

**Attenuation in the upper mantle beneath Southern California:
physical state of the lithosphere and asthenosphere**

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Abstract

We invert phase and amplitude data of Rayleigh waves for attenuation (Q^{-1}) and shear wave velocities beneath southern California using teleseismic sources recorded by the TriNet/USArray network. Fundamental mode surface wave studies from 25 to 143 sec period allow us to constrain the vertical variation of shear quality factor Q_μ in the upper mantle. We use 2-D sensitivity kernels for surface waves based on single-scattering (Born) approximation to account for the effects of scattering on amplitude. A one-dimensional shear velocity model reveals a pronounced low velocity zone (LVZ) from \sim 80 km to \sim 200 km underlying a high velocity lid. Q_μ shows a similar pattern; large Q_μ at depths shallower than 80 km and much smaller Q_μ at depths greater than 100 km. Models that attribute the variations of attenuation and shear velocities with depth solely to temperature and pressure effects predict too low Q_μ values if they match the shear velocities. Alternative models considering the presence of partial melt can explain the observed very low Sv velocities in the asthenosphere. Partial melt in the asthenosphere could be generated due to decompression and the reduced solidus for damp mantle when the asthenosphere rose to fill the space left by the subducted Farallon plate.

Introduction

Seismic velocity models provide one of most important tools to understand the physical state of the upper mantle such as temperature, and melt or fluid content. Temperature is one of the key parameters controlling upper mantle dynamics and rheology. A number of studies have converted seismic velocities to temperature [e.g., *Furlong, et al.* 1995; *Goes and Vacher* 2000; *Goes and var der Lee* 2002], but this conversion and the interpretation of seismic models require accurate information about

attenuation since the anelastic contribution to the temperature sensitivity of velocity is significant. *Karato* [1993] showed that at temperatures approaching the solidus, anelasticity may double the temperature sensitivity of seismic velocity and other authors have suggested even greater sensitivity [*Priestley and McKenzie*, 2006].

In southern California, a low velocity zone from ~80 to ~200 km in the upper mantle with the lowest velocity 4.1 km/s at 120 km is observed in regional Rayleigh wave tomography [*Yang and Forsyth*, 2006a]. *Yang and Forsyth* argued that the low velocity zone could be due to the presence of partial melt in the asthenosphere. Recently, several studies suggested, however, that the increase in temperature with depth is sufficient to explain the origin of the observed low velocity zone beneath both oceanic and continental plates if anelastic effects are taken in to account [*Stixrude and Lithgow-Bertelloni* 2005; *Faul and Jackson*, 2005; *Priestley and McKenzie*, 2006] and neither water nor melt is required. In contrast, by measuring attenuation with regional arrays of ocean-bottom seismometers, *Yang et al.* [2007] showed that attenuation beneath the East Pacific Rise is much less than is predicted by a model in which the velocity is controlled solely by the direct elastic and anelastic effects of changing temperature, and suggested that melt and water concentration do play an important role in the young oceanic upper mantle. Anelasticity that magnifies the effect of temperature on seismic velocities also causes attenuation of surface waves; thus measurements of the decay of the amplitude of seismic surface waves, which are sensitive to lithospheric and asthenosphere structure, provide a key to interpreting seismic velocity models correctly.

The attenuation of surface waves as a function of period has the potential of revealing the vertical distribution of the quality factor Q_μ in the upper mantle. But measuring attenuation on a regional basis can be extremely difficult because it is hard to separate intrinsic attenuation effects on amplitude from other effects such as scattering and focusing/defocusing caused by elastic heterogeneities, and because the energy lost

due to attenuation as the wave propagates across the study area is small when the wavelength of the waves is of the same order as the dimensions of the region of interest. In recent years, there have been a few 3-D global attenuation studies using surface waves [e.g., *Billien et al.*, 2000; *Selby and Woodhouse* 2002; *Gung and Romanowicz* 2004; *Dalton and Ekstrom*, 2006], but these global models can only provide a smoothed model that may add apparent attenuation due to scattering effects to the intrinsic attenuation since local focusing or defocusing from small-scale heterogeneities may be not accounted for. In regional attenuation studies, one common method is the two-station method, which measures the difference of amplitude between two stations lying along a great-circle path from an earthquake source [e.g., *Cong and Mitchell*, 1998]. The two-station method assumes that surface waves propagate along great-circle paths and there are no strong heterogeneities between the two stations or no multipathing between source and receiver. These conditions are not satisfied in southern California where the lateral variation of seismic velocities is large and scattering effects are strong [*Yang and Forsyth*, 2006a]. Recently, with the development of finite-frequency scattering theory for surface waves [*Zhou et. al.*, 2004}], *Yang and Forsyth* [2006b] developed a regional-scale surface wave tomography method that accounts for focusing effects within the study area. The scattering or focusing of surface waves by elastic structures are described by 2-D sensitivity kernels based on the Born approximation. By using many stations and sources, we can detect velocity heterogeneities that cause focusing and therefore we can separate the scattering effects on amplitude from attenuation.

In southern California, the dense coverage of the deployed TriNet seismic network, now incorporated in USAArray, make it possible to recover attenuation accurately. We simultaneously invert for attenuation coefficients and the lateral variation of phase velocities using phase and amplitude data of fundamental mode Rayleigh waves. The attenuation coefficients are used to invert for Q_μ in the upper mantle. Q_μ and shear

velocities are combined to constrain the physical state of the upper mantle.

Data processing and methodology

We use fundamental mode Rayleigh waves recorded at 40 broadband seismic stations selected from the TriNet/USArray network in southern California (Figure 1). About 120 teleseismic events that occurred from 2000 to 2004 with surface wave magnitudes larger than 6.0 and epicentral distances from 30° to 120° were chosen as sources (Figure 2). The large number of events and stations used in this study generate a very dense ray coverage, which allows us to accurately resolve attenuation coefficients and high-resolution phase velocity maps. Vertical components of Rayleigh waves are filtered with series of narrow-bandpass (10mHz), four-pole, double-pass Butterworth filters centered at frequencies ranging from 7 to 40 mHz. Fundamental mode Rayleigh waves are isolated from other seismic phases by cutting the filtered seismograms using boxcar time windows with a 50 s half cosine taper at each end. The width of the boxcar window is determined according to the width of the energy package of fundamental mode Rayleigh waves. Rayleigh waves in this 25-143 s period range sample the entire upper mantle with good sensitivity down to \sim 200 km depth. The filtered and windowed seismograms are converted to the frequency domain by Fourier transform to obtain amplitude and phase data. Details of the data processing procedure are described in *Yang and Forsyth [2006a,b]*. To effectively constrain attenuation and lateral variations of phase velocities, we need to carefully account for all other influences on amplitude, which include scattering, local site response or amplification, instrument response, source radiation pattern, and interference from other modes. The detailed technique to remove these effects is described by *Yang and Forsyth [2006a]*.

In this study, we use the tomography method developed by *Yang and Forsyth [2006b]*, which utilize finite-frequency kernels to account for the scattering of Rayleigh waves by heterogeneities within the array and a two-plane-wave technique [*Forsyth and*

Li, 2005] to account for the scattering and multi-pathing of Rayleigh waves outside the array. Recently, 2-D anelastic attenuation kernels for surface waves have been derived by *Dahlen and Zhou [2006]*. These kernels can be incorporated in surface wave tomography to resolve lateral variations of attenuation. However, since amplitude decay due to attenuation is relatively small within the study region in southern California with an aperture of \sim 500 km, we only solve for an average attenuation coefficient at each period. In this case, 2-D attenuation kernels are reduced to ray-theoretical results and the decay of the amplitude of surface waves due to attenuation can be expressed as $e^{-\gamma x}$, where γ is average attenuation coefficient and x is propagation distance. The quality factor of Rayleigh waves, Q_R , is related to γ_R by $\gamma_R = \pi/uTQ_R$, where T is period, and u is group velocity. Parameterization of the study region and the regularization of inversion are described in *Yang and Forsyth [2006a]*.

Attenuation coefficients and phase velocities

In the inversion, we solve for attenuation coefficients simultaneously with lateral variations in phase velocities in order to separate focusing/defocusing effects on amplitude. Detailed interpretation of the lateral variations in phase velocities within the arrays is the topic of another paper [*Yang and Forsyth, 2006a*]. Here we show an example of 2-D pattern of phase velocities for a period 25 s in Figure 3 to demonstrate the magnitude of the anomalies that will contribute to scattering. At this short period, a striking low velocity anomaly with north-south trend in the region of the southeastern Sierra Nevada and Walker Lane volcanic fields and a strong high velocity anomaly in the offshore Borderland region are imaged. The lateral variations of phase velocities within the extent of the array are large, more than 10%, which can cause substantial scattering for the incoming Rayleigh waves. If we do not consider scattering effects, scattering will obscure the amplitude variation due to attenuation. For example, we calculate that

focusing and defocusing will cause amplitude anomalies up to $\pm 20\%$ for a typical 25-s-period Rayleigh wave propagating through the velocity structure in the vicinity of the southern California array. Thus, simultaneous inversion for phase velocities and the structure of the incoming wavefield is essential to constrain attenuation.

To demonstrate that attenuation is resolvable and required by the data, we first perform an inversion with no attenuation factor. The inversion includes the wave parameters that describe interference and amplitude variations due to the complexity of the incoming wave fronts, station site corrections, and phase velocity variations that are adjusted to try to match both the observed phases and the observed amplitudes through the effects of focusing. Although a minor effect, the amplitudes are also corrected for the effects of geometrical spreading on a sphere. The observed amplitudes for each event are normalized by the rms value of all observed amplitudes for that event to keep all the amplitudes at the same scale independent of magnitude of the event. We then examine the average residual amplitudes as a function of distance across the array. In Figure 4, we bin the residuals in 50 km distance intervals measured from the nearest station to each source event at 25, 50 and 125 s periods. There is a clear, decrease in residual amplitude with increasing distance from the source at each period, indicating that attenuation is required by the data. Although there are some oscillations from bin to bin, when modeled with an attenuation function, the decrease in amplitude is significant at the 99% confidence level for each of the periods shown. The true decay with distance is somewhat underestimated in this experiment since all the other variables in the inversion have been adjusted as much as possible to match the amplitudes, and, thus, to mimic attenuation with a combination of other effects.

Figure 5 displays the average attenuation coefficients from 25 to 143 s. Attenuation coefficients at all these periods are less than $3 \times 10^{-4} \text{ km}^{-1}$. The standard deviations of the estimates are much smaller than the coefficients at all periods, indicating that attenuation is resolvably different than zero at greater than the 99% confidence level at all periods.

Attenuation coefficients first decrease quickly with increasing period from $2.6 \times 10^{-4} \text{ km}^{-1}$ at 25 s to $1.5 \times 10^{-4} \text{ km}^{-1}$ at 50 sec, which is expected due to the approximate inverse proportionality to period if Q_R is constant. Then, attenuation decreases slowly with periods at the periods longer than 50 sec. A more detailed discussion of the cause of the attenuation is presented in a later section.

Attenuation of Rayleigh waves is the integrated effect of intrinsic shear wave quality factor (Q_u) over a range of depths. The contribution of bulk attenuation to the attenuation of Rayleigh waves is very small and can be neglected. The attenuation coefficient of Rayleigh waves at an individual period can be represented as a sum of shear wave dissipation (Q_{ul}^{-1}) in each layer from the surface to great depth as given by,

$$\gamma_R = \frac{\pi}{C_R^2 T} \sum_{l=1}^N \left[\left(\beta_l \frac{\partial C_R}{\partial \beta_l} \right)_{\omega \rho \alpha} + \frac{1}{2} \left(\alpha_l \frac{\partial C_R}{\partial \alpha_l} \right)_{\omega \rho \beta} \right] Q_{ul}^{-1} \quad (1)$$

where ρ , ω and l are density, angular frequency, and layer number respectively [Mitchell,1995]; and the partial derivatives of phase velocity, C_R , with respect to shear velocity, β , or compressional velocity, α , include both the effects of intrinsic sensitivity and layer thickness.

In this inversion, we parameterize the depth range from surface to the depth of 410 km to 21 layers with layer thickness varying from 10 km at the surface to 50 km at the bottom. The model parameters of Q_{ul}^{-1} are slightly damped by assigning prior standard deviations of 0.06 in the diagonal terms of model covariance matrix, i.e., a priori standard deviations roughly comparable in size to the value of the starting model, and smoothed by adding off-diagonal terms to the model covariance matrix that enforce 0.3 correlation in changes of Q_{ul}^{-1} in the adjacent layers. Shear wave velocities (β_l) and compressional wave velocities (α_l) are inverted from the average phase velocities, and the phase velocity partial derivatives in each layer are computed using the algorithm of Saito

[1988]. The details about the shear velocity inversion are described in *Yang and Forsyth* [2006a].

Figure 6 shows an example of a particular Q_u model and the resolution kernels. The starting model for this example is the smoothed QL6 model [Durek and Ekstrom, 1996], which is plotted with dashed line in Figure 6. The resolution kernels become wider with increasing depth, because the sensitivity range of Rayleigh waves increases with period. The fit of the predicted attenuation coefficients to observed attenuation coefficients is good at all periods (Fig. 5).

Particular solutions to the inverse problem for Q_u structure may contain features that are suggested, but not required, by oscillations in the data or that are inherited from the starting model. It is tempting to interpret these details in terms of physical properties like temperature variation, differences in melt concentration or water content. But before doing so, it is important to consider what features are statistically resolvable. Because the resolution kernels for Q_u structure are reasonably compact (Fig. 6) **in the sense of not having significant side lobes**, we can resolve the required features in terms of average Q_u over resolvable depth ranges, i.e., depth ranges over which the diagonal elements of the resolution matrix sum to 1.0 piece of independent information about the system.

Since the inversion for Q_u is an underdetermined problem, we test the effect of initial values of Q_u on the resultant Q_u by using a variety of starting values. The effects of starting values on Q_u models are small at depths shallower than 250 km, but become larger at depths greater than 250 km. The detailed shapes of the particular model solutions do depend on the starting model at all depths, but if we consider the average Q_u over the resolvable depth ranges, the effects of starting models on the average Q_u are negligible.

These resolvable averages and their standard deviations are illustrated in Figure 6 . The initial inversion is actually in terms of Q_u^{-1} because there is a linear relationship between γ_R and Q_u^{-1} ; when we translate the results to plot the traditional Q_u versus depth, the uncertainties become asymmetric about the average value.

One required feature of the attenuation structure is a significant decrease in Q_u with increasing depth. At the 95% confidence level, the average Q_u in the 30-80 km depth range must be greater than ~ 120 . The best estimates for Q_u in ~ 50 -120 km range are about 100. In the asthenosphere at depths 80-200 km, the best estimate for Q_u is ~ 60 , somewhat lower than averages of 60-90 in the asthenosphere in various global 1-D models [see summary in *Romanowicz and Durek, 2000*], but in excellent agreement with the global average of ~ 63 in the 80-220 km depth range estimated by Resovsky et al. [2005]. The important point is that Q_u is not much lower than the global average despite anomalously low shear velocities in the asthenosphere.

The Q_u models and uncertainties above assume that intrinsic attenuation is independent of frequency, as is traditional for most seismological studies. Fundamental mode surface wave observations cannot distinguish between the effects of variations in attenuation caused by frequency dependence of the attenuation mechanism and those caused by depth dependence. The form of the frequency dependence is unclear and should depend on the physical mechanism responsible for the dissipation. Recent experiments suggest that temperature and grain-boundary-sliding effects may take the form $Q^{-1} \propto \omega^{-\alpha}$, with α in the range 0.1 - 0.3. *Jackson et al. [2002]* find best fitting value of α to be ~ 0.26 for olivine in the seismic frequency band. Since the mechanism and coefficients are still controversial, rather than correcting the data to a common reference frequency, we prefer the alternative approach of finding the apparent attenuation structure neglecting the frequency dependence, then comparing the structure to predictions that

incorporate the frequency dependence of the particular model according to the depth-sensitive kernels of fundamental mode Rayleigh waves. In a section below, attenuation structure is used to calculate the anelasticitic contribution to velocity reduction.

A 1-D shear wave velocity model inverted from the average phase velocities is plotted in Figure 7c (bold line). The most striking feature of this model is a low velocity zone from 80 km to 200 km with a minimum velocity of about 4.1 km/s at a depth of about 120 km. [The low-velocity zone model observed in this study is similar to a North American model \[Van der Lee and Nolet, 1997\]](#) for southern California. This low-velocity zone underlies an upper mantle lid about 50 km thick with average velocity 4.35 km/s. The velocity contrast between the low-velocity zone and the upper mantle lid is about 6%. *Yang and Forsyth* [2006a] argued that the combination of about 80-km-thick lithosphere (crust plus 50-km-thick lid) and average shear velocity of only about 4.35 km/s in this high velocity mantle lid suggests that composition or phase state, such as the presence of dissolved water and/or partial melt in the asthenosphere, may control the thickness of the lid and the existence of the pronounced low-velocity zone, rather than the temperature structure alone since simple thermal models do not predict an 80-km-thick lithosphere and low shear velocities simultaneously. It must be recognized that the lithosphere and low velocity zone are not laterally uniform. The average velocity in the high velocity lid is lowered by the replacement of delaminated lithosphere with upwelling, hot asthenosphere beneath the eastern edge of the Sierra Nevada and Owens Valley region [*Boyd et al.*, 2004; *Yang and Forsyth*, 2006a], just as the average velocity in the low velocity zone is increased by lithospheric drips beneath the Transverse Ranges and the southern Great Valley [*Aki*, 1982; *Humphreys and Clayton*, 1990; *Zandt and Carrigan*, 1993]. A study using SH phases also found variations in thickness of the lid from the coast to eastern California [*Melbourne and Helmberger*, 2001] Large areas, however, are within about 1% of the average velocity, so the 1-D

average profile is reasonably representative of the typical structure. In the following section, with the availability of attenuation structure, we investigate the origin of the relatively thick lithosphere and the presence of a very low velocity zone.

Problems with thermal models for the origin of the LVZ

Faul and Jackson [2005] developed a model to calculate shear wave velocity (V_s) and attenuation (Q_u) for melt-free polycrystalline aggregates of olivine at upper mantle temperature and seismic frequencies by fitting experimental shear modulus and attenuation data [*Jackson et al.*, 2002]. The model calculations include the approximate effects of the intrinsic frequency dependence of Q by associating a characteristic frequency at each depth with the frequency of the Rayleigh wave most sensitive to structure at that depth. Faul and Jackson showed that after grain size was adjusted to match the shear velocity and attenuation profile beneath 100 Ma seafloor, the changes in velocity with age of the seafloor and the contrast in velocity between continental and oceanic mantle were reasonably well predicted by simple thermal models, without invoking compositional changes, melting, or the presence of water in the asthenosphere. As we show below, these models fail to match the observed attenuation beneath southern California.

Temperature in continental crust and upper mantle can be simply calculated for a model that includes heat production and conduction in a layer overlying a convective asthenosphere if the upper mantle is in a thermal steady state. Temperature and heat flow Q for the conductive layer can be calculated from *Chapman* [1986]:

$$T(z) = T_t + \frac{Q_t}{k}z - \frac{\rho H}{2k}z^2 \quad \text{and} \quad Q_b = Q_t - \rho H \Delta z, \quad (2)$$

where the subscript t and b indicate top and bottom of the layer; Δz is the thickness of the layer; k is the thermal conductivity and ρH the volumetric heat production. This equation describes the temperature down to the intersection with the mantle adiabat. Of course,

with the complex, recent tectonic history of southern California and the apparent ongoing small-scale convection, the region is undoubtedly not in steady state and heat sources will be concentrated in the crust rather than distributed evenly throughout the conductive layer. Nevertheless, the family of geotherms represented by (2) probably gives a reasonable representation of the range of temperatures that might be expected.

One of the constraints on the temperature profile is that the temperature should not exceed typical mantle adiabatic temperatures if southern California is not underlain by a hot spot. We use a potential temperature of 1300 °C. Another constraint is from heat flow data, based on which *Humphreys and Hager* [1990] estimated the most likely temperature at the base of a 30-km-thick crust in southern California to be ~800 °C with possible range from 650–850 °C. With above two constraints on the temperature profile, we calculate a series of possible upper mantle temperature profiles allowing the temperature at the Moho to vary between 650 and 800 °C and assuming a steady thermal state in the upper mantle. In the calculation, the thermal conductivity in the mantle part of lithosphere is

$$k(T) = \frac{1}{0.174 + 0.000265T} + 3.68 \times 10^{-10} T^3 \quad (3),$$

from *Jaupart and Mareschal* [1999]. The value assumed for lithospheric mantle heat production is 0.03 uWm⁻³ [*Rudnick et al.* 1998]. The actual geotherm should lie among the range of temperature profiles, although not necessarily coincide with any single profile throughout the depth range.

Based on these series of temperature profiles, we calculate shear wave velocity V_s and Q_u using *Faul and Jackson*'s model, and then compare the results of these calculations with our seismological observations. The upper mantle grain sizes in the calculation are the same as those used for oceanic upper mantle by *Faul and Jackson* [2005], i.e., 1 mm from 0 to 150 km and then linearly increasing from 1 mm at 150 km to 5 cm at 350 km. Figure 7 shows the predicted temperature, Q_u , and shear velocity

calculated using *Faul and Jackson* model compared to observed shear wave velocity and attenuation from our Rayleigh wave observations. Except for the predicted gradient in the lithosphere, shear velocities are matched fairly well with the observed profile lying in the predicted ranges. However, all the models predict much greater attenuation (lower Q_u) in the asthenosphere in the depth range 80-200 km than we observe. (Fig. 7b). The predicted minimum values of Q_u are about 20 to 25, which are much lower than the 95% confidence limits of the seismic observations. The predicted high attenuation in the asthenosphere is required to match the observed low velocities with this purely thermal model through the anelastic effect. However, the observed low attenuation (high Q_u) suggests that temperature is not solely responsible for the very low velocity zone.

One could argue that our temperature models are too simple, that Laramide age subduction cooled the lower lithosphere and may have emplaced ocean lithosphere beneath southern California, so that the lithospheric geotherm may not be in conductive equilibrium. We agree. The basic point, however, is that any model that matches the velocities in the low velocity zone primarily through the anelastic effects of temperature predicts greater attenuation than is observed. Cooling of the lower lithosphere and/or asthenosphere just makes it more difficult to match the observed shear velocities.

A model for the origin of the LVZ in southern California

We construct a velocity and attenuation model in which there is a linear transition from dehydrated to damp mantle over the depth range 70 to 120 km (Figure 8). We assume that a small amount of dissolved water affects velocities only through the anelastic attenuation effect, not through a change in elastic coefficients.

Frequency-dependent shear velocities including both anharmonicity and anelasticity terms can be expressed as [*Minster and Anderson, 1981*]:

$$V_s(\omega, T, P) = V_0(T, P) \left[1 - \frac{1}{2} \cot\left(\frac{\pi\alpha}{2}\right) Q^{-1}(\omega, T, P) \right], \quad (4)$$

where $V_0(T, P)$ is the seismic wave velocity as a function of temperature and pressure. The anharmonic effects are independent of frequency and have been well constrained from laboratory experiments. The anelastic effect is frequency dependent, assumed to be of the form $Q^{-1} \propto \omega^{-\alpha}$.

For the anharmonic velocity, we adopt the new model of *Stixrude and Lithgow-Bertelloni* [2005] based on expected mineral abundance and reactions. In the vicinity of the LVZ, they find that the anharmonic shear velocity in the upper mantle can be approximated by the equation,

$$V_o = 4.77 + 0.0380(P, z/29.8) - 0.000378(T - 300), \quad (5)$$

with pressure P in GPa, depth z in km, temperature T in K, and velocity V_0 in km/s. *Stixrude and Lithgow-Bertelloni* showed that these predicted anharmonic velocities underpredict the extent of the LVZ, i.e., do not predict low enough velocities beneath seafloor of 100 Ma age. They concluded, without specifying the mechanism responsible for the high attenuation, that the anelastic effect of attenuation on velocity is probably responsible for the discrepancy. Adopting their approximate velocity model, we show that neither the direct effects of temperature on elasticity nor the indirect effects of anelasticity explain the low velocities observed beneath southern California.

Attenuation is thermally activated and frequency dependent, usually represented in the form

$$Q^{-1}(\omega, T, P) = A[\omega \exp\{(E+PV)/RT\}]^{-\alpha} \quad (6),$$

where ω is angular frequency, P is pressure, T temperature, E is activation energy, V activation volume and α is a constant that laboratory measurements and seismic observations indicate is between 0.1 and 0.3.

For simplicity, the enhancement of attenuation from the addition of water is assumed to occur through a change in A in equation (6), amounting to an enhancement of a factor of three from dry to damp, with no change in E, V, or α . The sensitivity of velocity to a change in temperature is given by

$$\frac{\partial V_s}{\partial T} = \frac{\partial V_o}{\partial T} - V_o \left[\frac{\alpha\pi}{2} \cot \frac{\alpha\pi}{2} \frac{Q^{-1}}{\pi} \frac{E + PV}{RT^2} \right] \quad (7).$$

In our model, the temperature sensitivity is reduced by choosing $E = 250 \text{ kJ mol}^{-1}$ smaller than in the model of *Faul and Jackson* [2005] and by the decrease in Q^{-1} associated with dehydration above 120 km. Activation volume is the same as in the model of *Faul and Jackson* with $V = 1.0 \times 10^{-5} \text{ m}^3 \text{ mol}^{-1}$. α is assigned a minimum value of 0.1 to maximize the effect of a given attenuation on velocity (eq.7).

The results are shown in Figure 9 with comparison to observations. The model agrees reasonably well with the constraints on Q and with the changes in velocity as a function of depth except in the depth range of 80-200 km. The anelastic correction for shear velocity is about 0.05-0.15 km/s with the largest value at 120 km corresponding to the largest dissipation (Q_u^{-1}). The calculated Vs can fit the average, observed shear velocities above ~ 70 km of 4.35 km/s if the hottest geotherm or thinnest thermal boundary layer is adopted. Between 80 and 200 km, observed shear wave velocities are much lower than the calculated Vs using any of this family of temperature profiles. The largest difference of about 0.2 km/s appears at ~ 120 km. The discrepancy would be larger if we use $\alpha = 0.26$. In the following section, we discuss possible origins for the velocity discrepancy in the asthenosphere.

Factors affecting velocity

One possible factor affecting predicted velocities in the low-velocity zone is the potential temperature of the upper mantle. Using equation 7, to reduce the velocity by 0.2 km/s at 120 km would require a potential temperature increase of about 290 °K. But, accompanying this temperature increase would be a decrease in the predicted Q to about 40, significantly below the observed value. To keep Q within observed bounds when increasing the temperature requires further alteration of some of the constants in equation 6 and an even larger temperature change, corresponding essentially to the change required for the purely anelastic effect (equation 5), or close to a 500 °K increase in potential temperature. These changes are much too high to be possible beneath southern California. Using accepted values of elastic constants, the expected form of frequency-and-temperature-dependent attenuation and reasonable values of mantle temperature, one can fit either the attenuation or the seismic velocities, but not both. Thus, temperature alone apparently cannot cause the very low shear velocities in the asthenosphere.

Variations of the bulk composition of the upper mantle can change seismic velocities as well. The major variation of the bulk compositions is due to partial melting since partial melting depletes the source rock substantially. *Schutt and Lesher* [2006] have experimentally shown that melt depletion cannot cause more than about a 0.5% change in mantle velocity. Similarly, *Stixrude and Lithgow-Bertelloni* [2005] show the variation of shear velocity is less than 1% within a plausible range of bulk compositions in the upper mantle. *Lee* [2003] shows the maximum shear velocity variation due to depletion is less than 1.5%. All of these studies support that bulk composition cannot be responsible for the 5% discrepancy between the observed and predicted shear velocity in the asthenosphere.

Shear velocities above the low velocity zone are in the range predicted by reasonable geotherms, but only if the thermal boundary layer is no thicker than 50 km. The transition from high velocity lid to low velocity zone centered at ~ 80 km is

apparently not controlled by temperature. The transition may be controlled by a compositional or state change, such as the presence of water or melt in the low-velocity zone with the base of the high-velocity lid representing a dehydration boundary or the solidus, or by a change in seismic anisotropy.

The presence of dissolved water can reduce seismic velocities through anelastic relaxation. Experimental studies of existing data on Q [Sato *et al.*, 1989 and Jackson *et al.*, 1992] and theoretical analysis suggest that water can enhance attenuation significantly. For example, Jackson *et al.* [1992] found that dunite specimens that were pre-dried have Q values 2-3 times larger than specimens that were not pre-dried. Karoto and Jung [1998] argued that the sharp velocity contrast around 60-80 km in the old oceanic upper mantle could be attributed to a sharp change in water content. The increase of attenuation from 70 to 150 km with increasing depth may be at least partially attributable to an increase in water content. Since the presence of water reduces seismic velocity via anelasticity with frequency dependence thought to be similar in form to the effects of temperature, the effect has been accounted for in the anelastic corrections (eq. 4) in Figure 9 that match the observed attenuation data. To lower the velocities further with addition of water would require lowering Q beyond the observed limits. Thus the presence of water cannot explain shear velocities as low as 4.1 km/s.

A change in anisotropy from the lithosphere to the asthenosphere could potentially contribute to the drop in apparent shear velocity. Our inversion for shear velocity effectively finds Sv , not the isotropically averaged S velocity assumed in the Stixrude and Lithgow-Bertelloni [2005] model, since we use only horizontally propagating Rayleigh waves. From the discrepancy between Love and Rayleigh waves, Polet and Kanamori [1997] showed that the upper mantle is radially anisotropic beneath southern California with $SH > Sv$ down to depths of about 220 km. If the anisotropy is due to the preferred horizontal alignment of olivine a-axes, however, the primary effect will be an increase in

the azimuthally averaged Love wave velocity compared to an isotropic earth, not a decrease in average Rayleigh wave velocity and inferred Sv velocity [Maupin, 1985; Montagner and Anderson, 1989]. The decrease in apparent Sv velocity should be less than 1%. In addition, Yang and Forsyth [2006a] found that the azimuthal anisotropy of Rayleigh waves in southern California decreased slightly over the 25 to 140 s period range, with no indication of an increase in anisotropy in the asthenosphere that could be responsible for the drop in velocity from the lid to the low velocity zone.

Grain size can also influence the velocity, as pointed out by Faul and Jackson [2005]. However, the grain size effect on velocity appears only indirectly through the anelastic effect. Since we have already corrected the elastic or anharmonic velocities for the actual observed attenuation, a change in grain size cannot be responsible for the anomalously low velocities.

The most plausible factor left to reduce seismic velocities is the presence of partial melt beneath the lithosphere. Partial melting has often been invoked to explain the origin of the low velocity zone in the oceanic upper mantle [Anderson and Sammis, 1969; Sato *et al.*, 1989, Yang *et al.*, 2007]. Partial melt has two different effects on seismic velocity: the direct effect due to the difference of elastic properties between melt and solid, and the indirect effects due to the enhanced anelasticity with the presence of partial melt. The anelastic effect depends sensitively on the mechanism of attenuation associated with melting. If it is melt squirt, then the dissipation peak may lie outside the seismic frequency band, so that the elastic moduli are relaxed but there is little attenuation of surface waves. The effect of melting-related grain boundary sliding is expected to affect the background attenuation in much the same way as temperature, although there may also be an attenuation peak in the seismic frequency band [Jackson *et al.*, 2004; Faul *et al.*, 2004]. . The direct velocity reduction for melt squirt depends on the aspect ratio of the melt pockets [Schmeling, 1985]. Hammond and Humphreys [2000a] have shown that direct Vs reduction per percent partial melt is at least 7.9% for realistic melt distributions

in peridotite with melt fraction around 1%. If we assume the discrepancy of shear velocities is totally due to the presence of partial melt and the effects of partial melt on seismic velocities are predictable by the melt squirt model of *Hammond and Humphreys* with no significant effect expected on attenuation [Hammond and Humphreys, 2000b] , we can estimate a melt fraction of less than 1% in the asthenosphere distributed from about 70 to 200 km (Figure 8).

The presence of partial melt can thus explain the observed low velocities in the asthenosphere beneath southern California (Figure 8 and 9). The specific model we propose is clearly ad hoc and non-unique; the point is simply to illustrate a possible way out of the dilemma posed by existence of very low shear velocities with only moderate attenuation. The presence of melt has also been suggested to explain the observed very sharp seismic velocity gradient at the base of lithosphere in eastern North America [Rychert *et al.*, 2005]. The mechanism for generating melt beneath eastern North America could be decompression melting of mildly hydrated asthenospheric material as it flows upward along the contours of the more rigid, shallowing lithosphere in response to the west-southwest absolute motion of the North American plate. The problem is how partial melt can be generated beneath southern California. In southern California, about 30 million years ago, a section of the Pacific-Farallon spreading center began to be consumed under the North American plate, resulting in the development of the San Andreas transform boundary. As the subduction of the spreading center continued, a slab window was left under southern California as the whole Farallon plate sank toward the east. The space occupied by the slab was replaced by upwelling asthenospheric mantle which could have begun to melt as it decompressed. The presence of a small amount of water would allow partial melting to begin at depths as great as 200 km [e.g., *Green*, 1971; *Wyllie*, 1971; *Hirth and Kohlstedt*, 1996], although the melt fraction would remain small until the dry solidus was passed at depths above ~ 70 km.

Conclusion:

A one-dimensional shear velocity model reveals a pronounced low velocity zone (LVZ) from ~80 km to ~200 km in the upper mantle beneath southern California. Models that attribute the variations of attenuation and shear velocities with depth solely to temperature and pressure effects predict too low Q values if they match the shear velocities. Partial melt may explain the observed low velocity zone and water in the asthenosphere may contribute to the attenuation. Partial melt in the asthenosphere could be generated due to decompression and the reduced solidus for damp mantle when the asthenosphere rose to fill the space left by the subducted Farallon plate. If partial melt and water have significant affects on velocity and attenuation, then the temperature sensitivity of shear velocity is probably less than has been estimated in recent models attributing the properties of the low velocity zone to temperature and pressure alone.

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Figure Captions:

Figure 1. Topography of southern California. Triangles represent broadband three-component seismic stations used in this study.

Figure 2. Azimuthal equidistant projection of earthquakes used in this study. The plot is centered on the center of the selected stations. The straight lines connecting each event

to the array center represent the great-circle ray paths. Note the good azimuthal coverage of the events.

Figure 3. Phase velocity variation at 25 sec. The velocity anomalies are calculated relative to average phase velocity at 25 sec in southern California.

Figure 4. Variation of residual amplitudes with propagation distance at periods of 25, 50 and 125 sec. Each circle represents the average residual amplitude in each 50-km distance interval over all used events. Error bars represent the standard deviations of the average residual amplitudes from an inversion that takes into account focusing, complexity of the incoming wavefield, and station corrections, but ignores attenuation. Note the systematic decay trend of residuals with distance indicating attenuation is required.

Figure 5. Attenuation coefficients of surface waves obtained from Rayleigh wave tomography. Thin line with circles is the observed values with error bars representing +/- one standard deviation. Bold line is the predicted values for the particular attenuation model shown in Figure 6.

Figure 6. A Q_u model (left) and resolution kernels (right). The resolution kernels are plotted as a function of depth for target depths of 30,60,80,120 and 150 km. Each kernel corresponds to one row of the resolution matrix. [The scale of each dotted panel is 0.4.](#) The average Q_u values in three well-resolved depth intervals are plotted on the left as the thin dash-dot lines. The horizontal extent of each rectangle represents +/- one standard error of the mean over that depth interval. The dashed line is the starting

model for Q_u inversion and the solid line is the particular solution, which has predicted attenuation coefficients shown as the bold solid line in Figure 5.

Figure 7. Predicted temperature (a), Q_u (b), and shear wave velocity (c) profiles (gray lines) based on the model of *Faul and Jackson* [2005]. The dashed line in (a) shows the 1300°C adiabat. Well-resolved average Q_u values in three depth intervals are plotted in (b). The meaning of each rectangle is the same as figure 6. Observed shear-wave velocities from our study are plotted in (c) using the bold line.

Figure 8. Schematic model for the origin of the low-velocity, high attenuation zone beneath southern California. The mantle lithosphere is dehydrated, with a transition to damp mantle between 70 and 120 km causing an increase in attenuation. A small amount of partial melt is present from 70 to 200 km, reducing the shear velocity in that depth range.

Figure 9. Predicted temperature (a), predicted Q_u (b) and shear wave velocity (c) based on our proposed model (eq. (6) and (7) and Figure 8). Thin black lines in (c) are the anharmonic velocities calculated based on equation (5); thin gray lines are the velocities calculated based on equation (4) including the anelastic effects; the bold gray line is the velocities for the hottest geotherm including partial melt effects, i.e., our preferred model. Vs reduction per percent partial melt is assumed 7.9% in this calculation. The amount of partial melt is shown in Figure 8.

















