Bulk Structure of the Crust and Upper Mantle beneath Alaska from an Approximate Rayleigh-Wave Dispersion Formula

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Abstract

We introduce a method for estimating crustal thickness and bulk crustal and uppermantle shear-wave velocities directly from high-quality measurements of fundamental-mode Rayleigh-wave dispersion in the period range from 10 to 40 s. The method is based on an approximate Rayleigh-wave dispersion formula and provides fast results with minimal model parameterization. We apply the method to Rayleigh-wave phase maps in Alaska to reveal first-order structure in a region that had not been systematically and densely instrumented prior to the Transportable Array (TA). To demonstrate the consistency of the results, we also apply the same method to existing Rayleighwave phase maps derived from TA data in the conterminous United States, where crustal and upper mantle structures are better known. We contrast features observed in maps of crustal thickness and bulk shear-wave velocity between the Cascadia and Alaska-Aleutian subduction zones to highlight differences in the two regions. Our results show that, contrary to conventional wisdom, first-order information on the location of major depth discontinuities (e.g., the Moho) can be extracted in a fast, straightforward manner from measurements of Rayleigh-wave dispersion alone.

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Supplemental Material

Introduction

Crustal thickness is a fundamental geophysical parameter needed to understand the dynamic forces responsible for mountain building and to constrain properties of the deep crust. Yet prior to the deployment of the Earthscope Transportable Array (TA) in Alaska (Incorporated Research Institutions for Seismology [IRIS] Transportable Array, 2003) knowledge of crustal thickness (i.e., depth of the Moho) across much of the state was limited. Figure 1 shows the expansion of the TA in Alaska between 2014 and 2017, with stations from 2016 and 2017 filling in regions of western and northern Alaska that had previously only been sparsely instrumented. Subsequently, research on crustal and upper mantle structure in Alaska has grown considerably in recent years with the availability of TA data (Jiang et al., 2018; Martin-Short et al., 2018; Miller and Moresi, 2018; Miller et al., 2018; Ward and Lin, 2018; Feng and Ritzwoller, 2019; Zhang et al., 2019; Berg et al., 2020). In addition to the structural implications of the Moho, its depth is also noteworthy because it formally marks the lower boundary of the transcrustal magma systems underlying volcanoes in Alaska. For example, deep long-period (DLP) earthquakes beneath Alaskan volcanoes are often observed to cluster around the depths of the Moho (Power et al., 2004, 2013). Tamura *et al.* (2016) have shown evidence for crustal thickness determining magma type in the western Aleutians as well as the Izu-Bonin arc.

Receiver functions are the most widely used approach for determining crustal thickness from passive seismic data (Schmandt *et al.*, 2015; Miller and Moresi, 2018; Zhang *et al.*, 2019). Based on multicomponent analysis of teleseismic bodywave scattering at the Moho, receiver functions provide estimates of crustal thickness and the ratio of *P*-wave velocity to *S*-wave velocity in the crust, which is itself related to the bulk Poisson's ratio of the crust. In this article, we present an alternative approach for obtaining the depth to the Moho, or crustal thickness, from the dispersion of Rayleigh waves. Traditionally, surface waves have been utilized to find

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Figure 1. Regional map of the Transportable Array in Alaska with stations color-coded by year of installation between 2014 and 2017. The color version of this figure is available only in the electronic edition.

the thickness of the crust (e.g., Evison et al., 1959); however, as comprehensively discussed by Lebedev et al. (2013), there are pitfalls and tradeoffs involved in mapping the Moho with surface waves. We develop a method suitable for obtaining a first-order approximation of the Moho interface and the bulk shear-wave velocity of the crust and upper mantle from fundamental-mode Rayleigh waves and discuss both the advantages and limitations of the method. Detailed Moho mapping, beyond a first-order approximation, requires more advanced surface-wave methods as described by Lebedev et al. (2013). Feng and Ritzwoller (2019), for example, have recently implemented a Bayesian Monte Carlo inversion of surface-wave dispersion in Alaska for depth models parameterized by 15 unknowns, including the Moho, at each lateral grid point. In contrast, our simplified method only solves for three parameters: the depth of the Moho and the bulk shear-wave velocities of the crust and upper mantle. In this way, our method bridges a gap between receiver functions, which estimate two parameters at each receiver (Moho and bulk crustal Poisson's ratio), and more advanced surface-wave methods that can image heterogeneities within the crust and upper mantle. The method is similar in principle to the grid search technique used by Pasyanos and Walter (2002) to map out crustal and uppermantle structure in North Africa, Europe, and the Middle East. We additionally use an approximation for Rayleigh-wave

dispersion that allows the grid search to run over only a single parameter, the thickness of the crust. In fact, the approximate Rayleigh-wave dispersion formula we use was discovered by Jeffreys (1935). The new aspect of our work lies in the use of this approximate formula, which Jeffreys (1935) referred to as the "first approximation," within an inversion scheme. The lack of full dispersion modeling and the reduced dimensions of the grid search make our method an extremely fast way to gain estimates of crustal and upper-mantle structure, including the Moho, from Rayleigh waves alone.

Data and Methods

We analyze fundamentalmode Rayleigh-wave phase velocity maps between 10 and 40 s period to determine the depth of the Moho as well as bulk crustal and upper-mantle

structure. The method we use is based on an approximate formula for Rayleigh-wave dispersion (Jeffreys, 1935; Haney and Tsai, 2015), which we describe later. In addition to Alaska, we also analyze phase maps determined with TA data from the conterminous United States to show the method we develop captures the known crustal and upper-mantle structure there. We select the period range from 10 to 40 s based on the findings of Lebedev et al. (2013) that Rayleigh waves have maximum sensitivity to the Moho within this band. This can be understood from the fact that the sensitivity depth of Rayleigh waves is roughly one-half of a wavelength (Haney and Tsai, 2015). For an average phase velocity of 3.5 km/s, the period range from 10 to 40 s corresponds to the typical depth range of the Moho from 18 to 70 km. The phase maps for the conterminous United States and Alaska have been published previously by Ekström (2014, 2017) and Ward and Lin (2018), respectively, and utilized correlations of ambient noise to recover the Rayleigh-wave portion of the interstation Green's functions for input to tomography. Finite-frequency effects and off-great-circle propagation, or ray-bending, were not taken into account in these phase maps.

To extract depth information from the phase velocity maps, we exploit recent results by Haney and Tsai (2015, 2017) demonstrating that an analogy to the Dix equation used in reflection seismology (Dix, 1955) exists for surface waves.

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The defining characteristic of the Dix equation for seismic reflections is the proportionality of squared stacking velocities, used to maximize hyperbolic stacking, with the squared velocities of the layers. Such a relation follows from the approximation that, above a reflector, the subsurface can be replaced by a homogeneous medium with velocity equal to the root mean square (rms) velocity over that depth interval. Thus, each reflector has a different effective homogeneous medium overlying it. For fundamental-mode Rayleigh waves in a medium with a Poisson's ratio of 0.25, a similar approach means that at each frequency the Rayleigh wave can be taken to propagate in a different effectively homogeneous medium with shear-wave velocity given by 1/0.9194 multiplied by the phase velocity at that frequency. Other values of Poisson's ratio can be used for this approximation besides 0.25 (Haney and Tsai, 2017); however, Poisson's ratio has a minor effect on Rayleigh-wave phase velocity compared to shear-wave velocity and so we take it to be 0.25 in this study. Chevrot and van der Hilst (2000) report crustal Poisson's ratio values between 0.23 and 0.29 in Australia, showing minor variability around 0.25. The analogy to the Dix equation for fundamental-mode Rayleigh waves leads to the following proportionality between squared phase velocity and squared shear-wave velocity (Haney and Tsai, 2015, 2017):

$$c^{2}(k) = \int_{0}^{\infty} \frac{\partial f(k, z)}{\partial z} \beta^{2}(z) \mathrm{d}z, \qquad (1)$$

in which $c^2(k)$ is the squared phase velocity as a function of wavenumber, $\beta^2(z)$ is the squared shear-wave velocity as a function of depth, and the kernel function is given by

$$f(k,z) = -2.8454e^{-1.6950kz} + 6.3096e^{-1.2408kz}$$

- 4.3095e^{-0.7866kz}. (2)

Notice that the numerical factors in the exponentials of equation (2) are either twice the value of the factor that appears in *P*-wave portion of the Rayleigh-wave eigenfunction $(2 \times 0.8475 = 1.6950)$, twice the value in the *S*-wave portion of the Rayleigh-wave eigenfunction $(2 \times 0.3933 = 0.7866)$, or the sum of those values (0.8475 + 0.3933 = 1.2408). This is because Rayleigh's principle, on which equation (2) is based, is quadratic in terms of the eigenfunctions. The other factors result from integrations of the Rayleigh-wave eigenfunctions (Haney and Tsai, 2015). All the numerical factors in equation (2) depend weakly on the Poisson's ratio, and more details can be found in Haney and Tsai (2017).

We consider a simplified model of a layer with thickness h and shear velocity β_1 over a half-space with shear velocity β_2 throughout this study. For a collection of m phase velocities and this simplified three-parameter model, we set $f_0 = -f(k, 0) = 0.9194^2 = 0.8453$ and $f_m = f(k_m, h)$ to find the

following expression from equation (1) for the *m*th phase velocity measurement (Haney and Tsai, 2015):

$$c_m^2 = (f_0 + f_m)\beta_1^2 - f_m\beta_2^2.$$
(3)

As can be seen from equation (3), squared phase velocity is linearly related to the squared shear-wave velocities; however, the dependence on the third parameter, the layer thickness, is nonlinear and embedded within the parameter f_m .

In Figure 2, we explore the accuracy of the Dix approximation for Rayleigh-wave dispersion for a crustal-scale model. Dix-approximate phase and group velocities for a model of a layer (thickness h = 38 km, $\beta_1 = 3.8$ km/s) over a halfspace ($\beta_2 = 4.2 \text{ km/s}$) are compared in Figure 2a to the true phase and group velocities from full dispersion modeling. Although not discussed here, Haney and Tsai (2017) have also described how to obtain expressions similar to equations (1)-(3) for group velocity. As can be seen in Figure 2a, the Dix approximation captures the main features of the dispersion curves. This comparison is extended in Figure 2b to test the accuracy of the approximation over a range of models with β_1 varying from 3 to 4 km/s and the velocity difference between the half-space and the layer, $\beta_2 - \beta_1$, varying from 0 to 1 km/s. Although the approximation becomes worse for larger values of the velocity difference, it provides reasonable accuracy not exceeding an rms error of 0.1 km/s over this range of the parameters. In Figure 2c,d, we plot the same comparison as in Figure 2a but for crustal thicknesses of 33 and 43 km, respectively, and observe similar accuracy.

The original form of the Dix-type relation for surface waves (Haney and Tsai, 2015) took density to be constant with depth and resulted in equations (1)–(3). However, a known density variation can be included in a straightforward way as we show here. When considering the crustal scale, this is applicable because a density contrast exists across the Moho. Following Haney and Tsai (2015), the modification of equation (1) when density varies with depth results in

$$c^{2}(k) = \int_{0}^{\infty} \frac{\partial g(k,z)}{\partial z} \rho(z) \beta^{2}(z) dz / \int_{0}^{\infty} \frac{\partial q(k,z)}{\partial z} \rho(z) dz, \qquad (4)$$

in which the kernel function in the numerator is given by

$$g(k, z) = -3.5306e^{-1.6950kz} + 7.8290e^{-1.2408kz}$$

- 5.3472e^{-0.7866kz}, (5)

and the kernel function in denominator by

$$q(k,z) = -1.0137 e^{-1.6950kz} + 2.9357 e^{-1.2408kz} - 3.1628 e^{-0.7866kz}.$$
(6)

Note that f(k, z) = -g(k, z)/q(k, 0) = g(k, z)/1.2408. It can be easily verified that equations (4)–(6) reduce to

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equations (1)-(3) when density is constant with depth. For a layer-over-half-space model with nonuniform density, the generalization of equation (3) becomes

$$c_m^2 = r(f_0 + f_m)\beta_1^2 - rf_m\beta_2^2(\rho_2/\rho_1),$$
(7)

in which the variable r is given by

$$r = 1/\left(1 + q_m \left(1 - \frac{\rho_2}{\rho_1}\right)\right),\tag{8}$$

and $q_m = q(k_m, h)$. Note that, for small density contrasts $(\rho_2/\rho_1 \sim 1)$, the second term on the right side of equation (7) means that the magnitude of the contrast, in terms of the density ratio, largely trades-off with the velocity of the underlying half-space.

We note that the approximation for a layer-over-half-space given by equation (7) was originally developed by Jeffreys (1935) for a Poisson's ratio of 0.25 (shown in equation 21 of the 1935 article). Although Jeffreys (1935) has subsequently been widely cited, the approximate dispersion formula in the article appears to have not received attention outside of two articles by Takeuchi *et al.* (1959) and Takeuchi and Kobayashi



Figure 2. (a) Comparison between true and Dix-approximate phase and group velocities for fundamental-mode Rayleigh-wave dispersion between 10 and 80 s period computed for a layer ($\beta = 3.8 \text{ km/s}$) with a thickness of 38 km over a half-space ($\beta = 4.2 \text{ km/s}$). The model has constant density and a Poisson's ratio equal to 0.25. (b) Root mean square (rms) error between Dix-approximate and true phase velocity dispersion curves over a range of models. The models are the same as in panel (a) except that the layer velocity ranges from 3 to 4 km/s and the increase of velocity in the half-space ranges from 0 to 1 km/s. The rms error overall increases with the velocity jump but remains below 0.1 km/s. (c) Same as (a) except for a thickness of 33 km. (d) Same as (a) except for a thickness of 43 km. The color version of this figure is available only in the electronic edition.

(1959). In these articles, Takeuchi and coauthors attempted to represent the Rayleigh-wave eigenfunction by a decomposition into exponentially decaying basis functions, which is one approach to extending the original approximation of Jeffreys (1935). Furthermore, since the article by Dix (1955) would not appear for another 20 yr, the connection to the analogous expression in reflection seismology was not possible in Jeffreys (1935).

Depending on whether density is constant or has an assumed depth variation, either equation (3) or (7) offers a

nonperturbational way to invert Rayleigh-wave phase velocities. For a layer-over-half-space model, there are three unknowns: the squared shear velocities of the layer and halfspace and the thickness of the layer. Our approach to solving this problem is a semi-analytic method involving the solution of a reduced linear system and grid search over the thickness of the layer, because it appears nonlinearly in equations (3) and (7). Given N measurements of phase velocity within a period band (e.g., 10–40 s), we choose a subset of N - 1 of the measurements and consider a candidate value for layer thickness h. From equation (7), we form a generally overdetermined matrix equation from the N - 1 phase velocities:

$$\vec{c}^2 = \mathbf{G}(h)\vec{\beta}^2,\tag{9}$$

in which the matrix **G** has (N - 1) rows and two columns and depends on the candidate value for the thickness of the layer. This matrix equation is solved in a least-squares sense for the two shear-wave velocities (e.g., crust and upper mantle) that are then substituted into the version of equation (7) for the other Nth remaining phase velocity measurement that was excluded in forming equation (9). Note that this process of excluding one of the measurements can be repeated for each of the N phase velocity measurements. This leads to N individual misfit or error functions, with the *m*th misfit function given by

$$e_m(h) = r(f_0 + f_m)\beta_1^2 - rf_m\beta_2^2(\rho_2/\rho_1) - c_m^2,$$
(10)

in which β_1 and β_2 in this expression are solutions of equation (9) and are functions of the phase velocities and candidate value for the thickness of the layer. Finally, we sum (stack) all N of the squared misfit functions and search over layer thickness to find the one with the least stacked misfit. Once the minimum misfit is identified, that value for layer thickness can be substituted back into each of the N versions of equation (9) to obtain N estimates of the layer and half-space shear velocities. We only accept a value for layer thickness if all N of its associated layer and half-space shear velocities are real-valued. For the final estimates of the two velocities, we simply take the mean over the N estimates. We have applied this technique to phase velocity maps in the period range from 10 to 40 s assuming a typical value for the Moho density contrast given by $\rho_2/\rho_1 = 3.3/2.8 = 1.17$ (Zhang et al., 2019). The Moho density contrast can vary spatially (Schmandt et al., 2015), but such variation cannot be resolved from Rayleigh-wave phase velocities. As mentioned earlier, in the worst case that Moho density contrast varies considerably from the assumed contrast, the main effect will be the trade-off with mantle shear-wave velocity seen in equation (7).

Results

Before delving into crustal and upper mantle structure in Alaska, we first apply the Dix-based method to phase velocity maps derived from TA data in the conterminous United States. This is done to test the new Dix-based method and observe its performance in a region with relatively well-constrained structure. The phase maps we analyze have been produced by Ekström (2014, 2017) using a spectral representation of the Rayleigh-wave Green's function at eight different periods between 10 and 40 s. For the inversion, we only consider locations where measurements exist at all eight periods. Across the United States, the inversion achieves an average rms misfit of $0.1 \text{ km}^2/\text{s}^2$ in terms of the difference between the modeled and observed squared phase velocities. We reject solutions that have an rms misfit greater than $0.3 \text{ km}^2/\text{s}^2$. These misfits are given in terms of squared phase velocities due to the Dix relation being in terms of squared phase velocities as shown in equation (9). To approximate the misfit in terms of phase velocity itself, we point out that the rms of squared phase velocity, $rms(c^2)$, is given in terms of the rms of phase velocity by $2 \times c \times rms(c)$. For a nominal phase velocity value of 3 km/s, the average rms misfit of 0.1 km^2/s^2 for phase velocity squared is equal to 0.017 km/s for phase velocity. Also, we do not apply any lateral smoothing to the results. Figure 3 plots maps of layer thicknesses and a histogram of its values (Fig. 3a,b) as well as maps and histograms for layer shear-wave velocity (Fig. 3c,d) and half-space (Fig. 3e,f) shear-wave velocity. As seen in the histograms, the peak in the distributions of layer thickness (38 km), layer shear-wave velocity (3.6 km/s), and half-space shear-wave velocity (4.6 km/s) is consistent with what would be expected for generic values of the Moho, crust, and upper mantle. Indeed, the layer thickness map in Figure 3a corresponds broadly to the pattern of Moho depth (Schmandt et al., 2015) observed over most of the United States. The simplicity of the Dix approach translates into fast (<10 min) total processing time for the entire United States once phase velocity maps are in hand. We note that uncertainties for the layer thicknesses can be estimated from the depth range over which the stack of the squared misfit functions shown in equation (10)is less than twice its minimum value, which yields an average uncertainty of ± 3.5 km across the United States. Recognizable features in the maps shown in Figure 3 include low crustal velocities near the Gulf of Mexico and in the Sacramento and San Joaquin basins (2–2.8 km/s), thin crust beneath the Great basin (~25 km), thick crust beneath the Rocky Mountain Front (40-50 km), and low upper-mantle velocities associated with the Yellowstone hotspot and Rio Grande rift (~4 km/s). Other features include a sharp increase in Moho depth in eastern Washington state, in an area where the Moho has been shown to be bifurcated (Gao, 2015). Relatively low uppermantle velocities are also seen beneath New England and Virginia, anomalies that have been discussed by Shen and Ritzwoller (2016). Haney and Tsai (2015) presented results from similar processing over the western United States using only three phase velocity maps at 8, 20, and 40 s period from Lin et al. (2008, 2009). The results from Haney and Tsai (2015)

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Figure 3. Result of Dix inversion of fundamental-mode Rayleigh waves for a layer-over-half-space model across the conterminous United States. (a) Depth of the dominant surface waveguide interface, which over most of the United States corresponds to the Moho. (b) Histogram of depth values in panel (a) showing a peak in the distribution at 38 km. (c) Bulk crustal shear-wave velocity and in (d) its histogram showing a peak at around 3.6 km/s. Features in (c) include the Gulf of Mexico, San Joaquin, and Sacramento Basins. (e) Bulk upper-mantle shear-wave velocity and in (f) its histogram showing a peak at around 4.6 km/s. The Yellowstone hotspot track stands out in panel (e) as well as a low-velocity zone in New England. The color version of this figure is available only in the electronic edition.

were a proof-of-concept in contrast to the results over the entire United States shown here using the larger set of phase velocity maps between 10 and 40 s period.

In Figure S1 (available in the supplemental material to this article), we show the misfit between the input phase velocities and the phase velocities modeled from the Dix approximation at three locations. These locations are meant to illustrate three different qualities for the data fit, ranging from very good $(0.017 \text{ km}^2/\text{s}^2)$, to average $(0.163 \text{ km}^2/\text{s}^2)$, to relatively poor $(0.314 \text{ km}^2/\text{s}^2)$. The errors for these three cases are distributed fairly randomly throughout the period band. To gain insight into any systematic errors, in Figure S2 we plot the rms misfit over the conterminous United States. Areas of higher misfit correspond to shallow basins, such as the Anadarko (Oklahoma), Powder River (Wyoming), and Williston (North Dakota) basins. Thus, we conclude misfit must be concentrated in the low end of the period range 10-40 s at those locations. Future efforts to improve the misfit within the basins would require modifying equations (3) and (7) to accommodate two or possibly even more layers.

From Figure 3, we conclude that the Dix approach has been able to obtain first-order estimates of crustal and uppermantle structure across most of the United States. An exception exists in central Washington state, where our method finds an unusually thin (20-25 km) crust. In this case, the presence of a high-velocity, thickened lower crust associated with a rift structure (Catchings and Mooney, 1988; their fig. 10) pulls the interface obtained with the Dix method above the true Moho. A second notable and intriguing complication exists in Cascadia as shown in Figure 4. Plotted side-by-side in Figure 4 are the computed thickness and shear-wave velocity of the overlying layer, with the locations of Quaternary and Holocene volcanoes overlain on the map. We observe that layer thickness becomes deeper to the west of the volcanic front and stops abruptly at the

southern end of the arc. This apparent thickening of the layer is consistent with prior observations of a weak Moho interface in the Cascadia forearc. Bostock et al. (2002) originally demonstrated from seismic data that the continental Moho becomes weak in the Cascadia forearc using an east-west seismic transect from Oregon between 44° and 45° latitude. Extensive serpentinization of the mantle wedge was implicated by Bostock et al. (2002) as the cause of the weak Moho that in some cases became locally inverted. Brocher et al. (2003) showed that controlled source reflections indicate a weak Moho beneath multiple transects across the Cascades forearc. More recently, Hansen et al. (2016) and Mann et al. (2019) report a weak Moho to the southwest of Mount St. Helens, and, as seen in Figure 4a, the layer thickness is particularly deep in that location. In Figure 4b, we observe that higher velocities of the overlying layer are correlated with the thickness anomaly to the west of the volcanic arc. We interpret this as the inclusion of the serpentinized mantle wedge into the overlying layer computed from the Dix method, which serves



Figure 4. Zoom-in of the surface waveguide interface and bulk-layer shear-wave velocity in Figure 3 for the Cascadia region. Holocene and Quaternary volcanoes are plotted as red triangles. (a) The surface waveguide interface, which typically corresponds to the Moho, shows a deep anomaly to the west of the volcanic front where, in (b), the layer shear-wave velocity also shows high-velocity anomaly. The patterns are indicative of widespread serpentinization of the mantle wedge in Cascadia (Bostock *et al.*, 2002; Brocher *et al.*, 2003; Abers *et al.*, 2017). In both (a) and (b), the Siletzia terrane boundary (Wells *et al.*, 1998) is shown with a thin black line and its location is uncorrelated with the anomaly to the west of the volcanic arc. In panel (a), contours of the depth to the Juan de Fuca slab interface from McCrory *et al.* (2012) are plotted at 20, 40, 60, and 80 km depth. The deepest surface waveguide interface lies between the 40 and 60 km contours. The color version of this figure is available only in the electronic edition.

to increase the overall velocity of the layer. Thus, the Dix method finds the dominant interface responsible for guiding the Rayleigh waves and, beneath the forearc, that interface is the slab instead of the weak Moho. The results in Figure 4 indicate that the weak Moho interface extends along the entire Cascades forearc. In addition, this complication needs to be kept in mind when associating the layer thickness obtained from the Dix method with the Moho, particularly in subduction zones where the Moho may be weak to nonexistent.

To give evidence that the anomalous pattern in Cascadia is not primarily due to accretion of the Siletzia oceanic terrane, we also plot the lateral extent of Siletzia from Wells *et al.* (1998) in Figure 4a,b. We observe that the Siletzia terrane boundaries cross the region of anomalous thickness and shear-wave velocity of the overlying layer. Thus, the forearc-wide anomalies of layer thickness and shear-wave velocity in Figure 4 are not dominantly produced by the Siletzia lower crust having higher velocities more typical of the mantle (Crosbie et al., 2019). They are more likely the result of reduced velocity of the forearc mantle due to hydration by post-Eocene subduction. This interpretation is consistent with the anomalies cross-cutting multiple crustal provinces that accreted to western North America prior to initiation of the Cascades arc (Dickinson, 2004). Moreover, the anomalous thicknesses in Figure 4a have values in the range of 40-50 km to the west of the volcanic arc, which are consistent with estimates of the depth to the Juan de Fuca slab in this portion of the forearc. We plot depth contours of the slab interface at 20, 40, 60, and 80 km depth from McCrory et al. (2012) in Figure 4a, and the anomalous waveguide interface thicknesses in the range of 40-50 km are observed to lie between the 40 and 60 km slab depth contours.

Having tested the Dix method in the conterminous United States, we turn to the analysis of phase velocity maps computed in Alaska using data from the TA. Estimates of the

Moho we derive from these maps can be compared to some of the first estimates of the Moho across the entire state of Alaska recently published by Miller and Moresi (2018), Zhang et al. (2019), and Feng and Ritzwoller (2019). Several different collections of Rayleigh-wave phase velocity maps from 10 to 40 s exist for Alaska. Ward and Lin (2018) imaged phase velocity at 14 periods between 10 and 40 s using traveltime information derived from ambient seismic noise correlations. Phase maps in Alaska have also been produced by Ward (2015), Feng and Ritzwoller (2019), and Berg et al. (2020). We apply the Dix method to the phase maps of Ward and Lin (2018), although the maps are similar to those of Ward (2015) and Berg et al. (2020). There are also similarities to the Rayleigh phase maps by Feng and Ritzwoller (2019). In fact, Ward and Lin (2018) compared their results to a Moho model computed with the Dix method applied to the phase velocity maps of Ward (2015) over a portion of Alaska and

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found general agreement with their 3D shear-wave velocity model.

In Figure 5, we plot maps and histograms of the Moho depth and the velocities of the crust and upper mantle in Alaska computed with the Dix method using the phase maps of Ward and Lin (2018). Similar to the conterminous United States, inversion is only applied at locations in Alaska where measurements exist at all 14 periods between 10 and 40 s, and we reject solutions that return an rms misfit in terms of squared phase velocity that is greater than $0.3 \text{ km}^2/\text{s}^2$. Because of these criteria, there is an area in the Yukon and a portion of northern Alaska where solutions are not obtained (Fig. 5). The Dix-based inversion achieves an average rms misfit for squared phase velocity of 0.12 km^2/s^2 in Alaska, slightly higher than the conterminous United States. In Figure 5e,f, the histograms indicate that the average crust beneath Alaska is overall thinner (31 km) and slower in velocity (3.4 km/s) than in the conterminous United States (Fig. 3) with values more in line with those observed in the western United States. The upper mantle is also slower, with a peak in its distribution at 4.3 km/s. Figure 5a-c shows maps of crustal shear-wave velocity, upper-mantle shear-wave velocity, and crustal thickness, respectively. Several first-order features are apparent in these maps, including in Figure 5c regions of thick crust south

Figure 5. Result of Dix inversion of fundamental-mode Rayleigh waves for a layer-over-half-space model across Alaska. Major terrane boundaries are plotted from Colpron et al. (2007). (a) Bulk crustal shear-wave velocity with labels indicating Yukon Flats basin (YF), Nenana basin (NB), Talkeetna Mountains (TM), and Cook Inlet basin (CIB). (b) Bulk upper-mantle shear-wave velocity with the location of the Wrangell Volcanic Field (WVF) shown. (c) Depth of the dominant surface waveguide interface, or Moho, with labels showing the Brooks range (BR), the Denali fault (DF), and the Ahklun Mountains (AM). Earthscope Transportable Array (TA) stations in Alaska shown with green circles. (d) Histogram of crustal thickness across Alaska with a peak in the distribution at 31 km. (e) Histograms of crustal (unfilled) and upper-mantle (filled) shear-wave velocities with peaks at 3.4 and 4.3 km/s. The color version of this figure is available only in the electronic edition.

of the Denali fault, within the Brooks Range, and beneath the Ahklun Mountains. Allam *et al.* (2017) have previously detected a sharp 10 km offset in the Moho across the Denali fault, and a similar jump is shown in Figure 5c. Sedimentary basins in interior Alaska, such as the Yukon Flats and Nenana basins (Van Kooten *et al.*, 2012), have overall lower bulk crustal shear-wave velocity in Figure 5a and overlie thin crust in Figure 5c. Relatively thin crust extends westward from these



Figure 6. Comparison between Moho depth from receiver functions by Miller and Moresi (2018) and the Moho interface derived from Rayleigh waves in this study. Holocene and Quaternary volcanoes are plotted as orange and red triangles, respectively, and Denali is indicated with a black triangle. The receiver function estimates are point measurements shown as circles filled in according to their depth. Overall agreement between the two Moho estimates is observed, especially in the Nenana basin, Ahklun Mountains, St. Elias Mountains, and the Brooks range. The color version of this figure is available only in the electronic edition.

basins, consistent with the lack of high topography in western Alaska (Zhang et al., 2019). In contrast, some of the deepest Moho in Alaska exists in portions of the highest regions of the St. Elias Mountains near the Canadian border, in an area where a deep Moho interface has been previously reported by Christeson et al. (2013). The Cook Inlet Basin displays low crustal velocity in Figure 5a, but the Moho there is not as shallow as seen below the Interior basins. An area of high crustal velocity is bordered to the north by the Talkeetna Mountains fault (Fig. 5a), on the western side of the Wrangellia Composite Terrane, and correlates with a mid-crustal high-velocity anomaly imaged in a similar location by Feng and Ritzwoller (2019). Finally, low mantle shear-wave velocity exists to the north of the Wrangell Volcanic Field (WVF) (Fig. 5b) as previously noted by Ward (2015). Interestingly, no large-scale structural features are observed to coincide with the active volcanic arc to the west of the Cook Inlet. Moho depths beneath the Cook Inlet volcanoes fall in the range of 30-35 km, similar to the locations of volcanic DLP earthquakes at these volcanoes (Power et al., 2004, 2013).

Discussion

Having applied the Dix method to both the Cascadia and Alaska subduction zones, we find significant differences in structure between the two regions in relation to the volcanic arc. Unlike Cascadia, where prominent thickness and shear-wave velocity anomalies are observed in the forearc (Fig. 4), the Cook Inlet forearc region exhibits a typical Moho depth and slower than normal bulk shear-wave velocity. Such lack of an anomalous forearc in Alaska is consistent with Cascadia being a younger, hotter subduction zone with extensive serpentinization of the mantle wedge. Abers et al. (2017) point out that, globally, Cascadia is an extreme endmember among subduction zones. Regarding the WVF, its relation to possible subduction and the presence of a slab beneath it are still under debate (Jiang et al., 2018; Martin-Short et al., 2018; Feng and Ritzwoller, 2019). From the Dix results, there are anomalously large values of thickness

to the southeast of WVF (Fig. 5c) but no associated anomaly for bulk crustal shear-wave velocity (Fig. 5a) as seen in Cascadia. Thus, the cause of the deep Moho in this region may be more a consequence of the general correlation between thicker crust and high topography (Zhang *et al.*, 2019). Assuming subduction occurs beneath WVF, there may also be contamination of the mantle wedge in this region with crustal material producing a thicker apparent crust (Abers *et al.*, 2017; Ward and Lin, 2018).

The complexity of the Moho near subduction zones warrants a comparison of the Dix-derived Moho with the recent Slab1.0 (Hayes *et al.*, 2012) and Slab2 (Hayes *et al.*, 2018) models of the slab interface beneath Alaska. Figures S3 and S4 show the areas of positive difference between Slab1.0 and Slab2, respectively, and our Moho interface; that is, the areas where either Slab1.0 or Slab2 lie above the Moho. Figure S3 shows that there is a limited region where Slab1.0 is above our Moho interface, but it is only at most about 5 km higher. In contrast, there is a broad region where Slab2 is higher than our Moho interface (by up to as much as 50 km), and it



Figure 7. (a) Histogram of the difference between the Moho obtained from surface waves and receiver functions (Miller and Moresi, 2018) in Alaska, showing the surface-wave Moho being most likely 2 km shallower than from receiver functions. (b) Crossplot of Moho from surface waves and receiver functions in Alaska, along with a best-fit line (red-dashed) and a line of perfect correlation (black-dashed). The crossplot has a relatively low R^2 equal to 0.14. (c) Same as (a) but for the conterminous United States using receiver functions by Schmandt *et al.* (2015), showing the surface-wave Moho being most likely 4 km shallower than from receiver functions. (d) Same as (b) but for the conterminous United States. The crossplot has a relatively higher R^2 than observed for Alaska given by 0.55. The color version of this figure is available only in the electronic edition.

corresponds to the area of flat subduction associated with the collision of the Yakutat microplate.

We compare the Moho interface derived from Rayleigh waves across Alaska with receiver function estimates by Miller and Moresi (2018) in Figure 6. General agreement exists between the two data sets, with the thickest crust in the Brooks Range, Chugach Mountains, and Wrangell-St. Elias Mountains. Both data sets find thin crust beneath the Nenana basin. These features are also exhibited in the Moho models for Alaska by Zhang *et al.* (2019) and Feng and Ritzwoller (2019). Although the average Moho we obtain from Rayleigh waves is 2 km shallower than that from receiver functions (Fig. 7a; compare to Fig. 7c for the conterminous United States), and the correlation coefficient between the two Moho determinations is fairly low ($R^2 = 0.14$, Fig. 7b; compare to $R^2 = 0.55$, Fig. 7d

for the conterminous United States), the somewhat poorer comparison in Alaska could have a number of explanations. For example, it may be related to the larger spacing of the TA in Alaska (85-100 km) compared to the conterminous United States. It may also be due to the presence of steeply dipping faults bounding active tectonic provinces in Alaska that are smoothed over using the Rayleigh-wave approach but not by receiver function analysis. The rapid variation of the Moho depth in places leads to the need to fit a smooth interface to receiver function data points (Miller and Moresi, 2018; Zhang et al., 2019), in contrast to the inherent lateral smoothness of the Moho derived from Rayleigh waves. The Moho model for Alaska developed by Feng and Ritzwoller (2019), being similarly based on surface waves, also displays such laterally smooth features. On the other hand, receiver functions are able to obtain Moho estimates in areas with sparse instrumentation such as the Aleutian Islands (Miller and Moresi, 2018). From this comparison, the Rayleigh-wave Dix method is seen to provide gen-

erally similar patterns of bulk crustal and upper-mantle structure as established techniques such as receiver functions.

Conclusion

We have developed a method to obtain first-order estimates of bulk crustal and upper mantle shear-wave velocity, as well as Moho depth, from fundamental-mode Rayleigh waves and applied it to phase velocity maps in Alaska derived from TA data. The method is based on an approximate Rayleighwave dispersion formula and yields fast results with a minimal amount of parameterization. We tested the method with highquality phase velocity maps from the conterminous United States and found general agreement with several previously known features. A complication for mapping of the Moho across the conterminous United States was encountered in the Cascadia forearc. It was due to the presence of weak Moho and revealed that, across much of the forearc in Cascadia, Rayleigh waves between 10 and 40 s period are primarily guided by the slab. No such weak Moho feature emerged from the analysis of phase velocity maps in Alaska, consistent with Alaska being an older and colder subduction zone. Across Alaska, the crustal and upper-mantle structure obtained from the new Rayleigh-wave method agreed overall with Moho depth estimates from recent studies that utilized receiver functions and advanced analysis of surface-wave dispersion. The convergence of various Moho depth models is bringing into focus a region that, prior to the TA, had been sparsely instrumented. The ability of the new method we developed indicates that first-order estimates of major discontinuities in depth such as the Moho can be obtained in fast, straightforward manner solely from Rayleigh-wave dispersion measurements.

Data and Resources

We provide supplemental material in the form of four supplemental figures, Figures S1–S4, and their captions. Data used in this study are available at the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC), which is supported by the National Science Foundation under Cooperative Support Agreement EAR-1851048.

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