8.1. Porosity and Permeability Enhancement in Unconfined Carbonate Aquifers as a Result of Solution

Stephen R.H. Worthington, Derek C. Ford, and Patricia A. Beddows

Abstract

Solution processes in unconfined carbonate aquifers result in a network of channels. To characterize the enhancement of porosity and permeability by dissolution, we examine four contrasting carbonate aquifers:

- Paleozoic dolostone.
- Paleozoic limestone.
- Mesozoic chalk.
- Cenozoic limestone.

In all four cases, the channels add little to the porosity, but they enhance the permeability of fractured rock by one to three orders of magnitude. Similar porosity and permeability changes are predicted for all unconfined carbonate aquifers, in both dolostone and limestone, in both allogenic and autogenic settings, and in carbonate rocks of all ages.

Introduction

Carbonate rocks are chemical precipitates. Chemical processes therefore affect their permeability and porosity at all phases, from deposition through shallow or deep burial to exposure at the surface, as explained in Chapter 3.2.

During burial, porosity reduces from an original 40 to 70 percent (Choquette and Pray, 1970) to only a few percent, or even less in the case of deep burial. Porosity enhancement and the formation of caves does occur to a limited extent in this setting (Chapter 5.2).

But porosity enhancement is much more important on a global scale once the carbonate rock is elevated above sea level and any low-permeability overlying rocks are removed, so that the aquifer becomes unconfined.

This chapter focusses on the setting of unconfined carbonate aquifers. These aquifers contain at least 90 percent of the world's explored caves (Palmer, 1991), provide important water resources, and host almost all groundwater contamination problems in carbonate rocks.

Dissolution by infiltrating meteoric water has long been recognized to result in a significant enhancement of permeability in at least some carbonate aquifers. This is commonly recognized where major caves have been explored or where abundant karst landforms lie at the surface. However, hydrogeologists have no consensus as to the proportion of carbonate aquifers with significant amounts of dissolution-enhanced permeability.

The creation of channels, conduits, and caves results in an increase in porosity, but we find widely contrasting views on the extent of this increase. For example, Freeze and Cherry (1979, p. 27) give a porosity range of 0 to 20 percent for ordinary limestone, and 5 to 50 percent for karst limestone. However, Bonacci (1987, p. 47) compiled data from 20 case studies of different karst aquifers around the world, and found that the effective porosity ranges from 0.17 percent to 10 percent, with a median value of only 1.6 percent.

Karst features such as dolines are often taken to be definitive evidence of a karst aquifer below. But in many areas with sur-

8.1 Plate 1. Infilled dolines in Jurassic limestone revealed by quarrying near Woodstock, England. Away from the quarry, obvious karst features are absent.

ficial sediment the dolines may be infilled and unrecognizable (8.1 Plate 1). Furthermore, caves can exist beneath insoluble caprock, so no evidence for karst features may appear at the surface. Mammoth Cave (Chapter 5.3.8) is a good example, because most of that cave underlies ridges capped by sandstone.

In this chapter we consider a range of case studies in order to identify the characteristic enhancement of porosity and permeability found in carbonate rocks in unconfined aquifers.

Terminology

Confusing and contradictory terminology is a major barrier to communication of ideas regarding flow in carbonate aquifers. Meteoric water circulating through an unconfined carbonate aquifer tends to produce an integrated network of dissolutionally enlarged fractures or fissures (see the chapters of Parts 3 and 4 in this volume). These have been called channels (Choquette and Pray, 1970), macrofissures (Reeves, 1979), or secondary fissures (Price et al., 1993). Larger examples are known as conduits (>1 cm diameter) or caves (~>1 m, enterable by people) (White, 1988; Ford and Williams, 1989). The interconnectivity of the enlarged fractures is of prime importance, resulting in a channel network.

In this account we will use the term channel to refer to all interconnected dissolutional enlargements along joints, faults, and bedding planes, so we include a wide aperture range, from much less than 1 mm up to several meters.

Conceptual Models and Measurement of Porosity and Permeability

In order to measure the porosity and permeability of a carbonate aquifer, we first need a conceptual model of it so that the different porosity and permeability components can be representatively sampled. The different modes of flow in carbonate aquifers have been classified in several ways. Burdon and Papakis (1963) introduced the term diffuse circulation to describe Darcy flow through pores and fissures in carbonates, differentiating it from a second mode of flow, concentrated circulation. White and Schmidt (1966) suggested the terms diffuse flow and conduit flow. More recently, Atkinson (1985) divided flow into diffuse, fissure, and conduit components, whereas Quinlan and Ewers (1985) suggested granular, diffuse, fracture, and conduit components. Thus a plethora of terms has been used!

The term double porosity is becoming commonly applied in hydrogeology to refer to fractured rock aquifers, where flow in both the fractures and the rock matrix need to be considered. If we add the third element of channels, then we can simply refer to carbonates as triple porosity aquifers (Worthington, 1994; Quinlan et al., 1996).

The rock matrix can be sampled from borehole cores or in fresh quarry faces. This avoids a potential problem of near-surface porosity changes that occur with weathering of natural rock outcrops. Samples can be tested for permeability in a permeameter, and their porosity can also be measured. Borehole packer tests with small separation of the packers (<1 m) can also measure the permeability of the unfractured bedrock.

Fracture permeability is usually assessed by pump, packer, or slug tests in boreholes. These tests measure the aquifer response (usually a change in water level) to a stress such as pumping water or adding or withdrawing a slug of water in a borehole. Fracture porosity can be calculated by using the cubic law with packer test data (Snow, 1968)

$$b = (12\, T \mu / \rho\, g)^{0.33} \qquad (1)$$

where b is fracture aperture, T is transmissivity (determined from packer tests), μ is dynamic viscosity (0.0179 g/cm/s at 0° C, falling to 0.0080 at 30°), ρ is fluid density (1.0 at 0°, 0.996 at 30°), and g is the acceleration due to gravity. Equation 1 assumes that all the flow is in a single smooth-walled fracture. The roughness of fracture walls means that true width will be somewhat larger than the equation indicates. An alternative method of determining fracture porosity is to use slug or pump test results combined with observations of fracture spacing in cores or quarry faces.

Channel porosity (including conduits and caves) can be calculated in several ways. Atkinson (1977) estimated it from the volume of prestorm water flushed out of a spring following heavy rains. Such water is usually clear and has lower suspended-sediment and higher dissolved solid concentrations than the floodwater that succeeds it. There are also accurate survey data for many caves, which can be used to estimate cave volume. A third method is to model a channel network, calculating channel aperture from

$$A = C\, R / v \qquad (2)$$

where A is the channel cross section, C is the catchment area of the channel, R is the runoff (precipitation minus evapotranspiration), and v is channel velocity. Channel velocity may be approximated by an average from tracer tests between sinks and springs in carbonate aquifers. In 8.1 Figure 1 the results of 2877 such tests reported from around the world are shown. The global average channel flow velocity from these tests is 0.022 m/s (78 m/hr).

A critical component of channeling in carbonate rocks is the interconnectivity of the channels. Equation 2 requires the assumption that channels form a tributary pattern. The theoretical studies in Part 4 in this volume and the examples of caves in Part 5 both show that channels are expected to be interconnected. However, patterns of flow often are not clear on cave maps, because the maps show both the passages with active streams and those that have been abandoned.

A clearer pattern emerges if a map displays just the active flowpaths. Three representative channel networks are presented in 8.1 Figure 2, showing the currently active cave streams in particularly well-mapped cave systems (though in each case future cave exploration and mapping may add substantially to each network). In each cave the widths of most passages are exaggerated to make them visible. The smaller headwater passages are exaggerated in width up to 20 times, with minimum true widths being about 0.5 m.

Exaggeration of passage width to aid comprehension is common in small-scale cave maps. But it leads to the false impression that explorable caves occupy a significant fraction of the volume of a limestone aquifer. In each case several tributary cave streams join together, with the combined flow emerging at a spring. In all three caves both discharge and usually passage size increase downstream. Thus channel networks in carbonate aquifers in some respects resemble the flow patterns of surface rivers.

Examples of Porosity and Permeability Measurement in Carbonate Aquifers

The four examples of carbonate aquifers described below illustrate some of the different ways in which the three porosity elements of matrix, fractures, and channels have been characterized. The four aquifers have been chosen to provide a wide range of rock type (limestone and dolostone), recharge (allogenic and autogenic), and age and diagenetic maturity (Paleozoic, Mesozoic, and Cenozoic).

Lower Paleozoic Dolostone: Smithville, Ontario, Canada

The Niagaran dolostone (Silurian) of southern Ontario is in a low-dip cratonic setting like that at Mammoth Cave (Chapter 5.3.8). Surficial karst features are rare, as the area has been glaciated, and the bedrock is mostly covered by several meters of glaciolacustrine silt, clay, and other deposits. Nevertheless, where the dolostone is exposed, it is readily karstified and exhibits good pavement features. Small springs are common and a young cave, 250 m in length, has been mapped close to Smithville.

The hydrogeology was extensively investigated at Smithville (40 km east of Niagara Falls) where PCBs (very toxic polychlorinated biphenyls from electrical transformers) spilled into the bedrock. Tests on 1043 core samples gave an average porosity of 6.6 percent (P. Lapcevic, 1998, personal communication). Packer tests with 2 m spacing in 15 boreholes that penetrate the 38 m thickness of the dolostone gave a geometric mean hydraulic conductivity of 10^{-6} m/s and an arithmetic mean of 10^{-4} m/s (Lapcevic et al., 1997). The geometric mean is used to represent the average permeability in porous media, and it would apply if the fractures were discontinuous and poorly connected (Domenico and Schwartz, 1990, p. 67).

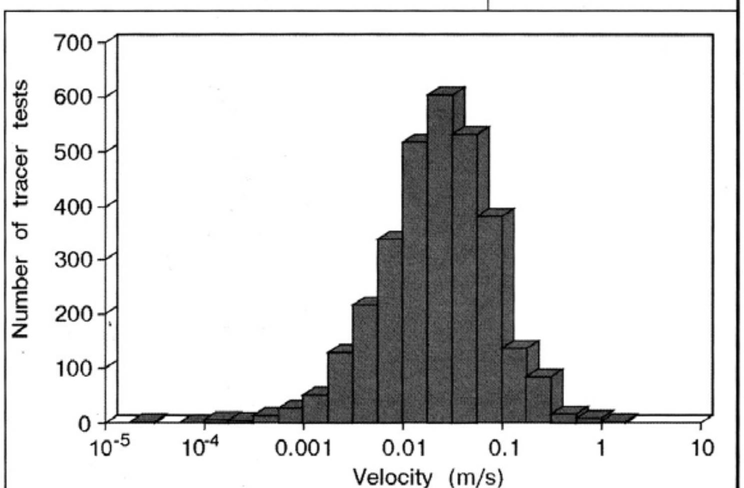
However, the bedding planes may be open for considerable distances here, as they are in most karst regions, so the fracture permeability is likely to be between the arithmetic mean and geometric mean.

The fracture porosity can be calculated using Equation 1, which gives 0.017 percent. This is a minimum value as it assumes smooth fractures and only includes bedding-plane porosity. Porosity due to jointing will add somewhat to this value, but in this low-dip cratonic tectonic setting, jointing is much less important than bedding planes with respect to both permeability and porosity. Some packer-test results were below the lower limit that the test could measure, 10^{-10} m/s, and these reflect the low permeability of the matrix.

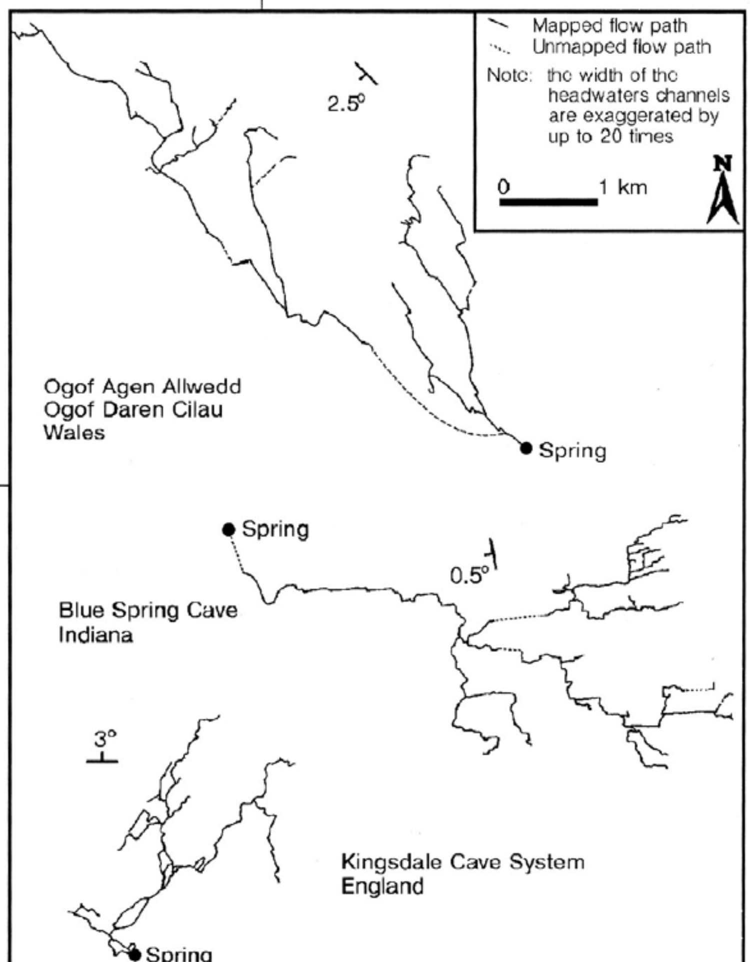
Point discharges at bedding planes in quarries and natural rock outcrops around Smithville suggest that channeling is focused on just four major bedding planes (8.1 Plate 2). With the assumption that 50 percent of runoff infiltrates through the overburden, Equation 2 was used to estimate that the mean channel porosity is 3 x 10^{-4} percent and the hydraulic conductivity is 3 x 10^{-4} m/s. Three pumping tests from a former municipal water well gave average hydraulic conductivities of 3 x 10^{-4} to 10^{-3} m/s. These high values support the theoretical calculations, suggesting that a high-permeability channel network surrounds the spill site.

Upper Paleozoic Limestone: Mammoth Cave Area, Kentucky, U.S.A.

Mammoth Cave (Chapter 5.3.8) is the most extensive known cave in the world. More than 500 km of interconnected passages have been



8.1 Figure 1. Velocities of 2877 traces in carbonate bedrock between sinking streams and springs.



8.1 Figure 2. Dendritic flow networks in three well-mapped groundwater catchments in carbonate bedrock.

8.1 Plate 2. Ice buildups in winter on a dolostone quarry face near Smithville, Ontario, Canada, showing point discharges from three bedding planes. In the foreground, a channel discharge of 0.3 L/s is so great that ice has not formed.

mapped in the cave, which underlies a surface area of about 100 km^2. There is a considerable body of knowledge on the explorable channels and their hydrogeology, both from studies within the cave and from measurements at springs (Palmer, 1981 and Chapter 5.3.8.; White and White, 1989). However, much less is known about matrix and fracture flow in the aquifer.

The limestone in the Mammoth Cave area is Mississippian in age, and dips gently to the northwest. The matrix porosity of the limestone, as measured from rock samples, averages 2.4 percent (Brown and Lambert, 1963; Worthington et al., 1998). Tests on cores from boreholes have given hydraulic conductivity values of 1 x 10^{-11} m/s for the rock matrix. (Brown and Lambert, 1963).

Fracture porosity and permeability have been determined from tests on a series of boreholes around Mill Hole, a karst window into the unexplored trunk channel drainage to Turnhole Spring, one of the largest springs in the region (8.1 Figure 3; Worthington et al., 1998). Downhole videos show that the spacing of fractures (mostly bedding planes) with visible enlargement is 0.7 m, but major open fractures with apertures of 2 to 12 cm have an average spacing of 8.5 m. Slug tests in nine boreholes yielded arithmetic and geometric mean hydraulic conductivities of 3 x 10^{-5} m/s and 6 x 10^{-6} m/s, respectively. Fracture porosity is about 0.03 percent , as calculated with Equation 1, using the slug-test results and fracture spacing from the downhole videos.

Channel storage has been evaluated in two ways for the Turnhole Spring catchment, which has a karst area without surface drainage of 159 km^2 (Worthington et al., 1998). Repeat tracer tests along a 6 km channel from a cave to the karst window gave velocities ranging from 0.0025 m/s to 0.13 m/s (Quinlan and Ewers, 1989), with a geometric mean close to the value of 0.022 m/s used in Equation 2. A total of 34 major flow paths were mapped in the Turnhole Spring catchment by Quinlan and Ray (1981), and the volume of all these major channels was calculated by Equation 2 to be 2.0 x 10^6 m^3.

A second method of calculating channel storage uses the displacement of water of low electrical conductivity following rainstorms. Hess and White (1988) determined an average storm-water travel time of 35 hours for an 11 km distance. The average length of 17 km for the entire groundwater catchment would give a travel time of 54 hours, and this can be combined with discharge data to give an estimated channel volume of 2.6 x 10^6 m^3 for the Turnhole Spring catchment (i.e. this is the approximate volume of active channels during this runoff event). The average of these two estimates for channel storage is 2.3 x 10^6 m^3. Cave passages have been recognized to descend to a depth of at least 23 m below the water table (Palmer, 1989, p. 311). If 23 m is taken to be the effective thickness of the aquifer, then the calculated channel volume represents 0.063 percent of the volume of the limestone aquifer.

The average flow in channels has been calculated by considering a cross section through the aquifer midway between the divide and the springs (8.1 Figure 3). At this point the mean discharge is 3.16 m^3/s. Using Darcy's Law with an average hydraulic gradient of 0.004 (Quinlan and Ray, 1981), matrix flow through the cross section is approximately 2 x 10^{-8} m^3/s, and fracture flow is 0.007 to 0.03 m^3/s, depending on whether the arithmetic or geometric mean of the slug-test data is used. Channel flow constitutes the remaining 3.12 to 3.15 m^3/s. From Darcy's Law, an equivalent hydraulic conductivity for the channel flow is 2.8 x 10^{-3} m/s (Worthington et al., 1998).

Mesozoic Limestone: The English Chalk

Chalk is a distinctive form of limestone that is composed mainly of skeletal calcite from algae. It accumulates in thick homogeneous formations in deeper marine environments and displays relatively little diagenetic alteration. It is rare at the global scale but is prominent in northwestern France, southern England, and a few other regions. The Cretaceous Chalk is the most important aquifer in Britain, and there have been a correspondingly large number of studies of this aquifer. The rock matrix has a high porosity, usually in the range of 20 to 45 percent. But the pore throats are very small (0.1 to 1 µm), which results in very low hydraulic conductivity values (10^{-9} to 10^{-7} m/s). In deep-burial situations (confined flow), such as in the North Sea oil fields, fracturing enhances the hydraulic conductivity to about 10^{-6} m/s (Price et al., 1993).

In the unconfined Chalk aquifer there is further permeability enhancement along fractures. Aquifer tests have shown that the uppermost 30 to 60 m of the saturated Chalk has the highest permeability, which has been attributed to dissolution (Foster and Crease, 1975; Headworth et al., 1982). Furthermore, it has been suggested that permeability is greater along valleys, and that this is due to a combination of stress relief and dissolution.

This results in increasing hydraulic conductivity, from about 4 x 10^{-6} m/s under interfluves to about 6 x 10^{-4} m/s underneath valleys, and an average fracture porosity of 0.01 percent (Price et al., 1993). However, the enhanced hydraulic conductivity values under valleys only represent a small fraction of the total area of the aquifer. If we use the spatial variations in transmissivity for the unconfined chalk in the County of Berkshire (given in Connorton and Reed, 1978), then the areal average for hydraulic conductivity in the upper 50 m of the aquifer is 6 x 10^{-5} m/s.

Some studies suggest that high permeabilities are not restricted to valleys. In Berkshire, for instance, Morel (1980) gives 49 transmissivity values, with the highest occurring under an interfluve. Furthermore, in Dorset, Houston et al. (1986) found that major north-trending troughs in the water table are discordant to the trends of the major dry valleys there. Gradients along the troughs were found to be up to 2 orders of magnitude lower than elsewhere in the region, implying that hydraulic conductivity in the troughs is correspondingly higher. The troughs terminate at the coast, at major springs. In the Chalk in Yorkshire, recent tracing from wells has demonstrated groundwater velocities of 400 m/day, both along dry valleys and from interfluves to the valleys (Ward et al., 1997).

The geometry of the high-permeability fractures is unclear. Price et al. (1989, p. 5) suggest that these fractures "should not be thought of as simple parallel-plate features." Occasionally, boreholes intersect openings with apertures >1 cm (Price et al., 1982; Price, 1996, p. 86), but more commonly the apertures are less. Dissolution has played the major role in enlarging apertures from their original (tectonic) widths (<0.1 mm) to typically 0.1 to 10 mm. It follows that the positive feedback processes that lead to channeling in other carbonate aquifers (8.1 Figure 2) should also occur here, resulting in interconnected networks of channels that are lenticular in cross section and are elongated along the fractures. This is supported by calculations using packer-test data. These indicate extensive interconnectivity of solutionally enlarged fractures in the Chalk (Price, 1994).

The porosity of these solutionally enlarged fractures, or channels, can be estimated from some studies. Foster and Crease (1975) considered that most of the permeability at a site in Yorkshire was imparted by "a small number (perhaps 4 or 5) of horizontal master conduits" with openings of perhaps 2 mm. Headworth (1978) found two to seven major fissures in each of a series of 19 boreholes in Hampshire.

Using the transmissivity of 2800 m^2/day measured there, one can show with Equation 1 that this transmissivity could be imparted by two fractures with 3 mm apertures or seven fractures with 1.9 mm apertures. If we postulate that the upper 50 m of the saturated zone represents the effective aquifer, then channel porosity from these two studies lies in the range 0.01 to 0.03 percent

Such channeling is supported by observations in quarries (Mortimore, 1993, p. 71) and by the occurrence of springs. Discharge from the Chalk is generally "from small springs and seepages into rivers" (Lloyd, 1993, p. 236). This concentration of discharge at specific points in places can be explained in structural terms (faulting) or lithologic terms (contact with an aquitard). However, the widespread occurrence of springs in the Chalk and in other carbonates most likely represents the discharge points from channel networks. Foster (1978, p. 92) suggested:

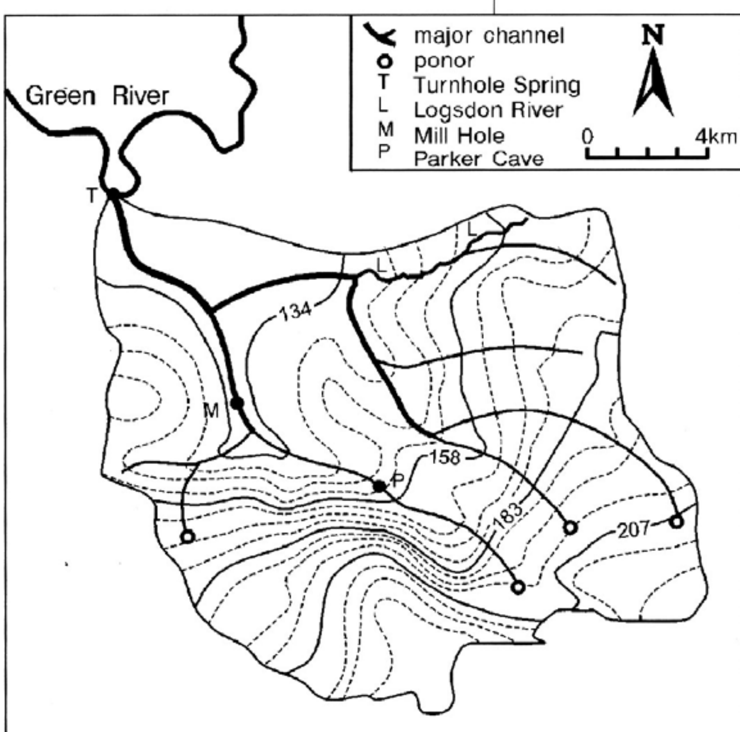
> High transmissivity appears to be due to the solution enlargement of fissures, which shows a general tendency to increase down groundwater flow lines to a maximum in the areas of current groundwater discharge.

The widespread intersection of solutionally enlarged fractures in boreholes in the Chalk suggests that these are not single channels such as are found in most caves in most types of limestone. Possibly they occur as series of interconnected anastomosing channels.

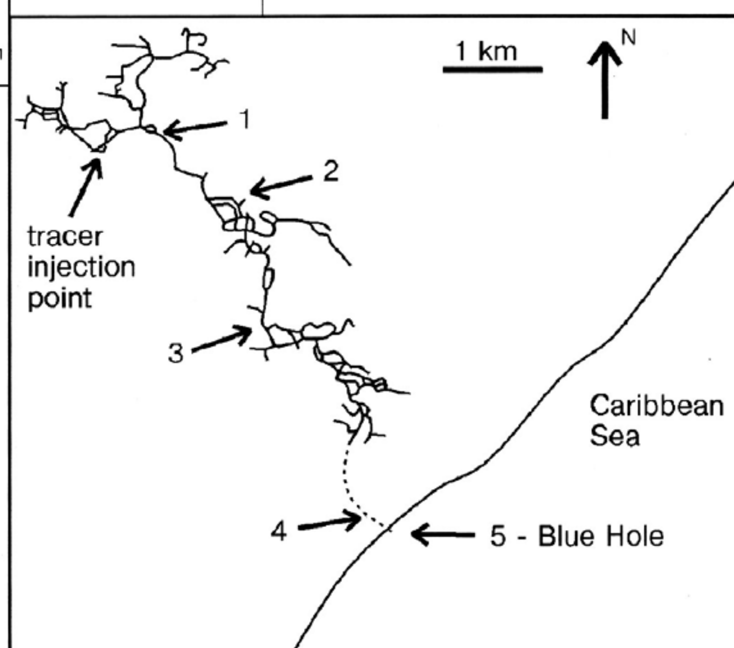
A small number of groundwater traces have been conducted from dolines or sinking streams to springs in the Chalk (Atkinson and Smith, 1974; Banks et al., 1995). These have yielded velocities of 0.02 to 0.07 m/s, which are typical of sink-to-spring velocities in limestone (8.1 Figure 2). However, most such dolines and sinking streams occur close to the contact between the Chalk and overlying lower-permeability strata, and thus are associated with the high recharge rates from allogenic runoff.

Cenozoic Limestone: Yucatan, Mexico

The Yucatan Peninsula in Mexico is a low-lying region of Cenozoic limestone, mostly Eocene to Pleistocene in age. Being young, it is diagenetically quite immature and displays high porosities. Harris (1984) measured an average porosity of 17 percent for near-coastal limestone, though the porosity may be somewhat lower for the older limestone farther inland. The pores are large and well connected, with the result that the hydraulic conductivity of rock cores measured in laboratories is in the range 10^{-6} to 5 x 10^{-3} m/s. Pumping tests in wells give much higher hydraulic conductivity values, ranging from 9 x 10^{-4} m/s to 10^{-2} m/s (Gonzalez Herrera, 1992). This increase in hydraulic conductivity of one to two orders of magnitude reflects the high permeability of fractures and possibly of channels. The fracture porosity can be estimated using Equation 1–if fracture apertures are less than 1 mm, then the minimum fracture porosity is 0.1 percent .



8.1 Figure 3. Groundwater catchment for Turnhole Spring, showing water-level contours and traced flow paths (after Quinlan and Ray, 1981).



8.1 Figure 4. Map of the major passages in Nohoch Nah Chich (after Hutcheson and Madden, 1994), showing traced flow paths. The whole cave is below the water table.

Two regional computer simulations have been produced for the northern part of the Yucatan Peninsula, with modeled hydraulic conductivity ranging from 0.1 to 1 m/s (Marin, 1990) and from 0.15 to 1.1 m/s (Gonzalez Herrera, 1992). These extremely high values are generated by the exceptionally low, real water-table gradients of 0.000005 to 0.00001 that are fed into the models. The two orders of magnitude increase over pumping-test values suggests that there is a high-permeability channel network that was not sampled by those tests. The difference between the geometric-mean hydraulic conductivities from pumping tests (10^{-3} m/s) and from numerical models (0.36 m/s) is 0.36 m/s, and this represents an approximation of the hydraulic conductivity due to channeling.

It has been known for some time that there are large coastal and submarine springs on both the northern and eastern coasts of the Yucatan Peninsula. Geochemical studies at coastal springs have demonstrated enhanced dissolution in the narrow coastal mixing zone between fresh and salt water (Back et al., 1984). Inland, there are many cenotes, which are roughly circular roof collapses into flooded conduits. Some cenotes are tens of kilometers from the coast, far removed from any simple mixing between fresh water and sea water. However, until recently there was no definite proof that large-aperture channels connected these inland cenotes to the coastal springs. In view of the lack of positive proof it has been assumed in a number of papers that no signficant channel networks are present.

Exploration of the flooded caves of the Yucatan Peninsula began in earnest in the mid-1980's. Since then more than 200 km of caves close to the east coast of the peninsula have been explored by scuba divers. At several locations, the coastal springs themselves have been dived inland into major channel networks. The most extensive cave known so far is Nohoch Nah Chich, which has been mapped from the coast to a distance 6 km inland (8.1 Figure 4, p. 467). Measurements have shown that there is a continuous channel over this distance through which there is a discharge >1.3 m^3/s; this channel may extend many kilometers further inland.

Tracer testing by Beddows (1998) has measured groundwater velocities of 0.01 to 0.03 m/s along the major channel, which has an average cross section of 120 m^2. The cave exploration and tracing testing demonstrate that dissolution has resulted in a network of channels which impart an extremely high permeability to this aquifer: This explains the hundred-fold increase in permeability over that obtained from pumping tests. A 1994 map of the cave showed that the volume of the 39 km of passages mapped at that time was 4 x 10^6 m^3. The cave underlay a surface area 5.5 km long and 1.9 km wide, and had a vertical range of 80 m. Thus the cave occupies 0.5 percent of that specific volume of limestone.

Discussion

Porosity Enhancement as a Result of Dissolution

Porosity values for the four aquifers described above are presented in 8.1 Table 1. The Paleozoic carbonates in Kentucky and Ontario are diagenetically mature and were buried to great depths beneath cover strata. They have much lower matrix porosities than the younger carbonates of the English Chalk and Yucatan Peninsula, which have not been buried to a significant extent. However, there are no consistent contrasts in fracture or channel porosity between the younger and older carbonates: the values are all much lower than the matrix porosities. Consequently, in all four aquifers almost all the storage is in the matrix (8.1 Table 2).

Permeability Enhancement as a Result of Dissolution

The permeability values for the four aquifers are given in 8.1 Table 3. At least 96 percent of aquifer storage is provided by the matrix of the rock (8.1 Table 1). However, the proportion of the groundwater flow (i.e. mass flux) that takes place through this matrix is minimal in all cases. It is calculated by

$$F_m = K_m / (K_m + K_f + K_c) \qquad (3)$$

where F is the fraction of flow, K is hydraulic conductivity, and the subscripts m, f, and c refer to the matrix, fractures, and channels, respectively. The proportion of flow that takes place through fractures and channels can be similarly calculated.

Results are shown in 8.1 Table 4. In each case almost all the flow takes place through the channel network, with fractures providing a small proportion of the flow, and the matrix a negligible amount.

Characterization of Channel Networks

These four sample aquifers display a very wide variety of initial conditions, yet as modern unconfined aquifers they are similar in many important respects. Almost all the storage is in the rock matrix, but almost all the flow takes place through channels. The formation of channel networks is the result of a positive-feedback loop between discharge and dissolution flux in developing channels (see Chapter 4.2.1). Networks are able to form because of the decrease in dissolution rate as groundwater approaches thermodynamic equilibrium with respect to calcite (Chapters 4.1 and 4.2). The implication of the models discussed in Chapter 4.2 is that channel networks should be found in all carbonates where there is a sufficient flux of infiltrating meteoric water, including all unconfined carbonates.

Major dissolutional enhancement of permeability in carbonate aquifers is dependent on the circulation of groundwater (Chapter 3.4). Thus, the capability must exist for:

- Chemically aggressive groundwater to recharge the system.
- The rock unit to transmit water through the fractures.
- Groundwater to drain from the system (Stringfield and LeGrand, 1966).

These three conditions are best satisfied in unconfined carbonate aquifers, which readily allow for recharge to and discharge from the aquifer. The four examples of carbonate aquifers described here represent a great range in rock type, recharge type (allogenic and autogenic), and age (Paleozoic, Mesozoic, and Cenozoic). These three factors will be considered in turn.

Dissolution rates are much slower for dolomite than for calcite. Where dolostone and limestone abut, then karst features are usually found to be better developed on the limestone (Buser et al., 1976). However, long caves are also known in dolostone, including:

- The longest cave in South America—Toca del Boa Vista, Brazil, with 78 km of passages mapped.
- Several long caves in Missouri, U.S.A., either partly or wholly formed in dolostone:
    - Mystery Cave, 27 km.
    - Berome Moore Cave, 27 km.
    - Rimstone River Cave, 22 km.
    - Carroll Cave, 18 km).

Furthermore, there have been many long-distance dye traces in dolostone. In Missouri, for instance, Aley (1978) described nine traces over distances greater than 40 km. Tracer velocities ranged from 0.009 m/s to 0.059 m/s. Thus the development of efficient dissolutional channel networks in dolostone is not uncommon.

Channeled allogenic water sources are optimum for rapid and efficient cave formation. Surface streams of highly aggressive water flowing from low-permeability rocks onto carbonates usually quickly disappear into sinkholes and caves. Many caves are readily accessible at sinking streams. In autogenic settings, the epikarst and other entry channels are small, so caves are less easy to explore, as noted in the cone karst at Caves Branch in Belize, for example (Chapter 5.3.12).

However, more than 50 caves have now been explored that are greater than 1000 m

Area	Porosity (%)		
	Matrix	Fracture	Channel
Smithville, Ontario, Canada	6.6	0.02	0.003
Mammoth Cave, Kentucky, U.S.A.	2.4	0.03	0.06
The Chalk, England	30	0.01	0.02
Nohoch Nah Chich, Yucatan, Mexico	17	0.1	0.5

8.1 Table 1. Matrix, fracture, and channel porosity in four carbonate aquifers. The channel porosity in the chalk is equivalent to that of the "secondary fractures" of Price et al., 1993.

Area	Proportion of storage (%)		
	Matrix	Fracture	Channel
Smithville, Ontario, Canada	99.7	0.3	0.05
Mammoth Cave, Kentucky, U.S.A.	96.4	1.2	2.4
The Chalk, England	99.9	0.03	0.07
Nohoch Nah Chich, Yucatan, Mexico	96.6	0.6	2.8

8.1 Table 2. Fractions of storage contributed by matrix, fracture, and channel porosity in four carbonate aquifers.

Area	Hydraulic conductivity (m s^{-1})		
	Matrix	Fracture	Channel
Smithville, Ontario, Canada	1×10^{-10}	1×10^{-5}	3×10^{-4}
Mammoth Cave, Kentucky, U.S.A.	2×10^{-11}	1×10^{-5}	3×10^{-3}
The Chalk, England	1×10^{-8}	4×10^{-6}	6×10^{-5}
Nohoch Nah Chich, Yucatan, Mexico	7×10^{-5}	1×10^{-3}	4×10^{-1}

8.1 Table 3. Matrix, fracture, and channel permeability in four carbonate aquifers.

Area	Proportion of flow (%)		
	Matrix	Fracture	Channel
Smithville, Ontario, Canada	0.000003	3.0	97.0
Mammoth Cave, Kentucky, U.S.A.	0.00	0.3	99.7
The Chalk, England	0.02	6.0	94.0
Nohoch Nah Chich, Yucatan, Mexico	0.02	0.2	99.7

8.1 Table 4. Fractions of flow contributed by matrix, fracture, and channel permeability in four carbonate aquifers. The channel flow in the chalk is equivalent to that of the "secondary fractures" of Price et al., 1993.

deep, and almost all of them have no allogenic recharge. In these caves, typically the uppermost 100 to 300 m of passages are small, as they were formed by the small amounts of water that drain from just a few dolines. At greater depths, these surface tributaries join, and passage size and discharge both increase. Thus channel networks and extensive caves are common in autogenic as well as allogenic settings.

As repeatedly noted in this book, the older carbonates have usually been subjected to deep burial, resulting in porosity reduction to a few percent or less. This applies to most Proterozoic and Paleozoic carbonates (e.g. Smithville, Ontario; Mammoth Cave, Kentucky) and to Mesozoic carbonates in alpine fold belts (e.g. Nort of Lake Thun Caves, Chapter 5.3.3; Sistema Cheve, Chapter 5.3.6). Younger carbonates that have escaped folding and deep burial have much higher porosities, such as the English Chalk, the limestone of the Yucatan Peninsula, and young carbonate islands (Chapter 5.1). Channel networks clearly develop in both low-porosity and high-porosity carbonates.

Sampling is a major problem in the hydrologic analysis of carbonate aquifers, because boreholes (the conventional exploration tool in hydrogeology) are unlikely to intersect the major channels that are conveying most of the flow and any contaminants. Some statistics for ten carbonate aquifers containing well-explored caves are provided in 8.1 Table 5, including the cave networks illustrated in 8.1 Figure 2. We will refer to the fraction of the bedrock that is occupied by an explored cave as the cave porosity. In all cases it is less than 0.5 percent. These data thus are comparable to the porosities in 8.1 Table 1, where channel porosities as low as 0.003 percent were calculated.

The final column of 8.1 Table 5 expresses the probability of a standard borehole (which is only 10 to 15 cm in diameter) intercepting the mapped caves. Most values lie between 0.5 percent and 2 percent. The highest value (7.5 percent) is for southern Gunung Api. This is an area that includes many very large cave passages and the largest explored cave room in the world (Sarawak Chamber, volume 1.2×10^7 m^3)—thus it represents an extreme value. Even in this extreme case, however, the probability of a borehole intersecting a known cave passage is only 1 in 14. In most cases, the probability ranges from 1 in 50 to 1 in 200. Clearly, boreholes cannot be relied on to detect the presence of caves.

Definition of Channel Networks

Four aquifers have been considered in detail in order to demonstrate the extent of dissolutional enhancement of porosity and permeability in unconfined carbonate strata. In all four cases we have shown that

Cave	Volume of rock: length x width x height m [1]	Volume of cave x 10^6 m^3 [2]	Length of cave km [2]	Cave porosity % [3]	Areal coverage of cave % [4]
Ogof Agen Allwedd - Ogof Daren Cilau, Wales	6200 x 1900 x 50	0.9	75	0.15	1.7
Blue Spring Cave, Indiana, U.S.A.	5100 x 2600 x 45	0.5	32	0.08	1.1
Kingsdale Cave System, England	2600 x 1500 x 100	0.17	20	0.04	1.8
Nohoch Nah Chich, Yucatan, Mexico	5500 x 1900 x 80	4	39	0.48	6.5
Mammoth Cave, Kentucky, U.S.A.	11000 x 9000 x 90	8	550	0.09	1.4
Castleguard Cave, Canada	6500 x 1200 x 400	0.12	20	0.004	0.51
Friars Hole System, West Virginia, U.S.A.	6000 x 2000 x 80	2.7	70	0.28	2.5
McFail's Cave, New York, U.S.A.	3500 x 2300 x 90	0.12	11	0.016	0.37
Skull Cave, New York, U.S.A.	1300 x 940 x 60	0.046	6	0.064	1.2
Caves in Southern Gunung Api, Malaysia	7000 x 2500 x 400	30	110	0.43	7.5

8.1 Table 5. Cave porosity and areal coverage for some well-mapped caves.
Notes:
(1) This represents the minimum rectangular block of rock that can contain the 3-D array of mapped passages in each cave.
(2) These refer to the explored and mapped cave passages. Increases in these values are likely as the caves become more completely explored.
(3) Cave porosity is defined as the volume of mapped cave divided by the minimum rectangular block of rock that can contain the cave, times 100.
(4) The areal coverage is the plan area of the cave divided by the minimum rectangular area which can contain the cave, times 100.

In all four cases we have shown that enhancement of porosity by dissolution is relatively minor. On the other hand, the enhancement of permeability is considerable, because dissolution created dendritic networks of interconnected channels that are able to convey 94 percent or more of the flow in the aquifer.

Opinions range as to whether such high-permeability networks of solutionally enlarged fractures should be termed a karst (or karstic, or karstified) aquifer. Atkinson and Smart (1981, p. 182-183) recognized the problem of defining the term, stating:

> There is no widely accepted definition of karstic groundwater flow, but in our opinion the term should be restricted to solution conduits in which a turbulent flow regime occurs. It is thus possible to envisage a spectrum of aquifer types from the wholly karstic, in which all groundwater flow and storage occurs in solution conduits, to the non-karstic fissured aquifer, in which solutionally-widened fissures are common but turbulent flow in conduits is a rarity.

A much broader definition has been proposed by Huntoon (1995, p. 344):

> A karst aquifer is an aquifer containing soluble rocks with a permeability structure dominated by interconnected conduits dissolved from the host rock which are organized to facilitate the circulation of fluid in a downgradient direction wherein the permeability structure evolved as a consequence of dissolution by the fluid.

Atkinson and Smart (1981) classify the English Chalk as being close to the nonkarstic fissured aquifer end of the spectrum, whereas the Carboniferous Limestone in Britain (in which most of the well-known caves are found) is classified as being closer to the karstic end of the spectrum. However, it is instructive to compare the differences in the responses of these two aquifers where adits have been driven into them below the water table. In the Chalk near Brighton, a total of 13.6 km of adits were driven to provide a water supply (Downing et al., 1993).

In some adits, water-yielding fissures were encountered every 7 to 12 m, but in others the interval was much larger, commonly 150 to 200 m, and in a few, over 500 m. Individual fissures yielded very large flows of the order of 5000 to 15,000 m^3/day (58 to 174 L/s) (Price et al., 1993, p. 48).

In North Wales the Halkyn Deep Level (12 km of adits) and Milwr Sea Level Tunnel (26 km of adits) were driven to dewater lead mines in the Carboniferous limestone. Appleton (1989) describes several inflows to the adits, with the largest inflow being 380 L/s. Many solution cavities were found, most of which were located along mineralized veins. The adits did not intersect any major caves, even though several are known in the limestone outcrop (Appleton, 1989).

In both the Cretaceous chalk and the Carboniferous limestone the inflow into the adits is concentrated at point inputs with discharges up to several hundred liters per second. In both cases dissolution has resulted in a channel network that contributes minimally to enhancing aquifer porosity, but has greatly enhanced aquifer permeability. Therefore these aquifers are very similar in terms of hydraulic functioning.

Wherever comprehensive evidence has been collected in unconfined carbonate aquifers, it suggests that dissolution inexorably results in a similar permeability structure, with channel networks providing most of the permeability of the aquifer, yet occupying a very minor fraction of aquifer volume. We believe that all unconfined carbonate aquifers function in a similar manner.

References

Aley, T., 1978, A predictive hydrologic model for evaluating the effects of land use and management on the quantity and quality of water from Ozark springs: *Missouri Speleology*, v. 18, p. 1-185.

Appleton, P., 1989, Limestones and caves of North Wales, in Ford, T.D., ed., *Limestones and caves of Wales*: Cambridge University Press, p. 217-254.

Atkinson, T.C., 1977, Diffuse flow and conduit flow in limestone terrain in the Mendip Hills, Somerset (Great Britain): *Journal of Hydrology*, v. 35, p. 93-110.

Atkinson, T.C., 1985, Present and future directions in karst hydrogeology: *Annales de la Société géologique de Belgique*, v. 108, p. 293-296.

Atkinson, T.C., and Smart, P.L., 1981, Artificial tracers in hydrogeology, *in A survey of British hydrogeology 1980*: London, Royal Society, p. 173-190.

Atkinson, T.C., and Smith, D.I., 1974, Rapid groundwater flow in fissures in the Chalk–An example from South Hampshire: *Quarterly Journal of Engineering Geology*, v. 7, p. 197-205.

Back, W.B., Hanshaw, B., and van Driel, J.N., 1984, Role of groundwater in shaping the eastern coastline of the Yucatan Peninsula, *in* LaFleur, R.G., ed., *Groundwater as a geomorphic agent*: Boston, Allen and Unwin, p. 280-293.

Banks, D., Davies, C., and Davies, W., 1995, The Chalk as a karstic aquifer: Evidence from a tracer test at Stanford Dingley, Berkshire, UK: *Quarterly Journal of Engineering Geology*, v. 28, p. S31-S38.

Beddows, P.A., 1999, McMaster University, *Conduit hydrology of a Tropical Carbonte Aquifer:Caribbean Coast of the Yucatan Peninsula*, M.Sc. thesis, McMaster University, 162 p.

Bonacci, O., 1987, *Karst hydrology, with special reference to the Dinaric Karst*: Berlin, Springer-Verlag, 184 p.

Brown, R.F., and Lambert, T.W., 1963, Reconnaissance of ground water resources in the Mississippian Plateau region, Kentucky: *U.S. Geological Survey Water Supply Paper* 1603, 58 p.

Burdon, D.J., and Papakis, N., 1963, *Handbook of karst hydrogeology*: Institute for Geology and Subsurface Research, Athens/United Nations Food and Agricultural Organization, 276 p.

Buser, S., Drobne, F., and Gospodarič, R., 1978, Geology and hydrogeology, *in* Gospodarič, R., and Habič, P., eds., *Underground water tracing–Investigations in Slovenia 1972-1975*: Postojna, Institute for Karst Research, p. 27-38.

Choquette, P.W., and Pray, L.C., 1970, Geological nomenclature and classification of porosity in sedimentary carbonates: *American Association of Petroleum Geologists Bulletin*, v. 54, p. 207-250.

Connorton, B.J., and Reed, R.N., 1978, A numerical model for the prediction of long term well yield in an unconfined chalk aquifer: *Quarterly Journal of Engineering Geology*, v. 11, p. 127-138.

Domenico, P.A., and Schwartz, F.W., 1990, *Physical and chemical hydrogeology*: New York, John Wiley, 824 p.

Downing, R.A., Price, M., and Jones, G.P., 1993, *Hydrogeology of the Chalk of North-West Europe*: Oxford, Clarendon Press, 300 p.

Ford, D.C., and Williams, P.W., 1989, *Karst geomorphology and hydrology*: London, Unwin Hyman, 601 p.

Foster, S.S.D., 1978, Discussion, in Thames groundwater scheme: *London, Institution of Civil Engineers*, p. 91-94.

Foster, S.S.D., and Crease, R.I., 1975, Hydraulic behaviour of the Chalk aquifer in the Yorkshire Wolds: *Proceedings of the Institution of Civil Engineers*, v. 59, p. 181-188.

Freeze, R.A., and Cherry, J.A., 1979, *Groundwater*: Englewood Cliffs, Prentice Hall, 604 p.

Gonzalez Herrera, R.A., 1992, *Evolution of groundwater contamination in the Yucatan Karstic Aquifer*: University of Waterloo, M.Sc. thesis, 149 p.

Harris, N.J., 1984, *Diagenesis of upper Pleistocene strand-plain limestones, northeastern Yucatan Peninsula, Mexico*: University of New Orleans, MSc thesis, 130 p.

Headworth, H.G., 1978, Hydrogeological characteristics of artesian boreholes in the Chalk of Hampshire: *Quarterly Journal of Engineering Geology*, v. 11, p. 139-144.

Headworth, H.G., Keating, T., and Packman, M.J., 1982, Evidence for a shallow highly-permeable zone in the Chalk of Hampshire, U.K.: *Journal of Hydrology*, v. 55, p. 93-112.

Hess, J.W., and White, W.B., 1988, Storm response of the karstic carbonate aquifer of southcentral Kentucky: *Journal of Hydrology*, v. 99, p. 235-252.

Houston, J.F.T., Eastwood, J.C., and Cosgrove, T.K.P., 1986, Locating potential borehole sites in a discordant flow regime in the Chalk Aquifer at Lulworth using integrated geophysical surveys: *Quarterly Journal of Engineering Geology*, v. 19, p. 271-282.

Huntoon, P. W., 1995, Is it appropriate to apply porous media groundwater circulation models to karstic aquifers? *in* El-Kadi, A.I., ed., *Groundwater models for resources analysis and management*: Boca Raton, Lewis Publishers, p. 339-358.

Hutcheson, E., and Madden, M., 1994, Unpublished map of Nohoch Nah Chich System.

Lapcevic, P., Novakowski, K., Bickerton, G., and Voralek, J., 1997, Hydraulic characterization of the fracture framework in carbonate rock underlying CWML site, Smithville, Ontario: *Air and Waste Management Association Annual Meeting, Proceedings,* Toronto.

Lloyd, J.W., 1993, The United Kingdom, *in* Downing, R.A., Price, M., and Jones, G.P., eds., *Hydrogeology of the Chalk of North-West Europe*: Oxford, Clarendon Press, p. 220-249.

Marin, L.E., 1990, *Field investigations and numerical simulation of ground-water flow in the karstic aquifer of northwestern Yucatan, Mexico*: Northern Illinois University, Ph.D. dissertation, 183 p.

Morel, E.H., 1980, The use of a numerical model in the management of the Chalk Aquifer in the upper Thames Basin: *Quarterly Journal of Engineering Geology*, v. 13, p. 153-165.

Mortimore, R.N., 1993, Chalk water and engineering geology, *in* Downing, R.A., Price, M., and Jones, G.P., eds., *Hydrogeology of the Chalk of North West Europe*: Oxford, Clarendon Press, p. 67-92.

Palmer, A.N., 1981, *A geological guide to Mammoth Cave National Park*: Teaneck, Zephyrus Press, 210 p.

Palmer, A.N., 1989, Stratigraphic and structural control of cave development and groundwater flow in the Mammoth Cave region, *in* White, W.B., and White, E.L., eds., *Karst hydrology: Concepts from the Mammoth Cave area*, New York, Van Nostrand Reinhold, p. 293-316.

Palmer, A.N., 1991, Origin and morphology of limestone caves: *Geological Society of America Bulletin*, v. 103, p. 1-21.

Price, M., 1994, A method for assessing the extent of fissuring in double-porosity aquifers, using data from packer tests: *International Association of Hydrological Sciences Publication*, no. 222, p. 271-278.

Price, M., 1996, *Introducing groundwater*: London, Chapman and Hall, 278p.

Price, M., Morris, B., and Robertson, A., 1982, A study of intergranular and fissure permeability in Chalk and Permian aquifers, using double packer injection testing: *Journal of Hydrology*, v. 54, p. 401-423.