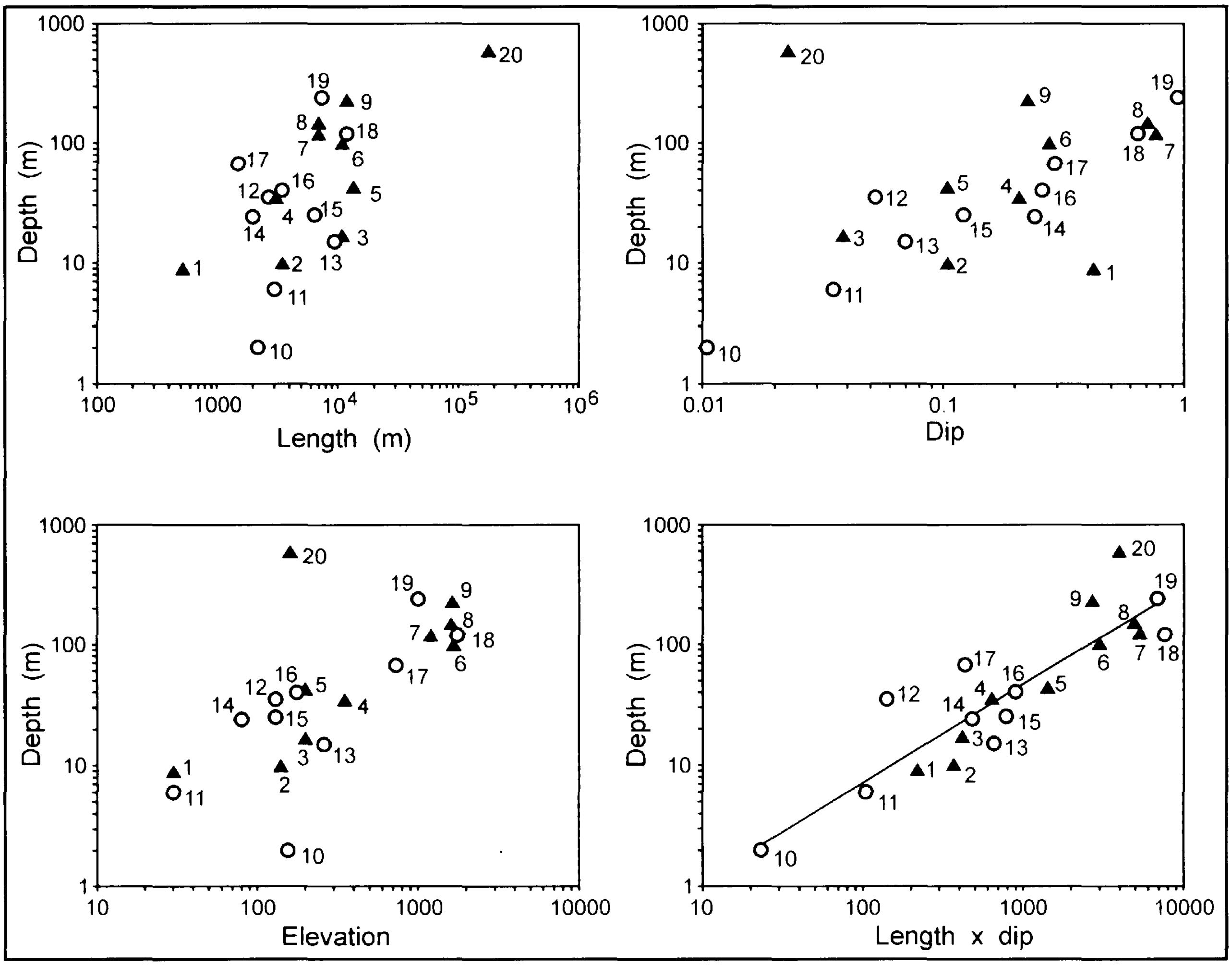


Fig.15. Correlation of conduit depth of flow below the water table with A) flow path length (top left); B) Stratal dip (top right); C) Range in elevation between recharge and spring (bottom left) and D) flow path length and dip (bottom right). Details of the 20 examples are given in Table 3.

The model presented above differs significantly from the Ford and Ewers (1978) model, which relates depth of flow to “fissure frequency”. The fissure frequency model attributes the occurrence of shallow or deep phreatic cave formation solely to the degree of fracturing. The model has been widely quoted in the literature, but suffers from three major problems. First, the model is based on the premise that shallow flow is the hydraulically most favoured pathway for conduit development, but Worthington (2001) showed that hydraulic factors favour deep flow in most karst aquifers (see Figure 1). The second problem is the lack of examples that fit the model and the many examples that contradict the model. For instance, mountainous areas typically have both substantial fracturing and deep flow, which is the opposite of the prediction of the model (Rossi *et al.,* 1997; Jeannin *et al.,* 2000). Conversely, limestones with little fracturing such as at Mammoth Cave and elsewhere in Kentucky have shallow flow, which again is the opposite of the model prediction. The third problem is that it appears that the hypothesis cannot be tested because Ford (2000) noted that the model does not provide predictions. If a model cannot provide predictions then it cannot be shown to have any applicability.

Fig. 16 gives a summary of how hydraulic and geological factors may interact. Figure 16a shows a limestone aquifer with a caprock such as shale. A simplifying assumption is made that the caprock has recently been eroded away from the limestone. Recharge is from a stream flowing off the caprock as well as from precipitation onto the limestone itself. The horizontal distance in the figure is in the range of several to several tens of kilometres. The vertical distance is in the range of several tens to several hundreds of metres, so there is substantial vertical exaggeration in the figure.

Fig. 16b shows flow lines in the limestone immediately after the removal of the caprock. If the limestone aquifer is of Palaeozoic age, then it is well-lithified, the porosity is up to a few percent, and most of the permeability is due to flow along bedding planes, joints and faults. The combined matrix and fracture hydraulic conductivity in such an aquifer is typically 10^{-6} – 10^{-5} m/s (Worthington *et al.,* 2000). Most of the recharge to the aquifer is assumed to be from the sinking stream, and so most of the flow lines emanate from the stream reach where it is sinking. Discharge from the aquifer occurs as a seepage zone in a valley.

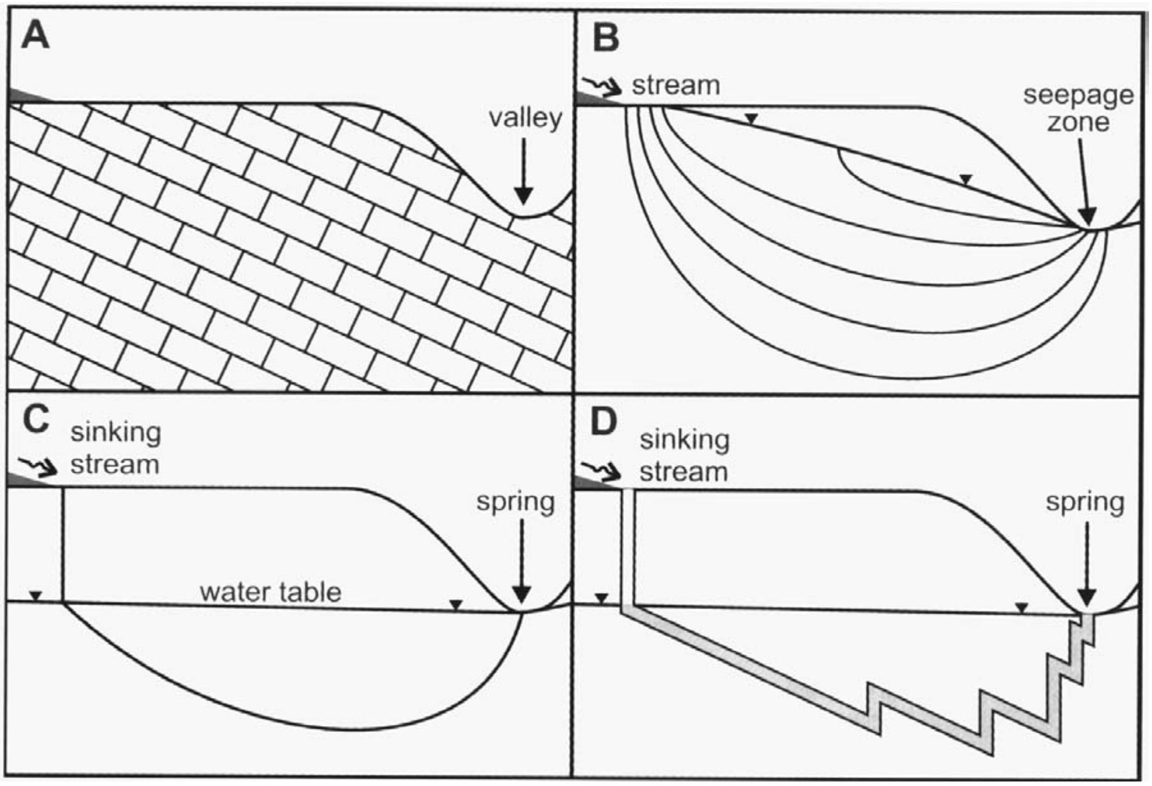


Fig.16. Factors influencing the initial conduit flow path in a limestone aquifer: A) geological contacts and location of valleys; B) initial flow field; C) theoretical flow path; D) actual flow path, following open fractures downdip and up joints.

The duration of the phase shown in Fig. 16b is extremely short (a few thousand years) because dissolution of bedrock fractures results in the formation of a conduit network that connects the recharge to the discharge area. The diagnostic features to differentiate the Figure 16c situation from that of Figure 16b in the field are the presence of discharge from discrete springs rather than from dispersed seepage, as well as the low hydraulic gradient. The low gradient is due to the increase in permeability, which is commonly two to three orders of magnitude (Worthington *et al.,* 2000). A series of simplified flow lines is shown in Figure 16b. The actual flow lines are less regular because they follow fractures, with the enlarged pathways being known as channels (Glennie, 1954) or primary tubes (Ford and Williams, 1989, p.253). Preferential enlargement of the channel that has the most favourable position and/or largest aperture results in the conduit shown schematically in Figure 16c. The most favourable depth for channel enlargement is a result of the balance between the competing factors of decreasing fracture aperture with depth and decreasing viscosity with depth; the former promotes shallow flow and the latter promotes deep flow. Equations 6 and 10 give approximations of the depth below the water table at which the conduit develops.

Fig. 16d shows a simplified scheme of the actual conduit flow path along fractures. In reality, caves typically follow individual joints for a few metres to a few tens of metres, and follow individual bedding planes for a few metres up to a few thousand metres. Consequently, many more fractures could be utilised than are shown in the figure.

The question was posed earlier as to why there are systematic differences between the shallow phreatic caves that occur in Kentucky, deeper phreatic caves in England, and much deeper phreatic caves that are common in mountainous areas. Equations 6 and 9 provide the answers. The Kentucky aquifers have shallow flow because of the platform setting, resulting in low dips and low permeability along joints. This is demonstrated by the small amount of passage development along joints and faults. Limestone aquifers in England have deeper flow because of the structural setting, resulting in steeper dips, open joints, many faults, and the concomitant cave development along joints and faults as well as along bedding planes. Mountainous areas commonly have even deeper flow because of a combination of open joints, steep dips and long flow paths.

Most conduit development in England is limited to a depth of about 100m because most flow paths are generally only a few kilometres long. From Equation 6 or Equation 10, the deepest flow is predicted to occur where there are steep dips and long flow paths. The Mendip Hills have steep dips and the Cheddar and Wookey Hole catchments are the longest in the Mendips, and so should have the deepest flow. The conduits feeding these spring are likely to descend to depths greater than 100m below the water table. Scuba diving has so far reached a depth of -58m at Cheddar and -70m at Wookey Hole and both conduits continue at depth beyond the limit of exploration (Farr, 2000). Similar conduit depths could occur along mineralised faults in Derbyshire such as on New Rake between P8 and Speedwell or on Faucet Rake between P9 and Speedwell (Gunn, 1991). Even deeper conduits may have formed in the Peak District along long regional flow paths which emerge at thermal springs (Worthington and Ford, 1995).

The model predicts that very deep flow is likely at major springs, which typically have long groundwater catchments. Mante Spring (Mexico) and Vaucluse Spring (France) are two prominent examples. These springs have been explored by scuba diving or submersible vehicles to depths in excess of 250m below the water table. Fish (1977) concluded that the flow path to Mante Spring is some 200km long and has a maximum depth of flow in excess of 1500m. The depth and length of the flow system feeding Comal Spring in Texas, which was described above, are of similar magnitude to those of the flow system feeding Mante Spring. Vaucluse Spring is the largest spring in France and its groundwater catchment is more than 60km long.

The model is applicable to most unconfined limestone aquifers and thus to most known caves and most limestone aquifers used for water supply. Its applicability in confined aquifers is limited because the depth of flow may be constrained by the structure. For instance, Bath Hot Springs are thought to be fed by geothermally-heated water flowing through a deep syncline (Waltham *et al.,* 1997, p.267). The model also does not apply to caves formed in special situations such as hypogene caves and sea coast mixing zone caves (Ford and Williams, 1989, 282–291).

The combination of flow path length and stratal dip (Equations 6 and 10) together with fracture anisotropy provide a satisfactory approximation that explains the depth of conduit flow in areas where the flow is at shallow depths below the water table (e.g. Kentucky), at moderate depths (e.g. England, Appalachian Mountains), or at great depths (e.g. many mountainous areas and major springs).

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