Wisconsin Geological and Natural History Survey
3817 Mineral Point Road
Madison, Wisconsin 53705-5100
TEL 608/262.7389 FAX 608/262.8086
www.uwex.edu/wgnhs/

James M. Robertson, Director and State Geologist

Investigating Groundwater recharge to the Cambrian-Ordovician Aquifer through fine-grained glacial deposits in the Fox River Valley

Final Report to the Wisconsin Department of Natural Resources;includes 2-page summary
Submitted September 18, 2008

Thomas S. Hooyer
David J. Hart
Kenneth R. Bradbury
William G. Batten

2008

Open-File Report 2008-07

48 p. [6 color]

This report represents work performed by the Wisconsin Geological and Natural History Survey or colleagues ans is released to the open files in the interest of making the information readily available. This report has not been edited or reviewed for conformity with Wisconsin geological and Natural History Survey standards and nomenclature.

Title:
Investigating groundwater recharge to the Cambrian-Ordovician Aquifer through fine-grained glacial deposits in the Fox River valley.

Project I.D.: *Agency no., PO NMH00000101; U.W-Extension Award Number, EXT100141*

Investigators:

Principal Investigators
T.S. Hooyer, Geologist; D.J. Hart, Hydrogeologist; and K.R. Bradbury, Hydrogeologist.
All at Wisconsin Geological and Natural History Survey, University of Wisconsin–Extension, 3817 Mineral Point Road, Madison, Wisconsin, 53705
D.M. Mickelson, Geologist
University of Wisconsin–Madison, Department of Geology and Geophysics, 1215 West Dayton Street, Madison, Wisconsin, 53706

Project Assistant*:*
C.A. Moeller, Graduate Student, University of Wisconsin –Madison, Department of Geology and Geophysics, 1215 West Dayton Street, Madison, Wisconsin, 53706

Period of Contract: *Beginning 07/01/07 through 06/30/08*

Background/Need:
Many municipalities, industries, and private residents in the Fox River valley rely on groundwater pumped from carbonate and sandstone rocks known as the Cambrian–Ordovician aquifer system. West of the Fox River, these rocks are overlain by glacial sediment consisting of silt and clay that limit infiltration of surface water to recharge bedrock aquifers. Because the thickness of this fine-grained sediment varies dramatically over the region, from a few to a hundred meters, surface water infiltration is thought to occur where the sediment is thin (LeRoux, 1957; Krohelski, 1986; Batten and Bradbury, 1996). An alternative idea is that the groundwater recharge occurs much further west of the Fox River valley where the overlying sediment is coarser-grained and thus is more conductive to surface water infiltration.

Objectives:
The objective of this study was to evaluate vertical groundwater flow through various thicknesses of fine-grained glacial deposits to assess recharge to bedrock aquifers that supply water to municipalities, industries and private residences.

Methods:
A series of boreholes were drilled in Outagamie County where the thickness of fine-grained sediment varied. Samples of core collected during drilling were analyzed in a consolidometer to determine the physical properties of the sediment including preconsolidation stress, hydraulic diffusivity, specific storage and hydraulic conductivity. Pore water was also extracted from core samples and analyzed for oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$) isotopes to determine the relative age of the groundwater. Nested wells were installed in the boreholes to (a) measure water levels to determine hydraulic gradient, (b) perform slug tests to obtain hydraulic conductivity values, and (c) obtain water samples for major ion and tritium analyses to fingerprint the origin of water.

Results and Discussion:
A total of four nested wells were installed across varying thicknesses of fine-grained sediment in Outagamie County. Two deep wells, RS-17 and RS-18, were installed in boreholes drilled to over 90 m (~300 ft) deep. The remaining two wells were shallow and were installed on two local farms were the sediment was about 15 m (50 ft) thick. The cores retrieved from the boreholes show that fine-grain sediment consisting of silt and clay dominates the geology. Consolidation testing of intact samples collected from the drilling core revealed hydraulic conductivity values that ranged from 1×10^{-8} ms^{-1} to 5×10^{-14} ms^{-1}. These values typically decreased exponentially as a function of applied load that is related to burial depth. This decrease is a result of the overlying weight of the sediment thus reducing effective porosity and permeability. The data revealed that within the first 15 m of the ground surface there is about an order of magnitude reduction in hydraulic conductivity, whereas it takes at least at least 60 m

(~200 ft) of burial to reduce hydraulic conductivity another order of magnitude. These results show that the thicker the fine-grained sediment the less likely it is to transmit water.

Hydraulic conductivity values determined from slug tests are similar to values obtained from the consolidation testing. In addition, they also show a decrease in hydraulic conductivity with depth. Using the values from the slug tests combined with long-term water level data, a vertical flux of 6 mma^{-1} (0.24 ina^{-1}) was calculated where the sediment was over 24 m (80 ft) thick. A much greater vertical flux of 163 mma^{-1} (6.4 ina^{-1}) was calculated where the sediment was much thinner on the farms.

The results of the tritium, major ion and $\delta^{18}O$ and $\delta^{2}H$ isotopic analyses on water samples also indicates that where the fine-grained sediment is thin, it appears to have a distinct chemical signature that may be representative of modern day infiltration. Where the sediment is thick, the water has a distinct chemical signature and may be relatively old (glacial age?). Use of a one-dimensional advection dispersion model to fit the $\delta^{18}O$ isotopic profiles collected at the deep boreholes showed that diffusion dominates with some minimal, but discernable contribution from advection.

Conclusions/Implications/Recommendations:

The fine-grained glacial deposits in the Fox River Valley limit surface water infiltration and groundwater recharge to bedrock aquifers. However, the thickness of the sediment is a critical variable in understanding groundwater recharge areas. This study has shown that thick sequences of fine-grained sediment (>30 m) prevent downward migration of modern precipitation to bedrock aquifers. However, where the sediment is thinner (<15 m), the bedrock aquifers are much closer to the surface and are susceptible to groundwater recharge. Potential recharge areas in Outagamie County were identified where the sediment is less than 15m thick. Using an average flux of 163 mma^{-1}and assuming that approximately 20% (206 km^2 or 128 m^2) of the bedrock area is within 15 m of the ground surface, results in over 1.8 x 10^8 ld^{-1}(40 x 10^6 gal. d^{-1}) of recharge. Approximately half of this area, 206 km^2 (64 m^2), is west of the Sinnipee Group subcrop and recharges the deep bedrock aquifer. Some recent work in the lower Fox River valley indicates that around 3.6 x 10^7 ld^{-1}(8 x 10^6 gal. d^{-1}) were pumped from municipal wells in 2007. Although recharge to the deep aquifer appears to be at least twice the pumping rate, a groundwater flow model is necessary to further understand the dynamics of the aquifer system in the Fox River valley.

Potential areas of recharge to the aquifer in other parts of the Fox River Valley could be identified using this technique. Even though many municipalities have been switching to surface water for drinking, understanding the spatial distribution of recharge in this region is still crucial for planning the future management of groundwater quality and quantity.

Related Publications (abstracts):

1. *Hooyer, T. S., Hart, D.J., Moeller-Eaton, C.A., and Batten, W.G., 2008. Vertical distribution of $\delta^{18}O$ in a clay-rich aquitard: Implications for groundwater recharge. American Water Resources Association Conference, Wisconsin Chapter, Brookfield, WI.*
2. *Moeller, C.A., Mickelson, D.M., Hooyer, T.S., Hart, D.J., and Batten, W.G., 2008. An analysis of hydraulic conductivity with depth and stress in a clay aquitard. American Water Resources Association Conference, Wisconsin Chapter, Brookfield, WI.*

Key Words:

Aquitard, groundwater recharge, hydraulic conductivity, oxygen isotopes, and Fox River valley.

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 Library, University of Wisconsin–Madison, 1975 Willow Drive, Madison, Wisconsin 53706; (608) 262-3069.

Investigating Groundwater recharge to the Cambrian-Ordovician Aquifer through fine-grained glacial deposits in the Fox River Valley

Final Report to the Wisconsin Department of Natural Resource
Submitted September 18, 2008

Thomas S. Hooyer, David J. Hart, Kenneth R. Bradbury and William G. Batten

INTRODUCTION

Unlithified glacial sediment covers a large part of the landscape along the Fox River valley in east-central Wisconsin. Most of this sediment consists of silt and clay which can limit infiltration of surface water to recharge important bedrock aquifers in the region. The thickness of this fine-grain sediment varies dramatically over the region, from a few to a hundred meters, and surface water infiltration is thought to occur only where the sediment is thin (LeRoux, 1957; Krohelski, 1986; Batten and Bradbury, 1996). To test this idea, a series of multilevel (nested) monitoring wells were installed at various locations in a well-defined area in Outagamie County where the thickness of fine-grained sediment varies. Data collected from the wells, including hydrogeologic and chemical analyses of groundwater samples, reveal that where the sediment is thick (>100 feet, ~30 m) there is virtually no infiltration of surface water to deep bedrock aquifers. However, where the sediment is thin (<50 feet, ~15 m), infiltration of surface water does occur and may be responsible for the majority of recharge to bedrock aquifers. This result also has implications for land use planning issues such as landfill siting and manure management. In this report we present the results of our study and use depth to bedrock geology data to identify potential areas of groundwater recharge in Outagamie County.

Geologic setting

A sequence of Cambrian and Ordovician sandstone and dolomites overlie crystalline Precambrian rock along the Fox River valley in east-central Wisconsin. These sandstones and dolomites vary in thickness and dip slightly eastward towards the Lake Michigan basin (Figure 1). The bottom-most layer consists of Cambrian sandstones that are overlain by the Ordovician Prairie du Chien, Ancell, and Sinnipee Groups. Both the Prairie du Chien and the Sinnipee Groups are primarily composed of dolomite, whereas the Ancell Group consists of sandstone of the St. Peter Formation (Batten and Bradbury, 1996). To the east of Lake Winnebago and the Fox River valley, younger bedrock units consisting primarily of the Ordovician Maquoketa shale and Silurian dolomite caps this entire bedrock sequence. To the west of the Fox River and beyond the subcrop of Silurian dolomite, the entire bedrock sequence is overlain by Quaternary-age glacial sediment that was deposited in glacial Lake Oshkosh that existed in front of the Green Bay Lobe of the Laurentide Ice Sheet during the last glaciation (Figure 2). This sediment varies in thickness across the landscape as a result of an undulating bedrock surface.

Hydrogeological setting

There are two bedrock aquifers in the Fox River valley. The first and most productive aquifer includes the deep Cambrian sandstone, the Prairie du Chien Group and the St. Peter sandstone (Figure 1). This aquifer is confined by the bottom, unweathered part of the Sinnipee Group. The second aquifer, used mainly by residential water-supply wells, is located in the upper, highly fractured dolomite of the Sinnipee Group and any local, unlithified sandy sediment that persists adjacent to the bedrock surface. This upper aquifer is typically not considered to be confined although it is usually covered with fine-grained glacial sediment in the Fox River valley.

The confining unit at the bottom of the Sinnipee Group limits vertical groundwater recharge to the lower, most productive aquifer where the recharge is understood to occur to the west where the aquifer subcrops (Krohelski, 1986; Batten and Bradbury, 1996; Conlon, 1998). However, the fine-grained glacial sediment that covers that area is generally greater than 50 feet (15 m) thick and may also act as a regional aquitard. The presence of this aquitard would shift the areas of recharge to the lower aquifer even further to the west beyond Outagamie County. Given the increasing demand on groundwater resources, it is important to understand how and where these aquifers are being recharged especially where they are covered by the sequence of fine-grained glacial sediment.

Study location

This study was conducted in Outagamie County primarily because (1) it is potentially an important recharge area for the deep sandstone aquifer, and (2) the landscape is covered with varying thicknesses of fine-grained glacial sediment. In addition, the geology is fairly well known because the Wisconsin Geological and Natural History Survey (WGNHS) is independently mapping the geology of the county including the bedrock and Quaternary-age deposits. As part of this mapping project, a series of boreholes were drilled in a buried bedrock valley that extends into Outagamie County (Figure 3). In two of these boreholes, RS-17 and RS-18, two nested wells were installed where the glacial sediment is thick (>300 feet, ~91 m). During this past year, two additional nested wells were installed on the Riehl and Lorenz farms to evaluate recharge to the deep aquifer where the glacial sediment is relatively thin (<50 feet, ~15 m).

As part of this study, water samples were collected from the municipal wells at the Villages of Black Creek and Shiocton, and a flowing well located on the Van Straten Farm. Water samples were analyzed at all the wells for various constituents aimed at chemically finger printing the source of water in the deep aquifer.

METHODS

Drilling, well installation and water level measurements

Of the four boreholes drilled as part of this project, two (RS-17 and RS-18) were drilled using rotosonic methods which limits the introduction of drilling fluids from the borehole into the surrounding formation. This method allows the collection of continuous 4-inch diameter core that remains intact for cohesive sediment such as clay and silt. The rotosonic drilling process includes advancing an 8-inch diameter casing as sediment core is collected, thus providing hole for the installation of nested wells. The other two boreholes, located on the Riehl and Lorenz farms, were drilled using the hollow-stem auger method that also limits the introduction of drilling fluids from the borehole into the surrounding formation. Although continuous samples are more difficult to collect, monitoring wells are easily installed though the augers once the borehole is completed. All core collected from the boreholes was described and sampled in the laboratory for various analyses.

At borehole RS-17, a Solinst CMT® (continuous multichannel tubing) multilevel well was installed (Einarson and Cherry, 2002). Predetermined 6-inch long ports were cut into the CMT tubing at ground surface and wrapped in wire screen before it was lowered into the borehole inside the 8-inch diameter steel casing. This casing was pulled back in 10-foot increments taking precaution to backfill with 5-foot thick layers of sand around the screened ports and separating these intervals with coated bentonite pellets that expand once hydrated, effectively sealing the borehole annulus.

The multilevel wells installed in borehole RS-18 consisted of seven 1-inch diameter PVC wells with 2.5-foot long screens. These wells were also installed through the center of the drill casing at predetermined intervals. The screens were surrounded with 5 to 6 feet (2 m) of sand. In between these sand packs, the borehole annulus was also backfilled with coated bentonite pellets.

The two well nests installed at the Riehl and Lorenz farms each consisted of three 1-inch diameter PVC wells with 2.5-foot long screens. Instead of installing three wells in one borehole, three individual boreholes were drilled immediately adjacent to each other and the wells installed independently to eliminate the chance of hydraulic connection along the annulus of the borehole.

Following installation, the wells were developed using an air-lifting technique that consisted of lowering flexible nylon tubing down the well and releasing compressed nitrogen that removed the water from the well casing. This process was repeated numerous times until a minimum of three times the volume of water in the sand pack pore space was removed. The Riehl and Lorenz nested wells were also purged numerous times using a Teflon® or stainless steel bailer.

In addition to these techniques, surging was performed on the wells at RS-18. This consisted of plunging each well with a rod slightly smaller in diameter than the well. The pressure induced by this surging was aimed at removing fine particles that may have been blocking the well screen.

After well development, water levels were measured using a hand-held electric tape. At RS-18, Riehl and Lorenz wells, automatic water level measuring devices were installed to collect data at various time intervals. Prior to collecting water samples, the standing water in the well was purged. Once the wells recovered, water samples were collected using a peristaltic pump.

Chemical analyses of pore water and well water

Samples from the RS-17, RS-18 and the Riehl core were collected at regular intervals to extract pore-water for oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$) isotope analyses. This was accomplished by first extracting an intact 6-inch section of core and trimming the outermost 1-inch to remove any sediment that may have been in contact with drilling fluids. The center portion of each sample was stored in an air-tight bag, refrigerated, and eventually placed into a high pressure stainless steel cylinder for pore-water extraction. Within this cylinder, there is a piston that is driven by compressed nitrogen that squeezes the sample. This squeezing or consolidation of the sample results in a reduction of pore space thus driving any pore-water through filters and a small hole in the bottom of the cylinder and into a sample container. To verify the results of the pore-water isotope analyses, water samples were collected from the nested wells for comparison. Since no porewater was extracted from the core samples collected at the Lorenz farm, isotope analyses were only performed on the water samples collected from the three nested wells. The samples were analyzed for $\delta^{18}O$ and $\delta^{2}H$ by direct pore-water equilibration at the University of Waterloo Environmental Isotope Laboratory.

In addition to the isotope analyses, the water samples collected from the nested wells were also analyzed for tritium and major ions (Table 1). Concentration of tritium in the atmosphere increased by about three orders of magnitude in response to bomb testing that started in the mid 1950s (Gaspar, 1987). Although there has been decay in bomb-induced tritium since testing in the 1960s, modern tritium concentrations are still 5 to 10 times higher then there pre-bomb levels. Thus, groundwater samples with elevated levels of tritium indicate relatively recent groundwater recharge. Analyses for tritium were conducted at the University of Waterloo Environmental Isotope Laboratory where they were electrolytically enriched to obtain a precision of 1 TU which is about the atmospheric tritium concentration in the United States prior to testing of nuclear bombs.

Hydraulic conductivity

Consolidation testing

Intact samples were collected from selected core for consolidation testing to determine preconsolidation stress, hydraulic diffusivity, vertical hydraulic conductivity, and specific storage (Grisak and Cherry, 1975; Keller et al., 1989). In a consolidation test, an axial load is applied incrementally to a confined, fully saturated sample and the amount of vertical displacement (consolidation) is measured as function of time. After each individual increase in load, excess pore-water pressure is allowed to dissipate as the sample consolidates.

A curve of the consolidation with applied load will give an indication of the maximum stress that the sample experienced during its geological history. Under initial loads and after the pore-water pressure dissipates, consolidation is due to elastic deformation of the grain skeleton. Under some larger load, the grain skeleton will begin to collapse signifying the start of permanent deformation. The stress at which this occurs is called the preconsolidation stress, σ_{pc}, and signifies the maximum effective stress (total stress minus pore pressure) that the sample experienced in its geologic past. When the load is removed the sample does not follow the same path of deformation since the grains have been realigned by application of stresses greater than the previous maximum effective stress. These deformation curves are often shown as the change in the void ratio with increasing and decreasing load.

A typical consolidation curve is shown in Figure 4 for sediment from boring RS-12 from a depth of 59.7 to 60 feet (18 m) below ground surface. The void ratio (pore volume/solid grain volume) is used here as a measure of consolidation. As the sample is loaded the pore volume decreases and the solid volume remains the same. The result is a reduced void ratio with increasing load. Using the Casagrande graphical method (Das, 1994), we determined a preconsolidation stress of approximately 500 kPa for this sample. We can compare this stress to the overburden stress from the overlying sediment. The overburden stress, σ, was calculated by applying the following equation

$$\sigma = \rho g h$$

where ρ is the density of the sediment, g is acceleration of gravity and h is the thickness of the overlying sediment. The resulting best estimate, 350 kPa, is close to preconsolidation stress and within the range of possible preconsolidation stresses between 150 kPa and 1000 kPa, respectively, determined by the Casagrande method. This result suggests that the entire mass of the glacial ice was not carried by the sediment but that significant pore fluid pressure reduced the effective stress to a value close to today's overburden stress.

During an application of any given axial load, there is a reduction in the sample volume (pore space) that occurs at a rate that is dependent on the expulsion of pore water. Given this relationship, the hydraulic diffusivity, defined as hydraulic conductivity divided by specific storage, can be calculated for each loading increment using the square root of time method for a semi-infinite length cylinder (Terzaghi, 1943; Wang, 2000; Hart and Hammon, 2002). The consolidation or vertical displacement as a function of time, $\Delta w(t)$, of the sample due to the draining of excess fluid pressure is:

$$\Delta w(t) = 2 c_m \gamma \sigma_z \sqrt{\frac{Dt}{\pi}}$$

where c_m is vertical compressibility of the sediment, γ is loading efficiency, σ_z the axial load, and D the hydraulic diffusivity. Using this relationship, the equation is solved for hydraulic diffusivity, D, by using the rate of consolidation, $\Delta w(t)$, for each incremental load. Figure 5 shows displacement with increasing time for sample RS-12, 59.7-60.0 feet (18 m). The entire test was conducted for around 100,000 seconds (~ 28 hours). When using the square-root time method, it is necessary to neglect longer time data. In this example, the calculated displacement matches the measured displacement up to 100 seconds because the assumption of a semi-infinite cylinder still applies. After that time, the pore pressure has diffused through the entire sample and the calculated and measured displacements diverge.

To estimate the hydraulic diffusivity, the vertical compressibility, c_m, the loading efficiency, γ, and the axial load, σ_z, must either be measured or estimated. The vertical compressibility, c_m, is the strain ($\Delta l/l$) divided by the axial load. The strain is determined from the long-term displacement and the sample length. In this example the axial load of 2×10^6 Pa caused a strain of 0.032 (0.051cm/1.61 cm). The calculated vertical compressibility is 1.6×10^{-8} Pa^{-1}. The loading efficiency was assumed to be 1 for soils; a common assumption, and so was not measured. These values were used along with the displacement-time curve to estimate a hydraulic diffusivity of 1.2×10^{-8} m^2/s for the sample.

The specific storage and hydraulic conductivity can be determined from these values. The specific storage (S_s) is approximated from the results of the consolidation testing by the relationship

$$S_s = c_m \rho_f g \,,$$

where ρ_f, is the fluid density and g is gravitational acceleration. The hydraulic conductivity of the sample can then be calculated as $K = D \cdot S_s$ (Freeze and Cherry, 1979).

Slug testing

Horizontal hydraulic conductivities of the glacial sediment were measured in piezometer nests at the RS-18, Riehl, and Lorenz sites using slug tests. Slug tests were used rather than pumping tests because the glacial sediment would not yield sufficient amounts of water in most of the piezometers. In slug tests, the water level in a well or piezometer is suddenly increased or lowered by a slug or by bailing water from the well. After the slug or baildown, the water levels are out of equilibrium with the hydraulic head in the surrounding sediment. The time response of the water levels depends on the hydraulic conductivity of that sediment. For typical well constructions, the water level in a well finished in sediment with high hydraulic conductivity will reach equilibrium quickly, usually in less than an hour, while the water level in wells finished in sediment with low hydraulic conductivity material may take days or even months to reach equilibrium.

The hydraulic conductivity can be determined from curves of water-level measurements with time. Different methods, each with slightly different assumptions, are available to analyze this data. We used the commercial software, Aqtesolv, and chose the Kansas Geological Survey (KGS) solution by Hyder et al. (1994) for the analysis because it accounts for the partial penetration of the piezometers. However, similar results were also found using the Hvorslev (1951) solution. A time-water level curve and the KGS solution fit for a slug test in Port 4 in RS-18 is shown in Figure 6. This test illustrates one of the difficulties when working with low-conductivity materials like this glacial lake sediment. The time needed for these tests will take much longer than those in higher conductivity materials like sands or gravels. In this example, the water-level in the piezometer doesn't come to equilibrium with the surrounding sediment water levels until nearly 10 days after the addition of the slug.

Slug tests were performed in the nested wells located in RS-18 and the Riehl and Lorenz farms to determine the in-situ hydraulic conductivity. Because of the presumed low conductivity of the sediment in many of the wells, a known volume (slug) of de-ionized water was placed into each of the wells and water levels were measured using automatic recorders (leveloggers). In those piezometers set in coarser grained sediment, a five-foot long 3/8-inch slug was inserted into each well to displace water.

RESULTS

Geology and water level data

The lithologic logs, well port locations, and water level data for RS-17 and RS-18 are shown in Figure 7 and for the Riehl and Lorenz Farm on Figure 8. The water level data presented on these figures only represent single measurements on the dates shown. A review of the logs reveals that RS-17, located in the Village of Black Creek, was drilled to a total depth of 307 feet (94 m) penetrating 300 feet (91 m) of fine-grained sediment (silt and clay) before encountering 7 feet (2 m) of fine sand. Further to the west along the axis of the buried valley near the city of Shiocton (Figure 3a), RS-18 was 308 feet (94 m) deep and penetrated a similar sequence with the exception of sand layers near the bottom of the borehole and a 20-foot thick layer about 40 feet (12 m) below ground surface. Unlike RS-17, the borehole ended at the top of the Precambrian bedrock surface providing a complete sequence of glacial sediment. The boreholes drilled at the Riehl and Lorenz farms were drilled to the top of the bedrock surface at depths of 55 feet (17 m) and 45 feet (14 m), respectively. Since the boreholes were drilled in upland areas (Figure 3), the sediment consisted primarily of fine-grained glacial till with some layers of local lake sediment.

The screens or ports associated with each nested well were positioned in the borehole so that water samples could be collected for analysis and water levels could be measured to assess the distribution of hydraulic head. The continuous records of water levels measured in the RS-18 ports are presented in Figure 9a with expanded time series for the last part of the records presented in Figure 9b. A review of the water levels (Figure 9b) indicates that each well recovers at a different rate once a purge event occurs. For example, it appears that the water level in port 4 recovers at a much faster rate than port 3 which has yet to equilibrate even after one year. Overall, the water level record indicates an upward gradient. One event of note occurred on April 25 (Figure 9b) when the water level in port 7 topped the casing and started to flow at rate of about 1 Lm^{-1}. Unfortunately, the protective casing that surrounds the nested well did not drain the water quickly enough so that the groundwater from port 7 flowed out, and filled the remaining well casings causing the water levels to rise artificially. The problem was discovered on May 2 when the problem was fixed by drilling holes into the protective casing to adequately drain the water.

Review of the water level data for RS-17 (Figure 7) shows that four of the seven ports (1, 5, 6 and 7) have exactly the same water level and vary together when purged. This indicates that the ports are connected and that parts of the CMT tubing failed during well installation. There are now cracks visible in the tubing at the wellhead. The remaining three ports (2, 3, and 4) may be operational as water levels have stabilized at slightly different elevations. Because this well is compromised, water level data was not collected nor analyzed during the past year.

The water level data for the nested wells on the Riehl and Lorenz farms is presented in Figure 10. In general, there is a downward hydraulic gradient at both locations. Only a few measurement were obtained from the deep well at the Riehl farm because the levelogger® wasn't always positioned below the water table.

Chemical analyses (stable isotopes, tritium and major ions)

To evaluate the relative age of pore-water in the fine-grained sediment, $\delta^{18}O$ and $\delta^{2}H$ were analyzed on 23 samples at RS-17 (Figure 11a), 30 samples from RS-18 (Figure 11b) and 13 samples from the Riehl boring (Figure 11d). Results of the $\delta^{18}O$ analyses in RS-17 and RS-18 show modern values near the surface (-9 ‰) gradually decreasing with depth (-16 ‰ to -18 ‰) before increasing towards the bedrock surface (-11 ‰ to -12 ‰). These bow-shaped isotopic curves are similar to another curve collected from rotosonic borehole RS-14 (Figure 11c) that was drilled prior to this project in the Village of Black Creek). For each sample the $\delta^{18}O$ value was also plotted verses its $\delta^{2}H$ value to assess potential fractionation (Figure 11f-j). The samples tend to cluster near the Wisconsin meteoric water line defined as $\delta^{2}H = 7.4\ \delta^{18}O + 4.7$ (Kendall and Coplen, 2001).

To verify the pore-water analyses, water samples were taken from the multilevel wells installed in RS-17, RS-18, and the Lorenz farm. Although no pore-water was extracted from the core retrieved at the Lorenz farm, water samples from each of the three wells were collected and analyzed for $\delta^{18}O$ and $\delta^{2}H$ (Figure 11e).

Additional samples were collected from the municipal wells located in the Villages of Black Creek and Shiocton for $\delta^{18}O$ and $\delta^{2}H$ analyses. The results indicate modern values of $\delta^{18}O$ and $\delta^{2}H$ in these wells that range between -9.95 ‰ to 10.3 ‰ and -67.31 ‰ to 68.54 ‰, respectively (Figure 12). A similar result was obtained for the water sample collected from the Van Straten flowing well ($\delta^{18}O$, 10.49 ‰ and $\delta^{2}H$, -70.13‰).

The results of the tritium analyses from select wells are presented in Table 2. Tritium is present in levels greater than 1 TU near surface port 1 at RS-18, the deep wells at both the Riehl and Lorenz farms, and Black Creek municipal well 1.

Concentrations of major ions for two sets of water samples collected in April and June, 2008 are presented graphically in two Piper diagrams (Figure 13). Both diagrams indicate that there are two groups of samples where the proportions of common ions are different from each other. One of these groups, consisting of all the samples from RS-18 except for the sample from port 1, appears to have a lower proportion of sodium (Na) and potassium (K). Plotting the concentration of the various ions with depth for RS-18 shows the increase in concentration in the sample collected from port 1 (Figure 14).

None of the samples analyzed in this study had Nitrate values greater than 1 mg/l.

Hydraulic conductivity

Preconsolidation tests

The preconsolidation stress for selected samples is plotted as a function of sample depth in Figure 15. Apart from two samples, RS-17 at 142.8 to 143.1 ft and RS-18 at 131.0 to 131.6 ft, there is a trend of increasing preconsolidation stress with depth.

Consolidation testing conducted on lake sediment samples yielded values of hydraulic diffusivity, hydraulic conductivity, and specific storage for various applied loads (Table 3). The hydraulic conductivity values, calculated using the hydraulic diffusivity and specific storage, range from 1 x 10^{-8} ms^{-1} to 5 x 10^{-14} ms^{-1}. For each load a corresponding burial depth can be calculated using the specific gravity of the sediment. The burial depth, a proxy for the applied axial stress, is plotted as a function of hydraulic conductivity (Figure 16). The data shows that hydraulic conductivity is generally greater at lower stresses and decreases with increasing depth of burial. Regression of the data set reveals that a power-law fit can approximate the dependence of the hydraulic conductivity with depth. In general, the hydraulic conductivity of the fine-grained sediment decreases by one or two orders of magnitude within the upper 50 (15 m) and 100 feet (30 m), respectively. At greater stresses the hydraulic conductivity still decreases but not as dramatically, decreasing another order of magnitude at a calculated depth of 250 feet (76 m). The variation in hydraulic conductivity for a given depth appears to be relatively small for all the samples and is usually within an order of magnitude.

Slug tests

Slug tests reveal that the fine-grained sediment consisting of laminated silt and clay and silt and clay with pebbles has values of hydraulic conductivity ranging from 1.6 x 10^{-8} ms^{-1} to 6.0 x 10^{-10} ms^{-1} (Table 4). In the coarse-grained sediment, the hydraulic conductivity ranges from 2.5 x 10^{-4} ms^{-1} to 6.0 x 10^{-5} ms^{-1}.

DISSCUSSION

Recent geological mapping reveals that the thickness of fine-grained glacial sediment varies throughout Outagamie County. Where the sediment is thick (>100 feet, ~30 m) there is very little surface water infiltration resulting in limited groundwater recharge to bedrock aquifers. Where the sediment is thin (<50 feet, ~15 m), surface water can infiltrate and recharge bedrock aquifers. The boreholes drilled as part of this study (Figure 7 and 8) show that the fine-grained glacial sediment makes up the majority (90%) of the unlithified material covering the bedrock. Layers of coarser material such as sand are present by they do not appear to be laterally extensive.

The fine-grained nature of the sediment is demonstrated by the low values of hydraulic conductivity determined by slug tests in the field and consolidation tests in the laboratory. The values, which ranged from 1 x 10^{-8} ms^{-1} to 5 x 10^{-14} ms^{-1}, are typical of fine-grained sediment such as lake sediment and some glacial tills. The consolidation tests reveal that hydraulic conductivity decreases exponentially with an increase in calculated burial depth. This is due to the overlying weight of the sediment that decreases the pore space in the sediment. Such a reduction must change the connections between pore spaces resulting in a more restrictive groundwater flow path. This reduction in pore space connectivity, as reflected in the decrease in hydraulic conductivity, is not linear with depth (Figure 16). Within 50 feet (15 m) of the ground surface, there is a one order of magnitude reduction in hydraulic conductivity from 1 x 10^{-9} ms^{-1} to 1 x 10^{-10} ms^{-1}. To achieve another order of magnitude reduction in hydraulic conductivity, the sediment must be at a burial depth of about 200 feet (61 m). These results show that the thicker the fine-grained sediment the less likely it is to transmit water.

Water level data from RS-18 indicates an upward hydraulic head loss of about 7 feet (2 m) between the upper and lower ports (Figure 7) although it is apparent that port 3 has not completely equilibrated (Figure 9a). In RS-17, the water level data indicates a slight downward loss of about 1 foot between ports 2 and 4 that are positioned over an 80 foot-thick section of lake sediment. However, the compromised ports at this well (1, 5, 6, and 7) and the lack of high resolution water level data make it difficult to evaluate. Using these hydraulic head difference and a representative hydraulic conductivity from the slug tests (1.4 x 10^{-9} ms^{-1}), a vertical flux of 4.1 x 10^{-10} ms^{-1} (12.7 mma^{-1}, 0.5 ina^{-1}) and 3.0 x 10^{-11} ms^{-1} (0.04 ina^{-1}) is calculated for RS-17 and RS-18, respectively. Such a low rate of advection suggests that diffusion, not advection, may be the primary mechanism for movement of water through the sediment where it is thick.

For both the Riehl and Lorenz nested wells where the glacial sediment is much thinner, the water level shows a downward gradient of about 7 feet (2 m) of head across a sediment thickness of 30 feet (9 m) ($\Delta h/\Delta z$=0.23) and 3 feet (1 m) of head over a sediment thickness of 24 feet (7 m) ($\Delta h/\Delta z$=0.12), respectively. Slug tests from the Riehl well reveal a hydraulic conductivity of 2.6 x 10^{-8} ms^{-1} excluding the deep port that is connected to the fractured dolomite bedrock. For the Lorenz well, the hydraulic conductivity is similar at 3.4 x 10^{-8} ms^{-1}. These values are similar to vertical hydraulic conductivity determined from the consolidation testing for similar depths. Using Darcy's law, a downward vertical flux of 6.1 x 10^{-9} ms^{-1} (7.5 ina^{-1}) and 4.2 x 10^{-9} ms^{-1} (5.3 ina^{-1}) is calculated for the Riehl and Lorenz wells, respectively. These fluxes are more than an order of magnitude larger than those in the thicker sediment at RS-17 and RS-18. Advection dominates transport at these locations where the sediment is less than 50 feet (15 m) thick.

These results have implications for land use planning and contaminant site investigations. When a more conductive unit is covered by a less conductive unit, groundwater flow is generally vertical through the overlying less conductive unit (Freeze and Witherspoon, 1967). This hydrogeology is common in the study area. The fine-grained lake sediments cover the more conductive sandstones or fractured dolomites. Where the depth to bedrock is less than 50 feet (15 m), groundwater will not discharge laterally in the sediment but will move vertically until it reaches bedrock. Site investigations should be designed with this in mind, placing monitoring points vertically as well as laterally since the flux will be mostly vertical in the sediment, not horizontal.

Water level fluctuations may also demonstrate that advection dominates transport where the sediment is thin while hydraulic diffusivity dominates transport where sediment is thick. An observation that may support this assertion is water levels that fluctuate at a higher frequency indicating a direct connection with infiltrating surface water. For example, at port 1 at RS-18, 20 feet (6 m) below ground surface, the water level changes more frequently than the deeper ports screened in the fine-grained sediment. The same observation is made at all three nested wells at the Lorenz farm and the shallow well at the Riehl farm.

In addition to the effects of less overburden stress, the increase in hydraulic conductivity in near surface sediment may be the result of shallow fractures that formed immediately following the draining of glacial Lake Oshkosh when the ground was permanently frozen (Attig et al., 1989) or even during more recent freeze/thaw cycles or the low effective stress due to the shallow depth of burial. While the presence of fractures can significantly increase the overall hydraulic conductivity of fine-grained sediment (Grisak and Cherry, 1975; Hendry, 1983; Simpkins and Bradbury, 1992), outcrops and cores collected from glacial Lake Oshkosh do not show any visible signs of fractures. However, this simply may be the result of failure to identify fractures in silt and clay.

The isotope data further supports the assertion that advection through the sediment is limited where it is thicker than 50 feet (15 m). $\delta^{18}O$ results from pore-water analyses at RS-14, RS-17, and RS-18 (Figure 11) indicate relatively old pore water. The analyses from all three sites yield bow-shaped curves typical of chemical diffusion with limited advection driven by differences in hydraulic head (Remenda et al., 1996). $\delta^{18}O$ values of about -30 ‰ are typical of glacial-age water, whereas values of -9 ‰ are indicative of modern precipitation. Because the pore water in the lake sediment at the time it was emplaced would have been derived from the proglacial lake that includes runoff and glacial meltwater, it is possible that the $\delta^{18}O$ values of the original pore water would be diluted and thus much lower than -30‰. The large difference between the depleted $\delta^{18}O$ values at depth in the aquitard and those at the ground surface suggest that little displacement of the sediment pore water has occurred, which is consistent with the low value of hydraulic conductivity determined from the consolidation and slug test data.

Although there are lighter $\delta^{18}O$ values near the middle of the sediment sequence, the heavier values close to the bedrock surface are difficult to explain especially since the values reflect modern day precipitation. These results are surprising in that several investigations into the stable isotopes of groundwater in the northern Midwest show that the geochemistry of the Cambrian-Ordovician aquifer system was modified during the Pleistocene by a large-scale emplacement of glacial meltwater, and that groundwater in this part of the aquifer system could be hundreds of thousands of years old (Siegel and Mandle, 1984; Siegel, 1989*).* Similarly, in a study of the stable isotope geochemistry of groundwater in the Cambrian Ordovician aquifer system of northern Illinois, stable isotope analyses provided evidence that a significant portion of the groundwater had been stored since the Pleistocene (Perry et al., 1982). If this is the case for the bedrock aquifers in the Fox River Valley, then lighter $\delta^{18}O$ values would be expected in the bottom part of the sediment sequence. The heavy $\delta^{18}O$ values at the sites we studied indicate that groundwater recharge must be occurring to the deep aquifer west of the lake basin or from areas where the fine-grained sediment is thin (<50 feet, 15 m).

One-dimensional advection dispersion models were fit to the oxygen isotope profiles in wells RS-14, RS-17, and RS-18. The models used a finite difference Crank-Nicolson Approximation to solve the advection-dispersion governing equation (Wang and Anderson, 1982),

$$D\frac{\partial^2 C}{\partial z^2} - v\frac{\partial C}{\partial z} = \frac{\partial C}{\partial t}$$

with the following boundary conditions,

$C(0,t)=C_1$	Constant concentration at the surface of the sediment,
$C(L,t)=C_2$	Constant concentration at the bottom of the sediment,
$C(z,0)=C_3$	Initial concentration throughout sediment,

where *D* is the chemical diffusion, *v* is the advective velocity, *C* is concentration, *z* is depth variable where *L* is the sediment thickness, and *t* is time (Figure 16a). MATLAB™ was used to implement the finite difference approximation to obtain a solution. Figure 16b illustrates solutions for several values of *t* with downward advection.

A least-squares best-fit inversion was used to fit the models to the measured profiles. The initial concentration, C_3, constant concentrations at the profile top and bottom, C_1 and C_2, respectively, chemical diffusion, *D*, and advective velocity, *v*, were all varied (Table 5) in the inversion to produce the best-fit plots shown in Figure 11. At RS-14 and RS-17, diffusion dominates with some discernable contribution from advection as seen by the shift in the minimum $\delta^{18}O$ downward in those plots. At RS-18 there is little shift. At this location, diffusion is dominates with very little contribution from advection.

The isotope data from the shallow borings and wells at the Riehl and Lorenz farms indicates that at least some of recharge may be occurring where the sediment is thin (<50 feet, 15 m). The $\delta^{18}O$ values on samples collected over the entire thickness of sediment at the Riehl farm (~50 feet, 15 m) show modern day values which suggests that no glacial age water remains in the sediment. The isotopic data is supported by the results of the major ion and tritium analyses. The major ion analyses indicates that water samples collected from RS-18 where the sediment is thick (Figure 13, group 2) are distinctly different from the samples collected where the sediment is thin such as the Riehl and Lorenz farms (Figure 13, group 2). The major ions in the water samples collected from the municipal wells in Villages of Black Creek and Shiocton fall within group 1 indicating that these wells are chemically similar. Two wells at the Village of Black Creek and one well at the Village of Shiocton pump water from shallow sand and gravel layers that are most likely hydraulically connected to the adjacent Cambrian sandstone that forms the deep aquifer. Therefore, the major ion analysis indicates that they may be recharged locally where the fine-grained sediment is thin.

This interpretation is supported by the tritium analysis of water samples. The presence of tritium above the detection limit in one of the Village of Black Creek's wells, RS-18 (port 1), and the deepest nested wells at the Riehl and Lorenz farms indicates relatively modern water (post 1950's). The lack of tritium at the Village of Shiocton Well 1 and in the deep ports (6 and 7) at RS-18 indicates no modern recharge which is expected because they are both located immediately above the granite bedrock. However, the lack of tritium at Shiocton Well 2 is surprising given its similarity in major ion chemistry to the shallow wells.

If the bedrock aquifers are recharged locally where the glacial sediment is thin, as indicated by the data interpreted in this study, then it is possible to identify potential recharge areas. Assuming that recharge to bedrock aquifers only occurs where the sediment is less than 50 feet (15 m) thick, recharge areas can identified for both the shallow (upper Sinnipee Group) and deep aquifers (Figure 17). The value of 50 feet (15 m) of sediment thickness was chosen because it corresponds to a decrease of an order of magnitude in the hydraulic conductivity in the consolidation testing and few documented fractures penetrate to this depth. The contrast of fluxes between the shallow Lorenz and Riehl sites to the deep sites at RS-17 and RS-18 supports setting the thickness where significant recharge occurs to 50 feet (15 m). The boundary between these two potential recharge areas is based on bedrock mapping recently completed by the WGNHS. Although this mapping does not distinguish between the upper and lower part of the Sinnipee Group, it was assumed that the lower contact with the St. Peter sandstone or the Prairie du Chien is the boundary between the two aquifers.

A goal of this study was to make measurements that could be used to test the idea that recharge areas to bedrock aquifers occur where the fine-grained glacial sediment is relatively thin (<50 feet thick). These areas are shown in Figure 17. A rough estimate of the rate of flow to the bedrock aquifers can be calculated by multiplying the flux by the area of shallow bedrock. The average downward flow at the Lorenz and Riehl sites is 6.4 inches per year. If we assume that the sediment is less than 50 feet (15 m) thick over 20%

of Outagamie County (640 mi^2), then the recharge occurs over an area of 128 square miles. After appropriate unit conversions, the calculated flow to the bedrock aquifers is approximately 40 million gallons per day (mgd). A review of the bedrock geology of Outagamie County shows that about half of the 128 square miles are located west of the Sinnipee Group. Thus, both the shallow and deep bedrock aquifers are recharged at a rate of about 20 million gallons per day. This estimated value of recharge to the deep aquifer is on the same order of magnitude as pumping in the lower Fox River Valley in 2007 (Hart et. al, 2008). The estimate is useful only in that it suggests that recharge in this area is important for the deep aquifer in the lower Fox River valley. There are many assumptions and uncertainties contained in the estimate. One uncertainty is whether the gradient and hydraulic conductivities measured at the shallow Riehl and Lorenz farms are applicable to the entire shallow area shown in Figure 17. Another uncertainty is that the existing slug tests determine the horizontal hydraulic conductivity rather than the vertical hydraulic conductivity that is needed in this estimate. A calibrated ground water flow model that included the sediment and bedrock geometries would be necessary to increase the certainty. This data would then provide essential targets for that model.

This rough estimate gives an indication as to whether or not the magnitude of flux is close to that pumped from the deep sandstone aquifer in the Fox River valley. Until last year, the pumping rate by municipalities from the deep sandstone aquifer in the lower Fox River Valley was approximately 8 mgd, a significant fraction of the estimated 20 mgd recharge. In August, 2007, the six municipalities of the Central Brown County Water Authority switched to water drawn from Lake Michigan and ceased pumping. The reduction in pumping was the majority of the previous rate of 8 mgd. It may be that because of the reduced pumping, the recharge to the deep sandstone aquifer will lessen and baseflow to streams in the recharge area will increase noticeably. The reduction in recharge may also become evident through a reduction in the downward gradient at the Lorenz and Riehl sites.

CONCLUSIONS

The fine-grained glacial deposits in the Fox River Valley greatly limit surface water infiltration and groundwater recharge to bedrock aquifers. The nature of the sediment and its thickness are critical variables in understanding flow patterns and recharge to the underlying aquifers. Thick sequences of fine-grained sediment prevent downward migration of modern precipitation to bedrock aquifers. However, where the sediment is thinner, the bedrock aquifers are much closer to the surface and are susceptible to groundwater recharge and contamination. Using sediment thickness as a proxy for potential groundwater recharge areas, potential recharge areas in Outagamie County were identified to the shallow and deep aquifers. The recharge areas were delineated using contact between the Sinnipee and the rock units below. Recharge areas to the aquifer in other parts of the Fox River Valley could potentially be identified using this technique. Even though many municipalities have been switching to surface water for drinking, understanding the spatial distribution of recharge in this region is still crucial for planning the future management of groundwater quality and quantity.

DATABASE DEVELOPMENT

The physical properties of unlithified sediment are important in many geology, hydrogeology, and engineering studies. As part of this project, the WGNHS updated and reissued a searchable database called Till*Pro* (Wisconsin Geological and Natural History Survey, 2008). This database initially only contained grain-size analyses on samples collected around the state, but was updated to include an additional 10,000 records on other physical properties such as hydraulic conductivity, shear strength, preconsolidation, and plasticity. The data was collected from many different sources including journal articles, miscellaneous databases at the WGNHS, and M.S. theses and Ph.D. dissertations. Users of this database can now search for information in many different ways including sample location and lithology. Till*Pro* is available for purchase at the WGNHS, 3817 Mineral Point Road, Madison WI, 53705. Copies of this database on CD-ROM has been included as part of this report.

ACKNOWLEDGEMENTS

The authors thank Carrie Moeller, a Ph.D. student at the University of Wisconsin-Madison, who collected and evaluated a large part of the data set presented in this report. We also thank Peter Chase who was instrumental in drilling and installing the various nested wells. This project would not have been possible without the cooperation of various landowners including the Lorenz and Riehl families, and the Villages of Black Creek (Joe Morrisey) and Shiocton (Dick Kominowski).

Funding for this project was provided by the Wisconsin Department of Natural Resources through the Bureau of Drinking Water and Groundwater. Additional funding to drill the deep boreholes was provided by STATEMAP, the state component of the National Cooperative Geologic Mapping Program of the U.S. Geological Survey.

REFERENCES

Attig, J.W., Mickelson, D.M., and Clayton, L., 1989. Late Wisconsin landform distribution and glacier-bed conditions in Wisconsin. Sedimentary Geology, 62, p. 399-405.

Batten, W.G., Bradbury, K.R., 1996. Regional groundwater flow system between the Wolf and Fox rivers near Green Bay, Wisconsin. Wisconsin Geological and Natural History Survey, Information Circular 75, 28pp.

Conlon, T.D., 1998. Hydrogeology and simulation of ground-water flow in the Sandstone Aquifer, northeastern Wisconsin. Water Resources Investigations, U. S. Geological Survey Report: WRI 97-4096, 60 pp.

Das, B.M. 1994. Principles of Geotechnical Engineering, 3rd edition, Boston, MA, PWS Publishing Company, p. 672.

Einarson M.D. and Cherry, J.A., 2002. A new multilevel ground water monitoring system using multichannel tubing. Ground Water Monitoring & Remediation, 22, 4, p. 52-65.

Freeze, R.A., and Witherspoon, P.A., 1967. Theoretical analysis of regional groundwater flow: 2. Effect of water-table configuration and subsurface permeability variation. Water Resources Res. 3, pp. 623-634.

Freeze, R.A., and Cherry, J.A., 1979. Groundwater. Prentice Hall, 604 pp.

Gaspar, E., 1987, Hydrological tracers. In: Gaspar, E., (ed) Modern Trends in Tracer Hydrology, v. 1, Boca Raton, Fl, CRC Press, p. 1-48.

Grisak, G.E., and Cherry, J.A., 1975. Hydrologic characteristics and response of fractured till and clay confining a shallow aquifer. Canadian Geotechnical Journal, 12, 1, p. 23-43

Hart, D.J., Luczaj, J. A., and Chase, P.M., 2008, A Large Scale Pumping Test in the Northeastern Wisconsin Groundwater Management Area, American Water Resources Association, Wisconsin Chapter. 32nd annual Meeting Program and Abstracts.

Hart, D.J. and Hammon III, W.S., 2002. Measurement of hydraulic conductivity and specific storage using the shipboard Manheim squeezer. In: Salisbury, M.H., Shinohara, M., Richter, C., et al., 2002. Proceeding of the Ocean Drilling Program, Initial Reports Volume 195.

Helmke, M.F., Simpkins, W.W., and Horton, R., 2005. Simulating conservative tracers in fractured till under realistic timescales. Ground Water, 43, 6, 877-889.

Hendry, M. J., 1983. Groundwater recharge through a heavy-textured soil. Journal of Hydrology, 63, p. 201-209.

Hvorslev, M.J., 1951. Time Lag and Soil Permeability in Ground-Water Observations. Bull. No. 36, Waterways Exper. Sta. Corps of Engrs, U.S. Army, Vicksburg, Mississippi, p. 1-50.

Hyder, Z., Butler, J.J., Jr., McElwee, C.D. and Liu, W., 1994. Slug tests in partially penetrating wells. Water Resources Research, vol. 30, no. 11, pp. 2945-2957.

International Atomic Energy Association (IAEA), 1981. J.R. Gat and R. Gontiafini, eds. Stable isotope hydrology: Deuterium and oxygen-18 in the water cycle: a monograph prepared under the aegis of the IAEA/UNESCO Working Group on Nuclear Techniques in Hydrology of the International Hydrological Programme. Vienna, 339 pp.

Keller, C.K., van der Kamp, G., Cherry, J.A., 1989. A multiscale study of the permeability of a thick clayey till. Water Resources Research, 25, 11, p. 2299-2317.

Kendall, C. and Coplen, T.B. 2001. Distribution of oxygen-18 and deuterium in river waters across the United States. Hydrological Processes, 15, p. 1363–1393.

Krohelski, J.T., 1986. Hydrogeology and groundwater use and quality, Brown County, Wisconsin. Wisconsin Geological and Natural History Survey, Information Circular 57, 42pp.

LeRoux, E.F., 1957. Geology and ground-water resources of Outagamie County, Wisconsin. U.S. Geological Survey Water-Supply Paper 1421, 57 pp.

McElwee, C.D., 2000. Implementation of a nonlinear model for analysis of slug tests. Kansas Geological Survey open file report 2000-01

Perry, E.C. Jr., Grundl, T., and Gilkeson, R.H., 1982. H, O, and S isotopic study of the groundwater in the Cambrian-Ordovician aquifer system of northern Illinois. In: Perry, E.C. Jr. and Montgomery, C.W., Eds. Isotopic Studies of Hydrologic Processes. Northern Illinois Univ. Press, p. 35-43.

Remenda, V.H., van der Kamp, G., Cherry, J.A., 1996. Use of vertical profiles of $\delta^{18}O$ to constrain estimates of hydraulic conductivity in a thick, unfractured aquitard. Water Resources Research, 32, 10, p. 2979-2987.

Siegel, D.I., 1989. Geochemistry of the Cambrian-Ordovician aquifer system in the northern Midwest, United States. U. S. Geological Survey Professional Paper, Report: P 1405-D, D1-D76.

Siegel, D.I. and Mandle, R.J., 1984. Isotopic evidence for glacial meltwater recharge to the Cambrian-Ordovician aquifer, north-central United States. *Quaternary Research*, 22, 3, p. 328-335.

Simpkins W.W. and K.R. Bradbury. 1992. Groundwater flow, velocity, and age in a thick, fine-grained till unit in southeastern Wisconsin. *Journal of Hydrology*, 132, p. 283-319.

Terzaghi, K., 1943. *Theoretical Soil Mechanics*. J. Wiley and Sons Inc., New York, 510 pp.

Wang, H.F. and M.P. Anderson, 1982. *Introduction to Groundwater Modeling*. W.H. Freeman and Company, New York, 237 pp.

Wang, H.F., 2000. *Theory of Linear Poroelasticity with Applications to Geomechanics and Hydrogeology*. Princeton University Press, New Jersey, 287 pp.

Wisconsin Geological and Natural History Survey, 2008, Till*Pro*: A database containing physical properties of unlithified sediment samples: Wisconsin Geological and Natural History Survey Open-File Report 2008-05, version 1.0, 1 CD-ROM.

Table 1. Summary of water samples collected for laboratory analyses

Sample Location	***Analysis*** $\delta^{18}O$ ***and*** $\delta^{2}H$	***tritium***	***Major ions April, 2008***[1]	***Major ions June, 2008***[1]
RS-17				
port 1	X		X	
port 2	X		X	
port 3	X		X	
port 4				
port 5				
port 6				
port 7				
RS-18				
port 1	X	X	X	X
port 2	X		X	X
port 3	X		X	X
port 4	X		X	X
port 5	X		X	X
port 6	X	X	X	X
port 7	X	X	X	X
VanStraten well	X		X	X
Schiocton wells				
well 1	X	X	X	X
well 2	X	X	X	X
Black Creek wells				
well 1	X	X	X	X
well 2	X		X	X
Riehl farm				
shallow	X		X	X
intermediate	X		X	X
deep	X	X	X	X
Lorenz farm				
shallow	X			X
intermediate	X			X
deep	X	X		X

[1] Includes Mg, Ca, Na, K, CL, SO_4, HCO_3, and N

Table 2. Results of water sample analyses for tritium

Sample location	Tritium, TU ± 1σ [1]
Monitoring wells	
RS-18, Port 1	1.2 ± 0.7
RS-18, Port 6	<0.8 ± 0.7
RS-18, Port 7	<0.8 ± 0.7
Riehl, deep	1.9 ± 0.8
Lorenz, deep	4.1 ± 0.8
Municipal Wells	
Shiocton well 1	<0.8 ± 0.8
Shiocton well 2	<0.8 ± 0.8
Back Creek well 1	1.6 ± 0.8

[1] 1 TU=3.221 Picocurries/L or 0.11919 Becquerels/L

Table 3. Hydraulic properties of fine-grained sediments as determined by consolidation testing.

Sample Id and Depth	Load (kPa)	Depth (ft)	Diffusivity (m^2/s)	Hydraulic Conductivity (m/s)	Specifice Storage (m^{-1})
RS-12 59.7-60	22	4.4	2.33E-07	8.59E-10	3.68E-03
RS-12 59.7-60	44	8.7	1.02E-07	2.42E-10	2.37E-03
RS-12 59.7-60	89	17.5	1.47E-07	2.38E-10	1.62E-03
RS-12 59.7-60	177	34.9	8.61E-08	9.13E-11	1.06E-03
RS-12 59.7-60	355	69.9	3.52E-08	2.98E-11	8.46E-04
RS-12 59.7-60	710	139.7	2.58E-08	1.55E-11	6.00E-04
RS-12 59.7-60	1065	209.6	1.68E-08	4.78E-12	2.84E-04
RS-12 59.7-60	1508	297.0	1.14E-08	2.40E-12	2.10E-04
RS-12 59.7-60	2063	406.1	1.16E-08	1.77E-12	1.53E-04
RS-12 59.7-60	2773	545.9	1.20E-08	1.41E-12	1.18E-04
RS-12 59.7-60	3483	685.6	1.38E-08	1.01E-12	7.35E-05
RS-12 59.7-60	4192	825.4	2.62E-08	1.14E-12	4.36E-05
RS-12 69.6-69.8	22	4.4	4.73E-07	1.91E-09	4.04E-03
RS-12 69.6-69.8	44	8.7	1.59E-07	2.53E-10	1.59E-03
RS-12 69.6-69.8	89	17.5	5.88E-07	8.79E-10	1.49E-03
RS-12 69.6-69.8	177	34.9	9.68E-08	6.93E-11	7.16E-04
RS-12 69.6-69.8	355	69.9	9.99E-08	6.03E-11	6.04E-04
RS-12 69.6-69.8	710	139.7	5.59E-08	2.62E-11	4.69E-04
RS-12 69.6-69.8	1109	218.4	4.45E-08	1.18E-11	2.66E-04
RS-12 69.6-69.8	1819	358.1	3.23E-08	6.21E-12	1.93E-04
RS-12 69.6-69.8	2773	545.9	2.94E-08	3.91E-12	1.33E-04
RS-12 69.6-69.8	3549	698.7	1.75E-08	1.37E-12	7.82E-05
RS-12 98-98.2	22	4.4	1.00E-06	8.11E-09	8.08E-03
RS-12 98-98.2	44	8.7	1.86E-07	2.63E-10	1.41E-03
RS-12 98-98.2	89	17.5	4.62E-07	5.60E-10	1.21E-03

RS-12 98-98.2	111	21.8	2.34E-07	1.69E-10	7.20E-04
RS-12 98-98.2	355	69.9	1.79E-07	8.74E-11	4.88E-04
RS-12 98-98.2	710	139.7	7.17E-08	2.54E-11	3.55E-04
RS-12 98-98.2	1287	253.3	6.13E-08	1.42E-11	2.31E-04
RS-12 98-98.2	1996	393.0	3.64E-08	5.97E-12	1.64E-04
RS-12 98-98.2	2773	545.9	2.13E-08	2.41E-12	1.13E-04
RS-12 98-98.2	3483	685.6	1.58E-08	1.05E-12	6.64E-05
RS-18 166.5	22	4.4	3.66E-08	1.74E-10	4.75E-03
RS-18 166.5	44	8.7	3.00E-08	3.90E-11	1.30E-03
RS-18 166.5	89	17.5	8.05E-08	9.98E-11	1.24E-03
RS-18 166.5	177	34.9	9.13E-08	5.90E-11	6.46E-04
RS-18 166.5	355	69.9	9.11E-08	4.06E-11	4.46E-04
RS-18 166.5	710	139.7	1.33E-07	3.81E-11	2.86E-04
RS-18 166.5	1420	279.5	5.53E-08	1.33E-11	2.40E-04
RS-18 166.5	2839	559.0	4.32E-08	7.41E-12	1.71E-04
RS-18 166.5	4569	899.6	3.74E-08	2.77E-12	7.39E-05
RS-18 225	22	4.4	1.26E-07	6.58E-10	5.24E-03
RS-18 225	44	8.7	6.57E-08	8.78E-11	1.34E-03
RS-18 225	89	17.5	6.03E-08	6.82E-11	1.13E-03
RS-18 225	177	34.9	4.90E-08	4.22E-11	8.61E-04
RS-18 225	355	69.9	7.33E-08	3.95E-11	5.38E-04
RS-18 225	710	139.7	3.95E-08	1.43E-11	3.62E-04
RS-18 225	1420	279.5	4.98E-08	1.01E-11	2.02E-04
RS-18 225	2839	559.0	5.76E-08	8.16E-12	1.42E-04
RS-18 225	4348	856.0	4.16E-08	2.24E-12	5.40E-05
RS-18 263	22	4.4	1.34E-07	3.93E-10	2.93E-03
RS-18 263	44	8.7	9.88E-08	1.41E-10	1.42E-03
RS-18 263	89	17.5	9.32E-08	9.66E-11	1.04E-03
RS-18 263	177	34.9	7.05E-08	4.64E-11	6.58E-04

RS-18 263	355	69.9	1.05E-07	4.36E-11	4.16E-04
RS-18 263	710	139.7	5.79E-08	1.89E-11	3.27E-04
RS-18 263	1420	279.5	7.69E-08	1.87E-11	2.44E-04
Ebben Quarry #1	22	4.4	1.53E-06	1.15E-08	7.50E-03
Ebben Quarry #1	44	8.7	1.15E-06	1.99E-09	1.73E-03
Ebben Quarry #1	89	17.5	1.94E-06	4.10E-09	2.11E-03
Ebben Quarry #1	177	34.9	4.58E-07	3.15E-10	6.88E-04
Ebben Quarry #1	355	69.9	3.33E-07	1.35E-10	4.04E-04
Ebben Quarry #1	710	139.7	2.47E-07	8.23E-11	3.33E-04
Ebben Quarry #1	1420	279.5	2.45E-07	4.36E-11	1.78E-04
Ebben Quarry #1	2617	515.3	2.09E-07	3.59E-11	1.72E-04
Ebben Quarry #1	2972	585.2	4.81E-08	1.62E-12	3.37E-05
Ebben Quarry #1	3194	628.9	5.49E-07	1.14E-11	2.07E-05
Ebben Quarry #2	22	4.4	2.85E-06	1.78E-08	6.24E-03
Ebben Quarry #2	44	8.7	1.60E-06	1.33E-09	8.35E-04
Ebben Quarry #2	89	17.5	2.89E-07	4.12E-10	1.43E-03
Ebben Quarry #2	177	34.9	1.34E-07	9.94E-11	7.42E-04
Ebben Quarry #2	355	69.9	1.35E-07	6.43E-11	4.76E-04
Ebben Quarry #2	710	139.7	1.47E-07	4.11E-11	2.79E-04
Ebben Quarry #2	1420	279.5	1.20E-07	2.03E-11	1.70E-04
Ebben Quarry #2	2839	559.0	2.66E-07	3.55E-11	1.33E-04
RS-15 47.6 ft	22	4.4	2.91E-04	3.65E-06	1.26E-02
RS-15 47.6 ft	44	8.7	3.77E-08	4.18E-11	1.11E-03
RS-15 47.6 ft	89	17.5	1.63E-07	1.88E-10	1.15E-03
RS-15 47.6 ft	177	34.9	5.45E-08	6.16E-11	1.13E-03
RS-15 47.6 ft	355	69.9	7.48E-08	4.04E-11	5.40E-04
RS-15 47.6 ft	710	139.7	8.55E-08	2.85E-11	3.33E-04
RS-15 47.6 ft	1420	279.5	8.85E-08	1.57E-11	1.78E-04
RS-15 47.6 ft	1863	366.8	6.37E-08	3.86E-12	6.05E-05

RS-15 47.6 ft	2573	506.6	8.05E-08	4.71E-12	5.86E-05
RS-15 47.6 ft	2928	576.5	1.04E-07	2.64E-12	2.54E-05
RS-15 47.6 ft	3194	628.9	1.08E-07	1.82E-12	1.69E-05
RS-16 80.5 ft (April)	22	4.4	8.95E-07	4.69E-09	5.24E-03
RS-16 80.5 ft (April)	44	8.7	2.05E-07	3.79E-10	1.85E-03
RS-16 80.5 ft (April)	89	17.5	8.32E-08	1.34E-10	1.61E-03
RS-16 80.5 ft (April)	177	34.9	8.67E-08	6.85E-11	7.91E-04
RS-16 80.5 ft (April)	355	69.9	1.58E-07	7.73E-11	4.88E-04
RS-16 80.5 ft (April)	710	139.7	1.44E-07	4.07E-11	2.82E-04
RS-16 80.5 ft (April)	1420	279.5	7.31E-08	1.09E-11	1.50E-04
RS-16 80.5 ft (April)	2129	419.2	1.22E-07	8.14E-12	6.67E-05
RS-16 80.5 ft (April)	2573	506.6	6.29E-09	1.18E-13	1.87E-05
RS-16 80.5 ft (April)	2928	576.5	1.82E-08	3.10E-13	1.70E-05
RS-16 80.5 ft (April)	3194	628.9	4.15E-09	4.62E-14	1.11E-05
RS-16 80.5 ft (March)	22	4.4	3.26E-06	6.48E-08	1.99E-02
RS-16 80.5 ft (March)	44	8.7	6.25E-07	7.36E-10	1.18E-03
RS-16 80.5 ft (March)	89	17.5	1.21E-06	1.36E-09	1.13E-03
RS-16 80.5 ft (March)	177	34.9	3.13E-07	2.17E-10	6.93E-04
RS-16 80.5 ft (March)	355	69.9	2.04E-06	1.83E-09	8.96E-04
RS-16 80.5 ft (March)	710	139.7	1.83E-06	7.64E-10	4.16E-04
RS-16 80.5 ft (March)	1420	279.5	2.66E-07	3.55E-11	1.33E-04
RS-16 80.5 ft (March)	2129	419.2	3.29E-07	1.77E-11	5.38E-05
RS-16 80.5 ft (March)	2573	506.6	1.86E-07	3.85E-12	2.07E-05
RS-16 80.5 ft (March)	2928	576.5	1.45E-07	2.10E-12	1.44E-05
RS-16 80.5 ft (March)	3194	628.9	7.70E-08	4.54E-13	5.90E-06

Table 4. Hydraulic conductivities obtained by slug tests from monitoring wells

Well number	Port number	Depth from Ground surface (ft)	Elevation (ft)	Sediment description	Hydraulic conductivity (m/s)
18	1	18	745	Laminated silt and clay	2.7×10^{-8}
18	2	47	726	Fine to medium sand	1.6×10^{-5}
18	3	89	676	Laminated silt and clay	1.4×10^{-10}
18	4	164	646	Fine grained silt and clay	2.5×10^{-9}
18	5	219	612	Laminated silt and clay	3.5×10^{-10}
18	6	285	527	Fine-grained sand	6.0×10^{-5}
18	7	304	487	Sand and gravel adjacent to fractured crystalline bedrock	2.5×10^{-4}
Riehl	shallow	23	777	Silt and clay with pebbles	5.1×10^{-8}
Riehl	medium	36	764	Silt and clay with fine sand	6.0×10^{-10}
Riehl	deep	51	749	Fractured dolomite	1.7×10^{-7}
Lorenz	shallow	23	802	Silt and clay with pebbles	1.7×10^{-8}
Lorenz	medium	34	791	Sandy silt with clay and pebbles	7.0×10^{-8}
Lorenz	deep	47	778	Sandy silt with clay and pebbles	1.6×10^{-8}

Table 5. Parameters used in the one-dimensional advection dispersion model

Borehole number	v (m^2)	D ($m^2 s^{-1}$)	C_1	C_2	C_3
RS-17	1.1×10^{-10}	1.6×10^{-9}	7.5	22	10.1
RS-18	1.6×10^{-11}	3.4×10^{-10}	8.8	17.9	12.3
RS-14	8.7×10^{-11}	1.4×10^{-9}	8.7	20.2	9.7

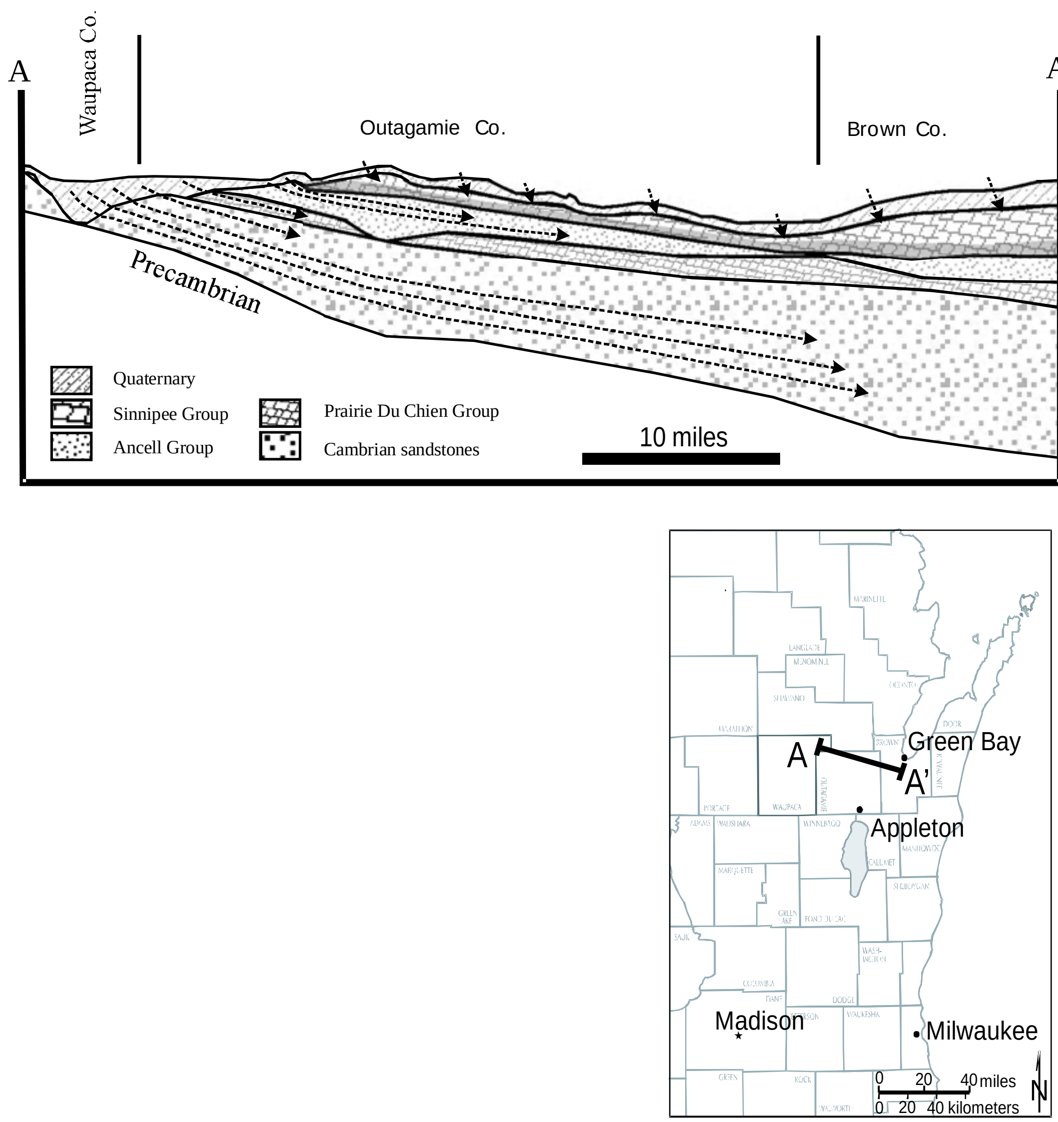


Figure 1. Idealized east to west cross section showing the bedrock geology and groundwater flow paths to the two bedrock aquifers that are divided by the confining unit of the lower Sinnipee Group (shaded area). General location of the cross section is shown in the inset diagram. Modified from Batten and Bradbury (1996).

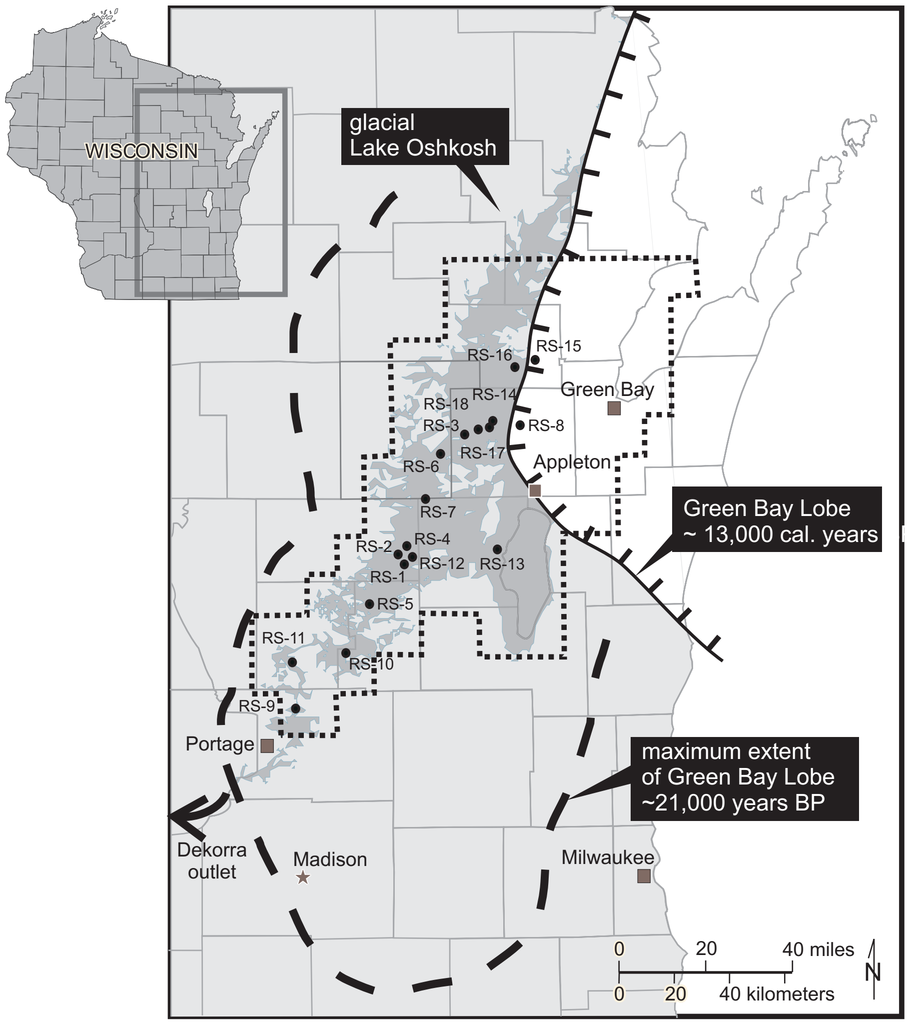
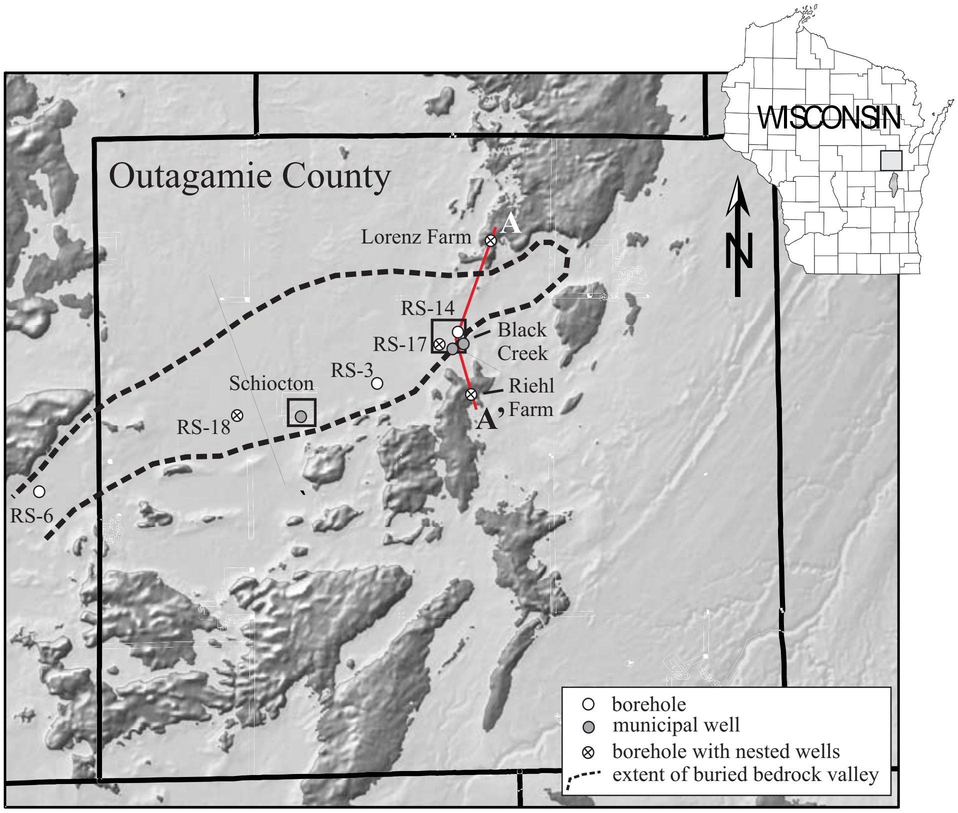


Figure 2. Map of east-central Wisconsin. The shaded gray area represents the extent of glacial Lake Oshkosh and the dotted line encompasses the area currently being mapped by the WGNHS. The dots represent locations of rotosonic boreholes drilled to study the geology of the fine-grained glacial sediment.

a.



b.

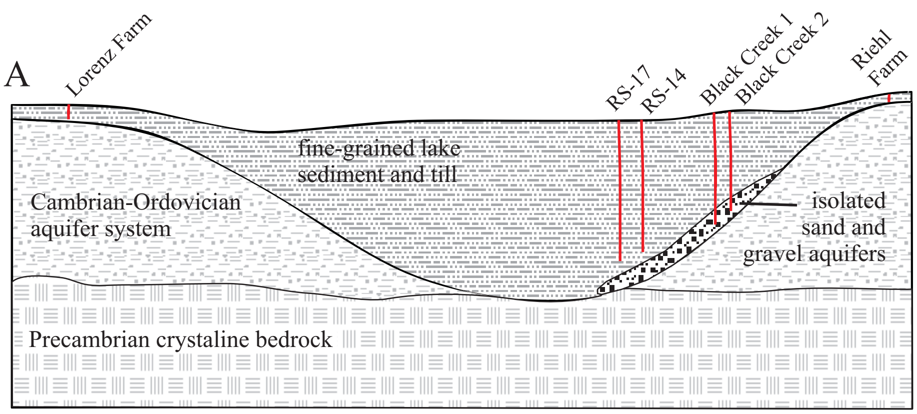


Figure 3. (a) Shaded relief map of Outagamie County showing the lateral margins of the buried bedrock valley, the locations of boreholes and wells used in this study, and the location of the generalized cross section. The light gray area represents the area covered by glacial Lake Oshkosh during the last glaciation. (b) A generalized cross section showing the position of the buried bedrock valley and some of the wells used in this study.

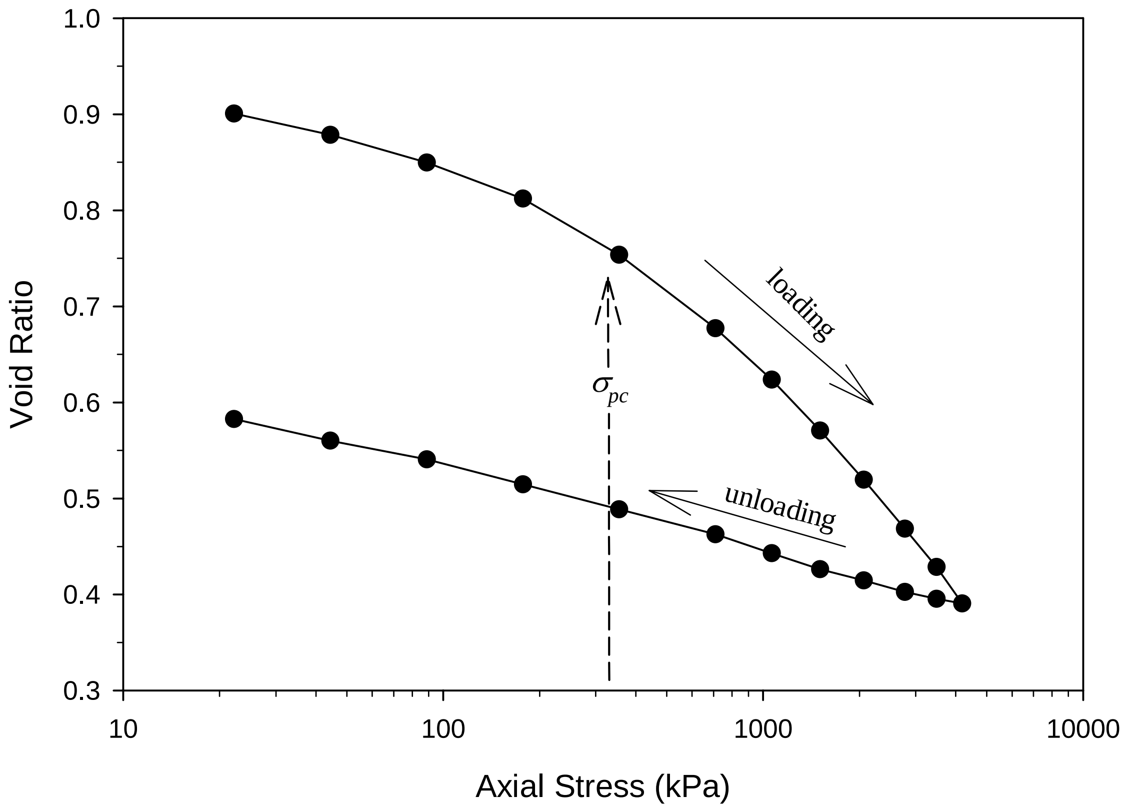


Figure 4. Consolidation curve of glacial Lake Oshkosh ediment sample (RS12, 59.7-60.0 feet) as a function of axial stress, determined from tests with a fixed-ring consolidometer. The arrows show the loading path and σ_{pc} represents the preconsolidation stress determined from the Casagrande method.

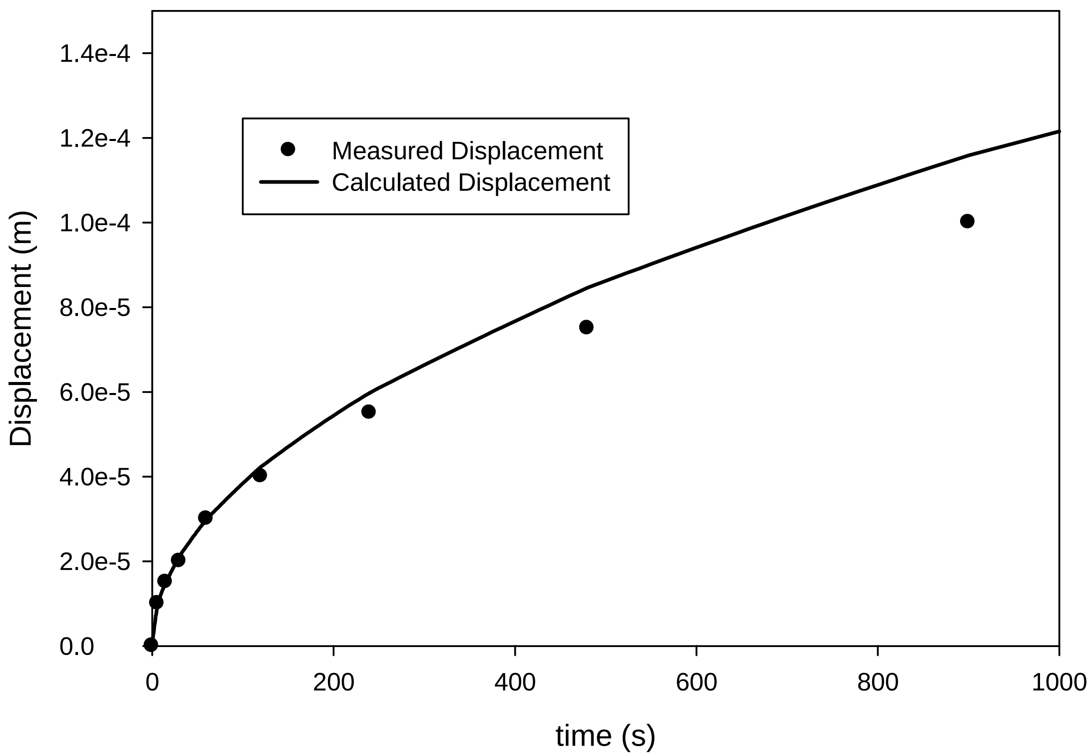


Figure 5. Measured and calculated displacement as a function of time for sample RS-12, 59.7-60.0 for a load of 2×10^6 Pa.

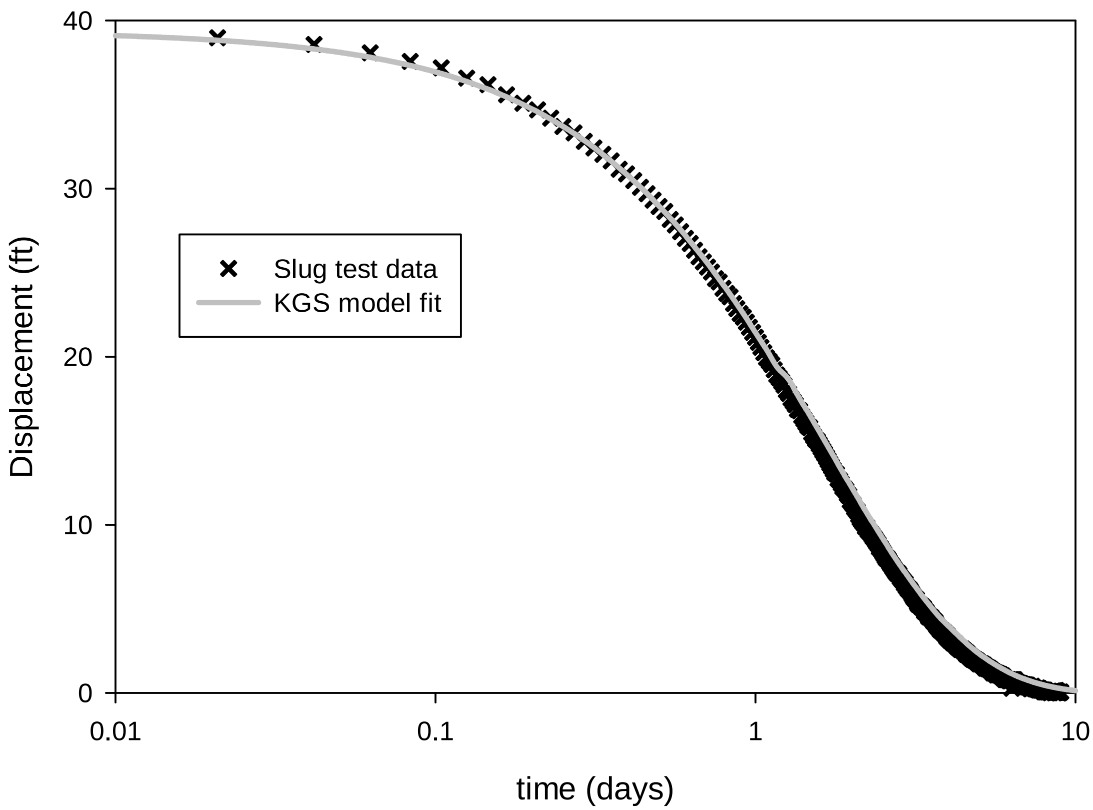


Figure 6. Example fo slug test data and KGS model file for RS-18, port 4.

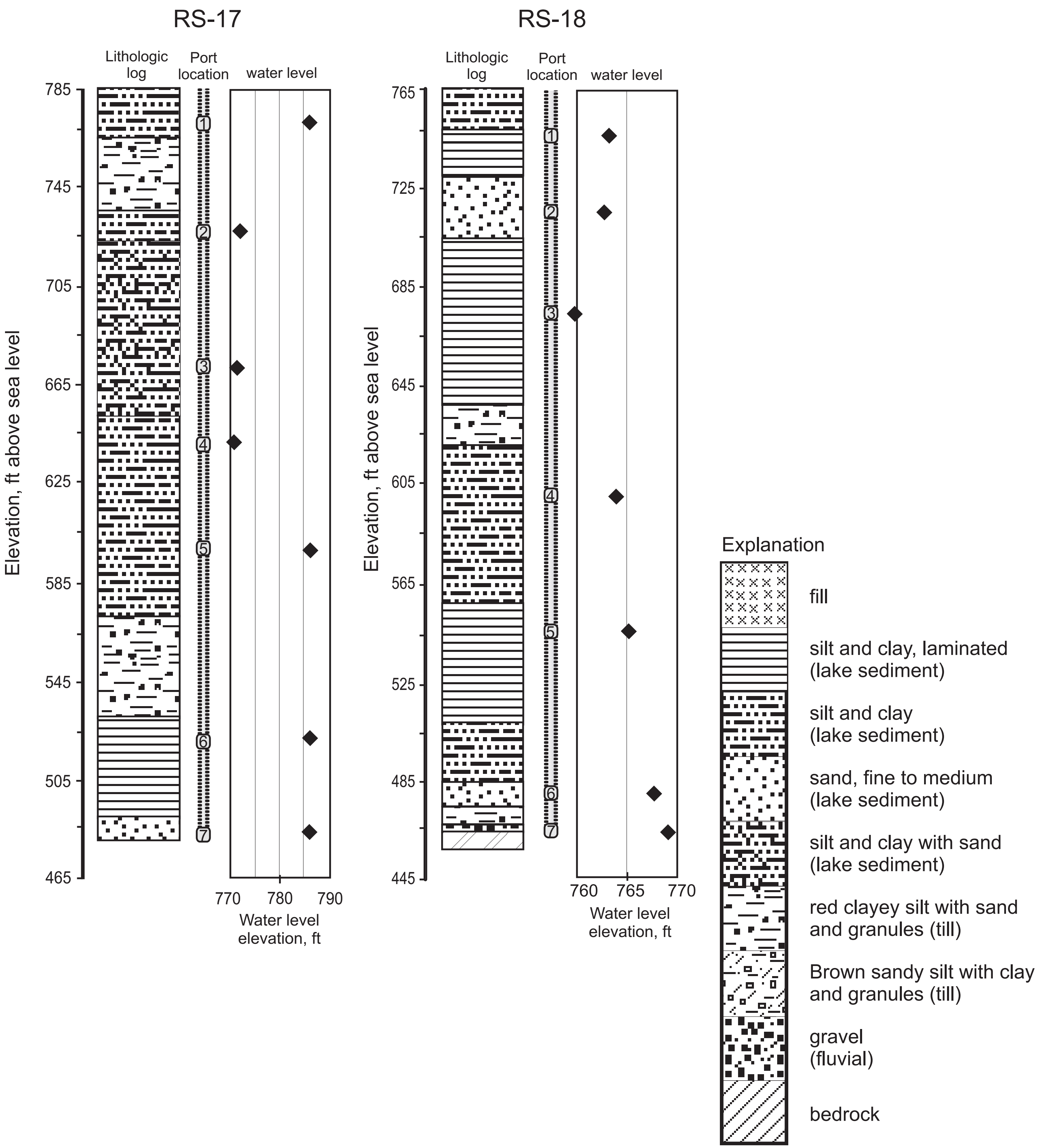


Figure 7. Lithologic logs, port locations, and water level elevations for rotosonic boreholes RS-17 and RS-18. The water level data for RS-17 and RS-18 represent single measurements made on June 8, 2007 and March 8, 2008, respectively. For RS-18, this was at least 8 months after ports were last purged. The borehole locations are shown on Figure 2 and 3.

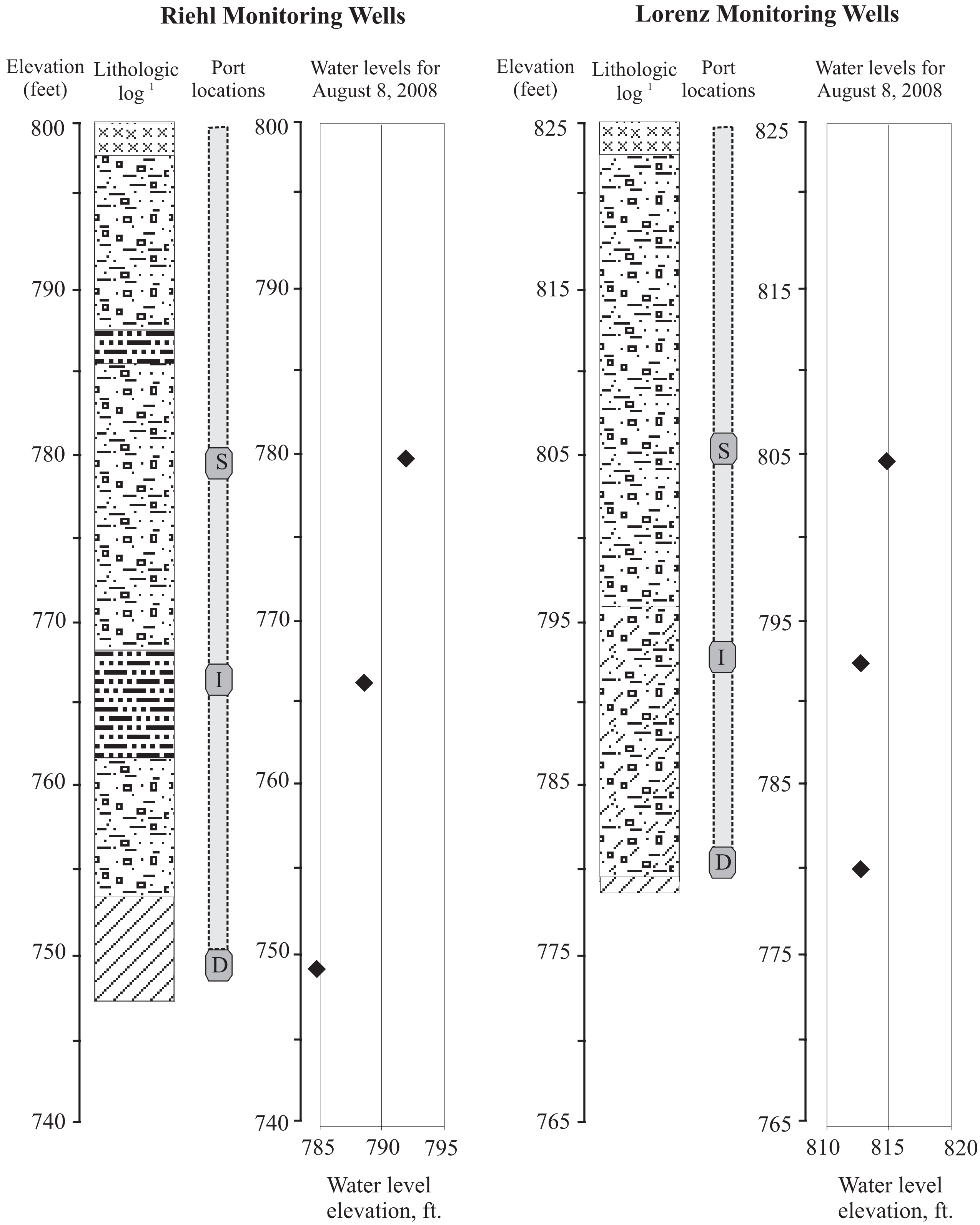


Figure 8. Lithologic logs, port locations, and water level elevations for Riehl and Lorenz monitoring wells. The water level data represents single measurements made on August 8, 2008, two months after ports were purged. The borehole locations are shown in Figure 2. The explanation for the lithologic log is included in Figure 7.

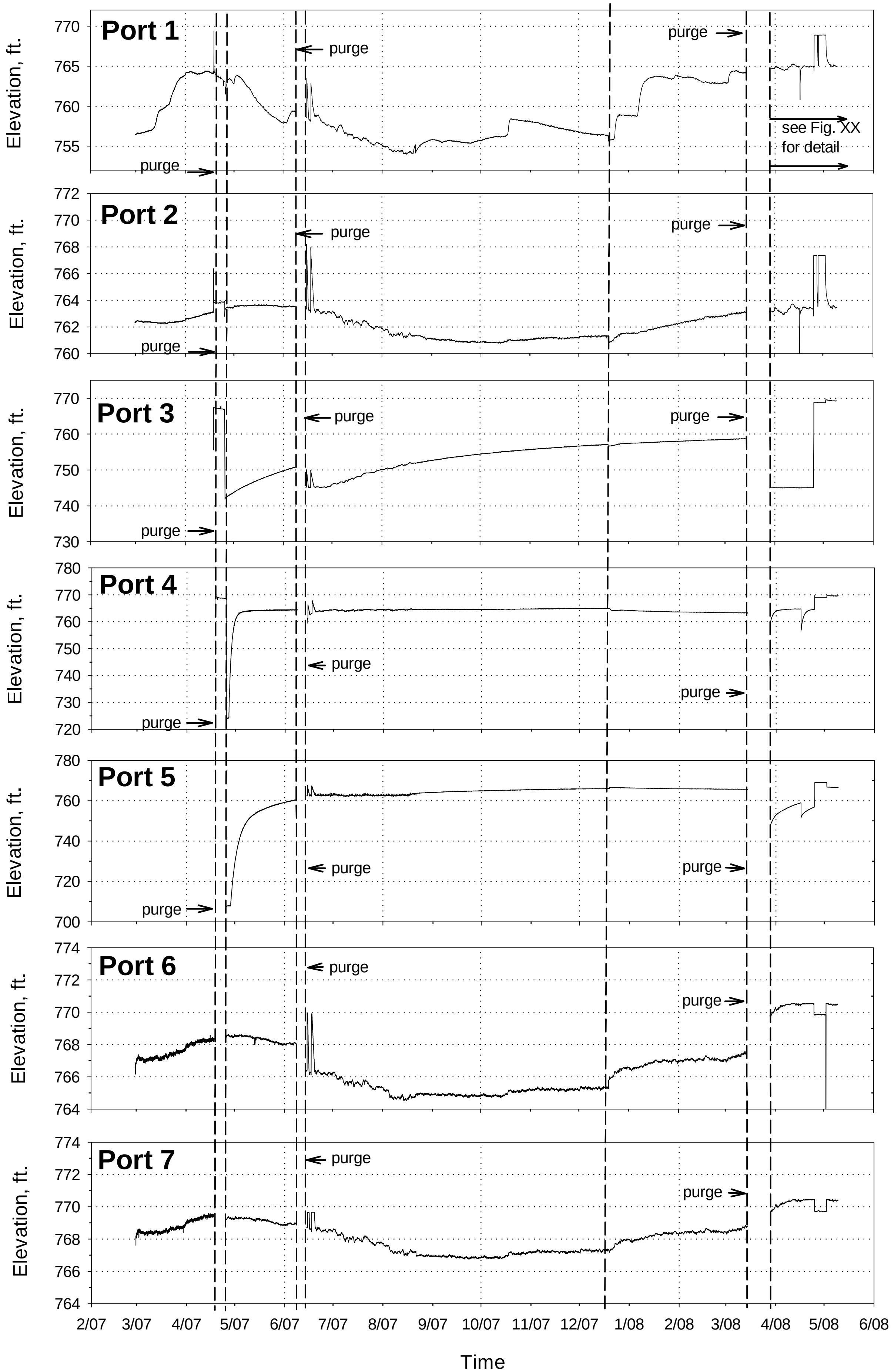


Figure 9a. Water level data for multilevel well RS-18

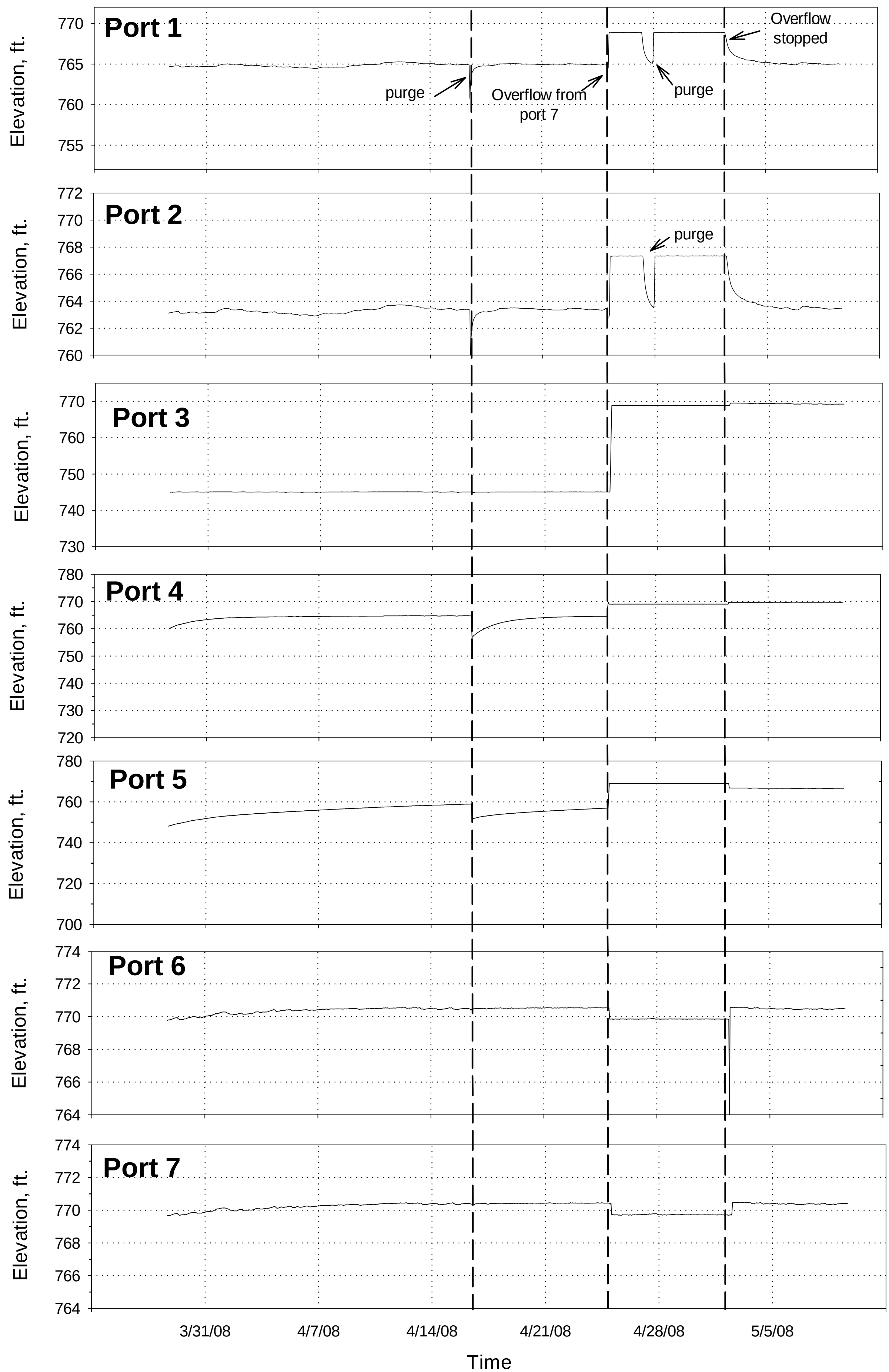


Figure 9b. Water level data for multilevel well RS-18 between March 25 and May 9, 2008.

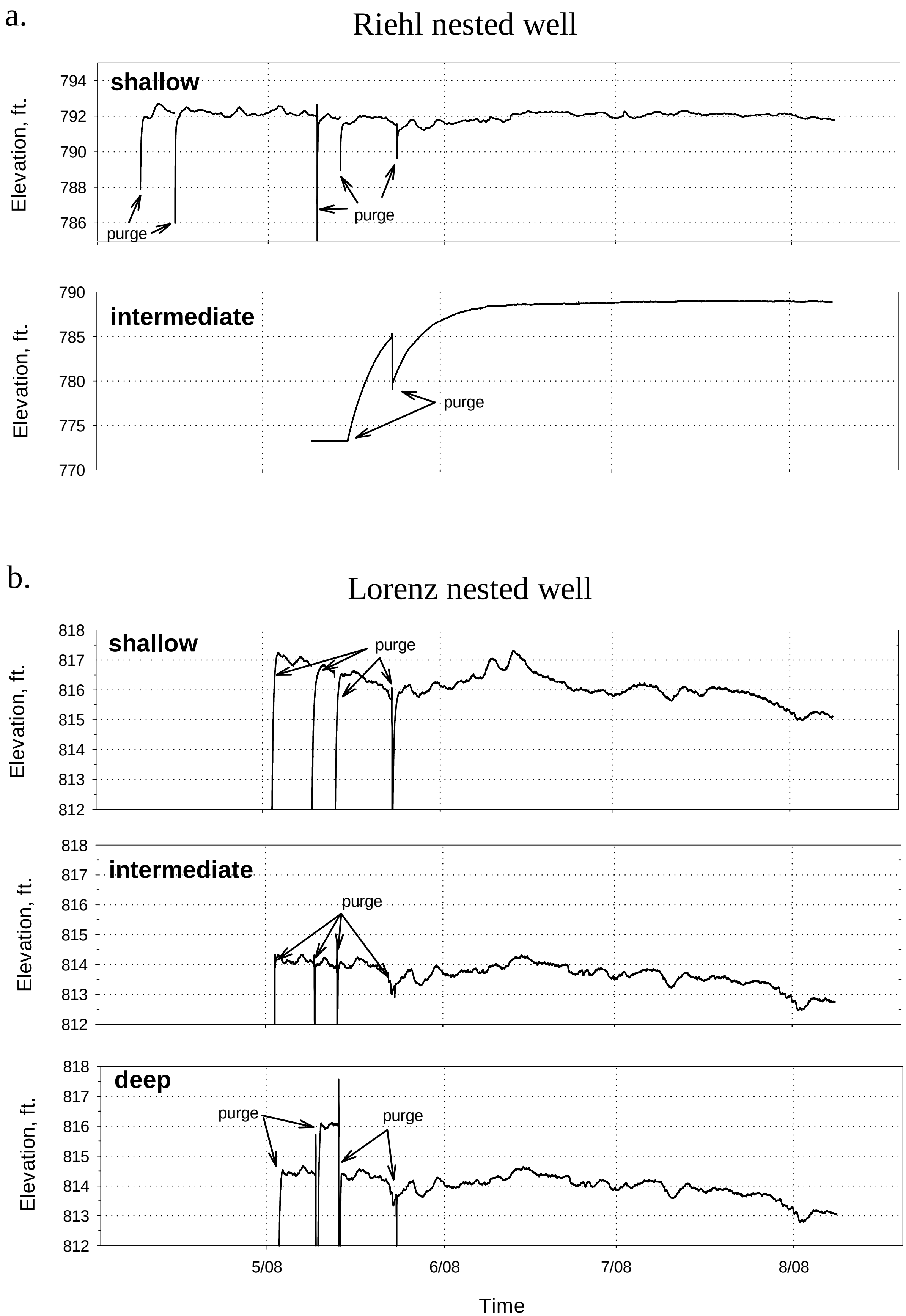


Figure 10. Water level data for nested wells at Riehl and Lorenz properties

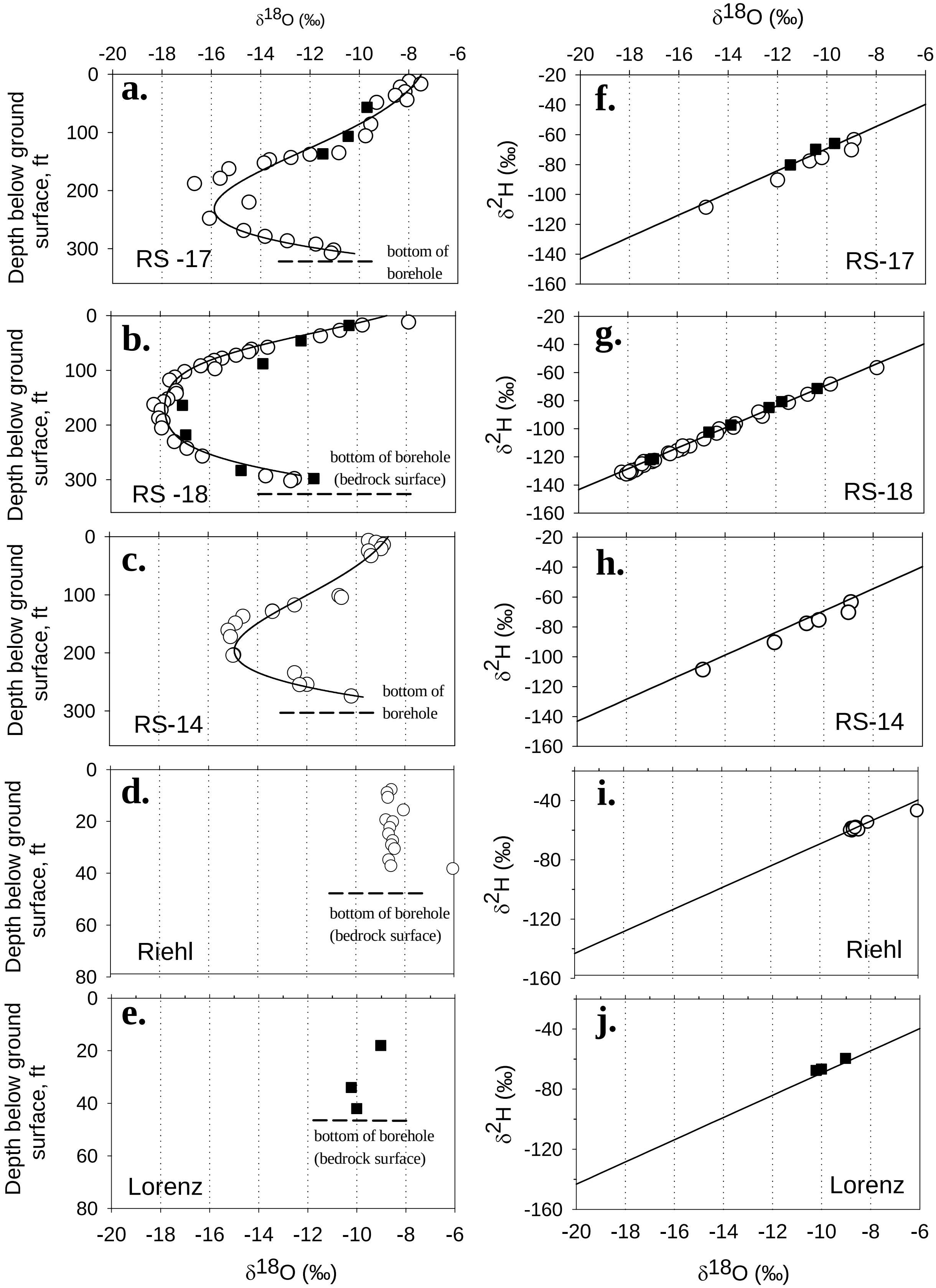


Figure 11. Oxygen isotope values as a function of depth below ground surface (a-e) and as a function of deuterium (f-j) for boreholes RS-17, RS-18, RS-14, Riehl and Lorenz. Sample analyses of pore water and well ports are shown as open circles and black squares, respectively. The black line in a-c represents a one-dimensional advection dispersion model fit to the data. The black lines in f-j represents the meteoric water line (Kendall and Coplen, 2001). The borehole locations are shown on Figure 3.

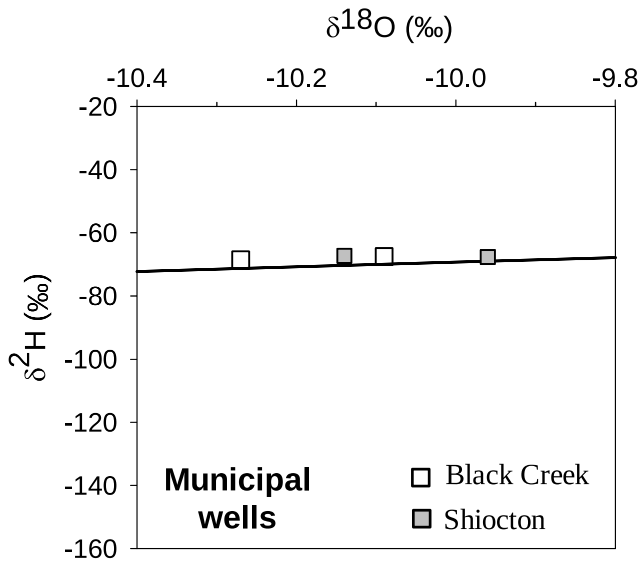


Figure 12. Oxygen isotope values as a function of deuterium for municapal wells located at the villages of Black Creek Shiocton. The black line represents the meteoric water line. The borehole locations are shown on Figure 3.

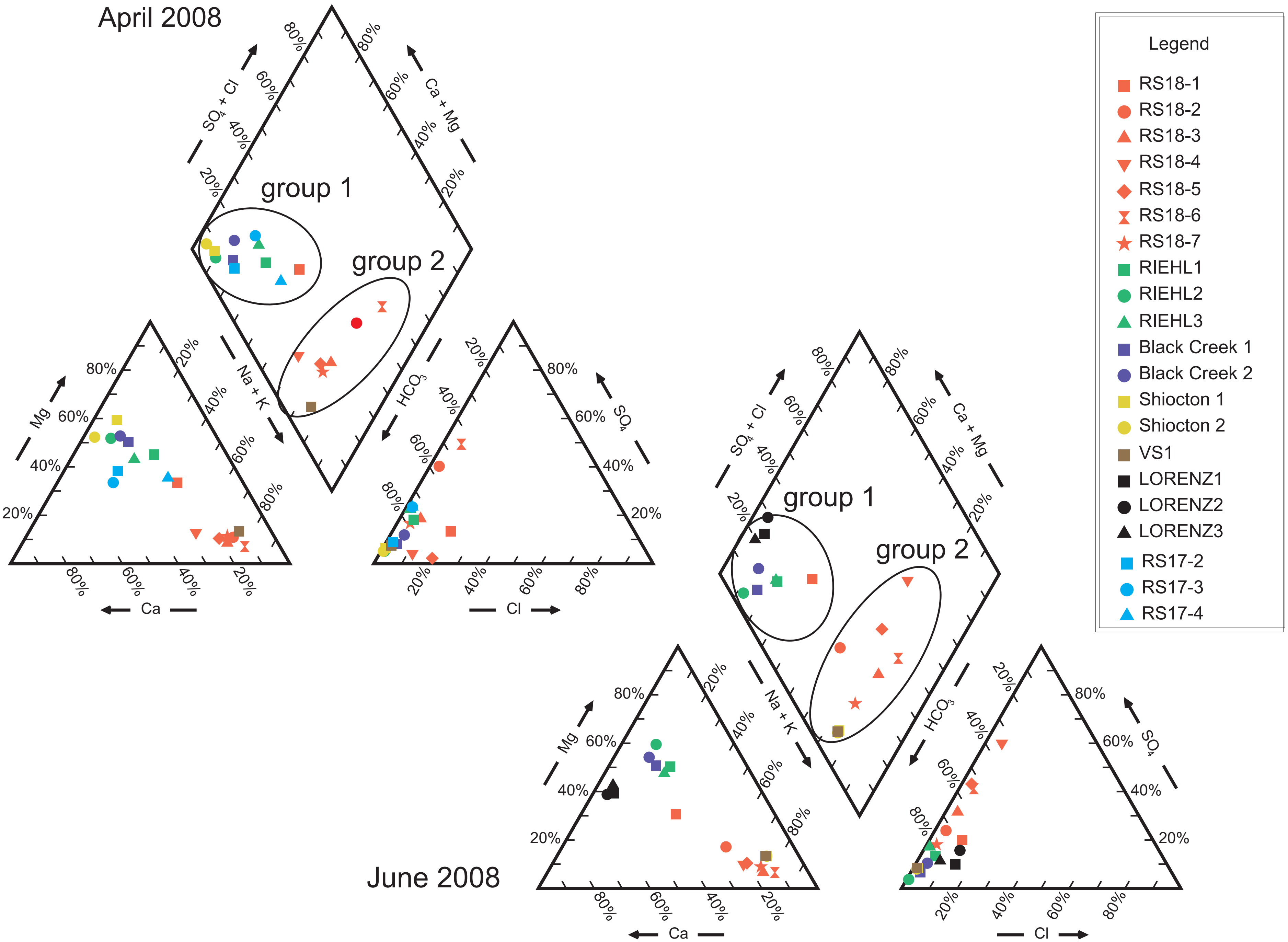


Figure 13. Piper diagram illustrating the proportions of common ions in water samples.

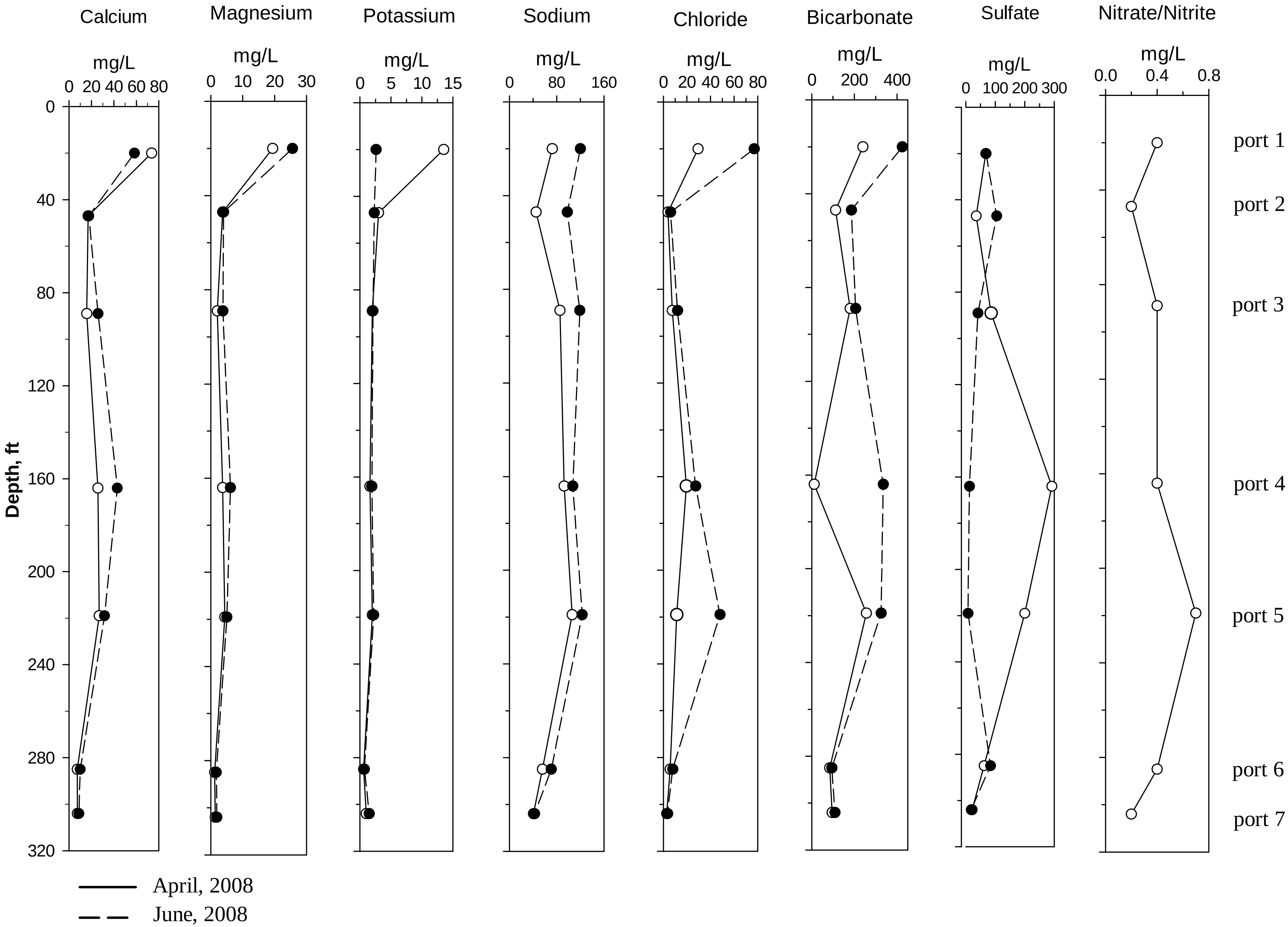


Figure 14. Concentration of major ions as a function of depth in the seven ports at RS-18. The two lines represent two different sampling dates

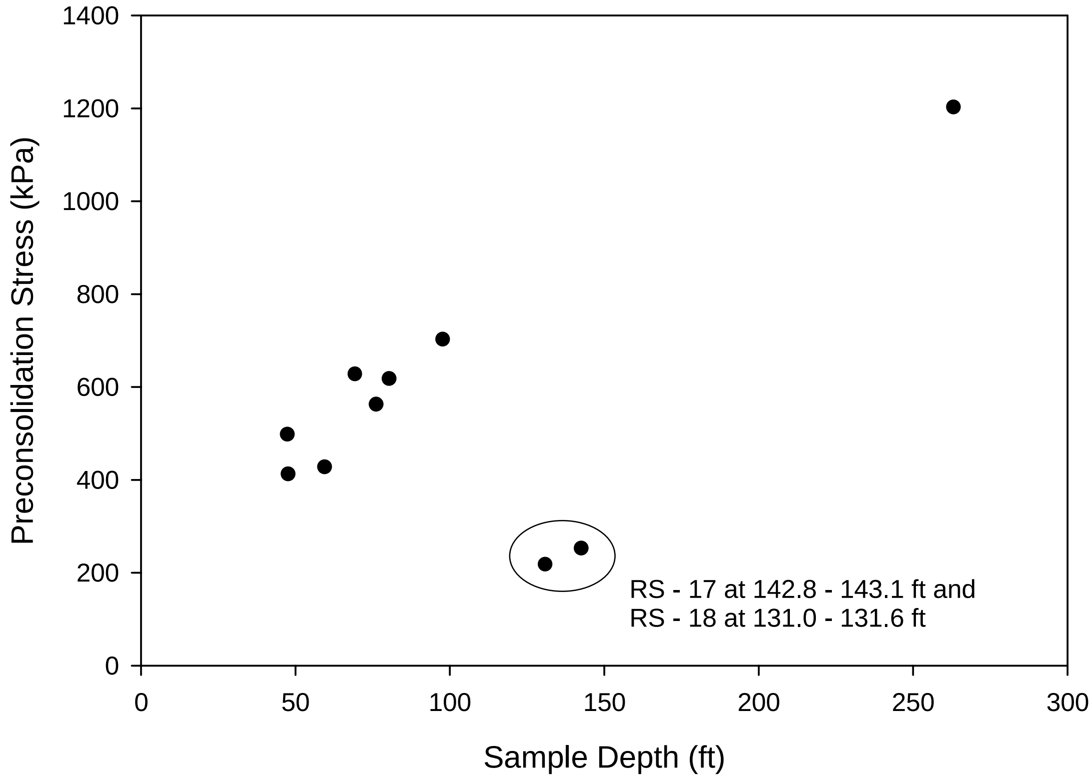


Figure 15. Preconsolidation stress as a function of sample depth.

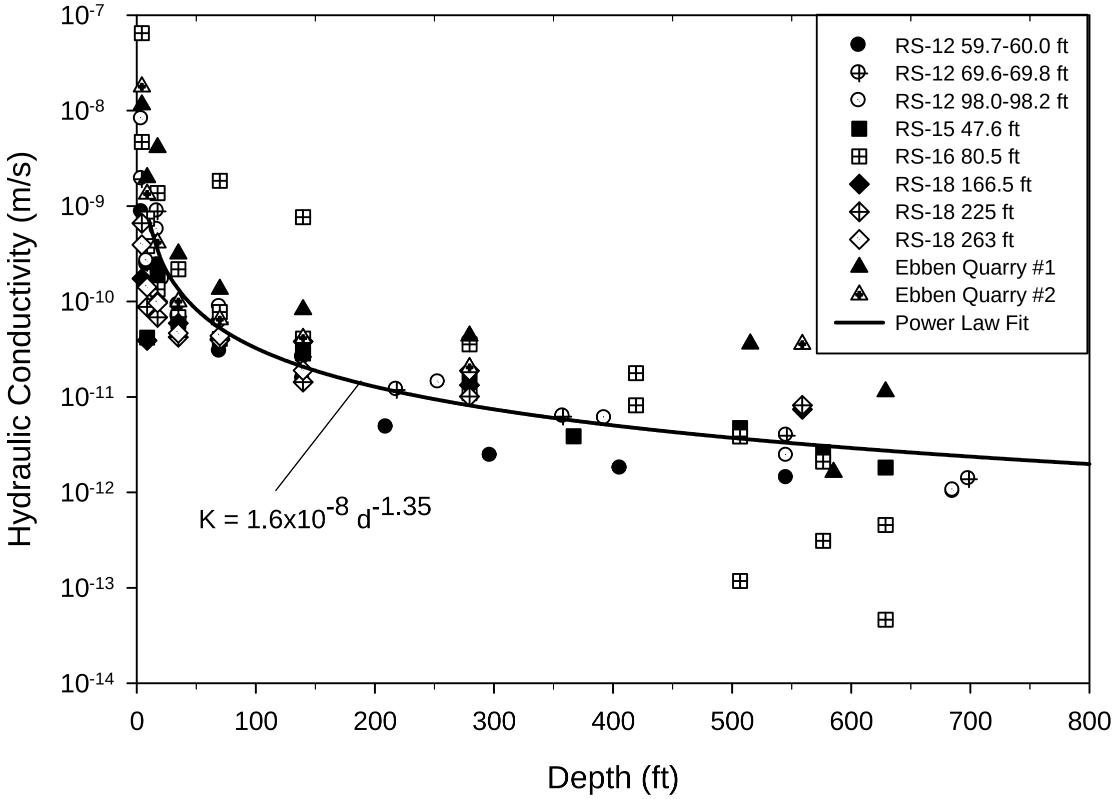


Figure 16. Hydraulic conductivity of the fine-grained sediment as a function of calculated burial depth determined from consolidation tests.

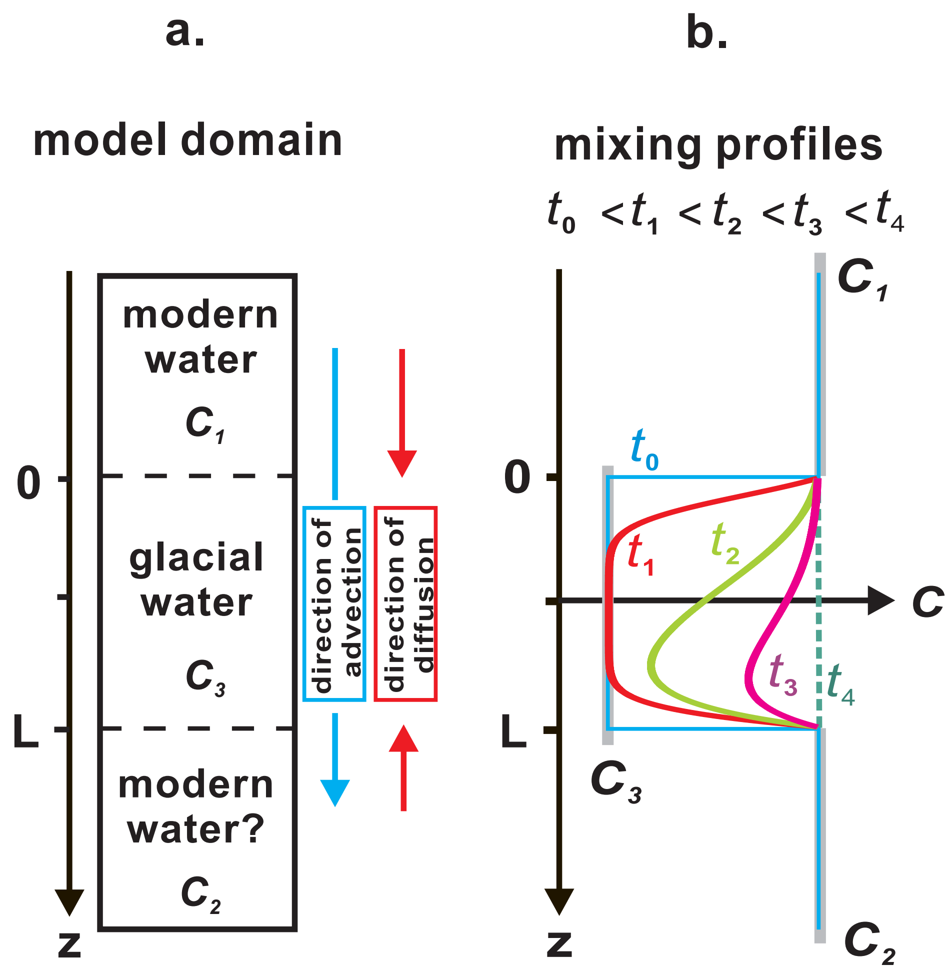
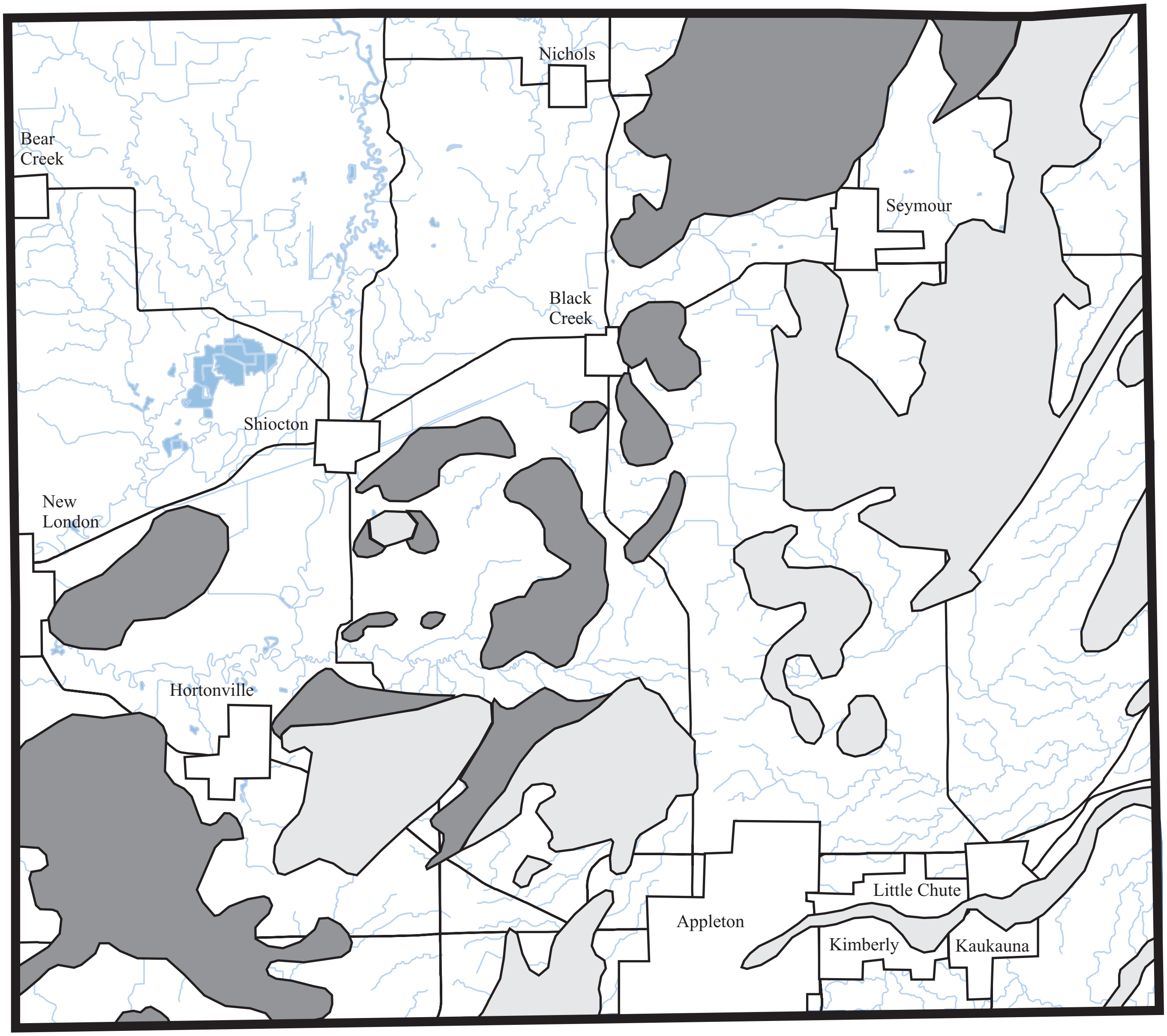


Figure 17 Mixing model schematic of (a) the model domain, and (b) the concentration of an index solute across the thickness of the aquitard



potential recharge areas to shallow aquifer (upper Sinnipee Group)

potential recharge areas to deep sandstone aquifer

Figure 18. Potential recharge areas to shallow and deep bedrock aquifers in Outagamie County based on depth to bedrock and the bedrock geology.