

Estimating water balance components of tropical wetland lakes in the Pantanal dry season, Brazil

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Abstract Water balance studies with stable water isotopes have rarely been conducted in remote and tropical wetland areas. As such, little is known regarding the water balance and groundwater–surface water interaction in the Pantanal, one of the largest and most pristine wetlands in the world. We applied MINA TrêS, a water balance model utilizing stable water isotopes ($\delta^{18}\text{O}$, $\delta^2\text{H}$) and chloride (Cl⁻) to assess the dry-season hydrological processes controlling groundwater–surface water interactions and the water balance of six floodplain lakes in the northern Pantanal, Brazil. Qualitatively, all lakes exhibited similarity in hydrological controls. Quantitatively, they differed significantly due to morphological differences in controlling groundwater inflow and lake volume. Our approach is readily transferable to other remote and tropical wetland systems with minimal data input requirements, which is useful in regions with sparse hydrometric monitoring.

Key words Pantanal; dry season; stable water isotopes; chloride; water balance; groundwater–surface water interactions

Estimation des composantes du bilan hydrique en saison sèche des lacs des zones humides tropicales du Pantanal (Brésil)

Résumé Les études sur le bilan hydrique utilisant des isotopes stables de l'eau ont rarement été menées dans les zones humides et tropicales reculées. En tant que tel, on connaît peu le bilan hydrique et les interactions entre eaux superficielles et souterraines dans le Pantanal, une des zones humides les plus importantes et les plus vierges du monde. Nous avons appliqué le modèle de bilan hydrique MINA TrêS, en utilisant des isotopes stables de l'eau ($\delta^{18}\text{O}$, $\delta^2\text{H}$) et les chlorures (Cl⁻), afin d'évaluer, en saison sèche, les processus hydrologiques contrôlant les interactions des eaux souterraines et superficielles et le bilan hydrologique de six lacs de la plaine d'inondation dans le Nord du Pantanal, au Brésil. Qualitativement, les contrôles hydrologiques de tous les lacs sont similaires. Quantitativement, ils diffèrent sensiblement en raison des différences morphologiques du contrôle de l'écoulement souterrain et du volume des lacs. Notre approche, qui ne demande que très peu de données, est facilement transférable à d'autres systèmes de zones humides tropicales reculées, ce qui est utile dans les régions où le suivi hydrométrique est rare.

Mots clefs Pantanal ; saison sèche ; isotopes stables de l'eau ; chlorures ; bilan hydrique ; interactions des eaux souterraines et superficielles

1 INTRODUCTION

Wetlands are key components within the hydrological cycle due to their role in reducing floods, recharging groundwater and increasing low flow (Bullock and Acreman 2003). One of the largest and most pristine freshwater wetlands in the world is the Pantanal,

located in Brazil, Bolivia and Paraguay and covering an area of approximately 140 000 km² (Hamilton *et al.* 1996). Declared as a Ramsar Site of International Importance by the Ramsar Convention (Ramsar Convention 2011), as a Biosphere Reserve by UNESCO and parts of it as a World Heritage Site

(Hamilton 2002), the Pantanal is one of the world's most biodiverse ecosystems, with hundreds of species of mammals, birds, reptiles and plants (Carter *et al.* 2004). Nevertheless, many aspects of this important landscape remain unclear (Junk 2002). The Pantanal is still in a rather pristine state (Da Silva 2000). However, human activities such as dam construction threaten the seasonal flood pulse (Zeilhofer and de Moura 2009). Hydrology can be seen as the major driver in wetlands (Junk 2002) and is linked to the ecological integrity of the ecosystem (Da Silva 2000). To protect this ecosystem, basic and applied research for sustainable management is necessary (Junk 2002) since research there is still in its infancy (Da Silva 2000).

The Pantanal is a remote area, markedly heterogeneous in its structure, consisting of large rivers and their deltas and thousands of lakes that are interspersed with grassland, forests and savannah (Por 1995). These floodplain lakes are dynamic systems, which are connected to the regional hydrological cycle through the inflow and outflow of surface water, groundwater, precipitation and evaporation (Rozanski *et al.* 2000). The quantification of such water balance fluxes is fundamental for understanding the regional hydrology (Rozanski *et al.* 2000).

For decades, water balance studies with environmental tracers have been used for the determination of the hydrological functioning of lakes (e.g. Gat 1981, Gonfiantini 1986, Rozanski *et al.* 2000). Such water balance tracer approaches typically involve the analysis of stable water isotopes (Gibson *et al.* 1993). Isotopic analysis of water has proven to be one of the most important tools for investigating the water budget of water bodies (Matsuyama *et al.* 2005) and is indispensable for water balance studies in fairly inaccessible areas (Turner and Townley 2006). Insight into lake hydrological processes can be gained even with only a small number of samples (Rozanski *et al.* 2000). The sampling is relatively easy and, therefore, suitable for remote areas. However, few studies have been undertaken to use stable water isotopes for the investigation of hydrological processes in wetlands (Clay *et al.* 2004).

The methods adopted in water balance studies that quantify lake water balance estimates are usually based on the kinetic fractionation model of Craig and Gordon (1965). Gonfiantini (1986) discusses the most important results of lake studies with stable isotopes. Reviews by Gat (1995) and Froehlich *et al.* (2005) highlight isotopes in lake studies from

a historical perspective. Recent water balance research focuses primarily on North America (e.g. Gibson *et al.* 1993, 1996, 2002, Gibson and Edwards 2002, Brock *et al.* 2007, 2008, Wolfe *et al.* 2007, Yi *et al.* 2008, Brock *et al.* 2009, Stets *et al.* 2010, Turner *et al.* 2010). Jolly *et al.* (2008) reviewed groundwater–surface water interactions in arid and semi-arid wetlands. However, the application of water balance tracer approaches in tropical wetland regions, such as in South America, has been scant. The water balance of tropical floodplain lakes in the Amazon basin was studied by Martinelli *et al.* (1989), who used isotopic compositions to differentiate between waters originating from inundations, from overflowing rivers and residual waters from previous cycles of lakes. However, corresponding water balance studies for the Pantanal are non-existent. Moreover, there is insufficient information available about the Pantanal water bodies for calculating water balance. Despite its hydrological importance, the Pantanal is one of the least known regions in the world (Por 1995). Hydrological studies in this wetland are rarely found due to difficulties in accessing the area and the fact that conventional methods for quantifying water balance fluxes are impractical and expensive, therefore inappropriate, and must be developed further for such large and remote ecosystems (Yi *et al.* 2008).

Therefore, the objectives of this study were to understand and compare the hydrological processes of different water bodies in the Pantanal dry season. In this context, groundwater–surface water interactions were assessed focusing on the role of groundwater as a source of inflow to the floodplain lakes during the dry season. In addition, the components of the water balance of different lakes were simulated with a simple model of stable water isotopes, Cl⁻ concentrations and meteorological data. The aim was to determine the whole water balance for these remote floodplain lakes. To our knowledge, this study is the first to quantify water balance components with stable water isotopes in the Pantanal wetland.

2 METHODS

2.1 Study area

The Pantanal is located between 16–20°S and 54–58°W in the centre of South America (Fig. 1). Most of the wetland area lies within the Brazilian states of Mato Grosso and Mato Grosso do Sul. Smaller parts

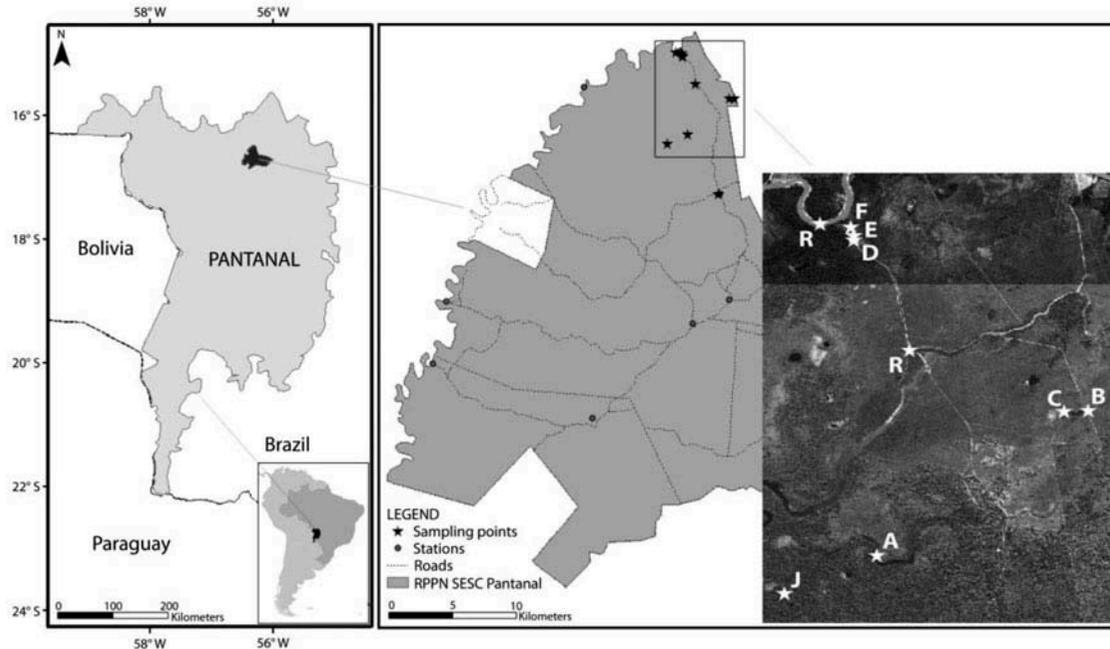


Fig. 1 Location of Pantanal wetland and sampling points within the RPPN SESC Pantanal.

can be found in Bolivia and Paraguay (Heckman 1998). The Pantanal is an inland river delta (Bullock and Acreman 2003) in a depressed area within the Upper Paraguay River basin. It forms a narrow panhandle where the Paraguay River and a great number of its tributaries gather from the surrounding plateau. Therefore, the Pantanal essentially consists of the alluvial fans of its main rivers, standing water bodies on these fans (e.g. wetland lakes) and a network of deltas of streams, which cross the whole wetland (Por 1995).

The Pantanal receives water from the highlands comprising an area almost 300 000 km² as well as from rainfall directly on the floodplain (Ponce 1995). Therefore, seasonal flooding, which is the most important hydrological and ecological event in the region (Girard 2011), may be of pluvial and/or fluvial origin (Pott and Pott 2004). During the rainy season, the floodplain is usually covered continuously with water (Por 1995), and the wetland acts like a large reservoir by storing surface and subsurface water for up to several years (Ponce 1995). During the dry season, smaller rivers and lakes as well as low lying sites dry up completely or are transformed to lentic water bodies (Girard 2011). However, a small part of the wetland retains standing water throughout the year (Heckman 1998).

Research took place within the Reserva Particular do Patrimônio Natural (Private reserve of national heritage) SESC (Serviço Social do Comércio, Commerce

Social Service) Pantanal, henceforth called RPPN SESC Pantanal, located in the northern part of the Pantanal (Fig. 1). The reserve is situated between 16°28' and 16°50'S and 56°00' and 56°30'W in the municipality of Barão de Melgaço, in the state of Mato Grosso, which is located in the central-western region of Brazil. The RPPN SESC Pantanal covers an area of about 106 307 ha and its elevation varies from 101 to 117 m a.s.l. (Ramsar Convention 2011).

The climate is characterized by dry autumns and winters, with heavy rainfall during the wet springs and summers (Hasenack *et al.* 2003). According to the Köppen-Geiger climate classification, the study area belongs to the type Aw, which is defined as equatorial savannah with dry winters and a minimum winter precipitation of below 60 mm/month (Kottek *et al.* 2006). The average precipitation in the rainy season (November–April) amounts to 1500 mm/year (Ramsar Convention 2011). From about mid-April until October, the dry season is characterized by very little precipitation, with less than 40 mm/month (Hasenack *et al.* 2003). The annual average temperature is 25.7°C (Ramsar Convention 2011) and afternoon temperatures during the dry season often reach the mid-40s°C (Heckman 1998). According to Girard (2011), the mean monthly relative humidity in the northern Pantanal ranges from 85% during the rainy season to below 60% when the floodplain is dry. In the dry season, relative humidity during daytime can drop below 30% (Heckman 1998). During at least six

months of the year, evapotranspiration is greater than precipitation. Generally, evapotranspiration amounts to between 1100 and 1400 mm/year (Girard 2011). The Pantanal soils are closely related to the type of sediments that are deposited, which in turn are a consequence of the type of source materials. Several alluvial deltas have led to the deposition of sediments, which can vary in their sand, mineral or nutrient content. These combinations cause different behaviour of water in the soil profile (Couto and De Oliveira 2011).

2.2 Water sampling and analysis

Fifty-five water samples were collected in the northern RPPN SESC Pantanal to assess groundwater–surface water interactions, as well as recharge conditions, and to provide information on source water flowing into the lakes. Samples were collected from two rivers and seven shallow lakes during the three-month study period ($n = 50$ water samples in total, with six to nine samples collected from each water body), as well as from groundwater ($n = 4$) and rainfall ($n = 1$, due to sparse rainfall during the dry season). All samples were taken within four measuring periods between August and October 2008 with intervals ranging from one day to three weeks. Within this context, water body samples were taken during all four sampling campaigns, the groundwater samples were taken during the first three campaigns and the single rain sample was taken during the third campaign. The surface water samples were collected at about 1 m from the shoreline and at a depth of about 10 cm.

All lakes studied are abandoned channels and are river-isolated in the dry season. Their morphometric parameters were estimated during the dry season 2008 (Table 1) derived from field observations at the beginning of the 3-month period. Unfortunately, it was not possible to estimate the size of Lake C due to limited accessibility.

All water samples were analysed for stable isotopic composition of water ($\delta^{18}\text{O}$, $\delta^2\text{H}$) and Cl^- concentrations. The analysis of Cl^- was conducted with an ion chromatograph (IC) (Dionex model DX 500, Sunnyvale, California). All samples were filtered using a 0.45- μm membrane and analysed in the same measuring series. The precision for the anion concentrations is approximately 5%. The analysis of $\delta^{18}\text{O}$ was conducted with a Delta S (Finnigan Mat, Bremen, Germany) isotope ratio mass spectrometer (IRMS) with a precision of 0.2‰ and that of $\delta^2\text{H}$ by a liquid water isotope analyser using tunable diode laser absorption spectrometry (TDLAS) with a precision of 0.6‰ (Los Gatos, Mountain View, California). Isotopic composition is calculated in per mil (‰) as the difference between the measured isotopic ratio and the Vienna Standard Mean Ocean Water (VSMOW), divided by this standard (Clark and Fritz 1997):

$$\delta_{\text{sam}} = \left(\frac{R_{\text{sam}} - R_{\text{ref}}}{R_{\text{ref}}} \right) \quad (1)$$

with:

$$R = \frac{^{18}\text{O}}{^{16}\text{O}} \quad (2)$$

or:

$$R = \frac{^2\text{H}}{^1\text{H}} \quad (3)$$

where δ_{sam} is the isotopic composition of a sample and R the measured ratio of less- and more-abundant isotopes in the collected water samples (R_{sam}) and the VSMOW (R_{ref}).

The δ notation allows differentiation of the origins of water (local precipitation or groundwater) and identification of water that has undergone an isotopic

Table 1 Sampling sites with morphometric characteristics (n.a.: not available) and coordinates of sampling points.

Sampling site	Latitude S	Longitude W	Length (m)	Width (m)	Area (m ²)	Mean Depth (m)	Volume (m ³)
Lake A	16°32'47.3	56°18'09.9	3200	120	181 000	2	362 000
Lake B	16°31'17.2	56°16'09.8	500	50	26 000	0.5	~13 000
Lake C	16°31'18.0	56°16'23.5	n.a.	n.a.	n.a.	0.3	n.a.
Lake D	16°29'31.7	56°18'20.7	50	10	300	<0.5	~120
Lake E	16°29'29.2	56°18'20.8	50	10	500	<0.5	~250
Lake F	16°29'24.1	56°18'22.5	20	6	120	<0.5	~60
Lake J	16°33'11.7	56°19'00.0	390	35	10 000	2	20 000

enrichment due to evaporation (surface water or groundwater). Stable isotopes show different fractionation during phase change due to different nuclear masses. Therefore, water of a particular environment has a characteristic isotope signature. Dansgaard (1964) formulated five different fractionation effects, with the continental effect the most important for this study area. It results in a depletion of isotopic compositions with increasing distance from the coast (1‰ per 100 km) and can therefore explain isotopic compositions below 0. Craig (1961) discovered that, on a global scale, worldwide measurements of precipitation plot on the Global Meteoric Water Line (GMWL):

$$\delta^2\text{H} = 8 \cdot \delta^{18}\text{O} + 10 \quad (4)$$

where δ values describe the isotopic composition of $\delta^{18}\text{O}$ and $\delta^2\text{H}$, and 10 is the deuterium excess. Local rainfall of a particular region plots on a Local Meteoric Water Line (LMWL), which deflects from the GMWL by another deuterium excess factor. Variations in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values that result in a deviation from the GMWL are mainly influenced by rain-out or evaporation (Dansgaard 1964). During the water evaporation process there is additionally a kinetic effect causing variation in the fractionation of $\delta^{18}\text{O}$ and $\delta^2\text{H}$. This leads to an enrichment of heavy isotopes in the residual liquid as evaporation moisture flux is depleted in $\delta^{18}\text{O}$ and $\delta^2\text{H}$. This evaporatively-enriched water plots on a local evaporation line (LEL) (Gat 1995).

For the RPPN SESC Pantanal, a LEL was constructed according to the approach described in Barnes and Allison (1983 cited in Brock *et al.* 2007).

2.3 Assessment of water balance components

On the basis of the mass conservation law, the water balance tracer approach is (Rozanski *et al.* 2000):

$$\begin{aligned} \delta_L \frac{dV_L}{dt} + V_L \frac{d\delta_L}{dt} = & \delta_{IS}I_S + \delta_{IG}I_G + \delta_P P \\ & - \delta_{OS}O_S - \delta_{OG}O_G \\ & - \delta_E E \end{aligned} \quad (5)$$

where δ values in ‰ represent the composition of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ for all components of the lake system; V_L is the lake volume with the isotopic composition of the lake δ_L ; I_S and I_G are the inflow of surface water and groundwater, with their isotopic compositions δ_{IS}

and δ_{IG} , respectively; P is precipitation, with its isotopic signature δ_P ; O_S and O_G are the outflow of surface water and groundwater with their isotopic compositions δ_{OS} and δ_{OG} , respectively; and E is evaporation, with the isotopic signature of evaporative flux δ_E .

For the application of mass balance equations, the following assumptions were made (Rozanski *et al.* 2000):

- The isotopic composition of the surface outflow is identical to the isotopic composition of the lake water ($\delta_{OS} = \delta_L$).
- The lake has to be well mixed, so the isotopic composition of the groundwater outflow is also equal to the isotopic composition of the lake water ($\delta_{OG} = \delta_L$).

Since not all parameters of the water balance equation are known, a simple mass balance model MINA TrêS (Model for INflow Assessment, trêS: Portuguese for three simulations) was developed to estimate the missing groundwater inflow component of the lake water balances. Conceptually, the model consists of a reservoir that receives a constant groundwater inflow and outflow but with time varying lake volumes and evaporation rates. Attributing a concentration to each flow component and the storage, solute or isotope mass balance can be expressed by:

$$\begin{aligned} c_{L(t)}V_{L(t)} = & c_{L(t-\Delta t)}V_{L(t-\Delta t)} + c_{I(t)}I \\ & - c_{L(t)}O - c_{E(t)}E(t) \end{aligned} \quad (6)$$

where V_L is the predicted lake volume at time t , c_I is the concentration of inflow I , O stands for outflow with the predicted concentration of the lake c_L , and c_E is the composition of evaporating water, where E stands for evaporation. This evaporative flux of water, c_E , is given in ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$, while c_E (chemical flux transported by evaporation) for Cl^- is equal to zero. The change in concentration relates to the volume and concentration of the lake at the time step $t - \Delta t$. Compositions are given in ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ and in mg/L for Cl^- concentrations.

With respect to the water balance equation, there was no need to consider precipitation as part of inflow during the study period, since there was only a single rainfall event (<5 mm), the term $\delta_P P$ drops out of equation (5), and is therefore not considered in the model equation (6). Furthermore, the waterbodies do not have river inputs or outlets in the dry season, so the terms $\delta_{IS}I_S$ and $\delta_{OS}I_O$ drop out of equation (5) and are

also not considered in the model equation (6). All lakes studied were cut off from their main river channels during the study period, without the presence of surface inflow and outflow during the dry season. Therefore, only subsurface inflow to the lakes needs to be determined, because it is assumed that outflow rates are equal to inflow rates. This assumption should be in correspondence with the natural wetland condition, as the wetland shows only a very small terrain gradient.

The MINA TrêS model uses initial conditions of stable water isotopes and Cl⁻ concentrations in the lake, together with estimates of daily evaporation in mm/d (ET) and chemical fluxes lost by ET for the simulation. The variable parameters are the inflow concentration (stable water isotope flux and Cl⁻ concentration) and the initial lake volume (depth), as these are adjusted until a good model fit is achieved. This is done by minimizing the sum of absolute differences between observed and simulated lake Cl⁻ concentrations and isotopic compositions for all water bodies. Therefore, the use of three different observed tracers enables all the unknown parameters to be constrained.

Note that due to different fractionation during the evaporation process, measured $\delta^{18}\text{O}$ and $\delta^2\text{H}$ compositions in the lakes provide independent information. However, regarding the inflow compositions, a linear dependence between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ can be assumed (equation (4)), and $\delta^2\text{H}$ of c_{I} was not allowed to vary as an independent variable. A minimum Cl⁻ concentration for the inflow was set to 0.97 mg/L, which is equal to the lowest measured concentration in the groundwater. That way, MINA TrêS is able to simulate the concentration c_{L} of a lake (stable isotope fractions of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ and Cl⁻ concentration) at a daily time step (subscript t). Even though the model includes some assumptions that may add some uncertainty to the calculations (e.g. the assumption of conservative Cl⁻), the combined use of three different tracers (Cl⁻, $\delta^{18}\text{O}$ and $\delta^2\text{H}$) should better balance the different error sources.

The lake volume, inflow, outflow and evaporation are calculated in mm/d for each day between the first and last water sample of each water body, which is defined by the study period. The values for initial lake compositions of $\delta^{18}\text{O}$ and $\delta^2\text{H}$, as well as Cl⁻ concentrations, correspond with the values of the first water sample. The lake depth at daily time steps is expressed in mm, which corresponds to the lake volume if surface area changes can be neglected. For small changes in water level, changes in lake areas are negligible as in the case of the application of MINA TrêS with daily time steps.

Evaporation losses, E , were determined by implementation of the Penman-Monteith approach for open water (Shuttleworth 1993) and were calculated for each day of the study period. For each water body, they were summed for the time period between each water sample on the basis of meteorological data measured in the RPPN SESC Pantanal.

The chemical flux transported by evaporation, c_{E} , is 0 in the case of Cl⁻, as there is no loss of Cl⁻ through evaporation. For stable isotopes, c_{E} values were calculated based on equation (7), where c_{E} corresponds to ε . In the literature, c_{E} is called the total isotopic separation factor; therefore, the common symbol ε is used (Gonfiantini 1986):

$$\varepsilon = \varepsilon_{\text{K}} + \varepsilon^* \quad (7)$$

where ε_{K} and ε^* are the separation factors for kinetic and equilibrium fractionation (in ‰), respectively, and ε_{K} is calculated as a function of relative humidity (h) for ^{18}O by:

$$\varepsilon_{\text{K}} = 14.2 \cdot (1 - h) \quad (8)$$

and for ^2H by:

$$\varepsilon_{\text{K}} = 12.5 \cdot (1 - h) \quad (9)$$

These ε_{K} values can be seen as representative of typical lake evaporation (Gonfiantini 1986). Data for relative humidity, h , were obtained from a climate station tower within the RPPN SESC Pantanal, at 15-min intervals. Mean values between 05:00 and 20:00 h, when evaporation mainly takes place, were derived for the study periods. Then, ε^* can be calculated by (Craig and Gordon 1965):

$$\varepsilon^* = (\alpha_{\text{v/l}} - 1) \quad (10)$$

where $\alpha_{\text{v/l}}$ is the fractionation factor of liquid water (subscript l) with respect to water vapour (subscript v). Majoube (1971) determined the temperature dependence of $\alpha_{\text{v/l}}$ experimentally by:

$$\begin{aligned} \ln^{18} \alpha_{\text{v/l}} &= -\ln^{18} \alpha_{\text{l/v}} \\ &= -2.0667 \times 10^{-3} + \frac{0.4156}{T_{\text{W}}} - \frac{1.137 \times 10^3}{T_{\text{W}}^2} \end{aligned} \quad (11)$$

for ^{18}O and for ^2H by:

$$\begin{aligned} \ln^2 \alpha_{v/l} &= -\ln^2 \alpha_{l/v} \\ &= -52.612 \times 10^{-3} + \frac{76.248}{T_W} - \frac{24.844 \times 10^3}{T_W^2} \end{aligned} \quad (12)$$

where T_W is the absolute water temperature in K. The mean value for each water body was measured on multiple days over the whole time period of the field study with a multiprobe (MS5 Hydrolab, Loveland, Colorado, USA).

3 RESULTS

3.1 Isotope analysis

The isotopic composition of water samples collected during the study ranged from -5.94 to 14.75‰ for $\delta^{18}\text{O}$ and from -38.8 to 58.9‰ for $\delta^2\text{H}$ (Fig. 2 and Appendix, Table A1). The GMWL could be used as the LMWL because isotopic composition of precipitation obtained from the GNIP (Global Network of Isotopes in Precipitation 1961–1987) Station in Cuiabá (100 km north of the study site) generally plotted close to the GMWL; its mean deviation from the GMWL was 0.6‰ for $\delta^{18}\text{O}$ and 4.8‰ for $\delta^2\text{H}$.

In relation to precipitation obtained from the GNIP Station in Cuiabá, the rainwater sample of the RPPN SESC Pantanal (21 September 2008) plotted close to the GMWL and within the heaviest range of isotopic compositions (Fig. 2). The δ values of

groundwater ranged between -5.9 and -4.8‰ for ^{18}O and between -38.8 and -36‰ for ^2H , showing only a very slight enrichment, or even plotted on the GMWL within the error range. All river water samples indicated very slight temporal enrichments: within their error range, they plotted very close to the GMWL and fairly close to the intersection of GMWL and LEL. The range of $\delta^{18}\text{O}$ was between -5.35 and -1.86‰ and $\delta^2\text{H}$ values ranged between -36.32 and -19.43‰ . Thus, the δ values of the river water encountered the least enrichment of all water samples and the smallest difference of δ values between the first and the last water sample, reflecting a groundwater source that has undergone evaporation enrichment. All lake water samples encountered stronger isotopic enrichment towards higher δ values. The slope of the LEL was calculated as 4.6, which is consistent with typical slopes of evaporating water for central South America (Martinelli *et al.* 1989). Thus, the generalized LEL equation for the study area is:

$$\delta^2\text{H} = 4.6\delta^{18}\text{O} - 9.51 \quad (13)$$

Considering the temporal evolution of the isotopic compositions and the Cl^- concentration of Lake A as an example, both stable isotopes indicated an evaporation enrichment of the lake water (Fig. 3). The Cl^- concentrations remained more or less stable within the measurement uncertainty of 5%. In

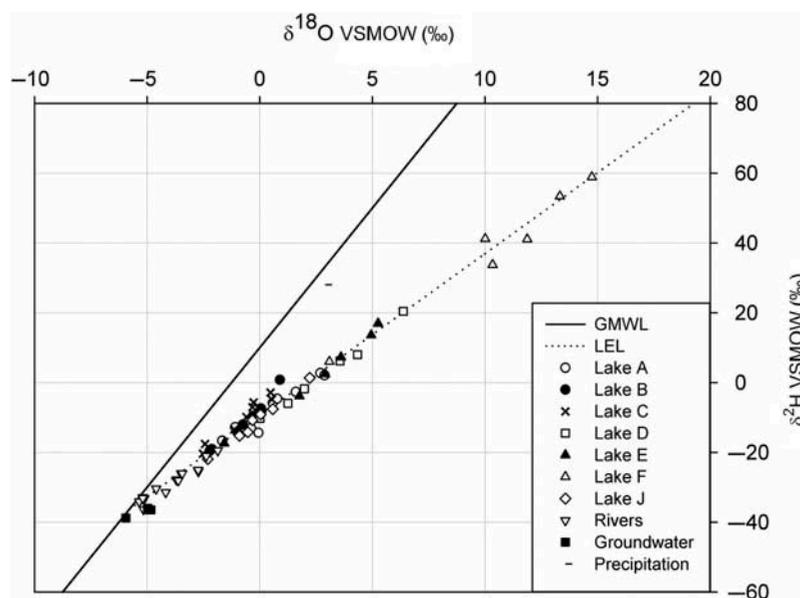


Fig. 2 The $\delta^{18}\text{O}$ – $\delta^2\text{H}$ relationship for all water samples collected from the RPPN SESC Pantanal including the Global Meteoric Water Line (GMWL) and the Local Evaporation Line (LEL).

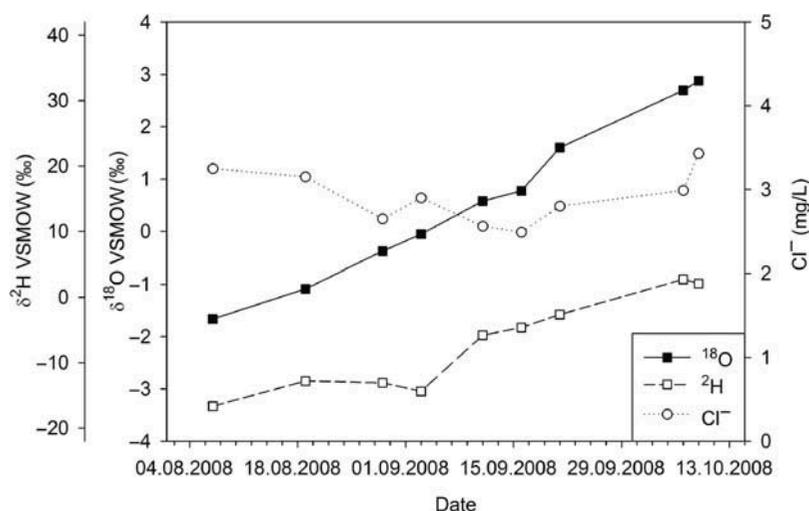


Fig. 3 Evolution of $\delta^{18}\text{O}$, $\delta^2\text{H}$ and Cl^- concentrations of Lake A over the time period of field study.

comparison with the measured Cl^- concentration of the rainfall sample, which was 11.41 mg/L, the Cl^- concentrations of the lakes were very low, except for Lake C. The mean values were 2.91, 1.7, 3.65, 2.2, 3.94 and 2.56 mg/L, for lakes A, B, D, E, F and J, respectively. Only Lake C showed a considerable high mean Cl^- concentration of 31.18 mg/L.

Furthermore, all samples of the water bodies plotted close to the LEL. With respect to their temporal evolution of the isotopic compositions, the first sample encountered a slight enrichment, the second a stronger enrichment relative to the first, and so on; therefore, the last sample of each water body exhibited the strongest enrichment in δ values. All water bodies showed this general trend in stable water isotope compositions. The river water also showed a temporal enrichment, but very slightly and always less enriched than the lake water, while the lakes revealed strong differences in the degree of evaporative enrichment. For Lake A, the absolute difference of the isotopic composition of $\delta^{18}\text{O}$ between the first and last water samples was 4.5‰, for Lake B it was 3.12‰, for Lake C 3.06‰, and for Lake J 4.53‰; for lakes D, E and F, it was 6.36‰, 6.83‰ and 11.65‰, respectively.

Relating evaporation enrichment, defined as the absolute difference of the isotopic composition of $\delta^{18}\text{O}$ between the first and last water samples, to lake properties (width-to-length ratio of the lakes, surface area, water depth or water volume, in Table 1, Lake C not determinable), the width-to-length ratio showed the highest correlation ($R^2 = 0.8244$), while the other correlations had R^2 values less than 0.0625.

3.2 Application of MINA TrêS

The MINA TrêS model was used to model the daily lake concentrations continually for all lakes studied. The predictions agreed with underlying assumptions for all water bodies with the exception of Lake J, where incompatible numerics were simulated. For the remaining six lakes, correlations between measured and modelled isotopic composition (R^2) ranged between 0.4 and 0.99, with a mean R^2 value of 0.8 (Table 2). Correlations between measured and modelled Cl^- concentrations were good for four lakes (B, D, E and F, with R^2 values between 0.56 and 0.82), but poorer for two of the lakes (A and C, with R^2 of 0.01 and 0.21, respectively) (Table 2). Figure 4 displays measured and modelled data for Lake D, as an example.

From the calibration of MINA TrêS, estimates of inflow rates were derived for all the lakes, with their isotopic compositions and Cl^- concentrations. The initial lake volume, daily inflow, total inflow for the period of study (7 August–10 October 2008) and

Table 2 Coefficient of determination (R^2) between modelled and measured inflow concentrations for $\delta^{18}\text{O}$ and Cl^- .

Waterbody	R^2 between measured and modelled	
	Cl^- concentrations	$\delta^{18}\text{O}$ values
Lake A	0.01	0.99
Lake B	0.59	0.99
Lake C	0.21	0.83
Lake D	0.67	0.96
Lake E	0.82	0.61
Lake F	0.56	0.4

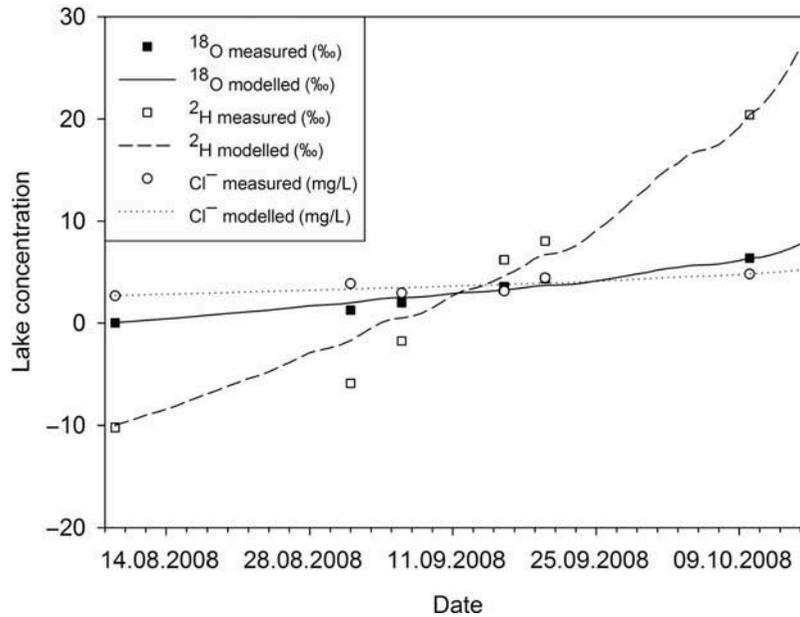


Fig. 4 Modelled and measured δ values of ^{18}O and ^2H and Cl^- concentrations of Lake D, where lines show modelled data and symbols show measured data.

inflow concentrations for all lakes are presented in Table 3. The initial lake volumes of lakes D, E and F were equal to the estimated lake volumes (see Table 1), but initial volumes of lakes A, B and C were adjusted during the calibration due to the iterative adaption of this parameter.

Concerning the inflow rates, there was a strong spatial variation between the different waterbodies. Groundwater inflow rates ranged between 9.7 and 18.5 mm/d, resulting in a total inflow during the study period of between 662.2 (Lake D) and 1278.1 mm (Lake C). The Cl^- inflow concentrations varied between 0.97 (Lake A) and 23.9 mg/L (Lake C). The Cl^- concentrations of all the lakes except for Lake C matched the measured Cl^- concentrations of groundwater. The isotopic signatures of modelled inflow values for $\delta^{18}\text{O}$ ranged from -5.53 to 20.31‰ and the Cl^- concentrations ranged from 0.97 to 23.89 mg/L. The $\delta^{18}\text{O}$ values of lakes B, C, D

and E matched the measured isotopic composition of the groundwater and river samples, and ranged between -5.5 and -4.2‰ . Lakes A and F had an inflow component with a more enriched isotopic signature of $\delta^{18}\text{O}$; for Lake A, the $\delta^{18}\text{O}$ value was modelled with 1.38‰ and for Lake F with 0.84‰ .

3.3 Final water balance

The final water balance was derived from calculated evaporation losses after Shuttleworth (1993) and the application of MINA TrêS (Table 4). On the basis of the difference in lake hydrology (Table 4), three categories of lake were identified.

In the first group, comprising lakes C and F, comparable proportions of water balance components were found. The water volume of these water bodies was lowest and evaporated almost entirely during the

Table 3 Initial lake volume (given as depth in mm due to the assumption that lake cross-sectional area is constant), inflow rates, as well as isotopic compositions and Cl^- concentrations of inflow simulated by MINA TrêS for the study period (7 August–10 October 2008).

Sampling site	Initial lake volume (mm)	Inflow rate (mm/d)	Total inflow (mm)	$c_1 \delta^{18}\text{O}$ (‰)	$c_1 \delta^2\text{H}$ (‰)	$c_1 \text{Cl}^-$ (mg/L)
Lake A	1800	11.1	776.2	1.4	21.1	1.0
Lake B	550	13.7	950.3	-5.5	-34.2	1.4
Lake C	400	18.5	1278.1	-4.8	-28.3	23.9
Lake D	500	9.7	662.2	-4.9	-29.2	2.6
Lake E	500	11.2	760.9	-4.2	-23.6	1.7
Lake F	383	16.2	1103.9	0.8	16.7	2.8

Table 4 Final water balance of the lakes for the field study period, 7 August–10 October 2008 (V_i : initial lake volume, V_p : volume at the end of the simulation period, where V_i and V_p are given as depths in mm due to the assumption that lake cross-sectional area is constant), I : inflow, E : evaporation, O : outflow).

Water body	V_i (mm)	V_p (mm)	I (mm)	E (mm)	O (mm)
Lake A	1800	1412	776	388	776
Lake B	550	167	950	383	950
Lake C	400	17	1278	383	1278
Lake D	500	122	662	378	662
Lake E	500	122	761	378	761
Lake F	383	5	1104	378	1104

study period. Evaporation amounted to 96% and 99% of initial lake volume for Lake C and Lake F, respectively. Therefore, the model predictions that the lakes would almost dry out by the end of the modelling period compared well with field observations. Inflow rates for lakes C and F were simulated as 1278 and 1104 mm, respectively, replacing the total initial water volume more than three times for Lake C and almost three times for Lake F.

The second group consisted of lakes B, D and E. For these lakes, evaporation amounted to 70% of initial lake volume for Lake B and 76% for lakes D and E. Their water volume was renewed around 1.5 times during the study time.

The third group consisted only of Lake A, where evaporation amounted to 22% of initial lake volume. Less than 50% of the initial water volume was replaced by inflow. Except for Lake A, the total initial water volume was replaced by groundwater exchange at least once during the study period.

4 DISCUSSION

4.1 Isotope analysis

The climatological conditions were very similar for all water bodies studied, as the study area is less than 10 km by 10 km (Fig. 1), with less than 5 mm of precipitation during the study period. Potential isotopic enrichment in lake water due to evaporation depends not only on the meteorological conditions, but also on the hydrological balance of the lakes. Groundwater inflow can significantly control the isotopic composition of a lake too and, therefore, it can be an important contribution to the lake balance (Froehlich *et al.* 2005). Relating evaporation enrichment to lake properties (except for Lake C, where morphometric parameters were not determinable), the width-to-length ratio showed the highest correlation. This indicates that orientation of the lakes to the local

groundwater flow field might play an important role. However, with the data presently available, there is insufficient information to describe this correlation with a causal mechanism.

In general, for all water bodies studied, the results of the isotope study correspond well with the general behaviour of water samples from different locations within the tropics. The enriched isotopic compositions of all of the lakes studied reflect the significant role of evaporation in the water balance of the lakes (Rozanski *et al.* 2000, Nyarko *et al.* 2010).

In terms of the isotopic composition of the groundwater samples, only slight enrichment occurred, indicating small evaporation effects. The range of values indicates that the groundwater was recharged from precipitation falling during the rainy season (December–March). Even though only four sampling campaigns could be conducted within the three months of the dry season, other studies have shown that isotope compositions in groundwater are stable over short time periods. Therefore, it is considered sufficient to sample groundwater once in every season to isotopically characterize the groundwater inflow component of the lakes (Rozanski *et al.* 2000).

Among all of the sampled water bodies, the river water exhibited the least enriched $\delta^{18}\text{O}$ values (Fig. 2). This corresponds with the typical behaviour of lower isotopic compositions of river water compared to lake water (Fontes *et al.* 1979), the latter being usually subject to isotopic enrichment by evaporation due to longer residence times in the lakes.

Particularly notable are the very high δ values of Lake F. In this case the very dry environment in the Pantanal during the dry season, in combination with the small water volume of the lake, seems to be responsible for such a strong enrichment. In North Africa, Fontes and Gonfiantini (1967) observed $\delta^{18}\text{O}$ values of +31.3‰ and $\delta^2\text{H}$ values of +129‰. They state that this very strong enrichment in heavy

isotopes is probably due to the very arid environment in that area. Another reason for the high δ values of Lake F could lie in a lower outflow than inflow, or even a zero outflow; hence, this lake would act more as a terminal lake.

Except for lakes A and F, $\delta^{18}\text{O}$ inflow compositions correspond well with $\delta^{18}\text{O}$ values of groundwater. This indicates that lakes B, C, D and E are all fed by local groundwater reservoirs. Lakes A and F show more enriched isotopic compositions of groundwater inflow. Their isotopic compositions suggest that inflow water could have contributed water with enriched δ values, which could serve as well as a useful indicator for the connection between surface water bodies in the wetland (Froehlich *et al.* 2005).

4.2 Application of MINA TrêS

The coefficients of determination for the modelled $\delta^{18}\text{O}$ inflow compositions indicate that MINA TrêS acceptably simulates the inflow component of the water balance for all lakes with the exception of Lake J. For Lake J, it seems that the model structure is too simple (e.g. assumptions of constant inflow rates and concentrations or equal inflow and outflow rates) to adequately simulate isotope or Cl^- enrichment.

In general, the results show a better performance for ^{18}O than for Cl^- . The concentration of Cl^- has often been used as a conservative tracer for the calculation of water fluxes. However, Svensson *et al.* (2012) found that on sites with low Cl^- deposition, Cl^- may not exhibit conservative behaviour, such as those located far inland away from marine influences. One mechanism suggested is related to vegetation dynamics, with Cl^- sequestration resulting from increased biomass growth (Lovett *et al.* 2005). Our study site in the Pantanal is located in an inland area that is recovering from a long period of ranching, with vegetation dynamics recovering from grazing since 1998. We do not have sufficient data to determine if Cl^- is conservative or non-conservative for the study area. Nevertheless, the aforementioned characteristics combined with sub-optimal model performance for Cl^- suggests the Cl^- balance as a future line of investigation.

A minimum Cl^- concentration for the inflow was set to 0.97 mg/L, which is equal to the lowest measured concentration in the groundwater samples. This assumption is in accordance with a study by Cross *et al.* (2001) where the input value represents a mean of measured inflow concentrations. Except for lakes

A and C, modelled Cl^- inflow concentrations match the measured concentrations in the groundwater samples, indicating inflow from local groundwater. The origin of the high Cl^- concentration of Lake C cannot only be explained by dry deposition and the balance between evaporation and precipitation. Dispersion of Cl^- by wind from salt marshes (Barbiéro *et al.* 2002), inflow of high-chloride groundwater (Hem 1992), accumulation of Cl^- from previous dry seasons, or measurement errors might be reasons for the high Cl^- concentrations of Lake C.

4.3 Final water balance

For all water bodies studied, the modelled inflow rates and calculated evaporation losses show that these fluxes play a significant role in the water balance, even though there are noticeable differences among the lakes. For all lakes studied, groundwater inflow is larger than evaporation in their water balance (Table 4). This is consistent with Jolly *et al.* (2008), who found that wetlands are strongly controlled by groundwater. In addition, lakes that could not be considered in this study showed a differing behaviour, since some dried out completely, while others kept their water levels more or less stable during the time period of field study. These field observations support the findings that groundwater plays an important role in the water balance of lakes.

The Penman-Monteith approach is the most commonly used method to determine potential evaporation on a physical basis and is sufficient to get a first approximation for estimating evaporation losses (Dingman 2008). Kohler and Parmele (1967) successfully applied a generalized version of the Penman equation and state that the method delivers good results if evaporation is calculated on a daily basis. Unfortunately, the data availability in the RPPN SESC Pantanal does not permit a more detailed approach.

In terms of the relative differences in water balance components, the lakes in the first group (lakes C and F) are the most dynamic. This is due to the small water volumes of those lakes. Even if almost the whole water volume is evaporated, inflow dominates their water balance. The initial water volume is replaced more than three times in the case of Lake C, and nearly three times for Lake F. Regarding absolute differences in groundwater inflow rates, lakes C and F show highest inflow rates, and perhaps counter-intuitively, they also had the lowest water volumes at the end of the study period (Table 4).

The second group of lakes (lakes B, D and E) are intermediary with respect to the water balance components. The quantity of groundwater inflow is double that of evaporation losses for lake D and E and three times higher in the case of Lake B. At the end of the modelling period, these lakes still have a considerable water volume.

For the third group (Lake A), both water balance processes of groundwater inflow and evaporation are present, but neither dominates the water balance (evaporation 22% and groundwater inflow 43%). Groundwater inflow (776 mm) is double the evaporation losses (388 mm). This lake is by far the largest of all the lakes studied and its inflow is relatively small, which influences the residence time of the water. Dense soils with low hydraulic conductivity due to depositional processes in the alluvia can be responsible for restricting the groundwater–lake flow pathway (Jolly *et al.* 2008).

5 CONCLUSION

Our study was aimed at identifying the dominant hydrological processes in the northern Pantanal wetland during the dry season. Stable water isotopes and hydrochemistry data served to provide qualitative and quantitative information regarding recharge conditions, evaporation enrichment of lake water, as well as groundwater inflow. It highlighted the differences between the wetland lakes studied.

We found that all lakes exhibited the same qualitative pattern of hydrological processes controlling their water balances, which is not surprising regarding their close proximity (study area ≤ 100 km²). However, quantitatively they differed significantly, even within a small area, and we found morphological differences in the lakes to be associated with groundwater inflow and lake volumes.

The MINA TrêS model adequately simulated groundwater inflow rates and $\delta^{18}\text{O}$ and $\delta^2\text{H}$ compositions, as well as Cl^- concentrations. On the basis of a simplified mass balance tracer approach, it provided quantitative estimates of water balance components, where prior information was scarce and hydrometric measurements are difficult to perform. Considering the simplicity of the model, it was capable of delivering promising results for a remote area with limited data availability and little *a priori* hydrological information. Limitations of this model were also evident due to the underlying assumptions.

Based on this approach, we can deduce that in the dry season groundwater inflow is a significant

process in the water balance of the lakes studied. The results showed that this groundwater inflow appeared to originate mainly from local groundwater without isotopic enrichment rather than from isotopically more enriched rivers. In addition, evaporative enrichment, which is controlled by the exchange between water volume and groundwater inflow, played a major role during the dry season. All lakes were closely connected to the groundwater and had significant groundwater inflow rates, which maintained lake water levels to a varying degree during the Pantanal dry season. Although there were differences in the groundwater inflow feeding the different lakes and its isotopic content, our study provides evidence that all these groundwater reservoirs were recharged from precipitation during the rainy season. The high exchange between groundwater and surface water may also determine the biogeochemical and ecological characteristics of these lakes, and potentially a large proportion of the lakes in the Pantanal. A comprehensive groundwater study including inflow measurements and flow direction as well as stable water isotope analysis of a larger number of lakes would serve to assess groundwater–surface water interactions on a larger scale in order to better understand this heterogeneous landscape.

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Appendix

Table A1 List of all water samples and measured compositions collected in the RPPN SESC Pantanal (the letter of water sample corresponds to the lake from which the sample was taken).

Water sample	Sampling date	$\delta^{18}\text{O}$ (‰)	$\delta^2\text{H}$ (‰)	Cl ⁻ (mg/L)
A1	07.08.2008	-1.67	-16.64	3.25
A2	19.08.2008	-1.09	-12.81	3.15
A3	29.08.2008	-0.38	-13.07	2.65
A4	03.09.2008	-0.05	-14.36	2.90
A5	11.09.2008	0.58	-5.79	2.56
A6	16.09.2008	0.77	-4.61	2.49
A7	21.09.2008	1.60	-2.63	2.80
A8	07.10.2008	2.70	2.73	2.99
A9	09.10.2008	2.88	2.09	3.43
B1	08.08.2008	-2.22	-19.60	1.51
B2	12.08.2008	-2.12	-19.06	1.41
B3	29.08.2008	-0.98	-14.30	1.48
B4	03.09.2008	-0.74	-11.99	1.42
B5	12.09.2008	-0.30	-9.54	1.90
B6	17.09.2008	0.06	-7.40	2.11
B7	18.09.2008	0.09	-7.97	1.59
B8	07.10.2008	0.90	0.79	2.16
C1	08.08.2008	-2.42	-17.58	19.02
C2	12.08.2008	-2.52	-20.36	11.38
C3	29.08.2008	-1.13	-13.44	12.03
C4	04.09.2008	-0.59	-9.86	27.95
C5	12.09.2008	-0.27	-5.70	33.14
C6	17.09.2008	0.49	-2.83	51.33
C7	18.09.2008	0.54	-4.80	80.18
C8	07.10.2008	-0.32	-7.03	14.44
D1	09.08.2008	0.01	-10.21	2.67
D2	01.09.2008	1.26	-5.91	3.87
D3	06.09.2008	1.99	-1.76	2.96
D4	16.09.2008	3.57	6.20	3.14
D5	20.09.2008	4.34	8.04	4.44
D6	10.10.2008	6.38	20.43	4.82
E1	09.08.2008	-1.57	-17.40	1.65
E2	01.09.2008	4.95	13.59	1.58
E3	06.09.2008	1.77	-3.88	1.99
E4	16.09.2008	2.91	2.45	2.38
E5	20.09.2008	3.61	7.27	2.12
E6	10.10.2008	5.26	16.94	3.47
F1	09.08.2008	3.10	6.03	3.33
F2	01.09.2008	10.34	33.74	3.79
F3	06.09.2008	11.88	41.04	2.68
F4	16.09.2008	13.33	53.40	4.52
F5	20.09.2008	14.75	58.87	3.83
F6	10.10.2008	10.01	41.16	5.50
J1	10.08.2008	-2.30	-21.92	2.12
J2	30.08.2008	-0.90	-15.22	2.18
J3	04.09.2008	-0.52	-14.19	3.20
J4	06.09.2008	-0.32	-10.74	2.22
J5	13.09.2008	0.05	-9.13	1.88
J6	19.09.2008	0.59	-7.59	4.90
J7	08.10.2008	2.23	1.33	1.43
Groundwater 1	09.08.2008	-4.83	-36.45	1.42
Groundwater 2	19.08.2008	-4.97	-36.03	5.94
Groundwater 3	29.08.2008	-4.88	-36.45	0.97
Groundwater 4	18.09.2008	-5.94	-38.80	1.39
Precipitation 1	21.09.2008	2.91	28.06	11.41