Computing near-surface S-wave velocity models using converted-wave data

Raul Cova and Kris Innanen.
CREWES/University of Calgary

Summary

One important product of the computation of static corrections is a near-surface velocity model. Refracted waves are the data often used for this purpose. However, in the case of PS-wave data, refracted S-waves are usually not available or hard to identify. Here, we propose an approach based on the inversion of τ-differences obtained by crosscorrelating τ-p receiver gathers from different locations. In this approach, the structure of the near-surface at a given location must be know. Results show that the inversion of the depth of the near-surface layer is very unstable. However, inverted velocities and dips displayed more stable behavours which combined provided a better estimation of the velocities in the near-surface.

Introduction

An S-wave velocity model for the near-surface is of utmost importance for imaging and full waveform inversion (FWI) considering elastic media. In the absence of S-wave refracted data, surface waves have been proposed as an alternative for this purpose. Nevertheless, very dense spatial sampling parameters are required to produce reliable and complete S-wave velocity models by using this data.

Here, we propose a method to compute S-wave near-surface velocities using the information captured by the crosscorrelation functions produced during the processing of the near-surface effects. This is achieved by inverting the intercept-time (τ) differences between two receiver stations in terms of the rayparameter (p) values. A two-layered velocity model, and the dip at the base of the near-surface are output by this inversion.

MODELLING TRAVELTIME DIFFERENCES IN τ-p DOMAIN

In a layered medium, the intercept time τ represents the aggregate vertical slowness-thickness product in equation 1 (Diebold and Stoffa, 1981):

$$\tau = \sum_{i=0}^{n-1} \Delta z_i \left(q_i^d + q_i^u \right) = \tau^d + \tau^u, \tag{1}$$

where q_i is the vertical slowness ($q_i = \cos(\theta_i)/v_i$) in the i-th layer and Δz_i is the layer thickness ($\Delta z_i = z_{i+1} - z_i$). The superscripts ^{d} and ^{u} denote the downgoing (P-wave mode) and upgoing (S-wave mode) legs of the raypath in Figure 1, respectively.

To understand the τ-contribution of the S-wave near-surface velocities to the total intercept time we expand the first two terms of the upgoing contribution. After rearranging terms and assuming the measurement surface z_0 is at depth $z_0 = 0$ m we can write,

$$\tau^u = \sum_{i=2}^{n-1} \Delta z_i q_i^u + z_2 q_1^u + z_1 \left(q_0^u - q_1^u \right). \tag{2}$$

The first term in equation 2 provides the total upgoing τ-contribution from the conversion point up to the base of the second layer. The second term represents the contribution from the base of the second layer

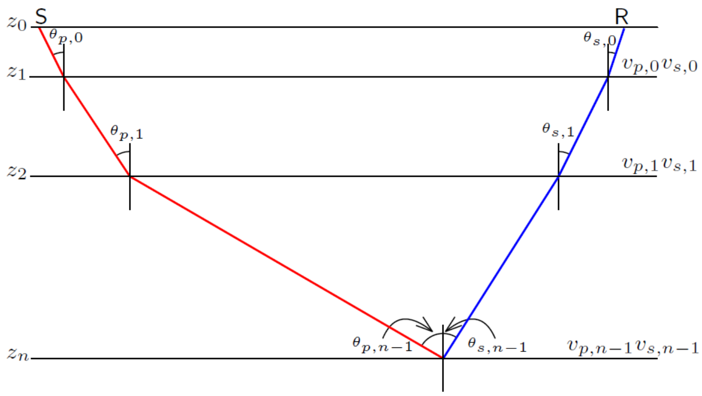

Figure 1. Schematic representation of a PS-raypath. Despite the asymmetry in the raypath the rayparameter *p* is constant in a horizontally layered medium.

up to the surface with velocity v_1 and raypath angle θ_1, as if the layer with velocity v_0 were not present in the model. Therefore, the effect of the near-surface layer with velocity v_0 and thickness z_1 is contained in the last term.

To remove the S-wave traveltime effects of the near-surface we must then apply a correction of the form,

$$\Delta\tau_{NS}^{u} = z_1\left(q_1^u - q_0^u\right). \tag{3}$$

Notice that in equation 3, the τ-contribution of the near-surface layer (z_1q_0) is removed and replaced by the τ-contribution obtained with the rayparameter controlling the propagation in medium 1. Since the correction is done in terms of rayparameter values, not only the velocities, but also the propagation angles in each medium are considered in the correction. Therefore, raypath consistency can be achieved by using equation 3 to remove near surface effects.

Dipping near-surface layer

For dipping interfaces equation 1 still holds (Diebold and Stoffa, 1981). However, the rayparameter value p is no longer constant, and its computation requires us to consider the dip at the base of the near-surface.

Assuming that near-surface effects on the P-wave leg have been removed and that the rest of the interfaces are flat, the vertical slowness in the near-surface layer can be computed as,

$$q_0^u = q_a \cos\phi - p_a \sin\phi. \tag{4}$$

Where the apparent horizontal slowness (p_a) along the base of the near-surface layer with dip angle ϕ is given by

$$p_a = p\cos\phi - q_1 \sin\phi, \tag{5}$$

and the apparent vertical slowness (q_a) is,

$$q_a = \sqrt{s_0^2 - p_a^2}, \tag{6}$$

where s_0 is the total slowness (s_0=1/v_0). Then, Equation 4 can be used to compute the vertical slownesses needed to obtain the near-surface correction in equation 3, when the base of the near surface is dipping.

τ-Differences

The interferometric processing of near-surface corrections proposed by Henley (2012) and extended by Cova et al. (2014) to the τ-p domain relies on the crosscorrelation between input traces and a set of pilot traces. Therefore, the τ-differences captured by the crosscorrelation functions represent the subtraction of the input traces traveltimes and the reference (pilot) traces. Using equation 1 and assuming that the downgoing τ-contributions are the same, this difference can be written as:

$$\Delta\tau_{xcorr} = \tau^{u} - \tau^{u}_{ref}. \tag{7}$$

Using equation 2 and assuming that only the depth and vertical slowness of the near-surface layer have changed we can write,

$$\Delta\tau_{xcorr} = z_1\left(q_0 - q_1\right) - z_{1,ref}\left(q_{0,ref} - q_1\right). \tag{8}$$

Therefore, the τ difference captured by the crosscorrelation operation contains the nearsurface correction at the reference conditions and the current location,

$$\Delta\tau_{xcorr} = \Delta\tau_{NS,ref} - \Delta\tau_{NS} \tag{9}$$

If the reference conditions at a receiver location are known then the near-surface effects can be computed from the τ-differences captured by the crosscorrelation operation.

To invert for the near-surface parameters we tried a quasi-Newton inversion approach. The objective function we used was the L_2 norm of the data misfit (δd). This objective function is minimized by applying iterative updates to the model parameters. These updates have the form,

$$\delta\mathrm{m} = \left[\mathrm{J(m)}^{\dagger}\mathrm{J(m)} + \mu^2\mathrm{I}\right]^{-1}\mathrm{J(m)}^{\dagger}\delta\mathrm{d}. \tag{10}$$

where, μ is a regularization weight, I is the identity matrix, ($\dagger$) denotes the transpose operator and $\mathrm{J(m)}$ is the Jacobian or sensitivity matrix $\mathrm{J(m)} = [\partial\mathrm{g(m)}/\partial\mathrm{m}]$. In our inversion problem there are four parameters we need to solve for (z, s_0, s_1, ϕ). Using the formulae developed in the previous section we derived each one of the derivatives in the sensitivity matrix.

Raytraced data tests

To test the inversion, we computed synthetic traces via raytracing using the velocity model in Figure 2a. Notice that no P-wave velocity contrast exists between the near-surface layer and the medium beneath to simulate the data as if P-wave statics were already removed. Also, only PS-traveltimes were modelled during the raytracing, no amplitude variations were included.

Figure 2b shows the crosscorrelation between the τ-p gather obtained at the reference location A (x = 1800m) and the current location B (x = 2020m). Notice how the modelled differences using equation 9 and the picked maximum of the crosscorrelation traces match very well.

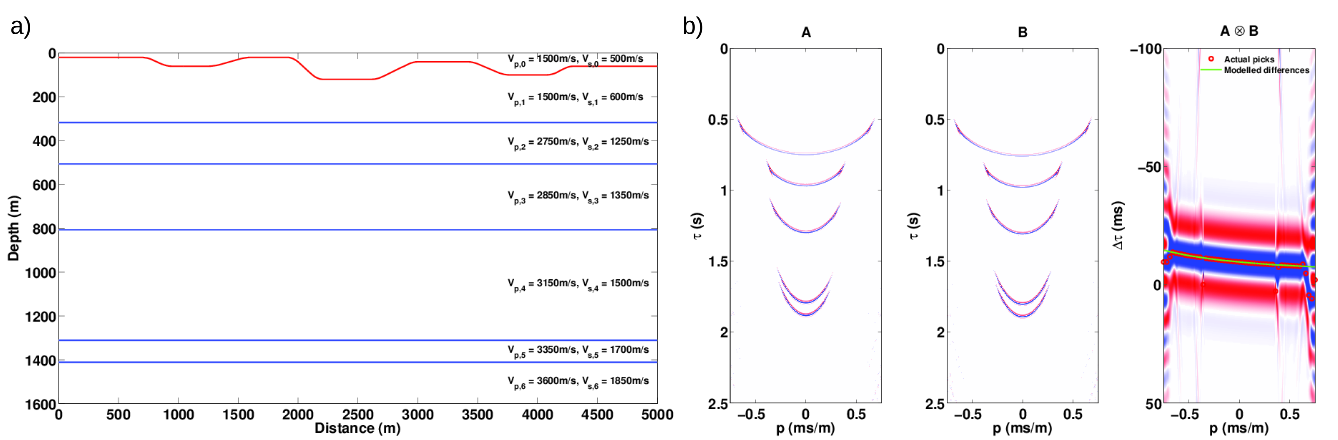

Figure 2. a) Velocity model used to compute synthetic converted-wave traces via ray-tracing. b) Crosscorrelation between reference τ-p gather (A) and τ-p data at receiver location x=2020m (B). Modelled and picked τ-differences are overlaid on the crosscorrelation panel.

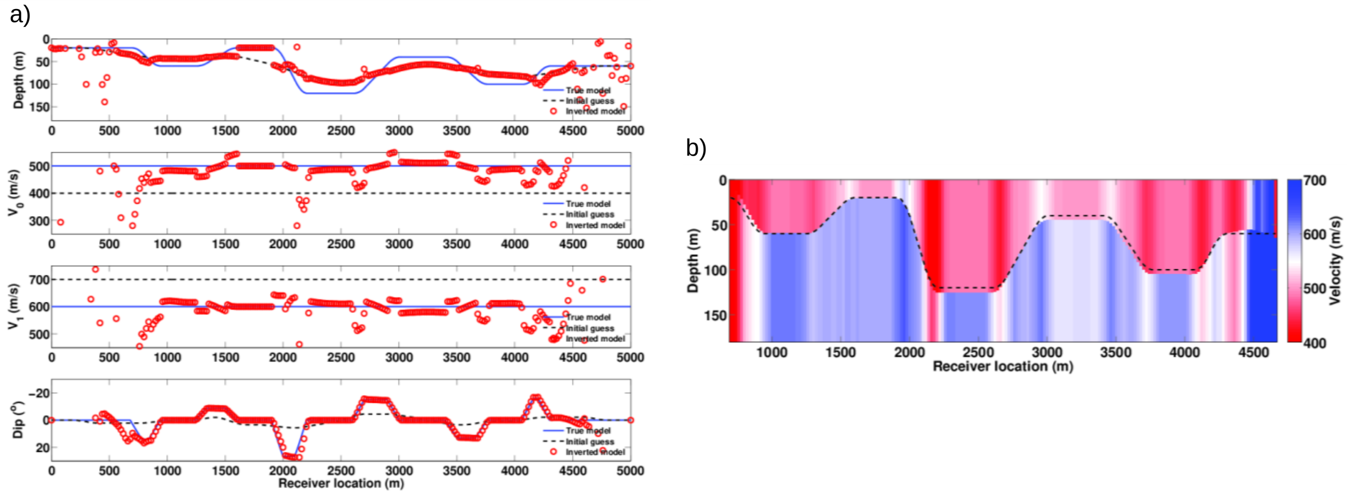

Figure 3. a) Inversion output for all the receiver locations. b) Reconstructed velocity model after inversion. The dashed line represents the actual depth of the near-surface layer. The depth of the velocity changes was reconstructed using the inverted dips.

Figure 3a displays the inversion results obtained for each model parameter at every receiver location. The regularization parameter was set at 0.1, since it provided the most stable results. Inverted depth values for the near-surface layer seems to be very constrained around the initial depth model. RMS average error for this parameter is 17.7 m. However, the inverted velocity values are relatively close to the actual velocities. RMS errors for the velocity of the near-surface and the medium beneath are 68.3 m/s and 97.6 m/s, respectively. On the other hand, the inverted dips display very good and stable results with an RMS error of just 2.1°.

Since the inversion results for the depth of the near-surface were unstable, we reconstructed an image of the velocity model (Figure 3b) using the projection of the reference depth along the inverted dips. The depth of the velocity changes computed in this way, provided values closer to the actual depths with an RMS error of 3.9 m.

Conclusions

The inversion presented in this study provides a method to produce S-wave near-surface velocity models that can be related to receiver static corrections during the processing of PS-data. Our approach requires that the near-surface parameters at one reference location must be known. Therefore, any result provided by the inversion will depend of the accuracy of this information.

The inverted dips displayed by far the most stable results. Since this parameter controls the shape of the data, it is less sensitive to errors in the individual picks. Computing the depth of the near-surface layer by using the inverted dips and the depth at the reference location provided better results that the inverted depths. Based on this observation a different parametrization of the inversion problem using the inverted dips as a constraint might be a good alternative for improving the inversion of the rest of the parameters.

Acknowledgements

The authors thank the sponsors of CREWES for continued support. This work was funded by CREWES industrial sponsors and NSERC (Natural Science and Engineering Research Council of Canada) through the grant CRDPJ 461179-13. We also thank David Henley for his suggestions and assistance with proof-reading.

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