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Structures of the oceanic lithosphere-asthenosphere boundary: Mineral-physics modeling and seismological signatures

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[1] We explore possible models for the seismological signature of the oceanic lithosphere-asthenosphere boundary (LAB) using the latest mineral-physics observations. The key features that need to be explained by any viable model include (1) a sharp (<20 km width) and a large (5-10%) velocity drop, (2) LAB depth at $\sim70 \text{ km}$ in the old oceanic upper mantle, and (3) an age-dependent LAB depth in the young oceanic upper mantle. We examine the plausibility of both partial melt and sub-solidus models. Because many of the LAB observations in the old oceanic regions are located in areas where temperature is $\sim1000-1200^{\circ}$ K, significant partial melting is difficult. We examine a layered model and a melt-accumulation model (at the LAB) and show that both models are difficult to reconcile with seismological observations. A sub-solidus model assuming absorption-band (AB) physical dispersion is inconsistent with the large velocity drop at the LAB. We explore a new sub-solidus model, originally proposed by *Karato* [2012], that depends on grain-boundary sliding. In contrast to the previous model where only the AB behavior was assumed, the new model predicts an age-dependent LAB structure including the age-dependent LAB depth and its sharpness. Strategies to test these models are presented.

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1. Introduction

[2] The lithosphere and the asthenosphere are defined through their mechanical properties. The physical reasons for a strong lithosphere over a weak asthenosphere are important, influencing how plate tectonics operates on Earth. Mechanical definitions based on seismicity, plate bending (flexure), and gravity anomalies have yielded "lithosphere" with thicknesses from 20 km to 200 km in the same regions [*McKenzie*, 1967; *Nakada and Lambeck*, 1989; *Peltier*, 1984; *Karato*, 2008]. For example, using post-glacial-rebound observations, *Peltier* [1984] inferred ~200 km continental lithosphere whereas *Nakada and Lambeck* [1989] inferred ~50-km continental lithosphere, making it difficult



to investigate causal influences on lithosphereasthenosphere boundary (LAB) depth.

[3] In addition to the mechanical constraints on LAB, other proxies have defined the LAB depth, e.g., petrology, temperature, electrical conductivity, and seismic wavespeeds [*Eaton et al.* 2009]. These different proxies have their advantages, but in this paper, we argue that the seismic wavespeeds best define the LAB [e.g., *Jones et al.*, 2010; *Behn et al.*, 2009].

[4] Puzzling results in both the oceanic and continental regions suggest that existing models for the "seismological" LAB need to be revisited. In the oceans, Kawakatsu et al. [2009] and Kumar and *Kawakatsu* [2011] observed a shallow (~60 km) and large velocity drop ($\sim 7\%$), in some old (~100-130 Ma) oceans. In the old continents, strong mid-lithospheric discontinuities (MLD) are observed at the depth of ~100-150 km, but no strong signals at depths ~200 km, where we would expect the LAB [Abt et al., 2010; Ford et al., 2010; Rychert and Shearer, 2009; Rychert et al., 2007; Thybo, 1997; 2006]. Karato [2012] argues that partial melting, a popular model for the LAB, is not expected at these depths in these regions, nor can the inferred large velocity drop be explained readily by a standard absorption-band (AB) anelastic model.

[5] In this paper, we review important shortcomings to the previous partial melt and sub-solidus models for the oceanic LAB. Then, we develop a new sub-solidus model, based on grain-boundary sliding, originally proposed by Karato [2012]. Our analysis extends the previous study via a new statistical analysis of experimental data to provide robust constraints on key parameters, allowing a more detailed discussion of their uncertainties. We introduce a new parameterization for grain-boundary-sliding attenuation and velocity reduction that captures better the features of theoretical models, while being functionally simple. We also explore a variety of published thermal models for the oceanic sea floor [Davies, 1988; McKenzie et al., 2005; Parsons and Sclater, 1977; Stein and Stein, 1992] to characterize the sensitivity of our results to uncertainties in thermal models, as well as exploring the age dependence of oceanic LAB structures. Models for the seismological observations of continental LAB will be discussed in a separate paper where we will discuss the origin and significance of MLD and the reasons for weak seismic signals from the supposed LAB using the mineral-physics models similar to those developed here.

2. Models of the Oceanic LAB

2.1. Difficulties with the Partial-melt Models

[6] A popular model to explain a sharp, large velocity drop at the LAB is the onset of partial melting [Anderson and Sammis, 1970; Anderson and Spetzler, 1970; Lambert and Wyllie, 1970]. In order for a partial-melt model of LAB to be valid, two conditions must be met. First, the temperature at the LAB must exceed the solidus. Second, the melt fraction below the LAB must be large enough to cause a large velocity reduction. To reduce velocity by 5–10% in the shallow upper mantle, one will need 3–6% melt fraction [Takei, 2002]. We evaluate below various partial-melt models with special attention to these two points.

2.1.1. Thermal Structures of the Lithosphere-asthenosphere

[7] In all partial-melt models, it is essential that the LAB temperature corresponds to the depth at which geotherm intersects the solidus. The simplest model that prescribes the temperature distribution in the oceanic upper mantle is the halfspace-cooling model [Davis and Lister, 1974; Turcotte and Schubert, 2002]. In this model, only one free parameter, the difference ΔT between the surface temperature and the potential temperature, prescribes completely the temperature distribution. As the plate moves away from the ridge, it cools conductively with time following the \sqrt{t} age dependence. The typical value for ΔT is 1350°K, which is within the bounds of petrological constraints on the ambient mantle potential temperature, $T_p \sim 1550 - 1600^{\circ} \text{K}$ [McKenzie] and Bickle, 1988]. The model predicts ~1030°K at 60 km in the 120 my old oceanic upper mantle. We conclude that the observed shallow depths of the LAB in the old oceanic upper mantle are difficult to reconcile with a partial-melt model with these geothermal structures.

[8] In the plate model, one assumes that the temperature at a certain depth is fixed, e.g., by the onset of small-scale convection [*McKenzie et al.*, 2005; *McKenzie*, 1967]. Consequently, plate models in general predict higher temperatures in the old oceanic upper mantle. The temperature at the LAB (e.g., 60 km in 120 Ma mantle) for plate models depends on the choice of plate thickness and the base temperature. *McKenzie et al.* [2005] provided one such plate model with a potential temperature of 1588°K (constrained by the crustal



thickness). Their model shows ~1100°K at 60 km in the 120 my old oceanic upper mantle, too low for partial melting to occur. However, if one chooses a higher potential temperature (1673°K), one could have partial melting at 60 km in 120 Ma old oceanic upper mantle (e.g., *Hirschmann* [2010], see Figure 1). If the system is closed and static, then the amount of melt that exists at these conditions is controlled by the pressure, temperature and composition (particularly the volatile content) and is less than ~0.3%. This is much below the melt fraction required by the velocity reduction (3–6%).

2.1.2. Processes of Melt Accumulation

[9] The above argument suggests that a viable partial-melt model must invoke an inhomogeneous melt distribution. *Kawakatsu et al.* [2009] used the laboratory observations of *Holtzman et al.* [2003] to suggest a layered asthenosphere within which thin layers of high melt fraction overlie melt-poor layers. If melt-rich layers occupy a substantial volume fraction and if the velocity reduction in these layers is large enough, one could explain the observed large

velocity reduction at the LAB. A large velocity reduction predicted by this model is associated with anisotropy induced by the layered structure. The velocity drop at the LAB equals the magnitude of V_{SH}/V_{SV} anisotropy. In other words, if this model were valid, one should expect $\frac{V_{SH} - V_{SV}}{\langle V_S \rangle} \sim 5-10\%$ in the asthenosphere. Such anisotropy is significantly larger than that indicated by seismological observations (~2%) [Ekström and Dziewonski, 1998; Montagner and Tanimoto, 1990; Nettles and Dziewoński, 2008], at least in the broad lateral averages that characterize surface-wave tomography. Another problem with this model is that the results of Holtzman et al. [2003] showed that their melt-rich layers are not parallel to the shear plane but tilted by $\sim 20^{\circ}$ relative to it. If the melt-rich layers are tilted, then buoyancy will remove the melt in these layers, and it will be impossible to maintain high melt fraction.

[10] An alternative melt-accumulation process involves vertical transport of melt in the asthenosphere, so that melt could accumulate near the LAB at or near the solidus. This localized mechanism does not predict large V_{SH}/V_{SV} anisotropy throughout the



Figure 1. Top: Recent seismic measurements of the LAB overlaid on temperature contours using three thermal models: halfspace cooling model (left panel), plate models of [*McKenzie et al.*, 2005] (middle panel), and [*Hirschmann*, 2010](right panel). Temperature contours are annotated in steps of 200°K, with symbols plotted at the seismologically inferred LAB depths. Squares are from [*Kawakatsu et al.*, 2009; *Kumar and Kawakatsu*, 2011], circles from [*Rychert and Shearer*, 2011]. Dark circles represent anomalous crust; open circles represent normal crust using the classification of [*Korenaga and Korenaga*, 2008]. Histogram plots are the temperature distribution calculated at each age, depth point. All three thermal models show a temperature structure confined by an upper bound of 1600°K. The plate model by [*Hirschmann*, 2010] was calculated using the mantle potential temperature of 1673°K and plate thickness of 125 km.



asthenosphere. In addition, there must be a process to maintain a large melt fraction (3–6%) near the LAB to a layer that is thick enough to produce a strong seismic signal. The structure of such a region (i.e., the melt fraction and its variation with depth) is controlled by the competition of gravity-induced compaction, melt accumulation, melt generation, and transport. Therefore, one needs to have a specific combination of solid-matrix viscosity, permeability, etc. to maintain a layer that explains the seismological observations. None of these parameters are constrained well, and therefore we conclude that this model remains speculative.

2.1.3. Seismological Consequences of an Accumulated Melt Model for the LAB

[11] As a further test, we compute Ps and Sp receiver functions using synthetic seismograms for velocity variations implied by a melt-accumulation model (Figure 2). To detect a melt-rich layer in the RF signals, several conditions must be met: (1) the wavelength of the seismic signal must be comparable to the layer thickness, or smaller, and (2) the velocity (impedance) jump must be strong enough to make the signal visible [e.g., *Leahy*, 2009; *Leahy and Park*, 2005]. With sensitivity



Figure 2. Synthetic receiver functions computed using a model given in (D), which represents a low-velocity layer (LVL) embedded at a depth of 70 km, as would be required by a melt layer with $\sim 7\%$ reduction in shear wave velocity. The lower panel describes schematically the conversions for an incident P phase (C) or an incident S phase (E). The synthetic seismograms for the P and S phases show conversions on the vertical (Z, in red) and radial (R, in blue) components. Receiver functions are then computed from this synthetic data using the method of [*Park and Levin*, 2000]. The Ps receiver function shows a closely spaced double polarity conversion following the initial pulse at time zero. The Psd₁ phase, which is generated from the top of the melt layer, is followed by the Psd₂ phase, generated from the bottom of the layer. In the case of the Sp receiver functions, the sequence is reversed (B). The double polarity signal is indicative of a low-velocity layer. In our computation of the receiver functions, we rotate the synthetic seismograms from the ZRT coordinates into the LQT coordinates to minimize the energy at zero time, which enables a better resolution of the converted phases. Amplitudes in the first 4 s for the radial and vertical component of the synthetic seismogram are scaled by 7% for display purposes. The Ps and Sp receiver functions are computed at a cut-off frequency of 1 Hz and a melt layer thickness of 20 km.



tests, we demonstrate the limitations of receiver functions for detecting a melt-rich (low-velocity) layer of variable thickness (Figure 3). Based on the receiver-function technique of *Park and Levin* [2000], (see also *Leahy et al.* [2012]), a cut-off frequency of 1 Hz would resolve a melt layer thickness >5 km. In a more general case, the peakto-peak amplitude depends on the magnitude of the velocity drop: the larger the magnitude of the velocity drop, the larger the peak-to-peak amplitude (Figure 4). In real data, high noise levels will make low amplitude RF signals difficult to detect.

[12] Another aspect of a melt-accumulation model is a possible signal from the bottom boundary. If both top and bottom boundaries are sharp (Figure 2), there will be a double-polarity signal: a negative phase (Psd₁ and Spd₂) generated from conversions at the top of the melt layer (the inferred LAB in most seismic studies), followed by a positive phase (Psd₂ and Spd₁), generated from conversions at the bottom of the low-velocity layer (LVL).

[13] However, the bottom boundary of the meltaccumulation zone could be gradational. Such a diffuse boundary would not produce sharp converted phases from the lower boundary of the meltaccumulation layer (Figure 5). For sharp velocity gradients (<15 km), the pulse from a bottom boundary should be detectable at frequencies >0.3 Hz. A broader velocity gradient >20 km requires computation of the receiver function in multiple frequency bands to clearly distinguish the pulse. For instance, a gradient width of 40 km at 0.3 Hz is so wide; such a pulse could be buried in the background noise. At higher frequencies, however (e.g., 1.5 Hz), the pulse



Figure 3. Sensitivity tests for Ps receiver functions through a low-velocity layer with varying thickness. The magnitude of the velocity drop is 7%. The peak-to-peak amplitude and time separation (A) are measured. For a thickness of 5 km and 0.3 Hz, the low-velocity layer is hard to detect (A, lower panel). For a thickness of 20 km and 1 Hz, the layer is easily identified (A, top panel). The cut-off frequency for detection becomes lower with larger thickness. The time separation between the negative and positive polarity, increases as thickness increase (C). The estimate of time separation also becomes stable at high frequency, since it is difficult to estimate the timing due to the broadening of the pulse in time.





Figure 4. Sensitivity tests for Ps receiver functions through a low-velocity layer showing how changes in the magnitude of the velocity drop $\delta V/V_{o}$ affect the peak-to peak amplitude. The larger the amplitude of the velocity drop, the larger the peak-to-peak amplitude (A and B). The detection of a low-velocity layer is dependent on the thickness of the low-velocity layer, and the amplitude of the velocity drop (C): the larger the amplitude, the easier it is to detect.

becomes discernable with a 6 s pulse width. Using multiple frequency bands, which take advantage of the high amplitude at low frequencies, and narrower pulse width at high frequencies, diffuse boundaries could be detected (Figures 3, 4, and 5).

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[14] Current RF studies show only a single converted pulse, either Sp or Ps, inconsistent with a meltaccumulation model with a sharp bottom boundary or with a thin melt-rich layer. The typical frequencies of LAB detection are lower than the cut-off frequency of 1 Hz for detection of a thin melt-rich layer. Re-evaluation of seismic data at higher frequencies would provide better test of the melt-accumulation model, but may be feasible only with Ps receiver functions. In comparison, receiver-function studies of the upper mantle above the 410 km discontinuity [Tauzin et al., 2010] show such a double-lobed polarity, consistent with a thick partial-melt layer.

2.2. Difficulties with the AB Version of Sub-solidus Model

[15] Karato and Jung [1998] provided a model to explain a sharp LAB velocity drop using a subsolidus model, exploring the role of a sharp change in water content across ~ 70 km depth. They hypothesized that water enhances anelastic relaxation and hence results in a sharp velocity drop at the LAB. Karato and Jung [1998] assumed a simple model of anelastic relaxation where velocity reduction is directly related to the seismic attenuation Q^{-1} , viz., the "AB" model of Anderson and Given [1982] and Minster and Anderson [1981]

$$\left(\frac{\delta V}{V_o}\right)_{ABM} = \frac{1}{2} \cot\left(\frac{\pi\alpha}{2}\right) Q^{-1}(\omega, T, P, C_w) \approx Q^{-1} \qquad (1)$$

where ω is the frequency, T, is the temperature, and C_w is the water content.

[16] In the asthenosphere, $Q \sim 80$ [Dziewonski and Anderson, 1981]; therefore, it is expected that the velocity drop will be small (< 1%). This result is not consistent with recent seismic observables that show a 5-7% drop in seismic wave velocities. Assuming the AB model, it will require much lower Q as pointed out by [Yang et al., 2007]. Since this is not the case, another mechanism must be required that does not affect Q but reduces seismic wave velocity in the observed frequency band. Such a mechanism has been proposed by Karato [2012]



Figure 5. Sensitivity tests for Ps receiver functions through a diffuse velocity gradient with varying gradient width (A). The pulse width and the amplitude are measured (B) and depend on the seismic frequency as well as the gradient width. For sharp velocity gradients (<15 km), a distinct pulse with maximum amplitude is resolved at frequencies >0.3Hz (C and D). Diffuse velocity gradients (>20 km) require that the receiver function be computed at multiple frequency bands for clear resolution, since lower frequencies have a broad pulse (B, lower panel) and higher frequencies have low-amplitudes but a sharper pulse (B, top panel). Combining multiple frequencies and checking persistence across the frequency band can help resolve the character of a melt layer with a gradient in melt fraction.

(see also [*Karato*, 1977]), in the form of a high-frequency dissipation peak due to grain-boundary sliding, which we discuss next.

3. A Modified Sub-solidus Model

3.1. Introduction

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[17] Theoretical studies on the deformation of polycrystalline material have demonstrated that anelastic relaxation and long-term creep in these materials occur as a successive process of elastically accommodated grain-boundary sliding followed by accommodation via diffusional mass transport [*Ghahremani*, 1980; *Kê*, 1947; *Morris and Jackson*, 2009; *Raj and Ashby*, 1971; *Zener*, 1941]. These studies also showed that the magnitude of the modulus reduction (relaxation) caused by the elastically accommodated grain-boundary sliding can be as much as $\sim 30\%$, and hence the influence of such a process on the reduction of seismic wave velocities can be substantial. However, until recently, this relaxation mechanism has been difficult to observe in experimental studies of upper-mantle materials. As a result, its implication for seismic structure of the LAB has not been explored.

[18] Due to significant improvements in experimental methodology and empirical parameterizations, some groups [*Jackson and Faul*, 2010; *Sundberg and Cooper*, 2010] have reported observations of the grain-boundary mechanism in mantle peridotites. The experimental evidence of a high-frequency



dissipation peak confirms theoretical studies and shows that the AB model alone is insufficient to characterize fully the anelasticity at upper-mantle conditions [e.g., *Karato*, 2012]. The relevant "seismological" effect of a high-frequency peak superimposed on the AB model is a significant reduction in velocity (caused by the modulus relaxation), as the characteristic frequency of grainboundary sliding ω_{gbs} shifts relative to the seismic frequency ω_s (see Figure 2 of *Karato* [2012]). A possible explanation for a large velocity drop (with a modest Q) at the LAB is for $\omega_{gbs} < \omega_s$ in the lithosphere, but $\omega_{gbs} > \omega_s$ in the asthenosphere.

[19] In its simplest form, the influence of grainboundary sliding can be understood using a single-Debye-peak model where the attenuation and associated velocity reduction are given by:

$$Q^{-1}(\omega) = \Delta_p \frac{\left(\frac{\omega_s}{\omega_{gbs}}\right)}{1 + \left(\frac{\omega_s}{\omega_{gbs}}\right)^2}$$
(2)

$$\left(\frac{\delta V}{V_o}\right)_{GBS} = \frac{\Delta_p}{2} \frac{1}{1 + \left(\frac{\omega_s}{\omega_{gbs}}\right)^2} \tag{3}$$

where Δ_p is the relaxation strength of the Debye peak, ω_s is the seismic frequency, and ω_{gbs} is the characteristic frequency of the peak mechanism (grain-boundary sliding). We focus on the following three points: (1) the sharpness of the velocity change, (2) the amplitude of the velocity change, and (3) the depth at which a sharp and large velocity change occurs. According to this model, the depth variation in seismic wave velocity is caused by the depth variation of the characteristic frequency of grain-boundary sliding (ω_{gbs}), in addition to the contribution from an anharmonic effect. When $\omega_{gbs} > \omega_s$, the velocity becomes lower. If ω_{gbs} changes abruptly with depth, one expects a sharp velocity change. Such abruptness is plausible for a water-based effect because mantle water content could change abruptly due to dehydration [Hirth and Kohlstedt, 1996; Karato and Jung, 1998]. If the frequency range covered by a shifting ω_{gbs} is entirely within the AB regime, then the variation in seismic wave velocity is limited by the value of Q. For a typical Q (~80), the amplitude of velocity change is <1%. When the characteristic frequency ω_{gbs} shifts across the seismic frequencies, then the GBS peak has greater influence on seismic velocities.

[20] To explain the relatively sharp and large velocity drop observed at the LAB, it is critical to constrain the details of the GBS relaxation (peak frequency, anelastic relaxation strength, and the sharpness of the peak). The temperature dependence of attenuation and velocity reduction comes from the temperature dependence of characteristic frequencies, ω_{AB} and ω_{GBS} . Because these characteristic frequencies are the inverse of characteristic times that are connected to microscopic "viscosity", ω_{AB} and ω_{GBS} also likely depend on water content as [e.g., *Karato*, 2003; *Karato*, 2006; *McCarthy et al.*, 2011], viz.,

$$\frac{\omega(T, C_w)}{\omega_o(T_o, C_{wo})} = \left(\frac{d}{d_o}\right)^{-m} \left(\frac{C_w}{C_{wo}}\right)^r \exp\left[-\frac{E_p}{RT_o}\left(\frac{T_o}{T} - 1\right)\right] (4)$$
$$\exp\left[\frac{V_p}{R}\left(\frac{P}{T} - \frac{P_o}{T_o}\right)\right]$$

where T_o , C_{wo} , d_o , and P_o are the reference temperature, water content, grain size, and pressure, respectively; E_p and V_p are the activation energy and the activation volume; and r and m are nondimensional constants that describe the sensitivity of the ω_{gbs} to water content and grain size. In the case of a dry mantle, r=0, the depth variation in ω_{gbs} is solely caused by temperature variation (with a small effect of pressure), and ω_{gbs} increases gradually with depth. Therefore, the LAB should be gradual, the width depending on the temperature gradient. However, for the case of a wet mantle, r=1 or 2, the change in characteristic frequency can be sharp if water content changes abruptly, which would cause a sharp drop in velocity (see Figure 3 of *Karato*, [2012]).

3.2. Influence of the Peak Width

[21] The actual GBS relaxation peak may be more diffuse than the simplest model, due to the distribution of relaxation peaks. This influences the sharpness of the velocity gradient. Therefore, we extend the approach taken by *Karato* [2012] by modifying $\left(\frac{\delta V}{V_o}\right)_{GBS}$ to include the broadness of the high-frequency peak observed in laboratory measurements [*Jackson and Faul*, 2010; *Sundberg and Cooper*, 2010]. The relaxation time for grain-boundary sliding is

$$\tau = \frac{\eta_b d}{G_u \delta} \tag{5}$$

where η_b is the grain-boundary viscosity, $_d$ is the grain size, G_u is the unrelaxed shear modulus, and δ is the grain-boundary width. The distribution of δ and η_b produces a distribution of relaxation times. In their experimental study of aluminum and magnesium oxide polycrystalline ceramic aggregates, *Pezzotti* [1999] observes a broadening of the Debye peak, which they attribute to the distribution



of grain-boundary types. Also, a theoretical study by *Lee and Morris* [2010] demonstrates that the grainboundary peak broadens when the viscosity and/or grain size varies within a bulk specimen.

[22] We apply a semi-empirical approach to parameterize the broadening of the grain-boundary peak. The physical model employed by *Jackson and Faul* [2010] incorporates this broadening with a lognormal distribution for the relaxation time around a dominant peak centered at ω_{gbs} :

$$D_{p}(\tau) = \sigma^{-1} (2\pi)^{-1/2} \exp\left\{\frac{-\left[\ln(\tau\omega_{gbs})/\sigma\right]^{2}}{2}\right\}$$
(6)

[23] This formulation of a relaxation timescale distribution $D_p(\tau)$ introduces an extra parameter σ , which describes the broadening of the grain peak around ω_{gbs} . The σ parameter broadens the attenuation function and weakens its peak intensity. In this study, we model this same effect by using a Lorentzian form for the attenuation function:

$$Q^{-1}(\omega,\beta) = \Delta_p(\beta) \frac{\left(\frac{\omega_s}{\omega_{gbs}}\right)^{\beta}}{1 + \left(\frac{\omega_s}{\omega_{gbs}}\right)^{2\beta}} \tag{7}$$

where the β parameter, similar to the σ parameter, defines the broadening around the peak, and the shrinking of the peak intensity is captured by Δ_p (β). This form provides us with a convenient way to explore plausible β values that are acceptable by the current experimental data. We describe the transformation [$\sigma \Rightarrow \beta, \Delta_p(\beta)$] by fitting our functional form to the physical parameterization reported by *Jackson and Faul* [2010], (Appendix A). This results in a modification of equation (3) to give:

$$\left(\frac{\delta V}{V_o}\right)_{GBS} = \frac{\Delta_p}{2} \frac{1}{1 + \left(\frac{\omega_s}{\omega_{gbs}}\right)^{2\beta}} \tag{8}$$

with this modification, we have a combined equation for relative velocity change parameterized by the variables: $(\alpha, T(z,t), \beta, r)$:

$$\left(\frac{\delta V}{V_o}\right)_{tot}(\alpha, T(z, t), \beta, r) = \left(\frac{\delta V}{V_o}\right)_{ABM} + \left(\frac{\delta V}{V_o}\right)_{GBS} \tag{9}$$

where α is the frequency dependence from the AB model: $Q^{-1} \sim \omega^{-\alpha}$, T(z,t) is the temperature calculated at depth *z*, *t* is age, β is the grain-boundary broadness parameter, and *r* is the water-sensitivity exponent. We fix the parameter $\alpha = 0.3$, constrained by experiments.

[24] Again, it is instructive to note that the velocity reduction $\left(\frac{\delta V}{V_o}\right)_{ABM}$ associated with the AB model alone is minimal, ~1%. Therefore, we argue that the most promising cause of a velocity reduction is $\left(\frac{\delta V}{V_o}\right)_{GBS}$, which is controlled by temperature and water effects through shifts in the characteristic frequency (4).

3.3. Experimental Constraints on Grainboundary Sliding Parameters

[25] Most previous experimental studies were made at water-poor conditions (except for an exploratory study by *Aizawa et al.*, [2008]), and, consequently, we lack strong experimental constraints on the water sensitivity parameter, *r*. However, as discussed by *Karato* [2006] and *McCarthy et al.* [2011], there is a close connection between the characteristic time of anelastic relaxation and the Maxwell time (characteristic time for plastic deformation). Maxwell time depends on water content as $\tau_M = \frac{1}{\omega_M} \propto C_w^{-r}$ with $r \sim 1-2$ (e.g., *Karato* [2008]), and we use r = 1 and 2 in this study.

[26] We determine the plausible range of subsolidus parameters that is compatible with the experimental observations. We employ a simplifying assumption that the activation energy for AB and GBS is the same, following *Jackson and Faul* [2010]. Uncertainties in these parameters will influence the interpretation of velocity-reduction model, as can be seen from equations (7) and (8). The LAB depth is controlled mainly by ω_{gbs} . The magnitude of velocity change caused by grain-boundary sliding is specified by Δ_p . The parameter β controls the sharpness of this peak.

[27] The reanalysis of the experimental data from *Jackson and Faul* [2010] showed that not all of these parameters are constrained well (Appendix B). We demonstrate the uncertainties in the model parameters with an analysis from pristine specimen #6585. A first-stage 3D grid search was made for the $[\Delta_p, \sigma, \omega_{gbs}]$ fixing the activation energy $E_p = 320$ kJ/mol. The results demonstrated the tradeoffs in the bivariate empirical probability distribution functions (PDFs) (Figure 6). This analysis shows that β is weakly constrained by the data.

[28] Consequently, we performed a 3D grid search for $[\Delta_p, \omega_{gbs}, E_p]$ using various β values. The result of this analysis shows that when the peak (gbs) activation energy E_p is similar to that of the high-temperature background (AB), then it is constrained tightly (Figure 7, and Table 1). The



Figure 6. Bivariate empirical probability distribution functions (PDFs) using a fixed $E_p = 320$ kJ/mol. The peak broadness parameter σ trades off with the other two parameters and is not constrained by the current data. For the peak position parameter τ_{pr} and peak intensity Δ_p , maximum probability gives a value of $10^{-2.3}$ s and $\sim 7\%$, respectively. We investigate this with a model search varying E_p , for a plausible value of σ .



Figure 7. Bivariate empirical probability distribution functions (PDFs) for fixed $\sigma = 1.5, 2.5, 3.5, 4.0$. These are 2D projections of the 3D probability density volume, integrated in the third dimension. Activation energy E_p is well constrained, as can be seen in the joint distribution of $[\Delta_p, E_p]$. The PDFs demonstrate the covariation of parameters. $[\Delta_p, \log \tau_{pr}]$ show a negative covariation, with lower peak position, $\log \tau_{pr}$, preferred at high peak intensities, Δ_p . The single variable probability density functions for $[\Delta_p, \log \tau_{pr}, E_p]$ are shown in the bottom row. The area under the probability density function integrates to 1. The blue lines show the mean values $\bar{\mathbf{m}}$ for the single specimen data (6585) while the black dotted line is value obtained from using the full data set. Only the activation energy shows significant difference between full data and single data inversions with a shift from 320 to 360 kJ/mol (Table 1).

peak intensity and peak position are constrained less well, but the optimal values are shown in Table 1, and sensitivity of the data to different model parameters in Figure 8.

^[29] We extend this analysis to all four different grain sizes from *Jackson and Faul* [2010] (Figure 9). We estimate grain-size sensitivity by using the following for the peak relaxation time:

$$\overline{\tau}_{gbs} = \tau_{pr} \left(\frac{d}{d_r}\right)^{-m_p} \exp\left[-\frac{E_p}{R} \left(\frac{1}{T} - \frac{1}{T_r}\right)\right] \exp\left[\frac{V_p}{R} \left(\frac{P}{T} - \frac{P_r}{T_r}\right)\right] \tag{10}$$

where m_p is the grain-size exponent, and $\tau_{gbs} = \omega_{gbs}^{-1}$. In the first search, we fix the grain-size exponent, $m_p = 1.3$, using the optimal value reported by *Jackson* and *Faul* [2010] and *Morris and Jackson* [2009].

Measure	σ	$\Delta_p(\%)$	$\log \tau_{pr}(s)$	E_p (kJ/mol)
m	1.5	7.42	-3.12	320.52
$\sigma(\mathbf{m})$		4.38	0.60	28.48
m	2.5	7.48	-3.11	320.67
$\sigma(\mathbf{m})$		4.35	0.60	28.60
m	3.5	7.56	-3.09	320.65
$\sigma(\mathbf{m})$		4.37	0.60	28.69
m	4.0	7.80	-3.06	320.40
$\sigma(\mathbf{m})$		4.44	0.61	28.91

Table 1. Statistical Results of Grid Search for Grain-boundary Sliding Parameters: $[\Delta_p, \tau_{pr}, E_p]$ for Fixed σ

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Figure 8. Shear modulus (top) and attenuation (bottom) data (circles), overlaid with model predictions (lines). Data is specimen 6585 from [*Jackson and Faul*, 2010]. Gray lines are for high-temperature data: $T > 1223^{\circ}$ K, while colored lines are for low-temperature data, from which the grain-boundary sliding parameters are constrained. Panels from left to right show the effect of varying parameters systematically, demonstrating the trade-offs. Model lines in the leftmost panel are calculated from the mean model parameters $\bar{\mathbf{m}}$. Center panel shows the effect of varying the peak broadening parameter, σ , while the rightmost panel varies σ as well as the peak intensity, Δ_p , and peak position log τ_{pr} . Shear-modulus data is not sensitive to peak position or peak broadening parameters, but is sensitive to peak intensity (compare left and center panel; with center and right panel).

The probability density functions for $[\Delta_p, \omega_{gbs}, E_p]$ differ only slightly from the results of the singlespecimen grid search (Figure 7, bottom row). For example, using the entire data set, the maximum likelihood value for the peak activation energy is $E_p = 356$ kJ/mol, similar to the single specimen study (Table 1). A full description for single and multiplegrain-size data is provided in Table 2. [30] We estimate that the peak frequency for grainboundary sliding depends on grain-size exponent $m_p = 1.4 \pm 0.47$ (Figure 10). This value is broadly consistent with the theoretical prediction that grain-boundary-peak relaxation time should have $m_p = 1$ (equation (5)). Our calculations are conducted while fixing the grain-size sensitivity of the steady-state viscous relaxation time Geochemistry Geophysics

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Figure 9. Data and model predictions for the full data set of [*Jackson and Faul*, 2010]. Grain size increases from left (2.9 μ m) to right (28.4 μ m), with shear modulus on top and attenuation on bottom. Gray lines are high-temperature data and are well described by the visco-elastic burger's model without the need for a grain-boundary sliding improvement. Colored lines emphasize the low-temperature data, which provides information on the grain-boundary sliding parameters. The pristine data from specimen 6585 is shown in panel B. The low-temperature data from the other specimen (A, C, and D) provide constraints on grain-size sensitivity of the grain-boundary relaxation time useful for extrapolation to mantle grain sizes.

Table 2. Statistical Results of Grid Search for Grain-boundary Sliding Parameters: $[\Delta p, \tau pr, Ep]$ for Fixed σ Using Full Data Set With Four Different Grain Sizes

Measure	σ	$\Delta p(\%)$	$\log \tau pr(s)$	Ep (kJ/mol)
m	1.5	6.67	-3.21	356.60
$\sigma(\mathbf{m})$		3.71	0.52	30.27
m	2.5	6.93	-3.18	357.10
$\sigma(\mathbf{m})$		3.73	0.52	30.24
m	3.5	7.09	-3.16	357.12
$\sigma(\mathbf{m})$		3.78	0.52	30.32
m	4.0	7.40	-3.12	356.77
$\sigma(\mathbf{m})$		3.87	0.53	30.60

(or Maxwell time), $m_v = 3$. This corresponds to the Coble-creep mechanism. Alternatively, if we use a value of $m_p = m_v$, we obtain $m_p = m_v =$ 1.59 (Figure 10). We conclude that the grainboundary-sliding parameters are reasonably well constrained by experimental data, but note that further experimental data would be useful.

4. Implications for the LAB Structure

4.1. LAB Depth: Temperature of Relaxation and the Dehydration Horizon

[31] In our model, the LAB corresponds to the depth at which seismic frequency becomes comparable to

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Figure 10. Probability density function for grain-size exponent, m_p , while fixing activation energy E_p , and steadystate diffusion creep grain-size exponent $m_v = 3$ (blue) and $m_p = m_v$ (black). The activation energy, $E_p = 360$ kJ/mol, is the maximum probability value calculated using the grid search on all data. Mean value of $m_p = 1.43$, 1.59, standard deviation = 0.47, 0.45 depending on the assumption for m_v .

 ω_{gbs} . With the full description of the parameters estimated from experimental data, we have a general expression for calculating the characteristic frequency as a function of temperature, water content, pressure, and mantle grain sizes (see equation (4)). The characteristic frequency curve (Figure 11), with its associated uncertainty, is calculated by using the statistical description of $[m_p, E_p, \log \tau_{pr}(s)]$, which was obtained from the data fit (Table 2) (with fixed values r=0 (dry), 1, and 2), and the values $\omega_{gbs,r} = 1/\tau_{pr}(Hz), d_r = 13.4 \mu m, T_r = 1173^{\circ} K, P_r = 0.2$ GPa, $C_{wo} = 10^{-4} \text{wt} \%$ and $V_p = 10 \times 10^{-6} \text{m}^3 \text{mol}^{-1}$ [Faul and Jackson, 2005]. This calculation is similar to that of *Karato* [2012], but instead of assuming an implicit mantle grain size, we allow for various grain sizes and its variation using the grain-size exponent m_p , estimated from our parameter search. We also demonstrate that the uncertainties in the model parameters do not invalidate the plausibility of the grain-boundary relaxation mechanism. We use the half-space cooling model to calculate T(z,t), and the water content-depth curve $C_w(z)$ from *Karato and Jung* [1998], noting that the major features of ω_{gbs} are similar for other geothermal models. Despite the uncertainties in the grain-boundary sliding parameters, the characteristic frequency still exceeds the seismic frequency at some depth defined by the geotherm and sea-floor age, both in the dry case (r=0) and in the wet case (r=1,2). Below this depth, the mantle materials become relaxed, and the seismic velocity drops, in our proposed model, leading to the observed seismic LAB.

[32] We can recognize two regimes where the LAB depth is controlled by different factors. In the old oceanic upper mantle where the geotherm is relatively cold, the characteristic frequency of GBS crosses the seismic band at ~70 km, the depth at



Figure 11. The variation of characteristic frequency with depth, for the half-space cooling model using the statistical description of the parameters: $[m_p, \log \tau_{pr}, E_p]$. The characteristic frequency is calculated at fixed mantle grain size d = 1 mm, age = 100 Ma, and r = 0,1,2 (left, middle, and right panels). The gray lines are calculated using 300 random realizations drawn from the normal distribution, $N(\bar{x}, \sigma)$, with average, \bar{x} , and standard deviation, σ , values: $m_p = N$ (1.43,0.5); $E_p = N(356 \text{ kJ/mol}, 30.2 \text{ kJ/mol})$; log $\tau_{pr} = N(-3.21 \text{ s}, 0.52 \text{ s})$. The most likely value (blue line) for the characteristic frequency is calculated using the average values: $[m_p = 1.43; E_p = 356 \text{ kJ/mol}; \log \tau_{pr}(s) = -3.21]$. Moderate to high water sensitivity, r =1, 2, will always result in relaxation beneath depths of 70 km, despite the uncertainties in the grain-boundary parameters. In a case where the upper mantle is dry (r=0), relaxation occurs over a wide depth range, in the average case, but the curves for model parameters that plot beneath this average reflect scenarios where relaxation may not occur.



which the water content changes abruptly. In these cases, the transition from unrelaxed to relaxed state occurs primarily due to the increase in water content at 70 km. Consequently, the LAB depth in this regime is insensitive to the geotherm and hence insensitive to the age of the ocean floor. In the young ocean where temperature is higher, the transition to unrelaxed state occurs at a depth shallower than 70 km when geotherm is close to the relaxation temperature (\sim 1300°K), where the rocks are essentially dry. In young ocean, the transition is solely due to the thermal effect and consequently, the LAB depth will depend on the age of the ocean floor and is less sharp compared to that in the old oceanic mantle.

[33] The depth of the LAB in the young oceanic mantle depends on the geothermal model. For example, the transition is predicted to occur at ~65 Ma for HSC, ~55 Ma for *McKenzie et al.* [2005] and ~75 Ma for *Hirschmann* [2010]. These predictions are only the maximum-likelihood values, as the uncertainties in the grain-boundary parameters permit a wider range of values (gray lines in Figure 17).

[34] The depth of the LAB in the young oceanic mantle depends also on the grain size. For grain sizes of 1 mm, the temperature of relaxation at dry

conditions is ~1230°K (at 3 GPa) and ~1350°K (3 GPa) for mantle grain size of 1 cm. The uncertainty in the temperature of relaxation shows a spread (1 σ) of about 100°K, for grain sizes of 1 mm or ~160°K for grain sizes of 1 cm (Figure 12). In our model, these uncertainties correspond to the uncertainties in the LAB depth of 15 km in 30 Ma oceanic upper mantle, see Figure 17.

4.2. Sharpness of the LAB: Age Dependence, Water Content, and the β Effect

[35] The sharpness of the LAB is an important quantity that characterizes various models of LAB. In this section, we quantify the sharpness of the LAB using the notion of the full width half maximum (FWHM) (Figure 13). The sharpness and velocity reduction in this study are calculated using $\beta = 1$.

[36] For a relatively young ocean, the large velocity drop at the LAB occurs when temperature exceeds the critical temperature for GBS (~1300°K). Because temperature changes with depth only gradually, the LAB in this regime is relatively broad. However, because the temperature gradient changes



Figure 12. Temperature at which relaxation occurs, T_{rel} . This is the temperature at which $\omega_{gbs} \ge \omega_{seismic}$. The seismic frequency, $\omega_{seismic} = 1Hz$, and characteristic frequency, ω_{gbs} , is calculated at mantle grain sizes d = 1 mm (blue), 1 cm (red), for pressures, P = 1,2,3 GPa (left to right). The temperature of relaxation is lower for smaller grain sizes than for higher grain sizes, but these differences are within the limits of uncertainty. The uncertainties are mapped from $[m_p, E_p, \log \tau_{pr}]$, by using 1000 random realizations of these parameters. There is very small sensitivity of relaxation temperature to pressure (results on the top).





Figure 13. The variation of attenuation, Q^{-1} (red line) and relative velocity reduction $\delta V/V_o$ (black line) with depth, for the absorption-band model (ABM), the grain-boundary sliding (GBS) model, and both contributions (ABM+GBS). The curves are calculated for a young ocean (t=45 Ma), using the half-space cooling model. The absorption-band model has one free parameter, the frequency dependence, α , which is fixed at 0.3. The optimal GBS parameters are used $m_p=1.4$, $E_p=356$ kJ/mol, $\log \tau_{pr}(s)=-3.2$, and r=0, d=1 cm. The relative velocity change for the absorption-band model is gradual, except around the 70 km dehydration horizon, where the velocity reduction <1%, due to the change in water content. The relative velocity change due to the grain-boundary sliding is much larger, ~7% and is much gradual—over 10 km (measured using the FWHM, full width half maximum, or the broadness of the dissipation peak); this number depends on the parameters of the grain-boundary sliding parameters, most especially β . The combined contributions (ABM+GBS) suggest that at this age, young oceans, the largest contribution to velocity reduction is due to the grain-boundary sliding (<7%), compared to the smaller prediction of <1% from the absorption-band model.

with age (steeper at younger age), the sharpness also depends on the age: sharper at younger oceanic regions (see Figure 14). In this regime, the sharpness is also sensitive to the details of the GBS peak. Consequently, we explore the effect of the peak broadening parameter, β , on the sharpness of the predicted LAB. For a young ocean of t=40 Ma, where the relaxation is gradual and is thermally induced, we calculate the attenuation and velocity reduction while varying the β value. As the attenuation peak broadens, the relative velocity reduction becomes less sharp, although the amplitude of velocity reduction remains essentially the same (Figure 15).

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[37] In the older ocean, the sharpness is controlled by the dehydration horizon (Figure 16). In this regime, the LAB lies at a constant depth and is always sharp (FWHM < 1 km). These results are summarized in a LAB depth-age-sharpness diagram for three thermal models (Figures 17 and 18). They show that for all the geothermal models, the sharpness of the LAB in young oceanic mantle is age dependent, with a diffuse LAB (~20 km) between 20–50 Ma, while in the older oceanic mantle, the LAB is at a constant depth and very sharp (velocity change occurs over a depth of < 1 km).

5. Discussion

5.1.1. A Comparison to Seismological Observations (Oceanic Upper Mantle)

[38] The large database of LAB measurements of the Pacific Ocean upper mantle provides us with constraints to test the validity of our new subsolidus model in comparison to the partial-melt model. These seismological studies of the LAB conducted using body-wave data cover a broad area with various sea-floor ages, with improved lateral, and vertical resolution. Most prominent is the P and S receiver functions [*Kawakatsu et al.*, 2009; e.g., *Kumar and Kawakatsu*, 2011; *Li et al.*, 2009; e.g., *Kumar and Kawakatsu*, 2011; *Li et al.*, 2009; *Li et al.*, 2000; *Rychert and Shearer*, 2009; *Rychert et al.*, 2007], and more recently, the SS precursor techniques [e.g., *Rychert and Shearer*, 2011; *Schmerr*, 2012], which offer almost complete coverage of the entire Pacific Ocean sea floor.

[39] For both the receiver function and SS precursor studies, the results agree with respect to the amplitude of the velocity change, which is $\sim 5-14\%$, consistent with earlier studies by [*Gaherty et al.*, 1996; *Tan and Helmberger*, 2007]. These results are in





Figure 14. The effect of the temperature gradient on the sharpness of the velocity reduction. The sharpness of the velocity reduction changes from 4 km (for 10 Ma) to 12 km (50 Ma). This effect is caused by the higher temperature gradient at younger oceans compared to older oceans.



Figure 15. The effect of β on the magnitude and sharpness of $\delta V/V_o$. For a single Debye peak, the velocity reduction is ~7%, while the sharpness is 8 km. As β varies from 1 to 0.75, 0.25, the predicted LAB caused by grain-boundary sliding becomes less sharp, 10 km, and 24 km, respectively, while the magnitude of $\delta V/V_o$ remains unchanged.

agreement with the maximum value of velocity reduction expected from the grain-boundary-sliding model, following relaxation. Although the partialmelt model can explain these results, the consequences of partial melt models prove difficult to reconcile with other seismological constraints such as anisotropy as well as the geodynamic difficulty, as already outlined.

[40] Another striking feature of these results is the reported age dependence of the LAB. Age

dependence is observed in the receiver-function studies of *Kawakatsu et al.* [2009] and *Kumar* and Kawakatsu [2011], as well as the SS precursor studies of *Rychert and Shearer* [2011]. Both studies observe a sharp boundary that shows an age dependence, with the depth following a thermal contour of ~1300°K. Although the prediction of age dependence is robust, the actual thermal contour is difficult to establish, due to the uncertainties in the depth estimates. The age-dependence is a robust





Figure 16. Predictions for Q^{-1} and $\delta V/V_o$ for old oceans, (t=100 Ma), using the half-space cooling model. The absorption-band model (ABM) has only a gradual reduction in velocity, with no contribution to a sharp velocity reduction. The grain-boundary sliding model (GBS) has a sharp-velocity reduction (FWHM <1 km) caused by the rapid change in water content at ~70 km. For all old oceans, where dehydration horizon is greater than relaxation temperature, the predicted LAB would be at this depth—70 km.

feature that strongly confirms our prediction for young oceans, and a thermal contour of $\sim 1300^{\circ}$ K is in broad agreement with the GBS model, within the bounds of uncertainty.

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[41] One of the important features of seismological observations is that the oceanic LAB is detected in some regions but not everywhere, e.g., the SS precursor results of Rychert and Shearer [2011] and Schmerr [2012]. Schmerr [2012] proposed that the oceanic LAB is visible in regions where melt accumulation is expected. In this model, a sharp and large velocity drop will occur only when a substantial amount of melt is accumulated at the LAB boundary (near the solidus). There is some correlation between visible LAB and hotspots, but the LAB is also visible in some old oceanic upper mantle near trenches. Enhanced melting in old oceanic mantle is not readily explained. One needs to invoke small-scale convection in the asthenosphere to explain enhanced melting on these regions. It is not clear why small-scale convection occurs in some old oceanic mantle not everywhere. Also, if the LAB were to correspond to the depth at which melt is accumulated, then the LAB depth near hotspots should be systematically shallower than those away from hotspots, since the temperature is expected to be hotter. Such a correlation is not clearly seen in these studies.

[42] A key to the sub-solidus model is the presence of a sharp water-content contrast at \sim 70 km depth.

This assumes an efficient melt extraction and resultant water removal from the mantle at ocean ridges. It is possible that the melt extraction and water removal are incomplete in some regions, and in these regions some water should be present in the oceanic lithosphere. In such a case, the transition from unrelaxed (in the shallow region) to relaxed (in the deep region) could occur at a depth shallower than ~70 km and the transition will not be sharp. An intermediate jump in water content with depth could shift the physical-dispersion transition into the seismic band, so that short-period Ps receiver functions would not detect an otherwise sharp LAB, while longer-period Sp receiver functions would detect it. Our model predicts that in regions where the LAB is invisible, the lithosphere would contain a substantial amount of water; electrical conductivity of the lithosphere in these regions would be higher than other regions.

[43] An alternative explanation could be made from the seismological perspective. The sharpness of a detected seismic interface is related to the wavenumber and frequency content of the wave that probes it. Ps receiver functions can detect interfacial transitions of 1 km thickness or less [*Park and Levin*, 2000; *Leahy and Park*, 2005; *Leahy et al.* 2012], and so should be capable of resolving thin LVLs. Teleseismic S waves lack the high-frequency content of teleseismic P waves, so Sp receiver functions are less capable of defining the sharpness of a boundary. SS waves average





Figure 17. Predicted depth of the LAB overlaid on the plausible thermal models. We describe these predictions using the statistical distribution of the grain-boundary sliding parameters (gray lines). The most likely value (dark line, with circles) for the depth of the LAB is calculated using the mean values given in Table 2, but for a $\beta = 1$. The size of the circles is scaled to the sharpness of the velocity drop (using the FWHM calculation). The blue lines are the temperature contours, displayed every 200°K. For each thermal model, the predicted LAB is age dependent at young oceans and constant for older oceans. The transition varies for each model, varying from 55 Ma, to 75 Ma, depending on the thermal model. For younger oceans, the *most likely* LAB is bounded by the thermal contour 1200°K and 1400°K, with a sharpness of ~10 km. The LAB is constant at ~70 km and sharp, caused by grain-boundary sliding from the intersection of the dehydration horizon.

over a large Fresnel zone at their bounce points, degrading the resolution of boundary sharpness. Although we attempt here to interpret the currently available LAB observations, further observations with diverse body waves will improve our perception of the variability of LAB throughout the world.

[44] In summary, the seismological results agree with our predictions in two important regards—the magnitude of the velocity reduction and the age dependence of the measured LAB including its depth and sharpness. However, due to the limited sensitivity of the SS precursor technique [e.g., *Schmerr*, 2012], the constraints on the sharpness of the velocity reduction is not as robust, making it difficult to make a comparison with our predictions on how LAB sharpness and amplitude should vary with age.

5.1.2. Uncertainties in Material Properties and Perspectives

[45] Our study demonstrates the potential importance of grain-boundary sliding to explain the





Figure 18. Predicted LAB depth and sharpness calculated for various peak broadness parameter, β , from sharp (black, $\beta = 1$) to relatively broad, (red, $\beta = 0.75$) and very broad (blue, $\beta = 0.25$). The size of the circle is scaled similar to Figure 23. The inset shows the expected receiver-function response to these cases (see also Figure 5).

oceanic LAB. In an oceanic-plate context, neither a partial-melt model nor a classic AB sub-solidus model provides an acceptable explanation of seismological observations. However, at the same time, our study illustrates large uncertainties in the relevant parameters. The sensitivity of anelastic relaxation to water content, both in the AB and for the GBS, is poorly constrained. Similarly, both the peak frequency (or temperature) and the breadth of frequency distribution for the GBS mechanism are poorly constrained. More extensive laboratory data on these processes are essential for our better understanding of key geophysical observations of the upper mantle (LAB and MLD).

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Appendix A

Peak Broadening, $\sigma \Rightarrow \beta$ Transformation

[47] The grain-boundary-sliding peak, observed by *Jackson and Faul* [2010], was modeled using a generalized Burgers model of viscoelasticity. In the frequency domain, this function has a real and imaginary part, given, respectively, as:

$$J_1(\omega) = J_U \left\{ 1 + \Delta \int_{T_L}^{T_H} \frac{D(\tau)d\tau}{1 + \omega^2 \tau^2} \right\}$$
(A1)

$$J_2(\omega) = J_U \left\{ \omega \Delta \int_{T_L}^{T_H} \frac{\tau D(\tau) d\tau}{1 + \omega^2 \tau^2} + \frac{1}{\omega \tau_m} \right\}$$
(A2)

where Δ is the relaxation strength of the mechanism being modeled, Δ_p for the grain-boundary peak, and Δ_B is the relaxation strength for the high-temperature background. τ_M is the Maxwell time for viscous relaxation, and $D(\tau)$ is the chosen distribution of anelastic relaxation times that most suitably fits the experimental data. The strain-energy dissipation Q^{-1} and shear modulus G are then given as functions of angular frequency:

$$G(\omega) = \left[J_1^2(\omega) + J_2^2(\omega)\right]^{-1/2}$$
(A3)

$$Q^{-1} = \frac{J_2(\omega)}{J_1(\omega)} \tag{A4}$$

[48] A distribution of anelastic relaxation time D_B (τ) that models the high-temperature background dissipation is given by

$$D_B(\tau) = \frac{\alpha \tau^{\alpha - 1}}{\tau_H^{\alpha} - \tau_L^{\alpha}}$$
(A5)

[49] This is a standard formulation and captures the general AB behavior at higher temperatures, *Anderson and Given* [1982]. At lower temperatures, *Jackson and Faul* [2010] identified the presence of a relaxation peak with the following distribution of relaxation time, which captures the peak mechanism:

$$D_p(\tau) = \sigma^{-1} (2\pi)^{-1/2} \exp\left\{\frac{-\left[\ln(\tau\omega_{gbs})/\sigma\right]^2}{2}\right\}$$
(A6)

where σ is the peak broadening parameter, and ω_{gbs} is the peak frequency. This distribution of relaxation times is associated with the peak intensity Δ_p , separate from the background relaxation strength Δ_B associated with the background relaxation time $D_B(\tau)$.

[50] With this parameterization, we can calculate the shear-modulus and attenuation pair [$G(\omega,\sigma)$,



 $Q^{-1}(\omega,\sigma)$]. This functional form requires a numerical integration, and the effect of the σ parameter on velocity reduction is not immediately apparent. In order to simplify, this we use a different parameterization for the grain-boundary peak:

$$Q^{-1}(\omega,\beta) = \Delta_p(\beta) \frac{\left(\frac{\omega}{\omega_{gbs}}\right)^{\beta}}{1 + \left(\frac{\omega}{\omega_{gbs}}\right)^{2\beta}}$$
(A7)

[51] The two functions $[Q^{-1}(\beta), Q^{-1}(\sigma)]$ are equivalent and fully capture the behavior given by σ . We observe that varying σ changes both the width and the peak height. We model this for the new parameterization in the following way:

1. For a specified number of σ values, we calculate $Q^{-1}(\omega,\sigma)$, and its FWHM and search for a corresponding β parameter using $Q^{-1}(\omega,\beta)$ that gives the same FWHM. This allows us to write $\beta(\alpha)$ (Table 3):

$$\beta = 0.0081\sigma^3 - 0.014\sigma^2 - 0.28\sigma + 1.07$$
 (A8)

2. A comparison of $Q^{-1}(\omega,\beta)$ with $Q^{-1}(\omega,\sigma)$ shows that changing σ also changes the peak height. To account for this, we make peak intensity a function of $\beta : \Delta_p(\beta)$. To constrain the change in peak height with peak broadening, we calculate the following ratio: $\frac{\Delta_p(\sigma)}{\Delta_p(0)}$, and use the relationship in (A8), $\beta(\sigma)$, to write this as $\Delta_p(\beta)$. $\Delta_p(0)$ is the peak height for a single Debye peak, i.e., with no broadening. This gives the following relationship:

$$\frac{\Delta_p(\beta)}{\Delta_p(1)} = 1.13\beta^3 - 2.017\beta^2 + 1.92\beta - 0.088$$
(A9)

Table 3. Best Fit Parameters for the Broad Dissipation Peak: $\Delta_p(\beta), \beta^{a}$

σ	β	FWHM ^b	$rac{\Delta_p(\sigma)}{\Delta_p(0)}$
0.25	1	1	0.94
1	0.78	1.6	0.72
3	0.33	3.5	0.36
4	0.25	4.4	0.29

^aThese results are obtained from the analysis discussed in Appendix A ^bFWHM (Full Width Half Maximum, in $\log_{10}\left(\frac{\omega_{s}}{\omega_{s}}\right)$). It prescribes how large ω_{gbs} has to be relative to ω_s for full velocity reduction (see Figure 13). ω_{gbs} varies with depth, prescribed by the geotherm, so FWHM allows us to map the effect of β into the broadness of velocity reduction with depth.

Appendix B Statistical Constraints on Grainboundary Parameters

[52] We perform a grid search model inversion for the grain-boundary parameters $[\Delta_p, \sigma, \omega_{gbs,r}, E_p]$ using the attenuation and shear-modulus data [$G(\omega), Q^{-1}(\omega)$] provided by Jackson and Faul [2010]. We fix the AB parameters with the optimal values given in Table 3 of Jackson and Faul [2010], where Δ_p is the peak intensity, σ is the peak broadening parameter, τ_{pr} is the relaxation period, at the reference conditions (i.e., temperature, $T_r = 900^{\circ}$ C, and grain size, $d = 2.9 \ \mu$ m (note $\omega_{gbs,r} = \frac{1}{\tau_{pr}}$), and E_p is the activation energy term that determines the temperature dependence of the grain-boundary relaxation time scale.

[53] To estimate the empirical probability of a given parameter (univariate PDF), or pair of parameters (bivariate PDF), we use the following approach. In the case of a 3D grid search in $[\Delta_p, \sigma, \tau_{pr}]$ for a fixed E_p , we define bounds on each parameter in the grid search, e.g.,

$$\begin{array}{l} 0 < \Delta_p < 0.4 \\ 0 < \sigma < 4 \\ -4 < \log \ \tau_{pr} < 0 \end{array} \tag{B1}$$

[54] Each parameter range is divided into *n* points: $\Delta_p^{1,...n}, \sigma^{1,...n}, \log \tau_{pr}^{1,...n}$, effectively dividing the 3D parameter space into a grid of n^3 points. With this discretization, we calculate the following measure of data fit, for all the n^3 in parameter space:

$$\chi^{2}_{G,Q} \left(\mathbf{m}_{i} = \left[\Delta_{p}, \log \tau_{pr}, \sigma \right]_{i} \right) \\ = \frac{\left(\left[G(\omega), Q(\omega) \right]_{m} - \left[G(\omega), Q(\omega) \right]_{d} \right)^{2}}{\left(G_{\sigma^{2}}, Q_{\sigma^{2}} \right)}$$
(B2)

where $[G(\omega), Q(\omega)]_d$ is the shear-modulus and attenuation data, $[G(\omega), Q(\omega)]_m$ is the model prediction, calculated using (A3) and (A4) at each model point $m_i = [\Delta_p, \log \tau_{pr}, \sigma]_i$, for all *i* in the n^3 space, and $(G_{\sigma^2}, Q_{\sigma^2})$ is the data variance, which estimates the uncertainty in the measurement of shear modulus and attenuation, as reported by [*Jackson and Faul*, 2010]. If $\chi^2_{G,Q}(\mathbf{m}_i) < \sim 1$, then the model parameters fit the data within $\sim 1\sigma$ of the reported data variance.

[55] A simple way to estimate the probability of a particular set of model parameters, $P(\mathbf{m}_i)$, is to define:

$$P(\mathbf{m}_i) = cL(\mathbf{m}_i) \tag{B3}$$

where $L(\mathbf{m}_i)$ is the likelihood function based on the data misfit function $\chi^2(\mathbf{m}_i)$ defined as:



$$L(\mathbf{m}_i) = \exp\left(-\frac{1}{2}\chi^2(\mathbf{m}_i)\right)$$
(B4)

and a normalization constant c that is defined to make

$$\int_{M} P(\mathbf{m}_i) d\mathbf{m} = 1$$
 (B5)

[56] It can be seen that the likelihood function weighs models \mathbf{m}_i that fit the data well with a high value, and those that don't (bigger χ^2) have a likelihood value that is effectively zero.

[57] With the probability density function, we calculate the model mean $\bar{\mathbf{m}}$ and variance $\sigma^2(\mathbf{m})$:

$$\bar{\mathbf{m}} = \int_{M} \mathbf{m} P(\mathbf{m}_i) d\mathbf{m}$$
(B6)

$$\sigma^{2}(\mathbf{m}) = \int_{M} (\mathbf{m} - \bar{\mathbf{m}})^{2} P(\mathbf{m}_{i}) d\mathbf{m}$$
(B7)

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