FINAL REPORT

ASSESSING THE EFFECT OF PLEISTOCENE GLACIATION ON THE SANDSTONE AQUIFER IN EASTERN WISCONSIN

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Introduction

Along the entire eastern portion of Wisconsin the primary municipal water supply aquifer is the confined Paleozoic sandstone aquifer. From a water supply/water planning perspective, it is important to know whether or not freshwater pumped from this aquifer is being replenished by modern recharge or whether it represents a finite resource emplaced during the Pleistocene. This project investigated the recharge history of the deep sandstone aquifer along the eastern edge of Wisconsin from Green Bay to Milwaukee. Knowledge of the recharge provenance is of paramount importance to developing strategies for long-term management. Pleistocene recharge would have been controlled by highly variable factors involving ice advance and retreat as well as permafrost formation and ice-induced pressure heads that likely drove recharge at rates that were much higher than at present.

Primary users of the information developed during this project will include water supply managers responsible for making long term water management decisions. In addition, the scientific community will be interested in the paleoclimatic information inherent to the noble gas temperature record and the companion δ^{18} O record. Information on subglacial meltwater recharge is also of critical importance to the understanding of glacial movement in general and the occurrence of periodic glacial surges especially in the context of global warming.

Approach

The current project builds upon the work of Klump, et al (2008) who provided clear evidence that deep groundwater in the southeast corner of the State was recharged during the Last Glacial Maximum (LGM). These authors sampled water from an east-west transect along the groundwater flow path in southeast Wisconsin near the terminus of the Green Bay Lobe of the Laurentide Ice Sheet (LIS). The current project collected groundwater along the flow path from similar transects located farther to the north near the center of the Green Bay Lobe and well removed from the LGM terminus. This yielded information on recharge conditions and timing as the ice advanced and retreated across the landscape.

Measurements of ionic composition (for basic geochemical modeling using PHREEQC (Parkhurst, 1995); δ^2 H and δ^{18} O (for ice melt signal and Pleistocene temperature record);

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 δ^{34} S_{sulfate} (for determination of sulfate source); δ^{13} C, ¹⁴C (age of water); noble gas content and fractionation patterns (recharge sources and an independent measure of the Pleistocene temperature) were made in the groundwater of the confined sandstone aquifer along two northeast-southwest transects between the latitudes of Green Bay, WI (44.5° N) and Cedarburg, WI (43.3° N). Except in the Green Bay area this aquifer is actively pumped and municipal water supply wells were used as sampling points. In the Green Bay area existing municipal wells are maintained for emergency use and sampling was performed during test pumping (performed on a quarterly basis). All wells were completed solely within the Paleozoic aquifer starting in the recharge area and continuing downgradient along the flow path. The north transect, near Green Bay, consisted of nine wells. The Klump, et al. (2008) transect was taken near Milwaukee (43.0° N) and will be referred to in this report as the south transect. The center transect, between Fond du Lac and Cedarburg, consisted of 8 wells. The aquifer within the center transect displays highly variable thicknesses because of significant topography on the underlying basement rocks and the flow pattern is not uniformly west to east therefore center transect data could not be reliably interpreted was not included in this study. Raw data collected from center transect wells are given in the appendices. For reference, the locations of center transect wells are included in Figure 1.

Note that it is only along the western edge of the Michigan Basin (Wisconsin/Illinois) that the deep sandstone aquifer is a) accessible, b) contains fresh groundwater with the requisite residence times needed for finding water recharged during the Pleistocene LGM and c) has a relationship between location of the recharge area and ice lobe movement that allows the collection and dating of glacial recharge from positions near to and well back from the maximum terminus of a single ice lobe.

Background

During the LGM (~11- 26 k.a.), the LIS affected regional to hemispheric atmospheric circulation, climate patterns, and the isotopic composition of continental precipitation, due to changes in albedo, surface topography, and temperature (eg. Clark, Alley, et al. (1999)). While the climatic effects of the LIS have long since vanished, an influence of the LIS still persists through its impact on regional groundwater flow systems throughout the United States (e.g.

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Boulton et al. (1995); Breemer et al. (2002); McIntosh and Walter (2006); Person et al. (2007)). There is compelling field evidence that injection of isotopically light, subglacial meltwater changed the salinity and geochemistry of groundwater in Wisconsin/Illinois (Siegel and Mandle (1984); Siegel (1990); Klump et al. (2008)).

Recent studies from different North American and European locations indicate that ice loading during the last glaciation strongly influenced groundwater flow patterns and origin (e.g. Filley (1984); Weaver and Bahr (1991a) (1991b); Young (1992); Clark et al. (2000); Edmunds (2001); Breemer et al. (2002); Person et al. (2003); Person et al. (2007); Hoaglund III et al. (2004); McIntosh and Walter (2006); Klump et al. (2008)), however the effects of the LIS may vary depending on the conductivity of the substrate (Carlson et al., 2007). Present-day water quality in southern Wisconsin and northern Illinois is evidently still affected by changes in the groundwater chemistry due to reversals of hydraulic gradients during glaciation and subglacial recharge (e.g. Siegel (1990), (1991); Siegel and Mandle (1984)). Breemer et al. (2002) demonstrated, using a 2-D groundwater model, that the LGM Lake Michigan Lobe probably reversed groundwater flow and increased groundwater velocity under Lake Michigan and in Illinois within ~1000 years of occupying its maximum position. Similarly drastic effects of the LIS on the underlying groundwater have been modeled at the continental scale (Lemieux et al. (2008a), Lemieux et al. (2008b)). Person et al. (2003) and Marksammer et al. (2007) showed that the LIS recharged aquifers along the New England Atlantic continental shelf explaining the disequilibrium between the freshwater/saltwater interface and modern sea level. In contrast, model simulations of the James Lobe of the LIS suggest that this lobe had little effect on groundwater flow in the western plains of North America because a poorly conductive Upper-Cretaceous shale layer occupies the upper layer of the bedrock and functioned as a regional aquitard (Carlson et al., 2007). These results were in close agreement with the isotopic spring data of Grasby and Chen (2005), which indicate reversal of groundwater flow in the aquifers only near surface exposures that were in direct contact with the James Lobe. Groundwater flow in glaciated regions underlain by more permeable bedrock like that of the Great Lakes region of North America was significantly influenced by the ice lobes whereas regions underlain by less permeable substrates such as the North American plains were not.

The geologic setting along the western coastline of Lake Michigan is ideal for studying the effects of Pleistocene glaciation on the underlying Paleozoic sandstone aquifer. The Paleozoic aquifer transitions from unconfined to confined conditions along a north-south line approximately 50 km inland from the coast of Lake Michigan (solid line in Figure 2). All stratigraphic units in the area dip eastward into the Michigan Basin and the Maquoketa Formation, a very effective aquitard ($K_z \sim 10^{-12}$ m/s), overlies the aquifer from this boundary eastward and serves as the confining unit. An extensive hydrostratigraphic framework developed for a hydrologic model of eastern Wisconsin near Lake Michigan (Feinstein et al., 2010) shows aquifer recharge occurring immediately west of the confined/unconfined boundary. Water moves slowly and continuously eastward in the confined portion of the aquifer since ice retreated out of Wisconsin ~11 k.a. (all ages calibrated, calendar years). This slow but continuous flow preserves a record of climatic conditions at the time of groundwater recharge.

In the region surrounding Green Bay, a veneer of Glacial Lake Oshkosh sediments overly the aquifer (red line in Figure 2) are draped over a variable bedrock surface and range in thickness from 150 m over bedrock valleys to as little as 5m in upland areas. These sediments are fine grained varve clays and display low vertical conductivities ($K_z \sim 10^{-9}$ m/s) using laboratory scale measurements (Moeller et al., 2007). Although field scale heterogeneities, especially in thinly covered upland areas will likely increase the vertical hydraulic conductivity, geochemical data support the notion that these sediments form an effective confining layer for the underlying sandstone aquifer. This has the effect of moving the recharge area to the western edge of the lake sediments.

Previous workers describe geochemical evolution from a Ca-HCO₃ character in upgradient areas to a Ca-SO₄ character downgradient, increasing salinity, dedolomitization reactions eastward along the flow path, ¹⁴C groundwater ages, noble gas and stable isotope data (Grundl and Cape (2006); Klump et al. (2008)) that are all in agreement with these modeling results. In the south transect distinct pulses of water are delineated that were recharged before, during and after the LGM (Klump et al., 2008). Similar trends have been reported in the sandstone aquifer in Illinois (e.g. Perry et al. (1982); Gilkeson et al. (1984)) and in shallower

dolomite aquifers (Ma et al. (2004); McIntosh and Walter (2006)) in other parts of the Michigan Basin.



Figure 1: Wisconsin bedrock map with the north and central transect areas delineated. (modified from Mudrey et al. (1982) and Krall (2010)). Dotted square is the south transect of Klump, et al (2008). Colors on main map represent the following stratigraphic units: Purple = Devonian shales. Grey = Silurian dolomite. Pink = Maquoketa Shale. Green = Sinnippee Group. Dark green = St. Peter Formation. Light blue = Prarie de Chien Group. Tan = Cambrian sandstones.



Figure 2: Ice age deposits of Wisconsin. Brown areas are moraines of the LGM (modified from Thwaites (1960)). The underlying Paleozoic sandstones become confined east of the thick black line. Areas inside the red line are confined by glacial Lake Oshkosh silts and clays (after Hooyer (2007)).

Methods

The Wisconsin Department of Natural Resources (WDNR) Drinking Water Database (http://www.dnr.state.wi.us/org/water/dwg/data.htm) and the associated Groundwater Retrieval Network (http://prodoasext.dnr.wi.gov/inter1/grn\$.startup) were used to identify potential sampling wells. Potential wells had to satisfy the following criteria: a) the well was in active status b) the well had historical record of major ion composition c) the well was open only in the deep sandstone aquifer d) the well owner would grant sampling access. Candidate wells were sampled for major ions, sulfur isotopes in the summer of 2009. From this data, wells were selected for further study.

Wells were sampled at the raw water tap that exits the well before any water treatment occurs. Wells were pumped for at least one hour before sampling. Field measurement of temperature pH, conductivity, and dissolved oxygen (DO) were taken by electrodes in a flow through chamber. Portable colorimetric test kits (Chemetrics, Inc.) were used at the outlet tubing to measure sulfide, ferrous iron and an additional DO measurement. Electrode based and colorimetric DO analyses always coincided within the accuracy limits of the colorimetric method.

Samples for discrete analyses were collected after exiting the flow-through chamber. Major anion, δ^{18} O and δ^{2} H samples were collected after filtering through a 0.2um filter. Major cation samples were treated with trace metal grade nitric acid (4N) immediately after filtration. Alkalinity was determined immediately by titration with 0.02N HCl to pH 4.5. Sulfur isotope samples were collected with no headspace in 4L collapsible bladders. pH was adjusted with HCl to ~3.0 and barium chloride added to quantitatively precipitate dissolved sulfate as BaSO₄. Carbon isotope samples were also collected with no headspace in 4L collapsible bladders to which barium chloride was added. pH was adjusted with NaOH to ~9.0 in order to quantitatively precipitate dissolved carbonate as BaCO₃. BaSO₄ and BaCO₃ precipitates were filtered, dried and sent to the University of Waterloo Environmental Isotope Laboratory for δ^{32} S analysis or the University of Arizona AMS Laboratory for ¹⁴C and δ^{13} C age dating analysis. δ^{18} O, δ^{2} H analyses

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were also performed at the University of Arizona AMS Laboratory. Noble gas samples were collected in 1m long copper tubes attached directly to the raw water sample tap via clear Tygon tubing. Much care was taken to avoid bubble entrainment and the ends of the copper tubes pinched shut to collect the sample. Noble gas analysis was performed at the Environmental Isotope Group at ETH University, Zurich Switzerland. Major ion analyses were done in the author's laboratory via ion chromatography (anions) or flame atomic adsorption spectroscopy (cations).

Results

North transect wells show increasing total dissolved solids (TDS) content along the flow path from northwest to southeast. The four upgradient (northeastern) wells (LT992, BF195, BF197, BF188) average 407(\pm 35) ppm TDS while the five downgradient (southwestern) wells (BF187, BF190, BF216, BF212, IE266) average 652(\pm 153) ppm TDS. In addition, the overall chemical character transitions from Ca-HCO₃ to Ca-SO₄ dominated waters along flowpath. Upgradient wells display water that is strongly bicarbonate in nature with nearly equal amounts of calcium and magnesium as expected in a dolomite-rich aquifer. Increasing amounts of sulfate and to a lesser extent chloride from the dissolution of gypsum and halite are evident in downgradient wells (Figure 3). An additional differentiation between upgradient wells and downgradient wells can be seen in the sulfate isotopic signature. Downgradient wells not only contain more sulfate, but the sulfate is isotopically heavy as is typical for sulfate that originates from marine gypsum. A well-known layer of pyrite and arsenopyrite exists immediately to the west of the study area which can serve as the source of isotopically light sulfate (Gotkowitz et al., 2004). All major ion data is tabulated in the Appendices.

A more telling attribute of geochemical evolution along flow path is the presence of dedolomitization - at least within aquifer systems such as this one that contain abundant carbonate minerals. The prevalence of relatively soluble calcite in the aquifer causes it to be at saturation with calcite, continuing dissolution of gypsum along the flow path drives a series of sequential reactions (dedolomitization) that lead to the net dissolution of dolomite and net precipitation of calcite. These reactions are:



Figure 3: Piper diagram of north transect wells. Red symbols are older waters that were recharged during the LGM, black symbols are younger waters that were recharged after the LGM (see text).

1) Gypsum dissolution adds Ca^{2+} to the water

2) Additional Ca²⁺ causes calcite oversaturation and the resultant precipitation

3) Calcite precipitation removes CO_3^{2-} from the water causing dolomite under saturation

and the resultant dolomite dissolution.

The overall reaction can be written as follows:

CaMg(CO₃)₂ [s] + CaSO₄ [s] + H₂O \rightarrow (dolomite) (gypsum) CaCO₃ [s] + Ca²⁺ + Mg²⁺ + SO₄²⁻ + HCO₃⁻ + OH⁻ (calcite) Wells in the north transect are all near saturation with calcite with calcite saturation indices averaging -0.19 ± 0.16 . Saturation indices within a few tenths of zero are considered to be at saturation. Saturation index data is tabulated in the appendices. Waters in which dedolomitization occurs plot along the charge balance derived from the overall reaction given above (Ca²⁺ + Mg²⁺ versus SO₄²⁻ + 1/2 HCO₃⁻; assume OH⁻ is very small). Strict dedolomitization requires that all sulfate is derived from sulfate minerals, primarily gypsum, however it has been demonstrated that within Green Bay and the surrounding environs, the oxidation of sulfide minerals, chiefly pyrite, are another source of dissolved sulfate (Gotkowitz et al., 2004). The δ^{34} S isotopic signature of pyrite-derived sulfate is isotopically light, displaying values that range from +10‰ to -15‰ (CDT) whereas sulfate derived from the dissolution of marine gypsum displays values that exceed +20‰ and can reach as high as +30 ‰ in early Paleozoic evaporates. (Clark and Fritz, 1974). The δ^{34} S isotopic content of measured sulfate was used to adjust for the portion that arises from gypsum dissolution by setting δ^{34} S_{pyrite} to the lowest measured value (4‰) and δ^{34} S_{gypsum} to the highest measured value (34‰).

Dedolomitization results are shown in Table 1 and Figure 4. The three farthest downgradient wells (BF212, BF216, IE266) fall close to the charge balance line and display cation and anion sums of 3.0 or greater indicating more extensive dedolomitization than in the remaining wells (cation and anion sums between 2.0 and 3.0)

Well #	Ca ²⁺ (mMol/ L)	Mg ²⁺ (mMol/L)	HCO ₃ ⁻ (mMol/L)	SO ₄ ²⁻ (total) (mMol/L)	δ ³⁴ S (‰)	SO ₄ ²⁻ (gypsum) (mMol/L)	SO ₄ ²⁻ (gypsum) (%)	$\begin{array}{c} Ca^{2+} + \\ Mg^{2+} \end{array}$	$1/2HCO_{3}^{-}$ + SO ₄ ²⁻
BF187	1.55	0.84	3.49	0.61	16.9	0.26	43	2.39	2.36
BF188	1.27	0.90	4.32	0.15	11.9	0.04	26	2.17	2.31
BF190	1.49	0.79	3.11	1.13	29.0	0.94	83	2.27	2.68
BF195	1.18	0.78	3.74	0.07	9.3	0.01	18	1.96	1.94
BF197	1.22	0.82	4.28	0.12	12.2	0.03	27	2.04	2.26
BF212	2.11	1.00	3.25	1.97	29.4	1.66	85	3.11	3.60
BF216	2.02	1.02	3.26	2.37	33.9	2.37	100	3.05	4.00
IE266*	2.94	1.10	2.87	2.38	29.2	2.00	84	4.04	3.81
LT992	1.33	1.19	5.07	0.02	5.6	0.00	5	2.52	2.55

Table 1: Dedolomitization analyses for north transect wells. IE266 is an anomalous well – see text.



Figure 4: Dedolomitization charge balance plot in north transect wells. Sulfate concentrations are adjusted to reflect the portion that is derived from gypsum dissolution. Red triangles are wells recharged during the LGM, black diamonds were recharged post LGM. Open red triangle is IE266.

Two wells exhibit anomalous behavior, BF191 and IE266. BF191 in particular is anomalous in that it shows the lowest measured TDS of any well, a strong Ca-HCO₃ character and little dedolomitization in spite of its location near the middle of the flow path. In addition its ¹⁴C age is negative, its isotopic signature is extremely heavy (resulting in an impossibly warm δ^{18} O temperature) and the noble gas data could not be rectified by the fitting program. All of these data are indicative contamination with another water source likely through a compromised well casing or localized faulting. Raw data for BF191 is listed in the appendices and it is located on the base map for reference (Figure 1) however it has been excluded from any further interpretation (data plots or contour maps). IE266 is anomalous only in its isotopic composition (and the resulting δ^{18} O temperature) and ¹⁴C age with respect to its position in the flow path; major ion data agrees with other wells in the transect. The noble gas sample was lost for this well. This is also likely to be caused by a compromised well casing or localized faulting that allows mixing of the aquifer with younger, warmer water within the well bore. IE266 data is included on all data plots and contour maps. The increasing age of these samples with distance downgradient was confirmed by 14 C age dating (Table 2). The 14 C ages and errors are averages and ±1 standard deviations obtained by using three different correction models for calculating the initial activity A⁰ of radiocarbon (Fontes and Garnier (1979); Pearson and White (1967); Tamers (1967)). Figure 8 contains a map of the same data. The 14 C data confirm that the eastern portion of this transect contains groundwater that was recharged during the time when the ice sheets of the Michigan and Green Bay Lobes last covered the area (Mickelson and Colgan, 2004). A clear demarcation between water older and younger than ~10 k.a. (LGM and post-LGM recharge respectively) is seen in the center of the transect roughly parallel to the modern day Fox River. Two analogous demarcation lines were observed in the south transect that separate pre-LGM from LGM wells and LGM from post-LGM wells (Klump et al., 2008). There are no municipal wells far enough eastward along the flowpath in the north transect to encounter pre-LGM water.

Well	$\delta^2 H$	δ ¹⁸ Ο	¹⁴ C	$\delta^{13}C$	\mathbf{A}^{0}	¹⁴ C age	$\delta^{18}O$ temperature
	(‰)	(‰)	(pmC)	(‰)	(pmC)	(yr)	(°C)
BF187	-110.30	-15.40	12.76(±0.15)	-10.9	47(±6)	10752(±1048)	-2.6
BF188	-88.96	-12.72	21.36(±0.20)	-12.6	54(±6)	7601(±887)	1.3
BF190	-125.65	-17.47	8.73(±0.15)	-11.1	48(±6)	14055(±1075)	-5.6
BF195	-84.70	-11.71	29.34(±0.20)	-11.6	50(±7)	4389(±1053)	2.7
BF197	-90.99	-12.75	18.31(±0.17)	-12.5	51(±2)	8528(±394)	1.2
BF212	-120.68	-16.85	4.43(±0.12)	-10.4	46(±8)	9326(±1430)	-4.7
BF216	-127.66	-17.31	4.52(±0.12)	-11.5	49(±5)	19684(±896)	-5.3
IE266*	-74.36	-10.35	28.25(±0.21)	-8.1	42(±16)	2801(±2964)	4.7
LT992	-78.10	-11.02	32.34(±.022)	-12.8	53(±3)	4051(±443)	3.7

Table 2: Stable isotope and radiocarbon analyses for north transect wells. ¹⁴C ages and errors are average values and ± 1 standard deviation as determined by making use of three different correction models for calculating the initial activity (A⁰) of radiocarbon. IE266 is an anomalous well – see text.

Stable isotope data also support the observation that Pleistocene age water is present in this aquifer. All wells exhibit isotopic values that lie along the global meteoric water line with no indication of isotope exchange with aquifer solids (Figure 5). Meteoric water is isotopically lighter if it evaporated in a cold climate and this systematic relationship between isotopic content

and temperature can be used to estimate the average annual temperature at the time of recharge from the relationship: $\delta^{18}O(\%) = 0.695*T(^{\circ}C)-13.6$ (Dansgaard, 1964). A plot of $\delta^{18}O$ temperature versus ¹⁴C age is shown in Figure 6. Also shown is the equivalent data for the south. Data from south transect wells transect (from Klump et al. (2008)) is tabulated in the Appendices. A minimum age is given for the oldest well in the south transect because this well has very little ¹⁴C activity (0.07 ± 0.2 pmC) and is at the edge of reliable ¹⁴C dating. Both temperature trends are consistent with modern day average annual temperatures of 6.4 °C in Green Bay and 8.5 °C in Milwaukee (data available at http://www.aos.wisc.edu/%7Esco/climhistory/index.html).

Obvious temperature minimums occur in both transects that broadly coincide with available literature dates for LIS advance and retreat. LIS chronology is not well constrained, largely due to a paucity of available radiocarbon dates (Attig et al., 2011). It is generally presumed that ice moved into the Wisconsin area about 26,000 years ago (Mickelson and Colgan (2004), Winguth et al. (2004)) and completely retreated out of northern United States by 9,800 years ago (Mickelson and Colgan, 2004). Estimates for the timing of the initial ice retreat of the Green Bay Lobe from its maximum position center around 22,000 years ago (Hooyer (2007); Winguth et al. (2004)) although more recent estimates have ice remaining at the maximum position for another 3500 years; until 18,500 years ago (Attig et al., 2011). The ice margin near the latitude of Fond du Lac (midway between the north and south transects) has been dated as early as 19,000 years ago although small re-advances occurred at 16,500 and 13,700 years ago (Hooyer (2007), Hooyer et al. (2009)). The Two Rivers terminal moraine is close to the north transect and has been dated at 13,700 years ago (Winguth et al., 2004). The south transect lies at a latitude midway between the Johnstown and Green Lake moraines which have been dated at 22,200 and 17,000 years respectively (Winguth et al., 2004). In Figure 6, ice presence is depicted as ~19,000-26,000 years ago in the south transect and ~13,000-26000 years ago at the north transect although the timing of retreat could be as much as 3500 years later.



Figure 5: Stable isotope data for north transect wells. Solid line is the global meteoric water line: $\delta^2 H = 8 \times (\delta^{18}O) + 10$. Red triangles are wells recharged during the LGM, black diamonds were recharged post LGM. Open red triangle is IE266.



Figure 6: Oxygen isotope temperature versus corrected ¹⁴C ages for north and south transects. South transect values taken from Klump et al. (2008). A minimum age is given for the oldest well in the south transect because this well has very little ¹⁴C activity ($0.07 \pm 0.2 \text{ pmC}$) and is at the edge of reliable ¹⁴C dating. Error bars are ±1 standard deviation as determined by making use of three different correction models for calculating the initial activity (A^0) of radiocarbon. Open red diamond is IE266.

Analysis of noble gas temperatures and the amount and fractionation of excess air are also consistent with a glacial origin for groundwater in the north transect. Both environmental parameters were calculated using the closed-system equilibrium (CE) model (Aeschbach-Hertig et al., 2000) which assumes that residual bubbles of air are trapped in groundwater as the water table rises during a recharge event. The air in these bubbles reaches a closed-system equilibrium with the surrounding water that is a direct function of the amount of air dissolved and hydrostatic head during equilibration and is an inverse function of soil temperature. Under these conditions the heavy noble gases preferentially dissolve and the aqueous noble gas content differs from atmospheric content. This fractionation of the different noble gases upon dissolution is described by the fractionated excess air, represented by F = 0. This fractionation results in a Ne \rightarrow Xe enrichment pattern in which successively heavier gases are preferentially dissolved. The pressure parameter q is the ratio of dry gas pressure in the entrapped gas to that in the free atmosphere and thereby is a measure of the hydrostatic pressure exerted on the entrapped air. The reader is referred to Kipfer, et al., (2002) for a detailed description of the process of excess air generation.

Groundwater almost universally contains dissolved air in excess of atmospheric equilibrium because of the extra pressure exerted on bubbles trapped during recharge. Normally recharged groundwater contains excess air, as defined by Ne in excess of atmospheric equilibrium (Δ Ne), of 10-50% (Kipfer et al., 2002). Pure glacial meltwater is known to contain as much as several hundred percent Δ Ne (Vaikmae et al., 2001). Excess air in glacial meltwater has very minimal fractionation because it originates as air trapped in firn – a process that does not significantly fractionate air (Severinghaus and Battle (2006); Huber et al. (2006)).

Noble gas data (Table 3) are unusual in several respects. The amount of excess air (ΔNe) ranges from ~60 to ~107% which is higher than is seen in normal groundwaters, but not as high as expected from pure melted glacial ice. Additionally, the pressure factor (q) is very high for normally recharged groundwater. Each 0.1 increment in q above 1.0 is equivalent to a recharge head of 1 meter (Aeschbach-Hertig et al., 1999) implying recharge heads of as much as 8 meters (Figure 7). Lastly, noble gas temperatures all lie between 1.6 and 3.2 °C. There is no agreement between $\delta^{18}O$ and noble gas temperatures; in particular there is no trend of decreasing.



Figure 7: Plot of excess air (Δ Ne) versus pressure factor (q) as calculated from the CE model. South transect values taken from (Klump et al., 2008). Black diamonds are post-LGM recharge wells; red diamonds are LGM recharged wells. Typical Δ Ne values seen in normally recharged groundwater lie between 10% and 50%. Δ Ne uncertainties are determined from the covariance matrix in the least squares fitting algorithm (Aeschbach-Hertig et al., 1999).

well #	noble gas temperature (°C)	Δ Ne (%)	q (CE model)	F (CE model)	δ ¹⁸ O temperature (°C)	¹⁴ C age (yr)
BF187	3.1(±0.6)	73.8(±1.4)	1.53	0.47	-2.6	10752(±1048)
BF188	3.2(±0.4)	67.8(±1.4)	1.44	0.43	1.3	7601(±887)
BF190			gas sample lost		-5.6	14055(±1075)
BF195	1.6(±0.3)	59.6(±1.4)	1.29	0.12	2.7	4389(±1053)
BF197	2.8(±0.4)	78.8(±1.4)	1.46	0.29	1.2	8528(±394)
BF212	3.2(±0.6)	106.8(±1.4)	1.78	0.37	-4.7	9326(±1430)
BF216	3.1(±0.8)	94.3(±1.4)	1.72	0.44	-5.3	19684(±896)
IE266*			gas sample lost		4.7	2801(±2964)
LT992	4.1(±0.4)	62.8(±1.4)	1.37	0.35	3.7	4051(±443)

Table 3: Noble gas data for north transect. ΔNe and noble gas temperature uncertainties are determined from the covariance matrix in the least squares fitting algorithm (Aeschbach-Hertig et al., 1999).

temperature in older, downgradient wells in the noble gas data. This is in contrast to the south transect in which δ^{18} O and noble gas temperatures largely agreed with each other (Klump et al., 2008).

Discussion

The confined nature of the deep sandstone and the resulting long flow paths are supported by the observed geochemical trends including increasing TDS, the transition from Ca-HCO₃ to Ca-SO₄ dominated water, and a well-developed dedolomitization pattern and increasing ¹⁴C age. This is consistent with eastward flow in a confined aquifer that is recharged to the west of the transect. Both the thick sequence of glacial lake clays and the Maquoketa Shale serve as confining units (cf. Figure 1).

 δ^{18} O data, in conjunction with ¹⁴C ages, display a cooling paleotemperature record that coincides with the LGM (Figure 6). Water recharged during the LGM displays δ^{18} O derived temperatures that indicate average annual temperatures as low as -5.3 °C; a temperature that is 12 °C cooler than the modern average annual temperature of 6.4 °C. This is much larger than previous estimates of -5 to -9 °C of cooling in North American paleogroundwaters (Clark et al. (1997); Stute et al. (1995); Aeschbach-Hertig et al. (2002); Ma et al. (2004)), and Europe (Andrews and Lee (1979); Stute and Deak (1989); Beyerle et al. (1998)). There could be no meteoric recharge in a climate this cold because of the presence of extensive permafrost and the lack of liquid water. This water must contain a large proportion of glacial ice that precipitated in a very cold climate and maintains its isotopic signature during later recharge. As such it does not reflect the climatic conditions prevailing during the time of recharge but it does indicate that water was injected into the aquifer during the LGM and that a large proportion of it originated as glacial meltwater. The demarcation between LGM and post-LGM recharged wells occurs at an age less than 10 k.a. and a δ^{18} O temperature above 0° C (Figure 8). The demarcation between LGM and post-LGM wells can also be seen in the major ion signature (Figure 3), and the extent of dedolomitization (Figure 4).

Glacially recharged water is a function of the hydrology of drainage above (supraglacial), within (englacial) and beneath (subglacial) glaciers. This is highly complex system that is a

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function of temporally variable supraglacial water supply, the direction and intensity of internal stress within the glacier and the existence of high conductivity zones (fractures and debris filled crevasse traces) (Gulley et al., 2009). The vast majority of supraglacial water that drains to the



Figure 8: Maps of north transect wells showing ¹⁴C ages and δ^{18} O temperatures. Bedrock units are shown using the color scheme of Figure 2. Dotted orange line (on ¹⁴C age map) denotes the demarcation between post-glacially recharged water (1) and glacially recharged water (2). Well IE266 is not included in the contouring.

base of the glacier is transported by a system of large (meter scale diameter), irregularly spaced moulins that preferentially occur in fast moving portions of the glacier where ice fracturing is prevalent (Gulley et al., 2009). Within alpine glaciers, water levels in these moulins can vary by hundreds of meters in as little as a day (Badino and Piccini (2002); Vieli et al. (2004)). A network of small interconnected fractures is also present that operates as a spatially extensive drainage system that transmits a much smaller volume of water. These two drainage systems are hydraulically connected with the rapidly fluctuating, high volume moulin system inducing transient pressure gradients in the less transmissive fracture network. Beneath continental

glaciers, direct basal melting is also a component of subglacial water. The relative proportion of basal met to surface drainage is specific to each ice sheet and is a function of the overall temperature regime, ice thickness and flow dynamics of the glacier (Bell, 2008). Although it is clear that two interconnected englacial drainage systems plus basal melting all contribute to subglacial water, the spatial and temporal variability of this complicated hydrologic system is poorly understood. Effects on the underlying aquifers are even less understood because most subglacial drainage studies focus solely on the effect that subglacial water has on the speed of ice movement.

The north transect is ~100 miles north of the maximum glacial extent but below the equilibrium line (where melting/ablation exceeds snowfall). As such, supraglacial lakes and streams containing a mix of meteoric water and glacial meltwater would have formed. The entire englacial drainage system began filling up early in the melt season when the moulins were still closed at the base by ice creep during the preceding winter and supraglacial influx began increasing. This was a very dynamic system with partially filled, frozen off sections of moulins and small fractures both of which contained water and pressurized air that acted like the pressurized bubbles of the closed system equilibrium model. As the melt season continued, the moulin-based drainage system re-opened and the englacial system was efficiently drained via surface water outlets at the terminus. As winter approached, supraglacial flux slowed, the moulins re-closed and the water table elevation became a function of slow drainage through relatively low conductivity subglacial sediments.

The contrast between the noble gas and isotopic data collected from the north transect and that reported for the south transect (Klump et al., 2008) contains information that indicates the physical mechanisms responsible for subglacial recharge at different positions within the glacier. In the north transect, the observed normal fractionation patterns (Ne \rightarrow Xe enrichment), pressure factors (q) indicating recharge heads of as much as 8 meters in the LGM recharged wells, Δ Ne values in excess of normal recharged groundwater (Figure 7) and noble gas temperatures that are all above 0 °C with no spatial trend (Table 3), are all consistent with recharge that was dominated by supraglacial melt and b) occurred at a water table that was in effect within the ice sheet. Recharge within the dynamic englacial system allows for the observed high pressure factors and the associated large Δ Ne values. Because the physics of pushing excess air into the water are the same in pressurized moulins as in the pore spaces of partially saturated sediment, the Ne \rightarrow Xe enrichment pattern is identical to normally recharged groundwater. A significant amount of direct basal melt recharge is not possible because direct recharge of melted ice would lead to a very small Ne \rightarrow Xe enrichment pattern. Noble gas temperatures reflect the air temperature within the moulins and this was just above freezing. δ^{18} O temperatures reflect the isotopic signal of the ice itself which does not get re-set when it melts. Note that in this case neither the δ 18O nor the noble gas temperatures reflect local climate. The conclusion is that recharge in the north transect is mostly supraglacial melt with lesser amounts of local meteoric water. This is a close analog to the modern day Greenland ice sheet in which surface melt drains completely to the subglacial hydrologic system. In contrast, modern day Antarctic ice sheets display little connectivity between surface meltwater and the base of the glacier but instead rely on basal melting to supply subglacial water (Bell, 2008).

In the south transect aquifer recharge occurred under conditions where the annual pressure change was approximately 0.5 to 4 meters and Δ Ne values ranging from 12%-72% both of which are within the typical range for meteoric waters (Figure 7). The south transect was at the very edge of the ice sheet and the recharge was dominated by meteoric water. Recharge occurred through a water table that was not subject to the dynamic conditions of englacial hydrology but was at the base of the ice or just below within the sediment. Any supraglacial lakes/streams had a large component of local meteoric water and moulins either did not freeze shut (it was too warm at this latitude) or the ice decayed so rapidly that active moulins did not form at all. Since most of the recharge in the south transect was meteoric water (less was supraglacial melt), the δ^{18} O temperatures and the noble gas temperatures match because they were set at the same time/place. In this case, both temperature records reflect the local climate at the time of recharge. The Ne \rightarrow Xe enrichment pattern remains normal therefore contributions from direct basal melt were small. Figure 9 is a diagrammatic representation of recharge mechanisms occurring along the axis of the Green Bay lobe.



Figure 9: Schematic of glacial recharge mechanisms as seen in the Green Bay lobe. In the south transect (Milwaukee) englacial hydrology is minimal and the water table resides at or just beneath the base of the ice. Recharge is dominated by meteoric water (recharge heads, pressure factors and fractionation are normal, δ^{18} O and noble gas temperatures reflect local climate). In the north transect (Green Bay) the water table resides within the glacier and englacial dynamics prevail. Recharge is dominated by supraglacial melt (recharge heads and pressure factors are high, fractionation is normal, δ^{18} O temperatures reflect moulin air temperature at time of recharge, noble gas temperatures reflect isotopic signature of the ice itself). See text for full explanation.

Conclusion

This study, in conjunction with earlier work (Klump et al., 2008), provides a better understanding of the paleohydrology of aquifers in the upper Midwest and by extension, in other glaciated parts of the world. It demonstrates the presence of glacially recharged water in the deep sandstone aquifer along the entire eastern coastline of Wisconsin. It is clear that Laurentide Ice Sheet did not seal off the aquifer but instead injected large volumes of glacial meltwater into the aquifer. This information is of use to water supply managers responsible for making long term water management decisions throughout eastern Wisconsin. For instance, the results of this study indicate that the majority of the water pumped from both the Green Bay and the Milwaukee areas has been Pleistocene in age and is not being replenished on a time scale relevant to human activities. This research also indicates that the LIS was analogous to the modern day Greenland ice sheet with subglacial water being primarily derived from supraglacial melt instead of basal melting. To the author's knowledge, this study is the first to use noble gas and isotopic data to deduce the provenance of water being delivered to the base of glaciers. Data of this sort can aid in determining potential forcings behind the rapid margin fluctuations that characterized the deglaciation of the southern LIS. In addition to addressing important water resource issues, our research on the ice-water system may lead to new insights regarding ice-stream and fast-ice behavior, of great significance not only for understanding the southern lobes of the LIS, but also for predictions of the response of the remaining ice sheets (Greenland and Antarctic), with attendant effects on sea level, to present and future climate change.

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APPENDIX A

North and Central transects: Physical well data, major ion and isotopic composition.

	General Information													
Well #	High Capacity #	Location (County, City)	Depth (ft)	Closed to (ft)	Contact	Elevation (masl)	static water level (mbgs)							
BF187	75557	Brown, DePere	863	265	Dan Carpenter	195	43							
BF188	75558	Brown, DePere	787	500	Dan Carpenter	191								
BF190	75560	Brown, Preble	950	311.5	Tom Landwehr	179	36							
BF191	75561	Brown, Preble	973	341	Tom Landwehr	182	27							
BF195	75565	Brown, Green Bay	777	235	Tom Landwehr	197								
BF197	75567	Brown, Green Bay	815	269.5	Tom Landwehr	178	46							
BF212	75582	Brown, Bellevue	855	547	Bill or Glen	180	52							
BF216	75586	Brown, Scott	1235	646	Mel Deprey	232	89							
IE266	1717	Brown, Bellevue	1130	605	Bill or Glen	230	96							
LT992	2895	Brown, Hobart	785	487	Rick Kinney	207	29							

North Transect (Brown County)

Center Transect (Fond du Lac, Washington & Ozaukee Counties)

	General Information													
Well #	High Capacity #	Location (County, City)	Depth (ft)	Closed to (ft)	Contact	Elevation (masl)	static water level (mbgs)							
AY379	1064	FdL, Byron	1150	373	Kathy Scharf	258	52							
BF794	78463	FdL, Campbellsport	1200	585	Mark Grueber	303	61							
BF800	78469	FdL, Fond du Lac	835	579	Kathy Scharf	229	18							
BG643	83701	Ozaukee, Cedarburg	1210	718.7	Tim Martin	242	33							
BH246	87902	Washington, Germantown	1282	531	Jim Driver	267	57							
BH275	87931	Washington, Allenton	735	440	Tom Gurecki	301	52							
KR925	1745	FdL, Campbellsport	1245	649	Mark Grueber	323	97							
MK423	2313	FdL, Waupun	866	326	Steve Schramm	271	16							

APPENDIX A continued

North Transect (Brown County)

						Chemica	l and Isoto	pe Data (n	ng/L unles	s otherwis	e denoted)											
Well #	Temp (°C)	Spec. Cond. (mmhos/cm)	pH (sU)	DO	Ca ²⁺	Mg ²⁺	Na ⁺	K^+	Fe ²⁺	HCO ₃	Cl	SO4 ²⁻	NO ₃	HS	δ ³⁴ S (‰)	δ ¹⁸ O (‰)	δ ² H (‰)	Alkalinit y (mEa/L)	measured TDS	calcite SI	Dolomite SI	Gypsum SI
BF187	11.8	0.689	7.48	1.6	62.1	20.4	22.3	4.5	0.25	213.2	27.0	136.2	0	0	16.89	-15.40	-110.30	3.5	427	-0.02	-0.97	-1.53
BF188	11.2	0.551	7.02	1.3	50.9	22.0	14.0	4.2	0.1	263.4	13.9	55.8	0	0	11.87	-12.72	-88.96	4.32	305	-0.45	-1.12	-1.96
BF190	11.9	0.552	7.35	1.1	59.5	19.1	43.4	5.2	0.4	189.9	54.5	130.0	0	0	28.97	-17.47	-125.65	3.11	421	-0.22	-0.77	-1.57
BF191	9.5	0.314	7.03	12.7	34.3	19.5	7.1	1.3	0.0	127.2	13.4	24.1	1.2	0	4.21	-5.85	-42.24	2.1	163	-0.9	-1.94	-2.41
BF195	10.7	0.252	7.19	2.0	47.3	18.9	12.3	3.0	0.0	227.9	12.2	37.8	0.34	0	9.35	-11.71	-84.70	3.8	239	-0.36	-0.99	-2.13
BF197	11.2	0.513	7.48	2.2	48.9	19.9	16.3	4.5	0.1	261.4	10.0	41.5	0	0	12.19	-12.75	-90.99	4.28	264	0	-0.25	-2.09
BF212	11.6	1.040	7.31	0.7	84.4	24.4	71.9	6.3	0.3	198.5	103.7	223.7	0	0	29.36	-16.85	-120.68	3.27	639	-0.14	-0.67	-1.26
BF216	14.3	1.036	7.34	0.7	81.1	24.9	78.2	7.7	0.3	199.1	99.8	228.5	0	0	33.94	-17.31	-127.66	3.26	647	-0.09	-0.49	-1.28
IE266	12.9	1.250	7.04	1.7	117.8	26.7	73.2	7.9	0.3	174.9	166.6	271.8	0	0	29.22	-10.35	-74.36	2.9	786	-0.33	-1.12	-1.08
LT992	10.6	0.543	7.31	0.9	53.4	28.9	6.6	3.4	0.4	309.6	4.9	33.0	0	0	5.57	-11.02	-78.10	5.08	300	-0.08	-0.29	-2.17

Center Transect (Fond du Lac, Washington & Ozaukee Counties)

						Chemica	l and Isoto	pe Data (n	ng/L unles	s otherwise	e denoted))										
Well #	Temp (°C)	Spec. Cond. (mmhos/cm)	pH (sU)	DO	Ca ²⁺	Mg ²⁺	Na ⁺	K^+	Fe ²⁺	HCO ₃	Cl	SO4 ²⁻	NO ₃	HS	δ ³⁴ S (‰)	δ ¹⁸ O (‰)	δ ² H (‰)	Alkalinit y (mEq/L)	measured TDS	calcite SI	Dolomite SI	Gypsum SI
BF794	10.0	0.834	7.19	4.5	90.9	26.5	10.8	4.2	0.6	372.5	10.7	121.9	0	0	15.98	-9.25	-62.80	6.11	474	0.05	-0.31	-1.46
BF800	11.2	1.139		3.5	92.3	35.3	40.3	4.4	0.0	247.0	102.8	192.1	0.34	0	19.57	-12.88	-89.98	4.05	671	0.04	-0.19	-1.30
BG643	10.4	0.758	7.06	1.4	72.0	34.8	12.0	1.4	0.0	358.1	29.6	44.6	2.3	0	1.60	-8.98	-61.51	5.9	411	-0.16	-0.52	-1.96
BH246	11.9	0.737	7.23	0.5	68.7	26.0	4.3	1.3	0.4	387.2	2.6	69.4	0	0	20.75	-9.12	-59.66	6.3	404	0.04	-0.19	-1.78
BH275	11.1	0.762	6.98	1.0	72.6	24.6	6.9	2.2	0.3	350.5	3.8	104.3	0	0	19.43	-9.22	-60.70	5.75	448	-0.25	-0.83	-1.59
KR925	13.0	0.948	7.2	0.8	101.0	30.8	23.2	5.9	0.9	299.8	34.3	216.2	0	0	19.41	-11.19	-79.21	4.94	587	0.02	-0.29	-1.21
MK423	10.0	0.661	7.18	1.3	69.0	31.3	3.8	1.5	1.5	381.9	4.9	21.0	0	0	n/a	-9.88	-66.01	6.26	365	-0.03	-0.28	-2.29

APPENDIX B

North and central transects: radiocarbon and noble gas data. ΔNe and noble gas temperature uncertainties are determined from the covariance matrix in the least squares fitting algorithm. Uncertainties in corrected ¹⁴C age are ±1 standard deviation as determined by making use of three different correction models for calculating the initial activity (A⁰) of radiocarbon. Uncertainties in percent modern ¹⁴C are analytical error as reported by the University of Arizona AMS analytical laboratory.

North Transect

well #	¹⁴ C (pcm)	¹³ C (‰ PDB)	uncorrected ¹⁴ C age (yr)	cor	rected ¹⁴ C (yr)	age	avg. corrected ¹⁴ C age (yr)	noble gas temperature (°C)	Ne (equ) (ccSTP/g)	Ne (m) (ccSTP/g)	Δ Ne (%)	q (CE model)	F (CE model)
				Tamers	Pearson	F. & G.							
BF187	12.8 (±0.2)	-10.9	16539	11962	10158	10137	10752 (±1048)	3.1 (±0.6)	2.10E-07	3.66E-07	74 (±1.4)	1.53	0.47
BF188	21.4 (±0.2)	-12.6	12400	8626	7097	7082	7601 (±887)	3.2 (±0.4)	2.10E-07	3.52E-07	68 (±1.4)	1.44	0.43
BF190	8.7 (±0.2)	-11.1	19588	15296	13445	13424	14055 (±1075)			gas samp	le lost		
BF191	105.0 (±0.4)	-4.6	-391	-4521	-14396	-14495	-11137 (±5730)		2.13302E-07	1.71E-07			
BF195	29.3 (±0.2)	-11.6	9850	5605	3789	3773	4389 (±1053)	1.6 (±0.3)	2.1395E-07	3.42E-07	60 (±1.4)	1.29	0.12
BF197	18.3 (±0.2)	-12.5	13638	8984	8304	8298	8529 (±394)	2.8 (±0.4)	2.1116E-07	3.78E-07	79 (±1.4)	1.46	0.29
BF212	4.4 (±0.1)	-10.4	25037	20978	18515	18487	19327 (±1430)	3.2 (±0.6)	2.10E-07	4.34E-07	107 (±1.4)	1.78	0.37
BF216	4.5 (±0.1)	-11.5	24875	20719	19180	19154	19684 (±896)	3.1 (±0.8)	2.10E-07	4.09E-07	94 (±1.4)	1.72	0.44
IE266	28.3 (±0.2)	-8.1	10154	6223	1133	1046	2801 (±2964)			gas samp	le lost		
LT992	32.3 (±0.2)	-12.8	9068	4562	3798	3792	4051 (±443)	4.1 (±0.4)	2.08008E-07	3.39E-07	63 (±1.4)	1.37	0.35

Center Transect

well #	¹⁴ C (pcm)	¹³ C (‰ PDB)	uncorrected ¹⁴ C age (yr)	corrected ¹⁴ C age (yr)		avg. corrected ¹⁴ C age (yr)	noble gas temperature (°C)	Ne (equ) (ccSTP/g)	Ne (m) (ccSTP/g)	Δ Ne (%)	q (CE model)	F (CE model)	
				Tamers	Pearson	F. & G.							
AY379	7.6 (±0.1)	-9.9	20722	16253	13667	13636	14519 (±1502)	0.8(±0.3)	2.16145E-07	3.82E-07	77 (±1.4)	1.38	0.10
BF794	19.1 (±0.2)	-11.7	13319	9190	7430	7418	8013 (±1019)	5.7 (±0.4)	2.04368E-07	5.98E-07	193 (±1.4)	2.07	0.09
BF800	42.6 (±0.3)	-11.8	6849	1319	842	838	999 (±277)	4.4 (±0.5)	2.07395E-07	2.87E-07	39 (±1.4)	1.28	0.62
BG643	69.3 (±0.3)	-13.5	2948	-1185	-2060	-2066	-1770 (±507)	1.8 (±0.5)	2.13538E-07	3.32E-07	55 (±1.4)	1.38	0.52
BH246	46.9 (±0.3)	-10.9	6084	1622	-601	-628	131 (±1291)	8.8 (±0.4)	1.98021E-07	2.39E-07	21 (±1.4)	1.13	0.59
BH275	28.8 (±0.2)	-11.2	9997	6262	3650	3623	4512 (±1516)	8.1 (±0.4)	1.99435E-07	2.38E-07	19 (±1.4)	1.12	0.68
KR925	22.5 (±0.2)	-12.3	11972	7722	6457	6439	6873 (±736)	4.2 (±0.4)	2.07798E-07	3.03E-07	46 (±1.4)	1.30	0.51
MK423	34.5 (±0.2)	-12.0	8560	4314	2742	2731	3262 (±911)	6.4 (±0.7)	2.02989E-07	2.69E-07	33 (±1.4)	1.26	0.68

APPENDIX C

South transect: Location map, noble gas and ¹⁴C age data as reported by Klump, et al. 2008.

well #	avg. corrected ¹⁴ C age (yr)	noble gas temperature (°C)	δ18O temperature (°C)	Δ Ne (%)	q (CE model)	F (CE model)
BR15		2.9 (±0.5)	3.2	72.4 (±4.7)	1.44	0.35
DO1	2065 (±828)	8.1 (±0.4)	6.2	12.5(±1.8)	1.06	0.00
FR5	27851 (±478)	3.4 (±0.4)	6.2	73.8 (±3.7)	1.40	0.2
MUS5		4.1 (±0.6)	7.7	20.1 (±2.5)	1.15	0.8
NB3	54974 (±23619)	3.3 (±0.6)	5.4	34.8 (±3.1)	1.24	0.62
NB7	29606 (±583)	2.6 (±0.2)	5.7	37.9 (±0.3)	1.23	0.51
PEW4		3.5 (±0.5)	5.3	30.6 (±2.9)	1.18	0.51
SUS1		4.5 (±0.6)	4.6	30.0 (±2.9)	1.20	0.65
WK10	20403 (±245)	1.4 (±0.5)	1.9	72.0 (±4.6)	1.40	0.29
WK5	16534 (±371)	2.3 (±0.5)	2.3	45.7 (±2.7)	1.31	0.57
WK9	6277 (±150)	5.5 (±0.5)	5.5	28.3 (±2.2)	1.17	0.58
WK1	8274 (±780)					

South Transect (Milwaukee and Waukesha counties)

