4-27-2009

Self-Organized Permeability in Carbonate Aquifers

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ground water

Issue Paper/

Self-Organized Permeability in Carbonate Aquifers

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Abstract

Advances over the past 40 years have resulted in a clear understanding of how dissolution processes in carbonate rocks enhance aquifer permeability. Laboratory experiments on dissolution rates of calcite and dolomite have established that there is a precipitous drop in dissolution rates as chemical equilibrium is approached. These results have been incorporated into numerical models, simulating the effects of dissolution over time and showing that it occurs along the entire length of pathways through carbonate aquifers. The pathways become enlarged and integrated over time, forming self-organized networks of channels that typically have apertures in the millimeter to centimeter range. The networks discharge at point-located springs. Recharge type is an important factor in determining channel size and distribution, resulting in a range of aquifer types, and this is well demonstrated by examples from England. Most carbonate aquifers have a large number of small channels, but in some cases large channels (i.e., enterable caves) can also develop. Rapid velocities found in ground water tracer tests, the high incidence of large-magnitude springs, and frequent microbial contamination of wells all support the model of self-organized channel development. A large majority of carbonate aquifers have such channel networks, where ground water velocities often exceed 100 m/d.

Introduction

Carbonate rocks are widespread globally, and it has been estimated that 20% or more of the world's population is partly or entirely dependent on ground water obtained from them (Ford and Williams 2007, 1). Carbonate rocks are unusual in being moderately soluble in infiltrating water, with typical dissolved solids concentrations of 100 to 300 mg/L, making them ideal potable water. Furthermore, based on sampling in wells, carbonate aquifers are commonly contaminated by microbes (Embrey and Runkle 2006). Such contamination is usually inferred to be associated with short travel times from the surface to well intakes, suggesting that ground water velocities may, on average, be higher in carbonate aquifers than in others.

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Received September 2008, accepted December 2008.

doi: 10.1111/j.1745-6584.2009.00551.x

Characterizing flow and transport in carbonates is challenging because of the often uncertain effects of dissolution processes. In many circumstances, dissolution of the bedrock results in solutionally enlarged pathways that are continuous and so provide vectors for pathogenic microbes and other contaminants to reach water supply wells. Consequently, it is important to understand how continuous, solutionally enlarged pathways do develop.

A generation ago, the processes that result in solutionally enlarged pathways were only poorly known. Then, in the 1970s, there was a major breakthrough in understanding the dissolution kinetics of limestone (Berner and Morse 1974; White 1977). This was followed by studies on aquifer evolution using empirical models (Ford 1971; Ford and Ewers 1978; Worthington 2001), physical models (Ewers 1982), and numerical models (Dreybrodt 1990; Palmer 1991; Dreybrodt et al. 2005). These advances have revolutionized our understanding of dissolution in carbonate aquifers. The modeling has shown that solutionally enlarged pathways will form dendritic, self-organized networks in the majority of carbonate aquifers.

In their classification of porosity in carbonate aquifers, Choquette and Pray (1970) defined solutional

porosity that is not fabric selective as being of three types: channel, vug, and cavern. Channels are laterally extensive openings that usually have elliptical shapes, vugs are openings with little lateral extension, and caverns are vugs or channels that are large enough for people to enter. Other terms such as *conduit* and *fissure* are also commonly used to describe solutionally enlarged pathways (Maurice et al. 2006). The term "channel" is used in this account to refer to solutionally enlarged pathways of any aperture or shape in cross section; thus it includes not only fissures and conduits but also caves.

Theoretical aspects of the dissolution of carbonate rocks are discussed first in this paper. The theory is then applied to some well-documented examples of the contrasting carbonate aquifers that are found in England. In conclusion, data from springs, tracer tests, and microbial contamination of wells are shown to provide good support for the theory.

Development of Self-Organized Permeability in Carbonates

Equilibrium and Kinetic Dissolution Processes

Meteoric water recharging carbonate aquifers typically dissolves 100 to 300 mg/L of calcite or dolomite. A key question concerns where this dissolution takes place. Calculations based on equilibrium chemistry alone would place all dissolution at the points of first contact on the bedrock surface, so that there would be no dissolution within the bedrock. However, experiments on dissolution kinetics showed that the dissolution rates slow substantially as chemical equilibrium is approached (Berner and Morse 1974: Plummer and Wigley 1976). This results in some of the dissolution taking place along fractures deep within the bedrock, as widely observed (White 1977). Modeling has shown that the positive feedback between increasing flow and increasing dissolution along a fracture results in increasing aperture along the whole flowpath. Eventually "breakthrough" is achieved, defined as the moment when dissolution has enlarged the fracture to the point where turbulent flow may occur throughout its length (Dreybrodt 1990; Palmer 1991).

Following breakthrough, there is rapid enlargement along the channel that results in a drop in head (Ewers 1982; Dreybrodt 1990; Palmer 1991). This causes a reorientation of the flow field so that flow is toward the channel and tributary channels then form (Figure 1; Ewers 1982; Ford and Williams 2007, 214–236). This process continues with the eventual formation of myriad interconnected channels. Thus, the aquifer permeability becomes dominated by self-organized channel networks. The smaller and more frequent channels, with apertures in the range 1 to 10 mm, are commonly seen in downhole videos. Some but not all carbonate aquifers also have very large channels that people can enter (i.e., caves).

Two-dimensional numerical modeling has shown how the patterns of channel networks that evolve depend on a number of factors. These include the CO_2 partial pressures of recharge water (Gabrovšek and Dreybrodt 2000), variations in the penetrability of fracturing (Hanna and Rajaram 1998; Kaufmann 2003), and whether constant head or constant recharge boundary conditions occur (Kaufmann and Braun 1999). However, the factor that has the most profound influence is the type of recharge to the aquifer, which is explained next.

Endmember Models of Aquifer Evolution

Recharge type is one of the major variables that influence how channel networks develop (Figure 2). The models shown represent the plan view of a confined aquifer 2000 m long, 500 m wide, and 1 m in thickness, with recharge at three points on the left at a constant head boundary ($z = 50$ m), discharge from all fractures on the right ($z = 0$ m), and two flanking no-flow boundaries. Initial fracture apertures have a truncated lognormal distribution with a geometric mean of 0.2 mm and a range between 0.05 and 0.5 mm. PCO_2 is 0.05 atmospheres. The ratio of the concentration of dissolved calcium (c) with respect to saturation values (c_{eq}) in recharge to the aquifer is varied between the two simulations, with the water being nearly saturated with respect to calcite when it first contacts the fractures in Figures 2A and 2B ($c/c_{eq} = 0.98$), but very undersaturated in Figures 2C and 2D ($c/c_{eq} = 0$). The former may be taken to represent percolation recharge to an aquifer because most dissolution

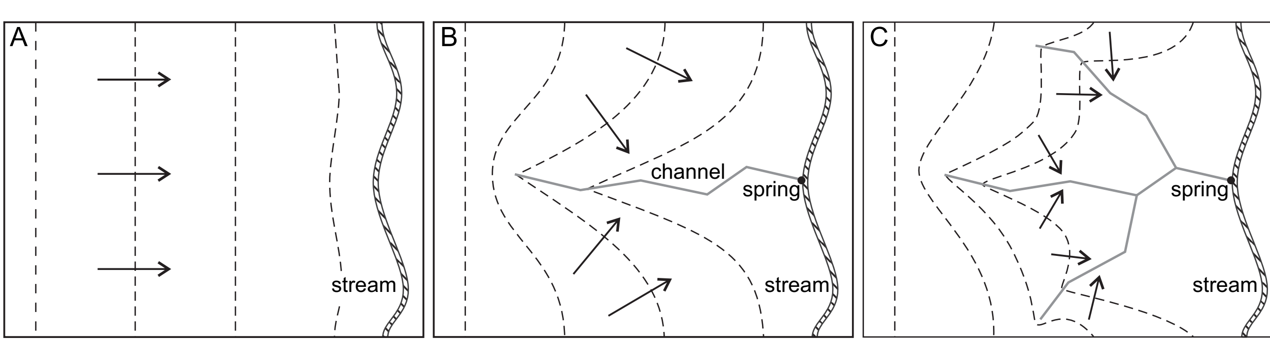

Figure 1. Development of self-organized permeability in carbonate aquifers, showing (A) the initial flow field, (B) the modified flow field after one channel has achieved breakthrough, and (C) after tributary channels have achieved breakthrough. Arrows represent flow lines (except in the channels) and dashed lines represent equipotentials.

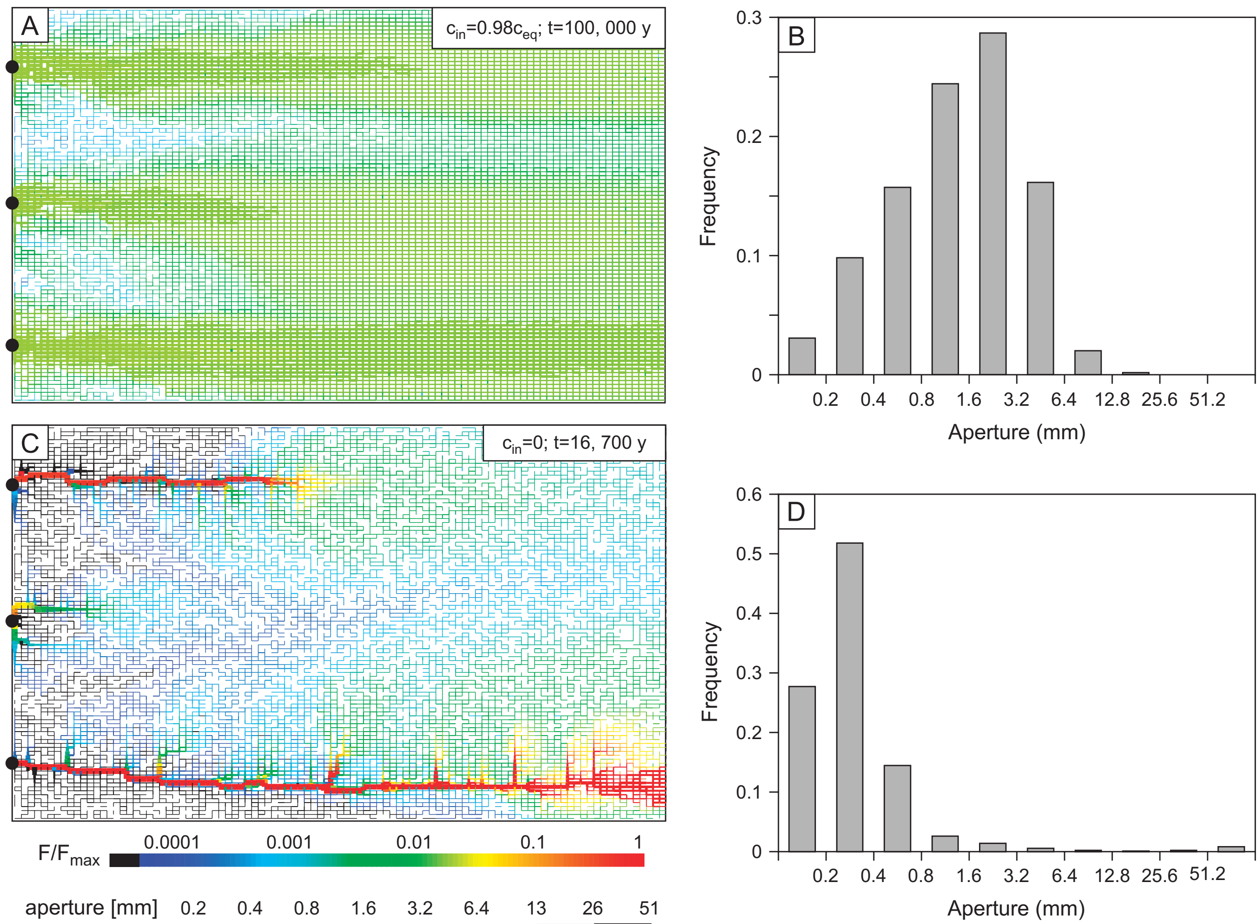

Figure 2. Numerical simulations of dissolution in fracture systems at the time of breakthrough, showing contrasts as a result of differing the saturation with respect to calcite for aquifer recharge. (A) c_{in} = 0.98 c_{eq}; (B) aperture distribution for (A); (C) c_{in} = 0; (D) aperture distribution for (C). F/F_{max} represent the dissolution rate with respect to the maximum rate of 4×10^{-12} mol/cm²/s (A and C after Dreybrodt et al. 2005).

takes place at or close to the bedrock surface, producing a weathered zone (epikarst). At the base of this weathered zone, the recharge is close to saturation (Ford and Williams 2007, 93–101). The model thus represents dissolution in the main part of the aquifer, excluding the weathered zone. In many formerly glaciated areas, a cover of as little as 1 m of glacial till containing local carbonate clasts has a similar effect (Ford 1983). The latter situation ($c/c_{eq} = 0$) represents point recharge from a sinking stream flowing off noncarbonate rocks, where the combination of high discharge and a high degree of undersaturation results in focused dissolution and the formation of caves in many instances.

Both simulations are shown at a time just before breakthrough. They have similar discharges of about 2 L/s at this time and both show the development of self-organized channel networks. Where the recharge to the aquifer is already close to saturation with respect to calcite (Figure 2a; where $c/c_{eq} = 0.98$), then many fractures are enlarged at similar rates, resulting in them having apertures in the range from 1.6 to 6.4 mm (Figure 2b). This simulation mimics the early development of solution porosity in limestone aquifers where the recharge is by percolation over wide areas and where caves are rare or absent.

Fracture enlargement is much more selective and faster where recharge is less saturated with respect to calcite (Figure 2c). Where the recharge to the aquifer has no dissolved calcium carbonate at all (i.e., $c/c_{eq} = 0$), then dissolution tends to result in preferential enlargement along one or a few main pathways, where dissolution of the rock surface proceeds at close to the maximum rate of 4×10^{-12} mol/cm²/s, and is thus shown as red in Figure 2C. This particular channel is enlarged to an aperture >51.2 mm along most of its length at a time just before breakthrough. Since discharge is proportional to the cube of fracture aperture (from the cubic law), it follows that this large aperture pathway will carry most of the discharge in the model. This simulation mimics the early development of aquifers with caves formed by recharge primarily via sinking streams.

The aperture distributions of the two simulations are approximately lognormal (Figures 2B and 2D). However, the first distribution is negatively skewed (Figure 2B), whereas the second distribution is positively skewed (Figure 2D), and their geometric means vary by a factor

of about 5. From the cubic law, flow (and hence equivalent hydraulic conductivity) is proportional to the cube of the fracture aperture, so these two simulations will vary in geometric mean hydraulic conductivity by a factor of about 5^3, or 125.

In porous media, it is generally considered that the geometric mean of hydraulic conductivity data gives a good representation of the permeability of an aquifer. That clearly cannot apply here. The discharge, hydraulic gradient, and aquifer cross section are almost identical in the two simulations, but their geometric mean hydraulic conductivities vary by a factor of about 125. The reason for the difference is that almost all the flow in Figures 2C and 2D is through the 2% of fractures that have apertures >1.6 mm. The key property that distinguishes these aquifers from porous medium aquifers thus is that the largest aperture fractures are integrated to form continuous large aperture pathways that provide much of the permeability of the aquifer.

Numerical modeling has shown that contrasting channel networks are formed by a number of other processes in addition to recharge type (Lauritzen et al. 1992; Kaufmann and Braun 1999; Dreybrodt et al. 2005). Recharge with high values of c/c_{eq}, densely fractured rock, and high hydraulic gradients all promote the development of networks with many channel of similar size, as shown by the example in Figure 2A. Conversely, low values of c/c_{eq}, sparsely fractured rock, and low hydraulic gradients promote the development of networks where a few large channels conduct most of the flow, as shown in Figure 2C.

The description given above is a rather simplified account of dissolution in limestone aquifers. The dissolution rate of dolostone is slower, but it is well established that dolostone aquifers have permeabilities that are similarly high. Consequently, it may be that the slower approach to equilibrium of dolostone results in more channels but of smaller size (geometries like those shown in Figure 2A) than would be found in limestone.

Examples of Contrasting Self-Organized Aquifers

Major Limestone Aquifers in England

The range of effects that dissolution has on the permeability of limestone aquifers is well displayed by the major limestone aquifers in England. Four of the most important English aquifers are in limestone, one of them (the Chalk) accounting for 55% of total ground water production in England and Wales. A compilation of the hydraulic properties of these four aquifers included data on porosity and hydraulic conductivity values from core tests and transmissivity and storage values from pumping data (Allen et al. 1997); Figure 3 is derived from it.

Atkinson and Smart (1981) recognized that there is a spectrum of aquifer types in British limestones. One endmember is represented by Paleozoic limestones such as the Carboniferous Limestone, which has many sinking streams, large numbers of closed depressions (sinkholes),

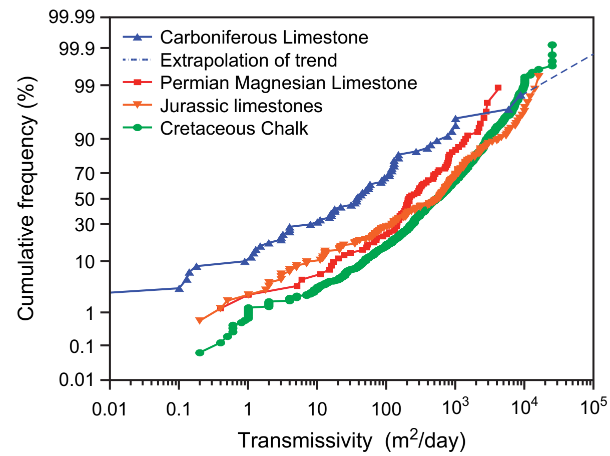

Figure 3. Distribution of transmissivity data from pumping tests in the four most important carbonate aquifers in England (based on data from Allen et al. 1997).

and much discharge through channels in which turbulent flow occurs. The other endmember is the Chalk, which has fewer channels with turbulent flow, more flow in smaller channels, and fewer sinking streams and closed depressions. Thus, the British limestone aquifers offer useful data for comparing the characteristics of contrasting limestone aquifers. The two end members, the Chalk and the Carboniferous Limestone, are discussed in detail in the following two sections.

The Chalk Aquifer

At a large scale, the Chalk has traditionally been considered to behave like a porous medium, especially for modeling regional water resources (Allen et al. 1997). However, recent and detailed site investigations, using borehole imaging and flowmeters, have found that most of the flow is through channels, with apertures that are commonly in the range from millimeters up to a few centimeters, but that larger apertures are rare (Price et al. 1982; Waters and Banks 1997). Rapid movement of tracers to pumping wells is common, indicating that there are well-developed channel networks in the Chalk (MacDonald et al. 1998, Schürch and Buckley 2002). Furthermore, modern 1:100,000 scale hydrogeological maps by the British Geological Survey record the presence of several hundred springs, the largest group of them having a discharge of more than 1 m^3/s and draining a ground water catchment of 64 km^2 (Day 2001). The simulation in Figure 2A provides a plausible analogue for channel development in the Chalk for two reasons: most of the recharge is by percolation and, second, because the resultant negatively skewed lognormal aperture distribution (Figure 2B) is similar to the negatively skewed distribution found in Chalk transmissivity values (Figure 3).

The Carboniferous Limestone Aquifer

The Carboniferous Limestone is considered to be a well-developed karst aquifer. There has been considerable research on hydrogeological aspects of its many

explored caves, and many tracer tests have been carried out between sinking streams and springs (Atkinson 1977; Atkinson and Smart 1981; Allen et al. 1997; Waltham et al. 1997). In these respects, it contrasts strongly with the Chalk.

As an example, Figure 4 shows 22 km of explored cave stream passages in the Carboniferous Limestone of the Yorkshire Dales. They form a dendritic pattern that links a line of discrete recharge points with one perennial spring and three adjacent springs that are slightly higher in elevation and act as intermittent overflows. The channel pattern is analogous to that of a surface river, and is similar in that both are self-organized to facilitate the efficient downgradient flow of water. The channel network shown in Figure 4 serves to integrate the drainage from 15 km^2 of shale caprock and more than 5 km^2 of the limestone aquifer and to discharge it at Leck Beck Head Springs (Waltham et al. 1997). This tributary ground water drainage pattern is typical of the patterns found in caves.

The spatial and volumetric density of the conduits as a fraction of the aquifer provides a convenient morphometric index of channel distribution. The channels shown in Figure 4 average 3 m in width, varying from <1 m in headwaters areas close to surface stream sinks into the cave entrances to >10 m close to the springs. Consequently, the channels cover an area of some 66,000 m^2, which is 1.2% of the limestone outcrop shown in Figure 4. Several kilometers of cave passage are below the water table and were explored and mapped by scuba divers. Most of these passages are within 40 m of the water table, but one passage reaches a maximum depth of 64 m. Assuming an average saturated aquifer thickness of 50 m and an average water-filled height of 2 m for these large channels, they then occupy just 0.048% of the volume of the aquifer. This low value is of the same magnitude as the volume of 0.02% that was calculated by Atkinson (1977) for channels in the saturated zone in the Carboniferous Limestone aquifer at Cheddar in southern England. These values show that the main solutional channels occupy just a very small fraction of the aquifer volume, indicating a very low "channel porosity."

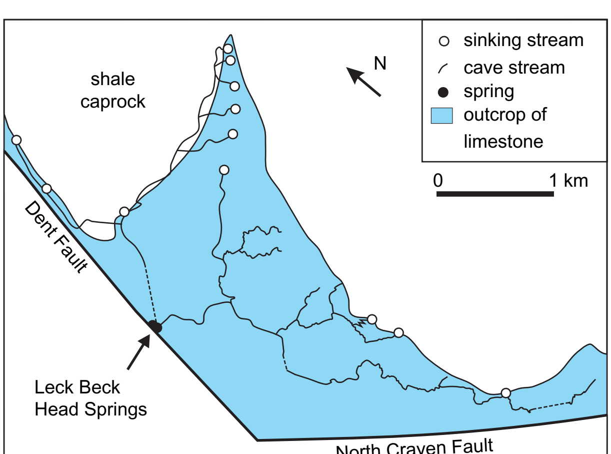

Figure 4. Major channels in the Carboniferous Limestone aquifer that drains to Leck Beck Head Springs (compiled principally from Waltham et al. 1997).

There are many hundreds of caves worldwide where similar tributary networks have now been mapped and most caves consist of such patterns (Courbon et al. 1989; Palmer 1991). Similarly, the proportion of 1.2% of the total area that the main channels occupy and the channel porosity of 0.048% are both similar to geometric means of these values (1.6% and 0.08%, respectively) in a sample of 10 other major caves ranging in mapped passage length from 6 to 550 km (Worthington 1999). These statistics demonstrate that caves only occupy a small fraction of the volume of an aquifer.

A randomly drilled borehole would have a probability of only 0.012 of intercepting one of the high-permeability channels that drain most of the flow in the Leck Beck aquifer. This very low probability is reflected in the transmissivity from 59 pumping tests in the Carboniferous Limestone, the highest value of which is 8800 m^2/d. The dashed line in Figure 3 shows an extrapolation of the log-normal distribution of the Carboniferous Limestone permeabilities: this indicates that if several hundred pumping tests could be made at different boreholes in the aquifer then it is probable that a few of these would have the very high transmissivities (10^4 to 10^5 m^2/d) that would reflect interception of the flow in the large channels. The permeability distribution in the Carboniferous Limestone (Figures 3 and 4) is consistent with the modeling in Figure 2C. In both cases, dissolution by sinking stream recharge with low dissolved calcium concentrations results in the creation of a small number of large channels.

Comparison of Aquifer Properties in the British and Some North American Aquifers

Summary transmissivity statistics of the four major limestone aquifers show that the greatest contrasts are between the Chalk and the Carboniferous Limestone, with the latter having the lowest geometric mean transmissivity and the highest standard deviation of log transmissivity (Table 1). The channel networks in the Chalk and the Carboniferous Limestone correspond approximately with the two scenarios modeled in Figure 2, where the only difference in initial conditions is the value of c/c_{eq}. However, it is likely that other factors such as the degree of fracturing and lithological differences also play important roles in forming the contrasting channel patterns of the two aquifers.

Recent numerical modeling has largely validated the concept of Atkinson and Smart (1981) that there is a spectrum of aquifer types in the British limestones from those with many small channels (Figure 2A) to those with fewer but larger channels (Figure 2C). Based on field transmissivity data, the Chalk represents the former, having a higher geometric mean transmissivity and a lower standard deviation of log transmissivity than the latter, which is represented by the Carboniferous Limestone. Applying these two parameters to the Magnesian Limestone and

Aquifer	Number of Pumping Tests	Geometric Mean Transmissivity (m^2/d)	Standard Deviation of Log Transmissivity
Cretaceous Chalk	1257	440	0.76
Jurassic limestone	162	286	1.02
Permian limestone	81	209	0.74
Carboniferous Limestone	59	22	1.31

Table 1
Transmissivity from Pumping Tests in the Four Major Limestones in England (Compiled from Data in Allen et al. 1997)

Jurassic limestone indicates that these aquifers are intermediate in characteristics between the Chalk and Carboniferous Limestone (Figure 3; Table 1). Atkinson and Smart (1981) similarly classified them as intermediate in behavior.

A comparison of carbonate aquifers in terms of average values of the porosity and hydraulic conductivity of the matrix and of the fractures and channels provides further insight into their characteristics (Foster 1993). Figure 5 provides approximate values for the four major English limestone aquifers considered earlier and for three important aquifers in North America, the Silurian Lockport dolostone aquifer in Ontario, the Mississippian limestone aquifer at Mammoth Cave in Kentucky (that contains the most extensive mapped cave in the world), and the Cenozoic limestone aquifer in the Yucatan Peninsula, Mexico, where very extensive underwater caves (tens of kilometers in length) in a sea water/fresh water mixing zone setting are being explored currently. The greatest uncertainty in Figure 5 is associated with the fracture and channel porosities, which are variously derived from data from packer tests, tracer tests, recession curves for spring discharge, and flowmeter data in boreholes (Worthington et al. 2000). In carbonates, the value for effective porosity for transport, which is also known as the dynamic or kinematic porosity (Allen et al. 1997, 16), corresponds with the porosity of channels and open fractures. Limited data suggest that it is likely to be typically in the range 0.0001 to 0.001, and it is usually much less than specific yield values from pumping tests because specific yield may include some contribution from the matrix as well as from fractures and channels (Allen et al. 1997, 28; Rayne et al. 2001). Where the fracture/channel porosity is poorly known, it is assigned a value of 0.0003 in Figure 5; this is somewhat less than typical values of specific yield but similar to values derived from tracer tests. Despite the uncertainties in these porosity values, it is emphasized that they are at least an order of magnitude lower than the matrix porosities in all cases.

There are substantial differences in absolute values between these different carbonate aquifers, but they do all share a number of common characteristics. The matrix of the rock has relatively high porosity and low hydraulic conductivity, whereas the fractures and channels have low porosity and high hydraulic conductivity. Thus, the matrix values and fracture/channel values plot in two distinct fields in Figure 5.

These concepts may be illustrated by the data from the Silurian dolostone aquifer at a polychlorinated biphenyl (PCB) spill site at Smithville, Ontario (Novakowski et al. 1999; Worthington 1999; Worthington and Ford 1999). The aquifer has been interpreted both as a fractured-rock aquifer (Zanini et al. 2000) and as a karst aquifer (Worthington 2002). The hydraulic conductivities of the matrix and of the channels/fractures at the site were derived from geometric mean values of 161 packer tests with a 0.5-m interval (in sections of boreholes that had no visible fractures), and from three pumping tests, respectively. The porosities of the matrix and of the channels/fractures were derived from mean values of 389 measurements from core samples and of 672 packer tests with a 0.5-m interval, respectively (Table 2).

These data can be used for either single- or double-porosity interpretations of the aquifer. The former situation would apply if the large aperture fractures and channels were discontinuous. Total values of porosity and hydraulic

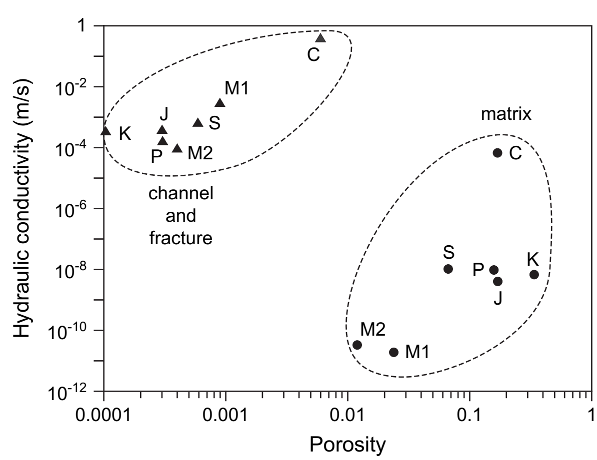

Figure 5. Porosity and hydraulic conductivity values for the matrix and for fractures and channels for Silurian dolostone (Ontario, S), Mississippian limestone (Kentucky, M1; England, M2), Permian limestone (England, P), Jurassic limestone (England, J), Cretaceous limestone (England, K) and Cenozoic limestone (Mexico, C), (from data in Worthington et al. 2000, Price et al. 1993, and Allen et al. 1997).

Table 2
Matrix and Fracture/Channel Properties in the Dolostone Aquifer at Smithville, Ontario

Parameter	Single Porosity Total	Double Porosity Matrix	Double Porosity Channel and Fracture
Hydraulic conductivity (K, m/s)	6E−4	1E−8	6E−4
Porosity (n)	0.0653	0.065	0.0003
Hydraulic gradient (i)	0.002	0.002	0.002
Area (A, m^2)	1	1	1
Discharge (m^3/s, $Q = KiA$)	1.2E−6	2E−11	1.2E−6
Velocity (m/d, $v = Ki/n$)	1.6	2.7E-5	350
Fraction of n (%)	100	99.54	0.46
Fraction of Q (%)	100	0.0017	99.9983

conductivity could then be used to calculate an average ground water velocity of 1.6 m/d, using Darcy's law (Table 2). However, if the fractures and channels form an interconnected network with continuous large aperture pathways through the aquifer, then these pathways should be considered separately from flow through the matrix. In this case, ground water velocities through the fractures and channels would be 350 m/d and velocities in the matrix would be 13 million times slower. Darcy's law can also be used to show that the flux of water through the matrix is very low and that more than 99.99% of flow is through the fracture and channel network (Table 2).

These two hypotheses were tested as Smithville by using tracer tests. Sixteen tracer tests over distances of 6 to 150 m to wells of the pump-and-treat system gave velocities of 15 to 150 m/d, while six traces between a sinking stream and a spring to the south of the site gave a geometric mean velocity of 2600 m/d (Worthington and Ford 1997; Worthington 2002). The rapid velocities measured in these tracer tests indicate that the fracture and channel pathways are interconnected and thus support the double-porosity model. The data from Smithville thus show that there is a proportionally large volume of slow-moving water in the matrix and a proportionally small volume of fast-moving water in the fractures and channels. The same conclusion applies to all the aquifers in Figure 5.

It is a challenge to incorporate these double-porosity concepts into numerical models. Occasionally, there is enough information available on the locations and flow characteristics of major channels (Figure 4) so that they can be specifically depicted in numerical models (e.g., Quinn and Tomasko 2000; Lindgren et al. 2004). In such cases, it would be possible to use pipe flow equations to depict both laminar and turbulent flow using MODFLOW (Shoemaker et al. 2007). However, more commonly, there is insufficient information to use this approach and single porosity models are used. These can often provide reasonable simulations of flow (e.g., heads at wells and flow at springs) but not of transport (Scanlon et al. 2003). If transport is the main concern, such as for identifying time of travel zones to municipal wells, then a low value can be used for effective porosity (e.g., 0.0001 to 0.001) to enable model velocities to be calibrated to velocities determined by tracer tests. Such model velocities would characterize the rapid flow through fractures and channels. It would be inadequate, however, for understanding the distributions of contaminants at locations such as the PCB spill site at Smithville, where transport in the channels, fractures, and matrix as well as exchange between them all need to be accounted for.

Evidence for Channel Networks from Springs, Tracer Tests, and Microbial Contamination of Wells

It has been stressed that the high conductivity of channels in carbonate aquifers allows highly concentrated flow with low hydraulic gradients, resulting in the formation of channel networks (Figure 1). As a consequence, carbonate aquifers might be expected to exhibit (1) rapid ground water flow (indicated by tracer velocities), (2) a high incidence of springs and especially of large springs, and (3) a high frequency of contamination by biota, including enteric microorganisms. These three hypotheses will be tested below.

Tracer testing has commonly been used in carbonate aquifers to determine travel time and flow direction. Quinlan (1986) estimated that more than 90% of all recorded ground water traces have taken place in carbonate aquifers. Most of these are natural-gradient traces from sinking streams to springs, which are relatively simple to execute because flowpaths converge on springs (Figure 4). The results from 3015 such traces, compiled from the published results from 34 countries, are shown in Figure 6. They display an approximately lognormal velocity distribution, with median and geometric mean velocities of 1940 and 1740 m/d, respectively. These rapid velocities imply that sink to spring tracer tests of tens of kilometers are feasible and, in fact, there are 595 traces over distances of at least 10 km included in the data of Figure 6. The rapid ground water velocities demonstrated in this large set of tracer tests clearly indicate that large aperture channels connecting sinking streams to springs are common in carbonate aquifers. These velocities reflect the combination of low kinematic porosity and high hydraulic

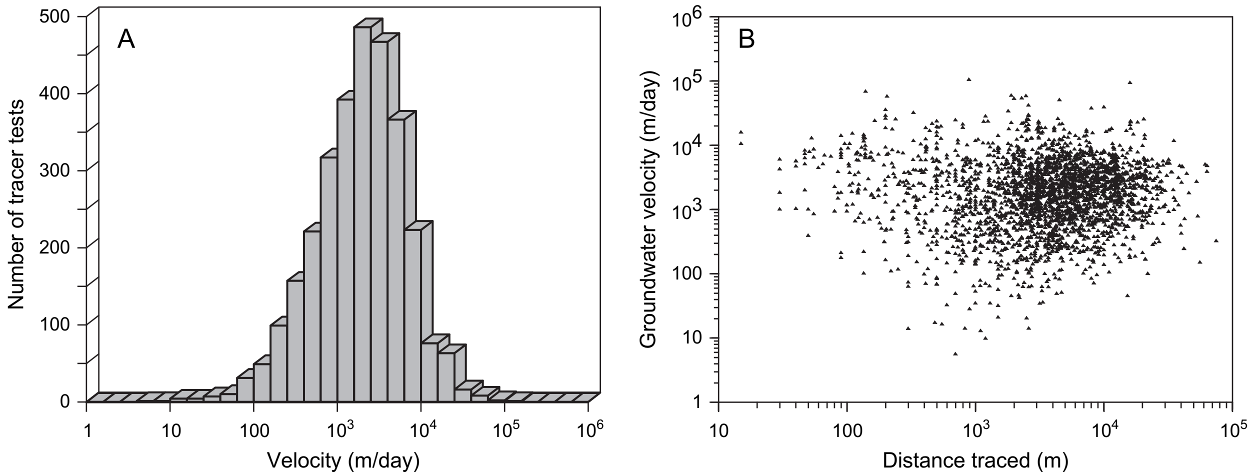

Figure 6. Ground water velocities for 3015 tracer tests along channels in carbonates: frequency distribution (A) and velocity as a function of distance traced (B).

conductivity (Figure 5). Rapid velocities over straight-line distances of many kilometers underscore the large scale of self-organization occurring in these aquifers.

Tracer tests between wells are far less common than those between sinking streams and springs. Where such interwell testing has been carried out in carbonate aquifers, however, results often show ground water velocities of hundreds of meters per day (e.g., MacDonald et al. 1998). These rapid velocities are also suggestive of substantial flow through channel networks with apertures of at least several millimeters.

Sanz Pérez (1996) gave details of an inventory of 8000 springs in Spain where carbonate rocks outcropped over 19% of the area surveyed. The springs in carbonates represented 49% of the total number of springs and accounted for 71% of aggregate spring discharge. The higher preponderance of springs (especially of large springs) in the carbonate aquifers contrasts strongly with the other aquifer types, where there are very few large springs (Figure 7). This contrasting behavior supports the hypothesis that channel networks are widespread in carbonate aquifers.

Embrey and Runkle (2006) gave summary results for an extensive survey of microbial contamination of wells in the United States. Data were compiled from 1174 wells in 16 major aquifers, 6 of which were partly or wholly composed of carbonate rocks. Carbonates had the largest proportion of wells that were positive for total coliform bacteria, for *Escherichia coli*, and for coliphage. Furthermore, public water supplies from carbonate rocks have been found to be especially susceptible to water borne disease (EPA 2006, 65596). Both these findings provide further support for the hypothesis that channel networks are common in carbonate aquifers.

Discussion

Characterizing Carbonate Aquifers

It has been shown that channel networks are predicted to develop in carbonate aquifers, and that there is substantial empirical evidence to show that such networks are common. This generalization is true for both limestones and for dolostones, and it extends from Precambian carbonates subject to amphibolite-grade metamorphism (Lauritzen 2001) to weakly consolidated Pleistocene limestones of the last glacial cycle (Beddows et al. 2007).

Parameterizing these carbonate aquifers using well data alone presents a major challenge because the high-permeability channel networks occupy much less than 1% of aquifer volume, even in aquifers where there are extensive caves (Figure 4; Worthington 1999). On average, only 12 wells per 1000 would probably intercept the large channels shown in Figure 4. There is a greater likelihood that smaller aperture channels will be intercepted, however. Measurements of discharge and chemical parameters at springs can provide complementary information on the

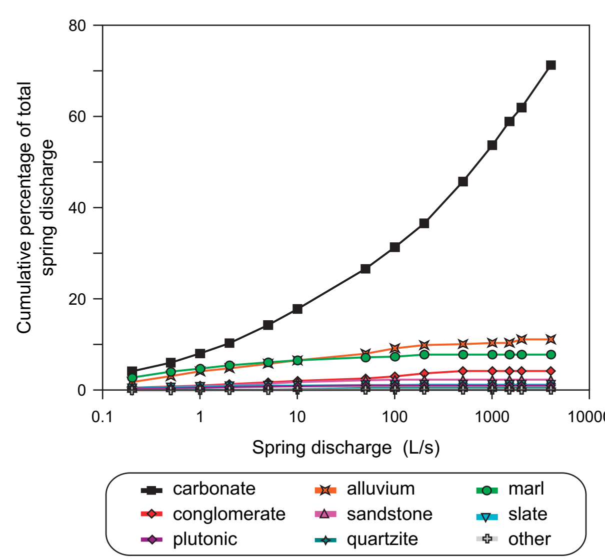

Figure 7. Spring discharge for differing aquifer types in Spain, showing the high incidence of large springs in carbonates (compiled from data in Sanz Pérez 1996).

degree of organization of the aquifer and its response to stresses such as precipitation (Ford and Williams 2007, 177–181, 188–191).

Definition of "Karst Aquifer"

Early limestone aquifer evolution models showed that the positive feedback between dissolution rates and flow rates should result in the formation of caves, that is, channels that are large enough for people to enter (Ford and Williams 2007, 214–221). However, it is only recently that modeling with large grids and a wide range of boundary conditions has shown that solution processes may lead to more numerous and extensive channels of the more modest apertures of Figure 2A. It has sometimes been thought that karst aquifers are those with explorable caves, but this recent modeling challenges that assertion as both types of aquifer shown in Figure 2 have channel networks with rapid flow.

The term *karst* usually refers to distinctive terrains attributable to the high solubility of the bedrock (e.g., Ford and Williams 2007, 1) but there is no widely accepted definition of what constitutes a karst (or karstified) aquifer. Definitions have ranged from the narrow to the broad. Atkinson and Smart (1981) offered a narrow definition:

> 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 (Atkinson and Smart 1981, 182–183).

This contrasts with the broad definition proposed by Huntoon (1995)

> Karst is a geologic environment 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 the downgradient direction wherein the permeability structure evolved as a consequence of dissolution by the fluid (Huntoon 1995, 343).

Turbulent flow is the defining criterion in the narrow definition. It can indeed be an important factor in transient modeling as head varies with the square of the discharge when flow is turbulent but head is proportional to discharge when flow is laminar. However, recognizing turbulent flow is less important in steady-state simulations or in quantifying contaminant transport. This latter point can be illustrated by the differing interpretations of a tracer test carried out to major springs in the Chalk that are used for the municipal water supply of the city of Portsmouth, England.

A trace from a closed depression to a spring traveled at 2200 m/d over a straight-line distance of 5.75 km (Atkinson and Smith 1974); these authors calculated that if all the flow from the tracer injection and flushing were along one circular pipe, then this would have a diameter of 74 cm. However, Price (1987) calculated that if, instead, the flow had been through a single smooth parallel-plate fracture, then its aperture could have been as small as 4.5 mm. The Reynolds numbers (Re) for these two cases are 14,000 and 87, respectively. Both Atkinson and Smith (1974) and Price (1987) suggested that their respective estimates were just end members and that flow was likely to be through a number of openings, but it is not possible to tell from the tracer test data whether flow in this case was in the turbulent regime ($Re > 2200$) or laminar regime ($Re < 2200$). This example illustrates the difficulty of assessing whether flow is turbulent, even when rapid ground water velocities have been demonstrated. This makes usage of the narrow definition problematic.

The advantage of the broad definition is that the challenging task of assessing channel aperture or Re is not necessary. Rather, if an aquifer is karstic, then there will be rapid velocities through the interconnected channels, an assumption that may be tested using tracer tests such as natural-gradient tests to springs or forced-gradient tests to pumping wells. In this case, we can define a karst aquifer as an aquifer with self-organized, high-permeability channel networks formed by positive feedback between dissolution and flow. There is no consensus on terminology to differentiate karst aquifers that have many small channels like the English Chalk from those with a wide range in channel size, like the English Carboniferous Limestone. Perhaps it would be appropriate to call the former "microkarstic" aquifers and the latter "macrokarstic" aquifers.

If the broad definition of a karst aquifer defined above is used, then most aquifers in carbonate rocks are karstic. However, achieving consensus on a definition for the term "karstic aquifer" is not important. Rather, what *is* important is to recognize that flow through carbonate aquifers inevitably leads to the formation of self-organized channel networks and that there is rapid ground water flow in these channels, very often exceeding 100 m/d.

Conclusions

There have been substantial advances in the past 40 years in understanding permeability development and distribution in carbonate aquifers. Laboratory experiments showed that there is a sharp decrease in the dissolution rate of limestone as equilibrium is approached. These dissolution rates were then incorporated into numerical models that simulate the evolution of carbonate aquifers from a fractured rock to one in which channel networks provide most of the permeability. A wide range of data supports such evolution patterns, including the high incidence of large-magnitude springs, frequent microbial contamination of wells, and the high velocities commonly measured from tracer tests.

The self-organized channel networks, produced by dissolution processes, result in the high permeabilities that characterize carbonate aquifers and make them highly productive. In our opinion, these chemical processes differentiate carbonate aquifers from fractured-rock aquifers, in which the permeability is primarily produced by physical processes such as tectonic movements.

Acknowledgments

We thank Franci Gabrovšek for providing us with high-quality images of numerical modeling results, Andrew McKenzie for the data set of transmissivity values of English limestones, and Chris Smart, Calvin Alexander, and an anonymous reviewer for providing helpful comments on the paper.

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