Chapter 3

Crosscorrelation of microseism noise

Passive seismic interferometry turns Valhall’s ocean-bottom cable geophones into virtual seismic sources by crosscorrelating microseism noise recorded between 0.175 – 1.75 Hz. Crosscorrelating all component combinations between the 3-component geophones of two stations retrieves a full virtual-seismic source matrix. The vertical-to-vertical and radial-to-radial elements are dominated by fundamental-mode Scholte waves, while the tangential-to-tangential element is dominated by fundamental-mode Love waves. A first overtone Scholte-wave mode is clearly retrieved and a first overtone Love-wave mode is fainter. Crosscorrelations converge to a long-term average more rapidly for shorter interstation offsets and lower frequencies. As little as one day of recording time may be sufficient to retrieve fundamental-mode Scholte waves at frequencies below 1 Hz at offsets smaller than 6 km. This provides the opportunity to survey the subsurface continuously using noise recordings with wave modes at frequencies not usually present in recordings made during controlled-source seismic surveying.
INTRODUCTION

In Chapter 2, I characterized the ambient seismic field recorded by Valhall’s Life of Field Seismic (LoFS) array. The ambient seismic field recordings are referred to as noise in the controlled-source community, but passive seismic interferometry (PSI) utilizes these recordings as signals. Although exploration-scale PSI is unlikely to retrieve gathers comparable to those from controlled-source seismic experiments, PSI has several advantages. Sources can be created at receiver stations directly adjacent to the platform where placing controlled sources may be undesirable. Moreover, virtual-seismic sources can be created using ambient noise recorded at any time, while a controlled-source survey is conducted typically once a year. Additionally ambient noise recorded by ocean-bottom stations contains wave-modes, such as Scholte and Love waves below 2 Hz (Stewart, 2006; Dellinger, 2008), at frequencies that are not illuminated by controlled-source seismic (Olofsson, 2010). These studies have successfully retrieved Scholte waves in a marine setting but have not analyzed the horizontal components for Love waves. Scholte waves from controlled-source data at Valhall are recorded only above 2 Hz (Hatchell et al., 2009), and Love waves are poorly (or even not at all) excited by airguns in the water column.

Studies of correlation convergence are scarce, and they lack empirical validation or neglect leading factors (such as frequency and interstation distance) in convergence rate analysis. Larose et al. (2008) used the Signal-to-Noise Ratio (SNR) based on maximum amplitude over background fluctuation level to create a model for a highly scattering medium. Weaver and Lobkis (2005) and Sabra et al. (2005c) used the variance of the crosscorrelation signals as a measure of convergence. Sabra et al. (2005c) neglected interstation offsets and frequency dependency but found a relationship for the variance of the correlation signal that is inversely proportional to the time bandwidth product. However, Weaver and Lobkis (2005) found that the amount of recording time needed to detect certain arrivals scales with the square of frequency and the square of interstation offsets. The SNR does not fully capture the convergence of crosscorrelations to a stable long-term averaged correlation signal (which may or may not estimate the Green’s function well). I quantify convergence
rate by comparing two crosscorrelation signals with a correlation coefficient following
the methods of Seats et al. (2012).

Seismic interferometry is employed to create virtual-seismic source matrices at
each receiver of Valhall’s LoFS array. Computing the correlation-coefficient between
two crosscorrelation signals provides a measure of similarity and is employed to de-
terminate if the crosscorrelations stabilized. Crosscorrelating different components at
different stations retrieves a full virtual seismic Green’s matrix that is interpreted in
terms of Scholte- and Love-wave modes. Although this chapter I present primarily
crosscorrelations of the 2010 recording, all four data sets introduced in Chapter 2
have been processed in a similar fashion into virtual-seismic surveys.

VIRTUAL SEISMIC SOURCES FROM PASSIVE SEISMIC
INTERFEROMETRY

The LoFS array has stations that consist of one hydrophone, one in-line (with the ca-
ble) geophone and two perpendicular (to the cable) geophones. The geophones record
particle velocity one meter below the sea floor. Station orientations (estimated from
controlled-source seismic data) were used to rotate the data to vertical, east, and north
components using direction cosines. Crosscorrelations between the recordings of par-
ticle velocity made by all components at all stations were computed. The derivation
for the seismic interferometry result of crosscorrelating recordings of particle veloc-
ities is included in Appendix A. Let $G_{i,j}^{v,f}(x_A, x_B, \omega)$ denote the frequency-domain
elastodynamic Green’s function of a receiver recording the i-th component of particle
velocity, $v_i$, at $x_A$ due to an external force density in the j-th direction, $f_j$, at $x_B$. An
estimate of the Green’s function, $G_{i,j}^{v,f}(x_A, x_B, \omega)$, and its reciprocal, $G_{j,i}^{v,f}(x_B, x_A, \omega)$,
can be retrieved as follows:

$$\left\langle v_i(x_A) v_j^*(x_B) \right\rangle \approx \left\{ G_{i,j}^{v,f}(x_A, x_B, \omega) + G_{j,i}^{v,f*}(x_B, x_A, \omega) \right\} S(\omega) \quad (3.1)$$
where the crosscorrelated signals $v_i(x_A, \omega)$ and $v_i(x_B, \omega)$ denote the particle velocity recordings made at $x_A$ and $x_B$ (the master station), respectively. Complex conjugation is denoted by $^*$, and $\langle \rangle$ denotes a spatial ensemble average. The power spectrum of the noise source signals is denoted by $S(\omega)$. The crosscorrelation signal approaches an equivalence of the superposition of the causal Green’s function and its anti-causal reciprocal counterpart, forming a purely symmetric signal. Thus, the crosscorrelation signal is referred to as an Estimated Green’s Function (EGF). An estimate for the phase of the Green’s function can be found by applying the Heaviside step function to the crosscorrelation signal either before or after symmetrizing. Repeating this procedure for each component at each station in the array with each component at a master station yields an estimated Green’s matrix (EGM) for each station pair, collectively called a virtual seismic survey.

VIRTUAL SEISMIC SOURCES IN VALHALL’S LOFS ARRAY

The processing sequence of noise correlations exists in many flavors. Following theoretical derivations for Green’s function retrieval, a simple cross spectra is deemed sufficient to retrieve an EGF colored by the power spectrum of the noise sources. I opted to compute straightforward cross spectra that, if needed, could still be whitened before stacking. See Appendix B for how to compute cross spectra, and their appropriate unit.

To extract the microseism noise and compress the data volume, the recorded pressure data were first filtered using a frequency-domain taper with a flat response for $0.2 - 1.5$ Hz and a Hann taper extending from $0.175$ to $1.75$ Hz. Filtering was done in 30-minute segments with 50% overlap and application of a time-domain Hann taper. This resulted in 116, 25, 7 and 486 segments for the 2004, 2005, 2008 and 2010 recordings, respectively. Traces that contained noise bursts, spikes or other data irregularities were detected and discarded. All available data were crosscorrelated for each segment. For each channel pair, the crosscorrelations from each segment
were stacked and normalized (averaged) to form a virtual seismic survey with EGMs between all stations in Valhall’s LoFS array.

Figure 3.1 shows an example of a virtual seismic source in vertical particle velocity located near the center of the array. This example was generated using all the data in the 2010 recording. Each frame is a snapshot corresponding to a certain correlation-time lag ($\tau$). Negative time lags correspond to the anti-causal EGF, while positive time lags correspond to the causal EGF. There is good retrieval of both causal and anti-causal EGFs without imposing symmetry conditions, resulting in an antisymmetric component that is approximately only 20% of the magnitude of the symmetric component. This is due to the azimuthal homogeneity of the directions in the ambient seismic field that is observed in the beam steering results in Chapter 2. The antisymmetric component changes polarity between stations south and north of the virtual source (apparent in the antisymmetric slice at $\tau = 5$ s).

**CROSSCORRELATION CONVERGENCE RATE**

The ensemble average in Equation 3.1 is evaluated by crosscorrelating half-hour long recordings and averaging them over (overlapping) windows. One advantage is that the cross spectra can be normalized or deconvolved prior to averaging. Because the spectrum of microseism at Valhall is relatively stable over time, no deconvolutionary division is applied to the cross spectra prior to stacking. Figure 3.2 shows a virtual seismic source gather (vertical-to-vertical) for six stack lengths of four cable lines for a virtual source at the first station. Correlating only half an hour of recording results in a correlation signal with weak signal and strong noise, this noise is named background-correlation fluctuations. Averaging half-hour crosscorrelations together makes the background-correlation fluctuation level decrease, while the coherent signals intensifies.

For a particular station couple separated by 471 m, we gather all the crosscorrelation signals from half-hour overlapping windows (Figure 3.3). Although we can observe a wave train in causal and anti-causal portions of the correlation signal for all
Figure 3.1: Symmetric and antisymmetric components of a virtual seismic source (vertical-to-vertical) at the center of the array generated by processing and stacking all of the 2010 recordings. The bottom row contains slices for the symmetric part at correlation-time lags $\tau = 0 \text{ s}$, $\tau = 5 \text{ s}$, $\tau = 10 \text{ s}$ and $\tau = 15 \text{ s}$. The bottom row contains slices for the antisymmetric part at correlation-time lags $\tau = 0 \text{ s}$, $\tau = 5 \text{ s}$, $\tau = 10 \text{ s}$ and $\tau = 15 \text{ s}$. [CR] [movie-v-snaps]
Figure 3.2: Virtual seismic source gather (vertical-to-vertical) for virtual source at the first station. a) crosscorrelation signal from 30-minute window, b) 2-hour stack, c) 12-hour stack, d) 1-day stack, e) 2-day stack, f) 5-day stack. [CR] shot-convergence
crosscorrelations, during certain times the background correlation fluctuation level is much higher. Stacking these traces will diminish background correlation fluctuations and result in a higher quality EGF.

Figure 3.3: Crosscorrelations of 30-minute recordings between vertical components of two stations separated by 471 m, shown as a function of the central time of the 30-minute recording window. [CR] [Joseph-Z-ests]

We would like to compare how far the stack has converged towards a stable result. Stacking crosscorrelations beyond the available recording time will change the stack. A similarity measure between two crosscorrelation signals is the correlation-coefficient between them. When two signals are equal, the correlation coefficient is 1, when two signals are uncorrelated (or perpendicular), the correlation coefficient is 0 and when two signals are equal (but have opposite sign), the correlation coefficient is $-1$. Figure 3.4 shows crosscorrelation stacks for a station couple separated by 471 m. In Figure 3.4, the correlation coefficient between (a) and (d) is 0.30, between (b) and (d) is 0.87, and between (c) and (d) is 0.94.

This comparison of two crosscorrelation signals can be performed for various normalized-stack lengths and averaged over station pairs with similar offsets (Figure 3.5). The crosscorrelation signals are first bandpass filtered in the frequency
domain with six different Hann tapers as follows: 0.25 – 0.50 Hz, 0.50 – 0.75 Hz, 0.75 – 1.00 Hz, 1.00 – 1.25 Hz, 1.25 – 1.50 Hz and 1.50 – 1.75 Hz (Figures 3.5a-f). If the convergence rate were linear with stacking time, we expect the 0.95 contour to lie at a stack of 0.95×(5 days+1.5 hours) = 4 days+19 hours+25.5 minutes long, and the 0.5 contour to lie at a stack of 0.5×(5 days+1.5 hours) = 2 days+12 hours+45 minutes long. The 0.95 contour line of the correlation coefficient is reached at a shorter stack length then at stacking 95% of the data. The 0.95 contour line of the correlation coefficient indicates that stacking more crosscorrelated recording time is necessary with larger offsets. Less stack is needed for lower frequencies than for higher frequencies, and the crosscorrelations more rapidly converge. In the discussion section I will formulate some rules of thumb.
CHAPTER 3. CROSSCORRELATION OF MICROSEISM NOISE

Figure 3.5: Correlation coefficient between full and partial stacks of crosscorrelations after bandpass filtering averaged as a function of offset and time corresponding to partial stack length. Bandpass filtered for: a) 0.25 – 0.50 Hz, b) 0.50 – 0.75 Hz, c) 0.75 – 1.00 Hz, d) 1.00 – 1.25 Hz, b) 1.25 – 1.50 Hz, and f) 1.50 – 1.75 Hz. The black vertical dotted line indicates a stack length of 50% of the data, and the black vertical dashed line indicates a stack length of 95% of the data. Red dotted line indicates a contour line of 0.5 and the red dashed line indicates a contour line of 0.95. [CR] corrcoef-ranges
APPORXIMATE SEPARATION OF LOVE AND
SCHOLTE WAVES BY COORDINATE
TRANSFORMATION

A coordinate system based on north, east and vertical components is an unnatural way to study the EGMs. In a perfectly stratified media the interface wave modes split into Love and Rayleigh-Scholte wave modes. Love waves have transverse polarization and only appear in the transverse components. Scholte waves appear in the vertical and radial components. Although the subsurface at Valhall is anisotropic and there are strong lateral inhomogeneities (Barkved and Kristiansen, 2005; Sirgue et al., 2010), these effects are secondary and the wave fields generated by the virtual seismic sources will, to first order, consist of Love- and Scholte-wave modes.

I exploit this first order behavior and rotate each virtual seismic source to a cylindrical coordinate system centered around the source. Thus, all virtual seismic source matrices were transformed from a coordinate system with north, east and vertical components to a cylindrical coordinate system centered at the source station with radial, tangential, and vertical components (Equation 3.2). In Figure 3.6, time slices (at $\tau = 5$ s) of the lower-left triangular set are the EGM elements ($G_{en}$, $G_{ee}$, $G_{ve}$ and $G_{vv}$) before rotation in a north, east and vertical coordinate system. The upper-right triangular set are time slices (at $\tau = 5$ s) of the EGM elements ($G_{rr}$, $G_{rt}$, $G_{rv}$, $G_{tt}$, $G_{tv}$ and $G_{vv}$) in a radial, tangential, and vertical coordinate system. The color scale varies per element. Now, the polarity and amplitude of the EGFs should no longer depend on geographic direction. Directionality of the virtual seismic sources can still be caused by the directionality of the energy in the ambient seismic field and by subsurface lateral inhomogeneities. The EGMs are rotated according to

$$G_{rtv} = M G_{nev} M^T,$$

(3.2)
CHAPTER 3. CROSSCORRELATION OF MICROSEISM NOISE

where

\[
M(x_r, x_s) = \begin{pmatrix}
\cos(\alpha) & \sin(\alpha) & 0 \\
-\sin(\alpha) & \cos(\alpha) & 0 \\
0 & 0 & 1
\end{pmatrix},
\]

\[G_{nev}(x_r, x_s) = \begin{pmatrix}
G_{nn}(x_r, x_s) & G_{ne}(x_r, x_s) & G_{nv}(x_r, x_s) \\
G_{en}(x_r, x_s) & G_{ee}(x_r, x_s) & G_{ev}(x_r, x_s) \\
G_{vn}(x_r, x_s) & G_{ve}(x_r, x_s) & G_{vv}(x_r, x_s)
\end{pmatrix}, \text{ and } \]

\[G_{rte}(x_r, x_s) = \begin{pmatrix}
G_{rr}(x_r, x_s) & G_{rt}(x_r, x_s) & G_{rv}(x_r, x_s) \\
G_{tr}(x_r, x_s) & G_{tt}(x_r, x_s) & G_{tv}(x_r, x_s) \\
G_{vr}(x_r, x_s) & G_{vt}(x_r, x_s) & G_{vv}(x_r, x_s)
\end{pmatrix}, \]

where alpha (\(\alpha\)) is the angle measured clockwise between north and a line connecting source to receiver.

The polarity of the observed wave front changes with azimuth for the various elements of the EGF matrix (Figure 3.6). In the \(G_{vv}, G_{nn}\) and \(G_{ee}\) elements the polarity is equal in all azimuths from the source, but the \(G_{nn}\) and \(G_{ee}\) elements, are relatively weak in the east-west and north-south directions from the source, respectively. The polarity flips in the north-south and east-west directions in the \(G_{en}\) and \(G_{ve}\) elements, respectively. In addition, \(G_{en}\) displays a butterfly pattern of flipped polarities and weak radiation. The \(G_{vv}\) element remains the same after rotation. The \(G_{tt}, G_{rr}\) and \(G_{rv}\) elements are now radiating with equal polarity and nearly equally strong in all directions from the source. The \(G_{vv}, G_{vr}\) and \(G_{rr}\) elements are dominated by Scholte waves, and the \(G_{tt}\) element is dominated by Love waves. The \(G_{rt}\) and \(G_{tv}\), which are expected to be zero in a perfectly stratified medium, display a complicated mixture of converted wave modes. The ambient noise field is actually stronger in the vertical component than in the horizontal components, so the amplitudes in the \(G_{rt}\) \(G_{tt}\) and \(G_{tt}\) elements are relatively weak, but the amplitude in the \(G_{vv}\) and \(G_{tv}\) elements are relatively strong.
Figure 3.6: Display of the action of coordinate-system transformation on the elements of the EGF matrices for time slices at correlation-time lag $\tau = 5$ s. The lower-left triangular set are the EGM elements ($G_{vn}$, $G_{ee}$, $G_{ve}$ and $G_{vv}$) before rotation to a north, east and vertical coordinate system. The upper-right triangular set are the EGM elements ($G_{rr}$, $G_{rt}$, $G_{rv}$, $G_{tt}$, $G_{tv}$ and $G_{vv}$) in a radial, tangential and vertical coordinate system. The color scale varies per element. [CR] sourcematrix
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DISPERSION ANALYSIS OF VIRTUAL SEISMIC SOURCES

Figure 3.7: Map of station locations in Valhall’s LoFS array. Each black dot denotes a station. The common midpoints for the dispersion analysis in Figure 3.8 are denoted by a blue square.

After transforming the virtual seismic survey to a midpoint and radial-offset domain, all offsets for a group of common midpoints in a patch in the middle part of the array were selected, UTM (525 – 526, 6236 – 6237) km. This patch, denoted by a blue square in Figure 3.7, was chosen for its good midpoint and offset coverage. Dispersion images (Figure 3.8) are calculated as the amplitude in the Radon \((\omega - p)\) domain balanced over frequencies. Relative amplitudes between panels are not preserved. Fundamental-mode Scholte waves dominate in the vertical-vertical, radial-radial and radial-vertical elements. The first overtone is also distinguishable and is more evident in the radial-radial than in the vertical-vertical element. The transverse-transverse element is dominated by Love waves that travel at a higher velocity (lower slowness) than the fundamental-mode Scholte waves. The transverse-vertical and transverse-radial components are much weaker than the other elements (before normalization) and do not contain well-defined modes.

The measured phase velocities for the 1st- and 2nd-Scholte and Love wave modes in the dispersion images in Figure 3.8 are shown with their associated wavelengths in
CHAPTER 3. CROSSCORRELATION OF MICROSEISM NOISE

Table 3.1: Measurements of phase velocity and wavelength as a function of frequency by picking maximums for the Love and 1st- and 2nd-Scholte wave modes in Figure 3.8.

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>0.4</th>
<th>0.5</th>
<th>0.6</th>
<th>0.7</th>
<th>0.8</th>
<th>0.9</th>
<th>1.0</th>
<th>1.1</th>
<th>1.2</th>
<th>1.3</th>
<th>1.4</th>
<th>1.5</th>
<th>1.6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scholte 1st c (m/s)</td>
<td>641.0</td>
<td>581.4</td>
<td>536.5</td>
<td>502.0</td>
<td>478.9</td>
<td>459.6</td>
<td>444.8</td>
<td>431.1</td>
<td>419.5</td>
<td>408.5</td>
<td>399.4</td>
<td>390.6</td>
<td>383.4</td>
</tr>
<tr>
<td>Scholte 2nd c (m/s)</td>
<td></td>
<td></td>
<td>838.9</td>
<td>766.9</td>
<td>726.7</td>
<td>698.3</td>
<td>668.4</td>
<td>641.0</td>
<td>615.8</td>
<td>592.4</td>
<td>578.7</td>
<td>563.1</td>
<td>555.6</td>
</tr>
<tr>
<td>Love 1st c (m/s)</td>
<td>586.9</td>
<td>558.0</td>
<td>534.2</td>
<td>518.7</td>
<td>502.0</td>
<td>486.4</td>
<td>473.5</td>
<td>466.4</td>
<td>459.6</td>
<td>456.2</td>
<td>451.3</td>
<td>443.3</td>
<td>440.1</td>
</tr>
<tr>
<td>Scholte 1st λ (m)</td>
<td>1662.5</td>
<td>1162.8</td>
<td>894.2</td>
<td>717.1</td>
<td>598.6</td>
<td>540.7</td>
<td>444.8</td>
<td>391.9</td>
<td>349.6</td>
<td>311.2</td>
<td>285.3</td>
<td>260.4</td>
<td>239.6</td>
</tr>
<tr>
<td>Scholte 2nd λ (m)</td>
<td></td>
<td></td>
<td>1398.2</td>
<td>1095.6</td>
<td>908.4</td>
<td>775.9</td>
<td>668.4</td>
<td>582.7</td>
<td>513.2</td>
<td>455.7</td>
<td>413.4</td>
<td>375.4</td>
<td>347.3</td>
</tr>
<tr>
<td>Love 1st λ (m)</td>
<td>1467.2</td>
<td>1116.0</td>
<td>890.3</td>
<td>741.0</td>
<td>627.5</td>
<td>540.4</td>
<td>473.5</td>
<td>424.0</td>
<td>384.0</td>
<td>359.9</td>
<td>322.4</td>
<td>295.5</td>
<td>275.1</td>
</tr>
</tbody>
</table>

Table 3.1. The Scholte wave fundamental tone travels at 640 m/s at 0.4 Hz to 380 m/s at 1.6 Hz. The higher-mode Scholte wave travels at 840 m/s at 0.6 Hz to 550 m/s at 1.6 Hz. The fundamental-mode Love wave travels at 590 m/s at 0.4 Hz to 440 m/s at 1.6 Hz. The higher-mode Scholte waves are always faster than the fundamental-mode Scholte waves. The fundamental-mode Scholte wave is faster than the fundamental-mode Love waves below 0.6 Hz, the Love waves travel faster than the fundamental Scholte waves above 0.6 Hz. The fundamental-mode Love waves are not well retrieved above 1.0 Hz.

None of the waves are spatially aliased at microseism frequencies for the in-line direction. For the cross-line direction, all three interface wave modes become aliased above a certain frequency for the cross-line direction: The fundamental Scholte wave above 0.9 Hz, the Scholte wave overtone mode above 1.2 Hz and the Love wave mode above 0.95 Hz. This corroborates the beam steering results (Chapter 2) that show the fundamental Scholte wave aliases above 0.95 Hz.

DISCUSSION

Passive seismic interferometry successfully turned microseism ambient seismic energy between 0.175 – 1.75 Hz into virtual seismic sources emitting interface waves. The gathers do not show an artifact from the platform energy, thus confirming our frequency analysis as a function of space (Chapter 2) that predicted that we could select a frequency range where the microseism noise is dominant in all of the stations of
Figure 3.8: Dispersion images for all elements of the virtual seismic source matrix. Images are arranged according to elements in Equation 3.5. Relative amplitudes between images are not preserved. The top- and bottom-left and bottom-right elements are expected to be dominated by Scholte waves. The center element is expected to be dominated by Love waves. The center-left and center-bottom elements are expected to be zero in a perfectly layered and isotropic medium. Fundamental and first overtone Scholte and Love waves are indicated by $S_0$, $S_1$ and $L_0$, $L_1$, respectively. [CR] dispersionmatrix
the LoFS array. The crosscorrelation signals were computed (after bandpassing) by a straightforward cross spectrum without amplitude balancing (in neither time nor frequency domain) or imposing symmetry conditions based on reciprocity. The microseism spectrum does not vary much over the array (Chapter 2) and the spectrum is relatively flat over the microseism bandwidth. Applying frequency-domain normalization is not needed, and I simply stack the crosscorrelations from overlapping time windows. This causes minimal spectral smoothing because the time-windows were chosen to be large. However, I neglect the transient nature of the power spectra of the noise sources. When the amplitudes of the EGFs are of interest, a method minimizing dependence on the transient nature of the ambient seismic field is important (Prieto et al., 2011). In later chapters I interpret only the kinematics of the EGFs.

The result after stacking all the 2010 data is remarkably symmetric. The antisymmetric component is approximately 20% of the magnitude of the symmetric component. This is consistent with the observation in Chapter 2 that the propagation direction of the microseism noise is generally uniformly distributed over azimuth. The antisymmetric component changes polarity between stations south and north of the virtual source (Figure 3.1). This reflects that any directionality of ambient energy persists throughout the field. When the orientations of source and receiver stations are interchanged, so are the causal and anticausal energy strengths. Consequently, the polarity of the antisymmetric component of the EGF changes.

The crosscorrelation signal contains a dominant arrival package and background correlation fluctuations. The crosscorrelations from recordings at UTC 0:00:00 on December 26th, 2010, exhibit particularly high background correlation fluctuations (Figure 3.3). This corresponds to relatively directional ambient seismic noise. Whether this relationship holds for other periods and is causal is beyond the analysis presented in this study.

By the law of large numbers, the stack of crosscorrelations will converge to the long-term average. However, the rate of convergence varies based on frequency and offset. There is no guarantee that the long-term average of crosscorrelations is equal to the exact Green’s function (Chapter 1). In fact, only those frequencies and wave
numbers present in the noise could be retrieved in virtual seismic sources. There is no guarantee that the long term average does not result in a non-uniform radiation pattern (e.g., EGFs with a consistent antisymmetric component).

For frequencies between $0.75 - 1.00$ Hz and offsets smaller than 6 km, stacking as little as 2 days of data equals within 5% a stack of 5 days of data (Figure 3.5). For frequencies between $1.50 - 1.75$ Hz and offsets smaller than 6 km, stacking as much as 4 days of data equals within 5% a stack of 5 days of data. For the higher frequency range, the 0.95 contour level lies close to the entire length of the dataset. The correlations did not stabilize yet, and more than 5 days of data is needed. Because the LoFS array is very dense, the phase and group velocities in certain regions of the subsurface may be sufficiently over-determined for imaging by noisy crosscorrelations, and less recording time would be sufficient at higher frequencies.

The sinusoidal nature of the 0.95 contour level at small interstation distances (Figures 3.5c-f) may be due to dispersion effects. The amplitude of the Green’s function is weaker for particular offsets such that the background correlation fluctuations are relatively strong, and thus the correlation-coefficient is smaller. Lower frequencies stabilize faster than higher frequencies. This is readily explained because the Fresnel zone is wider for lower frequencies than for higher frequencies. Fewer sources properly evaluate the ensemble average of crosscorrelations of responses of sources surrounding the station pair than at higher frequencies. These observations are consistent with Weaver and Lobkis (2005). This is apparent in Figures 3.5a and f. I interpret the 0.95 contour-line to be a measure of having converged when this correlation coefficient of 0.95 is achieved with stacking significantly less than the full 5 days of recording.

The 2010 recording is insufficiently long for the 5 day stack at the higher frequencies and longer offsets to have converged to its long-term average. But the following rules of thumb can be observed from Figure 3.5b to f:

- For a given desired crosscorrelation quality the required recording stack-length increases linearly with distance.

- For a given desired crosscorrelation quality about twice as much recording length
is needed at 1.125 Hz than at 0.625 Hz.

- At 0.625 Hz, after stacking 2.5 days, the crosscorrelations have converged for the entire offset range up to 10 km. At 0.875 Hz, a stack of about 3.5 days is required, while at 1.125 Hz a stack of about 5 days of data is required.

Different flavors of crosscorrelation will affect the convergence rate. Different inversion and imaging techniques may require a different crosscorrelation quality.

Virtual seismic sources are retrieved by crosscorrelation of microseism noise and organized into EGMs. It is difficult to judge the radiation polarization pattern in a coordinate system based on north, east and vertical components. This is because the radiation polarization changes as a function of azimuth for the horizontal components. These EGMs can be interpreted more naturally in terms of Love and fundamental and overtone Scholte waves after rotation to the cylindrical coordinate system centered at each source-station. This is an approximate separation due to lateral velocity variations and anisotropy, and the interface waves are only approximately decomposed into Love and Scholte waves. These virtual sources have an almost perfectly uniform radiation pattern over azimuth for the horizontal components, reflecting the uniform distribution over azimuth of incoming microseism noise. The higher-mode Scholte waves are always faster than the fundamental mode. Below 0.6 Hz, the fundamental-mode Scholte waves are also faster than the fundamental-mode Love waves, but above 0.6 Hz the fundamental-mode Love wave travel slower than the fundamental Scholte waves. According to the picked dispersion, in Valhall’s LoFS array with a 50 m in-line and 250 m cross-line spacing: the fundamental-mode Scholte waves become aliased above 0.9 Hz, the higher-mode Scholte waves above 1.2 Hz and the fundamental-mode Love waves become aliased above 0.95 Hz. Although the phase-velocity profiles change over space, small velocity values will not cause large differences in aliasing frequencies. The cross-line spacing is not consistent over the array and thus aliasing varies across the array.
CHAPTER 3. CROSSCORRELATION OF MICROSEISM NOISE

CONCLUSIONS

The microseism energy at Valhall proves sufficiently omnidirectional to be employed for passive seismic interferometry. Seismic interferometry applied at frequencies between 0.175 – 1.75 Hz to three component geophones between all stations in Valhall’s LoFS array yields an estimated Green’s matrix for a virtual seismic survey. This matrix must be rotated to a coordinate system with radial, tangential and vertical components with respect to the source-receiver couple in order to be interpreted. The vertical-to-vertical, radial-to-radial, and radial-to-vertical components are dominated by fundamental-mode Scholte waves but a first overtone is also visible. The higher-mode is stronger in the radial-to-radial component than in the vertical-to-vertical component. The tangential-to-tangential component is dominated by Love waves. Scholte and Love waves are dispersive, and lower frequencies travel faster than higher frequencies. The fundamental-mode Scholte-wave become aliased in the cross-line direction above 0.9 Hz and the fundamental-mode Love waves above 0.95 Hz. The fundamental-mode Scholte wave is faster than the fundamental-mode Love waves below 0.6 Hz, the Love waves travel faster than the fundamental Scholte waves above 0.6 Hz. The fundamental-mode Love wave are not well retrieved above 1.0 Hz. Background correlation fluctuations diminish when stacking more crosscorrelations and the coherent signal emerges. A convergence analysis shows that for frequencies below 1 Hz and offsets smaller than 6 km, stacking as little as two days of crosscorrelations converges to within 5% of a 5-day long recording. It is inconclusive whether the crosscorrelations at higher frequencies have converged to their long term average.

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