

Stable isotope profiling for estimation of net percolation through a reclaimed overburden dump

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Abstract

Characterising the hydrology and hydrogeology of reclaimed overburden fills at oil sands mines is required to evaluate contaminant loadings to surface water receptors. Conventional monitoring of water balance does not provide an accurate estimate of one of the most critical parameters, net percolation (groundwater recharge). Here we describe a novel method to estimate recharge rates in a saline-sodic overburden fill at the Syncrude Canada Ltd. Mildred Lake Mine in the Athabasca Oil Sands region of Alberta, Canada. This site has been used to study cover performance and monitor the surface water balance for different configurations of reclamation cover over the past decade. Core samples of the cover soil and overburden fill were collected from nine locations across the site. The core samples were analysed in the laboratory for the stable isotopes, $\delta^2\text{H}$ and $\delta^{18}\text{O}$, in the pore water. A local meteoric water line (LMWL) was developed from precipitation samples collected at the site in 2009 and was used to assess whether evaporation had altered the isotopic signatures of the pore water. Stable isotope profiles were generated from the soil sample analysis results at each sampling location. The one dimensional $\delta^2\text{H}$ profiles were mathematically simulated to estimate the long-term net percolation rates at different topographic positions. The estimated net percolation rates ranged from 32–50 mm/yr for the plateau and mid-slope bench to 0–12 mm/yr for locations on the slope.

1 Introduction

Surface mining of oil sands from the Athabasca oil sands region of northern Alberta typically requires stripping of overburden in order to access the bitumen-rich McMurray Formation. A significant proportion of the stripped overburden is a cretaceous shale that is both saline and sodic. This shale overburden is commonly placed in large upland overburden piles. Successful reclamation of these landscapes has been achieved by designing structures with adequate grading to prevent ponding, surface water diversion features to handle runoff, and vegetated soil covers which are intended to support the required land use. Characterising the fate and release rate of salts from the overburden is of critical importance to sustaining vegetation and to the long-term water quality of shallow aquifers and surface water bodies.

Previous research from the study site has helped to elucidate many of the parameters needed to characterise the water balance and salt transport in the cover and shale. These parameters include hydraulic conductivity (Meiers et al., 2003, 2006), effective diffusion coefficient (Kessler, 2007), pyrite oxidation rate in the shale (Nichol et al., 2006; Wall, 2005), matrix and fracture porosity (Kelln et al., 2009), and interflow mass transport rates (Kelln et al., 2009). However, to date, attempts to measure or calculate net percolation into the shale have been crude or unsuccessful (Kelln et al., 2008, unpublished tests by Syncrude Canada Ltd. [SCL]).

A critical parameter in this characterisation is the portion of the infiltrating precipitation that passes through the cover and into the shale overburden (i.e. groundwater recharge). To meet this objective, a novel approach was applied: the measurement of soil-water stable isotope profiles from soil samples. These isotopic profiles were interpreted using a one-dimensional contaminant transport model to estimate net percolation rates.

The primary objective of the numerical modelling in this study was to derive a more accurate estimate of net percolation through the cover and into the shale overburden.

2 Background information

2.1 Site description

The study site is located on the plateau and north slope of South Bison Hills (SBH), a reclaimed upland overburden pile at the SCL Mildred Lake Mine. The mine is situated 40 km north of Fort McMurray, Alberta in the surface mining region of the Athabasca oils sands. SBH was constructed over the course of two decades with final grading of the dump undertaken between 1996 and early 1999 (Kessler, 2007). The reclamation cover at the study site was constructed over a north facing slope in the winter of 1999. Three different cover prescriptions were constructed with total thicknesses of 50 cm, 80 cm and 100 cm. However, the focus area of the current study was the portion of the study site with the 100 cm thick cover. This includes the plateau and the D3 cover plot as shown in Figure 1. The 100 cm thick cover comprised 80 cm of salvaged glacial till overlain by 20 cm of a peat-mineral soil mixture. The cover was seeded with barley in June 1999 to minimise erosion. Alternating rows of white spruce and aspen seedlings were planted in the fall of 1999. Over the past decade, researchers, primarily from the University of Saskatchewan (U of S), have installed instrumentation at the SBH study site to measure soil and climatic conditions. This instrumentation includes two weather stations, two soil stations, numerous standpipe piezometers and an interflow collection and monitoring system. The layout of the instrumentation is included in Figure 1.

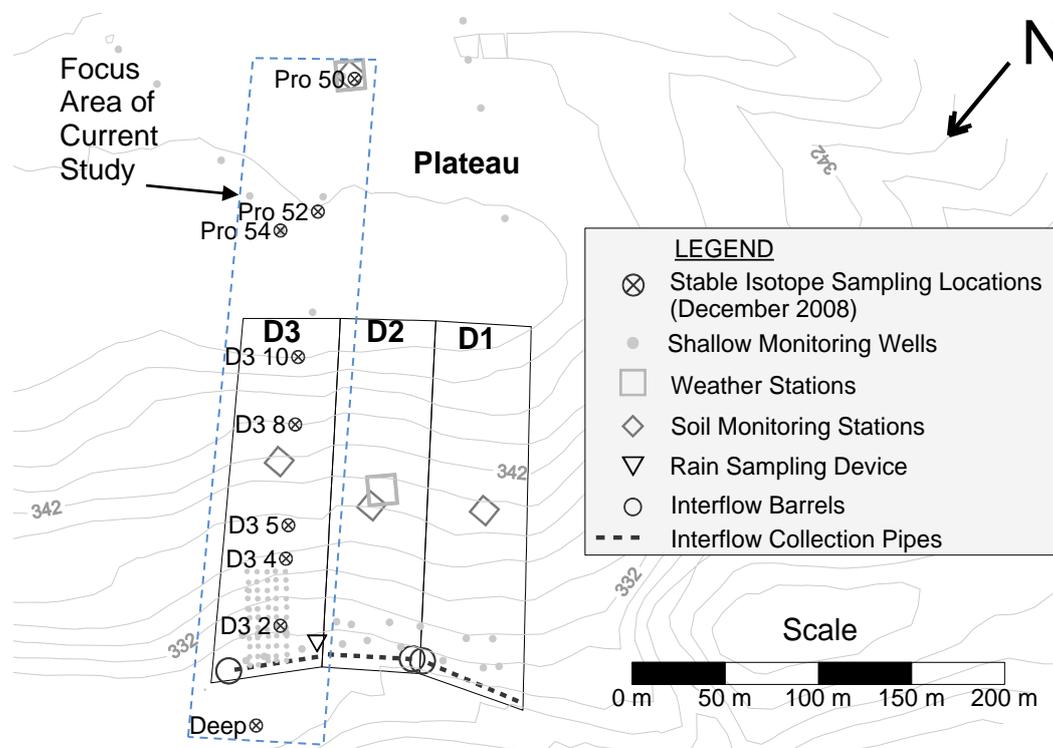


Figure 1 SBH study site map with instrumentation and sampling locations

2.2 Salt transport

The marine-origin shale contains naturally high concentrations of salts, dominated by sodium (Na^+) and sulphate (SO_4^{2-}). These ions are released into the pore water, through a series of geochemical reactions beginning with the oxidation of pyrite minerals (Wall, 2005). The salinity and sodicity of the shale overburden creates adverse conditions for the growth of most vegetation. The soil reclamation cover is intended to provide a suitable substrate for vegetative growth; however, previous research has shown that the salts from the underlying shale can be transported into the reclamation cover (Kessler, 2007; Merrill et al., 1983; Moran et al., 1990). Several mechanisms may be responsible for this contaminant transport into the cover soil. Molecular diffusion occurs immediately after cover placement, driven by the large salt concentration gradient created between the low concentration cover material and the high concentration shale overburden (Kessler, 2007). If the hydraulic conductivity (K) of the underlying waste material is sufficiently

low, molecular diffusion can be the dominant mechanism of upward salt transport (Merrill et al., 1983; Bailey, 2001). The other critical mechanism for salt transport from the underlying shale fill into the cover soil is via advection. Upward advective transport is generally a transient process created by periods of infiltration alternating with periods of evaporation or evapotranspiration during the summer months.

There are also two salt transport processes responsible for counteracting the upward diffusive and advective transport into the cover. Lateral interflow is gravity driven flow of groundwater that becomes perched along the interface of the cover soil and the underlying, low-permeability shale overburden (Kelln et al., 2007). This lateral flow of groundwater can carry with it salts that have accumulated above the cover-shale interface. The other salt flushing mechanism in the SBH soil cover is net percolation of precipitation that can transport salts downwards into the shale. These flushing mechanisms are critical in preventing the excessive accumulation of salts in the cover soil.

Although the fate and transport rate of salts is of critical importance to the successful reclamation of overburden piles, the use of salt profiles to characterise water dynamics in a cover soil system can be complicated by the non-conservative nature of the salt ions. This has led researchers to search for alternative conservative tracers. A method that has been used in aquitards for estimating groundwater flow rates and determining the mechanisms responsible for solute transport is the analysis of stable isotopes in pore water (Hendry and Wassenaar, 1999, 2004; Hendry et al., 2009, 2010; Desaulniers et al., 1986; Remenda et al., 1996; Simpkins and Bradbury, 1992). Stable isotopes are naturally occurring and are not subject to radioactive decay. The two primary stable isotopes of water are deuterium (^2H) and oxygen-18 (^{18}O). These stable isotopes are ideal tracers of water because they form part of water molecules that have essentially the same chemical properties as non-isotopic water molecules (Gat et al., 2001). Stable isotopes of water are conservative tracers in that they are not significantly affected by geochemical reactions at normal near-surface groundwater temperatures (Lawrence and Taylor, 1972; Remenda et al., 1996; Dowuona et al., 1993) and do not experience attenuation during transport. Stable isotopic profiles, therefore, can provide a better calibration tool for the 1D numerical transport model than dissolved ion profiles.

The concentration of stable isotopes from this study is presented as the relative difference between the sample and the internationally accepted standard, Vienna Standard Mean Ocean Water (VSMOW). The relative difference is signified by the ' δ ' symbol and is presented in units of per million (‰). The ' δ ' value is defined as (Gat et al., 2001):

$$\delta = \left(\frac{R_{\text{sample}} - R_{\text{standard}}}{R_{\text{standard}}} \right) \times 1000 \quad (1)$$

where R is the ratio of the abundance of the rare isotope to the dominant isotope.

3 Methodology

3.1 Field work

The field work component of this study was initiated in December 2008 and completed in the fall of 2009. For comparative purposes and trend identification, field measurements dating back to 2001 were also used. Figure 1 illustrates the layout of field instrumentation and sampling locations of importance for this study.

The collection of soil samples was undertaken in December 2008 using a small tracked auger drill rig. Three holes were drilled on the SBH plateau and five holes were drilled along a transect extending down the north facing slope. These eight sampling holes were drilled to a depth of approximately 3 m below ground surface. An additional hole was drilled below the access road at the base of the slope. The depth of this hole was 9.25 m below ground surface and was intended to provide background stable pore water isotope values for the shale at depth. The locations of the drill holes are illustrated in Figure 1.

Soil samples were collected by pushing a 35 mm ID split-spoon sampler into the soil over runs ranging from 0.2–0.6 m. Difficulties with sample recovery were encountered, primarily as a result of the frozen surface conditions and the heterogeneous nature of the cover, which has layers and pockets of media with varying soil stiffness and moisture content. Soft soils underlying hard or frozen soils were often compressed or not recovered. Efforts were made to minimise compression and core loss during drilling; however the potential

inaccuracy of assigned depths of samples is estimated to be up to ± 0.3 m in the zones of poor recovery. The recovery in the shale was generally better than the cover soil with an average recovery of approximately 70%. The range of error in sample depths for the shale is estimated to be as high as ± 0.1 m.

In addition to the soil samples, water samples from various sources were collected from the study site. One type of water sample collected was interflow. An interflow collection system was installed in the summer of 2000 by researchers from the U of S and SCL. The interflow collection system is a geomembrane-lined cut-off trench across the base of the slope with a 150 mm diameter, perforated, flexible polyethylene pipe placed at the bottom of the trench and backfilled with filter sand. The perforated pipes drain into three buried collection barrels. Collected interflow is pumped from the barrels using solar powered, automated submersible pumps. Cumulative flow rates are monitored at each interflow barrel using inline, turbine flow meters. Interflow water samples ($n = 56$) were collected in the spring and summer of 2009 and analysed for the stable isotopes, $\delta^2\text{H}$ and $\delta^{18}\text{O}$.

Shallow groundwater samples were also collected at the study site from a system of monitoring wells, installed between 2001 and 2006 by researchers from the U of S. The locations of these monitoring wells are shown in the site map in Figure 1. Groundwater samples ($n = 120$) were collected from the network of wells in May and August of 2009, and shipped to the laboratory for stable isotope analysis.

Rain water was collected using a plastic funnel located approximately 1.5 m above ground and affixed to a metal rod driven into the ground. The funnel drains precipitation into plastic sample bottles that were threaded into the bottom of the device. The precipitation collection device was located near the toe of the slope (Figure 1). Samples were collected less than 24 hours after precipitation events to minimise evaporation. Rain samples ($n = 26$) were collected after every rain event from May to September, 2009. Approximately 30 ml of water from each precipitation sample was poured into polyethylene bottles and shipped to the laboratory for stable isotope testing.

Snow samples ($n = 14$) were collected for $\delta^2\text{H}$ and $\delta^{18}\text{O}$ analysis on 30 March 2009. Eight of these samples were collected from the plateau and six from the slope. These samples were taken from cores through the full depth of the snow pack and, therefore, reflect the average stable isotope signature of the snowpack prior to snowmelt. Snow samples were stored in sealed plastic freezer bags and allowed to melt in the airtight bags in a refrigerator. Approximately 10 ml of water from each the melted snow samples was poured into polyethylene bottles and shipped to the laboratory for stable isotope testing.

3.2 Laboratory testing

Samples of snow, rain, interflow, and groundwater were analysed for the stable isotopes (^2H and ^{18}O). All water samples were stored in air tight containers in refrigerators or coolers from the day that they were sampled until the day they were delivered to the laboratory. Isotopic compositions of the water samples were measured at the National Water Research Institute (NWRI) in Saskatoon. This laboratory utilises an off-axis integrated cavity output spectroscopy laser (OA-ICOS) to directly measure absolute ratios of ^2H and ^{18}O in the liquid water molecules (Lis et al., 2007). The accuracy of the OA-ICOS laser for stable isotope testing of liquid water is 0.8‰ for $\delta^2\text{H}$ and 0.1‰ for $\delta^{18}\text{O}$ (Lis et al., 2007). Quality assurance of testing was ascertained by resubmitting 10% of samples as duplicates.

Isotopic analyses of soil samples was conducted at the U of S Aqueous Geochemistry Laboratory. The stable isotope composition of the soil pore water was measured using the $\text{H}_2\text{O}(\text{liquid})\text{--}\text{H}_2\text{O}(\text{vapour})$ equilibration laser spectroscopy method (Wassenaar et al., 2008). This direct equilibration method tests H_2O vapour that is isotopically equilibrated with the liquid pore water. Direct equilibration methods provide significant time and cost savings over methods that rely on physical extraction of pore water from the soil samples, and also avoid direct or indirect fractionation effects inherent in those methods (Koehler et al., 2000; Kelln et al., 2001; Wassenaar et al., 2008). Direct equilibration can also provide much faster results and better resolution than piezometer- or lysimeter-recovered pore water samples, especially in low permeability strata (Hendry et al., 2009).

Soil samples tested for this study were equilibrated for three days with atmospheric air in resealable, air-tight plastic bags. The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values of the water vapour were determined using the OA-ICOS technology.

The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values were calculated for the pore water using established fractionation factors between liquid water and water vapour.

3.3 Numerical modelling

One-dimensional contaminant transport models were constructed for the SBH site using the Geoslope® finite element software package Seep/W (for water flow) coupled with CTran/W (for contaminant transport).

The conceptual model, illustrated in Figure 2(a), was based on earlier modelling work by Kelln (2008) and Nichol et al. (2006). The soil cover is represented as a single 1 m layer overlying 5 m of shale. This total depth was judged to be a sufficient to ensure that the lower boundary conditions would not influence the model results.

Water ingress to the model occurs by two processes: interflow and infiltration. Infiltration can be either rain water or melted snow that percolates down through the cover. Some of this infiltrating water may contribute to lateral interflow and some of it may be removed from the model domain as evapotranspiration. The remainder of the infiltrating water that passes through the cover/shale interface is termed “net percolation”.

One-dimensional advective transport in the model was simulated by the following equation:

$$J_q = qC \quad (2)$$

where J_q is the mass flux due to advective contaminant transport ($\text{g}/\text{m}^2/\text{s}$), q is the specific discharge (m/s), and C is the concentration of contaminant in pore water (g/m^3).

Hydrodynamic dispersion of the isotopes by diffusive and dispersive transport was represented by Fick’s First Law as follows (Shackelford, 1991):

$$J_d = -\theta_e D_h \frac{\partial C}{\partial x} \quad (3)$$

where J_d is the mass flux transport by hydrodynamic dispersion ($\text{g}/\text{m}^2/\text{s}$), θ_e is the effective volumetric water content, $\delta C/\delta x$ is the concentration gradient ($\text{g}/\text{m}^3/\text{m}$), and D_h is the coefficient of hydrodynamic dispersion (m^2/s). D_h is defined as follows:

$$D_h = D_o\tau + \alpha v \quad (4)$$

where D_o is the free solution molecular diffusion coefficient for a particular solute at a particular temperature (m^2/s), τ is the tortuosity factor (unitless), α is the dispersivity (m), and v is the advective velocity (m/s) and is equal to q/θ . The first term in the above equation ($D_o\tau$) is referred to as the effective diffusion coefficient for the porous media, D_e .

The conceptual model for stable isotope transport is simpler than that for salts as illustrated in Figure 2(b). Rather than trying to accurately simulate the dynamics of all of the advective processes (infiltration, evapotranspiration, percolation), it was determined that a net percolation variable could account for all of these processes. The origin for this net percolation flow was set at the till-shale interface since it is the net percolation into the shale that is of interest in this study.

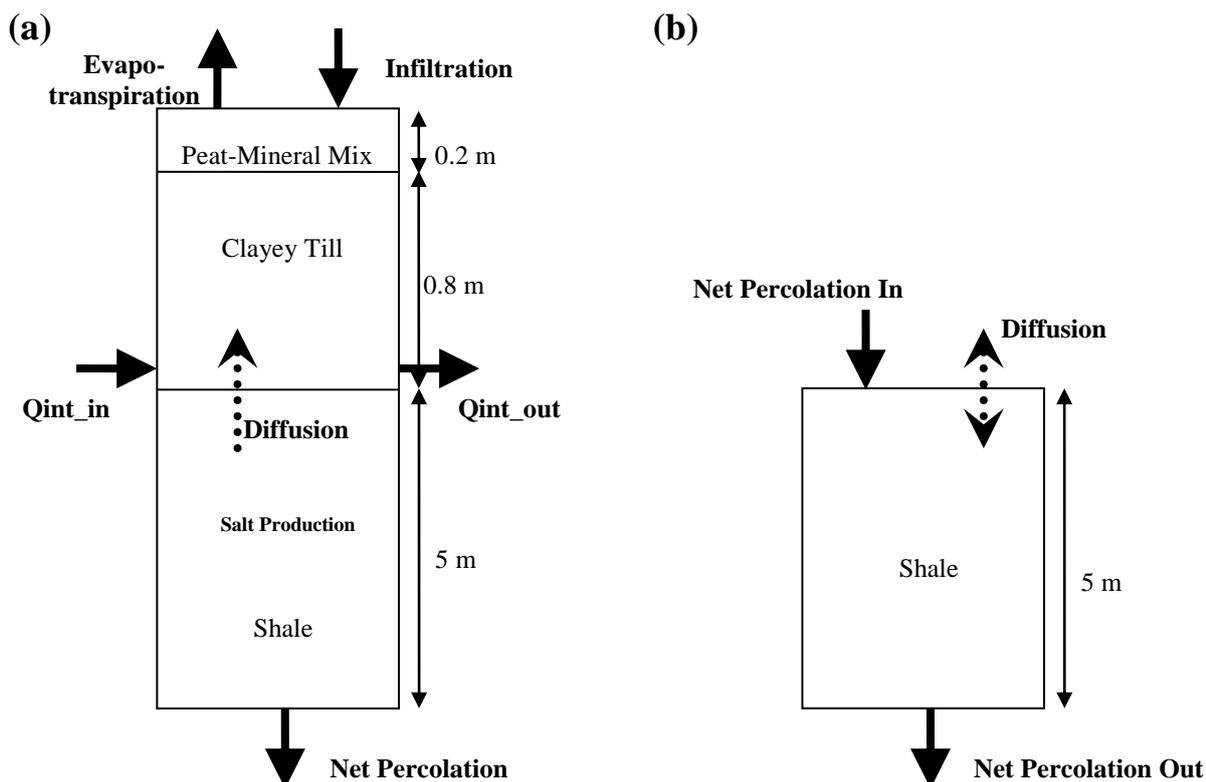


Figure 2 (a) Conceptual model for salt and moisture transport (adapted from Nichol et al., 2006); (b) Simplified stable isotope transport model used in this study

The use of a net percolation variable would not be appropriate when seasonal variations cause reversals of flow direction and cycling of solutes. This would apply to the cover soil, which experiences downward percolation during the spring snow melt and summer rains, and upward advection due to transpiration and evaporation during the summer growing season. However, it was assumed that the shale, with a low K and located at a depth of 1 m below ground surface, would not experience significant upward water flux during the summer growing season. This assumption was supported by previous studies that showed that a 1 m thick soil cover provides much more available water than is required by the vegetation over the growing season (Shurniak, 2003). As a consequence, the use of an annual average net percolation rate was deemed appropriate for modelling the shale.

The volumetric water content of the shale in the model was assumed to be constant year-round. A value of 0.4 was chosen based on recent volumetric water content readings from the SBH site and previous 1D transport modelling (Kelln, 2008). The use of a fixed net percolation variable and constant volumetric water content for the soils meant that the value of K and the pressure gradients had no bearing on the model results. This simplification allowed the K to be nominally set at a very high value (e.g. 1 m/s) with the boundary condition determining the flow through the model.

The value of D_e was varied, along with the net percolation rate, to fit the field measured profiles. An initial estimate of D_e was selected based on the values from published studies where the soil type was similar to the overburden shale. Kessler (2007) conducted diffusion experiments on till/shale half cells using site material and typical volumetric water contents (approximately 0.3). This work yielded a mean D_e for total salinity of 6×10^{-11} m²/s. Other measured D_e values from the literature include 1.7×10^{-10} m²/s for $\delta^2\text{H}$ in a Cretaceous shale with $n = 0.4$ (Hendry and Wassenaar, 1999) and 2.2×10^{-10} m²/s for $\delta^2\text{H}$ in shale with $n = 0.37$ (Hendry et al., 2010). A diffusion coefficient of 1.7×10^{-10} m²/s was also measured for $\delta^2\text{H}$ in clay till samples with an effective porosity of approximately 0.3 by Hendry and Wassenaar (1999) and van der Kamp et al. (1996). Shurbaji and Phillips (1995) suggested a higher diffusion coefficient of 7×10^{-10} m²/s for $\delta^2\text{H}$ transport in a saturated soil with $n = 0.35$. The D_e values used in our model were generally kept within the range of D_e

values established by these previous studies. However, D_e values were modified slightly between model runs to improve the fit to field data.

The mechanical dispersivity was not quantified for the conditions at the site. This value is difficult to quantify as it has been shown to vary with the scale of measurement (Pickens and Grisak, 1981; Gelhar et al., 1992). For the purposes of this modelling, α was assumed to be 0.01 m. This is roughly 1/100 of the scale of the simulated plume extent, consistent with the findings of other researchers (Pickens and Grisak, 1981; Gelhar et al., 1992).

The final important parameters in the $\delta^2\text{H}$ transport model were the background and source values. No previous isotopic profiles, pre-construction or otherwise, were collected from the study site to use in establishing baseline concentrations. However, during the 2008 drilling program a single deep hole was drilled into the shale (Figure 1) to provide a reasonable estimate of baseline conditions. Based on the isotopic profile of the deep hole a baseline concentration was established for the shale.

Unlike the shale, determining a source concentration for the net percolation water that migrates from the cover soil into the shale was difficult. Stable isotope analysis of snow and rain samples from the SBH site verified that the $\delta^2\text{H}$ values of waters contributing to net percolation vary over the year (data presented below). This variation is evident in the isotopic profiles within the cover soil (upper 1 m) (data not presented). Numerical modelling was conducted to understand how fluctuations in the isotopic content in the percolating water could affect the isotopic profiles in the shale. However, in the end, the source concentration for net percolation was based on the concentrations measured in samples taken from the interface at specific sites.

Interflow was not incorporated into the stable isotope transport model. This does not diminish the importance of interflow in defining the stable isotope profiles. In fact, the interflow water flowing down slope along the till-shale interface likely has the greatest impact on the source concentration of net percolation. However, by assigning the upper boundary as the till-shale interface, the stable isotope content of the interflow water is accounted for in the source concentration boundary condition in the model. The rate of interflow is not important in the stable isotope model because the interflow occurs above the model domain.

4 Results

4.1 Stable isotope analysis results

The results of the stable isotope analysis for water samples collected in 2009 are presented in Figure 3. A linear best-fit line was interpolated from the precipitation data to estimate of the local meteoric water line (LMWL) for the site. Ideally, the LMWL should be based on data collected over a longer period; however to the author's knowledge, such data do not exist in the Athabasca oil sands region.

The equation of the LMWL based on the 2009 rain ($n = 26$) and snow ($n = 14$) samples is as follows:

$$\delta^2\text{H} = 6.65 \delta^{18}\text{O} - 29.8 \quad (R^2 = 0.985, n = 40) \quad (5)$$

The surface water samples diverge slightly from the LMWL, suggesting that evaporation may have caused some isotopic fractionation. Conversely, the shallow groundwater and interflow water samples show no indications of fractionation and, with the exception of three samples, fall within a narrow range in values (e.g. $\delta^2\text{H}$ of -139 to -166‰).

The results of the stable isotope analysis for the soil samples are plotted in Figure 4 along with the LMWL to assess if fractionation effects are evident. Figure 4 shows that the majority of the soil samples plot close to the LMWL. A small percentage of the shale samples plot outside of the 95% prediction intervals for the LMWL. This may indicate that the pore water in these few samples has undergone some evaporation. However, the proximity of the plotted points to the LMWL suggests that the pore water in the cover soil and shale has not experienced fractionation by evaporation, sublimation, snow melt or frost front advance.

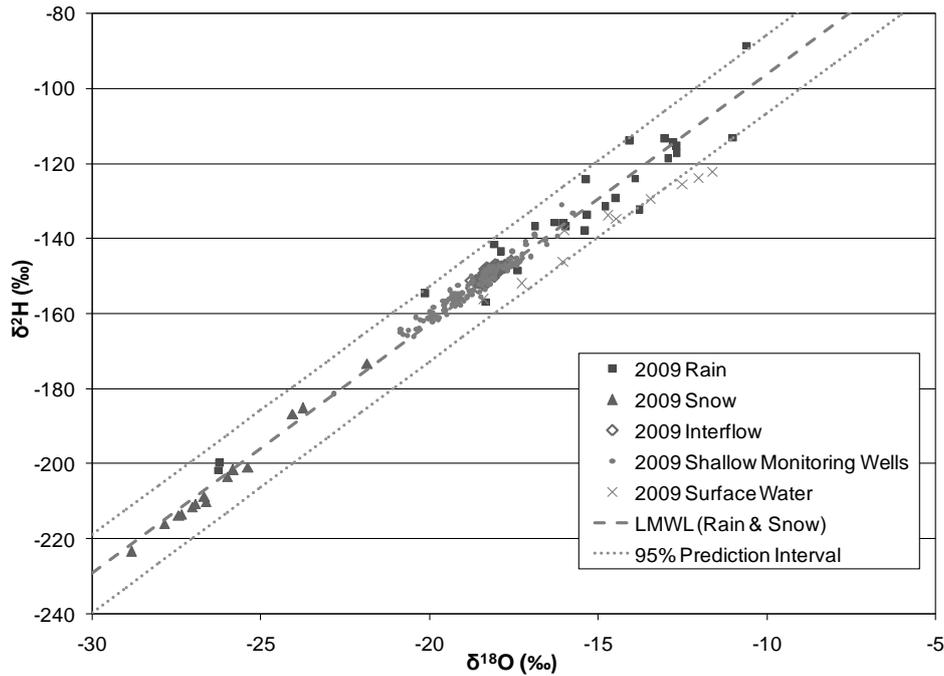


Figure 3 $\delta^2\text{H}$ and $\delta^{18}\text{O}$ cross plot for water samples collected in 2009

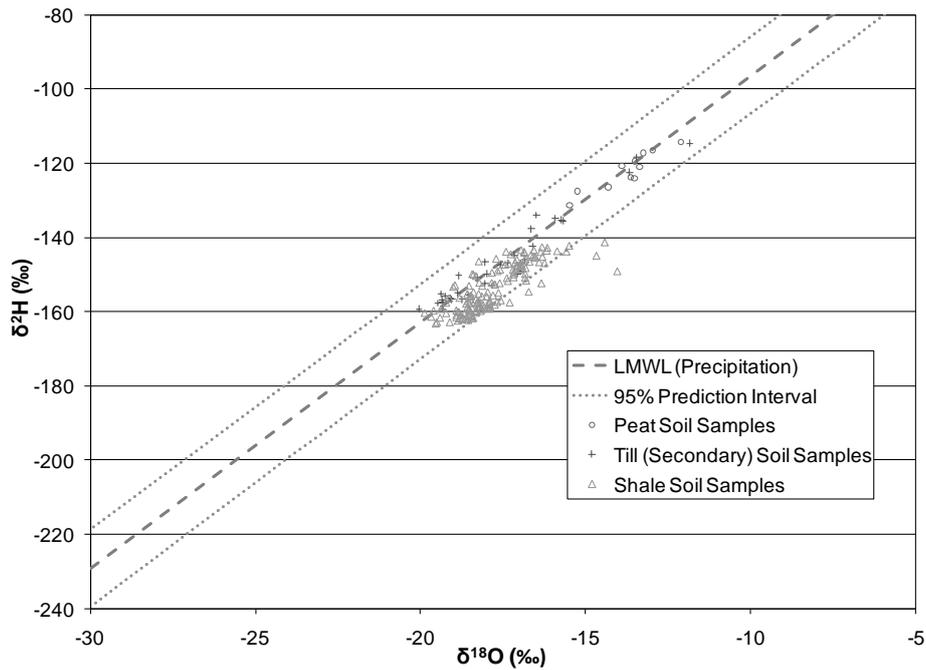


Figure 4 $\delta^2\text{H}$ and $\delta^{18}\text{O}$ cross plot for soil samples collected in 2009

4.2 Numerical model results for plateau locations

Measured and simulated isotopic profiles from the plateau sampling locations (Pro 50, Pro 52, and Pro 54; Figure 1) are plotted in Figure 5. The three measured profiles were similar in shape and thus were considered collectively. Two simulated profiles that bracket the measured values are also presented in Figure 5.

A background $\delta^2\text{H}$ value of -160‰ was used for the shale based on the deepest measured values at the plateau sampling locations and from the deep hole (data not presented). A source $\delta^2\text{H}$ value of -144‰ at the cover soil–shale interface was used for the plateau model, based on the measured values. The net percolation

rate and molecular diffusion coefficient were adjusted in the model to match the shape of the field profiles using a visual interpretation of goodness of fit. The range of net percolation rates that bracket the measured data was 35–50 mm/yr. The D_e value that provided the best fit to the measured data was $8 \times 10^{-11} \text{ m}^2/\text{s}$. This range of net percolation rates and D_e value result in a range of D_h values from 1.1×10^{-10} to $1.2 \times 10^{-10} \text{ m}^2/\text{s}$.

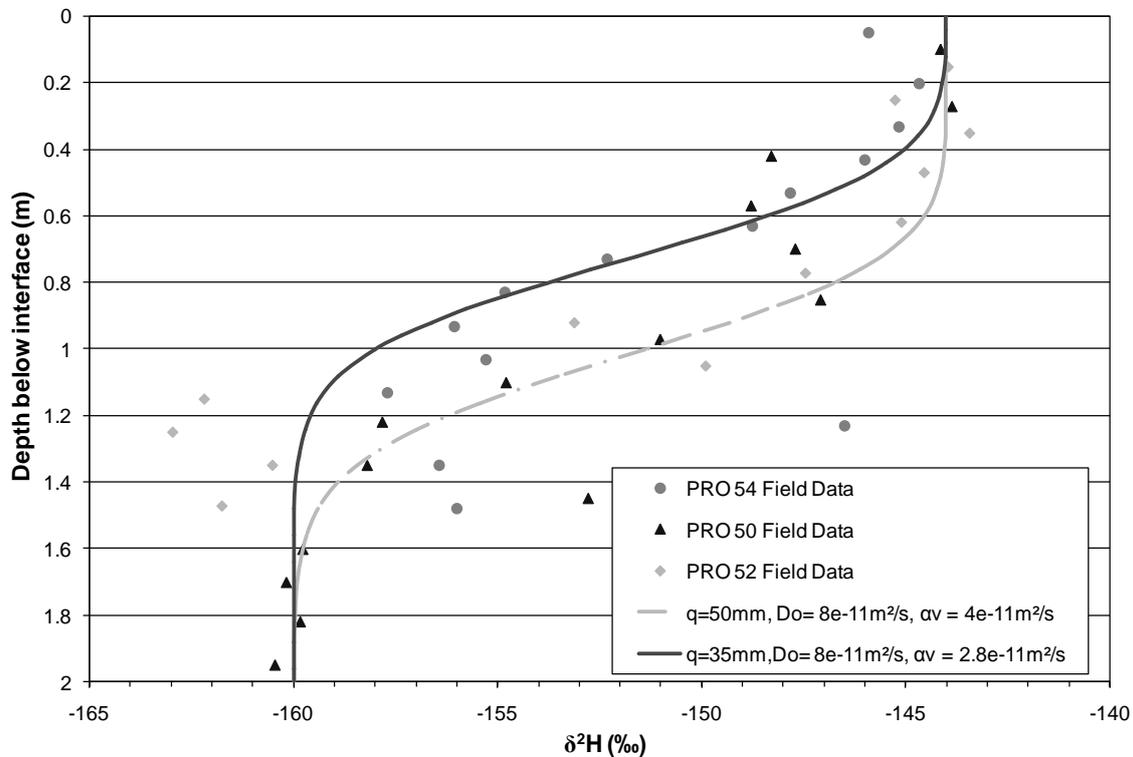


Figure 5 Measured and simulated $\delta^2\text{H}$ profiles for sampling locations on the plateau

4.3 Numerical model results for slope locations

The isotopic profiles along the slope are plotted in Figure 6. Unlike the sites on the plateau, the shapes of these profiles are not consistent and, thus, required independent assessment. The profiles can, however, be divided into two groups; Group A and B. The profiles in Group A, on the left side of Figure 6, are more depleted in $\delta^2\text{H}$ than the profiles of Group B. The Group B profiles have an enrichment bulge extending to a depth of up to 2 m below the cover-shale interface. These enrichment bulges are difficult to explain but may be related to the fact that at each of these three locations, till or lean oil sand was incorporated into the shale. No significant amounts of till or lean oil sand were observed in the shale at the other locations.

Simulated profiles for the sample locations along the slope are also included in Figure 6. Some of these sampling locations yielded more than one simulated profile that provided a good fit to the measured data; however, for clarity the simulated profiles were limited to one per sampling location. To obtain an acceptable fit to the measured data, we assumed that the presence of till or lean oil sand in the shale of the Group B locations enriched the initial $\delta^2\text{H}$ value. The degree to which the initial $\delta^2\text{H}$ value was increased was based on qualifying descriptors in the drill hole logs. The details of the simulated profiles for the sample locations along the slope are summarised in Table 1.

It is noted that the estimated net percolation rates for the slope locations are consistent at 8–12 mm/yr, with the exception of D3-05, which is located on a broad bench at midslope. These findings suggest that the bench provides a greater opportunity for infiltration and percolation, with reduced runoff and interflow. This would lead to conditions similar to the plateau, which might explain the similarity in net percolation rates. The value of D_e is also quite consistent, with most simulations varying from 5×10^{-11} to $1 \times 10^{-10} \text{ m}^2/\text{s}$.

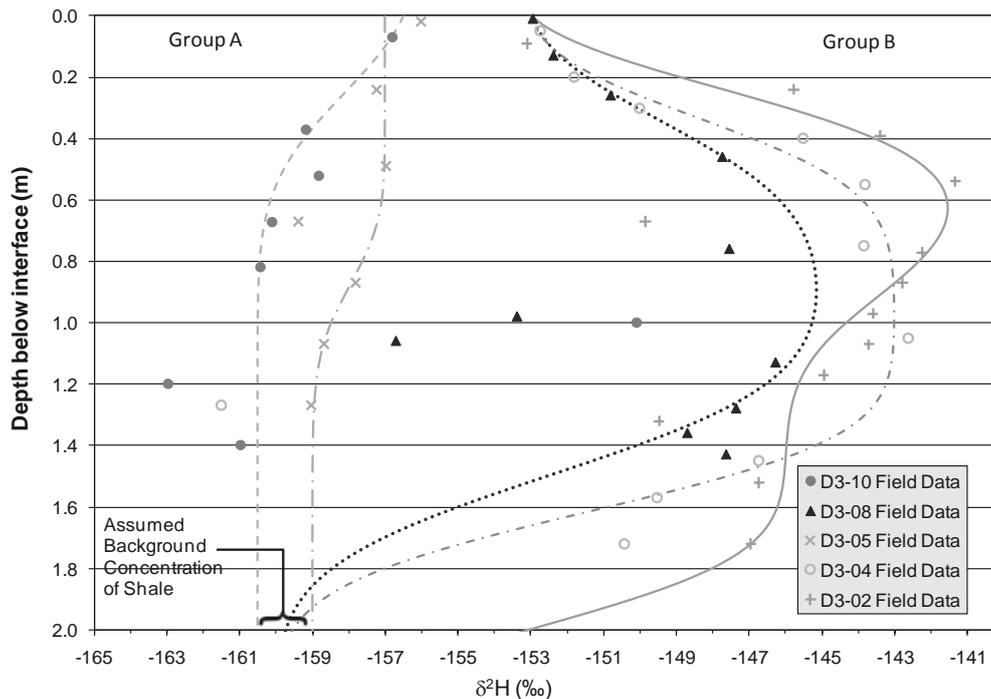


Figure 6 Measured and simulated $\delta^2\text{H}$ profiles for sampling locations along the slope

Table 1 Summary of estimated transport parameters from ^2H transport model

Sample Location	Net Percolation (mm/yr)	D_e (m^2/s)	αv (m^2/s)	D_h (m^2/s)
D3-02 – flat, toe of slope	8	5×10^{-11}	6×10^{-12}	6×10^{-11}
D3-04 – steep, mid-slope	12	5×10^{-11}	1×10^{-11}	6×10^{-11}
D3-05 – mid-slope bench ²	32 to 35	2×10^{-11} to 4×10^{-10}	3×10^{-11}	5×10^{-11} to 4.3×10^{-10}
D3-08 – slight bench near upper slope ²	0 to 12	8×10^{-11} to 1.5×10^{-10}	0 to 1×10^{-11}	9×10^{-11} to 1.5×10^{-10}
D3-10 – steep, upper slope	8	1×10^{-10}	6×10^{-12}	1.1×10^{-10}

Notes: (1) The value of αv varies only because the net percolation rate, and thus, velocity term has been varied. The value of α in the model simulations is fixed at 0.01 m. (2) Locations D3-05 and D3-08 each had two simulations that appeared to be equally good fits.

4.4 Evaluation of model results

A simple Darcian flux calculation was applied to determine if these net percolation rates were reasonable. The estimated net percolation from the plateau is 35–50 mm/yr or 1.1×10^{-9} to 1.6×10^{-9} m/s. A perched water table within the cover would create a vertical gradient within the shale of unity, which when applied over a typical period for which these conditions are observed to exist, would require a K of approximately 4.5×10^{-9} to 6×10^{-9} m/s to permit an equivalent 35–50 mm/yr of net percolation. Net percolation rates lower than 35 mm/yr in Table 5.1 (i.e. at slope locations) might suggest lower hydraulic conductivities, shorter ponding periods, or possibly both. For example 12 mm/yr at a K of 4.5×10^{-9} m/s would require approximately 31 days of ponded water. The presence of ponded water on the SBH study site slopes is typically 30 to 60 days as reported by Kelln et al. (2009) and supported by interflow measurement data.

The estimated range of K values above is reasonable based on the field measurements by Meiers (2002) who measured an average K of the upper shale (maximum depth of 1.7 m below ground surface) of 5×10^{-9} m/s in 2000. These estimated K values also compare well to estimates by Kelln et al. (2008). These authors observed the water table recession in ponded water within pockets of lean oil sand within the upper shale

following spring melt. By measuring the rate of drop of the perched water tables, Kelln et al. (2008) estimated that the saturated K for the shale was 3×10^{-9} m/s.

The estimated values of D_e are similar for the plateau and slope locations. Most values fall between 5×10^{-11} and 1.5×10^{-10} m²/s. These values are comparable but somewhat lower than diffusion coefficients published by other authors for similar conditions. The diffusion coefficients from the literature vary from 1.7×10^{-10} to 4×10^{-10} m²/s for diffusion of $\delta^2\text{H}$ in clayey soil with a saturated porosity of 0.4. The lower diffusion coefficient values calculated in this study are reasonable given that they are average values applied over the entire year in the model. Desaturation of even the upper few centimetres of shale overburden in the profile would greatly reduce the effective diffusion coefficient over dry periods, thus reducing the average diffusion coefficient for the year. Soil moisture content data collected from access tubes across the study site suggest that in many locations, the upper shale may indeed be desaturating seasonally.

5 Summary and conclusions

The analysis results from the 1D numerical transport model for deuterium demonstrate a notable difference in the estimated net percolation rates between the plateau region of the SBH study site and the sloped region of the study site. With the exception of the D3-05 sampling location, the net percolation rates estimated for the slope are considerably lower than those from the plateau. The range of net percolation rates calculated for the plateau is 35–50 mm/yr. The range of net percolation rate for D3-05 is only slightly lower at 32–35 mm/yr. The range of net percolation rates for the other slope locations is 0–12 mm/yr. Estimated values of D_e are similar for plateau and slope locations and are generally within a range of 5×10^{-11} to 1.5×10^{-10} m²/s.

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