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Quantification and modelling of water flow in rain-fed paddy fields in NE Thailand: Evidence of soil salinization under submerged conditions by artesian groundwater

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S U M M A R Y

Water flow and solute transport in soils forms an essential part in many groundwater hydrology studies. This is especially true for Northeast Thailand, where the agricultural land is affected by the soil salinity, which is a widespread and an increasing phenomenon affecting 25% of the agricultural land. Salinization appears as scattered discrete patches of 10–100 m^2 in the lowlands, illustrated by white efflorescences during the dry season and bare soil during the cropping season. A field study was undertaken in farm plots to measure the water flow and solute transport within the soil surface and the vadose zone, both inside and outside a saline patch. The water flow was measured on the soil surface with lysimeters and infiltration rings, and was derived in the soil from the hydraulic gradients measured with tensiometers placed at different depths. The salt transport was evaluated with water traps also placed at different depths, where the soil water's electrical conductivity was measured throughout the rainy season.

Field study results demonstrated that the accumulation of saline solutions in rain fed paddy fields, occurred mainly during the rainy season while the soil surface remained flooded. During this period the saline water table rose towards the soil surface independently of infiltration into the soil. It happened in specific places where the compacted soil layer, generally ubiquitous in the area at a depth of 40–50 cm, is interrupted. Therefore salinity appeareds in discret points as patches. Artesian upward flow already described in this area (Haworth et al., 1966; Williamson et al., 1989; Imaizumi et al., 2002) is most probably responsible for this water table rise, thereby affecting crop productivity. Numerical modelling of water flow using HYDRUS-3D further supported these results and showed that managing the depth of flooding within the plot can significantly reduce the outbreak of these saline plumes.

1. Introduction

Salinization is a soil degradation process occurring globally with approximately 20% of all irrigated soils affected by salt (Szabolcs, 1979; Ghassemi, 1995). It consists in an accumulation of soluble salts in the root zone preventing the adequate growth of most plants, mainly due to osmotic stress, as well as some toxicity related problems, and soil physical properties. This phenomenon is generally observed in arid climates where the water balance between rainfall and evaporation is negative, resulting in the capillary rise of saline groundwater during the dry season with the precipitation of salts on the soil surface (Patcharapreecha et al., 1989; Topark–Ngarm et al., 1989; Bressler, 1982; Raes et al., 1995; Ceuppens et al., 1997).

The undulating landscape of the north east of Thailand, often called the Khorat Plateau, comprises two distinctive landform components: (i) the lowlands where paddy soils are severely affected by salinity; and (ii) the uplands that were once covered by a native *Dipterocarpus* forest that have been massively cleared over the last several decades to extend the farming of upland cash crops that include cassava, sugar cane, and maize. It is argued that the evaporative discharge of saline groundwater in the lowlands is directly related to this land clearing as seasonal crops do not contribute as efficiently as trees towards evapotranspiration (Williamson et al., 1989; Imaizumi et al., 2002). Therefore the saline groundwater

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Nomenclature

α	empirical parameter of van Genuchten retention function (L^{-1})
θ	volumetric water content ($L^3\ L^{-3}$)
θ_r, θ_s	residual and saturated volumetric water content ($L^3\ L^{-3}$)
AET	actual evapo-transpiration rate ($L\ T^{-1}$)
c	solute concentration ($mol\ L^{-3}$)
c_r	solute concentration of the sink term ($mol\ L^{-3}$)
D_{ij}	dispersion coefficient tensor ($L^2\ T^{-1}$)
ET_0	reference evapo-transpiration ($L\ T^{-1}$)
h	pressure or tension head (L)
$h(t)$	variable pressure head (L)
h_0	constant surface pressure head (L)
H	hydraulic head $H = h + z$ (L)
i	infiltration rate ($L\ T^{-1}$)
I	Infiltrated water height (L)
K, K_s	hydraulic conductivity and saturated hyd. cond. ($L\ T^{-1}$)
l, m, n	empirical parameters of van Genuchten retention and hydraulic conductivity function (–)
q_z	vertical water flow ($L\ T^{-1}$)
R	rainfall (L)
S	sink term (T^{-1})
S_e	effective saturation (–)
t	time (T)
z, z_i	vertical spatial coordinate (L)
z_{lysi}	water height in lysimeter (L)
z_{ring}	water level in infiltration ring (L)

has risen closer towards the soil surface in the low lands and contributed to the salinization of the soils. Several hydrogeological studies have shown that the halite of the Cretaceous salt rock of Maha Sarakham formation (Hattori, 1993; Kohyama et al., 1993) is responsible for the salinity of the groundwater. These authors point out the presence of different levels of groundwater: (i) a deep very saline and confined groundwater leaching the Maha Sarakam formation, (ii) a more shallow groundwater located in Cenozoic deposits composed of sandstone, siltsone and claystone, generally unconfined which can be fresh or saline. These studies show that the two groundwater levels mainly connect in fractured zones where the network of faults helps to the ascension of the saline groundwater (at 12 to 15 m depth). Previous hydrogeological studies (Haworth et al., 1966; Williamson et al., 1989; Imaizumi et al., 2002) have already pointed out upward artesian flow from the deep groundwater level towards the superficial groundwater. The groundwater composition is mainly chloride and sodium, derived directly from the halite deposits (Fig. 1). However, salinity is not evenly distributed over the landscape but appears in discrete patches of 10 to 100m^2 where the soil surface is covered with white saline efflorescences (mainly halite) or is evidenced by bare patches within cultivated paddy fields. These visual features were confirmed with electromagnetic conductivity measurements, displaying areas of high conductivity (i.e. salinity) in these same areas (Grünberger, 2005). These results suggest that salt efflorescence has a deeper origin. Moreover, despite being submerged during the rainy season, when rice is cropped, these patches appear year after year at the same locations during the dry season. Although

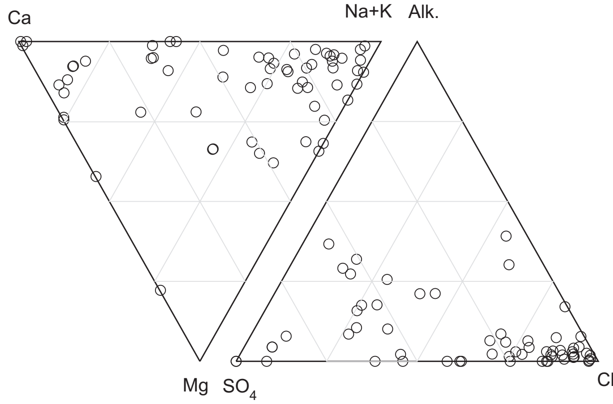

Fig. 1. Pipper diagram of the groundwater samples in the area.

hydrogeological studies describe the movements of saline groundwater, the actual process responsible for the soil salinization has not been studied. The objective of this study is to evaluate the relationship between salinity and water flow, which seems to be the main process responsible for the occurrence of salt patches, by monitoring and modelling the magnitude and direction of the water and solute flows in the soil during a cropping season.

2. Material and methods

2.1. The experimental site

Northeast Thailand is dominated by an undulating topography with uplands that rise to a height of 240 m and lowlands at 170 m, where paddy fields are traditionally cultivated (Fig. 2). Salinization has affected these soils for several decades inducing significant yield declines and in severe cases land abandonment. Salinity does not affect the soils uniformly, but develops in patches of approximately 10 m in diameter. A shallow saline water table present at a regional scale is described as being responsible for these saline outbreaks (Ghassemi, 1995). In an effort to assess the relationship between water flow within the soil and the development of salinity, the magnitude and direction of fluxes inside and outside a saline patch were monitored during the rainy and the dry season, in the region of Khon Kaen Province (16° 22′1.3″N, 102° 38′30.8″E).

An experimental site was instrumented in a paddy field to ascertain the magnitude and direction of flow, water table depth and salinity development during the growth of a rice crop from July to November 2004 (Fig. 3). The soil is part of a widely spread low terrace soil formation in the Northeast of Thailand and is classified as a palequult and belongs to the Kula Ronghai soil series. The soil profile is a 2 m thick uniform fine sandy loamy soil layer of aeolian origin (Lesturgez, 2005) mainly composed of quartz, clay minerals (mainly mixed kaolinite – smectite) and Fe and Mn oxides and hydroxides, attesting to periodic saturated conditions. In the lower part of the terrace the texture is loamy clayey sand with very poor drainage. The soil is saturated during most of the rainy season (during 4 to 5 months). The soil is acidic with pH ranging from 5 to 6, and saline and sodic (Saejiew, 2003). In a recent paper Boivin et al. (2004) described the pedogenesis of these soils as a result of translocation of clay minerals, facilitated by the saline environment and the waterlogged conditions. No evidence of ferrolysis, the commonly invoked process to describe the soil formation in this area (Brinkman, 1979), was found by the authors. Within the soil profile the clay content increased from 3% near the surface, to 10% at depths of 0.50–0.60 m. Similarly the bulk soil density

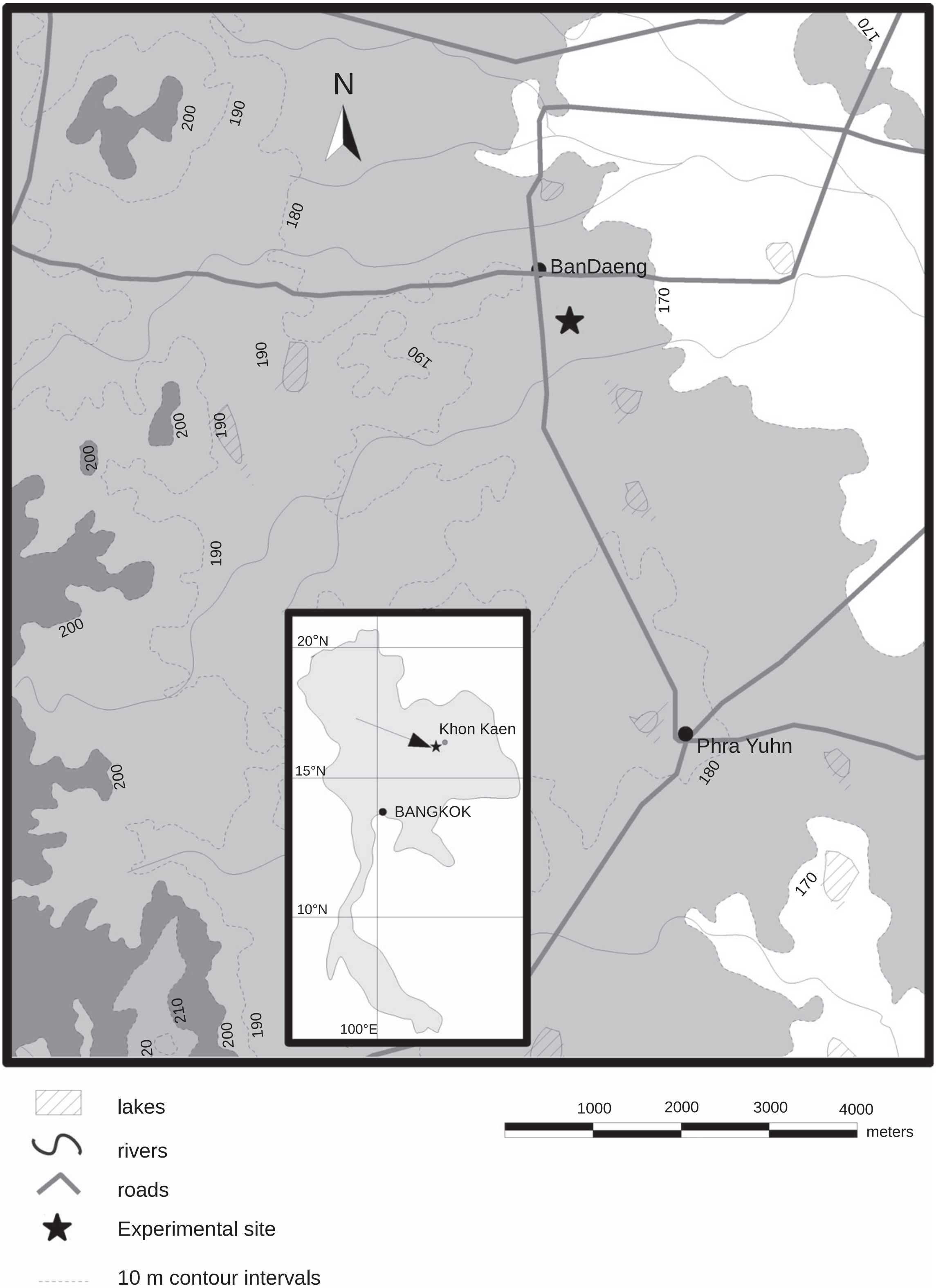

Fig. 2. Location of the experimental site.

varied from 1.52×10^3 kg m^{-3} in the surface layers to 1.91×10^3 kg m^{-3} at 0.50 m. In a more extended survey in the neigbouring plots, performed with a digital penetrometer, some general features appeared in this soil (Fig. 4) like (i) the presence of a plow sole at 20–40 cm in any situation, (ii) the presence of a dense layer at 50–70 cm apparently continuous except in discrete points, under the saline patches where it is systematically absent.

The saline water table is established in the sandy soil at a depth of about 1.75 m during the dry season, and the piezometric head reaches the soil surface during the rainy season. The field was

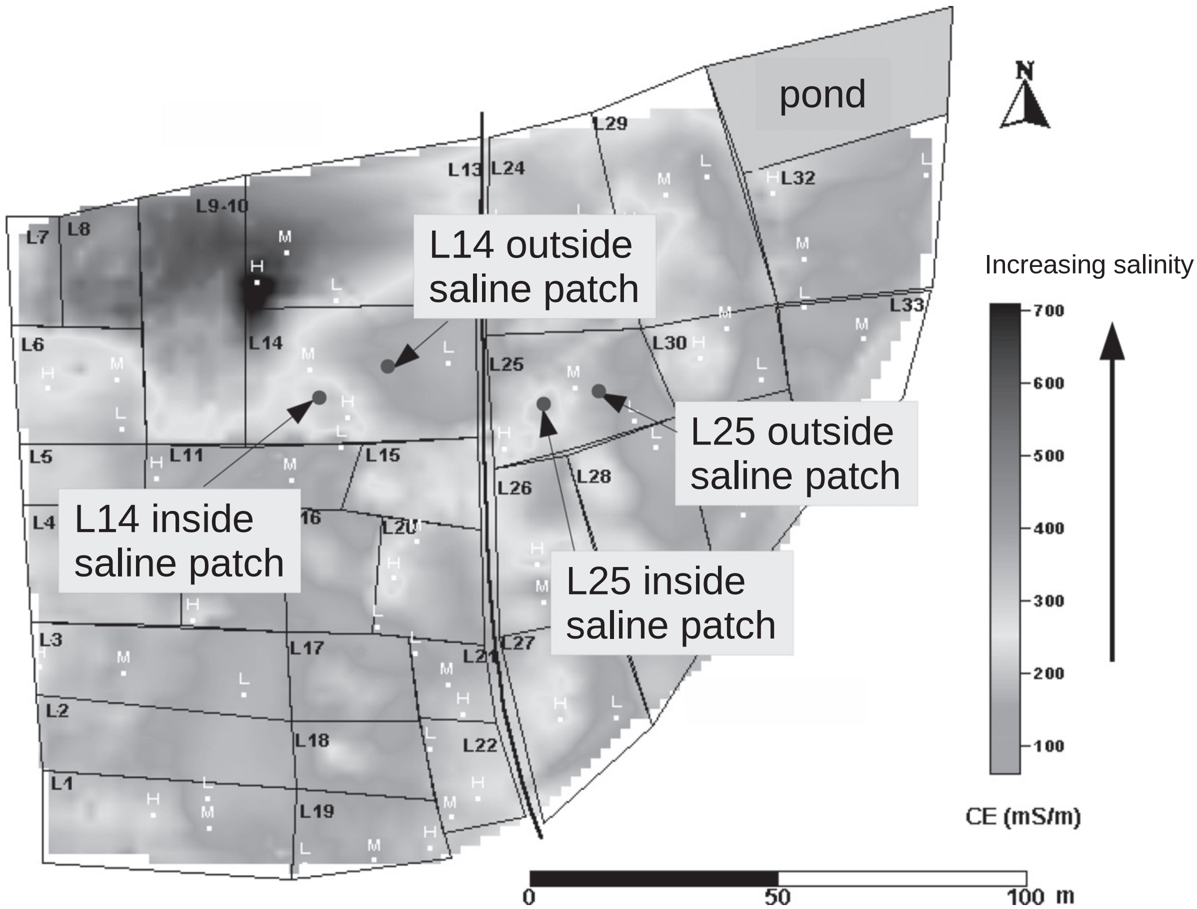

Fig. 3. Salinity map of the experimental site.

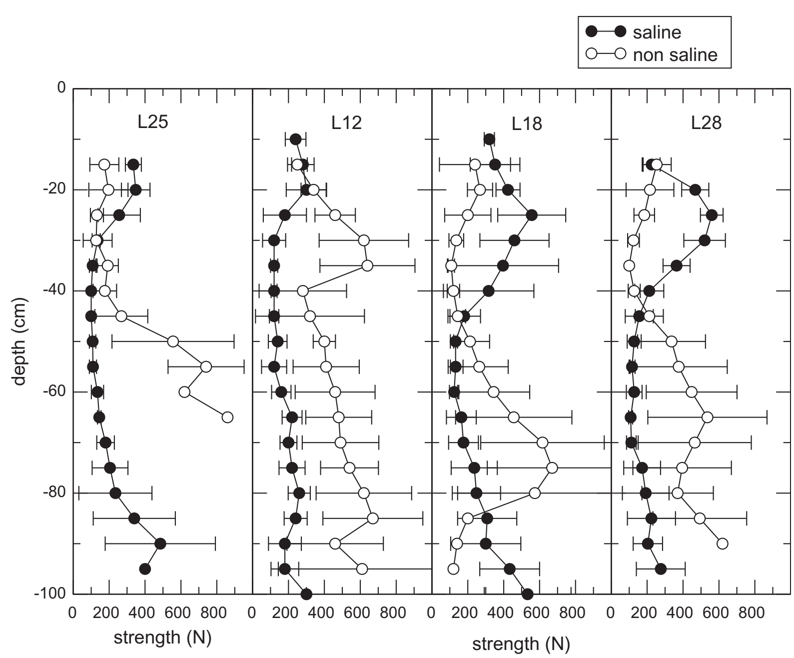

Fig. 4. Penetrometry profiles in different plots around the experimental plot (L25), inside and outside saline patches.

cultivated with rice from 16 June to 6 October 2004 under rain fed conditions, surface water level was maintained approximately at 0.15 m with ranges of 0.05 m to 0.25 m, both inside and outside the saline patch.

2.2. Site instrumentation

A complete water budget was undertaken to quantify the different fluxes inside and outside the saline patch during the entire cropping season. A meteorological weather station was installed in the experimental plot, where rainfall, relative humidity, temperature, wind speed and global solar radiation were recorded hourly. PVC piezometers, driven into the soil and sealed with a concrete base were installed at a maximum depth of 3.5 m in the area and the depth of the shallow water table was monitored manually every two days. The hydraulic head in the vadose zone was measured manually every other day with a portable electronic puncture tensiometer SMS 2500S (SDEC, France), in porous cups installed at 10, 25, 45, 65 and 110 cm depth. Soil moisture measurements could not be performed as the extremely saline conditions were unsuitable for TDR measurements. Consequently the volumetric water content θ was derived from the matric tension h with the corresponding retention curves. Although the salinity of soil solution was high the osmotic potential did not affect the retention properties. Salinity was most probably homogeneous inside and outside the porous cup and salinity was not important enough to affect water density (1007.6 kg m^{-3}) or fluidity. The components of the surface water balance, namely the actual evapo-transpiration (*AET*) and infiltration (*I*) of water from the flooded paddy field were monitored respectively with a cylindrical lysimeter and a cylindrical open infiltration ring, both 0.80 m in diameter and were sunk 0.25 m into the topsoil. The water level in these devices was measured on a scale inclined at 30° to double the precision of the readings, in order to calculate the net infiltration from the difference between both measurements (Hammecker et al., 2003). The soil water salinity (electrical conductivity) was measured in water traps placed in the soil at 0.10, 0.25 and 0.45 m depth. They consisted of cylindrical polyethylene containers (6 cm diameter,10 cm) perforated laterally in the upper part to let soil solution flow freely and accumulate in the lower part of the container, where the soil solution was sampled by suction through a capillary tube connected to the surface.

2.3. Calculation of surface water budget and soil water flow

The rainfall was monitored at 1.5 m above the soil surface, in an automatic rain gauge connected to data logger (Campbell CR10X). Actual evapo – transpiration (*AET*) was computed from the manual monitoring of the water level in the lysimeter, corrected from the rainfall, regardless of the soil porosity:

$$AET = \frac{dz_{lysi} + dR}{dt} = \frac{(z_1 - z_0) + R_1 - R_0}{t_1 - t_0} \tag{1}$$

where z_{lysi} represents the water level in the lysimeter, t the time, R the rainfall between two successive measurements defined by the subscripts 0 and 1. The net infiltration rate i is calculated by difference of the evolution of the water level in the infiltration ring and the *AET*:

$$i = \frac{dz_{ring} + dR}{dt} - AET = \frac{dz_{ring} - dz_{lysi}}{dt} \tag{2}$$

where z_{ring} is the water level in the ring. On the other hand, basing on the hydraulic gradient calculated between the tensiometers installed at different depths, the one dimensional (vertical) water flow in the soil was computed with Darcy law.

2.4. Soil physical properties measurement

The unsaturated hydraulic properties of the soil were measured partly *in situ* and partly in the laboratory. The saturated hydraulic conductivity was measured in the field along a soil profile with a disc infiltrometer (Perroux and White, 1988; Smettem and Clothier, 1989) according to Wooding's method (1968). However in case of saline soils with Na-saturated clays clay dispersion represents an important factor affecting water infiltration (Abu-Sharar et al., 1987; Abu-Sharar and Salameh, 1995; Amezketa and Aragües, 1995; Curtin et al., 1994). Laboratory experiments (Saejiew et al., 2004) performed on the soil from this site, clearly attested the high dispersivity of the clays. In a more recent study, *in situ* disc infiltration experiments performed with distilled water and saline solutions (Hammecker et al., 2005), clearly demonstrated that when performed with saline solutions the steady state infiltration could be reached quickly. When performed with distilled water the steady state infiltration was never reached completely as the infiltration rate continuously decreased. The dilution of the soil solution is most likely to be contributing to clay dispersion. In order to take into account realistic conditions, hydraulic conductivity measured with saline solution was considered for the two lower horizons of the soil profile, as only the saline water table reaches them. For the most superficial soil horizon, the saturated hydraulic conductivity was calculated from the infiltration experiment with distilled water (Hammecker et al., 2005), so that the infiltration of rain water could be simulated.

The retention curves of the soil samples were determined in the laboratory using an evaporation method (Wind, 1968), on undisturbed soil cylinders (5.5 cm height, 12 cm diameter). The hydraulic functions of unsaturated soils were adjusted to the experimental retention curves (Fig. 5) and the parameters were then determined with RETC code (Van Genuchten, 1991) described by Van Genuchten, 1980:

$$S_e = \frac{\theta - \theta_s}{\theta_r - \theta_s} = \left[1 + (\alpha \cdot h)^n\right]^{-m} \tag{3}$$

$$K = K_s \cdot S_e^l \left[1 - \left(1 - S_e^{1/m}\right)^m\right]^2 \tag{4}$$

where S_e is the effective saturation, θ_r and θ_s respectively the residual and saturated volumetric water content, h is the matric potential or tension head ($h = H - z$, where z is the depth of the measurement, counted negatively downwards), α, n, m, l are empirical parameters (where $m = 1 - 1/n$ and $l = 0.5$), K and K_s are respectively hydraulic conductivity and saturated hydraulic conductivity. Consequently from simple hydraulic head measurement in tensiometers, hydraulic conductivity could be calculated by combining Eqs. (3) and (4). Soil properties were measured along the soil profile inside and outside the saline patch. The deepest measurement outside the saline patch corresponded to the compact soil layer. The soil properties measured inside the saline patch at the same depth were considered as being representative of the soil underneath the compact layer. The measured soil hydraulic properties displayed in Table 1 were used for calibration and validation of the model during the dry season (Bernardeau, 2007).

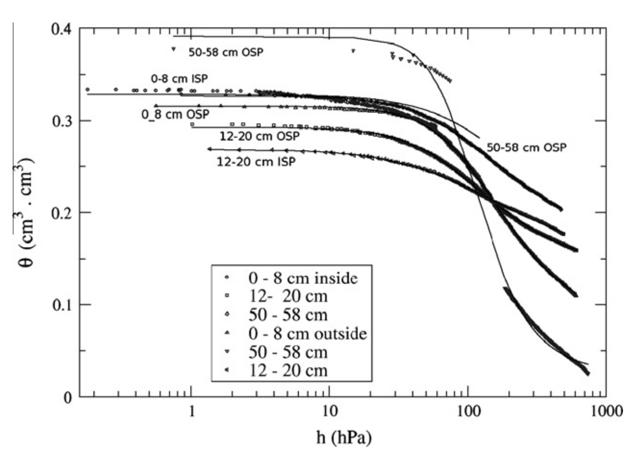

Fig. 5. Experimental retention data obtained by Wind's method (dots), and adjusted Van Genuchten curves (full lines).

2.5. Modelling

In order to quantify the water flow directions and magnitudes and solute transport during this period, experimental data were compared to numerical simulations performed with HYDRUS-3D (Šimůnek et al., 2008). This model simulates two- and three-dimensional movements of water, heat and solutes, in variably saturated porous media with the Galerkin finite element method to solve numerically the 2D or 3D form of Richards equation. As this equation is described as a function of the pressure head in HYDRUS-3D, it applies to both unsaturated and saturated conditions. It was therefore particularly well adapted to the simulation requirements of this situation.

Considering the size of the saline patch, and its almost circular geometry, a 2D simulation was perfomed over a domain representing a 10 m wide by 2 m deep vertical plane through the saline patch, formed by a regular rectangular grid of 100 by 100 nodes. Different soil materials with specific hydraulic properties, determined experimentaly, have been taken into account for this domain. Especially the presence of the compact layer mentioned earlier was simulated by a soil layer with lower hydraulic conductivity. The thickness of this layer was set to 0.5 m, it was located at a depth of 0.6 m and was interrupted over a distance of 2 m, as observed in the saline patch. However the genericity of the interruption size was not verfied as it has not been investigated precisely in the other saline patches present in the region.

Previous results showed that the surface water level and the piezometric head were closely related so that it was difficult to establish clearly the causes and consequences, namely if the evolution of the water table depth was controlled by the surface water level or inversely. Therefore modelling was performed with different experimental boundary conditions, in order to attempt to determine the governing processes responsible for the development of the pressure head $h(t,z)$ at different depth in the soil profile. For the lower boundary condition two different situations were tested: (i) a Dirichlet condition with the measured piezometric head sequence, namely with variable pressure head conditions $h_L(t)$, (ii) a Neumann condition with zero-flux. For the upper boundary several conditions were tested: (i) atmospheric conditions, namely the Neumann conditions with daily weather data of evaporation and rainfall, (ii) seepage face condition, (iii) variable pressure condition corresponding of the development of the water level at the soil surface $h_U(t)$.

The goodness of the modelling was quantified by the Root Mean Square Error (RMSE). The tensiometric data at five depths (0.1, 0.25, 0.45, 0.65, 1.10 m) during 66 days, hence 330 points were used to calculate RMSE. The partial differential equation governing two-dimension transport of non-reactive solutes was useed in this study:

$$\frac{\partial \theta c}{\partial t} = \frac{\partial}{\partial x_i}\left(\theta D_{ij}\frac{\partial c}{\partial x_j}\right) - \frac{\partial q_i c}{\partial x_i} - Sc_r \quad (5)$$

where c is the solute concentration, q_i the volumetric flux density of ith component, S a sink term, c_r the solute concentration of the sink term, and D_{ij} is the dispersion coefficient tensor. Cauchy type conditions (determined concentration flux) were imposed to both upper and lower boundaries. Dispersion was not measured but set artificially. Bernardeau (2007) tested several values for dispersion and found that for this situation 1.5 cm was adequate to fit to experimental data and to minimize Peclet number. In general a value of 1.5 cm is considered to be reasonable for this soil (Inoue, 2000; Thomasson and Wierengab, 2003).

3. Results and discussion

3.1. Piezometric data

The experimental results for the piezometric monitoring indicate that the water table was shallow, as piezometric head varied between 1.75 m during the dry season (November to April) and 0.10–0.20 m above the soil surface during the rainy season. The development of a piezometric head seemed to be closely related to the rain events and the response of the piezometric head to rainfall was immediate (Fig. 6). However when observed closely, Fig. 6 shows in some cases an important increase of piezometric head occurring before the actual rain event. This process can be explained by atmospheric pressure decrease before a big rain event (Acworth and Brain, 2008) or more probably that the water table followed regional dynamics and that the rainfall recharged the water table elsewhere in the watershed before the rain event occurred at the experimental site. The rapid response of the piezometric head to rain events is indicative of a rapid pressure transmission in the aquifer, considering the recharge area being distant from the experimental site (5–10 km). This behaviour is analogous to a confined aquifer. At the end of the rainy season the piezometric head dropped quickly reaching initial values within a few days.

3.2. Surface water flow

The evolution of the surface water level followed the same pattern as the piezometric head as depicted in Fig. 6, suggesting hydraulic equilibrium and continuity between groundwater and water in field surface. The water balance at the soil surface was quantified by monitoring the water level in the lysimeter and the infiltration ring and computed with equations (1) and (2). The average actual evapo-transpiration was found to be 6.4 mm d^{-1} inside and outside the saline patch during the monitored rainy season, and the potential evapo-transpiration ET_0 calculated with Pennman-Monteith equation (ASCE, 2004) during the same period, with the meteorological data, was 3.6mm · d^{-1}. The actual evapo-transpiration for rice in this region is consistent with the measurements of other authors (Tomar and O'Toole, 1980; Tyagi et al., 2000), with a crop growth coefficient factor reaching 1.4, suggesting an underestimation of ET_0, probably due to a malfunctioning of

Table 1
Unsaturated soil characteristics and bulk density at different depths, inside and outside the saline patch.

		θ_r m^3 m^{-3}	θ_s m^3 m^{-3}	α m^{-1} ($\times 10^{-1}$)	n	K_s m s^{-1} ($\times 10^{-7}$)	d kg m^{-3} ($\times 10^{3}$)
INSIDE (cm)	0–8	0.00	0.33	9.6	1.63	5.60	1.52
	12–20	0.08	0.29	12.9	1.51	1.48	1.68
	50–58	0.13	0.33	12.0	1.55	4.86	1.67
OUTSIDE (cm)	0–8	0.07	0.32	9.0	2.03	4.75	1.51
	12–20	0.13	0.27	13.4	1.56	1.30	1.73
	50–58	0.03	0.39	10.4	2.78	0.25	1.91

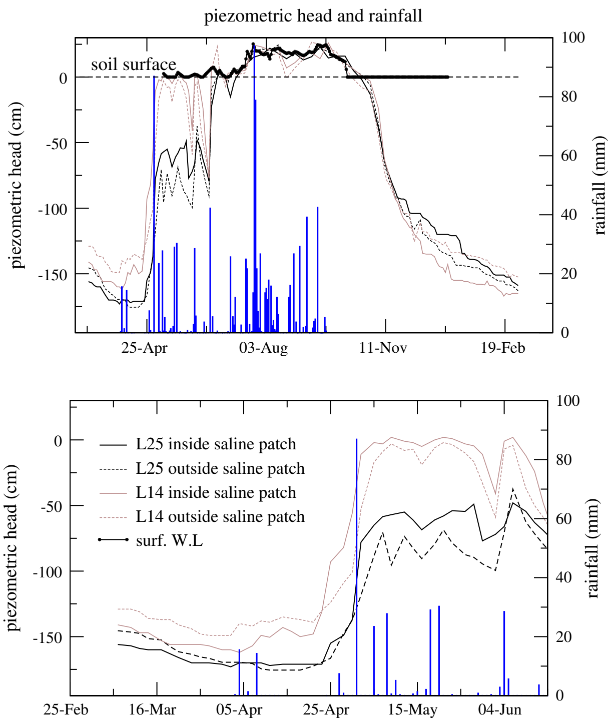

Fig. 6. Daily rainfall, surface water level (surf. W.L.) and piezometric head in two contiguous plots inside and outside saline patches: (a) during the rainy season 2004 and beginning of dry season, (b) at the beginning of the rainy season.

the anemometer. Water level at the soil surface in the plot was high enough (Fig. 6) to ensure non-saline conditions in the root zone, so that the development of the rice inside the saline patch was not affected by soil salinity during the monitoring period. Though an average value of AET for the entire cropping period has been evaluated, a slight increase in AET could be decteted, corresponding to the growth of rice. Nevertheless the precision of the measurements does not allow to follow acurately the evolution of evapo-transpiration during the different phenological stages of rice.

Water infiltration was found to be directed upward (positive water flow) inside the saline patch and slightly negative outside (Fig. 7). At the end of the cropping season the cumulative upward infiltration reached 50 cm inside the saline spot, and the cumulative downward infiltration reached only 10 cm outside the saline patch. These results suggest that inside the saline patch the water table reached the soil surface, whereas this phenomenon is limited or absent outside the patch.

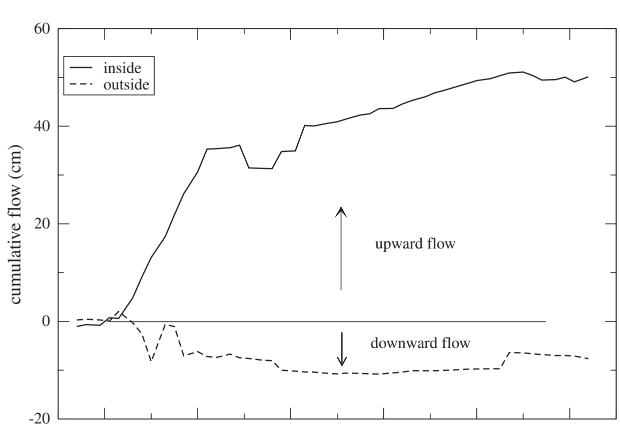

Fig. 7. Evolution of cumulative surface water flow.

3.3. Soil water flow

The tensiometric data, namely the hydraulic head (H), show contrasting behaviour inside and outside the saline patches, principally during the dry season (Fig. 8). Although the presence of solutes will affect the viscosity and the surface tension of the solution, these results illustrate differences in water flow conditions in the two situations. During the dry season, hydraulic heads at the different depths increase significantly outside the saline patch until they reach -500 cm, whereas these values remain low inside the saline patch i.e. between -100 and -200 cm. Development of the hydraulic head inside the saline patch is coherent with the development of the piezometric head, whereas outside the patch soil water seems to be completely disconnected from the dynamics of the water table. As depicted in Fig. 8b, the hydraulic gradient computed in the root zone between 10 and 45 cm, is found to be negative outside the saline patch and positive inside, during the cropping season (submersed conditions). This illustrates a contrasting situation with the possibility for water to flow upward inside the saline patches and downward elsewhere. During the following dry season, the hydraulic gradients showed sometimes extreme variations, mainly due to the refilling of the tensiometers when air had entered the ceramics. The actual water flow in soil, calculated with Darcy equation, the unsaturated soil characteristics (Table 1) and the hydraulic head data, are represented in Fig. 9. In both cases the development of calculated water flow in soil follows a similar trend as the surface water flow. Although these measurements were performed independently, with different techniques, the actual values for cumulative infiltration are similar: inside the saline patch 0.45 m cumulative upward flow with the soil water flow computation at the end of the cropping season and 0.50 m exfiltration with the surface water flow measurement; outside the saline patch with both techniques a cumulative infiltration of 0.05 m was found. Typically three periods could be defined: (i) the flooding period when the piezometric head increased to a maximum value and during which time all fluxes were directed upwards; (ii) the water table draw down, when the piezometric head fell drastically (by more than 1 meter in a week) and when the soil water suddenly flowed downward; (iii) the capillary rise period during the dry season, when the soil water was driven towards the surface. The capillary rise period was found to be effective only for two or three weeks until the beginning of January. After this period upward water flow is limited and almost nil after February, suggesting the absence of hydraulic continuity between the water table and the soil surface.

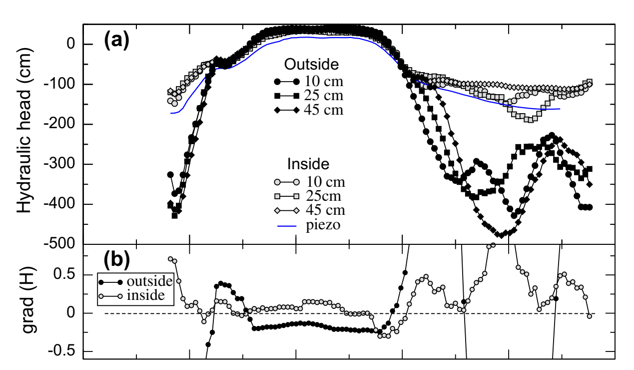

Fig. 8. (a) Development of the hydraulic head at different depth in the soil, and (b) development of hydraulic gradient between 10 and 45 cm, inside and outside the saline patch.

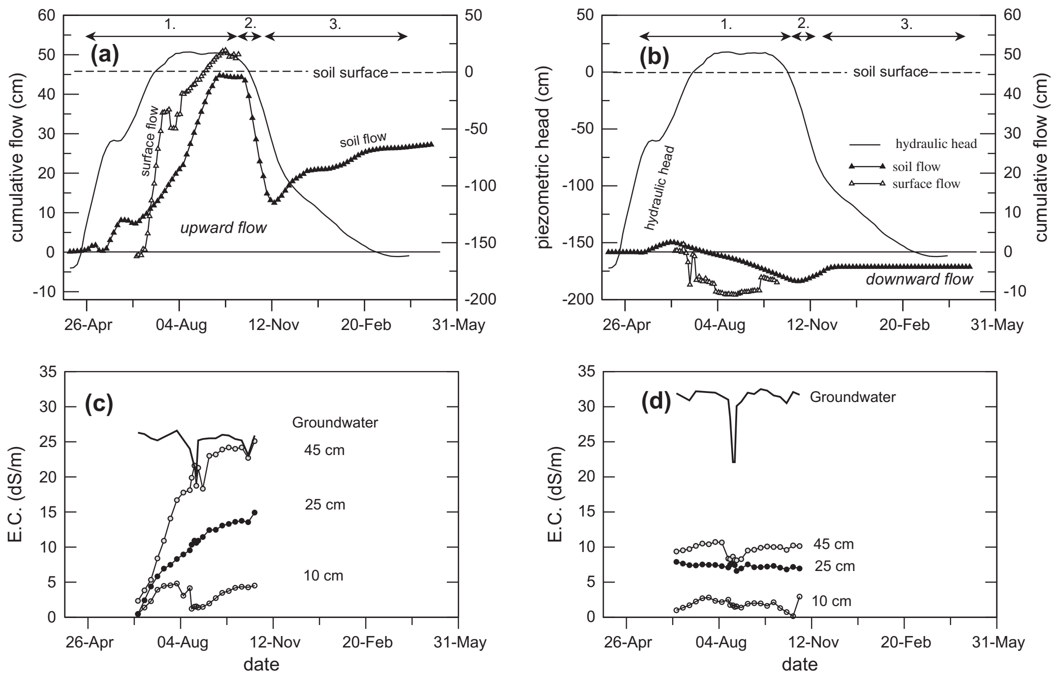

Fig. 9. The development of piezometric head, the cumulative water flow in soil between 10 and 45 cm (Soil flow) and the surface water flow (Surface flow) both inside (a) and outside (b) the saline patch. Three different periods can be identified according to the water table dynamics: (1) Groundwater uprise period; (2) Groundwater draw down period; (3) capillary rise period. At the same time the development of salinity is depicted as measured by the electrical conductivity of the soil solution during the cropping season (flooded soil surface) both inside (c) and outside (d) a saline patch.

3.4. Soil solution concentration

The electrical conductivity of the soil solution was monitored at different depths in water traps in soil and in piezometers. As shown in Fig. 9c and d, the salinity of the groundwater is high with EC values ranging from 25–30 dS m^{-1}, equivalent to 250–300 mmol l^{-1} NaCl, during the whole cropping season. In mid June, starting with the rainy season, the concentration of soil solution increased, with depth outside the saline patch, and remained unchanged during the whole cropping period, until early November. Contrasting this, inside the saline patch the concentration increased regularly with time at all depths, until reaching the same concentration as the water table for the deepest water trap at 45 cm. These results illustrated the progressive uprise of the saline groundwater during the rainy season inside the previously existing saline patch, whereas outside this uprise did not occur. When integrating the soil solution concentration over the depth of the root zone (0–45 cm) the average salt content in the soil profile for the 10–45 cm depth interval was derived from Fig. 9c and d. The initial salt content in solution at the beginning of the season was 9.15 g m^{-1}, whereas at the end of the rainy season it reaches 143 g m^{-1}. This suggests that while the surface of the plot was flooded, the salinity in the top soil increased by 1400%, unlike the usual observation in saline soils, where the rainy season corresponds to a salt leaching period. According to the previous results of the water flow, the development of the solutes suggested the presence of an artesian upward flow of saline groundwater where the dense layer was interrupted.

3.5. Results of modelling

The results for the simulations, displayed in Fig. 10 showed that for most of the cases the general trend was respected, but systematic underestimation of the pressure head during the flooding period was computed by the model. The underestimation of h is especially important in the upper part of the soil profile, indicating that the calculated infiltration rate was probably higher than the actual one. Evaluation of the modelling with RMSE (Table 2) invalidated the cases where the lower boundary condition was set to zero flux, namely assuming that the evolution of the groundwater level was controlled by surface water flow. It suggested that the development of the piezometric head and the evolution of the concentration of the soil solution were not governed by the surface water infiltration.

In the situations where the lower boundary condition was influenced by the groundwater dynamics, regardless of the upper boundary condition, the RMSE values showed better agreement. The best fit was achieved with the atmospheric conditions, though the computed values for pressure head in the upper part of the soil profile were underestimated. The evolution of the soil solution concentration was compared to the computed results obtained for the cases of atmospheric and variable pressure upper conditions. The results presented in Fig. 11-1. showed a clear difference in the two situations; with a dilution dynamic in the case of variable pressure conditions and a concentration dynamic in the case of atmospheric conditions, in accordance to the experimental results. These last results were illustrated in a 2D representation in Fig. 11-2 and showed an uprise of the saline groundwater where the compact layer was interrupted. During the first month of submersion, the model computed a downward water flow, whereas the experimental data showed limited observed downward water movement. This discrepancy is most probably due to the contribution of some physico-chemical mechanisms such as the presence of a second phase (air) and the influence of clay dispersion, which are not taken into account in HYDRUS-3D. For example, air entrapment between the shallow water table and the superficial wetting front, illustrated by the air-bubbles escaping from the soil surface during submerssion, is another potential mechanism leading to the reduction or blocking of water infiltration into soil (Hammecker et al., 2003). As clay dispersion happens gradually when fresh water infiltrates into saline soils, the model should consider spatial and temporal reduction of hydraulic conductivity.

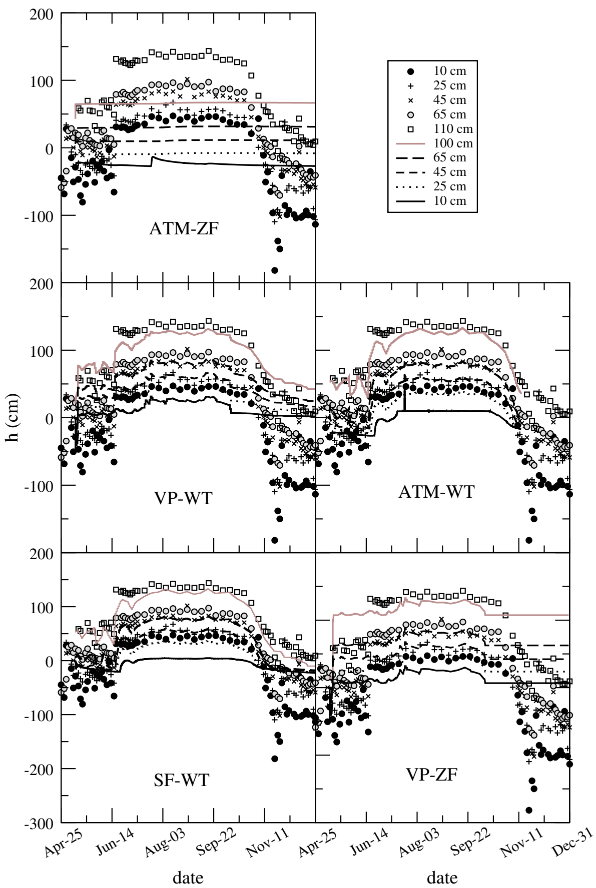

Fig. 10. Experimental (symbols) and simulated development (lines) of the pressure head for different boundary conditions. For the upper boundary the different conditions were: atmospheric (ATM), ponding conditions with variable pressure (VP), and seepage face (SF). For the lower boundary the conditions were: variable pressure according to the water table level measurements (WT) and zero (ZF).

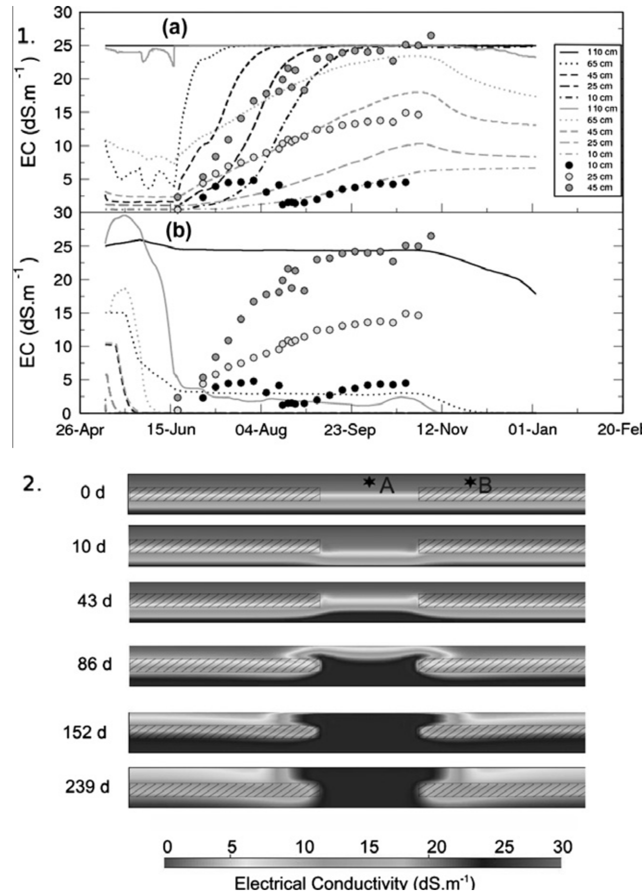

Fig. 11. Results of simulation. (1) Development of the experimental soil solution concentration at different depth (symbols) as compared to the computed data (lines) inside (black) and outside (gray) the saline patch with (a) atmospheric upper boundary conditions and (b) with variable pressure conditions; (2) 2D representation of the soil solution concentration development during the simulation of the rainy season with atmospheric upper boundary conditions. (A) and (B) are the observation points respectively inside and outside the saline patch.

Table 2
Evaluation of the modelling with root mean square error for different boundary conditions: horizontally the upper boundary conditions (BC) and vertically the lower boundary conditions. RMSE was calculated for hydraulic head (cm).

lower BC	Upper BC		
	Atmospheric	Seepage face	Variable pressure
Water table	22.19	32.42	28.32
Zero-flux	53.24	–	49.59

In order to evaluate the possibility of controlling the saline groundwater uprise, simulations with several constant pressure conditions have been tested. In the most superficial root zone (10 cm deep) as presented in Fig. 12, the effect of the height of the flooding water level was crucial to the water flow direction and to the development of the soil solution salinity. The cumulative water flow was directed downwards when the water level on soil surface was less than 12 cm during the entire cropping season, whereas surface water clearly infiltrates into the soil when the water level exceeded this height. The consequence on soil salinity was more profound as only 10 cm of water was necessary to prevent the saline groundwater reaching the upper root zone. The actual value of the required water level at the soil surface depends on the piezometric head developed during the rainy season. Hence it might vary from one year to another depending on the rainfall, and on the topography of the plot.

3.6. Discussion

Considering the experimental results for the surface and soil water flow, and the development of the solutes inside and outside the saline patch, it appeared that during the rainy season the main water and solute movement was directed upwards where the dense soil layer was interrupted. Moreover, according to the general hydrogeological environment mentioned by other authors, these results support the hypothesis of superficial outbreaks of the artesian upward flow of saline groundwater.

As most of the saline groundwater rose to the surface during the rainy season, the stock of salt in the top soil (0–45 cm) was constituted during this period. Though no measurements for the soil solution concentration were available during the dry season, it can be assumed that these solutes were only redistributed with the evaporative gradient towards the soil surface, delineating the saline patches with salt efflorescence in the area (Bolomey, 2002; Grünberger et al., 2008). Modelling demonstrated that artesian water flow occurred inside the saline patches i.e. where an underlying dense layer was interrupted. Outside the saline patches, this dense layer partially blocked the rise of the saline groundwater. Modelling also confirmed the presence of a confined water table erupting at the soil surface where the dense impervious soil layer was absent, when it was put under pressure from sources outside the immediate vicinity (in the uplands). Hence the scattered distri-

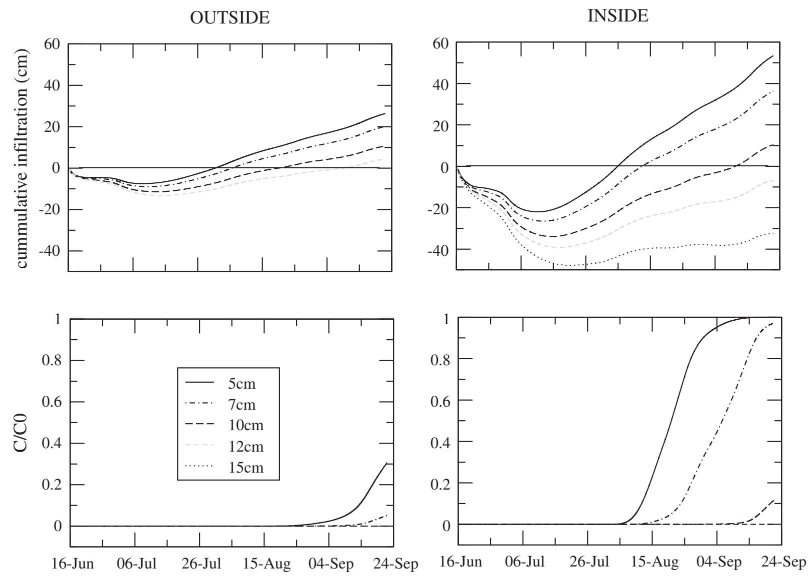

Fig. 12. Simulated cumulative infiltration and relative soil solution concentration at 10 cm, outside (b) and inside (a) the saline patch for different heights of ponding.

bution of salinity in the landscape, as finite spots could be explained by this mechanism. Unlike most of the cases of primary soil salinization usually described (Prathapar et al., 1992; Raes et al., 1995; Ceuppens et al., 1997), this case did not illustrate the typical wick effect, where the saline groundwater fed the soil surface by capillary rise, driven by evapotranspiration during the dry season. Here salinity was brought to the top soil (0–45 cm) during the rainy season, by general artesian groundwater flow, while the soil surface was submersed with ponded rainwater. It should nonetheless be mentioned that the dimensions and the geometry of the interruption of the dense soil layer have not been investigated in the study, though simulations with different widths produced equivalent results.

In contrast to what would normally be expected during the rainy season, the results indicate that soluble salts accumulated in the soil profile while the soil was submersed. During the rainy season, the confined aquifer was probably recharged in the uplands (alt. 240 m) area, where hydraulic head was transmitted rapidly towards the groundwater table in the lowlands. Consequently, as illustratd by the modelling results (Fig. 11.2) a plume of saline water rose towards the soil surface in a predetermined position where the confinement of the aquifer was locally interrupted. The salinization of this paddy field located in the lowlands, occurred mainly during the rainy season by ascendant water movement which was most probably related to regional artesian water flows described by other authors at a regional scale in deeper geological formations (Haworth et al., 1966; Williamson et al., 1989; Imaizumi et al., 2002). These results were found locally and within this specific study no data were available to extrapolate them further. However they apply at a larger scale where similar deep groundwater artesian flow has been described (Srisuk, 1994). A similar behaviour has been observed during an agronomic study more widely spread out in the area (Clermont-Dauphin et al., 2010). The authors described a systematic increase of soil salinity during the cropping period, where saline patches were identified previously, while the plots were flooded. On the other hand, the results obtained at a more extended scale with a digital penetrometer (Fig. 4), showed that when the dense soil layer at 0.5 m depth, was interrupted, it was systematically located under a saline patch. These results support the hypothesis of the presence of local breaks in the confinement of the saline groundwater, contributing directly to soil salinization.

The increase of the soil salinization at a regional scale during the last decades in this area, has been been attributed to an intensification of artesian upward flow, related to an intensive land clearing in the uplands by other authors (Williamson et al., 1989). However, they mainly described the water flow in the lower geological horizons, with no specific mention of the superficial process. Saline seeps have been described in the North American Great Plains (Miller et al., 1981; Miller et al., 1993) and Western Australia (George, 1991), but soil salinization under submersed conditions, like shown in this study was never described. In this study, the solutes pushed toward the top soil during the rainy season, and were most probably simply redistributed during the dry season as the piezometric head dropped rapidly at the end of the rainy season, breaking hydraulic continuity impeding a continuous capillary feed of the soil surface from the saline groundwater. This is confirmed in a recent study set up in the same area (Grünberger et al., 2010) where the authors assessed by isotope tracing that during the drying season, the redistribution of salts in the superficial layers was confined to the upper layers (above 40 cm) and that capillary rise was inadequate to explain surface salt increase inside the saline patches.

Drainage and surface flushing of the plots, as often prescribed by irrigation engineers in arid areas (Bressler, 1982), would not be adequate in this situation as there was a continuous seepage of saline water during the cropping season. The modelling results show that with an idoneous water management maintaining the water level in the plot to its maximum, salinity could be kept out of the root zone.

4. Conclusion

The quantification of the water budget at the plot scale in paddy fields of Northeast Thailand produced some unexpected results. By measuring the water flow and salt movements, it has been demonstrated that soil salinization predominantly occurred during the rainy season while the soil surface was completely submersed, which is contrary to what would have been expected. The mechanism was found to be regulated by artesian upward flow of a confined saline aquifer, breaking out towards the soil surface when the confinement layer located at 50 cm is interrupted. Consequently, saline patches, which can be seen in the landscape of this area, are evident at the soil surface. Water flow and solute transport modelling with HYDRUS-3D not only confirmed these measurements, but also provided some useful information on the possibility of controlling soil salinity in these situations regardless of commonly used remediation techniques.

More generally in other regions of the world, especially in hilly sub humid regions with saline subsoils, it can be expected that this kind of situation could also be observed even if not diagnosed, as this phenomenon remains discrete and might be overlooked by the visual outbreak of saline efflorescence during the dry season.

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