Group Velocity Tomography of South America and the Surrounding Oceans

Oleg Vdovin[1] vdovin@lemond.colorado.edu
José A. Rial[2] jar@wave.gphys.unc.edu
Anatoli L. Levshin[1] levshin@lemond.colorado.edu
Michael H. Ritzwoller[1] ritzwoller@lemond.colorado.edu

[1]Department of Physics and CIRES
Campus Box 390
University of Colorado
Boulder, CO 80309-0390
phone: 303-492-5243
fax: 303-492-7935

[2]Department of Geology
319 Mitchell Hall
University of North Carolina, Chapel Hill
Chapel Hill, NC 27599-3315

South American tomography web site at:

http://phys-geophys.colorado.edu/geophysics/south_america.dir/south_america.html

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Abstract

This paper presents the results of a study of the dispersion characteristics of broadband fundamental surface waves propagating across South America and the surrounding oceans. Broadband waveform data from about 765 events from 1977 through 1996 recorded at 48 individual stations have produced 7000 Rayleigh wave and 4800 Love wave dispersion curves. We present group velocity maps from 20 s to 150 s period for Rayleigh waves and from 20 s to 100 s for Love waves. Measurement uncertainties estimated from cluster analyses average about 0.025-0.030 km/s, but are larger for the short period Love waves. We estimate surface resolving kernels and show that average resolution across South America is about $6° - 7°$ for Rayleigh waves and $7° - 9°$ for Love waves below 100 s period but degrades at longer periods and near the periphery of the maps. The estimated maps produce a variance reduction relative to the Preliminary Reference Earth Model (PREM) of more than 90% at short periods and is more than 80% for Rayleigh and Love waves below about 100 s period but reduces to about 70% at longer periods. Synthetic experiments are presented to estimate the bias caused by azimuthal anisotropy using the recent global model of Trampert and Woodhouse (1996). This bias, which is worse for Love than Rayleigh waves, may be substantial (up to 5% in some places), but it is expected to be mostly off the continent and is not correlated with the main features that we interpret. Many known geological and tectonic structures are observed in the group velocity maps. Of particular note are the signatures of sedimentary basins (e.g., Maturin-Llanos Basin, Marañon-Ucayali-Madre de Dios complex, Chaco-Tarija Basin, S. Paraná Basin, the basins in the W. Caribbean and W. Gulf of Mexico), variations in crustal thickness (e.g., Andes, Altiplano, Brazilian Highlands), continental roots (e.g., Guyana and Guaporé shields, Sao Francisco craton), and the Galapagos thermal anomaly. Comparison of the estimated group velocity maps with those predicted by CRUST5.1/S16B30 is qualitatively good, but there are significant differences in detail that provide new information which should help to calibrate future crustal and upper mantle models of South America.

1. Introduction

Within this decade there have been vigorous efforts to produce new information about the structure of the South American crust and lithosphere. These efforts have concentrated largely on the use of data from several temporary regional seismic networks; namely, the Broadband ANdean JOint (BANJO) experiment in the central Andean Cordillera of Bolivia and northern Chile (1994-1995), the Seismic Exploration of the Deep Altiplano (SEDA) experiment (1994-1995), and the Brazilian Lithosphere Seismic Project (BLSP) in south-central Brazil (1992-1995). Data from these networks have provided new information about the crust (e.g, Zandt *et al.*, 1995; Beck *et al.*, 1996, 1997), lithosphere (e.g., James *et al.*, 1993; James and Snoke, 1994; Myers *et al.*, 1995; Snoke and James, 1997; Van der Lee *et al.*, 1997), and sub-lithospheric upper mantle (e.g., James, 1994; Clark *et al.*, 1995; VanDecar *et al.*, 1995; James and Assumpção, 1996) between 15°S-20°S latitude, a region that encompasses some of the most striking physiographic features on the South American continent, including the central Andean Cordillera of Bolivia and Northern Chile, the Altiplano, the Chaco and Paraná Basins, and the Sao Francisco Craton. In addition, teleseismic studies (e.g., Dorbath *et al.*, 1993; Comte *et al.*, 1994; Engdahl *et al.*, 1995, 1997) and refraction surveys (e.g., Wigger *et al.*, 1994) along the central Andes have provided complementary information about the crust and subducting lithosphere.

Even with these efforts, the seismic structure of the crust and upper mantle underlying South America remain rather poorly characterized as a whole, particularly in comparison with North America and Eurasia. This is partially ameliorated by the improved quality and resolution of global-scale studies such as recent phase velocity maps (e.g., Trampert and Woodhouse, 1995; Ekström *et al.*, 1997) and new models of the crust (e.g., Nataf and Ricard, 1996; Mooney *et al.*, 1997) and upper mantle (e.g., Zhang and Tanimoto, 1993; Grand, 1994; Su *et al.*, 1994; Trampert and Woodhouse, 1996; Laske and Masters, 1996; Li and Romanowicz, 1996; Masters *et al.*, 1996). Nevertheless, the deficit of information at sub-global resolutions outside of the band between 15°S and 20°S means that continental scale surface wave studies, which have been very successfully applied to North America (e.g., Alsina *et al.*, 1996; Van der Lee and Nolet, 1997a, 1997b) and Eurasia (e.g., Snieder, 1988; Curtis and Snieder, 1997; Griot and Montagner, 1997; Ritzwoller *et al.*, 1997; Ritzwoller and Levshin, 1998), should produce valuable information about the crust and lithosphere under much of South America. Surface wave studies dedicated to South America date back to James (1971a), which analyzed phase velocity curves for a few paths in and adjacent to the Andes. More recently, Souza (1995) used estimates of a number of surface wave dispersion curves to estimate crustal shear velocities in the continental shelf off of south-eastern Brazil and Snoke and James (1997) used BLSP data to constrain the crustal velocity structure beneath the Chaco and Paraná Basins.

The purpose of this study is to present the dispersion characteristics of broadband Rayleigh (20-150 s) and Love (20-100 s) waves propagating across all of South America and the surrounding oceans. These results are presented as group velocity maps which represent the local group velocity of a Rayleigh or a Love wave at each period. The main motivation for the study is that these group velocity maps provide new constraints on the shear velocity structure of the crust and uppermost mantle beneath South America as well as the location of internal boundaries such as the Moho. These maps display better resolution and should be more reliable (lower variance and bias) than globally estimated dispersion maps or such maps computed from current global models of the crust and mantle. They should also help to calibrate future generations of global dispersion maps and seismic models, to provide transportable data to be used in future inversions for the shear velocity structure of South America, and to help provide a larger scale context for further regional studies.

The methods of dispersion measurement and tomography that we employ here were applied initially to Eurasia and are discussed in detail by Ritzwoller and Levshin (1998) (hereinafter referred to as RL98) and Ritzwoller *et al.* (1997) (hereinafter referred to as RLRE97). Two innovations are introduced here, we have developed a new method of resolution and bias analysis and we have corrected for group source time shifts (Levshin *et al.*, 1998). South America is a more challenging target for surface wave dispersion studies than Eurasia. The quantity of continental seismicity is significantly lower and South America remains more poorly instrumented than Eurasia. Therefore, at least until the station distribution across the South American continent is improved substantially, surface wave studies across the entirety of South America must be based, in part, on the use of earthquakes and stations outside of the continent. Figure 1 displays the locations of the 765 earthquakes and the 48 stations (GSN, GEOSCOPE networks) used in this study. Most of the earthquakes occurred after 1987 and have $M_s \geq 5.5$, but we also processed some smaller earthquakes on the continent and high southern latitude events from 1977-1987. Nevertheless, only about 213 earthquakes and 15 stations are located on or very near to the South American continent (13°N - 56°S, 82°W - 34°W).

Many notable features of the South American crust and lithosphere are observed in the group velocity maps presented here. These include sedimentary basins at short periods (< 30 s), crustal velocity and thickness variations at intermediate periods ($\sim$ 50s), and continental roots beneath cratons and shields at the long period end of the study (≥ 100 s). Since the deepest penetrating waves of this study (Rayleigh, 150 s) do not sample below about 200 km depth, this study cannot resolve deep-seated sublithospheric structures such as the hypothesized fossil plume beneath southeastern Brazil (VanDecar *et al.*, 1995) or subducting slabs (e.g., Engdahl *et al.*, 1995, 1997).

The outline of the paper is as follows. Section 2 presents a brief discussion of the methods of measurement and tomography. The reader is referred to RL98 for a more complete treatment. We discuss resolution and bias in Section 3. In particular, we present a new method of estimating resolution and bias and investigate the bias caused by azimuthal anisotropy on our isotropic group velocity maps. Preliminary interpretation of the group velocity maps is the subject of Section 4.

2. Data, Measurement, and Surface Wave Tomography

We obtained group travel time, phase travel time, and spectral amplitude measurements by use of an interactive Frequency-Time Analysis (FTAN) method (e.g., Dziewonski *et al.*, 1969; Landisman *et al.*, 1969; Levshin *et al.*, 1972; Cara, 1973; Feng and Teng, 1983; Russell *et al.*, 1988; Levshin *et al.*, 1992; RL98). Group velocity and phase velocity are computed from the group and phase travel times and the distance between the receiver and the CMT location (Dziewonski *et al.*, 1981). Distances are measured on the reference ellipsoid and all times are referenced to the CMT time. RL98 describes the measurement procedure in detail. We measured about 7000 Rayleigh wave and 4800 Love wave dispersion curves using FTAN. Figure 2 displays the number of measurements for Rayleigh and Love waves as a function of period and Figure 3a shows the average path length of the measurements as a function of period and wave type. An analyst oversees every measurement.

The resulting data set exhibits considerable redundancy, which allows for consistency tests, outlier rejection, and the estimation of measurement uncertainties. We "cluster" and compare measurements from similar paths following the procedure described by RL98. About one-third of all measurements fall within some cluster. We produce a single measurement for each cluster at each period and wave type. Clear outliers are identified as part of the cluster analysis because they disagree with other measurements from similar paths. Figure 3b displays the standard deviation of the measurements within each cluster averaged over all clusters. We interpret this statistic as the average measurement uncertainty. Uncertainties are largely independent of period and average between about 0.025 km/s and 0.030 km/s, except for Love waves below about 40 s where uncertainties rise to about 0.040 km/s. The rise in uncertainties for short period Love waves is probably caused by scattering due to relatively small scale features which makes these measurements difficult to obtain reliably. The decreased uncertainty for the 20 s Love wave results from the fact that most 20 s Love wave measurements come from paths confined to the continent. This simplifies the paths considerably and tends to reduce scattering. We follow RL98 in assigning data weights from the cluster uncertainties.

We use the algorithm of Ditmar and Yanovskaya (1987) and Yanovskaya and Ditmar (1990) to construct the group velocity maps. (See also Levshin *et al.*, 1989, Ch. 6.) This method estimates a group velocity map $U(\theta, \phi)$ at each period and wave type by attempting to minimize the following penalty function

$$\sum_{i=1}^{N}[w_i(t_i^{obs} - t_i^{pred})]^2 + \lambda \int_S |\nabla U(\theta, \phi)|^2 dA, \tag{1}$$

where

$$t_i^{pred} = \int_{p_i} U^{-1}(\theta, \phi) ds + \delta t_i^{source}. \tag{2}$$

Here, p_i represents the i-th wave path. w_i is the weight associated with the i-th path which is a function of the measurement uncertainty, the number of paths within the cluster that generated the measurement, and a qualitative grade supplied by the analyst during measurement. t_i^{obs} and t_i^{pred} are the observed and the predicted group travel times along the i-th path and S is the region on which the inversion is performed. Choosing different values of the trade-off parameter, λ, changes the trade-off between the fit to the data and the 'smoothness' of the resulting group velocity map. 'Smooth' here is defined in terms of the spatial gradient of the model. The trade-off parameter is chosen by analyzing misfit and the visual characteristics of the resulting maps. Typically, we choose λ to produce a slightly underdamped map and, because the penalty function in equation (1) does not include a second-spatial gradient term, we smooth each map a posteriori with a 2D Gaussian spatial-smoothing filter. Both the resolution analysis and inversion occur separately for each period and wave type. In equation (2), we assume that each wave path, p_i, is along the great-circle linking the source and receiver and δt_i^{source} is the group time perturbation introduced by a source group time shift (e.g., Knopoff and Schwab, 1968; Levshin *et al.*, 1998) which is computed from the CMT.

3. Resolution and Bias

3.1 Errors Related to Path Coverage and Damping

Irrespective of theoretical errors, the resolution of the data set is dominantly a function of the path distribution (density of paths, azimuthal coverage, the average path length locally) and the weighting and damping applied during inversion. Path distribution depends strongly on geographical location, period, and wave type and derives principally from the nature of the data set used. Compared with Eurasia, South American path densities are lower, path lengths are longer, and azimuthal coverage is less uniform. Thus, we expect lower resolutions across South America than those presented in RL98 and RLRE97 for Eurasia. In particular, the longer paths make the shorter period measurements ($< 30s$) more difficult to obtain, and the South American maps below 30 s are probably less reliable than the maps at the same periods across Eurasia.

Figures 4a and 4b present examples of the path coverage, where we plot the path density, defined as the number of paths after the cluster analysis that intersect each square two degree cell ($\sim$50,000 km^2). Path densities are significantly higher for Rayleigh than for Love waves. Coverage is sufficient to construct Rayleigh wave maps up to 150 s period but only to about 100 s for Love waves. Due to our use of data from North and Central American stations, path densities are also higher in the northern (e.g., Caribbean, Central America, Cocos Plate) and western (e.g., Eastern Nazca Plate) parts of the studied area and are lowest in the east and southeast. In particular, Love wave path coverage in the south-western Atlantic Ocean is very poor.

Path density is only simply related to resolution if azimuthal coverage is relatively homogeneous. Unfortunately, azimuthal distribution is notoriously inhomogeneous across much of South America and, for this reason, it would be useful to estimate resolution and bias. Checker-board tests are commonly used to estimate resolution (e.g., RL98), but Lévêque *et al.* (1993) and others have argued persuasively that the results of these tests can be misleading and our own experience indicates that they are simply difficult to interpret unambiguously. Moreover, they do not provide information about the source or nature of the irresolution; e.g., if a cell is unresolved due to poor local coverage or due to 'leakage' from another nearby cell.

It is more useful to estimate spatial averaging kernels $K(\theta, \phi; \mathbf{r}_0)$ (e.g., Backus and Gilbert, 1968) for each point ($\mathbf{r}_0 = (\theta_0, \phi_0)$) on the map

$$\frac{\delta \hat{U}(\mathbf{r}_0)}{U} = \int K(\theta, \phi; \mathbf{r}_0) \frac{\delta U(\theta, \phi)}{U} d\Omega, \tag{3}$$

where $\delta \hat{U}(\mathbf{r}_0)/U$ is the estimated group velocity variation at $\mathbf{r}_0$ and $\delta U(\theta, \phi)/U$ is the 'true' group velocity variation across the entire map. The averaging kernel may be estimating from the resolution matrix, but we prefer to infer it a posteriori by determining our ability to retrieve a test function centered at $\mathbf{r}_0$. The test function we choose is a spatially-limited 2D Gaussian of fixed input amplitude, A, and known width, 2σ, whose centroid is moved to various positions, $\mathbf{r}_0$, around the region of study. We set A equal to a 10% velocity perturbation and σ is the standard deviation of the Gaussian, which we choose in the way described below. Let us call the input test function at position $\mathbf{r}_0$, $G^{in}(\theta,\phi;(A,\sigma,\mathbf{r}_0)) = \delta U(\theta,\phi)/U$. We compute synthetic travel times through this function and invert for the estimated velocity map, $M(\theta,\phi;G^{in}) = \delta\hat{U}(\theta,\phi)/U$, with the same ray paths, weighting, and damping as used at each period and wave type with the real data. If the input Gaussian is sufficiently delta-like, the estimated map is a good approximation to the averaging kernel, $M(\theta,\phi;G^{in}) \sim K(\theta,\phi;\mathbf{r}_0)$. In practice, it suffices if the width of G^{in} is somewhat less than the resolution at $\mathbf{r}_0$.

The averaging kernel at $\mathbf{r}_0$ can be characterized in a number of ways. We fit a 2D Gaussian, $G^{fit}(\theta,\phi;(\hat{A},\hat{\sigma},\hat{\mathbf{r}}))$, of variable amplitude, $\hat{A}$, width, $2\hat{\sigma}$, and centroid, $\hat{\mathbf{r}}$, to K and then identify the width of the best fit Gaussian with "resolution" and the difference between the centroids of the input and fit Gaussians with the "bias", $\hat{\mathbf{r}} - \mathbf{r}_0$. Figure 5 displays the results of attempts to estimate resolution and bias using this method at three diverse geographical points for the wave paths, data weights, and damping applied in the 150 s Rayleigh wave inversion. The small circles indicate the widths (2σ) of the input 2D Gaussian test functions (G^{in}) and the grey-shaded patterns are the estimated averaging kernels. The larger circles represent the widths ($2\hat{\sigma}$) of the Gaussians fit to the averaging kernels. The width of the input Gaussian is chosen so that the estimated averaging kernel is relatively well fit by the fit Gaussian. We use the criterion that the root-mean-square (rms) of the difference between the estimated surface and the fit Gaussian is about 40% of the rms of the estimated surface itself. That is, $\mathrm{rms}\left[K - G^{fit}\right]/\mathrm{rms}[K] \sim 40\%$.

The fidelity of the resolving kernels with respect to the input Gaussians is highly variable. In Figure 5a, where the test function is located on the continent near to the Colombian-Brazilian border, the resolving kernel is quite similar to the input Gaussian although its amplitude is approximately halved. The width of the fit Gaussian is increased to about 7° from 5° for the input Gaussian and the fit Gaussian is shifted to the south by about 1°. The fidelity of the resolving kernel in Figures 5b and 5c is worse, where the input Gaussians are located off the coast, near the Galapagos Islands and in the southwest Atlantic Ocean, respectively. This is reflected in poorer resolutions and larger bias in both cases. Resolution at the Galapagos is estimated to be 10° with a bias of 3° to the east-southeast and at the southwest Atlantic resolution is about 11° with bias of about 4° to the northeast.

We compile results such as those in Figure 5 to form maps of resolution and bias shown in Figures 6a and 6b. Not surprisingly, resolution is best where the majority of stations are located: on the continent, in the Caribbean, and in Central America. Resolution optimizes at about 40 s period for both Rayleigh and Love waves, although at any period resolution is typically better for

Rayleigh than for Love waves. Figure 7 displays the average resolution across the continent as a function of period and wave type. We estimate average resolution on the continent below 100 s period at between 6°-7° for Rayleigh waves and between 7°-9° for Love waves. Resolution degrades quickly for Love waves above 50 s and Rayleigh waves above about 100 s period. Bias can be a significant fraction of the estimated resolution at each point but is typically less than about half of the resolution. Regions of greatest bias, therefore, are those that display the worst resolution. Bias in the interior of South America is usually quite low, except in the far south particularly at long periods.

3.2 Theoretical Bias: Azimuthal Anisotropy

There are problems other than path distribution, weighting, and damping that affect the accuracy of the estimated maps which mainly derive from errors in approximations and assumptions. We call these problems 'theoretical errors'. There are three principal theoretical errors: (1) earthquake mislocations, (2) off-great-circle propagation, and (3) anisotropy. We have shown elsewhere that the effects of earthquake mislocations are not appreciable (RL98) away from source regions or in source regions with significant crossing paths that emanate from other source regions. Most of South America falls into one or the other of these two categories and, therefore, source mislocations are not likely to provide a significant error in the estimated group velocity maps except near the East Pacific Rise and the Mid-Atlantic Ridge. The second error, off-great-circle propagation, acts to smear the group velocity maps, but we believe that this smearing is typically within the stated resolution of the maps. This leaves anisotropy as a potential source of appreciable error. Because the azimuthal coverage is nonuniform across most of the region of study and because most paths possess a significant path-fraction in the oceans, this study should be more sensitive to biases caused by azimuthal anisotropy than our previous studies of Eurasia.

We have performed synthetic experiments designed to estimate the nature and magnitude of the bias to isotropic group velocity maps caused by azimuthal anisotropy. A study of the expected bias caused by azimuthal anisotropy on global phase velocity maps has been performed by Larson *et al.* (1997). To the best of our knowledge, there are no group velocity maps for azimuthal anisotropy across the studied region to be used in these experiments. There are, however, global phase velocity maps of azimuthal anisotropy (e.g., Montagner and Tanimoto, 1991; Trampert and Woodhouse, 1996). Because the radial sensitivity kernels of phase velocity are different from those of group velocity, the use of phase velocity azimuthal anisotropy maps as proxy for the group velocity maps is not optimal, but we do not expect that the general patterns of anisotropy should be greatly different. The global phase velocity maps should, therefore, provide useful information about the distribution and nature of azimuthal anisotropy in group velocity. However, because the geographical patterns of anisotropy and the local orientations of fast and slow axes are much better determined than the amplitudes (Trampert, personal communication, 1997), we should probably pay more attention to the distribution rather than the amplitude of bias.

Figure 8 displays a preliminary version of the model of the 2ψ component of azimuthal anisotropy produced by Trampert and Woodhouse (1996) for 40 s and 80 s Rayleigh and Love wave phase velocity. The patterns of anisotropy are similar although certainly not identical for Rayleigh and Love waves, but the amplitudes are uniformly larger for Love waves. Because phase velocity kernels sample deeper at each period than group velocity kernels, these maps perhaps correspond best to group velocities between 50 - 70 s and 100 - 125 s, respectively. To estimate the bias caused by these maps of azimuthal anisotropy, we use the paths, weighting and damping from the real inversions and apply them to synthetic travel times computed through the anisotropic maps. We use the paths, damping, etc. for the 50 s and 100 s Rayleigh and Love waves applied to the respective 40 s and 80 s maps in Figure 8. The resulting tomographic maps represent the isotropic perturbations that would result from the travel time anomalies imparted by the anisotropic structures. We use both the 2ψ and 4ψ components of Trampert and Woodhouse's model, but it is the 2ψ component that produces most of the bias.

The maps representing the bias caused by azimuthal anisotropy are shown in Figure 9. Bias levels of 1%-2% are common, but the bias is most severe off the continent, is stronger for Love than for Rayleigh waves, and is worse at long periods. This is because azimuthal coverage is more homogeneous on the continent, at short and intermediate periods, and for Rayleigh waves. Biases of several percent should not be a great concern at periods below about 50 s, but at and above 100 s they may correspond to a significant fraction of the observed anomalies. However, comparison with the observed maps in Figures 11a-11c does not reveal obvious correlations between the locations of bias and the major observed group velocity features. The major features that appear in Figures 11a-11c and the discussion of these features in section 5, therefore, should not be affected greatly by errors caused by azimuthal anisotropy. Nevertheless, the magnitude of the bias is appreciable and simultaneous estimation of isotropic and azimuthally anisotropic velocities is advisable in the future.

4. Group Velocity Maps

Using the tomographic method described in section 2, we construct group velocity maps for Rayleigh waves at the following periods: 20, 25, 30, 35, 40, 50, 60, 70, 80, 90, 100, 125, 150 s. The same periods are inverted for Love waves, except Love wave group velocity maps do not extend past 100 s period. Group velocity curves that represent the average of the group velocity maps across the studied region (latitude bounds: 60°S - 30°N; longitude bounds: 120°W - 0°W) for Rayleigh and Love waves are plotted in Figure 10. This figure also presents the group velocity curves predicted from PREM (Dziewonski and Anderson, 1981) and the average of a hybrid aspherical model of the crust (CRUST5.1: Mooney *et al.*, 1997) and mantle (S16B30: Masters *et al.*, 1996) across the studied region. Since the region of study is predominantly oceanic, these averages are more typical of oceanic than of continental crust and upper mantle. This is most clearly indicated by the high velocity short period Love wave. The laterally inhomogeneous model (CRUST5.1/S16B30) predicts average group velocities much better than PREM with a single exception. The group velocities predicted by CRUST5.1/S16B30 are too low for long period Love waves. This principally results from the fact that CRUST5.1/S16B30 is an isotropic model and the observed dispersion curves show clear signs of polarization anisotropy (transverse isotropy). The model CRUST5.1/S16B30 provides a better fit to long-period Rayleigh waves than Love waves. It would be difficult for any isotropic model to fit both types of waves well.

Figures 11a - 11c present a sampling of the estimated group velocity maps. These maps represent lateral group velocity variations relative to the regional average of Figure 10. The maps are smoothed a posteriori by applying a 2D Gaussian smoothing filter whose width at the e^{-1} point is 2.5°. The preliminary interpretation of these maps is the subject of section 5.

Figure 12 shows the fit to the measured dispersion curves delivered by the estimated group velocity maps expressed as two different measures of misfit. The first is variance reduction relative to PREM:

$$\text{Variance Reduction} = 1 - \frac{\sum_i (U_i^{obs} - U_i^{pred})^2}{\sum_i (U_i^{obs} - U_0)^2}, \tag{4}$$

where i is the path index, U_i^{pred} is the predicted group velocity for path i through the estimated group velocity map, U_i^{obs} is the measured group velocity for path i, and U_0 is the reference group velocity from PREM. The second measure of misfit is rms group velocity residual:

$$\text{RMS Velocity Misfit} = \left(\frac{1}{N} \sum_i (U_i^{obs} - U_i^{pred})^2 \right)^{1/2}, \tag{5}$$

where N is the number of measurements. For each measure of misfit, comparison is made between the estimated group velocity maps and those predicted by the hybrid model CRUST5.1/S16B30.

Variance reductions relative to PREM are more than 90% at short periods and more than 80% at all periods below about 100 s for both Rayleigh and Love waves. Variance reductions typically reduce at long periods because lateral variations in the group velocity maps have smaller amplitudes at longer periods and there is less signal to fit and therefore there is less reduction in variance. Consistent with this is the fact that at the short-period end of the spectrum, absolute misfit is the greatest even though variance reduction continues to be high. The degradation in misfit below about 40 s period is probably due to off-great-circle propagation and scattering caused by sedimentary basins and other small-scale features. The onset of this degradation in variance reduction is at longer periods for Love waves since they sample more shallowly than Rayleigh waves at every period. For example, the 30 s Love wave is more strongly sensitive to sedimentary features than the 30 s Rayleigh wave. Above 50 s period the absolute misfit is quite flat for both Rayleigh and Love waves, ranging between about 0.04 and 0.06 km/s.

Measurement uncertainties displayed in Figure 3b are also flat above 50 s period and average between 0.025 and 0.030 km/s. Thus misfits on average are about at the 1.5σ-2.0σ level relative to the average measurement uncertainties. The uncertainties reported in Figure 3b are estimates of the repeatability of the measurements, typically for events in the same region. This uncertainty in velocity results principally from difficulties in measuring the time of the arriving wave packet accurately. The remaining misfit between the measurements and the estimated group velocity maps over and above that caused by measurement errors is likely to result from theoretical errors which include event mislocations at all periods, off-great-circle propagation at the short-period end of the spectrum, and azimuthal anisotropy at longer periods.

Our confidence in the longest period maps is not as high as at shorter periods; in particular, above 150 s for Rayleigh waves and at and near 100 s for Love waves. There are two key reasons. First, accurate long period measurements can only be obtained from large earthquakes ($M_s \geq\sim 6.0$), and the distribution of these events around South America is not uniform and their numbers are not large. On average over the period of study, only about 11 events per year occurred with $M_s \geq 6.0$ and about 4 per year with $M_s \geq 6.5$ in the studied region. These events constitute less than 10% of the entire data set, which means that significant regions of South America are not well covered at long periods, as the path coverage maps in Figures 4a and 4b demonstrate. Global scale studies that utilize both major and minor arc wave packets demonstrate more uniform coverage across the continent (e.g., Trampert and Woodhouse, 1995, 1996; Ekstrom *et al.*, 1997) and should do a better job at these very long periods. Second, certain theoretical errors result in larger relative effects at long periods. Examples are event mislocation, as discussed in RL98, and azimuthal anisotropy. In summary, the advantages of continental-scale studies over global studies break down at periods beyond about 150 s for Rayleigh waves and about 100 s for Love waves.

5. Discussion

Because we believe that the corrupting effects of theoretical errors are at levels below the most striking features on the observed group velocity maps, it is worthwhile identifying and discussing some of the features that appear to have clear structural causes. However, because group velocity sensitivity kernels are complicated functions of radius, interpretation here should be considered preliminary in nature. In this interpretation, it is important to keep in mind path coverage and resolution (Figures 4 and 6) as well as the discussion of the effects of theoretical errors, in particular azimuthal anisotropy, in section 3.2 (Figure 9).

The interpretation of group velocity maps is based on the nature of group velocity sensitivity kernels which can be summarized by several rules of thumb. Group velocity sensitivity kernels are more complicated than phase velocity sensitivity kernels (e.g., they change sign) and are compressed nearer to the surface at each period. At a given period, Love waves sample more shallowly than Rayleigh waves and sensitivities for both types of waves compress toward the surface as period decreases. Consequently, everything else being equal, the best probe of sedimentary basins should be the shortest-period Love waves, which here is 20 s. Rayleigh waves between 30 s and 75 s are strongly sensitive to crustal velocities and thicknesses, and the 50 s Rayleigh wave map, to a fair approximation, can be seen as inversely related to Moho depth. That is, for a 50 s Rayleigh wave, low velocities result largely from thickened crust. Love wave sensitivity to crustal thickness maximizes between 80 s and 100 s period. At longer periods, the sensitivity of waves to crustal velocities and thicknesses diminishes. The uppermost mantle (80-150 km) is well represented in the 100 s Rayleigh wave map. The 150 s Rayleigh waves provide somewhat deeper sampling down to about 200 km.

Place names and the outlines of geological and tectonic structures are presented in Figure 13. Figure 14 presents Rayleigh wave group velocity maps predicted by the model CRUST5.1/S16B30 at a variety of periods. The model CRUST5.1 is defined on a 5° grid, and the blocky nature of the predicted group velocity maps results from the grid defining the model.

5.1 Crustal Structures

5.1.1 Sedimentary Basins

Everything else being equal, the best indicator of sedimentary basins should be the 20 s Love wave. However, off-pure-path propagation corrupts Love waves more than Rayleigh waves and the Rayleigh wave resolution is better at 20 s period. Thus, the best indicator of the location and nature of sedimentary basins across South America is probably the 20 s Rayleigh wave on Figure 11a. This map displays four strong low velocity anomalies on the South American continent in the following locations: (1) western Venezuela and eastern Colombia, (2) northeastern Peru and eastern Ecuador, (3) central and western Bolivia, and (4) eastern Paraguay and far southern Brazil. The first of these anomalies is coincident with the Llanos and Maturin basins, the second with the Marañon-Ucayali-Madre de Dios sedimentary complex, the third with the Altiplano and the Chaco-Tarija Basin, and the fourth with the southern Paraná Basin. Off the continent, a very strong low velocity anomaly appears in the western Gulf of Mexico associated with the large accumulation of sediments in that region. Low velocity anomalies appear in the western Caribbean and far eastern Brazil on the 20 s Love wave map in Figure 11c and appear to be associated with the western Caribbean and Maranhão Basins.

The sedimentary basins across South America are of two general types: those related to the contractional tectonics of the Andes (foreland basins) and those formed by extensional tectonics in the shield areas (rift basins and intracratonic basins). The foreland basins are mainly of Miocene age and occupy the eastern margin of the Andes, or sub-Andean foreland, from the southern tip of the continent to Venezuela. Typical examples of these are the Llanos Basin (Colombia-Venezuela) and the Marañon Basin (Peru). The Marañon-Ucayali Basin is part of a large structural depression that includes the Oriente basin in Ecuador, the Putumayo in southern Colombia and Madre de Dios in far southern Peru and northwestern Bolivia (Mathalone and Montoya, 1995). A prototypical member of this group, the Ucayali basin, described in detail by James and Snoke (1994), has a maximum depth of 8 km filled with a sequence of Tertiary molasse deposits consisting of low-velocity sediments which overlie a thickened ($\sim$45 km) crust. The Altiplano is also covered by up to 10 km of soft sediments (Wigger *et al.*, 1994). Most of the foreland basins produce observable group velocity anomalies.

The intracratonic basins of South America are commonly extensional basins caused by intra-continental rifts that cut across the Precambrian shields. Examples are the Paraná, Amazon, Maranhão, and Neuquén basins. Intracratonic basins do not produce as strong of group velocity anomalies, although the Chaco-Tarija Basin and the southern Paraná Basin are observable. The Paraná Basin is filled with Palaeozoic through Mesozoic sedimentary rocks and flood basalts. (James and Assumpção, 1996). The effect of the sediments on group velocity in the Paraná Basin is undoubtedly obscured by the flood basalts. However, the Amazon Basin cannot be distinguished at all in the tomographic maps. The 2,500 km long east-west Amazon basin is an intracratonic rift which may have originated in Cambrian time as a result of the Pan-African continental collision process (Burke and Lytwyn, 1993; Brito Neves and Cordani, 1991). The basin is filled with a thick sequence (4-6 km) of sedimentary rocks ranging in age from Late Ordovician to Carboniferous. These are in turn covered by thinner and more recent deposits accumulated after the opening of the Atlantic in a sequence that thickens (1-2 km) toward the mouth of the Amazon river.

It would be reasonable to conclude that the intracratonic basins are not as well represented on the group velocity maps because the accumulated sediments are older and therefore faster, and in some cases the sediments are penetrated by flood basalts. This is consistent with the argument RL98 for the older basins in Eurasia. However, because the intracratonic basins lie significantly east of the seismic source region along the Andes, it may be the case that we do not observe them simply due to poor resolution. Figure 15 presents a resolution analysis that aims to determine which basins are resolvable. The estimated structure, using the ray paths, data weighting and damping from the 20 s Love and Rayleigh waves, appears in Figures 15b and 15c. In this simulation we use the 3 km sedimentary contours from the model 3SMAC (Nataf and Ricard, 1996) and we assign the sediments a uniform 10% anomaly. This figure shows that, everything else being equal, the best resolved basins should be the foreland basins in the west as well as the Chaco-Tarija and Paranã Basins. We have little ability to resolve the Amazon or Maranhão Basins, particularly for the Rayleigh wave, and have virtually no ability to resolve sediments off the northeast coast of South America whose presence in the simulation appears to shift onto the continent. Thus, it appears that certain intracratonic basins may not appear on our short period maps simply because we cannot resolve them.

Inspection of Figure 14 reveals that the 20 s Rayleigh wave map predicted by the model CRUST5.1/S16B30 presents only minimal anomalies on the South American continent. The strongest anomalies are in the Gulf of Mexico and off the northeastern coast of Brazil, due to thick sediments in these regions in CRUST5.1. The low velocity anomalies of Figure 11a do not appear on the 20 s Rayleigh map from CRUST5.1/S16B30 since most of the basins across South America are too small in size to appear on a gridded $5^\circ \times 5^\circ$ model.

5.1.2 Crustal Thickness Variations on the Continent

Moho depth is well revealed by the 50 s Rayleigh and Love wave maps of Figures 11a and 11c. The Rayleigh wave map is probably the more reliable of the two due to better path coverage, and generally higher signal-to-noise levels in the data. The 50 s Rayleigh wave map is dominated by an elongate low velocity anomaly that runs along nearly the entire western edge of South America following the Andes. Very low velocities extend north from the Altiplano from Bolivia into southern and central Peru and also appear in western Colombia. The strongest low velocity anomaly and most prominent localized feature of the 50 s Rayleigh and Love wave maps is coincident with the Altiplano. The amplitude of these low velocity features is related to the thickness of the crust, although crustal velocities, which may be particularly low in regions of active volcanism, also contribute.

The Altiplano is a high intermountain plateau located between the volcanically active Western Cordillera and the intensely folded Eastern Cordillera. Underlying the Altiplano region is the thickest section (~70 km) of the Andean crust (e.g., James, 1971b). Recent crustal thickness determinations (Beck *et al.*, 1996) indicate that under the Eastern and Western Cordilleras the crust is 70-74 km thick, near La Paz 65 km thick, and towards the southeast, under the Altiplano, and for at least 500 km, crustal thickness averages 70 km. Between the latitudes of 20°S and 26°S results of seismic refraction profiles, gravity and magnetotelluric surveys (e.g., Wigger *et al.*, 1994) indicate that the average P-wave velocity in the crust under the Western Cordillera and the Altiplano-Puna regions is anomalously low with the lower crust and upper mantle characterized by dramatically reduced seismic velocities, suggesting a mechanically ductile crustal segment (Wigger *et al.*, 1994) possibly representing magmatic underplating by silicic melt intrusions in the crust whose source is the ascending hot fluids from the dehydration of the subducting Nazca plate. The ductile crustal segment, some 500 km wide, may extend for over 1,000 km or more from the latitude of La Paz (~16°S) to the southern end of the Puna (~26°S), flanked on the west and east by the Pre-Cordillera and the Chaco, respectively, both underlain by rigid crust. Body wave spectra of earthquakes on the subducting slab recorded at the BANJO and SEDA seismic networks corroborate the existence of a strongly attenuating uppermost mantle under the Altiplano (Myers *et al.*, 1995), but whether the anomaly is localized or extended is not determined. Thus, it should not be surprising that the Altiplano, with a slow thickened crust and perhaps with underplating magmatism, manifests as a large, broad low velocity surface wave anomaly on the intermediate period group velocity maps (30 - 70 s). The Altiplano anomaly has also been recently observed in global Rayleigh and Love wave phase velocity maps (e.g., Ekström *et al.*,1997; Trampert and

Woodhouse,1995).

In the central Andes of Peru and the northern Andes of Colombia, the 50s Rayleigh and Love wave maps also display pronounced low velocity anomalies. In the Central Andes, the low velocity anomaly at 50 s is shifted to the west of the anomaly on the 20 s map consistent with the interpretation that the 50 s anomaly results dominantly from the mountain roots and the 20 s anomaly results from sediments accumulated in folds to the east of the Andes. A similar shrinking and westward shifting of the major low velocity anomaly with period occurs for the Altiplano region, where the broad, short period anomaly induced by the Chaco-Tarija Basin shifts west as period increases to eventually give way to a narrow and more pronounced low velocity anomaly, as longer surface waves sense the thickened crust. In the northern Andes, the 50 s low velocity anomaly is in the area where the Andes range trifurcates ($\sim 7°$N) to compose the west, central and eastern Cordilleras and consequently becoming almost as wide as the Altiplano. In addition to the low velocity anomalies associated with the Andes and the Altiplano, more subtle low velocity anomalies on the 50 s Rayleigh wave map are also associated with the Paraná Basin and the Brazilian Highlands.

Figure 16a and 16b present an analysis of the ability of the 50 s Rayleigh wave to resolve the group velocity signal caused by Moho topography underlying the Andes. The amplitude of the input anomaly is constant and, therefore, variation of the amplitude of the retrieved anomaly away from the 10% input amplitude results from poor resolution. In particular, Figure 16b shows that the amplitude of the estimated anomaly is somewhat lower in the narrower northern Andes. However, the amplitude of the low velocity anomaly on the 50 s Rayleigh wave map in the northern Andes is actually quite large. It is, therefore, likely that at least part of the amplitude variation of the Andes anomaly observed in Figure 11a results from variations in crustal thickness along the strike of the Andes.

Figure 14 shows the group velocity map for the 50 s Rayleigh wave for CRUST5.1/S16B30. Within the limitations of a gridded 5° model, there appears to be pretty good agreement with the observed 50 s map across much of the South American continent. The major exception is that CRUST5.1 misses the very low velocity anomaly under Colombia which is apparently caused by very thick crust there.

5.1.3 Features Not Observed: Massive Flood Basalts and Active Volcanic Centers

Along the Andes there are three active volcanic centers, none of which appears to produce distinctive velocity anomalies in our short period group velocity maps. The Altiplano is not an exception since, as discussed above, its associated anomaly results from a thickened, ductile crust. Partial melt may contribute to the observed low velocity anomalies under the Altiplano at intermediate periods, but if so it is probably from deep, not shallow, magmatism. On the other hand, the volcanic centers appear to block the propagation of shorter period (2-10s) Lg waves (Rial and Ritzwoller, 1997) indicating that their associated magma bodies must be small (and shallow) as compared to the wavelength of 20s Rayleigh or Love waves.

RL98 argued that the 20 s maps across Eurasia demonstrated the high velocity signatures of massive flood basalts. However, the flood basalts of the Paraná Basin do not clearly imprint the short period maps in Figures 11a and 11c here. It is likely that the accumulated sediment in the basin obscures the signal from the flood basalts on the short and intermediate period maps as it does in the Siberian Tunguska Basin (RL98).

5.2 Lithosphere and Upper Mantle

5.2.1 Cratons and Shields

The Proterozoic areas of South America comprise the Amazonian Craton, the Rio de la Plata Craton, and the Patagonian Massif. The Amazonian Craton itself is distinguished into the Guyana and Guaporé shields and the Sao Francisco Craton. In southeastern Venezuela and north-central Brazil the Guyana and Guaporé shields are separated by the east-west Amazon Basin. Together, these five regions represent parts of the Gondwana supercontinent that many believe have remained largely unaffected by tectonism for $\sim$600 Ma except for the disturbance caused by the opening of the south Atlantic in mid-Mesozoic times ($\sim$130 Ma), and the consequent eruption of large basaltic lava flows that covered the Paraná Basin.

The most prominent high velocity anomalies in our long period group velocity maps (e.g., Rayleigh 100 s) are caused by the presence of the Amazonian Craton. The Rayleigh wave group velocity maps display what appears as a somewhat differentiated intracratonic structure on the 80-150s Rayleigh wave maps, where three separated high group velocity nuclei may be distinguished, which appear to correspond to the Guyana shield, the Guaporé shield, and the Sao Francisco craton. At longer periods these features seem to partially merge or increase in size, perhaps indicating a less differentiated, persistent high velocity structure with increasing depth. Caution must be exercised, however, because at these long periods ray density decreases substantially and the overall resolution is only about 7° for the 100 s Rayleigh wave and worse for the 150 s Rayleigh wave. Resolution analyses such as those in Figure 16c and 16d indicate, however, that long period Rayleigh waves should be able to differentiate at least certain aspects of these features.

Unlike the northern Proterozoic areas, the Patagonian Massif does not appear on the long period Rayleigh wave maps as a high velocity anomaly. In fact, there is a low velocity anomaly associated with it, particularly for Love waves at periods greater than 60 s. The persistence of this feature and its location under a recognized cratonic region is puzzling.

In Figure 14, the 100 s Rayleigh wave group velocity map from the model CRUST5.1/S16B30 displays the Amazonian Craton as an undifferentiated structure with significantly lower amplitudes than observed in Figure 11b.

5.2.2 Oceanic Ridges and the Galapagos Hot Spot

The low velocity signature of oceanic ridges only appears unambiguously above about 30 s period for Rayleigh waves and 50 s period for Love waves on Figures 11a - 11c. Apparently, the high temperatures associated with ridges are simply too narrow in the crust and uppermost mantle to affect group velocities appreciably. It is only the larger scale temperature anomalies at greater depths that contribute to the observed maps. The imprint of the ridges can be seen clearly to the longest periods of this study, although ridge signatures lose some of their continuity and appear to begin to break up somewhat on the 150 s Rayleigh wave map.

The most spectacular low velocity feature on any of the maps is the anomaly coincident with the Galapagos Ridge. This anomaly is most pronounced on the long period Rayleigh wave maps and is probably caused by the thermal anomaly associated with the Galapagos hot spot. The 150 s Rayleigh wave map provides compelling evidence that the thermal anomaly associated with this feature extends significantly deeper than the high temperatures associated with other ridges. It is unlikely that the deep structure of the Galapagos thermal anomaly is extremely narrow, at least above about 250 km in depth. Otherwise, it would be unable to affect the 150 s Rayleigh wave as significantly as it does. However, the feature may be narrower than the low velocity anomaly that appears on Figure 11b for the 100 s and 150 s Rayleigh waves, because the resolution at these periods in this region is believed to be no better than $7^\circ - 10^\circ$. Figure 5b shows the results of a simulation in which a 6° wide 2D Gassian-shaped anomaly is input and estimated with the ray paths, weighting and damping for the 150 s Rayleigh wave. The width of the feature on the estimated surface is about 10°. It is not strongly changed if the input Gaussian is narrowed; the feature merely becomes more irregularly shaped. These considerations lead us to conclude that the Galapagos thermal anomaly in the top 250 km is probably not much less than 500 km in width. If there is a high temperature plume producing the thermal anomaly, it could be significantly narrower than the thermal anomaly that it has created. Seismic tomography may, therefore, see only the result of a plume and not the plume itself.

The mantle model S16B30 possesses the Galapagos anomaly, but the amplitude is significantly lower than observed here. Presumably this is because of greater damping applied at these wavelengths during the construction of the global models.

5.2.3 Features Not Observed: Back-Arc Anomaly, Continental Plume, Subducting Plate

Neither subducting lithosphere beneath or adjacent to the Andes (e.g., Engdahl *et al.*, 1995, 1997) nor the hypothesized fossil plume beneath south-eastern Brazil (VanDecar *et al.*, 1995) are observed in the long period maps in Figure 11b. This is probably just a sensitivity issue. The 100 s - 150 s Rayleigh waves sample too shallow (<200 km) to sense these deep features. In addition, these features may also be too small to be resolved easily.

Another feature that is not observed is a low velocity anomaly in the back-arc running the length of the Andes. This feature may be expected because a back-arc low velocity anomaly is prominent along the entire eastern edge of Asia. The source of the Asian anomaly is probably enrichment in volatiles of the upper mantle from the partial melting of the subducting slab and/or the response of the mantle (up-welling) to lithospheric extension in the back-arc. The cause of the Asian anomaly is unclear, but the lack of a South American anomaly should shed light on it since the anomaly's appearance or absence must result from differences in the geometry or setting of subduction between the two continents.

The Nazca Plate subducting beneath South America is younger, has a shallower dip angle, and is closer to the deep continental roots than the Pacific and Philippine Plates subducting beneath Asia. It is reasonable to assume that volatilization is occurring above the Nazca Plate. Although there are differences in the volatile content of the two slabs, it is not unreasonable from this to conclude that the Asian anomaly is dominantly caused by mantle up-welling rather than volatilization. The absence of a South American anomaly, therefore, means that the mantle under South America is not behaving as it does in far eastern Asia. Perhaps the mantle does not have the freedom to well-up due to an interaction between the shallowly dipping subducting plate and the continental roots or perhaps the overlying lithosphere is simply not extending as rapidly as in Asia and, therefore, the mantle does not respond similarly. The latter is the more likely explanation because slab migration and lithospheric extension are much greater in eastern Asia than in western South America.

From these arguments we conclude that the Asian back-arc low velocity anomaly in the long period maps ($\geq$100 s) is caused by mantle upwelling due to overlying lithospheric extension. A similar South American back-arc anomaly is absent due to weaker extension in the back-arc.

5.3 Misfit Compared with CRUST5.1/S16B30

Figure 12 summarizes the fit to the group velocity measurements delivered by the model CRUST5.1/S16B30 and our group velocity maps. In sections 5.1 and 5.2 we discuss the ways in which the group velocity maps predicted by CRUST5.1/S16B30 differ from the observed maps. All of these differences affect the fit to the group velocity measurements, but there are probably three discrepancies that have the most significant affect on misfit. First, intermediate period Rayleigh waves (40 s - 80 s) are very sensitive to crustal thickness. CRUST5.1/S16B30 predicts group velocities that on average are too high at these periods, indicating that on average the crust is too thin in CRUST5.1 across South America. Because this is true over nearly the entire continent the cumulative impact on the misfit is appreciable. Second, the Galapagos anomaly is not well resolved on the mantle model S16B30. Therefore, the predicted intermediate and long period maps (50 s - 150 s) are also too fast over this large feature. Third, S16B30 is an isotropic model and cannot fit long period and Rayleigh and Love waves simultaneously. This is the reason for the increase in rms-misfit for the long period Love waves.

6. Conclusions

We report the results of a systematic study of broadband Rayleigh and Love wave dispersion across South America. There are three main reasons why we believe that this study represents a significant improvement in the understanding of South American surface wave dispersion. The first has to do with the data used. This study is broader band, displays denser and more uniform data coverage, and demonstrates higher resolution than previous studies that have been performed on this scale. Resolutions at most periods average between 6° and 8° across the continent with significantly better resolution in the west. Second, the group velocity maps reveal the signatures of known geological and tectonic features previously not revealed in surface wave studies of South America. This both lends credibility to the maps and spurs interest in their use to infer information about the features that are observed. Observations at short and intermediate periods ($\sim$20-80 s) provide entirely new constraints on crustal structures, and the long-period observations ($\geq$100 s) yield higher-resolution information about the deep lithosphere and upper mantle down to depths of 150 km-200 km. Finally, the group velocity maps provide a significant improvement in fit to the observed dispersion curves. This is particularly impressive when compared to misfits from existing mantle and crustal models.

Concerning the crust, observed group velocity anomalies include information about sedimentary velocities and thicknesses, crustal velocities, and Moho depths. The dispersion signatures of a number of sedimentary basins across the continent, the Caribbean, and western Gulf of Mexico are displayed clearly on the-short period (20-30 s) group velocity maps (e.g., Maturin-Llanos, Marañon-Ucayali-Madre de Dios, Chaco-Tarija, and Paraná Basins on South America). A high velocity anomaly is not associated with the Paraná flood basalts at short periods, unlike the Deccan Traps and Ethiopian flood basalts of Eurasia (RL98). The sediments in the basin appear to obscure the high velocity anomaly as they do for the Siberian Tunguska Basin (RL98). One of the most striking features on any of the group velocity maps is the low-velocity anomalies that appear on both the Rayleigh (30-90 s) and Love (40-125 s) wave maps associated with thickened crust beneath the Andes, the Altiplano, and, more subtly, the Brazilian Highlands. Such low-velocity anomalies beneath the Altiplano, in particular, are also observed on recent global phase velocity maps (e.g., Trampert and Woodhouse, 1995; Ekström *et al.*, 1997), but the present study provides significantly better resolution and more reliable amplitude estimates than previous studies. The breadth of the frequency band over which the dispersion maps are produced holds the promise for resolving these crustal structures from one another and from upper mantle structures during structural inversion for a shear velocity model (e.g., Das and Nolet, 1997).

Concerning the lower lithosphere and upper mantle, observed group velocity anomalies provide information about the thickness of the lithosphere, continental roots, the depth extent of mid-ocean ridges, and the Galapagos hot spot. The three components of the Amazonian Craton (Guyana shield, Guaporé shield, Sao Francisco Craton) are clearly imaged on the long-period maps, although the smaller southern cratons (Patagonian Massif, Rio de la Plata Craton) are not. The long period velocity anomaly associated with the Galapagos Ridge is probably the most striking feature on any of the maps. The Galapagos Ridge anomaly is more robust than those associated with the other ridges, and it appears that the thermal anomaly associated with the Galapagos hot spot penetrates deeper than other ridges. The large amplitudes of the velocity perturbation associated with the Galapagos Ridge probably mean that the Galapagos thermal anomaly is not much less than 500 km in diameter down to about 200 km depth. However, a high temperature plume, if it exists, could be significantly narrower. Finally, low velocities are not observed in the back-arc at long periods, which leads us to conclude that the east Asian back-arc anomaly (RL98) is predominantly the result of the mantle's response to back-arc extension rather than enrichment in volatiles.

The group velocity maps we present here provide a substantial improvement in fit to the observed data relative to recent models. They produce a variance reduction relative to the PREM of more than 90% at short periods and more than 80% for Rayleigh and Love waves below about 100 s period but reduces to about 70% at longer periods. RMS-misfit for the model CRUST-5.1/S16B30 is about 1.5-2 times larger than for the observed maps.

Although we have argued that this study represents a significant step toward understanding South American surface wave dispersion, there remain significant steps to be taken before the project will be completed. The effects of off-great-circle propagation (primarily at short periods) and azimuthal anisotropy (primarily at intermediate and long periods) need to be understood more clearly and probably incorporated in the inversion. Although the bias caused by azimuthal anisotropy is substantial, it is below the levels we interpret here. Second, the future use of data from temporary installations (e.g., BANJO, SEDA, BLSP) would improve resolution on the continent, especially in the east. Finally, the estimated dispersion maps (group and phase) may be inverted for shear velocity models across South America.

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Figures

Section 1: Introduction

1. Locations of (a) the 765 earthquakes (circles) and (b) the 48 stations (triangles) used in this study.

Section 2: Data and Measurement

2. The number of dispersion curved estimated for Rayleigh (solid) and Love (dashed) waves before (bold line) and after (thin lines) the cluster analysis.

3. (a) Average path length after the cluster analysis (Rayleigh - solid, Love - dashed). (b) Average measurement uncertainties defined as the standard deviation of the measurements that compose each cluster averaged over all clusters.

Section 3: Resolution and Bias

4a. Path density for Rayleigh waves at the indicated periods. Path density is defined as the number of rays intersecting a 2 degree square cell ($\sim 50,000$ km^2).

4b. Same as Figure 4a but for Love waves at the indicated periods.

5. Three examples of the ability to estimate an input 2D Gaussian test function using the paths, data weighting, and damping for the 150 s Rayleigh wave. The location of the input Gaussian (G^{in}) is the smaller white circle plotted at the 1σ point on each plot. The amplitude of the input Gaussian is 10%. The estimated averaging kernel (K) is grey-shaded with values $\delta U/U$ in percent. The fit Gaussian (G^{fit}) is the larger white circle plotted at the estimated $1\hat{\sigma}$ point, where $\hat{\sigma}$ is the standard deviation of the fit Gaussian.

6a. Estimated resolution and bias for Rayleigh waves at the indicated periods. Resolution is color coded from 3° to 18°. Bias is represented by black lines that begin at the centroid of the input Gaussian, marked by a black dot, and end at the centroid of the fit Gaussian.

6b. Same as Figure 6a but for Love waves at the indicated periods.

7. Estimated 'average' resolution versus period for Rayleigh (solid) and Love (dashed) waves where the average is taken across the South American continent.

8. 2ψ fast axes from the model of phase velocity azimuthal anisotropy of Trampert and Woodhouse (1996) for Rayleigh and Love waves at 40 s and 80 s period. A 1% velocity bar is presented for comparison in the center of the figure.

9. Azimuthal anisotropy bias. Estimated bias from azimuthal anisotropy expressed as percent error relative to the average group velocity across the region of study for each period and wave type. The $\pm 1\%$ contours are indicated as the white and black contours, respectively.

Section 4: Group Velocity Maps

10. Average estimated group velocity curve across the entire region of study (30°N - 60°S, -120°W - 0°W; bold solid line) compared with the prediction from PREM (long dashed line) and from CRUST5.1/S16B30 also averaged across the entire region of study (short dashed line).

11a. Estimated Rayleigh wave group velocity maps at the indicated periods presented in percent deviation from the averages across the displayed maps (Figure 10).

11b. Same as Figure 11a, but for Rayleigh waves at the indicated periods.

11c. Same as Figure 11a, but at the indicated periods.

12. Two measures of misfit to our group velocity measurements for Rayleigh (solid lines) and Love (dashed lines) waves for two different sets of group velocity maps. Thick lines are for our group velocity maps and thin lines are for the group velocity maps predicted by the model composed of the crustal model CRUST5.1 and the mantle model S16B30. (Top) Misfit is represented as variance reduction relative to the group velocity from PREM. (Bottom) Misfit is the RMS group velocity misfit [eq. (5)].

Section 5: Discussion

13. Schematic map of South America displaying key tectonic and geologic features.

14. Rayleigh wave group velocity maps predicted by the model CRUST5.1/S16B30 at the indicated periods. The velocity scales are the same as on the observed maps in Figures 11a and 11b, except perhaps for the maximum and minimum values.

15. Special resolution analysis for sedimentary basins. (a) The input model is the 3 km isopach assigned a velocity perturbation of 10%, taken from 3SMAC (Nataf and Ricard, 1996). (b) Estimated map using the paths, weighting and damping for the 20 s Rayleigh wave. (c) Estimated map using the paths, weighting and damping for the 20 s Love wave. Outlines of the input model are shown as contours in (b) and (c).

16. Special resolution analysis for Moho depression related to the Andes and continental shields. (a) The input Moho model is the 44 km contour, taken from 3SMAC (Nataf and Ricard,

1996) assigned a velocity perturbation of 10%. (b) Estimated map using the paths, weighting and damping for the 50 s Rayleigh wave. (c) The input cratonic model taken from 3SMAC assigned a velocity perturbation of 10%, (d) Estimated map using the paths, weighting and damping for the 100 s Rayleigh wave. Outlines of the input models are shown as contours in (b) and (d).