HYDRAULIC CONDUCTIVITY AND SPECIFIC STORAGE OF THE MAQUOKETA SHALE

A Final Report prepared for the

UNIVERSITY OF WISCONSIN – WATER RESOURCES INSTITUTE

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III. PROJECT SUMMARY

The Maquoketa Formation, a dolomitic shale, is an important regional confining unit between the Silurian dolomite aquifer and the deep Cambrian-Ordovician aquifer system in southeastern Wisconsin. Rapidly growing communities in the region rely on pumping municipal water supplies primarily from the deep aquifer system, which has caused the formerly upward vertical gradient across the Maquoketa Formation to be reversed. In addition, significant quantities of water are pumped from the Silurian dolomite aquifer which is the upper bedrock formation. Hence, the role that the Maquoketa confining unit plays in the regional multi-aquifer hydrogeologic system needs to be better understood for the purpose of long-term groundwater management and protection.

Two separate but related research projects were initiated in 1997 and 1998. The first project, entitled **Evaluation of the confining properties of the Maquoketa Formation in the SEWRPC region of southeastern Wisconsin** (P.I.s Eaton and Bradbury, 1998), focused on hydraulic and geochemical fieldwork at an observation well at one of two field sites (Minooka Park, Waukesha). The other project, entitled **Hydraulic Conductivity and Specific Storage of the Maquoketa Shale** (P.I. Wang and research assistant Hart, 1997) concentrated on laboratory measurements on core samples and computer modeling of poroelastic properties. The former project was planned for only one year, while the latter included a second-year budget component (FY99/00) to test some modeling hypotheses by installing a pumping well. We submitted a revised second year budget request to the second project (Wang, Eaton and Bradbury, 1998) in order to combine efforts and build on existing results with a slightly expanded joint fieldwork proposal. This report therefore summarizes early work by Wang and Hart, and presents the results of joint research conducted by all parties using the improved multiple-well design in the two years since.

The main objective of this research is to obtain hydrogeologic properties of the Maquoketa Formation in Waukesha County, Wisconsin. Due to the generally low conductivity of this confining unit, which is a lithologically diverse dolomitic shale (Eaton and Bradbury, 1998), hydrogeologic testing is considerably more difficult than for a conventional aquifer. We therefore employed two complementary approaches in our research. One is based on laboratory rock core tests and modeling, and subsequent field verification, using Biot♣s (1941) theory of poroelasticity, which accounts for coupled deformation of the rock mass with fluid pressure changes. Recent reviews are presented by Hsieh (1996) and Wang (2000).

The other approach is more conventional, relying on field hydraulic testing using multiple wells. One of the problems with single-well testing, particularly in low-conductivity formations, is that the volume of rock tested is limited to the immediate vicinity of the well. With a multiple-well configuration, a much larger and potentially more representative rock volume can be tested, and scaling effects can be evaluated. The major goal of the conventional approach was to conduct a pumping test in the underlying Sinnipee Group dolomite, and analyze resulting head change at multiple observation points in the Maquoketa Formation using the

■leaky aquifer• method of Hantush (1956) and Neuman and Witherspoon (1972) to estimate vertical hydraulic conductivity in the confining unit. The pumping was anticipated to provide the stress needed to induce the reverse water level fluctuations predicted by poroelastic theory. This was not the case, but we observed possible reverse water-level fluctuations due to drawdown from water sampling in a much closer well.

This study has investigated the hydraulic properties of the Maquoketa shale confining unit using a novel laboratory and poroelastic modeling approach as well as a more conventional field-based hydrogeological approach. Laboratory pulse-decay testing of rock core has established that hydraulic conductivity ranges between 6.2×10^{-14} and 4.3×10^{-12} ft/s, and specific storage ranges between 3.7×10^{-9} and 8.5×10^{-7} ft^{-1}, which we consider representative of unfractured rock matrix at small scales. Poroelastic modeling predicts a small reverse water-level fluctuation in response to pumping, and some of our field data may reflect such a coupled poroelastic response to our field testing.

However, prior field hydrogeologic testing (Eaton & Bradbury, 1998) resulted in considerably higher hydraulic conductivity values ranging between 1×10^{-9} ft/s and 1×10^{-4} ft/s. Multiple-well geophysical logging and hydraulic testing reported here indicate that significant bedding plane fractures occur in the upper 100 ft of the Maquoketa Formation, and that these conductive fractures are well connected vertically to the overlying Silurian dolomite aquifer. “Leaky aquifer” testing by pumping the adjacent formations failed to provide bulk hydraulic conductivity values for the Maquoketa Formation, in part because of the fractures but also because the underlying Sinnipee Group dolomite has a very low hydraulic conductivity of 2×10^{-9} ft/s at this site.

We suggest a new conceptual model of the hydrogeology of this important regional confining unit, consisting of a relatively high transmissivity, interconnected, but sparse fracture network embedded in a low conductivity rock matrix. Bulk hydraulic conductivity of the rock mass is therefore a complex function of matrix conductivity, fracture density and transmissivity, and fracture network interconnectedness. Areas of relatively low fracture density and interconnectedness, such as the shale-rich base of the formation, do not readily transmit head changes, and may account for the regional confining properties of the Maquoketa Formation. In contrast, the upper fractured 100 ft of the Maquoketa Formation have a good hydraulic connection to the overlying Silurian aquifer via this fracture network.

These results have significant implications for the role of the Maquoketa confining unit in the regional groundwater flow system. Although at large scales, the shale-rich base of the formation provides an effective confining unit, the upper part is hydraulically coupled to the overlying Silurian aquifer. This suggests that it is not a good assumption that the top of the Maquoketa Formation is an effectively “impermeable” or no-flow boundary to the Silurian aquifer. These findings also indicate that groundwater contamination (particularly DNAPLs) could migrate into the fractured top of the Maquoketa Formation.

IV. INTRODUCTION

The Maquoketa Formation, an Ordovician-age dolomitic shale, is an important regional confining unit between the upper Silurian dolomite aquifer and the deep Cambrian-Ordovician aquifer system in southeastern Wisconsin. Rapidly-growing communities in the region rely on pumping municipal water supplies primarily from the deep aquifer system, which has caused the formerly upward vertical gradient across the Maquoketa Formation to be reversed. In addition, significant quantities of water are pumped from the upper Silurian dolomite aquifer which, because it is the upper bedrock aquifer, is quite vulnerable to contamination from the land surface. Hence, the role that the Maquoketa confining unit plays in the regional multi-aquifer hydrogeologic system needs to be better understood for the purpose of long-term groundwater management and protection.

Concern about future groundwater management has led to the initiation of a joint project, under the auspices of the Southeastern Wisconsin Regional Planning Commission (SEWRPC), by the Wisconsin Geological and Natural History Survey, the U.S. Geological Survey, the Wisconsin DNR, and cooperating local water utilities to construct a three-dimensional groundwater flow model of the region. This computer model will be used to simulate the entire flow system, including the Silurian aquifer, the Cambrian-Ordovician aquifer system and the Maquoketa confining unit. It will provide guidance for local water utility managers on optimizing future municipal pumping while minimizing regional drawdown, which has reached several hundred feet in Waukesha County in the twentieth century. The results on the hydrogeology of the Maquoketa Formation described in this report are now being incorporated into the larger modeling effort.

V. BACKGROUND

Two separate but related research projects funded through the Groundwater Research Advisory Council/Joint Solicitation process were initiated in 1997 and 1998. The first project, entitled **Evaluation of the confining properties of the Maquoketa Formation in the SEWRPC region of southeastern Wisconsin** (P.I.s Eaton and Bradbury, 1998), focused on hydraulic and geochemical fieldwork at an observation well at one of two field sites (Minooka Park, Waukesha). The other project, entitled **Hydraulic Conductivity and Specific Storage of the Maquoketa Shale** (P.I. Wang and research assistant Hart, 1997) concentrated on laboratory measurements on core samples and computer modeling of poroelastic properties. The former project was planned for only one year, and a final report (Eaton and Bradbury, 1998) was submitted, while the latter included a second-year budget component (FY99/00) to test some modeling hypotheses by installing a pumping well.

Based on the results of these two related projects, we realized that a multiple-well field configuration at the second site (DOT Ryan Parcel, Pewaukee) would present major advantages in understanding the hydrogeology of this complex formation, from both a poroelastic perspective (Wang and Hart) and a conventional hydrogeologic perspective (Eaton and Bradbury). So we submitted a revised second year budget request to the second project (Wang, Eaton and Bradbury, 1998) in order to combine efforts and build on existing results with a slightly expanded joint fieldwork proposal. This report therefore summarizes early work by Wang and Hart, and presents the results of joint research conducted by all parties using the improved multiple-well design in the two years since.

A. Objectives

The main objective of this research is to obtain hydrogeologic properties of the Maquoketa Formation in Waukesha County, Wisconsin. Due to the generally low conductivity of this confining unit, which is a lithologically diverse dolomitic shale (Eaton and Bradbury, 1998), hydrogeologic testing is considerably more difficult than for a conventional aquifer. We therefore employed two complementary approaches in our research. One is based on laboratory rock core tests and modeling, and subsequent field verification, using Biot's (1941) theory of poroelasticity, which accounts for coupled deformation of the rock mass with fluid pressure changes. Poroelastic effects are expected to be more important and observable (notably through reverse water level fluctuations) in very low-conductivity formations (Wang, Eaton and Bradbury, 1998). Recent reviews are presented by Hsieh (1996) and Wang (2000).

The other approach is more conventional, relying on field hydraulic testing using multiple wells. One of the problems with single-well testing, particularly in low-conductivity formations, is that the volume of rock tested is limited to the immediate vicinity of the well. With a multiple-well configuration, a much larger and potentially more representative rock volume can be tested, and scaling effects can be evaluated. The major goal of the conventional approach was to conduct a pumping test in the underlying Sinnipee Group dolomite, and analyze resulting head change at multiple observation points in the Maquoketa Formation. We planned to use the "leaky aquifer" method of Hantush (1956) and Neuman and Witherspoon (1972) to estimate vertical hydraulic conductivity in the confining unit. The pumping was anticipated to provide the stress needed to induce the reverse water level fluctuations predicted by poroelastic theory. This was not the case, but we observed possible reverse water-level fluctuations due to drawdown from water sampling in a much closer well.

These approaches were based on a classical equivalent porous medium conceptual model of the Maquoketa Formation. However, as soon as we began installation of the additional pumping and observation wells, we observed head changes that contradicted this equivalent porous medium assumption, and pointed to a conceptual model of a discrete fracture network. Chemical evidence of preferential fracture flowpaths presented by Eaton and Bradbury (1998) had hinted at this possibility. Therefore, our objectives evolved slightly during this research to investigate and characterize the extent of this fracture network. Although water sampling was conducted and analyzed for major ion chemistry and isotopes during this project, the interpretation of these results has been postponed and will be presented as part of the final report of the ongoing research effort: **Verification and characterization of a fracture network within the Maquoketa shale confining unit, southeastern Wisconsin** (Eaton, Bradbury and Wang, 1999). Significant drawdown in the multi-level well during water sampling was used in the poroelastic modeling analysis to investigate possible reverse water level fluctuations.

B. Site preparation

Preparatory work consisted of the installation of a multi-level packer and monitoring system in a pre-existing corehole, and drilling of several wells, their logging and instrumentation at the field site known as the DOT Ryan Parcel near Pewaukee, in Waukesha County (Figure 1). We completed most of this preparatory work before January 2000, along with some preliminary poroelastic modeling, and were able to conduct the balance of the hydraulic testing in the spring and summer 2000.

1. Well drilling

We contracted with a private well-drilling company to drill three six-inch diameter wells at the field site as illustrated in Figure 1. The two observation wells (W-1 and W-2) are located approximately 30 ft to the east of the existing corehole (WK1376). Both are cased into the top of the Silurian dolomite and one well has a depth of 250 ft while the other has a depth of 360 ft. The third well (W-0) is situated about 58 ft to the west of the existing corehole, is cased through the Silurian dolomite, and is intended as a pumping well. It extends to a depth of 458 ft and is open to the Maquoketa Formation and the underlying Sinnipee Group dolomite.

2. Downhole geophysical logging

Prior to equipment installation, we conducted a full suite of downhole geophysical logs in each well. These tools included natural gamma, spontaneous potential (SP), single-point resistance, caliper, fluid temperature and resistivity, heat-pulse flowmeter, and in addition, video in the deep well. Natural gamma logs are very useful for distinguishing lithology because they are sensitive to gamma radiation associated with clay and feldspar minerals, which is measured in counts per second (cps) using a scintillation detector. In general, increasing radiation (increasing cps) is associated with increasing clay or shale content, and clean dolomite or sandstone shows low radiation readings. Spontaneous potential (SP) and single-point resistance respond to the electrical properties of the formation and groundwater.

Caliper logs measure the diameter of the well very precisely, and are very useful indicators of horizontal bedding-plane fractures and openings in carbonate formations such as the Maquoketa shale and Silurian dolomite. Temperature and fluid resistivity logs are sensitive to changes in fluid properties in the well, usually from water flowing into or out of the well at discrete locations such as fractures. Similarly, heat-pulse flowmeter logs measure changes in flow rates in a well at discrete inflow or outflow points. Finally, a video log was useful in verifying and directly observing fractured vs. unfractured zones in the deep well (W-0).

3. Equipment installation

In the first year of this multi-year research effort, rock core samples from the Maquoketa Formation were collected from two sites in Waukesha County, one at Minooka County Park and the other at the DOT Ryan Parcel. The first of these two cores was described in detail in the first final report (Eaton and Bradbury, 1998), and in the second set of core samples, we found the same six lithofacies with slightly different relative thicknesses. Both sets of rock core samples were used in the laboratory testing described in the next section. The resulting corehole at Minooka Park was instrumented with a multi-level packer and monitoring system, which has been described elsewhere (Eaton and Bradbury, 1998).

We installed a similar system more recently in the corehole (WK-1376) at the DOT Ryan Parcel site. It consists of six nitrogen-inflated rubber packers approximately 3 ft long (represented in black in Figure 1), which isolate different intervals of the corehole equipped with vibrating-wire pressure transducers and small sampling pumps. A datalogger at the surface continually records hydraulic head at different levels, while a pressure cylinder and valve system maintain packer pressures over time. The positions of the packers and interval lengths were chosen in order to isolate relatively different lithologies and fracture densities based on downhole logs. All packers are independently inflated and were found to adequately maintain pressure with the exception of Packer 4 at a depth of 230 ft, which turns out not to affect our results significantly.

We also installed monitoring instrumentation and dataloggers at the two easternmost six-inch diameter wells W-1 and W-2 (Figure 1). The monitoring well construction was different from that in the multi-level packer system because of a concern that head measurements in the long open intervals between packers might be insensitive to reverse water level fluctuations of a couple inches or less. In these six-inch monitoring wells, we installed pressure transducers embedded in short (2-3 ft thick) pea-gravel packs at different levels within the Maquoketa Formation and adjacent formations. In the shallower of the two wells (W-1), transducers were installed at depths of approximately 140ft, 180ft, and 230ft. In the deeper well (W-2), the transducers were installed at depths of approximately 280ft, 320ft and 360ft. The intervals between the gravel packs were carefully backfilled and sealed with coarse bentonite chips according to DNR well abandonment guidelines. An interval of approximately 50 ft was left open in the shallower well (W-1) just below the casing depth of 53 ft.

VI. LABORATORY AND POROELASTIC APPROACH

A. Laboratory measurements

1. Introduction

The laboratory measurements of the hydraulic conductivities and specific storage were conducted to complement and constrain the field tests. Although laboratory measurements have problems associated with scale and alteration of the sample during and after collection, they do have the advantage of allowing the experimenter strict control of the measurement environment. For example, field measurements of the hydraulic conductivity are limited to the heads and overburden stresses at which they are conducted. Laboratory experiments can simulate the behaviors of the shale at different pressure heads and overburden stresses, such as would be encountered by the Maquoketa shale at greater depth of burial to the east of the field site in Waukesha. In addition, the hydraulic conductivity anisotropy is measured more directly.

2. Methods

Core recovered by Eaton and Bradbury (1998) from both the Minooka Park and the DOT Ryan Parcel sites were tested. These samples had dried before testing and so the pore fluid and structure were possibly altered. For measurement of the vertical hydraulic conductivity, perpendicular to bedding, the 2-inch diameter cores were cut to lengths between 0.8 and 1.6 inches. Their ends were ground flat by hand using successively finer grit sandpaper and a grinding plate. The sample preparation was done without fluids to eliminate contamination of the shales by the cutting and grinding fluids and because many of the samples experienced fissile splitting upon rewetting. The samples were then saturated by placing them in a vacuum for at least 48 hours to evacuate the air. They were then submerged while still under vacuum in a water solution, created to mimic the pore fluid chemistry obtained by Eaton and Bradbury (1998) in the Minooka Park corehole, for the interval from which the sample was taken. Finally, the vacuum was released and the samples were allowed to saturate for at least 48 hours under atmospheric pressure. Effective porosity was calculated using the wet/dry mass-difference method. The samples for measurement of the horizontal hydraulic conductivity were one-inch diameter cores obtained by coring parallel to the bedding in the two-inch diameter cores. The samples were then jacketed with a Tygon sleeve and placed in the pressure vessel.

The hydraulic conductivity and specific storage were measured using the transient pulse decay method. This method allowed measurement of very low hydraulic conductivities and has been used previously to measure the hydraulic conductivity of shales (Aoki, 1996; Neuzil et al, 1981; Bredehoft et al, 1983). In the transient pulse test, the sample is placed between two reservoirs in the pulse decay apparatus (Figure 2). Before the test, the fluid pressure heads are equal in the two reservoirs and throughout the sample. At the beginning of the test, the fluid pressure is suddenly increased in the upper reservoir and held constant. This creates a head differential, which causes fluid to flow from the upper reservoir, into and through the sample, into the bottom reservoir.

The difference in pressure head between the upper reservoir, held at constant pressure, and the lower reservoir is recorded as a function of time. The pressure-time curve is then analyzed to determine K and Ss using a non-linear inversion (Wang and Hart, 1993) based on the analytical solution for the transient pulse decay (Hsieh et al, 1981). This analysis is similar to doing a pumping test and then inverting the data using a computerized algorithm. Figure 3 is an example of a pressure-time curve for sample MDOT-242-1. Note that the x-axis is logarithmic and that the y-axis is normalized head for the lower reservoir equal to the increase in the lower reservoir head divided by the initial step increase in the upper reservoir.

3. Results and Discussion

Table 1 summarizes the results of the experiments. The locations and depths of the samples are given in the first column. The vertical core scale hydraulic conductivities ranged in value from 8.2×10^{-14} to 4.3×10^{-12} ft/s. The horizontal core scale hydraulic conductivities ranged in value from 6.2×10^{-14} to 1.2×10^{-9} ft/s. Effective porosity ranges between 0.04 and 0.26. Although these values are extremely small, particularly hydraulic conductivity, they are within ranges given in the literature by Neuzil (1994) and Domenico and Schwartz (1998).

Table 1: Results of Laboratory Measurements on Rock Core.

Sample Location-depth (ft)	Vertical Hydraulic Conductivity (ft/s)	Horizontal Hydraulic Conductivity (ft/s)	Specific Storage (1/ft)	Effective Porosity
MDOT-180	4.3E-12	8.2E-13	3.7E-09	0.15
MDOT-212	1.4E-12	3.9E-10	6.7E-07	0.12
MDOT-242	4.3E-12	6.6E-13	8.5E-07	0.26
MDOT-267	3.1E-13	2.0E-12	8.2E-08	0.12
MDOT-297	3.9E-12	8.9E-13	2.7E-07	0.22
MDOT-340	1.9E-13	3.3E-13	5.2E-08	0.21
MMIN-251	3.6E-12	NM	2.8E-08	0.04
MMIN-288	2.4E-13	2.0E-12	4.0E-07	0.10
MMIN-319	3.9E-13	1.2E-09	1.1E-07	0.13
MMIN-350	8.2E-14	6.2E-14	4.9E-08	0.03
MMIN-375	1.8E-13	7.9E-13	3.7E-07	0.25
MMIN-397	NM	4.6E-13	7.0E-07	0.20
Mean*	7.2E-13	6.2E-13	3.0E-07	0.15

NM-not measured.
MDOT- DOT Ryan Parcel, Pewaukee; MMIN-Minooka Park, Waukesha.
*Means are harmonic for hydraulic conductivity and arithmetic for specific storage and effective porosity

Two samples had horizontal hydraulic conductivities with values greater than 1×10^{-10} ft/s.

These anomalous values are high due to dessication cracks that occurred along the bedding plane, which allowed for greater flow. When these values are eliminated from the data set, the core scale anisotropy (mean horizontal hydraulic conductivity/ mean vertical hydraulic conductivity = 0.86) is approximately unity. This does not imply that the Maquoketa Formation as a whole is isotropic. At a larger field-scale, downhole geophysical logs, core lithology and hydraulic testing show that while there is lateral continuity between wells, there is considerable vertical anisotropy and heterogeneity.

The laboratory effort also discovered little-to-no correlation between hydraulic conductivity and several parameters: effective stress, porosity, and gamma count. The absence of any role of effective stress: the difference between the confining pressure and the pore pressure, implies there must be relatively few microcracks present in these cores. An increasing effective stress would close any microcracks and thus reduce the hydraulic conductivity. Porosity, used by Neuzil (1994), as a determining factor for shale hydraulic conductivity, was also not correlated. We hoped that the natural-gamma downhole geophysical log would correlate with the hydraulic conductivity, but that too was not significantly correlated. This lack of correlation is likely due to the low hydraulic conductivity of both shale and dolomite found in the Maquoketa Formation. The clay minerals in the shale show a higher porosity and gamma count while the dolomite has a lower porosity and gamma count. The homogeneity of the dolomite implies few microcracks while the plasticity of the clay minerals in the shales will prevent microcracks from forming in the saturated cores.

B. Poroelastic modeling

1. Introduction

Reverse water level fluctuations are water level responses opposite normally expected behavior, i.e. water level increases near a pumping well (Langguth and Treskatis, 1989). This effect can be explained as a poroelastic effect, i.e., there is coupling between the stress and fluid pressure fields. Drawing down hydraulic head produces a local volumetric contraction in the rock matrix where the fluid pressure is reduced. This strain in turn produces deformation and stresses elsewhere in the aquifer, adjacent confining layers, or adjacent aquifers. This portion of the report gives some results of the modeling of this effect specific to the Maquoketa shale and Sinnipee Group dolomite formations and reports some initial field data suggestive of a reverse water level fluctuation. The planned leaky aquifer pumping test conducted in the underlying Sinnipee Group dolomite did not induce measurable reverse water level fluctuations in our observation wells, as described in later sections of this report. However, subsequent pumping to obtain samples for water chemistry resulted in several hundred feet of drawdown in the Sinnipee Group dolomite. This drawdown, which occurred in well WK1376, was much closer to observation well transducers, and did induce what appears to be a poroelastic response corresponding to our simulations.

2. Methods

A commercially available finite element code, ABAQUS, which can couple the stress and pore pressure fields, was used to create the model. The finite element grid is shown in Figure 4. The grid represents an axisymmetric domain with a horizontal radius of approximately 1300 ft and a vertical height of approximately 620 ft. The boundary conditions and aquifers and their respective parameters are included in Figure 4 as well. Note the location of the well nodes shown as a black band at the left hand side of the figure, immediately below the Maquoketa Formation. This configuration approximately represents the DOT Ryan field site, except that the Maquoketa shale in Figure 4 excludes 20 ft of the top fractured portion of the formation. Poroelastic parameters for the dolomites and the shale were taken from values for Indiana limestone (Hart and Wang, 1995) and Trafalgar shale (Aoki, 1996). The hydraulic conductivities, K, are from Stocks (1998), and Eaton and Bradbury (1998). The associated specific storage coefficients, Ss, are calculated from the poroelastic constants.

3. Results and Discussion

A head decrease of -7.5×10^5 Pa (~-250 ft of water) was applied to the well nodes and the model was allowed to run for approximately 120 hours model time. The model boundaries were chosen so that they are far enough away that they did not affect the results of the modeling of the reverse water level fluctuations. Figure 5 shows the deformed finite element mesh at a model time approximately 16 hours after the head decrease. The greatest volumetric contraction occurred at the well nodes. That contraction produces strain across the entire model domain in the same way that contraction is expected to affect an elastic rock matrix. Model elements far from the origin of the stress (head change due to flow into the well), are then expected to experience a corresponding head change proportional to the volume strain.

During pumping of water sample from the bottom interval in the multi-level monitoring system in well WK1376, the head in the Sinnipee Group dolomite, below the lowest packer, was drawn down by about 300 feet. Certain pressure transducers in wells W-1 and W-2 located 30 ft from WK1376 recorded head changes in the Sinnipee dolomite and Maquoketa Formation, following the head decrease in the sampled well. No significant head changes were observed in PZ-5 and PZ-6 or in any other levels of WK1376. Observed head variations are shown in Figure 6a. Head change in PZ-2 is not illustrated because of a signal interference problem in the transducer at that level. The change in sampling interval at approximately 7 hours corresponds to a resetting of datalogger collection interval to conserve battery power overnight. Fluctuation in measured head prior to this time may be caused by variable stress due to cycling of the double-valve sampling pump, or noise in transducer readings.

We used the poroelastic model to calculate head variations at elevations within the model domain corresponding to the positions of the pressure transducers in field monitoring wells relative to well WK1376, i.e. at a radial distance of 30 ft. These model-calculated head variations are plotted as a function of time in Figure 6b. Poroelastic modeling results indicate that the maximum head changes (approx. 3 inches of hydraulic head) due to rock matrix deformation would be expected to occur about 14 hours after initial stress due to pumping. The area of greatest reverse head fluctuation is predicted to occur in PZ-4 near the top of the Maquoketa Formation. This reverse head fluctuation forms almost immediately after the head decrease in the well nodes and persists for approximately 55 hours as the pressure decrease propagates radially outward.

There are many discrepancies between the measured and simulated data. The greatest difference is the lack of observed head decline in the Sinnipee (PZ-1) after the step decrease in head in WK1376. The modeled head in the upper Sinnipee decreases so rapidly that it cannot be shown with the choice of axes in Figure 6b. This may be explained by the fact that during field testing, the Sinnipee Group dolomite was found to be of considerably lower hydraulic conductivity than expected and simulated. It is in fact a confining unit not significantly different from the Maquoketa Formation at this site. Since PZ-4 is located near the top of the Maquoketa Formation where a reverse water level fluctuation is expected, the strong increase in observed head in Figure 6a might be due to the mechanism of reverse water level fluctuation. Tidal and barometric loading can be eliminated as causing this change because only PZ-4 shows a strong increase in heads; PZ-3 and PZ-2 (not shown) do not. The duration of the measured head rise also approximately corresponds to the duration of the modeled head rise (approximately 20 hours), which provides some evidence for this interpretation.

Unfortunately these preliminary results, although they are intriguing, remain uncertain and require further investigation. Of necessity, the poroelastic model is a highly simplified representation of the actual Maquoketa Formation in the field, using our best estimates of generalized hydraulic and rock mechanics parameters. In contrast, hydrogeologic field-testing during this project has shown that the Maquoketa Formation is a highly complex and heterogeneous formation, both hydraulically and mechanically. In fact, at the scale of the field site, field testing described in the next section indicates that the equivalent porous medium assumption may not be a valid conceptual model for the hydrogeology of the Maquoketa Formation.

VII. CONVENTIONAL FIELD-BASED APPROACH

A. Field hydraulic testing

Prior to planned testing, we first fortuitously observed head changes in various wells during drilling of others, and during preliminary test-pumping. We then carried out hydraulic testing using a submersible pump in the open interval of well W-0, then in well W-1. While testing well W-0, we isolated parts of the borehole using an inflatable packer.

1. Observations during drilling

During the drilling of the easternmost wells W-1 and W-2, we recorded head levels in the multi-level monitoring system at well WK1376. We observed no head changes until the bottom of the first drilled well W-1 reached 135 ft in the Silurian aquifer, the depth of a significant horizontal fracture feature we later observed on the caliper logs (Figures 7, 9). At that time, we were surprised to see simultaneous drawdown of heads at all levels within the underlying Maquoketa Formation, as indicated by the multi-level packer and monitoring system at WK1376. This drawdown increased to a maximum of 30 ft by the time well W-1 was completed, and was undoubtedly caused by the effective pumping of 50-100 gpm associated with drilling. Similar observations were made during drilling of well W-2.

2. Preliminary test pumping

In order to conduct the deep aquifer pumping test, we needed to install a packer in the deep pumping well (W-0) below which we planned to pump, since the well is open to both the Maquoketa Formation and Sinnipee Group dolomite. While lowering the packer assembly to its position at 340 ft depth for the deep pumping test, we decided to pump for short periods below the packer as it was lowered by 50 ft intervals. Resulting drawdowns occurred in several multi-level monitoring intervals in other wells until the packer reached a depth of 250 ft, below which there was no drawdown in other wells.

3. Deep aquifer pumping test

Having installed the packer at 340 ft in the deep pumping well (W-0), at the base of the Maquoketa Formation, we conducted a leaky aquifer test by pumping the underlying Sinnipee Group dolomite. Prior to pumping, we did not know what pumping rate could be sustained, although we hoped for 5-10 gpm. Unfortunately, the Sinnipee Group dolomite proved to be of significantly lower conductivity, and we were unable to sustain a pumping rate of more than 0.1 gpm. Nevertheless, we maintained that rate for approximately 96 hours and observed a drawdown of 230 ft in the pumping well. No resulting drawdowns were observed in any of the multi-level monitoring wells.

4. Shallow aquifer pumping test

Following the long-term deep aquifer test, we decided to try a pumping test in the open interval of well W-1, in hopes of reproducing the observations during drilling. We pumped from the Silurian aquifer for about 1.5 hours at a rate of 35 gpm. With a drawdown of 19 ft in the pumping well, we again observed immediate and simultaneous drawdowns of 0.5-2 ft in monitoring wells at most intervals throughout the Maquoketa Formation.

B. Results and Discussion

The results of the various field hydraulic tests we conducted were not exactly what we expected, but a careful analysis of downhole geophysical logs in conjunction with the pumping data provides some explanations, and suggests a new conceptual model for the hydrogeology of the Maquoketa Formation.

1. Downhole geophysical logs

The downhole geophysical logs for the coreholes in which the multi-level packer and monitoring systems were installed at Minooka Park, and more recently at the DOT Ryan Parcel (WK1376), were described in detail elsewhere (Eaton and Bradbury, 1998). This report focuses on logs from the six-inch diameter wells drilled for this study, which are very similar. An advantage of multiple wells drilled with different methods at one site is that it provides confidence that similarities seen in the logs represent regionally extensive features at specific depths, rather than particularities of individual wells.

a) Monitoring Well W-2

Downhole geophysical logs for the deep 6-inch diameter monitoring well W-2 are shown in Figures 7 and 8. The natural gamma log illustrates the shale-rich lithology of the Maquoketa Formation between depths of approximately 165 ft and 340 ft. On the caliper log, three distinct fracture features are present in the overlying Silurian dolomite, at depths of 60-70 ft, 105 ft and 130-140 ft. Additional fractured zones occur in the Maquoketa Formation at depths of approximately 170-175 ft, 185-195 ft and 235-240 ft. Variations in spontaneous potential (SP) and single point resistance reflect variations in lithology and water chemistry, some of which correspond to the fracture features.

On Figure 8, the logs of temperature and fluid resistivity indicate that the fractures at 105 ft and 185-195 ft cause only minor inflections in temperature, but major changes in fluid resistivity. The dual curves represent data collected while logging both downward and upward. The heat-pulse flowmeter log on the right indicates that water flows into the well at 105 ft and much of it exits at 135 ft as shown by the decrease in downward flow. Perturbations in vertical flow are apparent in the upper portion of the Maquoketa Formation above 250 ft but are considerably attenuated below.

b) Monitoring Well W-1

Downhole geophysical logs for the shallow 6-inch diameter well (W-1) are shown in Figures 9 and 10. It was not possible to collect a natural gamma log due to equipment problems. The caliper log shows significant fracturing at depths of 60 ft (just below the casing), 102 ft, and 130 ft in the Silurian dolomite. Numerous fractures are apparent within the Maquoketa Formation below 165 ft depth, some of which may have been washed out by complications during drilling. Variations in spontaneous potential (SP) and single point resistance are similar to those in other wells. In Figure 10, there is relatively little variation in temperature with depth, however a temperature inflection and a significant increase in fluid resistivity just below 100 ft depth correspond to a fracture. The lower decline in resistivity probably corresponds to flow into fractures just above 200 ft. A combined analysis of these logs suggests that fractures at 100 ft and just above 200 ft control most of the flow in the open well. A heat-pulse flowmeter log was not conducted in this well.

c) Pumping Well W-0

Downhole geophysical logs for the deep pumping well are shown in Figures 11, 12 and 13. In general, the same features are apparent as in the other well logs. Fractured zones are also present on the caliper log in this well at 165-175 ft, 185-200 ft and 230-245 ft, indicating that these features are certainly locally and possibly regionally extensive, probably related to bedding plane discontinuities. The temperature and fluid resistivity logs (Figure 13) show that they are hydraulically active. There appears to be a minor leak in the casing at about 110 ft, causing an offset in the fluid resistivity log, but no flow was recorded. The heat-pulse flowmeter log (Figure 12) shows that flow is exclusively downward in the open hole, reflecting the regional strong downward gradient. Since the Silurian dolomite aquifer is cased off in this well, all significant flow originates from the fractured upper part of the Maquoketa Formation.

In addition to the suite of logs conducted in other wells, we also conducted a downhole video log in this well to examine the Maquoketa Formation more directly. A sketch which summarizes the major details in the video log is presented in Figure 13. Numerous horizontal fractures are visible, some of which had up to 1/8 inch apertures, but many others were filled with blue-gray shale. These openings and occasional large holes are common above 250 ft, and below this depth the rock is relatively smooth and featureless.

2. Responses to drilling and test pumping

Simultaneous drawdown at all levels in the Maquoketa Formation during drilling in the overlying Silurian dolomite is incompatible with the conceptual model commonly assumed in hydrogeology. If the effects of pumping were transmitted through an equivalent porous medium, one would expect sequential and diminishing drawdowns with distance from the pumping center. The downhole geophysical logs provide additional evidence that instead, preferential flow occurs mainly via discrete fractures which form an interconnected network in the Silurian dolomite and Maquoketa Formation. Following installation of multi-level monitoring instrumentation in the 6-inch wells, more formal hydraulic testing was used to check this hypothesis.

a) Observations of hydraulic head equilibration

After instrumentation of the corehole and newly drilled wells, as described above, hydraulic heads were allowed to equilibrate over a period of months. As observed in the earlier multi-level packer system at Minooka Park (Eaton and Bradbury, 1998), heads in the multi-level packer system at the DOT Ryan Parcel first all equilibrated to approximately 847-852 ft above mean sea-level (msl), similar to the head in the overlying Silurian dolomite. Head in the lowermost interval (in the Sinnipee Group dolomite) briefly dropped to 560 ft above msl within a few weeks, then returned to the previous level, for reasons which may be related to difficulties regulating packer pressure. However, head in the lowermost interval in observation well W-2 (bentonite and gravelpack construction) equilibrated to 734 ft above msl, whereas head in other intervals in this well and in the shallower well W-1 (also bentonite and gravelpack) equilibrated to 840-866 ft above msl. Later, after the aquifer testing, when intervals in the multi-level packer well WK1376 were pumped for water chemistry samples, the head in the lowermost interval was drawn down and recovered to only 655 ft above sea-level, while head in the other intervals returned to their previous levels. Final equilibrated head levels for all wells are shown in Figure 14.

The regional head difference between the Silurian aquifer and the deep Cambrian-Ordovician aquifer system is on the order of several hundred feet because of significant pumping drawdown in the lower aquifer. In a homogeneous (equivalent porous medium) confining unit, this head loss would be expected to be linear across the thickness of the confining unit. The equilibrated head levels in bottom intervals in the underlying Sinnipee Group dolomite are clearly being influenced by the much lower head in the deep aquifer system, but the head loss appears to occur mainly across the lower contact of the Maquoketa Formation. Heads measured throughout the Maquoketa Formation are very similar to, and appear to be controlled by hydraulic head in the upper Silurian aquifer. Given the extremely low hydraulic conductivity measured for rock matrix during the laboratory core testing, it seems unlikely that head changes seen in drilling and preliminary test pumping could be transmitted near-instantaneously through such low-conductivity rock unless there are significant interconnections and short-circuiting via fractures. Subsequent deep and shallow aquifer testing provides indications of the extent and interconnectedness of such a fracture network.

3. Results of deep and shallow aquifer testing

Estimation methods for the vertical hydraulic conductivity of a confining unit by analysis of head changes resulting from pumping the adjacent aquifer were developed by Hantush (1956) and refined by Neuman and Witherspoon (1972). A recent overview of this “leaky aquifer” or “ratio” testing method is presented by Rowe and Nadarajah (1993). It is fundamentally dependent on a lag time between head changes in the aquifer and head changes at an observation point in the adjacent confining unit as the transient “pressure wave” migrates vertically through the confining unit (Rowe and Nadarajah, 1993). This lag time or rate of migration is a function of the bulk hydraulic conductivity of the confining unit. In a low-conductivity unit like the Maquoketa Formation, such a lag time would be expected to be observable under the equivalent porous medium assumption.

a) Deep aquifer test

We therefore pumped from the underlying Sinnipee Group dolomite for a long time (96 hours) in the hope of observing head changes in the overlying Maquoketa Formation. No significant drawdown was observed in this time-frame (Figure 15). We recorded very small head changes (~0.5 ft) which we originally thought could be reverse well fluctuations predicted by poroelastic theory, but a comparison to nearly identical atmospheric pressure variations (Figure 15) clearly explains these small head changes. Two explanations seem most likely for the lack of more significant head changes. One is simply that not enough time elapsed for pore pressures in the confining unit to adjust to the head changes in the underlying dolomite. The other is that the Sinnipee Group dolomite is of such low conductivity that the effects of the 230 ft drawdown are extremely localized near the pumping well. The "leaky aquifer" method depends on a significantly higher hydraulic conductivity in the aquifer being pumped compared to that of the confining unit, so that primarily vertical migration of the pore pressure change occurs in response to pumping.

Both explanations probably account in part for the lack of observed drawdown in the Maquoketa Formation. Estimated hydraulic conductivity of the Sinnipee Group dolomite on the basis of recovery of the 230 ft pumping drawdown is $2x10^{-9}$ ft/s, which is not significantly higher than previous lowest field estimates of hydraulic conductivity in the Maquoketa Formation of $1x10^{-9}$ ft/s (Eaton and Bradbury, 1998). On a regional scale for the purposes of flow modeling, the U.S.G.S. has considered the Sinnipee Group dolomite to be part of the "Maquoketa confining unit" where the Maquoketa Formation is present (Young, 1992). However, in that case, one would not expect that equilibrium head in the Sinnipee Group dolomite (as measured in the lowermost monitoring intervals - Figure 14) would be affected by regional drawdown due to pumping, and not head in the Maquoketa Formation. On the other hand, if head change migrates extremely slowly and radially from the pumping center rather than vertically, one should observe corresponding head change in the overlying Maquoketa Formation after some finite time, but it may be too long to observe in a practical field test.

b) Shallow aquifer test

Following the negative results to the pumping test involving the underlying Sinnipee Group dolomite, we repeated the same test by pumping from well W-1 in the overlying Silurian dolomite, which is clearly an aquifer. The "leaky aquifer" method (Neuman and Witherspoon, 1972; Rowe and Nadarajah, 1993) is equally applicable to pumping an aquifer overlying the confining unit in question, as long as it is adjacent. In this case, we observed significant drawdown in the Maquoketa Formation as a result of pumping from the Silurian dolomite aquifer (Figure 15), but there was no significant lag time. These results confirmed our observations during drilling and preliminary test pumping, but are not amenable to analysis using the "leaky aquifer" method (Neuman and Witherspoon, 1972; Rowe and Nadarajah, 1993).

One of the hallmarks of equivalent porous medium behavior is that the timing and magnitude of drawdown in observation wells in response to pumping in a production well is related to the distance between the observation points and pumping point. The time-drawdown and distance-drawdown analysis methods using the Theis equation (Theis, 1935) are based on this principle. The near-simultaneous immediate drawdown observed at all levels in well WK1376 (Figure 15) in response to pumping from the Silurian aquifer violates this principle. Individual monitored intervals should respond sequentially with depth in the Maquoketa Formation, the bottom of which is over 250 ft vertically from the center of pumping. A possible explanation is some sort of mass failure of the packers in the corehole monitoring system, but this is unlikely since all packers are individually inflated and maintain adequate pressure with only one exception: Packer 4. Also, similar immediate drawdowns were observed in two intervals in well W-1, a multi-level well of different construction.

Furthermore, a closer scrutiny of drawdown behavior in the multi-level monitoring systems (Figure 15) in WK1376 reveals that not only is the time-drawdown relationship violated, but also the distance-drawdown relationship. One might expect that the greatest drawdown would be observed in the upper part of the Maquoketa Formation closest to the Silurian aquifer, and the smallest drawdown at the greatest distance near the bottom of the Maquoketa Formation. In fact, the reverse is the case: the drawdown in the uppermost monitored interval is approximately 1 ft whereas the drawdown at the base of the Maquoketa is nearly 2 ft. Similar unpredictable drawdown behavior was apparent in monitored intervals during preliminary pump testing as the packer was being lowered in the deep well W-0 in preparation for the deep aquifer test. This is clear evidence for not only preferential flow bypassing the rock matrix, but the existence of somewhat independent fracture pathways of different transmissivity.

4. Conceptual models for hydraulic behavior

a) Fractured, porous medium

In the previous final report from the first year (Eaton and Bradbury, 1998), we hypothesized that hydraulic head in the Maquoketa Formation had not yet equilibrated with the reversal of pre-development vertical head gradient due to pumping in the twentieth century. We later established (Wang, Eaton and Bradbury, 1998) that these equilibrated heads could be explained as transient head conditions using equivalent porous medium flow modeling. However, a simpler explanation seems to be that the Maquoketa Formation is exceedingly hydraulically heterogeneous because of the contrast in transmissivity between fractures and rock matrix. More recent analysis has established that it reaches 10 orders of magnitude (Eaton, Anderson, Bradbury, and Wang, 2000a; Eaton, Anderson, Bradbury and Wang, 2000b).

A fracture network cross-cutting low-permeability matrix creates an extremely hydraulically heterogeneous 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 equilibrium head are highly dependent on scale and interconnectivity of the fracture network. If head is monitored in a relatively long open interval, such as in the multi-level packer well (WK1376), it will be dominated by head changes in any fracture which may intersect the interval. However, head monitored at smaller scales, such as the short intervals in wells W-1 and W-2, may be more representative of head in the rock matrix. This may explain the greater variance in head measured in short-interval monitoring wells compared to that measured in WK1376 (Figure 14). Furthermore, head in one interval (PZ-3, depth 280 ft) in well W-2 actually exceeds the elevation of the land surface (860 ft above msl), implying that no interconnected fractures intersect that interval. That head measurement may represent remnant confined head uninfluenced by current water table conditions as measured in the Silurian aquifer.

b) Hydraulic complexity in the Maquoketa Formation

A conceptual model involving a relatively high transmissivity, interconnected, discrete fracture network embedded in a very low-conductivity rock matrix appears to provide the best explanation for the results described in this report. Downhole geophysical observations of water-conducting fractures at similar stratigraphic elevations in multiple wells drilled at the DOT Ryan Parcel site provide evidence of horizontal continuity of the fracture network. Such observations have been made by others in the Silurian dolomite aquifer in Door County, Wisconsin, and in the Sinnipee Group dolomite aquifer in northeastern Wisconsin (Muldoon et al., 1999; Stocks, 1998) over much larger horizontal distances. However, in this study using hydraulic testing, we have also demonstrated significant vertical interconnectivity of these horizontal bedding plane fractures, in an otherwise low-conductivity dolomitic confining unit.

Such interconnectivity of the fracture network is exceedingly important because of the apparently much lower density of conductive fractures in the Maquoketa Formation compared to those dolomite aquifers, and the extreme contrast in transmissivity between fractures and matrix. Recent numerical simulation work (Taylor et al., 1999) has shown that when that contrast exceeds 6.5 orders of magnitude, the vast majority of flow (and hence head change) only occurs in the interconnected fracture network, bypassing considerable volumes of unfractured rock matrix. Vertical head distribution, as measured in the Maquoketa Formation, also reflects this complexity.

Hydrogeologic characterization of fractured rocks is challenging because explicit localization of fracture pathways is not possible. There is also some uncertainty in the performance of our multi-level monitoring wells. For instance, an alternate hypothesis of short circuiting through the packer system at well WK1376 could explain some of our results. However, the similarity in equilibrium heads measured in monitoring wells of different construction (Figure 14) over long periods of time supports the validity of our conceptual model. Slight differences in head are better explained by differing lengths of open interval. We plan to pursue further tests of our monitoring system to resolve these uncertainties.

VIII. CONCLUSIONS AND RECOMMENDATIONS

This study has investigated the hydraulic properties of the Maquoketa shale confining unit using a novel laboratory and poroelastic modeling approach as well as a more conventional field-based hydrogeological approach. Laboratory pulse-decay testing of rock core has established that hydraulic conductivity ranges between $6.2\text{x}10^{-14}$ and $4.3\text{x}10^{-12}$ ft/s, and specific storage ranges between $3.7\text{x}10^{-9}$ and $8.5\text{x}10^{-7}$ ft^{-1}, which we consider representative of unfractured rock matrix at small scales. Poroelastic modeling predicts a small reverse water-level fluctuation in response to pumping, and some of our field data may reflect such a coupled poroelastic response to our field testing.

However, prior field hydrogeologic testing (Eaton & Bradbury, 1998) resulted in considerably higher hydraulic conductivity values ranging between $1\text{x}10^{-9}$ ft/s and $1\text{x}10^{-4}$ ft/s. Multiple-well geophysical logging and hydraulic testing reported here indicate that significant bedding plane fractures occur in the upper 100 ft of the Maquoketa Formation, and that these conductive fractures are well connected vertically to the overlying Silurian dolomite aquifer. “Leaky aquifer” testing by pumping the adjacent formations failed to provide bulk hydraulic conductivity values for the Maquoketa Formation, in part because of the fractures but also because the underlying Sinnipee Group dolomite has a very low hydraulic conductivity of $2\text{x}10^{-9}$ ft/s at this site.

These combined results suggest a new conceptual model of the hydrogeology of this important regional confining unit, consisting of a relatively high transmissivity, interconnected, but sparse fracture network embedded in a low conductivity rock matrix. Bulk hydraulic conductivity of the rock mass is therefore a complex function of matrix conductivity, fracture density and transmissivity, and fracture network interconnectedness. Further parameters of the fractures and bulk properties of the formation are currently under investigation. Areas of relatively low fracture density and interconnectedness, such as the shale-rich base of the formation, do not readily transmit head changes, and may account for the regional confining properties of the Maquoketa Formation. In contrast, the upper fractured 100 ft of the Maquoketa Formation have a good hydraulic connection to the overlying Silurian aquifer via this fracture network.

These results have significant implications for the role of the Maquoketa confining unit in the regional groundwater flow system. Although at large scales, the shale-rich base of the formation provides an effective confining unit, the upper part is hydraulically coupled to the overlying Silurian aquifer. This suggests that it is not a good assumption that the top of the Maquoketa Formation is an effectively “impermeable” or no-flow boundary to the Silurian aquifer. In fact, groundwater contamination of the overlying Silurian aquifer, particularly due to DNAPL spills, is likely to migrate downward in the fractured top of the Maquoketa Formation, a situation that would be impossible to effectively remediate.

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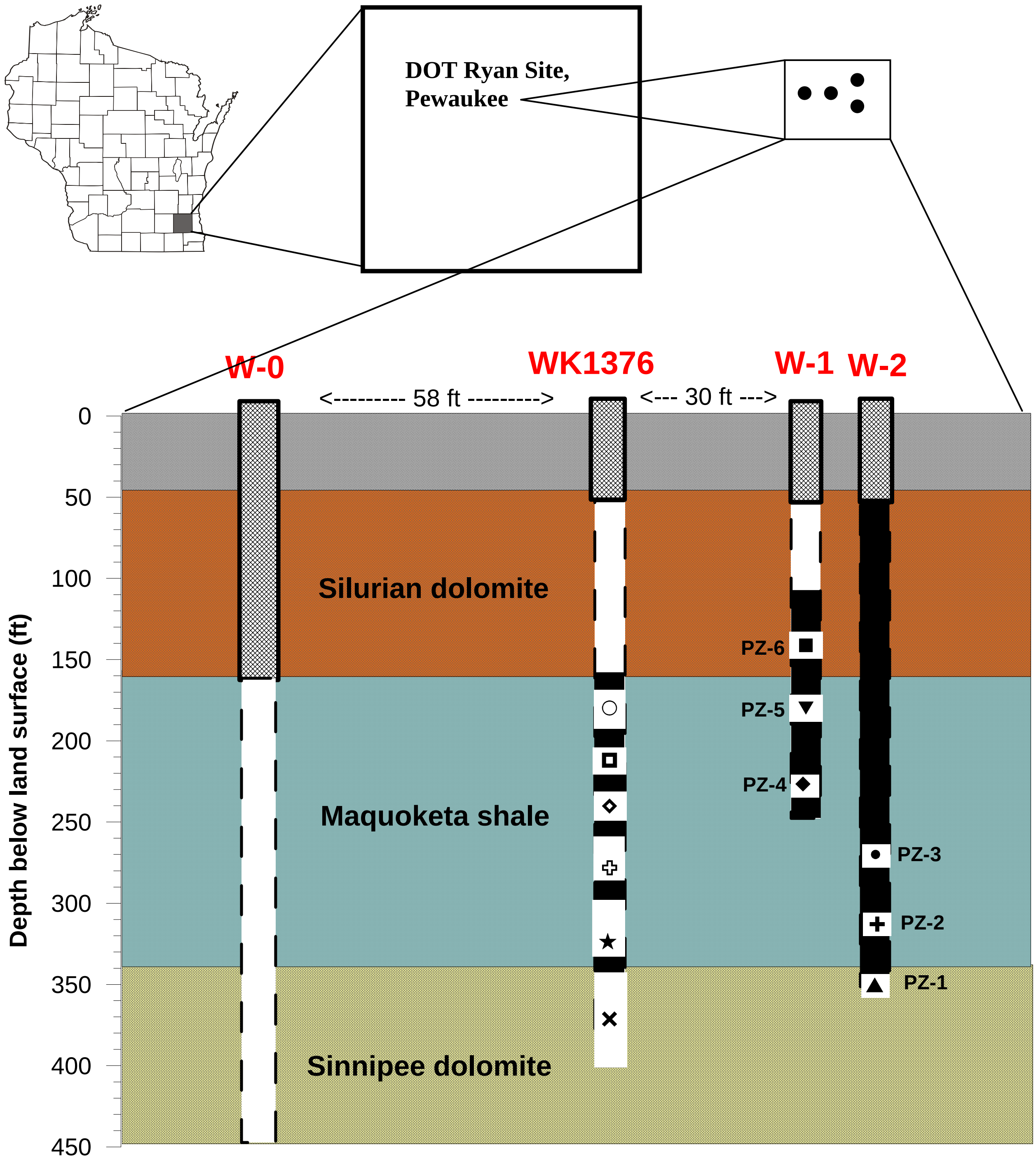
X. FIGURES

WAUKESHA COUNTY



NOTE: Symbols correspond to head data plotted in Figures 6a, 14 and 15
Intervals enlarged in well W-1 and well W-2 for clarity

Figure 1: Field site configuration showing location and elevation of monitored intervals during pumping tests

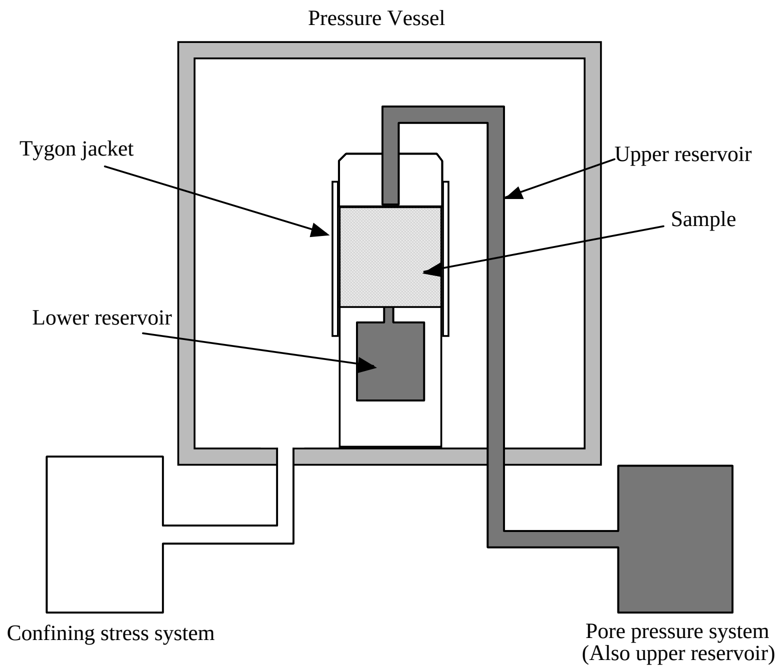


Figure 2: Diagram of the pulse decay apparatus.

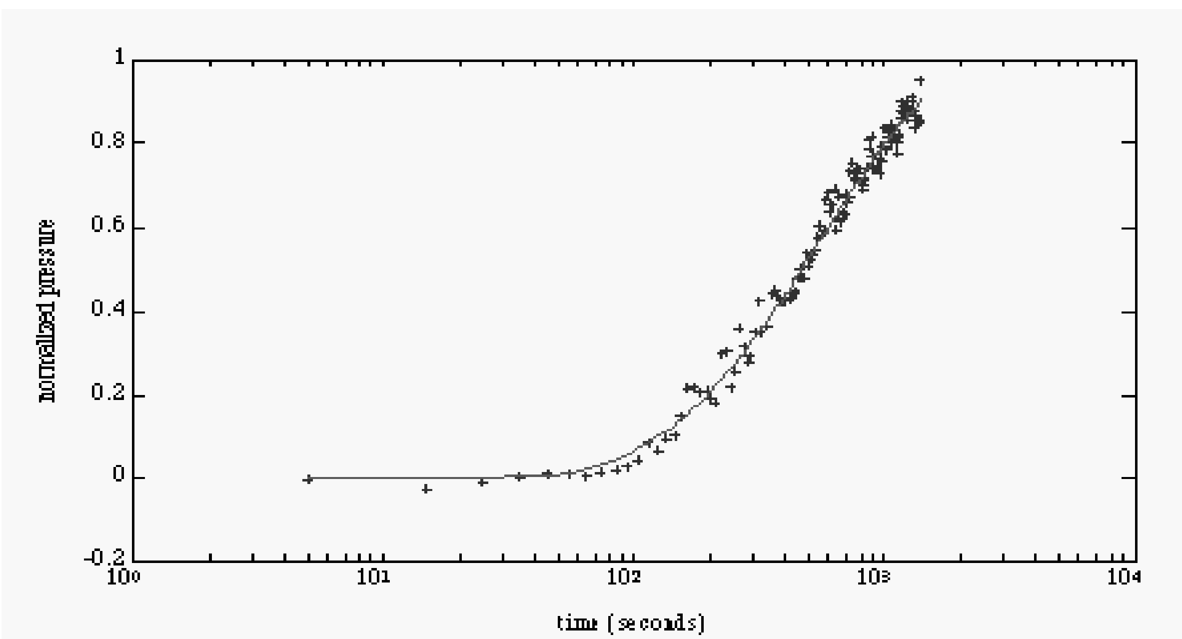


Figure 3: Normalized pressure head in the lower reservoir as a function of time for Sample MDOT-242-1

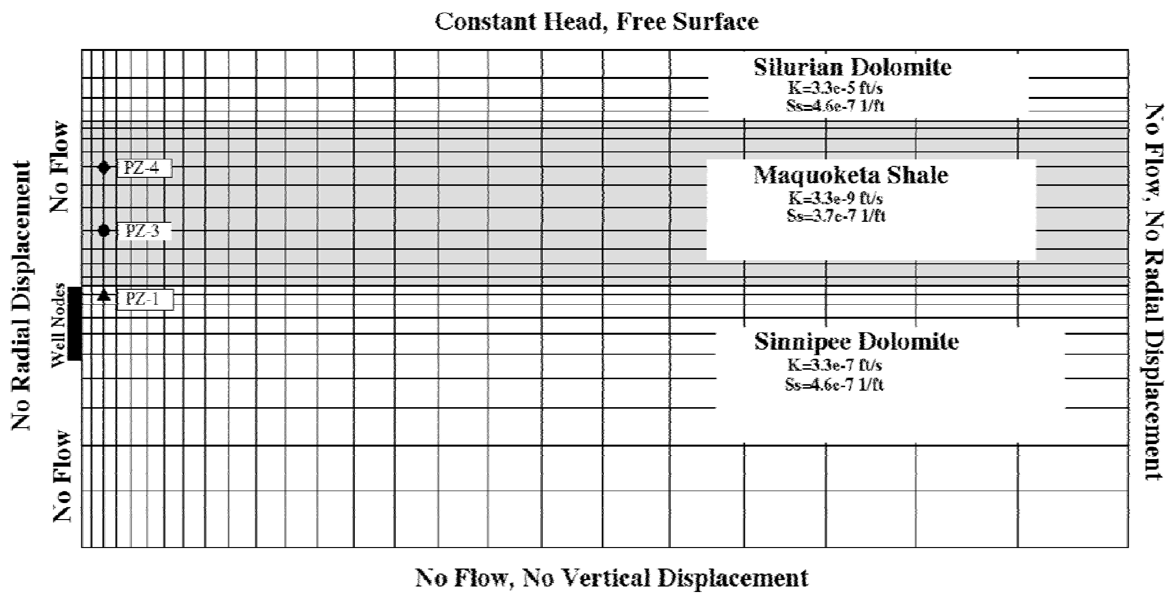


Figure 4: Finite element model mesh, boundary conditions and flow parameters. Symbols indicate locations of simulated head presented in Figure 6b.

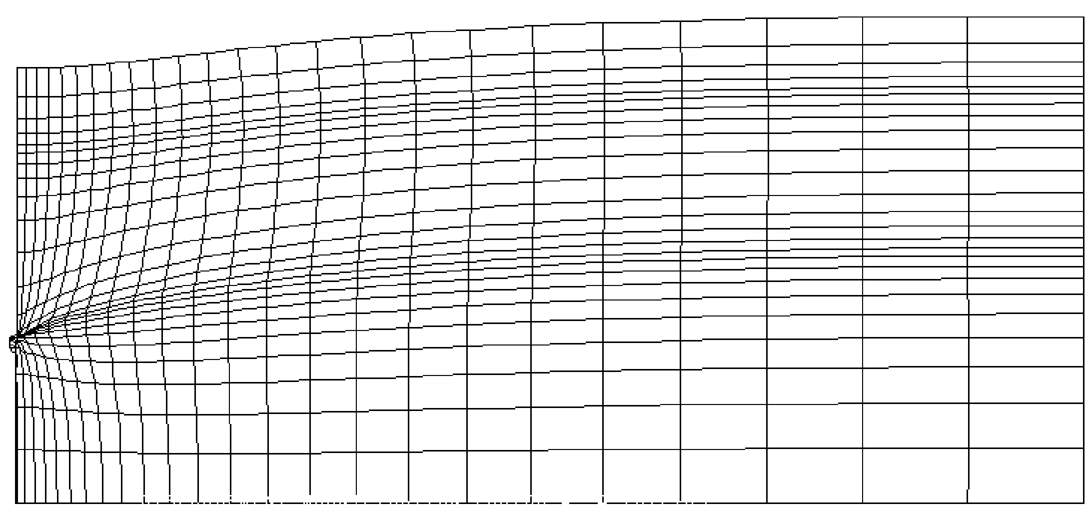


Figure 5: Deformed finite element mesh at 16 hours. Displacement exaggerated 2.5×10^5 times.

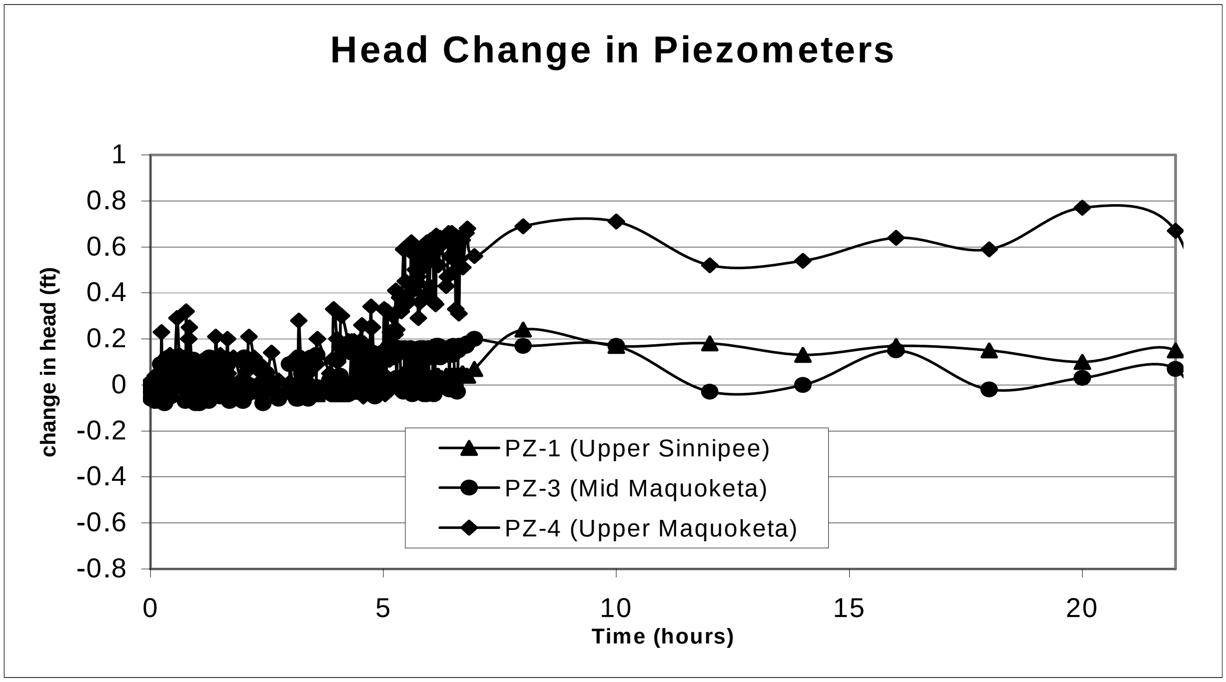


Figure 6a: Head variation in selected monitoring intervals in the Maquoketa and Sinnipee formations after head drop in the Sinnipee caused by sample pumping in well WK1376.

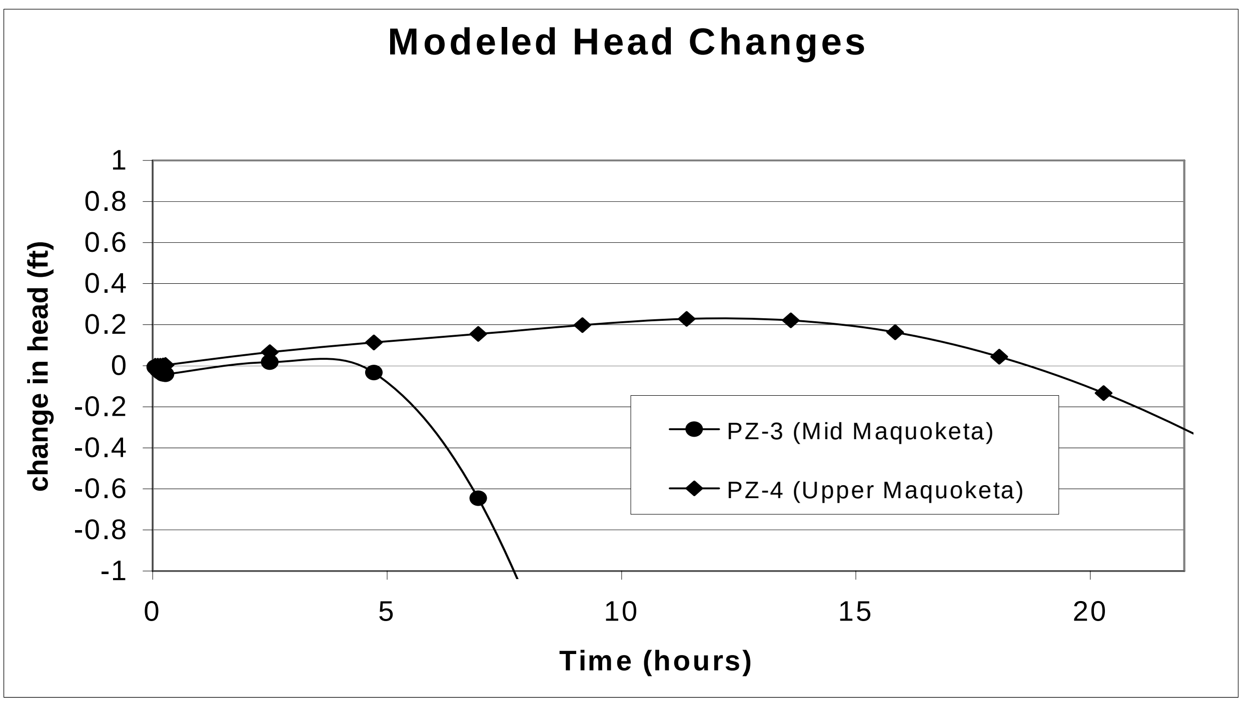


Figure 6b: Modeled head changes in the Maquoketa and Sinnipee formations due to a step decrease in the Sinnipee Group Dolomite. Head change not shown for PZ-1 at this scale.

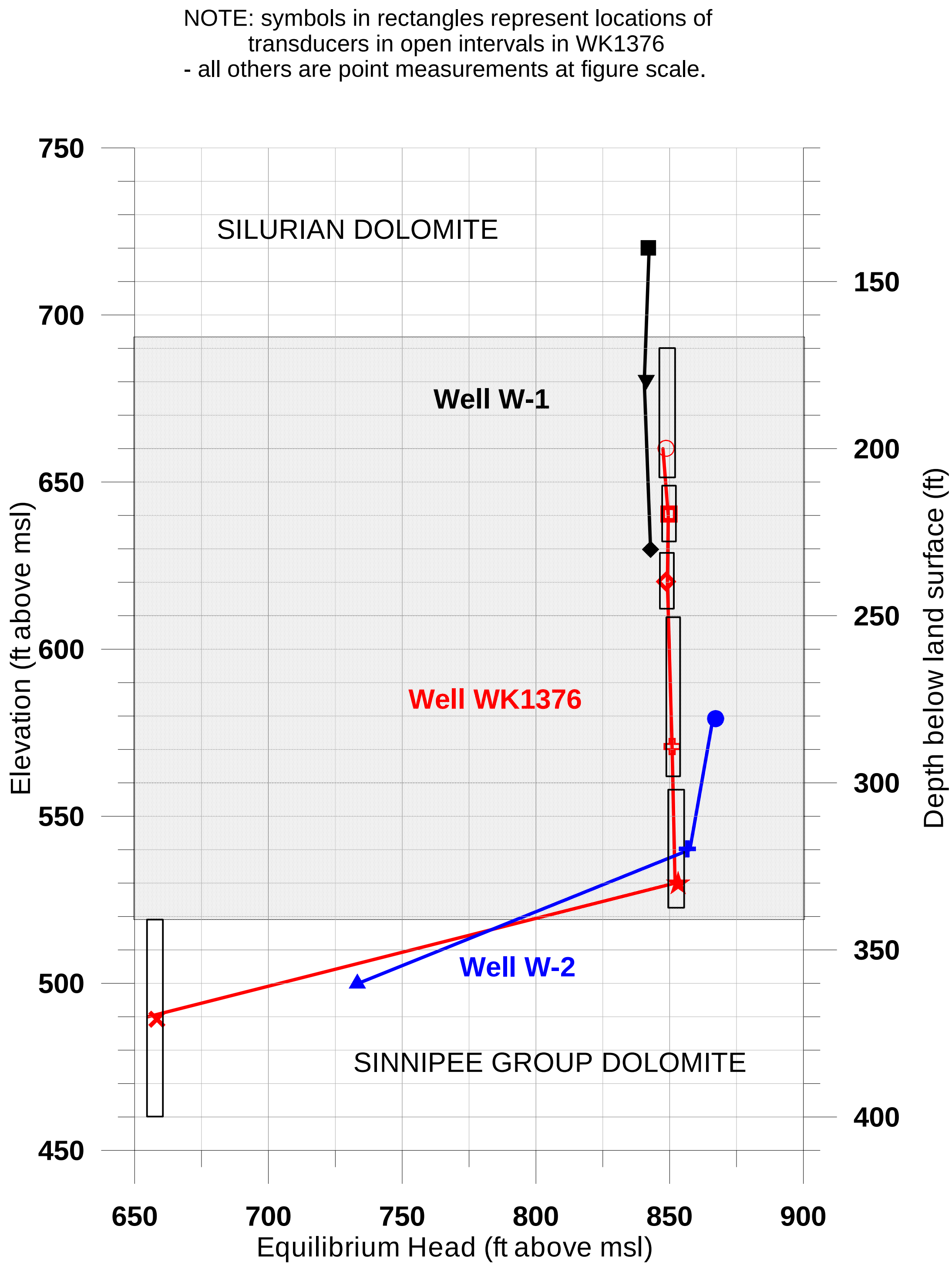


Figure 14: Equilibrium head throughout the Maquoketa Formation (stippled) in different monitoring wells. Symbols correspond to locations on Figure 1.

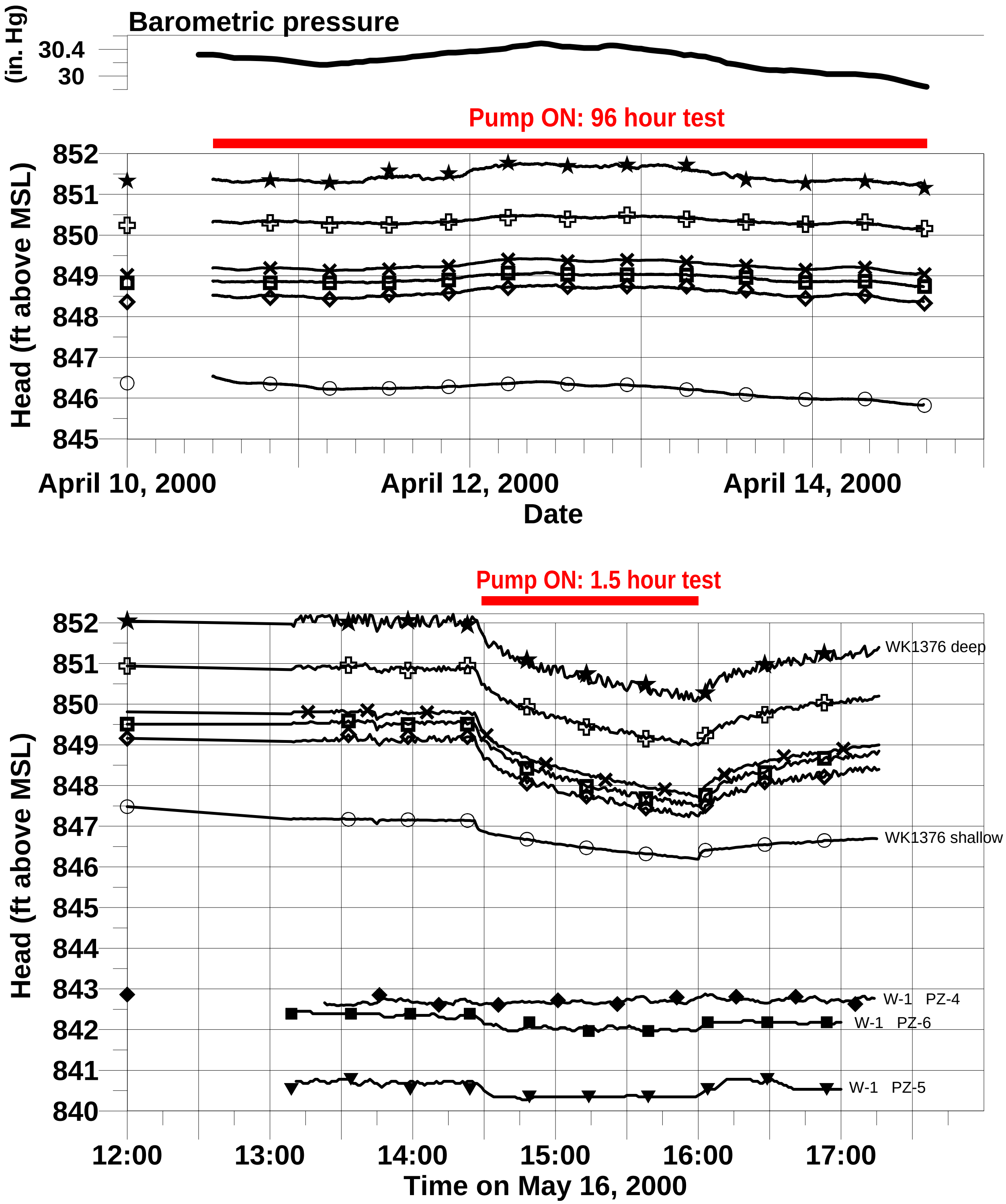
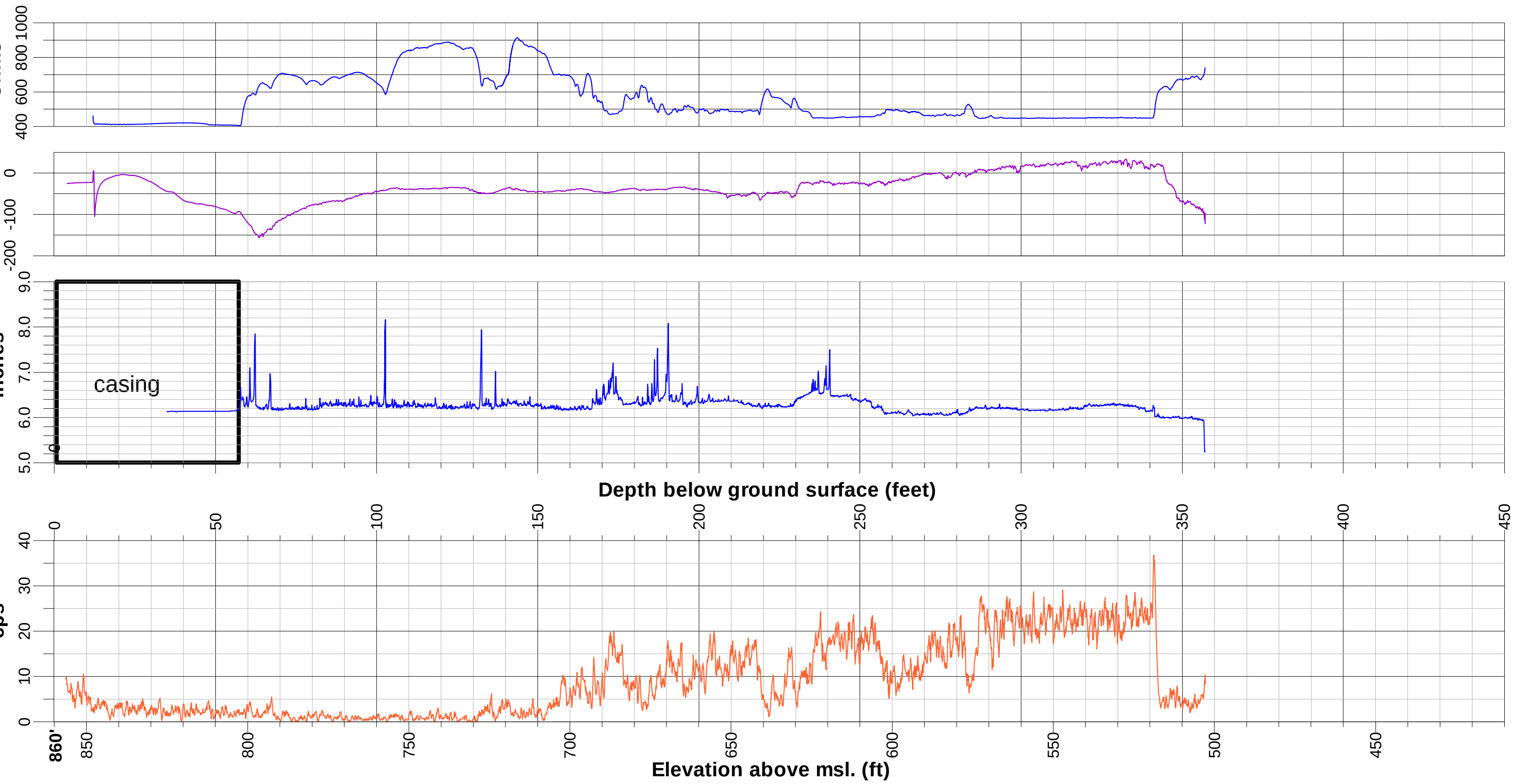


Figure 15: Selected observation well heads during deep aquifer (top) and shallow aquifer (bottom) testing. Symbols correspond to intervals located in Figure 1. No significant head change was observed in monitoring points not shown.

Figure 7: Geophysical Log
Deep observation well W-2
DOT Ryan Site, Pewaukee
T7N, R19E, section 16, NE1/4, NE1/4, SE1/4, NE1/4

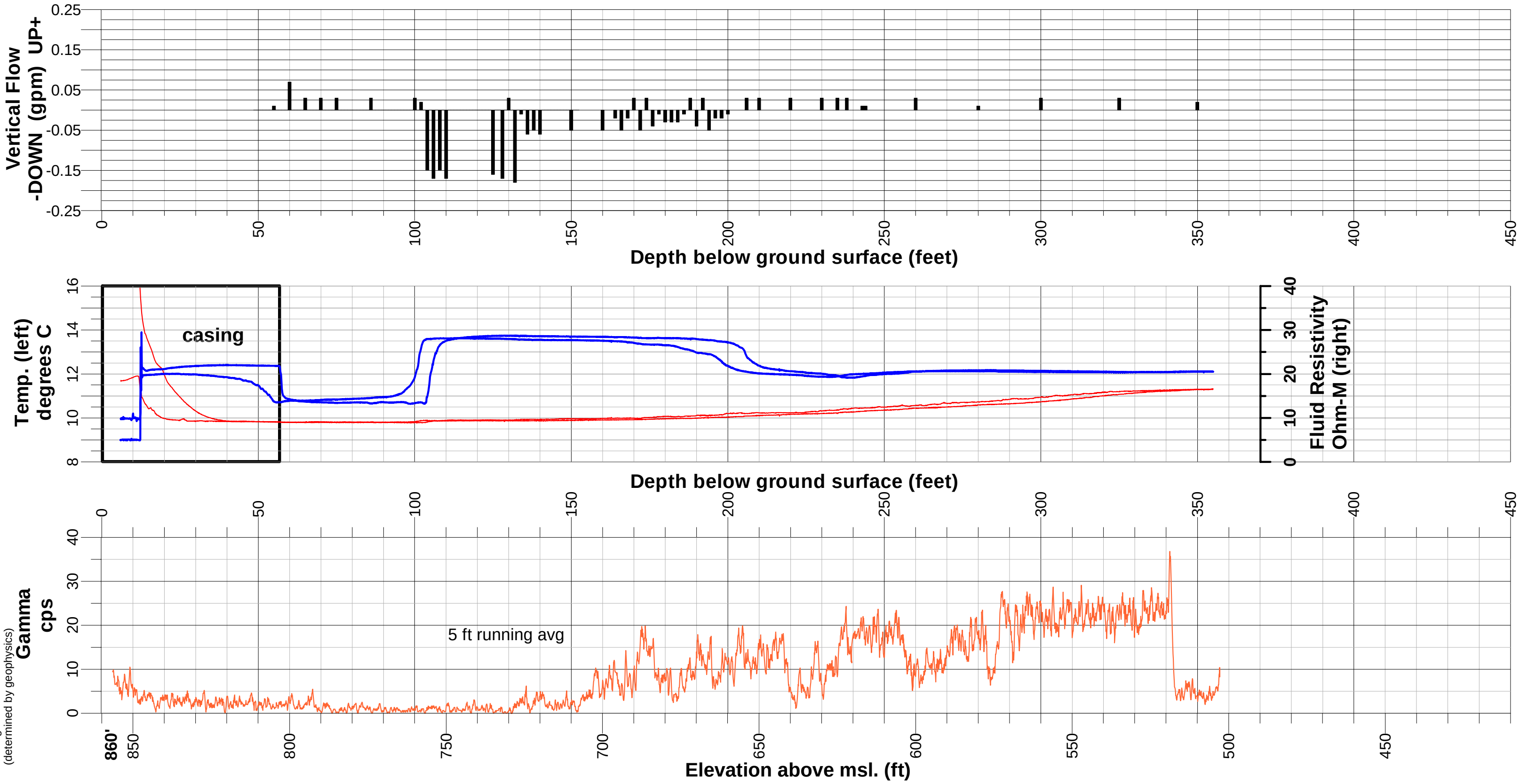


Wisconsin Geological and Natural History Survey
Geophysical Log
Files from Mt Sopris digital logger
Logged 9/24/99, 10/1/99 by T.Eaton

Elevation: 860 ft. from 7.5 min topo
Depth to water: 12 feet
Casing: 55 ft
(determined by geophysics): ft

Figure 8: Geophysical Log
Deep observation well W-2
DOT Ryan Site, Pewaukee

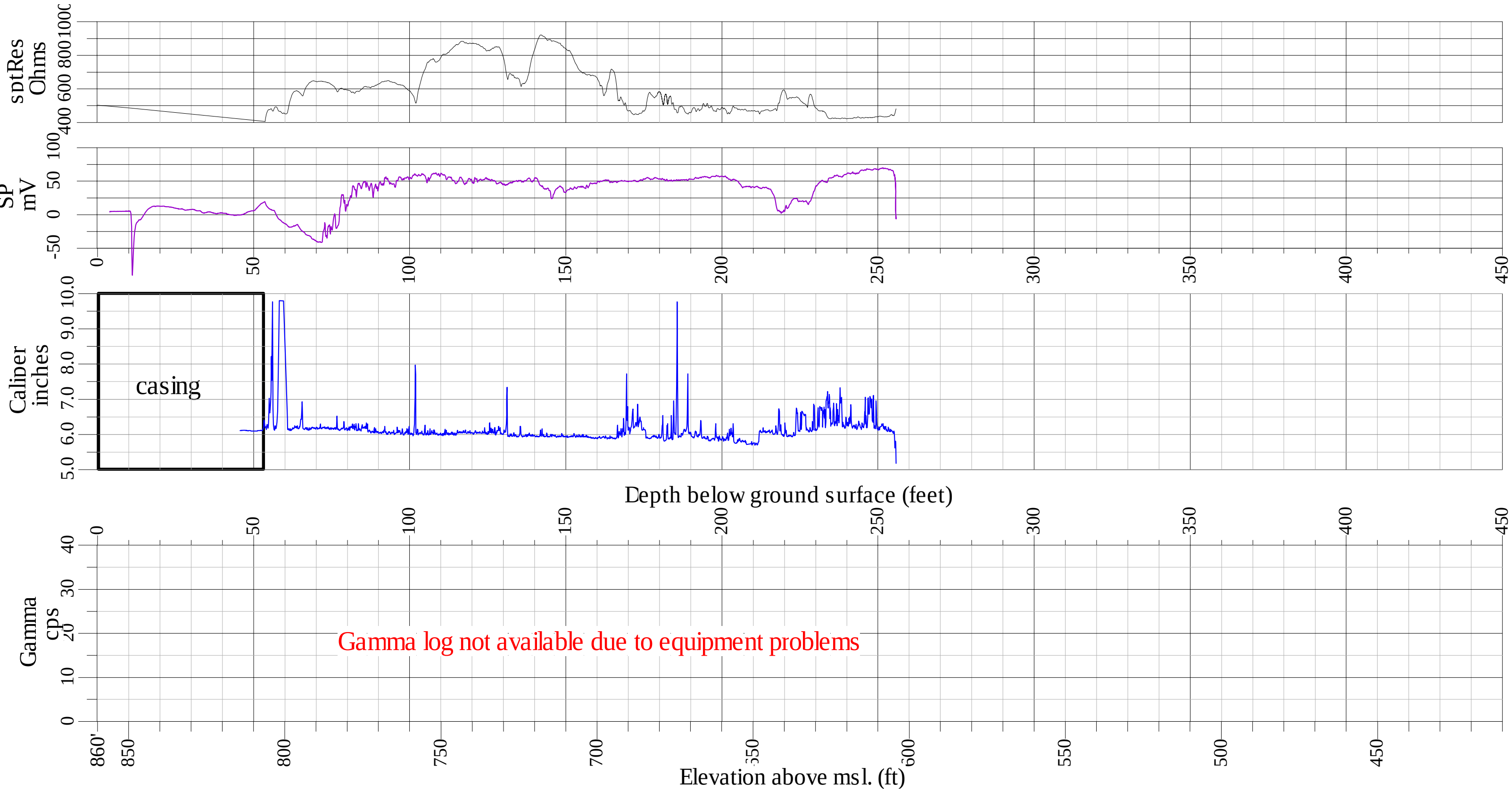
T7N, R19E, section 16, NE1/4, NE1/4, SE1/4, NE1/4



Wisconsin Geological and Natural History Survey
Geophysical Log
Files from Mt Sopris digital logger
Logged 9/24/99 by T.Eaton

Elevation: 860 ft. from 7.5 min topo
Depth to water: 12 feet
Casing: 55 ft
(determined by geophysics)

Figure 9: Geophysical Log
Shallow observation well W-1
DOT Ryan Site, Pewaukee
T7N, R19E, section 16, NE1/4, NE1/4, SE1/4, NE1/4



Wisconsin Geological and Natural History Survey
Geophysical Log
Files from Mt Sopris digital logger
Logged 3/9/00 by T.Eaton

Elevation: 860 ft. from 7.5 min topo
Depth to water: 11 feet
Casing: (determined by geophysics): 53 ft

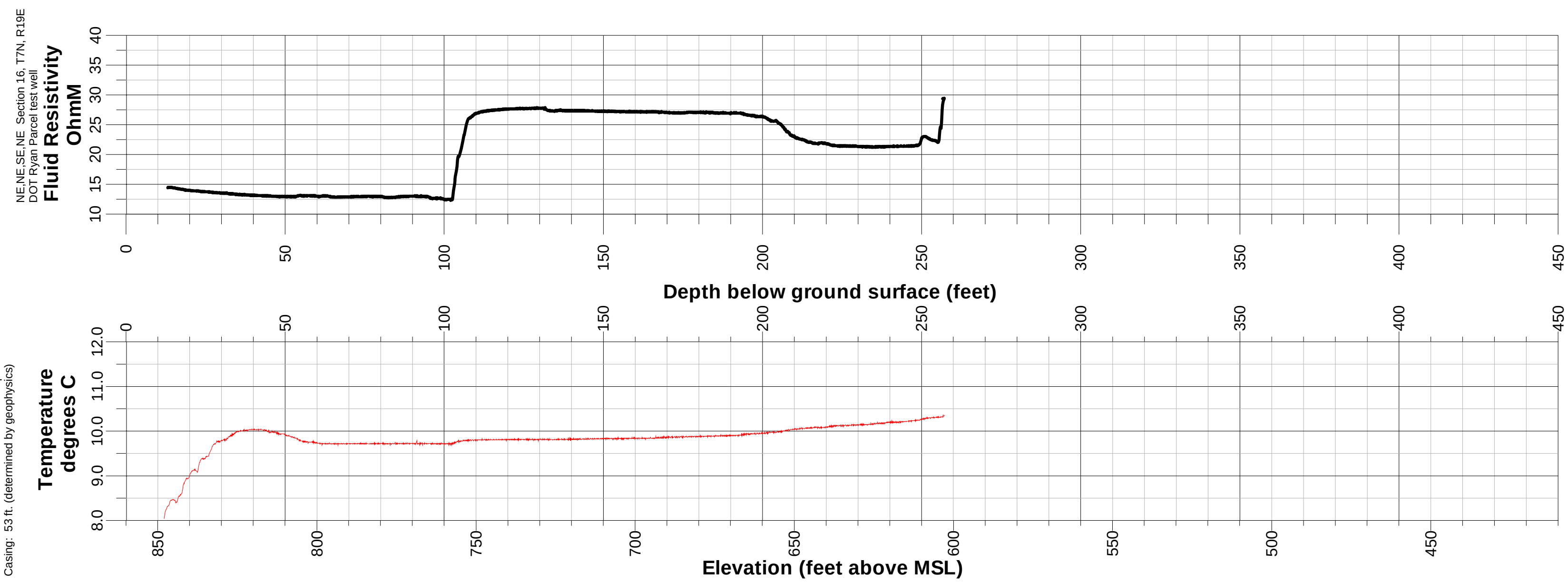


Figure 10: Geophysical Log
Shallow observation well W-1
DOT Ryan Site, Pewaukee

Wisconsin Geological and Natural History Survey
Geophysical Log
Mt. Sopris MGX 1000C digital logger
Logged 3/10/00 by T. Eaton

Wisconsin Geological and Natural History Survey
Geophysical Log
Files from Mt Sopris digital logger
Logged 12/23/99 by T.Eaton

Elevation: 860 ft. from 7.5 min topo
Depth to water: 15 feet
Casing:
(determined by geophysics): 165 ft

Figure 11: Geophysical Log
Deep pumping well W-0
DOT Ryan Parcel

T7N, R19E, section 16, NE1/4, NE1/4, SE1/4, NE1/4

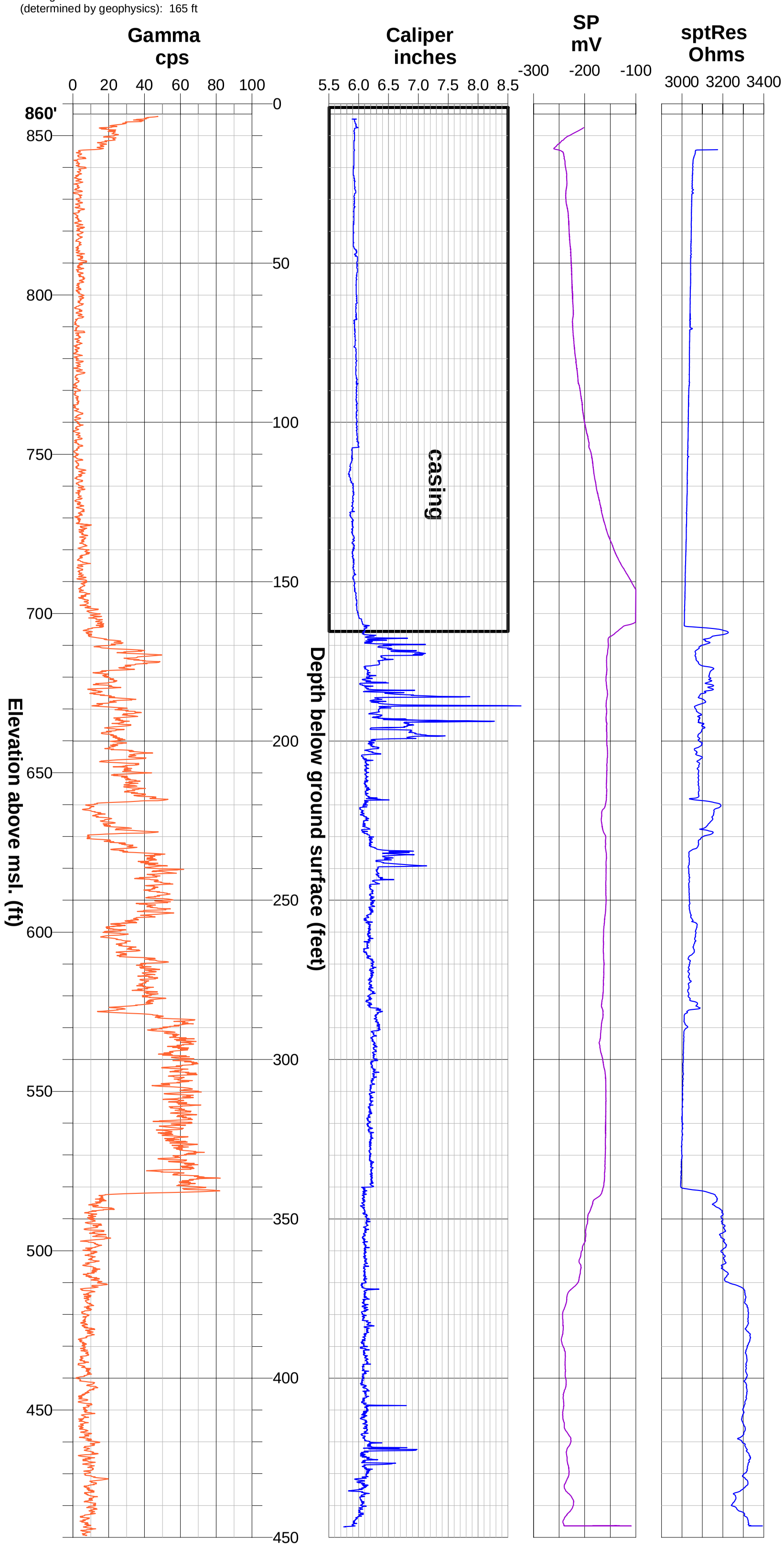
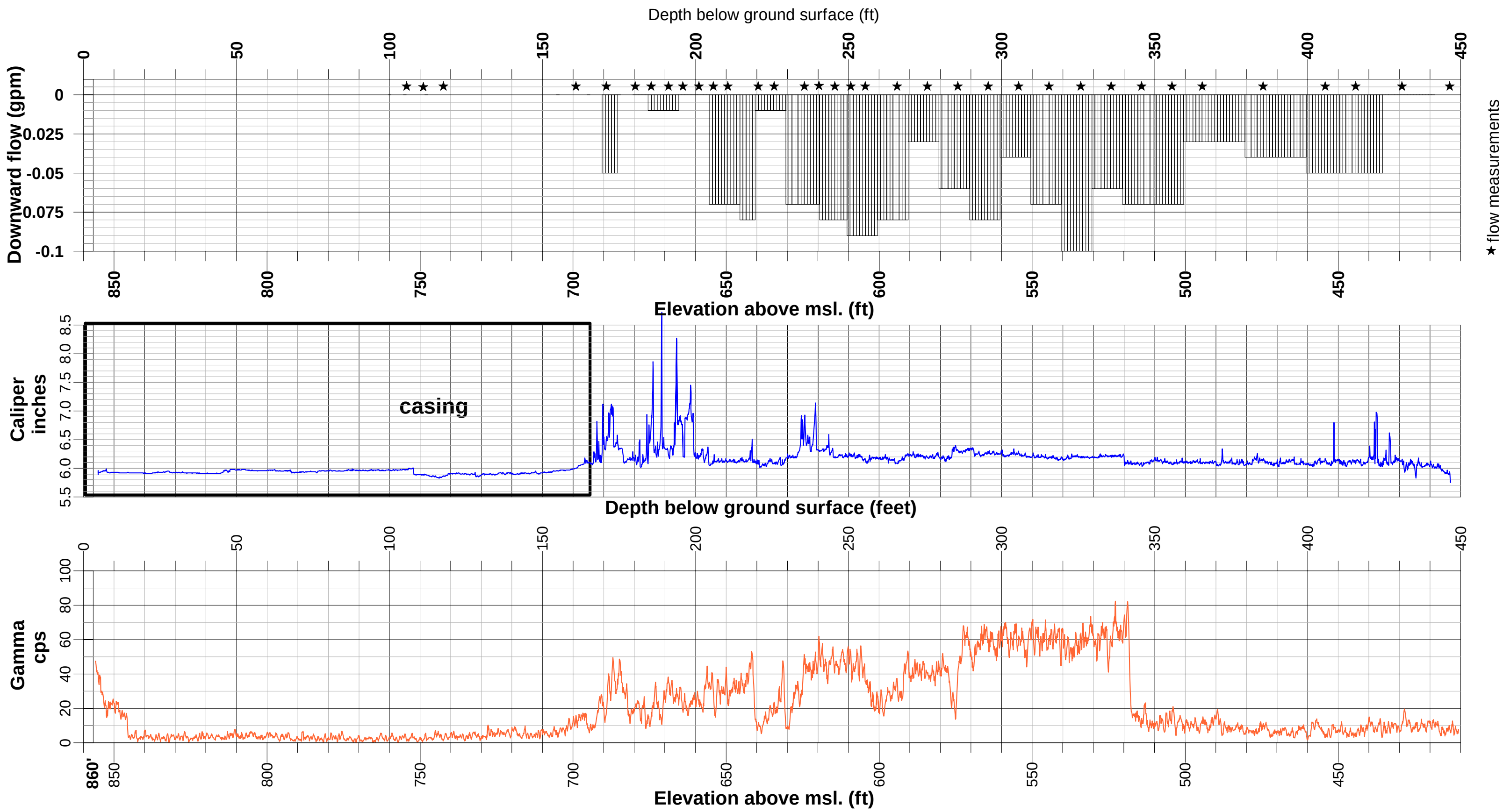


Figure 12: Geophysical log
Deep Pumping Well W-0
DOT Ryan Parcel

T7N, R19E, section 16, NE1/4, NE1/4, SE1/4, NE1/4



Wisconsin Geological and Natural History Survey
Geophysical Log
Files from Mt Sopris digital logger
Logged 12/23/99 by T.Eaton

Elevation: 860 ft. from 7.5 min topo
Depth to water: 15 feet
Casing: 165 ft
(determined by geophysics)

Wisconsin Geological and Natural History Survey
Geophysical Log
Mt. Sopris MGX 1000C digital logger
Logged 12/23/99 by T. Eaton

Elevation: 860 ft a.s.l. based on 7.5' topo
Casing: 165 ft. (determined by geophysics)

Figure 13: Geophysical Log
Deep Pumping Well W-0
DOT Ryan Site, Pewaukee

NE,NE,SE,NE Section 16, T7N, R19E
DOT Ryan Parcel test well