Verification and characterization of a fracture network within the Maquoketa Shale Confining Unit, southeastern Wisconsin: final report prepared for the Wisconsin Department of Natural Resources. 200...

Eaton, Timothy T. (Timothy Theodore); Bradbury, Kenneth R. (Kenneth Rhoads); Wang, Herbert
Madison, Wisconsin: Wisconsin Department of Natural Resources, 2001

https://digital.library.wisc.edu/1711.dl/K3BW3NS473SLL8B

VERIFICATION AND CHARACTERIZATION OF A FRACTURE NETWORK WITHIN THE MAQUOKETA SHALE CONFINING UNIT, SOUTHEASTERN WISCONSIN

A Final Report prepared for the

WISCONSIN DEPARTMENT OF NATURAL RESOURCES

Prepared by

Timothy T. Eaton[1,2]
Kenneth R. Bradbury[1]
Herbert F.Wang[2]

[1]WISCONSIN GEOLOGICAL AND NATURAL HISTORY SURVEY
3817 Mineral Point Road
Madison, Wisconsin 53705

[2]UNIVERSITY OF WISCONSIN-MADISON
DEPARTMENT OF GEOLOGY AND GEOPHYSICS
1215 W. Dayton Street
Madison, Wisconsin 53706

WGNHS Open-File Report 2001-04

Period of contract:

July 1, 2000-July 1, 2001

Conducted with the support of the

WISCONSIN DEPARTMENT OF NATURAL RESOURCES
UNIVERSITY OF WISCONSIN WATER RESOURCES INSTITUTE

December 15, 2001

I. TABLE OF CONTENTS

II. PROJECT SUMMARY

Background/Need: The Maquoketa Formation, a dolomitic shale, forms the most important aquitard in eastern Wisconsin, USA, isolating the water-table and Silurian aquifers from the underlying Cambrian-Ordovician aquifer. The hydraulic properties of the Maquoketa aquitard are of interest for input to a regional groundwater flow model, which will be used for better groundwater resource management and well-head protection.

Analysis of earlier results showed that while the Maquoketa Formation has very low rock-matrix hydraulic conductivity, simultaneous drawdowns occurred at multiple levels within this formation when the overlying Silurian aquifer was pumped. Such observations violate equivalent porous medium assumptions but they can be explained by local vertical fracture interconnections between the observed bedding-plane fractures within the Maquoketa Formation.

Objectives: The objective of the additional fieldwork described in this report was to characterize hydraulic and transport properties of a potential fracture network extending from the Silurian dolomite aquifer into the Maquoketa Formation. If significant vertical fracture interconnections exist, they could result in much higher bulk hydraulic conductivity in the upper Maquoketa Formation. The observed vertical head distribution showed that almost half of the total vertical head loss due to regional pumping occurs across the lower contact of the formation, which is not consistent with a homogeneous, very low-conductivity aquitard at steady-state with the heads in the adjacent aquifers.

Methods: Therefore, we proposed the use of more specialized methods, better suited to fractured rock characterization, to investigate the hydraulic properties and interconnectivity of discrete transmissive fractures in the Maquoketa Formation. These methods consist of short-interval packer testing and tracer experiments, which were conducted using wells at the DOT field site in Waukesha County. Preliminary geochemical and isotopic analysis of water samples showed no significant trends with depth, and a more comprehensive analysis is presented in this report.

Results and Discussion: The highest values of hydraulic conductivity are associated with the upper part of the Maquoketa Formation, where bedding-plane fractures associated with dolomitic beds had previously been identified. There is a broad range of values over almost 5 orders of magnitude, which results from the high transmissivity of these bedding-plane fractures, present in only some of the intervals tested. One five-foot interval that contained numerous small-aperture fractures is connected to a larger transmissive network, and yielded a sustainable pumping rate causing drawdown in other wells.

Water samples showed primarily calcium-magnesium-bicarbonate geochemistry, with sodium and sulfate more abundant in parts of the Maquoketa Formation. Variation in electrical conductivity and concentrations of major ions with depth was significant. We found no tritium, but high levels of strontium, normally a trace constituent, in water from the Maquoketa Formation. Convergent-flow tracer experiments were conducted using heated water, but no breakthrough (increase in temperature) was recorded by in-situ instrumentation in the Maquoketa Formation. However, drawdown observed in different wells due to pumping during these long-duration tracer tests was analyzed to estimate transmissivity of the fracture system.

Conclusions/Implications/Recommendations: The hydrogeology of the Maquoketa Formation is determined by significant geologic heterogeneity. This heterogeneity consists primarily of transmissive bedding-plane fractures, which are associated with interbedded shale and dolomite facies in the upper part of the formation. Oxygen and deuterium isotope data, and the lack of tritium shows that groundwater in the Maquoketa Formation is not modern and may date in part to Pleistocene times. The major ion chemistry plots along mixing trends, between endpoints interpreted to be characteristic of water in fractures and rock pores. The heterogeneity of water chemistry and variation in field-measured hydraulic conductivity are consistent with a three-dimensional flow system controlled by transmissive fractures within the aquitard.

Water chemistry and isotope data, and tracer tests indicate that no significant vertical flow occurs between the Silurian aquifer and Maquoketa Formation, although limited vertical interconnections may exist between bedding-plane fractures. The lateral continuity of the transmissive bedding-plane fractures and limited vertical interconnections increase the bulk hydraulic conductivity of the dolomitic upper part of the Maquoketa Formation. The current transient head field in the Maquoketa Formation coupled with the uniformly low hydraulic conductivity of the base of the formation account for the hydraulic effectiveness of this aquitard. However, the top of the Maquoketa Formation cannot be assumed to be virtually impermeable to contaminants, for example DNAPLs, whose migration is not dependent on head gradients.

Related Publications: Eaton, T.T. 2002. *Fracture heterogeneity and hydrogeology of the Maquoketa aquitard, southeastern Wisconsin.* unpublished Ph.D. dissertation (Geology and Geophysics), University of Wisconsin-Madison, 200 p.

Key words: Maquoketa, aquitard, fractures, heterogeneity

Funding: Wisconsin Department of Natural Resources

Final Report: A final report containing more detailed information on this project is available for loan at the Water Resources Institute, University of Wisconsin-Madison, 1975 Willow Drive, Madison, Wisconsin 53706. (608) 262-3069.

III. INTRODUCTION

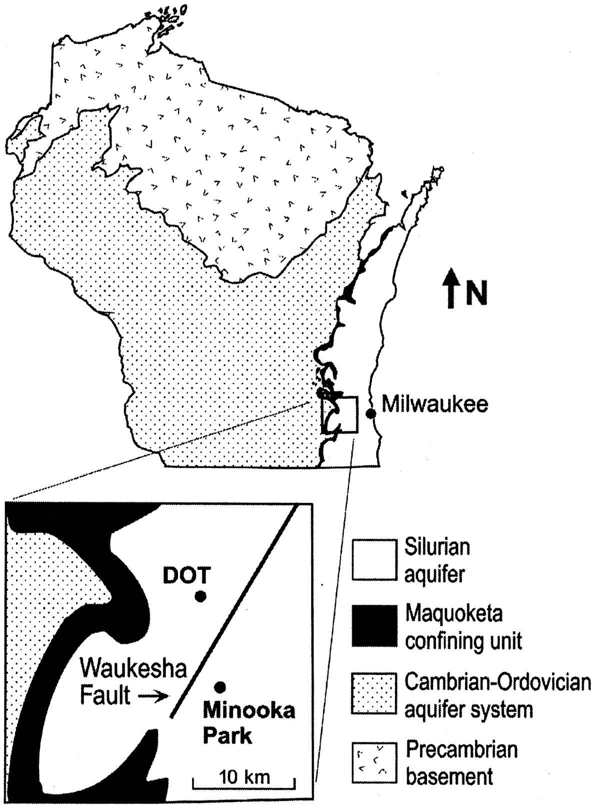
A. Background

This report represents the culmination of a series of research projects by a number of collaborating investigators over several years, funded by the Wisconsin Department of Natural Resources and the University of Wisconsin Water Resources Institute. Details on the different phases of the study can be found in previous reports (Eaton et al. 2000, Eaton and Bradbury 1998) which are also available as Wisconsin Geological and Natural History Open-File Reports. Further investigation of these results using local-scale groundwater flow modeling, and a comprehensive analysis of the entire research effort are presented by Eaton (2002).

The results in this report are presented in sections, two of which correspond to the methods of investigation proposed for this project: packer and tracer testing, and a third contains the results of geochemical and isotopic analysis of water samples collected from the Maquoketa and adjacent formations. Preliminary analysis of some of the sample data was presented earlier (Eaton and Bradbury 1998), but a more comprehensive analysis is presented here in light of the heterogeneity found in the Maquoketa Formation.

B. Need and Objectives

The Maquoketa Formation, a dolomitic shale, forms the most important aquitard in eastern Wisconsin, USA, isolating the water-table and Silurian aquifers from the underlying Cambrian-Ordovician aquifer (Figure 1). Prior to this series of research projects, hydraulic properties were not well known, and the formation was generally assumed to be hydraulically homogeneous. Since 1900, increased groundwater pumping for municipal supply in the vicinity of Milwaukee, Wisconsin, has caused more than 400 ft (120 m) of drawdown in the underlying Cambrian-Ordovician aquifer system. The reversed vertical head gradient across the aquitard since pre-development times makes leakage possible to the confined aquifer. The hydraulic properties of the Maquoketa aquitard are of interest for use in a regional groundwater flow model, which will be used for better groundwater resource management and well-head protection.



Simplified rock-stratigraphic nomenclature		**Hydro-geology**
Group	Formations	Symbol
(Quaternary)	(undif)	not shown
Silurian	(undif)	
Maquoketa	Maquoketa	
Sinnipee	Galena Platteville	
Ancell	Glenwood St. Peter	
Trempealeau	Jordan St.Lawrence	
Tunnel City	(undif)	
Elk Mound	Wonewoc Eau Claire Mt. Simon	
Precambrian (undifferentiated)		

(adapted from Ostrom 1967)

Figure 1. Field site locations in Waukesha County and generalized stratigraphy, southeastern Wisconsin. Note proximity to western subcrop of Maquoketa Formation and Waukesha Fault trend. In the area east of the Maquoketa subcrop, the underlying Sinnipee Group dolomite is part of the aquitard, but is not considered in this study.

Previous phases of this research effort, focused on investigation of the hydrogeology of the Maquoketa aquitard in southeastern Wisconsin, are described in earlier reports to the University of Wisconsin Water Resources Institute and Wisconsin Department of Natural Resources (Eaton et al. 2000, Eaton and Bradbury 1998). They included both field hydraulic testing as well as laboratory testing of rock core and poroelastic modeling (Eaton et al. 2000). Based on analysis of the earlier results, additional fieldwork was proposed and the results of the additional testing are presented in this report. Pumping a well open only to the overlying Silurian aquifer had caused drawdown at multiple intervals in a separate multi-level monitoring well open to the underlying Maquoketa Formation. These drawdowns all occurred in less than one minute and deeper intervals showed greater drawdown than shallower intervals closer to the pumped well. As discussed by Eaton et al. (2000), these observations violate equivalent-porous-medium assumptions about the Maquoketa Formation, which was shown to have very low rock-matrix hydraulic conductivity.

Description of core samples had indicated that the lithology of the Maquoketa Formation at two field sites (Minooka Park and DOT) in Waukesha County (Figure 1) is very heterogeneous and characterized by six lithofacies (Eaton 2002, Eaton and Bradbury 1998). In contrast to the base, a fissile greenish-gray dolomitic shale, the upper two-thirds of the formation is characterized by more indurated interbedded dolostone (mudstone through grainstone). Downhole geophysical logs and slug testing showed that transmissive bedding-plane fractures are associated with these interbeds. The simultaneous drawdowns observed during pumping of the overlying Silurian aquifer could be explained by local vertical fracture interconnections between the observed bedding-plane fractures.

The objective of the additional fieldwork described in this report was to characterize hydraulic and transport properties of a potential fracture network extending from the Silurian aquifer into the Maquoketa Formation. If significant vertical fracture interconnections exist, they could result in much higher bulk hydraulic conductivity in the upper Maquoketa Formation. Furthermore, the vertical distribution of hydraulic head measured in the Maquoketa Formation (Eaton 2002, Eaton et al. 2000, Eaton and Bradbury 1998) is not consistent with a homogeneous, very low-conductivity aquitard at steady-state with significantly different heads in the adjacent aquifers. We found that over half the 400 ft (120 m) of head loss occurs across less than 30 ft at the lower contact with the underlying Sinnipee Group dolomite (Eaton 2002). Higher bulk hydraulic conductivity in the upper Maquoketa Formation could account for this vertical head distribution, and would imply that the Maquoketa aquitard is significantly thinner than previously assumed.

Therefore, we proposed the use of more specialized methods, better suited to fractured rock characterization, to investigate the hydraulic properties and interconnectivity of discrete transmissive fractures in the Maquoketa Formation. These methods consist of short-interval packer testing and tracer experiments, described in the following sections, which were conducted using wells at the DOT field site (Figure 2). Collection of water samples for geochemical and isotopic analysis was originally planned as a way of estimating vertical hydraulic conductivity by analyzing vertical trends. Preliminary analysis (Eaton and Bradbury 1998) showed no significant trends with depth, and a more comprehensive analysis of geochemical and isotopic data is presented in this report.

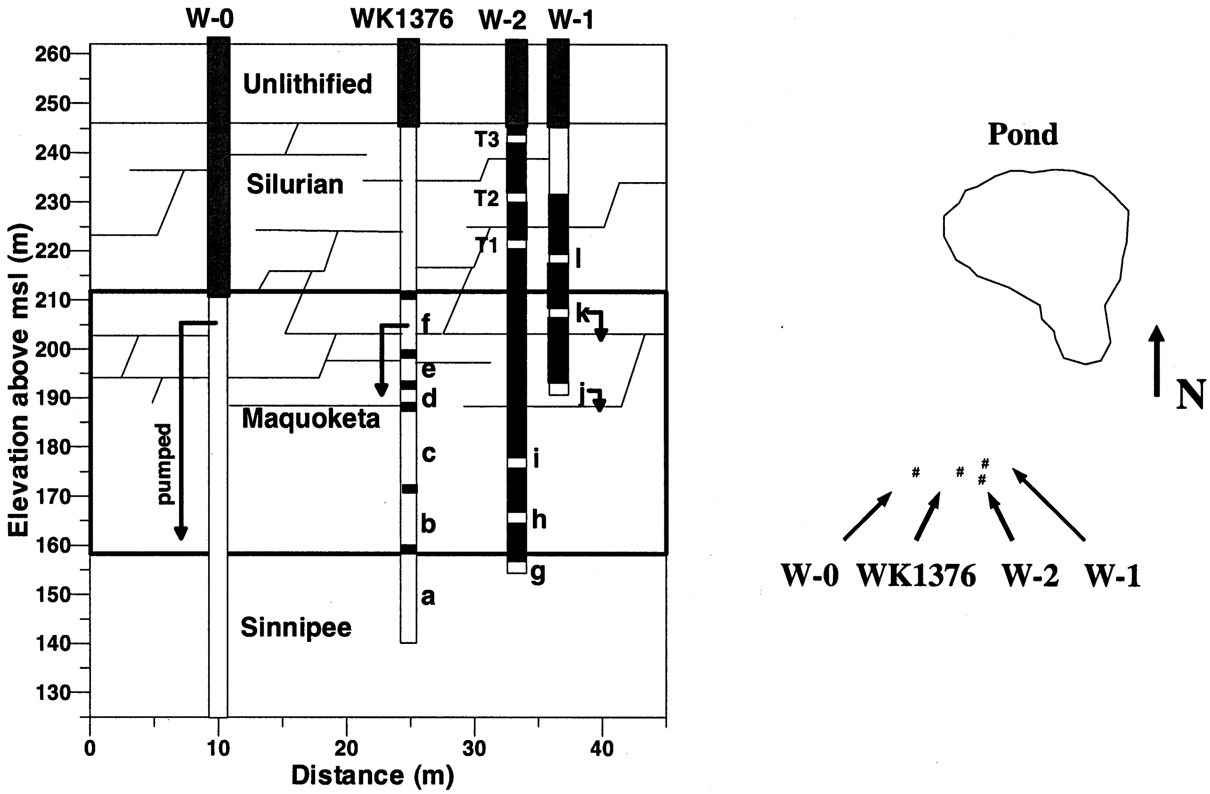


Figure 2. Schematic cross-section of monitoring wells (l) and plan view (r) at DOT field site. Black sections in wells indicate packers or bentonite backfill between monitored intervals. White sections in wells indicate open intervals, or instrumented monitoring zones (a-l). Arrows on section indicate magnitude of significant head change occurring during tracer test. The pond and long open intervals in wells WK1376, W-0 and W-1 as well as intervals T3, T2, T1 were sampled for geochemical analysis.

IV. SHORT-INTERVAL PACKER TESTING

A fracture network cross-cutting a low-permeability rock matrix creates an extremely heterogeneous hydraulic system in which head changes are very rapidly transmitted through the sparse, low-gradient fractures, but very slowly through rock matrix far from any fracture. Hence, measurements of head and hydraulic conductivity are highly dependent on the location, scale and interconnectivity of the fractures. If head is monitored in a relatively long open interval in the multi-level monitoring wells, it will be dominated by head changes in any fracture which may intersect the interval. Monitoring of head at smaller scales in unfractured intervals may produce values more representative of the rock matrix. Hydraulic testing in long intervals (30 ft) in the Maquoketa Formation, such as the slug testing reported by Eaton and Bradbury (1998), results in values that are averages of rock matrix properties and transmissivity of fractures intersecting those intervals. On the other hand, testing of short intervals may produce values more representative of fracture zones. This explains the range in values of field-measured hydraulic conductivity from $1x10^{-9}$ to $1x10^{-4}$ ft/s, and laboratory rock core measurements of $6.2x10^{-14}$ to $4.3x10^{-12}$ ft/s (Eaton et al. 2000).

A. Methods

To measure hydraulic properties that are more representative of fractures, we proposed short-interval straddle-packer testing at the DOT field site in well W-0 (WUWN PJ705) which is open to the entire Maquoketa Formation. We had planned to test as short intervals as possible, in the 1-2 ft range, but due to logistical and time constraints, we were only able to test 5 ft intervals. The apparatus (Figure 3) consists of two inflatable rubber packers, vertically separated by a screened interval, which are lowered into the well from a drill rig at the surface. The packers are inflated and the equilibrium water level in a standpipe from the screened interval is monitored with a pressure transducer. This water level is pumped down or perturbed abruptly, then allowed to recover, and the water level change over time is analyzed to determine hydraulic properties of the tested interval. Such methods have been used for testing fractured dolomite aquifers (Muldoon 1999, Bradbury et al. 1998).

For the Maquoketa Formation, we anticipated that transmissive bedding-plane fractures in the very low-conductivity rock matrix would result in a very broad range of hydraulic conductivity. We planned two different methods to perturb the water level in the standpipe of the short-interval packer system. For unfractured or less fractured low-conductivity intervals, we employed a wellhead manifold and nitrogen cylinder to pressurize the headspace in the standpipe (Figure 3a). For intervals intersecting high-transmissivity fractures, we pumped the screened interval using a submersible pump (Figure 3b). The change in head was imposed abruptly (less than 10 sec), satisfying requirements for common slug-testing methods (Butler 1998). Head change was monitored using pressure transducers, of which two were used for pressure testing, one measuring hydrostatic head in water and the other measuring gas pressure in the standpipe headspace (Figure 3a). A discussion of this pneumatic approach is given by Butler (1998). Similar test methods were described by

(a)

(b)

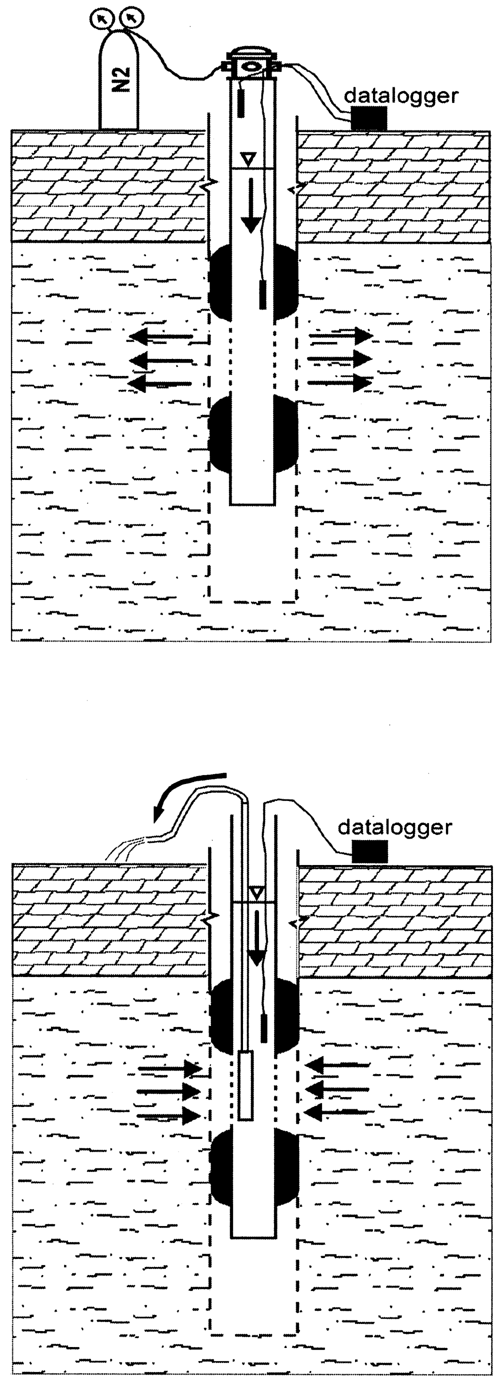


Figure 3. Schematic diagram showing straddle-packer testing methods: (a) for low conductivity intervals, standpipe is pressurized abruptly, creating head change and driving water into formation. Note that head is determined by differential between two pressure transducers. (b) for high conductivity intervals, head is drawn down by pumping test interval. Dimensions are not to scale.

Novakowski and Bickerton (1997) and Shapiro and Greene (1995), but full implementation of their analysis is beyond the scope of this study.

For pneumatic testing, hydraulic conductivity was estimated from head recovery following the release of pressure using primarily the Hvorslev (1951) method because of its simplicity and robustness. The Hvorslev equation (Butler 1998) is:

$$\mathbf{K = [(r_c^2)ln(R_e/r_w)]/(2BT_o)}$$

For this analysis, $\mathbf{r_c}$ is radius of casing or standpipe (0.125 ft), $\mathbf{R_e}$ is radius of influence, calculated as 200($\mathbf{r_w}$), where $\mathbf{r_w}$ is radius of well (0.250 ft), **B** is length of tested zone (5 ft), and $\mathbf{T_o}$ is basic lag time, read from the Hvorslev plot. $\mathbf{T_o}$ for the analyses presented here is given in Table 1 in the Results section.

The broad range of normalized-head curves with time shown in Figure 4 reflect different basic time lags ($\mathbf{T_o}$ at normalized head = 0.37) of response for the different intervals. The flatter curves (B and F) generally correspond to lower hydraulic conductivity while the steeper curves correspond to higher hydraulic conductivity (J and G). Some of the data are not very linear because of deviations from the assumptions (e.g., radial flow, homogeneity, negligible specific storage) of the Hvorslev model (Butler 1998). Estimates of hydraulic conductivity from fitting straight lines to these data are more uncertain. In some cases, the Cooper et al. (1967) method of slug-test analysis was applied without much success. Head in two intervals (J,N) did not recover after pressure displacement due to non-equilibrium initial conditions. For these intervals, data was corrected for the head trend prior to testing, and the initial head change was analyzed similar to a falling head test, but resulting values are probably overestimated. At a number of intervals (L,M,Q,T,U), no significant head displacement occurred and hydraulic conductivity was inferred to be 1E-9 ft/s or less, the assumed lower limit to the testing methods.

For more transmissive, fractured intervals, the Jacob semi-log method (Figure 5) was applied to recovery of head following cessation of pumping (Jacob 1963). The basic equations are:

$$\mathbf{K = T/b \qquad and \qquad T = (2.3Q)/(4p?s')}$$

T is transmissivity, **Q** is the pumping rate prior to cessation of pumping, **?s'** is extrapolated residual drawdown over one log cycle and **b** is length of interval tested (5 ft). Parameter values used in the Jacob analysis are given in Table 1. The Jacob method depends on linearity of late-time residual-drawdown values, and data in Figure 5 show that in some cases, the period of linear residual drawdown is quite short, increasing the uncertainty of the resulting estimates of hydraulic conductivity. For both pumping and pressure testing head recovery in the Maquoketa Formation, some uncertainty has been highlighted, a consequence of the difficulty of applying standard equivalent-porous-medium analysis methods to this

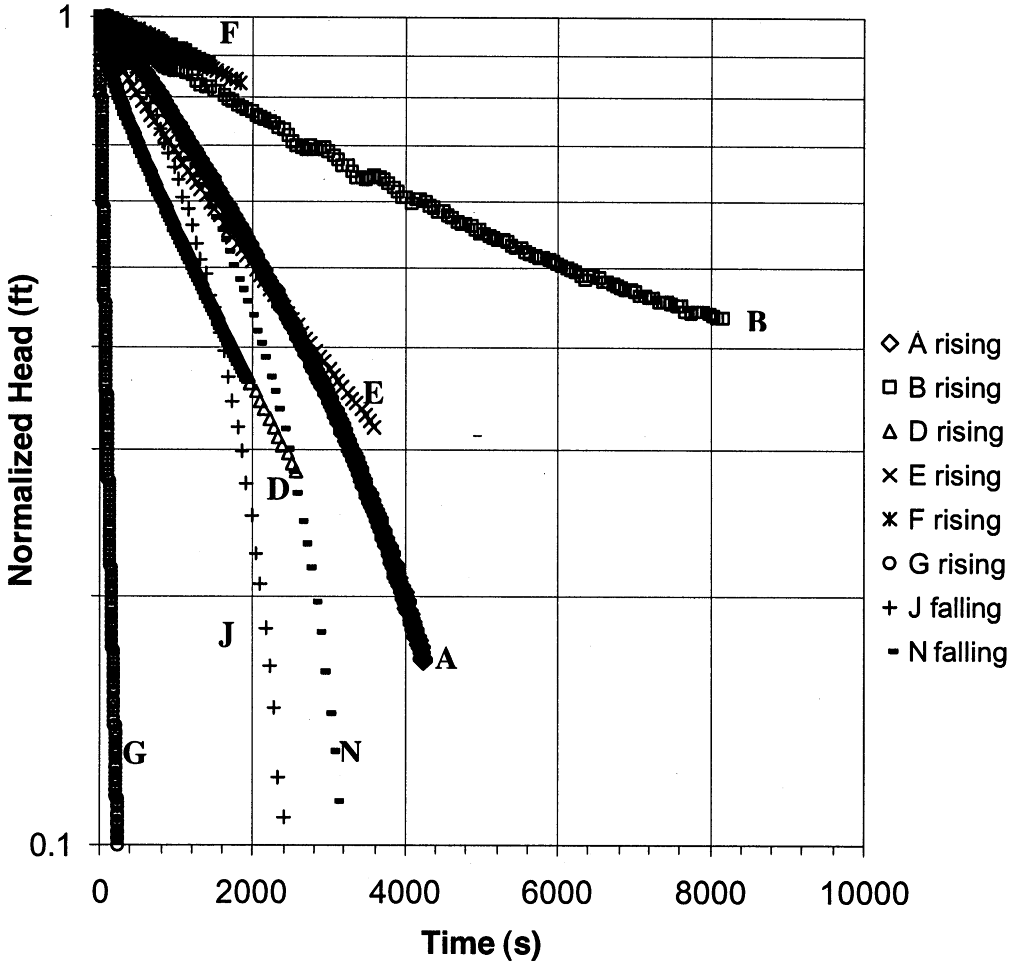


Figure 4. Change in normalized head over time as plotted for Hvorslev (1951) method for straddle-packer testing intervals (letters) using pneumatic technique. Refer to Table 1 for values estimated from straight lines fitted to each curve, plotted separately.

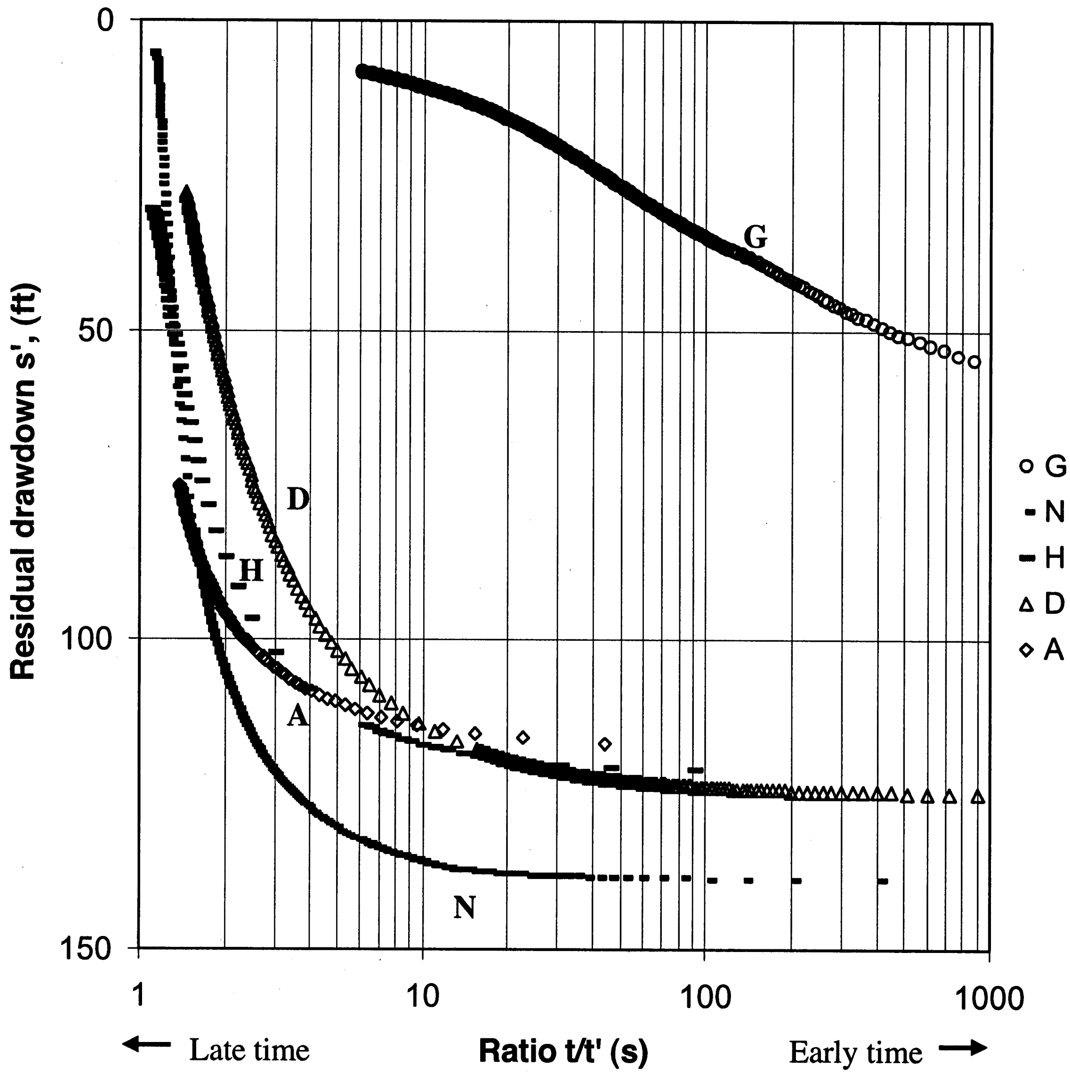


Figure 5. Head recovery, plotted for Jacob (1963) analysis, in intervals (letters) tested with submersible pump. Refer to Table 1 for values estimated from straight lines fitted to late-time linear portions of each curve, plotted separately.

heterogeneous, sparsely-fractured formation. However, the resulting hydraulic conductivity values are generally believed to be good estimates of the actual properties of the intervals tested despite this uncertainty.

B. Results

Table 1. Estimation methods, parameters and values of hydraulic conductivity from straddle-packer testing in the Maquoketa Formation

			Hvorslev (1951) analysis			Jacob (1963) analysis			
	Depth interval (ft)	Elev. (m.asl.)	T_o (s)	K (ft/s)	K (m/d)	Q (ft^3/s)	? s' (ft)	K (ft/s)	K (m/d)
A	167-172	210	2930	$2.8x10^{-6}$	$7.4x10^{-2}$	$1.8x10^{-4}$	187	$3.5x10^{-8}$	$9.2x10^{-4}$
B	172-177	209	9200	$9.0x10^{-7}$	$2.4x10^{-2}$				
D	182-187	206	1850	$4.5x10^{-6}$	$1.2x10^{-1}$	$8.2x10^{-4}$	220	$1.4x10^{-7}$	$3.7x10^{-3}$
E	187-192	204	3040	$2.7x10^{-6}$	$7.1x10^{-2}$				
F	192-197	203	9800	$8.4x10^{-7}$	$2.2x10^{-2}$				
G	197-202	201	90	$9.2x10^{-5}$	2.4	0.012	27	$1.6x10^{-5}$	$4.2x10^{-1}$
H	202-207	198	†			$2.0x10^{-4}$	270	$2.7x10^{-8}$	$7.1x10^{-4}$
J	217-222	195	1740	$4.8x10^{-6}$	$1.3x10^{-1}$				
L	227-232	192	*	$1x10^{-9}$	$2.6x10^{-5}$				
M	232-237	191	*	$1x10^{-9}$	$2.6x10^{-5}$				
N	237-242	189	21500	$4.0x10^{-7}$	$1.1x10^{-2}$	$6x10^{-5}$	680	$3x10^{-9}$	$8x10^{-5}$
Q	257-262	183	*	$1x10^{-9}$	$2.6x10^{-5}$				
R	272-277	178	‡						
S	281-286	176	?	$5.2x10^{-9}$	$1.4x10^{-4}$				
T	302-307	169	*	$1x10^{-9}$	$2.6x10^{-5}$				
U	320-325	164	*	$1x10^{-9}$	$2.6x10^{-5}$				

* no significant head displacement, K < equipment limit, assumed value $1x10^{-9}$ ft/s.

† no significant recovery, possible non-equilibrium starting head.

‡ very non-linear head displacement, possible non-equilibrium starting head.

? no significant recovery, tentative drawdown analysis using Cooper et al. (1967) curve-matching method.

Values of estimated hydraulic conductivity from packer-testing 5 ft intervals in the Maquoketa Formation are shown in Table 1.

The Maquoketa Formation is almost 180 ft thick in well W-0 (WUWN PJ705). All 5 ft intervals containing obvious fractures based on the downhole video log were tested, but not all possible intervals in the formation were tested. A selection of apparently unfractured intervals was tested and in most cases, we were unable to displace the equilibrated head level (intervals L,M,Q,T,U). Of the unfractured intervals not tested, most showed smooth shale lithology on the downhole video, and hydraulic conductivity is assumed to be similar to that of intervals where no displacement occurred.

The distribution of hydraulic conductivity values with depth is shown in Figure 6b compared to values (Figure 6a) calculated from slug testing in a well at a different field site, Minooka Park, which were reported previously (Eaton and Bradbury 1998). The highest values of hydraulic conductivity are associated with the upper part of the Maquoketa Formation, where bedding-plane fractures associated with dolomitic beds were identified (Eaton et al. 2000, Eaton and Bradbury 1998). There is a broad range of values over almost 5 orders of magnitude, which results from the high transmissivity of these bedding-plane fractures, present in only some of the intervals tested.

There is also some difference between values (Table 1) estimated using the Hvorslev (1951) and Jacob (1963) methods for the same intervals (gray bars adjacent to black bars in Figure 6b). This discrepancy is probably due to the difference in scale tested by the two methods, the Hvorslev slug technique being relevant to areas closer to the well than the Jacob pumping technique. Pumping recovery values are generally lower, perhaps reflecting boundary limits of the highest transmissivity fractures.

Values of hydraulic conductivity estimated for interval G are the highest recorded (Table 1). This was not expected because other intervals have one or more fractures whose apparent aperture in the downhole video log was much larger (up to 1/16 inch) but had much lower conductivity values. These large-aperture features are therefore not well connected to a larger transmissive fracture network. On the other hand, the multiple fine fractures observed in interval G appear to be highly connected to a larger transmissive network. Unlike for any other interval, we were able to sustain a pumping rate of 5.3 GPM (20 L/min) for over an hour and a half. In addition, when interval G was pressure-tested or pumped, drawdown was observed in many intervals of other wells (Figures 2,7), particularly well WK1376 (WUWN PJ702).

Furthermore, when interval H was pumped, head level above the upper packer was drawn down also, suggesting some vertical interconnection among the numerous bedding-plane fractures. We applied a numerical inverse analysis technique (Piggott et al. 1995) to estimate fracture transmissivity from the various inter-well drawdown data. However, this method assumes a very simple model of a single fracture connecting pumping and observation wells, and our observation of drawdown at multiple levels indicates that there are probably numerous vertical fracture interconnections, all of which contribute to transmissivity. Better estimates of fracture transmissivity using more conventional analysis methods are presented in a later section of this report.

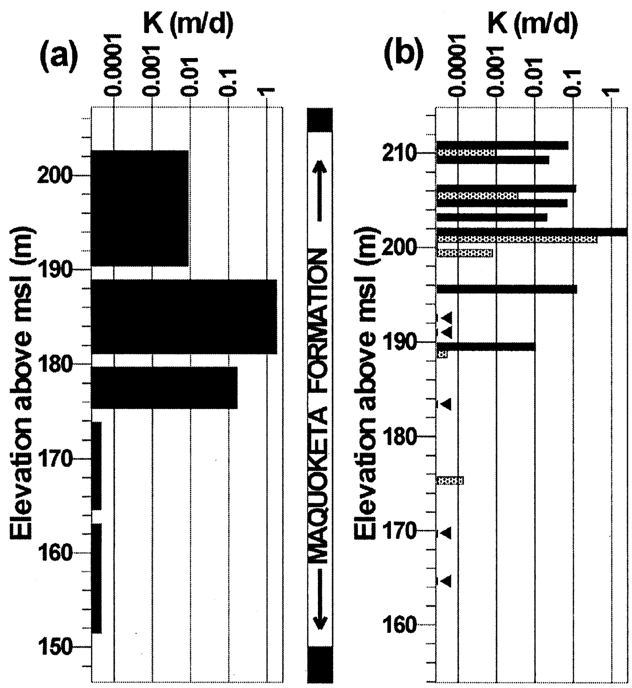


Figure 6. Detailed results of single-well hydraulic testing at two sites: (a) slug testing in WK1375 at Minooka Park, (b) straddle-packer testing in W-0 at DOT site. Triangles indicate values below testing limit. Gray bars indicate submersible pump used. Bar width indicates interval length tested.

V. GEOCHEMICAL AND ISOTOPIC ANALYSIS

A. Background

Prior to conducting tracer experiments, it is essential to determine the background chemistry in the groundwater system to be tested. In particular, one of the tracer substances we proposed for this study was bromide, which is naturally occurring, but usually present at only trace concentrations, which is why it is an attractive tracer. Only preliminary analysis of part of the geochemical and isotopic data we have collected in the Maquoketa Formation has been presented (Eaton and Bradbury 1998), and we omitted additional analysis from the previous report (Eaton et al. 2000) in order to present a more comprehensive overview here. For completeness, in addition to new data from the DOT field site (Table 2), data from the Minooka Park field site that was previously presented (Eaton and Bradbury 1998) is reproduced here (Table 3).

At the outset of our research effort, we considered the Maquoketa Formation to be a homogeneous low-conductivity aquitard. We expected to use hydrogeochemical sampling and vertical isotopic trends to evaluate slow vertical fluxes and hydraulic conductivity within this formation. However, the significant heterogeneity observed during field hydrogeologic testing (see also Eaton et al. 2000, Eaton and Bradbury 1998) causes interpretation of hydrogeochemical and isotopic data to be more complex. While a definitive analysis in a regional context is beyond the scope of this work, these data show possible effects of mixing by diffusion from the rock matrix into observed bedding-plane fractures and shed light on the relationship of the Maquoketa Formation to the overlying and underlying aquifer systems.

B. Methods

Groundwater samples were collected from each of the intervals in the multi-level packer systems within the Maquoketa Formation at the two field sites (Minooka Park and DOT). In addition, at the DOT field site (Figure 2), we collected water samples from a small surface-water pond, from well W-0 (WUWN PJ705) in the Maquoketa Formation, and the open intervals of wells W-1 (WUWN PJ703) and W-2 (WUWN PJ704) in the Silurian aquifer. Due to low sampling flow rates and depths of the sampled intervals, only limited purging was possible. We measured electrical conductivity, pH, dissolved oxygen, redox potential (Eh) and temperature in the field prior to filtering through 0.45 μm membrane filters. Samples were collected in polyethylene containers and preserved on ice for immediate delivery to the Wisconsin State Laboratory of Hygiene (Minooka Park samples) and EnChem, a private environmental testing laboratory (DOT samples) in Madison, Wisconsin, where they were analyzed for major ions using inductively-coupled plasma mass-spectroscopy (ICP-MS) and other standard methods.

We also sent samples to the Environmental Isotope Laboratory at the University of Waterloo, Ontario, Canada, for enriched tritium (^{3}H), deuterium (^{2}H) and oxygen-18 (^{18}O)

analyses. Tritium analyses used direct scintillation methods with a detection limit of 0.8 TU and uncertainty of +/-0.5 to +/-1.2, while deuterium and oxygen-18 are reported relative to VSMOW. Saturation indices of mineral phases and pCO_2 for major ion analyses were calculated using the geochemical speciation program PHREEQC (Parkhurst 1995) using the thermodynamic data provided in the associated database. Charge balances in all samples had less than 10% error.

C. Discussion of geochemical/isotope results

Samples consist of primarily calcium-magnesium-bicarbonate waters, with sodium and sulfate more abundant in the Maquoketa Formation, while chloride and magnesium are slightly elevated in the Silurian dolomite aquifer (Tables 2,3, Figure 7). Variation in electrical conductivity and concentrations of major ions with depth in the Maquoketa Formation was significant. Dissolved oxygen was generally low, below 1 mg/l, although levels up to 3 mg/l were recorded, possibly due to field error. Measured pH values ranged between 6.7 and 8.7. Redox potential (Eh), although not measured at the DOT site, followed a trend of increasingly reducing conditions with depth (-42 to –274 mV) at the Minooka Park site. Calculated saturation indices ($\log(IAP/K_{sp})$) show that groundwater in the Maquoketa Formation is oversaturated with $CO_2(g)$, and generally in equilibrium or slightly undersaturated with calcite and dolomite (Table 4). It is also undersaturated with respect to gypsum ($CaSO_4.2H_2O$) and the strontium-bearing minerals celestite ($SrSO_4$) and strontianite ($SrCO_3$). Although strontium is not usually considered a significant constituent in groundwater (Hem 1992), Nichols and McNall (1957) reported strontium levels exceeding 1 mg/l from wells in the Cambrian-Ordovician aquifer in southeastern Wisconsin. We measured levels up to 9.8 mg/l in water from the Maquoketa Formation and base of the Silurian dolomite.

High-sodium, low-chloride concentrations in the Maquoketa waters likely originate from cation exchange in the clay-rich shale (Na/Cl molar ratios 3-19). In contrast, high chloride concentrations in some samples (Figure 7) from the overlying Silurian aquifer probably result from road salt contamination (Na/Cl molar ratios 0.49-0.6). Trends shown on Figure 7 represent possible mixing lines between waters of different origin. We interpret Trend A to be a mixing line between Maquoketa-type sodium-bicarbonate porewater and calcium-magnesium-bicarbonate water that is more typical of the Silurian aquifer. Intermediate points, particularly evident on the cation part of Figure 7, may reflect mixing of water in fractures and less mobile porewater. The intermediate points along Trend B represent samples from the Maquoketa Formation at the DOT site. We infer a similar mixing origin for these data, whose deviation from Trend A is caused by relatively greater concentrations of sulfate than at Minooka Park.

Table 2. Geochemical Results for DOT field site

	Well WK1376 (WUWN PJ702)							W-2 (WUWN PJ704)			PJ703	PJ705	Pond
ID	None	Fig.2f	Fig.2e	Fig.2d	Fig.2c	Fig.2b	Fig.2a	Tube 3	Tube 2	Tube1	W-1	W-0	none
Interval	7	6	5	4	3	2	1	T3	T2	T1	open	open	
Depth, ft	20	180	210	240	280	325	370	61	105	130	75	250	0
elevation, m	256	207	198	189	177	163	149	244	230	222	239	186	262
sample date	8/3/00	8/1/00	8/1/00	8/1/00	8/1/00	8/1/00	8/1/00	8/2/00	8/2/00	8/2/00	8/2/00	8/3/00	8/3/00
D.O., mg/l	0.3	0.5	0.7	0.5	1.1	1	1	0.4	0.4	0.5	0.6	0.5	4
pH, units	NS	6.69	6.9	6.96	6.92	6.91	7.1	6.67	6.76	NS	7.36	8.09	NS
Eh, mV	NS	NS	NS	NS	NS	NS	NS	NS	NS	NS	NS	NS	NS
Cond. mS/cm	0.48	0.62	0.64	0.98	0.78	0.68	0.65	2.01	1.96	0.62	1.47	0.84	0.31
Samp. T °C	16.7	18.8	21.5	19.8	25.8	23.3	25.2	15.4	16.3	18.5	17.1	21.3	26.6
In-situ T °C	9	9.7	9.7	9.7	9.9	10.6	10.6	NS	NS	NS	NS	NS	NS
Ca (mg/l)	40	81	78	71	67	68	82	160	150	80	80	42	40
Mg (mg/l)	39	38	36	28	25	31	36	83	82	36	72	16	17
Na (mg/l)	5.1	7	20	110	120	37	8.8	130	140	6.1	100	120	0.73
K (mg/l)	2.2	2.2	4	9.3	10	6.7	7.2	2.6	2.9	3.6	2.5	16	2.5
Fe (mg/l)	4.8	0.27	0.46	0.3	0.25	0.48	0.26	0.6	0.41	0.035	20	0.081	0.17
Mn (mg/l)	0.11	0.0058	0.013	0.0088	0.034	0.034	0.0088	0.023	0.044	0.031	0.52	0.2	0.1
As (mg/l)	0.0008	0.011	0.0047	0.00068	0.00055	0.0012	0.0071	0.015	0.012	0.0059	ND	0.00017	0.001
Br (mg/l)	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND
Sr (mg/l)	3.7	7.9	7.6	7.5	6.6	6.7	7.6	2.7	2.8	9.8	1.8	6.7	0.2
Zn (mg/l)	0.0095	0.0078	0.0085	0.0036	0.0039	0.0035	0.004	0.0058	0.0072	0.0091	0.032	0.0047	0.0082
Alk* (mg/l)	240	340	340	350	400	360	330	350	370	340	240	210	160
HCO_3 (mg/l)	284.8	404.6	404.2	417.9	477.7	429.1	392.1	408.7	432.6	404.2	275.8	247.1	191.5
Cl (mg/l)	16	17	16	16	18	18	18	380	360	8.8	310	9.6	2.3
SO_4 (mg/l)	ND	26	27	160	110	5.1	21	48	52	27	20	180	1.6
NO_3 (mg/l)	ND	0.1	ND	ND	ND	ND	ND	ND	0.21	0.11	ND	0.11	0.08
S^- (mg/l)	0.67	0.4	0.8	0.53	1.1	0.4	NS	0.8	0.67	0.53	0.4	0.67	NS
Cu (mg/l)	0.0016	0.0011	0.0014	ND	ND	ND	0.0011	0.0016	0.0013	0.0014	0.0013	ND	0.002

* Alkalinity as $CaCO_3$, ND: not detected, NS: not sampled

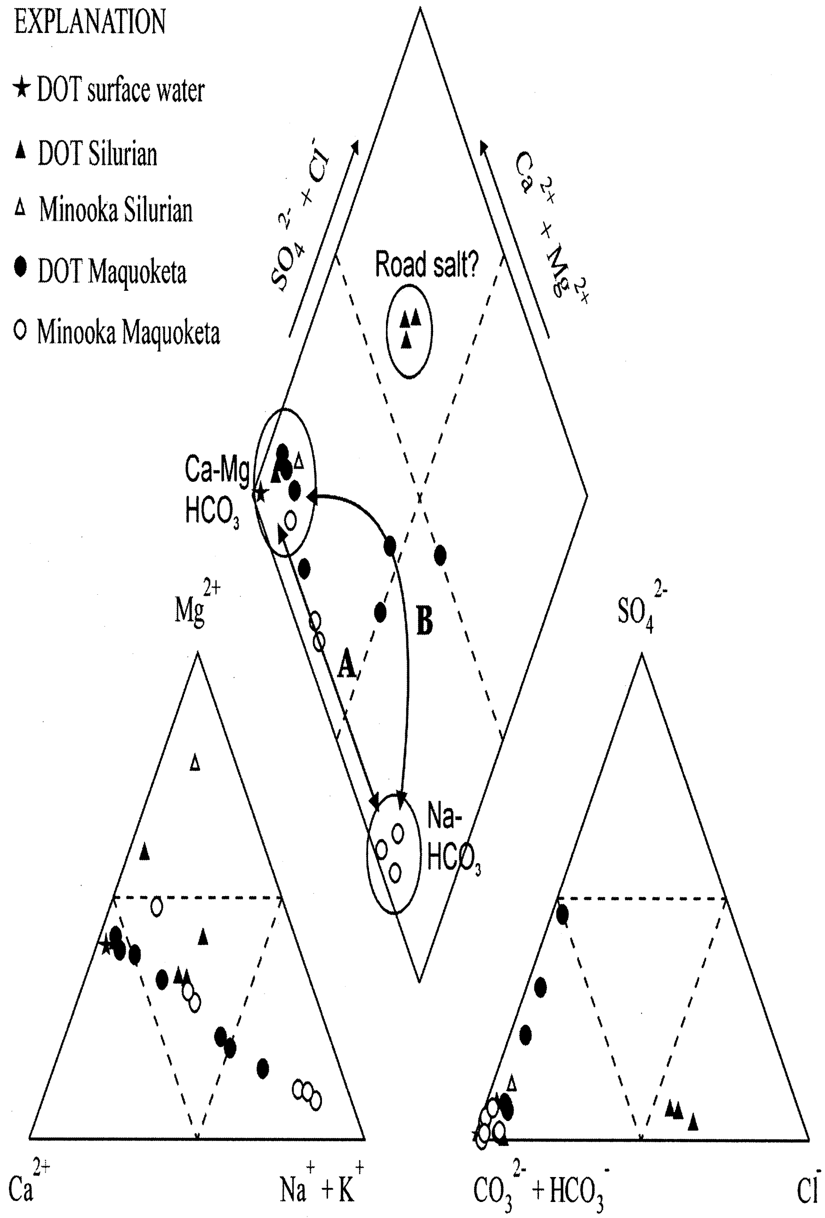


Figure 7. Tri-linear diagram showing hydrogeochemistry of water samples collected at DOT and Minooka Park field sites. Mixing trends A and B are inferred between end-member waters characteristic of fractures and rock matrix, as discussed in text.

Table 3. Geochemical Results for Minooka Park field site

	Well WK1375 (WUWN PJ701)						
Interval	**7**	**6**	**5**	**4**	**3**	**2**	**1**
Depth, ft	20	271.5	307	322.5	358	402.5	422
elevation, m	268	192	181	176	165	152	146
sample date	5/4/98	4/29/98	4/29/98	4/30/98	5/1/98	7/29/98	7/27/98
D.O., mg/l	2	2	0.6	2.7	3	0.1	0.8
PH, units	8.72	7.37	7.53	7.37	8.06	8.36	8.45
Eh, mV	20.6	-41.9	-53.2	-107.9	-87.1	-236	-274.3
Cond. mS/cm	0.344	0.673	0.869	0.673	0.484	0.83	0.684
Samp. T °C	12.9	12.5	9.7	11.7	13.2	15.5	14.8
In-situ T °C	NS	9.764	9.997	9.842	10.2	10.86	10.77
Ca (mg/l)	9.1	52	28	56	45	20	17
Mg (mg/l)	34	24	12	28	34	8.9	8.3
Na (mg/l)	6.4	54	160	52	17	170	120
K (mg/l)	3.7	5.4	9.8	5.1	3	8.8	7.3
Fe (mg/l)	0	0.16	0.33	0.19	0.28	0.04	0.08
Mn (mg/l)	0.087	0.0047	0.004	0.0096	0.005	0.0035	0.0059
As (mg/l)	0.0007	0.0007	0	0.0008	0.0008	ND	ND
Br (mg/l)	0.15	0.14	0.18	0.14	0.14	0.16	NS
Sr (mg/l)	0.058	2.768	2.232	1.912	1.743	NS	NS
Zn (mg/l)	ND	ND	0.009	ND	ND	ND	ND
Alk* (mg/l)	151	340	467	371	329	446	337
HCO_3 (mg/l)	166.8	405.7	560.6	441.9	329	521.4	392.5
Cl (mg/l)	6.95	3.45	6.98	3.58	4.5	8.8	16.4
SO_4 (mg/l)	17.6	12	0.38	16.9	19.1	6.7	7
NO_3 (mg/l)	ND	ND	ND	ND	ND	ND	ND
S^- (mg/l)	0.1	0.1	0.1	0.3	NS	NS	NS
Cu (mg/l)	ND	ND	ND	ND	NS	ND	ND

Alkalinity as $CaCO_3$, estimated from EPM balance for Interval 3; ND: not detected; NS: not sampled

Oxygen-18 (^{18}O) and deuterium (^{2}H) isotope data plot along a local meteoric water-line presented by Simpkins (1989) and show little evidence of fractionation by evaporative, biological or geothermal activity (Figure 8). Although data from the DOT site are clustered which may indicate mixing, some data from Minooka Park with more depleted ^{18}O signatures probably represent significantly older groundwater recharged during a cooler, Pleistocene climate (Eaton and Bradbury 1998). Tritiated water (Table 4) was found in the Silurian dolomite aquifer (6.1-11.9 TU) and the pond sampled at the DOT site (15.2 TU) but not within the Maquoketa Formation (<1.4 TU), indicating that groundwater within the confining unit was probably recharged prior to the peak of atmospheric tritium in the mid-1960s (Hendry 1988). These relatively old groundwater ages based on tritium data are consistent with the oxygen isotope data analysis.

Table 4. Saturation indices and isotope data for DOT and Minooka Park

Sampled interval‡	Elev. (m. amsl)	Sample source	pCO_2 (bars)	Calcite	Dolomite	Gypsum	Celestite	Strontianite	Tritium* (TU)	del 18O (permil SMOW)	del 2H (permil SMOW)
Minooka Park site											
7	268	Silurian	-3.57	0.29	1.33	-3.15	-3.63	-1.38	6.1	-8.58	-64.37
6	192	Maquoketa	-1.84	0.04	-0.09	-2.63	-2.21	-0.71	< 0.8	-10.38	-76.35
5	181	Maquoketa	-1.87	0.00	-0.25	-4.39	-3.78	-0.55	< 0.8	-12.44	-94.76
4	176	Maquoketa	-1.81	0.08	0.02	-2.47	-2.23	-0.85	1.2	-9.69	-71.66
3	165	Maquoketa	-2.64	0.57	1.20	-2.49	-2.20	-0.32	NS	NS	NS
2	152	Maquoketa	-2.71	0.73	1.34	-3.31	-7.01	-4.17	†	†	†
1	146	Sinnipee	-2.92	0.64	1.19	-3.32	-6.48	-3.71	†	†	†
DOT field site											
Pond	262	surface	-1.71	-0.51	-1.01	-3.55	-4.14	-2.36	15.2	-5.56	-35.69
W-2T3	244	Silurian	-1.18	-0.33	-0.85	-1.77	-1.84	-1.56	11.9	-9	-64.62
W-2T2	230	Silurian	-1.24	-0.25	-0.65	-1.76	-1.79	-1.43	11.8	-8.88	-63.22
W-2T1	222	Silurian	-1.49	-0.22	-0.68	-2.13	-1.34	-0.58	< 0.8	-9.04	-62.62
7	256	Silurian	-1.64	-0.63	-1.17	-4.79	-4.12	-1.12	NS	NS	NS
6(f)	207	Maquoketa	-1.18	-0.51	-1.24	-2.15	-1.46	-0.98	< 0.8	-8.99	-62.19
5(e)	198	Maquoketa	-1.39	-0.32	-0.86	-2.15	-1.45	-0.79	< 0.8	-9.01	-63.83
4(d)	189	Maquoketa	-1.44	-0.34	-0.97	-1.47	-0.74	-0.77	< 0.8	-9.17	-66.99
3(c)	177	Maquoketa	-1.34	-0.33	-0.97	-1.64	-0.94	-0.8	1.3	-9.02	-64.13
2(b)	163	Maquoketa	-1.37	-0.32	-0.85	-2.91	-2.22	-0.79	1.4	-8.9	-66.48
1(a)	149	Sinnipee	-1.6	-0.09	-0.41	-2.24	-1.57	-0.59	NS	NS	NS
W-1	N/A	Silurian	-2.8	-0.07	-0.08	-2.37	-2.31	-1.17	9.7	-9.24	-63.9
W-0	N/A	Maquoketa	-2.8	0.34	0.36	-1.59	-0.68	0.09	NS	NS	NS

NS: not sampled, partially due to difficulties pumping adequate volumes from deep intervals.

* Uncertainty from +/-0.5 for lowest values to +/-1.2 for highest values.

† Sampled, but only after partial packer failure, unreliable data.

‡ Letters in parentheses correspond to locations in Figure 2

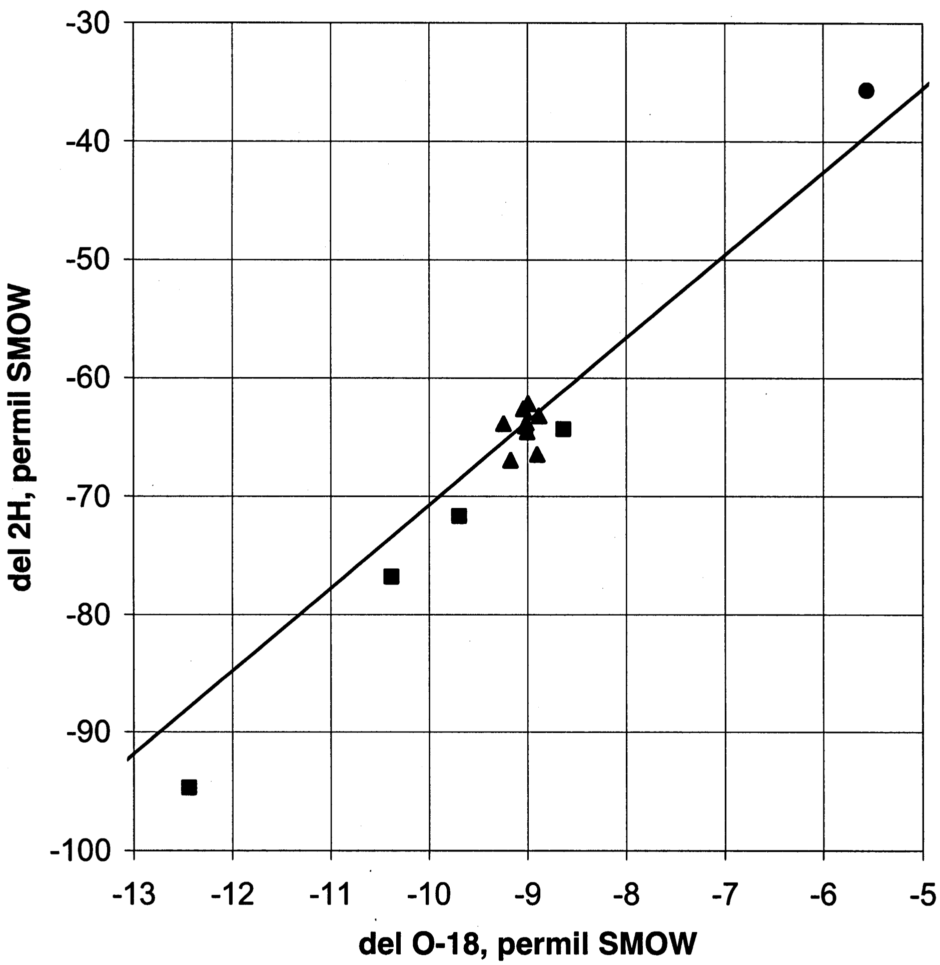


Figure 8. Oxygen-18 (^{18}O) and deuterium (^{2}H) data for DOT (triangles) and Minooka Park (squares) sites compared to local meteoric water-line for southeastern Wisconsin (Simpkins 1989). Circle represents surface water at DOT site.

Weaver and Bahr (1991) studied hydrogeochemistry of the Cambrian-Ordovician aquifer in eastern Wisconsin and suggested that the Maquoketa Formation could be a source of sulfate by gypsum dissolution. Siegel (1990) and others had earlier proposed, based on sulfur isotope data, that Na-SO_4-Cl water in the confined Cambrian-Ordovician aquifer is due to a reversed potentiometric gradient from the Michigan Basin during Pleistocene time. Maquoketa water samples containing elevated sulfate, sodium and strontium, which are constituents also found in the underlying aquifer, combined with undersaturated solubility conditions for gypsum, strontianite and celestite, supports the idea that the Maquoketa Formation may be a local source of solutes to the Cambrian-Ordovician aquifer system. An alternative possible source of sulfate in water of the Maquoketa Formation could be the oxidation of sulfide minerals prominent in rock core samples collected at both sites (Eaton 2002, Eaton and Bradbury 1998). However, oxidation of rock minerals seems unlikely given the apparent reducing conditions indicated by redox potential (Eh) measurements.

The geochemical mechanism for the high concentration of strontium in Maquoketa waters remains uncertain. Although it is possible that strontium may occur as a pure phase mineral (celestite or strontianite) in the Maquoketa Formation; as a trace element, it is more likely to form a solid-phase mineralogical association with calcium carbonate or dolomite. In this case, co-precipitation or co-dissolution processes (Bruno et al. 1998) may account for the aqueous concentrations. Alternatively, strontium concentrations could be governed by sorption-desorption processes from clay minerals in the shale lithology (Grutter et al. 1994). Further work, particularly on the mineralogy of the Maquoketa Formation, would be needed for confirmation of these processes.

VI. GROUNDWATER TRACER TESTING

A. Introduction

The hydraulic testing data reported above, and in earlier reports (Eaton et al. 2000, Eaton and Bradbury 1998) indicate that, although the rock matrix of the Maquoketa Formation has extremely low hydraulic conductivity, it contains a sparse network of fractures which are hydraulically connected to the overlying Silurian dolomite aquifer. We observed bedding-plane fractures directly from downhole geophysical logs, and inferred vertical interconnections between these fractures based on inter-well hydraulic responses to testing. Unambiguous evidence of this fracture system, and further characterization of it can only be obtained by the transfer of mass, i.e., a tracer substance, from one point to another in the fracture system. The chemical analyses reported above indicate that background levels of bromide are at trace to undetectable levels, making bromide a possibility for a tracer, as proposed. Any tracer in fractured rock is subject to considerable dilution, so we also proposed using Rhodamine WT, a fluorescent dye, which is detectable at much lower concentrations ($<1 \mu g/l$) using field fluorometry.

Tracer experiments are particularly uncertain in fractured rock because of the unpredictability of natural hydraulic head gradients (Muldoon and Bradbury 1998, NRC 1996). The well configuration at the DOT field site (Figure 2) is well suited to tracer experiments because we can attempt to control the head gradient by inducing flow from an injection point toward recovery points. In such an induced-gradient or convergent tracer test, we planned to pump well W-0 (PJ705), which is open to only the Maquoketa Formation, and induce flow from well W-2 (PJ704) where we proposed to inject tracer into the overlying Silurian dolomite. In this dipole flow configuration, the multi-level monitoring well WK1376 (PJ702), open to the Maquoketa Formation, is situated midway between the injection and pumping points and is well suited to intercepting potential tracer migrating along fracture paths. By monitoring and pumping the ports in this well, we hoped to detect and recover tracer.

In a letter report (7/7/00) to Richard Roth and Steve Karklins, Wisconsin DNR, we requested formal approval to conduct groundwater tracer tests using bromide and Rhodamine WT. We mapped the local water-table in the vicinity of the DOT field site and found that no private domestic wells were present in a downgradient direction from the proposed injection location at well W-2 (PJ704). We also estimated decreases in concentration due to dilution and dispersion, and determined that at the anticipated tracer injection concentrations, concentrations of bromide would decrease to 2 mg/l and 0.002 mg/l at distances of 100 ft and 1000 ft respectively from the injection well. For Rhodamine WT, we expected injected concentrations to decrease to 12 $\mu g/l$ at 100 ft and 0.01 $\mu g/l$ at 1000 ft. For these reasons, we demonstrated that use of the proposed tracers presented little if any risk to water quality or local drinking water wells, and obtained a conditional approval (7/28/00) from the DNR for use of these tracer substances.

B. Methods

One of the biggest operational difficulties in conducting groundwater tracer experiments is the logistics involved in sampling and analysis (Muldoon and Bradbury 1998). The two major aspects to this problem are: what methods are used for tracer detection, and how long after tracer injection is it feasible to monitor for breakthrough elsewhere in the groundwater system? Many workers have collected multiple small groundwater samples over long periods, for shipment to laboratories where they are analyzed for tracer concentrations. The alternative that we planned to use is to collect samples for periodic analysis in the field, even though field instruments such as specific ion electrodes often have higher detection limits than laboratory methods.

One of the reasons for proposing the use of fluorescent tracer dyes is that very low concentrations are detectable in the field using a portable field fluorometer. We borrowed a Turner Designs Model 10 field fluorometer from the University of Wisconsin-Madison Trout Lake Field Station for this purpose. We constructed a valve system to successively run samples from different monitoring ports in well WK1376 (PJ702) through the instrument while backflushing with deionized water between samples. Unfortunately, the instrument has an analog gauge indicating relative fluorescence, but with the assistance of technicians at the University of Wisconsin-Madison Dept. of Geology and Geophysics, we developed a telemetry link to a digital datalogger for continuous recording of instrument readings.

Even using field methods (specific ion electrode and fluorometry) for tracer detection, the second aspect to the logistical problem remains: how long to monitor for tracer breakthrough? Although inter-well response times to packer testing and earlier hydraulic testing (Eaton et al. 2000) were very short, less than one minute, tracer transport between injection and monitoring points could take much longer. Muldoon and Bradbury (1998) report monitoring times ranging from hours to over a day in fractured Silurian dolomite at a site in Door County, WI. However, their field site was in a shallow aquifer with distances between wells of about 30 ft (<10m), and little difference in relative elevation between tracer injection and detection points. We hoped to induce tracer transport over 30 ft (10m) horizontally, but from 50 to 150 ft (15-46 m) vertically between the Silurian aquifer and the underlying Maquoketa Formation (Figure 2). Tracer transport to the observation points at well WK1376 (PJ702) could take a day or more. Therefore, it is important to obtain some estimate of a timeframe in which tracer breakthrough could be detectable at this well.

Fortunately, the pressure transducers in place to monitor head in each of the deep multi-level well intervals (Figure 2) also contain thermistors which record in-situ temperature. Prior monitoring of these temperatures showed very little variation over time – less than 0.1 ^{o}C – even while pumping. We decided to use temperature as a tracer by introducing hot water at the tracer injection points while pumping well W-0 (PJ705), and monitoring the instrumentation at well WK1376 (PJ702) for any increase in temperature. Such in-situ detection would definitively indicate tracer breakthrough and provide an estimate of when chemical tracers might be expected to arrive. Muldoon and

Bradbury (1998) also attempted using temperature as a tracer, but found that natural fluctuations in temperature due to precipitation and pumping were significant at their site.

We therefore installed a submersible pump in well W-0 (PJ705), and made an initial attempt at hot-water tracer injection in the lowest Silurian interval in well W-2 (Figure 2). After 4-5 hours of pumping at a rate of approximately 20 L/min (5 GPM), causing a drawdown at well W-0 of over 100 ft, we began injection of previously pumped groundwater heated to a temperature of 60-65 °C over a camp stove. However, injection of heated groundwater was limited by the peristaltic pump rate and the narrow (3/8 inch) tubing leading to the injection interval. At an estimated injection rate of just under 0.2 L/min, we maintained hot-water tracer injection for another 13 hours despite occasional pump and stove problems. While continuing to pump well W-0, we then also pumped the individual intervals of well WK1376 (PJ702) for 1.5 hours to try to capture some of the 150 L of heated tracer we had already injected in well W-2 (PJ704). No definitive changes in temperature at intervals in well WK1376 were ever detected.

One drawback to using heated water as a tracer is that it eventually cools to ambient groundwater temperatures, and the slow injection rate using a long narrow tube accelerated this process. So we improved our technique by increasing the rate of injection of heated tracer and the temperature of injection. We also installed a larger submersible pump in well W-0. Another well (W-1) at the DOT field site is more suited to large injection rates because it has a 50 ft open interval in the Silurian aquifer (Figure 2). Using a kerosene-fueled pressurized-steam cleaner borrowed from the USGS, we were able to heat pumped groundwater to a temperature near boiling (100 °C) before injecting it into the open interval in well W-1. Prior to initiating this tracer injection, we pumped well W-0 at a rate of up to 30 L/min (8 GPM) for a period of 20 hours, causing drawdown of over 140 ft., and continued this rate of pumping for the duration of the experiment. Simulation of drawdown during these tests and comparison with measured drawdown is shown in Eaton (2002).

We monitored temperature in the injection well (W-1) and while we injected heated tracer at the base of the open interval (Figure 2), we measured water temperatures up to 39 °C near the top of the water column in the well. By diverting pumped groundwater from the pumped well W-0 to the steam cleaner, then to the injection well (W-1), we maintained an injection rate of 10-15 L/min (3-4 GPM) for 25 hours. Near the end of this time, we pumped the individual ports of well WK1376 for 4.5 hours to try to capture any of the 15000 L (4000 gal.) of heated water we had injected at well W-1. No significant changes in temperature at ports in well WK1376 or in any other multi-level wells were ever detected, during the experiment or afterwards.

C. Discussion and analysis of results

Failure to detect any temperature tracer breakthrough was disappointing, especially given the sensitivity of the in-situ thermistors, the constant background temperature, and the relatively much higher temperature of the injected tracer. However tracer testing in fractured rock is highly uncertain because tracer transport along fractures is dependent not just on head gradients but also on the morphology of fracture apertures. Fracture apertures vary along the plane of the fracture such that channels of larger aperture are formed which conduct most of the flow (NRC 1996). Unless a monitoring interval intersects the plane of a fracture in such a flow channel, it is quite possible for tracer transport pathways to bypass the monitoring interval. Our tracer experiment design attempted to avoid this situation by pumping at the monitoring well (WK1376) to capture any nearby tracer. However, variations in hydraulic gradient within these fracture planes of variable aperture are difficult to control through pumping, and tracer transport from one point to another in fractured rock depends on continuity of the hydraulic gradient along a very tortuous path.

A hydraulic connection between fractures in the Silurian aquifer and the Maquoketa Formation was demonstrated during pumping of the Silurian aquifer (Eaton et al. 2000). This was confirmed during the second tracer experiment when a slight increase in head occurred in the ports of WK1376 when the injection of the temperature tracer began in the open interval of well W-1. However, the pumping to establish the dipole flow configuration was apparently not sufficient to generate the appropriate gradients within the fracture system to induce tracer transport from the injection well to the observation well. Cooling of the temperature tracer and density-driven flow may also be factors, but given the length of the tests and volumes injected, these are unlikely to be the only explanations. It is probable that despite the pumping from the Maquoketa Formation, the temperature tracer simply migrated according to the much stronger flow within the Silurian aquifer towards the nearby wetland. In any case, planned chemical tracer experiments seemed unlikely to succeed under these circumstances, so we did not proceed with them.

Although no elevated temperatures indicating tracer breakthrough were detected, the long-term pumping in well W-0 caused significant drawdown in other monitoring wells, as shown schematically in Figure 2. This drawdown occurred in the multi-level packer well WK1376 and also in intervals j and k of well W-1, where during the second tracer experiment 10-15 L/min was being injected in the Silurian aquifer at this same well. Magnitudes of observed drawdown ranged from 5.9 to 46 ft (1.8-14 m). We analyzed these long-term drawdowns, and in two cases recovery data after pumping ceased. Figure 9 shows head change over time in observation wells for the two tracer tests, and Table 5 gives parameters and resulting transmissivity and storage values from Cooper-Jacob (1946) analysis. The Cooper-Jacob equations are:

$$\mathbf{T = (2.3Q)/(4p?s)} \qquad \text{and} \qquad \mathbf{S = 2.25Tt_o/r^2}$$

where $\mathbf{Q}$ = pumping rate (m^3/s), **?s** is extrapolated drawdown (m) over one log cycle, $\mathbf{t_o}$ (s) is projected intercept on abscissa, and **r** (m) is distance between pumping and observation well. Resulting values from this analysis are in the form of transmissivity **T** (m^2/d), which is not directly comparable to hydraulic conductivity **K** (m/d). Drawdown in observation wells in the Maquoketa Formation could not occur except

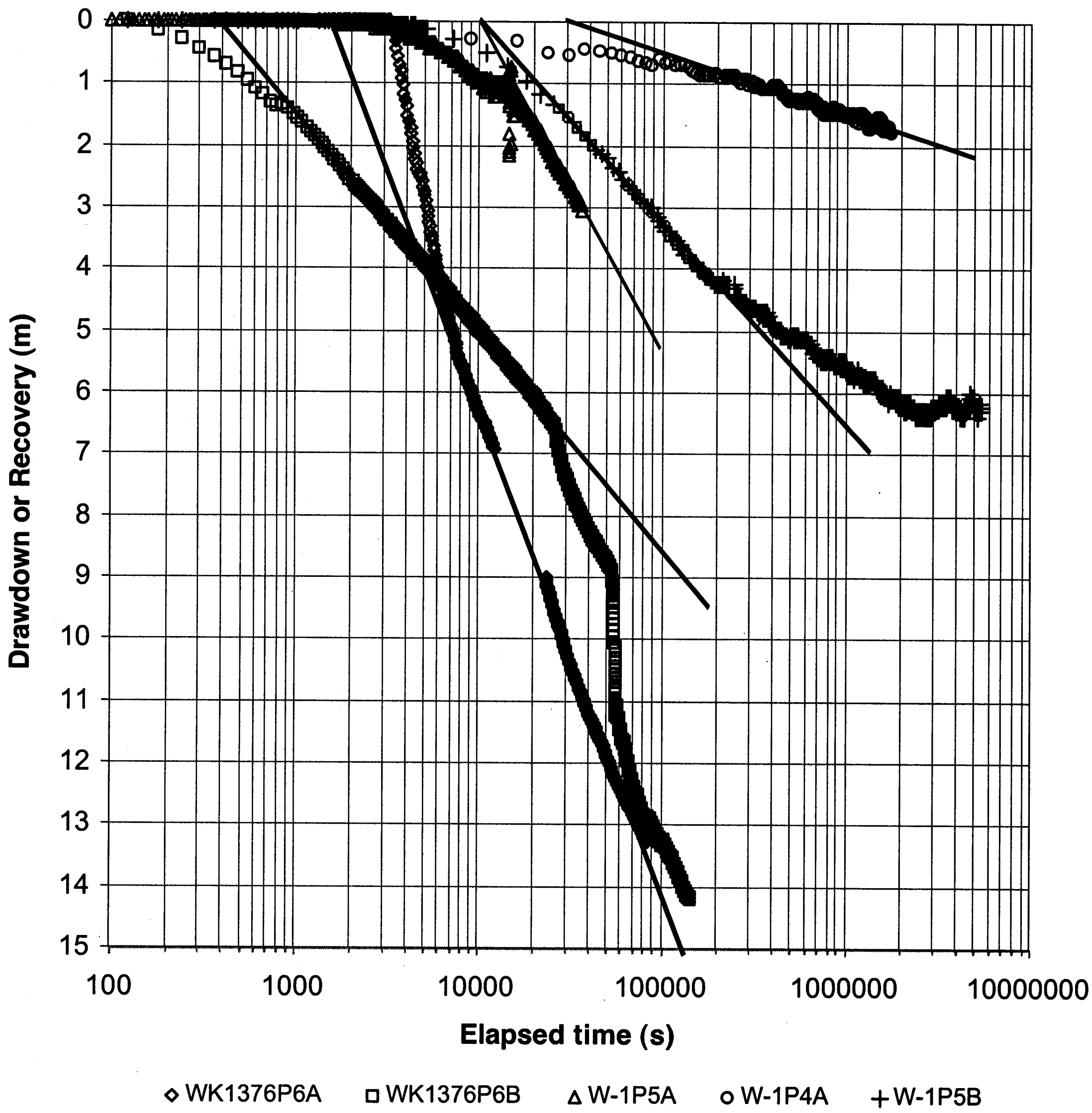


Figure 9. Head change in observation wells due to tracer test pumping in well W-0. Straight lines illustrate Cooper-Jacob analysis, actual values derived from separate plots and shown in Table 5. In legend, well IDs ending in ----A are from test on 11/1-2/00, well IDs ending in ----B are from test on 11/29-30/00. Data gap/disruption at s=15000 due to brief pump failure.

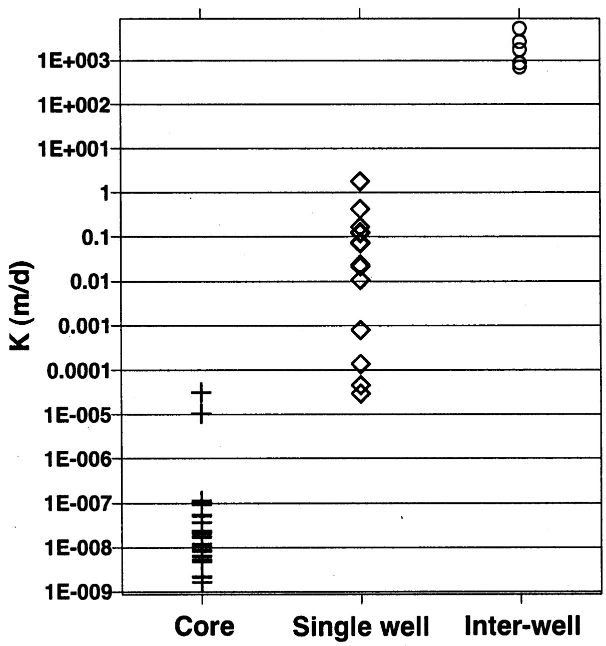


Figure 10. Range in hydraulic conductivity estimated from rock-core testing, slug and straddle-packer testing, and inter-well pumping for tracer tests. Fracture aperture b=0.001 m used to calculate K from fracture transmissivity for inter-well pumping.

through fractures, therefore transmissivity data from inter-well pumping pertains only to fracture flow. The number of individual fracture paths and their apertures which account for flow between wells is unknown. Therefore, a conservative assumption for converting **T** to **K** is to assume one fracture path and divide by the maximum aperture (b) observed in downhole video logging (Eaton et al. 2000) which is 0.001 m.

Hydraulic conductivity (Figure 10) estimated from transmissivity values contrasts sharply with previously measured conductivity values based on slug and straddle-packer testing, and especially laboratory rock-core testing. No apparent lag time was observed in observation well drawdown at different distances in response to pumping for tracer tests.

Table 5. Parameters and estimated fracture transmissivity from Cooper-Jacob analysis

Well data (refer to Fig.9)	Q (m^3/s)	? s (m)	t_o (s)	r (m)	T (m^2/s)	T(m^2/d)	S
WK1376P6A (f)	$3.3x10^{-4}$	7.5	1600	16	$8x10^{-6}$	0.7	$1x10^{-4}$
WK1376P6B (f)	$3.3x10^{-4}$	3.4	350	16	$2x10^{-5}$	2	$6x10^{-5}$
W-1P5A (k)	$3.3x10^{-4}$	5.0	9000	26	$1x10^{-5}$	1	$3x10^{-4}$
W-1P4A* (j)	$3.3x10^{-4}$	1.0	30000	26	$6x10^{-5}$	5	$6x10^{-3}$
W-1P5B* (k)	$5.3x10^{-4}$	3.4	10000	26	$3x10^{-5}$	2.6	$1x10^{-5}$

* Values calculated from plot of pumping recovery rather than drawdown due to data recording gaps. Letters in parentheses refer to intervals in Figure 2.

VII. CONCLUSIONS AND IMPLICATIONS

Research involving laboratory and field hydraulic testing, of which this report describes only a part (see also Eaton 2002, Eaton et al. 2000, Eaton and Bradbury 1998), has shown that the hydrogeology of the Maquoketa Formation is determined by significant geologic heterogeneity. This heterogeneity consists primarily of transmissive bedding-plane fractures, which are associated with interbedded shale and dolomite facies in the upper part of the formation. Hydraulic conductivity of the rock matrix measured in core samples was previously found (Eaton et al. 2000) to be uniformly very low ($3x10^{-14}$ to $3x10^{-12}$ ft/s). In contrast, the variable frequency and transmissivity of fractures causes bulk horizontal hydraulic conductivity measured at field scale, using slug or straddle-packer testing, to range over 5 orders of magnitude from $1x10^{-9}$ to $1x10^{-4}$ ft/s ($2x10^{-5}$ to >2 m/d). The continuity within and hydraulic connections among the transmissive bedding-plane fractures are of interest because most flow occurs in the interconnected fracture network, which would also determine potential contaminant transport.

Short-interval straddle-packer testing described in this report confirms that high conductivity zones result from the presence of transmissive fractures, which are not present in the uniform shale lithology at the base of the formation. Packer testing at 5 ft intervals based on visual assessment of fractures in a downhole video, showed that not all apparent large-aperture fractures are transmissive – the most conductive interval contained multiple small-aperture fractures. In contrast to other intervals tested, many of which have conductivities below the testing limit of our equipment ($1x10^{-9}$ ft/s), testing this conductive interval caused drawdown at multiple levels in other wells over 45 ft (15 m) away. This indicates that a sparse, laterally extensive and transmissive fracture network is present within the Maquoketa Formation and vertical interconnections may be locally important.

Analysis of groundwater samples shows that variation in electrical conductivity and concentrations of major ions with depth in the Maquoketa Formation are significant. Although it is usually a trace element, strontium concentrations in the mg/L range were found that may be related to reported incidence in the underlying Cambrian-Ordovician aquifer. Oxygen and deuterium isotope data, and the lack of tritium, indicate that groundwater in the Maquoketa Formation may be quite old (Pleistocene). Major ion chemistry data plot along mixing trends from a Ca-Mg-HCO_3 type to a Na-HCO_3 type, which are interpreted to be characteristic of water in fractures and water in the rock pores, respectively. These data are consistent with a three-dimensional flow system in the aquitard, where solutes from the rock matrix diffuse into the flow in the transmissive fracture network.

Prior hydraulic testing had suggested significant vertical interconnections between the Silurian dolomite aquifer and Maquoketa Formation. So two dipole, convergent flow, groundwater tracer experiments were carried out, using temperature as a tracer. No change in temperature was ever observed in a multi-level monitoring well in the Maquoketa Formation between the pumping and injection wells despite significant head changes observed in this and another multi-level well. We attribute this failure of tracer tests to our inability to generate the necessary head gradients along tortuous variable-aperture fracture pathways between the Silurian aquifer and Maquoketa Formation.

However, the inter-well drawdowns due to tracer test pumping were analyzed to estimate transmissivity of the interconnected fracture network. Values of five analyses of transmissivity based on drawdown or recovery in observation wells during the two tests were within a range of 7E-1 m^2/d to 5 m^2/d (7.5 ft^2/d – 54 ft^2/d), and storage values were between 1E-5 and 6E-3. These values represent estimates of the properties of the highest transmissivity features in the Maquoketa Formation: the fractures, which coincide primarily with bedding-planes. A comparison of these transmissivity data to the already broad range of hydraulic conductivity between laboratory (Eaton et al. 2000) and field data values illustrates the heterogeneity of this sparsely fractured, low-conductivity aquitard.

The implications of the transmissive fractures on the hydrogeology of the Maquoketa Formation depend not just on local pumping stresses, but on larger boundary conditions, the analysis of which is beyond the scope of this report. The rapid head responses in the Maquoketa Formation to pumping in the Silurian aquifer (Eaton et al. 2000) which prompted this study suggest that fractures may be locally interconnected in the vertical dimension, but geochemical and isotopic data and the failure of tracer tests indicate that gradients to cause major vertical flow are not present. Additional evidence for this is provided by analysis of the observed vertical head distribution (Eaton et al. 2000, Eaton and Bradbury 1998) and regional boundaries using numerical modeling (Eaton 2002) which is not presented here.

The lateral continuity of the transmissive bedding-plane fractures and limited vertical interconnections increase the bulk hydraulic conductivity of the upper dolomitic part of the Maquoketa Formation. The current transient head field in the Maquoketa Formation coupled with the uniformly low hydraulic conductivity of the base of the formation account for the hydraulic effectiveness of this aquitard. However, the top of the Maquoketa Formation cannot be assumed to be virtually impermeable to contaminants whose migration is not dependent on head gradients. For instance, a dense non-aqueous-phase liquid (DNAPL) contaminant mass at the base of the Silurian aquifer would probably penetrate the fracture system in the upper Maquoketa Formation by gravity flow. Localized pumping near the base of the Silurian aquifer could induce flow from and unpredictable gradients in the fractured upper Maquoketa Formation.

VIII. REFERENCES CITED

Bradbury, K.R., T.W. Rayne, M.A. Muldoon, and P.D. Roffers. 1998. *Application of a discrete fracture flow model for wellhead protection at Sturgeon Bay, Wisconsin.* Wisconsin Geological and Natural History Survey Open-File Report 92-2, 84 p.

Bruno J., L. Duro, J. de Pablo, I. Casas, C. Ayora, J. Delgado, M.J. Gimeno, J. Pena, C. Linklater, L. Perez del Villar, and P. Gomez. 1998. Estimation of the concentrations of trace metals in natural systems; the application of codissolution and coprecipitation approaches to El Berrocal (Spain) and Pocos de Caldas (Brazil). *Chemical Geology* 151(1-4), 277-291.

Butler, J.J., Jr. 1998. *The design, performance and analysis of slug tests*, Lewis Publishers, Boca Raton, FL, 252 p.

Cooper, H.H., Jr., J.D. Bredehoeft and I.S. Papadopulos. 1967. Response of a finite-diameter well to an instantaneous charge of water. *Water Resources Research* 3(1), 263-269.

Cooper, H.H., Jr., and C.E. Jacob. 1946. A generalized graphical method for evaluating formation constants and summarizing well-field history. *Transactions of the American Geophysical Union* 27(4), 526-534.

Eaton, T.T. 2002. *Fracture heterogeneity and hydrogeology of the Maquoketa aquitard, southeastern Wisconsin.* unpublished Ph.D. dissertation (Geology and Geophysics), University of Wisconsin-Madison, 200 p.

Eaton, T.T., D.J. Hart, K.R. Bradbury and H.F. Wang. 2000. *Hydraulic conductivity and specific storage of the Maquoketa shale*. Final Report for the University of Wisconsin Water Resources Institute, Wisconsin Geological and Natural History Survey Open-File Report 00-01, 31 p.

Eaton, T.T. and K.R. Bradbury. 1998. *Evaluation of the confining properties of the Maquoketa Formation in the SEWRPC area of southeastern Wisconsin*, Final Report for the Wisconsin Dept. of Natural Resources, Wisconsin Geological and Natural History Survey Open-File Report 98-11, 20 p.

Grutter, A., H.R.Vongunten, E. Rossler and R. Keil. 1994. Sorption of strontium on unconsolidated glaciofluvial deposits and clay minerals – mutual interference of cesium, strontium and barium. *Radiochimica Acta* 64 (3-4): 247-252.

Hem, J.D. 1992. *Study and interpretation of the chemical characteristics of natural water* (3rd edition), U.S. Geological Survey Water-Supply Paper 2254.

Hendry M.J. 1988. Do isotopes have a place in ground-water studies?. *Ground Water.* 26(4), 410-415.

Hvorslev, M.J. 1951. *Time lag and soil permeability in groundwater observations.* U.S. Army Corps of Engineers Waterway Experiment Station Bulletin 36. Vicksburg, MS.

Jacob, C.E. 1963. *The recovery method for determining the coefficient of transmissibility*, U.S. Geological Survey Water Supply Paper 1536-I.

Muldoon, M.A. 1999. *Hydrogeologic characterization of the Silurian dolomite in Door County, Wisconsin, at regional and site-specific scales: comparison of continuum and discrete-fracture approaches.* unpublished Ph.D dissertation (Geology and Geophysics), University of Wisconsin-Madison, 231 p.

Muldoon, M.A. and K.R. Bradbury. 1998. *Tracer study for characterization of groundwater movement and contaminant transport in fractured dolomite.* Wisconsin Geological and Natural History Survey Open-File Report 1998-02, 87 p.

National Research Council (Committee on Fracture Characterization and Fluid Flow). 1996. *Rock Fractures and Fluid Flow: Contemporary Understanding and Applications.* National Academy Press, Washington D.C., 551 p.

Nichols, M.S. and D.R. McNall. 1957. Strontium content of Wisconsin municipal waters. *Journal of the American Water Works Association* 49, 1493-1498

Novakowski, K.S. and G.S. Bickerton. 1997. Borehole measurement of the hydraulic properties of low-permeability rock. *Water Resources Research* 33(11), 2509-2517.

Ostrom, M.E. 1967. *Paleozoic stratigraphic nomenclature for Wisconsin.* Wisconsin Geological and Natural History Survey Circular 8.

Parkhurst, D.L. 1995. *User's guide to PHREEQC – A computer program for speciation, reaction-path, advective transport, and inverse geochemical calculations.* U.S. Geological Survey Water Resources Investigations Report 95-4227, 143 p.

Piggott, A.R, T.N.T. Huynh, P.A. Lapcevic and K.S. Novakowski. 1995. Automated analysis of hydraulic and tracer tests conducted in fractured rock. *Hydrogeology Journal* 4(3), 84-93

Shapiro, A.M., and E.A. Greene. 1995. Interpretation of prematurely terminated air-pressurized slug tests. *Ground Water* 33(4), 539-546.

Siegel, D.I. 1990. Sulfur isotope evidence for regional recharge of saline water during continental glaciation, north-central United States. *Geology* 18 (11), 1054-1056.

Simpkins, W.W. 1989. *Genesis and spatial distribution of variability in the lithostratigraphic, geotechnical, hydrogeological and geochemical properties of the Oak Creek Formation in southeastern Wisconsin.* unpublished Ph.D. dissertation (Geology and Geophysics), University of Wisconsin-Madison. 394 p.

Weaver T.R. and J.M. Bahr. 1991. Geochemical evolution in the Cambrian-Ordovician sandstone aquifer, eastern Wisconsin; 1, Major ion and radionuclide distribution. *Ground Water* 29 (3), 350-356.