HYDROGEOPHYSICAL INVESTIGATIONS OF UNSATURATED FLOW

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M.Sc.-Thesis
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PREFACE

The present M.Sc-thesis is the result of a one-year Master’s study in Geology-Geoscience at the Department of Geography and Geology, Faculty of Science, University of Copenhagen, Denmark. Supervisors for the project were Majken Caroline Looms, Karsten Høgh Jensen and Lars Nielsen. Field work for the thesis was carried out during 2008 and 2009 as two individual experiments at two different field sites in Denmark.

The thesis is comprised of firstly an introductory section regarding the theory that has been used, secondly a summary and discussion of two papers that are the result of the two field experiments. Appendices A and B contain the paper manuscripts, while statements of authorship regarding the papers are found as Appendix C.
## CONTENTS

### PREFACE

### ABSTRACT

1. INTRODUCTION AND OBJECTIVES 7

2. THEORY AND METHODS 9

2.1 The unsaturated Zone 9

2.2 Ground penetrating radar 13

2.3 Moisture content measurement 18

2.4 Preferential flow 19

3. PRESENTATION OF PAPERS 23

3.1 Paper I - Appendix A 23

3.2 Paper II - Appendix B 25

4. CONCLUSIONS AND PERSPECTIVES 29

ACKNOWLEDGEMENTS 30

REFERENCES 31

## APPENDICES

<table>
<thead>
<tr>
<th>A</th>
<th>Paper I</th>
<th>A1</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>Paper II</td>
<td>B1</td>
</tr>
<tr>
<td>C</td>
<td>Authorships statements</td>
<td>C1</td>
</tr>
</tbody>
</table>
ABSTRACT

Understanding the processes in the unsaturated zone becomes increasingly important as the struggle to protect groundwater resources intensifies, because the quality and quantity of groundwater is essentially determined by properties in the unsaturated zone. Unfortunately, unsaturated flow and transport are both difficult to predict and measure. This M.Sc. thesis presents the results of two field infiltration experiments that were monitored using ground penetrating radar measurements.

The first field experiment was a plane infiltration at Hjelm Heath, Denmark, for which we obtained a densely sampled ground penetrating radar (GPR) dataset before and after infiltration for subsequent analysis of the water movement. To map the preferential flow paths induced by the infiltration we added a dye tracer to the infiltrated water, and it was investigated whether those patterns could be recognised in the GPR data. Detailed analysis of the GPR data set demonstrated that the unsaturated flow at the field site is highly affected by layering and small geological changes in the subsurface. Reflections were delayed significantly within the infiltration area and a water balance based on the delay could account for almost all the infiltrated water. Amplitudes were shown to be reduced within the infiltration area due to attenuation of the electromagnetic signal.

The second field experiment was a point injection conducted at the field site Arrenæs, which is equipped with boreholes for cross-borehole geophysical measurements. The moisture content development was monitored by both GPR and Electrical Resistivity Tomography (ERT) during and after infiltration. Cross-borehole GPR data was obtained using two acquisition schemes: zero offset profiling and multiple offset gather. The data types rendered simple moisture content profiles and 2-dimensional velocity tomograms from inversion of arrival times, respectively. The GPR data showed that an infiltration plume developed as a result of the injection of water. Movement and expansion of the plume was not homogeneous and seemed to be susceptible to layering in the subsurface. Future analysis of the ERT data is expected to further elucidate the movement of the plume.
1. INTRODUCTION AND OBJECTIVES

As scientists and laymen all over the world become increasingly aware of the environmental and climatic impact human activity has on nature, groundwater is also becoming an important topic. Water resources as a life-sustaining necessity are threatened, not only in parts of the world where water scarcity is an everyday problem, but also in areas with otherwise stable water supplies. The unsaturated zone constitutes the link between atmospheric and surface hydrological processes and the groundwater. Processes in the unsaturated zone ultimately determine the quantity and quality of the groundwater, which is important both to scientists and decision-makers, and sustainable exploitation of water resources will be a theme on the political arena for many years to come.

The main objective of this M.Sc.-study is twofold: Firstly, we wanted to investigate the possibility of assessing spatial flow patterns in the unsaturated zone induced by a forced infiltration by means of high-resolution reflection ground penetrating radar (GPR). Second, cross-borehole ground penetrating radar and electrical resistivity tomography measurements during a point injection of water were analysed with the purpose of further highlighting these methods’ applicability in alluvial sediments.

Unsaturated flow and transport prediction is severely complicated by the prevalence of preferential flow caused by the spatial variability of hydraulic properties in the unsaturated zone. The existence of such preferred flow paths can pose a threat to the groundwater quality because they can carry not only water but also pollutants to great depths and thus bypass the natural filter, which the unsaturated soil profile constitutes. Monitoring of water movement in the unsaturated zone is a difficult task that is not carried out effectively using methods that are not able to capture the spatial variability and heterogeneity associated therewith. Point measurements of moisture content, using e.g. time domain reflectometry probing or sample collection are not particularly useful for tracing the wetting front on a
larger scale. Hydrogeophysics has evolved over the past decade as more methods become available for measuring and monitoring infiltration and moisture content at appropriate scales. Cross-borehole geophysical methods have become popular for assessing unsaturated zone processes as they are less destructive and provide data with a high spatial resolution.
2. THEORY AND METHODS

2.1 The unsaturated Zone

The unsaturated zone is found between the surface and the groundwater table, it is also commonly called the vadose zone, vadose being Latin for “shallow”. Processes in the unsaturated zone determine the quantity and quality of the water that eventually recharges to the groundwater at the water table. The unsaturated zone thus constitutes an important link between surface hydrological processes and groundwater hydrology, of which surface hydrological processes are highly variable whereas the groundwater usually is in a more stable condition. As the transition zone between two very different domains the unsaturated zone contributes to the redistribution of water through processes such as infiltration, evaporation, interflow and recharge as well as root water uptake by plants. (Dingman, 1994; Jensen, 2004). A soil profile is generally heterogeneous regarding texture, which causes other parameters to vary as well, such as hydraulic conductivity and soil moisture. The interaction between water and soil determines the distribution of water in the unsaturated zone.

In the unsaturated zone the subsurface consists of soil particles, air and water. Water is retained in the soil by forces of capillarity between individual water molecules and adsorption between water and soil particles (Jensen, 2004). The latter of these forces is by far the strongest and the combined effect of these determine the retention characteristics of a soil, which is highly dependent on soil texture. Sandy soils with large pore spaces retain water primarily by means of capillarity, whereas water in clayey soils adsorbs to the individual grains where it is held back more effectively by the capillary forces. Hence, a sandy soil will release more water at small pressures, i.e. it dries up easier, whereas water is retained better in a clayey soil. Although more water is retained in clayey soils, it is harder for plants to access because of the strong adsorptive forces of.

The gravimetric water content of a soil is the ratio of water to soil, and is denoted by the letter \( \theta \). As water can only exist in the pore space between soil particles, the maximal water content is when the entire porosity is water-filled, at which point the soil is saturated.
The lowest obtainable water content is of course $\theta = 0$, however this rarely happens in nature, because of capillarity effects. The water content of a soil can also be represented by the water saturation, which is merely the percentage of the porosity that is inhabited by water. Thus, by definition the water content can be between 0 and the porosity, whereas water saturation is between 0 and 100%.

As a three-phase system consisting of soil, air and water, the unsaturated zone is characterised by a pressure difference between the air and water phase, called the capillary pressure, $P_c$,

$$P_c = P_a - P_w$$  \hfill (1)

where $P_a$ and $P_w$ are the air and water pressures, respectively (Jensen, 2004). The air pressure, $P_a$ is assumed to be equal to atmospheric pressure and with this as reference level ($P_a = 0$), the capillary pressure is equal to the negative water pressure and equation (1) reduces to $P_c = -P_w$. The soil water tension, $h_c$ is the capillary pressure expressed in terms of length, also called capillary pressure head:

$$h_c = \frac{P_c}{\rho_w g} = \Psi$$  \hfill (2)

$\rho_w$ and $g$ are the the density of water and the gravitational force, respectively. Since the water pressure is negative, the soil water tension is also negative. $\Psi$ is the soil water pressure head, which is negative in the unsaturated zone and positive below the groundwater table (Jensen, 2004).

Attraction of water to soil particles is responsible for the capillary rise seen above the water table, where the soil is still saturated. For a soil of equally sized capillaries, desaturation will occur just above the capillary height, whereas a heterogeneous soil with capillaries of different sizes will desaturate gradually (Jensen, 2004). Retention parameters for a soil describe how wetting and desaturation take place. The retention curve for a soil thus relates water content, $\theta$ to soil water pressure head, $\Psi$. The most widely used equation for the retention curve is the van Genuchten retention function (Jensen, 2004):
Se is the effective saturation which is calculated based on the residual water content, \( \theta_r \) and the current saturation, \( \theta \), where \( \theta_r < \theta < \theta_s \). \( \alpha \) and \( n \) are parameters related to the particular soil. \( \alpha \) is related to the air entry suction and thus determines the suction or soil water potential at which the soil starts draining. \( n \) is related to the texture, i.e. pore-size distribution, and determines the slope of the retention curve (Jensen, 2004; Dingman, 1994).

\[
S_e = \left(1 + (\alpha \cdot P_c)^n\right)^{-\frac{1}{n}} \tag{3}
\]

\[
S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \tag{4}
\]

Figure 1.
Retention curves for three soil types.

Principal retention curves for three different types of soils; sand, loam and clay are seen in Figure 1. Each curve intersects the y-axis at the saturated water content, \( \theta_s \) of the soil. When suction, \( \Psi \) is increased the soil eventually reaches its air entry suction and full saturation can no longer be sustained by the soil; it starts draining and the retention curve breaks off from saturation. Given that suction is increasing the soil will drain further and the slope of the retention curve during drainage is determined by the rate at which the soil drains.
releases water. At very high suction rates the soil reaches its residual water content, where no more water will drain from the pores because it is held too tightly by adsorptive forces. In other words, $\alpha$ in Equation (3) determines when the soil starts releasing large amounts of water, and $n$ determines how fast the soil drains from saturation to residual water content. It should be noted that the retention curve for wetting does not match that of draining, due to hysteretic effects. When wetting a sample in stead of draining, the full saturation will not be achieved due to the presence of air bubbles, a condition called natural saturation (Jensen, 2004).

**Unsaturated flow equations**

Flow in the unsaturated zone can in general be described as for the saturated zone by the use of Darcy’s law, which states that water discharge is proportional to the hydraulic head gradient with the hydraulic conductivity and the cross-sectional area for flow as the proportionality factors. However, the simplicity of Darcy’s law is complicated by the fact that the hydraulic conductivity is water content dependant in the unsaturated zone, hence the retention parameters play a role in the flow equations. The governing differential equation for vertical flow in the unsaturated zone known as Richard’s equation was developed as a combination of conservation of mass and Darcy’s flow equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left( K(\theta) \frac{\partial \Psi(\theta)}{\partial z} \right) - S$$  \hspace{1cm} (5)

In this form of Richard’s equation, $\psi$ is the pressure head, $K$ is the hydraulic conductivity, $S$ is a sink term, $z$ is depth and $t$ is time (Dingman, 1994). The partial differential equation is non-linear and does not have any closed-form analytical solution (Jensen, 2004). Equation (5) is valid for vertical flow in the $z$-direction, but similar equations can be deducted for 3-dimensional flow. The first term on the right side represents the contribution of suction by the capillary pressure head, whereas the second represents the contribution of gravity to the unsaturated flow. Each of these terms vary and may be more or less dominating in a given soil depending on the initial water content.

In Richard’s equation, both $K$ and $\Psi$ are water content dependent, so in order to solve the equation many estimates of the relationship between $K$, $\Psi$ and water content have been developed.
produced. Some of these equations are only valid for specific soil types or for specific saturation intervals. In general, the hydraulic conductivity, $K$, decreases significantly with decreasing saturation. When pores in the soil empty the flow paths become more tortuous and at the same time, the cross-sectional area for flow decreases (Fitts, 2002). At very low moisture contents the hydraulic conductivity approaches 0, because the water is tightly bound to the soil particles.

Retention characteristics and subsequent soil moisture development and changes are important for both the crop yield in agricultural areas as well as for the recharge to the groundwater. Field capacity is commonly used to describe the water that is present in the soil and available to plants after natural gravitational drainage has carried precipitated water further into the subsurface. The permanent wilting point is another term often used to characterise soils. It denotes the very low moisture content, beyond which the plants cannot recover when soil moisture increases again (Shaw, 1994).

2.2 Ground penetrating radar

Ground penetrating radar has been used for glaciological purposes since the 1960s and became popular for archaeological and engineering applications in the 1980s (Reynolds, 1997). The possibility of adjusting the signal frequency depending on the type of study makes GPR suitable for many purposes within the geological sciences.

The GPR setup consists of a transmitter and a receiver. The two can be separated in the bistatic mode or the same antenna can be used both for transmitting and receiving, called monostatic mode (Reynolds, 1997). While the transmitter generates a pulse of electromagnetic waves, the receiver records the signal within an appropriate time window determined by the user. The time window must be large enough to encompass the two-way travel time range, for which data can be extracted from the received signal. The velocity of the electromagnetic wave through a material is given by

$$v = \frac{c}{\sqrt{\frac{\mu r F_r}{2} \left(1 + P_r^2\right)^{\frac{1}{2}}}}$$

(6)
where $c$ is velocity of light in vacuum ($= 0.3 \text{ m/ns}$), $\mu$ and $\varepsilon$ are the magnetic permeability and dielectric permittivity of the material, respectively (Reynolds, 1997). $P$ is the loss factor, which is dependent on the conductivity of the subsurface and the signal frequency. This parameter can be disregarded ($P \approx 0$) for low-loss materials with low conductivities (Reynolds, 1997). $\mu$ can be set to unity ($= 1$) for most unmagnetic materials, and so in most cases it is the dielectric permittivity which is the significant parameter and the electromagnetic velocity reduces to:

$$v = \frac{c}{\sqrt{\varepsilon}} \quad (7)$$

Most geologic materials have electromagnetic velocities of 0.05 - 0.2 m/ns (Reynolds, 1997). The dielectric permittivity of water is $\varepsilon = 80$, hence the presence of water in the subsurface will lower the electromagnetic velocity of the geologic material (Kearey et al., 2002). Topp et al. (1980) performed numerous experiments and collected empirical data to determine the relationship between the dielectric constant and the moisture content of a soil for a electromagnetic frequency range of 1 MHz - 1 GHz. They found a strong correlation between the dielectric constant, $\varepsilon$ and the moisture content, $\theta$ of samples and derived an empirical equation relating the two:

$$\theta = -5.3 \cdot 10^{-2} + 2.92 \cdot 10^{-2} \varepsilon - 5.5 \cdot 10^{-4} \varepsilon^2 + 4.3 \cdot 10^{-6} \varepsilon^3 \quad (8)$$

This equation has since provided the foundation for numerous studies in which GPR signal analysis is used to estimate moisture content (Binley, 2002a & 2002b; Looms, 2008; Huisman, 2003). Ferre et al. (1996) simplified the empirical equation of Topp et al. (1980) to a more easy-to-use format:

$$\theta = 0.1181 \cdot \sqrt{\varepsilon - 0.1841} \quad (9)$$

The advantage of equation (10) as compared to the original Topp equation is that conversion back and forth between the variables, $\theta$, $v$ and $\varepsilon$ is straight-forward. Equation (10) was shown to provide results for moisture contents between 0.05 and 0.40, which differed less than 5% from the original Topp equation (Ferre et al., 1996).

When propagating through a given media the electromagnetic wave loses energy due to attenuation effects, adsorption and spreading (Reynolds, 1997). Attenuation is a function
of the dielectric and electric properties of the media as well as the frequency of the signal itself. Adsorption occurs when parts of the electromagnetic signal is converted into heat. Because the electromagnetic signal is transmitted in a conical beam, this causes a reduction in energy due to spreading.

**GPR survey types**

When the transmitted electromagnetic signal encounters an interface between two dielectric permittivities, parts of the signal will be reflected back to the surface according to the reflection coefficient

\[ R = \frac{(v_2 - v_1)}{(v_2 + v_1)} \]  

(10)

\( v_1 \) and \( v_2 \) are the electromagnetic velocities on either side of the interface. Calculation of \( R \) is analogous to seismic reflection surveying. When such an interface is reached the signal loses some energy due to the reflection (Reynolds, 1997). Large penetration depths of the signal can be achieved in areas of low conductivity, and with low frequency antennas. It is also important to notice that there is a trade-off between penetration depth and resolution when working with reflection GPR. The higher the frequency the better the resolution but the lower the penetration depth and vice versa.

![Figure 2. Overview of wave propagation in soil with interface between 2 layers of differing dielectric permittivity.](image)

An overview of the different waves and their travel paths from transmitter to receiver is depicted in Figure 2. The figure shows the two antennas at the surface and how the emitted signal encounters an interface between two layers of different dielectric properties.
The first waveform to arrive at the receiver is the direct wave (or couple of waveforms) that has travelled through the air between transmitter and receiver. Next comes the ground wave, after which reflections from the subsurface arrive. Just as in seismic surveys, the radargram will contain multiples of prominent reflections which must not be confused with real reflections. Migration can also be applied as a processing tool for restoration of real dips of reflections, when structures in the subsurface create diffraction effects in the data. The signals that are recorded by the receiver are usually displayed in a radargram for subsequent processing and analysis.

Surface based GPR survey data such as those obtained from reflection GPR have the disadvantage of losing substantial resolution with depth and cannot be used for determining small scale changes below the first couple of meters of the subsurface. Cross-Borehole GPR was developed in an attempt to overcome this problem (Looms, 2007) because the method allows for more high-resolution assessment of the velocity distribution in the subsurface. As the antennas are lowered into previously installed boreholes, this provides the method with yet another advantage, namely that it can be used in areas with high-conductive topsoil without signal loss to obtain usable data even at depth.
Two main data acquisition schemes for cross-borehole GPR exist (Figure 3). In the Zero Offset Profiling (ZOP), the antennas are lowered simultaneously in to each borehole so that their midpoint is always at the same depth (Huisman et al., 2003). The distance between measurements can vary, but is often 0.25 m (Binley, 2002a; Looms et al., 2008). The acquisition time for a complete ZOP profile between 2 boreholes is low, which makes the method suitable for measuring transient processes in the unsaturated zone, for example as a result of an infiltration experiment (Huisman et al., 2003). In the Multiple Offset Gather (MOG) (also called Multiple Offset Profiling, MOP) one antenna is kept at a fixed depth, while the other is moved down the other borehole. The procedure is then repeated with the fixed antenna at a new depth. MOG data can be used to reconstruct a tomographic velocity distribution between the two boreholes and thus provide detail about the two-dimensional distribution of moisture content that cannot be attained from ZOP data (Huisman et al., 2003).
Figure 3 shows an overview of the cross-borehole acquisition schemes. Notice that it takes two people to carry out the ZOP, because the antennas are lowered simultaneously, whereas one person can carry out the MOG, since one antenna is fixed in the borehole. Acquisition of a MOG between two boreholes is more time consuming than the corresponding ZOP and thus MOG is not as suitable for monitoring transient changes that happen over small time scales. However, in the case where multiple boreholes exist and ZOP data is being collected between these as well as between the two used for MOG, acquisition time for the two methods is almost the same.

2.3 Moisture content measurement

There are several methods, with which it is possible to measure soil moisture. The most precise method is measuring the gravimetric water content by comparing the original and dry weight of a sample. The volumetric water content can subsequently be calculated if the density of the soil is also known. Measuring the water content using this method, however precise it might be, is laborious, destructive, unrepeatable and it is also difficult to obtain samples from large depths. Indirect measurements of soil moisture include neutron probing and Time Domain Reflectometry (TDR). Neutron probing measures the amount of water in a soil by emitting fast neutrons into the matrix where they lose kinetic energy by colliding with hydrogen atoms in the water molecules. The amount of slow moving neutrons counted around the probe is translated into water content using a calibration curve which is dependent on the soil’s bulk density (Jensen, 2004). There is an almost linear relationship between water content and the amount of slow moving neutrons. Because emission of fast neutrons requires a source of radioactive decay, the equipment for neutron probing must be handled with some caution. Also, the method can not be used close to the surface because the fast moving neutrons will interfere with the atmosphere. In very dry soils the neutron cloud will be allowed to expand far into the formation and thus measurements of soil moisture will be less point specific.
Time domain reflectometry (TDR) is a method which measures changes in dielectric permittivity which is strongly related to moisture content (Topp et al., 1980). The TDR probe consists of a pair of metallic rods of a known length which are inserted into the porous medium of interest. An electromagnetic pulse is directed along the rods and the travel time through the porous medium is used for calculation of the dielectric permittivity (Evett, 2003; Jensen, 2004). The TDR method can measure moisture contents up to 0.5 with high accuracy and repeatability (Evett, 2003). TDR probe length can vary from a couple of centimeters to more than a meter. TDR measurements are essentially the integrated moisture content over the length of the probe, i.e. a mean value. Long probes will therefore not capture the spatial variation in soil moisture, because they measure mean moisture content (Evett, 2003). For TDR to provide detailed information about the spatial variability of small scale moisture content differences, installation of many probes at many depths is needed, which is not always a viable possibility.

2.4 Preferential flow

It has long been acknowledged that water flow in soils is rarely uniform and that preferential flow often occurs as the bypassing of certain parts of the subsurface (Simunek, 2003) (Beven et al., 1982). Unstable wetting fronts occur where the subsurface is not homogeneous and are further facilitated by the presence of cracks, old roots, wormholes and the like. Preferential flow is responsible for the unpredictable movement of both water and solutes, which can carry both to greater depths than what is predicted by the commonly used flow equation such as Richard’s equation. Water in the unsaturated zone will percolate downwards due mainly to the gravitational pull and will tend to move through the largest continuous pores that are filled with water. This is resembled in the unsaturated hydraulic conductivity which increases dramatically as the moisture content increases (Jensen, 2004; Fitts, 2003). Simunek et al., (2003) summarise the causes of preferential flow to include structural features such as macropores, flow instability caused by soil heterogeneities, and finally channelling due to sloping layers in the subsurface (Simunek et al., 2003). Preferential flow happens on scales ranging from very small-scale
differences in grain size, which might not even be visible to the naked eye to the much larger macropores, which are easily distinguishable. In this context macropores which facilitate fast percolation of vast amounts of water is of great importance for the estimation of both recharge and pollution threats (Beven et al., 1982)

Effective visualisation of infiltration patterns is of great importance for the studies of preferential flow, and for this purpose dyes are very valuable tools. An ideal tracer behaves like the infiltrating water and is characterised by being conservative, i.e. no retardation; the tracer should not be present in any background concentrations and finally it should not be subject to any chemical or biological degradation. Different types of tracers are appropriate for different purposes. Dye tracers consist of large organic molecules, which are capable of interacting with the solid matrix in the soils. The food dye Brilliant Blue has commonly been used as a tracer although studies have shown that the dye is subject to non-linear adsorption and thus exhibits retardation (Kasteel et al., 2002). This makes the dye unsuited for tracing the travel times of water, however it can still be used for illustrating different flow paths in soils. Wang et al. (2002) compare the performance of Brilliant Blue as a dye tracer with that of a pH-tracer and found that it did not follow the actual wetting front but experienced retardation. They concluded that BB is not a conservative tracer, and suggest that new tracers be used for detecting water movement. Nevertheless, the advantages of Brilliant Blue make it one of the preferred dye tracers for studies of unsaturated zone flow.

Many laboratory and field experiments have shown that preferential flow dominates the flow field for soil types of all kinds, ranging from unstructured sandy sediments to layered clayey soils (Flury et al., 1994; Wildenschild et al., 1994; Schmalz et al., 2002). One of the most comprehensive field studies of preferential flow was performed by Flury et al. (1994), who conducted dye tracer experiments on 14 different types of soils, representing both coarse and fine-grained sediments as well as structured and non-structured soils. They infiltrated 40 mm of water containing a dye tracer during 8 hours and 1 x 1 m profiles were excavated and analysed for dye-staining after 1 day. They found that structured soils were more prone to bypass flow and pulse splitting, and that the water generally penetrated deeper into the subsurface in these types of soils, thus making them more hazardous
regarding pesticide and pollution leaking to the groundwater than otherwise considered (Flury et al., 1994). Weiler and Flühler (2004) compared infiltration patterns from three different field sites at two irrigation rates using Brilliant Blue as a dye tracer and concluded that there were 5 prevailing flow types of which 3 were dominated by macropore flow and 2 by either heterogeneous or homogeneous matrix flow. Javaux et al. (2006) performed a Brilliant Blue infiltration experiment on a macroscopically heterogeneous soil column and showed that the transport of BB was not uniform and included both varying travel depths and concentrations within the columns. The presence of a clay layer within the sand column dispersed the infiltration even more and caused the concentration of BB to be double-peaked. Kung (1990) performed a dye tracer experiment on a larger scale at a natural field site consisting of sandy outwash sediments. The site was excavated very thoroughly after infiltration to a depth of almost 7 m. The excavation revealed bypassing of small-scale lenses containing coarser material, and funnelling due to the presence of layers with varying grain sizes. Kung (1990) also examined the three-dimensional structures of the dye-staining and found that the water would both exhibit sheet flow on top of coarse layers and infiltrate in cylindrical columns or fingers. Lateral flow was initiated in the top of the soil profile as water flowed on top of a coarse layer instead of infiltrating vertically.

The use of ground penetrating radar for monitoring water flow has been investigated by e.g. Trinks et al. (2001) and Truss (2007). Trinks et al. (2001) conducted a laboratory experiment in a large scale sand box, in which a point injection was made into dry sand. Very densely sampled GPR measurements were obtained using an automated system and changes in GPR data due to the increased moisture content were investigated. The high spatial resolution of the data set allowed for three-dimensional assessment of the development of the infiltration plume, which was seen to experience some degree of preferential flow. The contrast in moisture content at the tip of the infiltration finger caused a diffraction hyperbola (Trinks et al., 2001). Truss et al. (2007) used time-lapse ground penetrating radar to monitor the natural infiltration at a site consisting of limestone with scattered sand-filled dissolution sinks at the surface. Comparison of GPR radargrams before and during infiltration showed that rainwater drained preferably through the
dissolution sinks and that infiltration at depth followed the internal layering in the limestone.
3. PRESENTATION OF PAPERS

3.1 Paper I - Appendix A

Title: Visualising unsaturated flow phenomena using reflection ground penetrating radar
Field site: Hjelm Hede

Field work and data analysis
At the area Hjelm Heath in northern Denmark we conducted an infiltration experiment, in which we infiltrated 100 mm of water across an area of 3x3 meters. The subsurface at the site consists of sandy alluvial sediments and infiltration was performed below the topsoil and directly in to the underlying formation. The commonly used dye Brilliant Blue was added to the infiltrated water for visualisation of flow patterns. Prior to infiltration 100 reflection GPR lines (250 MHz) were acquired in a quasi-three-dimensional manner with line and trace spacing of 0.05 m. The GPR survey area encompassed the entire infiltration area and covered an area of 4.95 x 4.7 m. The GPR survey was repeated 24 hrs after infiltration so that the two data sets were comparable line to line. Next, the area was excavated to a depth of 2 m to reveal the infiltration patterns highlighted by the dye tracer. The densely sampled GPR survey allowed for a detailed investigation of the changes in electromagnetic signal caused by the infiltration. It was expected that the BB dye-staining would be more or less visible in the GPR data, both as attenuated areas and because the increase in moisture content in the subsurface would cause scattering and diffraction of the GPR signal. This turned out not to be the case, because the increase in moisture content within the dye-stained area was not as significant as expected and therefore did not affect the GPR signal much.

Nevertheless, the GPR data sets were analysed in terms of delay and amplitude changes for seven prominent reflections. Delay of reflections was expected due to the increase in moisture content in the subsurface, which causes the electromagnetic velocity to decrease (the dielectric permittivity increases). Amplitude changes are a result of the increased conductivity of the subsurface due to both the presence of extra water as well as the Brilliant Blue dyed water’s high conductivity.
Results

The main delay of reflections takes place within the infiltration area but also show that lateral flow is prevalent at all depths. Amplitude changes calculated for intervals between prominent reflections show that the most pronounced changes take place within and just outside the infiltration area, which further elucidates the prevalence of lateral flow. Calculation of the increase in moisture content corresponding to the delay of reflections can account for 75% of the infiltrated water if lateral flow up to 0.5 m is considered. The remaining water is believed to have percolated deeper into the subsurface, or it may have moved laterally further out of the area.

The experiment carried out at Hjelm Hede is in some ways similar to the work carried out by Truss et al. (2007), who used time lapse GPR to monitor natural infiltration, however they did not assess the infiltration patterns on smaller scale using tracers. The preferential flow observed in the excavation at Hjelm Hede are mostly controlled by macropores and soil heterogeneities, although roots as macropores here did not favour flow and the latter was not visible to the naked eye. Based on the GPR reflection amplitude analysis and the delay assessment, we could conclude that the final cause of preferential flow as summarised by Simunek et al. (2003), namely funnelling due to sloping layers, was also prevalent in the area. The dye tracer patterns found at Hjelm Heath are similar to the findings of Kung (1990), whose field experiment was carried out in an environment much similar.

We hoped to extract more information from the dye-staining patterns found at Hjelm Heath, but as the results demonstrate, we had to go about a different approach when analysing the reflection GPR data. Had there not been many distinct and easily recognised reflections in the subsurface, assessment of the moisture content development would have been severely obstructed. If the experiment at Hjelm Hede, or somewhere similar, was to be recreated, it would definitely be worthwhile to assess the temporal moisture content development by increasing the number of GPR datasets. As the acquisition time is rather long for the type of GPR survey we carried out (100 lines ∼ 3.5 hrs), it would also ease the process significantly if an automated system for the GPR equipment could be
implemented. Such an installation would also ensure that the detailed geometry of the survey was maintained so that lines will be comparable with less uncertainty.

### 3.2 Paper II - Appendix B

**Title:** Unsaturated flow inferred from cross-borehole ground penetrating radar monitoring of point injection of water

**Field site:** Arrenæs

**Field work and data analysis**

At the field site Arrenæs we performed a point injection of water which was monitored using cross-borehole ground penetrating radar (GPR) and electrical resistivity tomography (ERT). Background data were obtained using both methods prior to the infiltration, which continued for 5 days. After infiltration had ceased, movement of the infiltrated water was monitored for another 5 days. There are 8 GPR boreholes and 4 ERT boreholes at the field site and infiltration took place in the middle of the area. Cross-Borehole GPR measurements were obtained using two different setups. One is the Zero Offset Profiling (ZOP), in which we obtain data for every 0.25 m to a depth of 12 m for all combinations of boreholes, in this case there are 20. The Multiple Offset Gather (MOG) consisted of measurements where one antenna was kept at a fixed depth (for every m), while the other was moved downward in steps of 0.25 m. As this is a more time consuming method, MOG data were only obtained between 2 boreholes in a profile intersecting the injection point. Arrival times for both ZOP and MOG data were picked, and for ZOP data directly converted to moisture content profiles using the relationship between electromagnetic velocity and moisture content found by Topp et al. (1980) and simplified by Ferre et al. (1996). The arrival times for the MOG data were inverted using a least-squares inversion scheme to produce a two-dimensional velocity distribution in the subsurface between the two MOG boreholes.

**Results**

The field experiment carried out at Arrenæs is part of a longer series of experiments, which have all been with the purpose assessing the applicability of the cross-borehole GPR and
ERT methods to both monitor infiltration but also map spatial soil moisture variability (Looms et al., 2008). Other studies involving improving inversion techniques have also benefited from the experiments carried out at Arrenæs, as data from the field experiments have been used to verify the theoretical approaches and insights (Cordua et al., 2008 & 2009).

The three-dimensional development of the infiltration was captured well by the ZOP data, and in combination with the 2-dimensional velocity distributions obtained from inversion of MIG data it was concluded that large volumes of the subsurface where bypassed during infiltration. Layering in the subsurface diverted the infiltration and caused lateral flow several meters away from the injection point. Movement of the injected water was not uniform and a clearly delineated infiltration plume did not develop. Moisture content increase simulated by HYDRUS2D are much higher than the increases seen in both the ZOP and MOG data. More than half of the infiltrated water could be accounted for by crude calculations based on the field data, however, given the obvious prevalence of lateral flow as well as the inherent difficulty in determining the three dimensional extend of the infiltration plume this result is highly uncertain.

The results obtained at Arrenæs demonstrate that single analysis of the GPR data from ZOP and MOG measurements can capture the movement of the injected water, especially since the 20 available ZOP lines comprise a, almost three-dimensional dataset from which details of the moisture content development could be extracted. Two-dimensional inversion of the ZOP data for each measurement depth could be a future application which can illuminate the three-dimensional movement of the infiltration plume further. The heterogeneous infiltration that was observed from the point injection was in many ways resolved better in the ZOP data than by MOG data, especially since the main plume did not seem to develop exactly along the MOG line. As such, the MOG survey type and two-dimensional inversion of the results is probably a more suitable application for plane infiltrations which behave in a more two-dimensional manner, that is, apart from the obvious prevalence of preferential and lateral flow that is also observed.

A major inconvenience concerning the data interpretation of cross-borehole GPR data from Arrenæs is that they were not compared to the ERT measurements, which have been
shown to add extra detail to the analysis (Looms et al., 2008). Future work on this dataset will surely involve detailed investigation of the ERT data to fill some of the gaps in the GPR data, namely the moisture content distribution and development around the injection site and just below, where the tomographic data coverage is poor.

Future studies at Arrenæs will be carried out later this year. The intention is to perform a plane infiltration of a considerable amount of water at an area equipped with new boreholes for GPR just south of the existing sites. Infiltration and movement of the water front will be monitored using microgravimetry and the results hopefully validated by GPR surveys. It is always interesting to couple new geophysical methods and hopefully the experiment can help explore the advantages and limitations of microgravimetric investigations in the unsaturated zone.
4. CONCLUSIONS AND PERSPECTIVES

Although we were not able to recognise the infiltration patterns on a small scale in the otherwise densely sampled GPR data set, the analysis method provided interesting insights into the application of geophysical methods for assessment of unsaturated flow. Reflection ground penetrating radar as applied at Hjelm Heath can be used for visualising heterogeneous flow developments, provided there are prominent reflections to analyse. and GPR remains one of the key methods within hydrogeophysics.

Results of a plane infiltration at Arrenæs presented in Looms et al. (2008) showed that half of the water escaped from the survey area through lateral flow on top of an impermeable clay layer. The point injection presented here took place beneath the clay layer in order to make sure that all water was accounted for. We expected to be able to delineate the infiltration plume, as others have analysed similar point injections (Binley et al., 2002a) in different geological settings with much success. Our results show that the geology at Arrenæs, although seemingly simple, does not favour development of such a plume.

In conclusion, the work presented here underlines the versatility of GPR as a key method within hydrogeophysics. Some applications of the method have inherent drawbacks, which must be addressed, for example in terms of obtaining extra data using other geophysical methods. Nonetheless, the results of the two field experiments Results of the field experiments both illustrate that water flow in the unsaturated zone is not easily assessed, it is highly heterogeneous in the geological settings investigated here. Neither field experiment has been carried out before in the here presented manner, meaning that not only are the results interesting for those scientists, engineers and others working with local Danish field sites, but the results are relevant for a much broader group of people working with hydrology and hydrogeophysics in other parts of the world.
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I am very grateful for being given the opportunity and responsibility of presenting preliminary results of the data analysis from the experiment at Hjelm Heath both at the EGU General Assembly 2008 in Vienna and the AGU Fall Meeting in San Francisco. Both conferences were rewarding experiences, which I had not expected to be able to take part in beforehand.

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To the Alpha group (you know who you are): You have made these last couple of months a lot of fun, and I will miss it.

Finally, to my family, and in particular Emil: Thanks for all the love, support and patience.

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REFERENCES


APPENDIX A

PAPER I

VISUALISING UNSATURATED FLOW PHENOMENA USING REFLECTION GROUND PENETRATING RADAR

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ABSTRACT

Complex unsaturated flow phenomena, such as unstable wetting fronts and preferential flow cannot be investigated using small-scale sampling of the investigated soils. Dye tracer experiments can help visualize the dynamics of water flow, but are destructive and therefore irreproducible. In this study, we investigated the applicability of high-resolution Ground Penetrating Radar (GPR) for non-destructive visualization of unsaturated flow patterns in the shallow subsurface arising from forced infiltration experiments.

Synthetic studies using a reflection GPR Finite Difference Time Domain modeling code, revealed that differences in water content caused by preferential flow and fingering could be resolved using reflection GPR. The full waveform GPR modeling indicates that areas with increased water content can be distinguished for water content contrasts down to approximately 5% within the top 2 meters, and that attenuation of the signal within areas of increase moisture content will be prevalent.

We conducted a field experiment in which 100 mm of Brilliant Blue dyed water was infiltrated in relatively homogenous and undisturbed sandy alluvial sediments. GPR data sets were collected before and after infiltration. Dye-staining patterns, revealed by excavating a 2 m deep trench through the infiltration area, were compared with changes in the GPR data. The dye-staining patterns could not, as hoped for, be used to verify the GPR results, and a different approach was therefore taken. Detailed analysis of reflection amplitude changes as well as reflection delay revealed significant differences within the dye-stained area, however, delayed reflections below the extent of the dye-staining (to a depth of approximately 5 m) were also found indicating significant displacement flow of the initial soil water. This was corroborated by gravimetric soil moisture measurements taken after infiltration that revealed little or no difference between dye-stained and dye-free soil samples.

Dye-staining patterns were not as useful as expected and could not be used directly as a means of ground-truthing the GPR data. The results of the synthetic GPR modeling, as well as the observed changes in the real GPR data set, nonetheless, underlines the potential of the reflection GPR as a non-destructive method to visualize unsaturated flow phenomena.
1. INTRODUCTION

The quality and quantity of the groundwater resources depend highly on the flow and transport properties of the unsaturated zone. Traditionally, unsaturated hydraulic parameters are estimated using retention and hydraulic conductivity experiments performed on small soil samples in the laboratory. These small-scale analyses are not able to describe preferential flow paths or instable wetting fronts that are observed at field scale. Prediction and characterisation of flow in the unsaturated zone was found to be difficult and highly complex in several studies using dye tracers for visualisation of flow paths (Flury et al., 1994; Schmalz et al., 2002; Weiler and Flühler, 2004). Large-scale experiments on both homogeneous and heterogeneous soil columns have shown that fingering and preferential flow exist and dominate the flow field for soil types ranging from clayey soils to unstructured sandy sediments (Flury et al., 1994; Wildenschild et al., 1994, Schmalz et al., 2002).

Dye tracing infiltration experiments are powerful techniques for visualizing the dynamics of unsaturated water flow and have been widely used for many purposes (Flury & Wai, 2003). The popularity of Brilliant Blue as a dye tracer can be attributed to its excellent visibility in most soil types, its non-toxicity, and its easy solubility in water (Flury & Wai, 2003). Brilliant Blue has been found to be subject to retardation because of non-linear absorption and in consequence it cannot be considered an ideal, conservative tracer (Kasteel et al., 2002). Thus Brilliant Blue is not suitable for estimation of water flow properties but is ideal for illustrating water flow pathways in soils. A major drawback in using dye tracing for unsaturated flow characterization is that the method is destructive and unrepeatable at the same location as the survey area needs to be excavated. Alternatively, borehole sampling and other coarsely sampled point measurements do not provide the same detailed information about small-scale unsaturated flow patterns. Dense point sampling is effectively as destructive as excavation of a profile and furthermore such sampling may potentially generate artificial flow paths.

Reflection ground penetrating radar (GPR) is a non-destructive geophysical method which is sensitive to changes in moisture content in the unsaturated zone (Topp et al., 1980; Reynolds, 1997) and has, as a result, been used for soil moisture estimation in several studies (Truss et al., 2007; Trinks et al., 2001; Huisman et al., 2003). Trinks et al. (2001)
used time-lapse GPR to monitor the movement of an induced plume of water in dry sand in the laboratory. By collecting densely sampled GPR lines every 30 minutes and comparing the time-lapse data sets to the baseline signal, the authors were able to follow the infiltration plume in three dimensions over time, and distinguish preferential flow. Truss et al. (2007) conducted a field experiment, in which they used reflection GPR to monitor both natural and forced infiltration of water in oolithic limestone containing sand-filled dissolution sinks at the surface. For natural infiltration bypass flow occurred which was highly dependent on sediment type and geological boundaries, while the forced point injection of water in the dissolution sinks allowed for three-dimensional monitoring of the downward water migration (Truss et al., 2007).

The aim of this study was to assess the applicability of densely sampled reflection GPR surveys for quantitative monitoring of water movement in unsaturated sandy alluvial sediments after forced infiltration of water. Addition of Brilliant Blue to the infiltrating water and subsequent excavation of a trench allowed for visual assessment of the migration of the wetting front, and these results were then compared to the high-resolution GPR data. To investigate the sensitivity of the recorded GPR signals to various infiltration patterns forward modeling of synthetic subsurface scenarios was made. The simulation results demonstrate how changes in moisture content affect the reflected GPR signal and provided insights in to how small changes in moisture content in the subsurface it is possible to resolve using reflection GPR.
2. MATERIALS AND METHODS

2.1 Ground Penetrating Radar

Ground penetrating radar is a geophysical method in which high-frequency electromagnetic waves are emitted from a transmitter and recorded by a receiver. Ground penetrating radar wave velocity, $v$, depends on the dielectric permittivity, $\varepsilon$, and magnetic permittivity, $\mu$, of the subsurface, whereas attenuation of the signal is dependent on permittivity, electrical conductivity, $\sigma$, and frequency of the signal (Reynolds, 1997). The magnetic permittivity can be disregarded in non-magnetic materials, including sandy sediments. The dielectric permittivity is related to electromagnetic wave velocity, $v$ in the following manner:

$$\sqrt{\varepsilon} = \frac{c}{v}, \quad (1)$$

where $c$ is the speed of light (~ 0.3 m/ns) (Reynolds, 1997). If the electromagnetic wave velocity or dielectric properties are known, the moisture content of a given subsurface can be calculated using the empirical relationship of Topp, which relates moisture content $\theta$ to electromagnetic properties (Topp et al., 1980):

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} \varepsilon - 5.5 \times 10^{-4} \varepsilon^2 + 4.3 \times 10^{-6} \varepsilon^3. \quad (2)$$

In this study we will use the simpler form of Topp's equation developed by Ferre et al. (1996):

$$\theta = 0.1181 \sqrt{\varepsilon} - 0.1841 \quad (3)$$

which has shown to provide results that differ less than 5% from Topp's original equation for volumetric water contents between 5 % and 40 % (Ferre et al., 1996).

Radar wave velocity contrasts of the subsurface cause parts of the emitted electromagnetic signal to reflect back to the receiver at the surface. The magnitude of the reflection is determined by the reflectivity coefficient, which is the ratio between the difference between and the sum of velocities on either side of the reflection (Reynolds, 1997). Following equations (1) and (2), water in a porous material lowers the electromagnetic wave velocity by increasing the dielectric permittivity. Large moisture content contrasts produce high reflection coefficients, because of large GPR wave velocity differences, and these show up in the resulting radargram as reflections. Reflection GPR is useful in highly-resistive areas, where the penetration depth can extend to 20 m or more in
extreme low-conductive sediments. Presence of low-resistivity layers, such as clay, will absorb the GPR signal energy and thus decrease the penetration depth. The vertical resolution depends on frequency and can be approximated by one quarter of the pulse length (Reynolds, 1997). The horizontal resolution decreases with depth as the radius of the first Fresnel zone increases and as the signal moves downward in a conical beam of increasing width (Reynolds, 1997).

2.2 Field Site

Hjelm Heath, a heath area situated in the northern part of Jutland in Denmark (see Figure 1), was chosen as a field site for this study. The area was formed as a glacial outwash plain during the last phases of the Weichselian glaciation when large amounts of melt water was released from the melting glaciers and covered the underlying till with tens of meters of meltwater sediments, predominantly sand (Houmark-Nielsen et al., 2005). The vegetation cover consists of heather, wavy hairgrass and black berries, the root zone extends to a depth of 60 cm (Ladekarl et al., 2005), and the soil development is typical of a podzol (Dalsgaard, 1998). The top 6 m of soil below the podzol consists mainly of medium to coarse sand (between 68% and 99%). The texture varies with depth; some layers being rich in silt and clay and some layers rich in gravel and coarse sand (Ladekarl et al., 2005). With sand of different grain sizes being the dominant texture of the subsurface at Hjelm Hede, the area is suitable for reflection GPR surveys and penetration depth using 250 MHz antennas is expected to be ~5 - 10 m. The area around Hjelm Heath has not been subject to significant deformation after deposition, and the sandy sediments are assumed to be fairly undisturbed. The groundwater table is found 21 m below the surface (Ladekarl et al., 2005).

A well was installed in 1992 equipped with Time Domain Reflectometry (TDR) probes down to a depth of 7 m. The well is equipped with 11 TDR probes, some of which are placed at the same depth, but on opposite sides of the well. Distance from TDR well to survey area is approximately 20 m. Continuous measurements of moisture content over time have shown that duplicate probes at the same depth do not exhibit similar moisture content development. Wetting fronts were observed to bypass certain probes, and observed responses to infiltrations were in some circumstances recorded at deeper probes.
before it was observed closer to the surface (Ladekarl et al., 2005). The observations clearly suggested that for all soil moisture conditions, water movement arising from natural precipitation was highly non-uniform and heterogeneous.

2.3 Infiltration experiment

The topsoil (upper 10 cm) as well as vegetation including roots were removed prior to data collection. An additional 30 cm of soil was removed, so that the infiltration took place 46 cms below the original surface and directly in to the alluvial sand beneath the podzol horizons. A Brilliant Blue solution was made by dissolving 5 kg of Brilliant Blue powder in 1000 L water, resulting in a concentration of 5 g BB/L. The conductivity of the solution was 0.287 S/m, which is considerably higher than fresh water (∼0.001 S/m) and what can be expected of the natural moisture in the subsurface at Hjelm Heath. The infiltration experiment was inspired by the works of Flury et al. (1994) and Javaux et al. (2006) who both performed infiltration experiments using Brilliant Blue as tracers. Flury et al. (1994) infiltrated 40 mm of Brilliant Blue dyed water (concentration: 4 g/L) in 14 different soil types and found that after 8 hrs the penetration depth was in many cases considerably less than 1 m, but that for some soils extreme bypassing was seen. Javaux et al. (2006) infiltrated 180 mm of Brilliant Blue dyed water (concentration: 1 g/L) into a 1 m sand column and within 8 hours dyed water had not reached the maximal depth. We chose to infiltrate a total amount of 100 mm BB-solution during 2 hours across an area of 3x3 m. Watering cans equipped with simple sprinklers for more uniform spreading of the water were used for water application. After infiltration the area was left uncovered, and a small rain shower delivered a couple of mm precipitation during the night, which is a negligible amount compared to the infiltrated 100 mm. Evapotranspiration is of no significance as the infiltration experiment was conducted in early April at which time evapotranspiration is low. One day after infiltration, a trench was excavated through the infiltration area to a depth of 2 meters for visual assessment of the dye tracing patterns outlined by Brilliant Blue (see Figure 4 for location and extend of trench excavation).
2.4 GPR data acquisition

Ground Penetrating Radar lines were collected using a Sensors and Software PulseEKKO Pro System (Sensors & Software Inc., Mississauga, ON, Canada) equipped with shielded 250 MHz antennas. For precise data collection we constructed a setup consisting of two hollow PVC-slides, along which the GPR instrument was pulled. Prior to carrying out the actual field work, it was investigated whether the slides affected the GPR data acquisition process and it was verified they do not significantly influence the data collected by the shielded antenna. The slides were of dimensions 10x10x600 cm and were fixed to the ground manually using bamboo sticks, which were thought not to adversely obscure the data collection either. Careful re-positioning of the GPR instrument for each line along the PVC-slides assured that the repeated data sets were comparable line to line. Both before
and after infiltration, a total of 100 GPR lines were collected with a line spacing of 0.05 m. For all lines the trace spacing was also 0.05 m thus providing an uniformly sampled 3D dataset of the subsurface. Each line consisted of 94 traces, making the final data acquisition area 4.95 x 4.7 m (23.3 m²). Because Brilliant Blue was infiltrated over a 3 x 3 m area within the GPR-area (see Figure 1), the GPR data set collected after infiltration contains both lines affected and not affected by direct infiltration, and each GPR line crossing the infiltration area has traces at the beginning and end that are not affected by BB infiltration. For details of the field site layout, see Figure 1. The collected GPR data was processed using the Ekko View Deluxe and Matlab software packages. For all lines a devow correction was applied to remove low frequency noise.

By repeating GPR surveys with exactly the same geometry, it is possible to distinguish in detail the moisture content changes occurring within the geologic media. Differences between the two GPR datasets collected must arise from changes in moisture content and soil water conductivity, since the geology of the subsurface can be assumed to remain unchanged within this short time frame. The GPR signal will be attenuated by the extra amount of water, and the higher conductivity of the infiltrating BB solution will further attenuate the signal (Reynolds, 1997). The decrease in electromagnetic wave velocity will produce time shifts between the two data sets because reflected waves arrive later at the surface. Also, there is a chance that some reflections will become more pronounced due to ponding at geological boundaries, which will cause the reflectivity coefficient to increase.

For detailed analysis of single reflections in the GPR data, matlab was used to pick reflections in all lines for both data sets, thus obtaining the arrival time and amplitude for each reflection in the entire dataset. The arrival time differences can hereafter be translated into electromagnetic wave velocity changes and converted into moisture content changes using equation (3). Analyses of multiple reflections provide a basis for estimating the spatial distribution of the water in the subsurface and can potentially be validated against the Brilliant Blue dye-staining patterns.
3. SYNTHETIC STUDIES

In order to evaluate the possibilities and limitations of using reflection GPR to monitor unsaturated flow phenomena a synthetic analysis was performed using the 2 dimensional finite-difference time-domain (FDTD) code developed by Irving et al. (2005). This code is readily available from the authors, and is matlab-based making it easy to modify to fit user-specific needs. Synthetic subsurface models are defined by specifying distributions of electrical conductivity, as well as dielectric and magnetic permittivity. Any type of synthetic model can theoretically be created and modeled using the code, however, the computation time increases with the complexity of the chosen subsurface characteristics and model discretization.

A synthetic geological model was created for Hjelm Heath based on information from nearby borehole lithology as well as TDR measurements on site. The nearest borehole containing information about the geology is located several hundreds of meters from the field site, and it indicates that to a depth of approximately 10 m, the subsurface consists of sandy and gravelly sediments, below which there are indications of layers of finer material (GEUS, 2009). The initial electrical property model for dielectric permittivity, $\varepsilon$, used for the FDTD simulation is shown in Figure 2 (top left). In the top of the model, an air-earth interface was included by adding a layer with dielectric permittivity $\varepsilon = 1$ and electrical conductivity $\sigma = 0$ S/m. The four layers in the subsurface have different values of $\varepsilon$ ranging from 4 to 6 corresponding to moisture contents values between ca. 0.05 and 0.12, respectively. The dielectric permittivity values were calculated, based loosely on soil moisture measurements from the nearby TDR well using Equation (3). Since, for simplicity, we are only interested in the delay of reflections, the conductivity is set to $\sigma = 0.001$ S/m and the magnetic permittivity is represented by its free space value, $\mu = 1$ throughout the model. During a real forced infiltration experiment, changes in electrical conductivity are expected to occur, caused both by the added amount of water and the potentially different electrical conductivity of the infiltrated water (compared to the incipient soil water), Thus, the amplitude variations and attenuation of the modeled signal in the synthetic tests will not be affected as much as during a true infiltration experiment.
Several additions to the dielectric permittivity model in Figure 2A were tested: First, a continuous layer of increased moisture content with a thickness of 1.3 m was added in the top of the model (Type A) and it was investigated whether the delay of reflections could be translated into the correct moisture content increase using equation (3). Next, the continuous layer of increased permittivity was combined with an approximately 1 m deep infiltration “finger” of varying width and shape (Type B and C). The purpose of these model
types was to evaluate how preferential flow pathways affect the collected GPR data, and whether they could be resolved to provide meaningful results. Finally, we added a horizontal wedge of increased moisture content to the model (Type D) to assess for how small changes in moisture content the length and shape of the wedge could still be distinguished. Inspiration for this last forward modeling test came from Widess (1973), who investigated the effect of a thinning bed on seismic reflections and concluded that it was impossible to distinguish reflections thinner than 1/8 of the pulse wavelength. Horizontal wedges like the in Type D models could occur during an infiltration experiment due to lateral flow. The type A dielectric permittivity model and forward modelling result are seen in Figure 2 (middle), whereas Figure 4 shows modeling results for type B, C and D. Both the dielectric permittivity model and a synthetic radargram are pictured for all model types. In order to make the models more realistic noise was added as 5% of the RMS of the model and the permittivity contrasts were smoothed by interpolation for type B, C and D models before running the forward simulation. All subsequent analyses and results presented here were performed on unmigrated data. For more reasonable computation time, only parts of the dielectric permittivity models were subjected to forward modelling of the GPR signal. The modeled part of each model has been marked in Figure 2 and 3 by a greyish rectangle.

The resulting radargram corresponding to the type A model having $\theta = 0.25$ in the wet area (i.e. a moisture content contrast of 0.20) is shown in Figure 2. Reflections below the wet area are delayed by 13.7 ns. Reflections from top and bottom of the wet area are very distinct (green) and cause multiples at depth. Assuming that we know the depth to the bottom of the wet area, as well as the depth to the geological boundary causing the first reflection, the electromagnetic velocity can be calculated for the wet area both before and after adding the water. Subsequently, the corresponding moisture contents for the given volume in the subsurface can be calculated using equation (3). For the type A model shown in Figure 2, the calculated moisture content increase is slightly underestimated ($\sim 0.01$). In other model tests having lower moisture content contrasts the moisture content increase was also underestimated (not shown here). All calculated moisture content increases for synthetic tests of this kind, were within 20% of the actual moisture content increase.
The type B and C models in Figure 3 contain moisture content contrasts that produce diffraction hyperbolas which obscure the radargram image, making it difficult to distinguish the actual shape and extent of the wet area of interest. For type B models it was possible to locate the bottom of the 1 m wide high-moisture-content area (red arrow), and calculations of moisture content based on travel time to the bottom of the “finger” provided reasonable results. This was not possible for the spike-shaped infiltration area in type C.
models nor for rectangular areas (model B) with a width of 0.5 m. The green arrow in the type C radargram (Figure 3) marks the diffraction structure arising from the sides of the spike-shaped “finger”. A similar response is seen in for the square infiltration “finger” in type B (marked by green arrow in type B radargram), however it is not as pronounced, probably due to the smaller moisture content contrast. Of the three models depicted in Figure 3, the type C model has the largest moisture content increase (Δθ = 0.12), so it is expected that here is distinct attenuation of the signal in the top of the radargram. The infiltration wedge, in type D models, was distinguishable down to changes in moisture content around Δθ = 0.05 (blue arrow in Figure 3). For moisture content changes below Δθ = 0.01 the radar reflections completely drown in noise from the surroundings. The signal is also damped within the wedge, due to the higher permittivity and thus higher attenuation therein (not pictured).

Based on the results of the numerical modeling, we expect to be able to delineate the bulk changes in moisture content arising from a forced infiltration experiment using reflection GPR. In the case of a moisture content increase in large coherent volumes of the subsurface it will be possible to see attenuation in the wet areas. Furthermore, if preferential flow occurs in the form of fingers > 0.5 m they will be discernible, while smaller fingers of increased moisture content may be identified due to scattering of the wave field, and lateral flow around 0.20 m thickness (for a moisture content increase of 0.05) will appear in the radargram.

It should be noted that all the synthetic analyses in this study were made using a 2 dimensional code, the underlying assumption being that the constructed subsurface models continue indefinitely into the subsurface (perpendicular to the presented image). In other words the “fingers” are more like trenches than cylinders. Wave scattering is expected to increase in true 3D environments and the findings presented here are therefore perhaps overly optimistic.
4. RESULTS

4.1 Dye-Staining patterns

After measuring the second GPR data set different kinds of excavations were carried out. First, a 1 x 1 m hole was dug to succeeding depths of 0.08, 0.18, 0.28, 0.36 and 0.71 m and the horizontal dye-staining patterns were evaluated. Figure 5A shows the horizontal infiltration patterns at a depth of 0.28 m as well as a close-up of a root structure in the corner of the excavation. For location of the horizontal excavation, please refer to Figure 1. According to previous studies involving macropores (Beven et al., 1982), roots are in many cases the cause of preferential flow as they create a flow path for the water. The root structure illustrated in Figure 5A and many others found in the horizontal vertical excavation differs from the idea of roots as macropores. The roots create hardened volumes of sediments which do not favor water flow. They are seen in the horizontal excavation as black circular features surrounded by an outer periphery of non-stained sand. The BB dye-staining effectively illustrates that there is no water flow in the near vicinity of the root structure. In stead of acting as a source of macropore flow, these roots are acting as no flow barriers which infiltrating water has to circumvent.

The main excavation created a trench, which was dug to a maximal depth of 2 m, as it was not safe to excavate the profile beyond this depth because of the risk of collapsing walls. Only one side of the excavation was kept as a vertical profile which can be seen in its full length and depth in Figure 4. The vertical profile reveals that although the subsurface at Hjelm Heath is considered to be rather homogeneous, it consists of several different types of sand with varying grain sizes. This concurs with grain size analyses results from the field site reported by Ladekarl et al. (2005). The top 0.20 m of sand are highly disturbed by bioturbation and old root remnants (Figure 5A), but it is possible to distinguish cross-bedding. From approximately 0.2 m to 2 m depth, the sediments are mostly planar dipping slightly (−2°) toward Southeast. At 1.2 m depth there is a horizon that has a dark orange color, probably stemming from precipitated iron oxides. The horizon does not appear particularly hardened, however it seems it has a minor effect on the infiltration in some parts of the profile (Figure 5B) Just below 2 m depth the grain size decreases and the
sediments become cross-bedded. The cross-bedding can be seen in Figure 5C toward the bottom of the profile (near the lowest blue finger).

Figure 4. Excavated profile at XLINE44. Profile is 4 m wide and 2.2 m deep. Extend of infiltration area is marked with blue arrow. Brilliant Blue-stained area is outlined in blue and used in Figure 6.

In Figure 4 we observe that the top 1 m of the profile below the infiltrated area is almost entirely blue due to dye-staining. Note that the trench crosses the infiltration area (blue arrow in Figure 4), meaning that the edges of the profile have not experienced infiltration. However, there exist areas within the top 1 m where bypass flow is obvious as marked in Figure 4 (D). Some of these bypassed areas are slightly darker than the surrounding sand and consist of hardened sediments and precipitates which are possible remnants of roots or bioturbation. Below 1 m depth the dye-staining pattern becomes more irregular and uniform matrix flow appears no longer to be the dominant flow field. In the middle section of the profile there are three large flow fingers marked in Figure 4 (E) which have facilitated the deepest dye infiltration. The fingers have a maximal width of 0.2 m in the excavated plane, however their three-dimensional structure and extend of them is not revealed by the main excavation. Towards the edges of the infiltration area water has infiltrated laterally to the sides (Figure 4 (F)). The dye-staining patterns show that BB-stained water moves along small sedimentary or structural boundaries in the otherwise homogeneous sediments. Part of the irregularity of water flow is initiated by grain size
differences in the sand acting as small capillary barriers. Mostly the small scale lateral flow seems to be affected by capillary barriers and not as much the infiltration patterns at a larger scale. Small-scale flow along horizontal or dipping boundaries between changing lithologies as well as preferential flow can thus be responsible for transport of water out of the infiltration area.

Sedimentary structures within the large fingers (Figure 4 (E)) did not reveal any variation that would make them more prone to water flow. The “floor” of the excavation pit at 2 m depth revealed large blue stains as well, which indicates that there has been a substantial percolation of infiltrated water deeper than 2 m into the subsurface. Infiltration patterns seen in the excavation at Hjelm Heath are similar to the flow type “heterogeneous matrix flow and fingering” as defined by Weiler and Flühler (2004) with most of the fingers being rather large in diameter and there being a high connectivity between the dyed areas. Capillary barriers on cm scale are also observed many places in the excavation in areas where coarser sand lenses are present within the finer surrounding sand (for an example see Figure 5B).

With the purpose of illustrating the 3-dimensional differences in infiltration patterns, it was decided to excavate the 2 m deep profile laterally. 30 cm sediment was therefore removed from the middle meter of the profile (for exact location, see Figure 1). The 1 meter wide and 2 m high profile, can be seen in Figure 5C. Brilliant Blue dye-staining in the top meter of this profile is a lot less coherent compared to Figure 4 with almost 1/4 of the area completely un-dyed. Notice also that the three large infiltration fingers seen in the main profile (Figure 4) are not visible anymore at this position. Rather, there is one large wide infiltration path to the right which seems to extend well below the depth of the excavation.

The large non-stained/bypassed areas in the top 1 m are the horizontal view of the root structures seen in Figure 5A. The black root remnant causing by-pass flow here extends to a depth of 0.35 m, (0.81 m below surface), while the surrounding un-stained sand extends to a depth of 0.65 m (1.11 m below surface). According to Ladekarl et al., (2005) the root zone extends only to a depth of 0.6 m, but the dye-staining illustrates how roots affect water flow at even deeper in to the subsurface. It can be speculated whether the bypass of such large volumes or hardened roots and sand can be responsible for the deep penetration in certain areas of the profiles, by forcing water to penetrate deeper in to the subsurface.
4.2 Ground Penetrating Radar data

The 100 background pre-infiltration GPR lines overall show the overall same distribution of reflections in the subsurface. This is exemplified by XLINE14, XLINE44 and XLINE84 in Figure 6. The three lines were collected 0.7, 2.2 and 4.2 m into the area (See Figure 1). All three lines were migrated using the EKKO View Deluxe software (Sensors & Software, Mississauga, ON, Canada) and a true-amplitude scaling function was applied. A migration velocity of 0.125 m/ns was chosen based on the average soil moisture measurements
from the nearby TDR well. From the migrated radargrams the subsurface can be divided into 3 sections. The top section consists of parallel reflections dipping slightly towards the SE which corresponds to the planar bedding found in the top 2 m of the excavated profile. Below the parallel reflections we find a more heterogeneous collection of less continuous reflections with several distinct dipping reflections across the entire profile, c.f. section 2. In XLINE44 this structure could resemble several generations of a buried river channel. Due to the short length of each line it is difficult, based on migrated data, to determine what this structure really is, and also the presumed channel is at a depth where the GPR signal has been dispersed and attenuated. The important point is that the pronounced reflections can be recognized throughout the dataset. Below 6 m, in section 3, the reflections become less pronounced and the radargrams are mostly dominated by multiples and noise.

![Figure 6](image)

GPR radargrams from XLINE14, XLINE44 and XLINE84.
All sections have been processed using the EKKO View Deluxe software. Migration was done using a velocity of 0.125 m/ns and the applied SEC gain function was with the following parameters: Start value: 15, Attenuation: 0.2, Max value: 1000. Note: Depths are approximate, and only based on the migration velocity.

The depths associated with the GPR data in Figure 6 are not exact as they are based on the migration velocity. They are, however, a good approximation, since the reflection in the
bottom of Section 1 (indicated by the top punctured line) is most likely to be the interface between the two textures of sand seen in the excavation at approximately 2 m depth (Figure 4 and 5C).

Figure 7 shows XLINE44 and XLINE84 before and after infiltration as well as a plot of the difference between the two. The same true amplitude scaling function has been applied to both lines, which have otherwise not been processed. The 3 sections introduced in Figure 6 still apply to these unmigrated data. XLINE44 is aligned with the excavated profile and the outline of the BB dye-stained area as seen in the excavation (Figure 4) has therefore been superimposed on Figure 7B. In the top of the radargram the most obvious change is seen around 50 ns as attenuated areas, and at depth we find attenuation in the middle of the radargram from 125 to 200 ns.

A closer look at the single traces of each radargram reveals that amplitudes for traces affected by infiltration are diminished after infiltration while traces from outside the infiltration area do not show the same tendencies. On several traces, specially the ones inside the infiltration area, the two data sets differ somewhat in the very top, i.e. within the first couple of waveforms. This could be attributed to minor changes in the characteristics of the surface between the acquisitions of the two data sets. While measuring the first data set and during infiltration, people walked on the experimental area, which will likely have affected the upper-most sediments, which were completely undisturbed prior to the experiments. Also, before measuring both data sets and during the infiltration the surface was raked, which has disturbed the upper few centimeters of sand. These effects, as well as the general uncertainty in repositioning the GPR-equipment along the slides can probably explain the difference in signal at shallow depths. The GPR signal is most focused in the beginning of the trace and therefore small variations in e.g. the surface characteristics will have great influence on the onset of the signal. Further into the subsurface the energy wave becomes less focused and these small-scale effects disappear.
In Figure 7B it is worth noticing that the outline of the BB dye-stained area is not reproduced in the GPR data. Although dye-staining was clearly visible in the excavations (Figure 4 & 5), the infiltrated water has not had significant influence on the GPR data, neither as heavy attenuation in the dye-stained area nor as scattering effects which would be expected based on the GPR forward modelling (Figure 6 & 7).
simplest reason for this discrepancy between expected and actual GPR signal is that the infiltrated water is not sitting in the dye-stained areas. In this case, the difference in permittivity and conductivity parameters between the two GPR surveys is not significant enough to create the expected differences.

4.3 Moisture content measurements

Before conducting the vertical excavation 10 undisturbed soil samples were collected from the excavated profile in the middle of the infiltration area (for exact location see Figure 1) for gravimetric measurement of moisture content. The samples were collected every 0.20 m, the first at a depth of 0.20 m from the top of the infiltration area (i.e. 0.66 m below the surface). The results of the moisture content measurements are shown in Figure 8 as the blue line. The moisture content increases from $\theta = 0.05$ at the top location to $\theta = 0.12$ at a depth of 2.5 m. The samples that were collected from the excavated profile ranged from being completely dyed to un-dyed. However, upon measuring the moisture contents of said samples, no connection between measured moisture content and degree of dye-staining could be made.

Moisture content profiles obtained from TDR measurements in the nearby well are also depicted in Figure 10. We have included minimum/maximum profiles for 2008 as well as for the specific week in 2008 where the field work was carried out. The TDR data shows an increase in moisture content from 0.06 to 0.11 within the first couple of meters, and then decreases to 0.05 between 2 m and 2.5 m, after which it then increases again with depth. The gravimetric measurements after infiltration are slightly higher than the moisture content measured by the TDR probes. This difference may be caused by the forced infiltration experiment, but could also arise from changes in the subsurface between the two measuring points as there is approximately 20 m from the TDR well to the experimental area.
Figure 8
Moisture Content at Hjelm Hede estimated from TDR, soil samples taken after infiltration, calculation of moisture content increase based on GPR reflection delay. Deepest extend of observed dye-stained area is at 2.5 m depth.
See Figure 4 for location of soil samples.

The largest difference in moisture content is found at 1 m depth, where the samples show a moisture content of 0.09, while the TDR data show values less than 0.07. Assuming this change represents the true increase in moisture content in the infiltration area, the decrease in electromagnetic velocity is 0.01 m/ns, which corresponds to an increase in dielectric permittivity of 0.8. This permittivity contrast is smaller than the examples from GPR modeling that are shown in Table 1, hence such a minor increase in moisture content do not give rise to significant reflectivity contrasts and will not show up in the GPR data. The realization that the Brillant Blue dye-staining does not highlight the presence of the infiltrated water leads to a different approach, in which the 7 reflections marked in Figure 7 are investigated in detail for more information and insights into the nature of the infiltration at Hjelm Heath.
4.4 Reflection delay

The seven prominent reflections have been marked by red in Figure 7A and 7C and superimposed onto Figure 7B and 7D, respectively, to illustrate if and how reflections have changed position from before infiltration to after. The uppermost four reflections are almost horizontal, while the three deeper reflections have a distinct dip. Reflection 5 dips toward SE and reflections 6 and 7 both dip towards NW. The seven reflections are more or less identifiable in all recorded radargrams, but there are some differences between them. The dip of reflections 5 and 6 varies across the area and reflection 6 is more pronounced and easier to recognize in some GPR lines. Reflection 6 and 7 are furthermore found at the depth of the channel-like structure seen in the migrated data (Figure 6) and thus both reflections are highly affected by diffraction.

Comparison of Figure 7B and 7D shows that the reflections are delayed in XLINE44 after infiltration, whereas the reflections are more or less the same for XLINE84. This result corresponds well with XLINE44 being situated within the infiltration area, whereas XLINE84 is situated outside and as such the latter should not be affected by the infiltration. Since the radargrams have not been migrated these delays are “real” because no velocity bias has been applied, and thus they represent a decrease in GPR wave velocity, which must be caused by an increase in moisture content arising from the infiltration experiment. The changes are most prominent for reflection 4, 5 and 6, and furthermore the delays increase with depth. This can partially be explained by the delay accumulating throughout the depth of radargram but could also indicate that the bulk amount of water already has percolated through the top section and is now found at deeper depths. In some isolated areas the reflectivity is seen to increase after infiltration, which can be attributed to ponding of water on one side of a geological boundary, thus causing the reflectivity coefficient to increase due to the higher contrast in GPR velocity between the two media.

The arrival time and amplitude of the seven reflections shown in Figure 7 were selected for the entire dataset. The difference in arrival time for each reflection was subsequently calculated, and the result is illustrated in Figure 9. For reflection 1 the figure illustrates the delay for the entire volume between the surface and the reflection, whereas the other delays represent the delay between each reflection and the previous. Extreme values of arrival time differences seen in Figure 9 as “spots” are probably due to errors in the
manual picking of reflections in the data set. Note that for reflection 6 and 7, timing of the reflections in the western part of the survey area was difficult to pick due to diffraction effects and thus interpretation the results for these reflections should be careful. In spite of these inherent picking difficulties and apprehensions toward the reflection delay analysis, Figure 9 also shows continuous areas experiencing the same delay, which can not be dismissed as errors.

Figure 9.
Delay of reflections. For reflections 2-7 the delay is calculated by subtracting the delay of the previous reflection. Black box indicates infiltration area.

Approximate depths to reflections:
Reflection 1-4: 1.5 - 2 m
Reflection 5: 2 - 2.5 m
Reflection 6: 3 - 4 m
Reflection 7: 4 - 5 m

There is a significant delay in arrival time for reflection 1, 3, 5 and 6, and the delay is more or less confined to the infiltration area. Reflections 2 and 4 show much less delay, which is furthermore not confined to the infiltration area but scattered throughout the whole survey area. In general, however, the area experiencing the most overall delay corresponds to the
area directly below the infiltration area (indicated by a black box). The delay of reflection 1 and 3 is mostly confined to the southern part of the area, whereas the delay of reflection 5 is seen in a S-W trending belt, and the delay of reflection 6 and 7 is mostly confined to the northern part of the area. This delay pattern and subsequent moisture content change can be correlated with the general geology observed in the radargrams as dipping reflections, as well as in the excavation, where measurements confirmed that the sediments in the top 2 meters dipped towards SE. Hence, the water movement at Hjelm Heath even on this rather small scale seems to be controlled by the geology and dip of sedimentary structures in the subsurface.

Since reflection 2 and 4 have not been particularly delayed the infiltrated water must have percolated beyond these reflections. Also, reflection 7 is not delayed compared to reflection 6, indicating that the moisture content has not increased beyond reflection 6. The results presented in Figure 9 suggest that large amounts of water have moved deeper into the subsurface than the dye-staining patterns in Figure 4 and 5 suggest. The large stained area noticed in the bottom of the excavation can have acted as a flow path deeper into the subsurface, and thus contributed to the transport of water below 2 m depth, but as Figure 9 shows the moisture content has increased across larger areas in the subsurface. Particularly for reflection 1, but also for reflection 3, 5 and 6 it is observed that the delay extends 0.5-1 m outside the infiltration area, corresponding to lateral flow, which was also observed in the dye-staining in Figure 4 and 5.

4.5 Amplitude Analysis

An analysis of amplitude changes in the entire GPR data set before and after infiltration can provide an insight into the overall impact of the infiltration experiment on the GPR wave energy distribution. Amplitude changes in deep reflections are caused by both moisture content and conductivity changes, both of which affect the dielectric constant. An increase in moisture content near the top, however, will also deprive the signal of some of its strength thus causing the signal to have less strength at depth.

Amplitude analysis was carried out between reflection 1 and 4, and between reflection 4 and 6 (see Figure 6 and 7). The first interval incorporates all the horizontal reflections in the top of the radargram (section 1), whereas the second interval contains the part of the
data set which is not yet influenced severely by diffraction and bow-tie effects associated with the channel-like structures in the unmigrated data (section 2). RMS values were calculated from the amplitude values of each GPR line in the given time interval, and the difference between the before and after data set was calculated by simple subtraction. We used the arrival time of the specific reflections obtained previously to determine the intervals of interest, thus automatically taking into account the time shifts occurring in the radargram. Granted that the same reflections 1 and 4 have successfully been identified in all lines both before and after infiltration, this method renders results that merely show the change in GPR signal amplitude within the part of the subsurface between the two reflections. Needless to say there are some uncertainties in determining the exact location of reflection 6 and 7 in unmigrated data in the area that is influenced by the channel (as discussed previously). As for reflection timing, changes in amplitude RMS between the two datasets can be interpreted to represent changes in moisture content and/or conductivity, because the geology of the sediments in the subsurface otherwise remains unaltered between the two data sets.

Figure 10 shows the difference in amplitude RMS obtained from the amplitude analysis as well as the extent of the infiltration area. For both analysed sections of the subsurface it is evident that the GPR signal is subject to change, both within and up to almost 1 m outside

A: Reflection 1-4

B: Reflection 4-6

Figure 10.
A; Difference in GPR signal RMS between reflection 1 and 4.
B: Difference in GPR signal RMS between reflection 4 and 6.
Red and blue colours represent increase and decrease in RMS value, respectively.
Black box indicates infiltration area.
the infiltration area, probably induced by the lateral movement of water along sedimentary boundaries as was also observed in Figure 9. This effect is most obvious along the Southeastern edge of the infiltration area, although changes in amplitude also occur in other areas outside the infiltration area. Again, the direction of the presumed lateral flow corresponds to strike-dip measurements on-site after excavation of the profile at XLINE44, and the amplitude analysis overall show the same tendencies as the reflection delay.

Since the amplitude analysis was performed by subtracting the second data set from the first, we would expect areas experiencing attenuation of the signal to have negative (i.e. blue in Figure 9) RMS changes. This is seen to be the case for most of the infiltration area, whereas an increase in RMS is more dominating along the outskirts. This is perhaps a result of smaller scale lateral flow increasing the reflectivity in the GPR data, because of the very unpredictable moisture content increase associated with lateral flow.

An explanation for the areas outside the infiltration area being subject to changing GPR signal strength is the three-dimensional nature of the propagating wave. In any GPR line, the signal reflected back to the receiver represents a volume of the subsurface that becomes progressively larger with depth. Because of scattering of the signal in all 3 dimensions, the change in RMS value on the edges of the infiltration area could in fact be caused by increased moisture content within the infiltration area. This effect will increase with depth and it can be argued that both part of the reflection delay as well as amplitude RMS differences at depth along the edges of the infiltration area arise from this.
5. DISCUSSION

The estimated moisture content increase based on reflection delay in Figure 8 can be used to make an approximate water balance for the infiltration experiment, assuming that the average delay corresponds to a uniform moisture content increase across the entire area. By multiplying the moisture content increase with both the estimated depth to the reflection and the infiltration area we can calculate the moisture content increase in liters for that particular reflection. Summing up the profile yields an estimate of the total amount of water in the subsurface. This method is a crude approximation including many uncertainties, however, it is an illustrative way of assessing the reflection delays in terms of moisture content. By assuming that the water is situated in a $3 \times 3$ m column of the subsurface the total amount of water adds up to 510 L, which is only little more than half of the infiltrated amount (900L). If the size of the infiltration area is increased to $3.5 \times 3.5$ m and we thereby take into account the lateral flow observed in Figures 3, 8 and 9, this yields a total of 694 L extra water in the subsurface. If this number represents the true amount of water that is stored within the top 4 m, the reflection delay has captured 75% of the infiltrated water. The remaining water has then either percolated deeper in to the subsurface or has moved laterally away from the survey area, that is, further away than 0.5 m. Granted, the calculations mentioned here are influenced by substantial uncertainty, however they provide an indication of where the water is actually situated and that it is not necessarily in accordance with what the dye patterns illustrate in the excavation.

The data analyses show that there is some lateral movement of water out of the infiltration area, however this is not assumed to be responsible for 25% of the water being unaccounted for. The remaining question is then: Where is the rest of the water? Considering that some of the water has probably moved out of the area laterally, this can account for some of the discrepancy, however it is very likely that the water has also moved deeper into the subsurface along preferential flow paths. Whether the BB dye has moved as far into the subsurface as the water is so far unknown, because we only have visual assessment of the water and dye movement from the excavation. Kept in mind that other studies show that Brilliant Blue as a tracer experiences retardation (Ketelsen et al.,...
1999; Kasteel et al., 2002) it is likely that there is a moisture content increase further into the subsurface than what is outlined by the tracer. Assuming that the average moisture content between 0 and 2 m was 0.075 before infiltration (value based on TDR measurements in Figure 8) and a formation porosity of 0.4, then the amount of water in this volume would be 1350 L. Addition of 900L of water to this through infiltration will increase the moisture content to 0.125, which is far from what was measured in any of the samples from the excavation. Only reflection 4 experiences delay corresponding to a sufficient increase in moisture content, while the delay of the other reflections does not come close (Figure 8). As the expected amount of water is not present in the top 2 meters, displacement flow has most likely caused some of the initial water to be pushed further into the subsurface thus elevating both the moisture content in the near surface sediments as well as at depths beyond the reach of the infiltrated water containing tracer. What is measured in the GPR data set is then the result of a combination of flow types consisting of both preferential flow and displacement flow. This explains why there are moisture content changes at depth and at the same time dye-staining in the excavation which suggests that the infiltrated water has only reached depths of a couple of meters.

Many previous studies have attempted to describe unsaturated flow in terms of preferential flow paths and unstable wetting fronts. These investigation have rarely included moisture content measurements and subsequent assessment of the whereabouts of the infiltrated water. In Flury (1994) and other studies, the soils are classified as either initially “dry” or “wet” but the actual moisture contents of these soils neither before nor after infiltration are not measured (Flury et al., 1994) (Weiler & Flühler, 2003). Hence, results of these studies merely illustrate the movement of the infiltrated dyed water and as such do not shed light on the overall movement of soil water in the surrounding subsurface. Our results show that the infiltrating water is not sitting in the dyed areas and that displacement flow is responsible for distributing the water. Moisture content increases are seen deeper than 4 m in to the subsurface although the dye-staining seen in the excavation is most prominent in the top 2 meters of the soil.

According to the texture analyses performed by Ladekarl et al., (2005) the sand between 3 and 5 m is much coarser than the surrounding geology and contains much gravel compared to the other samples analyzed. Assuming that this geology also represents the
specific experimental site used for the present study, this can mean two things: i) the coarser layer has facilitated a much faster infiltration to deeper layers, which cannot be distinguished in the GPR data set because of noise, multiples and loss of energy. Or, ii) The coarser layer is acting as a large capillary barrier (because the general moisture content is still low) and thus water is exiting the infiltration area on top of the coarser layer, which means that lateral flow IS accountable for the missing water. The latter is in agreement with the assumption that the water is likely to move along the geological boundaries as the reflection analysis also highlighted.

Since it was necessary to walk inside the infiltration area while the GPR data sets were collected and during infiltration it is possible that this could have caused compaction of the upper sediments and thus influence the infiltration. Cracks and small deformation could have caused the water to bypass some areas while flow could be facilitated in others. Walking on the infiltration and GPR area could also have caused slight compaction in the upper-most sediments affecting the flow at shallow depths. Compaction has been shown to induce preferential flow in sandy loams (Mooney & Nipattasuk, 2003), but upon excavating the profile we did not find evidence that such factors govern the water flow. Also, the infiltration experiment performed for this work did not reflect naturally occurring conditions in Denmark as it is highly unlikely that a rain shower would produce 100 mm of precipitation within 2 hours (annual precipitation is 781 mm (DMI, 2009)). The purpose of this experiment, however, was not to simulate natural conditions, but rather to assess whether we could obtain useful information regarding unsaturated flow using this method. Future investigations of the type presented here could include a lower infiltration rate and conducting the infiltration over a longer period without removing the topsoil. An experiment of this kind will mimic natural conditions. Also, it would be interesting to see the effect of the BB dye-staining on the data by conducting a similar experiment without the tracer, thus removing the effect of the increased conductivity of the applied water, as well as the unknown effect the viscosity of the BB solution. Such an infiltration experiment with fresh water only would allow for an assessment of the naturally occurring preferential flow patterns and thus provide further insight into the nature of unsaturated flow at the field site. For even more precise data collection and higher reproducibility between the before and after data, a more automated system for GPR acquisition and positioning is needed. This
would also make way for the possibility of increasing the number of datasets obtained, thus allowing for a more detailed assessment of the moisture content development and flow in the subsurface.
6. CONCLUSION

The results presented here show how the infiltrating dye-stained water causes attenuation of the GPR signal, and how reflections at depth are delayed by the increase in soil moisture. Both of these effects are mostly confined to the infiltration area, with GPR data from outside being less or not affected at all. Conversion of the decrease in electromagnetic wave velocity to an increase in moisture content shows that not all of the infiltrated water can be accounted for. The percentage of water that is not accounted for is believed to have moved deeper than the deepest reflection analysed or followed lateral flow paths away from the primary area of interest. The existence of lateral flow was illustrated both by amplitude analysis and reflection delay which show that there is some degree of lateral flow out of the infiltration area toward SE. This is also observed in the delay of the most prominent reflectors indicating that lateral flow, at least in the top couple of metres, is important. Delay of reflections at depths beyond the extend of the dye-staining patterns in the excavation outlines the increase in moisture content at depth. GPR reflections, which become more pronounced after infiltration suggest that there are some areas where geological boundaries between different types of sediment cause an increase in reflectivity because of ponding on either side of the boundary.

While dye-staining patterns on a small scale seem to be slightly predictable as it is affected by the presence of roots and by minor grain size variations between different geological layers, the water flow on a larger scale seems generally unaffected by this at Hjelm Hede. This illustrates the complexity and randomness of preferential flow and that the scale of measure is important to consider when working with experiments and data sets like this. Brilliant Blue dye-staining is helpful for illustrating and highlighting the existing preferential flow paths, however in this case the flow patterns are more complex than the dye-staining is able to highlight due to lateral flow and displacement flow. Other studies do not attempt to describe the infiltration in terms of quantitative calculations of soil moisture changes, and it is therefore difficult to compare the results obtained at Hjelm Heath to those of others.

The changes in GPR signal turned out not to be as significant as expected based on the numerical modelling as well as other experiments described in the literature. We did not manage to resolve small infiltration patterns created by preferential flow and outlined by
the dye tracer using reflection GPR. However, the sample density and the quasi-3-
dimensional dataset provided an excellent means of assessing both reflection delays
caused by electromagnetic velocity changes as well as amplitude differences and these
results provided more knowledge of the flow regime at the field site. We have used the
GPR data to obtain an overview of the overall movement of the water and have found that
at this field site the unsaturated flow regime is dominated both by preferential flow, lateral
flow and displacement flow. The results highlights the overall applicability of using a non-
destructive method as high-resolution reflection GPR to describe unsaturated flow.

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APPENDIX B

PAPER II

UNSATURATED FLOW INFERRED FROM CROSS-BOREHOLE GROUND PENETRATING RADAR MONITORING OF POINT INJECTION OF WATER

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ABSTRACT

Cross-borehole Ground Penetrating Radar is used to illustrate the development of an infiltration plume arising from a point injection of water. Infiltration proceeded for 5 days and was monitored both during infiltration and for 5 days after it had ceased. The two GPR data acquisition schemes, Zero Offset Profiling (ZOP) and Multiple Offset Gather (MOG) each contribute to the overall analysis of the water movement with different insights, and combined evaluation of results from both methods are used to establish an overview of the water movement in the subsurface. The largest increase in moisture content of 0.06 was found at 2 m depth, where an infiltration plume is seen to develop. Instead of moving downward as an entity the infiltrated water disperses, some moves into the surrounding formations, and out of the survey area while some moves downward and causes minor increases in moisture content at depth. To further illuminate the flow properties and response to the point injection, analysis of electrical resistivity tomography data obtained during the field experiment is needed.
1. INTRODUCTION

The use of cross-borehole geophysical methods such as electrical resistivity tomography (ERT) and ground penetrating radar (GPR) have great advantages compared to standard surface deployed geophysical methods and point measurements such as time domain reflectometry (TDR). The penetration depth of surface GPR is greatly influenced by the near-surface sediments which in many cases have strong attenuating effects on the transmitted signal. Point measurements have inherent drawbacks as they cannot represent the spatial heterogeneity of the subsurface which occur at much larger scale.

The use of cross-borehole geophysical methods to estimate unsaturated hydraulic parameters and flow have been investigated in several studies such as Binley et al. (2002a & 2002b) and Looms et al. (2008b). Binley et al. (2002a) report the results of a tracer injection experiment in the vadose zone at a field site consisting of sandstone. Both ERT and GPR measurements were taken during the monitoring period and the authors were able to clearly follow the migration of the infiltration plume. The hydraulic conductivity at the site was estimated based on moment analysis of the data. Binley et al. (2002b) present the seasonal variation of moisture content at the same field site as inferred from cross-borehole ERT and GPR and demonstrate how the moisture content in the shallow subsurface respond to the net infiltration with hardly any delay.

Looms et al. (2008) performed a plane infiltration experiment at the field site Arrenæs, Denmark and used cross-borehole GPR and ERT to analyse the downward movement and spreading of the water. Migration of the water in the subsurface was susceptible to layering and small differences in grain size resulted in lateral flow. Only half of the infiltrated water could be accounted for both by moment analysis and the calculated water accumulation rate was a factor 2 lower than the applied infiltration rate. Looms et al. (2008) concluded that since no other causes of this discrepancy could be determined it was most likely the case the layers in the top meter diverted the water laterally. The objective of the present study was to perform an infiltration beneath the impermeable layer found in the top meter of the soil profile at Arrenæs. Hopefully such an experiment will shed light on the flow properties at the field site.
We here present the first results from a point injection experiment made beneath the impermeable clay layer at Arrenæs. Cross-borehole GPR data is obtained using two different data collection methods; zero offset profiling (ZOP) and multiple offset gather (MOG). The background moisture content of the subsurface is described based on the cross-borehole data obtained prior to infiltration as well as reflection GPR surveys. Movement of the injected water is monitored by repeated cross-borehole GPR surveys carried out each day during the injection period and thereafter. The migration of the injected water is expected to be dependent on small changes in the geology of the subsurface such as grain size variations within the sand (Looms et al., 2008).
2. MATERIALS AND METHODS

2.1 Field site

Arrenæs is situated in the Northeastern part of Zealand, Denmark on a peninsula surrounded by the largest Danish lake, Arresø. At the field site there is a 20-30 m thick unsaturated zone. Information from wells and boreholes in the close vicinity of the area indicate that the top 20-25 m is almost entirely sandy alluvial sediments, although layers of moraine clay have been observed in the top of wells that are only 50 m away (GEUS, 2009). Due to the glacial origin of the sediments a certain degree of heterogeneity in the subsurface sediments should be expected, such as intercalations of clay (Looms et al., 2008; Cordua et al., 2008). Prior to previous studies (Looms et al., 2008) the field site was equipped with boreholes drilled to a depth of 12 m and instrumentation for cross-borehole ERT and GPR. Two experimental sites were created, both of which consisted of 4 ERT and GPR boreholes placed in a square 7 m and 5 m apart, respectively (see Figure 2). Both borehole types have PVC casing and the ERT boreholes are equipped with electrodes every 50 cm. Site B, which is furthest to the North was in 2008 equipped with additionally 4 GPR boreholes along the edges of the previous GPR square (see Figure 2).

2.2 Cross-borehole GPR

In the Zero Offset configuration (ZOP) transmitter and receiver are each placed in one of the boreholes and lowered simultaneously with steps of 0.25 m thus obtaining 49 measurements of the horizontal electromagnetic wave between the two 12 m deep boreholes. The complete ZOP data set contains profiles from each possible configuration of boreholes in the area which adds up to 20 (See Figure 2). ZOP lines between the corner boreholes and those halfway between have been disregarded as the amount of extra information gained is considered too small. In the multiple offset gather (MOG) one antenna is fixed in one borehole while the other is lowered into the other borehole in steps of 0.25 m at a time. The fixed
antenna is then moved 1 m into the borehole and the procedure repeated. Finally the gather is reversed so that the moving antenna becomes the fixed. In total, this creates 1176 measurements for each combination of boreholes. Because of the large amount of measurements in a MOG, this data collection scheme is only used for one combination of boreholes, in this case borehole 1 and 3. A complete MOG data set can be acquired within 1 hour 45 minutes, while the complete ZOP gather (all 20 lines) takes approximately 2 hours.

The ZOP data provides one-dimensional information about the velocity distribution between all boreholes, i.e. across most of the area, whereas a two-dimensional tomographic image of the velocity distribution between borehole 1 and 3 is achieved by inversion of the MOG data. For all ZOP and MOG lines the first arrival time of each individual trace is picked as well as for calibration traces obtained above the boreholes. The calibration traces are obtained above ground where the electromagnetic velocity is known and can therefore be used to establish an “absolute time zero” for each line, which is essentially the time when the signal was emitted from the transmitter. Using the “absolute time zero” the true travel time for the electromagnetic wave through the subsurface can be calculated using the known distance between boreholes. The signal passes through a smaller volume of the subsurface for short ZOP lines (e.g. ZOPAB) compared to longer lines (e.g. ZOP13), hence short ZOP lines can exhibit larger contrasts in electromagnetic velocity because the calculations average over a smaller volume. For ZOP data the electromagnetic velocity at all depths is converted to moisture content using Topp’s equation (Topp et al., 1980) in the form suggested by Ferre et al. (1996).

A two-dimensional velocity distribution between the utilised boreholes is acquired through inversion of the picked arrival times of the MOG data. Here, we will use the linearised least-squares inversion code which incorporates both correlated and un-correlated data errors presented in Cordua et al. (2008 & 2009). A newer version of the code was made...
available to the authors by Cordua et al. For large angles between transmitter and receiver the electromagnetic wave will tend to move along the antennas and within boreholes as well as along the surface, instead of propagating through the geologic formation, therefore editing of the original dataset is necessary. The number of measured traces during each MOG is narrowed down from 1176 to 744 by removing those traces having an acquisition angle >45° and those closer to the surface than 1 m, leaving the final straight ray grid as seen in Figure 1. Data coverage is best in the middle of the area between 3 and 11 meters and towards both the top and bottom of the profile, less rays pass through each cell. The eventual bending of rays due to linearisation, however, can change the ray coverage in the final inversion result. In picking the arrival times for input to the inversion uncorrelated error of 0.8 ns and 2 ns of correlated data error is assumed (Cordua et al., 2008 & 2009). The horizontal and vertical correlation lengths are 8.5 and 2.4 m, respectively (Cordua et al., 2009).

![Diagram of GPR and ERT borehole positions and ZOP and MOG setup](image)

**Figure 2.**
LEFT: Overview of GPR and ERT borehole positions. Distance between GPR1 and GPR3 is 5 m. Distance between ERT1 and ERT3 is 7 m. North is up.
RIGHT: ZOP and MOG setup.

Development of the moisture content in the subsurface and movement of the infiltrated water is investigated by comparing moisture content during and after the infiltration period to the baseline moisture content at Day 0 obtained before infiltration
commenced for both ZOP and MOG data. Soil moisture values estimated using data from the two survey methods provide comparable values, although it must be expected that the ZOP results in general render lower moisture content contrasts than the inversion results, because the method averages over the entire volume between two boreholes. Inverting the MOG data yields a result in which the moisture content of single cells between the boreholes is estimated and this allows for a greater variation in moisture content in the two-dimensional tomogram.

2.3 Field experiment

To make sure that no water was lost in the uppermost sediments it was decided to drill the infiltration borehole to a depth of 1.5 m thus avoiding lateral flow on top of the upper clay layer as observed in Looms et al. (2008b). Drilling was carried out using a motorised hand drill and the infiltration borehole was kept open by a 70 mm plastic tube. The bottom 0.10 m of the plastic tube was perforated to allow water flow out of the sides as well as in the bottom.

Tap water for infiltration was kept in a 1 m$^3$ rain water tank, which was covered with a plastic tarpaulin to avoid evaporation and algae growth. A peristaltic pump connected to two 12 V batteries ensured a constant infiltration rate during the experiment. However, the batteries ran out of power on two occasions for 1 and 3 hours, respectively. After the second episode an extra battery was connected to the pump and the rest of the infiltration went without further interruptions. Pressure transducers were placed in the water tank as well as in the infiltration hole to record the water outflow and monitor the eventual build-up of water in the infiltration borehole. The infiltration was initiated on October 21 2008 and lasted for a total of 5 days (120 hrs), during which the infiltration rate was kept at an average of 0.145 L/min. The infiltration rate was measured several times during the experiment and found to be reasonable constant. After infiltration was brought to an end data collection continued for another 5 days to monitor the movement of the infiltrated plume of water. Due to equipment malfunctions some GPR data from day 5 and 6 are missing. 2D- and 3D-ERT data was obtained for all 10 days, however they will not be presented here, as they have not yet been analysed. The last day of measurement was October 31,
2008. During the entire period, the area was covered with a tarpaulin to avoid natural infiltration.
3. SYNTHETIC MODELLING

Synthetic modelling of the point injection experiment were performed using the HYDRUS2D software (Simunek et al., 1999). A geological model with 7 layers was created based on the lithological analysis of GPR borehole C (see Figure 2) as well as retention experiments on selected samples performed by Bakmand-Mikalski & Karlsson (2008). Distribution of the layers can be seen in Figure 3-I. Cell size is 0.10 x 0.10 m and the entire model is 5 m wide and 15 m deep. The dominating lithology is layer 3, which is sand. Layer 4 is an other kind of sand with a much higher hydraulic conductivity and different retention parameters.

An injection with an infiltration rate equivalent to the one used in the field experiment was applied for 120 hours in the top cell layer, and the model was run for 5 days of infiltration and 5 days after. In Figure 3 the injected water is seen to move downward as a gravity driven plume and become dispersed into the surrounding sediments just beneath the injection site. After 120 hours of infiltration the water has reached 2 m into the subsurface and the radius of the plume is little more than 1 m. The moisture contents are elevated significantly within the plume, the maximal increase being approximately 0.25. Upon ending the injection after 120 hours the infiltrated water continues to move downward and the moisture content in the top sediments slowly returns to normal. The dipping layer at 2-3 m depth disrupts the downward migration of water and in Figure 3-III the onset of lateral flow is recognized on top of the layer.
The heterogeneity of the subsurface models does not severely disrupt the downwards migration of water and development of a plume. It cannot be disregarded, though, that thicker layers of more impermeable character could divert the water flow to a larger extend. The moisture content development in the HYDRUS2D simulations resemble the results from the field experiment carried out by Binley et al. (2002a), in which an infiltration plume also developed and moved downward in the same way as seen here.
4. RESULTS

4.1 Background moisture content (Day 0)

Assessment of the larger geological structures at Arrenæs can be made by analysis of reflection GPR surveys carried out at the field site. Figure 4 shows two such reflection GPR lines, of which one passes through site B, where the infiltration took place, and the other 20 m south of site B (see Figure 2 for overview). The reflection GPR data were obtained using a Sensors and Software PulseEKKO PE100 system (Sensors & Software Inc, Mississauga, ON, Canada) system equipped with 100 MHz antennas. Trace spacing was 0.25 m, and sampling interval was 0.8 ns. A SEC gain function has been applied to the data, which is otherwise unprocessed.

The GPR lines pictured in Figure 4 show highly variable penetration depths which increase toward S and E, probably due to the lack of clay just beneath the top soil in this direction. This has been confirmed by shallow soil samples obtained around the area at depths of approximately 1 m. The reflections from the upper 10 m of sediments are chaotic and probably arise from moraine sediments of different kinds. In GPR line 5 (Figure 4) deeper structures are visible as continuous reflections that dip toward W, however no such structures are seen in Line 1, probably because these structures are intersected at a different angle. Since the geology of the area is expected to consist of glacial moraine sediments such as sand with intercalations of clay (Looms et al., 2008; Cordua et al., 2008) it is not surprising that the reflection GPR shows that the subsurface is heterogeneous.
Figure 4.
Reflection GPR lines obtained using 100MHz antennae (unmigrated data). Line 1 passes through area A and B, whereas line 5 passes by the southern edge of Area A as seen in the overview (right). Depth in meters is based on velocity of 0.13 m/ns. (Note that the length axis is not the same for the radargrams.)

The mean moisture content as well as the calculated moisture content for all ZOP lines are shown in Figure 5. The average moisture content in the area is 0.07, but in many regions the moisture content is very different from this value. At approximately 3 m we find very low moisture contents in all ZOP lines. The variation in soil moisture at this depth is only a few percent, but from 3 to 8 m, the variation in moisture content increases. At 8 m we find a layer of significantly higher soil moisture in almost all ZOP lines. This layer must be horizontal within the area covered by the ZOP survey.
as it is situated at the same depth in all lines. Below 8 m the variation in soil moisture becomes more significant as lines show soil moistures from below 0.05 to 0.10. An other horizontal layer of increased soil moisture is found at 10 m depth, however the increase is not as significant as at 8 m. From 10 to 12 m the variation in soil moisture between ZOP lines increases as particular lines experience either even higher or lower soil moisture values.

The soil moisture profiles shown in Figure 5 show the same tendencies as those presented in Looms et al. (2008), although more detail about the distribution of the soil moisture can be read from the new ZOP data because of the extra lines in the data set. The lines which exhibit the lowest soil moistures are ZOPAB, -A4, -B4, -A3, whereas lines ZOP2D, -12, -2C, and -23 exhibit the highest soil moisture values. All 4 “wet” ZOP lines are situated around borehole 2. Since ZOPAB, which intersects all the “wet” lines shows low moisture contents, this indicates that just around borehole 2 the moisture content is high.

The velocity distribution obtained from inversion of MOG data between borehole 1 and 3 at day 0, during the five days of infiltration and on three days after infiltration had ceased are shown in the tomograms in Figure 6. Data from day 6 is missing due to equipment malfunctions. The background velocity distribution at Day 0 is seen to be far from uniform, as was also the case for the background ZOP data. From 2 m to 8 m the velocities are rather high, which corresponds to low moisture contents. A low-velocity area dominates at 8 m depth in the vicinity of borehole 1, which corresponds to the high moisture content seen in many of the ZOP lines (Figure 5). The second peak in moisture content found in the ZOP data at 10 m depth is not as significant in the
tomographic section of the subsurface, but there is a section in the middle of the profile at a depth of 10 m, which probably represents this area of higher soil moisture. Toward the very bottom of the tomographic section, velocities decrease, just as the soil moisture is seen to increase slightly for the deepest ZOP data. In general the velocities seen in the tomogram vary less near borehole 3 than near borehole 1.

On the surface of the area there are no visible indications of a variation in geology of the subsurface, however this must be the case because of the heterogeneous soil moisture distribution. As the area is a moraine hill it is plausible that there are intercalations of e.g. clay in the subsurface, which can account for the heterogeneity seen in the velocity tomogram.

Figure 6. Velocity distribution in the subsurface from cross-borehole GPR data. Borehole 1 and 3 are to the left and right, respectively. Note that the velocities in the top 1 m have been interpolated and are not part of the inversion result. Black dot marks injection point.
4.2 Infiltration experiment (Day 1 - 10)

Data collection using both the ZOP and MOG survey continued throughout the infiltration period and 5 days after, however due to equipment failure ZOP data from day 6 and 7 are missing, as well as MOG data from day 6. Figure 7 shows the moisture content increase at Day 1, Day 4 and Day 7, which is calculated by simple subtraction of the moisture content at Day 0 from the moisture content measured for the specific day. After 1 day of infiltration the lines that intersect the infiltration borehole respond to the infiltration, which would also be expected, although the increase in moisture content for both ZOP24 and ZOP13 is surprisingly insignificant (Figure 7). ZOP lines in the Southeastern part of the survey area such as ZOP12, ZOP1B, ZOP1C and ZOPAD also respond to the infiltration at Day 1, and it is ZOPAD which shows the largest increase in moisture content of 2%. The here mentioned lines are colored in Figure 7, whereas the remaining ZOP lines are kept in grey for clarity. The increases in moisture content are seen primarily at depths around 2 m, although the moisture content increase for ZOPAD takes place at 2.5 m depth. The results of the ZOP survey for Day 4 and Day 7 depicted in Figure 7 show significantly larger values of soil moisture increase, up to 6% at Day 4 for ZOPAD. Notice at Day 4 that for some lines (e.g. ZOPAD, -1B, -A4) the moisture content increase is high both at 2 and 4 m depth. This tendency is even more pronounced at Day 7, where especially the moisture content increase of ZOPAD and ZOP1B shows this tendency. The ZOP lines experiencing an increase in moisture content at 2 m depth are slightly different from those where the moisture content increases at 4.5 m. An example of this is ZOPAB, for which the first response to the infiltration happens at Day 7 at 4.5 m depth. Three out of the 4 intersecting lines clearly show the double-peak in Figure 7, whereas ZOP13 shows a more uniform increase in water content between 1.5 and 5 m, which might indicate that the water is moving vertically between the two preferred depths along ZOP13 somewhere.
ZOP data provide one-dimensional information about the velocity in the subsurface, but as the moisture content increase distribution is seen to vary across the area, we here try to establish a quasi-three-dimensional image of the water movement in the subsurface. Illustrated in Figure 8 is the moisture content change at 2 depths for each day of measurement for all 20 ZOP lines across the entire area. Yellow lines indicate that no change has taken place compared to Day 0, blue and red lines represent ZOP lines in which an increase in moisture content is found at a depth of 2-3 m (blue) or 4-5 m (red). The depth-distinction has been chosen because of the double-peak in moisture content change, which is found in many of the ZOP lines (Figure 7). The shaded areas in Figure 8 illustrate the maximum extend of the area of increased moisture content for the two depths associated with the double-peak in moisture content increase. It should be noted that in outlining the blue and red areas, the quantitative increase in moisture content has not been taken into account as the one-dimensional nature of the ZOP gather and the different lengths of ZOP lines make
this difficult. The shaded areas are merely representing the areas, in which there could be infiltrated water present.

Since lines ZOPAD and ZOP12 respond to the infiltration as early as on Day 1, the lateral movement of water is prevalent even at this early stage of infiltration. The maximum extend of the blue area (i.e. moisture content increase at 2-3 m) does not change significantly throughout the remaining days of infiltration. On day 9 and 10, however, moisture content is back to normal in ZOP1C and the blue area has shrunk. The increase in water content at 4.5 m is not evident until Day 3 where the increase in moisture content in ZOP12 and ZOP1B extends to a depth of almost 4 m and we see the double-peak start to evolve.

The presence of the double-peak in moisture content increase indicates that the infiltration plume does not expand homogeneously and that the infiltrating water moves along preferred flow paths thus bypassing certain parts of the subsurface. This was also the case in the data analysed by Looms et al. (2008). Although the data presented therein were the result of a uniform two-dimensional infiltration across the entire area, they also found that an area between 2 and 4 m depth at borehole 1
was bypassed. This emphasises that the development of the infiltration plume is not uniform and that the same areas are bypassed in this infiltration experiment. Because of the lack of data from day 5 and 6 we do not know the maximum water content increase near the infiltration borehole. The largest water content increase is measured on Day 4 of the infiltration for ZOPAD which experiences a delay in first arrival time corresponding to an increase in water content of 6%. The travel time for each line is representative of the entire volume the electromagnetic signal has travelled through, hence the water content increase in single layers in the subsurface may be larger. The different development of the shallow and deeper part of the infiltration plume is indicative of an infiltration which probably follows some geological features or layers in the subsurface. As mentioned earlier, small changes in geology can greatly affect infiltration of water due to minor differences in background moisture content.

Figure 9 shows the result of the 2D-inversion of MOG data as moisture content change for each day of measurement compared to Day 0. The inversion results shown in Figure 6 have been converted to moisture content using Topp’s equation (Topp et al., 1980) and the moisture content change calculated by subtracting the moisture content at Day 0.

There are some inherent problems when interpreting the results of the 2D inversion of MOG data. Because of editing of the dataset, the velocity distribution in the top is almost entirely extrapolated, i.e. there is no initial ray coverage (see Figure 1). This area has been outlined in Figure 6 and 9 and it is evident that the low data coverage area makes capturing and delineating the initial development of the infiltration plume difficult, since injection took place 1.5 m below the surface in the middle of the profile. The moisture content within the extrapolated area is almost constant, which can be probably be attributed to the a priori data which was used in the inversion.

The first distinct response to the infiltration is seen near borehole 1 at a depth of 2 m at Day 1. From here the developing plume moves back toward the middle of the profile while the moisture content increase rises further. After three days of infiltration the moisture content increase reaches all the way from borehole 1 to 3 at a depth of 2-3 m and has also moved further into the subsurface to a depth of 4 m around the middle of the profile. The moisture content increase in these areas is between 0.02
and 0.05. This is all in agreement with the developments seen in the ZOP data, where the double-peak in moisture content increase appeared at Day 4. The double-peak in the ZOP data is most pronounced on Day 7, and here we see in the tomogram that two somewhat isolated areas of increased soil moisture have developed. The upper area is between 2-2.5 m close to borehole 1 and the lower is at 4 m depth where the area of increased soil moisture expands toward the middle of the profile, which is a development also observed in Figure 8, where the ZOP data show that the area of increased soil moisture at 4 m depth changes shape and moves away from the line between borehole 1 and 3, along which the MOG data was obtained.

![Figure 9. Moisture content increase Day 1 to Day 9 compared to Day 0. Contours indicate moisture content increases of 0.02, 0.03 and 0.04.](image-url)
Deeper parts of the profile (from 4 m and down) respond to the infiltration at Day 3, and at Day 5 we find that the soil moisture has increased with up to 0.02 close to borehole 5 at depths of 6 and 10 m. In the remaining parts of the deeper profile we find increases in moisture content of around 0.01, however at Day 7 this effect has disappeared and moisture contents are back to normal. Figure 9 furthermore shows that the largest increase in moisture content at depth takes place at 10 m and not at 8 m, where the first low velocity anomaly was found in the background data. Instead of moving downward as a plume the infiltrated water disperses into the surroundings and only isolated areas are left wetter than when infiltration commenced. On Day 9 we still observe wet isolated areas at both 2 and 4 m depth, while the remaining profile has more or less returned to normal background values of soil moisture. This again concurs with the ZOP data (Figure 8) which show that several lines at Day 10 still exhibit increased soil moisture values (not shown).

The area of increased moisture content has an unusual shape, it is surprising that the earliest response to the infiltration happens close to borehole 1, although we did notice in the ZOP data that ZOP24 does not respond much to the infiltration. This could mean that the infiltrating water is diverted just after it has moved out of the borehole and into the subsurface. Since we don't see any response in the tomographic data to the very early infiltration we suspect that the water is moving in to the subsurface toward borehole 2 right at the infiltration depth. Subsequently and approximately 0.5-1 m further down the water moves back towards borehole 1 and is intersected by the MOG line and captured in the tomogram. Other reasons for the unlikely soil moisture development illustrated by the tomographic data could be inversion artefacts, which for some reason creates increased moisture contents near the side of the model. However, this doesn't seem likely, given that the data coverage in the middle of the profile is better than at the sides. Also, the results are not in direct opposition to the ZOP data, because they show that the primary moisture content increase happens close to borehole 1.
5. DISCUSSION

As the infiltration took place below the upper clay layer it is assumed that all this water has made it into the subsurface at 1.5 m depth. Below 4 m (Figure 9) we find an average increase in soil moisture of 0.01, whereas the moisture content increase in the top of the profile near borehole reaches values of up to 0.04. Assuming that the maximum extend of the infiltration plume at Day 5 corresponds to a sphere with a diameter of 4 m with an average increase in soil moisture of 0.03 the this can account for 550 L of water, which is little more than half of the infiltrated amount. If we add the increase at depth of 0.01 for a cylindrical volume 5 m high and with a 4 m diameter we can account for approximately 80% of the infiltrated amount. Given that water will likely have moved laterally out of the area especially near borehole 1, this number is probably not entirely misguided. Unlike the results presented in Looms et al. (2008) we are thus able to account for the infiltrated water. Further assessment of the moisture content increase can be made by moment analysis of the data, from which the hydraulic conductivity in the area can be calculated based on the movement of the centre of mass.

For a more thorough assessment of the early water movement nearest to the surface, it will be essential to analyse the Electrical Resistivity Tomography data obtained throughout the experimental period. In particular the two dimensional ERT setup between borehole 1 and 3 will supplement the velocity tomogram obtained from MOG data well. The data coverage of the ERT setup is much better in the top of the profile due to the inclusion of surface electrodes in the survey.

During the 2 hour acquisition time for each of the GPR configurations it is obvious that water will have moved within the subsurface. The data error associated with this is not considered a significant problem. However, the combinations of boreholes and the order in which we measured each line did not change from day to day. Thus, when comparing ZOP data from different days the timing of each measurement is actually comparable. Regarding MOG data, the problem is that all data are inverted in the same inversion routine, which can not take the temporal difference between measurements of up to 2 hours into consideration. However, because the inversion
results are both consistent from day to day and in agreement with the ZOP data, it is assumed that this issue is not critical to the data quality.

![Figure 10](image)

Figure 10.
Moisture content change from Looms et al. (2008a) for Day 1-4 during infiltration over 8x8 m area. Data were obtained using a MOG survey scheme. Borehole 1 and 3 are on the left and right, respectively. Modified from Looms et al. (2008).

The field experiment carried out at Arrenæs in October 2008 is similar to the one described in Looms et al. (2008), which took place in 2005. They performed a uniform infiltration across an area of 8x8 m, whereas we here perform a point injection in the middle of the same area. Looms et al. (2008) found that the infiltration did not take place as uniform sheet flow as can be seen in Figure 10, which depicts the soil moisture change for their experiment at Day 1 to 4 inferred from MOG GPR data obtained between borehole 1 and 3 as well. Large volumes of the subsurface are bypassed, for example in the vicinity of borehole 3 between 2 and 4 m depth as well as close to borehole 1 below 6 m. The infiltration pattern is very similar to the one seen in Figure 9 with the same structures developing.

To shed further light on the actual development of the infiltration “plume” it would be worthwhile to perform an inversion of the ZOP data for each measurement depth. In combination with the 2D tomogram from MOG data this will enable a more detailed 3-dimensional assessment of the infiltration. Such results for depths of 2 and 4 meters will probably be along the lines of what is shown very crudely in Figure 8, and could be used to delineate the true three-dimensional structure of the infiltration “plume”. Such results will supplement the two-dimensional tomograms well although the uncertainty will be higher due to the poorer data coverage.
6. CONCLUSION

Arrival times for ZOP data were converted to moisture content and the spatial distribution of ZOP lines allowed for at quasi-3-dimensional assessment of the development of soil moisture in the subsurface. The results of the ZOP data showed the overall same distribution of soil moisture as the MOG data, for which the arrival times where inverted and a 2-dimensional tomographic image of the velocity distribution in a profile through the injection point was obtained.

Both ZOP and MOG data show that the injected water does not move downward as a coherent and homogeneously developing plume of increased soil moisture as seen in other studies (Binley et al., 2002a). Rather, the water follows the same flow paths as observed by Looms et al. (2008) during a plane infiltration experiment. Calculations of the amount of extra water in the subsurface could account for almost all the infiltrated water. The discrepancy is assumed to be due to a combination of calculation uncertainties and the lateral movement of water out of the survey area.

The work presented here effectively illustrates that for this field experiment more than one type of data is required to fully understand the results. We find that for the infiltration scheme designed for this experiment that GPR measurements from the here presented types of MOG and ZOP surveys are somewhat inadequate in describing the early stages of infiltration, due to bad data coverage in the MOG inversion results. In order to improve the quality of the data from this experiment it will be necessary to process and interpret the ERT data as well. This will provide useful information and shed further light on the development of the point injection plume.

The injected plume of water is dispersed and disturbed during its development and never evolved in the same homogeneous manner shown by Binley et al (2002a). Their experiment took place in a geological environment much different than what is found at Arrenæs. Although the flow field is seen to be heterogeneous, we expect that further analysis of the remaining data will show that cross-borehole geophysical methods (GPR and ERT) can successfully be used for monitoring and assessment of point injections at field sites like Arrenæs containing alluvial sediments.
REFERENCES


APPENDIX C

AUTHORSHIP STATEMENTS
AUTHORSHIP STATEMENT

PAPER 1

Title of paper: Visualising unsaturated flow phenomena using reflection ground penetrating radar

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Journal: Not yet submitted

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Description of the individual authors’ contribution to the paper:

E.B. Harder, M.C. Looms, K.H. Jensen and L. Nielsen carried out the field experiments.

E.B. Haarder carried out the data processing and analysis that lay the basis for the paper and is the main author.

M.C. Looms provided the matlab code for visualising GPR data and picking arrival times.

M.C. Looms K.H. Jensen, and L. Nielsen contributed with helpful suggestions for the data analysis, and to the writing and reviewing of the manuscript.
Authorship Statement

Paper 2

Title of paper: Unsaturated flow inferred from cross borehole ground penetrating radar monitoring of point injection of water

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Description of the individual authors' contribution to the paper:

E.B. Haarder and M.C. Looms carried out the field experiments.

E.B. Haarder carried out the data analysis and bulk work, and is the main author.

M.C. Looms provided the matlab code for picking arrival times.

M.C. Looms, K.H. Jensen, and L. Nielsen contributed to the ongoing discussion of the results.