Interferometry by deconvolution of GPR data

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***Abstract*—Directional decomposition of GPR data is possible when multi-component multi-offset GPR data is available on a surface. The decomposed wave fields are propagating in the two opposite directions perpendicular to the recording surface. These two decomposed wave fields are related to each other through a reflection operator. This operator represents the reflection response of the medium on the side of the surface that is source free, and with the receivers acting as both sources and receivers. Here we illustrate the concept in a one-dimensional setting and show how this concept can be used for different GPR applications.**

***Index Terms*—interferometry, deconvolution, multi-offset, GPR.**

I. INTRODUCTION

The possibility of bringing the principle of daylight imaging, as we do with our eyes, over to other diffuse wave fields is known for more than 20 years, [1], [2]. During the last eight years many interferometric methods have been developed for random fields and for controlled-source data. The underlying theories have in common that the medium is assumed to be lossless and non-moving. The main reason for this underlying assumption is that the wave equation in lossless and non-moving media is invariant for time-reversal. For an overview of the theory of seismic interferometry or Greens function retrieval and its applications to passive as well as controlled-source data, we refer to a reprint book [3], which contains a large number of papers on this subject.

Until 2005 it was commonly thought that time-reversal invariance was a necessary condition for interferometry, but recent research shows that this assumption can be relaxed. The first analysis of the interferometric method for ground penetrating radar (GPR) data, in which losses play a prominent role, can be found in [4]. It was shown that losses lead to amplitude errors as well as the occurrence of spurious events. By choosing the recording locations in a specific way, the spurious events arrive before the first desired arrival and can thus be identified [5]. An example of applying this type of interferometry to borehole GPR data is found in [6]. In many applications in natural soil, the losses are too high to use cross correlation methods and this problem can be circumvented using cross convolution techniques when the sources are located on an irregular boundary [7]. The number of acquisition configurations where cross convolution methods can be used is rather limited. By choosing one receiver in a lossless medium, e.g. air, and a configuration with all dissipative parameters outside the surface distribution of noise or transient sources, cross correlation methods work without strong amplitude errors in the time window of interest [8]. The occurrence of spurious events depends on the irregularity of the boundary surface and on the strength of the subsurface heterogeneity. This approach is valid for waves and diffusive fields in dissipative media. Independently, it was shown that a volume distribution of uncorrelated noise sources, with source strengths proportional to the dissipation parameters of the medium, precisely compensates for the energy losses [9], [10]. As a consequence, the responses obtained by interferometry in such configurations are error free. Also this approach holds for waves in dissipative media and for pure diffusion processes. In natural media, the presence of spatially uncorrelated volume noise sources with correct strength is not considered probable. Recently it was shown that interferometry by cross correlation, including its extensions for waves and diffusion in dissipative and/or moving media, can be represented in a unified form [11], [12]. From the result of these studies we conclude that the above described methods provide a rather limited practical approach to applications where intrinsic loss factors play a significant role. For specific configurations interferometry can be applied through a deconvolution of the up going field by the down going field at the same depth level [13]. Here we demonstrate that this concept of 'interferometry-by-deconvolution' is suitable for GPR applications.

II. RADAR WAVE INTERFEROMETRY IN DISSIPATIVE MEDIA

Let us start by illustrating the concept in a one-dimensional setting. The 1D version of interferometry-by-deconvolution was introduced by Riley and Claerbout for seismic waves [14]. It is valid for dissipating media and therefore it is of interest for GPR applications. It starts with a source in a stack of layers and the fields that define the energy state are recorded at a particular depth level away from the source level. These recorded fields can be decomposed into flux-normalized down going and up going parts. At any source-free depth level, z_l, we can write the electric field, $\hat{E}(z_l,\omega)$, and magnetic field, $\hat{H}(z_l,\omega)$, in terms of the up going potential, $\hat{p}^-(z_l,\omega)$, and down going potential, $\hat{p}^+(z_l,\omega)$, as

$$\begin{bmatrix} \hat{E}(z_l,\omega) \\ \hat{H}(z_l,\omega) \end{bmatrix} = \frac{1}{\sqrt{2}} \begin{bmatrix} a & a \\ 1/a & -1/a \end{bmatrix} \begin{bmatrix} \hat{p}^+(z_l,\omega) \\ \hat{p}^-(z_l,\omega) \end{bmatrix}, \quad (1)$$

where $a = \sqrt{\hat{Z}}$, and the plane wave impedance $\hat{Z}$ is given by $\hat{Z} = \sqrt{\mu/\hat{\mathcal{E}}}$, where $\hat{\mathcal{E}} = \hat{\varepsilon} - \mathrm{j}\hat{\sigma}/\omega$ and the plane wave admittance $\hat{Y}$ is the reciprocal of the plane wave impedance. Note that for physical reasons we allow both the electric permittivity $\hat{\varepsilon}$ and the conductivity $\hat{\sigma}$ to be functions of

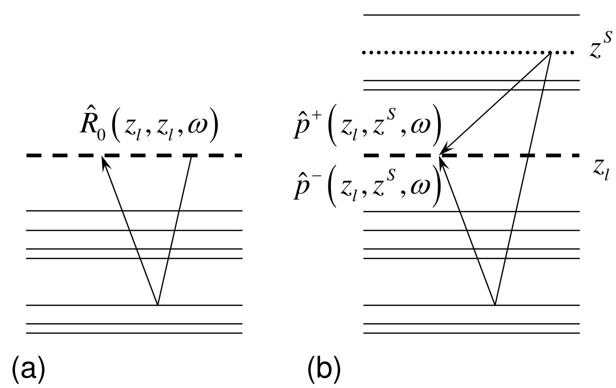


Fig. 1. (a) State A: the response of the medium inside $\mathbb{D}$ with a homogeneous upper half space, above $\partial\mathbb{D}_1$; (b) State B: the measured response of the real earth.

frequency, although mathematically a frequency dependent permittivity can contain a frequency dependent conductivity term implicitly. For a source located above the receiver level it was shown that the up going part deconvolved in the time domain by the down going part yields the reflection response of the stack of layers that are below the receivers, while all other interfaces have disappeared and a homogeneous half space exists above the receiver level. Deconvolution in the time-domain corresponds in the frequency domain to division, which can be written as

$$\hat{R}_0(z_l, z_l, \omega) = \hat{p}^-(z_l, z^S, \omega)/\hat{p}^+(z_l, z^S, \omega), \qquad (2)$$

where $\hat{R}_0(z_l, z_l, \omega)$ denotes the reflection response of the layered earth that is homogeneous above the receiver level z_l, which is now also the source level. The source location, z^S, is now put explicitly in the expressions for the up going wave field $\hat{p}^-(z_l, z^S, \omega)$ and down going wave field $\hat{p}^+(z_l, z^S, \omega)$, which are again flux-normalized recorded wave potentials in the measurement configuration with heterogeneities below and above the source level and receiver level. This is illustrated in Figure 1 where Figure 1(b) shows the measurement configuration with reflectors above and below the source and receivers and Figure 1(a) shows the result after division in the frequency domain. In the right-hand side of equation (2) two recorded wave fields occur, the up going field at the receiver level and the down going field at the receiver level, both due to a source at a higher level. In the left-hand side of equation (2) a reflection coefficient occurs at the receiver level due to a source at that same level. We have hereby created a new response through deconvolution that can be understood as a weighted form of cross correlation, for which reason we call it interferometry-by-deconvolution (IbD) [15]. A straight division as described in equation (2) can be numerically unstable and a stabilized version is given by,

$$\hat{R}_0(z_l, z_l, \omega) = \frac{\hat{p}^-(z_l, z^S, \omega)[\hat{p}^+(z_l, z^S, \omega)]^*}{|\hat{p}^+(z_l, z^S, \omega)|^2 + \epsilon^2}, \qquad (3)$$

where the superscript * denotes complex conjugation and ϵ is the stabilization factor. The following two important observations can be made from equation (2) and Figure 1. Firstly, we have eliminated the original source without needing to know its location other than that it must be located above the receiver level. Secondly, this newly created response does not contain the direct field, which is usually strong in GPR applications, nor does it contain any information from the level above the receivers.

Consequentially above surface objects, whose presence show up as reflections that can hamper unshielded GPR data interpretation [16] are eliminated in this process. Eliminating the overburden can be achieved through IbD at any depth level. This implies that it is very suitable in monitoring studies when a horizontal borehole is available and the changes in the target zone can be studied using time-lapse GPR data, without being bothered by all changes in the subsurface above the target zone, e.g., changed weather conditions or changes due to anthropogenic activities. A last interesting detail in this procedure is the possible presence of an interface at the receiver level, such as the earth surface. Figure 1 is just a sketch and the exact location of the up going and down going wave fields is not specified. If the receivers are located in a homogeneous layer, the up going wave field just above the receiver level is the same as the up going wave field just below the receiver level and the same applies for the down going wave field. However, the decomposed wave fields are not continuous across an interface. This implies that when the receivers are located at an interface between two different layers, different up and down going wave fields occur at both sides of the interface as shown in Figure 2. In principle it does not matter whether the up going and down going fields above or below the interface are used. Of course the presence of the interface remains in the created data when the up going and down going wave fields above the interface are used, while it has vanished when the up going and down going wave fields below the interface are used. Especially the earth surface is usually a strong reflector and for data recorded at the earth surface it is worthwhile to use the fields below the surface in the deconvolution. To carry out the decomposition of the data into up going and down going wave fields below the earth

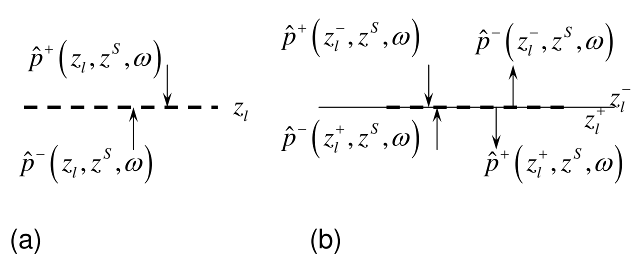
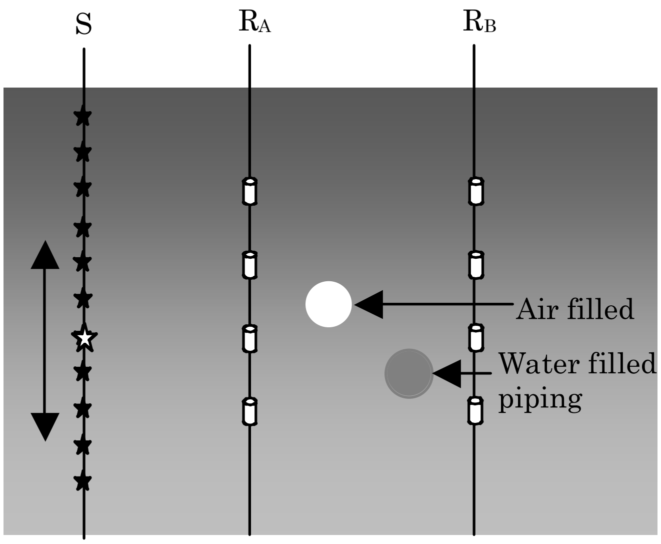
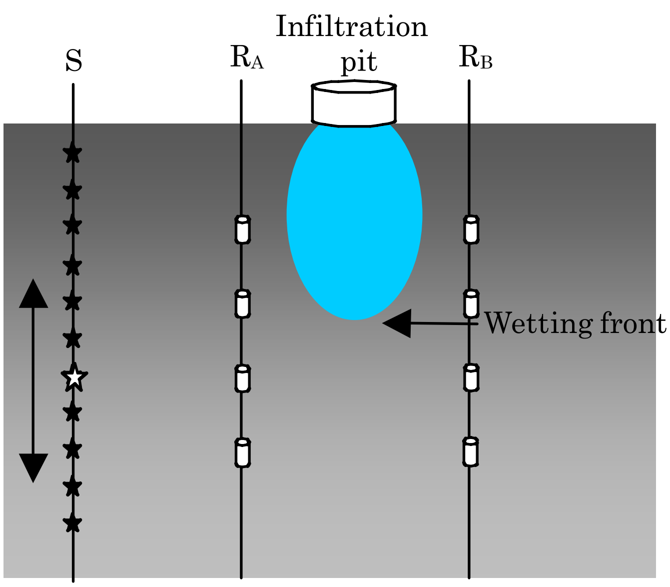


Fig. 2. (a) Receiver at a homogeneous level, up going waves only come from below the receiver level, while down going waves only come from above the receiver level. (b) Receiver at an interface of discontinuity, up and down going waves occur at both sides of the receiver level.

surface it is necessary that the medium parameters just below the surface are known and they can be obtain by independent measurements or be estimated from the data.



(a)



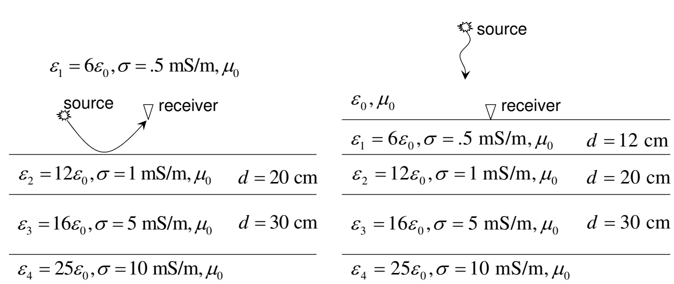
(b)

Fig. 3. Possible scenarios for (a) monitoring piping or the occurrence of preferential flow paths; (b) monitoring infiltration wetting fronts for vadose zone hydrology (b).

Here we are interested in the situation as depicted in Figure 3 and we can perform the left going and right going separation of the recorded wave field. Let us say we do not use directional antennas in any of the holes and that we can measure the vertical electric field with a dipole antenna and the cross-plane magnetic flux with a suitable loop antenna. Let us assume that the zone of interest is the area in between the two receiver holes, R_A and R_B. By deconvolving the left going wave field by the right going wave field the whole subsurface left of the receiver hole is homogenized and has the properties of the medium just right of the borehole. The advantage of this approach that possible changes over time that have occurred in the zone to the left of the zone of interest have disappeared from the data. This provides a much cleaner data set and all changes over time are certain to occur to the right of the receiver hole where IbD has been applied to the recorded data. Similarly, to investigate if any changes to the right of the zone of interest have occurred, it is sufficient to perform this IbD on the data in the right most borehole and repeat this procedure for the time-lapse monitoring data. In these data sets, only reflections from heterogeneities that occur to the right of the zone of interest remain in the data. If these reflections change in the time lapse monitoring data, it implies that changes have occurred to the right of the zone of interest. These can then be imaged and characterized using the data after IbD from the right most bore hole and this helps the imaging and characterization of the time lapse processes that we are interested in and that occur in the zone of interest. Suitable applications can be found in monitoring small geo-environmental changes like occurrence of preferential flow paths or piping as depicted in Figure 3(a), or in monitoring of infiltration wetting fronts, e.g., for vadose zone hydrology as shown in Figure 3(b) . It is interesting to note that in case the middle borehole indicated in Figure 3(a) or (b) is used as the transmitter hole, data in the two other boreholes can then be used to generate new GPR data between the two receiver wells by cross-convolving the data from one borehole with the data from the other borehole [17].

III. NUMERICAL EXAMPLES

To give a numerical example of the technique consider a one-dimensional earth model with a normal incidence plane wave excitation from a source at some unknown height above the surface. The receivers measuring the scalar electric and magnetic field strengths are located at the earth surface. The earth is modeled as a layered medium with layer properties given in Figure 4. Figure 5 shows the earth response for the



(a) (b)

Fig. 4. The model that results after the deconvolution procedure with the first layer extended as the upper half space (a) and the measurement configuration with a plane wave source at unknown location in the air and receivers at the earth surface (b).

model of Figure 4(b) using a 750 MHz center frequency Ricker wavelet. The response from the first subsurface interface is partly masked and relatively strong earth surface multiple reflections are visible at the tails of all reflections, while internal multiples are much weaker. After the deconvolution procedure, where we retained the bandwidth of the original source wavelet, the upper half space has the properties of the first layer and no direct wave is present. The absence of both the original earth surface and the direct wave is clearly visible in Figure 6 where the reflections form the three interfaces are clearly visible and are not contaminated with earth surface related multiples. To demonstrate the results are exact the direct modeled reflection response is also shown in Figure 6 with no visible differences.

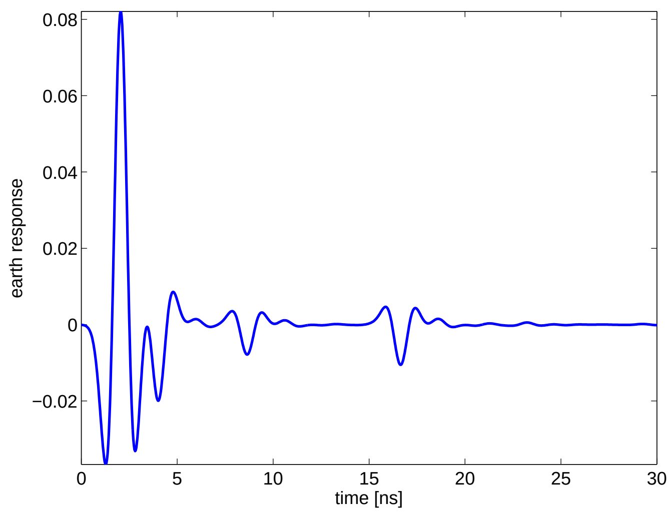


Fig. 5. The electric field earth response measured by one of the receivers at the earth surface of the model shown in Figure 4(b).

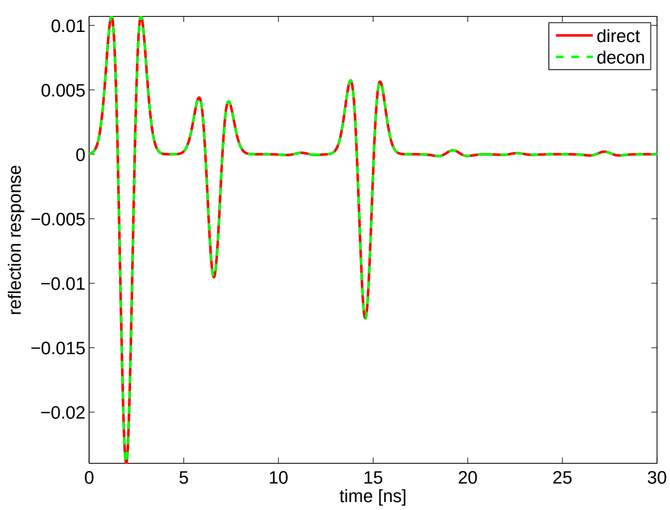


Fig. 6. The reflection response after deconvolution (green dashed line) and the directly modeled reflection response (red solid line) of the model in Figure 4(a).

IV. CONCLUSIONS

We have illustrated how the concept of interferometry-by-deconvolution can be applied to GPR data. We have discussed possible practical measurement configurations for surface GPR and borehole GPR applications. We have argued that the two main advantages of IbD are the homogenization of half space starting at the receiver line and containing the source locations, and the absence of the direct field that usually masks other reflection events. These advantages make this approach especially interesting for monitoring studies.

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