

# Chapter 11

## Management of Carbonate Aquifers

Stephen R.H. Worthington

**Abstract** Carbonate aquifers are common globally and are widely utilized due to their high permeability. Advances in recent decades in understanding dissolution kinetics have facilitated the numerical modeling of dissolutional enhancement of permeability. This has shown how the dissolution results in an interconnected network of channels that not only results in high permeability but also in rapid groundwater velocities. The high permeability often results in a lack of surface water and thick unsaturated zones, so utilization of groundwater is often from low-elevation springs, especially in mountainous areas. Groundwater divides may not coincide with surface-water divides, sometimes resulting in jurisdictional issues over exploitation of the groundwater. Contaminant transport in carbonates is more complicated than in porous medium aquifers. Transport through the channels may be several orders of magnitude faster than transport through the matrix of the rock. This results in complicated contaminant plumes and makes carbonate aquifers more susceptible to bacterial contamination than other aquifer types.

### 11.1 Introduction

The approach to managing aquifers in carbonate rocks is (or should be) somewhat different from that used with porous medium aquifers such as sand due to the different permeability structure. In carbonate aquifers, dissolution produces a network of solutionally-enlarged pathways that enhance the permeability. These aquifers are sometimes called karst aquifers.

There is no widely agreed definition of what constitutes a karst aquifer. A narrow definition includes only the small fraction of carbonate (limestone and dolostone)

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S.R.H. Worthington (✉)

Worthington Groundwater, 55 Mayfair Avenue, Dundas, ON L9H 3K9, Canada

e-mail: sw@worthingtongroundwater.com

aquifers that have known caves, demonstrable turbulent flow in conduits with diameters  $>1$  cm, or a surficial karst landscape. A broad definition focuses on the permeability structure of the aquifer, defining a karst aquifer as having “self-organized, high-permeability channel networks formed by positive feedback between dissolution and flow” (Worthington and Ford 2009, p. 334). However, it is less important to classify carbonate aquifers than to recognize the dissolution-enhanced permeability structure of these aquifers which results in high permeability and rapid groundwater velocities.

Infiltrating precipitation will usually dissolve 100–400 mg/L of calcite or dolomite in a carbonate aquifer. The combination of the high permeability that results from this dissolution combined with the widespread distribution of carbonate rocks result in these aquifers being important globally. It has been suggested that at least 20% of the world’s population depends largely or entirely on groundwater obtained from carbonate aquifers (Ford and Williams 2007).

This chapter will describe how dissolution enhances permeability in carbonates and the permeability and porosity characteristics of these aquifers, and then will address the management issues specific to these aquifers.

## 11.2 Dissolution Processes in Carbonates

There are two aspects to the dissolution of carbonate rocks. These are equilibrium solubility and kinetic solubility effects.

### 11.2.1 *Equilibrium Solubility Effects*

Equilibrium solubility refers to the concentration at which a mineral species is at equilibrium when dissolved in a solute, which in a carbonate aquifer is the groundwater. At lower concentrations, the solution is under-saturated with respect to that mineral and so dissolution will occur. At higher concentrations, the solution is supersaturated with respect to that mineral and so precipitation of the mineral will occur.

The solubility of calcite in pure water at 25°C is 14 mg/L. However, its solubility increases substantially with increased carbon dioxide partial pressure. Concentrations of  $\text{CO}_2$  in soil air are typically one to two orders of magnitude greater than in the atmosphere. This results in groundwater in limestone aquifers typically having dissolved  $\text{CaCO}_3$  concentrations of 100–400 mg/L. Variations in carbon dioxide partial pressure are usually the most important factor in determining dissolved  $\text{CaCO}_3$  concentrations, but there are several other factors. Calcite solubility decreases as a function of temperature and the common ion effect. It increases as a function of pressure, the ionic strength effect, open rather than closed system conditions and the presence of organic complexes (Ford and Williams 2007, pp. 45–62).

While the dissolution potential of  $\text{CaCO}_3$  is greater for cold than warm environments, tropical regions typically have more productive soils which produce more dissolved  $\text{CO}_2$  and also have higher precipitation amounts, leading to greater amounts of  $\text{CaCO}_3$  being dissolved.

### 11.2.2 Kinetic Solubility Effects

Kinetic solubility effects refer to the rate at which reactions occur. It has commonly been assumed that reaction rates of carbonate and evaporate mineral dissolution are rapid with respect to groundwater flow. The implication for dissolution is that the water would come to equilibrium on first coming into contact with a soluble mineral near the ground surface or shallow in the bedrock, and that no dissolution would subsequently occur as the water flows through the deeper bedrock, assuming that the equilibrium solubility does not change. Limited experimental data supported this conclusion for limestone (Weyl 1958).

This conclusion posed a problem as it appeared that karst aquifers could not develop. If the groundwater solution comes to full equilibrium on first contact with the carbonate rock, then there is no further dissolution potential further along groundwater flow paths. However, karst aquifers clearly do exist and sometimes have deep flow paths extending to great distances from the infiltration site where water first comes into contact with the carbonate bedrock. It was therefore clear that there was an incomplete understanding of dissolution processes.

Further laboratory experiments on the dissolution of calcite provided a resolution to this conundrum. It was found that dissolution rates diminish precipitously as chemical equilibrium is approached for calcite and dolomite and so only slowly reach full equilibrium (Berner and Morse 1974; Morse and Arvidson 2002). Consequently, most groundwater is slightly undersaturated with respects to calcite and so dissolution is able to proceed throughout most carbonate aquifers even deep within the aquifer and distant from the site of infiltration. Dissolution rates can be expressed as

$$F = k(1 - c / c_{eq})^n \quad (11.1)$$

where  $F$  is the dissolution rate,  $k$  is the reaction coefficient,  $c$  is the solute concentration,  $c_{eq}$  is the equilibrium solute concentration, and  $n$  is the reaction order (Dreybrodt 1996). The reaction order far from equilibrium is one, but it increases to between 4 and 11 where  $c/c_{eq}$  exceeds 0.6–0.8 (Eisenlohr et al. 1999).

These results turned conventional wisdom on its head. Rather than karst aquifers being rather rare, it now became clear that the majority of, if not all, carbonate aquifers should be karstic, at least in their upper parts where most flow is likely to be concentrated. This theoretical evidence provided an explanation for the high permeability that is common in carbonate aquifers. These findings have been supported by the results from the many numerical simulations of the development of karst aquifers, which are described in the following section.

### ***11.2.3 Reactive Transport Numerical Models of Karst Aquifer Development***

Reactive transport numerical modeling of karst aquifers commenced with one-dimensional models (Dreybrodt 1990; Palmer 1991). These early models showed how karstification leads to increases in permeability of several orders of magnitude over periods of  $10^3$ – $10^6$  years. Early two-dimensional models with small grids (i.e., a small number of large dimension cells) showed how preferential flow would result in most of the flow becoming focused on one main channel. Many of the early papers had simulated recharge at sinking streams, where the water is often significantly under-saturated with respect to calcite. However, Dreybrodt (1996) showed that karstification occurs even where there is just percolation recharge to a carbonate aquifer. In the late 1990s, larger grids (with a large number of smaller dimension cells) were employed that gave more insights to competition between different pathways.

Dreybrodt et al. (2005) systematically examined many aspects of dissolution of limestone, using percolation networks and networks with log-normal aperture distributions and large grids (10,000 nodes). Liedl et al. (2003) developed a numerical model based on MODFLOW, adding subroutines to simulate dissolution and to simulate turbulent flow. The latter subroutine was subsequently added to the US Geological Survey release of MODFLOW 2005 (Shoemaker et al. 2008; Reimann and Hill 2009).

Evidence from numerical modeling, preferential flow directly observed in wells, and the distribution of caves supports the concept that carbonate aquifers vary between macrokarstic and microkarstic end members (Worthington and Ford 2009). Macrokarstic development commonly occurs where there are sinking streams; recharge is substantially undersaturated with respect to calcite, which results in a small number of large channels and in some instances caves. Microkarstic development is common where there is percolation recharge; most dissolution occurs in the upper few meters of bedrock, resulting in a high-permeability weathered (or epikarst) zone. Infiltrating water below this zone is close to saturation and follows many pathways, resulting in a dense network of many small channels. Other factors promoting macrokarst include sparsely fractured rock and low initial hydraulic gradients. Conversely, densely fractured rock and high hydraulic gradients both promote the development of microkarstic networks.

## **11.3 The Permeability Structure of Carbonate Aquifers**

### ***11.3.1 Local-Scale Characteristics***

Measurements in boreholes are important for investigating the local variability of permeability in carbonate aquifers. Pumping tests are commonly used to determine aquifer properties and give average values over the whole depth of the borehole for transmissivity, hydraulic conductivity, and storage. However, flow into boreholes in

carbonates is rarely represented by seepage from the whole saturated thickness. Instead, preferential flow from a limited number of horizons is more common (Morin et al. 1988; Schürch and Buckley 2002). Packer testing and measuring flow while pumping from the borehole at a low rate are the most useful methods of determining preferential flow into boreholes. Optical or acoustic televiewer, video, temperature, and conductivity are also helpful for identifying and characterizing preferential flow, and gamma ray logs facilitate comparisons between boreholes to be made (Fig. 11.1).

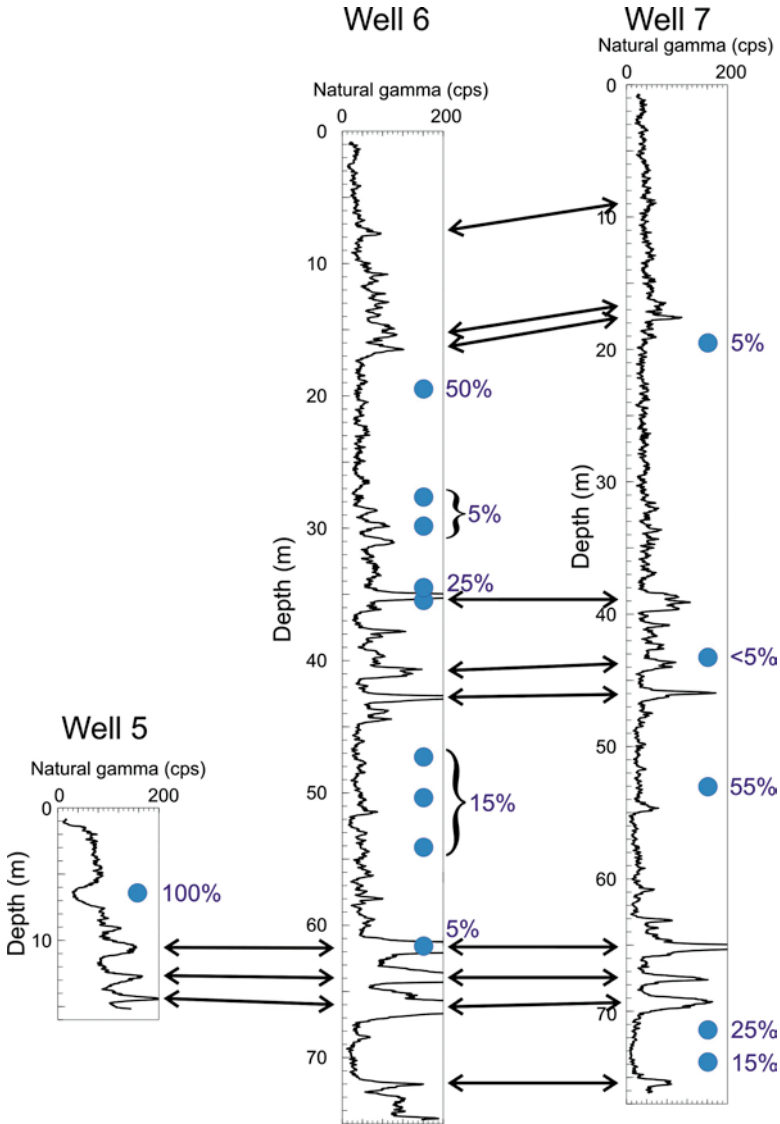
Ideally, porosity and hydraulic conductivity measurements are made separately for the matrix, fractures, and solutionally-enlarged channels. However, in many cases it is challenging to measure hydraulic conductivity and porosity for these three components. A somewhat simpler approach is to group fracture and channel properties together. Plotting porosity and hydraulic conductivity together shows that matrix and fracture/conduit values plot in two distinct fields (Fig. 11.2). In all the examples shown, the matrix of the rock has high porosity and low hydraulic conductivity. Furthermore, the fractures and channels account for a smaller fraction of total porosity but a much higher fraction of total hydraulic conductivity than the matrix. Consequently, most of the storage is in the matrix but most of the flow is in the channels.

### ***11.3.2 Aquifer-Scale Characteristics***

The development by dissolution of interconnected channel networks in carbonate aquifers makes them distinctly different from porous medium aquifers. The most common ways to investigate these differences are tracer tests and spring studies. Tracer tests have been most commonly used where there are sinking streams, and many thousands of such tests have been carried out over distances of up to tens of kilometers (Fig. 11.3). These tests have been used to determine groundwater velocities and the existence of flow paths between the injection and the observation sites. The geometric mean velocity for the 3,015 tracer tests shown in Fig. 11.3 is 1,740 m/day (Worthington and Ford 2009). These tests are all between sinking streams and springs and represent flow along large channels.

Springs in carbonates represent the outlets of channel networks, and heads in large channels are lower (except at high flow) than in the surrounding bedrock. Consequently, mapping of heads in boreholes can help identify the location of large channels and thus the major flow paths in an aquifer. Figure 11.4 shows a number of contrasting characteristics at a scale of kilometers or more. These differences can be used to estimate the nature of karstification in an aquifer, for instance, whether an aquifer is closer to macrokarstic or microkarstic end members.

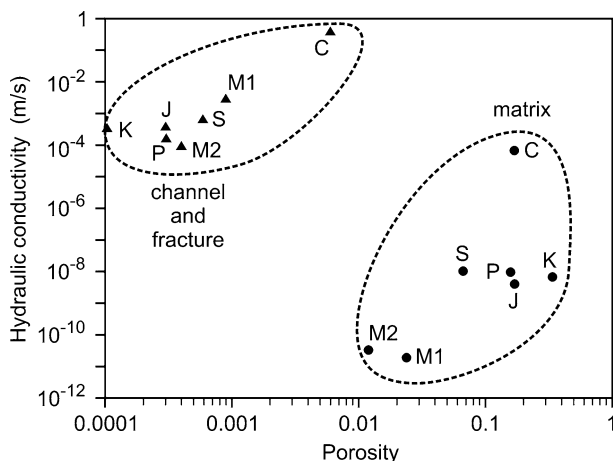
The specific discharge ( $L/s/km^2$ ) from an area can be calculated from the gauged discharge of river basins. If the mean discharge for a spring is measured, then the area of the groundwater basin feeding that spring may be determined. The boundaries of that groundwater basin can be determined by a combination of measuring heads in wells together with tracer tests to the spring (Fig. 11.5).



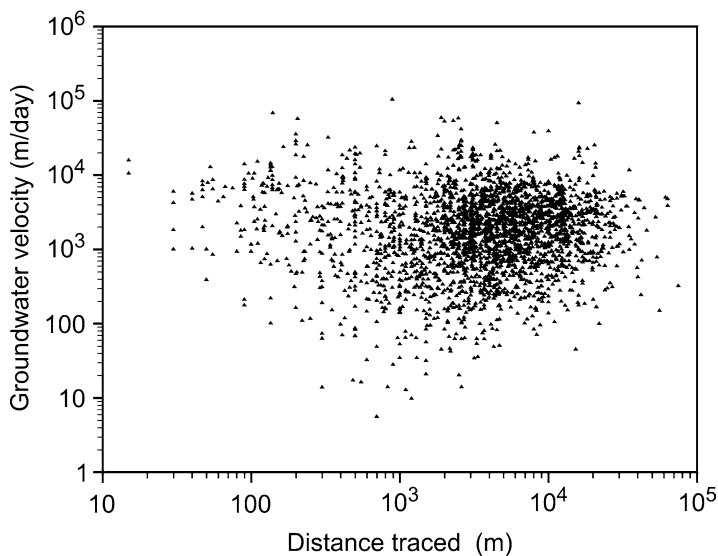
**Fig. 11.1** Stratigraphic correlation from gamma ray logs (black line) and inflow from flow meter measurements (blue dots) at the three municipal wells in use at the time of the Walkerton Tragedy (after Worthington et al. 2002)

### 11.3.3 Scaling Effects

In carbonate aquifers, most randomly-drilled boreholes are unlikely to intersect major channels. Consequently, localized hydraulic testing such as packer tests will have lower average permeabilities than tests such as pumping tests that sample

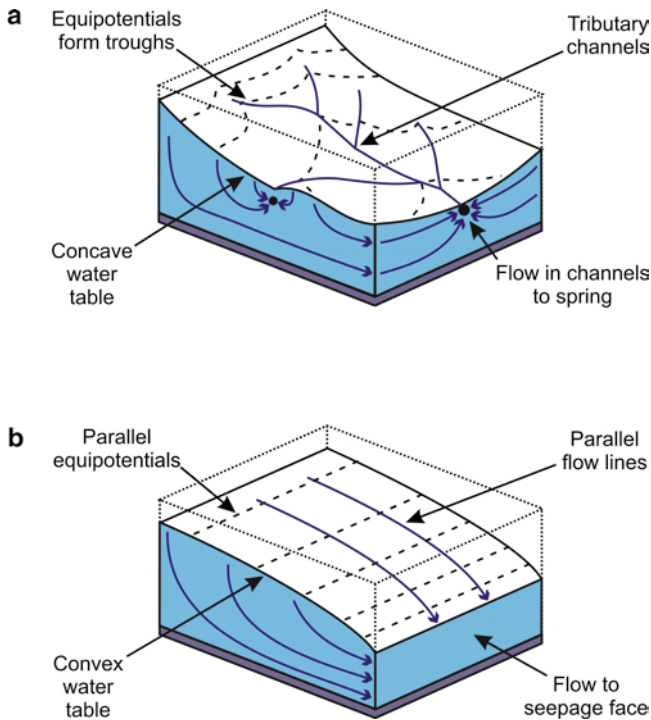


**Fig. 11.2** Porosity and hydraulic conductivity values for the matrix and for fractures and channels for Silurian dolostone (Ontario, S), Mississippian limestone (Kentucky, M1; England, M2), Permian limestone (England, P), Jurassic limestone (England, J), Cretaceous limestone (England, K) and Cenozoic limestone (Mexico, C), (after Worthington and Ford 2009)



**Fig. 11.3** Groundwater velocities for 3,015 tracer tests along channels in carbonate aquifers (after Worthington and Ford 2009)

a larger volume of the aquifer. This property is known as the scaling effect. It was introduced by Kiraly (1975), and his data for limestones in the Jura Mountains in Switzerland are shown in Fig. 11.6, as well as data from several other carbonate aquifers.

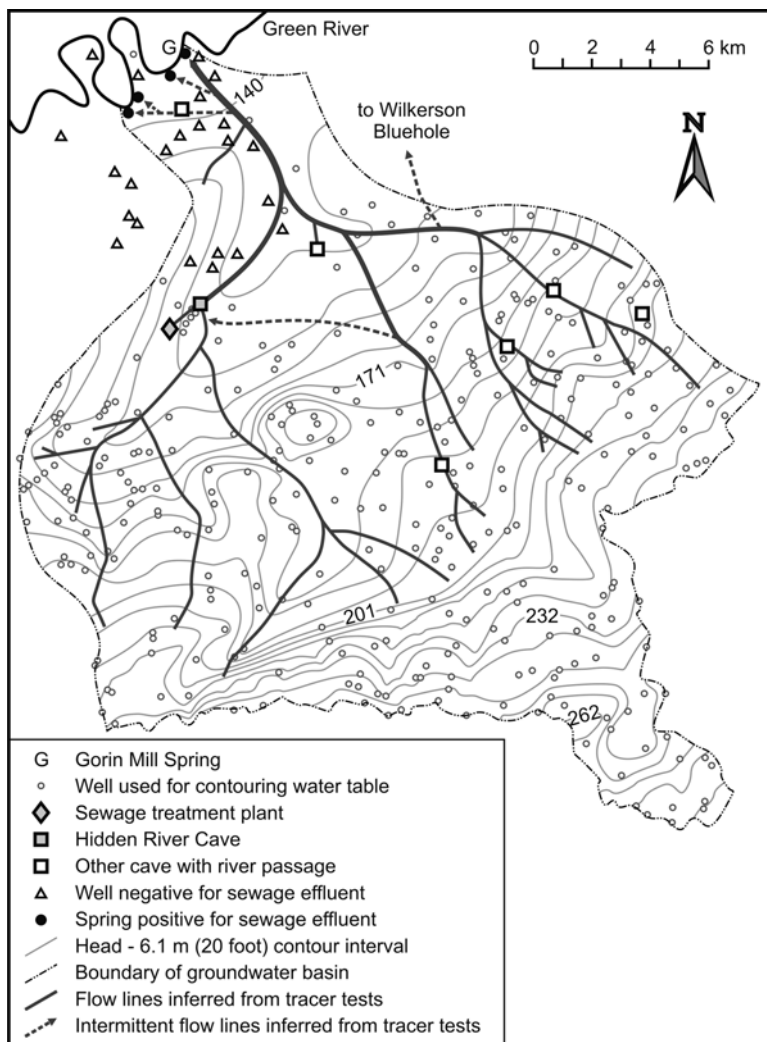


**Fig. 11.4** The principal aquifer-scale differences between (a) an ideal karst aquifer and (b) a homogeneous porous-medium aquifer (after Worthington 2009)

There are large differences in the hydraulic conductivity for tests on core rock samples (effective test radius assumed to be 0.1 m) because the matrix permeability of poorly compacted rocks (e.g., Cenozoic carbonates) is much higher than for highly compacted rocks that were deeply buried (e.g., most Paleozoic carbonates). However, as the test radius increases by moving from fist-sized rock samples to methods like pumping tests that sample ~100 m of the rock surrounding the borehole, the permeability values tend to converge, as all the carbonate aquifers shown have channel networks and thus have broad similarities at the scale of the whole aquifer. Scaling effects are substantial where karstification has occurred. However, where karstification is absent and bulk permeability is very low (e.g.  $<1^{-10}$  m/s), then the bulk permeability is likely to be similar to the matrix permeability and scaling effects may be absent. This broad similarity between different carbonate aquifers is also illustrated by Fig. 11.2, where the results are grouped into two well defined areas, irrespective of the age, climate and recharge, folding, or other geological history characteristics

The contrasts between the two distinct fields on Fig. 11.2 can be described as double-porosity behavior. Water moves quickly through the channel network, but slowly through the matrix. This means that the residence time of the groundwater in





**Fig. 11.5** Contour map of water table heads and groundwater flow paths (*black lines*) in the groundwater basin that drains to Gorin Mill Spring, Kentucky. The streams in the southern part of the area (not shown) flow across impure limestones and then sink on reaching purer limestones. Consequently, there is no surface drainage in the northern two-thirds of the drainage basin (compiled from Quinlan and Ray 1989; Quinlan and Ewers 1989; and Ray 1997)

the channels and fractures is likely to be substantially less than the matrix water. Water discharging from the aquifer will be a blend of all these waters and so will have a mixed age. The implication is that age dating using environmental tracers such as tritium, He/H, or CFCs will give much older ages than age dating using tracers that are introduced to the channel network and sampled at a spring in a

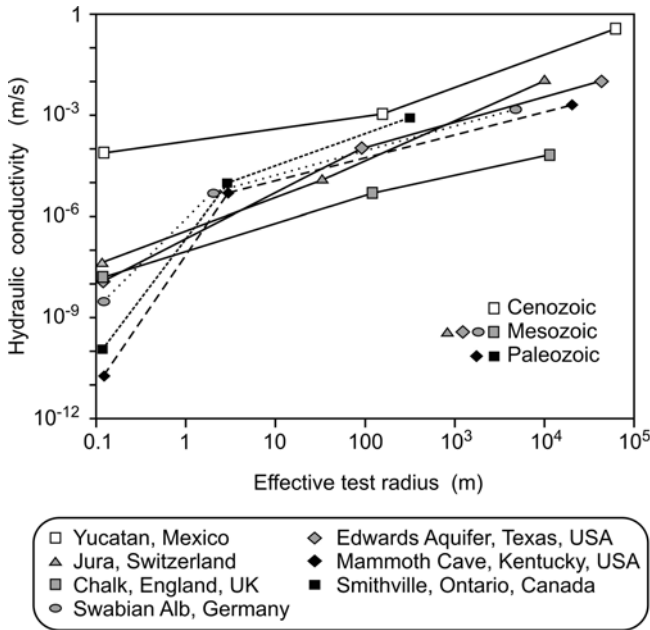
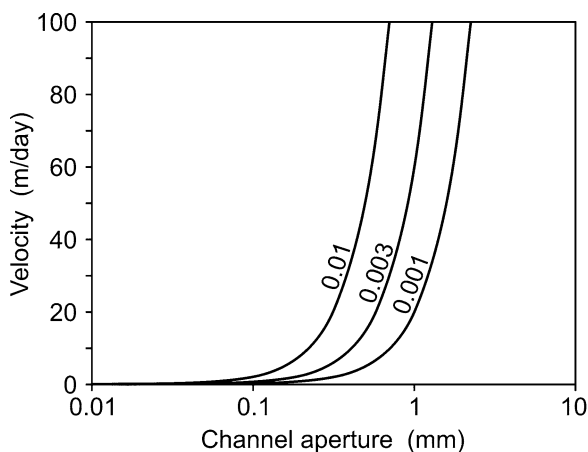


Fig. 11.6 Hydraulic conductivity as a function of testing scale (after Worthington 2009)

natural-gradient test or at a well in a pumping test. There have been a number of case studies where spring water has been dated using both environmental tracers and introduced tracers. The introduced tracers gave average groundwater ages that were about two orders of magnitude less than the environmental tracers (Worthington 2007). For instance, <sup>3</sup>H/<sup>3</sup>He dating gave a groundwater residence time of 39 years for the discharge at Wakulla Springs (Florida), assuming that the aquifer behaves as a porous medium (Katz 2001). However, subsequent tracing using fluorescent dyes showed that the travel times are of only days to weeks over distances of some 10 km between sinking streams and Wakulla Springs (Loper et al. 2005). The dye tests demonstrated rapid flow along channels whereas the <sup>3</sup>H/<sup>3</sup>He dating gave an average residence time value that includes the much slower flow through the matrix of the rock. Consequently, the two tracing methods yield complementary information about transport characteristics in these aquifers.

Groundwater velocities through the channel networks can be estimated using Darcy’s law, using channel/fracture porosity as the effective porosity (Fig. 11.2). However, direct measurement using introduced tracers has been widely used and is highly recommended where any questions of contaminant transport are being addressed. Tracer test results can also be used to estimate channel apertures (Fig. 11.7). It is notable that channel apertures as small as 1 mm will give groundwater velocities of tens or even hundreds of meters per day. Such velocities are usually much greater than estimates derived from assumptions that an aquifer behaves as an equivalent porous medium.

**Fig. 11.7** Groundwater velocity along a circular channel, calculated from Poiseuille's Law



## 11.4 Water Supply Issues

Carbonate rocks outcrop over large areas globally and their high permeability ensures that their use for water supplies is widespread. However, the high permeability has a number of adverse consequences. One is that contamination can move quickly through these aquifers; this will be discussed in [Sect. 11.5](#). A second consequence of the high permeability is that infiltration is high, often resulting in an absence of stream flow on the surface. The high permeability also results in low hydraulic gradients and in mountainous areas, this results in thick unsaturated zones. The lack of surface flow and thick unsaturated zones together mean that it is expensive to access groundwater since deeper wells are required, and the water has to be pumped to a higher relative level. Therefore, communities often rely on springs for their water supplies.

Groundwater divides in carbonates may differ significantly from surface water divides, especially where the unsaturated zone is thick. Administrative boundaries often coincide with surface drainage divides, and aquifers that cross such boundaries may create jurisdictional problems due to competing interests. In carbonate aquifer, tracer tests are often used to define groundwater divides. A further consideration in macrokarstic aquifers is that there is a low probability of a well intersecting the major channels so that most wells have low yields. These conditions make drilling wells in carbonate aquifers often more problematic than in simpler porous medium aquifers. Examples from England and from Texas will be described below to illustrate some of the problems associated with water supply from limestone aquifers.

### 11.4.1 Limestone Aquifers in England

Contrasts in groundwater abstraction from the Carboniferous Limestone and the Chalk in England exemplify the range of conditions that may occur. The Carboniferous Limestone outcrops in hilly areas such as the Mendip Hills and the Peak District,

where the unsaturated zone is up to 200 m thick. Surface flow is absent in the hilly areas but there are streams and rivers in deep valleys that are incised down to the water table. There are many sinking streams and discharge from the aquifer is often from large springs (Atkinson 1977). The geometric mean transmissivity from 59 pumping tests in the Carboniferous Limestone is 22 m<sup>2</sup>/day and the standard deviation of log transmissivity is 1.31 (Worthington and Ford 2009). The thick unsaturated zone, low average transmissivity, and high standard deviation mean that using wells to exploit the water supply is expensive and unpredictable. As a result, there are few wells on the limestone that outcrop and water supplies are often drawn instead from springs.

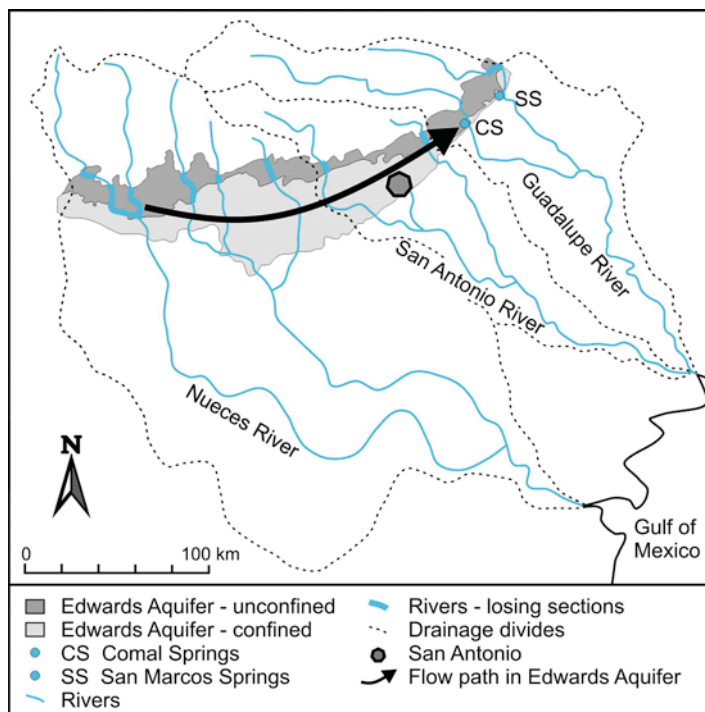
The Cretaceous Chalk outcrops in south-east England in a series of cuestas (Downing et al. 1993). The unsaturated zone is often >50 m thick in areas at higher elevation. There are few major sinking streams and discharge from the Chalk is predominantly from a large number of small springs at the margins of the Chalk outcrop. There are many water-supply wells over the outcrop of the Chalk although springs are also used. The Chalk provides more than half of groundwater abstraction in all of England due to it being the predominant aquifer in the south-east coincident with the highest population densities. The geometric mean transmissivity from 1,257 pumping tests is 440 m<sup>2</sup>/day and the standard deviation of log transmissivity is 0.76. The high average transmissivity and low standard deviation are a result of the microkarstic nature of the aquifer, with many small channels (Schürch and Buckley 2002; Maurice et al. 2006). This also means that high-productivity wells can be more reliably drilled in the microkarstic Chalk than in the macrokarstic Carboniferous Limestone aquifers.

#### ***11.4.2 Edwards Aquifer, Texas, U.S.A.***

The Edwards Aquifer is a Cretaceous limestone aquifer located in a semi-arid area with a growing population. It provides almost the whole water supply for San Antonio, a city with a population of 1.7 million people. The aquifer is about 150 m thick and dips to the south-east at about 1°. Rivers flowing from the north lose much of their recharge to the aquifer when they cross its outcrop (Fig. 11.8). To the south, the confined aquifer extends to depths of more than 1,000 m below the surface. The major natural points of discharge from the aquifer are at Comal Springs and San Marcos Springs, where mean flows are 8.2 and 5.0 m<sup>3</sup>/s, respectively. Pumping from wells accounts for about half the total discharge from the aquifer (Hamilton et al. 2009).

Water flowing to Comal Springs comes from as far as 225 km to the west. This results in major inter-watershed transfers of groundwater from the Nueces and San Antonio watersheds to Comal and San Marcos Springs, which are in the Guadalupe watershed (Fig. 11.8). This is a remarkable example of how groundwater divides may differ from surface water divides.

Increases in pumping from the aquifer resulted in some of the springs becoming intermittent and the viability of federally-listed endangered species in Comal and



**Fig. 11.8** Inter-watershed transfer of groundwater in Texas. Riverbed losses in the Nueces and San Antonio watersheds flow in the Edwards Aquifer to springs in the Guadalupe watershed (modified from Hamilton et al. 2009)

San Marcos Springs was threatened. Following a lawsuit brought by the Sierra Club in the 1990s against the U.S. Fish and Wildlife Service, the Edwards Aquifer Authority was formed to manage the aquifer. A permit system is now in use to limit withdrawals from the aquifer so that flows are maintained at the springs. In addition, a “critical period” plan is in place to reduce aquifer withdrawals when the discharge from Comal Springs or San Marcos Springs or the water levels in index wells in San Antonio or Uvalde falls below trigger levels ([www.edwardsaquifer.org](http://www.edwardsaquifer.org)).

## 11.5 Contamination Issues

The presence of interconnected channels networks in carbonate aquifers means that groundwater can move rapidly through them (Figs. 11.3 and 11.4). It also means that contamination can move equally rapidly through them. For instance, a study of water quality was carried out in 1,174 wells in 16 major aquifers in the U.S., six of which were partly or wholly in carbonate rocks (Embrey and Runkle 2006).

Wells in carbonate rocks were more likely to be positive for total coliform bacteria, for *Escherichia coli*, and for coliphage than wells in any other rock type. Given the rapid die-offs of these bacteria in groundwater, this provides evidence of rapid flow of surface waters from a source of contamination through the aquifer system to the wells that were sampled.

Contaminant plumes in carbonates are substantially different from plumes in porous medium aquifers because the contamination will spread quickly along the channel network towards springs. At the same time, contaminant movement into fractures and the matrix of the rock may be several orders of magnitudes slower. Consequently, contaminant plumes in carbonate rocks are much more complicated than the simple oval shapes that are found in homogeneous porous media. Examples from Kentucky and Ontario will be described below to illustrate some of the problems associated with contamination in carbonate aquifers.

### ***11.5.1 Gorin Mill Spring Groundwater Basin, Kentucky, U.S.A.***

Detailed water table maps based on well data and using information from tracer tests and cave maps have given many insights on how the flow in carbonate aquifers is organized. The most comprehensive such map covers an area of more than 1,000 km<sup>2</sup> in Kentucky, and is based on more than 400 tracer tests and water levels in 1,500 wells (Quinlan and Ray 1989). Figure 11.5 shows an excerpt from this map. It depicts the 380 km<sup>2</sup> area that drains to Gorin Mill Spring, the largest spring in Kentucky. To the east and west, there are similar, though smaller, groundwater basins and these also drain to large springs on the Green River.

In the southern part of the Gorin Mill Spring catchment area, streams flow across impure limestones and then sink on reaching purer limestones. Consequently, there is no surface drainage in the northern two-thirds of the drainage basin. The flow lines represent the inferred paths of tracer tests. These are located largely along troughs in the water table, though in a few places it is possible to access underground streams. The largest of these is at Hidden River Cave, a show cave in the center of the city of Horse Cave. The tracer injections are at sinking streams, wells, or cave streams. The low heads along the troughs result in flow converging on these troughs, and all the tracer tests flowed to Gorin Mill Spring, a large perennial spring. Tracer tests in high flow demonstrated that there are overflow conduits that allow water to spill over into different parts of the basin. One of these conduits drains to Wilkerson Bluehole, a large spring on the Green River 6 km east of Goring Mill Spring (Ray 1997). There are a number of intermittent springs that discharge some of the flow from the basin during high-flow periods. Flow in the Gorin Mill Spring basin converges on large conduits that have diameters of 5–10 m. These form what is essentially a tributary network of conduits feeding one main spring, though with some distributary conduits.

Until 1912, the town of Horse Cave obtained its water from Hidden River Cave, which is located in the center of the town. Subsequently, municipal water

was obtained from wells, and more recently from a spring 26 km away. Waste disposal from residences and industry was by means of septic tanks or directly into wells or dolines. Hidden River Cave was a show cave from 1916 to 1943, when it was forced to close due to pollution. In 1964, a sewage treatment plant went into operation. However, the plant was not capable of adequately purifying the incoming waste stream. This included wastes from a creamery and from a metal-plating plant. The contamination of Hidden River Cave continued and for many years the stench from the polluted water was noticeable through much of the center of the town. In 1989, a new regional sewage treatment system went into operation and this effectively ended the half century of gross contamination of the aquifer.

Quinlan and Rowe (1977) sampled 23 wells between the sewage treatment plant and the Green River and also a number of springs close to the river. Effluent from the sewage treatment plant in Horse Cave had elevated concentrations of chromium, copper, nickel, and zinc, with chromium concentrations that at times exceeded 10 mg/L. Gorin Mill Spring and the nearby intermittent springs were all positive for these metals, but the 23 wells between the sewage treatment plant and Green River were all found to be negative. Tracer testing convincingly demonstrated why the apparently down-gradient wells situated in between the source and the springs were negative; this showed that flow from the sewage treatment plant was to several springs, but not to any of the 23 wells (Fig. 11.5). Flow in the aquifer converges on the major conduits, which provide pathways with very high permeability through the aquifer and thus occupy troughs where hydraulic gradients are very low. The springs represent the terminal points for the conduits where the groundwater discharges to the surface.

The groundwater studies carried out in this area of Kentucky demonstrate a number of ways in which carbonate aquifers differ from porous medium aquifers. Springs can be effective monitoring points since they may discharge the water recharge over considerable surface areas. Conversely, wells that are apparently down-gradient and even situated very close to a facility may not provide reliable monitoring points unless they can be shown by tracer testing to be on the flow path from that facility. Flow in most of the aquifer is convergent to conduits, but distributaries may be found close to the output point.

It is a common practice in North America to place monitoring wells down-gradient from a contaminated site and to sample them at fixed intervals such as monthly or quarterly. Such a monitoring program is unlikely to provide adequate sampling of contaminant transport where there are substantial conduits as the down-gradient conduit is most unlikely to be intercepted by a randomly-placed well. Furthermore, the rapid flow in conduits means that water quality sampling can be extremely variable in the short term, especially following major recharge events. Quinlan (1990) recommended that springs or wells that have been shown by tracer testing to drain the facility should be monitored, and that there should be frequent sampling through major runoff events. The advent of inexpensive submersible data loggers that measure head, temperature, and electrical conductivity has made this task of monitoring during major events much easier.



### ***11.5.2 Bacterial Contamination at Walkerton, Ontario, Canada***

In May 2000, about 2,300 people became ill and seven people died following bacterial contamination of the municipal water supply at Walkerton. Following groundwater and other investigations, a public inquiry, the Walkerton Inquiry, was held to determine the reasons why the water supply became contaminated (O'Connor 2002). It was found that the pathogenic bacteria were derived from cow manure. Such bacteria have rapid die-off in groundwater, implying that the travel time from the surface source to the wells was probably only days or less.

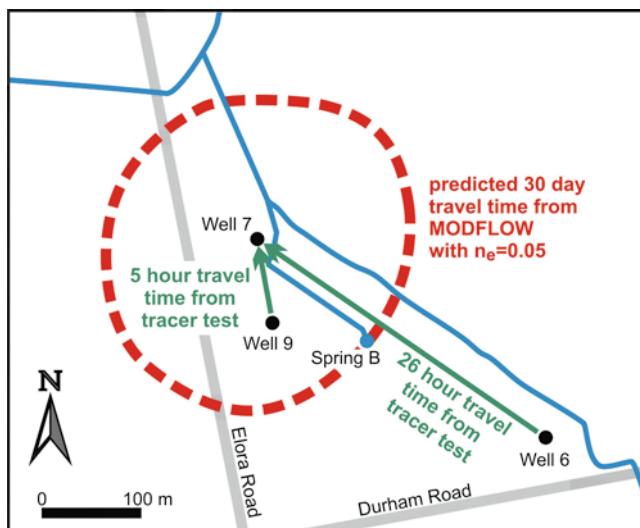
At the inquiry, two conceptual models of flow through the aquifer were presented. The first conceptual model was the one most commonly employed by hydrogeologists, that the aquifer behaves as an equivalent porous medium and that the effective porosity of the aquifer was similar to the total porosity of the aquifer, which in this case was about 5% (Golder 2000). This model also implicitly assumed that any productive horizons found in boreholes were isolated and did not form part of a high-permeability fracture/channel network. Modeling using MODFLOW showed that the 30-day capture zone at Well 7, the most productive municipal well, would extend about 150 m from the well (Fig. 11.1).

A second conceptual model assumed that the aquifer had a network of dissolutional channels. In this case, the relevant effective porosity for the transport of pathogenic bacteria through the aquifer would be closer to 0.1%. This porosity is known as the kinematic porosity (Worthington and Ford 2009). This estimate was based first on the cross-sections of the productive channels and fractures intersected in the wells. In Wells 5, 6, and 7, the number of horizons with measurable flow varied between one and nine and the sum of the cross-sections of these channels and fractures was estimated to be less than 0.1% of the cross-section of the boreholes (Fig 11.1). This indicated that the kinematic porosity of the aquifer would be <0.1%, assuming that the channels formed an interconnected network in the aquifer. The kinematic porosity estimate was also based on experience from many tracer tests (Fig. 11.3). If the kinematic porosity in the aquifer at Walkerton was as low as 0.1%, then groundwater velocities would be 50 times faster than with the first conceptual model and would likely exceed 100 m/day (Worthington 2001).

A post-audit was carried out in October 2001, after the hearings of the Walkerton Inquiry had ended. Tracers were introduced into Wells 6 and 9 and both travelled rapidly to Well 7 (Fig. 11.9). This demonstrated that the kinematic porosity of the aquifer was very low (<0.1%) and showed that the channels encountered in the boreholes formed part of an interconnected network, as is expected from theory (see Sect. 11.2 above). Calculations showed that the hydraulic apertures of the channels between injection wells and Well 7 were at least 3 mm (Worthington et al. 2002).

The tracer testing at Walkerton showed the importance of directly measuring travel times and how uncalibrated estimates based on porous medium assumptions may be extremely poor. The problems with such estimates were noted by Freeze and Cherry (1979, p. 427), who stated “velocity estimates based on the use of these parameters [hydraulic conductivity, hydraulic gradient, and porosity] in Darcy-based





**Fig. 11.9** Predicted 150 m diameter 30-day travel time zone to Well 7 at Walkerton, assuming an effective porosity of 0.05, and results from the post-audit introduced tracer tests that measured travel times from Well 6 and Well 9 respectively (from Worthington et al. 2002)

equations have large inherent uncertainties that generally cannot be avoided". The Walkerton Tragedy demonstrates how carbonate aquifers may be susceptible to bacterial contamination because of the presence of interconnected channel networks with concomitant rapid groundwater velocities.

## 11.6 Conclusions

Carbonate aquifers are common globally and are widely utilized due to their high permeability. Advances in recent decades in understanding dissolution kinetics have facilitated the numerical modeling of dissolutional enhancement of permeability. This has shown how the dissolution results in an interconnected network of channels that not only results in high permeability but also in rapid groundwater velocities.

In mountainous areas, the high permeability results in a lack of surface water and thick unsaturated zones, so utilization of groundwater is commonly from low-elevation springs. Groundwater divides may not coincide with surface-water divides, sometimes resulting in jurisdictional issues over exploitation of the groundwater.

Contamination from point sources travels down-gradient through channel networks to springs, so that springs are useful monitoring locations. Monitoring using wells is challenging as much of the contamination may travel along major channels and bypass wells. Contaminant movement into fractures and the matrix of the rock may be several orders of magnitudes slower than movement through the channels.

Consequently, contaminant plumes are much more complicated than the simple oval shapes that are found in homogeneous porous media. The rapid groundwater velocities in the channel network also make carbonate aquifers more susceptible to bacterial contamination than other aquifer types.

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