

## REPORTS

## GEOPHYSICS

# Tomography reveals buoyant asthenosphere accumulating beneath the Juan de Fuca plate

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The boundary between Earth's strong lithospheric plates and the underlying mantle asthenosphere corresponds to an abrupt seismic velocity decrease and electrical conductivity increase with depth, perhaps indicating a thin, weak layer that may strongly influence plate motion dynamics. The behavior of such a layer at subduction zones remains unexplored. We present a tomographic model, derived from on- and offshore seismic experiments, that reveals a strong low-velocity feature beneath the subducting Juan de Fuca slab along the entire Cascadia subduction zone. Through simple geodynamic arguments, we propose that this low-velocity feature is the accumulation of material from a thin, weak, buoyant layer present beneath the entire oceanic lithosphere. The presence of this feature could have major implications for our understanding of the asthenosphere and subduction zone dynamics.

The physical causes of the lithosphere-asthenosphere boundary (LAB), possibly representing a zone that mechanically decouples tectonic plates from the asthenospheric mantle beneath (1, 2), remain poorly understood (3). The LAB beneath continents appears deep and somewhat obfuscated by other discontinuities (4, 5), so distinguishing the LAB has been hindered by the complicated nature of deep continental structures. Oceans are tectonically simpler than continents, so both the observation and description of the LAB should be simpler there. However, because oceans are poorly instrumented, serious difficulties remain in resolving the LAB beneath oceanic lithosphere (6). Of particular interest is this lithosphere-asthenosphere interaction at convergent margins. The geometry of an oceanic plate changes as it dips into the mantle at a subduction zone, and the response of the uppermost mantle remains debated (7–9). New constraints on processes beneath a dipping plate may provide insights into subduction zone dynamics as well as large-scale asthenospheric flow and its role in the evolution of tectonic plates.

First arrivals of *P* waves from distant earthquakes recorded on large seismic arrays can be used to illuminate large parts of the mantle. One such array, the Cascadia Initiative (10), was a 4-year (2011–2015) amphibious seismic deployment that covered the Juan de Fuca plate and the Cascadia subduction zone (Fig. 1). Using 61,559 *P*-wave arrivals observed on the Cascadia Initiative, the Transportable Array, and other regional seismic networks, we generated a *P*-wave velocity

model of the region through finite-frequency tomographic inversion following the method of Obrebski *et al.* (11). Our application of finite-frequency sensitivity kernels means that our inversion takes into account the frequency-dependent volume that is sampled by a *P* wave traveling from the source to the seismometer and obviates the need to smooth the final model. The noise characteristics of the ocean bottom seismometers (OBSs) require that we use long-period (9.1- to 12.5-s) arrivals.

Our model (CASC16-P) shows an expected north-striking, east-dipping, high-velocity (+3% *P*-wave velocity change,  $dV_p/V_p$ ) Juan de Fuca slab, seen at 150 km depth as a continuous north-south structure at about 122°W between 40°N and the northern edge of our model at 50°N (Fig. 2A and fig. S14). Vertical cross sections at 47°N (Fig. 2B) and 41°N (Fig. 2C) indicate that the slab is continuous down to the transition zone at ~410 km, or deeper, consistent with previous land-based studies in the region (11–14). A previously unidentified strong low-velocity anomaly (–2 to –3%  $dV_p/V_p$ ) is seen at 150 km depth with similar strike as the Juan de Fuca slab, just to the west. Vertical cross sections through the models and synthetic tests (see figs. S6 and S7 and Materials and methods) indicate that this feature is restricted to the top 300 km of the model directly beneath the Juan de Fuca slab, meaning that it does not follow the slab all the way down to the transition zone. Further tests with various station correction terms (fig. S5) indicate that this feature is not a shallow structure being incorrectly mapped to depth. This truncated feature appears to take the shape of a horizontal cylinder, slightly elongated vertically in cross section beneath the high-velocity slab. The addition of data from the Cascadia Initiative has provided the offshore resolution needed to confidently

identify the full extent of this feature, though evidence exists for the feature in previous land-based tomographic models of the region (11–14).

We propose that this low  $dV_p$  feature is related to previously reported observations of a layer of partial melt beneath the oceanic lithosphere. A range of techniques—including receiver functions from borehole OBSs on the Pacific and Philippine Sea plates (15), precursors to teleseismic SS-phase arrivals spanning the Pacific ocean (16), magnetotelluric inversion on the Nazca plate (17), and explosion-generated reflected *P* waves offshore of New Zealand (18)—resolve a narrow (10 to 25 km) region, immediately below the oceanic lithosphere, characterized by low seismic wave velocities (–6 to –10%  $dV_s/V_s$  and  $dV_p/V_p$ ) and high conductivity (4 to 6 ohm-m). The interpretation for each of these studies is slightly different, but they all indicate that partial melt fractions of ~1 to 4% in the uppermost asthenosphere are consistent with their findings, with variations arising due to different geometries of melt layers, composition of the melt, and the crystal structure of the materials in the layer. The slow seismic feature we observe in CASC16-P similarly coincides with a region of high conductivity. This thin layer is not resolvable in tomographic studies that use relative travel times from teleseismic events (11–13) because such variations affect each ray path in the same way (19). At a subduction zone, however, this layer might become observable where it changes geometry with the slab as it descends, thus becoming detectable via teleseismic body-wave tomography.

Predicting a priori how this layer might behave beneath a subduction zone requires knowledge of its physical properties. Because we cannot resolve the layer to the west of the subduction zone, CASC16-P does not provide a direct observation of the source of the material in this layer. Previous reports (15–18) attribute the layer to volatiles and/or hydrated mantle increasing partial melt fraction in the uppermost asthenosphere, decreasing density and viscosity, separating into lenses (20) or channels (15, 21), and ponding beneath the rigid, impermeable lithospheric lid (22). Our observations, as well as the land-based observation of a high-conductivity region that roughly coincides with our low velocities (23), could be explained by the accumulation of material from this horizontal layer due to its own buoyancy and low viscosity.

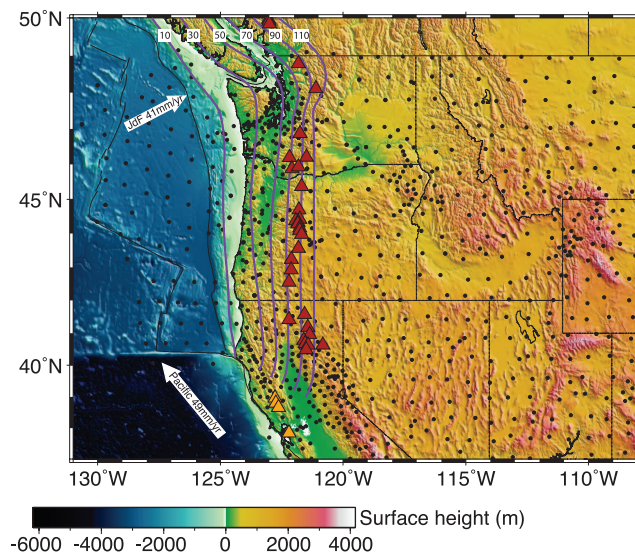
Here we use two straightforward fluid-mechanical scaling calculations to demonstrate the plausibility of the accumulation hypothesis: First, for a thin, buoyant, low-viscosity layer to accumulate beneath the downgoing slab the ratio of the upward mass flux due to the layer buoyancy (Poiseuille flow) must exceed the downward flux due to drag from the downgoing slab (Couette flow) (24). Referring to the notation in Fig. 3, this means that

$$\frac{\Delta\rho g \sin(\theta) h^2}{6\nu_0 \mu_1} \geq 1 \quad (1)$$

Here,  $\Delta\rho$  is the difference in density,  $g$  is gravitational acceleration,  $\theta$  is the dip angle,  $h$  is the

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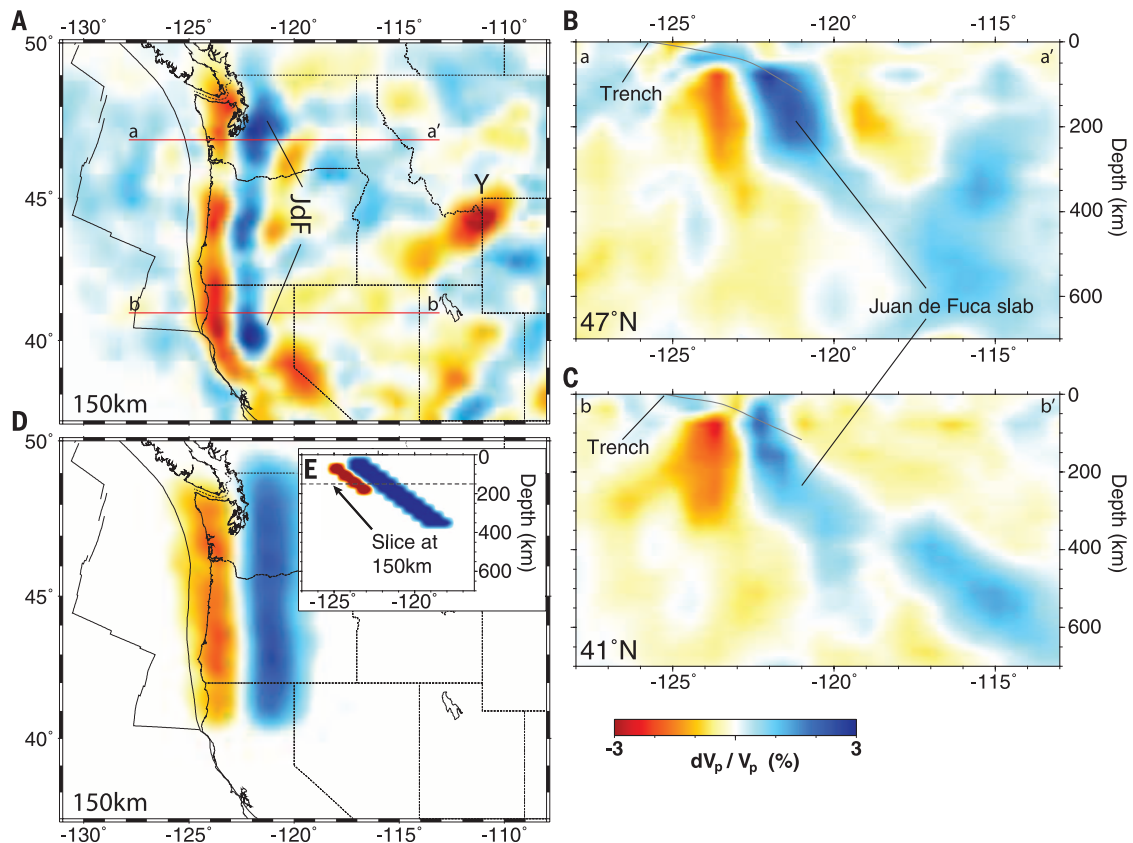
**Fig. 1. Stations used in our inversion (black dots) overlay on topography.** Cascade volcanoes are denoted by red triangles, and northward age-progressive volcanism of the California Coast Ranges is shown by orange triangles. White arrows indicate plate motions relative to North America. The estimated depth of the top of the subducting slab is shown by the purple contours labeled in kilometers (36).

layer thickness,  $v_0$  is the plate velocity, and  $\mu_1$  is the thin-layer viscosity. The second condition is that the horizontal gravitational spreading velocity of the accumulated low-velocity volume due to its own buoyancy cannot exceed the horizontal advection velocity due to plate motion (25), otherwise the accumulating low-viscosity body would spread out instead of accumulating to the observed thickness of  $H \approx 50$  to 100 km, as inferred from our tomographic images. This means that

$$\frac{\Delta\rho g H^2}{v_0 \mu_m} \leq 1 \quad (2)$$

Here,  $H$  is the thickness of the accumulated material, and  $\{\mu_m\}$  is the viscosity of the underlying mantle. In obtaining Eqs. 1 and 2, we assumed that the thin-layer viscosity controlling return flow is much smaller than the underlying mantle viscosity governing gravitational spreading—that is,  $\mu_1 \ll \mu_m$  (see supplementary materials for details).

Assuming that  $\Delta\rho \approx 5$  to 20  $\text{kg/m}^3$  [1 to 4% partial melt, with 500  $\text{kg/m}^3$  density contrast between melt and solid (26)],  $g = 9.8 \text{ m/s}^2$ ,  $\theta = 40^\circ$ ,



**Fig. 2. Slices through CASC16-P.** (A) A slice at 150 km depth. The slow Yellowstone (Y) and fast Juan de Fuca (JdF) slab anomalies are indicated. A strong low-velocity anomaly is seen west of the Juan de Fuca slab. Lines a-a' and b-b' mark the locations of vertical slices through CASC16-P at 47°N (B) and 41°N (C). These slices show the slab extending through the transition zone, whereas the low-velocity anomaly is confined to shallower depths ( $\sim 300$  km).

The top of the slab (36) is shown with a gray line that begins at the trench. There is no vertical exaggeration in the vertical slices. (D) A slice at 150 km depth shows the recovery of a synthetic test with a 120-km-thick +4% slab anomaly underlain by a 40-km-thick  $-10\%$  layer anomaly. The input synthetic velocity model is shown in (E). The recovery of synthetic features (D) is similar to that observed (A).



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