Assessment of global phase velocity models

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SUMMARY

We construct new Love and Rayleigh wave phase velocity models based on measurements made from an aspherical starting model and strict data quality control derived from cluster analyses. These new models are in good agreement with previous ones and the question arises whether the slight changes show an improved capacity to explain the data. To this effect, we propose an objective method to compare different phase velocity models published in recent literature. The method is based on comparing calculated synthetics to raw seismograms. We find a reassuring convergence, between all the models we tested, at the longest periods and more scatter at the shorter periods. At 40 s, the different models show gains of up to 3.5 cycles over PREM. Generally, the higher the gain over PREM, the smaller the period considered, which confirms that the Earth's heterogeneity is strongest in the uppermost parts of the Earth. Apart from assessing different models against each other, our method gives an estimate, comparable to cluster analyses, of the underlying data errors that went into the construction of the models themselves. Moreover, ray coverage is still far from perfect for constructing phase velocity models. As a result, we find that without precaution, degree zero is biased through spectral leakage by 0.1 to 0.2 per cent with respect to PREM.

Key words: error analyses, phase velocity, spectral leakage, surface waves.

1 INTRODUCTION

The earliest information concerning the Earth's 3-D structure came from surface waves, when Tams (1921) noticed clear velocity differences for waves propagating along continental versus oceanic paths. Gutenberg (1924) attempted an explanation of such differences in terms of regional variations in crustal structure when he noticed that short-period surface waves are more sensitive to crustal structure than long-period ones. This led the way to intensive studies, both theoretical and observational, to link dispersive properties of surface waves to the internal structure of the Earth. The first regional models of the Earth's dispersive properties of crust and upper mantle are summarized in papers by Brune (1969) and Dorman (1969), respectively. With the start of digital recording of seismograms in the late seventies, new methods of analysis were at the disposal of seismologists. This led to a rather precise knowledge of global distributions of phase velocities, typically for periods longer than 150 s and the first few degrees in the spherical harmonic expansion of heterogeneity (e.g. Masters et al. 1982; Nakanishi & Anderson 1982, 1983, 1984; Montagner & Tanimoto 1990). The strength of these phase velocity distributions is that they may directly be used to infer the Earth's 3-D velocity field (e.g. Nataf et al. 1986; Montagner & Tanimoto 1991). Recent contributions to global phase velocity models exploit the fast

growing number of digital seismograms (Zhang & Lay 1996), automate the measuring technique and extend it to much shorter periods (Trampert & Woodhouse 1995, 1996; Ekström *et al.* 1997), add polarization data (Laske & Masters 1996) and extend the measurements to higher modes (van Heijst & Woodhouse 1997, 1999).

Phase velocity models have two main advantages: they are a compact way of representing large amounts of phase data as a function of frequency and they are to first order a linear combination of the underlying 3-D velocity structure. Before using phase velocity information, we would like be certain that it correctly explains observed seismograms, not only the ones selected to construct the maps, but equally importantly, the large majority that has not been used in the mapping process. The higher the predictive power of the models, the better they constrain the true Earth structure. Assessment of phase velocity models, so far, has relied on correlations between existing models, comparing variance reductions between different studies and matching dominant features against expectations based on surface tectonics. None of these assessment techniques is entirely satisfactory: correlations between existing models cannot make statements about their relation to the true earth model; variance reductions are highly dependent on the set-up of a specific inverse problem, which varies from study to study; matching against expectations gives at best a qualitative judgement.

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We propose here to test existing models against a set of raw seismograms. The approach is inspired by the study of Ritzwoller & Lavely (1995) in which they compared 3-D earth models against measured structure coefficients derived from observed mode splitting. Our emphasis is mainly on the shortperiod end, which led us to select the models of Trampert & Woodhouse (1995, 1996), Ekström *et al.* (1997) and van Heijst & Woodhouse (1999) (referred to as TW95, TW96, ETL97 and VHW99) for comparison. We also included the longer-period study of Laske & Masters (1996) (LM96), mainly because they used phase and polarization data to construct their model. In addition, we used a new model, presented in the next section, where the measuring technique of TW95 has been refined.

2 CONSTRUCTION OF A NEW GLOBAL PHASE VELOCITY MODEL

The main improvements over previous models are due to cleaner phase velocity measurements. Measurements are somewhat dependent on the starting model used in the inverse process. Here we start from an existing aspherical model that speeds up convergence and makes the result less dependent on the starting phase. Great care has been taken to resolve the 2π phase ambiguity at short periods, and outliers in the measurements have been identified by cluster analyses.

Determining the phase velocity in a seismogram is a highly non-linear problem, particularly at higher frequencies. Using a classical iterative scheme (e.g. Tarantola & Valette 1982) can make the solution dependent on the starting model. In TW95 and TW96, we have chosen this starting model from an approximate estimation of group velocities. The gross features of the models obtained in TW95 and TW96 are quite similar, suggesting taking one of these models as a starting point. We have chosen the latter. The number of iterations in the nonlinear inversion process dropped significantly compared to starting from a group velocity estimation, indicating that the aspherical model brings the phase closer to the final solution required by the observation. Measuring phase velocities from an aspherical starting model is also used in ETL97.

Another major concern in phase velocity measurements is the 2π phase ambiguity arising from the multivalued nature of the Fourier phase. While at periods longer than 150 s it is possible to determine the exact number of full cycles in phase without ambiguity, at shorter periods this becomes increasingly difficult. One way to solve the problem is to acknowledge that dispersion curves are intrinsically smooth, owing to the fact that lateral and vertical heterogeneities are averaged during surface wave propagation. The usual approach is then to anchor the full number of cycles at long periods and extrapolate to the correct number of cycles at shorter periods requiring a smooth dispersion curve. In all automatic measuring techniques, the dispersion curve is usually parametrized on some spline basis. To impose a smooth curve, two possibilities exist: choosing a small number of splines resulting in an implicit smoothing, or using many splines with an explicit derivative smoothing constraint. TW95, TW96 and VHW99 opted for the latter, while ETL97 chose the former. We compared measuring results based on six B-splines with no explicit smoothing and on 36 B-splines with explicit Laplacian smoothing. All measurements have been mapped onto phase velocity models parametrized in spherical harmonics up to degree 40. Fig. 1 shows the different amplitude spectra obtained in the case of 40 s Love waves. The spectrum of the map based on differential measurements (36 B-spline minus six B-spline parametrization measurement) has little amplitude at low degrees and becomes flat from middegrees onwards. By visual inspection, the maps derived from 36 or six B-spline measurements alone differ little since the main power in these maps is found to be in the lower degrees (Fig. 1 and Fig. 2 bottom). Differences occur mainly for smaller scale lengths, as can be seen in Fig. 2 (top). The differences in the measurements cannot be regarded as measurement noise since they map efficiently (with a 30 per cent variance reduction)



Figure 1. Amplitude spectra of two 40 s Love wave phase velocity models constructed with measurements parametrized on a different number of B-spline knots. The spectrum of the model constructed from the differential measurements is also shown. In the six B-spline measurements π phase-shifts remained undetected, which explains the power in the model calculated from differential measurements.





Figure 2. 40 s Love wave phase velocity model. Top: model constructed from differential measurements. This model explains 30 per cent of the differential data variance, indicating that cycle errors efficiently map into the models. Bottom: model constructed from 'clean' measurements. This model explains 91 per cent of the data variance. The maps represent relative variations with respect to PREM in per cent.

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into structures. For comparison, the inversion of random noise instead of measurements achieves a variance reduction of only 3 per cent. The reason for these differences is that the 2π jumps at short period are not always possible to avoid if the smoothing is not appropriate for a given path (Fig. 3). With a small number of splines, the solution will interpolate evenly through the two sides where the jump occurred and the bias will remain unnoticed, resulting in a π phase-shift in the vicinity of the actual 2π jump. With an overparametrized dispersion curve, the jump will be recognized as a sharp peak in the residual curve (observed phase minus predicted final phase), even with a strong Laplacian damping, simply because the spline knots are closely spaced. We therefore favour a spline overparametrization that leads to reliable 2π jump detections in the automatic procedure. Moreover, spline computations are so efficient that they do not lead to noticeable penalties in computation time. The measurements of ETL97, on the other hand, based on six B-splines only, could suffer from such unnoticed biases at the short-period end as our results below suggest.

We remeasured all GDSN and GEOSCOPE seismograms, available on the Oxford WORM jukebox, recorded between 1980 and 1993 and extended the period until the end of 1995. All events had magnitudes M_w between 5.8 and 6.6. Rather than discarding the measurements from TW95 and TW96, we constructed a combined data set. Some path may thus occur several times with possibly conflicting phase velocity. The idea is that concatenated measurements starting from the spherical model PREM (Dziewonski & Anderson 1981) with different control parameters and measurements starting from an aspherical model should cover the uncertainties of the measuring technique itself well. Other errors include, for instance, source location, focal mechanism and timing (for a more detailed discussion see ETL97). We assess the total error by cluster analyses, where similar paths (within 0.5° at both endpoints) are compared. Only variances for clusters with more than 10 paths are considered. To eliminate outliers, all data in a given cluster for which the measurements falls outside one standard deviation around the mean are excluded. This resulted in keeping a total of 54 249 Rayleigh wave measurements (45 938 R1 and 8311 R2) and 41 016 Love wave measurements (36 819 G1 and 4197 G2). The data coverage (number of rays passing a cell with an area of 10° squared) is shown in Fig. 4 (bottom). The rays belonging to clusters with fewer than 10 measurements were all kept since they cross the more sparsely sampled parts of the Earth and are thus important for overall resolution. We assigned mean errors, derived from all clusters with sufficient paths, to these phase measurements. These mean errors as a function of frequency for relative phase velocities $\Delta(dc/c)$ result in a phase error proportional to distance $\Delta(d\phi)$. Using $\Delta(d\phi) =$ $kx\Delta(dc/c)$, k being the wavenumber and x the distance, mean errors in phase for a distance of 10^4 km are 21° , 41° and 73° for Rayleigh waves and 30° , 43° and 83° for Love waves for periods of 150, 80 and 40 s, respectively.

We constructed phase velocity maps from our cleaned data. Relative phase velocities are expanded in spherical harmonics up to degree and order 40. Explicit Laplacian smoothing is introduced to stabilize the solution and prevent most of the spectral leakage (Trampert 1998). The variance reductions achieved are high, indicating a consistent data set: 92, 79 and 47 per cent for Rayleigh waves and 91, 79 and 51 per cent for Love waves for periods of 40, 80 and 150 s, respectively. Some studies fit phase rather than phase velocities (ETL97 and VHW99). We found that it made no difference to our final results. The maps are similar to previous ones and are not shown here, except for Love waves at 40 s (Fig. 2, bottom). Note that the blue velocity area in Tibet persists, consistent with our previous models. This was criticized by ETL97 as measurement noise. The data that contribute with highest sensitivities to this area all show rapid variations in their dispersion curves. As discussed earlier, this could have been missed by ETL97 because of their six B-spline parametrization. A recent global traveltime model by Bijwaard et al. (1998) shows a continental subduction zone at precisely this location, corroborating our observation. It is interesting to see the evolution of our models through their overall resolution.



Figure 3. Example of a 36 and six B-spline measurement for event 012594E in the Harvard catalogue for the vertical component of station BJI. The epicentral distance is 13 923 km. At a period of 45 s (0.022 Hz), a sudden 2π jump is seen on the 36 B-spline measurement. Using six B-splines, the measurement shows a smooth interpolation through the jump, which will be difficult to detect on the residual curve.



Log(Ray density)

Figure 4. Ray density plotted on a base-10 logarithmic scale. The ray density is calculated by dividing the Earth into cells with constant area $(10^{\circ} \times 10^{\circ})$ and counting the number of rays passing through each cell. Top: coverage used in the assessment of the phase velocity models. Bottom: coverage used in the construction of the phase velocity model presented in this study. There is roughly a factor of 10 between the ray density in the two plots. Both maps are for Rayleigh waves.

From TW95 to TW96 the improvement in resolution was dramatic, mainly in the Southern Hemisphere, due to the inclusion of major-arc data. The present model shows further improvements in all areas of lower ray density (light grey colours in Fig. 4, bottom) but also a degradation in the Pacific where ray density is highest. This leads to a much more uniform resolution over the globe and will make interpretations more straightforward. The radius of the averaging kernel is about 700 km, which corresponds to an overall resolution up to spherical harmonic degree 28. The model can be retrieved from ftp://terra.geo.uu.nl/pub/people/jeannot/twgji00iso.tar.gz. In the construction of all models and in the assessment part below we

assume that surface waves follow great-circle paths. We are not going to repeat a discussion of this assumption here but refer the reader to TW95, LM96, ETL97 or VHW99.

3 ASSESSMENT OF THE MODELS

We arbitrarily selected data from 40 events with a magnitude M_w between 6.1 and 6.3 recorded in 1994. This resulted in approximately 9000 seismograms equally divided between vertical and transverse components. The only selection criterion we required was that seismograms had a signal-to-noise ratio greater than 5. This eliminates approximately 50 per cent of the raw data. Most parts of the world are adequately covered by the resulting rays. The ray density corresponds to roughly onetenth of that used in the construction of the models described in the previous section (Fig. 4). The rejection rate in the automatic phase velocity measuring technique is much higher due to many in-built safeguards and as a result not more than 15 per cent of these data went into any of our own models. We assumed that this should approximately hold for other authors as well (for example, ETL97 report that the models of TW95 and TW96 explain 90 and 87 per cent, respectively, of their variance). The idea is to adjust the phase of a synthetic seismogram calculated for PREM with a given model and compare the result to the phase of the observed seismogram. With our Fourier transform convention this phase adjustment is given by

$$d\phi = \omega/c_0 \left[(dc/c_0) ds \right], \tag{1}$$

where c_0 is the reference phase velocity at radial frequency ω and ds the incremental path along the ray. The exact amplitude ratios between observed and synthetic seismogram are used to make the comparison amplitude independent. All models are taken at four–six different frequencies along the dispersion curve of interest and spline interpolated to obtain a smooth variation as a function of frequency. We checked that the exact number of model points and the spacing did not have a significant effect on the results.

All source parameters for the events are from the Harvard catalogue. The seismograms were sampled at one point every 16 s but a higher sampling rate made no difference to the results. Once the total phase (PREM plus phase velocity model) of the synthetic seismogram is determined, synthetic and observed seismograms are narrow-band filtered around certain fixed periods (40, 60, 80, 100 and 150 s). This leaves two shifted sinusoids, and their phase-shift α is readily determined from the corresponding unexplained variance using the expression

$$\frac{\sum_{\text{samples}} [d_i - s_i]^2}{\sum_{\text{samples}} d_i^2} \simeq \frac{\sum_{\text{samples}} [\sin x - \sin(x + \alpha)]^2}{\sum_{\text{samples}} \sin^2 x} = 4\sin^2(\alpha/2), \quad (2)$$

where d_i and s_i are the samples of the narrow-band filtered real and synthetic data, respectively. An important issue is to isolate the fundamental mode from the overtones in a seismogram. Since we do not want to favour our own measuring technique, we do not use time-variable filtering as in previous studies. Instead, the variance reductions are measured on cross-correlation functions with purely fundamental-mode synthetics. This enhances the signal-to-noise ratio and thus the importance of the fundamental mode in the measurement. This is similar to the branch cross-correlation functions used in VHW99. The measured phase-shift α is always positive because of the square in the right-hand side of expression (2) and lower than 180° because the argument of the sine is $\alpha/2$. Hence, we cannot determine whether the data are fast or slow with respect to the synthetics, but only that there is a phase-shift up to $\pm 180^{\circ}$. According to eq. (2), the phase-shift α between narrowband-filtered observed and synthetic data is directly derived from the variance reduction achieved. There are of course many good reasons why a synthetic phase will not correspond to the observed one, so that analysing individual paths will show complicated trends. We suggest looking at mean phase-shifts for paths having the same epicentral distance. This will reveal

global trends in the model shortcomings (for example, better agreements for shorter paths). For a given distance bin (taken to be 2000 km), the measured phase-shifts α for all individual source-receiver pairs will cover a certain range. Although this range of measured phase-shifts will always be contained between 0 and 180° , the true range can of course be much bigger. This true range is unknown, because we cannot unambiguously determine the initial phase of the sinusoids. The question is what happens to our mean phase-shift for true ranges in excess of 180°. If one assumes that a range is sampled more or less homogeneously, the behaviour of the mean phase-shift can be modelled as a function of range. Such a simulation can be seen in Fig. 5. The straight line up to a range of $\pm 180^{\circ}$ is easily understood and the oscillations around 90° for bigger ranges are due to the half-cycle periodicity of the right-hand side of eq. (2). In reality, this range is of course not sampled uniformly owing to an inhomogeneous ray density, biases in the models towards certain paths, etc. What remains approximately true, though, is that a mean phase-shift smaller that 90° corresponds to a range of phase-shifts smaller than $\pm 180^{\circ}$. Mean phaseshifts with values of approximately 90° correspond to a true range of at least $\pm 180^{\circ}$.

4 RESULTS

Figs 6 and 7 show the mean phase-shifts as a function of epicentral distance. The standard deviations are not shown because they are fairly similar for the different models, varying roughly between 30° and 40° . Note that this standard deviation corresponds to measured phase-shifts α that always lie between 0 and 180° . The means are calculated for 2000 km distance bins, resulting in a total of 20 distance bins. Each bin contains between 200 and 600 data, except the 12th bin, with only 35 data points. The results for this bin are likely to be unreliable. As in our previous studies, we discarded data with minor-arc distances smaller than 20° and larger than 160° . The results for LM96 are only shown from 80 s onwards.

Before analysing the phase velocity models themselves, we can make some comments on the reference model. We find that the longer the period, the better the seismograms are explained by the reference model PREM (Dziewonski & Anderson 1981). For a given period used in the construction of PREM (from 75 s onwards), Rayleigh waves are better modelled than Love waves. This could be due to higher-quality vertical measurements that entered the construction of PREM. An alternative explanation could be that different path coverages were used for Rayleigh- and Love-wave-sensitive data.

The predictions from the different phase velocity models show an increasing scatter in data fits with decreasing period. This is best illustrated by the cumulative standard deviation obtained from the different model predictions seen in Figs 6 and 7 as a function of distance (see Fig. 8). This change in scatter with period is more severe in Rayleigh waves than in Love waves. Overall it is reassuring to see how well the different studies converge at the longest periods (within 6° for Rayleigh and 10° for Love at 150 s). At 40 s, model predictions lie within 15° (Love) and 28° (Rayleigh) of each other, except for ETL97, which makes noticeably worse predictions in the distance range $4000-12\ 000\ \text{km}$. These latter averages are higher because the ETL97 model gives significantly more phase-shifts around 180° in this distance range. This is consistent with the suggestion



Figure 5. Mean phase-shift as a function of range. The range corresponds to that of individual phase-shifts, which may be bigger than 180° . The mean is calculated from measured phase-shifts using the right-hand side of eq. (2), which always lies between 0 and 180° .

that many of their measurements in this predominant distance range contain 180° phase-shifts due to unrecognized cycle slips within the dispersion curve, as discussed previously.

At 40 s, PREM shows an average phase-shift of approximately 90° for any distance. This indicates that individual

PREM predictions are falling in a range larger than $\pm 180^{\circ}$. The phase velocity model predictions quickly align with PREM and only show significant improvements over PREM for the shortest epicentral distances. Does this mean that we do not gain anything by using these models? To answer this question,



Figure 6. Mean phase-shift as a function of distance obtained for different model predictions. Results for Love waves at different frequencies.



Figure 7. Same as Fig. 6, but for Rayleigh waves.



Figure 8. Cumulative standard deviation of model predictions at a given period. For a given distance bin, seen in Figs 6 and 7, the different model predictions are averaged and the standard deviation is calculated. L stands for Love and R for Rayleigh waves; the number in the label indicates the period.

it is interesting to analyse the individual improvements over PREM as given by eq. (1). Fig. 9 shows 2 standard deviations of this gain, which should cover more than 95 per cent of the individual data. It is clear from this figure that for Love waves at 40 s, we gain up to 3.5 cycles over PREM depending on the distance. From 80 s period onwards, mean phase-shifts lower than 90° are being noticed on major-arc distances, indicating that these predictions are starting to fall within a cycle of the real data. Where mean phase-shifts are lower than 90°, all data are predicted within a cycle ($\pm 180^\circ$). This holds for all models with periods longer than 80 s. The largest improvements over PREM are for the smallest periods, and vice versa, indicating that the strongest heterogeneities are located in the uppermost mantle and/or the crust.

Even for the longest periods, the mean phase-shifts are quite significant. It is interesting to note that the mean phase-shifts in Figs 6 and 7 are comparable to the phase errors estimated from cluster analyses described in an earlier section. The mean phase-shifts obtained here are thus a good measure of the errors in the data that went into the construction of the models themselves. Using these average error estimates as a function of epicentral distance, the χ^2 values of our models are close to



Figure 9. Range of the gain over PREM for the Love wave phase velocity models of this study as a function of distance. For each path, the gain over PREM is calculated using eq. (1). The gains for all paths are averaged and the standard deviation is calculated. 2 standard deviations are shown, which should cover 95 per cent of the data.

one. Assuming that cluster analyses or mean phase-shifts, as calculated in this work, represent realistic error estimates for the data, we find that our phase velocity models explain all information on 3-D structure contained in the data.

5 DEGREE ZERO

It has been noted in the past that the different phase velocity models disagree significantly (a few tenths of a per cent) on the estimation of the degree zero, or spherical average, perturbation with respect to PREM (van Heijst 1997). For the construction of a new spherical average model (e.g. Masters et al. 1999) it is particularly important to understand this finding. Our assessment approach can remeasure the degree zero contribution of the models quite easily by looking at all individual phase corrections with respect to PREM as given in eq. (1). The spherical average will show as a linear trend if the gains are plotted as a function of distance. Fig. 10 shows an example for Rayleigh waves at 40 s. The linear trend corresponds to PREM being 1 per cent slow in this case. We will show now that these degree zero estimations can be biased and the problem is best illustrated for 40 s Love waves. The models of ETL97 and VHW99 give spherical averages of -0.05 and -0.14 per cent, respectively, whereas we have a spherical average of +0.09 per cent. Looking at these results we cannot decide whether PREM is on average too fast or too slow. The reason for these discrepancies is the sampling of the Earth by seismic waves. We made different measurements of degree zero, as shown in Fig. 10. Using synthetic data calculated for model VHW99, we find for a random sampling -0.14 per cent, for minor-arc sampling -0.1 per cent and for minor- and major-arc sampling +0.006 per cent. The random sampling comes quite close to a homogeneous ray coverage on the sphere and we find of course the same value as given by the model. For minor-arc sampling, as used in the study of VHW99, we find a somewhat smaller value. Finally, a minor- and major-arc coverage changes the sign of the recovered spherical average. This is easily under-

stood if one considers the structure in this model (Fig. 2, bottom). The Love wave 40 s phase velocity distribution is dominated by a continent-ocean signal, where oceans are faster and continents slower than average. Minor arcs mainly cover the Northern Hemisphere, where most of the continents are located. Adding the major arcs contributes a lot of faster structure due to the Southern oceans, which changes the sign of the spherical average. This is a real example of spectral leakage (Trampert & Snieder 1996), where ocean-continent structure (degrees 2-6) contaminate degree zero through improper sampling. Originally, spectral leakage was identified in cases where certain degrees of real structure were not permitted in the model parametrization. Here, the low degrees that contaminate are present in the parametrization, but owing to damping, the full signal is not allowed to map onto these degrees, and hence leakage towards degree zero occurs. Although the spherical average of 40 s Love waves is small, spectral leakage can be as large as the spherical average itself in this case. We checked systematically Love and Rayleigh phase velocity models between 40 and 150 s and found that spectral leakage can introduce errors between 0.1 and 0.2 per cent in the determination of degree zero. We suggest thus that in the next generation of spherical models, spectral leakage should not be ignored. We routinely use Laplacian damping in the construction of our phase velocity models. Next to full anti-leakage calculations, which are quite tedious, this is the most effective way to counteract spectral leakage.

The question arises whether the observed scatter between the different model predictions is due to aspherical differences or disagreements in degree zero. To investigate this, we made simulations giving all models the same degree zero. At the shortest periods, the scatter remained unchanged, suggesting that aspherical structure was different in the models considered. At the long-period end, the scatter was reduced, indicating that degree zero is the main culprit. This is easily understood if we recall that the strength of the heterogeneous part diminishes with increasing period.



Figure 10. Gain of the Rayleigh 40 s phase velocity model of this study over PREM as a function of distance. The solid line is the linear regression of the data corresponding to the individual paths. The slope is a direct measure of degree zero present in the model.

6 CONCLUDING REMARKS

We constructed a new phase velocity model and presented a new assessment technique that is entirely objective as it compares model predictions directly against raw data. The predictions from different models show increasing scatter with decreasing period. At the longest periods considered here, all predictions fall within 10° of each other. This is quite reassuring, and we furthermore found that the differences were mainly due to degree zero. Between 80 and 150 s, predictions for all paths are within a cycle of the data. At increasingly shorter periods, predictions within a cycle are distance-dependent but considerable gains over PREM are still achieved (up to 3.5 cycles). At these short periods, the main reason for differences in model predictions lies in the aspherical parts of the models. Where smaller than 90°, the mean phase-shifts measured here are reliable estimates of the phase errors in the data used for the construction of the models themselves. Degree zero determinations of current phase velocity models are prone to spectral leakage. Without precaution, for Love and Rayleigh waves between 40 and 150 s, spectral leakage can introduce biases between 0.1 and 0.2 per cent with respect to PREM.

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