Towards a method for attenuation inversion from reservoir-scale ambient noise OBS recordings

Cornelis Weemstra*, Spectraseis AG, ETH Zürich, Alex Goertz, Spectraseis AG and Lapo Boschi, ETH Zürich

SUMMARY

We analyze ambient seismic and acoustic noise from a broadband passive OBS survey acquired over an oil reservoir in the norwegian North Sea. We analyze the data with respect to azimuthal variation of the incident wavefield in order to find directions of possible dominant sources to evaluate the validity of the assumption of a diffuse wavefield. Whitening of the spectra yields the most clear Green’s functions for higher frequencies and, more important, gives us the complex coherency. Following the approach of Prieto et al. (2009) we are able to determine the phase velocities by fitting Bessel functions to the real part of the complex coherencies. Fitting the distance-dependent coherency to a Bessel function, we are able to estimate the surface-wave slowness of the area; an important strength of this method is that it will ultimately allow us to evaluate the quality factor Q as well.

INTRODUCTION

Passive seismic interferometry involves the cross-correlation of ambient noise recordings. By virtue of this technique, the Green’s function associated with the location of two seismic stations can be measured from the cross-correlation of the continuous ambient signal recorded at the two stations. Surface waves extracted from the ambient seismic wave field via interferometry can be used for velocity inversion (Shapiro and Campillo, 2004; Sabra et al., 2005; Bussat and Kugler, 2009). A fully equipartitioned wavefield is a prerequisite for obtaining perfectly symmetric Green’s functions (Snieder et al., 2007). This means that the energy flux over the array is isotropic. Such a wavefield can be generated by a homogeneous distribution of uncorrelated sources surrounding the array (e.g. Larose et al. (2006) and Wapenaar et al. (2010)). Multiple scattering among heterogeneities in a complex medium also approximates an equipartitioned wavefield (Campillo and Paul, 2003).

The passive seismic data set we use was acquired in April/May 2007. It is recorded over a ∼ 220 km² survey area at an average depth of 360 m, offshore Norway. Figure 1 shows the configuration of the array and the duration of recording at each location. The stations at these locations were not all recording synchronously though. Data was recorded at the 117 seabed locations by 16 ocean-bottom seismometers (OBS). The stations were systematically redeployed at new locations after 1 to 2 days of recording except for two stations that were recording continuously (denoted by black symbols in Figure 1). OBS’s were equipped with a broadband seismometer and a differential pressure gauge (DPG). The instruments have a flat response to particle velocity between 240 s and 50 Hz, and data were acquired with a sampling rate of 125 Hz. The main energy in the data below 5 Hz stems from swell noise, ocean microseisms and Scholte waves, i.e. waves arising at a fluid-solid interface.

Our eventual goal is to derive a 1-D attenuation profile for the area covered by the array. We partly follow the approach of Prieto et al. (2009), who determine the phase velocities for different frequencies by fitting Bessel functions to the real part of the stacked cross-spectra. We apply the same technique to our dataset, working mainly with the DPG component. The initial results are promising for our final objective of inverting for the quality factor Q. We have mainly looked at the DPG components, because the quality of these recordings is shown before by Bussat and Kugler (2009). They have used the DPG component of the same data for ambient-noise surface Wave tomography (ANSWT).

THEORY

Aki (1957) derived that the spatial correlation of the ground motions equals a Bessel function of the first kind with integer order zero, \( J_0 \). We will simply refer to this as ‘Bessel function ’in the rest of this abstract. This is valid for any pair of stations in an equipartitioned wavefield:

\[
\langle u_A(\omega)u_B^*(\omega) \rangle = |F(\omega)|^2 J_0(kr)
\]  

where \( |F(\omega)|^2 \) is the average spectral density of the equipartitioned field. \( u_A \) and \( u_B \) are the Fourier transformed recordings at the two arbitrary stations, in this case A and B. The angular frequency is represented by \( \omega \). \( k \) is the wave number, i.e. \( 2\pi/\lambda \).
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and so depends on velocity and frequency. The brackets ⟨⟩ represent the ensemble average and the asterisk denotes complex conjugation. For vertical components of motion, Yokoi and Margaryan (2008) show that

$$\Re [\gamma_{AB}] = J_0(kr)$$

(2)

where

$$\gamma_{AB}(\omega) = \frac{u_A(\omega)u_B^*(\omega)}{\langle |u_A(\omega)| \rangle \langle |u_B^*(\omega)| \rangle}$$

(3)

is dubbed complex coherency. \(\Re\) means that the right side of equation 2 only equals the real part of the complex coherency. The normalization applied to the cross spectrum \(\langle u_A(\omega)u_B^*(\omega) \rangle\) in equation 3 has the same effect as whitening the data prior to cross-correlation. A form of equation 1 is used by Ekström et al. (2009) on USArray data. For a more thorough explanation and relation to the Green’s function we refer to Prieto et al. (2009). The key message in this short theory section is equation 2. This is the fundament of the method applied and implicates that the real part of the complex coherency is proportional to a Bessel function of the first kind for an equipartitioned wavefield with no intrinsic attenuation, no multiple scattering and no dispersion (Prieto et al., 2009).

APPLICATION TO OBS-RECORDINGS

Based on the work of Prieto et al. (2009), we implement a new algorithm to extract from continuous seismic data the complex coherence described above, and compare it to the Bessel function \(J_0\); importantly, we work at a completely different scalelength (hence seismic frequency range) than Prieto et al. (2009), so that a number of adaptations are needed.

Preprocessing

We whiten our recordings prior to cross-correlation in order to obtain the complex coherence \(\gamma\). Bensen et al. (2007) give a nice overview of the different preprocessing steps commonly used in the seismological community. They advice to whiten the spectra before cross-correlation in order to obtain a broader measurement band and get rid of contamination by (resonance) peaks in the spectra. For our data this means that the higher frequencies are amplified with respect to the lower frequency content of the microseism’s peak. Inspired by a talk of Seats et al. (2010) and the limited amount of data, we have made use of an overlap of 75% of the time windows.

Diffusivity

Figure 2 gives an overview of the temporal change of the incident wavefield for the frequency band between 0.25 and 0.45 Hz. This frequency range corresponds to the first dispersive mode as shown by Bussat and Kugler (2009). Each pair of plots (one station map plus one polar plot) is associated with one recording day. For each recording day, all cross-correlations based on synchronous recordings of more than 4 hours, interstation distances of more than 3.2 km and a signal to noise ratio (SNR) higher than 4 are taken into account. Each dot in the polar plot represents one station couple. The difference between the amplitude of the causal and anti-causal Green’s function is proportional to the distance of the dot from the origin. This gives a measure of asymmetry of the Green’s function for that interstation path. The azimuth of the dot with respect to the origin coincides with the azimuth between the two stations. Essentially, the overall azimuth of the dots gives an idea of the direction where most energy is coming from on that particular day. A fully equipartitioned wavefield would show only dots in the center. Stations couples for which the cross-correlation fulfilled the criteria, are connected by blue lines in the configuration insets. Differences between causal and anti-causal amplitude are normalized with respect to their maximum value, observed on day 6.

The SNR mentioned above is an empirical one. The cross correlations are first normalized such that their maximum values have an amplitude of one. The SNR is defined as the maximum amplitude in the velocity range of interest divided by the standard deviation on the noise windows. Figure 2 shows results for the frequency range 0.25-0.45 Hz for which the velocity range of interest is 350 m/s to 750 m/s. The noise windows are defined as the windows corresponding to higher and lower velocities than the velocity range of interest. A transitional margin outside of the velocity range of interest is employed.
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The width of this margin depends on the frequency range of the cross correlations. Longer periodic cross correlations have longer margins between 'signal window' and 'noise window'.

The dominating back-azimuth of the ambient wavefield clearly changes over time. On day 6 most energy is propagating towards an ENE direction, while on days 7 & 8 the wavefield seems to be more diffuse. Considering the NNW-SSE aligned line of stations recording on day 11, most cross-correlations show a higher amplitude Green's function for North-South traveling energy than for energy propagating from South to North.

**Bessel function fitting**

For an equipartitioned wavefield with no intrinsic attenuation, no multiple scattering and no dispersion (velocity is constant with respect to frequency), the real part of the complex coherency is proportional to a Bessel function in the frequency as well as the interstation distance dimension (equation 2). As these conditions are generally not fulfilled in reality, we try to approximate them with appropriate data processing. Different frequencies travel with different velocities. However, keeping the frequency fixed, the real part of the complex coherency should fit a Bessel function with distance. Averaging over azimuth to approximate the condition of an equipartitioned wavefield is supported by the plots in Figure 2 and is in line with the method of Prieto et al. (2009). We force the cross-correlations to be symmetric by stacking the causal and anti-causal parts. This means that the complex coherency is not complex anymore. This averaging over interstation azimuths is done by dividing the range of interstation distances into bins of 100 meters. For each and every bin, the coherencies of stations separated by a distance within that bin are stacked.

To arrive at these coherencies the following processing sequence is executed:

1. Traces are cut in time-windows of 60 seconds with an overlap of 75%.
2. The time-windows are detrended.
3. Cosine taper of 2.5% of the trace length.
4. Fourier transformation of the traces.
5. Whitening of the amplitude spectra
6. Multiplication of the spectra of the synchronous time-windows, i.e. actual cross-correlation.

Some of the distance bins only contain complex coherency values based on a few hours of synchronous recording of one station couple. On the other hand, other distance bins contain a stacked coherency based on days of synchronous recordings and tens of station couples. This is simply due to the configuration of the array and cannot be overcome. Figure 3 shows the coherency as a function of interstation distance and frequency. To distinguish to a very first order between stable and less stable coherency stacks, we only take into account the bins based on 2 or more interstation paths and more than 4 hours of synchronous recordings.

Variations in phase velocities explain variations in oscillation rates of the coherency with frequency. In particular, below 0.8 Hz, the phase velocities are lower which gives rise to higher oscillation rates of the coherency with frequency. The lower phase velocities and strong non-dispersiveness of the wavefield below 0.8 Hz is shown by Bussat and Kugler (2009). A vertical cross-section of Figure 3 for the interstation distance bin centered around 1450 m and extended to a frequency of 10 Hz is shown in Figure 4. Below approximately 1 Hz, the different modes cause the coherency to be relatively scattered, i.e. there is interference of signals of the same frequency traveling with different velocities.

Bessel functions can be fit to the horizontal cross-sections of Figure 3. Fixing the frequency, we perform a grid-search over a 2-D grid with the scaling of the Bessel function in one dimension and the velocity in the other. The scaling is needed because the coherency values are proportional to a Bessel function and not equal. This means that the maximum of the best fitting Bessel function is a variable. We perform the grid search for velocities from 500 to 5000 m/s with an increment of 2 m/s from one node to the next. The scaling factor is changed from 0 to 1 with an increment of 0.01. For each node in the grid, we calculate the error as the sum of the differences between the

![Figure 3: Real part of the complex coherency for interstation distances 0-10000 meters and 0-4 Hz. The triangles on the color bar indicate that some values are off-scale.](image)

![Figure 4: Stacked coherency versus frequency for stations having an interstation distance between 1400 and 1500 meters. The stack is based on 21 interstation paths and 336180 time windows.](image)
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Bessel function values and the data coherency values, i.e. the L1-norm of the difference vector. The minimum L1-norm in the grid corresponds to the velocity and scaling factor of the best fitting Bessel function. We use the L1-norm to mitigate the effect of outliers due to limited amount of data in some distance bins.

Figure 5: Best fitting $J_0$ to the real parts of the coherency for a frequency of 0.325 Hz (top graph) and 1.60 Hz (bottom graph). Note the difference in scale along the vertical axis.

The grid search is performed for all frequencies between 0 and 2 Hz. The best fitting Bessel function (green line) is plotted on top of the coherency values and shown for two frequencies in Figure 5. The frequency, velocity, scaling factor and L1-norm are designated in the top right corner of the graphs. The real part of the complex coherency decays faster with interstation distance than the best fitting Bessel function which suggests the effect of intrinsic and scattering attenuation. This is to be expected as the derivation by Aki (1957) does not take into account intrinsic and scattering attenuation of the wavefield. A search over a range of Q-factors will be done to arrive at a 1-D attenuation profile for the area covered by the array.

If the coherency values are very close to zero, the fit to the Bessel function is of course relatively poor. Nevertheless, because we vary the scaling factors, the L1-norm corresponding to the best fitting Bessel function is very low and it seems like a good fit if we were only to consider the value of the error. To correct for this we multiply each of these misfits by the inverse of the scale factor corresponding to the best fitting Bessel function. We show in Figure 6 the misfit as a function of velocity (and thus slowness) and frequency. Along the vertical axis, the best fitting Bessel functions are centered in the blue throughs. The similarity of this contour plot with the pf-spectrum in Bus-sat and Kugler (2009) is striking. In fact, Figure 6 is also a dispersion plot, but with amplitude being the L1-norm of the difference between the coherency and the Bessel function.

Figure 6: Error normalized for the scale factor as a function of frequency and slowness.

CONCLUSIONS

Despite a very limited acquisition geometry and a not well equipartitioned wave field, we are able to extract stable green’s functions from this dataset. The change of the wavefield over time supports averaging over interstation azimuth by binning the coherencies by interstation distance. Whitening of the spectrum prior to cross-correlation and binning by interstation distance yields a very reasonable fit of the Bessel functions to the real part of the complex coherencies. Phase velocities can be obtained by fitting the Bessel function to the real part of the complex coherency by minimizing the L1-norm of the differences. The real part of the complex coherency decays faster with interstation distance than the best fitting Bessel function which suggests the effect of intrinsic and scattering attenuation.

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