Groundwater recharge through a thick sequence of fine-grained sediment in the Fox River Valley, east-central Wisconsin. [DNR-194] 2007

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DRINKING WATER & GW

Jeff Helmuth
Department of Natural Resources
Bureau of Drinking Water and Groundwater
101 S Webster Street
Madison, WI 53702

RE: Final Report - Groundwater recharge through a thick sequence of fine-grained sediment in the Fox River Valley, East-Central Wisconsin.

Dear Jeff,

The attached report presents the results of the above referenced study and provides a brief summary of the data with some interpretations. We appreciate your agency funding this project and hope the information is useful in understanding the ground water resources of Wisconsin. Please contact me should you have any questions.

Sincerely,

Tom Hooyer
Geologist
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Madison, WI 53705
608.263.4175

Groundwater recharge through a thick sequence of fine-grained sediment in the Fox River Valley, east-central Wisconsin

Introduction

Groundwater in the Fox River Valley has been of significant interest for the past several decades due to the concern of groundwater availability in rapidly growing metropolitan areas like Green Bay, Appleton, Menasha and Oshkosh (LeRoux, 1957; Olcott, 1968; Krohelski, 1986; Batten and Bradbury, 1996). The water used by many municipalities, industries, and private residents in the Fox River Valley is pumped from carbonate and sandstone rocks known as the Cambrian-Ordovician aquifer system (Figure 1). These rocks lie on top of relatively impermeable Precambrian crystalline bedrock and dip eastward towards the Lake Michigan basin. Capping the Cambrian-Ordovician aquifer system is both the Maquoketa shale and dolomite of the Sinnipee Group, and/or fine-grained glacial sediment. Where present along the eastern margin of the Fox River Valley, the Maquoketa shale and the Sinnipee Group act as aquitards, confining to the aquifers below. Recharge to these aquifers is traditionally understood to occur west of the Maquoketa shale and Sinnipee Group subcrop (Krohelski, 1986; Batten and Bradbury, 1996; Conlon, 1998). However, a thick sequence of fine-grained glacial sediment covers large parts of this area and may act as an additional regional aquitard. The presence of this unit might shift the areas of recharge even further to the west than is currently assumed.

The fine-grained sediment that covers the Fox River Valley is the result of a large lake that once existed in front of the Green Bay Lobe of the Laurentide Ice Sheet as it retreated northward from its maximum extent around 21,000 years ago (Chamberlin, 1883; Alden, 1918; Thwaits, 1943). The extent of this lake, called glacial Lake Oshkosh, was controlled by the position of the receding ice margin and changing outlet elevations (Figure 2). The sediment deposited in the lake varies in thickness up to 500 ft as a result of an undulating bedrock surface. The distribution of this lake sediment is well known due to mapping efforts by the Wisconsin Geological and Natural History Survey (WGNHS). As part of this effort, a series of boreholes (Figure 2) have been drilled to recover a complete sequence of glacial sediment within the lake basin. The cores revealed a detailed record of thick, fine-grained sediment, containing lake sediment and till within a buried bedrock valley system. The regional extent and thickness of this find-grained sediment will act as a control of recharge to the underlying aquifer system.

Given the importance of groundwater resources in the Fox River Valley, this study was aimed at understanding vertical groundwater flow through the fine-grained glacial sediment to deeper bedrock aquifers. To characterize the lake sediment sequence, two boreholes were drilled to collect continuous core and install multilevel wells. Representative samples of the core were used to determine physical properties of the sediment, including hydraulic conductivity. To evaluate vertical groundwater flow across the sediment sequence we installed a multilevel well in each borehole to determine the distribution of hydraulic head across the sediment sequence

Pore water was also extracted from core sub samples to evaluate the relative age of the pore water using stable isotope of oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$). Analyses of pore water using these isotopes can be used to determine groundwater flow paths and estimate residence times in fine-grained sediments (Perry et al., 1982; Desaulniers, 1988; Hendry and Wassenaar, 1999; Remenda et al., 1994). For example, a light $\delta^{18}O$ value compared to a modern precipitation value might indicate the presence of glacial-age water in a given sediment sequence. Such old water has been observed in lake sediments from Manitoba and Ontario, Canada (e.g., Remenda et al., 1994; Birks et al., 2007). The presence of such old water in the sediment in the Fox River Valley would reveal that little to no groundwater recharge is occurring through the sediment to bedrock aquifers.

The results of this study show that the extensive nature and low hydraulic conductivity of the sediment limits vertical recharge to underlying bedrock aquifer systems. Instead, recharge to these aquifers is probably originating from areas where the sediment cover is thin or further to the west beyond the extent of the area once occupied by glacial Lake Oshkosh.

METHODS

Drilling and well installation

Two boreholes were drilled using rotosonic drilling methods which limits the introduction of drilling fluids from the borehole into the surrounding formation. It also allows the collection of continuous 4-inch diameter core that remains intact for cohesive sediment such as clay and silt. In addition, the drilling process also includes driving an 8-inch diameter casing is down the borehole forming an annular space suited for multilevel well installation. The two boreholes drilled for this study, RS-17 and RS-18, are 307 feet deep and 308 feet deep, respectively. Both boreholes were drilled in Outagamie County along the axis of a buried bedrock valley. This valley cuts through the Cambrian-Ordovician aquifer system to the Precambrian crystalline bedrock (Figure 3b). The continuous core collected from each borehole was described and sampled in the laboratory. Upon reaching the final drilling depth at each borehole, the natural gamma radiation was measured using downhole geophysical equipment to distinguish sand layers from clay sequences.

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 through the middle of the steel casing. This casing was removed 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 well 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 of sand. In between these sand packs, the borehole annulus was also backfilled with coated bentonite pellets.

Following installation, both 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. In addition to this technique, 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, automatic water level measuring devices, level loggers®, were installed to collect data at fifteen-minute intervals. Water samples were also collected from most of the wells using a peristaltic pump.

Consolidation experiments

Several intact samples were collected from each core for consolidation testing to determine preconsolidation stress and hydraulic diffusivity (Grisak and Cherry, 1975; Keller et al., 1989). In this 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. Under initial loads, 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.

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:

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

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, this equation is solved for hydraulic diffusivity, D, by using the rate of consolidation, $\Delta w(t)$, for each incremental load. 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).

Extraction of pore water for stable isotope analyses

Samples from the core were collected at regular intervals to extract pore-water. 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. At one end of the cyclinder there is a piston, driven by compressed nitrogen that consolidates the sample. This consolidation 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 analyses, water samples were collected from the multilevel wells for comparison. Prior to sampling, the wells were purged to remove water present in the well casing. All samples were analyzed for $\delta^{18}O$ and $\delta^{2}H$ by direct pore-water equilibration at the University of Waterloo Environmental Isotope Laboratory.

RESULTS

The lithologic logs and well port locations for RS-17 and RS-18 are shown in Figure 4. Water level data is also included but only represents a single measurement made on June 8, 2007, six weeks after the ports were purged. A review of the logs reveals that RS-17, located in the city of Black Creek, was drilled to a total depth of 307 feet penetrating 300 feet of fine-grained sediment (silt and clay) before encountering 7 feet of fine sand. Further to the west along the axis of the buried valley near the city of Shiocton, RS-18 was 308 feet 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 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 screens or ports associated with the multilevel wells were positioned in each borehole so that water samples could be collected and water levels could be measured to assess the distribution of hydraulic head. The water levels measured in RS-17 and RS-18 are presented in Figure 5 and Figure 6, respectively. Review of the data for RS-17 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) appear to be operational as water levels have stabilized at slightly different elevations over a seven month period. Water levels measured in RS-18 do not appear to show cross connection. In three ports located in more conductive sand and gravel (2, 6, and 7), the water levels were stable and appear to recover within an hour after purging. However, for the ports located in the silt and clay (3, 4, and 5), the water levels appear to take longer to equilibrate, on the order of weeks, following purging. Although the water levels has recovered in Ports 4 , the water level in Port 3 and 5 are still rising. For the near surface Port 1, the water level initially fluctuates before gradually decreasing.

Consolidation testing conducted on lake sediment samples yield values of hydraulic conductivity that range from 1 x 10^{-8} ms^{-1} to 1 x 10^{-11} ms^{-1} depending on the applied load (Figure 7). A more representative value for hydraulic conductivity is obtained when the specimen was subjected to its maximum effective stress (σ_{pc}) defined as the maximum load the specimen was subjected to during its geological history. The average hydraulic conductivity associated with these values is 3.5 x 10^{-10} ms^{-1} (Table 1). This value is about an order of magnitude smaller than the hydraulic conductivity determined from consolidation tests in previous studies that averaged 3.8 x 10^{-9} ms^{-1}.

To evaluate the relative age of water pore-water in the fine-grained sediment, stable $\delta^{18}O$ and $\delta^{2}H$ analyses were analyzed on 23 samples at RS-17 (Figure 8a) and 30 samples from RS-18 (Figure 8b). Results of the $\delta^{18}O$ analyses 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 that was drilled prior to this project in the Village of Black Creek (Figure 8c). For each sample the $\delta^{18}O$ value was also plotted verses its $\delta^{2}H$ value to assess potential fractionation (Figure 8d-f). 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 and RS-18. These values are shown in Figures 8a and b as the dark squares. Where collected, the $\delta^{18}O$ values from the multilevel wells appear to mimic the pore-water extracted from the sediment.

DISCUSSION

While the Maquoketa shale and Sinnipee Group have been considered to be the primary confining units for the Cambrian-Ordovician aquifer system throughout the Fox River Valley, results from this study indicate that the fine-grained sediment is an aquitard that impedes downward groundwater flow and recharge. This is especially the case where the sediment is thick such as the buried bedrock valleys. The two boreholes drilled as part of this study confirm previous studies that show limited sand and gravel deposits throughout the sequence. Even readvances of the Green Bay Lobe into glacial Lake Oshkosh appear to produce fine-grained tills that consist primarily of reworked lake sediment (clay and silt).

The fine-grained nature of the sequence is demonstrated by the low values of hydraulic conductivity determined in consolidation tests. These values could potentially be as high as 2×10^{-8} ms^{-1} depending upon the effective stress that is a function of sample depth and sediment pore-water pressure where the sample was collected. If the pore-water pressure at a given depth equals the overburden pressure (zero effective stress), then the appropriate hydraulic conductivity would be associated with low axial stress in the consolidation tests. For example, in sample RS-18, 263.2 to 263.8 feet, the hydraulic conductivity could potentially range from 2×10^{-8} ms^{-1} to 2×10^{-10} ms^{-1}. Despite not knowing the exact pore-water pressure at this sample location, hydraulic conductivity values would never be greater than 2×10^{-8} ms^{-1}. The effect of elevated pore-water pressures in the sediment will be evaluated in a future project along with conducting slug tests to determine hydraulic conductivity in the field.

Water level data from RS-18 indicates an upward head gradient of about 10 feet between the upper and lower ports although it is apparent that not all ports have equilibrated. In RS-17, the water level data indicates a slight downward gradient of about 1 foot over an 80 foot-thick section of lake sediment. However, the comprised ports at this well and the lack of high resolution water level data make it difficult to evaluate. Using these gradients and the average hydraulic conductivity from the consolidation experiments, a vertical flux of 4.4×10^{-12} ms^{-1} and 1.3×10^{-11} ms^{-1} is calculated for RS-17 and RS-18, respectively. Such a low rate of advection suggests that diffusion may be the primary mechanism for movement of water through the sediment. If this is the case, then it should be reflected in the chemistry of the pore water.

Given that advection through the fine-grained sediment may be limited, the fluctuating water level in Port 1 at RS-18 indicates that the upper part of the sequence might transmit water. This may be the result of near-surface 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 the difficultly of identifying fractures in silt and clay. The $\delta^{18}O$ results of the water samples taken from the multilevel wells indicate that fractures do not penetrate to any significant depth in the glacial sediment. If fracture flow were the dominant mechanism moving water through the sediment, you would not expect the stable isotope values of the multilevel wells to match those of the sediment pore water throughout the entire thickness of the lake sediment. However, the results of the well samples match those taken from the pore water at RS-17 and RS-18 and show a similar $\delta^{18}O$ signature throughout the profile (Figures 8a and 8b). This suggests that diffusion may indeed be the primary mechanisms of flow throughout the section and that fractures are limited with depth.

$\delta^{18}O$ results from pore-water analyses at RS-14, RS-17, and RS-18 indicate relatively old pore water supporting the assertion that vertical advection is limited. 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 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 at., 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. However, this is not the situation indicating that groundwater recharge must be occurring elsewhere.

The bow-shaped stable isotope profiles indicate that the light $\delta^{18}O$ signal is present where the aquitard is sufficiently thick. A review of oxygen isotope profiles from four other boreholes (RS-12, RS-13, RS-15 and RS-16), drilled prior to this study, show vertical profiles indicative of pore water with heavier $\delta^{18}O$ values similar to modern precipitation throughout the thickness of the sediment (Figure 9). In all of these boreholes the fine-grained sediment is not as thick. Thus, one explanation is that sufficient time has elapsed for the pore-water to diffuse through the sequence. It may also be the case that some advection is occurring or that fractionation of the $\delta^{18}O$ is affecting the isotopic signature observed at these locations.

CONCLUSIONS

The fine-grained glacial deposits in the Fox River Valley greatly limit surface water infiltration and groundwater recharge to the bedrock aquifers, particularly west of the Fox River where the Maquoketa shale and Sinnipee Group subcrop. The nature of the sediment is a critical variable in understanding flow patterns and recharge to the underlying aquifers. The thick sequence of fine-grained sediment in the buried bedrock valley in Outagamie County prevents downward migration of modern precipitation to the bedrock underneath. However, groundwater flow is still uncertain in areas where the sediment is much thinner over bedrock uplands. Thus, we are currently planning to collect hydrological data and investigate upland areas with thinner glacial sediment to define local conditions and improve our understanding of the regional groundwater flow system. Where the glacial sediment may only be 15 to 30 feet thick, the bedrock aquifers are much closer to the surface and might be susceptible to groundwater recharge (LeRoux, 1957; Olcott, 1968). In future work we will continue monitoring our existing deep wells and install two new shallow wells where the sediment is significantly thinner and potentially fractured (Helmke et al., 2005). We expect these shallow wells to yield information regarding the flux of groundwater to the bedrock aquifer so that distribution of recharge to the deep aquifer can be determined. Since large portions of the Cambrian-Ordovician aquifer lie beneath the fine-grained till and lake sediment, understanding the spatial distribution of recharge in this region is crucial for planning the future management of groundwater quality and quantity.

REFERENCES

Alden, W.C., 1918. The Quaternary geology of southeastern Wisconsin with a chapter on the older rock formations. U.S. Geological Survey Professional Paper 106, 356 p.

Attig, J.W., Mickelson, D.M., and Clayton, L., 1989. Late Wisconsin landform distribution and glacier-bed conditions in Wisconsin. Sedimentary Geology, 62, 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, 28p.

Birks, S.J., Edwards, T.W.D., Remenda, V.H., 2007. Isotopic evolution of Glacial Lake Agassiz: New insights from cellulose and porewater isotopic archives. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 246, 1, 8-22.

Chamberlin, T.C., 1883. General geology, *in* Geology of Wisconsin, v. 1: Madison, Wisconsin, Commissioners of Public Printing, p. 3-300.

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.

Desaulniers, D. E. and Cherry, J.A.1988. Origin and movement of groundwater and major ions in a thick deposit of Champlain Sea clay near Montreal. *Canadian Geotechnical Journal,* 26, 80-89.

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

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

Gerber, R. E., J. I. Boyce, et al. (2001). Evaluation of heterogeneity and field-scale groundwater flow regime in a leaky till aquitard. *Hydrogeology Journal,* 9, 1, 60-78.

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, 23-43

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, 201-209.

Hendry, M.J., Wassenaar, L.I., 1999. Implications of the transport of *D* and ^{18}O in porewaters on groundwater flow and the timing of geologic events in a thick aquitard system. *Water Resources Research*, 35, 6, 1751–1760.

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, 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, 1363–1393.

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

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

Olcott, P.G., 1968. Water resources of Wisconsin, Fox-Wolf river basin. Hydrologic Investigations Atlas, U. S. Geological Survey. Reston, VA, United States, 4 sheets.

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, pp. 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, 2979-2987.

Remenda, V.H., Cherry, J.A., Edwards, T.W.D., 1994. Isotopic composition of old ground water from Lake Agassiz: implications for Late Pleistocene climate. *Science,* 266, 1975–1978.

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, 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, 283-319.

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

Thwaites, Fredrik Turville, 1943. Pleistocene of part of northeastern Wisconsin. *Geological Society of America Bulletin*, 54, 1, 87-144.

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

Table 1. Results from consolidation experiments

Sample location	Depth (ft.)	Maximum effective stress (σ_{pc})	Hydraulic diffusivity	Calculated hydraulic conductivity
Consolidation tests from this study				
RS-17	47.7 – 48.3	410 kPa	1×10^{-6} m^2s^{-1}	7×10^{-10} ms^{-1}
RS-17	142.8 – 143.1	250 kPa	3×10^{-6} m^2s^{-1}	2×10^{-10} ms^{-1}
RS-18	131.0 – 131.6	215 kPa	1×10^{-6} m^2s^{-1}	3×10^{-10} ms^{-1}
RS-18	263.2 – 263.8	1200 kPa	8×10^{-7} m^2s^{-1}	2×10^{-10} ms^{-1}
Consolidation tests from previous studies				
RS-12	59.7 – 60.0	425 kPa	2×10^{-4} m^2s^{-1}	3×10^{-9} ms^{-1}
RS-12	69.6 – 69.8	625 kPa	4×10^{-4} m^2s^{-1}	3×10^{-9} ms^{-1}
RS-12	98.0 – 98.2	700 kPa	7×10^{-4} m^2s^{-1}	4×10^{-9} ms^{-1}
RS-15	47.6 – 48.0	495 kPa	1×10^{-6} m^2s^{-1}	8×10^{-9} ms^{-1}
RS-16	76.0 – 77.0	560 kPa	1×10^{-6} m^2s^{-1}	4×10^{-9} ms^{-1}
RS-16	80.5 – 81.0	615 kPa	3×10^{-6} m^2s^{-1}	1×10^{-9} ms^{-1}

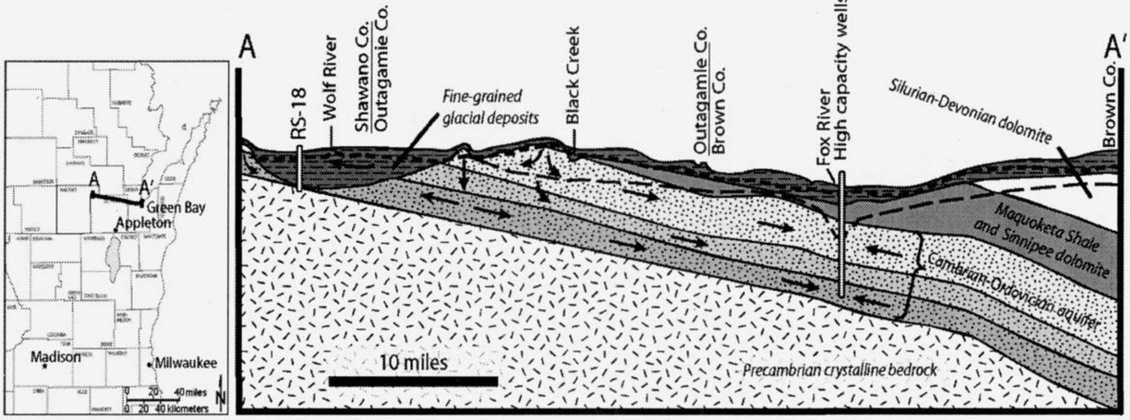

Figure 1. Idealized cross section showing the extent of the Cambrian-Ordovician aquifer and the overlying confining unit of the Maquoketa Shale and the Sinnipee Group. The entire sequence is capped by relatively impermeable fine-grained sediment (modified after Conlon, 1998; Batten and Bradbury, 1996).

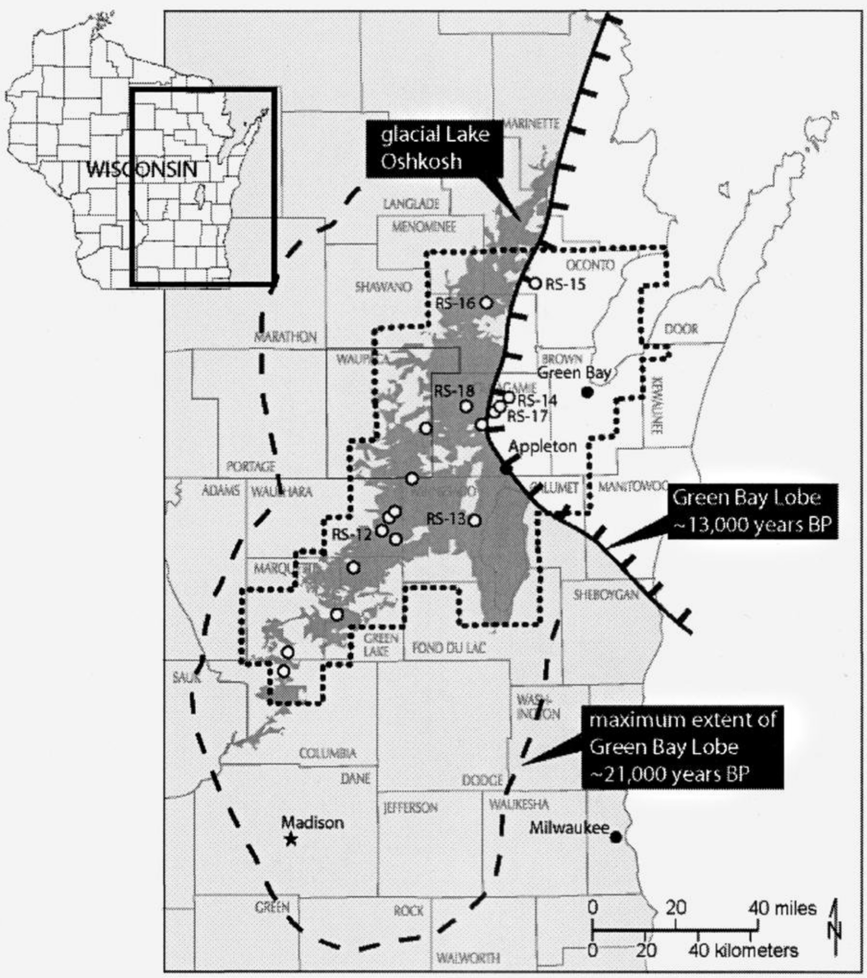

Figure 2. Map of east central Wisconsin. The shaded gray area represents the extent of glacial Lake Oshkosh and the dotted line encompasses the area currently being mapped by the WGNHS. The dots represent locations of rotosonic boreholes drilled by the WGNHS.

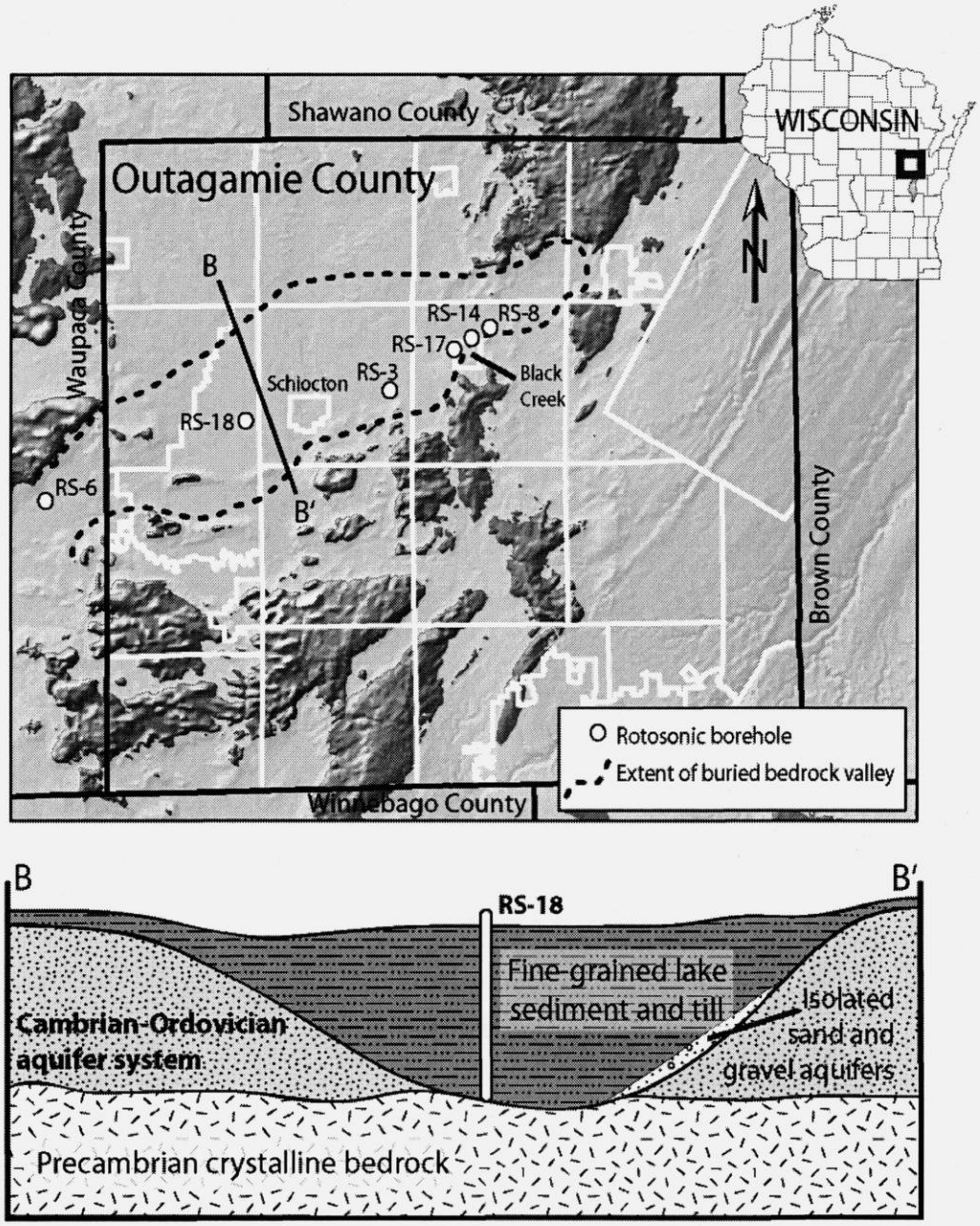

Figure 3. Shaded relief map of Outagamie County, showing the lateral margins of the buried bedrock valley. The lighter gray shaded surface is area once covered by glacial Lake Oshkosh. Boreholes RS-17 and RS-18 contain multilevel well systems currently being monitored.

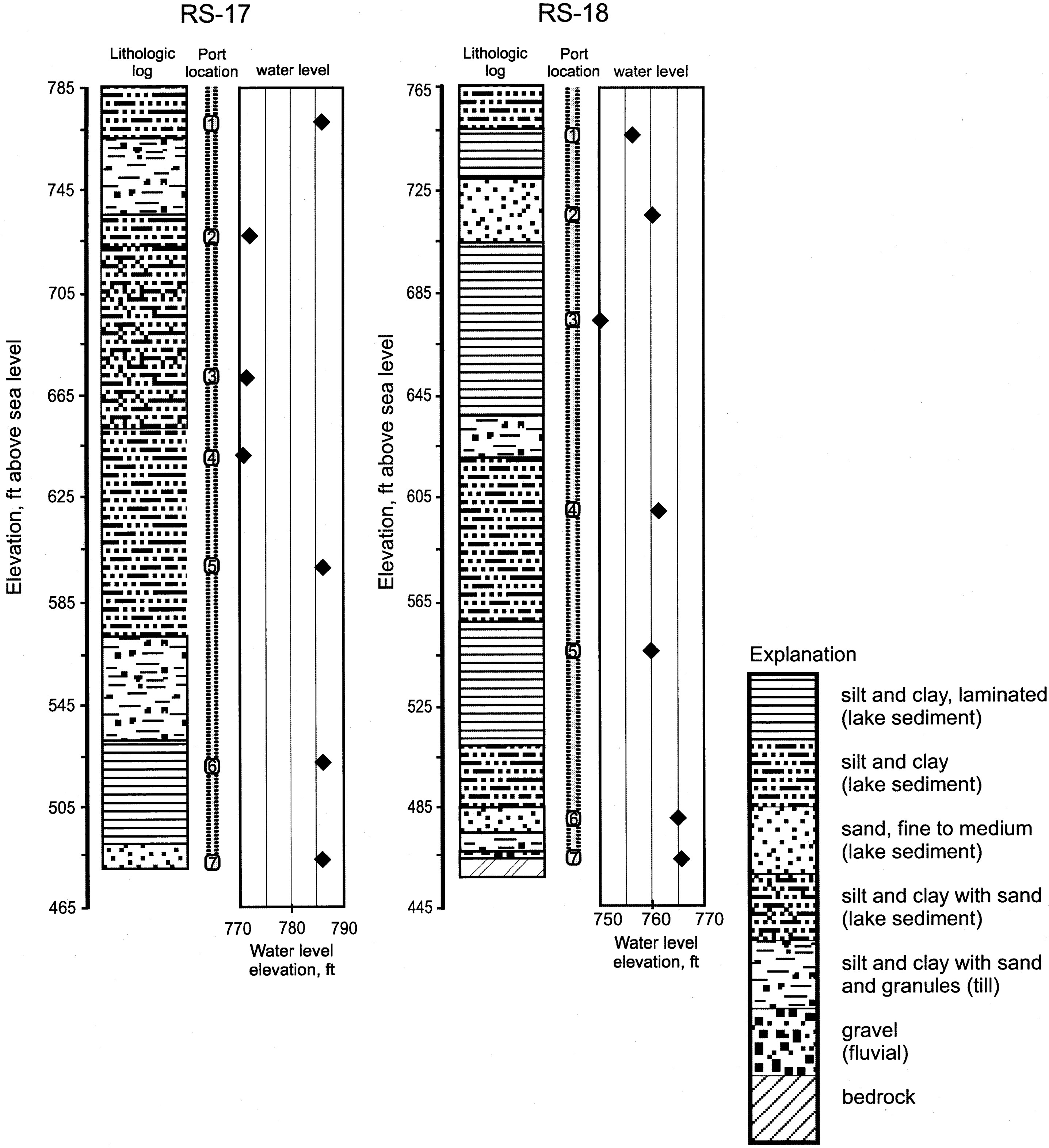

Figure 4. Lithologic logs, port locations, and water level elevations for rotosonic boreholes RS-17 and RS-18. The water level data represents single measurements made on June 8, 2007, six weeks after ports were purged. The borehole locations are shown on Figure 2 and 3.

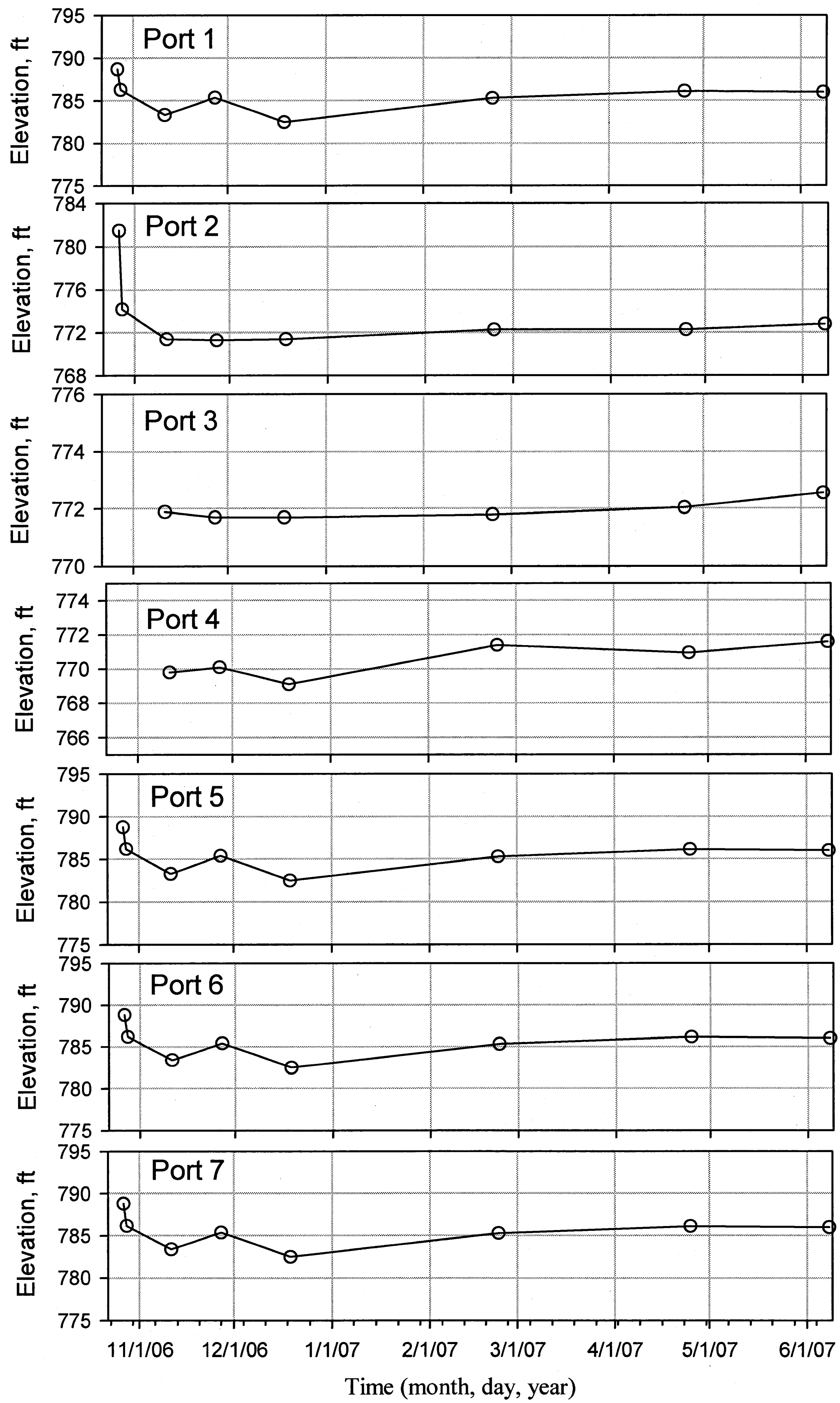

Figure 5. Water level data for multilevel well RS-17

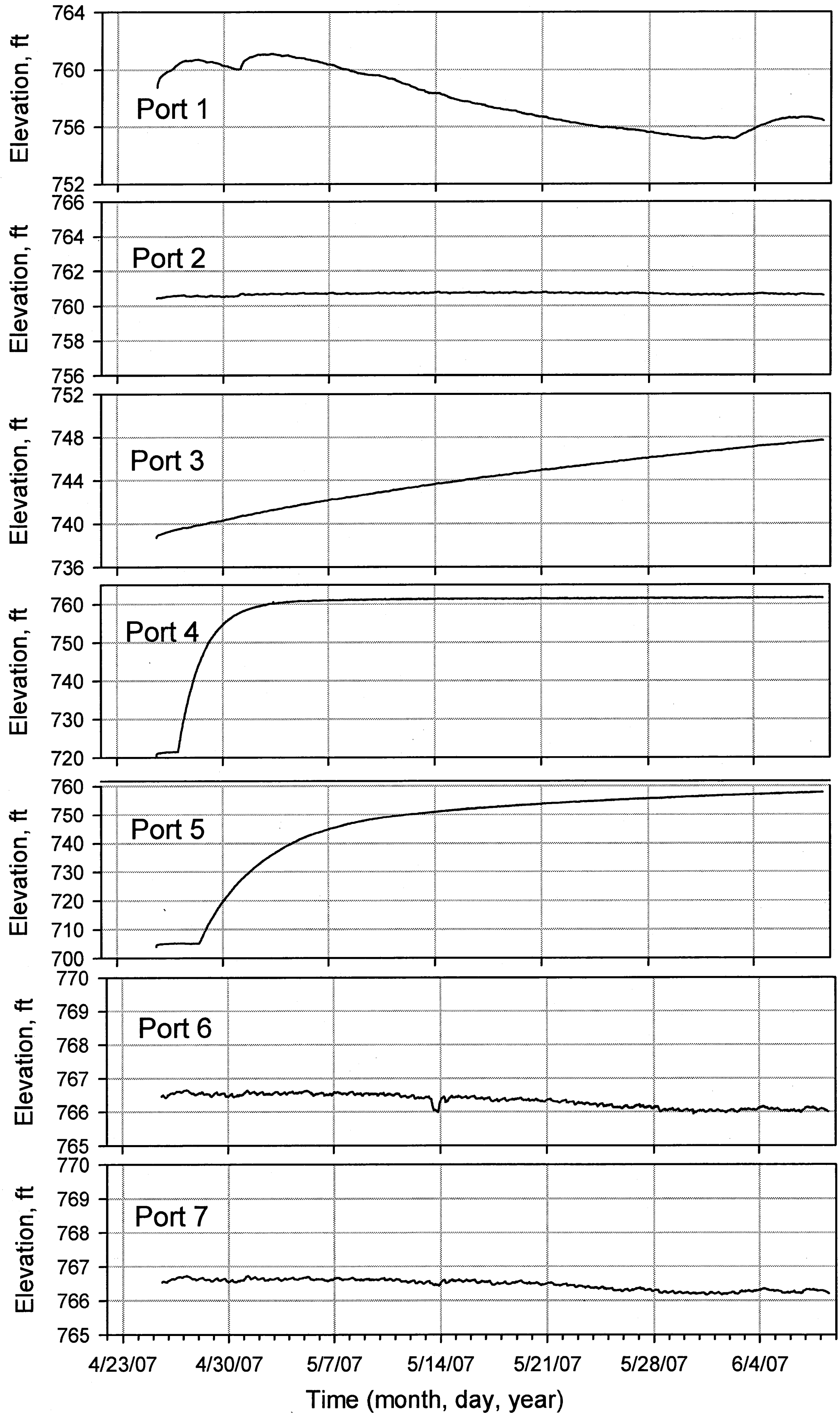

Figure 6. Water level data for multilevel well RS-18

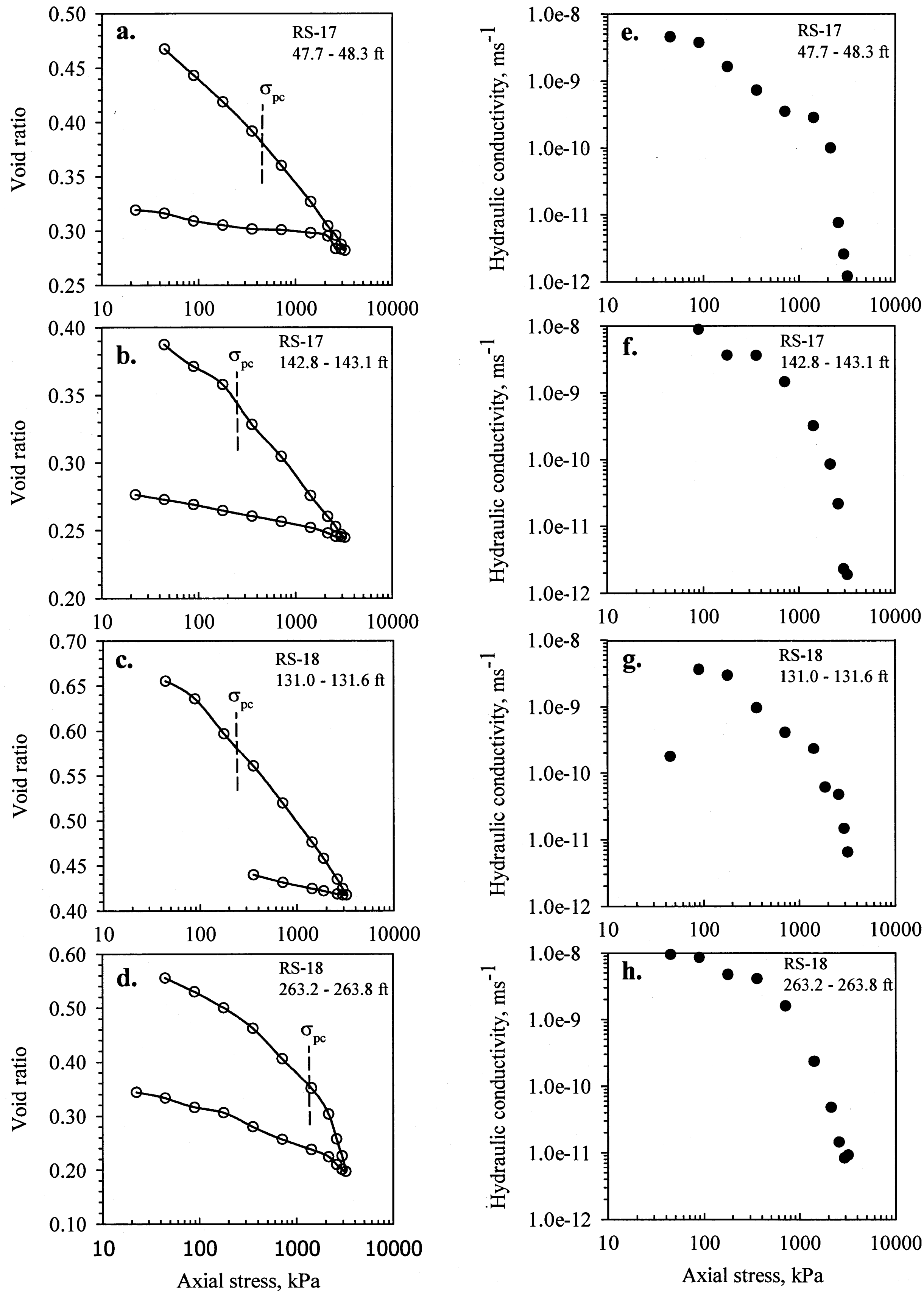

Figure 7 Consolidation curves (a-d) and hydraulic conductivity values (e-h) of glacial Lake Oshkosh sediment samples as function of total normal stress, determined from tests with a fixed-ring consolidometer. The samples were collected from large-diameter core whose locations are shown on Figure 3.

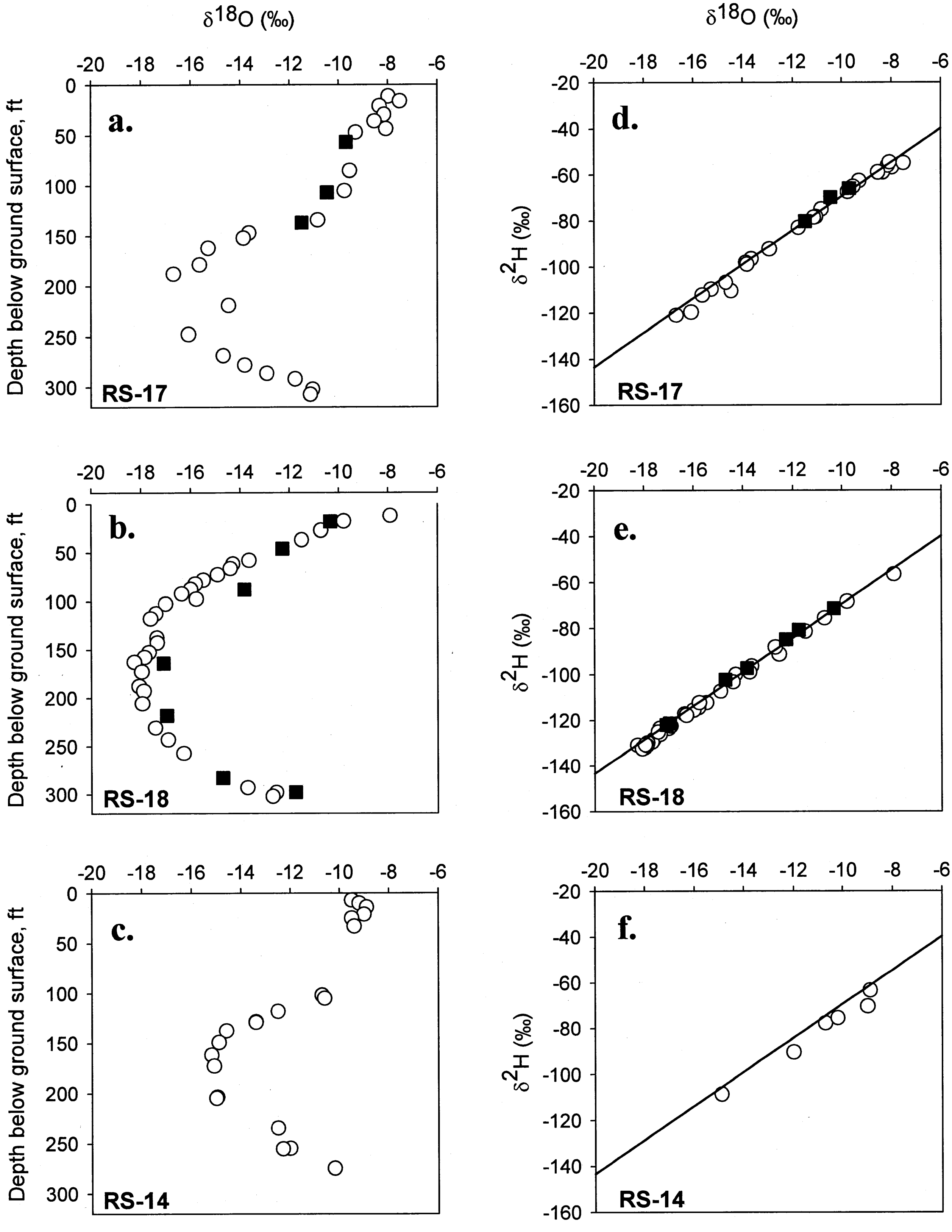

Figure 8. Oxygen isotope values as a function of depth below ground surface (a-c) and as a function of deuterium (d-f) for boreholes RS-17, RS-18, and RS-14. Sample analyses of pore water and well ports are shown as open circles and black squares, respectively. The black line in d-f represents the meteoric water line. The borehole locations are shown on Figure 3.

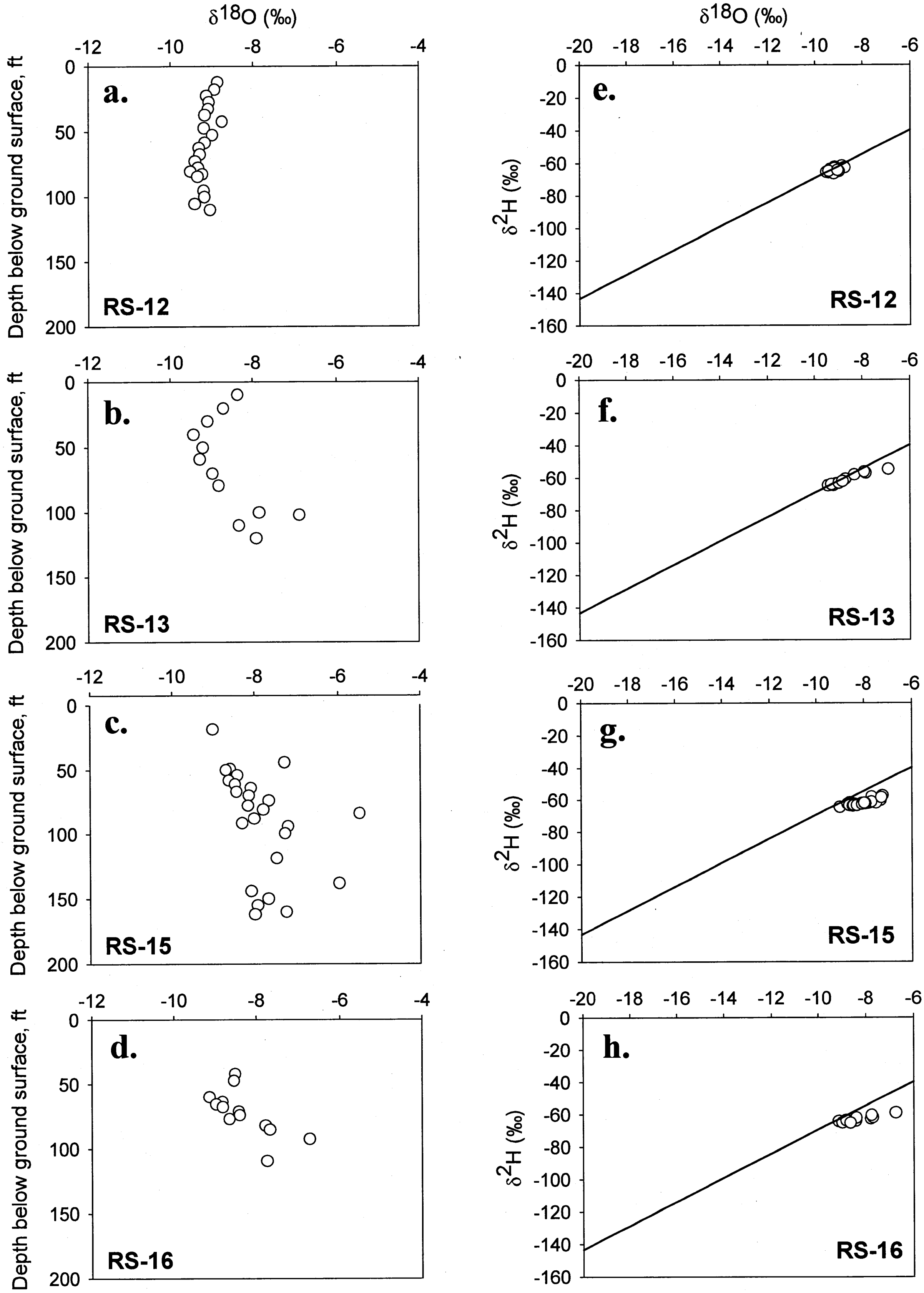

Figure 9. Oxygen isotope values as a function of depth below ground surface (a-d) and as a function of deuterium (e-h) from boreholes RS-12, RS-13, RS-15 and RS-16. The black line in e-h represents the meteoric water line. The borehole locations are shown on Figure 3.

Appendix D: Fecal Source Tracking primers and TaqMan probes designed for the specific amplification of HAdV, group I BAdV and group II BAdV. Anchor primers designed to amplify all HAdV and/or BAdV (for use when targeting source samples). 'Fluor" indicates the position of a user-defined fluorescent molecule; 'BHQ,' black hole quencher. I, inosine. An asterix has been placed next to the TaqMan probes that were labeled with HEX (and, thus, did not generate an amplification signal). These probes must re-made and reevaluated.

BAdV group I FST primers: Group I BAdV share the same RVS primer, but required different FWD primers based on greater similarity between particular BAdV with other livestock AdV than between all the group I BAdV.
BAdV2/OAdV2-5 FWD 3' DNM primer: 5'-GATGCCTCCGAGTATCTCTC-3'
BAdV 1, 3 and PAdV 3&5 FWD 3' DNM primer: 5'-GATGC(C/G)TC(C/A)GAGTACCTGTC-3'
BAdV 1, 3 & 10; PAdV 3&5 FWD 3' DNM primer: GATGC(C/G/T)TCCGAGTAC(C/T)TGTC
BAdV RVS 3' DNM primer (BAdV 1, 2, 3 10): 5'- GTTGTA**I**GC(G/T/C)GTGCCGCT-3'

BAdV group I FST probe:
***BAdV group I FST antisense probe (BAdV 1, 2 and 3)**:
Fluor-5'-CG(A/G)ATGTCAAAGTA**I**GTGCT(G/C)GCCATGTCCA-3'-BHQ

HAdV FST primers: HAdV are divided into subgroups based on genetic similarity; these subgroups were aggregated when possible to facilitate the amplification of as many HAdV as possible with a single primer.
HAdV FST FWD primer (species A, B, C, D, E, F): 5'-ACGC(C/T)TCGGAGTA(T/C)CTGAG-3'
HAdV species A,D,E,F RVS 3' DNM primer: 5'-**I**GGCAC(G/A)AA**I**CGCAGCGT-3'
HAdV species C RVS 3' DNM primer: 5'-AGGGATGAACCGCAGCGT-3'
HAdV species B RVS 3' DNM primer: 5'-GGGCACGAAGCGCA(A/G)CAT-3'

HAdV FST probes:
HAdV sense FST probe: Fluor-5'-CTGGTGCA**I**TT(T/C)GCCCG(C/T)GCCAC-3'-BHQ
***HAdV 40/41 antisense FST probe:** Fluor-5'-TCGCTG(C/T)GACCTGTCTGTGGTTACATC-3'-BHQ

Anchor primers: Anchor primers are designed to amplify all HAdV or group I BAdV in source materials. These sets share the same FWD primer.
FWD Anchor primer: 5'-CA**I**TGG(T/G)C**I**TACATGCACATC-3'
RVS HAdV anchor primer: 5'-AC**I**GTGGG**I**TT(C/T)CT(G/A)AACTTGTT-3'
RVS BAdV group I anchor primer: 5'- AC**I**GT(C/T/G)GGGTTTCTAAA(C/T)TTGTT-3'

MAdV Primers
FWD: 5'-GGCCAACACTACCGACACTTG-3'
RVS: 5'-TTGTCCTGTGGCATTTGA-3'

MAdV Probes:
***MAdV Probe A**: Fluor-5'- CGCCAATGTGGCTCAGTATATGCCGG -3'–BHQ
***MAdV Probe B**: Fluor-5'-GGAAAGGGAAACATGGCTGCCATGG-3'–BHQ

BAdV group II Anchor primers:
Atadenovirus FWD Anchor: 5'-CACAT(T/C)GC(G/T)GGTAGAAATGC-3'
Atadenovirus RVS Anchor A: 5'-CCATGG(A/C/T)AC(G/A)CT**I**GAATCC-3'
Atadenovirus RVS Anchor B: 5'-GCTTGATTATAA(T/C)T**I**GC(T/A)GCCATTTG-3'

BAdV group II FST primers:
BAdV group II FST FWD: 5'-G(A/G)AATGCTAC(T/A)AATGATC-3'
BAdV group II FST RVS A: 5'-GCTTT**I**A(A/C)TCT(A/G)TTAAA(A/G)CTCC-3'
BAdV group II FST RVS B: 5'-CCATGG(C/T)AC**I**CT(A/T)GAATCC-3'

BAdV group II probes:
FST Probe BAdV 4, 5 & 8: 5'-TTTGC(A/T)GACTATTTGGGAGCTGTTAA(C/T)AAT-3'
FST Probe BAdV 6 & 7: 5'-TTTGC(A/T)GATTACTTAGG(A/T)GCAGT**I**AATAATCTT-3'

Deer AdV FWD primer:
FST deer AdV FWD: 5'GAAATGCCACCAATGATC-3 (OdAdV)

Deer AdV probe:
Deer AdV Probe: Fluor-5'-TTTGCTGATTTCCTTGGCGCTGTAAATAAT-3'