Empirical Green’s functions calculated from the inversion of earthquake radiation patterns

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SUMMARY
In this paper, the empirical Green’s function is calculated based on a method that inverts the radiation patterns of earthquakes. The method utilizes several close-lying earthquakes with known focal mechanisms that are received by a common seismic station. The empirical Green’s function between the given source area and the seismic station is then calculated by extracting common features and phases from earthquake waveforms. This is achieved through removing the effect of focal mechanisms from the waveforms and combining the remaining terms based on a damped least-squares approach. The resulting function represents the main characteristics of the propagation path after the waves have left the source area until they are received by a given seismic station. The application of the method is not limited by any requirement to make any separations of seismic phases and the earthquakes used are not required to have the same focal mechanisms. In this research the method is applied to two different data sets, including local Icelandic and regional Iranian earthquakes. In the latter data set the 1990 June 20 Iranian earthquake (Mw = 7.4) was modelled at broad-band seismic station KIV in Russia.

Key words: earthquakes, focal mechanisms, Green’s functions, Iceland, inversion, Iran.

INTRODUCTION
In classical applications of empirical Green’s function methods, normally the waveform of a small earthquake occurring in the source region of a large earthquake is taken as an empirical Green’s function of the large earthquake for a given seismic station (e.g. Capuano et al. 1994). These methods have two major limitations. First, the corner frequency of the small event has to be high enough in comparison with the corner frequency of the large event and at the same time the signal-to-noise ratio of the two events should be reasonably high at a given seismic station. Second, in order to avoid the complexity of having different focal mechanisms, the two events should have the same focal mechanisms. Plicka & Zahradnık (1998) introduced a new method for calculating the empirical Green’s function, which is the motivation of the method introduced in this article. According to their method, empirical Green’s tensor spatial derivatives are calculated based on the representation theorem (Aki & Richards 1980) by removing the focal mechanisms of earthquakes from the waveforms. Contrary to Plicka & Zahradnık (1998), who determined the Green’s function in the usual manner as a response to excitation by elementary dipoles along the coordinate axes of a fixed Cartesian system, in the current research the Green’s function is defined as a response to excitation by vertical strike-slip, vertical dip-slip and 45°-dipping dip-slip faults (three fundamental faults). The basic idea behind the two methods is the same but the way they approach the problem is different. Below, the theory of the method is first explained and the applications are then discussed.

METHODOLOGY
The Fourier transforms of the observed displacements at the free surface of the Earth due to an arbitrarily oriented double-couple point source are defined as (Herrmann & Wang 1985; Jost & Herrmann 1989)

\[
\begin{align*}
\omega r, 0, w & = Z_{SS}A_1 + Z_{DS}A_2 + Z_{DD}A_3, \\
\omega r, 0, q & = R_{SS}A_1 + R_{DS}A_2 + R_{DD}A_3, \\
\omega r, 0, v & = T_{SS}A_1 + T_{DS}A_2
\end{align*}
\]

where \( w \), \( q \) and \( v \) are vertical, radial and tangential displacements, respectively, \( r \) stands for the epicentral distance, 0 indicates that measurements are made on the free surface and \( \omega \) implies that the expressions are evaluated in the frequency domain. The functions \( Z_{SS}, R_{SS} \) and \( T_{SS} \) are the components of the Green’s function required to evaluate a whole waveform excited by a double-couple source (the isotropic components of the Green’s function are ignored). The subscripts SS, DS and DD are the three fundamental faults—vertical strike-slip, vertical dip-slip...
and 45°-dipping dip-slip, respectively. The coefficients \( A \) are the azimuthal radiation patterns of the waveform for a double-couple source and are defined as

\[
A_1 = 1/2(M_{xz} - M_{yz}) \cos 2\phi + M_{xy} \sin 2\phi,
\]

\[
A_2 = M_{xz} \cos \phi + M_{zx} \sin \phi,
\]

\[
A_3 = -1/2(M_{xx} + M_{yy}),
\]

\[
A_4 = 1/2(M_{xx} - M_{yy}) \sin 2\phi - M_{xy} \cos 2\phi,
\]

\[
A_5 = -M_{zz} \cos \phi + M_{zz} \sin \phi,
\]

(2)

where \( \phi \) is the forward azimuth. The components of the moment tensor are related to the orientation of the fault and auxiliary planes (Aki & Richards 1980) via

\[
M_{xx} = -M_0(\sin \delta \cos \lambda \sin 2\varphi + \sin 2\delta \sin \lambda \sin^2 \varphi),
\]

\[
M_{xy} = M_0(\sin \delta \cos \lambda \cos 2\varphi + 1/2 \sin 2 \delta \sin \lambda \sin 2\varphi) = M_{yx},
\]

\[
M_{xz} = -M_0(\cos \delta \cos \lambda \cos \varphi + \cos 2 \delta \sin \lambda \sin \varphi) = M_{zx},
\]

\[
M_{yy} = M_0(\sin \delta \cos \lambda \sin 2\varphi - \sin 2 \delta \sin \lambda \cos^2 \varphi),
\]

\[
M_{yz} = -M_0(\cos \delta \cos \lambda \sin \varphi - \cos 2 \delta \sin \lambda \cos \varphi) = M_{zy},
\]

\[
M_{zz} = M_0 \sin 2 \delta \sin \lambda.
\]

(3)

where \( \varphi, \delta, \lambda \) and \( M_0 \) are the strike, dip, rake and scalar seismic moment, respectively. The first three parameters specify the geometry and the fourth determines the amount of slip associated with a given faulting. In general, these four parameters determine the focal mechanism of a fault movement. For an extended source, the radiation pattern coefficients and consequently the components of the moment tensor become time-dependent quantities. This time dependence can be omitted by working on frequencies well below the corner frequency of the spectra, which allows the source time function to be assumed to be Heaviside function.

The system of equations introduced in (1)–(3) are usually used for seismic moment tensor inversion (Saikia & Herrmann 1985; Shomali & Slunga 2000), in which several seismic stations are distributed in at least two quadrants of the focal sphere and employed to invert the waveforms based on a pre-computed Green’s function in order to calculate the fault parameters (strike, dip and rake) and the scalar seismic moment.

In the current study it is necessary to assume that the focal mechanisms of some earthquakes are known. Given the focal mechanisms, the components of the moment tensor for a double-couple point source can be calculated from eq. (3). Using these components, the radiation pattern coefficients can be computed for a given seismic station from eq. (2). Thus, the only terms that remain unknown are the components of the Green’s function associated with the three fundamental fault types. It has to be added that eq. (1) is only valid for one earthquake and seismic station. Expanding eq. (1) for earthquakes occurring close to each other and received by a common seismic station can be achieved as follows:

\[
w'(r, 0, \omega) = ZSSA'_1 + ZDSA'_2 + ZDDA'_3,
\]

\[
g'(r, 0, \omega) = RSSA'_1 + RDSA'_2 + RDDA'_3,
\]

\[
v'(r, 0, \omega) = TSSA'_4 + TDSA'_5 \quad (i = 1, \ldots, N),
\]

(4)

where \( N \) is the number of earthquakes in a studied area. Eq. (4) implies that the Green’s function components are common between the earthquakes, which is the fundamental assumption in the current research. The vertical component of eq. (4) can be written in the following matrix form

\[
\begin{bmatrix}
w^1(r, 0, \omega) \\
w^2(r, 0, \omega) \\
\vdots \\
w^N(r, 0, \omega)
\end{bmatrix} =
\begin{bmatrix}
A_{11} & A_{12} & A_{13} \\
A_{21} & A_{22} & A_{23} \\
\vdots & \vdots & \vdots \\
A_{N1} & A_{N2} & A_{N3}
\end{bmatrix}
\begin{bmatrix}
Z_{SS} \\
Z_{DS} \\
Z_{DD}
\end{bmatrix}.
\]

(5)

For the radial components, the corresponding matrix can be obtained by replacing the Green’s function components \( R_{SS}, R_{DS} \) and \( R_{DD} \) with \( Z_{SS}, Z_{DS} \) and \( Z_{DD} \). The radiation pattern coefficients for the vertical and radial components are the same, indicating the coupling between \( P \) and \( SV \) phases. For the tangential component the following form is given:

\[
\begin{bmatrix}
v^1(r, 0, \omega) \\
v^2(r, 0, \omega) \\
\vdots \\
v^N(r, 0, \omega)
\end{bmatrix} =
\begin{bmatrix}
A_{14} & A_{15} \\
A_{24} & A_{25} \\
\vdots & \vdots \\
A_{N4} & A_{N5}
\end{bmatrix}
\begin{bmatrix}
T_{SS} \\
T_{DS} \\
T_{DD}
\end{bmatrix}.
\]

(6)

Eqs (5) and (6) imply that the system of linear equations introduced in eq. (4) can in principle be solved based on a least-squares algorithm if at least theoretically more than three independent earthquakes for the vertical and radial components and more than two independent earthquakes for the tangential component are used. The system of linear equations in (4) was solved in the frequency domain for the real and imaginary parts of the time-series independently according to a damped least-squares method. The inversion was carried out at frequencies below the corner frequency (between 0.1 and 7.0 Hz for the Icelandic data and between 0.01 and 0.1 Hz for the Iranian data) in order to allow for the assumption of a Heaviside function as the source time function.

LOCAL ICELANDIC EARTHQUAKES

The method was applied to selected local earthquakes, \( 1.4 < M_L < 2.8 \), taken from the South Iceland Seismic Zone (SISZ). Earthquakes within the area down to magnitude \( M_L = -0.5 \) are selected and located automatically by the South Iceland Lowland (SIL) seismic network (Stefánsson et al. 1993). The focal mechanisms of the earthquakes are then determined based on the spectral amplitude method. This is routinely performed in the SIL network (Rögnvaldsson & Slunga 1993; Slunga et al. 1995). The earthquakes and seismic stations used in the current study are depicted in Fig. 1 and the source parameters of the earthquakes are summarized in Table 1. The coordinates of the Icelandic seismic stations used are given in Table 2.

The empirical Green’s functions between the given source area (Fig. 1) and the four seismic stations HEI, GYG, SAU and SOL were calculated based on the first seven earthquakes listed in Table 1. These Green’s functions were then used to model another earthquake at the four seismic stations. Fig. 2 shows the vertical components of the earthquakes at seismic
station HEI. Fig. 3 shows a comparison between observed and calculated time-series for those earthquakes used for calculating the empirical Green’s function (events #1–#7) at seismic station HEI. The synthetic time-series shown in Fig. 3 are obtained by back-substituting the Green’s function calculated into expression (4). All time-series are bandpassed between 0.5 and 3.5 Hz. The wavelength associated with the upper limit of the filter is expected to be higher than the size of the cluster. With regard to Fig. 3, it can be seen that the method can predict the main features and phases of the observed time-series reasonably well. The method in fact tries to enhance those phases that are common among the time-series and damps those that are not. The empirical Green’s function between the given source area and seismic station HEI was then used to model the eighth earthquake (event #8). The modelling was performed by substituting the resulting Green’s function into expression (1), recalling that at this stage the radiation pattern coefficients are also known because the focal mechanism of the earthquake is known. Fig. 4 shows the modelling of the eighth earthquake at seismic station HEI. In Fig. 4, observed and calculated time-series are shown on the left and the spectral amplitudes are displayed on the right. The numbers next to the

Table 1. Source parameters of the Icelandic earthquakes calculated by the spectral amplitude method.*

<table>
<thead>
<tr>
<th>Event</th>
<th>Date (ymmd)</th>
<th>Origin Time (UTC)</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
<th>Depth (km)</th>
<th>( \text{M}_0 ) (10^{12} , \text{Nm})</th>
<th>( \text{M}_L )</th>
<th>( \varphi ) (°)</th>
<th>( \delta ) (°)</th>
<th>( \lambda ) (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>#1</td>
<td>911216</td>
<td>103103.7</td>
<td>64.082</td>
<td>-20.560</td>
<td>6.5</td>
<td>2.01</td>
<td>2.3</td>
<td>265 (1)</td>
<td>74 (72)</td>
<td>-19 (-164)</td>
</tr>
<tr>
<td>#2</td>
<td>911216</td>
<td>103230.9</td>
<td>64.080</td>
<td>-20.563</td>
<td>6.0</td>
<td>0.77</td>
<td>1.9</td>
<td>274 (8)</td>
<td>80 (71)</td>
<td>-19 (-169)</td>
</tr>
<tr>
<td>#3</td>
<td>911216</td>
<td>103515.1</td>
<td>64.080</td>
<td>-20.561</td>
<td>6.8</td>
<td>1.65</td>
<td>2.2</td>
<td>271 (11)</td>
<td>78 (51)</td>
<td>-40 (-165)</td>
</tr>
<tr>
<td>#4</td>
<td>911216</td>
<td>110558.5</td>
<td>64.080</td>
<td>-20.564</td>
<td>6.2</td>
<td>0.67</td>
<td>1.8</td>
<td>283 (19)</td>
<td>75 (71)</td>
<td>-20 (-164)</td>
</tr>
<tr>
<td>#5</td>
<td>911216</td>
<td>131534.9</td>
<td>64.084</td>
<td>-20.553</td>
<td>6.3</td>
<td>5.82</td>
<td>2.8</td>
<td>357 (139)</td>
<td>24 (71)</td>
<td>126 (76)</td>
</tr>
<tr>
<td>#6</td>
<td>911216</td>
<td>131630.6</td>
<td>64.076</td>
<td>-20.567</td>
<td>6.4</td>
<td>0.06</td>
<td>0.8</td>
<td>268 (358)</td>
<td>86 (88)</td>
<td>-2 (-176)</td>
</tr>
<tr>
<td>#7</td>
<td>911216</td>
<td>144039.0</td>
<td>64.082</td>
<td>-20.558</td>
<td>3.6</td>
<td>0.69</td>
<td>1.8</td>
<td>3 (272)</td>
<td>81 (88)</td>
<td>-178 (-9)</td>
</tr>
<tr>
<td>#8</td>
<td>911216</td>
<td>131717.1</td>
<td>64.071</td>
<td>-20.572</td>
<td>5.2</td>
<td>1.66</td>
<td>2.2</td>
<td>3 (272)</td>
<td>81 (88)</td>
<td>-178 (-9)</td>
</tr>
</tbody>
</table>

*Data from the SIL network.

\( \text{M}_0 \): scalar seismic moment; \( \text{M}_L \): local magnitude; \( \varphi \): strike; \( \delta \): dip; \( \lambda \): rake.

Table 2. Icelandic seismic stations used in the research.*

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Elevation (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GYG</td>
<td>64.281</td>
<td>-20.215</td>
<td>0.113</td>
</tr>
<tr>
<td>HEI</td>
<td>64.200</td>
<td>-21.237</td>
<td>0.160</td>
</tr>
<tr>
<td>SAU</td>
<td>63.990</td>
<td>-20.416</td>
<td>0.077</td>
</tr>
<tr>
<td>SOL</td>
<td>63.929</td>
<td>-20.944</td>
<td>0.031</td>
</tr>
</tbody>
</table>

*Data from the SIL network.

Figure 1. Epicentral map and focal mechanisms of the studied events (circles) within the SIL network. The SIL seismic stations are denoted by triangles. Only those used in the present study are named. The shaded areas show volcanic zones and the thin lines indicate active faults.
Figure 2. Vertical components of the Icelandic earthquakes at seismic station HEI. The sampling rate is 100 Hz and no filter has been applied to the data. The first seven earthquakes (Ev.1–Ev.7) were used for calculating the empirical Green’s function. The eighth earthquake (Ev. 8) was modelled using the estimated function. The source parameters of the events are given in Table 1. The starting time is arbitrary and the data are in counts.

Figure 3. Comparison between observed (solid lines) and calculated (dashed lines) vertical components of the Icelandic earthquakes (Ev.1–Ev.7) at seismic station HEI. All time-series are bandpassed between 0.5 and 3.5 Hz. The starting time is arbitrary and the data are in counts.
Figure 4. Comparison between observed (solid lines) and calculated (dashed lines) data of the Icelandic earthquake #8 at seismic station HEI. To the left the data in the time domain are shown with arbitrary starting time and to the right their spectral amplitudes are depicted. All time-series are bandpassed between 0.5 and 3.5 Hz. The numbers are the maximum values of the normalized cross-correlation functions between the corresponding spectral amplitudes of the observed and calculated data.

Figure 5. As Fig. 4 but for seismic station SOL.
spectral amplitude part of the figure are maximum values of a cross-correlation function between the spectral amplitudes of the observed and calculated data. The spectral amplitudes for the observed and calculated data are more similar than the time-series. The correlation value is higher for the tangential component than for the other two components.

The empirical Green’s functions for the given fracture area and other seismic stations are then calculated in the same way as before. The modelling of event #8 at seismic station SOL is illustrated in Fig. 5. Figs 4 and 5 show reasonable matches between the spectral amplitudes of the observed and calculated time-series. The dominant oscillations in the frequency domain are predicted well.

**REGIONAL IRANIAN EARTHQUAKE**

On 1990 June 20, a major earthquake occurred in northwest Iran near the town of Rasht. The earthquake, which is known as the Rudbar earthquake, was followed by a large number of aftershocks. The main shock ($Ms = 7.7$) caused more than 40,000 deaths (Gao & Wallace 1995). A detailed search was made to find a seismic station that would have recorded the main shock and some other earthquakes from the area including aftershocks with known focal mechanisms. The broad-band seismic station KIV in Russia was found to be suitable to be used for calculating the empirical Green’s function between the source area and seismic station KIV (Fig. 6). The source parameters of the earthquakes are listed in Table 3. Fig. 7 shows the tangential components of the main shock and the eight earthquakes from the source region, including five aftershocks. All the time-series were bandpassed between 0.01 and 0.08 Hz. The upper limit of the filter provides a sufficiently large wavelength in comparison with the spatial separation of the events. Similarity between the time-series is less pronounced than for the Icelandic earthquakes. This might be due to the more scattered spatial distribution of the earthquakes and larger magnitude differences.

In Fig. 8 modelling of the eight earthquakes (events #1–#8) from Table 3 is shown for tangential components. The synthetic time-series are calculated by substituting the empirical Green’s function into expression (4). Prediction of the main phases on calculated time-series is considered reasonable. In order to reduce the effect of the complexity of the source time function, the inversion was performed up to 0.1 Hz, which is expected to be lower than the corner frequency of the main shock. Fig. 9 illustrates the modelling of the main shock of the 1990 June 20 Iranian earthquake at seismic station KIV for vertical and tangential components. Modelling of the main shock is achieved by inserting the empirical Green’s function between the fracture area and seismic station KIV into expression (1), based on the focal mechanism of the main shock given in Table 3. In the figure, the spectral amplitudes show a reasonable similarity of dominant features between the observed and calculated time-series, especially for the tangential component with the higher correlation value. For the main shock of the Iranian earthquake, some differences are apparent between the observed and calculated time-series and corresponding spectra. The tangential components show higher correlation. The lower correlation between the data may be due to the fact that the source time function of the main shock (or modelled earthquake) is different from the source time function of the earthquakes used for calculating the Green’s function. Uncertainties in the focal mechanisms of the events used, their

**Table 3.** Source parameters of the Iranian earthquakes.

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Origin Time (UTC)</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Depth (km)</th>
<th>$M_0$ (N m)</th>
<th>Mag</th>
<th>$\phi$ (°)</th>
<th>$\delta$ (°)</th>
<th>$\lambda$ (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Main shock*</td>
<td>900620</td>
<td>210031.1</td>
<td>36.95</td>
<td>49.52</td>
<td>15.0</td>
<td>1.35e20</td>
<td>7.7</td>
<td>200 (300)</td>
<td>59 (73)</td>
<td>160 (32)</td>
</tr>
<tr>
<td>#1↑</td>
<td>900621</td>
<td>020851.0</td>
<td>36.79</td>
<td>49.83</td>
<td>12.0</td>
<td>0.92e17</td>
<td>5.4</td>
<td>291 (23)</td>
<td>80 (80)</td>
<td>−10 (−170)</td>
</tr>
<tr>
<td>#2*</td>
<td>900621</td>
<td>090218.6</td>
<td>36.51</td>
<td>49.77</td>
<td>15.0</td>
<td>4.95e17</td>
<td>5.4</td>
<td>204 (351)</td>
<td>26 (68)</td>
<td>121 (76)</td>
</tr>
<tr>
<td>#3↑</td>
<td>900621</td>
<td>121727.5</td>
<td>36.99</td>
<td>49.48</td>
<td>6.0</td>
<td>0.80e17</td>
<td>5.3</td>
<td>302 (171)</td>
<td>51 (51)</td>
<td>54 (126)</td>
</tr>
<tr>
<td>#4↑</td>
<td>900621</td>
<td>212739.6</td>
<td>36.68</td>
<td>49.75</td>
<td>8.0</td>
<td>1.10e17</td>
<td>4.9</td>
<td>103 (11)</td>
<td>81 (80)</td>
<td>10 (171)</td>
</tr>
<tr>
<td>#5*</td>
<td>900624</td>
<td>094555.6</td>
<td>36.08</td>
<td>48.91</td>
<td>15.0</td>
<td>1.17e17</td>
<td>4.6</td>
<td>234 (138)</td>
<td>69 (75)</td>
<td>−163 (−22)</td>
</tr>
<tr>
<td>#6*</td>
<td>900706</td>
<td>193459.3</td>
<td>37.03</td>
<td>49.48</td>
<td>15.0</td>
<td>9.58e16</td>
<td>4.3</td>
<td>94 (359)</td>
<td>37 (86)</td>
<td>6 (127)</td>
</tr>
<tr>
<td>#7*</td>
<td>911128</td>
<td>172001.1</td>
<td>36.88</td>
<td>49.33</td>
<td>15.0</td>
<td>3.24e17</td>
<td>5.0</td>
<td>219 (354)</td>
<td>36 (63)</td>
<td>130 (65)</td>
</tr>
<tr>
<td>#8*</td>
<td>951015</td>
<td>065637.8</td>
<td>37.06</td>
<td>49.53</td>
<td>15.0</td>
<td>6.61e16</td>
<td>5.2</td>
<td>66 (158)</td>
<td>49 (88)</td>
<td>178 (41)</td>
</tr>
</tbody>
</table>

* Data from Harvard centroid moment tensor (CMT) catalogue.
↑ Data from Gao & Wallace (1995).
$M_0$: scalar seismic moment; Mag: magnitude; $\phi$: strike; $\delta$: dip; $\lambda$: rake.
Figure 7. Tangential components of the Iranian earthquakes at seismic station KIV. MS and Ev. stand for the main shock and other earthquakes, respectively. The sampling rate is 20 Hz and the data are bandpass filtered between 0.01 and 0.08 Hz. The starting time is arbitrary and the data are in counts.

Figure 8. Observed (solid lines) and calculated (dashed lines) tangential components of the Iranian earthquakes (Ev.1–Ev.8). The time-series are bandpassed between 0.01 and 0.08 Hz. The starting time is arbitrary and the data are in counts.
scattered spatial distribution and the larger-magnitude differences are other sources of mismatch between the observed and calculated data. The method introduced assumes that the focal mechanisms of the events are known and possible error in the focal mechanism cannot be revealed through the process.

DISCUSSION AND CONCLUSIONS

In the current study a method for calculating an empirical Green’s function is introduced that is based on inversions of earthquakes radiation patterns. The method introduced was motivated by the work of Plicka & Zahradnık (1998). The method is mainly based on the assumption that Green’s function or propagation path effects are common between close-lying earthquakes registered at a single seismic station. The method starts by extracting the focal mechanisms of the earthquakes and combines the remaining terms according to a damped least-squares approach. The application of the method to two different data sets, local and regional earthquakes, was discussed. According to these results the method extracts common features and phases among the time-series and damps those that are not common. The method assumes known focal mechanisms of the earthquakes. Therefore, uncertain focal mechanisms might degrade the results. As for other empirical Green’s function approaches, the method introduced needs no instrument correction. Furthermore, in this method it is no longer necessary to separate any phases from each other. Thus, the resulting Green’s function may contain both body and surface waves. The method also does not require the use of earthquakes with the same focal mechanisms. Indeed, the method demands the use of different focal mechanisms as much as possible to place more constraints on the inversion. The method provides the possibility of calculating the empirical Green’s function for each component of a seismic station individually. This can be considered as a step towards separating the source from propagation path effects, which is the fundamental aim of any empirical Green’s function approach.

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