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Core Ideas

Soil water stable isotope profiles can be used to quantify groundwater recharge
Climate conditions are recorded in water stable isotope profiles of the unsaturated zone
Soil water isotope profiles provide insight into the seasonality of recharge events

Abstract

A reliable estimate of recharge is needed for the sustainable management of groundwater resources. Water stable isotope ($\delta^{18}O$ and $\delta^{2}H$) profiles in the unsaturated zone are frequently used to quantify groundwater recharge, based on the seasonality of water isotopic compositions in precipitation. A very simple approach consists of integrating soil water content between peak values of soil water isotopic compositions, typically corresponding to precipitation signatures from warm and cold seasons. When precipitation isotopic compositions are available, a conceptual surface water isotopes budget and lumped parameter dispersion model can be computed. These models were applied on two field sites with similar permeable soils with grass cover but contrasting recharge regimes and seasonality: one in the Paris Basin (France) with continuous recharge from autumn to spring, and another in the St. Lawrence Lowlands (Quebec, Canada) with episodic recharge in fall and after snowmelt. For the two sites, the peak to peak method and isotope surface budget led to comparable recharge intensities. At least at the Paris Basin site, evaporation is shown to slightly modify the average unsaturated zone and hence groundwater isotope composition. The proposed parameterization of isotope fractionation due to evaporation allows qualitative estimation of the fraction of evaporation, at least during the recharge seasons. In spite of its simplifications and limitations, the proposed parsimonious model can give estimates of recharge in a variety of sites

even if they are not well characterized, as it benefits from the large availability of monthly isotopic compositions in precipitation.

1. Introduction

The UNESCO World Water Assessment Program identified a gap in knowledge on the impact of climate change on groundwater resources. While numerous recent studies have attempted to bridge this gap, they have also demonstrated that we currently do not have the ability to quantitatively predict the impact of climate change on groundwater resources with a satisfying degree of confidence (Crosbie et al., 2013, Kurylyk and MacQuarrie, 2013). Recharge is a key parameter in groundwater management, and thus a reliable estimate of recharge is necessary for sustainable groundwater resource development (Rivard et al., 2014). To achieve accurate water budgets, recharge quantification has been investigated at local scales using lysimeters (Pfletschinger et al., 2012, Xu and Chen, 2005), water-table fluctuations (Hagedorn et al., 2011, Healy and Cook, 2002, Liang and Zhang, 2012), and baseflow analysis (McCallum et al., 2014). Independently of the method, the impact of all possible environmental stresses, meteorological variability, and changes in climate, land use, and water extraction on the recharge regime cannot be completely taken into account (Kurylyk and MacQuarrie, 2013).

Variability in the stable isotope signature of soil pore water collected at depths with suction lysimeters has been used to investigate subsurface processes (Darling and Bath, 1988, Stumpp et al., 2009a, 2009b, Thomas et al., 2013). As an alternative method, water stable isotopes in the unsaturated zone offer a time-integrated fingerprint of recharge and of subsurface water pathways (Darling and Bath, 1988, Bengtsson et al., 1987, Koeniger et al., 2016). Unsaturated zone stable isotope profiles under temperate climate conditions have thus been used to study groundwater recharge mechanisms (Lee et al., 2007, Li et al., 2007, McConville et al., 2001, Mueller et al., 2014, Saxena, 1984, Song et al., 2009, Stumpp and Hendry, 2012, 2009a, 2009b). Suction lysimeters are quite time consuming to employ, with a sampling effort of at least one year, while cryogenic extraction of soil water allows the collection of comparable stable isotope data within one or a few field work days. New, faster methods, such as direct equilibration, are also becoming widely used, and raised issues regarding the possible biases between the various methods of soil water isotope analysis employed (Orlowski et al., 2016).

The isotopic composition of precipitation is strongly correlated with air temperature. A distinct seasonal pattern therefore occurs in precipitation under temperate climates, with summer rainfall enriched in heavy isotopes and winter rainfall depleted (Rozanski et al., 1993). In addition to $\delta^{18}O$ and $\delta^{2}H$, lc-excess values (Landwehr and Coplen, 2004) are very sensitive to local evaporation and associated isotope fractionation occurring at the surface. These are therefore used in addition to $\delta^{18}O$ and $\delta^{2}H$ to evaluate recharge processes and apparent mean residence times of soil water (Lee et al., 2007).

Even if transpiration predominates under temperate climates and is considered to not fractionate water stable isotopes, evaporation from plant interception or from the upper soil leads to fractionation and enrichment in heavy isotopes (Braud et al., 2005, 2009a, Sprenger et al., 2016, Sutanto et al., 2012). Water recharging during a given season can therefore be identified in deep unsaturated zone profiles, and recharge rates can be obtained from the displacement between successive seasonal inputs (Gehrels et al., 1998, Maloszewski et al., 2006, Maloszewski and Zuber, 1993, McConville et al., 2001, Saxena, 1984). On the one hand, this simple peak-shift method is easily applied (Adomako et al., 2010), even if isotopic compositions of precipitation are not available (Gehrels et al., 1998). On the other hand, numerical models, such as SiSPAT-Isotopes (Braud et al 2005, 2009b), Soil-Litter-Iso (Haverd et al., 2010) or a modified version of HYDRUS-1D (Stumpp et al., 2009a, 2009b, 2012), are increasingly being used to understand the dynamics of water stable isotopic composition in the unsaturated zone (Sprenger et al., 2015). Such level of model refinement is associated with high data requirement that can only be achieved for specific study sites, but less for more numerous and less known sites at the regional scale where there is a lack on recharge rates estimation.

The aim of this study is to apply a simple and parsimonious method to quantify the rate and seasonal pattern of recharge, based on water stable isotope depth profiles, applicable to sites under contrasted climate conditions, and requiring limited amount of additional data. First, the two study sites are presented, in the Paris Basin (France) and in the St. Lawrence Lowlands (Quebec, Canada), including water stable isotopic compositions of both precipitation and the unsaturated zone for each. Recharge is then calculated based on a simple peak-shift approach. A model coupling a surface water and isotope budget to a lumped parameter model is proposed for a more robust quantification of recharge, but also to explain the evolution of water isotopic composition in the unsaturated zone, and particularly to discuss the influence of evaporation.

2 Materials and methods

2.1 Study sites and sampling

2.1.1 The Paris Basin site (PB)

The Paris Basin site (PB) is located 35 km south of Paris (France), in the very fine and well-sorted sands of the Oligocene Fontainebleau aquifer. This regional unconfined aquifer has a maximum thickness of 50-70 m. The total porosity of the aquifer is 25-40%, with volumetric water content in the unsaturated zone in the range of 7-28 vol% (Schneider, 2005). The hydraulic conductivity reaches 1.10^{-5} to 6.10^{-5} m/s (Corcho Alvarado et al., 2007, Renard and Tognelli, 2016).

This study site is covered with grass, with a 60 cm-thick sandy soil layer overlying a clayey-sand layer from 60 to 100 cm depth, and homogeneous sands below 100 cm (Figure 1a). In May 2006, a 3 m-deep trench was dug in the unsaturated zone using an excavator. Soil sampling for pore water extraction and isotopic analysis was conducted immediately after opening the trench. Soil was sampled using a spatula, with a 2.5 cm spacing from 0 to 137 cm depth, and a 10 cm spacing from 140 to 280 cm depth (Figure 1a). Soil samples weighting on average 100 g were collected in polypropylene bottles, and were stored in these air-tight bottles until stable isotope analysis. Gravimetric water content was measured on the grab samples and converted into volumetric water content assuming a bulk density of 1500 kg m^{-3}.

The water table below the site lies on average 6 to 8 m below the surface. Several groundwater samples were collected from nearby wells for stable isotope analysis.

Meteorological data, precipitation and temperature, were obtained with a 15 day time step from the nearby Trappes station (Meteo France). The average annual precipitation is 668 mm/yr, while average monthly temperatures fluctuate between 0 and 25 °C. An average recharge rate of 100 to 150 mm/yr was calculated for the previous decades based on environmental tracers and at the regional scale around the study site (Corcho Alvarado et al., 2007) (Table 1), and confirmed by hydrological modelling in the Fontainebleau Sands (Renard and Tognelli, 2016).

Precipitation was collected 10 km away, at the GEOPS laboratory (Université Paris-Sud/Paris-Saclay, Orsay, France), from August 2002 to July 2012, using paraffin oil to prevent evaporation of water in the collector and isotopic fractionation. Cumulated bulk precipitation was sampled twice a month, and its isotopic composition was measured (Figures 2-4). d-excess (d-exc = $\delta^2H - 8\ \delta^{18}O$) in precipitation was calculated to look for variable origins of air masses. The Local Meteoric Water Line (LWML) was defined based on isotopic composition of precipitation. Its equation ($\delta^2H = a\ \delta^{18}O + b$) was used to calculate lc-excess values (lc-exc = $\delta^2H - a\ \delta^{18}O - b$) for the soil pore water profile.

2.1.2 The St. Lawrence Lowlands site (SLL)

The St. Lawrence Lowlands site (SLL) is located 70 km southwest of Montreal, in the Vaudreuil-Soulanges area (Quebec, Canada), and in the medium sands of the Saint-Télésphore esker. These glacio-fluvial sediments can reach up to 40 m thickness, and lie above the regional bedrock aquifer (Larocque et al., 2015). The aquifer is locally unconfined, with a total porosity of 40%, and a hydraulic conductivity on the order of 10^{-5} to 10^{-4} m/s. The study site is located in a flat area close to a sand quarry. The area is covered mainly by woodland, except on the site outcrops where the soil is covered with grass and where the unsaturated zone was sampled.

Taking advantage of an outcrop recently exposed on the border of the sandpit excavation, the entire unsaturated zone was sampled in May 2013, for the measurement of water content and pore water isotopic composition. The vertical outcrop was refreshed immediately before sampling, by removing the 5 to 10 cm of sand exposed to the atmosphere. Sand samples were taken using a spatula, with a 5 cm spacing below the soil surface down to a depth of 2.5 m, and then a 10 cm spacing down to the water table at 4.5 m depth (Figure 1b). Soil samples weighted on average 100 g, and were stored in air-tight polypropylene bottles until stable isotope analysis. The sand was homogeneous over the whole depth profile. Gravimetric water content was measured on the grab samples and converted into volumetric water content assuming a bulk density of 1500 kg m^{-3}. Daily weather data (minimum and maximum temperature, and precipitation) were obtained from the Coteau du Lac station (http://www.mddelcc.gouv.qc.ca/climat/donnees/index.asp), 20 km from the study site. Average annual precipitation in this area for the period 1980-2010 is 960 mm/yr (Larocque et al., 2015), with average monthly temperatures ranging from -11 to 23°C. An average recharge rate of 189 mm/yr was calculated for the period 1990-2010 using a regional surface water budget (Larocque et al., 2015) (Table 1).

Monthly isotopic composition values and precipitation amounts were obtained from the closest station of the GNIP network (http://www-naweb.iaea.org/napc/ih/IHS_resources_gnip.html), namely Ottawa, from January 2010 to June 2012 (Figure 4). Even though Ottawa is relatively far (130 km) from the study site, the climate is similar at Ottawa and the SLL site, with similar average temperatures and precipitations as well as altitude, and Ottawa precipitation isotope compositions can thus be used for the SLL site model. Discrete snowpack and rain samples were collected at SLL for measurement of isotope composition, in 2012 and 2013. As for PB, the LMWL was determined and used to calculate the lc-excess for the soil pore water profile.

2.2 Water extraction and isotopic analysis

Stable isotopic analysis of soil pore water was conducted on water extracted from the collected soil samples using cryogenic vacuum extraction. As soils sampled for this study were mainly of sand type and relatively wet, water isotope composition from cryogenic extraction can be assumed to have limited bias. A small soil sample of around 100 grams for PB (resp. 40 g for SLL) was introduced into a vacuum line. Pore water was extracted during 6h for PB (resp. 2h for SLL) and condensed in a collection tube maintained in liquid nitrogen (Araguás-Araguás et al., 1995). The soil sample was heated around 60 °C during the extraction, under a vacuum around 10^{-3} mbar. After extraction, the soil sample was weighed, heated overnight at 100 °C, and then re-weighed to determine the water extraction yield, as well as the initial soil water content.

The isotopic signature of water condensed in the collection tube was shown to follow a Rayleigh distillation curve (data not shown), and more than 98% of pore water must be extracted to ensure the absence of isotopic fractionation in the pore water sample. For this study, extraction yields were monitored to be above 98% for all samples.

Oxygen and hydrogen isotopic compositions of the extracted soil pore water as well as of the groundwater and precipitations were measured with a Thermo Finnigan Delta +™ isotope ratio mass spectrometer in dual inlet mode, coupled to an equilibration bench, at GEOPS Laboratory (Université Paris Sud, France) for PB, and on a Micromass Isoprime™ isotope ratio mass spectrometer in dual inlet mode, coupled to an Aquaprep™ system, at GEOTOP-UQAM (Montreal, Canada) for SLL. Each analysis required 200 µl of water, equilibrated at 40^0C for 7 hours with CO_2 for $\delta^{18}O$, and equilibrated at 40^0C for 4 hours with H_2 with a platinum catalyst for δ^2H.

Raw data were corrected using three internal working water standards, expressed in δ notation and normalized to the VSMOW-SLAP scale:

$$\delta^{18}O = \frac{R_{sample}}{R_{VSMOW}} - 1, \text{ where } R_i = \left(\frac{[^{18}O]}{[^{16}O]}\right)_i \quad (1)$$

Where i corresponds to sample or VSMOW. Hydrogen isotopic composition is expressed similarly to δ^2H, with $R_i=([^2H]/[^1H])_i$. $\delta^{18}O$ and δ^2H are reported in ‰ versus the VSMOW international standard. The precision of liquid water analysis (1σ) was 0.05 ‰ for $\delta^{18}O$, and 1 ‰ for δD. To check that the uncertainty due to cryogenic extraction was small, dried soil samples were rewetted with standard water. The measured isotopic composition for the extracted water was equal to that of the standard water in the range of the analytical precision of the liquid water analysis.

2.3. Recharge estimation from water isotope profiles in the unsaturated zone

The seasonal variation in isotopic composition of precipitation (Figure 2), combined with a seasonal pattern of evapotranspiration, were both directly linked to atmospheric temperatures. This led to a seasonal pattern of the amount and isotopic composition of infiltration and recharge. Even if the input function (i.e., rain isotopic composition) was unknown, a simple interpretation of water stable isotope depth profiles would allow the amount and periods of recharge to be quantified. The successive recharge seasons (summer/autumn vs. winter/spring) were identified in water isotope profiles in the unsaturated zone due to their contrasting water isotopic compositions (enriched versus depleted in heavy isotopes) (Figures 2, 5 and 6). The amount of recharge *R* (in m/yr) that occurred during a certain time period *T* (in yr) was obtained by integrating the volumetric water

content $\theta_w(z)$ over the depth interval [Z_1, Z_2] (in m), corresponding to the time period T (equation after and Leibundgut et al., 2009) :

$$R = \frac{\int_{Z1}^{Z2} \theta_w(z) \cdot dz}{T} \quad (2)$$

For this method, the time period could only be roughly estimated, typically to one hydrological season or one year, from the evolution of water isotope composition with depth.

2.4. Lumped parameter models for water isotopic composition in the unsaturated zone

In order to explain the observed evolution of water isotopic composition in the unsaturated zone, a surface water and isotopic mass balance was used, combined with a lumped parameter model for isotope transport in the unsaturated zone. The successive steps of the model, from time series of precipitation and its stable isotopic composition to the depth profile of water isotopes in the unsaturated zone, are described below.

2.4.1. Water budget at the subsurface

As a reference point from which to discuss the water budget at the surface, the traditional Thornthwaite method (Thornthwaite, 1948) was used to calculate the potential recharge fluxes. This conceptual box model applied to the soil layer from the surface to the root depth, according to:

$$P(i) = ET(i) + R(i) + I(i) + (S(i) - S(i-1)) \quad (3)$$

Where i is the time step, P is precipitation (rain), ET is actual evapotranspiration, R is surface and subsurface runoff, I is percolation below the root depth i.e. recharge, and S is soil storage, all expressed in mm.

This water budget was calculated at the smallest time step available for stable isotopes in precipitation (15 days for PB; one month for SLL). Potential evapotranspiration (PET) was calculated following Turc (1961). Even if it was shown to tend to underestimate evapotranspiration (Fisher et al., 2013), this method was originally developed for a monthly time step, close to the one used here, and could be applied solely with temperature data, hence well adapted for poorly instrumented sites. For both PB and SLL sites, surface and subsurface runoff was assumed to be negligible due to the flat topography and very permeable sandy soils. The conceptual soil storage reservoir, with a maximum capacity S_{max}, corresponded to the upper soil down to the maximum rooting depth (0.1 to 0.2 m at both sites), where evaporation and transpiration occurred, and below which there was no more water abstraction. Water "overflowing" and leaving this soil reservoir percolated vertically in the unsaturated zone and finally recharged the underlying aquifer. At each

time step *i*, precipitation P(*i)* and soil water *S(i-1)* were considered as available water for evaporation and transpiration. Actual evapotranspiration was calculated based on the demand on potential evapotranspiration and this amount of water available in the precipitation and soil reservoir: $ET(i)= min[PET(i), S(i-1)+P(i)]$. If some water remained available after subtraction of the actual evapotranspiration, it was added to the soil storage. Recharge was then calculated as overflow, water exceeding the threshold capacity S_{max} of the soil storage reservoir. This could be summarized as $I(i) = max[S(i-1) + P(i) – min[PET(i), S(i-1)+P(i)]$, 0] and $S(i) = max (S(i-1) + P(i) – min[PET(i), S(i-1)+P(i)], S_{max}]$. Such simple conceptual box model was assumed valid as the calculation time step was long, typically 15 to 30 days, and as the objective was to build a data parsimonious model.

For SLL, recharge was expected to occur in autumn and with snowmelt in the spring, the soil usually being frozen from mid-December to March. Hence, the water budget was modified, with evapotranspiration and recharge set to zero when average monthly temperature remained below zero. The potential excess water then accumulated in the soil reservoir until positive temperatures occurred.

2.4.2. Hydro-isotopic surface budget

Following the traditional surface water budget (Equation 3), a simple mass balance approach was applied to determine the isotopic composition of percolating water. At each time step, the water isotopic composition of the soil storage reservoir was calculated by weighting the isotopic composition of water in the soil reservoir at the previous time step and the isotopic composition of the precipitation at the current time step:

$$\delta_{mix}(i) = \frac{\delta_S(i-1)\cdot S(i-1)+\delta_P(i)\cdot P(i)}{S(i-1)+P(i)} \quad (4)$$

where δ is the isotopic composition of oxygen $\delta^{18}O$ or deuterium, δ^2H, δ_S and δ_P are the isotopic compositions of soil storage and of precipitation respectively, and *S(i)* and *P(i)* are the amounts of water in soil storage and in precipitation respectively at time step *i*.

Partitioning between evaporation and transpiration was not included in the model, neither based on field properties nor directly based on water stable isotopes as done by Sutanto et al. (2012). However, for grassland under northern latitudes as for the two sites, transpiration was shown to dominate with more than 60 to 65 % of total actual evapotranspiration (Wei et al., 2017). As will be discussed in Section 4.4., the isotopic composition of water in the unsaturated zone was used as a simple proxy to identify whether some evaporation occurred during the recharge period(s), without going further into quantification. At PB, isotopic compositions clearly followed an evaporative

trend, with negative lc-excess (Figure 6a), while it was not the case at SLL with null or slightly positive lc-excess (Figure 6b). In the model, isotopic fractionation was therefore assumed to occur due to evaporation of water in the soil reservoir. Soils of the two study sites were wet (Figure 5, water content above 10 vol%), consistent with medium intensity of evapotranspiration fluxes, and the equations of isotopic fractionation for open water (Gonfiantini, 1986) were used, including equilibrium fractionation as well as kinetic enrichment. At each time step, the isotopic composition of water in soil storage was calculated by:

$$\delta_S(i) = \left(\delta_{mix}(i) - \frac{A(T,Hr)}{B(T,Hr)}\right) \cdot f^{B(T,Hr)} + \frac{A(T,Hr)}{B(T,Hr)} \quad (5)$$

where f is the fraction of water remaining after evaporation, and $A(T,Hr)$ and $B(T,Hr)$ are parameters defined in Gonfiantini (1986), which depend on both atmospheric temperature, T, and atmospheric relative humidity, Hr. Following Barnes and Allison (1983), in A and B terms the kinetic enrichment is equal to $(1-Hr)$, the relative difference of transport resistance in air between isotopes being neglected.

The outputs of the water budget (Sect. 2.4.1., Equation (3)) were used to calculate $\delta_{inf}(i)$, the isotopic composition of water percolating from soil storage into the unsaturated zone. $\delta_{inf}(i)$ was determined for the time steps during which recharge occured (Figure 3), and was equal to $\delta_s(i)$, the isotopic composition of water in soil storage at this same time step.

The model was applied to PB from January 2002 to May 2006. For SLL, the hydro-isotopic surface budget was calculated from January 2010 to May 2013, with a monthly time step.

2.4.3. Transport in the unsaturated zone: piston flow with dispersion

The second step of the model was the transport of water stable isotopes through the unsaturated zone. Considering the limited amount of data available with which to constrain the model, a conceptual input-output lumped parameter model based on transit time distribution was chosen in this study to identify the main processes. According to the literature (Barnes and Allison, 1988, Lindström and Rodhe, 1992) and in agreement with the observed variability in $\delta^{18}O$ and $\delta^{2}H$ with depth (Figure 5), transport of water stable isotopes in the unsaturated zone occurred by convection or piston flow, as well as dispersion, leading to attenuation at depth,. The convolution integral method of Maloszewski and Zuber (1993) was chosen to reproduce and interpret soil depth profile data.

First, the amount of recharge, $I(i)$ (Equation (3)), and the isotopic composition of the percolating water, $\delta_{inf}(i)$, obtained from the water and isotope budgets as a function of time (Figure 3) were combined and transformed into $\delta_{inf}(q)$, a series of isotopic compositions for each increment of recharge water q (in mm). Water content was not constant with depth (Figure 5), but the spatial and

temporal variations of soil water content were not included in the model as it was intended only as a simple model for validation of the surface budget and main recharge processes and not as an investigation of transient processes. The assumption of homogeneous and constant water content by volume θ_w was done, corresponding to steady-state vertical flow of water. This assumption was required to apply the transit time distribution model. A simple mass balance, or piston flow model, was applied to propagate the series $\delta_{inf}(q)$ into a series $\delta'_{inf}(z)$ of isotopic composition as a function of depth in the unsaturated zone. This was done using the simple conversion:

$$z = \frac{Q_{max} - q}{\theta_w} \quad (6)$$

where z is the depth (in mm) reached by the parcel of water after a cumulated amount of recharge of q (in mm) and Q_{max} the cumulated recharge amount (in mm) calculated for the entire time series (2002-2006 at PB).

Second, the relationship between input and output soil water isotopic compositions was based on a transit time distribution function. The dispersion model (Maloszewski et al., 2006) was chosen as the most appropriate for solute transport in soil columns. This dispersion model was retained also for its parsimony, with dispersion parameter as the single parameter. However, the limitations of this model have to be kept in mind, especially through the steady-state assumption. The model proposed here aimed at giving a general estimate of water fluxes, not at explaining the transient dynamics of soil water isotopes.

Traditionally, the dispersion model was used to determine the temporal evolution of the isotopic composition of unsaturated zone water collected at a fixed depth with a lysimeter, based on the temporal evolution of the isotopic composition in precipitation. For the current application, the isotopic composition of unsaturated zone water for all depth had to be calculated at a fixed time, namely time of sampling. The lumped parameter dispersion model was therefore modified as follows. For each depth Z, the isotopic composition after dispersion $\delta'_{out}(Z)$ was obtained by a convolution between the transit time distribution function with the theoretical non attenuated depth profile $\delta'_{inf}(z)$:

$$\delta'_{out}(Z) = \int_{z=0}^{Zmax} \delta'_{inf}(z) \cdot g(Z, z) \cdot dz \quad (7)$$

where $Z_{max} = Q_{max} / \theta_w$ (in m) is the maximum depth reached by water that recharged at the really beginning of the time series, and g is the transit time distribution function of the dispersion model:

$$g(z) = \frac{\sqrt{Z} \cdot z^{-3/2}}{\sqrt{4\pi \cdot D}} \exp\left(-\frac{(Z-z)^2}{4D\, z\, Z}\right) \quad (8)$$

where D is the dispersion parameter (-), that corresponds to the inverse of the Peclet number (Maloszewski and Zuber, 1993) and hence can theoretically be linked to dispersivity.

The depth profile of water content was not modeled, as it was not the main focus of the study, and not adapted to the lumped parameter model.

342 The transport model was applied only at PB. For SLL, the surface budget was less well constrained,
343 due to the lack on time series on snowpack and snowmelt, and the transport segment was not
344 pursued.
345

346 **2.4.4. Sensitivity analysis and calibration of the model parameters**

347 Four years of data were available prior to sampling at PB. The model was thus run from July 2002
348 to May 2006. Isotopic compositions were observed to be constant below 2 m at PB (Figure 5a).
349 With such dispersion the available 4-year period of data was long enough so that the first and oldest
350 recharged water isotopic compositions were attenuated at 2 m depth. Especially, any data older than
351 2002 would have been attenuated. To ensure initialization of the model but keep its simplicity, the
352 recharge time series, $\delta'_{inf(z)}$, was artificially extended into the past (corresponding to the deepest
353 part of the unsaturated zone) using the measured average isotopic composition of the deep
354 unsaturated zone. For cases where shorter time series of meteorological data were available or if
355 dispersion was lower, monthly average data of precipitation amounts and isotope compositions
356 should be used for a spin up period.
357 A simple sensitivity analysis was conducted to determine the influence of the hydro-isotopic budget
358 parameters on the average isotopic composition and intensity of recharge (Table 2). The value of
359 atmospheric relative humidity *Hr* did not influence much the isotopic composition of recharged
360 water, but controlled the slope of the unsaturated zone data points relative to the meteoric water line
361 in $\delta^{18}O$-$\delta^{2}H$ space. The value of *f* strongly influenced the isotopic composition of recharged water
362 (Table 2), and the value of S_{max} also influenced it, albeit to a lesser extent. The values of these three
363 latter parameters were determined by a simple inverse approach. The objective function φ that was
364 minimized was defined as the difference between the calculated volume weighted average $\delta^{18}O$ of
365 recharge water and the measured volume weighted average $\delta^{18}O$ in the unsaturated zone:

366 $$\varphi = \left| \frac{\sum_i \left(I(i) \cdot \delta_{inf}(i) \right)}{\sum_i I(i)} - \overline{\delta^{18} O_{UZ}} \right| \quad (9)$$

367 The calibration of the two parameters of the transport model, namely the dispersion coefficient, *D*,
368 and soil water content, θ_w, was then performed. The transport parameter values depended on those
369 of the parameters obtained from the surface budget. Soil water content θ_w, was not very sensitive
370 and was estimated by manual adjustment, For the dispersion coefficient, a coupled multi-parameter
371 calibration as in Sprenger et al. (2015) was not conducted, but rather a simple parameter adjustment
372 based on the minimization of the objective function φ' defined as the sum of the differences
373 between the measured and calculated $\delta^{18}O$ in the unsaturated zone at each depth where
374 measurement was available:

$$\varphi' = |\sum_{Z0}^{Zmax}(\delta'_{out}(Z) - \delta^{18}O_{UZ}(Z))| \quad (10)$$

The transport model applied only to water percolating below the active rooting depth. The upper 10 to 20 cm of the soil, which corresponded to the conceptual soil reservoir or at least where transient surface water fluxes occurred, were thus not handled by the transport model and not considered for the fit and interpretation.

3 Results

3.1. Water budget

For PB, the average annual precipitation and potential evapotranspiration were 668 mm/yr and 746 mm/yr respectively. The surface water budget was calculated for a fitted S_{max} value of 52 mm, giving the best agreement for isotopic composition (see Section 3.5.). The actual evapotranspiration was 455 mm/yr, for an average recharge of 203 mm/yr (Table 1). Recharge was obtained almost continuously from September/October until April/May (Figures 2 and 3). In that case, the value of S_{max} could be compared to soil moisture capacity, according to $S_{max} = RD\ (\theta_{FC}\text{-}\theta_{WP})$ (Renard and Tognelli, 2016) where RD is the rooting depth (0.2 m for PB site), θ_{FC} (resp. θ_{WP}) the water content of soil at field capacity (resp. at wilting point) (around 5 %, resp. 2 %). Physical soil properties gave a S_{max} value of 6 mm, which was lower but compatible with the fitted value. The obtained intensities for the various components of the water budget were in agreement with those used by Renard and Tognelli (2016) for another site nearby in the Paris Basin.

For SLL, the average annual precipitation (rain and snow) and potential evapotranspiration for the 2010-2013 period were 922 mm/yr and 630 mm/yr respectively. The S_{max} value leading to the best agreement between modelled and measured volume weighted average isotopic composition was 300 mm, a large value that could not be linked to a physical interpretation of soil moisture content. It has to be emphasized here that S_{max} was a fitting parameter of the conceptual water budget, and that the link with field observations was not straightforward, especially as S_{max} depended on the model time step. The large value of S_{max} was thus consistent with the long time step (one month) of the model at SLL. The corresponding recharge was 200 mm/yr, in the same order of magnitude as the literature value (Table 1), but occurred only for snowmelt. The actual evapotranspiration was 608 mm/yr, almost equal to the potential evapotranspiration. At the regional scale, Larocque et al. (2015) obtained a lower value of 381 mm/yr, but this had to be considered carefully as they considered runoff. The water budget thus appeared to be somehow too simple to fully handle

processes occurring in cold climate and thus to be affected by large uncertainty. Larocque et al. (2015) used a more complex surface water budget better adapted to cold climate and showed that recharge occurred during two separate periods for SLL, from October to December and then from March to May. A S_{max} value of 50 mm was also tested, similar to that for PB site. This led to an actual evapotranspiration of 517 mm/yr and recharge of 372 mm/yr, with an acceptable fit of isotopic composition but this time with the two expected periods of recharge.

3.2. Stable isotopic composition of precipitation

Monthly-averaged amount-weighted isotopic compositions of precipitation for the two sites are presented in Figure 2. The entire time series of isotopic compositions and d-excess in precipitation from 2002 to 2006 for the PB site is presented in Figure 3. A marked seasonal fluctuation existed for both sites, due to temperature effects, with correlation coefficients (R^2) of 0.36 and 0.60 between $\delta^{18}O$ in precipitation and temperature for PB and SLL sites respectively. Isotopic compositions of both oxygen and hydrogen were more depleted in heavy isotopes during winter, and more enriched in heavy isotopes during summer. The amplitude of this seasonal fluctuation was greater for SLL (97 ‰ for δ^2H and 12 ‰ for $\delta^{18}O$) than for PB (29 ‰ for δ^2H and 5 ‰ for $\delta^{18}O$) due to the larger underlying temperature amplitude. For SLL, the isotopic compositions of the local rain samples were fully consistent with the values from the GNIP station in Ottawa. The average isotopic composition of the snow samples was -17.3 ± 0.1 ‰ for $\delta^{18}O$ and -126.1 ± 1.0 ‰ for δ^2H, also in the range of the GNIP station data.

For both sites, the local meteoric water lines (LMWL) were close to the global meteoric water line (Figure 4a, $\delta^2H = 7.12 \cdot \delta^{18}O + 2.19$ for PB, Figure 4b $\delta^2H = 7.93 \cdot \delta^{18}O + 10.1$ for SLL). They compared well with LWML equations in the same regions (see for example Millot et al., 2010 for PB and Arnoux et al. 2017 for SLL). For PB, the d-excess varied throughout the year according to the variable sources of vapour masses and the continental recycling of water vapour. The lc-excess was therefore used to overcome this variability and to investigate local evaporation effects.

3.3. Stable isotopic composition of pore water in the unsaturated zone and of groundwater

Depth profiles of the water isotopic composition in the unsaturated zone are presented in Figures 5a and 6a for PB, and Figures 5b and 6b for SLL. Only $\delta^{18}O$ profiles are reported, as δ^2H profiles show similar patterns of variability. At both sites, at least one cycle of pore water isotopic composition

variation was observed: higher isotopic compositions corresponding to recharge at the end of summer and during autumn, and lower values corresponding to recharge during winter. The amplitude of soil water isotopic composition variability was dampened compared to that of precipitation, with 12 ‰ for δ^2H and 2 ‰ for $\delta^{18}O$ for PB, and with 51 ‰ for δ^2H and 7 ‰ for $\delta^{18}O$ for SLL. At PB, the 4.5 ‰ $\delta^{18}O$ amplitude in precipitation was attenuated to 1 ‰ at 1 m depth and to less than 0.1 ‰, in the range of noise, at 2 m depth. Depth profiles of soil water content also showed some variability (Figure 5), with higher water content in the finer layer at PB and at the surface at both sites, as sampling was conducted in May following winter infiltration.

The average isotopic composition of water in the unsaturated zone, weighted by the water content at each depth, was equal to $\delta^{18}O_{uz}$ = -6.9 ‰ and lc-excess = -3.7 ‰ for PB, and $\delta^{18}O_{uz}$ = -11.3 ‰ and lc-excess = 2.3 ‰ for SLL (Figures 5 and 6).

At PB, the average isotopic composition of groundwater was $\delta^{18}O$ = -6.3 ‰ and δ^2H = - 48.3 ‰ (Figure 4a), while the measured isotopic composition of groundwater in the regional surface aquifer of SLL was $\delta^{18}O$ = -11.1 ‰ and δ^2H = -78.5 ‰ (Larocque et al., 2015) (Figure 4b).

Even if the two sites had contrasted recharge seasonality and mechanisms, the average isotopic composition of the unsaturated zone was close to that of local groundwater (Figure 4). For SB, it was quite close to the weighted average isotopic composition of local precipitation, reflecting the continuous infiltration and recharge of precipitation from fall to spring. On the contrary, for SLL the average isotopic composition in the unsaturated zone was more depleted in heavy isotopes compared to the weighted average isotopic composition of local precipitation, corresponding to the large amount of heavy isotope-depleted snowmelt water infiltration as freshet. Comparison between $\delta^{18}O$ values in precipitation and in the unsaturated zone indicated that recharge occurred predominantly during late fall ($\delta^{18}O$ around -10 ‰) and spring ($\delta^{18}O$ of snow around -15 ‰) at SLL (Figures 2 and 5b) and more continuously during fall, winter and spring at PB (Figures 2, 3 and 5a). At PB, lc-excess values in the unsaturated zone ranged between 0 and -5 ‰ (Figure 6a), indicating that at least some evaporation occurs before or during infiltration, leading to a slight enrichment of the isotopic composition in recharge in the unsaturated zone. At SLL, average lc-excess values in the unsaturated zone were positive (Figure 6b), showing no evidence of evaporation, and suggesting existence of heterogeneity in precipitation sources between recharge and non recharge periods.

3.4. Quantification of recharge for the two contrasted sites from unsaturated zone profiles

Annual amounts of recharge were calculated following Equation (2) for both the PB and SLL profiles, and are reported in Table 1. The depths between which soil water content was integrated were chosen based on the visual observations of maximum / minimum as well as smooth or more

abrupt changes in the isotopic composition vs. depth (Figure 5), and on the expected seasonality of isotopic composition in recharge, from higher values in autumn to more negative values in spring. At PB, one cycle between two minima of $\delta^{18}O$ was identified between 22 and 160 cm depth (Figure 5a), corresponding to an annual recharge of 304 mm/yr (Table 1). More specifically, infiltration during autumn –winter 2005 could be identified by relatively higher isotope composition between 50 and 125 cm depth, while recent infiltration in spring 2006 would lie on top of the profile, down to 50 cm depth (Figure 5a). At SLL, relatively abrupt changes of $\delta^{18}O$ vs. depth allowed identification of spring snowmelt and autumn recharge for the two hydrological years preceding sampling (Figure 5b). In the upper part of the profile, snowmelt recharge in 2013 was identified with low $\delta^{18}O$ values between 22 and 72 cm and amounted to 48 mm, and a recharge in autumn 2012 of 66 mm was identified with high $\delta^{18}O$ values between 77 and 142 cm. This led to an annual recharge of 114 mm/yr (Table 1). Below that, low $\delta^{18}O$ are again observed from 147 to 212 cm, corresponding to spring snowmelt in 2012, with a calculated recharge of 68 mm. Below 217 cm, $\delta^{18}O$ values increased slightly and probably corresponded to recharge from autumn 2011, but the amplitude of variations were too dampened to allow quantification. Quantification of recharge based on integration of water content versus depth was based on a subjective identification of the depth range corresponding to a given period of recharge, and thus subjected to a large uncertainty. But the results obtained here provided a straightforward first order estimate of the amount of recharge and also a confirmation of the expected seasonality of recharge.

3.5. Modeling the isotopic composition evolution from precipitation to infiltration and recharge

For PB site, Equation (5) and isotope fractionation were at first not included in the model, but the minimization of the objective function φ never led to a good fit of the data (data not shown). As already identified with the negative lc-excess value in the unsaturated zone, this confirmed that some evaporation occurred at PB. Isotope fractionation was therefore included in the model. Minimization of the objective function φ then led to the determination of the surface water budget parameters (S_{max} and evaporation parameters, f and Hr). The best fit (Figures 5a and 6a) was obtained for S_{max} =52 mm, Hr=0.3 and f=0.97. The obtained average annual recharge was then 203 mm/yr (Table 1), not fully similar to the values obtained from the depth integration method (Section 3.4) and the literature, but of the same order of magnitude and acceptable regarding the simplifications of the model and associated uncertainties. It has to be noted that the surface hydro-isotopic budget was self-sufficient to quantify amount and seasonality of recharge. The transport model served as validation and confirmation of major recharge processes.

Regarding the lumped parameter transport model for PB, the distance between the two $\delta^{18}O$ minima in the soil profile (Figure 5a) was compared to the amount of recharge between the two $\delta_{inf}(r)$ series minima, and Equation (6) was used to deduce an average water content of $\theta_w = 0.42$. This value was higher than the field value of 0.15 to 0.25 (Figure 5a), but this parameter was not optimized automatically, was not very sensitive and thus affected by a large uncertainty. A dispersion parameter D of 0.06 gave the best fit between observed and calculated profiles (Figures 5a and 6a). This dispersion parameter was an effective bulk parameter for the whole profile, disregarding the spatial variations of dispersivity with depth, and was not interpreted in terms of dispersivity.

For SLL, as already guessed by the positive lc-excess values in the unsaturated zone, a model without evaporation and without isotope fractionation allowed to minimize the objective function and was thus retained. There was thus only one parameter, S_{max}, to calibrate. The best agreement between average $\delta^{18}O$ and $\delta^{2}H$ calculated for recharge versus measured values in the unsaturated zone profile was obtained for a S_{max} of 300 mm, leading to an annual recharge of 200 mm/yr (Table 1), again different from but of the order of magnitude of values obtained from the simple depth integration method (Section 3.4) and from the literature. However, as already discussed above (Section 3.1), a S_{max} value of 50 mm, similar to that for PB site and more physical, led to a higher recharge of 372 mm/yr, and emphasized the current limitations of the model for cold climate.

4 Discussion

4.1. Processes to include in models of soil water isotope profiles

In this study, a conceptual water isotope budget was used at the surface, based on the traditional water budget, but also taking physical processes that alter soil water isotopic composition into account, namely mixing and isotopic fractionation associated with evaporation. Regarding transport in the unsaturated zone, the choice of a lumped dispersion model to reproduce the measured $\delta^{18}O$ in the unsaturated zone was based on the limited availability of data with which to constrain the model. The simplified surface budget was consistent with a simplified transport model that did not incorporate the variations of water content with depth and with time. The influence of soil heterogeneity as well as of immobile water on the average isotopic composition of recharge water, were considered to be negligible by several authors (Barnes and Allison, 1988, Lindström and Rodhe, 1992), but shown to occur by others (Gazis and Feng, 2004, Schoen et al., 1999). The use of a lumped parameter model allowed these processes to be taken somehow into account through the effective dispersion parameter.

Other modeling approaches with more physically based numerical models of surface water, isotopes, and heat budgets have been proposed, coupled with water and isotope transport in the unsaturated zone based on Richard's equation (Braud et al., 2005, Gehrels et al., 1998, Haverd and Cuntz, 2010, Melayah et al., 1996, Rothfuss et al., 2012, Sprenger et al., 2015, Stumpp et al., 2012). Such physically based models handle transient dynamics of flow and transport in the unsaturated zone as well as soil heterogeneity, and help improve the calibration of soil transport parameters and the understanding of water residence time. However, they often lack detailed presentation of surface isotope water budget. Non-fractionating transpiration is usually included in the models for sites in temperate areas (e.g. HYDRUS-1D, Stumpp et al., 2012), but isotope fractionating evaporation is less often taken into account (Haverd and Cuntz, 2010, Rothfuss et al., 2012, Sprenger et al., 2017). These modeling approaches can be more or less complex but always require many data and parameters. On the opposite, the model proposed here is much more conceptual and simplified, but includes this isotope fractionation in a simple way while requiring few data and parameters.

4.2. Parsimonious models for recharge quantification and uncertainties

The two sites considered in this study have contrasted recharge mechanisms and seasonality, with episodic recharge in autumn and from snow melt at SLL, and more continuous recharge from autumn to spring at PB. The two models proposed here prove to allow quantification of recharge amount and possibly seasonality, for sites that are not so well characterized and where only limited data are available to interpret the isotope composition of pore water in the unsaturated zone.

The two methods give estimates or recharge intensity that are of the same order of magnitude as values given in the literature and obtained with hydrogeological modeling (Table 1). However, even if a precise quantification was not the scope of the study and uncertainties on annual recharge estimated from depth integration of water content as well as from our simple model are large, noticeable differences remain with literature values. Such uncertainties could not easily be handled by water managers, and the proposed models are not intended for their use. More generally, uncertainties on recharge estimations remain significant, especially when comparing methods and even with complex models, which emphasizes the need to further improve recharge quantification methods.

4.3. Limitations of the proposed water isotope budget and transport model for recharge quantification

The main limitation of the use of unsaturated zone water isotope profiles is the attenuation of the precipitation isotope signal at depth (Cook et al., 1992). As opposed to arid or semi arid regions

with deep unsaturated zone where several years of recharge intensity are stored, temperate areas generally have shallower unsaturated zone and higher recharge intensity that limits the preservation of the signals. The seasonal variability can clearly be observed at SLL, but is smaller for PB (Figure 5). The model shown here is intended to be theoretically applicable to any site under a temperate climate, even if sites characterized by strong seasonality, and especially by snow cover, are better suited to the quantification of recharge from soil water isotopes profiles.

Another limitation of the interpretation of soil water isotope profiles is that a long-term time series of precipitation isotopic composition is required. Monthly time step time series have been used as they are more easily acquired or available from GNIP stations almost all over the world. A monthly time step is shown here to enable the application of a hydro-isotopic surface budget and the identification of seasonal variations in recharge. If associated with a soil sampling at fine spatial resolution, typically every 5 cm or so, high resolution precipitation data could allow using a two components lumped parameter model, with both piston flow and dispersion models, to investigate the existence of preferential infiltration events, as proposed by Stumpp and Maloszewski (2010).

Uncertainties in the surface water budget propagate also into the transport model. The major limitation of the transport model is the assumption of steady-state flow (homogeneous and constant soil water content), that is required to apply the lumped parameter dispersion model, but that prevents any investigation of transient processes.

Finally, the surface water budget and transport models are calibrated here on one single profile taken at one time, which is one of its strength but also a limitation as it increases the uncertainties, from sampling and cryogenic extraction to model calibration.

Despite these limitations, the simple surface and transport models are complementary to the peak to peak method and are applicable to sites where little information is available.

4.4. Impact of evaporation on water isotopic composition in the unsaturated zone

Isotopic fractionation is associated with evaporation at the soil surface, and is known to be an important control of the isotopic composition of water in soils (Barnes and Allison, 1983, 1988, Sutanto et al., 2012). The average negative lc-excess value in the unsaturated zone for PB indicates the role of local evaporation, and this is confirmed by the hydro-isotopic surface budget, where isotope fractionation has to be included in order to fit the data. For SLL, this is not the case, average lc-excess is positive and a good fit of the data is obtained without fractionation. The fitted f coefficient of 0.97 for evaporation at PB might not be directly transposed into evaporation partitioning but confirms that transpiration largely dominates water uptake for soil under a temperate climate and covered with grass. Evaporation occurring during the hot summers clearly

leads to isotope fractionation, but as there is no recharge during summer, this fractionated signal is barely preserved in a small amount of soil moisture. When autumn rain is mixed with this small amount of fractionated water remaining in the soil reservoir, the evaporated signal is largely dampened. These values of *f* and evaporation therefore correspond to the recharge periods, namely autumn and spring. At SLL, where average temperatures are lower, evaporation from the unsaturated zone is negligible during the recharge seasons, spring and autumn, and hence does not impact the isotopic composition of recharge.

For both sites, lc-excess appears to be a robust indicator of local evaporation intensity. Water stable isotopes in the unsaturated zone have thus already been used to partition evaporation and transpiration fluxes (Sutanto et al., 2012). Surface budgets for water stable isotopes, including fractionation associated with evaporation, as proposed in this study, should be further developed and tested under temperate climates.

Equation (5) is proposed as a simple parameterization of the isotope fractionation associated to evaporation in soils, but it would require improvement and validation by future experimental and numerical studies. Equation (5) is originally established for open water, and is adapted to sites such as PB where evaporation fluxes are small and kinetic fractionation is limited. Future work should focus more specifically on the validation of such isotope fractionation equation. One major issue would be to include a parameterization or a fit of a temporal evolution of the *f* parameter, which would require to include partitioning of evapotranspiration.

5. Conclusion

Based on the results from two field sites with contrasting hydrologic and climate conditions, water stable isotope ($\delta^{18}O$ and $\delta^{2}H$) profiles in the unsaturated zone are shown to permit the quantification and identification of seasonality in recharge, based on the seasonality of the water isotopic composition of precipitation. A very simple approach consists of integrating soil water content between extreme values of soil water isotopic composition, typically corresponding to precipitation signatures during warm and cold seasons. When a time series of precipitation isotopic composition is available, even only at a monthly time scale, a water isotope budget can be calculated at the surface using a conceptual overflowing soil reservoir model including mixing of successive precipitation but also isotope fractionation due to evaporation. The intensity of evaporation, associated with enrichment in heavy isotopes, can be constrained by the lc-excess measured in the unsaturated zone. A first and simple parameterization is proposed to take this fractionation into account in the surface water budget. This water budget can be constrained with the volume

weighted average isotopic composition of pore water in the unsaturated zone, and thus allows determination of the intensity and seasonality of recharge. This model and the peak to peak integration lead to similar values, consistent at first order with more complex methods, but emphasizing the uncertainties remaining on recharge quantification. A lumped parameter dispersion model is then used to represent percolation of water and attenuation of water isotope signal at depth. It allows the main variations of water isotope composition with depth to be reproduced and soil parameters and transport processes to be discussed.

Even if the attenuation of the precipitation isotopic signal during recharge limits the investigation to the two first meters of the unsaturated zone and to sites with contrasted recharge periods, the approaches proposed in this study could be applied to a large number of sites, and are very promising for the investigation of spatial and temporal variability of recharge as well as for drawing recharge maps at the local to regional spatial scales.

Data availability

All data are available upon request to the corresponding author.

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Figure captions

Figure 1: Locations and photographs of the trenches where sand samples were taken for soil water extraction and water isotope analysis, for PB (a.) and for SLL (b.).

Figure 2: Monthly average oxygen isotopic composition of precipitation at PB (black) and SLL (grey). For each site, dashed lines represent the volume weighted average oxygen isotopic composition of the unsaturated zone profiles sampled in May 2006 for PB and May 2013 for SLL. Periods of recharge obtained from the water budget are represented by horizontal bars.

Figure 3: Time series of water and isotope surface budget for the PB site from 2002 to 2006: isotopic composition measured in precipitation (black and grey solid lines with monthly average as dashed lines), isotopic composition calculated in infiltrating water (red and orange), d-excess in precipitation (dashed grey line), measured precipitation (blue), calculated potential evapotranspiration (green) and infiltration into the unsaturated zone (red). See text for details.

Figure 4: Stable isotopic composition of water in precipitation (blue circles), in the unsaturated zone pore water (red triangles), and in groundwater (black square) for PB (a) and SLL (b). For each site, filled symbols correspond to volume weighted averages of individual empty data point. For each site, the Local Meteoric Water Lines (LMWL) and Global Meteoric Water Lines (GMWL) are represented by a black solid line and grey dashed line respectively.

Figure 5: Depth profiles of pore water oxygen isotopic composition (black) and volumetric water content (grey) at PB (a) and SLL (b). Vertical dashed lines correspond to volume weighted average isotopic composition in groundwater (black) and precipitation (blue). Annual or seasonal recharge is calculated through depth integration of water content in shaded areas. For PB, the modeled depth profile of isotopic composition (dashed line) is calculated from the hydro-isotopic surface budget and the 1D lumped parameter model (see text for details).

Figure 6: Depth profiles of lc-excess in pore water at PB (a) and SLL (b). For PB, the modeled depth profile of lc-excess (dashed line) is calculated from the hydro-isotopic surface budget and the 1D lumped parameter model (see text for details).

Tables

Table 1: Estimate of the amount of annual groundwater recharge at PB and SLL, using several methods. See text for details.

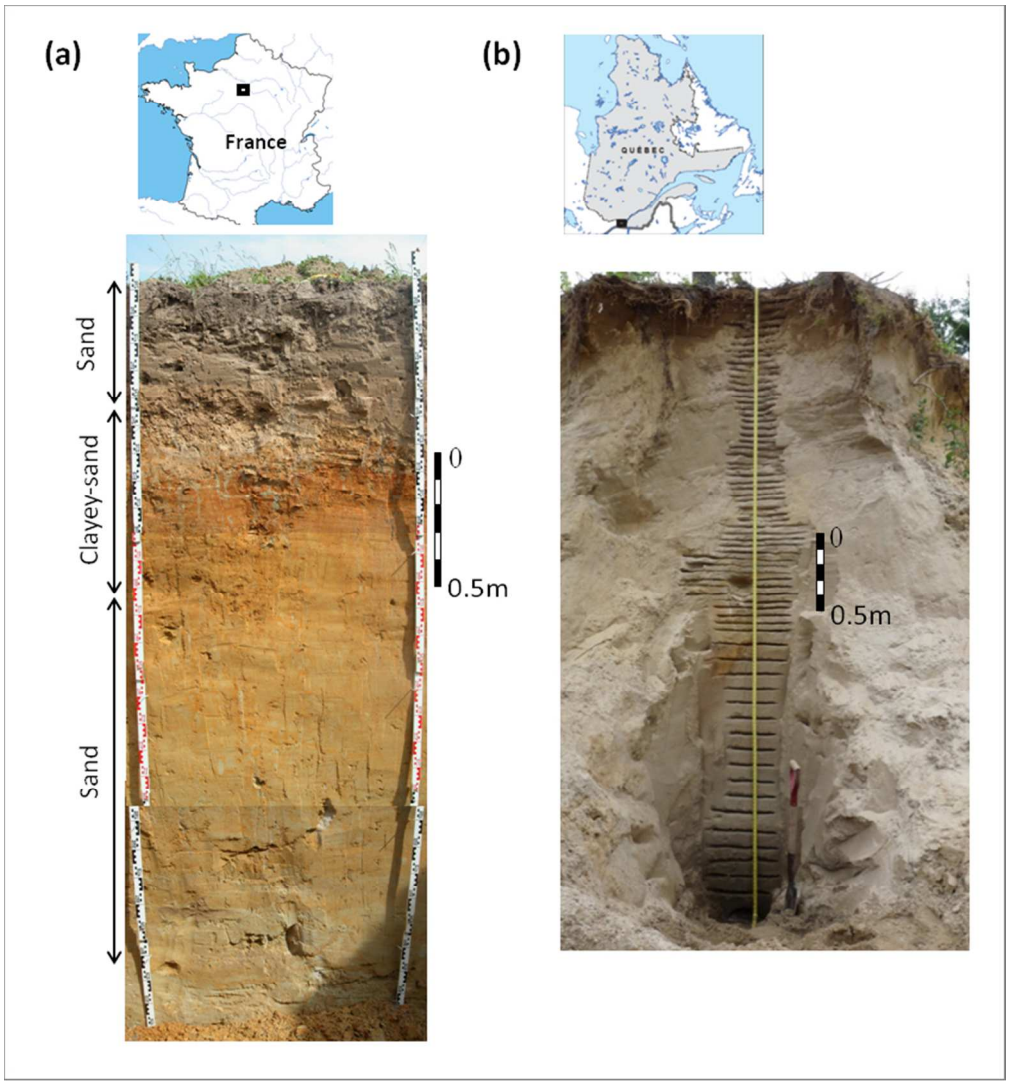
	PB	SLL
UZ isotope profile	304 mm/yr	114 mm/yr
Hydro-isotopic water budget	203 mm/yr	200 - 370 mm/yr
Regional Modeling	150 mm/yr[1]	189 mm/yr[2]

[1] Corcho Alvarado et al. (2007)
[2] Larocque et al. (2015)

Table 2: Sensitivity analysis of the hydro-isotopic surface budget to the parameters *Hr*, S_{max} and *f*, and to potential evapotranspiration (PET) for PB. See text for details.

	Average Recharge (mm)	$\delta^{18}O$-weighted average (‰)	$\delta^{2}H$-weighted average (‰)	lc-excess (‰)
Ref. case*	**203**	**-6.86**	**-50.00**	**-3.35**
Hr = 0.1	203	-6.95	-50.84	-3.55
Hr = 0.5	203	-6.72	-48.56	-2.90
S_{max} x2	147	-6.02	-46.31	-5.64
S_{max} /2	231	-7.41	-52.52	-1.95
f 0.99	203	-8.13	-55.60	0.10
f 0.90	203	-2.52	-30.40	-14.65
PET +10%	192	-6.88	-50.16	-3.36
PET - 10%	217	-6.82	-49.75	-3.38
UZ profile (PB site)		***-6.86***	***-50.39***	-3.74

* parameter values : S_{max} = 52 mm, *Hr*=0.30, *f*=0.97



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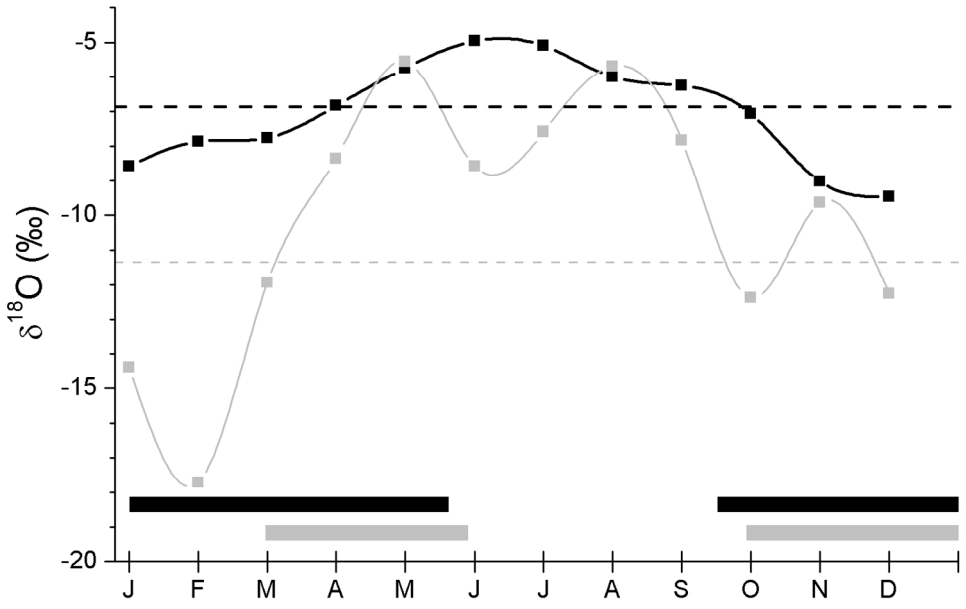

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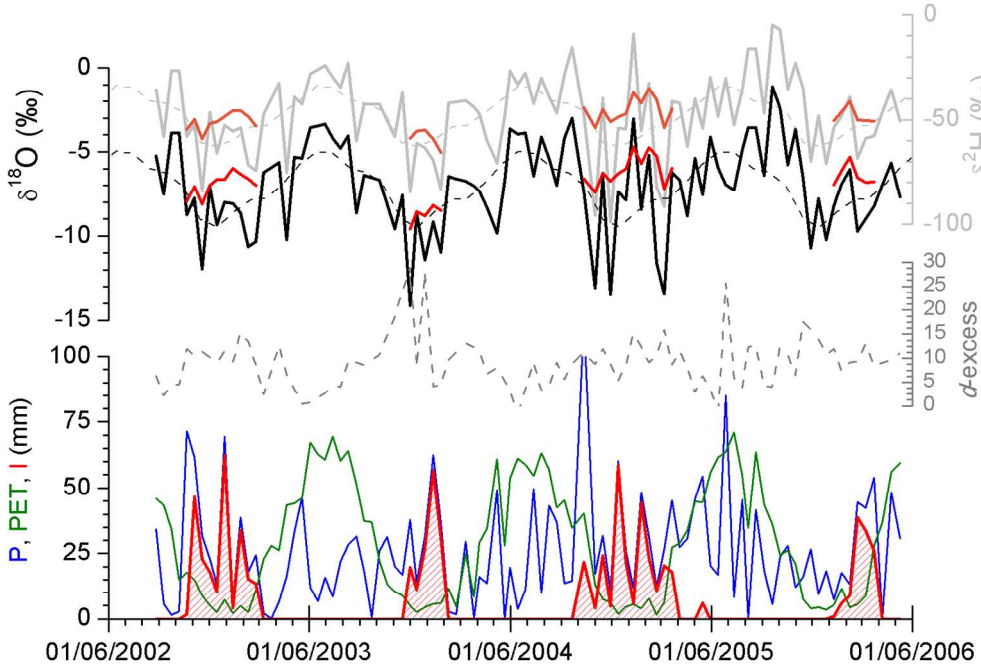

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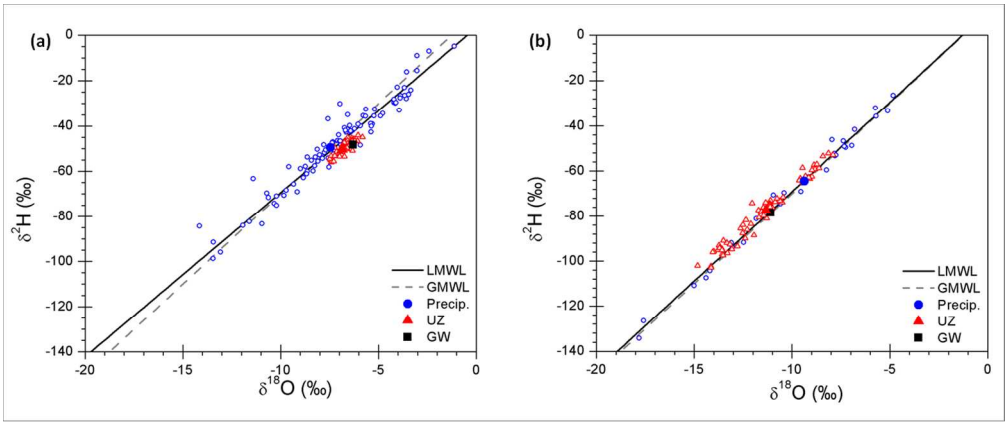

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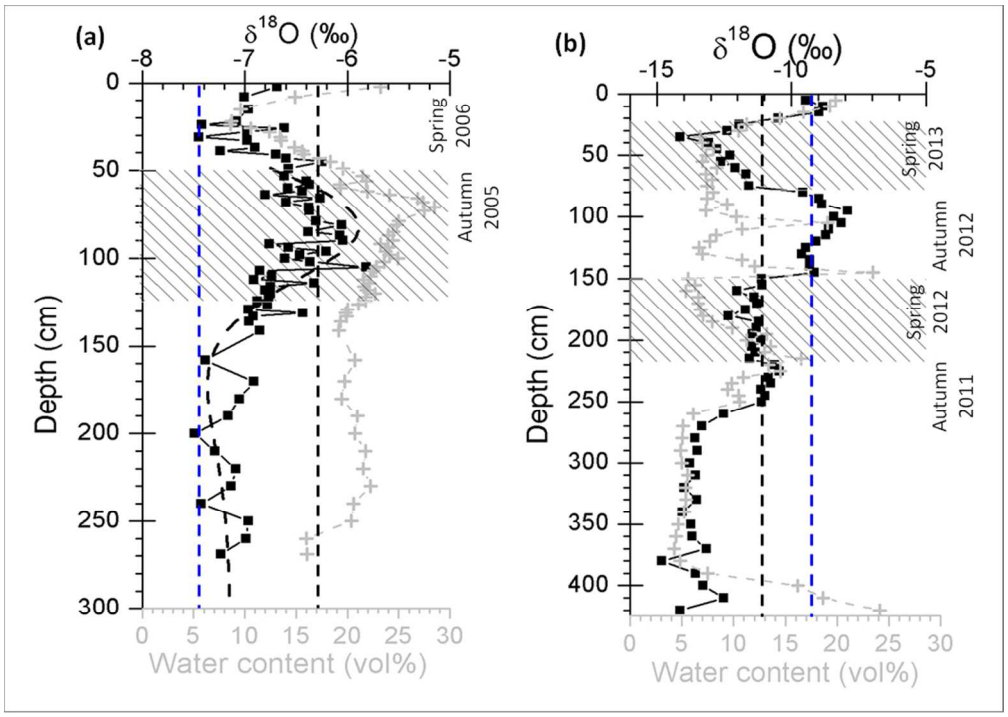

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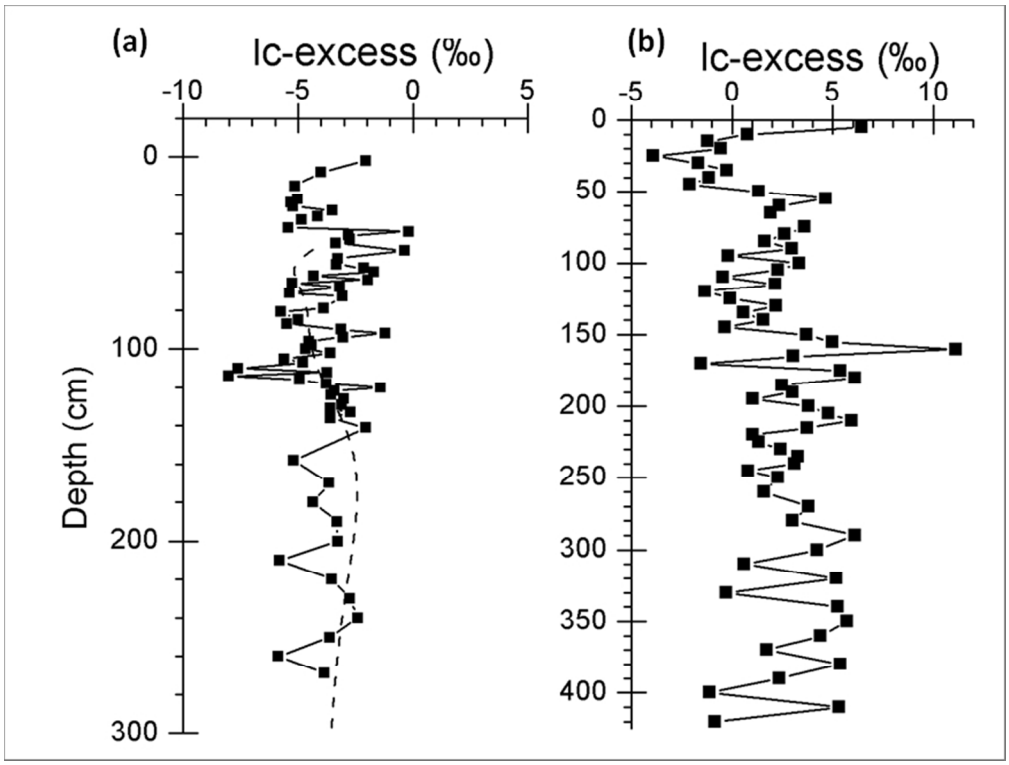

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