

2014

Upper Mantle Structure of the Cascades from Full-Wave Ambient Noise Tomography: Evidence for 3D Mantle Upwelling in the Back-Arc

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Citation/Publisher Attribution

Gao, H. & Shen, Y. (2014). Upper mantle structure of the Cascades from full-wave ambient noise tomography: Evidence for 3D mantle upwelling in the back-arc. *Earth and Planetary Science Letters*, 390(15), 222-233. <http://dx.doi.org/10.1016/j.epsl.2014.01.012>

Available at: <http://dx.doi.org/10.1016/j.epsl.2014.01.012>

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1 Upper mantle structure of the Cascades from full-wave ambient noise tomography:
2 Evidence for 3D mantle upwelling in the back-arc

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12

13 ABSTRACT

14 Melt generation and volcanism at subduction zones may result from several possible
15 processes: hydration of the mantle wedge by fluid released from the slab, subduction-
16 induced mantle upwelling beneath the back-arc, and heating of downgoing
17 sediments/oceanic crust atop the slab. Each process predicts a distinctly different spatial
18 pattern of melt generation and can thus be distinguished with high-resolution seismic
19 imaging. Here we construct an upper mantle model of the Pacific Northwest using a full-
20 wave ambient noise tomographic method. Normalized vertical components of continuous
21 seismic records at station pairs are cross-correlated to extract empirical Green's functions
22 at periods of 7-200 s. We simulate wave propagation within the 3D Earth structure using
23 a finite-difference method and calculate sensitivity kernels of Rayleigh waves to

24 perturbations of V_p and V_s based on the Strain Green's Tensor database. Phase delays
25 are extracted by cross-correlating the observed and synthetic waveforms at multiple
26 frequency bands.

27 Our tomographic result reveals three separate low shear-wave velocity anomalies
28 along the back-arc in the upper mantle ~200 km east of the Cascade volcanic arc, with the
29 central one being the largest in size and lowest in velocity. These back-arc low-velocity
30 anomalies are spatially correlated with the three arc-volcano clusters. The geometry of
31 the low-velocity volumes relative to the slab and arc is consistent with the pattern of
32 subduction-induced decompressional melting in the back-arc. Their along-strike variation
33 suggests that the large-scale plate-motion-induced flow in the back-arc mantle wedge is
34 modulated by small-scale convection, resulting in a highly 3D process that defines the
35 segmentation of volcanism along the Cascade arc.

36

37 Keywords: full-wave ambient noise tomography, the Cascadia subduction zone, low-
38 velocity anomaly, decompressional melting

39

40

41 1. Introduction

42 The mechanisms of melt generation in the upper mantle wedge (Figure 1) have been
43 the focus of numerous studies, as they are fundamental to our understanding of arc
44 volcanism along subduction zones. In general, melt production is positively correlated to
45 water content in arc basalt (Kelley et al., 2006), which supports flux melting by fluid
46 released from the subducting slab (van Keken, 2003). Subduction-induced mantle

47 upwelling and decompressional melting explains existence of nearly anhydrous lavas, as
48 well as low seismic velocities and high attenuation at the back-arc spreading center
49 (Conder et al., 2002; Wiens et al., 2008). Melting of the oceanic crust/sedimentary layer
50 atop the slab also contributes significantly to the trace element signatures at arc volcanoes
51 and the thermal structure of the mantle wedge (Conder, 2005). The melting processes and
52 their relative importance at various subduction zones may depend on the slab age,
53 sediment thickness, subduction rate, and other factors.

54 Melt generation also varies along strike. At the Honshu subduction zone, northeast
55 Japan, body-wave tomography reveals strong spatial correlation between the along-strike
56 segments of arc volcanoes and the low-velocity anomalies at the back-arc in the upper
57 mantle (named as ‘hot fingers’ by Tamura et al. (2002)). Numerical experiments with a
58 low-viscosity mantle wedge, presumably due to water released from the slab by
59 dehydration reaction, produce the finger-like small-scale convection (Honda and Yoshida,
60 2005). To our knowledge, the Honshu subduction zone is the only place where this
61 distinctive hot-finger structure has been imaged and correlated with arc volcanism. Are
62 the hot-finger structure and, by inference, small-scale mantle convection a general
63 phenomenon at subduction zones or unique to those with a possibly hydrated mantle
64 wedge in the back-arc produced by an old and cold slab subducting at a fast rate, such as
65 the western Pacific plate?

66 With a relatively young and thin slab and presumably shallow dehydration (van
67 Keken et al., 2011), the subduction of the Juan de Fuca plate beneath western North
68 America represents an end member in the subduction zone system. Along strike, the
69 subducting slab, the overriding plate and seismicity show clearly segmented signatures

70 (Tréhu et al., 1994; Brocher et al., 2003; Burdick et al., 2008). In particular, the
71 Quaternary volcanoes are spatially clustered and different in the composition of primitive
72 basalts (Figure 2, Schmidt et al., 2008). The arc basalts show evidence for both flux and
73 dry melting, with the maximum water content lower than those found at other arcs
74 (Elkins Tanton et al., 2001; Ruscitto et al., 2010). Unlike the subduction zones with well-
75 developed back-arc spreading centers (e.g., Wiens et al., 2008), the Cascades have only
76 volumetrically minor and sparse Quaternary volcanic activities behind the arc (Till et al.,
77 2013). So the extent and form of subduction-induced mantle flow and decompressional
78 melting in the back-arc of the Cascades and those of similar continental subduction zones
79 remain enigmatic.

80 A well-defined crust and upper mantle velocity model is needed to understand the
81 melting processes and the causes of along-strike segmentation of volcanism in the
82 Cascadia subduction zone. Although there have been many velocity models for the
83 Cascades (Lee and Crosson, 1990; Symons and Crosson, 1997; Parsons et al., 1999; Zhao
84 et al., 2001; Brocher et al., 2001; Calvert et al., 2001; Shapiro and Ritzwoller, 2002;
85 Tréhu et al., 2002; van Wagoner et al., 2002; Crosson et al., 2002; Graindorge et al.,
86 2003; Ramachandran et al., 2004; Burdick et al., 2008, 2010; Roth et al., 2008; Yang et
87 al., 2008; Abers et al., 2009; Audet et al., 2009; Moschetti et al., 2010; Schmandt and
88 Humphreys, 2010; Calkins et al., 2011; Gao et al., 2011; Delorey and Vidale, 2011;
89 Porritt et al., 2011; Wagner et al., 2012; Shen et al., 2013), none of these models covers
90 the entire subduction zone, has high enough resolution at the depth of melt generation,
91 and is adequately accurate to explain the observed waveforms as illustrated by Gao and
92 Shen (2012) with full-wave simulation. In addition, the magnitudes of velocity

93 perturbations in the existing models resolved from body-wave tomography in the Pacific
94 Northwest (Roth et al., 2008; Burdick et al., 2008, 2010; Obrebski et al., 2010, 2011;
95 Schmandt and Humphreys, 2010, 2011; James et al., 2011; Sigloch, 2011) vary within a
96 wide range (Becker, 2012). Furthermore, the resolution gap in the upper mantle between
97 surface-wave tomography and body-wave tomography limits geodynamic interpretations
98 at the depth of melt generation.

99 In this study, we invert for the upper mantle structure of the Pacific Northwest using
100 an advanced full-wave tomographic method based on simulation of wave propagation
101 within the 3D Earth structure (Zhang et al., 2012). Compared to previous studies, three
102 factors significantly improve the model resolution: First, there has been a huge increase
103 in broadband seismic data in the Pacific Northwest. We have processed ambient noise
104 waveforms from ~1000 stations (see station distribution in Figure 2); Secondly, we have
105 developed a new waveform normalization method (Shen et al., 2012) that improves the
106 quality of surface waves extracted from ambient seismic noise and are able to obtain
107 much more broadband and higher quality Rayleigh waves than in previous studies
108 (Figure 3); and thirdly, the full-wave tomographic method is based on wave propagation
109 simulation in 3D models, a more accurate theory that relates seismic data to the Earth
110 structure (Zhao et al., 2005; Chen et al., 2007; Shen and Zhang, 2010). Unlike previous
111 studies of the Pacific Northwest, we calculate the sensitivities of Rayleigh waves to
112 perturbations of both V_p and V_s and jointly invert for the velocity model. The velocity
113 model is then improved by iteratively reducing the misfit between the observed and
114 synthetic waveforms.

115

116 2. Data and Methods

117 The procedure of the full-wave ambient noise tomography includes extraction of
118 empirical Green's functions (EGFs) from continuous ambient noise waveform, finite-
119 difference wave propagation simulation in the 3D Earth structure, measurement of phase
120 delays between observed and synthetic waveforms, calculation of sensitivity kernels and
121 inversion for velocity perturbations. The first three steps are fully described by Gao and
122 Shen (2012), so we only briefly summarize here.

123

124 2.1. Extraction of empirical Green's functions

125 To retrieve Rayleigh-wave EGFs between station pairs, we process the vertical
126 component of continuous seismic data recorded between 1995 and 2012 by about 1000
127 stations in an area extending from northern California to Vancouver Island, Canada
128 (Figure 2). We include seismic stations from the EarthScope USArray Transportable
129 Array (TA), the Canadian National Seismograph Network (CN), the Plate Boundary
130 Observatory borehole seismic network (PB), the Portable Observatories for Lithospheric
131 Analysis and Research Investigating Seismicity (PO), the University of Oregon regional
132 network (UO), the Pacific Northwest regional seismic network (UW), the United States
133 national seismic network (US), the Berkeley digital seismography network (BK), the
134 Cascade chain volcano monitoring (CC), the Caltech regional seismic network (CI), and
135 many other flexible arrays including High Lava Plains broadband seismic experiment
136 (XC), Mendocino experiment (XQ), Cascadia arrays for EarthScope (XU), Flexarray
137 along Cascadia experiment for segmentations (YW), and Wallowa broadband experiment
138 (ZG). To our knowledge, this is the first time that all these networks are combined for

139 surface-wave tomography, resulting in a dataset that provides the most comprehensive,
140 dense coverage of the study area.

141 Prior to cross-correlating the vertical-component waveforms from station pairs, we
142 remove instrument response, normalize ambient noise data with a frequency-time-
143 normalization method (Shen et al., 2012), and eliminate time segments of large ($M > 5.5$)
144 earthquakes. To increase the signal-to-noise ratio, we stack daily cross-correlations for
145 each station pair, producing high-quality Rayleigh waves at periods of 7-200 s (Figure 3).
146 As this study focuses on structures from the mid-crust to upper mantle, for computational
147 reasons we only use 15-200 s periods in the analysis described here. The EGFs are then
148 recovered as the time derivative of the stacked cross-correlations (e.g., Sabra et al., 2005;
149 Snieder, 2004). In addition, we obtain monthly stacks of cross-correlations, whose
150 variations provide estimates of the uncertainties of EGFs and their travel times.

151 The conditions to equate EGFs with Green's functions of the Earth include a uniform
152 distribution of noise sources around the seismic stations and zero attenuation (e.g.,
153 Wapenaar, 2004; Wapenaar and Fokkema, 2006). These conditions are usually not
154 strictly satisfied in ambient noise seismic tomography (e.g., Yang and Ritzwoller, 2008)
155 and the Pacific Northwest is no exception as most of the ambient noise comes from the
156 Pacific Ocean. Following helioseismological practices in dealing with similar issues,
157 Tromp et al. (2010) suggested construction of ensemble-averaged cross-correlation and
158 corresponding ensemble-averaged sensitivity kernels. This method requires the power
159 spectral distribution of the ambient noise sources, which is highly variable spatially and
160 temporally (e.g., Uchiyama and McWilliams, 2008; Bromirski and Gerstoft, 2009). To
161 construct ensemble cross-correlations, we must know the global power spectral

162 distribution of ambient noise sources for the various overlapping recording periods of
163 station pairs. This detailed knowledge of the global power spectral distribution of
164 ambient noise sources is currently unavailable and requires substantial work that is
165 beyond the scope of this study. On the other hand, it has been suggested that the non-
166 uniformity of noise sources would significantly affect the surface-wave amplitude (Tsai
167 and Moschetti, 2010) but not the velocity (Snieder, 2004). Numerical experiments show
168 that the non-uniform distribution of noise sources leads to less than 0.5% error in travel
169 times and phase velocity (Yang and Ritzwoller, 2008). This level of error is much less
170 than the lateral velocity variations in the Cascades (e.g., Porritt et al., 2011; Gao et al.,
171 2012). Furthermore, for the study area with an average travel time of ~ 150 s, a 0.5% error
172 is equivalent to a ~ 0.75 -s error in travel time, less than measurement errors and the RMS
173 data misfit. In the following, we consider the effects of non-uniform noise source
174 distribution on travel times secondary to those of the Earth structure, and EGF a close
175 approximation to the Green's function of the Earth for velocity inversion.

176

177 2.2. Finite-difference wave simulation

178 We implement a nonstaggered-grid, finite-difference method to simulate wave
179 propagation in the 3D spherical Earth structure (Zhang et al., 2012). Each seismic station
180 is considered as a virtual source and all others as receivers. The regional 3D shear-wave
181 velocity model by Gao et al. (2011) is chosen as the initial reference model, merged with
182 CUB (Shapiro and Ritzwoller, 2002) at locations beyond the coverage of the original
183 model. Deeper than 400 km, we use the AK135 model (Kennett et al., 1995). P-wave
184 velocity is converted from shear-wave velocity with a V_p/V_s ratio of 1.74 in the crust

185 (Brocher, 2005) and the depth-dependent relationship of V_p and V_s of AK135 in the
186 mantle (Kennett et al., 1995). Density is calculated as a function of V_p (Christensen and
187 Mooney, 1995). Constraints on the Moho depth (Lowry and Pérez-Gussinyé, 2011) are
188 added to the initial reference model. No anisotropy and attenuation are included in the
189 simulation, though the effect of attenuation is considered in the interpretation of the
190 tomographic results.

191 For computational reasons, we carry out two levels of finite-difference wave
192 simulation, starting from a coarser grid for longer-period waves. The horizontal grid
193 spacing is 10 km and 5 km for level 1 and level 2, respectively, along the geographic
194 longitude and latitude. The vertical grid spacing is about one-third of the horizontal
195 spacing near the surface and increases with depth to approximately the same as the
196 horizontal spacing at ~ 100 km depth. Such grid sizes are sufficient to accurately simulate
197 waves at periods greater than 40 s and 21 s for level 1 and level 2, respectively (Zhang et
198 al., 2012). The total wave propagation time is 1000 s as the longest inter-station distance
199 is ~ 3000 km. To maintain numerical stability, we use a time step of 0.5 s for level 1 and
200 0.25 s for level 2. To calculate Green's functions, we use a Gaussian pulse with a half
201 width of 7.5 s and 4 s, respectively, as the source-time function of the vertical force
202 applied at the virtual station. The wave simulations are executed on a Linux cluster with
203 17 nodes (each with 24 cpu-cores). It takes about 0.3 and 1 hour per simulation with two
204 nodes for level 1 and level 2, respectively.

205

206 2.3. Cross-correlation of EGFs and synthetics

207 The phase delay times between the EGFs and synthetics are measured by cross-
 208 correlation at multiple overlapping period bands, with the central periods of 55 s, 75 s,
 209 112.5 s, 150 s, and 200 s for level 1 and 22.5 s, 37.5 s, 56 s, 75 s, and 112.5 s for level 2.
 210 The corresponding Rayleigh waves have peak sensitivities to structures from the mid-
 211 crust to 250 km depth. In this paper, we focus our interpretations on the mantle structure
 212 (50-140 km). Before the delay measurement, the EGFs are convolved with the source-
 213 time function used in the calculation of Green's functions to account for the finite-
 214 frequency nature and initial time shift of the simulated Green's functions. To ensure high-
 215 quality signals, the signal-to-noise ratio of EGFs is required to be at least eight, the inter-
 216 station distance at least 1.5-wavelength, and the cross-correlation coefficient between the
 217 EGFs and synthetics greater than 0.90. The number of measured phase delays varies from
 218 ~3,000 to 20,000 within different frequency bands.

219

220 2.4. Sensitivity kernels and inversion method

221 In previous surface-wave tomographic studies (e.g., Yang et al., 2008; Gao et al.,
 222 2011; Porritt et al., 2011; Wagner et al., 2012), the inversion of phase velocity to shear-
 223 wave velocity is carried out under the assumption that Rayleigh waves are not affected by
 224 P-wave speed. This is not accurate, especially at shallow depths (Figure S1). We
 225 represent the Rayleigh wave phase delay time δt as a joint Vp and Vs inverse problem,

$$226 \quad \delta t = \int [K_\alpha(\mathbf{m}_0, x)\Delta\mathbf{m}_\alpha + K_\beta(\mathbf{m}_0, x)\Delta\mathbf{m}_\beta]dV \quad (1)$$

227 where \mathbf{m}_0 , $\Delta\mathbf{m}_\alpha$ and $\Delta\mathbf{m}_\beta$ are the 3D reference model, Vp and Vs perturbations, and K_α
 228 and K_β the Rayleigh-wave sensitivity kernels to Vp and Vs, respectively. The integration
 229 is for the 3D volume of the model. Although the velocity structure in the shallow crust is

230 not well constrained in this study due to the intermediate- to long-period data used, the
231 inclusion of V_p in inversion provides additional degrees of freedom that minimize the
232 extent to which V_p anomalies in the shallow crust are mapped into the deep crust and
233 upper mantle. The effect of density is not explicitly expressed in the equation, but density
234 is recalculated based on V_p (Christensen and Mooney, 1995) after each iterative model
235 update. The sensitivity kernels are calculated with the strain-Green-tensor-based,
236 scattering-integral method (Zhao et al., 2005; Chen et al., 2007; Zhang et al., 2007). The
237 inverse problem is solved with damping and smoothness constraints. The best-fit
238 damping and smoothing parameters, which are gradually reduced with iterative
239 inversions, are chosen from the tradeoff of the normalized chi-squared value and the
240 model variance (Gao and Shen, 2012).

241 We start wave propagation simulation from level 1 to construct a large-scale
242 framework, which provides the reference model for level 2. The model is then iteratively
243 updated by alternating the two-level full-wave tomographic imaging. The solution does
244 not change significantly after 3-4 iterations. In total, we run five iterations for each level.
245 Compared to the initial reference model, the updated model yields synthetic waveforms
246 that match the observed EGFs much better (Figure 4), with the standard deviation of
247 phase delays decreasing from 2.5 s to 0.7 s. The phase delays of the updated model as a
248 function of inter-station distance are less scattered and centered around zero (Figure 5).
249 We observe that, on average, our model has much stronger velocity perturbations
250 compared to previous models (Shapiro and Ritzwoller, 2002; Yang et al., 2008; Gao et
251 al., 2011). It appears that more data and/or a more accurate methodology result in a
252 stronger contrast of the velocity anomalies. This has also been observed in the western

253 U.S. among various body-wave tomographic studies (as compared by Becker (2012)),
254 Iceland (Hung et al., 2004), and the Lau Basin (Wiens et al., 2006, 2012).

255

256 3. Seismic Results and Discussion

257 We focus our discussion on the mantle structure, in the depth range that is best
258 constrained by the ambient noise data and most relevant to melt generation (40-140 km).
259 Because the Rayleigh-wave sensitivity to V_p concentrates primarily in the shallow depth
260 (Figure S1), V_p in the mantle is not well constrained and thus not interpreted. Our shear-
261 wave velocity model, as shown in Figure 6, has features that are similar to those in the
262 previous tomographic models. For example, we image the low-velocity Yellowstone
263 hotspot at all the depths (e.g., Moschetti et al., 2010; Gao et al., 2011; Wagner et al.,
264 2012), an observation that is consistent with the active magmatism and high surface heat
265 flow (Lowry and Pérez-Gussinyé, 2011). The geometry of the subducting slab is better
266 resolved in this study at depths greater than 80 km than in previous surface-wave studies
267 (e.g., Yang et al., 2008; Gao et al., 2011; Porritt et al., 2011; Wagner et al., 2012). The
268 seismic velocity of the subducting slab is heterogeneous along strike. At depth of ~ 110
269 km, we image a slab hole (or weak slab) in northern Oregon, which spatially correlates
270 with a similar gap or weak slab in the body-wave tomographic models at depths greater
271 than 160 km (Roth et al., 2008; Burdick et al., 2008; Schmandt and Humphreys, 2010).

272 The most striking features of our Cascade model are three low shear-wave velocity
273 volumes, with velocities as low as ~ 3.6 km/s, in the upper mantle along the back-arc
274 (Figures 6 and 7). The back-arc low-velocity anomalies are about 200 km away from the
275 arc, segmented along strike and correlate spatially with the three volcano clusters along

276 the Cascades (Figure 2). The inter-spacing of these anomalies (center to center) is ~300
277 km. Large-scale, separate low-velocity anomalies have also been imaged in the upper
278 mantle wedge along the Izu-Bonin-Mariana arc (where the spacing of the anomalies is
279 ~500 km), and are interpreted as heterogeneous along-strike mantle flow (Isse et al.,
280 2009). Among the three anomalies, the central one at the Oregon back-arc is the largest in
281 size and lowest in velocity and corresponds to the arc segment with the largest
282 Quaternary eruption volume (Sherrod and Smith, 1990).

283 Analysis of model resolution in full-wave inversion is complicated by several factors,
284 including the non-linear relationship between the data and model and the multiple model
285 iterations carried out to reach the final model. The prohibitive cost of forward wave
286 propagation simulation makes the probabilistic approaches commonly used to deal with
287 nonlinear inverse problems impractical. Fichtner and Trampert (2011) proposed a method
288 based on the Fréchet derivatives of the misfit function. Their method, however, depends
289 on the condition that the model is in the vicinity of an optimal Earth model and the global
290 minimum of the misfit function has been found, an assumption that is difficult to verify in
291 large-scale, nonlinear full-wave inversion.

292 Synthetic inversion of various input models is a common practice in tomographic
293 resolution analysis. This approach has limitations and can be misleading in the sense that
294 synthetic inversion explores only a limited model subspace (Lévêque et al., 1993).
295 Nevertheless, it is useful if the limitations are understood and interpretations are
296 restricted to the model subspaces explored. With this caveat in mind, we run multiple
297 resolution tests. We first use our preferred model as the input (Figure 6). The synthetic
298 phase delay times are calculated with the sensitivity kernels (Eq. 1). The uncertainties of

299 individual observed phase measurements are estimated from the variations of monthly-
300 stacked EGFs and added to the synthetic phase delays. The velocity perturbations of the
301 input model are well reconstructed in inversion at the depth of our interest (Figures 9 and
302 10). We then run the 3D checkerboard resolution tests with a maximum of $\pm 10\%$ velocity
303 perturbations for both V_p and V_s (Figures S2-S4). The velocity variation within each
304 checkerboard cell is a cosine function. The dimensions of the checkerboard cells vary
305 from 100-200 km along the geographic longitude and latitude, and from 90-150 km
306 vertically. For the small checkerboard, although the pattern of the velocity perturbation
307 can be fairly well reconstructed, the magnitude is underestimated. For the larger
308 checkerboards, both the pattern and smooth variation of the magnitude can be well
309 recovered. The sizes of our observed back-arc low-velocity volumes are comparable to
310 the largest checkerboard. In the above tests, the recovered structures are obtained in a
311 single model iteration. A fully non-linear inversion with multiple iterations, as for the
312 observed data in this study, will further minimize the residual, resulting in a sharper
313 reconstruction of the model. The results in Figures 9 and 10 can thus be considered
314 conservative. Taken together, Figures 9 and 10 and the checkerboard tests indicate that an
315 Earth model with a structure that resembles the inferred back-arc low-velocity anomalies
316 is well resolved.

317 The three distinct, segmented low-velocity volumes along the Cascade back-arc in the
318 upper mantle are resolved for the first time. We attribute the resolution to the dense data
319 coverage, an EGF dataset with a broad frequency band well suited for imaging the crust
320 and upper 200 km mantle, and an advanced full-wave tomographic method. This allows
321 us to gain new insight into the dynamic processes of the Cascadia mantle wedge.

322

323 3.1. What controls the reduction of shear-wave velocity?

324 The seismic velocity can be affected by a few factors, including temperature, water
325 content and presence of partial melt. As shown in Figure 8, the shear-wave velocity of the
326 back-arc anomalies is 3.6-4.0 km/s within the depths of 80-120 km, which are deeper and
327 lower compared to where melting is inferred for the 0-4 Ma Pacific mantle (Nishimura
328 and Forsyth, 1989), the Lau back-arc basin (Wiens et al., 2006, 2008), and the Izu-Bonin-
329 Mariana arc (Isse et al., 2009). This strongly suggests presence of melt beneath the
330 Cascade back-arc, although temperature, water content, and grain size can also contribute
331 to the reduction of seismic velocity. The young, subducting Juan de Fuca plate leads to
332 shallow dehydration and less subduction-related water input into the back-arc mantle
333 wedge to affect seismic velocities. This inference is supported by the observation that the
334 primitive basalts erupted on the back-arc side of the Cascades are nominally dry (Ruscitto
335 et al., 2010). The back-arc lithosphere inferred from our model is relatively thin (Figure
336 7), consistent with the receiver function images of the lithosphere-asthenosphere
337 boundary (Kumar et al., 2012; Hopper et al., 2013) and the high surface heat flow
338 (Blackwell et al., 1990; Ingebritsen and Mariner, 2010). However, mantle temperature
339 alone cannot explain the observed back-arc low-velocity anomalies. Using the method of
340 Jackson and Faul (2010) for the geothermal profiles estimated from the Cascade heat
341 flow (based on Currie et al. (2004), Figure 8c), we find that the predicted shear-wave
342 velocities are all much higher than the observed (Figure 8b). The lowest velocity
343 calculated with the geotherms is ~ 4.2 km/s within the depth range of 80-110 km, which is
344 more than 0.4 km/s higher than the observed velocities beneath the back-arc low-velocity

345 anomalies (blue lines in Figure 8b). Correction for attenuation ignored in forward wave
346 simulation assuming a low Q value of 50 (Dalton et al., 2008) reduces the velocity
347 mismatch by ~0.1 km/s to 0.3 km/s. The additional velocity reduction needed to match
348 the observed is indicative of the presence of partial melt, which can drastically reduce
349 shear-wave velocities. The electromagnetic study in the region (Egbert, 2012) also
350 supports the possible presence of melt in the back-arc.

351 Using S-to-P converted phases (Sp), Hopper et al. (2013) map the lithosphere-
352 asthenosphere boundary (LAB) beneath the Pacific Northwest. The depth of their LAB is
353 consistent with the base of the imaged high-velocity mantle lid of the upper plate (Figure
354 7), including the deepening of the LAB from the back-arc in south and central Oregon to
355 the back-arc of Washington, where the latest magmatism occurred more than 15 Ma ago.
356 Hopper et al. (2013) attribute the consistent Sp observed beneath the Cascades back-arc
357 to the negative LAB velocity gradient created by a layer of partial melt ponding beneath a
358 solidus-defined boundary. Interestingly, strong Sp phases - in other words a large and/or
359 sharp negative velocity gradient at the LAB - are clustered in three areas in the back-arc
360 that roughly overlap with the three low-velocity volumes (Figure 7). A simple
361 explanation of the spatial overlap between the strong Sp phase and low-velocity volumes
362 is that the low-velocity volumes represent the regions of partial melt production. Melts
363 migrate upwards to collect at the base of the lithosphere, causing a relatively large and/or
364 sharp negative velocity gradient to generate the strong Sp phase.

365

366 3.2. What processes contribute to the pattern of the back-arc anomalies?

367 The geometry and magnitude of the velocity anomalies in the mantle wedge and their
368 spatial correlation with the arc volcanoes provide constraints on the mechanism of melt
369 generation at the Cascades. Compositionally buoyant small-scale diapirs (Figure 1)
370 triggered by fluid released at the slab arise from the top of the slab with a more or less
371 vertical geometry beneath the arc (Hasenclever et al., 2011). These clearly do not match
372 the observed back-arc low-velocity volumes. Nevertheless, we cannot exclude the
373 possibilities of melting related to small-scale diapirs (Hall and Kincaid, 2001) because of
374 the difficulties in tomographically imaging such small-scale features. Beneath the Oregon
375 Cascade arc, where a hydrated uppermost mantle wedge has been suggested previously
376 (Bostock et al., 2002), the low-velocity zone atop and approximately parallel to the slab
377 extending from depth of ~100 km upward to ~60 km near the Cascade arc (Figure 7c) is
378 consistent with flux melting (Wiens et al., 2008; Zhao et al., 1997). Note that the plate
379 interface terminates at depth of 100 km and is poorly defined at greater depths (McCroory
380 et al., 2004, 2012). The lack of such a low-velocity zone along the entire arc (Figures 7b,
381 7c and 7d), however, indicates a variable strength of flux melting along the arc.

382 The pattern of the back-arc low-velocity anomalies (Figures 7b, 7c and 7d) is most
383 consistent with subduction-induced mantle upwelling and decompressional melting
384 (Figure 1). The asthenospheric flow from beneath the old and thick North America
385 continental lithosphere towards the mantle wedge must undergo decompression. The fact
386 that the lowest velocities are at 80-110 km depth suggests that melting may involve damp
387 (50-200 p.p.m H₂O) peridotite and/or carbonated peridotite (Dasgupta et al., 2013).
388 Bifurcation of the Yellowstone plume driven by subduction-induced mantle flow
389 (Kincaid et al., 2013) may contribute to an excess mantle temperature and the stronger

390 anomaly beneath the Oregon back-arc, though bifurcation of the plume cannot explain all
391 three back-arc low-velocity volumes. We suggest that the decompressional melts in the
392 back-arc are not only responsible for the volcanism in the back-arc (Till et al., 2013), but
393 also a likely source of the low-water-content magmas at the arc (Elkins Tanton et al.,
394 2001; Ruscitto et al., 2010) ~200 km away. Melts may migrate upslope along a dipping
395 decompaction channel near the base of the lithosphere (Sparks and Parmentier, 1991),
396 though the exact mechanism of melt migration remains unclear.

397 Extension of the northern Basin and Range may contribute to the velocity reduction
398 in the southern Cascades (Ingebritsen and Mariner, 2010; Wang et al., 2002). However,
399 this mechanism cannot explain the back-arc low-velocity anomalies to the north.
400 Furthermore, the Basin and Range goes beyond the study area, while the southern low-
401 velocity anomaly extends only to near the southern end of the slab (Figure 2). The spatial
402 mismatch with the Basin and Range and the close correlation with the slab and the arc
403 volcanoes suggest that the low-velocity anomaly beneath the southern Cascade back-arc
404 reflects primarily the subduction processes.

405 Laboratory experiments with slab rollback and small back-arc extension that mimic
406 the Cascadia subduction zone indicate that upwelling above the slab in the mantle wedge
407 is influenced by the large-scale plate motion (Druken et al., 2011). The laboratory
408 experiments show the strongest upwelling in the center of the slab, which matches the
409 largest low-velocity anomaly in our seismic imaging. However, the magnitude of
410 upwelling varies gradually along strike from the center to the edge in the laboratory
411 experiments, different from the segmented pattern of our observed low-velocity
412 anomalies. Thus, besides the plate-motion-controlled processes other mechanisms must

413 also be at work in Cascadia. One likely mechanism is small-scale convection due to
414 buoyancy associated with melting, differences in the temperature of the upwelling
415 mantle, negative buoyancy from the cooling of the lithosphere above, and pre-existing
416 lithosphere structure. Between the central and northern low-velocity volumes, the SKS-
417 splitting anisotropy pattern is complex and deviates from the plate motion direction
418 (Yuan and Romanowicz, 2010), possibly the consequence of the disturbance of lattice-
419 preferred orientation by small-scale convection in the mantle wedge (Morishige and
420 Honda, 2011).

421 The tight spatial correlation between the low-velocity volumes in the Cascade back-
422 arc and the volcano clusters is similar to the hot-finger structure at the Honshu subduction
423 zone (Tamura et al., 2002), which may be a consequence of small-scale convection
424 within a low-viscosity wedge in the Japan back-arc (Honda and Yoshida, 2005).
425 However, there is a notable difference between the Cascade back-arc anomalies and the
426 hot-finger structure beneath Honshu: The spacing of the volcano clusters and the finger-
427 like feature in Honshu is 50-100 km, much smaller than at the Cascades. Numerical
428 simulations show that the formation and dimension of small-scale convection depend on
429 the slab age, subducting rate and water flux into the mantle released from the slab, which
430 affects mantle viscosity (Honda and Yoshida, 2005). Furthermore, long-wavelength rolls
431 dominate in the early development of small-scale convection and/or for cases with a
432 small subduction speed, while short-wavelength rolls take over in late stages and/or for
433 cases with a large subduction speed (Honda, 2011). Compared to Honshu, the Cascadia
434 subduction zone is younger (subduction started about 48 Ma versus over 130 Ma) and has
435 a lower subduction speed (3.5 cm/yr versus 9 cm/yr). Thus, the link between the

436 wavelength of small-scale 3D convection and clustering of arc volcanoes at different
437 subduction zones may reflect the different subduction histories and the relatively dry or
438 wet nature of the mantle wedge.

439

440 4. Conclusions

441 A new upper mantle shear-wave velocity model in the Pacific Northwest has been
442 constructed in this study using an advanced full-wave ambient noise tomographic
443 method. We have imaged three segmented low-velocity ($\sim 3.6\text{-}4.0$ km/s) volumes along
444 the Cascade back-arc within the depth range of 80-115 km, which provides new insights
445 into the melting processes at the Cascadia subduction zone. Such low-velocity anomalies
446 require presence of partial melt within a hot upper mantle wedge. The correlation of the
447 three back-arc low-velocity anomalies in the mantle wedge with the volcanic arc and the
448 subducting slab is consistent with the pattern predicted by subduction-induced
449 decompressional melting. Furthermore, the along-strike variation suggests existence of
450 small-scale convection with a scale of ~ 300 km. Whether and how the melts generated
451 within the back-arc low-velocity volumes supply the low-water-content magmas at the
452 arc (Elkins Tanton et al., 2001; Ruscitto et al., 2010) remain unknown.

453 Whether small-scale mantle convection and 3D decompressional melting are
454 ubiquitous in the back-arc of subduction zones requires further understanding of the
455 connections between large-scale plate-driven processes and small-scale convection.
456 Although our shear-wave velocity model provides important constraints on the melting
457 processes in Cascadia, several other lines of work should be carried to further constrain
458 the complex 3D processes at the subduction zone. Our interpretation of the existence of

459 back-arc decompressional melting is based mainly on the shear-wave velocity model.
460 Poisson's ratio is in fact more sensitive to melt and fluid than Vp and Vs. Attenuation
461 also reflects temperature, water content, and grain size. Therefore, a Vp/Vs model and
462 attenuation structure (e.g., Lawrence and Prieto, 2011) will help distinguish the
463 contributions of melt, fluid, temperature and chemical composition, and understand the
464 melting generation processes (e.g., Wiens et al., 2008).

465

466

467 Acknowledgements

468 All the seismic waveforms used in this study are obtained from the IRIS Data
469 Management Center and the Canadian National Seismograph Network. We thank
470 Katherine Kelley, Mark Parmentier, Kelsey Druken and Chris Kincaid for helpful
471 discussions. We thank Ulrich Faul for providing the code to calculate shear-wave
472 velocities. This research is supported by the National Science Foundation (grant number
473 1144771) and Air Force Research Laboratory (grant number FA9453-10-C-0217).

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767

768 Figure Captions

769 Figure 1. Schematic illustration of possible melt generation and migration processes at
770 general subduction zones, which may depend on the slab age, sediment thickness,
771 subduction rate, and other factors. Thin lines with open arrows indicate melt paths, while
772 those with solid arrows represent mantle flow lines. VF stands for the volcanic front.

773

774 Figure 2. Segmentation of the Cascade arc volcanoes (red triangles), defined by
775 geochemical signatures (Schmidt et al., 2008), and distribution of seismic stations used in

776 this study (white dots). The depth contours of the Juan de Fuca plate interface at 20-100
777 km are from the model of McCrory et al. (2004). The time-progressive Newberry (NB)
778 and Yellowstone (YS) rhyolite eruptive progression across Snake River Plain (SRP) and
779 High Lava Plains (HLP) are shown as thin brown lines (in Ma). CB stands for the
780 Columbia basin.

781

782 Figure 3. Example of Empirical Green's Functions. (a) The lines connect the receivers
783 (blue triangles) to the "virtual source" (red dot). (b-c) EGFs from the "source" to all the
784 other stations derived from ambient noise cross-correlation of vertical-to-vertical
785 components are plotted by the inter-station distance, filtered at 100-200 s and 10-25 s,
786 respectively.

787

788 Figure 4. Observed EGFs vs. synthetics, filtered at periods of 10-25 s. (a) The lines
789 connect the "source" (red triangle) to the receivers (blue triangles). (b-c) Comparison of
790 observed waveforms (black lines) and synthetics (red lines), sorted by the distance
791 between the "source" and each receiver. Synthetic waveforms are generated from the
792 initial reference model (b) and our full-wave tomographic model (c), respectively.

793

794 Figure 5. Comparison of phase delay time between observed and synthetic waveforms
795 from the initial reference model (black) and our full-wave ambient noise tomography
796 (red). (a-b) The phase delay time versus inter-station distance at periods of 50-100 s and
797 25-50 s, respectively. (c-d) The histogram of phase delay time at corresponding period
798 ranges. It shows that the improvement in fitting the data with our updated model.

799

800 Figure 6. Shear-wave velocity structure (in km/s) at multiple depths. All the panels share
801 the same velocity scale, as denoted by the color bar. Other symbols are the same as in
802 Figure 2.

803

804 Figure 7. Segmented low-velocity anomalies along the Cascade back-arc. (a) Horizontal
805 slice at depth of 94 km (V_s in km/s). The black dashed lines outline the amplitude of
806 largest negative S_p phase from receiver functions in the back-arc (Hopper et al., 2013).
807 The magenta lines mark the profile locations in (b), (c), (d) and (e), respectively. The
808 three white dots mark the point locations in Figure 8. All the panels share the same color
809 bar. (b-d) W-E profiles across the back-arc anomalies. The y-axis has the approximate
810 same length scale as the x-axis. The triangles mark the volcano centers. The Juan de Fuca
811 plate interface at depths of 20-100 km from the model of McCrory et al. (2004) is
812 projected. At greater depth, the plate interface is poorly defined. (e) S-N profile along the
813 back-arc low-velocity anomalies, which spatially correlate with the three volcano clusters
814 as in Figure 1. The length scale of y-axis is exaggerated two times of the x-axis.

815

816 Figure 8. Point velocity versus depth. (a) Comparison of velocities for points above the
817 back-arc anomalies (blue) with the velocities of the Lau Basin (red) and the 0-4 Ma
818 Pacific upper mantle (black) (based on Wiens et al. (2008)), and the calculated velocity
819 for the Cascade geotherm (gray, the same as the black line in (b)). (b) Comparison of the
820 observed and calculated shear-wave velocities. The blue lines are the observed back-arc
821 velocities, the same as in (a). The other four lines are calculated (with the code from

822 Jackson and Faul (2010)) for the four geothermal profiles in (c) based on Figure 4 of
823 Currie et al. (2004). In the legend of (c), the first number indicates the surface heat flow
824 (in mW/m^2) and the second one is the potential temperature of the mantle (in $^{\circ}\text{C}$). The
825 difference between the calculated velocity and the observed is suggestive of the presence
826 of partial melt.

827

828 Figure 9. Resolution test for the low-velocity anomalies along the back-arc imaged in this
829 study. (Upper panels) Input model is our preferred tomographic model; (Lower panels)
830 Recovered velocities at the corresponding depths. Note that the amplitude of the velocity
831 perturbation is not fully recovered partly because of the damping and smoothing factors
832 used in the inversion.

833

834 Figure 10. Vertical resolution test along three W-E profiles that cross the back-arc low
835 velocity anomalies. (Upper panels) Input model is our preferred tomographic model,
836 same as in Figure 9. The three profiles correspond to the cross-sections in Figures 7b-7d,
837 respectively. The velocity perturbation varies within a range of $\pm 10\%$. (Lower panels)
838 The recovered velocity models along the corresponding vertical profiles.