Monitoring Unsaturated Flow and Transport Using Cross-Borehole Geophysical Methods

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Improved understanding of unsaturated flow and transport processes is limited by the lack of appropriate in situ measurement techniques. This study was conducted to determine whether two noninvasive cross-borehole geophysical methods combined could be used to estimate two important unsaturated zone transport parameters, namely the pore water velocity and longitudinal dispersivity. Cross-borehole electrical resistivity tomography and ground penetrating radar were used to estimate temporal and spatial variation of electrical resistivity and water content, respectively, during a 20-d forced infiltration experiment. The resulting one-dimensional profiles and two-dimensional images of moisture content and electrical resistivity were subsequently combined to estimate solute tracer concentration. The results were used to analyze the downward migration and vertical spreading of water and tracer mass. The two geophysical methods provided independent estimates of soil moisture content and electrical resistivity that were spatially and temporally consistent. The observed changes in moisture content and electrical resistivity were, as a first approximation, used in a one-dimensional moment analysis. The transport behavior was found to be very susceptible to layering of the subsurface. Even slight reductions in grain size apparently lead to flow barriers and associated lateral flow, resulting in tracer mass loss, reduced vertical pore water velocity, and increased longitudinal dispersivity. Synthetic data showed that the estimated unsaturated transport parameters (i.e., pore water velocity and longitudinal dispersivity) and the mass estimate were influenced by the selected electrical resistivity tomography inversion routine. In effect, an overprediction of all three parameters was observed.

ABBREVIATIONS: EM, electromagnetic; ERT, electrical resistivity tomography; GPR, ground penetrating radar; MOG, multiple-offset gather; ZOP, zero-offset profiling.

Recent research has shown that cross-borehole geophysical methods offer promising alternatives to traditional techniques for hydrologic characterization. The measurements collected by geophysical methods are on a more appropriate scale, with high spatial resolution and with minimum intrusion of the subsurface. In particular, cross-borehole ground penetrating radar (GPR) and electrical resistivity tomography (ERT) can provide estimates of soil moisture content and electrical conductivity variations in the vadose zone between boreholes located up to ~10 m apart (Hubbard et al., 1997; Daily et al., 1992; Binley et al., 2001; Alumbaugh et al., 2002; Ferré et al., 2003). Such data provide valuable information on flow and transport and may be used for identifying or constraining hydraulic and transport parameters of hydrologic models.

A number of studies, aimed at identifying unsaturated hydraulic parameters using GPR and ERT surveys, have been reported. Binley et al. (2002) and Winship et al. (2006) used moisture content changes arising from a point tracer injection experiment, monitored by cross-borehole GPR and ERT, to estimate the first and second spatial moment of the tracer mass. These values were then compared with forward hydrologic simulations of the tracer test to estimate the effective hydraulic conductivity of the subsurface of interest (a UK Sherwood Sandstone). Cassiani et al. (2004) identified effective hydraulic conductivities using moisture content profiles estimated by GPR in a Monte Carlo inverse approach, while Cassiani and Binley (2005) and Binley and Beven (2003) identified probability distributions of the parameters in functional relationships for the unsaturated hydraulic functions. Kowalsky et al. (2005) performed joint inversion of GPR and neutron probe data collected during a point injection flow experiment to estimate hydraulic conductivity and important petrophysical parameters.

Identifying an optimal set of hydraulic parameters using hydrogeophysical methods is often not trivial. This is partly due to technical limitations of the geophysical methods associated with resolution and inversion artifacts, but also due to the experimental circumstances and the type of data collected to constrain the hydraulic parameters. In general, hydrogeophysical measurements have been conducted during two experiment types: (i) point injection and trench experiments, where mass balance considerations are limited by assumptions concerning the Fresnel zone of the GPR measurements (Winship et al., 2006); and (ii) natural loading of the subsurface, where moisture content changes are minimal and multiple measurement profiles only provide limited additional information to constrain the hydraulic parameters (Binley and...
Beven, 2003). A synthetic study by Rucker and Ferré (2004b) stressed the importance of collecting additional data types (in their case pressure head measurements) to further constrain the parameter identification.

In this study we have used cross-borehole geophysical methods to monitor and analyze a forced water and tracer infiltration experiment. As opposed to previous investigations, our experiment was designed to approximate one-dimensional conditions by applying water and tracer over an area at the surface. Furthermore, a substantial loading of the system was imposed to produce a considerable dynamic response to enable simple moment analysis for identifying effective pore water velocity and longitudinal dispersivity, as well as the possibility of evaluating if the data honored the water and tracer mass balances.

In line with Winship et al. (2006), we have made simultaneous use of high-resolution images based on cross-borehole GPR and ERT. By combining estimates of moisture content (obtained from GPR-determined relative permittivity) with electrical resistivity (through ERT measurements), tracer concentration estimates were obtained. With this approach, two independent data types (moisture content and tracer concentration) were collected to constrain the flow and transport parameters.

Field Site

The field site was established in Denmark on a 20- to 30-m unit of unsaturated alluvial sand sediments. A schematic of the field site setup is illustrated in Fig. 1. The experimental setup consisted of four ERT and four GPR boreholes drilled to a depth of 12 m. The eight boreholes formed a cross consisting of two lines. Along each line, the outer two boreholes (7 m apart) were equipped with ERT-instrumented polyvinyl chloride tubes (electrodes every 50 cm), while the inner two boreholes (5 m apart) had access tubes for GPR antennae.

Figures 2b and 2c present well logging results (natural gamma and neutron logging) collected in the GPR access tubes. Apart from a topsoil layer (approximately 1 m thick), the subsurface characteristics appear not to vary substantially with depth; however, the results of a grain size analysis of sediment samples taken at a well located 18 m from the center of the field site indicated that a slight layering exists (Fig. 2a). The grain size percentiles show that the top 1 m consists mainly of silt, with just a minor fraction of clay (14%). Below this topsoil, a layered sequence can be observed: coarse sand, followed by finer sand, and finally coarse sand again. In the sand, the measured clay content was <1%.

Materials and Methods

A forced tracer infiltration experiment was initiated at the field site in October 2005. Cross-borehole GPR (measuring the relative permittivity) and cross-borehole ERT (measuring electrical resistivity) were used to monitor the downward migration of water and tracer.

Cross-Borehole Ground Penetrating Radar

Measurements were taken using a Sensors and Software PulseEKKO PE100 system (Sensors & Software Inc., Mississauga, ON, Canada) equipped with 100 MHz antennae. Two measurement techniques were conducted: zero-offset profiling (ZOP) and multiple-offset gather (MOG) (Binley et al., 2001). Zero-offset profiling enabled an estimation of the one-dimensional electromagnetic wave velocity distribution between two boreholes, while MOG resulted in a two-dimensional tomographic image of the electromagnetic wave velocity distribution. During the ZOP measurements, two antennae were lowered simultaneously into a set of boreholes with measurements taken every 0.25 m. A total of six ZOPs were collected during approximately 30 min. All possible sets of pairs between the four boreholes in Fig. 1 were sampled, i.e., GPR1–2, GPR1–3, GPR1–4, GPR2–3, GPR2–4, and GPR3–4; however, only one MOG was collected using borehole pair GPR1–3. This collection routine was more time consuming (1 h, 30 min) due to a larger amount of measurements (1176 measurements in total). The measurements were collected by fixing an antenna at a specified depth in one borehole and lowering the other antenna by 0.25-m increments throughout the depth range of the second borehole. The fixed antenna was then moved to a new position where the procedure was repeated. In each borehole, the antenna was fixed every 1 m, resulting in 24 fixed antenna positions. Some of the data collected were later
disregarded, such as the measurements collected near the surface and traces having an acquisition angle $>45^\circ$. Attenuation of the signal and wave guiding along the antennae have been observed to deteriorate the quality of the high-angle traces (Peterson, 2001; Irving and Knight, 2005). The editing resulted in a data set with just 745 traces.

The first arrival time of each electromagnetic (EM) wave was picked individually, from which the velocity distribution was determined. The one-dimensional calculation associated with the ZOP data set is straightforward because the distance the EM wave had traveled was equal to the borehole spacing. On the other hand, the velocity distribution resulting from the MOG was achieved through two-dimensional tomographic inversion to account for all the crossing rays passing through each portion of the subsurface. For this procedure, the straight-ray MIGRATOM code (Jackson and Tweeton, 1994) was used. Straight-ray assumptions have been used successfully to analyze data where velocity contrasts are $<20\%$ (Peterson, 2001). If velocity contrasts exceed this value, the velocity may be misinterpreted. The first arriving ray is, in this case, no longer the direct ray, but instead rays traveling faster across longer distances.

The area between the boreholes was discretized by 50- by 50-cm uniform pixels, resulting in 25 × 11 model parameters. The total amount of cells or model parameters (275) was therefore less than the number of data values (745). This criterion was suggested by Jackson and Tweeton (1994) to remain within the resolution capability of the data and to provide robust estimates, even in the presence of noise.

The one- and two-dimensional velocity distributions were subsequently converted to relative permittivity and moisture content using

$$\sqrt{\varepsilon_r} = \frac{\varepsilon}{\nu}$$

and the empirical relationship of Topp et al. (1980):

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} \varepsilon_r$$
$$-5.5 \times 10^{-4} \varepsilon_r^2 + 4.3 \times 10^{-6} \varepsilon_r^3$$

where $\nu$ is the resulting velocity, $\varepsilon$ the radar wave velocity in air ($\sim 0.3 \text{ m ns}^{-1}$), $\varepsilon_r$ the bulk permittivity, and $\theta$ the moisture content at a given depth. Given that the subsurface at the field site was coarse-textured mineral soil with clay contents $<1\%$ in the measurement area, the Topp model was expected to give reliable results (Lesmes and Friedman, 2005).

Cross-Borehole Electrical Resistivity Tomography

Before initiating the experiment, 96 electrodes were installed in the field: 84 borehole electrodes and 12 surface electrodes. The latter were aligned along the dotted lines on Fig. 1 with 1-m spacing. Four electrodes were used for each resistance measurement. A measurement scheme, consisting of 2315 measurements and 2315 reciprocals, was constructed having only horizontal borehole dipoles (Winship et al., 2006). Horizontal dipoles result in higher values of measured transfer resistances and are therefore more robust toward noise (Binley and Kemna, 2005). The noise level was estimated by comparing the original measurements with their corresponding reciprocal measurement. In the reciprocal measurements, the current and potential electrodes used in the original measurements were interchanged. This value of noise has been found to represent the true noise level better than using duplicate measurements, which tends to underestimate noise levels (LaBrecque et al., 1996; Slater et al., 2000).

An IRIS SYSCAL Pro Switch 96 10-channel system (IRIS Instruments, Orleans, France) was used to record data during a 2-h, 15-min sampling period. A common data set having reciprocal errors $<10\%$ (1767 measurements) was used in the final data inversion, as suggested by French et al. (2002). The volume of interest was discretized into a numerical grid of 108 × 52 × 52 cells and solved for $29 \times 24 \times 24$ model parameters. In the area of interest (i.e., the infiltration area), the cells had a dimension of 50 by 50 by 50 cm, whereas the surrounding area was discretized more coarsely, up to 56 m. The finite-element-based, Occam-type, three-dimensional electrical resistivity inversion program R3 (Binley, 2007) was used to produce three-dimensional electrical resistivity tomograms. This inversion approach has been widely documented and applied in the literature (e.g., for tests of robustness; LaBrecque et al., 1996).

The obtained bulk electrical resistivity, $\rho_b$, is by Archie's empirical law (Archie, 1942) related to the pore water electrical resistivity, $\rho_w$, the porosity, $\Phi$, and the saturation degree, $S = \theta/\Phi$:

$$\rho_b = \rho_w \Phi^{-m} S^{-n}$$

where $m$ and $n$ are empirical constants set to 1.3 and 2, respectively. The parameter $m$ indicates the cementation of the soil. Unconsolidated sands, as found at the study site, are typically assigned a value of 1.3 (Lesmes and Friedman, 2005), while $n$ is most generally set to 2. This parameter accounts for the large increase of apparent electrical resistivity with decreasing soil moisture (Hendrickx et al., 2002). Surface conduction was disregarded as a first approximation since the subsurface below 1 m consisted of coarse-grained sands with a negligible amount of fine-grained material ($<1\%$). The top 1 m did, however, contain up to 14% fine-grained material ($<2 \mu m$) and the electrical conductivity may therefore be underpredicted in this area (Lesmes and Friedman, 2005). The porosity was assumed constant at 0.41 cm$^3$ cm$^{-3}$. This value was obtained from core samples withdrawn from the sand at 1- to 2-m depth.

By rearranging Eq. [3], the pore water electrical resistivity, $\rho_w$, can be found using the bulk electrical resistivity determined by cross-borehole ERT, the moisture content estimated from cross-borehole GPR, and porosity:

$$\rho_w = \frac{\theta \rho_b}{\Phi^{-m} S^{-n}}$$

Changes in soil water electrical conductivity were converted to meq L$^{-1}$ assuming that 100 $\mu$S cm$^{-1} = 1$ meq L$^{-1}$ (Appelo and Postma, 1999).

Infiltration Experiment

The forced infiltration tracer experiment lasted 20 d from 18 Oct. to 7 Nov. 2005. The infiltration experiment was designed to mimic one-dimensional water and solute infiltration within the measurement area to enable a one-dimensional analysis of the results. During 20 d, clean water (electrical conductivity = 760 $\mu$S cm$^{-1}$) was applied at a rate of 88.4 mm d$^{-1}$ over a 7.33-
7.33-m area using drippers spaced every 33 cm over the entire surface (484 drippers). Each dripper consisted of 85-cm-long, 1-mm tubing having enough resistance to ensure constant flow in all drippers. Tap water supplied from a nearby tap was used as irrigation water. The water was supplied to a water tank in which it was maintained at a constant level and from which it was distributed to the drippers. An independent water meter was connected to the system to verify that a constant flow rate was imposed. Furthermore, flow measurements were made at the individual drippers to ensure that water was uniformly applied to the field site. The field site was covered by a tent to avoid precipitation and to minimize evapotranspiration. The potential evapotranspiration was estimated to 0.8 mm d$^{-1}$ using the Penman–Monteith equation (Dingman, 1994) and data from a nearby weather station. The estimated evapotranspiration constituted <1% of the applied infiltration.

Geophysical measurements were taken on a daily basis at the start of the experiment and reduced to every 2 to 3 d toward the end. The entire measurement routine, comprised of MOG, ZOP, and ERT, was completed in 4½ to 5 h. During this time frame, a water front migrating with a vertical velocity of 1 m d$^{-1}$ will move up to 21 cm, a distance comparable with the vertical discretization of the ZOP, but half that of the MOG and ERT results. As a result, moisture content estimates can appear slightly smeared at the water front, but this effect should be negligible compared with the spatial smoothing of the geophysical methods.

After 4 d of infiltration, a saline tracer, 890 L containing 75 kg NaCl with an electrical conductivity of 105.4 mS cm$^{-1}$, was added during 150 min through the drippers. The applied tracer corresponded to 16.5 mm of water. Density flow was assumed to be insignificant, considering the high dilution of the tracer during the downward migration and the already dominant vertical gravitational flow.

**Results and Discussion**

Figure 3 shows the background (pretracer and preinfiltration) cross-borehole GPR- and ERT-derived moisture content profiles. The six collected ZOP (Fig. 3a) and six similar cross-sections withdrawn from the three-dimensional ERT parameter volume (Fig. 3b) exhibit some variability with depth. The overall trends, however, such as increased moisture content around 8 m followed by a drier sequence, are easily recognizable in each profile. The average variability of the moisture contents at each depth are 0.026 and 0.015 cm$^3$ cm$^{-3}$ for the six GPR and six ERT profiles, respectively.

The GPR data from 0.00 to 1.50 m is not included in Fig. 3. In this region, the first arriving electromagnetic signal may be a refracted wave traveling at the air–soil interface and not the direct wave passing through the subsurface. It can be difficult to distinguish between the different waveforms, rendering estimation of the moisture content at these depths problematic. To calculate moisture content from ERT, the soil water electrical conductivity and porosity were assumed to be constant throughout the measurement volume. The groundwater electrical conductivity ($=760$ μS cm$^{-1}$) sampled below the field site was assumed to be representative of the soil water electrical conductivity, while the porosity was set to 0.41 (the porosity determined for the sand layer at 1–2 m depth). Local-scale variations in porosity and soil water electrical conductivity may contribute to "noisy" results, while a general bias of the soil water electrical conductivity will bias the soil moisture content.

As expected, the ERT and GPR methods resulted in moisture profiles having similar trends and magnitudes (Fig. 3c). The ERT profile, however, indicates slightly higher moisture contents near the surface and lower moisture contents below approximately 8 m. To illustrate how well the ERT method, as applied in this study, can reproduce vertical resistivity variations of the subsurface, a synthetic ERT test is presented in Fig. 4. The ERT profile in Fig. 4 was obtained by constructing a synthetic resistivity model and using the ERT inversion algorithm to solve the forward problem for a selected measurement scheme. The synthetic measurements were then used as input for an ERT inversion. The synthetic resistivity model was constructed to represent the background GPR moisture content profile in Fig. 3c, having slightly larger moisture content values at selected depths (i.e., 6, 8, and 10 m) where thin layers of finer sand appeared likely. The top 1.5 m was also assigned a high moisture content ($\sim0.35$ cm$^3$ cm$^{-3}$) to mimic

![Fig. 4. The result of a synthetic electrical resistivity tomography (ERT) test is presented. The synthetic input resistivity profile and the inverted resistivity profile are converted to moisture content using Eq. [1].](image-url)
the higher retention of the topsoil indicated by the background ERT profiles in Fig. 3b.

The synthetic analysis of the electrical resistivity profile shows that the magnitudes of thick, low-resistance layers tended to be overpredicted by the selected ERT inversion algorithm. Furthermore, Fig. 4 illustrates that the ERT method, used in this study, suffered in prediction capability near the bottom of the profile due to fewer measurements that resulted in a low resolution. At this depth, the estimated resistivities, corresponding to the moisture content in Fig. 3b, are >1000 Ω m. High resistivity values, caused by low moisture contents, are intrinsically prone to higher measurement errors since the electrical contact resistance between the electrode and the surrounding subsurface may be very high. The cross-borehole ERT-derived moisture content values are therefore quite uncertain at this depth.

The vertical resolution of the ERT method is also lower than the GPR ZOP data, due to the larger vertical spacing of the electrodes (50 cm) compared with the vertical measurement spacing of the GPR (25 cm), and slight variations in moisture caused by layering will not, as a result, be as pronounced.

The vertical variations of the drained profiles in Fig. 3 are consistent with the layered changes in grain size distribution (see Fig. 2a). The top 1 m differs from the rest of the profile, having higher moisture content, and thereby indicating finer material with high retention. Below the topsoil the moisture content does not vary much, but slight changes exist. Most evident are the higher moisture content around 8-m depth and the low moisture content below this depth. This confirms the observed smaller grain sizes in Fig. 2a at 8 m and the underlying coarser material. But also the coarse material observed around 2- to 4-m depth in Fig. 2a can be recognized in Fig. 3. At this depth, low moisture contents down to 0.08 are observed.

Soil Moisture Content

The GPR-derived moisture content profiles (Fig. 5) and moisture content tomographic images (Fig. 6) collected throughout the infiltration experiment elucidate the downward water movement. During the first 7 d of infiltration, the water front migrated 0.50 to 1.50 m d$^{-1}$. After approximately 7 d, the infiltrating water reached 8 m, where further migration was temporarily halted. For the next 6 d, an increase in moisture content above the 8-m boundary is observed instead. At some time between Day 13 and Day 15, however, the water continued the downward movement. The tomographic images in Fig. 6 furthermore suggest that initially preferential flow took place and flow was not entirely one-dimensional, as would be expected due to the applied infiltration mode. Water flowed from the top 1.5 m through the left portion of the subsurface (at 2–5-m depth), bypassing almost completely a 2- by 2-m area at the very right of the image (near borehole GPR3). Below 4 m, the water moved laterally toward the right side of the image, whereafter the infiltration became laterally more uniform.

The estimated one-dimensional GPR moisture data can be used to evaluate the amount of water accumulated within the measurement volume (see Fig. 7). The mean moisture content profile is, for this estimation, assumed to be representative of the entire infiltration area. In order for this assumption to be valid, water infiltration should be uniform within the infiltration area.
The tomographic results in Fig. 6 as well as the six background ZOP in Fig. 3a did, however, show that this assumption is not valid. To illustrate the variance present in the infiltration area, the water accumulation estimated from the individual ZOPs are also plotted in Fig. 7 (gray curves). The individual profiles do exhibit some variance. In particular, ZOP34 and ZOP23 (indicated with arrows) have substantially lower water accumulation than the remaining four ZOPs, perhaps caused by the area bypassed by the water infiltration at 2- to 4-m depth, as discussed above. All curves show consistency in the accumulation trends, however, having high and fairly linear water accumulation during the first period of the infiltration experiment, after which the accumulation rate decreased during the last 10 d of measurement. It is especially evident from the average ZOP curve in Fig. 7 that the increase in infiltrated water was nearly constant during the first 6 to 7 d. After 7 d, the water accumulation stagnated, coinciding with the time the water front reached 8 m. The sediment at 8-m depth has small grain sizes and overlies coarser material (see Fig. 2a). A capillary barrier was most likely created at this depth. The water built up above the capillary barrier and at the same time lateral transport out of the infiltration area took place. This continued until the moisture content became great enough (or suction low enough) for the water to penetrate into the underlying coarser sand. After 13 d, the water broke through the capillary barrier and the water accumulated in the measurement volume increased again. The accumulation rate was, however, diminished compared with the first 7 to 8 d, indicating continued lateral flow.

Figure 7 also suggests that the average water accumulation rate in the measurement volume during the first 8 d of 2.16 m$^3$ d$^{-1}$ differs by more than a factor 2 compared with the rate at which water was infiltrated through the drippers of 4.75 m$^3$ d$^{-1}$ (as measured by the water meter at injection). The difference between these two infiltration rates is substantial. Several plausible causes were examined to determine the reason for this difference. Among others, the moisture content estimates could be erroneous due to refraction, as suggested by Rucker and Ferré (2004a). They found water balance errors up to 41% when conducting synthetic tests of layered sediments without accounting for refracted arrivals of the EM wave. The proposed methodology was applied to the ZOPs collected at our field site but refraction could not explain the large water balance error. It only led to small corrections in moisture content near the surface (top 2 m) and at the water front.

Inadequate petrophysical relationships could perhaps also explain the disagreement. Binley et al. (2001) and West et al. (2003) discussed the necessity of determining site-specific relationships to convert relative permittivity values to soil moisture; however, using different petrophysical models did not improve the water balance result. Furthermore, it has been shown that changes in volumetric moisture content, when using the complex refractive index method, are only dependent on the background permittivity value, the permittivity at time $t$, and the permittivity of air and water (Binley et al., 2002) and are therefore not dependent on the permittivity of the soil in question.

Furthermore, as shown in Fig. 3c, the moisture content profiles derived from GPR and ERT are in good agreement, suggesting that the large quantities of missing water are not due to flaws in the measurement or conversion procedures, but were more likely caused by uncontrolled subsurface diversion of the infiltrated water. In fact, the particle size data suggests that a sequence of fine sand overlying coarser sand is present at 1- to 2-m depth, which can create a hydraulic barrier and associated lateral diversion of the infiltrated water. This effect is observed at the 8-m depth where the contrast in grain size distribution, according to Fig. 2a, is smaller than at the 1- to 2-m depth.

Unfortunately, the design of the infiltration experiment does not allow a quantification or identification of lateral flow since the geophysical measurements were constrained within the infiltration area. The three-dimensional electrical resistivity parameter volume does extend beyond the infiltration area. The electrical resistivity estimates in this area are highly uncertain, however, and detection of small-scale lateral flow mechanisms is not possible.

![Fig. 7. Accumulated difference in water volume estimated from the individual zero-offset profiles (ZOPs) (gray curves) and an average value (black curve).](image)

![Fig. 8. Average electrical resistivity profiles measured using electrical resistivity tomography at various days. The gray curve is the background electrical resistivity profile, i.e., Day 0.](image)
Figures 8 and 9 illustrate the changes in bulk electrical resistivity values determined by the ERT method. In Fig. 8, the electrical resistivity is averaged at each depth from the three-dimensional tomograms to represent a one-dimensional profile, while the two-dimensional electrical resistivity images (Fig. 9) were extracted from the three-dimensional inversion results. The tomographic images represent the same area (between Boreholes 1 and 3) sampled with the GPR method in Fig. 6. The water infiltration during the first 4 d only lowered the electrical resistivity slightly near the surface. There is a clear reduction in electrical resistivity (an increase in electrical conductivity) after 5 d, however, corresponding to the time the tracer was added. The subsequent nine profiles and images indicate the tracer’s downward movement and vertical spreading.

The large range of electrical resistivity values of the subsurface, from tens to several thousands of ohm meters, and the coarser vertical discretization mask the details of the infiltration dynamics somewhat. It is possible to distinguish the water front movement ceasing at approximately 8-m depth at Day 7 to 8, however, and the following buildup of water above this depth. The electrical resistivity results also indicate a breakthrough of water below the 8-m boundary around Day 13 and thereby corroborate the infiltration dynamics observed using GPR.

Electrical Resistivity Tomography and Ground Penetrating Radar Image Comparison

The two-dimensional images of the moisture content and electrical resistivity shown in Fig. 6 and 9 clearly show some structural differences. This is to be expected, as the two methods have very different resolution and sensitivity patterns within the plotted area. Ground penetrating radar has a higher resolution in areas with many crossing ray paths (i.e., in the central part of the interwell region) and ERT has the highest resolution around the electrodes (located in the boreholes and at the surface; see, among others, Day-Lewis et al., 2005). Furthermore, changes in electrical resistivity occurred due to changes in moisture content as well as changes in soil water electrical conductivity (Eq. [3]).

Nonetheless, the infiltration patterns are quite consistent. The tomographic images of the two methods both visualize an initial infiltration slightly skewed toward the left of the image (Borehole 1). This becomes especially evident when changes in moisture content for the first 4 d are plotted (see Fig. 10).

Tracer Mass

When the cross-borehole ERT and GPR results are combined to estimate the tracer concentration, following Eq. [4], the tracer movement and spreading is better visualized. Figure 11 and Fig. 12 show the one-dimensional tracer profiles and the two-dimensional tracer images, respectively. Since GPR moisture content estimates were not available for the top 1.50 m, the ERT moisture content data from the day before the tracer application (Day 4) were used as an approximation. Electrical resistivity data in Fig. 8 and 9 illustrate that this approximation is not too rough, as there does not appear to be further electrical resistivity changes in the top 1.50 m after 4 d. Instead the resistivity changes are observed deeper, around 2- to 4-m depth, indicating that the water front had passed the top layer and steady-state conditions could be expected to be present.
Figure 11 illustrates how the shape of the tracer pulse became more dispersed as it migrated downward in the profile. At Day 13, only a little spatial variation in concentration remains, and after 15 d the pulse is not recognizable any more. A better understanding of the tracer movement can be achieved from examining the two-dimensional images in Fig. 12. The tracer infiltration behaves fairly one-dimensionally the first 2 d, i.e., Days 5 and 6. In the following images, however, some of the tracer appears to move laterally out of the area of measurement toward the left of the image. This observation may be caused by the temporary standstill of water at 8-m depth, but it could also be an artifact as the ERT method is more sensitive toward low values of electrical resistivity near the edges of the images. After Day 10, the tracer migrates downward again, but has by this time become so diluted that it is difficult to distinguish the tracer plume from background noise.

These results illustrate that the cross-borehole geophysical methods enable a nondestructive mapping of the migration and spreading of an applied tracer, and the methods provide data on a refined spatial scale comparable to the scale of traditional water sampling from suction probes. The varying spatial resolution of the computed images resulting from cross-borehole ERT and GPR, as mentioned above, is, however, still an issue that needs to be dealt with.

In an attempt to overcome these shortcomings, transfer functions (e.g., random field averaging [Day-Lewis et al., 2005]), apparent petrophysical relations (Singha and Gorelick, 2006; Singha and Moysey, 2006), and full-inverse statistical calibration (Moysey et al., 2005, 2006) have been applied to cross-borehole geophysical data. Joint inversion of multiple data types can also be used to produce more reliable geophysical models (Kowalsky et al., 2005; Linde et al., 2006).

**Moment Analysis**

The one-dimensional mass profiles in Fig. 11 were used for a one-dimensional moment analysis. In moment analysis, the location of the center of mass and how the tracer pulse disperses at consecutive times is quantified. By computing the zeroth, first, and second moments for the profiles, it is possible to estimate how mass balance, pore water velocity, and dispersivity develop with time. (For a description of moment analysis, see, e.g., Singha and Gorelick [2005].) The results of this analysis are plotted in Fig. 13a to 13c. The results clearly reveal a temporal development of both mass balance and transport parameters. The pore water velocity of the tracer declines throughout the measurement period, decreasing most substantially during the first 4 to 5 d. After 5 d, the pore water velocity stabilizes at approximately 0.3 m d$^{-1}$. The dispersivity value is around 0.2 m for the first 5 d and then increases to 0.5 m during the last period of the infiltration experiment. The changes in both pore water velocity and dispersivity coincide with the standstill of the water front at approximately 8 m. At the same time, the tracer mass begins to decrease, which complicates the interpretation of the results in a one-dimensional framework, as water and tracer are diverted laterally out of the infiltration area.

The amount of mass distributed over the area was 17.3 meq m$^{-2}$. This amount is double the amount of mass estimated using the cross-borehole geophysical techniques. The lack of mass is in accordance with the previously discussed error in water balance, and again confirms the speculation that tracer and water moved laterally out of the measurement area in the top 1.0 to 1.5 m.

It is important to note that, since the basic assumption behind the one-dimensional moment analysis is violated due to the occurrence of lateral flow, the temporal development of
the transport parameters cannot be considered to represent the true physical properties of the subsurface. The observed reduced vertical velocity cannot therefore be directly associated with the observed increased vertical longitudinal dispersivity.

A two-dimensional moment analysis was used to determine the center of mass of the tomographic images in Fig. 12. The location of the first moment is illustrated with a red circle in Fig. 12. The tracer is observed to move slightly toward the left of the image from Day 8 to Day 13. After Day 13, the center of mass moves back toward the middle of the image. The changes in the zeroth moment (not shown here but identical to the results of the one-dimensional moment analysis in Fig. 13a) show substantial mass loss after Day 8. As the majority of the mass is no longer contained within the images in Fig. 12 after Day 8, the moment analysis results cannot therefore be assumed to represent the complete transport regimes.

**Synthetic Test**

A synthetic numerical experiment was performed to examine whether the ERT inversion method would inherently alter the shape and magnitude of the estimated tracer plume and thereby influence the resulting transport parameters. A forward one-dimensional hydrologic simulation of flow and transport was performed using the code HYDRUS-1D (Simunek et al., 1998) for the soil and transport parameters listed in Table 1 (i.e., typical values for sand in HYDRUS-1D). The development of tracer concentration and moisture content profiles in a 12-m profile were calculated for a 16-d-long forced tracer infiltration experiment with constant loading. The electrical resistivity profiles were calculated using Eq. [3] and used as input for a forward model calculation using the R3 ERT inversion program (assuming the petrophysical relationships as discussed above). Resistance data for the same measurement configuration as used in the field were thereby obtained and subsequently used as input for inversion. Finally, the electrical resistivity profiles were combined with the simulated moisture content profiles to compute tracer concentration (Eq. [4]) and these profiles were then used in a moment analysis.

The purpose of the synthetic test was not to represent reality; it was designed only as a means to identify errors associated with the ERT method. Random errors were not added to the synthetic data and one-dimensional flow through a homogeneous medium was assumed. The synthetic test cannot, as a result, be compared with the moment analysis results conducted on the real data, where lateral flow and loss of tracer mass greatly influenced the results.

The true and estimated values of mass, pore water velocity, and dispersivity based on the one-dimensional model are shown in Fig. 13d to 13f, where the true values are represented by a gray horizontal line. Figure 13d illustrates that the tracer mass is poorly determined by the ERT method. Apart from Days 1 and 2, the mass was generally overpredicted by 30%. This systematic overprediction of tracer mass can be explained by the results of the synthetic analysis presented in Fig. 4. Low electrically resistive anomalies extending horizontally across the entire measurement volume will be underpredicted using the adopted ERT inversion routine. As a result, too-high electrical conductivity and tracer concentration estimates were produced locally.

Although the pore water velocity was also slightly overpredicted for the majority of the simulated days (Fig. 13e), the estimated values are nonetheless quite close to the true value (difference ≤10%). The determination of the pore water velocity is therefore considered quite robust. On the other hand, the longitudinal dispersivity appears to be difficult to determine correctly (Fig. 13f) due to smoothing of the model parameters and the chosen vertical discretization. Initially the tracer plume was narrow and the vertical discretization too coarse to capture the plume properly. As the plume penetrated deeper into the profile and became more dispersed, however, the method did a better job in capturing the shape of the plume and the dispersivity was estimated more accurately. On the other hand, when the plume was located deeper in the subsurface and came close to the bottom of the measurement volume, where the resolution of the ERT method is low, the error increased again.

The correct estimation of the transport parameters is seen to depend highly on how well the vertical changes in electrical resistivity are captured by the ERT method. The survey design,
i.e., electrode spacing, borehole separation, and measurement configuration, should therefore be optimized toward the vertical resolution to produce reliable results.

Conclusions

Cross-borehole ERT and GPR provide a promising alternative to existing methodologies designed to monitor tracer infiltration. We have demonstrated that when both cross-borehole methods are combined, tracer mass profiles and images can be obtained. The one-dimensional profiles enable a determination of important transport parameters through moment analysis, while the two-dimensional images provide a means to better understand and visualize transport dynamics. The moment analysis also helped understand the observed water loss and clearly illustrated how even small structural changes in layered sediments can result in capillary barriers and associated lateral flow, thus affecting the downward migration dramatically. Lateral flow was not expected to be such a dominant flow mechanism at the selected field site and the infiltration design, aimed at analyzing one-dimensional properties, was therefore not ideal.

The findings in this work stress the importance of investigating the limitations of the geophysical methods. We illustrated this by examining how the chosen ERT inversion method inherently changes the shape and magnitude of the tracer plume. This is one of many factors that could have an impact on the results. Future work should therefore concentrate on finding ways to deal with these shortcomings. This could be attempted by further developing the work concerning transfer functions, and thereby accounting for the spatial variability of the resolution of the geophysical methods, or by adopting completely new approaches, such as integrated data fusion (described in Moysé et al. [2006] and explored in, e.g., Binley and Beven [2003]). In this methodology, the collected geophysical data is used to evaluate the likelihood of a series of plausible geophysical models by computing forwarding data for these models and comparing the misfit of the collected and modeled data. By using the data directly, without first producing tomographic images through inversion, uncertainties introduced by the inversion procedure are avoided.

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