



# Inferring crustal viscosity from seismic velocity: Application to the lower crust of Southern California

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## ABSTRACT

We investigate the role of composition on the viscosity of the lower crust through a joint inversion of seismic P-wave ( $V_p$ ) and S-wave ( $V_s$ ) velocities. We determine the efficacy of using seismic velocity to constrain viscosity, extending previous research demonstrating robust relationships between seismic velocity and crustal composition, as well as crustal composition and viscosity. First, we calculate equilibrium mineral assemblages and seismic velocities for a global compilation of crustal rocks at relevant pressures and temperatures. Second, we use a rheological mixing model that incorporates single-phase flow laws for major crust-forming minerals to calculate aggregate viscosity from predicted mineral assemblages. We find a robust correlation between crustal viscosity and  $V_p$  together with  $V_s$  in the  $\alpha$ -quartz regime. Using seismic data, geodetic surface strain rates, and heat flow measurements from Southern California, our method predicts that lower crustal viscosity varies regionally by four orders of magnitude, and lower crustal stress varies by three orders of magnitude at 25 km depth. At least half of the total variability in stress can be attributed to composition, implying that regional lithology has a significant effect on lower crustal geodynamics. Finally, we use our method to predict the depth of the brittle–ductile transition and compare this to regional variations of the seismic–aseismic transition. The variations in the seismic–aseismic transition are not explained by the variations in our model rheology inferred from the geophysical observations. Thus, we conclude that fabric development, in conjunction with compositional variations (i.e., quartz and mica content), is required to explain the regional changes in the seismic–aseismic transition.

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## 1. Introduction

The viscosity structure of the lower crust and upper mantle controls deformation processes such as post-seismic creep (e.g., Freed et al., 2007), lower crustal flow (e.g., Clark et al., 2005), post-glacial isostatic rebound (e.g., Jull and McKenzie, 1996), and orogenesis. Estimates of crustal and upper mantle viscosity are typically constrained by analyses of post-seismic relaxation, paleo-lake shoreline deflection, global flow and stress models, and laboratory derived flow laws for crust-forming rocks and minerals (e.g., Thatcher and Pollitz, 2008; England et al., 2013; Shi et al., 2015; Humphreys and Coblenz, 2007; Bürgmann and Dresen, 2008). In most previous studies based on laboratory data the rheology of the lower continental crust was approximated using a flow law for a dominant mineral phase (e.g., quartz or feldspar). However, the lower crust is composed of multiple mineral phases with variable strength and abundances that are controlled by the bulk composition, pressure, and temperature conditions. Thus, to accurately

calculate crustal viscosity, composition, pressure, and temperature must be constrained.

Seismic P-wave ( $V_p$ ) and S-wave ( $V_s$ ) velocities are frequently used as a proxy for crustal composition. Christensen and Mooney (1995) measured  $V_p$  for a variety of igneous and metamorphic rocks and compared the measurements to average continental crustal seismic profiles. With knowledge of field geology, they created a petrological model of the continental crust. Rudnick and Fountain (1995) estimated average mid and lower crustal compositions by comparing the predicted  $V_p$  of granulites to  $V_p$  profiles constructed for typical tectonic environments. Christensen (1996) added  $V_s$  to the analysis of crustal composition and showed that crustal  $V_p/V_s$  ratios are nominally independent of temperature, but sensitive to quartz abundance, with a strong linear correlation between  $V_p/V_s$  and silica content (between 55 and 75 wt%  $\text{SiO}_2$ ). More recently, Shillington et al. (2013) and Jagoutz and Behn (2013) combined thermodynamic modeling,  $V_p/V_s$ , and  $V_p$  to constrain the composition of arc lower crust, showing that the additional information provided by  $V_p/V_s$  gives a much tighter constraint on composition than is obtained by  $V_s$  alone. Other studies have linked seismic data to the mechanical properties of the crust. For example, Lowry and Pérez-Gussinyé (2011) found

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a relationship between  $V_p/V_s$  and effective elastic thickness in the western United States, pointing toward a relationship between seismic velocity and viscosity.

Here, we investigate the relationship between aggregate viscosity and seismic velocities by calculating a fit between viscosity (estimated using multi-phase rheological mixing theory) and seismic P-wave and S-wave velocity for a wide range of crustal compositions at typical crustal pressures and temperatures. Our analyses show that while neither  $V_p$  nor  $V_s$  alone robustly predict aggregate viscosity, together  $V_p$  and  $V_s$  predict aggregate viscosity within a factor of 2 at most crustal conditions. We apply our method to calculate the viscosity of the lower crust in Southern California, taking advantage of data from a densely instrumented region. We discuss the benefits and limitations of our methodology in the context of constraining compositional controls on upper and lower crustal viscosity – and compare the predicted depth to the brittle–ductile transition with observed seismic–aseismic transitions.

## 2. Methods

To estimate aggregate crustal viscosity from  $V_p$  and  $V_s$  data, we use the following 4-step approach: 1) create a rock database that spans crustal composition space; 2) calculate equilibrium mineral assemblages and seismic velocities for each composition in the rock database over a range of relevant pressure ( $P$ ) and temperature ( $T$ ); 3) calculate the aggregate viscosity for each composition at all  $P$ – $T$  conditions; and 4) compile our estimates of  $V_p$ ,  $V_s$ , and aggregate viscosity over crustal  $P$ – $T$  space to derive a generic relationship between viscosity and  $P$ ,  $T$ ,  $V_p$ ,  $V_s$ , and stress or strain rate. Finally, because viscosity is strongly sensitive to temperature, one must incorporate additional constraints when applying our methodology to specific geologic settings. Below, we describe each of these steps as well as our methodology for constraining crustal temperatures in Southern California.

### 2.1. Compositional database for crustal rocks

To create a compositional database for crustal rocks, we take all available whole rock compositions for the continental US and Alaska from Earthchem ([www.earthchem.org](http://www.earthchem.org)). We filter out carbonates (>2 wt%  $\text{CO}_2$ ) and samples with oxide analyses that sum to less than 95 wt% or more than 102 wt%. To this dataset, we add arc rocks from the compilation of Kelemen and Behn (2016), which includes Archean and post-Archean massif data from Hacker et al. (2015), crustal xenolith and amphibolite data from Huang et al. (2013), and lavas and plutonic rocks from the Aleutians, Izu–Bonin–Marianas, Kohistan (Jagoutz and Schmidt, 2012) and Talkeetna (Kelemen et al., 2014) arcs. The final crustal database includes 96,388 samples. To facilitate the thermodynamic calculations presented below, we sub-sampled this database by first sorting the compositions by wt%  $\text{SiO}_2$  and then taking every 28th sample, creating a representative database of 3442 samples. Sub-sampling does not alter the frequency distribution for any of the major oxides (Supplemental Fig. 1); a principal component analysis of the total and sub-sampled data sets produces similar sub-spaces (Supplemental Tables 1 and 2). Thus, the relationships derived from the sub-sampled data accurately reflect those present in the total composition space.

### 2.2. Calculation of equilibrium mineral assemblages and seismic velocity

To calculate the equilibrium mineral assemblage for each composition, we use the Gibbs free energy minimization routine *Perple\_X* (Connolly, 2009). We calculate mineral assemblages over

crustal temperatures (300–1000 °C) and pressures (0.1–1.2 GPa). We assume a minimum equilibrium temperature of 500 °C, a reasonable lower bound for net transfer reactions under hydrous crustal conditions (Austrheim, 1998). The lower continental crust typically contains 0–1 wt%  $\text{H}_2\text{O}$  (Huang et al., 2013). In this study, we only consider anhydrous mineral assemblages and incorporate the influence of  $\text{H}_2\text{O}$  through its effect on the viscosity of nominally anhydrous mineral phases (see below). The role of hydrous phases (e.g., amphibole and mica) is ignored in our calculations due to the complexities of quantitatively incorporating their effects on rheology (Shinevar et al., 2015); however, we discuss their potential role on crustal viscosity in the Section 4.

For all *Perple\_X* calculations, we assume that 25 mol% of the total iron oxide is ferric (Cottrell and Kelley, 2011; Kelley and Cottrell, 2012); variations in this value have little influence on aggregate viscosity and seismic wave speeds. Solution models for crustal minerals are taken from Hacker (2008). Seismic velocities are calculated using a compilation of mineral properties (Abers and Hacker, 2016) implemented into *Perple\_X* with the  $\alpha$ – $\beta$  quartz implementation used by Jagoutz and Behn (2013).

### 2.3. Calculation of aggregate viscosity

To calculate crustal viscosity at a given condition, we employ mixing theory to determine the aggregate viscosity for each equilibrium mineral assemblage. We only consider deformation via wet dislocation creep. This assumes that diffusion creep is not an important mechanism controlling the rheology of the lower crust (see Section 4 for further discussion). Theoretical and experimental investigations show a power law relation between stress ( $\sigma$ ) and strain rate ( $\dot{\epsilon}$ ) with the form:

$$\dot{\epsilon} = A f_{\text{H}_2\text{O}}^r \sigma^n \exp\left(\frac{-(E + PV)}{RT}\right), \quad (1)$$

where  $A$  is the pre-exponential factor,  $f_{\text{H}_2\text{O}}$  is water fugacity,  $r$  is the fugacity exponent,  $n$  is the stress exponent,  $E$  is the activation energy,  $V$  is the activation volume, and  $R$  is the gas constant. To estimate water fugacity, we use an exponential fit to water fugacity values along crustal geotherms (Shinevar et al., 2015):

$$f_{\text{H}_2\text{O}} = a_{\text{H}_2\text{O}} B_1 \exp\left(-\frac{B_2 + PB_3}{RT}\right) \quad (2)$$

where  $a_{\text{H}_2\text{O}}$  is the water activity and  $B_1$ ,  $B_2$ , and  $B_3$  are empirically fit constants. We explore fugacity values using water activities between 0.01 and 1. Aggregate viscosities are calculated for strain rates ranging from  $10^{-12}$  to  $10^{-16}$   $\text{s}^{-1}$  and shear stresses ranging from 1 to 100 MPa.

The effective viscosity of major crust-forming minerals varies by almost 3.5 orders of magnitude at 650 °C – between the strongest minerals (olivine and pyroxene) and the weakest anhydrous mineral (quartz) (Fig. 1). Here we consider five major crust-forming minerals: quartz, feldspar (plagioclase + alkali feldspar), pyroxene (orthopyroxene + clinopyroxene), garnet, and olivine. Wet dislocation creep flow laws for quartz, feldspar, pyroxene, and olivine are taken from Hirth et al. (2001), Rybacki et al. (2006), Dimanov and Dresen (2005), and Hirth and Kohlstedt (2003), respectively. The pyroxene flow law was modified by introducing a linear water fugacity term, in which the published pre-exponential coefficient is recalculated accordingly ( $A = A_{\text{experimental}} / f_{\text{H}_2\text{O}}$  for the experimental fugacity). The pre-exponential coefficients are corrected to express the relationship between the second invariants of stress and strain rate rather than uniaxial or simple shear. All garnet flow laws (e.g. Karato et al., 1995; Xu et al., 2013) found in the literature predict effective viscosities lower than plagioclase at lower crustal

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