

The lithosphere–asthenosphere boundary beneath the western United States

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SUMMARY

S receiver functions from 67 broad-band seismic stations in the western United States clearly reveal the existence of a mantle discontinuity with velocity reduction downward, which we interpret as the lithosphere–asthenosphere boundary (LAB). The average depth of the LAB is ~ 70 km. The boundary is relatively sharp with an overall sharpness of less than 20 km. The boundary is more prominent south of the Mendocino Triple Junction, where the Farallon Plate has completely subducted. This may indicate partial melts at the base of the lithosphere caused by the upwelling of the asthenospheric flow through the slab window. A double low velocity zone is observed at base of the lithosphere beneath southern Sierra Nevada, implying a second melting zone at a depth of ~ 100 km, well correlated with previous studies of lithospheric delamination in the area.

Key words: asthenosphere, converted waves, Farallon Plate, lithosphere, receiver functions.

1 INTRODUCTION

The Farallon Plate began to subduct beneath the North American Plate more than 100 Ma. Approximately 30 Ma the Pacific–Farallon ridge system began to contact North America (Atwater 1970). Currently most parts of the Farallon Plate have completely subducted into the mantle. The rest of the Farallon Plate is separated by the Pacific Plate into the Juan de Fuca and Gorda plate in the north and the Cocos plate in the south, resulting in the formation of Mendocino Triple Junction (MTJ) and Rivera Triple Junction (RTJ), respectively. Between the two triple junctions, subduction was replaced by transform motion of the Pacific Plate relative to the North American Plate along the western margin of North America (Furlong & Schwartz 2004). As the two triple junctions moved apart from each other, the San Andreas Fault (SAF) system was born. The coast range of the western US is a tectonically very active region. Many of the tectonic features are directly related to the northward migration of the MTJ along the western edge of North America. The MTJ separates two different tectonic units. North of the MTJ, the Juan de Fuca and Gorda plate are subducting under the North American Plate. South of the MTJ is a transform margin, where the Farallon slab has completely subducted beneath the North American Plate, leaving a slab window which is thought to be filled by the upwelling asthenospheric materials. The distinct tectonic differences should leave significant imprints on the mantle lithosphere underlying different geological settings at the surface. Accurate knowledge of the structure and thickness of the lithosphere is important to unravel the tectonic evolution of the western US.

Abundant studies have focused on the extraordinary tectonic activity in this area, especially on the northward migration of MTJ, the

associated scenarios of local crustal deformation and crust–mantle interactions (e.g. Verdock & Zandt 1994; Beaudoin *et al.* 1998; Das & Nolet 1998; Zhu & Kanamori 2000; Zandt *et al.* 2004). In contrast, direct constraints on lithospheric thickness are rare. High temperature in the upper mantle estimated from seismic tomography and xenolith data indicates a thin (< 100 km) thermal lithosphere (Artemieva & Mooney 2001). Global and regional seismic tomography reveals a thin lithosphere (less than 100 km thick) beneath most of the western US (Zandt & Furlong 1982; Benz *et al.* 1992; Dueker *et al.* 1993; Grand 1994; Humphreys & Dueker 1994; Bijwaard *et al.* 1998; Humphreys *et al.* 2003). By surface wave inversion, a seismically visible lithosphere is not observed beneath the western United States while a 200-km thick lithosphere is seen in the North American craton (van der Lee & Nolet 1997). Instead, an area of extreme low V_s in the order of -10 per cent is imaged in the uppermost mantle down to 300 km depth beneath the western US. Based on tomographic models (V_p and V_s), a high mantle temperature (1300°C) was estimated at a shallow depth in the western US with a minimum depth as shallow as 50 km in the Basin and Range, indicating a thin lithosphere in this region (Goes & van der Lee 2002). Beneath the Coast Range, an upwarp of low-velocity asthenosphere to unusually shallow depths (30–60 km) is observed by inversion of teleseismic traveltimes data (Zandt 1981). In the thermal models of Zandt & Furlong (1982), thickness of the thin lithosphere reaches 60 km beneath the northern Great Valley and eastern Transverse Ranges. Based on P -wave residuals, there is an eastward increase of the lithospheric thickness from ~ 70 km beneath Basin and Range to ~ 100 km beneath the Colorado Plateau (Zandt *et al.* 1995). In California a remarkable decrease of the lithospheric thickness from the coast to inland area was inferred by anomalies in regional shear

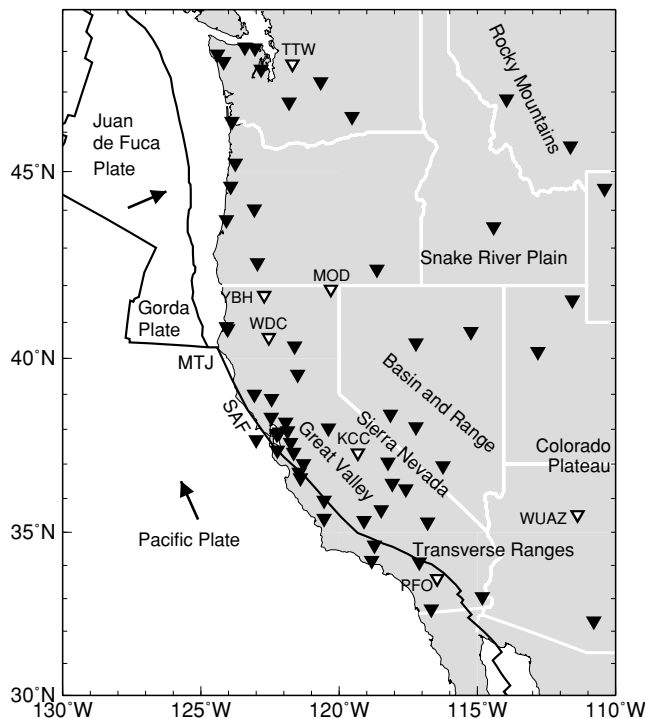


Figure 1. Map of broad-band seismic stations (filled and open triangles) used in this work. Open triangles denote seven stations (WDC, YBH, KCC, MOD, PFO, WUAZ and TTW) that are referred in Figs 3 and 5. Black lines are plate boundaries. Arrows indicate motions of the Pacific, Gorda and Juan de Fuca plates relative to North America. MTJ—Mendocino triple junction; SAF—San Andreas Fault.

wave arrival times (Melbourne & Helmberger 2001). The composition of late Cenozoic basalts across the Basin and Range province indicates a strong depth variation of the mantle melting zone (Wang *et al.* 2002), possibly representing the lithosphere–asthenosphere boundary (LAB).

Over decades the receiver function technique has been developed into one of the most powerful tools to observe seismic discontinuities in the crust and upper mantle (e.g. Vinnik 1977; Langston 1979). The technique isolates the *P*-to-*S* converted waves from the *P*-wave coda generated at seismic discontinuities (*P* receiver function). Although there are successful applications of the *P* receiver function for observation of the LAB (Rychert *et al.* 2005; Chen *et al.* 2006; Heuer *et al.* 2006), the method is generally limited for detecting interfaces in the mantle lithosphere due to interference of the primary conversions with strong crustal reverberations. The recently developed *S* receiver function technique appears to be able to overcome this problem, because it analyses *S*-to-*P* converted phases, which arrive earlier than the incident *S* phase and are consequently separated from the later arriving *S* reverberations (Farra & Vinnik 2000; Vinnik & Farra 2002; Vinnik *et al.* 2003, 2004; Yuan *et al.* 2006). This characteristic enables the *S* receiver functions to effectively detect the LAB (Li *et al.* 2004; Kumar *et al.* 2005a,b, 2006; Sodoudi *et al.* 2006).

In this paper, we present results of *S* receiver functions from 67 permanent broad-band seismic stations distributed in the western US (Fig. 1). The stations belong to diverse seismic networks, including the Global Seismograph Network, the Geoscope, the Pacific Northwest Regional Seismic Network, the Berkeley Digital Seismic Network, the Caltech Regional Seismic Network, the U.S. National Seismic Network, the Leo Brady Network and University

of Oregon Regional Network. Many of these stations are part of the ongoing USArray project. The 2–18 yr long continuous seismic data recorded by the dense broad-band station networks offer an ideal opportunity for studying the seismic structure of the mantle lithosphere by *S* receiver functions. As will be demonstrated later, *S* receiver functions have longer periods than *P* receiver functions and consequently have lower resolution for the crust. Here, we focus on detecting boundaries in the mantle lithosphere rather than discussing the crustal heterogeneity that may be better resolved by *P* receiver functions. In most cases we use the term *S* receiver functions for the receiver functions generated either by *S* or by *SKS* phases. In some special cases we also distinctly use *S* and *SKS* receiver functions for the receiver functions generated by *S* and *SKS* phases, respectively.

2 METHOD

2.1 *P* and *S* receiver function methods

There are a few differences between *P* and *S* receiver functions. (1) *S* receiver functions are free of *S* multiples and therefore can be very useful for the depth range of the mantle lithosphere where *P* receiver functions are sometimes dominated by crustal multiples. (2) The *S*-to-*P* converted waves cannot be observed over all teleseismic distances as they do not occur at overcritical incidence angles (Yuan *et al.* 2006). The distance range in which *S*-to-*P* conversions can be observed depends mainly on the depth of the discontinuity. For the base of the lithosphere the useful distance range for *S* receiver functions is about 60°–85° for the *S* phase. We also use *SKS* phase to calculate *S* receiver functions. The useful distance range is 85°–110°. Beyond 110° there are not many records with clear *SKS* phase strong enough for *S* receiver functions. (3) Within these distance ranges, *S* waves often interfere with diverse other phases. Since *S* waves are not the first arrivals, they sometimes arrive within *P*-wave coda and interfere with multiple mantle *P* waves (Bock 1994). At distances of 80°–90° there is strong interference among *S*, *SKS* and *ScS* phases. At distances larger than 60°, the *SPn* phase appears closely after the *S*, significantly contributing to waveform complication. Other possible interfering phases include a series of depth phases that appear in the time window of interest. (4) *S* waves are characterized by lower frequencies than *P* waves due to strong *S*-wave attenuation in the mantle, resulting in a lower spatial resolution. *S* receiver functions cannot resolve fine crustal structures in some cases, but may be more suited for gradient transition zones, which are generally thought to be the case for the LAB. (5) *S* receiver functions sample a broader area than the *P* receiver functions. At a given depth, piercing points of *S*-to-*P* conversions are located farther away from the stations than those of the *P*-to-*S* conversions. *S* receiver functions can be used to expand the sampling area, which is useful for ocean island stations (Li *et al.* 2004). However, in the case of strong lateral heterogeneities, for example, the existence of dipping layers, *S* receiver functions lose coherence in individual traces and the results may be biased by variable piercing point locations.

2.2 Synthetic receiver functions

In Fig. 2 we show synthetic *P* and *S* receiver functions for two simplified models, approximately representing the first-order crust and upper-mantle structure beneath the western US (Fig. 2a). Model 1 contains a 30-km thick homogeneous crust and a 40-km thick mantle lithosphere. In model 2 we will discuss the effect on *P* and *S*

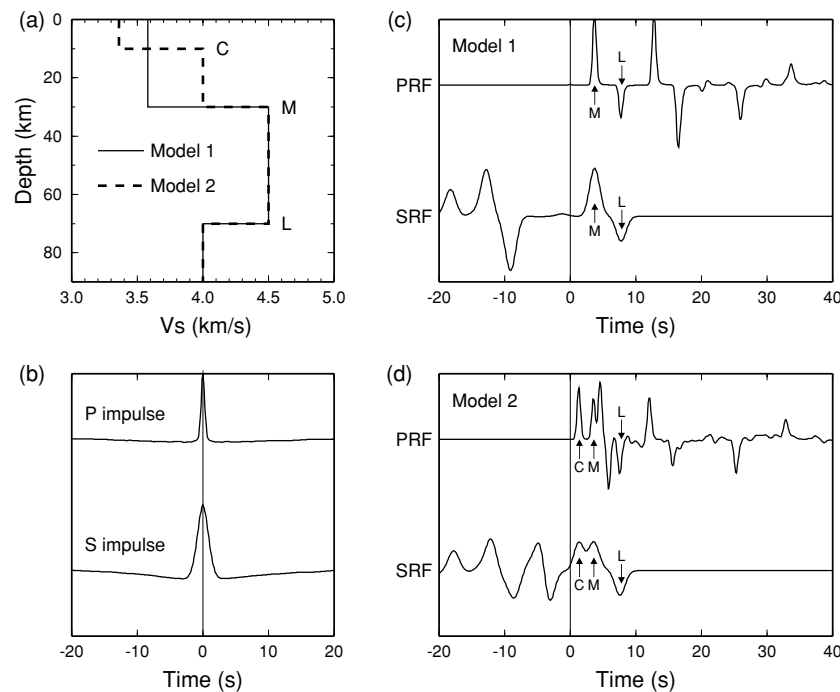


Figure 2. Comparison of synthetic P and S receiver functions for two simplified models. (a) Shear wave velocities of the two models. We fix the V_p/V_s ratio to 1.73 in the crust and 1.80 in the mantle. Model 1 (solid line) has a homogeneous crust with the crust–mantle boundary (M) at 30 km depth and the lithosphere–asthenosphere boundary (L) at 70 km depth. An intracrustal interface (C) is added in model 2 (dashed line). (b) P and S impulses derived from stacks of all deconvolved individual P and S waveforms, respectively. (c) P and S receiver functions for model 1. Timing is referred to a slowness of 6.4 s deg^{-1} . We use a spectral division deconvolution method and choose water level and Gaussian parameter so that the deconvolved P and S wave impulses well simulate the impulses obtained from the data (b). Therefore, the frequency content of the synthetic receiver functions is approximately identical with that of the data. For the S receiver function we reversed the amplitudes as well as the time axis to enable a direct comparison with P receiver function. Arrows (labeled C, M and L) indicate primary P -to- S conversions in the P receiver function and S -to- P conversions in the S receiver function. Unmarked phases in the P receiver function are multiples. The multiples with large amplitudes in the S receiver function appear at the negative time. (d) Same as (c) but for model 2.

receiver functions from adding an intracrustal interface at a depth of 10 km. Synthetic P and S receiver functions are calculated by a Haskell matrix propagation approach with an assumption of plain wave incidence. We use the spectral division method for deconvolution. We choose water level and Gaussian parameter so that the widths of deconvolved P - and S -wave impulses approximately simulate the impulses derived from the field data (Fig. 2b), respectively. The P -wave impulse is significantly narrower ($\sim 2 \text{ s}$) than the S -wave impulse ($\sim 5 \text{ s}$). The corresponding P and S receiver functions are generated for both models 1 and 2 (Figs 2c and d). In P receiver functions, primary conversions appear together with multiple phases at positive times, while they occur on either side of 0 s in the S receiver functions. For model 1, which has a homogeneous crustal layer, it is possible to observe the Moho as well as the LAB with both P and S receiver functions. The multiples in P receiver function arrive later than 10 s and are separated from the LAB conversion phase (marked L). Complicated structures in the crust may cause interfering multiple phases in the P receiver function, as shown in Fig. 2d for model 2. Here the multiples from an intracrustal interface (C) appear close to the primary converted phases of the Moho (M) and LAB (L), resulting in difficulty in detecting the Moho and LAB. In contrast, the LAB phase can be clearly identified in the S receiver function. We note that due to lower frequency content S receiver functions cannot easily resolve fine crustal structure. Depending on velocity contrasts of C and M interfaces, S receiver functions may sometimes observe only one of them, rather than resolve both interfaces, as will be demonstrated with field data in Fig. 5. P receiver functions may suffer from problem of multiples from interface C,

which could possibly mask the primary conversions of the Moho (Fig. 2d).

2.3 S receiver function processing

The P receiver function algorithm is primarily composed of two steps: first, a coordinate rotation isolates the P -to- S converted waves from the P wave; and secondly, deconvolution removes the source-site complications and the propagation effects. Further processing steps, such as moveout correction, common-conversion-point (CCP) stack and migration can be used to construct receiver function profiles for dense seismic arrays (e.g. Duecker and Sheehan 1997; Yuan *et al.* 1997; Jones & Phinney 1998; Bostock *et al.* 2001; Kind *et al.* 2002). While computing of P receiver functions has been already a standard routine process, calculation of S receiver functions is still under development (Li *et al.* 2004; Kumar *et al.* 2005a,b, 2006; Yuan *et al.* 2006). Here, we briefly summarize the processing method for S receiver functions, which we applied. We use waveform data of S and SKS phases for calculation of S receiver functions. As stated above, S waveforms are usually noisier than P records due to influence of multiple P reflections at the Earth surface and a series of other phases, such as S , SKS , ScS and SPn as well as depth phases. To identify S -to- P converted phases, we visually inspect S -wave data and select records with clear S or SKS phases. Weak S or SKS phases may be caused by unfavourable earthquake source mechanism or strong interfering phases. Data with strong interference of S and SKS are excluded to avoid contamination of converted phases

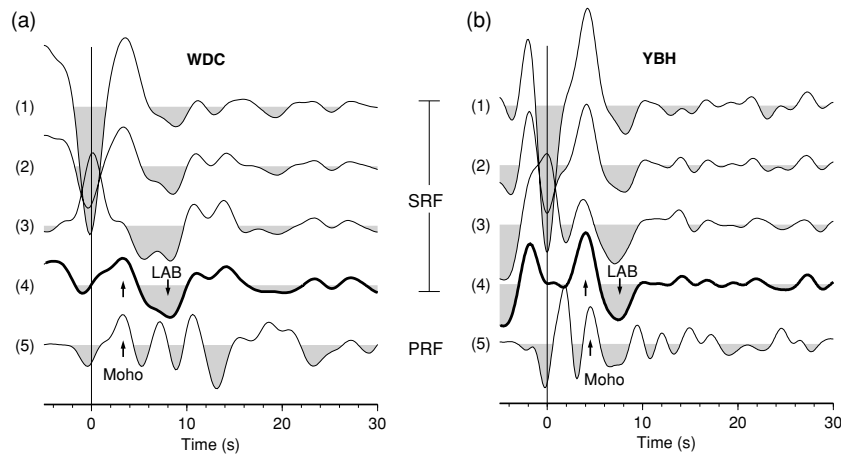


Figure 3. Stacks of individual *S* receiver functions (traces 1–4) and *P* receiver functions (trace 5) of station WDC (a) and YBH (b), experimented for rotation with different incidence angles. 131 *S* and 279 *P* receiver functions are stacked at station WDC; 110 *S* and 255 *P* receiver functions are stacked at station YBH. Individual *S* receiver functions and *P* receiver functions are bandpass filtered with 3–20 s and moveout corrected with a reference slowness of 6.4 s deg^{-1} prior to summation. Each *S* receiver function stack (traces 1–4) is a summation of all available individual *S* receiver functions rotated in a different way. For traces 1 and 2, rotation is done using theoretical incidence angles of *Sp* and of *S* phase, respectively. Trace 3 is the vertical component. Trace 4 uses calculated incidence angles by analysis of covariance matrix that minimizes *S* wave amplitudes at 0 s on each individual *P* component. Trace 5 is stacked *SV* component of *P* receiver functions, which are calculated using theoretical incidence angles.

from different reference phases. Visual inspection and selection of *S* and *SKS* waveform data is an important step and can hardly be replaced by any automated approach. Even through visual inspection, there are still quite a few records that do not produce meaningful *S* receiver function result. After deconvolution a quality control is carefully issued to remove the bad traces that are obviously different from the majority of the *S* receiver functions. These bad traces are thought to be probably caused by interference with disturbing phases. This relatively subjective approach ensures an optimal isolation of the *S*-to-*P* converted waves. We rotate records of *S* and *SKS* phases from *Z*, *N* and *E* components to *P*, *SV* and *SH* components. Because the incidence angles of most *S*-to-*P* conversions are as large as 40° – 50° , using vertical component is not effective here although it is often used for *P* receiver functions. The large residual *S* wave amplitudes projected on the vertical component can mask the real conversion phases in the first several seconds (see Fig. 3). The *S* receiver function is obtained by deconvolving the *SV* component from the *P* component. To display *S* receiver functions in the same way as to *P* receiver functions for a direct comparison, we further (1) reversed time scales of *S* receiver functions so that the converted phases appear at positive times; (2) reversed amplitudes of *S* receiver functions, so that a positive amplitude in *S* receiver functions indicates a positive boundary and (3) used a reference slowness of 6.4 s deg^{-1} for moveout correction.

Improper coordinate rotation may project a few per cent of *S* or *SKS* energy on the *P* component, interfering with converted phases in the first several seconds due to sidelobes. Converted phases at shallow depths such as the Moho are more sensitive to rotation angles than later mantle phases. In Fig. 3 we test the influence of rotation on the *S* receiver functions with stations WDC and YBH as examples. We found that using calculated incidence angles by minimizing the *S*-wave amplitudes at 0 s on each individual *P* component (trace 4) gives reliable *S* receiver functions that are less influenced by the strong reference *S* waves, especially for the phases at shallow depths like the Moho. The Moho phases are comparable on *S* receiver functions (trace 4) and *P* receiver functions (trace 5). *P* receiver functions have higher resolution, but *S* receiver functions exhibit a significant negative phase below the Moho (labelled

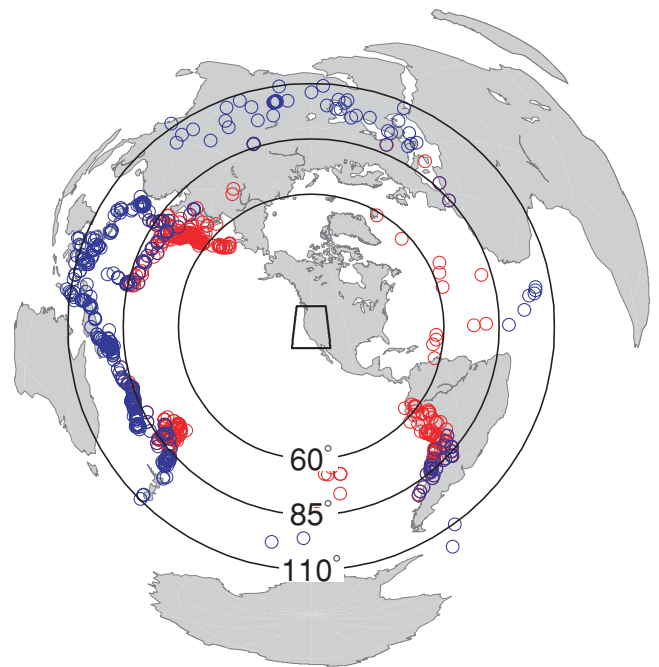


Figure 4. Map of teleseismic earthquakes used for *S* receiver functions. Earthquakes at epicentral distances of 60° – 85° (red) and 85° – 110° (blue) are used for *S* receiver functions and *SKS* receiver functions, respectively.

LAB), which is possibly masked by crustal reverberations in *P* receiver functions. Comparing traces 1–4, the LAB phase is weakly dependent on different rotation methods and should reveal a real structure in the uppermost mantle.

We analysed a large amount of data in this way and selected 447 events for *S* receiver functions and 360 events for *SKS* receiver functions (Fig. 4). We obtained 4460 *S* receiver functions and 3430 *SKS* receiver functions. The number of *S* receiver functions at each individual station varies from about 30 to more than 250. The number of traces exceeds 100 at more than half of the stations. We also

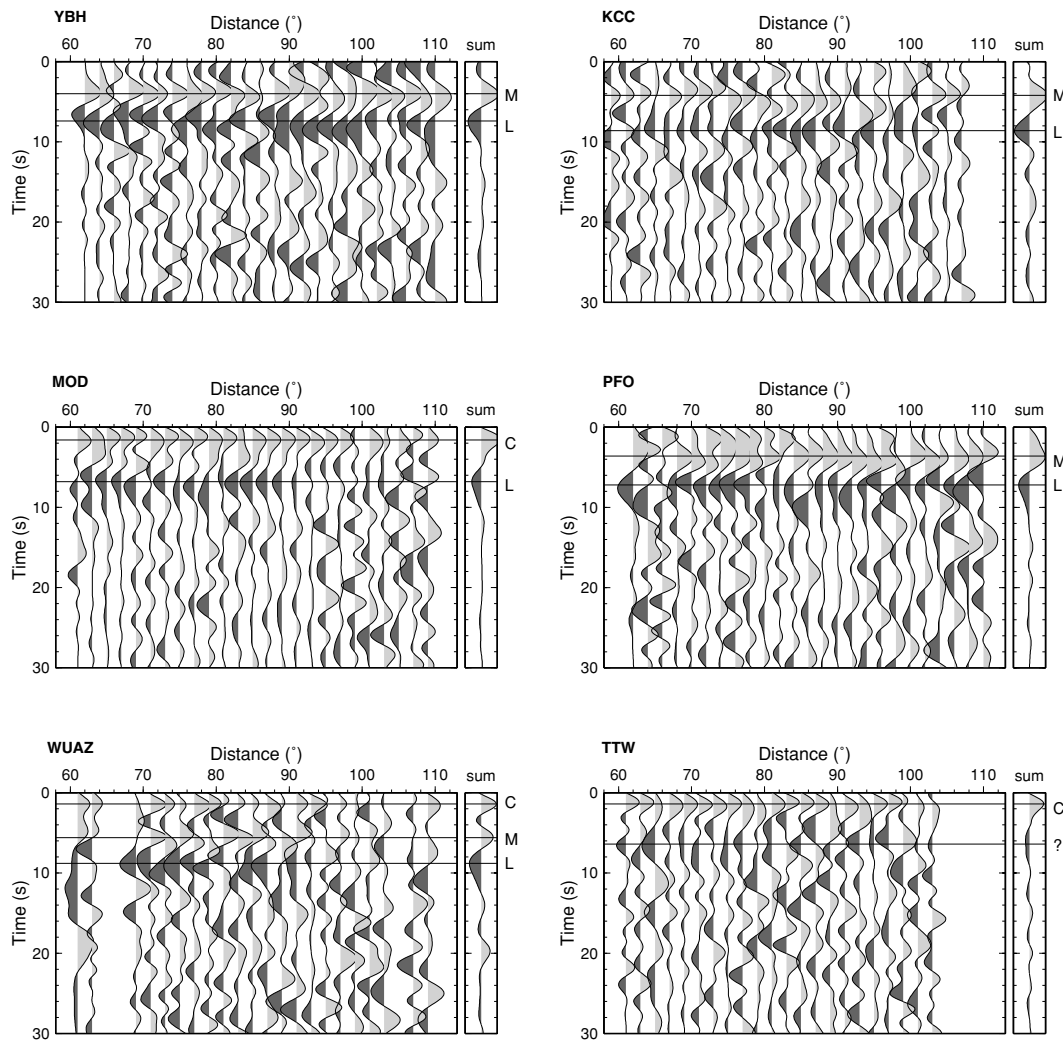


Figure 5. Binning stacks of *S* receiver functions of six stations (YBH, KCC, MOD, PFO, WUAZ and TTW) according to epicentral distances. Individual *S* receiver functions were moveout corrected and stacked in bins of epicentral distances of 2° without overlapping. *S* receiver functions are calculated from *S* waveforms for distance less than 85° and from *SKS* waveforms for distance larger than 85° . The ‘sum’ trace is obtained by directly stacking all the individual *S* receiver functions. At time 0 s is the arrival of the reference phase (*S* or *SKS*). Positive and negative amplitudes are plotted in grey and black, respectively. C, M and L mark an intracrustal interface, the Moho and the LAB, respectively.

calculated *P* receiver functions for all the 67 stations using method described by Yuan *et al.* (1997). The *P* receiver function processing is a standard procedure and is more or less automated. Altogether we obtained a total of 9440 *P* receiver functions.

3 OBSERVATIONS

3.1 Sample *S* receiver functions

We calculated *S* receiver functions from *S* and *SKS* waveform data for 67 stations shown in Fig. 1. In Fig. 5 we show the *S* receiver functions of six stations (for station locations see Fig. 1). The number of original individual *S* receiver functions varies from 82 at TTW to 253 at PFO. We stack individual *S* receiver functions in bins of 2° epicentral distances from 60° to 85° for *S* receiver functions and from 85° to 110° for *SKS* receiver functions. Summation traces show mainly two phases, one positive and one negative. We mark the positive phase around 4 s as the Moho (M) at stations YBH,

KCC, PFO and WUAZ and mark the earlier phase at 1–2 s as an intracrustal interface (C) at stations MOD, WUAZ and TTW. After these positive phases, a negative phase (L) at ~ 8 s is coherently seen at five stations, except station TTW, in much of the distance range of 60° – 110° . The negative phase indicates an upper boundary of a low velocity zone in the mantle at a depth of ~ 70 km. Incorporating previous knowledge, the boundary can be conveniently interpreted as the LAB. We will confirm this interpretation afterwards with stacked *S* receiver functions from the entire data set (Figs 6–8). At most stations the Moho and LAB phases are consistently observed with *S* receiver functions and *SKS* receiver functions. The arrival times of the LAB phase varies from less than 7 s at station MOD to more than 9 s at station WUAZ. Station TTW is located in the Juan de Fuca subduction zone. The weak LAB phase at TTW is possibly caused by a reduced LAB velocity contrast and sharpness and/or by interference of the LAB with the subducted Cascadean slab. As shown by the synthetics (Fig. 2) *S* receiver functions might not be able to resolve fine crustal structures. At most stations, there is an agreement in the Moho observations between *P* and *S* receiver functions.

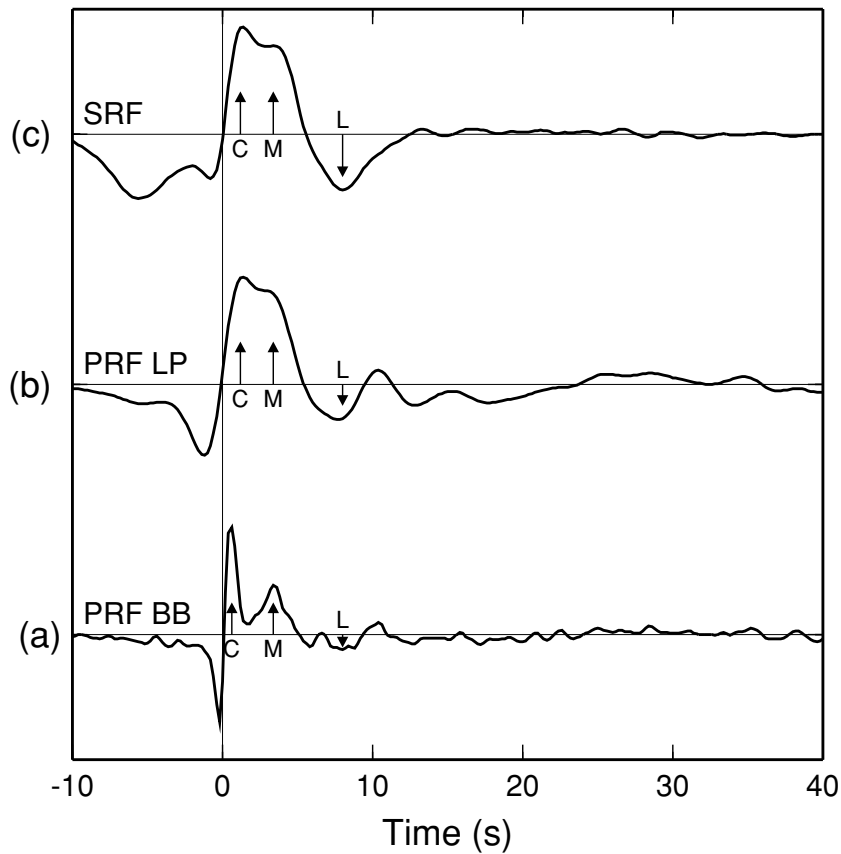


Figure 6. Stacks of all processed receiver functions in the western US. (a) Stack of 9440 broad-band *P* receiver functions. (b) 4-s lowpass filtered *P* receiver function stack. (c) Stack of 4460 *S* receiver functions. Arrows indicate primary *P*-to-*S* conversions in the *P* receiver function stacks (a and b) and *S*-to-*P* conversions in the *S* receiver function stack (c) and are labelled C, M and L, respectively.

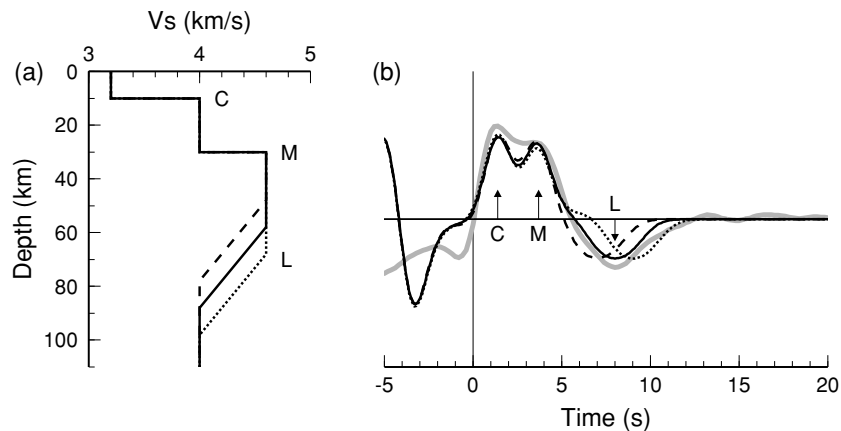


Figure 7. (a) Simple models found to model the stacked *S* receiver function. We use sharp velocity changes for the intracrustal interface at 10 km and the Moho at 30 km. The LAB is represented by a 30 km transition zone at three different depths. (b) Waveforms of stacked *S* receiver function data (thick grey line) and the best fitting synthetic *S* receiver function (black solid). Dashed and dotted lines are *S* receiver functions calculated for different LAB depths, 10 km shallower and 10 km deeper, respectively.

Our Moho observations generally agree with the Earthscope Automated Receiver Survey (<http://www.seis.sc.edu/ears/>) (Crotwell & Owens 2005) and the North American Moho map (Chulick & Mooney 2002). The Moho is missing at some stations, probably due to interference with energy from the much stronger interface C at these stations. However, we cannot rule out the possibility of no

large velocity contrast existing at the crust–mantle transition. We note that *P* receiver functions at station TTW also show a strong converted phase at 2.3 s along with a weak indication for the existence of the Moho contrast. At station MOD, the Moho phase can be seen at 4.3 s in the *P* receiver functions, while it is missing in the *S* receiver functions.

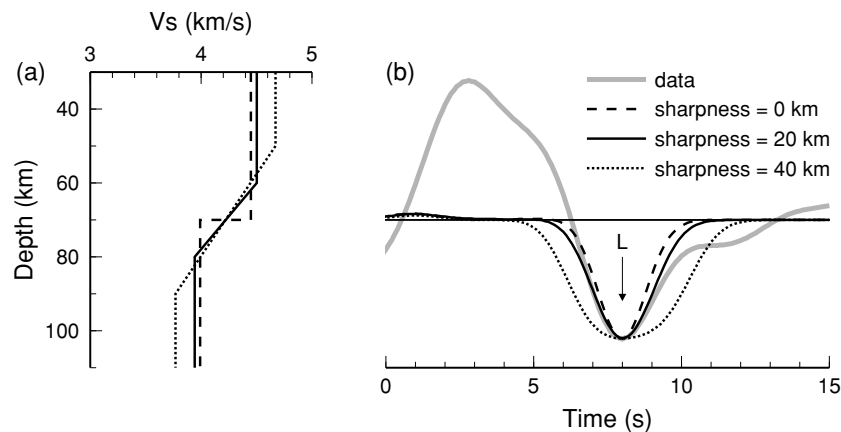


Figure 8. Modelling of the LAB phase at 8 s. (a) Models with different sharpness and velocity contrast of the LAB. Dashed line: LAB as a first-order discontinuity; solid line: LAB as a 20-km thick gradient transition zone and dotted line: LAB as a 40-km thick gradient transition zone. (b) Waveforms of data and synthetic *S* receiver functions. The thick grey line is the summation of all the *S* receiver function stacks for each station after an alignment of the LAB phase. A 3–20 s bandpass filter is applied prior to alignment and summation for a better phase correlation. Synthetic *S* receiver functions correspond to the models shown in (a).

3.2 Comparison of *P* and *S* receiver function stacks

We stacked all the *P* and *S* receiver functions separately and compared them to see the general characteristics of the crust and upper mantle beneath the western US (Fig. 6). The *P* receiver function stack (Fig. 6a) shows two significant phases, the Moho *P*-to-*S* conversion which arrives at 3.5 s (M) and an intracrustal converted phase at 0.6 s (C). In the *S* receiver function stack (Fig. 6c), the M and C phases join together to form a broad signal due to long periods. A significant phase (L) with negative amplitudes appears at 8.0 s in the *S* receiver function after this positive phase. We interpret this negative phase as *S*-to-*P* conversion at the base of the lithosphere. The *P*-to-*S* conversion at this boundary is hard to identify on the broad-band *P* receiver function (Fig. 6a). Two factors can mainly be responsible for the differences between *P* and *S* receiver functions: the different frequency content and the existence of crustal multiples in the *P* receiver function. It appears by first inspection and comparison of the *P* and *S* receiver functions that there is significant discrepancy between them. However, after a lowpass filter with a corner frequency of 4 s, the *P* receiver function (Fig. 6b) looks very similar to the *S* receiver function. The C and M phases are almost identical. The negative phase (L) after the Moho phase becomes more significant than on the broad-band *P* receiver function stack (Fig. 6a). In Fig. 6b this phase is broader and is shifted slightly earlier than in the *S* receiver function (Fig. 6c), due to the interference with multiple phases. Multiples are weak in the *P* receiver function stack (Figs 6a and b), because they are more influenced by lateral heterogeneities and are effectively suppressed. With this comparison and considering the synthetic models in Fig. 2, we conclude that the negative phase L is a representative feature for the study area, indicating a low velocity interface at an average depth of 70 km. We interpret this boundary as the LAB. In the follow section, we further investigate its sharpness, velocity contrast and the lateral variation over the whole study area. We will also see that the Moho may not be an ideal object of *S* receiver functions in this study due to low resolution. It might be better resolved by *P* receiver functions or by a joint inversion of *P* and *S* receiver functions.

3.3 *S* receiver function modelling

We model the *S* receiver function stack (Fig. 6c) with a simple model containing a Moho at a depth of 30 km, an intracrustal interface at a

depth of 10- and a 30-km thick gradient transition zone of velocity reduction at depth of 60–90 km, representing the LAB (Fig. 7a). The synthetic *S* receiver function simulates the data well (Fig. 7b). In order to see the depth resolution of the *S* receiver functions, we also calculate two other models with ± 10 km variations in the LAB depth. Synthetic *S* receiver functions for these two models clearly differ from the stacked *S* receiver function from the real data, indicating that a 10 km variation in the LAB depth can be well resolved.

The waveform of the LAB phase in the *S* receiver functions provides constraints on nature of the boundary. We compare the waveform of the LAB phase on the *P* component stack with that of the reference phases (*S*/*SKS*) on the *SV* component stack. We found that the LAB phase is broader than the reference phase (Fig. 2b). The broadening of the LAB phase can be caused both by reduction of sharpness and by the LAB topography. The 30-km thick LAB transition zone as modelled in Fig. 7 does not necessarily reflect the sharpness of the LAB. To eliminate the topography effect as much as possible, we stacked individual *S* receiver functions and *SKS* receiver functions separately at each station, shifted station stacks to align the LAB phase and summed them to create the final stack. A 3–20 s bandpass filter was applied in order to better identify the LAB phase. In this way the waveform of the LAB phase is preserved, while other phases like the Moho are distorted. Therefore, we only model the LAB waveform here. The resulting *S* receiver function stack is shown in Fig. 8b, together with three synthetic *S* receiver functions, calculated for LAB sharpness values of 0, 20 and 40 km, respectively. The stack of reference phase (*S* + *SKS*) on the *SV* component is used as the input signal (identical with the dashed line in Fig. 8b). While the synthetic waveform of the LAB as a first order discontinuity is narrower than the real data, a gradient zone 20-km wide models the LAB data well. In this model the LAB represent a transition zone of shear wave velocity from 4.5 to 4.0 km s⁻¹ downwards. The LAB waveform for a gradient zone 40 km wide is much broader than the real data. Synthetic *S* receiver functions also reveal the ability of the *S* receiver functions for resolving the sharpness of the LAB. The *S* receiver function for a LAB 20-km wide is slightly broader than the *S* receiver function for a LAB 0 km wide, whereas it significantly differs from the *S* receiver function for a LAB 40 km wide. The dominant wave period of the *S* receiver function is longer than 5 s, corresponding to a wavelength close to

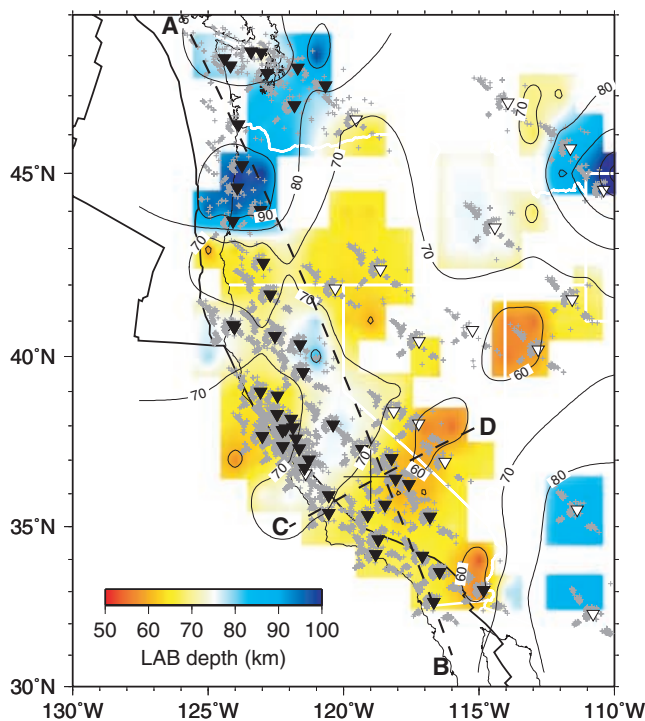


Figure 9. Map of the LAB depth. Crosses are piercing points at 70 km depth. The dashed lines AB and CD mark locations of the sections shown in Figs 10 and 11, respectively. Black triangles denote the stations in the coast range of the western US used for the AB section. Cross in southern California marks the region of proposed mantle delamination (Zandt *et al.* 2004).

20 km. A velocity boundary with a sharpness less than 20 km might not dramatically broaden the *S* receiver function waveform and will, therefore, be detected by the *S* receiver function as close to a first order discontinuity.

3.4 Mapping the LAB depth

To map the topography of the LAB, we geographically divided the study area into grid points with intervals of 1° latitude and 1° longitude. At each grid, we stacked the moveout corrected *S* receiver functions with piercing points at the 70 km depth within a cell of $2^\circ \times 2^\circ$ size centred on that point. Arrival times of the LAB phase are measured on the stacked *S* receiver functions and converted to depths using the IASP91 model (Kennett & Engdahl 1991). The choice of reference model is not very important, because the uncertainty of depth estimation due to different reference models is a few kilometres and is smaller than what we can resolve with the *S* receiver function data (~ 10 km). No measurement would be obtained if no clear conversion phase could be identified or when there are not enough traces stacked (the minimum number of receiver functions stacked in each cell is 20). In some region there are double negative phases after the Moho phase. In this case we pick the earlier negative phase as the LAB phase. The real LAB might be a complicated structure composing of multiple velocity contrasts (see next paragraph and Fig. 10). The depth measurements are interpolated and plotted in colour on the horizontal plane (Fig. 9). The LAB depth is better constrained in the western part of the study area due to high station density.

The LAB map generally agrees with previous *P*- and *S*-wave tomography works (e.g. Grand 1994; Humphreys and Dueker 1994;

Humphreys *et al.* 2003; Bijwaard *et al.* 1998), although local differences still exist, which are probably due to different methods used. The depth of the LAB is relatively homogeneous in the central part of the area. It is less than 70 km beneath much of California and Nevada and increases to the north and to the east to more than 80 km. The LAB is slightly deeper in central and northwestern California than in the close surroundings. Humphreys *et al.* (2003) shows a high velocity anomaly at a depth of 100 km in northwestern and central California, approximately where the lithosphere is thicker than the adjacent regions. To the north in Oregon and Washington the deeper LAB also coincides with the high velocity anomaly in the tomography (Humphreys *et al.* 2003). In the Basin and Range province the LAB is elevated to ~ 60 km, whereas the LAB is much deeper, at a depth of >80 km, beneath western part of the Colorado Plateau. Zandt *et al.* (1995) observed a similar LAB topography. They observed a lithospheric thickness of ~ 70 km in the Basin and Range and ~ 100 km in the Colorado Plateau with a sudden change from west to east across the western border of the Colorado Plateau. Surface wave tomography (van der Lee & Nolet 1997) did not observe the LAB in the western US, probably due to low resolution. Instead, an extremely low velocity was found in the upper mantle, in a good agreement with our observations of a thin lithosphere and a sharp LAB.

3.5 CCP sections

To explore the lateral variations of the LAB depth and sharpness, we built a depth section along the coast area, where most of the stations are located. We projected the individual *S* receiver functions to the space and constructed a section by CCP stack along the coast (Fig. 10a). Only stations within a distance range of 300 km from the coast (black triangles in Fig. 9) are used for the CCP stack. This includes stations in California, coast range of Oregon and Washington. The CCP stack was then smoothed over a length scale of 40 km using a moving Gaussian window (Fig. 10b) to better correlate the LAB phase. Clear LAB phase is observed in the southern part beneath SAF and Gorda subduction zone. To the north the LAB becomes less clear beneath Juan de Fuca subduction zone. We also compared *S* receiver function stacks at individual stations and found that the LAB phase is significantly weaker at stations in Oregon and Washington. For instance it is hard to identify the LAB phase at station TTW (Fig. 5). Our observation may indicate a sharper LAB with strong velocity contrast beneath SAF zone compared to the LAB beneath the Juan de Fuca subduction zone. South of the MTJ, the Farallon Plate has completely subducted into the mantle, leaving a slab window with upwelling flow of the asthenospheric material. This is evident from low heat flow in the north and high heat flow in the south (Lachenbruch & Sass 1980) and from high temperature and low velocity (Goes & van der Lee 2002; van der Lee & Nolet 1997) in the uppermost mantle. The emplacement of hot asthenospheric mantle at shallow levels would be expected to generate melts at the base of the lithosphere, consequently increasing the LAB contrast. Similarly, a sharp LAB was also observed beneath eastern North America and is interpreted as partial melts at the base of the lithosphere (Rychert *et al.* 2005).

A double LAB phase can be identified in central California between 35° – 38° N (Fig. 10a). Spatial smoothing results in a depression of the LAB in the region of the doublet phase. The region is characterized by the Great Valley to the west and the southern Sierra Nevada mountain range to the east. To localize the double LAB phase we construct a CCP section (line CD in Fig. 9) of *S* receiver

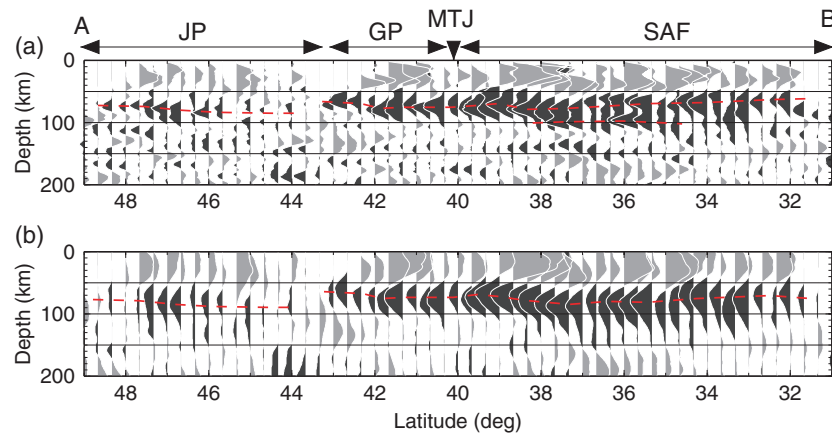


Figure 10. (a) *S* receiver function section along the coast of the western US (line AB in Fig. 9). Positive and negative amplitudes are shaded in grey and black, respectively. The LAB phases with negative amplitudes are marked. JP, Juan de Fuca Plate; GP, Gorda Plate; SAF, San Andreas Fault; MTJ, Mendocino Triple Junction. (b) Same as (a) but spatially smoothed with a 40 km window.

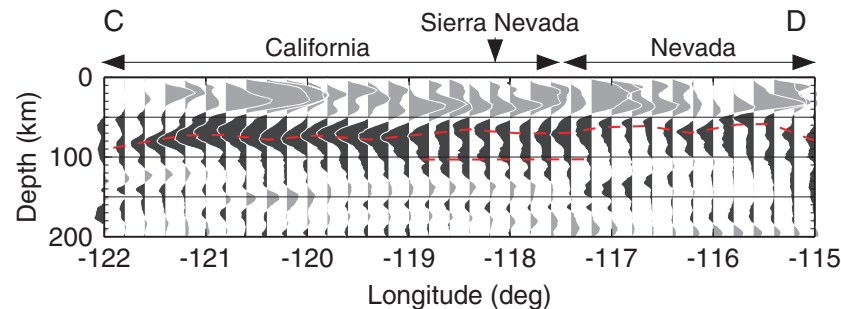


Figure 11. *S* receiver function section across California and Nevada (line CD in Fig. 9). The section width is 200 km. Positive and negative amplitudes are shaded in grey and black, respectively. The LAB phases with negative amplitudes are marked.

functions across California and Nevada (Fig. 11). The section width is 200 km, meaning that only the receiver function data, whose piercing points are located within a distance range of 100 km from the section line, have contributed to the CCP section. In Fig. 11 we can see that the LAB phase is much stronger in the western part of the section beneath much of California and becomes defused towards the east. The depth of LAB decreases from the west (~ 80 km) to the east (60 km). A double LAB phase can be observed in the central part of the section, beneath the Sierra Nevada. The Sierra Nevada of California is the highest mountain range in the United States associated with Cenozoic volcanism. Ongoing mantle delamination was suggested by numerous studies to explain the lack of a crustal root (e.g. Wernicke *et al.* 1996; Manley *et al.* 2000; Zandt *et al.* 2004). The double LAB phase might indicate multiple melting zones at the base of the lithosphere, associated with the mantle delamination processes. One can also speculate that the second low velocity zone might represent a zone of chemical differentiation or melting triggered by metamorphism and dehydration within the proposed lithospheric drip that extends to more than 150 km deep into the asthenosphere (Zandt *et al.* 2004).

4 CONCLUSION AND DISCUSSIONS

Our observation reveals a negative discontinuity in *S* receiver functions within the western US at an average depth of ~ 70 km, which is interpreted as the base of the lithosphere. The observation generally

agrees with previous global and regional *P*- and *S*-wave tomography works (Zandt & Furlong 1982; Benz *et al.* 1992; Dueker *et al.* 1993; Grand 1994; Humphreys & Dueker 1994; Bijwaard *et al.* 1998; Humphreys *et al.* 2003) and the estimate of the thermal lithosphere thickness (Artemieva & Mooney 2001; Goes & van der Lee 2002) and is consistent with the image of extremely low velocity in the uppermost mantle by surface wave tomography (van der Lee & Nolet 1997). Local discrepancies can be attributed to different methods used, which have different sensitivities and resolutions. Seismic tomography using body waves has a relatively high resolution, however, the method is more sensitive to lateral than to vertical velocity variations. Surface wave tomography is good to measure the thickness of the lithosphere, however, the vertical and lateral resolution is usually poor. The resolution of the *S* receiver function method is higher than the tomography method, however, *S* receiver functions are easily contaminated by noise and diverse interfering phases. In some cases the nature of the LAB itself is unfavourable to produce strong converted waves. The way we stack *S* receiver functions with lateral smoothing may help to enhance the converted phases at the cost of a reduction of the lateral resolution.

The dominant period of the *S* receiver functions is longer than 5 s, corresponding to a wavelength of ~ 20 km. Theoretically the spatial resolution may optimally reach ~ 6 km vertically and ~ 60 km (Fresnel zone at a depth of 70 km) horizontally. Synthetic tests show that a 10 km depth variation of the LAB can be clearly identified (Fig. 7). The resolution for the LAB sharpness is about 20 km, close to the dominant *S*-wave wavelength. Our data is not

sensitive enough to distinguish a first order discontinuity from a gradient transition zone with a sharpness of less than 20 km (Fig. 8). We observed a sharp (less than 20 km) seismic velocity gradient zone with V_s reduction of ~ 15 per cent, corresponding well to the results of surface wave inversions (van der Lee & Nolet 1997). For such a strong velocity reduction, the asthenosphere should contain a few percent partial melt or be enriched in volatiles associated with the subduction history (van der Lee & Nolet 1997; Rychert *et al.* 2005). We did not observe significant variation of the LAB depth, as reported by regional shear wave arrival times (Melbourne & Helmberger 2001) and by composition of the Cenozoic volcanic rocks (Wang *et al.* 2002). Due to summation of S receiver functions with lateral smoothing, we cannot rule out the local existence of short-wavelength strong anomalies in the LAB depth, which could probably be resolved by the ongoing USArray project with larger station density.

A 30-km thick crust with an intracrustal interface at a depth of 10 km is an average feature of the crustal structures in the western US, although there are strong lateral variations as shown by receiver functions at individual stations. It is difficult to accurately determine the Moho depth and fine crustal structures using S receiver functions alone due to the long periods of the data. Along the North American margin, conversion phases from the Gorda slab north of the MTJ (Verdonck & Zandt 1994) and the mafic lower crustal layer south of the MTJ (Furlong & Schwartz 2004) may interfere with the Moho phase in S receiver functions, possibly resulting in a misinterpretation of the Moho depth. This disadvantage poses a limitation to the application of S receiver functions for detailed imaging of crustal structure.

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