Econ. Environ. Geol.
Vol. 27, No. 1, p. 93-100, 1994.

Lead Transport in Groundwater in Door County, Wisconsin

Nam C. Woo*

ABSTRACT: This study examined the mechanisms of lead transport in the groundwater system and those of irregular detections in groundwater in Door County, Wisconsin. During the spring recharge period in 1991, water-level movement and water-quality change were monitored from two monitoring wells equipped with three piezometers each and from five house wells, respectively.

Water-level responses to recharge events were fast with a relatively short lag time ranging from 3 to 10 days, indicating that recharge of groundwater occurs through the high hydraulic conductivity (K) zones in the Silurian dolomite aquifer system. Lead was detected only on particles filtered from groundwater, but not in dissolved state. Concentrations ranged from 0.2 to 7.1 $\mu g/mg$, converted into the total lead concentration in groundwater ranging from 0.3 $\mu g/l$ to 4.7 $\mu g/l$.

A lag time between recharge events and peak particle movement at the sampled wells was estimated to range from 19 to 22 days. Due to the particulate nature of lead in groundwater, only the wells connected with the high K zones detect lead, causing the spatial variation. In a given well, lead concentration varies at different sampling times due to the variation in the initial amounts of lead-carrying particles introduced into the groundwater system during recharge events, the lag in particle transport and the dispersion of lead-carrying particles along the advective flowpaths.

INTRODUCTION

Groundwater in Door County, Wisconsin (Fig. 1), has been contaminated by lead from the lead-contaminated soils at abandoned lead-arsenate mixing sites. Since 1983, the Wisconsin Department of Natural Resources (WDNR) sampled 333 drinking water supply wells and found irregular detections of lead in groundwater, either temporally and spatially (Stoll, 1988). Although many studies (Sherrill, 1978; Nauta, 1987; Blanchard, 1988; McKereghan, 1988; Craig, 1989; Wiersma and Stieglitz, 1989; Bradbury and Muldoon, 1992) characterized the groundwater flow system in Door County, they could not resolve the mechanism of contaminant transport to the groundwater and the question of irregular distribution of lead contamination. Therefore, this project was initiated to identify the mechanisms of lead transport in the groundwater system and those of irregular detections in groundwater by studying the groundwater movement and its change in quality during the spring recharge events.

HYDROGEOLOGIC SETTINGS

Geologic units in the county range in age from Precambrian to Holocene. Precambrian crystalline rocks occur at depths of greater than 450 m and are overlain by sedimentary rocks of Cambrian, Ordovician and Silurian age (Fig. 2). The rocks generally strike about 20° to the northeast and dip to the southeast at about 8.6 m/km (Sherrill, 1978; Craig, 1989).

The Silurian dolomite aquifer system is the most important source of water supply to wells in Door County and includes the Niagaran aquifer and the underlying Alexandrian aquifer. Unconfined conditions predominate in the upper part of the Niagaran aquifer due to the abundance of vertical joints. Confined conditions generally prevail in the lower part of the Niagaran aquifer and the Alexandrian aquifer due to the dominant bedding-plane joints. The major orientations of the vertical joints are at azimuths of approximately 70° and 155° (Rosen, 1984; Craig, 1989). Bradbury and Muldoon (1992) noticed near-vertical fractures on the bedrock surface which were filled with fine-grained soil. The fractures are spaced about 3 to 7 m apart in the Town of Sevastopol.

METHODS

Groundwater Monitoring and Sampling

The period of groundwater monitoring and sampling was scheduled to overlap the spring recharge

* Dept. of Geology, Natural Sciences Research Institute, Yonsei Univ., Seoul 120-749, Korea

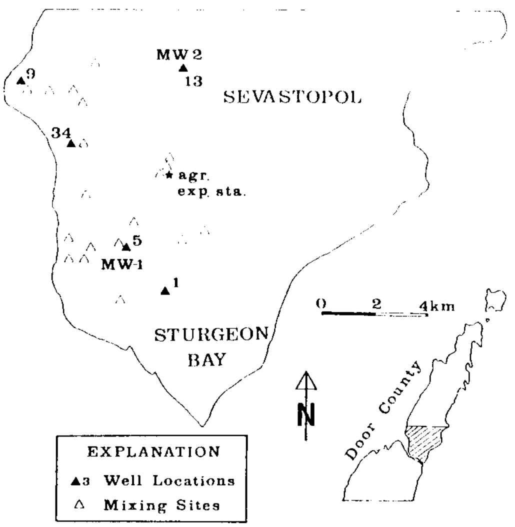

Fig. 1. Location of sampled wells in Sevastopol and Sturgeon Bay areas, Door County: Monitoring wells, MW-1 and MW-2, are at locations of wells 5 and 13, respectively.

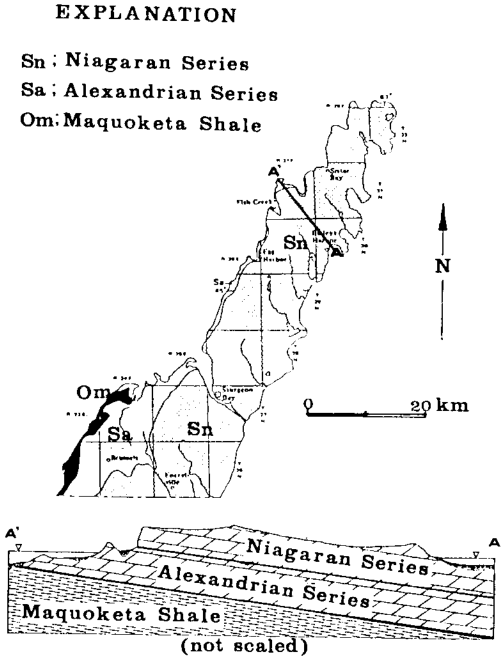

Fig. 2. Bedrock geology (top) and general cross section (bottom) of Door County, Wisconsin (Modified from Sherrill, 1978 and Craig, 1989).

in the study area, because during this period groundwater recharge is maximum, thus producing large variations in water level and possibly groundwater chemistry. Based on the climatological data obtained from the Peninsula Agricultural Experimental Station at Sturgeon Bay (Fig. 1), the spring-melt generally occurs during the last week of March or the first week of April in this area. Therefore, water levels were measured and groundwater samples collected on a daily basis from March 20 to March 26, 1991, for the detailed information of water-level movement and water-quality change during the spring-melt period. After that, three more sets of samples were collected at approximately one week interval.

Water levels were monitored from five piezometers in two monitoring wells (MW-1 and MW-2), located approximately 10 m away from house wells 5 and 13, respectively (Fig. 1). The monitoring wells were constructed by CTW Corp. in 1986, and have been maintained by the WDNR (Stoll, 1992). MW-1 encloses three piezometers (MW-1A, -1B, -1C) at 166 m, 158 m and 148 m above mean sea level (MSL), respectively. MW-2 also holds three piezometers (MW-2A, -2B, -2C) at 158 m, 148 m and 139 m above MSL, respectively. All piezometers were equipped with 1.5 m long screens. Water levels were measured with a fiberglass tape equipped with a plopper. Due to an installation problem, MW-1B was filled with cement grout (CTW CORP., 1986), subsequently no water-level movement was monitored at MW-1B.

Groundwater samples were collected from five house water-supply wells (1, 5, 9, 13 and 34 in Fig. 1) in the Towns of Sturgeon Bay and Sevastopol areas. The samples included three components: water samples filtered through 0.45 micrometer(μm) pore size Whatman membrane filters, unfiltered counterparts and filter cakes. Prior to the collection of samples from the nearest tap to the pump, water was flushed for at least 10 minutes to remove possible contaminant residues in the pump tank. During the groundwater survey for lead and arsenic by the WDNR (Stoll, 1988), water samples were collected after three to five minutes of flushing. This study's protocol was established to exceed the WDNR's flushing but to remain relatively consistent with it.

After flushing, water samples were collected from the tap into a 1-gallon (3.8 liter) plastic jug and immediately filtered through a pre-weighed 0.45 μm filter. The amount of filtered water varied from 1.5 to 3.5 liter depending on filtering time. Pre-weighed filters were stored in sealed plastic Petri dishes to prevent hydration before use. Approximately 500 milliliter (ml) of filtered water was acidified with sufficient concentrated nitric acid (trace metal grade) to bring the solution's pH below 2. Filter cakes were saved in the original Petri dishes. Unfiltered samples were taken without treatment and refrigerated at about 4°C in an ice-cooler. The samples were analyzed for Pb.

The pH, Eh (in mV), temperature and electrical conductivity (EC) were measured on site. The pH meter was calibrated against standard buffer solutions at pH 4.01 and 7.00 before use. For quality control of the sampling procedure, complete field blanks were also taken using distilled deionized (DSDI) water from a 5-gallon plastic jug, which was brought to the field for cleaning equipment and mixing titration reagents. These blank samples could indicate if sampling procedures contributed any Pb.

Analysis of Groundwater and Particles on Filter

Water

Groundwater samples were analyzed for Pb by graphite furnace atomic absorption spectrometry (FAAS) with the detection limit of 0.3 $\mu g/l$. Chemical analysis for Pb conformed the APHA standard methods #3113 (American Public Health Association, 1989). Concentrations of Pb were calculated from calibration curves prepared with standard stock solutions.

Filters

The filters were digested in 125-ml Erlenmyer flasks with concentrated nitric acid according to APHA standard method #3030E (American Public Health Association, 1989). The resulting solutions were analyzed by FAAS for lead and arsenic. Laboratory blanks with DSDI water and blank filter were also run through the same steps for quality control.

RESULTS AND DISCUSSION

Observation of Groundwater Movement

Groundwater levels were monitored from five piezometers at two monitoring sites during the spring of 1991. After precipitation events, water moves downward through the unsaturated zone. Because elevations of the ground surface and the water level at MW-1 are about 198 and 184 m, respectively, recharge water should move downward about 14 m to reach the water level. As the infiltrating water recharges the aquifer, the pressure on the water level increases due to both compressed air and hydraulic head. The elevated pressure is transmitted through the aquifer system, causing water-level movement over time.

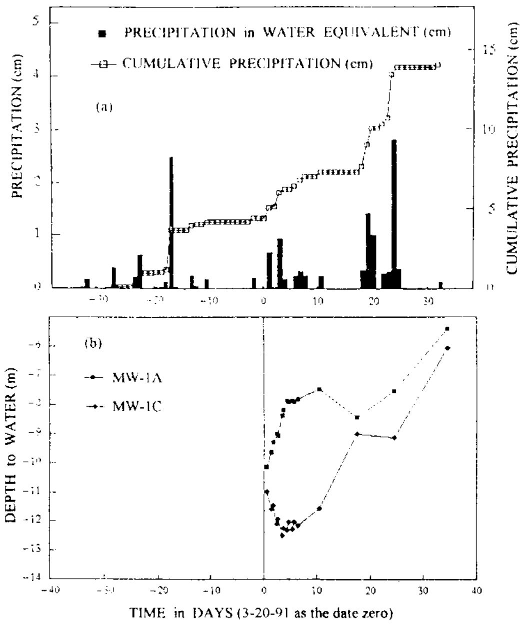

Fig. 3. Comparison of water-level responses at monitoring well (MW-1) with precipitation records in Sevastopol area. Elevations of screens at MW-1A and -1C are 166 and 148 m above mean sea level, respectively. Negative depths indicate below ground surface.

Water-level responses at two piezometers in MW-1 were compared through the monitoring period (Fig. 3). The responses were also compared with precipitation records to estimate the lag time between the responses and recharge events. Precipitation data represent the amount of daily precipitation as water equivalent.

Water levels in MW-1A and MW-1C showed distinct responses during the monitoring period. The water level in MW-1A moved up by 2.4 m from March 20 to 25 (from day 0 to day 5 in Fig. 3-(b)). On March 30 (day 10), water-level movement reversed. Water-level movement was again reversed on April 6 (day 18), and continued rising until April 23 (day 34). In contrast, the water level in MW-1C dropped by 1.5 m during the wet days from March 20 to 25 (from day 0 to 5 in Fig. 3-(b)), with the average rate of 0.3 m/day. The rate of water-level drop seems too high to be the rate of water-level decline during winter time. Therefore, the drop was considered as a water-level decline after a previous big recharge event, which was able to raise water-level higher than the drop.

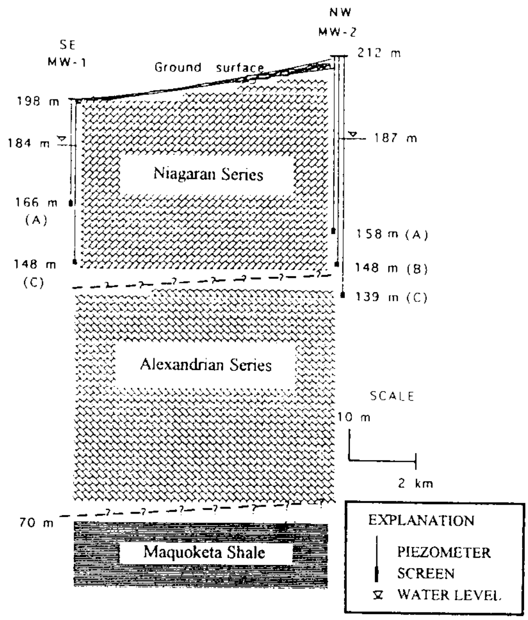

Fig. 4. Elevations of piezometers in monitoring wells, water table, and bedrocks above mean sea level.

In recharge events, 1 cm of recharge can produce 1 to 2 m of water-level rise in the dolomite aquifer which has an effective porosity of 0.5 to 1.0 % (Bradbury and Muldoon, 1992). To have water-level drop of at least 1.5 m in MW-1C, the preceding recharge event should record at least 0.75 cm. The climatological data showed that the soil temperature at 15 cm depth was always above 0℃ after February 18, 1991 (day -30). Therefore, the ground probably thawed sometime after February 18. The recharge event greater than 0.75 cm after February 18 occurred on March 2 (day -18) with 2.49 cm/day.

During the dry period between March 10 and March 18 (days -10 and -2), water-level in MW-1C had probably dropped continuously. Thus, the first drop of water level in MW-1C from March 20 to 25 (from day 0 to 5) seems to show water-level decline during the tail of the dry period. The subsequent rise of water level in MW-1C from March 26 (day 6) probably resulted from cumulative recharge events starting on March 18 or 21 (day -2 or 1). Therefore, the water-level response in MW-1C was estimated to occur about 5 to 9 days behind recharge events.

After March 26 (day 6), the water level in MW-1C kept moving up until April 6 (day 17), probably due to the recharge events from March 18 to 31 (day -2 to 11). A water-level drop was observed on April 13 (day 24), which could have been caused by the dry period from April 1 to 7 (day 12 to 18). On April 23 (day 34), the water level was the highest observed during the monitoring period. The water-level rise probably started with recharge events beginning on April 8 (day 19). From the water-level responses, lag time was estimated to be about 9 days, consistent with the previous estimation of 5 to 9 days.

Except for the first water-level drop, the other inflections of water-level movement in MW-1C occurred also in MW-1A, but on different days. The first reversal in MW-1A on March 30 (day 10 in Fig. 3) could be matched with the inflection in MW-1C on April 6 (day 17). The second reversal of water-level movement in MW-1A on April 6 (day 17) could be matched with one in MW-1C on April 13 (day 24). The two matching inflections gave an estimation of lag time in water-level response between MW-1A and -1C of about 7 days. However, after March 26 (day 6) the water level was monitored approximately once a week. The 7-day lag time could be partly a result of sparse water-level measurements. Therefore, the real lag time was estimated to be about 7 ± 3 days.

As the screen of MW-1A is located at a shallower depth than that of MW-1C (Fig. 4), water-level responses in MW-1A should precede those in MW-1C. Subsequently, the first big rise in water level at MW-1A from March 20 to 26 (day 0 to 6 in Fig. 3) does not match with the recharge event on March 2 (day -18). Instead, the rise should be matched with recharge events starting from March 18 (day -2). Therefore, the maximum lag time between recharge events from March 18 and water-level rise in MW-1A from March 20 was 2 days. The relatively fast response of water level in MW-1A implies that the recharge probably occurs through highly conductive zones in dolomite.

From March 20 to 31 (day 0 to 11 in Fig. 3-(a)), recharge accumulated about 2.9 cm. The minimum water-level rise in MW-1C responding to these events was about 3.48 m (Fig. 3-(b)). Therefore, the maximum effective porosity of the dolomite was estimated as (2.9 cm/3.48 m)=0.0083 (0.83%). This agreed with Bradbury and Muldoon's (1992), ranging from 0.5 to 1.0 %.

Vertical hydraulic conductivity (Kv) of the dolomite between MW-1A and -1C was estimated from the lag time in water-level response. The screen of MW-1C located 18 m below that of MW-1A. The maximum difference in water-level between MW-1A and -1C appeared about 4.7 m (Fig. 3). Therefore, the maximum vertical gradient between MW-1A and -1C

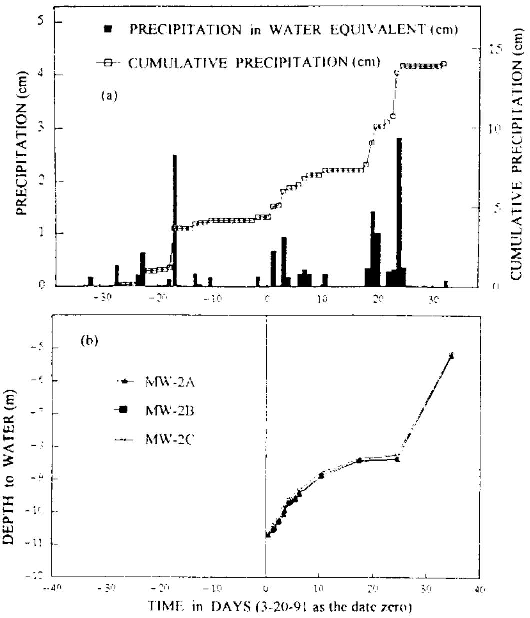

Fig. 5. Comparison of water-level responses at monitoring well (MW-2) with precipitation records in Sevastopol area. Elevations of screens at MW-2A, -2B and -2C are 158, 148, and 139 m above mean sea level, respectively. Negative depths indicate below ground surface.

was (4.7 m/18m)=0.26. As the maximum lag time was 10 days, the minimum vertical hydraulic conductivity (*K*v) was {(18 m/10 day)×(0.0083/0.26)}=0.06 m/day (=10 $^{4.3}$ cm/sec).

The *K*v for the dolomite above the screen of MW-1A was calculated by the same procedure. The maximum vertical hydraulic gradient was assumed as 0.26 as previously estimated. Lag time between recharge events and water-level responses was 2 days. The minimum *K*v was about 0.29 m/day (10 $^{3.5}$ cm/sec). This agrees with Rovey's (1990) conclusion on high conductivity zones, of which geometric mean of hydraulic conductivities was greater than 10 ^{4}cm/sec, in the Silurian dolomite aquifer in southeastern Wisconsin. In addition, the *K*v of the dolomite above the screen of MW-1A and that between MW-1A and -1C show a difference of less than an order of magnitude, thus implying that the screens of both MW-1A and -1C are located in a relatively homogeneous aquifer.

At the site of MW-1, the boundary between the Silurian dolomite and the Maquoketa shale occurs at about 70 m above MSL (Fig. 4; Sherrill, 1978). The maximum thicknesses of the Niagaran series and the Alexandrian series were estimated as about 105 m and 70 m, respectively. As the screens of MW-1A and -1C are located 96 m and 78 m above the boundary to the Maquoketa Shale, they open to most likely the Niagaran dolomite aquifer system.

At MW-2, water levels in the three piezometers (MW-2A, -2B and -2C) fluctuated almost identically during the monitoring period (Fig. 5). The lack of vertical gradient indicates two possibilities. The first possibility is that vertical high conductivity zones like joints extend to the full depth of MW-2, subsequently the three piezometers become interconnected in the aquifer. If three piezometers are vertically interconnected, water levels in MW-2 should move similarly to those of MW-1A showing fast responses to recharge events. Secondly, improper construction of the three piezometers may have left them unseparated by cement grout. The possibility of well-construction failure is supported by the well construction report (CTW CORP., 1986), which describes problems with cement grout during the sealing process. In this case, MW-2 with three piezometers will actually function like an open bore-hole, and water-level responses should also look like those of MW-1A (Fig. 3-(b)). Water-level rise would be fast due to recharge through MW-2A. However, water-level drop during the dry period would become slower than that of MW-1A, because recharge would continue through MW-2B and -2C.

Water-level responses in MW-2 piezometers (Fig. 5) generally looked like those of MW-1A. However, reversals of MW-1A on March 30 and on April 6 (day 10 and 17 in Fig. 3) were not observed in MW-2. The absence of reversals in water level could be due to the weekly monitoring frequency after March 26 (day 6). Based on water-level measurements alone, it could not be determined that which possibility is more reasonable for water-level responses in MW-2.

MW-1 and MW-2 are located almost along the strike of the bedrock in the study area, approximately N20°E (Sherrill, 1978; Fig. 2). Therefore, the vertical sections of the dolomite aquifer at the MW-1 and MW-2 would be comparable. Screens of the three piezometers of MW-2 are possibly located in the lower part of the Niagaran series or the upper part of the Alexandrian series (Fig. 4). Sherrill (1978) described the observations of discontinuous vertical joints diminishing both in size and number with depth in the Niagaran series. Therefore, well construction failure is more probable than the vertical fracturing, and is proposed as the cause of the observed water-level responses of the three piezometers in MW-2.

Table 1. Field conditions and Pb levels of ground water at the study wells in Door County, Wisconsin.

House Wells		1 (AVG)	1 (SD)	5 (AVG)	5 (SD)	9 (AVG)	9 (SD)	13 (AVG)	13 (SD)	34 (AVG)	34 (SD)	Total
Temperature	(C)	9.3	0.8	8.9	0.7	9.0	1.0	8.4	0.8	8.7	1.0	
pH		6.9	0.29	6.9	0.28	6.9	0.33	6.9	0.31	7.0	0.41	
Eh	(mV)	210	57	149	44	186	61	156	38	178	44	
EC	(mmbos)	0.357	0.016	0.417	0.047	0.400	0.021	0.358	0.053	0.396	0.012	
Pb in filtered samples		0(8)*		0(15)		0(9)		0(15)		0(9)		0(56)
Pb on particles		2(8)		9(15)		9(9)		15(15)		9(9)		44(56)
Concentration of Pb on particles (μg/mg)		1.1	0.39	0.4	0.27	4.3	1.91	0.8	0.37	1.4	0.51	

The number in ( ) indicates the total number of samples and the number outside ( ) donotes the number of samples with Pb detection

The water-level rise in MW-2 from March 20 to 26 (day 0 to 6 in Fig. 5-(b)) could not be matched with recharge event on March 2 (day -18 in Fig. 5-(a)), because no significant water-level drop due to the dry period from March 10 (day -10) to March 17 (day -3) was recorded after the rise. Therefore, the water-level rise should be matched with the recharge events beginning on March 18 (day -2). Subsequently, the maximum lag time is estimated to be three days. In addition, the inflections on April 6 and 13 (day 17 and 24 in Fig. 5-(b)) probably match with the beginning and ending of the dry period from April 1 to 7 (day 12 to 18 in Fig. 5-(a)). Because of the weekly monitoring interval, the general lag time between recharge events and water-level rise in MW-2 was estimated as ranging from 3 to 10 days.

In summary, unconfined aquifer conditions were observed in the study area. Water-level responses to recharge events were fast with a relatively short lag time ranging from 3 to 10 days, indicating that groundwater recharge occurs through the high hydraulic conductivity zones, such as fractures well developed in the Silurian dolomite aquifer system. The dolomite responds as a homogeneous unit to the depth of about 50 m. The effective porosity of the dolomite was 0.83 %. The minimum vertical hydraulic conductivity (Kv) of the dolomite was approximately 10 ^{4} cm/sec.

Analysis of Groundwater and Particles

A total of 136 groundwater samples and 86 filter cakes were collected from five house wells and two monitoring wells. However, samples from both monitoring wells showed anomalously high pH's, ranging from 11.9 to 12.5. The high pH's indicate that cement grout materials may have been introduced into the monitoring wells. Therefore, chemical analyses of those samples were not done. For Pb, each filtered groundwater sample was analyzed three times and the standard deviation was calculated by the built-in program in the FAAS. The results were examined using a Q-test (Skoog and West, 1971) to identify outliers at the 95 % confidence level. No lead was detected in field and laboratory blanks.

Table 1 summarizes the general field condition and lead levels of groundwater at the five house wells. Lead detections occurred on particles filtered from groundwater. Concentrations ranged from 0.2 to 7.1 μg/mg at the detection limit of 0.1 μg/mg, converted into the total Pb concentration in groundwater ranging from 0.3 μg/l to 4.7 μg/l. However, no lead was found in the filtered groundwater samples. This indicates that Pb is not in the dissolved state, but sorbed onto particles in groundwater.

Particle Transport of Lead in Groundwater

Because of the lead association with particles in groundwater, particle movement was examined to determine the transport of lead in groundwater during the spring recharge. About 80 % of significantly lead-contaminated wells in the county (15 out of the 19 wells for which well construction information is known; Stoll, 1988) were cased to 45 m below the ground surface. Therefore, groundwater movement in MW-1C, screened 50 m below the surface, was considered to represent the hydraulic system at the study area.

After precipitation events, recharging water carries soil particles into the groundwater system via advective flows. Water-level responses at wells are initiated by a kinematic pressure wave transmitted through the aquifer. If the pressure change causes the partic-

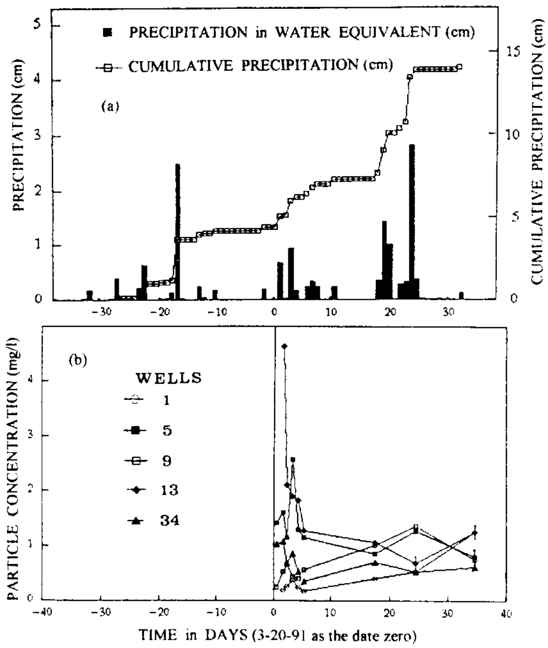

Fig. 6. Variation of particle concentration in ground water at the sampled wells in Door county druing the spring recharge: Error bars represent typical uncertainty of particle concentrations. (a); precipitation data and (b); particle concentrations.

les to dislodge, the variations of particle concentrations at sampled wells (Fig. 6) should be consistent with the water-level changes at MW-1C in Fig. 3. However, the two vary inconsistently, thus implying that the particles generated by the pressure change can be ignored. In addition, particle movement should be registered later than water-level responses at sampled wells, because particles move slower than the kinematic pressure wave in groundwater.

The advective flows arrive at the wells with peaks of particle concentration. These peaks occurred between March 21 and 24 (day 1 and 4 in Fig. 6). The water level declined in MW-1C from March 20 to 25 (day 0 to 5 in Fig. 3) because of the dry period after the recharge event on March 2 (day -18). Recharge events starting on March 18 (day -2) caused the water-level rise in MW-1C beginning on March 26 (day 6). Therefore, the recharge event responsible for those advective flows and particle peaks could not be the ones starting from March 18 (day -2), because of the relatively slower movement of advective flow than that of a pressure wave. Instead, the particle peaks were caused by the recharge event on March 2 (day -18). Consequently, a lag time between the recharge event and peak particle movement at the sampled wells was estimated to range from 19 to 22 days.

The lag time could vary at different wells depending on factors including the thickness of soil, the sizes and development of preferential flow paths in soil, the depth to water level, the extent and connectivity of vertical and horizontal fractures in dolomite aquifer, the depth to openings of wells, the flow rate of groundwater, precipitation conditions, vegetation and land use. In the study area, the spatial variation of lead detection could be caused by the advective transport of lead-carrying particles along the high K flowpaths such as fractures in the Silurian dolomite aquifers.

Temporal variation of lead concentration in a given well could be caused by several factors including: initial amounts of particles introduced into the groundwater system after recharge events, the lag of particle transport in groundwater flow, and the following dispersion of lead-carrying particles along the advective flows. The lead concentration will be maximum at the peak time of particle concentration, and become low and various at different sampling times.

SUMMARY AND IMPLICATIONS

Water-level responses to the recharge events indicate unconfined conditions of the aquifer and verify the existence of high hydraulic conductivity (K) zones in the Silurian dolomite aquifer system. Fast recharge of groundwater in ranging from three to 10 days through the high K zones in the aquifer confirms that groundwater in the county can be easily contaminated from various pollutants in soils such as pesticide residues, agricultural wastes, leaked materials from underground septic tank and leaking underground fuel tanks. Therefore, to protect from contamination the groundwater serving as the major source of water supply in the County, possible pollutants in soils should be identified and removed from the groundwater recharge area.

Lead was found on the particle fractions in total Pb concentrations in groundwater ranging from 0.3 $\mu g/l$ to 4.7 $\mu g/l$. The particulate nature of lead signifies that the lead contamination of groundwater should be monitored based on the total lead concentration in unfiltered samples, but not in filtered water samples.

A lag time between recharge events and peaks of particle concentration at the wells screened 50 m below the surface ranged from 19 to 22 days. Measured lag times can be used to notify the county resident the periods of possibly elevated lead levels in groundwater after recharge events. In addition, groundwater monitoring of lead contamination should be intensified during those periods.

The spatial and temporal variation of lead detections could be caused due to the particulate nature of lead in groundwater. Lead detections in groundwater occurs only at the wells connected with the high *K* flowpaths. In addition, the lead concentration in a given well varies at different sampling times due to the variation in the amounts of particles introduced into the groundwater system during recharge events, the lag time of Pb transport in groundwater and the dispersion of Pb-carrying particles along the paths.

ACKNOWLEDGEMENTS

Thanks to Ms. Graff in the Department of Geosciences, UW-Milwaukee, and Mr. Anderson and Mr. Paddock in the Center for Great Lake Studies for their invaluable help during laboratory studies and analysis.

REFERENCES

American Public Health Association (1989) Standard methods for the examination of water and wastewater. Clesceri, L., Greenberg, A. and Trussel, R., eds., 17th ed, 1524 p.

Blanchard, M. C. (1988) Investigation of the shallow fractured dolomite aquifer in Door County, Wisconsin. MS Thesis, University of Wisconsin-Madison, 185 p.

Bradbury, K. R. and Muldoon, M. A. (1992) Hydrogeology and groundwater monitoring of fractured dolomite in the upper priority watershed, Door County, Wisconsin. Wisconsin Geological and Natural History Survey.

Craig, K. D. (1989) Determination of Ground-Water Discharge into Green Bay and Lake Michigan in Door County, Wisconsin. The University of Wisconsin-Milwaukee, MS Thesis, 162 p.

CTW CORP. (1986) Well Construction Report. Waukesha, Wisconsin.

McKereghan, P. F. (1988) Digital Simulation of Chloride Transport in the Silurian Dolomite Aquifer in Door County, Wisconsin. The University of Wisconsin-Milwaukee, MS Thesis, 137 p.

Nauta, R. (1987) A three-dimensional groundwater flow model of the Silurian Dolomite aquifer of Door County, Wisconsin. M. S. Thesis, The University of Wisconsin-Madison, 105 p.

Rosen, C. J. (1984) Karst geomorphology of the Door Peninsula, Wisconsin. M. S. Thesis, University of Wisconsin-Milwaukee, Geography, 119 p.

Rovey, C. W. (1990) Stratigraphy and sedimentology and diagenesis of Silurian and Devonian carbonates in the southeast Wisconsin with implications for regional ground-water flow into Lake Michigan. Ph.D. dissertation, Department of Geosciences, UW-Milwaukee.

Sherrill, M. G. (1978) Geology and groundwater in Door County, Wisconsin, with Emphasis on Contamination Potential in the Silurian Dolomite. Geological Survey Water-Supply Paper 2047, 38 p.

Skoog, D. A. and West, D. M. (1971) Principles of instrumental analysis. 710 p.

Stoll, R. C. (1988) Lead concentration variabilities of private water supplies located in Door County, Wisconsin. Wisconsin DNR, 95 p.

Stoll, R. C. (1992) Personal Communication. Wisconsin DNR, Green Bay, Wisconsin.

Wiersma, J. H. and Stieglitz, R. D. (1989) Lead Migration from contaminated sites in Door County, Wisconsin. The University of Wisconsin-Green Bay, 89 p.

Manuscript received 16 August 1993

위스컨신주 도어지역의 지하수내 납성분의 이동

우 남 칠

요 약: 이 연구는 위스컨신주 도어지역의 지하수계에서의 납성분의 이동과 불규칙한 검출에 대한 메카니즘을 규명하였다. 1991년 봄 충진기간 동안에, 세개씩의 수위정(piezometers)을 내장한 두곳의 관측정과 다섯군데의 가정우물에서 수위와 수질의 변화를 측정하였다.

충진현상에 대한 수위의 변화는 지하수의 충진이 약 4일에서 10일 정도 연체되면서 비교적 빠르게 일어남을 보여주며, 이것은 지하수의 충진이 실루리안 돌로마이트 대수층 내에서 높은 수리전도율(*K*)을 가진 지대를 통해서 일어나고 있음을 지시해준다. 지하수내의 납성분은 미립자에서만 발견되었고, 용융된 상태로는 검출되지 않았다. 그 농도는 0.2에서 7.1 μg/mg인데, 이는 지하수내 총 납성분으로는 0.3에서 4.7 μg/*l*이다.

충진사건부터 최대의 미립자가 가정우물까지 이동하는데 걸리는 연체시간은 약 19일에서 22일 이었다. 지하수내에서 납성분이 미립자와 더불어 있으므로, 가정우물 중에서도 높은 수리전도대와 연결된 우물에서만 납성분이 검출되고, 결과적으로 지역적 불규칙성을 유발한다. 또한 우물에서는, 충진사건에 의한 지하수계로의 미립자의 유입량의 변화, 미립자의 이동에 따른 연체현상, 그리고 이동중의 분산 등으로 인해서 납성분의 농도가 채취하는 시간에 따라서 다르게 나타난다.