Imaging subsurface objects by seismic P-wave tomography: numerical and experimental validations

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ABSTRACT
We tackle the problem of characterizing the subsurface, more specifically detecting shallow buried objects, using seismic techniques. This problem is commonly encountered in civil engineering when cavities or pipes have to be identified from the surface in urban areas. Our strategy consists of processing not only first arrivals, but also later ones, and using them both in tomography and migration processes, sequentially. These two steps, which form the basis of seismic imaging, can be carried out separately provided that the incident and diffracted wavefields are separated in the data space. Tomography is implemented here as an iterative technique for reconstructing the background velocity field from the first-arrival traveltimes. The later signals are then migrated by a Kirchhoff method implemented in the space domain. To study the reliability of this methodology, it is first applied to synthetic cases in the acoustic and elastic approximation. Both the background velocity field and the local impedance contrasts are reconstructed as defined in the predicted model. An experimental case, specifically designed for the purpose, is then considered in order to test the algorithms under real conditions. The resulting image coincides well with the predicted model when only P-waves are generated. In the elastic mode, surface waves make P-wave extraction difficult, so that the reconstruction remains incomplete. This is confirmed by the real data example. Finally, we demonstrate the appropriateness of the proposed method under such circumstances, provided that suitable preprocessing of data is carried out, in particular, the removal of surface waves.

INTRODUCTION
Among the many difficulties encountered in the environmental and civil-engineering fields, characterization of the subsurface from a structural and a mechanical point of view constitutes the major challenge. Seismic sounding is an appropriate geophysical method to tackle this problem because of its ability to produce high-resolution images of the first 100 m of the subsurface. Numerous examples illustrate the diverse applications where seismic methods are effective, whether for detecting cavities (Grandjean et al. 2002), imaging active faults (Pratt et al. 1998) or delineating aquifer structure (Liberty 1998; Grandjean 2006). Such seismic images are generally difficult to obtain, especially when the subsurface is characterized by strong velocity variations with heterogeneities similar to the seismic signal wavelength. Nevertheless, we show that such media can be studied by this means, provided that both the first and the later arrivals are considered. Our work is particularly focused on applications related to civil engineering, such as the detection of shallow buried objects, like cavities or pipes.

As Mora (1989) has already described, the complex problem of tomography and migration in subsurface imaging needs to be tackled. The former is dedicated to estimating large wavelength velocity contrasts, here called the background velocity. By extending the normal moveout (NMO) analyses, the tomography concept has the advantage of being applicable to laterally heterogeneous velocity fields. Moreover, seismic antenna geometry can account for complex configurations making the data independently indexed to a source–receiver pair, in contrast to classical seismic profiling where data refer to CDP gathers. Experimental seismology provides good examples of the possibilities, as shown by Nolet (1987). The migration process is often considered for the step following tomography because it relies upon the background velocity field to re-allocate the seismic events to their real positions. Numerous techniques can be used, depending on the complexity of the sounded medium, the kind of data considered, i.e. prestack or post-stack, and the available a priori information on the velocity field. For complex media, the prestack depth migration is generally used for ensuring the focusing of reflectors (Robein 1999). In particular, reverse-time migration has been successfully tested for detecting cavities from radar-wave measurements (Hui Zhou and Sato 2004).

New methods have been developed more recently, either for reducing the computation times (Ecoubet et al. 2002) or improving the inverse problem conditioning. For example, the
stereotomography method (Gosselet et al. 2003) inverts the velocity field and the medium reflectivity at the same time. This technique requires the picking of locally coherent seismic events by determining their arrival time and their slope in common receiver and source gathers. Because this work can be very detailed and demands major computational resources, we opted for a strategy where tomography and migration processes are carried out separately. This two-step approach, taken as the basis of seismic imaging, can be carried out sequentially, provided that the incident and diffracted wavefields are separated in the data space (Dessa et al. 2004). Here, we chose to use a simultaneous iterative reconstruction technique (SIRT) tomography algorithm for reconstructing the background velocity field from the first-arrival traveltimes. The late arrivals are then re-allocated by a Kirchhoff migration technique computed directly in the space domain to image the local velocity anomalies. This imaging technique is tested with both with synthetic cases and with a real example, in order to study its robustness, as well as possible associated difficulties.

THEORETICAL APPROACH

In the tomography and migration processes, we first need to define and solve the forward problem of wave propagation for computing traveltimes. According to previous work in this field, several alternatives can be considered, for example, ray tracing or the finite-difference time-domain (FDTD) method. Ray tracing (Čeverný et al. 1977) is defined within the framework of the asymptotic-ray theory in the high-frequency approximation. Using the Born approximation, ray tracing is a fast and robust way to compute wavepaths, provided that the medium is sufficiently smooth. This condition rules out media that are too rough, thus making ray tracing unsuitable for our purpose, as we are considering the highly heterogeneous subsurface domain. In our approach, we aim to represent wavepaths by fat rays, i.e., Fresnel volumes, because they depend intrinsically on the wave resolution (Husen and Kissing 2001). Fresnel volumes are computed by solving the eikonal equation, which can be numerically resolved, for example by the well-known FDTD method (Virieux 1986). This scheme uses a numerical integration of the equation to compute the wavefield propagation on a grid that is spatially discretized according to the signal wavelength divided by ten. Each grid node is characterized by the properties of the medium, making it possible to work with highly heterogeneous media. Unfortunately, the FDTD method is time-consuming, so we decided to tackle the forward problem by using another method based on a faster numerical technique. In the following sections, we describe the principles of the proposed imaging technique and establish the specific conditions of our approach: the use of Fresnel wavepaths in the forward step of the inverse problem, a probabilistic reconstruction algorithm for reconstructing the background velocity field, and a Kirchhoff-based migration in the space domain to solve the diffraction imaging problem.

FMM and Fresnel wavepath approach

The solution of the eikonal equation on a discrete lattice by a fast tracking scheme was presented by Cao and Greenhalgh (1994). More particularly, the fast marching method (FMM), proposed by Sethian and Popovici (1999), solves this equation by an upwind scheme for reducing the computing times. Implemented numerically to the second order (Grandjean and Sage 2004), the algorithm consists of successively searching for the lattice gridpoint to which the wavefront has to be propagated. This procedure, which mimics a monotonously expanding wavefront, is repeated until the lattice has been completely scanned. In fact, this procedure can be considered as another expression of the Huygens’ principle.

Čeverný and Soares (1992) used a ‘paraxial ray theory’ to define Fresnel volumes. In a medium between source $S$ and receiver $R$, the Fresnel volume is defined by the set of points $M$, where the waves are delayed after the shortest traveltime $t_{sr}$ by less than half a period. These waves can thus be added constructively to form the first arrival of the wave. This approach considers the centre frequency of the wave in the analysis, thus enabling the evaluation of the tomography resolution and reducing the sparseness of the ray distribution. Frequency band-limited waves propagating in the ground are thus affected not only by structures along the raypath, as assumed by the ray theory, but also by structures located in the vicinity of the raypath. Using this approach, Watanabe et al. (1999) proposed an algorithm for calculating a Fresnel volume:

1. Compute the traveltimes from $S$ to $R$ through each gridpoint $M$ of the grid: $t_{SMR} = t_{SM} + t_{MR}$.
2. At each gridpoint $M$, estimate the difference $\Delta t$ (Harlan 1990) between $t_{SMR}$ and $t_{SM}$; where $t_{sr}$ is the shortest traveltime from $S$ to $R$.
3. At each grid point $M$, test whether $\Delta t$ is less than half a period:

$$\Delta t = t_{sa} + t_{sm} - t_{sr} < \frac{1}{2f},$$

where $f$ is the centre frequency of the propagating wave.

Watanabe et al. (1999) also proposed a numerical definition of Fresnel volumes, characterized by a weighting function $\omega$ that depends linearly on the delay expressed in (1):

$$\omega = \begin{cases} 1 - 2f\Delta t, & 0 < \Delta t < \frac{1}{2f} \\ 0, & \frac{1}{2f} < \Delta t \end{cases}. $$

This formulation will be used in the tomographic reconstruction to compute the seismic path from a source to a receiver at each iteration. One important point in the inverse problem of seismic imaging is related to underconditioning. Because the seismic wave interacts with subsurface objects only from the

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surface, their shadow side is poorly investigated, creating divergence in the solution space. This effect can be damped by using regularization techniques during the reconstruction (Gosselet 2004). With our approach, regularization is straightforward since the spatial influence of the local velocity variations on Fresnel wavepaths depends on the centre frequency of the propagating wave: the lower the frequency, the fatter the Fresnel wavepath and the more the velocity variations far from the ray are taken into account in the reconstruction.

Tomography of background velocities

In the following, the indices $i$, $j$ and $k$ are related to the source–receiver number, the cell number and the iteration number, respectively. The fundamentals of the proposed SIRT state that a slowness perturbation $\Delta s$ in cell $j$ generates a travelt ime perturbation $\Delta t$, according to

$$\frac{\Delta s_j}{s_j} = \frac{\rho_j - \rho^\prime}{\rho^\prime} = \frac{\Delta t_j}{t_j} \Rightarrow \Delta s_j = s_j \frac{\Delta t_j}{t_j},$$

(3)

where the superscripts $O$ and $C$ refer respectively to observed and calculated first arrivals. We can now state that the Fresnel weights defined in (2) represent the propagating wave: the lower the frequency, the fatter the Fresnel wavepath between the observed and the calculated traveltimes. Rather than using an arithmetic mean, Watanabe et al. (1999) averaged the slowness update over the sum of all the Fresnel weights in cell $j$, i.e. $\sum_{t} w_i$. Consequently, low-weight cells are updated in the same manner as high-weight cells. Because low-weight cells are defined as low probability zones that can explain the differences between observed and calculated traveltimes, logically we relate the weight to the slowness perturbation applied to each cell: the lower the probability, the weaker the slowness perturbations. With JaTS-SIRT, the cell weight $w_i$ is normalized by the sum of all the weights $\sum_{t} w_i$ that Fresnel wavepath. This increases the influence of small-scale Fresnel wavepaths compared to larger-scale ones, which makes intuitive sense.

Diffraction imaging

After the JaTS-SIRT algorithm has been used to reconstruct the background velocity, we can re-allocate late arrivals by using a migration process. We consider a point $(x,z)$ located in an elastic homogeneous medium bounded by a surface $\Sigma$. The backpropagation of the pressure field $P$ from $\Sigma$ to the point $(x,z)$ is given by the Kirchhoff integral (Berkhout 1987):

$$P_{x,z} \propto \int_{\Sigma} \alpha P \left( \tau - \frac{r}{V} \right) + \beta V \cdot n \left( \tau - \frac{r}{V} \right) d\Sigma,$$

(8)

where $\tau$ is the propagating time, $r$ is the wavepath, $V$ is the velocity of the medium, $P_{x,z}$ and $V \cdot n$ are respectively the pressure field and the normal component of the velocity vector recorded at the surface, and $\alpha$ and $\beta$ are weighting parameters introduced in the Green’s function. This expression can be simplified by considering the acoustic case and neglecting amplitude effects in the propagation. The reflectivity $R$ at the point $(x,z)$ is then deduced by a double summation, respectively related to the Kirchhoff integral and to the data redundancy due to the multifold acquisition (Robein 1999). For $N_j$ sources and $N_i$ receivers, we have

$$R_{x,z} = \frac{1}{N_j} \sum_{j} \sum_{r} P_j \left( X_r, Z = 0, \tau - (t_{s_j} - t_{r_j}) \right),$$

(9)

where $X$ and $Z$ are the coordinates of the receiver $i$ at the surface, and $t_{s_j}$ and $t_{r_j}$ are the traveltimes from the source $j$ to the point $(x,z)$ and from this point to the receiver $i$, respectively. For convenience, (9) can be rearranged as
we can use either frequency filtering or the Karhunen–Loève technique, the latter being well-suited to sepa-

trating dispersive waves from others (Bitri and Grandjean 2004). In the following sections, we consider three cases of increasing complexity. Two synthetic cases, realized in the acoustic and elastic approximations, are used to investigate the efficiency of the algorithms when either guided waves or wave conversions are not generated but are observed; a discussion presents some solutions for filtering non-acoustic wave components. The last case presents results obtained for real data.

**Synthetic cases**

We first consider a synthetic case in the acoustic approximation. In this model, a high velocity (and density) object is embedded in a two-layer medium, 5 m long and 1 m deep. This case represents a scaled-down model of a pipe located in the subsurface. The low-

velocity upper layer, velocity 200 m/s (density 1800 Kg/m$^3$) and thickness 0.1 m, represents the weathered zone. The high-velocity lower layer, i.e. the host material, has a velocity of 250 m/s (density 1900 Kg/m$^3$). A circular anomaly of velocity 700 m/s (density 2200 Kg/m$^3$), 0.4 m deep with a radius of 0.2 m, is introduced into the host material (Fig. 1a). A 600 Hz Ricker wavelet is used as source signal in order to maintain the ratio of approximately unity

\[
R_{ij} = \frac{1}{N_j} \sum_j \sum_i |S_i(x,z) e^{-2\pi t}|
\]

where $t(f_x, z)$ is the traveltine computed from the source $f$ to points $(x, z)$ of the medium and from these points to the receiver $i$. If the background velocity field can be differentiated from the velocity tomography, the traveltine map can be easily computed by FMM as described in the previous section.
between wavelength and the anomaly depth, as is commonly observed in reality (Grandjean and Leparoux 2004). Figure 1(b) shows the Fresnel wavepath related to source #1 and receiver #15. Note that two distinct waves arrive at the same time: one is refracted in the second layer, the other from the anomaly.

Seismic records were simulated by the FDTD code implemented in JaTS in the acoustic approximation and described by Keiswetter et al. (1996). The modelling scheme is based on the scalar wave equation in two dimensions:

\[
\frac{\partial^2 p}{\partial t^2} - \rho \frac{\partial}{\partial x} \left( \frac{1}{\rho} \frac{\partial p}{\partial x} \right) + \frac{\partial}{\partial z} \left( \frac{1}{\rho} \frac{\partial p}{\partial z} \right) = \delta(x_s, z_s) f(t),
\]

where \( p \) is the acoustic pressure, \( \rho = \rho(x,z) \) is the density, \( V_p = V_p(x,z) \) is the P-wave velocity, \( \delta(x_s, z_s) \) is the Dirac delta function, which takes the value of unity at the source position and zero elsewhere, and \( f(t) \) is a time-dependant source signal defined for a given number of iterations.

The finite-difference solution of (11) uses a second-order difference scheme. Here, the second derivative with respect to time is approximated using a standard central differencing equation. To reduce numerical dispersion, the sampling interval is reduced to ten gridpoints per wavelength. For numerical solutions to be stable, stability criteria are imposed according to Kelly et al. (1982). The boundary conditions are those described by Reynolds (1978).

The seismic shot corresponding to source #1 is shown in Fig. 1(c). The direct and refracted waves clearly appear as first arrivals, and the diffracted signal related to the presence of the P-wave velocity anomaly is indicated by red arrows. A total of nine sources, spaced every 0.5 m, were shot using this model. For each simulation, vertical displacements were recorded in time at receiver points spaced every 0.15 m.

The generic processing flow applied to the data is summarized in Fig. 2, except that no surface-wave suppression was applied, because the synthetic data were computed in the acoustic approximation. Processing was thus reduced to trace amplitude normalization and manual first-break picking. Using the source and receiver positions, picks were used to reconstruct the background velocity field (Fig. 3a). After muting the first arrivals, the late arrivals were migrated, assuming the background velocity field previously reconstructed by tomography. This step allows the reconstruction of local impedance contrasts related to the local anomaly, as shown in Fig. 3(b). This resulting image can be compared with the predicted model of Fig. 1(a). We note that the upper velocity layer has been correctly imaged, even if the transition to the deeper one is not as sharp. This lack of resolution comes from the limited frequency bandwidth of the seismic signal, as is commonly observed in subsurface seismic. Its origin lies in the band-limited source frequency content. The local velocity anomaly is correctly imaged, as we can identify its shape and depth, but the lateral contours remain blurred. This artefact, also commonly observed in surface seismics, is the result of the distortion induced by the source and geophones geometry, which is classically spread on the surface side only, making diffractions from the lateral sides undetectable. This synthetic case demonstrates the stability of the algorithms presented and their ability to image subsurface velocity structures when only P-waves are generated. Nevertheless, we note certain limitations, essentially due to wave resolution and source/geophone geometry. These limitations are well-known in seismic imaging subsurface object with seismic tomography

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engineering and can be attenuated under two conditions: working with a signal bandwidth as wide as possible in high frequencies, and using additional seismic sources and receivers gathered around the study zone, for example in boreholes.

In order to present a more realistic scenario, we now consider an elastic propagation in the same velocity model. This model had shear-wave velocities that were 0.7 times the P-wave velocities. To maximize the ray coverage, we doubled the number of geophones. The FDTD scheme used to compute seismograms was adapted from Virieux (1986), Levander (1988) and Juhlin (1995), and is based on the stress–velocity wave equation in two dimensions. In the 2D Cartesian system, the P-SV equations of motion are

\[ \rho \frac{\partial u}{\partial t} = \frac{\partial \tau_{xx}}{\partial x} + \frac{\partial \tau_{xz}}{\partial z}, \quad \frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \tau_{zz}}{\partial z}, \quad (12a) \]

with the constitutive equations for an isotropic medium:

\[ \tau_{xx} = \left( \lambda + 2\mu \right) \frac{\partial u}{\partial x} + \frac{\partial \sigma_{zz}}{\partial z}, \quad \tau_{xz} = \mu \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial z} \right), \quad (12b) \]

where \( u, v \) and \( u_x, u_y \) are respectively the displacement and velocity components along the \( x \) and \( z \)-axes, \( \lambda, \mu \) and \( \rho \) are the elastic constants, and \( t_x \) and \( t_y \) refer respectively to the normal and tangential stress components. Equations (12a) and (12b) are differentiated and solved using a 2D staggered finite-difference grid (Virieux 1986; Levander 1988). Boundary conditions are classically defined to model a semi-infinite space, satisfying the free-surface conditions at \( z = 0 \). The other boundaries at the grid periphery are coded to satisfy the absorbing conditions of Clayton and Engquist (1977). A spatially localized source is initiated with a second Gaussian derivative function of stress components, using the source insertion principle of Alterman and Karal (1968).

Figure 4(a) shows a seismic synthetic shot representing the elastic wavefield generated by a source located at \( x = 2.25 \mathrm{~m}, \ z = 0 \mathrm{~m} \), and recorded at 51 geophones spread over the surface. On this shot, the P-wavefield, consisting of direct and diffracted waves, is difficult to identify because of the presence of surface waves. In order to extract the P-wavefield, we used a Karhunen–Loève (KL) filter to separate dispersive waves from volume waves (Birgi and Grandjean 2004). This method consists of computing the dispersion image of the seismic shot that represents the distribution of the seismic energy in the phase-velocity versus frequency domain (Park et al. 1998). From this image, the dispersion curve can be identified as the set of maximum values (the black line in Fig. 4d). This curve is then used to filter the seismic shot by a KL technique so that surface waves (Fig. 4b) are separated from non-dispersive waves (Fig. 4c). On this filtered data, it is easier to observe not only the direct wave but also the diffracted waves that are related purely to the local heterogeneities of the medium. Finally, the imaging technique described in the previous section can be carried out on the filtered data.

![Figure 5](image)

**Figure 5**

(a) Background velocity field inverted from the first-arrival traveltimes: they were computed from the model in Fig. 1(a) in the elastic approximation; the two layers and their respective velocities are clearly visible. (b) The anomaly reconstructed from the late arrivals using the velocity field shown in (a) is highlighted by the dotted circle.

Figure 5 shows the results of both the reconstructed velocity field (Fig. 5a) and the migrated late-arrival signals (Fig. 5b). Compared with the previous acoustic case, the diffraction patterns are not imaged with sufficient accuracy to be correctly interpreted. The reasons that the circular velocity anomaly is unfocused can be related to (i) the presence of multiple diffractions or P–S wave conversions that are not filtered by the KL technique; (ii) damage to the diffracted signals during the filtering processes that introduces phase perturbations. Both these phenomena alter the stacked signals during the imaging step, making it impossible for the migrated section to image local anomalies correctly. We should also mention that a low \( V_p/V_s \) ratio was used in these simulations in order to restrict the generation of higher modes of guided waves and P–S wave conversions. The presence of such complex wavefields could introduce additional complications into our method, particularly when extracting P-waves by filtering. The next section deals with this type of challenge by considering real data recorded on a test site.

**EXPERIMENT ON A TEST SITE**

In this section, we apply the above-described methodology to the data set derived from Grandjean and Leparoux (2004). The principle of this experiment was to reproduce, at a reduced scale, the interaction between seismic waves and a void. The geometry and characteristics of the seismic line were thus adapted to the problem considered, i.e. to detect a miniaturized cavity embedded in a host material with a constant P-wave velocity of 200 m/s (Fig. 6). We selected a filling material composed of coarse limestone gravel with a mean diameter of 0.02 m. In order to simulate
a void in this medium without using masonry, we buried at 0.2 m depth a cylinder made of cellular polystyrene, 0.4 m in diameter and 2 m long. The acquisition system consisted of a series of 32 wide-band accelerometers (0.1–4800 Hz) gathered into a connecting stream and linked to a high-frequency acquisition system. The seismic source was generated by dropping a small (1 cm in diameter) iron ball from a height of 0.2 m on a small stone anvil laid on the surface of the test site. More details on the site and the acquisition system characteristics can be found in Grandjean and Leparoux (2004).

Figure 7 shows an example of a shot for which the low- and high-frequency components have been separated by using band-pass frequency filtering. This separation is efficient enough on this record to determine P-wave first arrivals and diffractions on the high-frequency panel. The same processing as above was applied to the whole data set in order to reject surface waves and retain only volume waves. The first arrivals were manually picked and constituted the input data for the tomographic reconstruction. The initial model was estimated from the direct-arrival slope in the seismic record shot-gathers, and taken as homogeneous. Ten iterations were necessary to converge to a realistic model, explaining 95% of observations.

Figure 8 shows the different results obtained after the tomographic and imaging steps. It can be seen that the velocity field, assumed constant at 200 m/s, is affected by some local anomalies ranging from 130 to 220 m/s. These anomalies were interpreted by Grandjean and Leparoux (2004) as local compaction effects related to the building works. The migrated section shows a series of seismic events for which the strongest one is related to the cylinder. Although this event is similar to the cylinder shape and located at the same position, we note that the final reconstruction is not as clear-cut as for the synthetic acoustic cases. We attribute this problem to the poor quality of the diffractions extracted from the seismic records, as shown in Fig. 7(b). As in the elastic synthetic case, the presence of strong surface waves makes it difficult to recover the diffracted signals that, in addition, are intermingled with the first arrivals, and thus are also reduced to zero during the muting process.

This example, illustrating our P-wave imaging method applied to the subsurface, indicates the difficulties in filtering surface waves. Surface-wave scattering on the topographic roughness could also generate perturbing signals that are difficult to remove. As mentioned by Snieder (1986), the influence of topography on the propagation and scattering of surface waves produces focusing and defocusing effects making them difficult to filter. Nevertheless, Blonk and Hermann (1994) and Blonk et al. (1995) studied this phenomenon and have proposed some solutions for removing such scattered waves.

CONCLUSION
A method of P-wave seismic imaging based on travelt ime tomography and Kirchhoff migration has been developed. This method is tested to determine its efficiency in tackling the
problem of detecting and imaging buried heterogeneities in highly contrasted subsurface media. We implement previously known tomography and migration concepts to develop a two-step imaging technique using algorithms adapted for imaging highly heterogeneous media. This technique consists of processing first arrivals as well as later arrivals, and using them both in tomography and migration processes, sequentially. Tomography is implemented here as an iterative technique based on probabilistic SIRT for reconstructing the background velocity field from the first-arrival traveltimes. In this formulation, Fresnel wavepaths constitute the forward step of travelt ime inversion. Then, from the inverted velocity field, the seismic signals are migrated by a Kirchhoff method implemented in the space domain. In particular, we studied solutions for improving the quality of the seismic imaging of high-contrast media: the use of Fresnel wavepaths instead of classical seismic rays to take into account the seismic resolution; a migration based on Kirchhoff summation in the space domain, which is equivalent to a prestack depth migration; the identification of an adapted filter based on the Karhunen–Loève transform for removing the P-wave signal-to-noise ratio.

The proposed methodology was first applied to a synthetic acoustic case, where both the background velocity field and local impedance contrasts were reconstructed as defined in the predicted model. To examine a more complex situation, a second synthetic case was considered in the elastic approximation. In spite of using adapted filtering for removing them, the surface waves reduce the diffracted signals to a poor quality. Consequently, the reconstructed model shows the anomaly with an incomplete and unfocused shape. Finally, an experimental case specifically designed for the purpose was considered, to test the algorithms under real conditions. The results are in agreement with reality, even if the surface waves are again a problem. This work demonstrated the theoretical suitability of the proposed method. The tests carried out on synthetic and real data sets highlighted the restrictions to be considered, particularly suitable data preprocessing to remove surface waves that could mask P-wave signals.

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