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Implications of the distribution of δD in pore waters for groundwater flow and the timing of geologic events in a thick aquitard system

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Abstract. A detailed vertical profile of the stable isotope δD in pore water was measured through a thick aquitard system consisting of surficial Quaternary clay-rich till (80 m thick) and an underlying Cretaceous marine clay (76 m thick). Numerical modeling was used to simulate one-dimensional (vertical) groundwater flow and transport of δD. Best fit simulations to the data provided an independent estimate of long-term groundwater velocity through the aquitard and estimates of the timing of late Pleistocene and Holocene events. Best fit simulations to the measured isotope profile across the till-clay interface yielded a groundwater velocity of 0.75–1.0 m per 10 ka for a transport time of between 20 ka and 30 ka. The estimate of velocity agreed well with that calculated from hydraulic data and suggested that hydraulic conductivities of these aquitards are independent of volume tested. The 20–30 ka time frame required for the δD profile to develop across the till-clay interface reflects the timing of till deposition and shows that the till is the Battleford Formation, a younger till than previously believed. Numerical transport modeling of δD in the upper 30 m of the profile yielded a nonunique fit. Assuming a similar groundwater velocity to that determined across the till-clay interface, a best fit was obtained for a transport times of 7.5–10 ka. This range compared favorably with that reported for the start of the Holocene (about 10 ka B.P.). This study shows that the application of δD, and by analogy $\delta^{18}O$, to the study of thick aquitard systems not only can provide independent, long-term estimates of very low groundwater velocities but can also provide insight into the timing of major geologic events such as glaciations.

1. Introduction

Nonfractured, clay-rich glacial tills deposited by the late Pleistocene continental glaciations are widespread throughout Canada, the northern United States, and Europe. Despite their widespread occurrence and importance as protective covers for regional aquifers and as potential waste repositories, few detailed groundwater studies have been conducted on these nonfractured, low-permeability tills. A few studies in western Canada show tills there typically have groundwater velocities of less than a few tens of meters per 10 ka, with downward flow [*Hendry*, 1988; *Keller et al.*, 1989; *Remenda et al.*, 1996; *Shaw and Hendry*, 1998].

Vertical profiles of environmental isotopes (δD and $\delta^{18}O$) in pore waters have provided valuable information on the hydrology of surficial till aquitards. For example, δD and $\delta^{18}O$ were used to show molecular diffusion was the dominant transport mechanism in several surficial aquitards [*Desaulniers et al.*, 1981; *Remenda et al.*, 1994, 1996]. Stable isotopes were further used to constrain hydraulic conductivities (K) in nonfractured clay tills, and results suggested conventional permeability testing in clay till aquitards can overestimate K [*Remenda et al.*, 1996]. Previous investigations using δD and $\delta^{18}O$ have focused on the upper 30 m of surficial aquitards. In these studies, vertical changes in δD and $\delta^{18}O$ were attributed to a shift from cooler climatic conditions of the late Pleistocene to warmer postglacial conditions between 12 ka and 10 ka B.P. However, the applicability of detailed δD and $\delta^{18}O$ profiles collected through thicker aquitard sequences in order to provide independent estimates of groundwater velocities and information on the timing of major geologic and climatologic events has not been tested.

The objectives of this study were to (1) instrument a thick, nonfractured till aquitard system for δD and $\delta^{18}O$ sampling, (2) use δD and $\delta^{18}O$ profiles to estimate the vertical groundwater velocity through the aquitard system, (3) compare the velocities determined from the isotopic data with velocities determined hydrodynamically, and (4) determine if δD and $\delta^{18}O$ modeling could provide information on the timing of major climatic and geologic events.

Selection of a suitable aquitard test site is critical to successful and accurate interpretation of stable isotopic data. In this study, five constraints were imposed to improve the potential for success. First, the clay till aquitard was to be underlain by a thick, nonfractured, clay-rich, low-permeability bedrock unit, making high vertical hydraulic gradients and advective fluxes across the till aquitard less likely. If the aquitard were underlain by an aquifer, it is probable that δD and $\delta^{18}O$ at this lower boundary would have changed since till deposition. Thus the

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Figure 1. Location of the study site (triangle).

presence of an underlying bedrock clay aquitard provided a well-defined and stable lower concentration boundary for the overlying till aquitard. Second, on the basis of theoretical transport modeling, both the till aquitard and underlying bedrock clay aquitard had to be relatively thick (>60 m) to avoid irresolvable overlap of diffusion profiles. Third, because of possible differences in geotechnical and geochemical properties among variable age tills [cf. *Remenda et al.*, 1996], it was desirable that the thick till aquitard consist of a single geologic unit. Fourth, the till and clay aquitards were to be laterally extensive and relatively homogeneous, allowing one-dimensional transport modeling. Finally, the aquitard system needed to be well characterized hydrogeologically. An aquitard system located about 140 km south of Saskatoon, Canada (Figure 1), met these conditions (discussed in section 2) and was selected for this study.

2. Site Hydrogeology

The stratigraphy of the study site consists of 80 m of plastic, clay-rich Pleistocene till (35% clay-size fraction) unconformably overlying 76 m of massive, plastic clay (55% clay-size fraction) of the Cretaceous (72 Ma–71 Ma B.P.) Snakebite Member of the Bearpaw Formation, hereafter called clay (Figure 2). These deposits are underlain by the laterally extensive (22 m thick) Ardkenneth Member of the Bearpaw Formation (74.5 Ma–72 Ma B.P.), an aquifer which acts as a drain to the overlying clay. The upper 3 to 4 m of the till is fractured, weathered, and brown in color. Below 3–4 m the till is massive, unoxidized, dark gray, and nonfractured. Drilling logs show that the till is relatively homogenous with depth except for a sand layer (1.5 m thick) and oxidized zone encountered at 65 m and from 61 to 62.5 m below ground, respectively (Figure 2). *Christiansen* [1986] did not report the presence of this sand layer in a stratigraphic borehole drilled at the study area. This suggests that the sand layer is a localized pocket and not a significant geologic feature within the till. The oxidized zone is discussed below.

Bulk hydraulic conductivity (K) for the till and clay were estimated to be of the order of 10^{-11} m s^{-1} and 10^{-12} m s^{-1}, respectively [*Shaw and Hendry*, 1998]. The very low K of the unoxidized till and clay was reflected by the ability to install piezometers in the till and clay in dry boreholes, extending to depths of more than 90 m below the water table. The low K of the clay was further evidenced by the fact that it took more than 1 year to yield 500 mL of water from piezometers installed in the clay. The average downward groundwater velocity through the unoxidized till was calculated to be between 0.5 and 0.8 m per 10 ka based on laboratory and field determined K values, a measured hydraulic gradient of 0.014, and a measured porosity of 0.31 [*Shaw and Hendry*, 1998]. In contrast to the unoxidized till, groundwater flow in the upper oxidized till zone is dynamic, responding to spring snow melt and precipitation events. Detailed descriptions of the geology and hydrogeology at the site are presented by *Shaw and Hendry* [1998].

3. Materials and Methods

The δD and δ^{18}O profiles of pore waters were obtained from 21 piezometers installed in the till and upper 16 m of clay (Figure 2) in the fall of 1995 ($n = 18$) and summer of 1996 ($n = 4$). Samples were collected on seven occasions from July 1995 to April 1998. Piezometer construction details are given by *Shaw and Hendry* [1998]. Water samples were also obtained on three occasions from a domestic well completed in the Ardkenneth aquifer. The well is located about 2 km south of the site. All water samples were analyzed for δD and δ^{18}O by standard CO_2-water and H_2-water equilibration techniques at the National Water Research Institute, Saskatoon. The δD and δ^{18}O results are reported as the relative difference between the $^{18}O/^{16}O$ and $D/^{1}H$ abundance ratios of the samples and standard mean ocean water expressed in per mil. The precision of the δD and δ^{18}O analyses were ±2 and ±0.1‰, respectively.

Water samples collected from all piezometers during the first sampling were sent to the University of Waterloo for enriched tritium analyses by liquid scintillation counting. Tritium analyses are reported as tritium units (TU), where 1 TU equals 1 tritium atom (^{3}H) in 10^{18} ^{1}H atoms. The precision of the ^{3}H analyses was ±0.8 TU.

Installation of piezometers, especially at depths greater than 100 m, was not feasible because of time constraints and costs. As a result, an alternate method was used to obtain additional

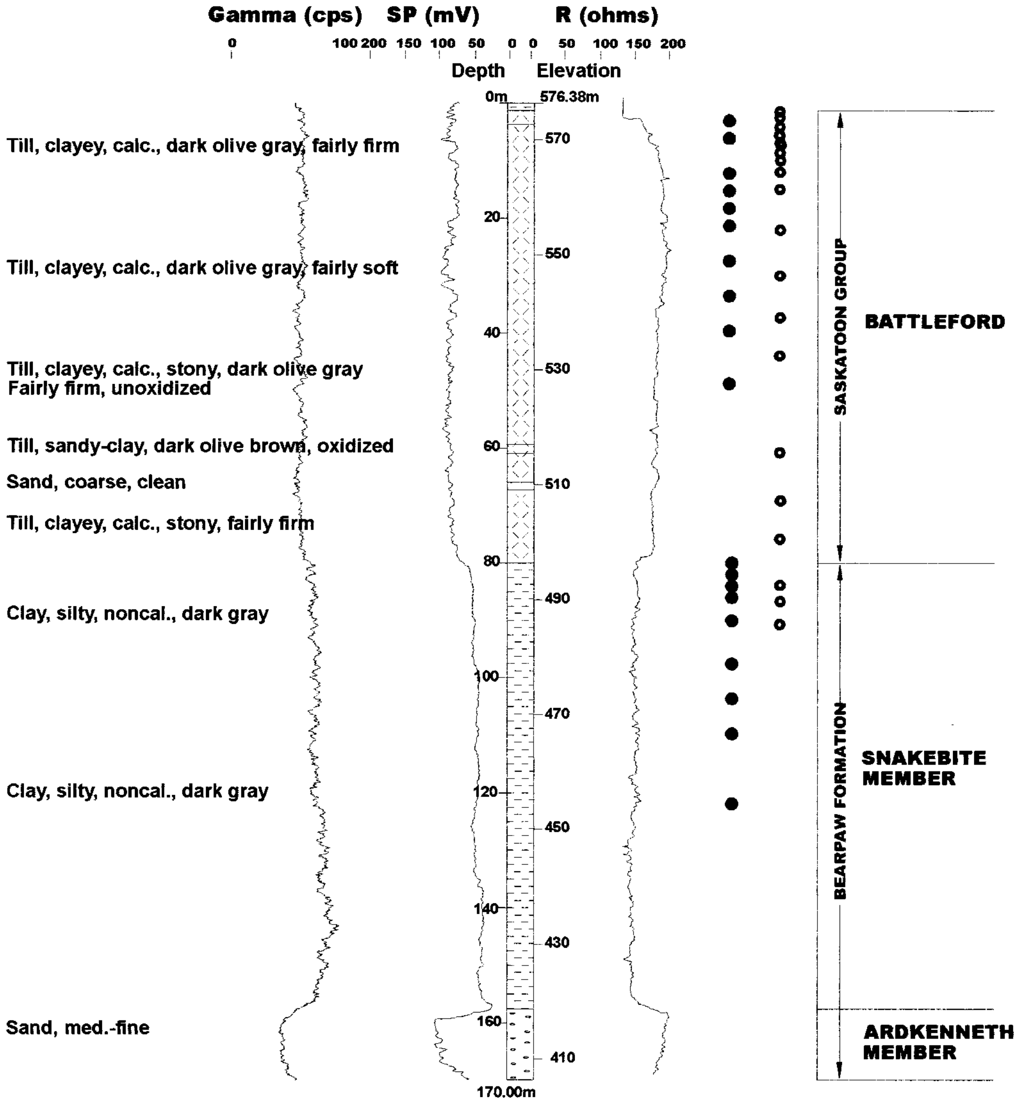

Figure 2. Borehole and geophysical logs at the study site. R is resistance; SP is spontaneous potential. Location of core samples (solid circles) and midpoint of the sand pack for the piezometers (open circles) are shown.

pore water isotopic data. The radial diffusion cell technique [*van der Kamp et al.*, 1996] was used to obtain isotopic data from six core samples collected from the till and seven samples from the clay (Figure 2). The radial diffusion cell approach was also used to measure the effective diffusion coefficients (D_e) and effective porosities (n_e) for δD in three core samples from the till and three from the clay [*Novakowski and van der Kamp*, 1996]. All radial diffusion testing was conducted at a temperature of 4°C, which was similar to the in situ pore water temperature of between 4.5°C and 5.5°C.

4. Results and Discussion

4.1. Distribution of δD, δ^{18}O, and ^{3}H

The δD and δ^{18}O values of pore water from piezometers, domestic wells, and diffusion cells lie on the Saskatoon local meteoric water line (Figure 3). This showed that δ^{18}O and δD profiles are not the result of either evaporation or isotopic exchange with minerals but can only be attributed to mixing between isotopically enriched pore waters and isotopically depleted pore waters. The meteoric correlation between the δD and δ^{18}O allowed us to focus discussion on hydrogen isotopes, knowing that these findings apply equally to δ^{18}O.

A summary of the δD measurements on water samples collected from piezometers showed little variability in each piezometer over 4 years of sampling. In all cases, measurements were close to or within the error of measurement (Figure 4). The range in the δD values from the piezometers in the upper oxidized till zone reflected the water-level fluctuations in this fractured zone (from at or near ground surface in the spring of 1996, 1997, and 1998 to about 3 m below ground surface in the winters of 1997 and 1998) [*Shaw and Hendry*, 1998].

Results of pore water δD obtained using the radial diffusion cells exhibited the same δD trends in concentrations with depth as those from the piezometers (Figure 4), with the exception of the sample from 34.4 m. The δD value obtained from the sample at 34.4 m was 3‰ greater than values from an adjacent piezometer. We attribute the anomalous δD value from the diffusion cell test at 34.4 m to sampling error during the diffusion testing. The overall good fit between the pore

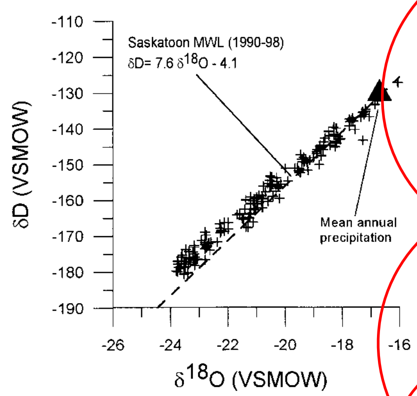

Figure 3. Plot of δD versus $\delta^{18}O$ for pore water samples (crosses) collected from 21 piezometers installed at the study site. The dashed line represents the Saskatoon local meteoric water line.

water δD values obtained from core samples using the radial diffusion cell technique and piezometer samples demonstrates that the radial diffusion cell technique is a viable alternative method to obtain estimates of the δD and $\delta^{18}O$ from clay-rich aquitards.

Altogether, well-defined trends in δD were observed through the 156 m of the nonfractured aquitards. A decrease in δD values with depth was observed in the unoxidized till from −136‰ at the top of the unoxidized till to −178‰ at about 30 m depth. Between 30 and 46 m the δD reached a minimum at about −178‰. The δD values thereafter increased across the till-clay interface to −144‰ at 105 m. Below 105 m the δD values in the clay and the underlying aquifer were constant at about −144‰.

The δD profile through the till aquitard cannot be attributed to the upward leakage of water from the underlying Ardkenneth aquifer because the hydraulic head in the aquifer is about 50 m lower than that in the till [*van der Kamp and Jaworski*, 1989]. Tritium results showed the trends in δD were not the result of downward leakage from the more permeable oxidized till zone along the piezometers as ^{3}H concentrations in all water samples from the unoxidized till were below detection. In contrast, water samples collected from the oxidized and fractured till near ground surface (2.3 and 2.8 m depths) were tritiated (12.4 and 9.2 TU, respectively), indicating the presence of post-1952 recharge water at these depths, and emphasized the dynamic nature of groundwater movement in fractured till.

Sauer et al. [1993] states that tills in southern Saskatchewan were deposited by temperate glaciers. As temperate glaciers are characterized by meltwater throughout their thickness [*Ahlman*, 1935], the low δD values (−178‰) from 30 to 46 m were attributed to glacial meltwater introduced into the till during its deposition in the late Pleistocene. These low δD values were very similar to the inferred δD of glacial meltwaters impounded in Lake Agassiz (about −180‰) [*Remenda et al.*, 1994], a proglacial lake that occupied parts of North Dakota and southern Manitoba at the end of the last glacial maximum [*Fenton et al.*, 1983]. This similarity provided additional support for assuming that the pore water between 30 and 46 m is predominantly Pleistocene glacial meltwater with a δD of about −178‰.

The mean δD value measured at the top of the unoxidized till (−136‰) was identical to average precipitation at Saskatoon, reflecting present-day tritiated input to the system (−136‰, $n = 66$) (Figures 1 and 3). The decreasing trend in δD from the top of the unoxidized till to about 30 m depth is similar to that observed in other surficial, clay-rich deposits [*Desaulniers et al.*, 1981; *Hendry*, 1988; *Simpkins and Bradbury*, 1992; *Remenda et al.*, 1994, 1996]. These studies attribute the presence of such a profile to a rapid change in climatic conditions that occurred after the last deglaciation in the area, whereby the low temperatures of the late Pleistocene were replaced by higher temperatures of the Holocene between 12 ka and 10 ka [*Sauchyn*, 1990; *Greenland Ice-Core Project Members*, 1993].

The difference in the δD values between the top of the unoxidized till (−136‰) and the glacial meltwater (−178‰) translates to a temperature of formation of the meltwater about 8°C lower than the present-day air temperature of about 0°C [*Dansgaard*, 1964]. However, simply translating the measured δ values into an air temperature relationship for the study area is speculative because the ice sheet that emplaced the glacial water did not originate at the site but originated rather farther to the north in the vicinity of Hudson Bay.

The δD values of −144‰ at depths greater than 105 m (in the clay) do not reflect connate waters because the water present in the western interior Cretaceous seaway had δD values of about −14‰ [*Kyser et al.*, 1993]. This suggested that the water in the clay entered the clay as postdepositional recharge, possibly during an interglacial period. The δD profile that has developed between the clay (about −144‰) and the till (about −178‰) was attributed to long-term diffusive mixing of deuterium in the clay and the till.

4.2. Conceptual Transport Model

Numerical simulations of the upper and lower δD profiles were conducted to provide an independent estimate of long-term groundwater flow velocity in the aquitards for comparison

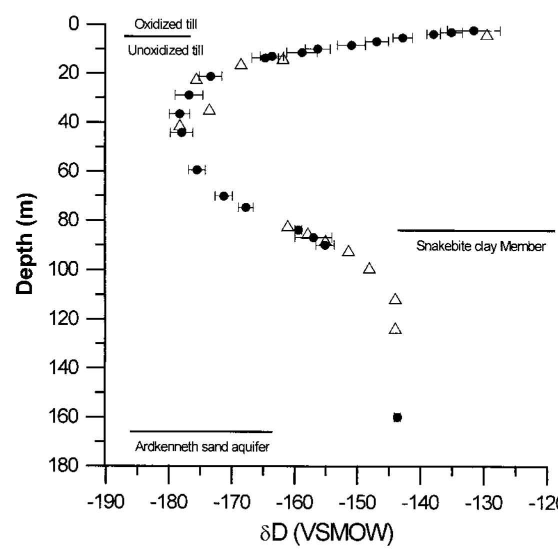

Figure 4. Pore water δD values versus depth through the entire aquitard system. The dots and bars represent the mean δD and standard deviation of seven water samples collected from 22 piezometers over a 4 year period. The triangles represent the pore water δD values determined using radial diffusion cells.

with velocity determined using conventional hydraulic data. These simulations also provided estimates of the time required to develop the δD profile across the till-clay boundary (and thus an estimate of the timing of till deposition) and the profile at the top of the unoxidized till (and thus an estimate of the timing of the start of the Holocene). The applicable transport equation for δD is

$$\frac{\partial C}{\partial t} = D\frac{\partial^2 C}{\partial z^2} - V\frac{\partial C}{\partial z} \tag{1}$$

where V is the average linear pore water velocity, D is the coefficient of hydrodynamic dispersion, C is the δD concentration in water, z is the depth, and t is time. In (1), $D = D_e + \alpha V$ and $V = v/n_e$, where D_e is the effective diffusion coefficient for δD, α is dispersivity, v is the Darcy velocity, and n_e is effective porosity. If V is small, D devolves to D_e. In our analysis we assumed that advection follows Darcy's law and that there is no threshold hydraulic gradient [*Neuzil*, 1986]. The validity of these assumptions is discussed in section 4.4. In addition, we also assumed that long-term groundwater flow was at or near steady state and that transport through the aquitard system was dominated by one-dimensional vertical flow.

In low K, relatively compressible, clay-rich sediments, such as the till and clay at the study site, the time to dissipate transient groundwater flow may require geologic timescales. This caused researchers to question the likelihood of steady state flow conditions in media with low hydraulic diffusivities [*Bredehoeft et al.*, 1983; *Neuzil and Pollock*, 1983; *Tóth and Millar*, 1983; *Neuzil*, 1986; *Belitz and Bredehoeft*, 1988; *Corbet and Bethke*, 1992]. *Shaw* [1998] estimated the time required to dissipate the pore water pressure disturbance caused by the last deglaciation in the till and clay at the study area. His simulations suggest that pressure disturbance decayed rapidly in the first 1500 years after deglaciation and was largely (95%) completed after about 1500 to 2700 years. Thus, with the exception of a relatively short period of time after deglaciation, we assumed that near steady state conditions existed in the till and clay since till deposition.

Assuming transport through the aquitard system was dominated by vertical transport allowed us to apply a one-dimensional vertical model to the aquitard system. This assumption was considered valid because groundwater flow is predominantly in the vertical direction as indicated by high vertical gradients between the water table and the underlying Ardkenneth aquifer compared with modest lateral head gradients in a large area around the study site [*van der Kamp and Jaworski*, 1989].

Table 1. Measured Effective Diffusion Coefficients for δD, Total Porosities, and Effective Porosities for δD From Core Samples

Sample Depth, m	Effective Diffusion Coefficient, m²/s	Total Porosity (n)	Effective Porosity (n_e)
	Unoxidized Till		
16	1.7×10^{-10}	0.29	0.29
22	1.7×10^{-10}	0.30	0.31
34	1.7×10^{-10}	0.31	0.33
	Clay		
83	1.7×10^{-10}	0.37	0.41
111	1.7×10^{-10}	0.38	0.40
123	1.7×10^{-10}	0.36	0.38

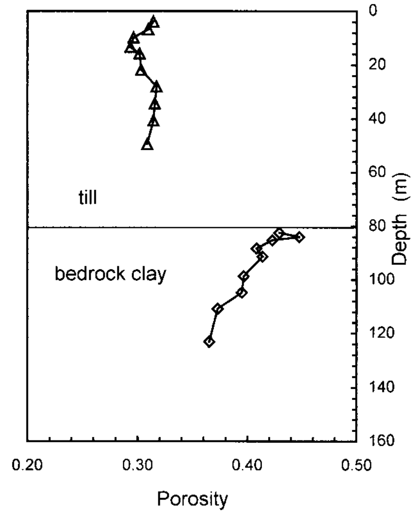

Figure 5. Distribution of porosity at the study site [after *Shaw and Hendry*, 1998].

4.3. Numerical Model and Initial and Boundary Conditions

The numerical model, POLLUTEv6 [*Rowe and Booker*, 1997], was used to solve (1). To determine advective velocities and timing of geologic and climatologic events in the aquitard system using (1), it was necessary to know D_e and n_e for δD in the till and the clay as well as the initial and boundary conditions in the till and clay. Results of D_e measurements for δD determined using the radial diffusion testing showed that D_e for the till and clay was 1.7×10^{-10} m² s⁻¹ (Table 1). This value was identical to that determined for $\delta^{18}O$ in an older, clay till aquitard near Saskatoon (1.7×10^{-10} m² s^{-1} [*van der Kamp et al.*, 1996]).

Results of the n_e measurements for δD and total porosity measurements determined using the radial diffusion testing are presented in Table 1. In general, n_e measurements were slightly greater (typically <5%) than the measured total porosity. This small systematic overestimation was attributed to analytical error in the determination of n_e. The overall good agreement between the total porosity and the n_e for the stable isotopes of water was similar to findings by *van der Kamp et al.* [1996]. As a result, n_e was assumed equal to total porosity (Figure 5), and V for δD was assumed equal to that computed using total porosity.

The vertical distribution of δD in the till and clay at the time of till deposition is unknown. Homogenization of the till during its deposition was assumed to have resulted in an initially uniform vertical δD profile throughout the till. We assumed the initial δD value in the till to be that measured value between 30 and 46 m (−178‰). The uniform distribution of several geotechnical parameters and K throughout the till [*Shaw and Hendry*, 1998] and the uniformity in the distribution of major, trace, and rare earth elements in the till solids [Yan et al., 1999] support this assumption. On the basis of consistent δD values below 105 m we assumed an initial δD of −144‰ throughout the clay at the time of till deposition. Review of geochemical and geotechnical data collected above and below the till-clay interface [*Shaw and Hendry*, 1998; *Yan et al.*, 1999] suggested that the interface is well defined.

Ideally, the transport of δD through the aquitard system should be modeled as a whole. Because of the large differences in the timing of the transport across the till-clay boundary and downward from the top of the till, we modeled the transport of δD in two separate simulations. First, the transport of δD across the clay-till interface was modeled. Second, the transport of δD from the oxidized till into the underlying unoxidized till was modeled. On the basis of the boundary conditions used in the simulations (described in section 4.4) we could have combined the two profiles using simple algebraic substitution. For ease of presentation, however, we chose to present the simulations individually.

In the first set of simulations the lower boundary was defined as the base of the clay (top of Ardkenneth aquifer) and defined by a fixed δD concentration of −144‰ in keeping with the uniform pore water values in the lower clay and the Ardkenneth aquifer. We assumed that an infinite thickness of till with a fixed δD concentration of −178‰ was rapidly deposited on the bedrock and that transport began after deposition. In the second set of simulations the upper boundary condition was defined at the contact between the oxidized and unoxidized till (4.0 m below ground surface) and set at the average present-day δD of precipitation (−136‰), and the lower boundary was of infinite extent. At the onset of these simulations the δD concentration in the till was −178‰. The use of a constant upper boundary condition was in keeping with the present knowledge of the retreat of the glacier about 12 ka B.P. [*Christiansen*, 1992], transition from cold climatic conditions of the late Pleistocene to warm climatic conditions of the Holocene 12 ka B.P.–10 ka B.P. [*Greenland Ice-Core Project Members*, 1993], and relatively stable climatic conditions throughout the Holocene [*Greenland Ice-Core Project Members*, 1993]. The best fits between the modeled profiles and measured data were determined using trial-and-error visualization.

4.4. Simulations of the Till-Clay Isotopic Profile

To encompass pore water velocities determined from head differences and measured K (0.5–0.8 m per 10 ka), we initially assumed the downward advective velocity ranged from 0.0 (diffusion only) to 2.5 m per 10 ka in increments of 0.5 m per 10 ka. Results of these transport simulations are presented in Figure 6. The simulated δD profiles yielded unique model curves representing a balance between downward advection and diffusion. The best fit model in Figure 6 with measured data was obtained using a simulated velocity of 1.0 m per 10 ka and for a time of transport of between 20 ka and 30 ka. On the basis of this best fit simulation the model was rerun with velocities of 0.75 and 1.25 m per 10 ka. Little difference was observed in the results for the 0.75 m per 10 ka and the 1.0 m per 10 ka fits. The fit was, however, less satisfactory for the 1.25 m per 10 ka simulation. This suggested that the groundwater velocity through the till-clay aquitard system is between 0.75 and 1.0 m per 10 ka over a transport time of 20 ka to 30 ka. Further refinement of the transport times was not possible with the existing data.

The best fit simulation across the till-clay boundary (i.e., 0.75–1.0 m per 10 ka between 20 and 30 ka) was considered significant for several reasons. First, on the basis of the assumptions used in the current study and the methods used to determine velocity from the hydraulic testing, we considered the estimated range in velocity determined from the isotopic model to be the same as that calculated from laboratory and field testing (0.5–0.8 m per 10 ka). This similarity suggested that the K of both aquitards is independent of volume tested, ranging from <100 cm^3 (volume of laboratory samples) to about 6×10^4 m^3 (volume of simulations). This observation is in keeping with the findings of *Neuzil* [1994], who concluded that scale dependence of permeability is not widespread in clayey geologic media and when present may affect flow at only the regional scale. Our estimate of velocity (and thus K) was in good agreement with the laboratory estimates of K. In contrast, others have concluded that conventional permeability testing often overestimates K and thus velocity [*Remenda et al.*, 1996; *Rudolph et al.*, 1991].

The similarity between the calculated velocity from the isotope model simulations and that determined from hydraulic testing suggested that the vertical hydraulic gradient across the aquitards have been fairly uniform for much of the time period in question, that Darcy's law is valid at very low velocities, and that there is no apparent threshold gradient effect at this site. Although *Neuzil* [1994] has shown that Darcy's law is valid at moderate to low hydraulic gradients, there is a lack of experimental data on the validity of Darcy's law in low-permeability media ($K < 10^{-9}$ m s^{-1}) under low hydraulic gradients such as those measured at this study site. In fact, the calculated downward velocities at this site were the lowest values that we could find in the literature.

Prior to our investigations, two tills were believed to exist in the study area: the upper 3 to 4 m of oxidized till was originally identified as Battleford Formation, whereas the remaining till was considered Floral Formation [*Christiansen*, 1986]. The time required for the development of the best fit isotopic profile suggested the till was deposited between 20 ka and 30 ka B.P. This time agrees with the current understanding of the timing of deposition of the Battleford Formation and not the Floral Formation. Radiocarbon ages on two samples of organic material deposited immediately below the Battleford Formation suggested that this till was deposited between 18 ka and 38 ka [*Christiansen*, 1971]. By contrast, the Floral Formation is estimated to be >128 ka B.P. [*SkaraWoolf*, 1981]. The strong indication that the till aquitard is Battleford and not Floral is also supported by measured preconsolidation pressures measured in the till to a depth of 50 m [*Shaw*, 1998]. Preconsolidation pressures ranged from 400 to 600 kPa, a range of values indicative of lower, basal meltout till from the Battleford Formation [*Sauer and Christiansen*, 1991] and much lower than reported values for Floral Formation (1700 to 2200 kPa) [*Sauer and Christiansen*, 1991; *Sauer et al.*, 1993]. It could be argued that the presence of the remnant, oxidized zone from 61 to 62.5 m below ground (Figure 2) defines the boundary between the Floral and Battleford Formations. The similarity in total carbonate content and grain-size distributions [*Shaw and Hendry*, 1998] and detailed rare earth element analyses of till samples [*Yan et al.*, 1999] above and below the oxidized zone suggested that the oxidized zone does not represent a break in the till sequence. The presence of the oxidized zone is attributed to redeposition during the last glaciation (E. A. Christiansen, personal communication, 1997).

4.5. Simulations of the Holocene Isotopic Profile

As in section 4.4, we assumed a range of long-term downward advective velocities ranging from 0.0 to 2.5 m per 10 ka for the upper Holocene isotopic simulations. Results of numerical modeling of the Holocene isotopic profile are presented in Figure 7. It can be argued that very good fits between the modeled profiles and the measured data could be obtained

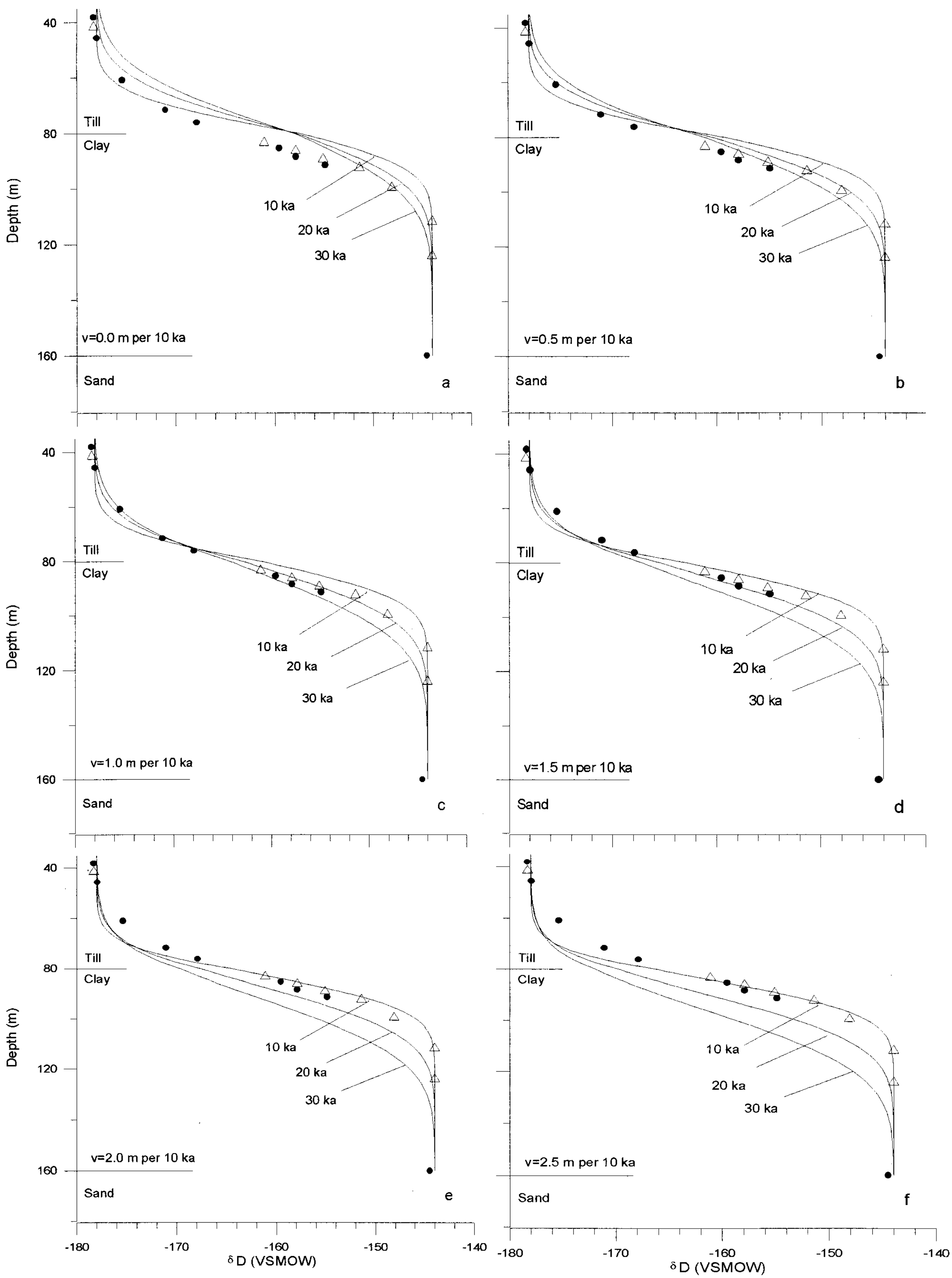

Figure 6. Pore water δD measurements between 35 and 175 m depth and the time required to evolve the observed profile by transporting pore water deuterium between the till (Battleford till) and clay (Snakebite Member). See Figure 4 for the definition of symbols.

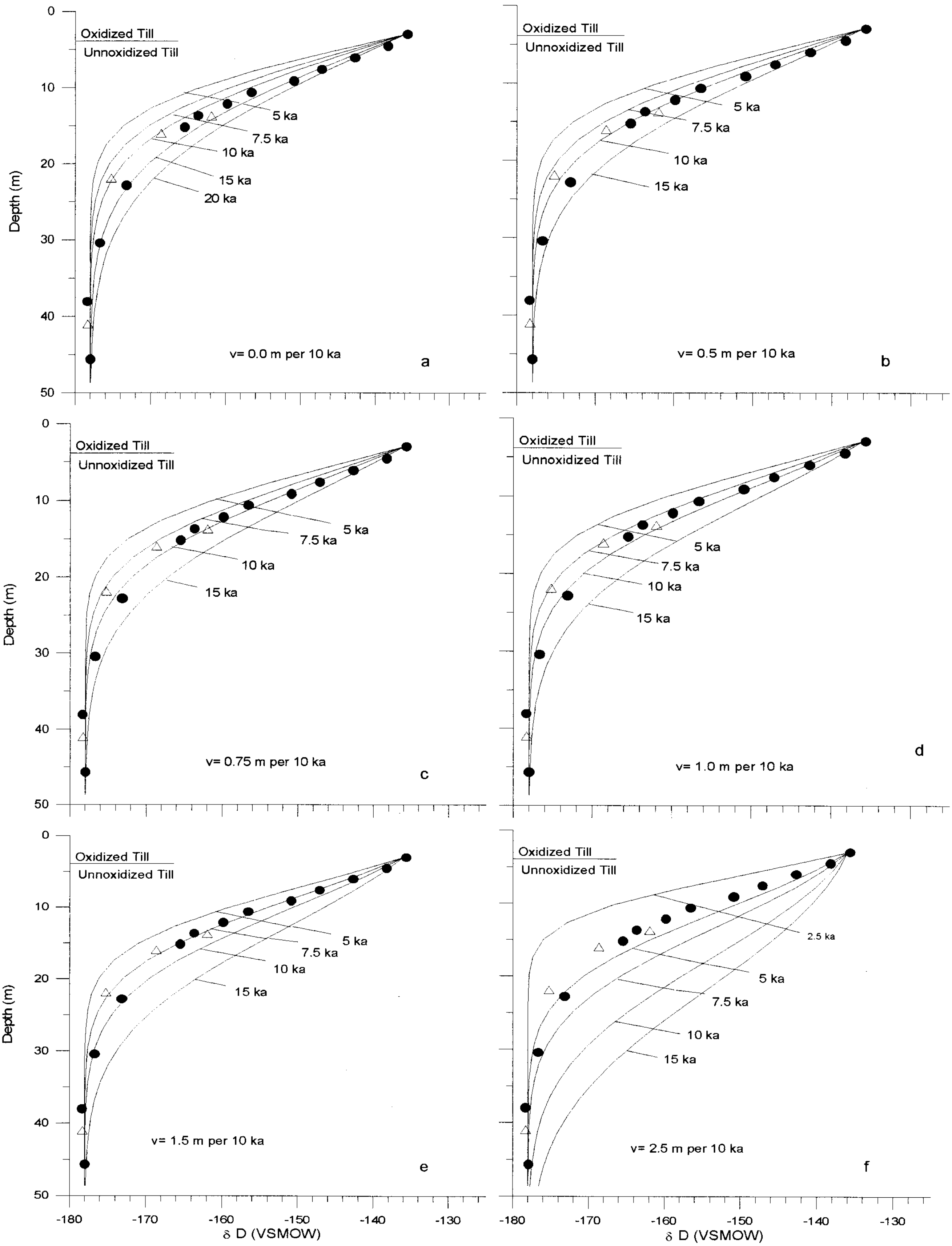

Figure 7. Pore water δD measurements in the upper 50 m of the profile and the time required to evolve the observed profile by transporting pore water deuterium downward into the till from the water table. See Figure 4 for the definition of symbols.

for all velocities tested. For example, using a velocity of 0.0 m per 10 ka the best fit time of transport was between 10 ka and 15 ka. Similarly, for a velocity of 2.5 m per 10 ka the best fit transport time was about 4 ka. These results indicated that there is a nonunique fit between the model results and the Holocene profile. The lack of a unique fit between the simulated isotopic profiles and the measured data was attributed to the fact that, unlike the till-clay profile, both diffusive and advective transport in the upper till are in the downward direction. The dominance of diffusive transport relative to advection in the transport of δD in the Holocene profile was shown using the dimensionless Peclet number (Pe) [*Love et al.*, 1995]:

$$Pe = (V^*z)/(D_e) \qquad (2)$$

where V ranges from 0.75 to 1.0 m per 10 ka, z is the thickness of the aquitard profile (~80 m), and D_e is 1.7×10^{-10} m^2 s^{-1}. For Pe values <5, diffusion is the dominant transport process, whereas for Pe values of >9 advection dominates. On the basis of (2), solute transport in the aquitard system is dominated by diffusion if $V < 3.4$ m per 10 ka, which is the case in the aquitard system.

Because of the nonunique fit between the modeling and the measured data, we could estimate the velocity by assuming the timing of the Holocene event or estimate the timing of the Holocene event by assuming the velocity. We assumed the long-term velocity determined from numerical modeling of the δD concentrations across the till-clay boundary (0.75–1.0 m per 10 ka) was applicable to the Holocene profile because it agreed with hydraulic data. Using these data, the best fit between the simulated profile and the measured Holocene profile was obtained for a transport time of between 7.5 ka and 10 ka (Figure 7). This compares favorably with the start of the Holocene (10 ka) [*Greenland Ice-Core Project Members*, 1993; *Sauchyn*, 1990].

In these model calculations we assumed that the isotope profile began at the base of the oxidized till and not at ground surface. Research suggests that the oxidation and fracturing of the till occurred after deglaciation and was caused by a lowering of the water table at the start of the Holocene [*Hendry et al.*, 1986; *van Stempvoort et al.*, 1994]. We attributed the difference between our estimated timing of the Holocene (7.5 ka–10 ka) and the start of the Holocene [*Greenland Ice-Core Project Members*, 1993; *Sauchyn*, 1990] to the time required to oxidize the upper 3 to 4 m of till in the study area. The length of time (up to 2.5 ka) to develop the oxidized zone from the unoxidized parent material is not unreasonable [*Hendry et al.*, 1986].

5. Summary and Conclusions

Detailed isotopic profiles of δD and $\delta^{18}O$ in pore water were obtained from a thick, tiered aquitard system located in Saskatchewan, Canada. The aquitard system consists of 80 m of clay-rich Pleistocene till unconformably overlying 76 m of clay of the Cretaceous Snakebite Member of the Bearpaw Formation (72 Ma–71 Ma B.P.). These deposits are underlain by a laterally extensive sand aquifer (22 m thick), the Ardkenneth Member of the Bearpaw Formation (74.5 Ma–72 Ma B.P.). The upper 3 to 4 m of the till is fractured and weathered. Below 3 to 4 m the till is massive, unoxidized (dark gray color), and nonfractured.

Numerical modeling of the transport of δD conducted using measured values for the effective diffusion coefficient for δD across the till-clay interface (i.e., downward advection and upward diffusion) yielded best fit simulated velocities of between 0.75 and 1.0 m per 10 ka for a transport time period of 20 ka to 30 ka. These results provided insight into the hydraulics of the aquitard system and on the timing of glaciation. First, the long-term groundwater velocity determined in this study was similar to that calculated from hydraulic testing (0.5–0.8 m per 10 ka), suggesting that the K of these aquitards are independent of volume tested. The close agreement between the calculated velocities from modeled isotopic profiles and hydraulics suggested that Darcy's law is valid at very small velocities in porous media and that there is no apparent threshold effect at these low velocities. Second, the estimated time required for the δD profile to develop across the lower till-clay interface (20 ka–30 ka) was indicative of the timing of till deposition and showed that the till is Battleford Formation and not a much older till as previously believed.

Numerical modeling of δD transport in the upper 30 m of the unoxidized till yielded a nonunique fit between the model results and the measured data. Assuming a groundwater velocity between 0.75 and 1.0 m per 10 ka, the best fit between the simulated profiles was obtained for transport times of 7.5 ka–10 ka. This age range compares favorably with that reported for the start of the Holocene (10 ka B.P.).

In this study we showed that the application of stable isotopes of water to the study of thick aquitard systems not only can provide independent, long-term estimates of very low groundwater velocities but can also provide insight into the timing of major geologic events such as glaciations. The study also showed that under the hydrogeological conditions at the site (which represent many areas in the interior Great Plains of North America), clay-rich till and clay bedrock may effectively contain contaminants from entering the biosphere for at least several tens of thousands of years.

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