Chapter 2

Surface wave tomography of North America and the Caribbean using global and regional broadband networks: phase velocity maps and limitations of ray theory

Abstract

We present phase velocity maps of fundamental mode Rayleigh waves across the North American and Caribbean plates. Our dataset consists of 1846 waveforms from 172 events recorded at 91 broadband stations operating in North America. We compute phase velocity maps in 4 narrow period bands between 50 and 150 s using a non-linear waveform inversion method that solves for phase velocity perturbations relative to a reference Earth model (PREM). Our results show a strong velocity contrast between high velocities beneath the stable North American craton, and lower velocities in the tectonically active western margin, in agreement with other regional and global surface wave tomography studies. We perform detailed comparisons with global model results, which display good agreement between phase velocity maps in the location and amplitude of the anomalies. However, forward modelling shows that regional maps are more accurate to predict wave-
forms. In addition, at long periods, the amplitude of the velocity anomalies imaged in our regional phase velocity maps is three time larger than in global phase velocity models. This amplitude factor is necessary to explain the data accurately, showing that regional models provide a better image of velocity structures. Synthetic tests show that the raypath coverage used in this study enables to resolve velocity features of the order of 800 to 1000 km. However, only larger length scale features are observed in the phase velocity maps. The limitation in resolution of our maps can be attributed to the wave propagation theory used in the inversion. Ray theory does not account for off-great circle ray propagation effects, such as ray bending or scattering. For wavelengths less than 1000 km, scattering effects are significant and may need to be considered.

2.1 Introduction

The main goal of this study is to obtain better phase velocity maps with improved lateral resolution across North America using the extensive surface wave database available nowadays. In the last decades, the number of high quality broadband seismic stations deployed in the region has increased considerably. In addition to the global and national networks, the North American continent and in particular the United States will be densely covered in the near future with permanent and temporary seismic stations from the ANSS (USGS, 1999) and USArray (Levander et al., 1999) projects. As a consequence, high resolution velocity models of the region are expected to be derived from this extensive data collection. The present research investigates how the increasing density of seismic stations improves the imaging of velocity structure.

This study, covering the entire North American region, attempts to close the gap between local and global tomography studies. The latest surface wave investigations of North America were performed for subregions of the continent: Canada (Frederiksen et al., 2001), the Arctic region (Levshin et al., 2001), and the continental United States (Alsina et al., 1996; Van der Lee and Nolet, 1997). Alsina et al. (1996) imaged the United States with a smaller dataset, using the same method applied here, followed by a linear inversion which included scattering effects. Van der Lee and Nolet (1997), and Frederiksen et al. (2001) used the partitioned waveform inversion method of Nolet (1990). As this method includes higher-mode Rayleigh waves, deeper velocity structures are imaged.

The sensitivity of the fundamental mode Rayleigh waves for the periods considered in this study allows to relate phase velocity maps to shear velocity structure, which can be correlated to features in the uppermost mantle, (e.g. Curtis et al. (1998)).

Another application of phase velocity maps is the determination of source mechanisms using moment tensor inversion methods (e.g. CMT). Even for globally recorded large earthquakes, corrections for the aspherical Earth structure need to be taken into account (Dziewonski et al., 1992), although long wavelength 3D velocity models are generally sufficient (e.g. Woodhouse and Dziewonski (1984)). However, for smaller events, higher resolution models are required to predict surface wave travel times with the necessary accuracy (Arvidsson and Ekström, 1998; Pasyanos et al., 1996). This study should
provide phase velocity maps of North America with the appropriate resolution for this application.

In this study, we also perform a detailed comparison of regional and global models in terms of velocity structures and resolution. Chevrot et al. (1998) and Larson and Ekström (2001) previously analyzed the agreements between global and regional models. A systematic underestimation of velocity amplitude is observed in global models (Nolet et al., 1994; Larson and Ekström, 2001). Here we compare in particular our velocity maps to the global model of Trampert and Woodhouse (2001).

2.2 Data

Figure 2.1: Study area showing the distribution of broadband seismic stations (open triangles) and earthquakes (solid circles) used to determine variations in phase velocity across North America. The great circle paths sampling the area are depicted by grey lines. Also labeled are the events recorded in 2000 (stars) and used in the forward modeling of section 2.6.1.

The waveforms used in this study were extracted from seismograms recorded between 1995 and 1999 by a number of global and regional networks: Global Seismograph Network (GSN), Geoscope, U.S. National Seismograph Network (USNSN), Canadian National Seismic Network (CNSN), Berkeley Digital Seismic Network (BDSN), TERRAs-
cope and other individual stations in North America. In total, we collected waveforms for 172 events recorded by a subset of 91 permanent broadband stations. In California, where a large number of broadband stations are available, only a subset of well distributed stations was selected. The earthquake source parameters were taken from the Harvard Centroid Moment Tensor catalog (Dziewonski and Anderson, 1981). The earthquakes used were all shallow, most of them being located at crustal depths. We first measured group velocities for the fundamental mode Rayleigh waves using an implementation of the time-frequency analysis method (e.g. Levshin et al. (1992)) developed by Charles Ammon, which is described in Pasyanos et al. (2001). In order to select only the fundamental mode Rayleigh wave and eliminate heavily scattered energy (coda) and higher modes, we apply a phase-matched filter which uses the group velocity information. The resulting clean seismograms are used in the non-linear waveform inversion.

![Map of ray density within the study region. Ray density is defined as the number of rays crossing each 2° by 2° cell.](image)

Figure 2.2: Map of ray density within the study region. Ray density is defined as the number of rays crossing each 2° by 2° cell.
2.3 Inversion method

We only selected seismograms for which we were able to obtain reliable group velocity measurements in the complete period range 50-150 s. As a result, the source-receiver geometry is identical for all periods and consists of 1846 raypaths. Figure 2.1 shows the distribution of seismic sources, stations and raypath coverage used in this study. Well-distributed seismic sources and stations throughout the region result in homogeneous raypath coverage over most of North America, except for the Arctic Ocean and the easternmost part of the Caribbean basin and Mid-Atlantic ridge. This coverage represents a large improvement with respect to previous surface wave studies of North America: 275 recordings used in Alsina et al. (1996), and 685 in Van der Lee and Nolet (1997).

While raypath coverage provides insight into the areas that can be imaged, raypath density provides more details on how well the region is sampled. The raypath density map shown in Figure 2.2 is calculated as the number of rays hitting each 2° by 2° cell. The entire continental United States and the Gulf of Mexico are well sampled, while poorer sampling is observed around the edges of the studied region.

2.3 Inversion method

The procedure used in this work is based on a non-linear iterative inversion method developed by Nolet et al. (1986) for waveform fitting. The same inversion technique was previously applied by Alsina et al. (1996) to the United States. The phase velocity maps are computed by minimizing a penalty function in the time domain (Snieder, 1988) through Equation 2.1:

\[ F(m) = \sum_{i=1}^{N} \int [u_i(t) - s_i(m, t)]^2 dt + \gamma \int |\nabla_h m|^2 dW \]  

(2.1)

The \( N \) synthetic seismograms \( s_i \) are computed through the predicted phase velocity model \( m \) and are compared to the observed seismograms \( u_i \). The minimization is achieved using a conjugate gradient method. A smoothing constraint related to the horizontal gradient of the model \( |\nabla_h|^2 \) is applied. The damping parameter \( \gamma \) is chosen to get the optimal trade-off between model smoothness and data fit. The study region covers an area of 10,340 by 12,100 km, which consists of 48 by 56 nodes distributed on a rectangular grid of 2° spacing. At each grid point of the model \( m \), phase velocity perturbations \( \delta c/c \) relative to the reference model PREM (Dziewonski and Anderson, 1981) are computed. The phase velocities between the nodes are interpolated using bi-cubic splines. Synthetic seismograms that fit poorly the data after a preliminary inversion are removed, resulting in the elimination of 17% of the original dataset. The distribution of the discarded paths does not present any systematic relation to a specific source or station.

We carry out the inversion in three steps in order to obtain smooth convergence towards the minimum of the penalty function. First, we solve for velocity perturbations by inverting the normalized envelope of the waveforms, which is insensitive to cycle skipping. This step is equivalent to inverting for group velocities and provides a model that
fits the overall phase observations without considering the contribution of both amplitude and phase. Second, the normalized waveforms of the seismograms, which contain phase information, are inverted. Finally, the true waveforms (i.e. not normalized amplitude) containing amplitude and phase information are inverted to obtain the final phase velocity maps. For each of the three steps, ten iterations are performed. The value of the damping parameter ($\gamma = 0.01$) is identical for all periods and for each of the three inversion steps. The waveforms are extracted and inverted individually over four narrow period bands: 50-66 s, 66-88 s, 88-115 s and 115-150 s. The width of the period bands is calculated in order to obtain equal spacing in both period and frequency logarithms ($\ln dT = 0.26$). In each period band, we assume a constant relative phase velocity perturbation. For the 115-150 s period band, PREM was used as starting model. The phase velocity map obtained for the 115-150 s period band was used as the starting model of the inversion in the 88-115 s period band. This process is applied to successively shorter period bands.

### 2.4 Resolution Tests and Error Estimates

The raypath coverage and ray density per cell (Figures 2.1, 2.2) provide a qualitative insight into the resolution of the model. For a more quantitative analysis, we perform checkerboard and spike anomaly tests. First, an input model is created and synthetic seismograms are computed by forward modelling through the given model using the source and receiver geometry of this study (Figure 2.1). The resulting synthetic dataset is then inverted using the same procedure and model parameterization described above. Comparing the input model with its reconstruction provides an assessment of model resolution, in particular on the amplitude, location and size of the velocity anomalies. Lévêque et al. (1993) showed that these tests could be misleading for certain parameterizations and source-receiver configurations. In our case, the well-distributed combination of sources and stations should not lead to this problem. We perform several tests for input anomalies of different size and location. The ideal case would be to perform a spike test on each cell of the model, providing information on point by point resolution as performed by Ritzwoller and Levshin (1998). However, this is not practical given the size of our model, which consists of 2,688 grid points.

To illustrate the synthetic experiments, checkerboard and spike tests results corresponding to the 66-88 s period band are displayed on Figures 2.3 and 2.4. Similar results are observed for all periods with minor degradation at longer periods. The left panels of Figures 2.3 and 2.4 show the input velocity models used and the right panels present the reconstructed models after inversion. Alternating velocity perturbations of +5% and -5% are used to produce a checkerboard pattern on which we applied a sinusoidal smoothing. Checkerboard tests for velocity anomalies of $16^\circ$, $12^\circ$ and $8^\circ$ are shown on Figures 2.3a-b, 2.3c-d and 2.4a-b respectively. The checkerboard shown in Figure 2.3c-d is rotated by $45^\circ$ compared to the others (Figures 2.3a-b, 2.4a-b) in order to assess the influence of azimuthal coverage. The $8^\circ$ spike test is displayed on Figure 2.4c-d and consists of five discrete anomalies characterized by an amplitude variation of +5%.
Figure 2.3: Results of the synthetic tests: input (a,c) and output (b,d) phase velocity maps of the sensitivity tests for the period band 66-88 seconds. Cell sizes of the checkerboards are 10° (a-b) and 12° with transversal orientation (c-d).
Figure 2.4: Results of the synthetic tests: input (a,c) and output (b,d) velocity maps of the sensitivity tests for the period band 66-88 seconds. (a-b): checkerboard test with 8° cell size. (c-d): spike test.
The distribution of the anomalies is similar to the one used by Van der Lee and Nolet (1997) with an additional anomaly located beneath Iceland. The spatial location and the amplitude of the anomalies are well recovered in our synthetic reconstructions in most of the study area. All reconstructed models display lower resolution along the edges of the target area, where anomalies are smeared out because of limited ray coverage and poor azimuthal sampling. Alsina et al. (1996) observed in their phase velocity maps a bias toward smaller reconstructed amplitudes, which they attributed to the smoothing parameter used in the inversion. In our study, the data set is six times larger, requiring less smoothing and, as a result we achieve a better reconstruction of the amplitude of the synthetic anomalies.

For the long wavelength anomalies, a good match in amplitude and location is achieved across the entire North American continent and the Caribbean. In Greenland, Iceland and in the Pacific and Atlantic oceans reconstruction is poorer due to the lack of crossing rays. For intermediate and small wavelengths (12° and 8°, Figures 2.3c-d and 2.4a-b), reconstruction discrepancies between regions are observed. Beneath the continental United States, the amplitude and location of the checkerboard patterns are still well recovered. For the rest of the study area, the resolving power for 12° and 8° anomalies is lower and is related to inadequate ray sampling. In the Caribbean, smearing occurs in a E-W direction (Figure 2.3d), and for Canada, Alaska and the Arctic we obtain low resolution. For the 8° anomalies, we reconstruct well the locations of the heterogeneities, but the amplitudes are 30% lower in average. Figure 2.4c-d shows results for the five discrete spike anomalies. In the continent, all four anomalies are well reconstructed in amplitude (100% in the center of the spike) and location. For the same test, the model obtained by Van der Lee and Nolet (1997) displays underestimated amplitudes (a maximum of 60% in the center of the spike). However, the anomaly beneath Iceland is slightly shifted to the southwest by about 1000 km due to poor sampling in the region. These results show that the source-receiver geometry is adequate for imaging structures of approximately down to 800 km in the continental United States and in most of the North American continent.

Another measure of the performance of our inversion is the variance reduction, which is computed for each period band using:

\[
\text{variance reduction} (%) = \left( 1 - \frac{\sum_{i=1}^{N} (u_i - s_i)^2}{\sum_{i=1}^{N} (u_i - u_i^0)^2} \right) \times 100 \tag{2.2}
\]

where \( N \) is the number of seismograms, \( u_i^0 \) is the synthetic seismogram computed through the laterally homogeneous model (PREM), \( u_i \) the observed waveform (data) and \( s_i \) the synthetic seismogram computed though the model obtained from the inversion. The variance reduction for each of the period bands is shown in Table 2.1. Variance reduction systematically decreases for shorter periods due to the complexity of the waveforms. Overall, from 50 to 150 s, the nonlinear inversion accounts for up to 70% reduction in variance. This represents a substantial improvement with respect to the study of Alsina et al. (1996) for which the maximum variance reduction was 40%.
Figure 2.5: Waveform fits for the Colombia event of February 8, 1995 ($M_w = 6.4$) recorded by station CMB (California). The epicentral distance is 5800 km. Solid lines: data. Dotted lines: synthetic seismograms computed through the laterally homogeneous model PREM (input seismograms). Dashed lines: output synthetic seismograms after waveform inversion.
2.5 Inversion Results

<table>
<thead>
<tr>
<th>period band (s)</th>
<th>variance reduction (%)&lt;sup&gt;a&lt;/sup&gt;</th>
<th>variance reduction (%)&lt;sup&gt;b&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>115–150</td>
<td>75.2</td>
<td>42.7</td>
</tr>
<tr>
<td>88–115</td>
<td>64.9</td>
<td>32.4</td>
</tr>
<tr>
<td>66–88</td>
<td>61.6</td>
<td>14.3</td>
</tr>
<tr>
<td>50–66</td>
<td>41.8</td>
<td>−40&lt;sup&gt;c&lt;/sup&gt;</td>
</tr>
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</table>

<sup>a</sup>obtained for our phase velocity maps

<sup>b</sup>obtained for phase velocity maps with amplitude reduced to 33%

<sup>c</sup>negative variance reduction indicates that the output synthetics present a poorer fit to the data than the input synthetics (PREM).

As mentioned before, final synthetics presenting worse fit than the initial synthetics (PREM) are excluded from the final inversion. In this process, 380 seismograms were eliminated. In Figure 2.5, we show an example of waveform fits for one event at all period bands considered in this study. The initial synthetic seismograms computed through the homogeneous starting model (PREM) predict poorly both the amplitude and phase of the observed data. After the three-step inversion, both amplitude and phase of the observed seismogram are well matched by the synthetic seismogram.

2.5 Inversion Results

Figure 2.6 shows the phase velocity maps obtained for the four period bands considered: 115-150 s, 88-115 s, 66-88 s and 50-66 s. Our results generally agree with previous regional surface wave studies (e.g. Alsina et al. (1996); Van der Lee and Nolet (1997); Frederiksen et al. (2001)), but we provide new information across a larger area of North America, which includes the entire western Cordillera, the Canadian Shield, Alaska, and the Caribbean basin.

The phase velocity maps display a remarkable consistency across North America for all periods (Figure 2.6). Although the resolution tests show that we are able to image anomalies of at least 800 km, the average length scale of the features observed in our models is larger than 1000 km. The two most prominent features imaged in our models are the relative high velocities (+3 to +5%) present beneath northeastern North America and the slower velocities (-3 to -6%) imaged beneath the active western margin of North and Central America. The two anomalies are well resolved and extend over all periods with a maximum contrast of 10% at short periods to a lower value of 6% at long periods. The high velocities beneath cratonic North America reflect old and cooler upper mantle. These two velocity structures are consistent with global models (Trampert and Woodhouse, 1995; Bijwaard and Spakman, 2000; Larson and Ekström, 2001) and regional studies (Grand, 1994; Alsina et al., 1996; Van der Lee and Nolet, 1997).
Figure 2.6: Estimated phase velocity maps after inversion for 115-150 seconds (a), 88-115 seconds (b), 66-88 seconds (c), 50-66 seconds (d). For a colour version of this Figure, see Appendix, Figure A.1
Depending on the period, the phase velocity perturbations observed can be related to Earth structure at different depths. For short periods (less than 50 s), Rayleigh waves are sensitive primarily to crustal structures, while at longer periods they sample the upper mantle from 50 to 300 km. This allows the interpretation of the phase velocity maps in terms of shear velocity structure.

Beneath the Canadian shield and the northeastern United States, we observe fast velocity anomalies of up to +7%. At long periods, the amplitude decreases to 2%, suggesting that the craton extends down to 150 km. Besides the two main anomalies that characterize continental North America, small scale features are also imaged. A weak lower velocity feature (-1%) is observed at long periods along the eastern Atlantic coast of North America from the Florida peninsula to 45°N (Figure 2.6a-b) and suggests low shear velocities at a depth of approximately 200 km. This feature in our model does not remain at shallower depths, as observed by Van der Lee and Nolet (1997) and Frederiksen et al. (2001), and does not extend to the Appalachians.

The second dominant feature observed below the western Cordillera presents strong negative perturbations (-6%) from the Aleutian Islands to Panama, which are associated with the active tectonic processes in the area. Beneath Alaska, this feature persists at long period along the west coast due to warm mantle materials. Along the Pacific coast of Canada, smaller amplitudes (-2%) are observed at short periods. In the western United States, it is a broad structure of -5% expanding into the interior, to the Basin and Range province. No specific low velocity signature is observed beneath Yellowstone, unlike in the model of Van der Lee and Nolet (1997). Alsina et al. (1996) interpreted relative high velocities of about +2 to +3% (50-100 s period band) beneath the western United States (between 120°W and 125°W) as the subducted Juan de Fuca plate in western Washington. However, in our model, clear low velocities are observed beneath this region. This discrepancy can be explained by the fact that the region lays at the edge of their model where lower resolution is expected.

In Mexico and Central America, we also observe slow heterogeneities in relation with the complex tectonic activity and volcanism in the region. At short periods, low velocities are observed as far south as Panama and Nicaragua, as a signature of a warmer than average mantle. The Gulf of Mexico, is imaged at short periods (50-60 s) as a low velocity zone of -1 to -2%, due to the effect of a thick sedimentary basin. The recent work of Bassin et al. (2000) displays low phase velocities in the Gulf of Mexico, and the study of Vdovin et al. (1999) also shows low group velocities in the region, particularly for periods shorter than 50 s. The crust of the Caribbean basin is younger than the adjacent Atlantic Ocean, and it is surrounded by subduction zones, leading to relatively slower phase velocities. The Caribbean plate is made of the Venezuelan and Colombian basins separated by the Beata ridge (Burke et al., 1978). The Venezuelan basin is characterized by faster anomalies (+1%) relative to the Colombian basin, which is underlain by slower velocity anomalies (-2%). This velocity differentiation agrees with the anomalously thick oceanic crust observed in the region related to the basaltic flow episode which took place in the pre-Mesozoic (80 Ma). In the eastern part (Venezuelan basin), thinner crust, similar
to average oceanic crust is observed (Diebold et al., 1981) and the velocity signature is similar to the Atlantic Ocean. Thicker crust is reported in the Colombian basin leading to lower velocity anomalies. The persistence of this feature at long periods suggest that it extends through the upper mantle.

Relative slow velocities of -1 to -3% are observed beneath Iceland and off the south-eastern coast of Greenland. These relative low velocities are attributed to warm upper mantle and crust at the Mid-Atlantic ridge and the Iceland hotspot. Relative low velocity anomalies beneath Iceland are weaker at longer periods. The resolution in this region is lower, as shown by the checkerboard and spike tests, and no information on the extent of the hotspot at larger depths can be inferred from our models. Along the eastern edge of the North American plate, the Atlantic basin is characterized by relative high velocities (+2 to +3%). The velocity decreases from the continent shore to the Mid Atlantic Ridge to reach relative low velocities beneath the ridge, correlating with the oceanic lithosphere age. The Ridge slow anomalies are clearly seen at shorter periods (66-88 s, corresponding to depths of 50-120 km), but diminish and become wider at longer periods.

2.6 Discussion

2.6.1 Phase prediction: comparison of global and regional models

Our results show similarities with the recent global phase velocity model of Trampert and Woodhouse (2001), hereafter referred to as TW01. Using 46,000 Rayleigh wave measurements recorded primarily by GSN and Geoscope stations, they computed global phase velocity maps in the period range 40 s to 150 s. Shown in Figure 2.7 are the phase velocity maps of TW01 for periods of 60 s and 130 s using spherical harmonic expansion up to degree and order 40. The phase velocity maps given in TW01 are obtained for narrow frequency bands (2.5 mHz wide) around a number of central frequencies. Therefore, we compare our maps for a period band (e.g. 115-150 s) with TW01 maps at the closest central period (e.g. 130 s). Major features, such as the large velocity contrast between the North American craton and the western Cordillera are clearly observed in both models. A comparison of the two models for periods of 50 s to 100 s shows good agreement in both amplitude and lateral variations in phase velocities. The correlation between our model and TW01, computed over the whole inversion region, displays values of 70% at short periods (Table 2.2).

For periods greater than 100 s, the agreement between the two maps (Figure 2.7) is good, but the amplitude of the anomalies differs by approximately a factor three and the correlation decreases to 46%. For shorter periods, although the correlation between the two maps is good, a significant difference is observed in the Gulf of Mexico and the Caribbean basin. One of the new results obtained in our study is the slow velocity region imaged beneath the Gulf of Mexico and the Caribbean plate. While both models show slower velocities for the western part of the Caribbean basin, our results suggest large variations between the western and eastern portions of the basin. Differences also occur
along the edges of our model, where we lack ray coverage. For example, the Mid Atlantic Ridge is imaged as a slow region at 60 s while in our model, such a negative anomaly is only observed above 70 s.

Table 2.2: Correlation between our regional model and the global model TW01.

<table>
<thead>
<tr>
<th>period band</th>
<th>correlation (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>115 – 150</td>
<td>46.1</td>
</tr>
<tr>
<td>88 – 115</td>
<td>66.0</td>
</tr>
<tr>
<td>66 – 88</td>
<td>71.1</td>
</tr>
<tr>
<td>50 – 66</td>
<td>70.9</td>
</tr>
</tbody>
</table>

To determine the influence of the differences in the amplitude of the velocity anomalies in the data fit, we compute the variance reduction for a velocity model identical to the one obtained by inversion, but with the amplitude of the anomalies reduced to one third. A large decrease of the variance reduction is observed at all periods (right column of Table 2.1). For the shortest periods (50-66 s), the synthetic seismograms exhibit poorer fit to the data than the PREM seismograms, showing the importance of imaging accurately the amplitude of the velocity perturbations. The factor three in amplitude observed between our model and TW01 is therefore significant and suggests that global models underestimate phase velocities at long periods. This may be attributed to the strong regularization applied in TW01 to compensate for noisy global data. Other surface wave global studies (Ekström et al., 1997; Zhang and Lay, 1996) report lower amplitude at long periods than regional models. Larson and Ekström (2001) for Eurasia and Nolet et al. (1986) previously investigated the underestimation of velocity anomalies by global models in comparison with regional models. They suggest that the observed discrepancy may be due to three reasons: the different phase velocity measurement techniques used in the global and regional inversions; the variable signal to noise ratio observed especially for short path lengths; and the damping parameters applied.

The accurate estimation of phase arrival times is an important application of phase velocity maps. It is particularly relevant for moment tensor methods that use surface wave waveforms. In order to investigate the accuracy of our regional model and TW01 in predicting phase arrival times, we perform a forward modelling test. A total of 100 seismograms from 8 events recorded in 2000 (stars on Figure 2.1), which were not included in our inversion dataset, were processed using the same method described in section 2.2 (measurement of the group velocity followed by a phase-matched filter to isolate the fundamental mode). The waveforms are compared with the predicted seismograms computed by forward propagation through the two phase velocity maps (regional and global). As TW01 was obtained by inverting phase velocities measured from normalized seismograms, we compare the predictions in terms of normalized waveforms. Examples of the obtained predictions are given on Figure 2.8 for three different seis-
mograms at 80 s period. The synthetics computed through the two models exhibit significant differences. Predictions computed through TW01 present phase arrival time differences of almost 25 s while the time differences displayed by the regional predictions do not exceed 10 s. From the 100 seismograms used for this test, 97 were better predicted by our regional velocity map. Moreover, stronger discrepancies would be observed if true waveforms (not normalized) were compared. Almost 60% of the synthetics computed through our regional model explain the data with RMS residuals between 0 and 0.4 (root mean square of the difference between observed and predicted normalized waveforms). In contrast, the minimum RMS residual observed for TW01 synthetics is 0.9. In Figure 2.9, the RMS residual averaged over all seismograms is displayed as a function of period. The discrepancy between the two sets of predictions is significant. The average RMS residual obtained with our regional model is always lower than 0.6 for all periods, while for TW01 it reaches values of 3.2. This shows that despite the similarity observed between the two phase velocity maps, slight variations in velocity structures are important to explain the data. This is also supported by the systematic comparison of phase velocity models done by Trampert and Woodhouse (2001).
Therefore, to obtain accurate phase predictions, it is essential to use regional phase velocity maps. This is particularly important for regional moment tensor calculation.
The phase velocity maps of North America obtained in this study, which predict phase shifts with better accuracy, should significantly improve CMT determination in the region.

![Data Fit](image)

Figure 2.9: Comparison of the phase prediction for regional (solid line) and global (dashed line) models: root mean square of unexplained variance to the data after forward modeling. A total of 100 seismograms recorded in 2000 were used to compute the mean.

### 2.6.2 Limitations of ray theory

The previous sections showed that we are able to reconstruct velocity features with better accuracy than global models (which has been illustrated for the TW01 model) and obtain better or comparable resolution than previous regional models (Alsina et al., 1996; Van der Lee and Nolet, 1997; Frederiksen et al., 2001). However, the length scale of the velocity features actually observed in the phase velocity maps is larger than expected from the synthetic tests (800 km). In order to understand the reasons why we do not obtain models with smaller features, we need to consider the approximations used in the inversion of the surface waves between 50 s and 150 s. Our method is based on ray theory and assumes that the ray path follows the great circle linking the source and the receiver. In ray theory, the wave that travels through an inhomogeneous medium is assumed to be sensitive to velocity perturbations along the raypath. In reality, the wave is also sensitive to and gets scattered by structures off the great circle. For example, Laske (1995) showed that using off great circle arrival angles from polarization analysis improves significantly
the fit of long periods surface waves. If we take scattering effects into account in the forward direction (Snieder, 1988), we can define scattering kernels (or Fréchet kernels) as the sensitivity of the wave to velocity perturbation at any geographical point (Marquering et al., 1998; Snieder and Lomax, 1996; Dahlen et al., 2000; Hung et al., 2000). An example of phase velocity sensitivity kernel is shown in Figure 2.10, computed using the formalism described in Spetzler et al. (2002). Equivalent behavior of the delay time is reported by Nolet and Dahlen (2000) using Gaussian beam solution. Constructive interference at the receiver occurs for the waves scattered by anomalies inside a region defined as the first Fresnel zone which corresponds to the width of the main peak in Figure 2.10 (e.g. Kravstov and Orlov (1990); Spetzler and Snieder (2001)). It is an ellipsoidal area along the source receiver path with a maximum width given by:

\[
L_F = \sqrt{\frac{3\lambda}{2}} \frac{\tan \frac{\Delta}{2}}{2}
\]

(2.3)

Figure 2.10: Cross-section of the scattering sensitivity kernel for relative phase velocity perturbations using finite frequencies. The epicentral distance is 100°, the viewpoint is 50° away from the source along the great circle and the frequency is 100 s.

The maximum width of the Fresnel zone \((L_F)\) increases with the period \((T = \frac{\lambda}{v})\) and the epicentral distance \(\Delta\). For more details on the derivation of sensitivity kernels and width of the Fresnel zone using Rytov approximation in a uniform medium, we refer to Spetzler et al. (2002). In our study, at 100 s period, the Fresnel zone width can be as large as 1860 km for an epicentral distance of 10,000 km (Figure 2.11). Inversions based on
ray theory allow for reconstruction of velocity perturbations only along the raypath. For velocity perturbations much larger than the width of the Fresnel zone \( (a \gg L_F, \ a \text{ being the characteristic length of the heterogeneity}) \), the velocity reconstruction obtained using scattering theory is equivalent to the one computed with ray theory but for structures on the order of or smaller than this width \( (a \lesssim L_F) \), ray theory is no longer applicable.

Figure 2.11: Fresnel zone width for four different paths at 100 seconds period. The width is computed using Equation 2.3.

To understand the consequences of diffraction effects on phase velocity structures imaging, Spetzler et al. (2001) computed the bias induced by diffraction on phase velocity measurements and compared it with the relative measurements error (see Figure 2 in Spetzler et al. (2001)). For a surface wave of 150 s, the error induced by the use of ray theory is not significant for velocity perturbations larger than 1250 km but is already of the order of 50% for velocity perturbations of 1000 km. Recently Yoshizawa and Kennett (2002) suggested that the zone of influence (e.g. the region to which a wave is sensitive to velocity anomalies) around the surface wave path is only one third of the first Fresnel
2.7 Conclusions

Regional Rayleigh wave phase velocity maps were produced in the period range 50 s - 150 s for North America and the Caribbean. These maps predict phase delays and long period velocity amplitudes with higher accuracy than global models and are therefore reliable for source mechanism estimation or as starting model to S-wave structure studies. Despite strong improvements in data coverage and lateral resolution compared to previous investigations of the region, ray theory approximation is a limiting factor to reconstruct short wavelength structures. For the scale of velocity features we aim to image, off-great-circle propagation and scattering might be important. The influence of scattering effects in regional tomography requires further evaluation. The implementation of diffraction kernels in inversion procedures will determine if higher order wave propagation approximation is necessary. These results have broad implications for proposed large-scale seismic experiments in North America. Significant increase in station density may not translate into better models without improving the methodological assumptions and limitations.