

1           **Along-Strike Variations of Alaska Subduction Zone Structure and Hydration**  
2                           **Determined From Amphibious Seismic Data**  
3

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12   **Key Points:**

- 13       • Crustal thickness of the inner forearc (35-42 km) generally exceeds that of the volcanic  
14       arc, but becomes variable in the Shumagin segment.
- 15       • The Shumagin segment has more incoming plate mantle hydration than the Semidi  
16       segment, aligning with abundant plate bending normal faults.
- 17       • Hydration extends to depths of 18 km below the Moho, indicating more water subducts  
18       than most previous estimates.  
19

## 20 **Abstract**

21 We develop a 3-D isotropic shear velocity model for the Alaska subduction zone using data from  
22 seafloor and land-based seismographs to investigate along-strike variations in structure. By  
23 applying ambient noise and teleseismic Helmholtz tomography, we derive Rayleigh wave group  
24 and phase velocity dispersion maps, then invert them for shear velocity structure using a  
25 Bayesian Monte Carlo algorithm. For land-based stations, we perform a joint inversion of  
26 receiver functions and dispersion curves. The forearc crust is relatively thick (35-42 km) and has  
27 reduced lower crustal velocities beneath the Kodiak and Semidi segments, which may promote  
28 higher seismic coupling. Bristol Bay Basin crust is relatively thin and has a high-velocity lower  
29 layer, suggesting a dense mafic lower crust emplaced by the rifting processes. The incoming  
30 plate shows low uppermost mantle velocities, indicating serpentinization. This hydration is more  
31 pronounced in the Shumagin segment, with greater velocity reduction extending to  $18 \pm 3$  km  
32 depth, compared to the Semidi segment, showing smaller reductions extending to  $14 \pm 3$  km  
33 depth. Our estimates of percent serpentinization from  $V_S$  reduction and  $V_P/V_S$  are larger than  
34 those determined using  $V_P$  reduction in prior studies, likely due to water in cracks affecting  $V_S$   
35 more than  $V_P$ . Revised estimates of serpentinization show that more water subducts than  
36 previous studies, and that twice as much mantle water is subducted in the Shumagin segment  
37 compared to the Semidi segment. Together with estimates from other subduction zones, the  
38 results indicate a wide variation in subducted mantle water between different subduction  
39 segments.

40

## 41 **Plain Language Summary**

42 This study uses seismic data from the 2018-2019 Alaska Amphibious Community Seismic  
43 Experiment and other land stations to image the 3-D seismic velocity structure of the Alaska  
44 subduction zone. The analysis combines constraints from both Rayleigh waves and converted  
45 body waves. The results provide insight into the distinct lateral variations observed for many  
46 properties of the subduction zone. Thick, low-velocity forearc crust is found beneath the Kodiak  
47 and Semidi segments, which may be related to the higher seismic coupling in these regions. The  
48 Bristol Bay Basin has a thin crust with a high velocity lower layer, suggesting a dense mafic  
49 lower crust emplaced by the extensional processes that formed the basin. Low velocities in the  
50 incoming plate near the trench in the Shumagin segment indicate pronounced mantle hydration,  
51 extending to about 18 km below the Moho. Together with estimates from other subduction  
52 zones, the results indicate a wide variation in subducted mantle water between different  
53 subduction segments.

54

## 55 **1 Introduction**

56 Subduction zones are the locus of many of the most important geological processes,  
57 including earthquakes, volcanism, sediment accretion, and the formation of new crust. It is  
58 particularly useful to study the along-strike variability of these parameters within a single large  
59 subduction segment. The Alaska subduction zone is one of the most tectonically active plate  
60 boundaries worldwide, with numerous large earthquakes and active volcanoes. In the Alaska  
61 Peninsula region, the subducting plate has an intermediate age ( $\sim 50$ -55 Ma) and a relatively  
62 uniform convergence rate ( $\sim 63$  mm/yr) (DeMets et al., 2010), but the seismicity, geodetic

63 locking, and earthquake rupture zones show distinct along-strike variations (e.g., Davies et al.,  
64 1981; Shillington et al., 2015; Xiao et al., 2021) (**Figure 1**). These along-strike variations make  
65 it an ideal place to study many subduction zone processes, including earthquake and geodetic  
66 properties, forearc and backarc tectonics, and the pathways of water through the subduction  
67 system. Many aspects of the along-strike variations in these processes can be revealed by  
68 detailed imaging of variations in subduction zone structure.

69 Subduction zones are also the key to understanding the long-term water cycle on Earth,  
70 since subducting oceanic plate serves as the only mechanism to carry water into the deep interior  
71 of the Earth. The water flux from hydrated oceanic plates is essential for arc volcanism and may  
72 control along-arc changes in magma chemistry (Sadofsky et al., 2008; Manea et al., 2014).  
73 Furthermore, water greatly lowers the viscosity of the surrounding mantle, which is important for  
74 the dynamics of subduction (e.g., Hebert et al., 2009). Hydration of downgoing slabs is likely to  
75 be ubiquitous in the worldwide oceanic trench regions, but the degree and extent of mantle  
76 hydration is poorly constrained, and remains the main uncertainty in the global water cycle (van  
77 Keken et al., 2011). A previous study of the Alaska subduction zone suggested strong along-  
78 strike variations in mantle hydration of the incoming plate in the offshore Alaska Peninsula  
79 region (Shillington et al., 2015).

80 Active-source studies have revealed low mantle velocities in the plate-bending regions of  
81 many subduction zones that are interpreted as indicating serpentinization of mantle peridotite  
82 (Ranero et al., 2003; Van Avendonk et al., 2011; Shillington et al., 2015; Fujie et al., 2018;  
83 Arnulf et al., 2022), including offshore of the Shumagin segment (Shillington et al., 2015;  
84 Acquisto et al., 2022b; Shillington et al., 2022). However, in most cases active-source studies  
85 provide only limited constraints on the depth extent of mantle serpentinization, leaving large  
86 uncertainties in the amount of bound water subducted. Passive-source studies, on the other hand,  
87 are able to image the deeper structure of the incoming plate and place constraints on the depth of  
88 serpentinization. A passive source study in the Mariana subduction zone found that the mantle  
89 hydration at the Mariana trench extends to ~24 km below the Moho (Cai et al., 2018), suggesting  
90 the total amount of water input into the Mariana trench is at least 4.3 times more than previous  
91 estimates (van Keken et al., 2011). Since the Alaska Subduction Zone differs from Mariana in  
92 terms of incoming plate age (~50 Ma, as opposed to ~150 Ma for Mariana), and shows extensive  
93 along-strike variation in incoming plate fabric and faulting, it represents an excellent target to  
94 test the possible widespread occurrence of deeper incoming plate hydration.

95 The 2018-2019 Alaska Amphibious Community Seismic Experiment (AACSE) deployed  
96 an amphibious array of 75 ocean bottom seismographs (OBS) and 30 land seismographs,  
97 complemented by a temporary nodal array on Kodiak Island and an active-source seismic  
98 experiment offshore the Semidi segment and Kodiak Island (Barcheck et al., 2020). The  
99 amphibious array of AACSE, along with several other simultaneously operating land seismic  
100 networks, provide a unique opportunity to image subduction zone structure. In this study, we use  
101 this dataset to image the along-strike variations in the Alaska subduction zone structure and  
102 provide important constraints on the hydration of the downgoing plate, as well as the structure  
103 and tectonics of the forearc and backarc regions. In contrast to previous surface wave  
104 tomographic studies that carry out large-scale imaging of the entire Alaska region, or analyze  
105 only one type of seismic data, our analysis includes both ambient noise tomography and  
106 teleseismic earthquake Helmholtz tomography, as well as P-wave receiver functions, to better  
107 constrain the Alaska subduction zone with higher resolution.

108

## 109 2 Tectonic Setting and Previous Work

110 From west to east, the plate boundary can be divided into three main segments, named  
111 Shumagin, Semidi, and Kodiak segments after the corresponding forearc islands (**Figure 1b**). The  
112 Shumagin segment, also called the Shumagin Gap, is a region with abundant interplate and  
113 intermediate-depth earthquakes (**Figure 1a**), but has not ruptured during a great earthquake ( $M \geq$   
114 8) for at least 150 years (Davies et al., 1981). The Semidi and Kodiak segments, however, have  
115 much less seismicity at all depths and have ruptured in several historical great earthquakes (e.g.,  
116 1938 M8.2; 1946 M8.6; 1964 M9.2) (Davies et al., 1981) (**Figure 1b**). In recent years, two  
117 megathrust events (July 22, 2020 M7.8 Simeonof earthquake and July 29, 2021 M8.2 Chignik  
118 earthquake) occurred in the Semidi segment (**Figure 1b**). The slip model of the 2021 M8.2  
119 Chignik earthquake suggests that its rupture zone is within the estimated 1938 M8.2 aftershock  
120 zone (Chengli Liu et al., 2022; He et al., 2023). The 2020 M7.8 Simeonof earthquake, however,  
121 expanded westward and is considered to have ruptured the eastern Shumagin Gap and a little bit  
122 of the westernmost Semidi segment (Liu et al., 2020), and was followed by an unusual strike-slip  
123 M7.6 event within the Shumagin Gap on October 19, 2020 (Y. Jiang et al., 2022). In addition,  
124 geodetic results suggest that the plate motion along the megathrust changes from creeping nearly  
125 aseismically in the western portion of the Shumagin segment, to weakly coupled in the eastern  
126 portion of the Shumagin segment, to intermediate coupled in the Semidi segment, to strongly  
127 locked in the Kodiak segment (S. Li & Freymueller, 2018; Drooff & Freymueller, 2021; Xiao et  
128 al., 2021). The characteristics of double seismic zones (DSZs) also show clear variations  
129 between each segment (Wei et al., 2021; Aziz Zanjani & Lin, 2022).

130 The distribution of incoming plate fabric, sediment thickness, and plate bending faults are  
131 also highly variable along the strike. The incoming plate fabric has a variable orientation  
132 resulting from the spreading along the Kula plate and Kula-Farallon ridge (e.g., Lonsdale, 1988;  
133 Bradley et al., 2003). Magnetic anomalies on the incoming Pacific plate show that the paleo-  
134 spreading direction changes from sub-parallel to the trench axis in the Shumagin segment to sub-  
135 perpendicular to the trench axis in the Semidi and Kodiak segments (**Figure 1a**). As for the  
136 sediment thickness of the incoming plate, the Shumagin segment has disrupted sediments (~0.5  
137 km) while the Semidi segment has relatively thick and stable sediment (~1 km) (Shillington et  
138 al., 2015; J. Li et al., 2018). Furthermore, the Shumagin segment shows abundant outer-rise  
139 faults whereas the Semidi and Kodiak segments have few outer-rise faults (Shillington et al.,  
140 2015).

141 A previous active source study compared profiles between Shumagin and Semidi  
142 segments, finding much stronger velocity reduction and thus hydration in the Shumagin segment  
143 (Shillington et al., 2015; Shillington et al., 2022). They suggest that hydration is controlled by  
144 the intensity of plate-bending faults, which is in turn strongly influenced by the plate abyssal hill  
145 fabric formed during spreading (e.g., D. H. Christensen & Ruff, 1988; Masson, 1991; Ranero et  
146 al., 2003). Small faults from the plate spreading fabric are parallel to magnetic anomalies, being  
147 oriented roughly trench-parallel in the Shumagin segment but at high angles to the trench axis in  
148 the Semidi segment (**Figure 1**; **Figure 8a**). However, other studies suggest that factors like the  
149 trench-ridge angle (Fujie et al., 2018), or slab curvature (Naliboff et al., 2013) may be more  
150 important in controlling the along-strike variation.

151 The overriding plate in the Alaska Peninsula is built from a series of accreted terranes,  
152 including the Peninsula Terrance, the Chugach Terrane, and the younger Prince William Terrane  
153 (Bruns et al., 1985; Horowitz et al., 1989) (Figure 1c). Active-source projects EDGE (Moore et  
154 al., 1991) and ALEUT (Shillington et al., 2015; Bécel et al., 2017) have revealed variations in P-  
155 wave velocity and upper plate structures associated with these accreted terranes. The crustal  
156 structure of the Kodiak Shelf shows a series of arched reflectors in the lower crust that coincide  
157 with low-velocity rocks, providing evidence for large-scale underplating between Kenai  
158 Peninsula and Kodiak Island (Moore et al., 1991; Ye et al., 1997). The forearc structure is  
159 spatially complex as a result of this accretion history, and may provide important controls on the  
160 seismogenic characteristics of megathrust earthquakes, such as their down-dip limit and the  
161 seismogenic extent (Kuehn, 2019; Shillington et al., 2022). In the Shumagin segment, the outer  
162 forearc has a small frontal prism and hosts active crustal-scale splay faults, indicating a typical  
163 tsunamigenic structure (Bécel et al., 2017; von Huene et al., 2021). Downdip variations in the  
164 seismic reflection character of the plate interface at the eastern Shumagin segment have been  
165 linked to the changes in fault structure and corresponding seismogenic behaviors (J. Li et al.,  
166 2015). Along-strike variations in pore-fluid pressure and sediment thickness appear to correlate  
167 with changes in seismicity, locking, and earthquake history (J. Li et al., 2018).

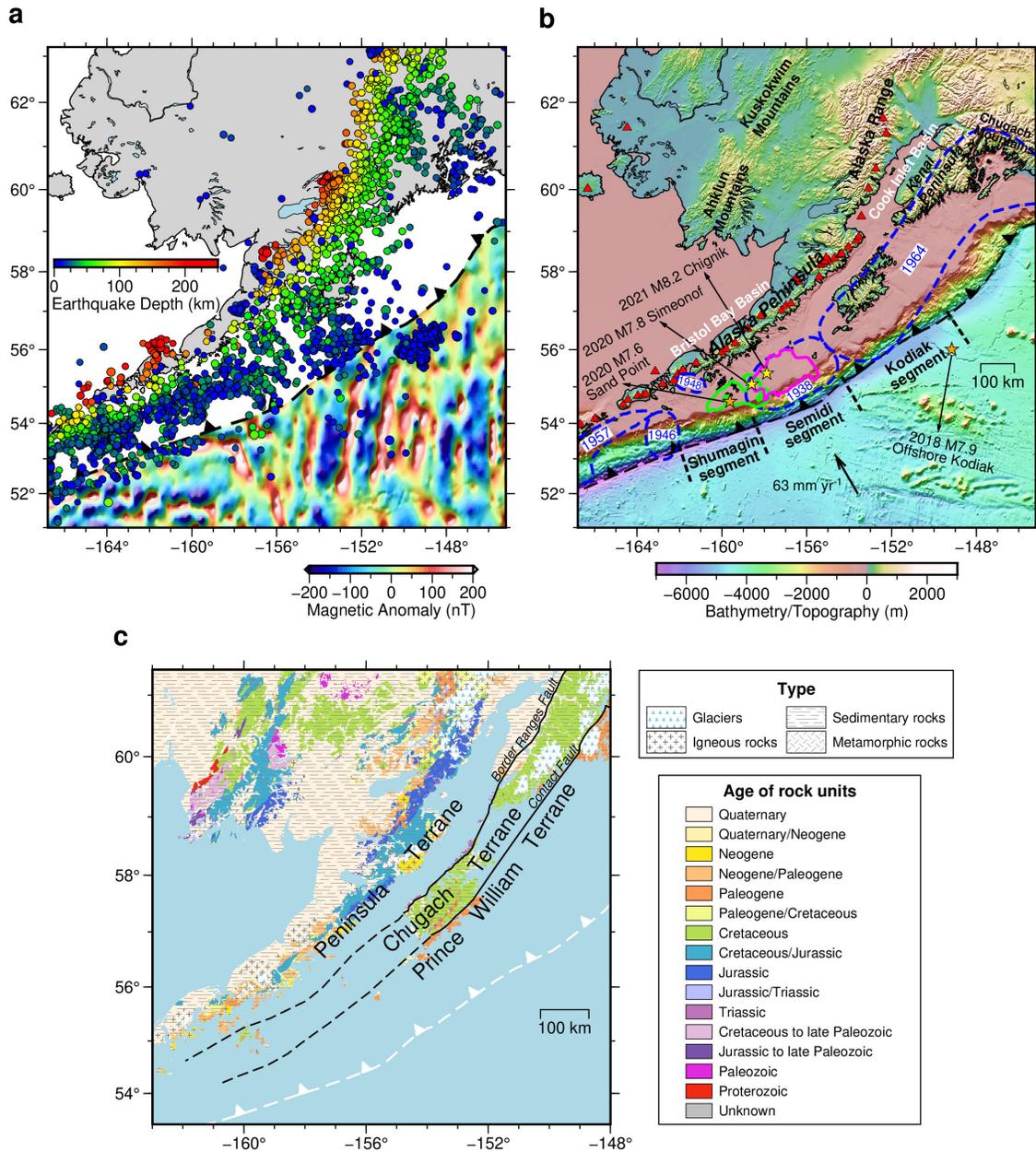
168 The volcanic arc and backarc also show variations between each segment. Active  
169 volcanoes are widely distributed along the Alaska Peninsula and the southwestern Alaska Range  
170 (Figure 1b). Alaska has about 100 volcanoes active in the past 11,000 years and more than 50  
171 volcanoes considered historically active since 1760 (Cameron et al., 2018). In spatial  
172 distribution, both Semidi and Kodiak segments have strong arc volcanism, especially at the arc  
173 front of the Kodiak segment. The composition of arc lavas varies along the Alaska Peninsula,  
174 which has been attributed to variations in water and sediments entering the subduction zone (Wei  
175 et al., 2021). The Bristol Bay Basin, located on the north side of the Alaska Peninsula in the  
176 backarc of the Shumagin and Semidi segments, contains a significant thickness of sediments.  
177 Active-source studies and numerical models suggest that Bristol Bay formed mainly through two  
178 stages. In the early or middle Eocene through a late Miocene phase, extension led to fault-  
179 controlled subsidence. Then a late Eocene through Holocene phase of volcanic-arc loading or  
180 northward prograding delta led to flexural subsidence (Walker et al., 2003).

181 Previous passive source seismic studies have mostly investigated the structure of  
182 continental portions of southern and central Alaska (Eberhart-Phillips et al., 2006; Qi et al.,  
183 2007; Y. Wang & Tape, 2014; Ward, 2015; Martin-Short et al., 2018; Yang & Gao, 2020) or  
184 along the Alaska Peninsula (You Tian & Zhao, 2012; Janiszewski et al., 2013). Benefiting from  
185 the Earthscope Transportable Array that deployed in Alaska from 2014 until 2021, the entire  
186 Alaska region has also been imaged for both isotropic and anisotropic velocity structures (C.  
187 Jiang et al., 2018; Ward & Lin, 2018; Feng & Ritzwoller, 2019; Gou et al., 2019; Berg et al.,  
188 2020; Feng et al., 2020) as well as seismic velocity interfaces (Gama et al., 2022). However, the  
189 lack of seismic array coverage offshore the Alaska Peninsula and the large-scale  
190 parameterization of the studies leads to the lack of resolution in the incoming plate and forearc  
191 area of the Alaska subduction zone.

192 Several recent studies have taken advantage of the 2018-2019 AACSE dataset to carry  
193 out more detailed studies of the Alaska Peninsula region. A nodal seismograph array installed on  
194 Kodiak Island was used to image the structure immediately beneath Kodiak Island (Onyango et  
195 al., 2022). Airgun shots recorded by AACSE ocean bottom seismographs provide improved

196 estimates of shallow crustal structure in the offshore peninsula area (Acquisto et al., 2022a).  
 197 Body wave tomography (Gou et al., 2022; F. Wang et al., 2022), and surface wave tomography  
 198 (Feng, 2021; Chuanming Liu et al., 2022) provide improved images of the structure beneath the  
 199 region. Constraints on azimuthal anisotropy are provided by a recent shear-wave splitting study  
 200 (Lynner, 2021).

201



202

203 **Figure 1.** The geological setting of the Alaska subduction zone, with the plate boundary marked  
 204 as the dashed black line with triangles. **(a)** The magnetic anomaly of the incoming Pacific plate  
 205 and earthquake distribution. The incoming plate magnetic anomalies are from EMAG2v3 (Meyer

206 et al., 2017). The earthquakes with  $M \geq 4$  from the Alaska Earthquake Information Center  
207 (AEIC) catalog from 1990 to 2022 and from the AACSE catalog during the AACSE deployment  
208 (Ruppert et al., 2022) are plotted as circles colored by their epicentral depths. **(b)** Great  
209 earthquake rupture zones and prominent geological regions (e.g., peninsulas, mountains, basins).  
210 Dashed blue contours show the rupture zones of historical earthquakes (Davies et al., 1981).  
211 Yellow stars indicate the epicenters of megathrust events: the 2020 M7.8 Simeonof earthquake  
212 and 2021 M8.2 Chignik earthquake, and their rupture zones are shown as green and magenta,  
213 respectively (Liu et al., 2020; Chengli Liu et al., 2022). Orange stars show the epicenters of the  
214 2018 M7.9 Offshore Kodiak earthquake and the 2020 M7.6 Sand Point earthquake, both of  
215 which are intraplate strike-slip events. The division of the Shumagin segment, Semidi segment,  
216 and Kodiak segment is labeled on the incoming Pacific plate. The convergence rate between the  
217 Pacific plate and the North American plate is relatively uniform in the study area, thus a black  
218 arrow with the average value ( $\sim 63 \text{ mm yr}^{-1}$ ) is marked on the incoming plate (DeMets et al.,  
219 2010). **(c)** Locations of the Peninsula Terrane, the Chugach Terrane, and the Prince William  
220 Terrane (Horowitz et al., 1989) on the geologic map of the Alaska subduction zone (F. H. Wilson  
221 et al., 2015). These terranes have distinctly different rock ages, bounded by the Border Ranges  
222 Fault and the Contact Fault, respectively. The well-determined positions of faults are shown as  
223 solid lines and locations that are only approximate are shown as dashed lines.

224

### 225 **3 Data and Method**

#### 226 **3.1 Seismic data**

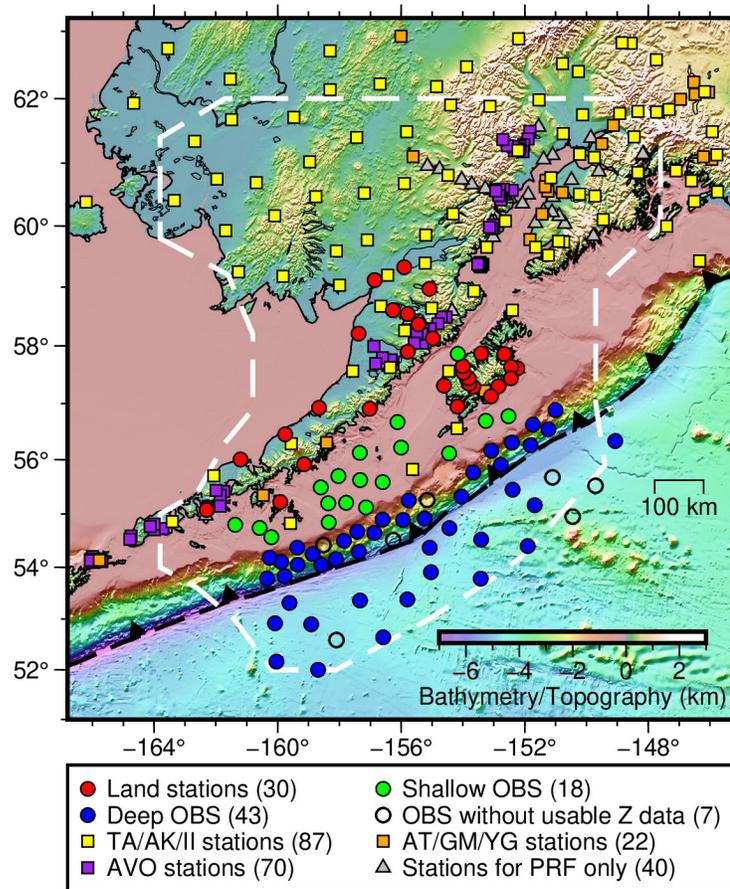
227 Data for this analysis come from the Alaska Amphibious Community Seismic  
228 Experiment (AACSE; May 2018 – September 2019) as well as several land seismograph  
229 networks. The AACSE deployed an array of broadband seismic stations covering the Alaska-  
230 Aleutian Trench and the Alaska Peninsula from May 2018 to September 2019, including 75  
231 ocean bottom seismographs (OBSs) and 30 land stations (Barcheck et al., 2020). At the same  
232 time, the EarthScope Transportable Array (TA) was operating many seismic stations throughout  
233 Alaska. Therefore, land stations from TA and several other networks also augment the  
234 amphibious seismic array.

235 After excluding those stations with non-broadband sensors, or poor quality, this study  
236 includes 61 AACSE ocean bottom seismographs (OBSs), 30 AACSE land stations, and 179 land  
237 stations of other networks (network codes: TA, AK, II, AT, GM, YG, AV) to analyze the  
238 Rayleigh wave dispersion. The quality of stations is examined for their long-term noise levels by  
239 calculating the Power Spectral Density (PSD) Power Density Function (PDF). Poor-quality  
240 stations are those with strange shapes or extreme values in the PSD PDF plots. Altogether, we  
241 use a dense amphibious array of 270 seismic stations to perform tomographic inversion (see  
242 Section 3.3 to 3.5). In addition, we use 40 land stations that operated outside the AACSE  
243 deployment time period for P-wave receiver functions only (see Section 3.6). The detailed  
244 distribution of stations that contribute to this study is shown in **Figure 2**.

245 Some AACSE land sites on Kodiak Island and the Alaska Peninsula experienced bear  
246 attacks on their GPS antennae and lost time synchronization. Some OBSs have possible clock  
247 drift due to the early shutdown of their dataloggers and clocks. We test and correct the time drift

248 using a cross-correlation technique (Stehly et al., 2007; Gouédard et al., 2014); please see the  
 249 supplement of Barcheck et al. (2020) for the details on the method and correction results.

250



251

252 **Figure 2.** A map of the seismic stations used in this study. The stations are plotted with different  
 253 shapes and colors according to their networks and types. Red, green, and blue circles represent  
 254 land stations, shallow ocean bottom seismographs (OBS), and deep OBS, respectively, that were  
 255 deployed by the AACSE project (Barcheck et al., 2020), OBSs without usable vertical  
 256 component data are shown as blank circles. Yellow squares are the land stations of the  
 257 Transportable Array (TA), Alaska Network (AK), and Global Seismic Network. Orange squares  
 258 are the land stations of the Tsunami Warning, and U.S. Geological Survey Networks (GM).  
 259 Purple squares are the stations belonging to the Alaska Volcano Observatory (AVO) network,  
 260 which are clustered around volcanoes. Gray triangles are those stations that are used for receiver  
 261 function analysis only. Note that the 4 OBSs that were not recovered and 3 OBSs that failed to

262 record any seismic or pressure data are not included here (Barcheck et al., 2020). The dashed  
263 white line encloses the study region defined in Section 3.5.

264

### 265 3.2 Pre-processing the OBS data

266 OBS data usually contain noises that are not present in land seismic records, since the  
267 seafloor environment generally has higher noise levels in the long-period seismic band. Previous  
268 work has shown that for OBS vertical component data, there are two main noise sources: one is  
269 tilt noise resulting from variable ocean-bottom currents tilting the instrument and causing  
270 horizontal noise to be recorded by the vertical component; the other is compliance noise resulting  
271 from the vertical movements of seafloor due to sea bottom pressure changes resulting from  
272 infragravity waves (Webb & Crawford, 1999; Crawford & Webb, 2000). The similarity of tilt  
273 noise on vertical and horizontal components, and the similarity of compliance noise on the  
274 vertical component and the differential pressure gauge (DPG) suggest that both noises on the  
275 vertical component can be largely removed by estimating spectral transfer functions (Bell et al.,  
276 2014). For various types of AACSE OBSs, there are three different types of pressure channels:  
277 differential pressure gauge (DPG), absolute pressure gauge (APG), and hydrophone. Through  
278 tests and comparisons, we found that both DPG and APG are able to remove the compliance  
279 noise, while the hydrophone is not very successful.

280 We use equations 1-7 in Bell et al. (2014) to calculate the spectral transfer function used  
281 for noise removal. The transfer functions are best calculated from time series without  
282 earthquakes, so we select time windows for transfer function estimation by combining methods  
283 described in previous studies (Ye Tian & Ritzwoller, 2017; Janiszewski et al., 2019; Ma et al.,  
284 2020). First we predict the arrival time of earthquakes with  $M_S/M_W \geq 4.5$  using the International  
285 Seismological Centre (ISC) catalog where the Rayleigh wave time window is taken from 20 s  
286 before a predicted  $4.0 \text{ km s}^{-1}$  arrival to 600 s after it, and exclude any time windows that overlap  
287 with the Rayleigh wave windows (Ma et al., 2020). Then we check the remaining time windows  
288 in an earthquake band (10-40 s) and remove those with suspicious high amplitudes (either small  
289 earthquakes or signal singularities). Furthermore, the remaining time windows are then evaluated  
290 using a norm outlier rejection method (Janiszewski et al., 2019). In this way, the selected time  
291 windows will contain purely noises.

292 We use the coherence of the transfer function between the vertical and horizontal seismic  
293 components as well as the vertical and the pressure time series to determine the frequencies for  
294 noise removal. To avoid over-corrections that would distort the signals, we follow Ye Tian and  
295 Ritzwoller (2017) and only apply corrections for periods where the transfer function coherence is  
296 above 0.4, which mostly lies in a period range between 15 and 150 s. After removing the tilt and  
297 compliance noise, the surface wave signals extracted from both ambient noise cross-correlations  
298 and teleseismic earthquakes are distinctly improved (**Figure S1 in the Supporting Information**).

### 299 3.3 Ambient noise tomography

300 With the data of all land stations and pre-processed OBSs, the interstation empirical  
301 Green's functions are then determined by ambient noise cross-correlation procedures described

302 in Bensen et al. (2007). First we cut the continuous data to daily length and down-sample them to  
303 one sample per second. Then we calculate the ambient noise cross-correlations over the vertical-  
304 vertical components of daily length time series using both time-domain normalization with an  
305 earthquake filtering band of 10-40 s and spectral whitening. Daily cross-correlations are stacked  
306 for each station pair over the entire time period of the deployment.

307 We then apply an automated Frequency-Time Analysis (FTAN) with a phase-matched  
308 filter to measure the Rayleigh wave phase and group velocity dispersion curves from the  
309 symmetric Green's functions of each station pair (Bensen et al., 2007; Lin et al., 2008). The  
310 FTAN method directly measures group velocity dispersion, but requires reference phase velocity  
311 dispersion curves to avoid cycle-skipping problems in determining the phase velocity dispersion.  
312 To avoid the cycle-skipping problems, we use a two-step process similar to Lin et al. (2008): (1)  
313 We apply FTAN using the reference interstation phase velocity dispersion curves from global  
314 Rayleigh wave dispersion model GDM52 (25-250 s) (Ekström, 2011), resulting preliminary  
315 measurements where most station pairs have resolved the cycle-skipping problems. Using the  
316 tomographic method and selection criteria described below, we invert for preliminary phase  
317 velocity maps at periods between 8 and 36 s. We use these maps to estimate the dispersion  
318 curves for every station pair which we then use as the revised reference curves. (2) We repeat the  
319 FTAN using the revised reference interstation phase velocity curves, resulting all interstation  
320 measurements without cycle-skipping problems. As there are rapid changes from oceanic to  
321 continental structures, the FTAN measurements for land-land station pairs, OBS-land station  
322 pairs, and OBS-OBS pairs also vary a lot (**Figure S2 in the Supporting Information**). The oceanic  
323 paths generally show extremely low group/phase velocity at periods < 16 s but increase rapidly  
324 to high velocity at periods > 20 s.

325 To quantify the strength of signals for each station pair, we define the frequency-  
326 dependent signal-to-noise ratio (SNR) as the ratio of the signal peak in the predicted arrival  
327 window to the root mean square (rms) of the noise trailing the arrival window, in each period  
328 band for the symmetric component cross-correlation. The prediction window is defined by  
329 assuming surface waves travel between 0.5 and 5.0 km/s, and the trailing noise window starts  
330 500 s after the predicted window until the end of lag time. There are relatively few good  
331 measurements below 8 s, and the SNR decreases rapidly for oceanic paths at periods greater than  
332 36 s, thus we invert for phase and group velocity maps from 8 to 36 s using a Gaussian ray-  
333 theoretical tomography method (Barmin et al., 2001). The grid spacing is  $0.3^\circ \times 0.2^\circ$ , which is  
334 roughly equally spaced in longitude and latitude. The isotropic cell size in the tomographic  
335 inversion is  $0.5^\circ$ , which could recover checker sizes ranging from  $3^\circ \times 2^\circ$  to  $1.8^\circ \times 1.2^\circ$  in  
336 checkerboard tests (**Figure S3 in the Supporting Information**).

337 For each frequency, we only keep station pairs with distances larger than twice the  
338 wavelength. To exclude the unreliable measurements while considering the relatively high noise  
339 of OBS records, we excluded measurements with  $\text{SNR} < 7$ . To further constrain the 2-D  
340 inversion results, we apply quality control based on the travel-time residuals from the previous  
341 inversion. The paths with residuals outside two standard deviations, about 2 % to 6 % of the total  
342 measurements for each period, are removed after three times of quality control. The remaining  
343 measurements are used to finalize the Rayleigh phase and group velocity maps from 8 s to 36 s  
344 (**Figure S4 in the Supporting Information**). At short periods (between 8 to 14 s), the group and  
345 phase velocity maps reflect the very shallow structure and water depth, where incoming plate  
and trench are dominated by low-speed anomalies and mountain ranges show high-speed

347 anomalies. At longer periods (20-30 s), the group and phase velocity maps reflect the crust and  
348 uppermost mantle structure, where the incoming plate is dominated by high-speed anomalies and  
349 low-speed anomalies cover the forearc region.

### 350 3.4 Teleseismic earthquake tomography

351 At longer periods ( $T > 20$  s), we analyze the Rayleigh wave phase velocity from  
352 teleseismic waves traversing the array using the Helmholtz tomography method (Lin &  
353 Ritzwoller, 2011) implemented in the ASWMS package (Jin & Gaherty, 2015). We select  
354 earthquakes with  $M_w \geq 5$  and epicentral distances between  $20^\circ$  and  $160^\circ$  from the International  
355 Seismological Centre (ISC) catalog for analysis. The events are chosen to be high-quality,  
356 relatively evenly distributed with respect to the seismic array, and also separated enough in time  
357 from each other to avoid overlapping on seismograms. High-quality events refer to those that  
358 pass the automatic quality control in the ASWMS package based on the coherence of nearby  
359 stations and reasonable misfit in the Eikonal and amplitude inversions. Finally, 265 earthquakes  
360 are used to determine the phase velocities.

361 The implementation of Helmholtz tomography involves Eikonal tomography plus the  
362 amplitude term correction, where Eikonal tomography inverts the phase delays for spatial  
363 variations in apparent phase velocity via the Eikonal equation (Lin et al., 2009) and amplitude  
364 Laplacian term correction accounts for the local amplification due to wavefield focusing and  
365 defocusing effects (Lin et al., 2012; Eddy & Ekström, 2014; Russell & Dalton, 2022). The  
366 amplitude term corrects for the influence of non-plane wave propagation on the apparent phase  
367 velocities, allowing for the recovery of the true structural phase velocity via the Helmholtz  
368 equation. The waveforms of all events and stations are cut from the earthquake origin time to  
369 10800 s after. Based on multichannel cross-correlations of station pairs within 410 km, the phase  
370 velocity variations of a series of periods are estimated for each event at node spacing  $0.3^\circ \times 0.2^\circ$ .  
371 We estimate the local amplification term (Eddy & Ekström, 2014), calculate the smoothed  
372 Laplacian term of corrected 2-D amplitudes, and finally convert the apparent phase velocity to  
373 structural phase velocity for each period. The final structural phase velocity dispersion maps are  
374 stacked over maps of all events. The checkerboard tests show that the inverted velocity maps show  
375 distinct checkers and generally recover more than 80% of input anomaly amplitudes, suggesting  
376 that the parameters above work well ([Figure S5 in the Supporting Information](#)).

377 The tomographic results produce isotropic phase velocity at node spacing  $0.3^\circ \times 0.2^\circ$  for  
378 periods from 23 s to 100 s ([Figure S6 in the Supporting Information](#)). The phase velocity maps at  
379 these longer periods constrain the lower crust and uppermost mantle structure. At the 40 s period,  
380 the high-velocity anomalies still dominate the incoming plate region, and also extend north  
381 across the Aleutian Trench a little bit compared to the 25 s phase velocity map. At even longer  
382 periods (e.g., 60 s, 100 s), the trench region is replaced by low velocity, and high-speed  
383 anomalies gradually occupy the volcanic arc.

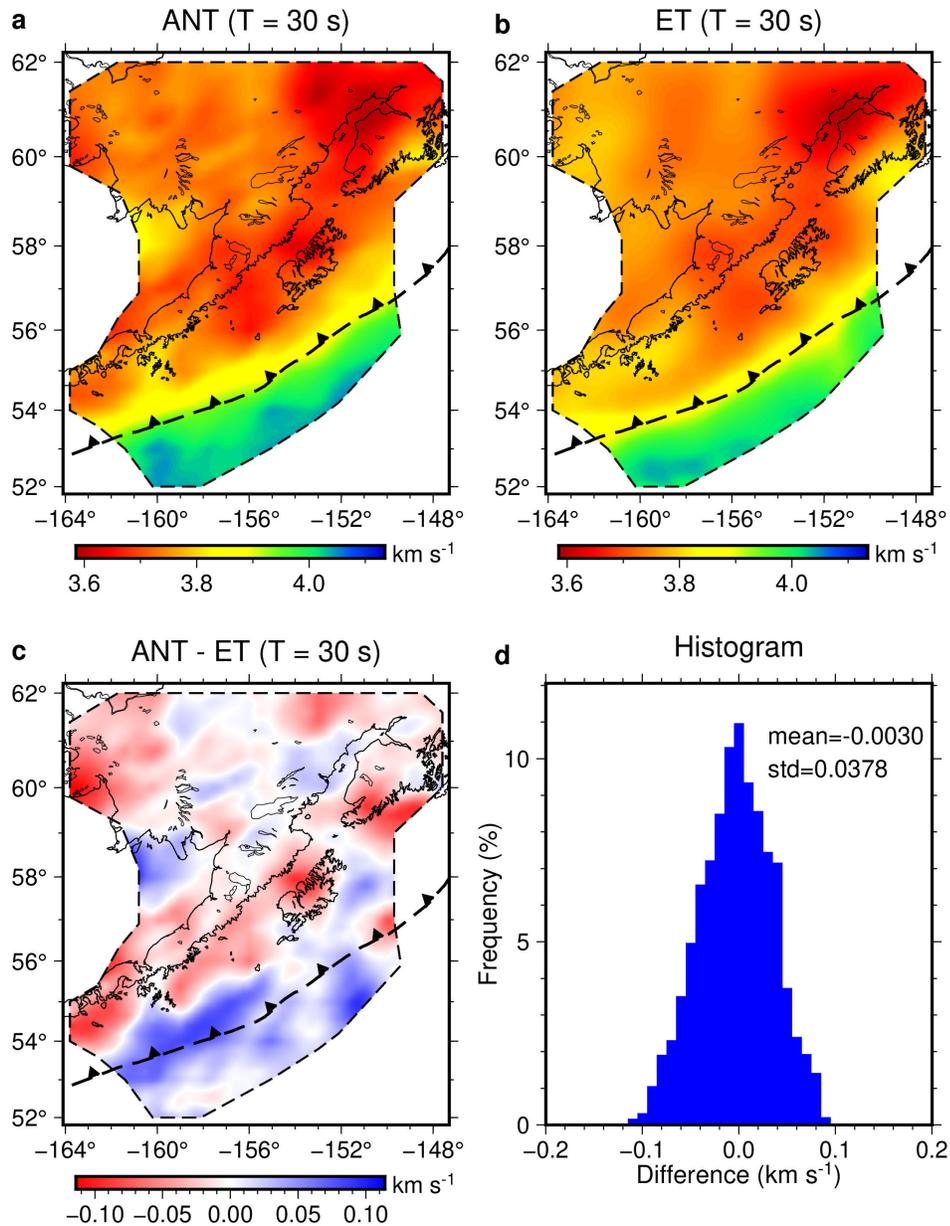
### 384 3.5 Local Rayleigh wave dispersion curves

385 The Rayleigh wave phase velocity dispersion curves measured from ambient noise and  
386 earthquake data are then evaluated in their overlapping period band. The study region is defined  
387 by the areas that are well-recovered in checkerboard tests of both ambient noise tomography and  
388 teleseismic earthquake tomography. Comparisons between the phase velocity maps show that

389 their measurements are generally consistent at 24-34 s periods. For example, at 30 s, most of the  
390 phase velocity differences are less than  $0.1 \text{ km s}^{-1}$  (Figure 3). The incoming plate, trench, and  
391 northern Bristol Bay area generally have slightly higher phase velocity from ambient noise  
392 tomography than from earthquake tomography, whereas other areas show the opposite  
393 relationship (Figure 3). Furthermore, we define the reliable range of nodes to extract local  
394 dispersion curves as those with high ray path coverage in both ambient noise tomography and

395 teleseismic earthquake tomography, as well as small phase velocity differences in their  
 396 overlapping phase velocity maps.

397



398

399 **Figure 3.** Comparison of Rayleigh wave phase velocity at 30 s period of the study region  
 400 estimated from ambient noise tomography (ANT) and earthquake Helmholtz tomography (ET).  
 401 **(a)** The phase velocity map at 30 s from ANT. **(b)** The phase velocity map at 30 s from ET. **(c)**  
 402 The difference between phase velocity maps from ANT and ET at 30 s. **(d)** Histogram of the

403 differences in (c), showing that the results from the two tomography results are generally less  
 404 than  $0.1 \text{ km s}^{-1}$ .

405

406 Therefore, the final phase velocity dispersion curves for each node are constructed in the  
 407 following way: 1) For periods less than or equal to 23 s, the phase velocities come from ambient  
 408 noise tomography. 2) For periods greater than or equal to 36 s, the phase velocities come from  
 409 Eikonal tomography. 3) For periods larger than 23 and less than 36 s, we take a weighted  
 410 average of the measurements from the two methods, with the weights changing linearly in  
 411 between.

412 Uncertainty estimation of the local phase and group velocity curves is important for the  
 413 shear velocity inversion. The tomographic methods used here do not provide an estimation of  
 414 uncertainties directly, but we can estimate the local uncertainties for short periods from the local  
 415 resolution in the method of Barmin et al. (2001). Similar to Shen et al. (2016), we use an  
 416 empirical scaling relationship:

$$417 \quad \sigma(\mathbf{r}) = kR(\mathbf{r}) \quad (1)$$

418 where  $\sigma(\mathbf{r})$  is the uncertainty estimate at location  $\mathbf{r}$ , and  $R(\mathbf{r})$  is the estimate of resolution, which  
 419 is the standard deviation of the resolving kernel at the location (Barmin et al., 2001). We  
 420 estimate the value of  $k$  for each period, so that a local resolution of  $\sim 50 \text{ km}$  (i.e., minimum  
 421 resolution value in the data-rich region) produces a phase velocity uncertainty estimate of  $0.027$   
 422  $\text{km s}^{-1}$  for 8 s,  $0.021 \text{ km s}^{-1}$  for 16 s,  $0.016 \text{ km s}^{-1}$  for 24 s,  $0.021 \text{ km s}^{-1}$  for 32 s,  $0.024 \text{ km s}^{-1}$  for  
 423 periods  $> 32 \text{ s}$ . Uncertainties for other periods are interpolated based on these anchor points.  
 424 From an empirical relationship, the uncertainties of group velocity are estimated to be twice that  
 425 of phase velocity (e.g., Moschetti et al., 2010; Shen et al., 2016). Considering that the group  
 426 velocity measurements in this region have even larger uncertainty at shorter periods, we use a  
 427 factor of 2.5 to calculate the group velocity uncertainties. The local uncertainties of phase  
 428 velocity from the Helmholtz tomography are scaled from the corresponding standard deviation  
 429 values by multiplying a factor of 0.3. In this way, the phase velocities from two datasets at  
 430 overlapped periods have similar uncertainties.

### 431 3.6 P-wave receiver functions for land stations

432 Contrasting to the surface wave analysis that requires a concurrent deployment of seismic  
 433 stations, the P-wave receiver functions (PRFs) analysis is performed on each station individually.  
 434 To use joint inversion to better constrain the continental Moho, we try to include all land stations  
 435 within the study region that operated sufficient dates from May 2014 to December 2021. The  
 436 longer date range is chosen to make the best use of TA stations and other temporary stations with  
 437 enough data outside the AACSE deployment period. For all land stations with sufficient data  
 438 quantity and quality, we first prepare the P-wave seismic data from earthquakes with  $m_b \geq 5.0$   
 439 and epicentral distances between  $30^\circ$  and  $90^\circ$ . The seismograms are decimated to a sample rate  
 440 of 10 Hz and cut to a time window from 30 s before and 60 s after the P-wave onset. The  
 441 horizontal components are cosine tapered and pre-filtered with a bandpass filter of 0.02 to 2 Hz,  
 442 then rotated into radial and transverse components. Using a time-domain iterative deconvolution

443 algorithm (Ligorria & Ammon, 1999), we perform 200 iterations to estimate the PRFs, using a  
444 Gaussian low pass filter with a corner frequency of  $\sim 1$  s.

445 We apply automated quality control to the PRFs in two steps. First we correct the time to  
446 align the Ps phase and check individual PRFs to exclude those problematic ones (e.g., extreme  
447 amplitude, negative polarity at  $t = 0$ ). Then we use the similarity of PRFs over the range of back-  
448 azimuths to further constrain the quality and generally retain more than 30 PRFs for each station.  
449 If the individual PRFs have good azimuthal coverage, a “harmonic stripping” method is applied  
450 to determine the isotropic or average PRF, which represents the common component over all  
451 azimuths (Shen et al., 2013). For stations lacking a good azimuthal distribution of individual  
452 PRFs, we use a weighted stack of all PRFs to get a single PRF for the station site. In total, we  
453 obtain 188 land stations with quality-controlled stacked PRFs. The stacked PRFs of stations  
454 along profiles suggest that the overall quality of the PRFs is reasonably good to constrain the  
455 interface structures ([Figure S7 in the Supporting Information](#)).

### 456 3.7 Bayesian Monte Carlo inversion

457 The resulting local Rayleigh wave dispersion curves with group velocity from 8-36 s and  
458 phase velocity from 8-100 s are then inverted for the azimuthally averaged vertically polarized  
459 shear wave velocity ( $V_{SV}$ ) structure using a Bayesian Monte Carlo inversion method (Shen et al.,  
460 2013; Shen & Ritzwoller, 2016). The Bayesian inversion requires the proper construction of the  
461 model space and the estimation of prior information, which is based on the location of the nodes.  
462 We divide the nodes into three groups: the incoming plate group to the south of the trench axis,  
463 the inner trench slope group just north of the trench axis, and the forearc and backarc group. The  
464 boundary between the inner trench slope and the forearc/backarc region is taken as the 20 km  
465 depth contour of the slab interface from the Slab2.0 model (Hayes et al., 2018).

466 For the nodes in the ocean, we include a water layer with a starting thickness from the  
467 125-km Gaussian-filtered bathymetry ( $m_{w0}$ ) and allow a 100% thickness but no more than 1.5  
468 km perturbation. The incoming plate nodes include a 0-1 km sedimentary layer in the Shumagin  
469 segment and a 0-2 km sedimentary layer in the Semidi and Kodiak segments, based on the  
470 previous active-source results (Shillington et al., 2015). For inner trench nodes, the starting  
471 crustal thickness of ( $m_{c0}$ ) is calculated following the depth of slab interface as well as the slab  
472 dip angle in the Slab2.0 model, with an assumption of a 6 km oceanic crust atop the subducting  
473 slab. The crustal thickness of most nodes in the inner trench slope region then allows a 30%  
474 thickness perturbation with respect to  $m_{c0}$ . One exception is the Kenai Peninsula nodes, which  
475 have a slab interface less than 20 km in the Slab2.0 model, but tend to have a deeper Moho than  
476 that predicted by the slab interface through tests. We thus allow those nodes within the Kenai  
477 Peninsula to have a 60% thickness perturbation with respect to  $m_{c0}$ . For all nodes, the uppermost  
478 mantle structure from the Moho discontinuity down to 300 km depth is represented by a 6-knot  
479 B-spline curve. The bottom 100 km is gradually merged into the STW105  $V_{SV}$  model  
480 (Kustowski et al., 2008). The parameterization and search range of the velocity models in  
481 different regions are defined by a series of variables for three groups of nodes ([Table 1](#)). Each

482 variable in different regions is set accordingly based on our a priori information of the study  
 483 region.

484

485 Table 1. Parameterization and search range for the velocity models in different regions.

		Incoming plate	Inner trench slope	Forearc and backarc
Water layer (for oceanic nodes only)	Thickness	$m_{w0} \pm \min(m_{w0}, 1.5)$ (km)		
Sedimentary layer	Thickness	Shumagin: 0-1 km Semidi & Kodiak: 0-2 km	0-6 km	0-6 km
	$V_{SV}$ (top: 1.0 km s <sup>-1</sup> ; bottom: 2.0 km s <sup>-1</sup> )	Linear velocity increase, with top and bottom allows 1.0 km s <sup>-1</sup> and 1.5 km s <sup>-1</sup> perturbation, respectively		
Crustal layer	Thickness	4-8 km	For nodes within Kenai Peninsula: $m_{c0} \pm 0.6 m_{c0}$ (km)  For others: $m_{c0} \pm$ 0.3 $m_{c0}$ (km)	20-50 km
	$V_{SV}$ (top: 3.1 km s <sup>-1</sup> ; bottom: 3.8 km s <sup>-1</sup> )	Linear velocity increase, both variables allow 20% perturbation	3 cubic B-spline coefficients, each allows 20% perturbation	4 cubic B-spline coefficients, each allows 20% perturbation
Mantle layer	$V_{SV}$ (top: 4.2 km s <sup>-1</sup> ; bottom: 4.4 km s <sup>-1</sup> )	6 cubic B-spline coefficients, each allows 25% perturbation		

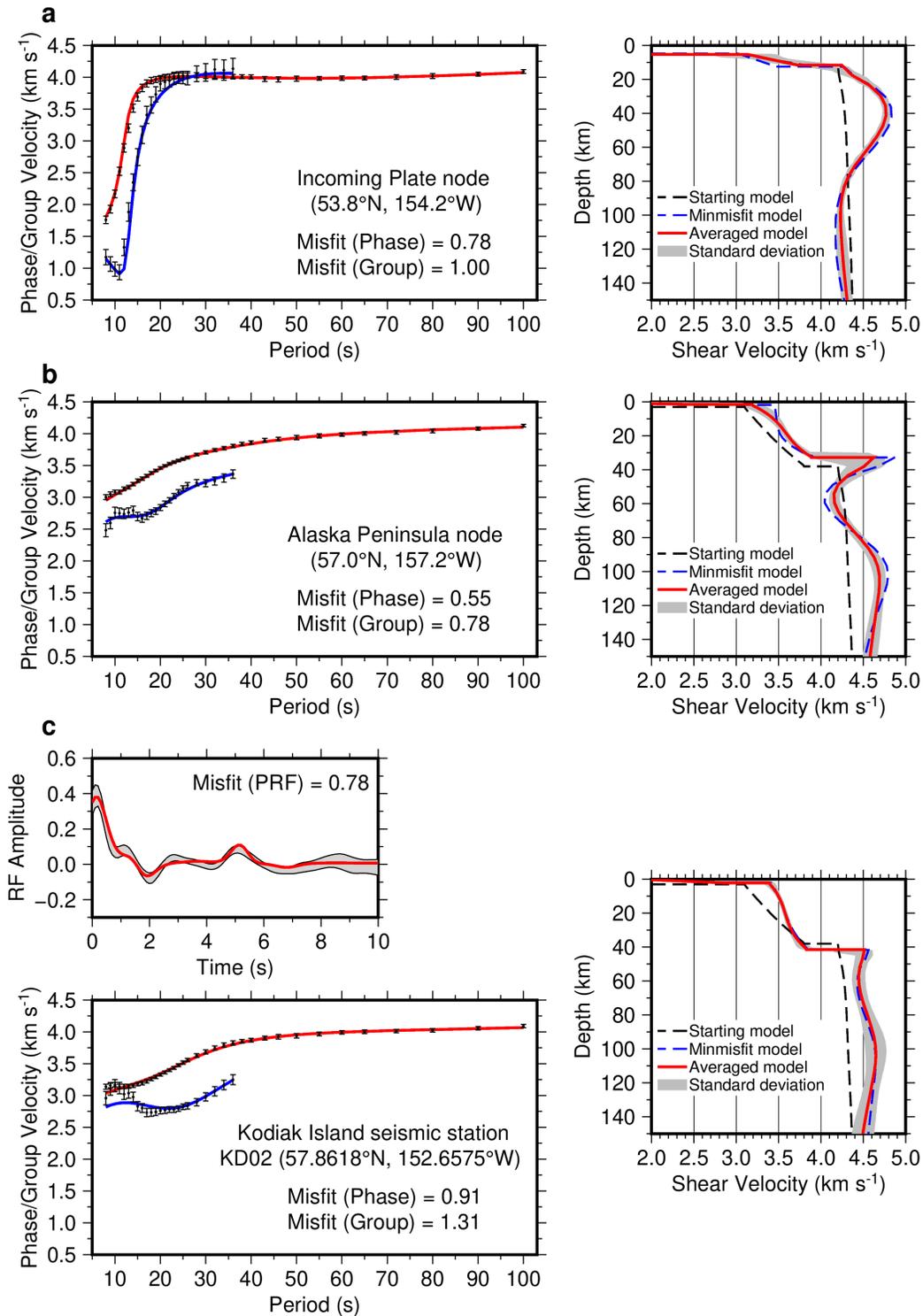
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487 The Bayesian Monte Carlo inversion is performed with a grid of  $0.3^\circ \times 0.2^\circ$  spaced nodes  
 488 which have phase and group velocity measurements from both ambient noise data and  
 489 teleseismic earthquake data. Examples of inversion results show that the Bayesian Monte Carlo  
 490 inversion can well fit the measured group and phase velocity dispersion curves ([Figure 4a](#);

491 **Figure 4b**). Finally, all 2015 evenly spaced nodes give structures that are based on the mean of  
492 at least 5000 acceptable models.

493 For land stations with high-quality PRFs, their local structures at stations are jointly  
494 determined using the Rayleigh wave dispersion and PRFs. The Moho conversion in the PRF  
495 between 3 and 7 s is very helpful to invert the Moho depth and resolve potential trade-offs  
496 between Moho depth and velocity structure (**Figure 4c**), so we fit the first 10 s of the PRFs.  
497 Among all 188 stations with high-quality PRFs, we finally get 180 well-constrained joint  
498 inversion results and their structures are generally based on the mean of at least 500 acceptable  
499 models. The joint inversion requires fewer accepted models to achieve meaningful and stable  
500 results since the receiver function helps reduce the model space that fits the datasets. The 3-D  
501 structural model is constructed on the grid of evenly spaced nodes by combining the structure  
502 from the Rayleigh wave inversion with the PRF joint inversion results for all well-constrained  
503 stations within a 75 km distance, using an inverse distance weighting scheme (Shen et al., 2018).  
504 The structure for nodes lacking nearby land seismographs with good PRF results is based solely  
505 on the Rayleigh wave inversion results. The final 3-D azimuthally-averaged vertically-polarized  
506 shear velocity model is determined using all the well-constrained nodes by interpolating with a  
507 simple kriging algorithm (Shen & Ritzwoller, 2016; Shen et al., 2018).

508



509

510 **Figure 4.** Examples of the Bayesian Monte Carlo inversion for different geological regions show  
 511 the resulting 1-D shear wave velocity structure beneath each node and how the predicted phase  
 512 and group dispersion curves (and receiver function, if applicable) fit the measurements. **(a)**

513 Incoming plate node (53.8°N, 154.2°W). The derived  $V_{SV}$  profile is an oceanic structure with  
514 reduced velocity in the uppermost mantle. (b) Alaska Peninsula node (57.0°N, 157.2°W). The  
515 derived  $V_{SV}$  profile is a typical volcanic arc structure with a low-velocity zone (LVZ) beneath  
516 the Moho. (c) Joint inversion result for Kodiak Island seismic station KD02. The structure shows  
517 a thick, low-velocity crust and a strong subducting oceanic Moho discontinuity.

518

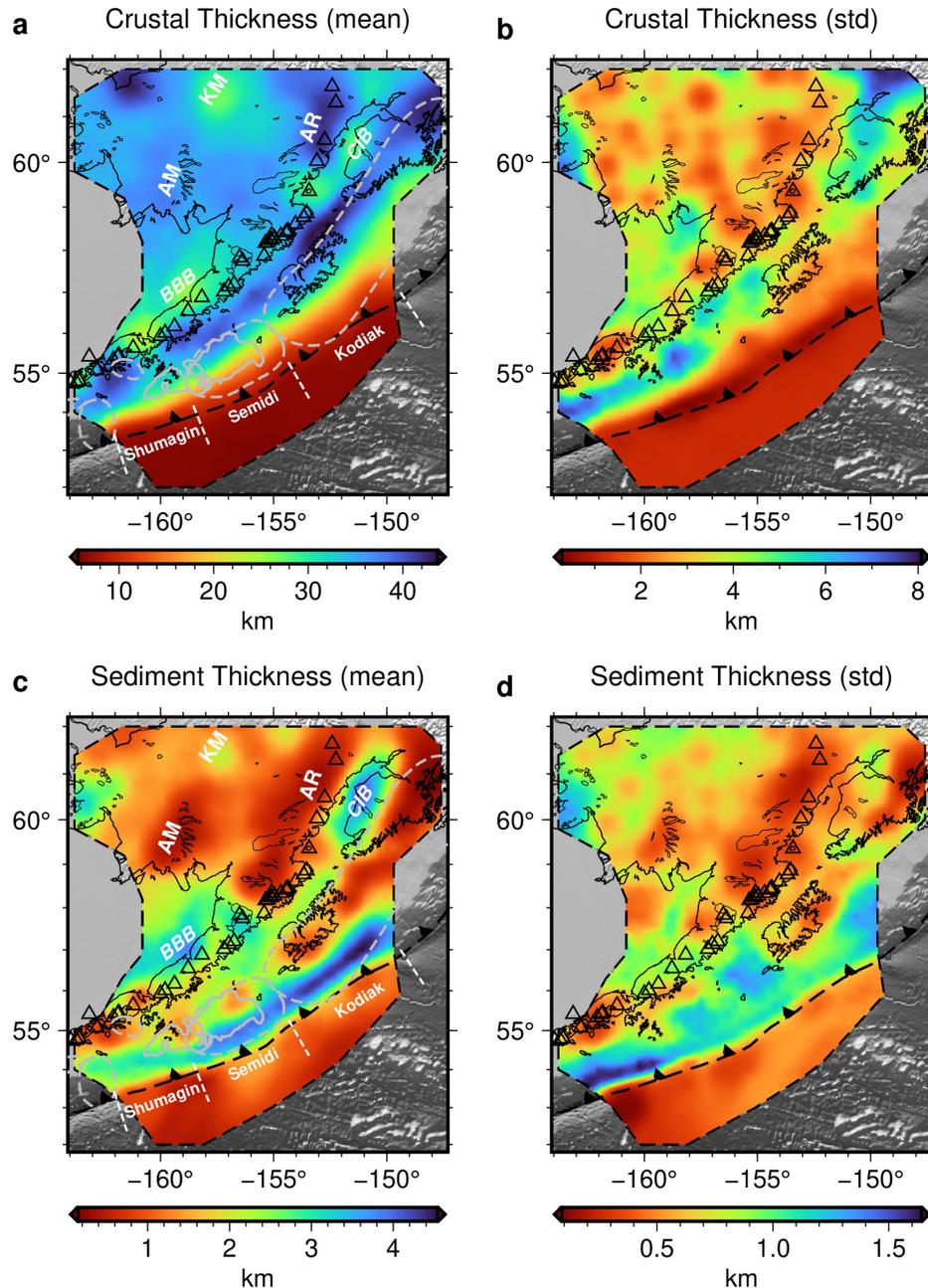
## 519 **4 Results**

### 520 4.1 Thickness of crust and sediment

521 The posterior distribution of the crustal thickness and sediment thickness provides their  
522 preferred values and uncertainty maps (Figure 5). Note that the inner forearc structure is rather  
523 complex as there are possibly two Moho discontinuities (an overriding plate Moho and a  
524 subducting plate Moho), but we only set one Moho in the parameterization since the resolution  
525 of the methods does not allow for reliably determining a complex structure. The inversion  
526 generally picks the shallowest Moho. Therefore, the observed Moho in the seaward part of the  
527 forearc represents the subducting plate Moho and the observed Moho in the arcward section of  
528 the forearc represents the overriding plate Moho, with a section in between where the overriding  
529 and subducting Mohos are in close proximity and the identity of the Moho discontinuity from the  
530 inversion is uncertain. The thickest crust is a band in the inner forearc from the Kodiak segment  
531 to the eastern edge of the Shumagin segment, whose crustal thickness exceeds that in the arc and  
532 backarc regions (Figure 5a). The comparison between the final uncertainty map and that from  
533 surface wave inversion only (Figure S7 in the Supporting Information) clearly shows how the  
534 PRFs help reduce uncertainty (Figure 5b; Figure S7b). The crustal thickness along the arc is  
535 relatively constant (32-36 km) to the west of the Alaska Range (>40 km), slightly thinner than  
536 indicated by a previous receiver function analysis of the stations along the Aleutian arc mostly  
537 west of the study region (Janiszewski et al., 2013).

538 Sediment as defined in this study includes both recent pelagic and terrigenous sediment  
539 as well as deformed and potentially older sediments in forearc basins and in the accretionary  
540 prism. The mean distribution of sediment thickness is generally less than 2 km in the incoming  
541 plate and the continental regions. Though the inversion method is not highly sensitive to thin  
542 sedimentary cover, the model clearly resolves thicker sediment along the outer forearc of the  
543 Kodiak and Semidi segments and in the Bristol Bay and Cook Inlet basins. The very thick (up to  
544 4.5 km thickness) low-velocity sediments in the outer forearc basin (Figure 5c) are consistent  
545 with the outer forearc basin structure in the Shumagin segment (Shillington et al., 2022) and to  
546 the south of Kodiak Island (Fisher & von Huene, 1982) determined using active source methods,  
547 though thickest sediments of Shillington et al. (2022) are in the accretionary prism. Bristol Bay  
548 Basin and the Cook Inlet Basin both show about 3 km of low-velocity sediments. The  
549 distribution of sediment thickness in the Cook Inlet Basin has a similar pattern to the map of  
550 depth to the base of Cenozoic strata (Shellenbaum et al., 2010; Silwal et al., 2018). The previous  
551 active-source survey in Bristol Bay Basin shows a boundary at about 3 km depth for the faulted  
552 basement (Walker et al., 2003), consistent with the sediment thickness results here.

553



554

555 **Figure 5.** Map views of the posterior distribution for the crustal thickness and sediment  
 556 thickness of the study region. The background image is the topography/bathymetry in gray  
 557 scales. **(a)** Map view of the mean of the crustal thickness. The dashed gray lines are the contours  
 558 of earthquake rupture zones shown in **Figure 1b**. The dashed white lines marked the range of the  
 559 Shumagin, Semidi, and Kodiak segments. Annotation of geological features: AM = Ahklun  
 560 Mountains; KM = Kuskokwim Mountains; AR = Alaska Range; BBB = Bristol Bay Basin; CIB  
 561 = Cook Inlet Basin. **(b)** Map view of the uncertainty of the crustal thickness. **(c)** Map view of the  
 562 mean of the sediment thickness. Other labels are the same as that in **(a)**. **(d)** Map view of the  
 563 uncertainty of the sediment thickness.

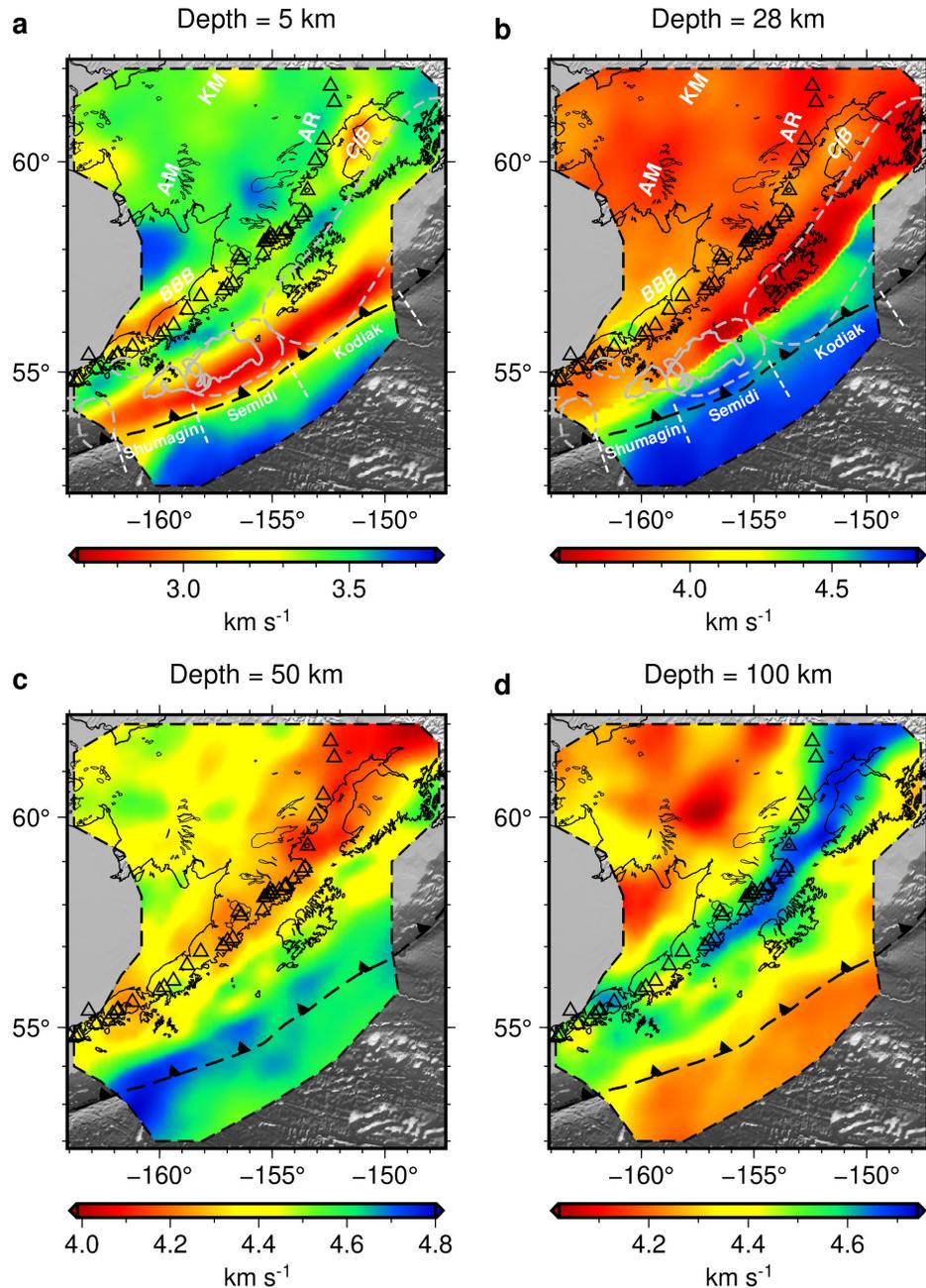
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## 565 4.2 Shear velocity structure

566 The shear wave velocity structure is presented as a series of map views (Figure 6) as well  
567 as cross-sections normal to the trench along the Shumagin, Semidi, and Kodiak segments (Figure  
568 7). The downgoing Pacific plate, featured by high velocity in the mantle, dominates the shear  
569 velocity model. The depth of the slab interface changes from less than 10 km near the trench to  
570 greater than 100 km beneath the volcanic arc, similar to the slab geometry from previous studies  
571 (e.g., Abers et al., 2017). The uppermost mantle beneath the incoming plate shows a clear  
572 velocity reduction from the seaward end to the near-trench region (Figure 7).

573 The 3-D shear velocity model successfully resolves features like the accretionary prism,  
574 forearc crust, shallow basins, arc volcanic magma, and major mountains. At very shallow depths  
575 (~5 km), the outer forearc is dominated by low-velocity sediment of the accretionary prism while  
576 most regions show a typical crystalline upper crust (Figure 6). In the Semidi and Kodiak  
577 segments, the outer forearc shows a low-velocity (~3.5 km/s) lower crust with a larger thickness,  
578 which is not found in the Shumagin segment. A very similar contrast is also seen in recent  
579 active-source imaging, where results show a lower velocity crust in the Semidi segment than that  
580 in the Shumagin segment (Burstein et al., 2022). The Bristol Bay Basin is featured by thinner  
581 crust, high-velocity lower crust, and low-velocity upper mantle. The low-velocity anomalies  
582 beneath volcanic arcs are observed for all segments but are most prominent in the Kodiak  
583 segment (Figure 7). The Ahklun Mountains and Kuskokwim Mountains show similar crustal  
584 velocities but quite different Moho depths and upper mantle structures (Figure 5; Figure 7).

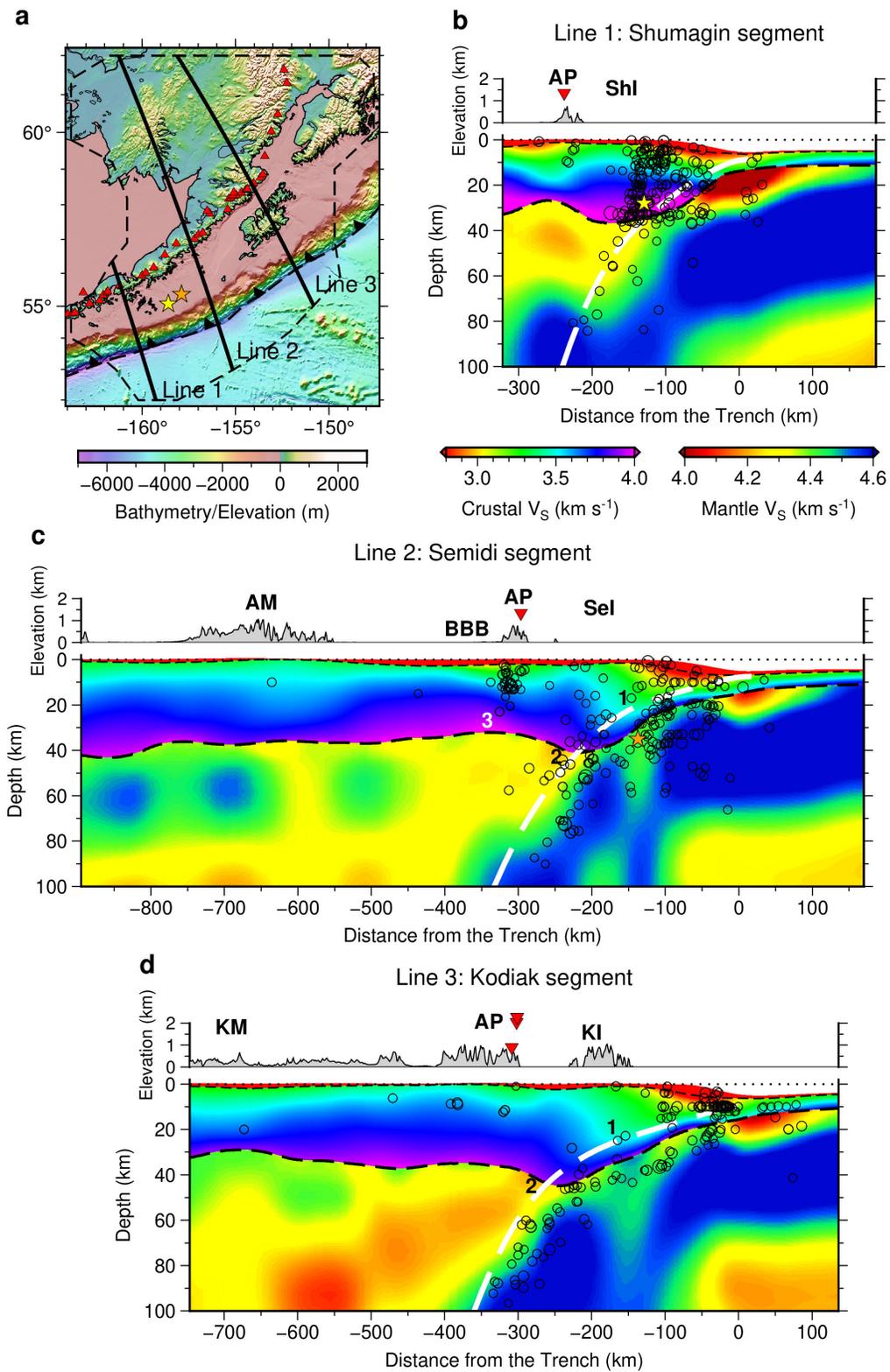
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586

587 **Figure 6.** Horizontal slices for azimuthal averaged shear wave velocity ( $V_{SV}$ ) of the study region  
 588 at different depths. The background image is the topography/bathymetry in gray scales. **(a-d)**  
 589 Map view of the shear velocity at 5, 28, 50, and 100 km depths, respectively. The depth is  
 590 defined here as relative to the solid surface, either the seafloor or the continental surface. Active  
 591 volcanoes are marked as triangles along the volcanic arc. Note that there are different velocity  
 592 color scales for each sub-figure. In (a) and (b), the dashed gray lines are the contours of  
 593 earthquake rupture zones shown in Figure 1b and the dashed white lines marked the range of  
 594 Shumagin, Semidi, and Kodiak segments. Annotation of geological features: AM = Ahklun  
 595 Mountains; KM = Kuskokwim Mountains; AR = Alaska Range; BBB = Bristol Bay Basin; CIB  
 596 = Cook Inlet Basin.

597



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599

600

**Figure 7.** Vertical cross-sections of the 3-D shear velocity model along three lines crossing the Shumagin, Semidi, and Kodiak segments, respectively. **(a)** Location of the three lines shown on

601 the bathymetry and elevation map. The arc volcanoes are plotted as red triangles. The epicenter  
 602 of the 2020 M7.8 Simeonof megathrust earthquake (~28 km depth) is marked by a yellow star.  
 603 The epicenter of the 2021 M8.2 Chignik megathrust earthquake (~35 km depth) is marked by an  
 604 orange star. The dashed black line encloses the study region. **(b-d)** Cross-sections along each  
 605 line. The x-axis is the distance relative to the trench axis with a positive distance in the direction  
 606 of the incoming Pacific plate. The elevation along the profile is plotted above the velocity image  
 607 with geological features labeled (Annotation: AP = Alaska Peninsula; AM = Ahklun Mountains;  
 608 KM = Kuskokwim Mountains; BBB = Bristol Bay Basin; ShI = Shumagin Islands; SeI = Semidi  
 609 Islands; KI = Kodiak Island). Active volcanoes near the profiles (within 25 km) are plotted as red  
 610 triangles, above the topography image. The  $V_{SV}$  images show the structure along each cross-  
 611 section, where  $V_{SV}$  values in the crust and mantle use different color bars. Depths are plotted  
 612 relative to sea level here (shown as a black dotted line). The thin black dashed line is the  
 613 boundary between sediment and crust, while the thick black dashed line is the Moho depth in the  
 614 3-D  $V_{SV}$  model. The white dashed line depicts the depth of the slab interface from Slab2.0  
 615 (Hayes et al., 2018). The  $M \geq 4$  earthquakes from the AEIC catalog since 1990 and the AACSE  
 616 catalog (Ruppert et al., 2022) are plotted on the cross-section image as open black circles if they  
 617 are within 50 km from the projection line. The hypocenter of the 2020 M7.8 Simeonof  
 618 megathrust event is projected to the cross-section along Line 1 and shown as a yellow star in (b).  
 619 The hypocenter of the 2021 M8.2 Chignik megathrust event is projected to the cross-section  
 620 along Line 2 and shown as an orange star in (c). Labels 1, 2, and 3 are for reference of some  
 621 velocity features discussed in the text.

622

## 623 5 Discussion

624 The 3-D shear velocity images clearly reveal systematic along-strike variations in the  
 625 subduction zone structure. Here we will focus on the predominant features, discussing the  
 626 forearc crustal thickness, the backarc basin structure, and the incoming plate hydration.

### 627 5.1 Thick forearc crust associated with the Chugach Terrane

628 The crustal structure of the forearc shows a band of thick crust (35-42 km) extending  
 629 from the eastern edge of the study region, beneath Kodiak Island, to just east of the Shumagin  
 630 Islands. Crustal thickness in this region generally exceeds the crustal thickness of the volcanic  
 631 arc and backarc regions (**Figure 5a**). Immediately seaward of the thickest crust, the velocity  
 632 profile is characterized by a thick section of reduced lower crustal velocities (~3.5 km s<sup>-1</sup>)  
 633 extending to the plate interface at depths of about 30 km at many locations (label 1 in **Figure 7c**  
 634 and **Figure 7d**). This feature may be the along-strike continuation of the 20-km-thick low  
 635 velocity lower crustal units imaged just to the east of Kodiak Island by the EDGE active source  
 636 study (Ye et al., 1997). The shear velocities in this study are consistent with the lowermost  
 637 crustal P wave velocity of about 6.0 km s<sup>-1</sup> observed by Ye et al. (1997). They interpreted the  
 638 low-velocity region as underplated sediments and upper crustal rocks of subducted terrains,  
 639 associated with the uplift of the Kodiak region during the Eocene to Oligocene (Moore et al.,  
 640 1991). An alternative interpretation is that those lower velocities represent the Paleogene-aged  
 641 Prince William Terrane and/or the Mesozoic-aged Chugach Terrane (e.g., Horowitz et al., 1989)  
 642 (**Figure 1c**); these terranes are dominantly composed of lightly metamorphosed accretionary  
 643 complexes (e.g., Sample & Moore, 1987), which would also be expected to have relatively low

644 velocities. In this case, the reflections observed by Moore et al. (1991) may arise from layering  
645 within Prince William/Chugach/Peninsula Terranes from accretionary complexes or intrusions  
646 (Figure 1c). Intracrustal reflections are observed on all the Alaska/Aleutian arc profiles and  
647 bright bands of lower crustal reflections are even observed farther west in the oceanic part of the  
648 arc (Calvert & McGeary, 2013). However, these studies did not show a clear anticlinal structure  
649 as described by Moore et al. (1991) and Ye et al. (1997).

650 The results presented here suggest that the distinctive thickened crust just inboard of the  
651 slow lower crustal forearc material extends to the eastern edge of the Shumagin Islands region.  
652 The Chugach Terrane is often thought to extend to Sanak Island west of the Shumagin Islands  
653 due to the similarity of the accretionary and intrusive surficial rocks (e.g., Bradley et al., 2003)  
654 (Figure 1c). However, crustal thickness becomes variable within and westward of the Shumagin  
655 Islands region, being reduced in places to about 30 km, and the lower crust has a higher velocity  
656 ( $\sim 3.9 \text{ km s}^{-1}$ ) beneath the Shumagin Islands. Thus the Shumagin islands may represent a major  
657 change in forearc morphology, defining a transition from thick accretionary crustal material and  
658 thicker forearc crust to the east to more conventional forearc crust to the west. Alternatively, the  
659 forearc crust in the Shumagin region may have been thinned and modified by deformation at the  
660 edge of the Beringian margin that led to the formation of a series of extensional basins in this  
661 region in the Oligocene-Miocene (e.g., Horowitz et al., 1989; Bécel et al., 2017; von Huene et  
662 al., 2019; Kahrizi et al., 2024).

663 The along-strike and down-dip variations in the thickness and velocity of the forearc  
664 crust above the megathrust could be important for the mechanical properties of the megathrust,  
665 with implications for strain accumulation and release (Sallarès & Ranero, 2019; Bassett et al.,  
666 2022). Although the recent 2020 M7.8 Simeonof and 2021 M8.2 Chgnik earthquakes occurred at  
667 similar depth ranges in the Shumagin and Semidi forearc regions (e.g., Liu et al., 2023), the  
668 differences in crustal thickness along strike suggest that rupture zone of the 2021 M8.2  
669 earthquake in the Semidi segment is overlain by continental crust (orange star in Figure 7c), but  
670 that part of the rupture zone of the 2020 M7.8 in the Shumagin segment may have occurred  
671 below the continental Moho (Shillington et al., 2022) (Figure 5a; yellow star in Figure 7b);  
672 complexity of megathrust properties near the continental Moho are speculated to contribute to  
673 the patchiness of the M7.8 event (Shillington et al., 2022; Liu et al., 2023). In the Semidi  
674 segment, the recent rupture in the 2021 M8.2 event is confined to depths of  $\sim 26\text{-}42$  km (Liu et  
675 al., 2023), the portion of the megathrust overlain by relatively high-velocity crust. The shallower  
676 megathrust overlain by low-velocity crust has recently been observed to host a slow slip event in  
677 2018 and aseismic afterslip following the 2021 event (Brooks et al., 2023; He et al., 2023). These  
678 correlations suggest that the overriding plate could influence megathrust slip behavior. Downdip  
679 changes in bulk rigidity or permeability of the overriding plate and/or in frictional properties on  
680 the megathrust due to the overriding plate are proposed to influence megathrust behavior in other  
681 locations (Sallarès & Ranero, 2019; Bassett et al., 2022). Finally, differences in overriding plate  
682 structure and present-day inputs to the subduction zone could also influence megathrust frictional  
683 properties and heterogeneity. Low velocities in the outer forearc of the Semidi and Kodiak  
684 segments shown here could indicate significant underplating in the past (Moore et al., 1991), and  
685 thicker sedimentary sections are subducting in these segments today than farther west (e.g., von  
686 Huene et al., 2012; J. Li et al., 2018). Extensive sediment subduction is likely to reduce the  
687 inherent roughness of the plate interface and produce a large, smooth megathrust fault zone  
688 favorable to great earthquakes (Bangs et al., 2015; Scholl et al., 2015). Global studies suggest

689 higher seismic coupling and propensity for great earthquakes in regions with substantial  
690 sediment subduction and underplating (Ruff, 1989; Heuret et al., 2012).

## 691 5.2 Volcanic arc, mantle-wedge corner, and backarc structure

692 The mantle wedge structure is characterized by low shear velocities ( $4.1\text{-}4.3\text{ km s}^{-1}$ ) in  
693 the upper mantle beneath the volcanic arc. The Kodiak and Semidi segments have adequate  
694 resolutions in the backarc and both reveal continuous low-velocity anomalies sloping upward  
695 from more than 100 km depth beneath the backarc to the Moho beneath the volcanic front  
696 (Figure 7c-d). Similar inclined low-velocity zones have been observed at many volcanic arcs  
697 around the world, and are generally interpreted as the zone of hydrous partial melting and melt  
698 transport above the slab (e.g., Zhao et al., 2007; Wiens et al., 2008; Ward & Lin, 2018; Yang &  
699 Gao, 2020). These results suggest that a significant portion of the partial melt formation in the  
700 Alaska mantle wedge occurs beneath the backarc rather than immediately beneath the volcanic  
701 arc. Melt is transported along the inclined zone by porous and channelized flow to the Moho  
702 beneath the volcanoes (C. R. Wilson et al., 2014; Cerpa et al., 2018).

703 Unlike for some arcs, identified in the compilation of Abers et al. (2017), there is no  
704 indication of a high-velocity mantle wedge seaward of the volcanic arc. Instead, low mantle  
705 velocities extend seaward from the volcanic front into the corner of the mantle wedge (label 2 in  
706 Figure 7c-d). Partial melt does not provide a good explanation for these low velocities, since  
707 melt is highly buoyant and there is no magmatism observed on the forearc side of the volcanic  
708 front. In addition, heat flow anomalies characteristic of magma supply to the crust are limited to  
709 the volcanic arc and backarc regions in most arcs (Furukawa, 1993; Rees Jones et al., 2018). The  
710 low velocities in the wedge corner instead could result from serpentinization of the mantle  
711 peridotite by water released from the slab immediately below (Hyndman & Peacock, 2003;  
712 Reynard, 2013). The reduced shear velocities in the wedge corner of  $4.1\text{-}4.3\text{ km s}^{-1}$  are  
713 compatible with P-wave velocities of  $7.3\text{-}8.0\text{ km s}^{-1}$  from P-wave tomography in the Shumagin  
714 region (Abers, 1994).

715 The amount of forearc mantle serpentinization can be estimated from the velocity  
716 reduction along Lines 1, 2, and 3 (Figure 7b-d). The low velocity at the inner forearc uppermost  
717 mantle is  $4.29 \pm 0.08\text{ km s}^{-1}$  for Line 1,  $4.21 \pm 0.10\text{ km s}^{-1}$  for Line 2, and  $4.19 \pm 0.09\text{ km s}^{-1}$  for  
718 Line 3. Because the velocities are similar to one another within the uncertainty estimates, the  
719 inferred serpentinization is comparable for the three lines and we cannot infer along-strike  
720 variations. The experimental relationship between shear velocity  $V_S$  and serpentine volume  
721 fraction ( $\Phi$ ) at 600 MPa is  $V_S = 4.51 - 2.19\Phi$  for lizardite and chrysotile, and  $V_S = 4.51 - 0.84\Phi$   
722 for antigorite (Ji et al., 2013). Experimental work suggests that serpentinization of mantle  
723 peridotite forms mostly lizardite at temperatures below  $500^\circ\text{C}$  (e.g., Nakatani & Nakamura,  
724 2016), as expected for the uppermost mantle with ages around 50 Ma (Stein & Stein, 1992;  
725 McKenzie et al., 2005). The assumption of lizardite mineralogy also results in a conservative  
726 estimate of the serpentinization percentage and water content of the mantle. Assuming an  
727 average value of the mantle wedge corner velocity of  $4.23 \pm 0.09\text{ km s}^{-1}$ , the corresponding  
728 mantle serpentinization above the slab in the wedge corner is roughly  $13 \pm 4\text{ vol}\%$ . Such a range  
729 of forearc mantle serpentinization is lower than the value of 20-35 vol% estimated by Yang and  
730 Gao (2020) along the Aleutian arc, where they observed a low velocity of  $3.7\text{-}4.1\text{ km s}^{-1}$  in the  
731 forearc mantle. Also this estimate of forearc mantle serpentinization is distinctly lower than a  
732 previous estimate by Bostock et al. (2002) for Cascadia, inferred from the velocity change across

733 the crust-mantle boundary beneath the forearc. However, it is consistent with Abers et al (2017),  
734 who concluded that forearc mantle wedges show only modest degrees of serpentinization.

735 The backarc crust shows significant along-strike variations. In the northeast part of the  
736 study region, the backarc is characterized by a relatively typical continental crust with a  
737 thickness of about 35 km. However, in southern Bristol Bay, crust with significantly reduced  
738 thickness (~28 km) is found just to the north of the Alaska Peninsula and the active volcanic arc  
739 (Figure 5a). The entire region of Bristol Bay is underlain by a ~10 km thick layer of high  
740 velocity (~4.0 km s<sup>-1</sup>) lower crust (label 3 in Figure 7c) that we interpret as mafic underplating  
741 (Thybo and Artemieva, 2013). The ~4.0 km s<sup>-1</sup> shear velocity is consistent with the typical V<sub>P</sub>  
742 observed for mafic underplating given that lower crustal mafic rocks have high V<sub>P</sub>/V<sub>S</sub> ratios  
743 (Thybo & Artemieva, 2013). The high-velocity lower crust indicates a dense mafic composition,  
744 resulting in a crust with a higher average density. Negative buoyancy caused by this denser crust  
745 as well as the reduced crustal thickness result in lower elevations in Bristol Bay relative to  
746 backarc regions farther to the northeast.

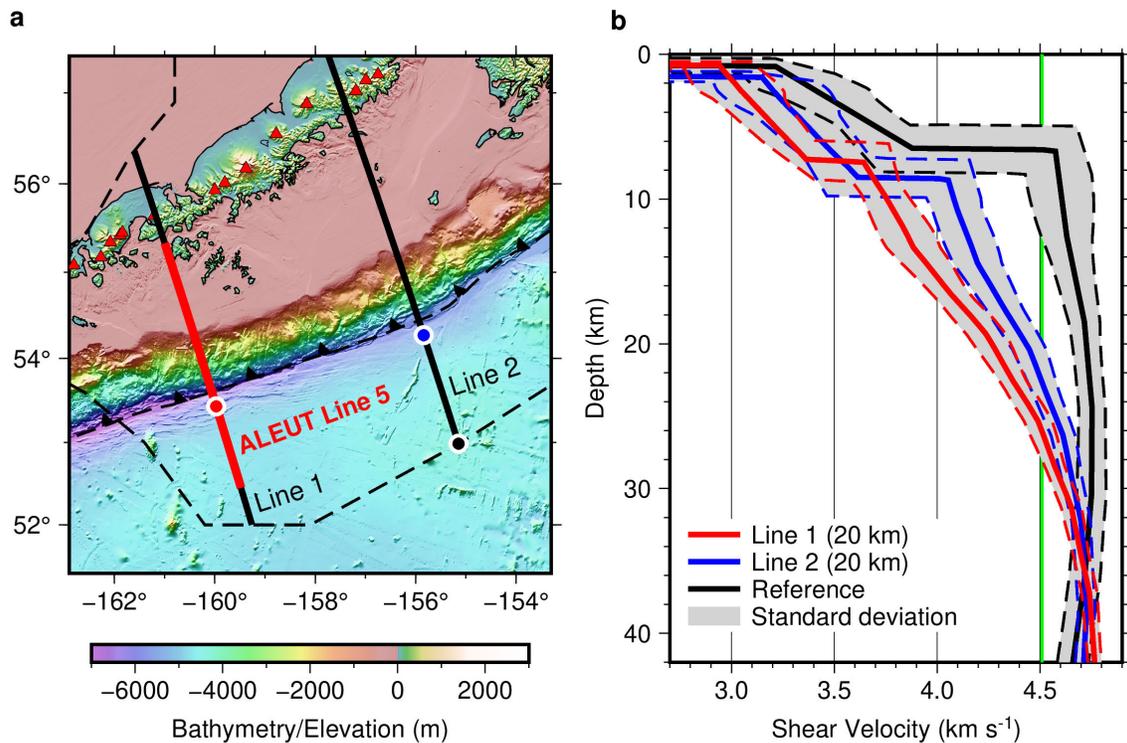
747 Active-source surveys have shown a significant sedimentary sequence in the Bristol Bay  
748 Basin (Marlow et al., 1994; Walker et al., 2003). Bond et al. (1988) suggested that the basin  
749 formed primarily by flexural subsidence caused by the Oligocene to present crustal thickening of  
750 the Alaska Peninsula. In contrast, Walker et al. (2003) proposed that the basin was initially  
751 formed through tectonic extension. In the Walker et al. (2003) model, an early or middle Eocene  
752 through late Miocene phase of extension led to fault-controlled subsidence, then a late Eocene  
753 through Holocene phase of volcanic-arc loading or northward prograding delta led to flexural  
754 subsidence. Our observation of the thinner crust with a dense high-velocity mafic lower crust  
755 beneath the Bristol Bay basin supports Walker et al. (2003) model, with the dense lower crustal  
756 underplate emplaced during tectonic extension and associated magmatism.

### 757 5.3 Incoming plate hydration

758 The incoming plate structure clearly shows a low-velocity zone at the top of the  
759 subducting oceanic mantle (Figure 7). The velocities decrease towards the trench and show  
760 significant along-strike variations. Low-velocity zones at the top of the incoming plate mantle  
761 have been observed at many other subduction zones and are generally attributed to the  
762 serpentinization of mantle peridotite (Ivandic et al., 2008; Van Avendonk et al., 2011;  
763 Shillington et al., 2015; Cai et al., 2018; Mark et al., 2023) and/or to the effects of water in plate-  
764 bending faults (Miller & Lizarralde, 2016; Korenaga, 2017). The Shumagin segment shows a  
765 distinct low-velocity zone (~3.65 km s<sup>-1</sup>) at the top of the incoming plate mantle, suggesting  
766 strong hydration if the velocity reduction is due to serpentinization. In contrast, the Semidi  
767 segment and Kodiak segment show much weaker hydration of the incoming Pacific plate, with a  
768 velocity reduction only to ~4.05 km s<sup>-1</sup> and ~4.0 km s<sup>-1</sup>, respectively.

769 The extent of incoming plate hydration can be better compared using velocity profiles at  
770 locations near the trench axis (Figure 8). Since the resolution of the incoming plate in the Kodiak  
771 segment (Line 3; Figure 7d) is limited by the small number of nearby OBSs returning data  
772 (Figure 2), we only make the comparison between the Shumagin segment (Line 3; Figure 7b)  
773 and the Semidi segment (Line 3; Figure 7c). We choose the trench profiles of both Shumagin and  
774 Semidi segments at locations 20 km seaward from the trench axis to limit the smoothing effect of  
775 surface waves. To evaluate the magnitude of mantle hydration, we also need a reference profile

776 that represents an unaltered oceanic plate structure. The seaward end points of those projection  
 777 lines are the possible candidates. The resolution at the seaward end of Line 2 is the best among  
 778 all segments and at a significant distance away from the trench axis, thus we choose the seaward  
 779 end of Line 2 to be a reference profile giving the velocity structure of the unaltered incoming  
 780 plate in the region (Figure 8a). By comparing the shear velocity profiles, we observe that the  
 781 magnitude of velocity reduction and thus incoming plate hydration is stronger in the Shumagin  
 782 segment than that in the Semidi segment, consistent with the active source results from  
 783 Shillington et al (2015). Although the reference velocity profile also shows a small velocity  
 784 reduction ( $\sim 0.1 \text{ km s}^{-1}$ ) atop the mantle, tests in which a series of synthetic dispersion curves  
 785 were inverted using the same parameterization suggest that the magnitude of such a small  
 786 velocity reduction was not well constrained by the dispersion data. Besides, we note that the  
 787 active source profile in this region also shows a significant increase in P-wave velocity with  
 788 depth beneath the mocho farther from the trench, consistent with the reference profile.  
 789



790

791 **Figure 8.** Velocity profiles on the incoming plate show the upper mantle hydration of the  
 792 Shumagin segment and the Semidi segment. (a) The locations of velocity profiles on cross  
 793 sections. The red circle on Line 1 and blue circle on Line 2 are chosen 20 km seaward away from  
 794 the trench axis so that they represent the hydration status at the trench and minimize the spatial  
 795 smoothing of surface waves. The black circle at the seaward end of Line 2 gives the location of  
 796 the velocity profile representing the unaltered oceanic plate structure. ALEUT Line 5, shown as  
 797 the red line, is part of Line 1 (Shillington et al., 2022). (b) The 1-D shear velocity profiles of the  
 798 reference, Line 1 near the trench, and Line 2 near the trench. Here the profile near the trench is  
 799 chosen at 20 km seaward from the trench axis. The uncertainty contours of each are shown as

800 gray zones. The experimental velocity value of unaltered upper mantle peridotite,  $4.51 \text{ km s}^{-1}$ , is  
801 marked as a green line. The depth is relative to the seafloor.

802

803 Assuming the velocity reduction atop the mantle is purely due to serpentinization, we  
804 could use the shear velocity reduction to constrain the hydration. We still follow the method  
805 described in Section 5.2 to estimate the serpentinization (McKenzie et al., 2005; Ji et al., 2013;  
806 Nakatani & Nakamura, 2016). Although the reference velocity profile shows uppermost mantle  
807 velocities ranging from  $4.55 \text{ km s}^{-1}$  to  $4.7 \text{ km s}^{-1}$ , here we use the experimental value of  $4.51 \text{ km}$   
808  $\text{s}^{-1}$  as the reference velocity of unaltered upper mantle peridotite, which provides a further  
809 conservative estimate of the degree of serpentinization and facilitates comparison with other  
810 studies. In the Shumagin segment, the shear velocity reduction is  $0.87 \pm 0.12 \text{ km s}^{-1}$  immediately  
811 below the Moho (Figure 8b). The velocity reduction then becomes smaller with depth until there  
812 is no velocity reduction at  $18 \pm 3 \text{ km}$  below the Moho. The corresponding hydration is roughly  
813 equivalent to a  $40 \pm 6 \text{ vol}\%$  serpentinization at the top of the mantle, reducing to no  
814 serpentinization at  $18 \pm 3 \text{ km}$  below the Moho.

815 Similar calculations can be made for the Semidi segment. Using  $4.51 \text{ km s}^{-1}$  as the  
816 reference value, a  $0.46 \pm 0.13 \text{ km s}^{-1}$  shear velocity reduction is observed right below the Moho,  
817 decreasing to no velocity reduction at  $14 \pm 3 \text{ km}$  below the Moho. This gives an estimate of  $21 \pm$   
818  $6 \text{ vol}\%$  serpentinization at the top of the mantle, decreasing to zero at  $14 \pm 3 \text{ km}$  below the  
819 Moho.

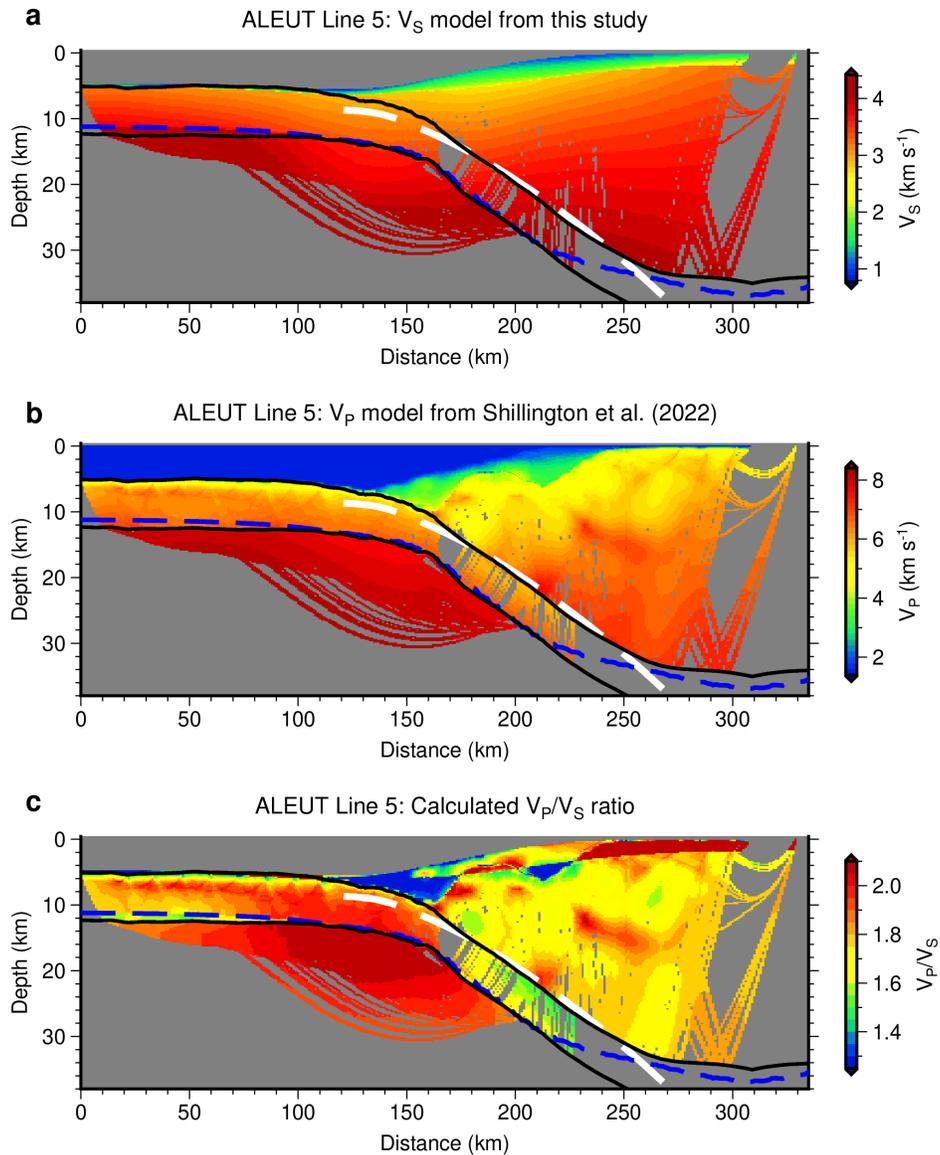
820 The maximum degree of serpentinization, as well as the total thickness of the  
821 serpentinized layer, is larger in the Shumagin segment compared to that in the Semidi segment,  
822 consistent with previous active source results (Shillington et al., 2015). Using the shear velocity  
823 reduction at the uppermost mantle, we find that the serpentinization in the Shumagin segment is  
824 approximately two times greater than that in the Semidi segment. Carried by the hydrous  
825 minerals, more water is expected to input into the deep Earth through the Shumagin segment in  
826 the Alaska subduction zone.

827 The distribution of mantle velocity reduction is similar to the distribution of seismicity  
828 located by AACSE ocean bottom seismographs (Matulka & Wiens, 2022) as well as mapped  
829 fault scarps in seafloor bathymetry (Clarke, 2022) and seismic reflection images (Shillington et  
830 al., 2015), consistent with the idea that velocity reduction is caused by hydration from plate-  
831 bending faults penetrating into the upper mantle. A recent magnetotelluric study along the  
832 Shumagin segment suggests a source of fluids at depth of 15-25 km beneath the Moho in the  
833 forearc that they interpreted as due to dehydration of serpentinized mantle (Cordell et al., 2023).  
834 The Shumagin section shows numerous plate bending faults and has a high seismicity rate,  
835 compared to the near-absence of seafloor faults and a lower seismicity rate in the Semedi  
836 segment. The depth extent of inferred serpentinization along the Shumagin segment coincides  
837 with the depth range of normal faulting earthquakes along plate bending faults (Matulka &  
838 Wiens, 2022). The maximum incoming plate seismicity rate occurs 5-10 km below the Moho,  
839 and earthquakes are largely limited to depths less than 15-20 km below the Moho, coinciding  
840 with the lower limit of serpentinization from this study. The depth range of seismicity in the  
841 Semidi segment is similar, but with a much lower seismicity rate.

842 It is worthwhile to compare the Alaska Trench results with the central Mariana Trench  
843 (Cai et al., 2018) and the southern Mariana Trench (Zhu et al., 2021). Since the studies use

844 similar techniques, we can directly compare the  $V_S$  profiles. The near-trench regions of the  
845 central Mariana Trench show uppermost mantle  $V_S$  reduced to  $\sim 3.5 \text{ km s}^{-1}$ , lower than the  $\sim 3.65$   
846  $\text{km s}^{-1}$  observed in the Shumagin segment. In addition, the lowered seismic velocities in Mariana  
847 extend to about 24 km below the Moho, compared to only 18 km below the Moho in the  
848 Shumagin segment. Similar, but less well-constrained velocities and depths were found by Zhu  
849 et al. (2021) in the southern Mariana Trench. Cai et al. (2018) interpreted the extremely low  
850 incoming plate mantle velocities at the trench as partly due to pore water in the bending faults,  
851 and used a velocity of  $4.1 \text{ km s}^{-1}$  found beneath the forearc after pore water would have been  
852 expelled for calculating the degree of serpentinization. The lower velocities and greater depth  
853 extent of the mantle velocity reduction suggest a larger percent and extent of mantle  
854 serpentinization for Mariana than in the Shumagin or Semidi segments of the Alaska Subduction  
855 Zone. The difference in the hydration of the incoming Alaska and Mariana plates is largely due  
856 to the differences in their oceanic plate age (Alaska  $\sim 50\text{-}55 \text{ Ma}$ ; Mariana  $\sim 150 \text{ Ma}$ ). The thicker,  
857 colder lithosphere at the Mariana Trench results in a deeper neutral plane and greater extensional  
858 strain above the neutral plane. Another factor is the distribution of outer rise plate-bending faults  
859 (abundant in the Mariana and the Shumagin segment of Alaska; fewer in the Semidi segment of  
860 Alaska). Finally, the overall geological setting of the Mariana subduction zone is more  
861 extensional, further enhancing shallow extensional faulting and deepening the neutral plane  
862 (Emry et al., 2014; Eimer et al., 2020).

863



864

865 **Figure 9.** Comparison of the  $V_S$  model from this study, the  $V_P$  model from Shillington et al.  
 866 (2022), and the calculated  $V_P/V_S$  ratio using two models. The white dashed line shows the depth  
 867 of slab interface in the Slab2.0 model (Hayes et al., 2018). The two solid black lines show  
 868 boundaries of the  $V_P$  model that are constrained by reflections (Shillington et al., 2022), which  
 869 mark either the ocean bottom, the interface between the subducting plate and the overriding  
 870 plate, Moho depth of the overriding plate, or Moho depth of the subducting plate. The dashed  
 871 blue line shows the Moho depth of the  $V_S$  model constrained by this study. **(a)** The shear velocity  
 872 ( $V_S$ ) model from this study, which is just a cross-section of the 3-D shear velocity model along  
 873 the ALEUT Line 5. **(b)** The P-wave velocity ( $V_P$ ) model from Shillington et al. (2022), which is  
 874 a 2-D model of the ALEUT Line 5 determined by joint refraction and reflection 2-D  
 875 tomographic inversion. **(c)** The calculated  $V_P/V_S$  ratio using two models along ALEUT Line 5.  
 876 The colorbar is limited to show the 5- to 95-percentile range of all  $V_P/V_S$  values.

877

## 878 5.4 Comparison with previous active source results

879 The incoming plate hydration has been previously examined using active-source data  
 880 (Shillington et al., 2015). In the Shumagin segment, the P-wave velocity of the upper mantle is  
 881 reduced from  $8.2 \text{ km s}^{-1}$  to  $7.5 \text{ km s}^{-1}$ . In the Semidi segment, the P-wave velocities range  
 882 between  $8.0 \text{ km s}^{-1}$  to  $7.7 \text{ km s}^{-1}$ , but do not show a systematic trenchward decrease; these  
 883 variations may result from heterogeneity in intermediate spreading crust (Shillington et al.,  
 884 2015). The experimental relationship between  $V_P$  and serpentine volume fraction ( $\Phi$ ) at 600 MPa  
 885 is  $V_P = 8.10 - 3.00\Phi$  for lizardite and chrysotile (Ji et al., 2013). For the Shumagin segment, the  
 886 lowest  $V_S$  of  $3.64 \text{ km s}^{-1}$  is equivalent to  $\sim 40 \text{ vol\%}$  serpentinization, whereas the lowest  $V_P = 7.5$   
 887  $\text{ km s}^{-1}$  is equivalent to  $\sim 20 \text{ vol\%}$  serpentinization. Clearly, the serpentinization immediately  
 888 beneath the Moho estimated from shear velocity reduction tends to be higher than estimates from  
 889 P-wave velocities.

890 To investigate the differences in hydration estimated from  $V_P$  and  $V_S$ , we compare the  
 891 shear velocity model with the active-source P-wave model along the ALEUT Line 5 (Shillington  
 892 et al., 2022) (Figure 9). The  $V_P$  and  $V_S$  structures along the same projection line show similar  
 893 features, though the active source P-wave model exhibits more details and the shear velocity  
 894 model shows the smoothing effect of surface waves (Figure 9a; Figure 9b). The  $V_P/V_S$  ratio  
 895 (Figure 9c) calculated from the two models shows some small features with extreme values due  
 896 to the higher resolution and different settings of the slab in the active source P-wave model, but  
 897 are in general consistent with expected ratios for oceanic and forearc crust. For example, most of  
 898 the forearc crust has  $V_P/V_S$  ratios between 1.65 and 1.85, which is typical for the continental  
 899 crust (N. I. Christensen, 1996).

900 The crust of the incoming plate shows a distinct region of high  $V_P/V_S$  ratio in the plate  
 901 bending region near the trench. Previous active source studies show large  $V_S$  reductions and  
 902  $V_P/V_S$  ratio increases in the crust of plate-bending regions of various subduction zones (Fujie et  
 903 al., 2013; Fujie et al., 2018; Grevemeyer et al., 2018). This is generally interpreted as due to the  
 904 hydration of crustal rocks as well as the additional effect of water in joints and cracks. The high  
 905  $V_P/V_S$  ratio of the crust of the incoming plate is generally greater than 1.9 for both Kuril Trench  
 906 and Japan Trench (Fujie et al., 2018), also quite similar to what we observe here (Figure 9c).

907 The incoming plate mantle shows an extremely high  $V_P/V_S$  ratio of greater than 2.05 near  
 908 the trench axis. The experimental relationship between  $V_P/V_S$  ratio and serpentine volume  
 909 fraction ( $\Phi$ ) at 600 MPa is  $V_P/V_S = 1.77 + 0.38\Phi$  for lizardite and chrysotile, and  $V_P/V_S = 1.77 +$   
 910  $0.04\Phi$  for antigorite (Ji et al., 2013). Using the  $V_P/V_S = 1.77 + 0.38\Phi$  relationship and  $V_P/V_S =$   
 911 2.05, we can estimate the serpentinization from  $V_P/V_S$  for the uppermost mantle in the Shumagin  
 912 segment as 73 vol%. This value, of course, is unrealistic but suggests that the serpentinization  
 913 implied by the  $V_S$  reduction,  $V_P$  reduction, and  $V_P/V_S$  increase are inconsistent.

914 The discrepancy between estimates of serpentinization from  $V_P$ ,  $V_S$ , and the  $V_P/V_S$  ratio  
 915 may result from the effect of water in joints and cracks. Poroelastic calculations by Takei (2002)  
 916 show that for water-filled cracks with large aspect ratios, as expected in partially serpentinized  
 917 peridotite, the fractional velocity reduction in  $V_S$  is significantly larger than the fractional  
 918 reduction in  $V_P$ . Korenaga et al. (2017) also showed that modest porosity in crack-like pore  
 919 spaces with large aspect ratios lowers  $V_S$  more significantly than  $V_P$  and increases the  $V_P/V_S$   
 920 ratio. Cai et al. (2018) attributed part of the large  $V_S$  reduction in the Mariana outer-rise mantle  
 921 to water in cracks and joints. Mark et al. (2023) found evidence from seismic anisotropy for

922 water in crack-like pores in the upper 1 km of the Mariana outer rise mantle using active-source  
 923 data. If the water in crack-like porosity exists in the mantle, the percent serpentinization  
 924 determined by  $V_S$  values and the  $V_P/V_S$  ratios will be overestimated.

925 In the following discussion, we assume that the percent serpentinization of the mantle  
 926 immediately below the Moho is better estimated by the  $V_P$  reduction determined by active source  
 927 data (Shillington et al., 2015), since we observe that the water porosity has a more limited  
 928 influence on  $V_P$ . However, the maximum depth of serpentinization is determined by this study  
 929 due to the limited depth penetration of the active source results. This discussion assumes that  
 930 percent serpentinization can be estimated using formulas for bulk serpentinization, as has  
 931 traditionally been done in previous studies (e.g., Grevenmeyer et al., 2018). The actual situation  
 932 may be more complex, as the serpentinization may be localized in narrow regions surrounding  
 933 discrete faults (Hatakeyama et al., 2017). In this case, there will be frequency-dependent wave  
 934 propagation through the mantle at the frequencies used in active source studies (Miller &  
 935 Lizarralde, 2016; Miller et al., 2021; Mark et al., 2023). Estimates of serpentinization taking this  
 936 effect into account generally result in smaller percentages of serpentinization, but require  
 937 analysis of azimuthal anisotropy, which is not available in this case. Therefore we will use the  
 938 serpentinization estimates based on bulk serpentinization given by Shillington et al. (2015).

### 939 5.5 Quantitative estimates of subducted water

940 The amount of bound water carried into the Alaska subduction zone by the subducting  
 941 mantle can be assessed, given estimates of the percentage serpentinization as a function of depth  
 942 on the incoming plate, since both lizardite and antigorite contain 13% water by weight. The  
 943 water content of the mantle by weight is calculated from

$$944 \quad w_h = w_s \alpha_s \rho_s / \rho_m \quad (2)$$

945 where  $w_s$  is the weight fraction of water in serpentine,  $\alpha_s$  is the volume fraction of serpentine in  
 946 the mantle determined from seismic measurements, and  $\rho_s$  and  $\rho_m$  are the densities of serpentine  
 947 and the mantle, respectively (Carlson & Miller, 2003). Here we assume that serpentinization is  
 948 maximum at the Moho, where the percent serpentinization is determined from the  $V_P$  velocity  
 949 reduction, and decreases linearly to the maximum depth of serpentinization determined from this  
 950 study. We do not include any liquid water in pore spaces, since this water will be eliminated with  
 951 increasing pressure (David et al., 1994) and will not be subducted to significant depths. We also  
 952 do not explicitly include possible hydrous minerals other than serpentine, such as chlorite and  
 953 brucite, but note that these other hydrous minerals will also lower the seismic velocity in a  
 954 similar way to serpentine. Experimental evidence indicates that the dominant hydrous mineral in  
 955 the incoming plate mantle is likely to be lizardite serpentine (Okamoto et al., 2011).

956 For the Shumagin segment, the  $V_P$  reduction from Shillington et al (2015) gives 20 vol%  
 957 serpentinization at the Moho, decreasing to zero at 18 km below the Moho. The total water  
 958 content of the hydrated mantle at the Shumagin segment is then equivalent to an 18 km thick,  
 959 partially serpentinized (10 vol% serpentine, thus 1.0 wt% water) slab mantle layer. Applying the  
 960 convergence rate of 66 mm yr<sup>-1</sup> (DeMets et al., 2010), the amount of mantle water input into the  
 961 Shumagin segment is 37 Tg Myr<sup>-1</sup> m<sup>-1</sup>.

962 A similar calculation for the Samedí segment is more uncertain because the evidence of  
 963  $V_P$  reduction from hydration is less clear in the active-source data.  $V_P$  is apparently reduced to  
 964 7.7 km s<sup>-1</sup>, but it is unclear whether this is due to hydration or to variability associated with

965 intermediate spreading crust. The  $V_S$  reduction observed in this study suggests the reduction is  
966 likely due to hydration, in which case we can calculate a serpentinization of 13 vol% from the  $V_P$   
967 reduction using the relationships in Ji et al. (2013). Assuming that the serpentinization decreases  
968 linearly from 13 vol% at the Moho to zero at a depth of 14 km below the Moho, this is  
969 equivalent to a 14 km thick, partially serpentinized (6.5 vol% serpentine, thus 0.6 wt% water)  
970 hydrated mantle layer. With the convergence rate of 63 mm yr<sup>-1</sup> (DeMets et al., 2010), this  
971 provides an estimate of 17 Tg Myr<sup>-1</sup> m<sup>-1</sup> for the flux of mantle water into the Semidi segment.  
972 This indicates that the subducting mantle carries more than twice as much water into the  
973 Shumagin segment compared to the Semidi segment.

974 These estimates necessarily involve a number of assumptions and are thus only very  
975 approximate, but are improvements on previous estimates that made ad-hoc assumptions about  
976 the hydration of the subducting mantle (e.g., van Keken et al., 2011), which had no constraint on  
977 the depth extent of the serpentinized layer. The largest uncertainty in these estimates is  
978 associated with the volume percent of serpentinization, due to the uncertainty of interpreting the  
979 discrepant estimates from  $V_P$ ,  $V_S$ , and  $V_P/V_S$ , as well as the possible effects of liquid water in  
980 crack-like porosity (Korenaga et al., 2017) and anisotropy (Miller & Lizarralde, 2016; Mark et  
981 al., 2023). All of the assumptions made in our estimations are conservative and thus result in a  
982 minimum estimate of subducting water in each segment.

983 These new estimates of subducted mantle water can be combined with previous estimates  
984 of water subducted in the crust and sediments to estimate the total water flux. van Keken et al.  
985 (2011) did not divide the segments, but estimated that 18 Tg Myr<sup>-1</sup> m<sup>-1</sup> subducts in the crust and  
986 sediments into the Alaska subduction zone offshore the Alaska Peninsula. Adding this to the  
987 mantle estimates gives total subducted water estimates of 55 Tg Myr<sup>-1</sup> m<sup>-1</sup> for the Shumagin  
988 segment and 35 Tg Myr<sup>-1</sup> m<sup>-1</sup> for the Semidi segment. Because the degree of mantle hydration  
989 was nearly unconstrained, van Keken et al. (2011) calculated three scenarios for mantle  
990 hydration. These estimates were 18 Tg Myr<sup>-1</sup> m<sup>-1</sup> for no hydration, 26 Tg Myr<sup>-1</sup> m<sup>-1</sup> for 2 wt%  
991 water in a 2 km thick mantle layer beneath the Moho, and 53 Tg Myr<sup>-1</sup> m<sup>-1</sup> for full  
992 serpentinization of a 2 km thick layer. The new estimates exceed the intermediate scenario for  
993 both the Shumagin and Semidi segments, and the new estimate for Shumagin is almost identical  
994 to the full serpentinization scenario of van Keken et al. (2011).

995 The water flux estimates for both Shumagin and Semidi segments are much less than the  
996 94 Tg Myr<sup>-1</sup> m<sup>-1</sup> estimated for the total water flux at the Mariana Trench (Cai et al., 2018). This  
997 difference results partly from the greater inferred percent serpentinization and the greater depth  
998 extent of serpentinization for Mariana. The greater depth extent, as indicated by both the velocity  
999 structure and the greater depth of plate bending earthquakes for the Mariana incoming plate  
1000 (Eimer et al., 2020), may result at least in part from the greater age, and thus greater thickness, of  
1001 the Mariana lithosphere. An older plate has a colder thermal condition and the serpentine could  
1002 be stable to a greater depth. Antigorite is the main stable phase of serpentine at high  
1003 temperatures, up to ~630°C at 1 GPa (Reynard, 2013; Schwartz et al., 2013). From the recent  
1004 plate cooling model (Richards et al., 2018), the thermal condition limit of 600°C is 25 km below  
1005 the seafloor for a 50 Ma plate, and 45 km below the seafloor for a 150 Ma plate. Moreover, the  
1006 neutral plane is deeper for older lithosphere and produces a mechanism that could cause deeper  
1007 stable depth (e.g., Sandiford & Craig, 2023). In addition, the overall extensional stress field of  
1008 the Mariana arc may be a contributing factor; the slab in the Marianas is dipping more steeply  
1009 than in the Alaska subduction zone (Nishikawa & Ide, 2015; Hayes et al., 2018).

1010 Comparisons of the mantle water flux estimates for Shumagin and Semidi segments and  
1011 the central Mariana subduction zone suggest that hydration of the uppermost mantle at  
1012 subduction zones is highly variable, not only for different subduction zones, but also for different  
1013 segments of the same subduction zone. In Section 5.3, we have discussed the strong correlation  
1014 between the distribution of mantle hydration, seismicity, and outer rise faults (Clarke, 2022;  
1015 Matulka & Wiens, 2022), and that the depth extent of mantle hydration coincides with the depth  
1016 range of normal faulting earthquakes along plate bending faults (Matulka & Wiens, 2022). The  
1017 along-strike variation of mantle hydration in Alaska and Mariana is highly correlated with  
1018 seismicity and earthquake ruptures, where strong mantle hydration leads to an abundance of  
1019 small earthquakes and the absence of large megathrust earthquakes. Thus, along-strike changes  
1020 in hydration can have major effects on intermediate depth and shallow thrust zone seismicity  
1021 (Shillington et al., 2015; Wei et al., 2021; F. Wang et al., 2022).

1022

## 1023 **6 Conclusions**

1024 We determine a 3-D isotropic shear velocity model of the Alaska subduction zone from a  
1025 Bayesian Monte Carlo inversion of Rayleigh wave dispersion data using OBS and land station  
1026 data acquired by the AACSE project and other nearby land networks. A joint inversion including  
1027 P-wave receiver functions is carried out for land seismic stations.

1028 The 3-D model shows major along-strike changes in structure. The forearc structure,  
1029 including Kodiak Island, appears to have a relatively thick crust (35-42 km) and reduced lower  
1030 crustal velocities ( $\sim 3.5 \text{ km s}^{-1}$ ) from the Kodiak segment to the eastern edge of the Shumagin  
1031 segment. The eastern portion with distinctive thickened crust is just inboard of the slow lower  
1032 crustal material extended from the Chugach Terrane. The crustal thickness becomes variable  
1033 westward of the Shumagin Islands, suggesting that the Shumagin Islands may represent a major  
1034 change in forearc morphology. The continuous low-velocity anomalies observed in the mantle  
1035 wedge likely represent the hydrous partial melting and melt transport above the slab. The low  
1036 mantle velocities that extend seaward of the volcanic front into the mantle wedge corner,  
1037 however, are likely due to approximately 13 vol% serpentinization of the mantle peridotite by  
1038 water released from the slab immediately below. As for backarc structure, most regions in the  
1039 northeast are characterized by a relatively typical continental crust. The Bristol Bay Basin,  
1040 however, shows a significantly reduced crustal thickness and a high-velocity lower crust,  
1041 indicating a dense mafic composition emplaced during the tectonic extension process that formed  
1042 the basin.

1043 The incoming plate structure shows a low-velocity zone at the top of the subducting  
1044 oceanic mantle, which results from the serpentinization of mantle peridotite due to water  
1045 penetrating into the mantle through outer-rise plate-bending faults. Velocity reduction is greater  
1046 and the thickness of the low-velocity region is larger in the Shumagin segment compared to the  
1047 Semidi segment. Estimates of serpentinization percentage from  $V_S$  reductions and  $V_P/V_S$  ratios  
1048 are larger than that estimated from  $V_P$  reduction in Shillington et al. (2015), suggesting that  $V_S$   
1049 may be strongly affected by liquid water in crack-like pores. Therefore we estimate the  
1050 serpentinization percentage from the previous  $V_P$  results, but use the  $V_S$  results to constrain the  
1051 thickness of the hydrated region. The amount of mantle water input into the strongly hydrated  
1052 Shumagin segment is about  $37 \text{ Tg Myr}^{-1} \text{ m}^{-1}$ , while the amount of mantle water input into the

1053 Semidi segment is about  $17 \text{ Tg Myr}^{-1} \text{ m}^{-1}$ . Thus the amount of mantle water input into the  
1054 Shumagin segment is more than twice the mantle water flux into the Semidi segment. However,  
1055 the amount of water input in both sections is much less than previously estimated for the Mariana  
1056 incoming plate using similar methods. Water input into subduction zones bound as hydrous  
1057 minerals in the mantle is highly variable, both between different subduction zones as well as  
1058 between different segments of the same subduction zone.

1059

1060

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1075

## 1076 **Open Research**

1077 All seismic data were downloaded through the EarthScope Consortium Web Services  
1078 (<https://service.iris.edu/>), including the following seismic networks: (1) XO (2018-2019)  
1079 (AACSE) (Abers et al., 2018); (2) TA (USArray Transportable Array) (IRIS Transportable  
1080 Array, 2003); (3) AK (Alaska Earthquake Center, 1987); (4) II (GSN – IRIS/IDA) (Scripps  
1081 Institution of Oceanography, 1986); (5) AT (NTWC Alaska) (NOAA, 1967); (6) GM (U. S.  
1082 Geological Survey, 2016); (7) YG (2016-2018) (WVLF) (D. H. Christensen & Abers, 2016); (8)  
1083 AV (Alaska Volcano Observatory, 1988). The ambient noise processing code, FTAN analysis  
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1085 (<http://ciei.colorado.edu/Products/>). The ASWMS package is from IRIS DMC Data Services  
1086 Products (<https://ds.iris.edu/ds/products/aswms/>). We also used open-source software including  
1087 ObsPy (Beyreuther et al., 2010) and GMT (Wessel et al., 2019) for data analysis and  
1088 visualization.

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