Regional distance shear-coupled PL propagation within the northern Altiplano, central Andes

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SUMMARY

Properties of the shear-coupled P wavetrain (SPL) from regional earthquakes provide important information about the structure of the crust and upper mantle. We investigate broad-band seismic data from intermediate-depth earthquakes and develop a grid search technique using synthetic seismograms to study the sensitivity of SPL and to model the crustal structure of the northern Altiplano, central Andes.

Waveforms from an earthquake that occurred on 1994 December 12 within the Nazca slab beneath the Altiplano display a clear SPL wavetrain at the temporary stations deployed during the BANJO and SEDA experiments. We relocate this event and determine the moment tensor by inverting the complete long-period waveforms. With these source parameters fixed, we perform sensitivity analyses using a reflectivity technique to compute synthetic seismograms at a distance of 313 km (BANJO station 2, SALI). We find that, at this distance, the long-period SPL wavetrain is sensitive to the following model parameters, in order of decreasing sensitivity: crustal \( V_P/V_S \), mantle \( V_P/V_S \), average crustal velocity, crustal thickness, focal depth, distance (location), crustal \( Q_a \) and \( Q_b \), and mantle velocity.

We develop a grid search method to investigate the four parameters of the crust/upper mantle model to which the synthetic seismograms are most sensitive at SALI (crustal \( V_P/V_S \), mantle \( V_P/V_S \), average crustal velocity, crustal thickness). Trade-offs exist among all four of the model parameters, resulting in a range of acceptable crustal models that provide excellent fits between the data and synthetic seismograms in the passband of 15–100 s at a single station. However, by using data at a range of distances (150–450 km) we find that the model that provides the best overall fit between the data and synthetic seismograms, and thus best approximates the average characteristics of the crust and upper mantle structure of the northern Altiplano, is characterized by an average crustal velocity of 6.0 km s\(^{-1}\), a crustal Poisson’s ratio (\( \sigma_{\text{crust}} \)) of 0.25 and a mantle Poisson’s ratio (\( \sigma_{\text{mantle}} \)) of 0.27. The resulting models confirm previous estimates of low Poisson’s ratios, low average crustal velocity and thick crust in the Altiplano.

Key words: body waves, crustal structure, seismic wave propagation, sensitivity, synthetic seismograms.

1 INTRODUCTION

1.1 Previous geophysical studies of the Altiplano

The Altiplano of the central Andes is a high plateau with an average elevation of 3.6 km which is part of an active continental mountain belt. The Altiplano is bounded to the west by an active volcanic arc (Western Cordillera) and to the east by fold and thrust belts (Eastern Cordillera and sub-Andean zone). Recent geophysical studies focusing on the crustal and upper mantle structure of the Altiplano indicate that the crust is thick with low average crustal P-wave velocities. An east–west seismic refraction profile at 21.2°S (Wigger et al. 1994) revealed that the Altiplano has thick crust and low average crustal P-wave velocities (5.9–6.1 km s\(^{-1}\)).

The receiver function analyses performed by Beck et al. (1996) suggested that the central Altiplano crust is 60–65 km thick with a low crustal Poisson’s ratio (0.25). These results are...
consistent with the study of Zandt et al. (1996), who used waveform modelling to determine low average crustal P-wave velocities (6.0 km s\(^{-1}\)), a low crustal Poisson's ratio (0.25) and a crustal thickness of 65 km for the Altiplano. The results of Schuessler (1994) and Myers et al. (1998) also suggest low average P-wave velocities and a thick crust for the Altiplano of the central Andes. In this study we: (1) utilize broad-band waveforms recorded at regional distances to investigate the characteristics and sensitivity of the shear-coupled P wavetrain to certain parameters of the crust and mantle; (2) develop a single-station grid search technique in order to analyse efficiently a large number of combinations of crust and upper mantle properties; and (3) place constraints on the structure of the Altiplano.

1.2 Previous studies of shear-coupled P waves

Shear-coupled P waves are a long-period wavetrain resulting from the coupling of shear phases with PL waves in the crustal waveguide. As demonstrated by Oliver (1961) and Chander et al. (1968), the concept of an SPL wavetrain can be explained by a combination of ray theory and leaking-mode theory. Elastic strain energy radiated from the source as SV waves reaches the surface where it becomes partially trapped in the waveguide created by the crust. After it has become trapped, it travels the remaining distance from source to receiver as a leaking mode of the waveguide.

In the following discussions of the specific phases that constitute SPL, we make use of the nomenclature outlined in Båth & Stefánsson (1966): ray segments that are up-going in the crustal layer are represented by lowercase letters; ray segments that are down-going in the crustal layer are represented by uppercase letters; and the up-going rays (direct phases) from the mantle are represented by uppercase letters.

At regional distances, the SPL wavetrain contains SsPn (subsequently abbreviated as sPn), SsPmp, and multiples of SsPmp (e.g. SsPmpPmp). Several studies in recent years have noted the presence of and exploited the information contained in shear-coupled P waves (e.g. Singh et al. 1995; Langston 1996; Zandt et al. 1996; Zhu & Helmberger 1997). We build upon the results of Zandt et al. (1996), who documented the presence of the shear-coupled P wavetrain in regional seismograms from intermediate-depth earthquakes in the Nazca slab and modelled the waveforms to estimate an average crustal structure for the northern Altiplano.

Figure 1. Map of west-central South America showing the locations of the BANJO (black triangles) and SEDA (black squares) portable broadband seismic stations, PDE locations (white circles) and associated Harvard CMT solutions for selected intermediate-depth events that occurred during the BANJO and SEDA deployments. Also shown is the Harvard CMT location for event 94346 (white triangle), the location for event 94346 from Zandt et al. (1996), and the location for event 94346 determined in this study. Stations used in this study to determine the hypocentral location for event 94346 are circled. Numbers adjacent to event locations correspond to Julian day (YYDDD), depth/mb. The 3500 m elevation line is contoured in white.
We explore the behaviour of shear-coupled $P$ waves at regional distances by first conducting a series of sensitivity tests to identify the trade-offs among parameters such as crustal thickness, average crustal velocity, and crust and mantle Poisson’s ratios. In order to utilize regional phases to determine velocity models, we must next understand the trade-offs among parameters at single stations. We therefore developed a single-station grid search technique to analyse efficiently a large number of combinations of crust and upper mantle properties, and investigate the structure of the northern Altiplano. We then expand our study to include multiple stations for more robust results.

2 DATA

Many of the earthquakes recorded at regional distances (100–1400 km) on the portable broad-band stations in the BANJO and SEDA arrays (see Beck et al. 1996) occurred at intermediate depths within the subducting Nazca slab; the epicentral locations of six of these are plotted in Fig. 1. Ray paths from earthquakes in the Nazca slab beneath the Altiplano travel from the slab, through the mantle and crust and directly to the BANJO and SEDA stations located on the margin of the plateau. These are relatively pure Altiplano ray paths, and, because these earthquakes were recorded at regional distances,

Figure 2. Example of data recorded by the BANJO and SEDA broad-band stations from events 94167 and 94341 (refer to Fig. 1). The instrument responses have been deconvolved from the data; traces are aligned on the $P$ wave. (a) Vertical (solid line) and radial (dashed line) component data from event 94167, band-passed between 1 and 100 s. (b) Vertical (solid line) and radial (dashed line) component data from event 94341, band-passed between 1 and 100 s. (c) Vertical (solid line) and radial (dashed line) component data from event 94346, band-passed between 1 and 100 s. (d) Vertical (solid line) and radial (dashed line) component data from event 94346, band-passed between 15 and 100 s.
their waveforms are sensitive to details of the crust–mantle waveguide. Source-station geometries such as these will provide us with dense ray-path coverage for the central Andes.

The broad-band waveforms of intermediate-depth earthquakes recorded by BANJO and SEDA stations are rather complicated because, at regional distances, many phases arrive close together in time. At regional distances, the crust acts as a waveguide, and hundreds of reflections between the surface and Moho are important for waveform character (Lay & Wallace 1995). Figs 2(a) and (b) show the broad-band waveforms recorded on the BANJO and SEDA STS2 and Guralp sensors for the 1994 June 16 (mL 5.6) and 1994 December 7 (mL 5.6) events. In the following discussions, we will use the year and Julian day when referring to an earthquake; for example 1994 June 16 = 94167, and 1994 December 7 = 94341. Figs 2(c) and (d) show broad-band and band-passed data from an earthquake that occurred on 1994 December 12 (event 94346, mL 5.9) (see also Fig. 1).

Note that all of the broad-band seismograms in Fig. 2 display a clear P-wave arrival, followed by a low-amplitude SV-wave arrival, which was converted to P at the Moho (Sp), and then a complicated set of large-amplitude phases. These large-amplitude arrivals display P-wave apparent velocities. This wavetrain has prograde elliptical particle motion, which suggests that the energy left the earthquake source as a S and converted to PL before reaching the recording stations. These independent analyses allowed us to identify the SPL wavetrain on the majority of seismograms recorded on the Altiplano. The dominant phase in the SPL wavetrain is SsPmp and its multiples (Fig. 2), which are created when the direct SV wave converts to a P wave at the free surface and becomes trapped in the crust because of post-critical P-wave reflection at the Moho (Langston 1996) (Fig. 3). Like all waves that have been supercritically reflected or transmitted at a discontinuity, SsPmp (and its multiples) undergoes a phase shift. The resulting waveform is a linear combination of the original waveform and its Hilbert Transform (Aki & Richards 1980). The shape and polarity of this post-critically reflected wave is controlled by the angle of incidence. The amplitude of SsPmp (and, at further distances, its multiples) is large because of the almost complete reflection at the Moho (Langston 1996).

Our study of shear-coupled PL propagation focuses on the data from event 94346 recorded at BANJO station SALI (station 2, distance 313 km). In the broad-band record at SALI (Fig. 3), we see a clear P-wave arrival, followed by low-amplitude Ps and Sp arrivals, and then the dominant SsPmp arrival. As we will demonstrate, the Sp arrival is an important phase in waveform modelling; P–Sp traveltimes provide important information regarding crustal thickness, average crustal velocities and mantle VP/VS ratios.

3 SOURCE AND SENSITIVITY STUDY

3.1 Determination of event hypocentre, and moment tensor inversion

An accurate hypocentral location is required for accurate fitting of regional seismograms. To constrain the event 94346 epicentre, P- and S-wave arrival times were picked from the vertical and radial component seismograms from the BANJO and SEDA stations on the Altiplano (Figs 1 and 2). We were careful to pick the S arrival using all three components so as not to confuse it with the SsPmp arrival. Using the epicentre and foci depth reported by the PDE catalogue as a starting point, we used the relocation program HYPO (Lienert 1994) to constrain the latitude, longitude, and origin time of event 94346. We used a simplified crustal/mantle model of the Altiplano in the relocation (65-km thick crust, VP = 6.0 km s\(^{-1}\), VS = 3.2 km s\(^{-1}\), crust = 0.25; from Zandt et al. 1996). The location that resulted in the lowest P and S residuals was 17.59°S, 69.80°W (Fig. 1).

To constrain the focal depth and to determine the source mechanism for this event, we performed a linear least-squares inversion of the complete long-period waveforms, following the method of Randall et al. (1995). We used the same crustal model as described above and in Zandt et al. (1996).

We make use of the vertical, radial and transverse components of the time domain displacement seismograms from the BANJO stations on the Altiplano (1 to 4). The data are bandpass-filtered between 15 and 100 s because we want to reduce the sensitivity of the regional waveform to details of the crustal structure. A series of inversions is performed with different fixed depths between 125 and 210 km to provide a depth estimate for the earthquake. The quality of each data–synthetic seismogram fit is assessed visually and by an analysis of the least-squares error calculated by the inversion. A focal depth of 160 km resulted in the optimal data–synthetic fit; at this depth, the non-double-couple CLVD mechanism makes up 6.4 per cent of the total solution. The resultant focal
mechanism at a depth of 160 km (Fig. 4a) is similar to the Harvard CMT solution, and indicates normal faulting with a small component of strike-slip motion. The focal mechanism solution does not appear to be significantly sensitive to depth (Fig. 4a).

We perform an ellipticity analysis to confirm our depth estimate independently. We notice that, in the data, the largest peak in the SPL wavetrain on the vertical component leads the same peak on the radial component by approximately 3 s (Fig. 2d). This relationship in the long-period data is a result of the prograde elliptical particle motion of SPL. We use a reflectivity technique (described below) to compute synthetic seismograms for a range of focal depths. For each depth, we plot the particle motion of the large-amplitude SPL arrival in the synthetic seismograms (examples are shown in Fig. 4b). The depth at which the particle motion of the band-passed synthetic seismograms most closely approximates the particle motion of the data is 160 km, consistent with our depth estimate from the moment tensor inversion.

### 3.2 Waveform sensitivity

The sensitivity of the SpPmp arrival and the SPL wavetrain (interfering SpPmp arrivals) to parameters of the continental crust and mantle at a given distance must be understood before event 94346 can be used for modelling purposes. We have chosen a distance and azimuth of 313 km and 136°, respectively (SALI event-station geometry), for these sensitivity analyses. We computed approximately 500 synthetic seismograms, utilizing a reflectivity technique implemented by Randall (1994) to investigate the sensitivity of the complete waveforms to crustal thickness, focal depth, epicentral location (source–receiver separation), average crustal velocity, crustal Vp/Vs ratio, Pn, mantle Vp/Vs ratio, and crustal Qx and Qg. Furthermore, we describe the trade-off among the aforementioned parameters and use these analyses to characterize fully the behaviour of SPL propagation through the Altiplano.

Our standard starting model is the 8-layer approximation for the structure of the crust and upper mantle of the Altiplano that was used in the moment tensor inversion (65-km thick crust, Vp = 6.0 km s\(^{-1}\), f\(_{0.9\%}\) = 0.25; from Zandt et al. 1996). We use the focal mechanism computed in the moment tensor inversion for the double-couple source in our model, placed at a depth of 160 km.

The resulting synthetic seismograms display the shear-coupled P waves we are investigating, and indicate the same prograde elliptical particle motion as is present in the data. Vertical and radial components of the broad-band (1–100 s) synthetic records and longer-period (15–100 s) synthetic records were analysed. The timing and amplitude of P, Sp, and SpPmp on the broad-band waveforms, and an analysis of SPL at longer periods, provide important information about properties of the crust and upper mantle structure. We make the following caveats with respect to the following analyses: the conclusions drawn in each subsection are restricted to the source–station geometry of SALI, and to the range of selected parameters.

In the following discussions of waveform sensitivity, we refer to the long-period shear-coupled P wavetrain (band-passed 15–100 s) as SPL, and refer to the individual components of the SPL wavetrain by their individual names (e.g. SpPmp, sPn, Sp).

We test eight parameters of the crust and upper mantle. Our sensitivity analyses suggest that the relative effects of these eight crust and upper mantle model parameters, in order of

![Figure 4](image-url). (a) Results from the moment tensor inversion for event 94346. The figure shows the depth versus normalized error curve for the range of focal depths tested. The arrow indicates the depth at which we obtained the best data–synthetic fit (160 km). The solid line represents the normalized error calculated by the inversion for the data–synthetic fit at these depths. Black dots indicate those depths for which focal mechanisms are shown. Note the relative insensitivity of the focal solution to depth. (b) Variation of particle motion with depth (130–180 km) in synthetic waveforms computed at a distance of 313 km. In each plot, the particle motion is shown for the large SPL arrival. Radial particle motion (x-axis) is plotted against vertical particle motion (y-axis). Particle motion for the data is shown with the solid line, and particle motion of the synthetic waveform is shown with the dashed line. Note the increase in ellipticity with increasing depth. As depth increases, particle motion becomes less rectilinear and more elliptical, and the amount of P energy relative to SV energy trapped in the crust–mantle waveguide approaches unity. The particle motion of SPL in the synthetic waveforms computed with a source depth of 160 km (shaded box) best fits the particle motion of SPL in the data.
decreasing effect on the computed synthetic waveform, are: crustal \( V_p/V_s \), mantle \( V_p/V_s \), average crustal velocity, crustal thickness, focal depth, distance (location), source \( Q_s \) and \( Q_p \), and upper mantle \( Pn \) velocity. In the following paragraphs, we discuss in detail the four most influential model parameters, as well as trade-offs that exist among these four parameters. We highlight those model perturbations that result in noticeable changes in synthetic seismograms. Such relationships will be particularly valuable when attempting to fit synthetic seismograms to the data from SALI, for fine-tuning models, and also for future studies in this region incorporating additional event–station geometries.

3.2.1 Sensitivity to crustal thickness

We fixed the average crustal velocity at 6.0 km s\(^{-1}\) and computed synthetics for a source–receiver separation of 313 km for crustal thicknesses of 45, 55, 65 and 75 km to test the sensitivity of the waveforms to variations in crustal thickness. This range reflects the range of crustal thickness determined by previous geophysical studies (e.g. Beck et al. 1996). The broad-band and band-passed vertical component results of this sensitivity analysis are presented in Fig. 5(a) (note: in this and all subsequent waveform figures, seismograms have been aligned on the \( P \) wave peak). Because an increased crustal thickness causes the \( S \) wave to reach the Moho sooner, the most notable effect of the thickening of the crust is a decrease in the time between the \( P \) and \( Sp \) arrivals. Hence, identifying the \( Sp \) arrival in the data can be very important and diagnostic of crustal thickness.

3.2.2 Sensitivity to average crustal velocity

We analyse the effect on the waveforms of varying the average crustal velocity by varying the \( P \)-wave velocity, maintaining average crustal \( P \)-wave velocities of 5.75–6.75 km s\(^{-1}\). We maintain the \( V_p/V_s \) of our standard model (~1.73) by varying \( V_p \) (5.75–6.75 km s\(^{-1}\)) and then calculating the associated \( S \)-wave velocity in each model (\( V_s = 3.32–3.90 \) km s\(^{-1}\)). Selected results of this sensitivity analysis are presented in Fig. 5(b). The \( P–Sp \) relative traveltime decreases as the average velocity increases, suggesting that the change in the path of the \( Sp \) phase is more sensitive to the crustal velocity than the change in the path of the direct \( P \) wave. The \( P \) leg becomes shorter as crustal \( V_p \) increases because the \( S \)-to-\( P \) angle of refraction steepens. Increasing the average crustal velocity decreases the amplitude of the \( S_{SPL} \) peak more on the radial component than on the vertical component (band-passed records). This increase also decreases the \( P–SPL \) traveltime and decreases the relative traveltime between the peak and trough.

![Figure 5](https://example.com/figure5.png)

**Figure 5.** (a) Synthetic seismograms computed to test sensitivity to variations in crustal thickness. Top: broad-band vertical components. Bottom: band-passed vertical components. (b) Synthetic seismograms computed to test sensitivity to variations in average crustal \( P \)-wave velocity. Top: broad-band vertical components. Bottom: band-passed vertical components.
3.2.3 Sensitivity to crustal V<sub>p</sub>/V<sub>s</sub>

We examine the sensitivity of the waveforms to V<sub>p</sub>/V<sub>s</sub> in the crust by fixing the P-wave velocity in each of the two crustal layers (5.5 km s<sup>-1</sup> and 6.1 km s<sup>-1</sup>, respectively) and allowing V<sub>p</sub>/V<sub>s</sub> to vary between 1.5 and 1.9 in each layer by changing the S-wave velocity. Increasing the V<sub>p</sub>/V<sub>s</sub> ratio increases the amplitude of the Sp arrival on the broad-band vertical component (Fig. 6a), but does not significantly affect the amplitude of the Sp arrival on the broad-band radial component. On the filtered records, increasing V<sub>p</sub>/V<sub>s</sub> has a greater effect on the amplitude of the large arrival on the radial component than on the vertical component. This is to be expected, since the radial components contain more S-wave energy. A low V<sub>p</sub>/V<sub>s</sub> (high V<sub>s</sub>) in our model results in less contrast in V<sub>s</sub> between the mantle and the crust, giving rise to lower amplitudes on our seismograms.

3.2.4 Sensitivity to mantle V<sub>p</sub>/V<sub>s</sub>

We analyse the sensitivity of the waveforms to the mantle V<sub>p</sub>/V<sub>s</sub> ratio by fixing Pn and allowing V<sub>p</sub>/V<sub>s</sub> to vary between 1.6 and 2.0 in each mantle layer. The starting crustal model was adjusted for this test by fixing Pn and calculating new Sn velocities for each V<sub>p</sub>/V<sub>s</sub> ratio. Selected results of this sensitivity analysis are presented in Fig. 6(b). In general, the magnitude and timing relations are similar to those observed in the crustal V<sub>p</sub>/V<sub>s</sub> analyses (Fig. 6a). However, because the S wave travels further in the mantle than it does in the crust, varying V<sub>s</sub> in the mantle results in longer traveltime differences than are seen in the crustal V<sub>p</sub>/V<sub>s</sub> sensitivity test. The analysis described in this section also suggests that the waveforms are more sensitive to Sn velocities (through a variation in V<sub>p</sub>/V<sub>s</sub>) than to Pn velocities (variation of Pn). A comparison of the synthetic seismograms computed to test sensitivity to crustal velocities and crustal V<sub>p</sub>/V<sub>s</sub> results in a similar conclusion.

3.2.5 Investigation of trade-offs among parameters

The trade-off between crustal thickness and average crustal velocity is commonly encountered in waveform modelling. The synthetic seismograms computed in this study for a range of crustal thicknesses and average crustal velocities also display this trade-off. This trade-off necessitates careful analysis of the long-period P–SPL and the broad-band P–SsPmp traveltimes when investigating the effect of crustal thickness and average velocity variations on shear-coupled P waves in synthetic seismograms. We find that these waveforms are affected slightly more by changes to average crustal velocity than they are by changes to the crustal thickness (compare Figs 5b and a). We interpret this observation as a result of the complex interaction of phases (S, S<sub>n</sub> and S<sub>sPmp</sub>) at this distance. We also note that the P–Sp traveltine is slightly more sensitive to crustal thickness than to average crustal velocity.

![Figure 6](https://example.com/figure6.png)

**Figure 6.** (a) Synthetic seismograms computed to test sensitivity to variations in crustal V<sub>p</sub>/V<sub>s</sub>. Top: broad-band vertical components. Bottom: band-passed vertical components. (b) Synthetic seismograms computed to test sensitivity to variations in mantle V<sub>p</sub>/V<sub>s</sub>. Top: broad-band vertical components. Bottom: band-passed vertical components.

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It may also be difficult to isolate the effect of changing the crustal $V_p/V_S$ ratio from that of changing the mantle $V_p/V_S$ ratio (Figs 6a and b). The band-passed synthetic seismograms display this trade-off especially well. The behaviour of $Sp$ on the broad-band synthetics is the key to isolating the differences between varying the crustal $V_p/V_S$ and varying the mantle $V_p/V_S$. Increasing the mantle $V_p/V_S$ ratio while keeping the crustal $V_p/V_S$ ratio fixed increases the $P-\text{Sp}$ traveltime and decreases the amplitude of the $Sp$ arrival on the broad-band vertical and radial synthetic seismograms. Increasing the crustal $V_p/V_S$ ratio while keeping the mantle $V_p/V_S$ ratio fixed does not affect the $P-\text{Sp}$ traveltime, and affects the amplitude of the $Sp$ arrival only on the broad-band vertical component.

4 GRID SEARCH TO DETERMINE CRUSTAL STRUCTURE

We have used sensitivity analyses to characterize the behaviour of the regional SPL wavetrain and to isolate those four properties of the crust and upper mantle that most significantly affect regional SPL propagation. We next develop a comprehensive grid search method to test different combinations of the four most important model parameters. We use visual inspection and correlation coefficients to isolate the combination(s) of parameters that provides a best fit at SALI, and at the other BANJO and SEDA stations on the northern Altiplano, thereby constraining the crustal structure of the northern Altiplano.

Our starting model is based upon the simplest model that crustal $V_p/V_S$ ratio fixed increases the $P-\text{Sp}$ traveltime and decreases the amplitude of the $Sp$ arrival on the broad-band vertical component. Our grid search does not affect the $P-\text{Sp}$ traveltime, and affects the amplitude of the $Sp$ arrival only on the broad-band vertical component.

Figure 7. Results of the grid search using data recorded at SALI. Correlation coefficients are the averages of those calculated for the data and synthetic vertical and radial components. $\sigma_{\text{crust}}$ increases from left to right; $\sigma_{\text{mantle}}$ increases from top to bottom. A correlation coefficient of 1.0 indicates a perfect data–synthetic fit. Note the trade-offs between crustal thickness and velocity, as predicted by our sensitivity analyses.

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This grid search results in the generation of 324 synthetic seismograms (vertical, radial and tangential components) from 108 unique combinations of grid search variables. The data and synthetic seismograms are band-passed between 15 and 100 s, cut to a length of 85 s and aligned on their pre-filtered P-wave peaks. We align the data and synthetic waveforms \textit{a priori} to eliminate the possibility of our results being biased by a possible location error because the epicentres of the intermediate-depth earthquakes we analyse lie outside our network of stations.

Using the method outlined in Wallace et al. (1981), each vertical, radial and tangential component of the synthetic seismograms is numerically correlated with the corresponding component of the data from SALI. A correlation coefficient of 1.0 indicates a perfect fit between data and synthetic seismogram. We then average the correlation coefficients calculated for the vertical and radial components to give a single value for each model. The tangential component data and synthetic seismograms are very simple, and do not provide any constraints additional to those from the vertical and radial data; hence, the tangential data are not used in the grid search.

The averaged vertical and radial component numerical correlation results for all combinations of crustal thickness, average crustal P-wave velocity, $\sigma_{\text{crust}}$ and $\sigma_{\text{mantle}}$ are displayed in Fig. 7. The results clearly illustrate trade-offs that exist for these crustal and upper mantle parameters, as was predicted by our sensitivity analyses. For example, for a given set of $\sigma_{\text{crust}}$, $\sigma_{\text{mantle}}$ and average crustal P-wave velocities, the data–synthetic fit decreases very slightly as the crustal thickness increases. As $\sigma_{\text{crust}}$ increases, we obtain better fits with greater average crustal velocity. This correlation shifts to even larger average crustal velocities for higher $\sigma_{\text{mantle}}$. In general, correlation coefficients are lower for higher $\sigma_{\text{mantle}}$.

For the purposes of the following discussion, we divide our nine best models (correlation coefficients at SALI $\geq 0.95$) into three classes of three models each, based on average crustal P-wave velocity. Class I models have an average crustal P-wave velocity of 6.0 km s$^{-1}$, crustal thicknesses of 55, 60 and 65 km, and $\sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.25/0.27$. Class II models have an average crustal P-wave velocity of 6.25 km s$^{-1}$, crustal thicknesses of 55 and 60 km, and $\sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.27/0.29$ and 0.29/0.27. Class III models have an average crustal P-wave velocity of 6.5 km s$^{-1}$, crustal thicknesses of 55, 60 and 65 km, and $\sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.29/0.29$.

To illustrate the complexity in the trade-offs that exist among these four model parameters further, consider the trend of the correlation coefficients as average crustal velocity is increased.

(1) Within the Class I models ($V_p = 6.0$ km s$^{-1}$), we find it difficult to constrain the crustal thickness within the range of our search (Fig. 7). Crustal thicknesses of 55, 60 and 65 km provide very similar data–synthetic fits. Our difficulty in constraining the crustal thickness was predicted by the results of our sensitivity analysis, which suggested that, out of the four grid search parameters under consideration, perturbations to crustal thickness in the model result in the least amount of variability in $SPL$.

(2) If the average crustal P-wave velocity is increased to 6.25 km s$^{-1}$, a best fit is achieved using crustal thicknesses of 55 or 60 km and slightly higher $\sigma_{\text{crust}}$ and $\sigma_{\text{mantle}}$ (Class II models). We note that at this higher velocity, although the first half of the data and synthetic waveforms match better, we sacrifice some of the good fit between the data and synthetics in terms of the amplitude of $SPL$.

(3) If the average velocity is increased to 6.5 km s$^{-1}$ (Class III models), we match the initial portion of the waveforms much better, but we sacrifice the overall fit (timing and amplitude) of $SPL$. These examples illustrate the trade-offs in model parameters arising from the complexity of the $SPL$ waveform at intermediate distances.

Synthetic seismograms computed for each Class I, II and III model discussed above provide excellent and similar fits to the data at SALI. Hence, at a single station with this source-station geometry we can determine a family of models that fit the long-period data. This can be very useful information, especially if additional constraints are available to eliminate some of the models. For example, as an independent check of the crustal and mantle $V_p/V_S$ (or Poisson’s ratios), we constructed a Wadati diagram using P- and S-wave picks at Altiplano stations for events similar in location to event 94346 (from Myers et al. 1998). The Wadati diagram suggests an average $V_p/V_S$ ratio for the crust and mantle of 1.78, corresponding to an average Poisson’s ratio of 0.27 (Fig. 8). Because direct P- and S-wave rays from events in the Nazca slab spend more time in the mantle than they do in the crust, we would expect that the average Poisson’s ratio would be weighted towards the mantle value. Thus, the Poisson’s ratio of 0.27 suggested by the Wadati diagram is a reasonable average for the crust and mantle in this region. Dorbath & Granet (1996) used P- and S-wave travel times picked from vertical component short-period records of local earthquakes to calculate an average $V_p/V_S$ for the crust and mantle of the Altiplano

\[ V_p/V_S = 1.78, \sigma = 0.27 \]

\[ V_p/V_S = 1.91, \sigma = 0.31 \]

\[ V_p/V_S = 1.73, \sigma = 0.25 \]

\[ V_p/V_S = 1.84, \sigma = 0.29 \]

Figure 8. Wadati diagram constructed from P- and S-wave picks from Nazca slab earthquakes recorded by BANJO and SEDA stations located on the Altiplano. The solid black line is the slope of the best-fitting line through the phase picks. The slopes of the dashed lines represent several different $V_p/V_S$ ratios. The best-fitting straight line corresponds to an average $V_p/V_S$ ratio for the crust and mantle of 1.78, corresponding to an average Poisson’s ratio of 0.27.
1.68, corresponding to a Poisson’s ratio of approximately 0.23. Beck et al. (1996) determined a crustal Poisson’s ratio of 0.25–0.26 for the Altiplano from receiver functions. Hence, there is no indication of a high crustal Poisson’s ratio for the Altiplano. The models of our preferred set (Class I; see classification scheme above) have low average crustal velocities, crustal Poisson’s ratios of 0.25, and mantle Poisson’s ratios of 0.27.

Fig. 9(a) is a record section showing the fit between the vertical and radial component data and synthetics at the Altiplano stations for our preferred model, a member of the Class I category (average \( V_p = 6.0 \) km s\(^{-1}\); crustal thickness 60 km; and \( \sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.25/0.27 \)). For comparison, we show in Fig. 9(b) a record section showing the fit between the vertical and radial component data and synthetics at the Altiplano stations for one model in the Class III category (average \( V_p = 6.5 \) km s\(^{-1}\); crustal thickness 60 km; and \( \sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.29/0.29 \)). The data–synthetic waveform fit is clearly worse at distances of 150 km to 250 km for the model with the higher average \( V_p \) and higher \( \sigma_{\text{crust}} \) and \( \sigma_{\text{mantle}} \).

Utilizing data from a range of distances when possible will help to eliminate models because trade-offs vary as a function of distance. We note that, in Figs 9(a) and (b), the quality of the data–synthetic waveform fit at Altiplano stations varies with distance, presumably reflecting the variations from a 1-D model. The crust of the Altiplano is represented as two simple layers; neither gradients or other subtle vertical variations are included. We conclude that, over all, the fit between the data and synthetic seismograms at some distances is better for the models with lower average crustal velocities (\( V_p = 6.0 \) km s\(^{-1}\); Class I models) than for models with high average crustal velocities (\( V_p = 6.5 \) km s\(^{-1}\); Class III models) and higher Poisson’s ratios (compare Figs 9a and b).

We conclude that, of the 108 models tested in our grid search, the three models that provide the best fit between the data and synthetic seismograms at all of the Altiplano stations are characterized by low average crustal velocities, \( \sigma_{\text{crust}} = 0.25 \) and \( \sigma_{\text{mantle}} = 0.27 \) (Class I models).

The results of our sensitivity analyses suggest that crustal thickness will be most easily constrained in future studies by using the phase \( Sp \); the timing and amplitude of this phase are particularly sensitive to changes in crustal structure. We infer from the results of Zandt et al. (1996) that the lack of a large-amplitude \( Sp \) arrival in our data is due to the presence of

Figure 9. (a) Record sections showing the fit between the vertical and radial data (solid lines) and synthetic seismograms (dashed lines) at the Altiplano stations. Synthetic seismograms were computed using an average \( V_p = 6.0 \) km s\(^{-1}\), crustal thickness of 60 km and \( \sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.25/0.27 \). (b) Record sections showing the fit between the vertical and radial data and synthetic seismograms computed using an average \( V_p = 6.5 \) km s\(^{-1}\), crustal thickness of 60 km and \( \sigma_{\text{crust}}/\sigma_{\text{mantle}} = 0.29/0.29 \).

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a velocity gradient at the Moho beneath the Altiplano. It is possible that different source–station geometries and/or a localized sharper Moho would result in a more well-defined Sp arrival; inclusion of these events in future structure analyses will allow us to constrain the crustal thickness better.

5 CONCLUSIONS

Broad-band regional waveform modelling is a useful technique with which to image the crust and upper mantle structure in the central Andes in Bolivia. Complicated arrivals near the S-wave arrival on regional broad-band data from event 94346 prompted us to use synthetic seismogram techniques to study the sensitivity of regional distance shear-coupled P waves to changes in the earth model and earthquake source parameters. After determining the depth, location and focal mechanism for this event, we perform synthetic analyses using a reflectivity technique to compute synthetic seismograms. We find that the long-period SPL wavetrain is sensitive to the following model parameters, in order of decreasing sensitivity: crustal $V_P/V_S$ ($\sigma_{crust}$), mantle $V_P/V_S$ ($\sigma_{mantle}$), average crustal velocity, crustal thickness, focal depth, distance (location), crustal $Q_s$ and $Q_p$, and upper mantle velocity ($Pn$).

We used a grid search method to investigate the four parameters of the crust/upper mantle model to which the synthetic seismograms are most sensitive at this distance (crustal thickness, average crustal P-wave velocity, $\sigma_{crust}$, $\sigma_{mantle}$). Reflectivity synthetic seismograms are computed in our grid search to determine a best crustal model for BANJO station SALI. Large trade-offs exist among all four of the model parameters, and there is a range of acceptable crustal models that provide excellent fits between the data and synthetic seismograms. The models that provide the best overall fit between the data and synthetic seismograms, and thus best approximate the average characteristics of the crust and upper mantle structure of the northern Altiplano, are characterized by an average crustal P-wave velocity of 6.0 km s$^{-1}$, $\sigma_{crust} = 0.25$ and $\sigma_{mantle} = 0.27$. With these parameters, crustal thicknesses of 55, 60 and 65 km provide very similar data–synthetic fits at SALI and at the other Altiplano stations. Future studies incorporating additional events with a variety of source–receiver geometries will allow us to constrain the crustal structure further.

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