Local Thickening of the Cascadia Forearc Crust and the Origin of Seismic Reflectors in the Uppermost Mantle

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ABSTRACT

Seismic reflection profiles from three different surveys of the Cascadia forearc are interpreted using P wave velocities and relocated hypocentres, which were both derived from the first arrival traveltime inversion of wide-angle seismic data and local earthquakes. The subduction decollement, which is characterized beneath the continental shelf by a reflection of 0.5 s duration, can be traced landward into a large duplex structure in the lower forearc crust near southern Vancouver Island. Beneath Vancouver Island, the roof thrust of the duplex is revealed by a 5-12 km thick zone, identified previously as the E reflectors, and the floor thrust is defined by a short duration reflection from a <2 km thick interface at the top of the subducting plate. We show that another zone of reflectors exists east of Vancouver Island that is approximately 8 km thick, and identified as the D reflectors. These overlie the E reflectors; together the two zones define the landward part of the duplex. The combined zones reach depths as great as 50 km. The duplex structure extends for more than 120 km perpendicular to the margin, has an along-strike extent of 80 km, and at depths between 30 km and 50 km the duplex structure correlates with a region of anomalously deep seismicity, where velocities are less than 7000 m s^{-1}. We suggest that these relatively low velocities indicate the presence of either crustal rocks from the oceanic plate that have been underplated to the continent or crustal rocks from the forearc that have been transported downward by subduction erosion. The absence of seismicity from within the E reflectors implies that they are significantly weaker than the overlying crust, and the reflectors may be a zone of active ductile shear. In contrast, seismicity in parts of the D reflectors can be interpreted to mean that ductile shearing no longer occurs in the landward part of the duplex. Merging of the D and E reflectors at 42-46 km depth creates reflectivity in the uppermost mantle

with a vertical thickness of at least 15 km. We suggest that pervasive reflectivity in the upper mantle elsewhere beneath Puget Sound and the Strait of Georgia arises from similar shear zones.

INTRODUCTION

The Cascadia subduction zone, where the oceanic Juan de Fuca plate descends beneath the overlying North American plate, extends 1100 km from northern California to northern Vancouver Island. Along much of this subduction zone, sedimentary rocks originally deposited on the oceanic plate are scraped off and accreted to the western edge of the continent; the initiation of this process is indicated by the deformation front in Figure 1. The intersection of the deformation front with the SW-trending Nootka transform fault marks the approximate northern limit of the subduction zone. This fault is characterized by high levels of seismicity caused by relative motion between the Juan de Fuca plate and the small oceanic Explorer plate to the north. Earthquakes within the oceanic plate occur predominantly in two well-defined, ~50 km wide, geographic bands beneath the western and eastern edges of Vancouver Island that merge further south. These earthquakes have been attributed to dehydration embrittlement of the oceanic mantle (Peacock, 2001; Preston et al., 2003) and the gradual transformation of basalt to eclogite; this latter process may begin at depths as shallow as 40 km (Hacker et al., 2003; Preston et al., 2003).

The latest stage of subduction began along the margin of North America's Pacific Northwest in the Early Eocene (Engebretson et al., 1985; Haeussler et al., 2003). During this period, the Pacific Rim and Crescent terranes were accreted to the western margin of the continent. The Pacific Rim terrane, comprising mainly Mesozoic sediment and metasedimentary rocks, was thrust beneath the more inboard Wrangellia terrane, which at that time formed the edge of the continent, and now constructs most of Vancouver Island. The predominantly volcanic Crescent terrane of early Eocene age was also thrust beneath the Pacific Rim terrane. Rocks of both terranes crop out on the southwest corner of Vancouver Island, and the Crescent terrane is exposed as far south as Oregon, where it is known as the Siletz terrane. In Oregon, the thickness of the Crescent terrane is between 18 km and 28 km (Parsons et al., 1998), and at the latitude of northern Puget Sound it is imaged as an east-dipping slab that extends to a depth of approximately 25 km (Ramachandran et al., submitted); at 15-25 km depth, P wave velocities are 6.8-7.2 km s^{-1} implying that the Crescent terrane is largely gabbroic in this region.

Deep seismic reflections were first clearly observed above the subducting Juan de Fuca plate by a Vibroseis survey acquired across Vancouver Island in 1984 (Yorath et al., 1985). The reflectors beneath Vancouver Island occur in a 5-8 km thick zone that dips landward and increases in depth from 20 km to 33 km. Referred to as the E reflections, they were interpreted to arise from imbricated mafic and sedimentary rocks (Green et al., 1986) or sedimentary rocks accreted to the base of an underplated mafic unit (Clowes et al., 1987a). A marine survey shot off the west coast of Vancouver Island in 1985 (Clowes et al., 1987b) showed that the high-amplitude E reflectors truncate the Leech River fault, which separates the mafic Crescent and predominantly metasedimentary Pacific Rim terranes. This truncation led to the suggestion that the E reflections represent a regionally extensive shear zone developed in the crust above the subducting plate (Calvert and Clowes, 1990). A subsequent integrated interpretation of the 1985 data and another offshore survey shot in 1989 showed that beneath the continental shelf the deepest E reflections merge into other reflections that originate at, or just above, the top of the subducting Juan de

Fuca plate (Calvert, 1996). Beneath Vancouver Island the E reflectors may be associated with the presence of fluid-filled porosity of approximately 2%, because a magnetotelluric survey identified a landward-dipping conductivity anomaly at depths similar to the E reflectors (Kurtz et al., 1986).

In 1998, the SHIPS (Seismic Hazards Investigation in Puget Sound) program acquired seismic reflection data through the Strait of Juan de Fuca, Strait of Georgia and Puget Sound with the objective of delineating the crustal architecture to provide constraints on earthquake hazard analyses (Fisher et al, 1999). In this paper, we combine seismic data from this survey with the earlier reflection data to map the top of the subducting plate from 5 km depth beneath the deep ocean basin to at least 50 km depth east of Vancouver Island. Furthermore we identify bands of reflections that we interpret to be distributed shear zones that delineate a major duplex structure that has thickened the forearc crust near the seaward region of the forearc mantle wedge. By correlating the seismic reflection data with earthquake hypocentres relocated using a 3-D P wave velocity model (Ramachandran et al., 2005), we infer that the duplex structure encloses crustal rocks that have been derived either from the base of the forearc crust or from the downgoing oceanic crust itself, and then carried down to the forearc mantle. We also infer that one of the recently discovered episodic tremor and slow-slip (ETS) seismic events, namely that of February-March 2003, may have initiated in the deepest region of this duplex structure.

SEISMIC REFLECTION DATA

We show results from three different seismic reflection surveys, two shot by the Geological Survey of Canada in 1985 and 1989, and the SHIPS survey shot in 1998. The acquisition parameters differ somewhat among these surveys. In terms of the appearance of the stacked reflection data, the character of the airgun source array is probably the most significant, with the SHIPS reflection lines exhibiting a lower dominant frequency. The shot point and receiver spacing was 50 m and 25 m respectively in all the surveys, and the streamer lengths varied between 2575 m (SHIPS survey) and 3758 m, yielding 24 to 36-fold data. Although specific parameters differ slightly, the prestack processing flow is generally similar for all three surveys, consisting of amplitude recovery, deconvolution, frequency-wavenumber (F-K) filter to attenuate coherent noise, stacking velocity-analysis, normal moveout and stack. For the SHIPS lines, the F-K filter was applied in the constant offset domain as part of a dip-moveout correction. Further details can be found in Yorath et al. (1987), Spence et al. (1990), and Fisher et al. (1999). The sections presented here are coherency filtered; both unmigrated and migrated profiles are shown. Additional large format sections and specific line locations can be found in Fisher et al. (in press) and Calvert (1986).

High levels of coherent noise due to out-of-plane scattering near the seafloor are present in much of the seismic data. Water-layer multiple reflections also affect the data at late record times where the water depth increases down the continental slope into the deep ocean basin. Although most of the scattered coherent noise was removed during processing, some scattered noise is still present in the displayed sections (e.g. Figure 2). The noise appears mostly as steeply dipping coherent arrivals at late times that can be excluded from consideration during data interpretation. Some scattered energy, for example that associated with the apex of a scattering hyperbola, can appear subhorizontal, and might in some instances be misinterpreted as a primary reflection. To avoid this misidentification, reflectors are identified on unmigrated sections using several reflection segments, and where possible more than one seismic section. Only clear terrane boundary faults or regionally extensive reflecting surfaces are interpreted in this paper. Migrated sections are used to position reflectors in the subsurface, for example for correlation with the velocity-depth model. This positioning was not done on the strike section because event migration would occur mostly in the up-dip cross-line direction. In practice, the relatively low dip, less than 10^{o}, of most of the interpreted reflectors results in a relatively small lateral displacement, so a regional interpretation can be undertaken on either the stacked or migrated section.

To correlate seismic reflection data with earthquake hypocentres and interval velocity data, the reflection sections were converted to depth using smoothed interval velocities extracted from a 3-D P wave velocity model derived from both SHIPS wide-angle data and local-earthquake first arrivals (Ramachandran, 2001). In the composite sections presented later, the part of the velocity model that has a non-zero ray density is superimposed on the seismic reflection data. Checkerboard tests indicate that in the eastern part of the Strait of Juan de Fuca, the edges of structures with a horizontal dimension of 40 km are resolved at depths as large as 30 km (Ramachandran, 2001).

REFLECTION DATA AND INTERPRETATION OF INTER-PLATE BOUNDARY

We limit ourselves here to the interpretation of the primary features of the seismic reflection profiles, namely reflections from the inter-plate boundary and the thick bands of high amplitude reflections below 15 km. These features delineate the deep structures that are the focus of this paper.

Continental Slope

The deformation front at the base of the continental slope marks the location on the seafloor where the incoming subhorizontal sedimentary strata of the Juan de Fuca plate become involved in convergent tectonics (Figure 2). High amplitude reflections at the base of the sedimentary section, which are also associated with diffraction arrivals, are from the top of the igneous oceanic crust. On some seismic lines, a 0.4 s-thick series of reflections from the deepest incoming sedimentary strata can be traced landward below the continental slope to where they become high amplitude reflections beneath the continental shelf, suggesting that the subduction decollement occurs at the base of this sequence (Calvert, 1996). More commonly, as demonstrated by line 85-01 (Figure 2), these reflections are obscured by multiple reflections and other coherent noise beneath the slope.

Continental Shelf

Terrane boundaries beneath the continental shelf have been interpreted elsewhere (Calvert, 1996), and here we review only those characteristics of the plate boundary that are relevant to the interpretation of structures further landward. In Figure 2, the inter-plate boundary (labelled JdF) is interpreted to be the discontinuous 0.5 s-thick package of reflections that increase in time from approximately 5 s near the edge of the continental shelf to 9 s near the west coast of Vancouver Island. The discontinuity of these reflections is partly associated with overlying lateral velocity variation due to folding of the accretionary wedge. Because the upper part of the reflection package has been traced into the deepest levels of the incoming sedimentary section on other lines, the reflection from the top of the igneous oceanic crust is probably included in the lower part of this reflection package. It is unclear whether the inter-plate boundary corresponds to a thin interface within these reflectors or is a broad zone of distributed shear that could be as much as 2 km thick (Nedimovi_ et al., 2003).

Near the west coast of Vancouver Island, the approximately 2 s-thick band of E reflections extend as far west as 20 km seaward of Vancouver Island. The lower reflections merge into the upper part of the reflection package from the inter-plate boundary (Figure 3). Near the west coast of Vancouver Island, a 1-2 s thick non-reflective zone separates the base of the E reflections and a thinner, 0.5 s-thick, reflection, labelled F (Figures 2 and 3). Nedimovi_ et al. (2003) have interpreted the F reflector as the Moho of the subducting plate, but we favour the interpretation that the F reflector corresponds to the top of the descending oceanic plate (Calvert and Clowes, 1990), perhaps the top of the igneous crust. Three observations that support this interpretation are: first, P wave velocities just beneath the F reflector in the 3-D P wave velocity model of Ramachandran (2001) are 6.8-7.4 km s^{-1}, which are too low to be upper oceanic mantle; second, the time interval between the deepest of the E reflectors and the F reflector is as little as 0. 5 s, too small to be typical of oceanic crust; third, hypocentres of earthquakes within the subducting plate occur within 3 km of the JdF reflection package beneath the continental shelf, i.e. within the oceanic crust, but increase in depth landward, following the increase in depth of the F reflector rather than the geometry of the deepest of the E reflectors (Figure 4). In addition, the F reflection further landward is approximately 6 km above the oceanic Moho, which has been located using wide-angle reflections from the SHIPS survey (Preston et al., 2003). Thus we propose that the F reflector marks the top of the Juan de Fuca plate.

Vancouver Island

We combined six seismic lines (85-01, 85-05, JDF-1, JDF-5, JDF-3, and SG-1; see Fisher et al. (in press) for further details of individual lines) into a composite section that has been projected onto an azimuth of 063^{o} to simulate a dip line across the forearc region near southern Vancouver Island (Figure 5). This projection assumes limited along-strike structural variation. Although this is not true of upper crustal structures such as the Leech River fault, along-strike variation does not appear to be significant at the depth of the inter-plate boundary. The interpreted composite section in depth is shown in Figure 6.

The most dramatic structure in the dip section, which extends from the continental shelf west of Vancouver Island to the east side of the Strait of Georgia, is the band of E reflections that reach a maximum vertical thickness of 4 s (two-way time), or 12 km. This band of reflectors dips at less than 4^{o} to the east-northeast beneath Vancouver Island, but its dip increases sharply to approximately 15^{o} beneath the island's east coast. The reflectors can be followed downward to 16 s at the easternmost end of the section. The Leech River fault (LRF in Figures 5 and 6), which at the surface marks the boundary between the Crescent and Pacific Rim terranes, is truncated at depth by the E reflectors (Calvert and Clowes, 1990). Beneath Vancouver Island, the discontinuous F reflection occurs 1.8 s after the deepest of the E reflections (Figure 5). On high-gain displays, the F reflection is 80% continuous beneath western Vancouver Island, but can only be observed as localized reflection segments further east. Near the west coast of Vancouver Island, the E and F reflections define a wedge that narrows and merges westward into the JdF reflection from the inter-plate boundary. Below the west end of the section, the F reflector dips at approximately 15^{o}, but the dip decreases to 4^{o} below Vancouver Island (Figure 6). At the east coast of Vancouver Island, the F reflection is more difficult to identify owing to elevated levels of coherent noise. No indication is observed that the F reflector crosscuts the E reflectors, and the F reflector probably merges into the deepest of the E reflectors (Figure 6).

Puget Sound and Strait of Georgia

SHIPS seismic lines (PS-2 and SG-1; see Fisher et al. (in press)) have been combined to form a composite section approximately parallel to the margin (Figures 1 and 7). The distance of this seismic section from the deformation front varies, because the marine survey had to avoid islands in the Strait of Georgia and Puget Sound. The seismic section has not been projected onto any azimuth, because the strike of the margin changes along the profile's 330 km length.

Two parts of the margin-parallel composite section near southern Vancouver Island are shown in Figure 8. To demonstrate the similarity in the geometry of the deep reflections, north is plotted to the left in Figure 8a from Puget Sound, and north is to the right in Figure 8b from the Strait of Georgia. Although the data contain coherent noise scattered from the surrounding islands, the E reflections can be identified having an apparent dip to the right in both parts of Figure 8, i.e. away from the trench owing to the orientations of the two profiles (Figure 1). The top of the subducting plate is interpreted to be located at the deepest of the E reflectors, because this reflector corresponds to the top of inslab seismicity (Figure 7). In Figure 8b, another approximately 1.5 s-thick band of reflections can be identified above the E reflections beneath the Strait of Georgia. This band, which is labelled D in Figure 8b, appears to be separated from the E reflections by a less reflective region. However, the D and E reflections merge together near shot point (SP) 4000, forming a zone of reflectivity that is 4 s, >15 km, thick. A similar anastomosing geometry is seen in Figure 8a from Puget Sound, but the D reflections cannot be traced as far along the section. Both seismic reflection sections show that a >15 km thick band of reflectivity exists deeper than 40 km, depths which would normally correspond to the upper forearc mantle, because the crustal thickness is estimated to be 35 km (Spence et al., 1985; Clowes et al, 1995).

Below the northern end of the composite along-strike section, where obscuring coherent noise is low, the sedimentary section is less than 1 s thick, and the underlying upper and midcrust lack laterally coherent reflections (Figure 9). However, subhorizontal reflections are evident below 9 s, extending to the base of the section at 16 s. As the thickness of the forearc crust is approximately 35 km, i.e. 11-12 s two-way time, this zone of reflectivity, which is more than 26 km thick, exists in the lowermost crust and upper forearc mantle down to depths greater than 50 km.

NON-VOLCANIC TREMOR

Episodic slow-slip on the inter-plate boundary has been inferred from monitoring of surface deformation using GPS (Dragert et al., 2001). Also non-volcanic tremors that occur during episodic slow-slip have been identified as low-amplitude wavetrains with a peak frequency between 1 and 5 Hz that propagate with S wave velocity (Rogers and Dragert, 2003). The emergence of the tremors makes it impossible to pick P and S wave phases, so an alternative approach to locating their sources has been employed (Kao and Shan, 2004). A predicted arrival time is calculated at each seismograph station for each possible subsurface source position and for many different origin times using a 3-D S wave velocity model. The amplitude envelope of the seismograms near the predicted arrival times are then stacked to compute a measure of coherence. The origin time and 3-D spatial location that correspond to the maximum coherence is interpreted to be that of the tremor source. This approach assumes that the tremor's largest amplitude is associated with the direct S wave arrival, as is observed for local earthquakes at epicentral distances less than approximately 200 km.

Tremors that occurred during the episodic tremor and slip (ETS) events of February-March 2003 have been thus located (Kao et al., 2005). The S wave velocity model employed is a scaled version of the 3-D P wave velocity model (assuming a constant Poisson ratio of 0.25) that was used to depth convert the seismic reflection data and relocate the earthquake hypocentres. Tremor locations within 15 km of the seismic reflection profiles are shown in Figures 6 and 7.

Although the non-volcanic tremors occur at the same time as slow-slip, which is inferred to take place on the inter-plate boundary, the tremors are distributed over a broad depth range, extending from within 5 km of the surface to depths greater than 50 km. Tremors from four days of the 2003 ETS event, when activity was high, are shown in Figure 6. The migration of tremor activity to the northwest, given the location of the seismic lines (Figure 1), produces the apparent updip movement over time shown in Figure 6. In the latter part of the ETS event, shallow tremors also occurred, but this activity is beneath central Vancouver Island and some distance from the seismic profiles shown. Unlike the earthquakes, some tremors occur within the E reflectors, but they do not appear to occur preferentially in one subsurface region. Some of the early tremors of the 2003 ETS event occurred in the low velocity region of the upper mantle, including within the E reflections (Figure 7). Tremor activity of the ETS event began in northern Puget Sound, and then migrated to the northwest, away from the location of the seismic profiles. Thus no tremors later than Feb 27th are shown in Figure 7.

LOWER CRUSTAL DUPLEX

The presentday inter-plate boundary is interpreted to be the JdF reflector beneath the continental shelf and slope, the F reflector beneath Vancouver Island, and to be located at the base of the E reflectors east of Vancouver Island, defining a ramp-flat-ramp geometry (Figure 6). The deepest of the E reflectors merges into the top of the JdF reflector approximately 20 km west of Vancouver Island. Beneath Vancouver Island, the E reflectors dip shallowly landward, but near the east coast they dip more steeply, defining a ramp-flat geometry, and they are interpreted to merge with the F reflector further to the east. Our preferred interpretation is that the E reflectors are shear zone because they truncate the lower end of the Leech River terrane boundary fault at approximately 20 km depth (Calvert and Clowes, 1990; Nedimovi_ et al., 2003). We interpret the E and F reflectors together to be a 110 km-wide duplex structure, with the F reflector being the floor thrust and the E reflectors marking the roof thrust. The E reflectors may be a zone where ductile deformation is distributed vertically over 5-12 km. Alternatively, this thick zone of shallowly NE-dipping reflectors may arise from sheared rocks with the active roof thrust confined to a much thinner surface within the thicker reflectivity.

Although the two seismic sections displayed in Figures 8a and 8b make an oblique angle to the strike of the downgoing plate, they show the existence of a shallower band of reflections that dip landward and merge into the E reflections beneath the Strait of Georgia. The dip of these D reflections decreases seaward, but they cannot be traced further seaward into the E reflections once more (grey region in Figure 10), perhaps due to high levels of coherent noise. The observations that the D reflections merge into the E reflections and have a similar vertical extent (Figure 8) suggests that the D reflectors also arise from a shear zone, defining another part of the interpreted duplex structure, which we show graphically in Figure 10.

Velocities in the upper mantle between the base of the continental crust and the top of the subducting plate typically range from 7000-7500 m s^{-1}, which may indicate the presence of partially serpentinized mantle peridotite (Bostock et al., 2002, Brocher et al., 2003). However between SP 1500 on line PS-2 and SP 4400 on line SG-1, i.e. in the less reflective unit between the D and E reflectors, P wave velocities are 6800-7000 m s^{-1} (Figure 7). This low velocity region extends 80 km along the strike of the margin, and is sufficiently large to be well-resolved by the tomographic inversion. On the basis of the likely uncertainty in the absolute values of velocity, these rocks could be more highly serpentinized mantle, mafic rocks derived from the subducting slab, mafic rocks from the lower forearc crust, e.g. Crescent terrane, or possibly rocks derived from the sedimentary wedge (Salisbury and Iulucci, 2003). The existence of some earthquake hypocentres in the low velocity rocks at depths as great as 40 km implies that they are anomalous with respect to the rest of the upper mantle, which is notably aseismic. The presence of this seismicity suggests that these rocks are not partially serpentinized peridotite, and implies that they have a crustal origin. Furthermore, the lack of seismicity in the lower part of the accreted sediments suggests that mafic rocks that can support brittle faulting at these depths probably dominate the region between the D and E reflectors.

We suggest two alternative models to explain the formation of the large duplex that has thickened the forearc crust in this region: either underplating of the continental crust with blocks of oceanic crust or subduction erosion of the lower forearc crust:

1) Continental underplating: If the subduction thrust cuts down into the crust of the oceanic plate (dashed line in Figure 11a) as an out-of-sequence thrust, a block of oceanic crust is transferred to the overriding plate. If there is some relative motion between this block and the continent, then the block will be carried downward relative to the continent. (This downward motion is not required to create a thickened forearc crust if the block is detached from the oceanic plate at greater depth.) Subsequently, the subduction thrust again cuts down into the oceanic crust (dashed line in Figure 11b), adding another block to the continent, seaward of the first. If some motion also continues on the former subduction thrust, then the second block will be carried downward, underthrusting the first (Figure 11c).

2) Subduction erosion: Erosion could be initiated by a thrust cutting up into the forearc crust (dashed line in Figure 11d), perhaps controlled by a zone of weakness in the forearc crust due to fluids rising from the subduction plate (thick grey line in Figure 11d) and/or strong coupling between the two plates such as at a large asperity (Wells et al., 2003). If this thrust is the subduction megathrust, a block of lower forearc crust will be transferred to the oceanic plate and carried downward. If this thrust is a new, second thrust, then the enclosed block will move downward relative to the overriding continent, and the oceanic plate will also subduct beneath the block. If the locus of thrusting moves back to the underside of the block, perhaps when it reaches the corner of the forearc mantle, then subduction can proceed beneath a thickened forearc crust. If a new fault cuts up into the forearc crust farther seaward (dashed line in Figure 11e), then a second block will be transported downward in the same fashion. Thrusting along the top of the second block will cause it to be transported downward, underthrusting the first (Figure 11f). In this

model, lower crustal rocks from the Crescent terrane could have been transported eastward and become imbricated in the large duplex structure.

In both these scenarios, we view the first block as corresponding to the transparent zone presently found between the D and E reflectors, and the second block to the transparent zone between the E and F reflectors. At lower crustal depths, it is unlikely that deformation will be accommodated by the motion of rigid blocks along thin faults as might be expected in a brittle upper crust. In the lower crust, the crust will deform in a ductile fashion with strain focussed in certain regions. Thus the thrusts shown in Figure 11 as the D and E reflectors may have evolved over time into highly strained ductile shear zones with deformation distributed vertically over several kilometres, and the less reflective units that they bound may be zones of lower strain. The distribution of strain is controlled by lithology and the distribution of water; for example, felsic sediments in the accretionary wedge are more likely to be the focus of strain than mafic blocks detached either from the oceanic crust or the base of the Crescent terrane. The underplating model proposed above is somewhat similar to that suggested by Clowes et al. (1987a) and Calvert (1996), but includes simultaneous slip both above and below the detached block (Figure 11b). Our model also differs from standard models of duplex formation where only one thrust is active at any one time.

DISCUSSION

In general, the forearc crust is approximately 35 km thick, and the upper forearc mantle probably consists of partially serpentinized peridotite. However, the presence in the mantle of fragments of mafic crust, which have been detached from the subducting oceanic slab or from the lower forearc crust, cannot be excluded on the basis of the seismic tomography if their lateral dimensions are less than 20 km. When crustal rocks are transported downward on a significantly larger scale, the forearc crust may become locally thickened. Near the south end of Vancouver Island, the crustal thickness is approximately 50 km, because the lower crustal blocks of the duplex structure have been transported downward, and lie above the presentday inter-plate boundary. This thickened crust can be identified on the basis of its low P wave velocity relative to the surrounding mantle. If subsequent metamorphism, such as the partial transformation of basaltic rocks to eclogite (metamorphism will probably not be complete at depths as shallow as 40-50 km) occurs, then this local crustal root may become indistinguishable from the surrounding serpentinized upper mantle on the basis of its P wave velocity.

The merging of the D and E reflectors to create a thicker band of reflectivity, for example beneath Puget Sound, suggests that the pervasive reflectivity seen throughout the upper mantle beneath the northern Strait of Georgia also arose from merging shear zones. The continuation into the upper mantle of two zones of crustal reflectors mapped beneath Vancouver Island has been proposed (Clowes et al., 1995), and the reflectivity in Figure 9 could be the expression of the merging of these two zones at approximately 35 km depth. Although the seismic velocities are not well constrained in this region, they are not anomalous with respect to the surrounding mantle. If the reflectivity arose from the downward transport of crustal rocks, as we have inferred further south, then any velocity anomaly appears to have disappeared.

Inslab seismicity occurs in two geographic bands along the western and eastern sides of southern Vancouver Island (Figure 1). These two bands merge further south, and the eastern band only extends to the north part way up the Strait of Georgia. The two bands of inslab seismicity correlate with the two locations where the inter-plate boundary (F reflector) dips most steeply.

Near the west coast of Vancouver Island, this seismicity occurs over a depth range of some 15 km, in both the oceanic crust and mantle. Focal mechanisms of all types occur here, indicating that deformation is quite complex. High fluid pressures generated by dehydration reactions in the subducting plate will reduce the effective normal stress on faults in the subducting plate, and facilitate the occurrence of earthquakes. Near southern Vancouver Island, however, it appears that the geometry of the upper plate is also an important factor (Calvert, 2004). Deformation in the oceanic plate as it accommodates the block trapped between the E and F reflectors may be a trigger for much of this seismicity.

The focal mechanisms are not consistent with simple flexure of the plate, because they do not show a landward change from normal to reverse faulting in the upper part of the plate, and a more complex process, in which the upper part of the plate fragments under oblique convergence might explain the observed mechanisms, which have a significant component of strike-slip motion. The increase in the dip of the plate near the east coast of Vancouver Island might be attributable to the plate deforming around the crustal block between the D and E reflectors or the effect of the increase in the density of oceanic crust undergoing the basalt to eclogite phase transition. The downdip tension indicated by the focal mechanisms here is consistent with the latter hypothesis. However, the northward termination of this eastern band of seismicity could also be a consequence of the limited along-strike extent of the lower crustal duplex structure.

Concentrations of crustal earthquakes occur to depths as great as 33 km (Figure 6); however, it is notable that the maximum depth of crustal seismicity occurs immediately above the shallowest of the E reflectors, and shallows to the west in a similar fashion. The absence of seismicity in the

E reflectors is striking, and suggests that accumulating stress is being released by an aseismic mechanism. The E reflectors are approximately isothermal with a temperature of 400^{o} C, which marks the onset of ductile behaviour in most crustal rocks (Hyndman, 1988). Therefore it is possible that the E reflectors represent active, aseismic ductile deformation occurring above the subducting plate, as has been suggested in small orogens (Beaumont et al., 1994).

The lower apparent amplitude of the D reflections, and the occurrence of earthquakes within them (Figure 7) distinguishes them from the E reflections, and suggests that the D reflections represent a ductile shear zone that is no longer active, consistent with both our models of evolution of the duplex described earlier. Although the origin of non-volcanic tremor activity is still unknown, the occurrence of tremors within the E reflectors demonstrates that the source mechanism is different from local earthquakes. An association of tremor activity with fluid flow has been suggested (Kao et al., 2005), and might be supported by the inference of connected fluids within the E reflectors by magnetotelluric investigations (Kurtz et al., 1986) if there were a greater density of tremor along the reflectors. However, most tremor is distributed over a broad depth range (Kao et al., 2005), and such an association remains equivocal. The region of the northwestward, along-strike migration of tremor corresponds approximately with the location of the lower crustal duplex structure. We therefore tentatively suggest that the existence of this crustal root may control the segmentation of slip along the Cascadia margin, causing slow slip on the megathrust to occur at different times both north and south of the thickened forearc crust.

CONCLUSIONS

Deep seismic reflection profiles show the existence of reflectors in the corner of the continental mantle wedge of the northern Cascadia subduction zone. In some locations, reflectivity extends throughout the uppermost mantle; in others, it is confined to 5-10 km thick zones that are interpreted to be either active or former locations of distributed ductile shear. We suggest that the cause of the upper mantle reflectivity arises from shearing associated with the downward tectonic transport of crustal rocks derived either from the subducting plate or from the lower forearc crust. A low velocity zone in the uppermost mantle extending from the southern Strait of Georgia to northern Puget Sound indicates that such transported crustal rocks have locally thickened the forearc crust to 50 km above the descending oceanic plate.

Near southern Vancouver Island, seismic reflectors define a >100 km long duplex structure that may have an along-strike extent of 80 km. We speculate that the landward part of the duplex is no longer actively deforming, and that the duplex has grown seaward with both the aseismic E reflectors and the F reflector at the top of the descending slab now being locations of active thrusting. Seismicity in the subducting slab occurs where the dip of the top of the plate increases to 15^{o} as it deforms around imbricated crustal rocks that have been transported downward above the plate.

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REFERENCES CITED

Beaumont, C., Fullsack, P. and Hamilton, J., 1994, Styles of deformation in compressional orogens caused by subduction of the underlying lithosphere. Tectonophysics, 232, 119-132.

Bostock, M.G., Hyndman, R.D., Rondenay, S. and Peacock, S.M., 2002. An inverted continental Moho and the serpentinization of the forearc mantle. Nature, 417, 536-538.

Brocher, T.M., Parsons, T. Tréhu, A.M., Snelson, C.M. and Fisher, M.A., 2003. Seismic evidence for widespread serpentinized forearc upper mantle along the Cascadia margin, Geology, 31, 267-270.

Calvert, A.J. and Clowes, R.M., 1990. Deep, high amplitude reflections from a major shear zone above the subducting Juan de Fuca plate. Geology, 18, 1091-1094.

Calvert, A.J., 1996. Seismic reflection constraints on imbrication and underplating of the northern Cascadia convergent margin. Can. J. Earth Sci., 33, 1294-1307.

Calvert, A.J., Fisher, M.A., Ramachandran, K. and Tréhu, A.M., 2003, Possible emplacement of crustal rocks into the forearc mantle of the Cascadia subduction zone. Geophys. Res. Lett., 30, 2196, doi:10.1029/2003GL018541.

Calvert, A.J., 2004. Seismic reflection imaging of two megathrust shear zones in the northern Cascadia subduction zone. Nature, 428, 163-167.

Clowes R.M., Brandon, M.T., Green, A.G., Yorath, C.J., Sutherland Brown, A., Kanesewich, E.R. and Spencer, C., 1987a. LITHOPROBE - southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections. Can. J. Earth Sci., 28, 542-556.

Clowes R.M., Yorath, C.J. and Hyndman, R.D., 1987b. Reflection mapping across the convergent margin of western Canada. Geophys. J.R. astr. Soc., 89, 79-84.

Clowes R.M., Zelt C.A., Amor J.R., and Ellis R., 1995. Lithospheric structure in the southern Canadian Cordillera from a network of seismic refraction lines. Can. J. Earth Sci., 32, 1485-1513.

Dragert, H., Wang, K. and James, T.S., 2001. A silent slip event on the deeper Cascadia subduction interface. Science, 292, 1525-1528.

Engebretson, D.C., Gordon, R.G. and Cox, A., 1985. Relative motions between oceanic and continental plates in the Pacific Basin. Geol Soc. Am. Spec. Pap. 206.

Fisher, M.A., and 19 others, 1999. Seismic survey probes urban earthquake hazards in Pacific Northwest. Eos, 80, 13-17.

Fisher, M.A., Hyndman, R.D., Johnson, S.Y., Brocher, T.M., Wells, R.E., Calvert, A.J., ten Brink, U.S., and the SHIPS working group, 2005. Crustal structure and earthquake hazards of the subduction zone in southwestern British Columbia and western Washington state. U.S., Prof. Paper 1661-B, U.S. Geol. Surv., Reston, in press.

Green, A.G., Clowes, R.M., Yorath, C.J., Spencer, C., Kanesewich, E.R., Brandon, M.T. and Sutherland Brown, A., 1986. Seismic reflection imaging of the subducting Juan de Fuca plate. Nature, 319, 210-213.

Hacker, B.R., Peacock, S.M., Abers, G.A. and Holloway, S.D., 2003. Subduction factory 2. Are intermediate-depth earthquakes linked to metamorphic dehydration reactions? J. Geophys. Res., 108, doi:10.1029/2001JB001129.

Haeussler, P.J., Bradley, D.C., Wells, R.E., Miller, M.L., 2003. Life and death of the Ressurrection plate : Evidence for its existence and subduction in the northeastern Pacific in Paleocene-Eocene time. Bull. Geol. Soc. Am., 115, 867-880.

Hyndman, R.D., 1988. Dipping reflectors, electrically conductive zones and free water beneath a subduction zone. J. Geophys. Res. 93, 13391-13405.

Kao, H. and Shan, S.-J., 2004. The source-scanning algorithm : mapping the distribution of seismic sources in time and space. Geophys. J. Int., 157, 589-594.

Kao, H., Shan, S.-J., Dragert, H., Rogers, G., Cassidy, J.F. and Ramachandran, K., 2005. A wide depth distribution of seismic tremors along the northern Cascadia margin. Nature, 436, 841-844.

Kurtz, R.D., De Laurier, J.M. and Gupta, J.C., 1986. A magnetotelluric sounding across the subducting Juan de Fuca plate. Nature, 321, 596-599.

Nedimovi_, M.R., Hyndman, R.D., Ramachandran, K. and Spence, G.D., 2003. Reflection signature of seismic and aseismic slip on the northern Cascadia subduction interface. Nature, 424, 416-420.

Parsons, T., Trehu, A.M., Luetgert, J.H., Miller, K., Kilbride, F., Wells, R.E., Fisher, M.A., Flueh, E., ten Brink, U.S., Christensen, N.I., 1998. A new view into the Cascadia subduction zone and volcanic arc: Implications for earthquake hazards along the Washington margin. Geology, 26, 199-202.

Peacock, S.M., 2001. Are the lower planes of double seismic zones caused by serpentine dehydration in subducting oceanic mantle? Geology, 29, 299–302.

Preston, L.A., Creager, K.C., Crosson, R.S., Brocher, T.M. and Tréhu, A.M., 2003. Intraslab earthquakes: Dehydration of the Cascadia slab. Science, 302, 1197-1200.

Ramachandran, K., 2001. Velocity structure of S.W. British Columbia and N.W. Washington from 3-D non-linear seismic tomography, Ph.D. dissertation, University of Victoria, Victoria.

Ramachandran, K., Hyndman, R.D., & Brocher, T.M., 2005. Northern Cascadia subduction zone: Tomographic 3-D P wave velocity structure. Geophys. J. Int., submitted.

Rogers, G. and Dragert, H., 2003. Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip, Science, 300, 1942-1943.

Salisbury, M., and R. Iulucci, 2003. GSC Atlantic / Dalhousie University rock properties database, http:// www.gsca.nrcan.gc.ca/pubprod/rockprop, Geol. Surv. Can., Dartmouth.

Spence, G.D., Clowes, R.M. and Ellis, R.M., 1985. Seismic structure across the active subduction zone of western Canada. J. Geophys. Res. 90, 6754-6772.

Spence, G.D., Hyndman, R.D., Langton, S.G., Davis, E.E. and Yorath, C.J., 1990. Multichannel seismic reflection profiles across the Vancouver Island continental shelf and slope. Geol. Surv. Can. Open File Rep. 2391.

Wells, R.E., Blakely, R.J., Sugiyama, Y., Scholl, D.W. and Dinterman, D.A., 2003. Basin-centered asperities in great subduction zone earthquakes: A link between slip, subduction and subduction erosion. J. Geophys. Res. 108, doi:10.1029/2002JB002072.

Yorath, C.J., Green, A.G., Clowes, R.M., Sutherland Brown, A., Brandon, M.T., Kanesewich, E.R., Hyndman, R.D., and Spencer, C., 1985. Lithoprobe, southern Vancouver Island: Seismic reflection sees through Wrangellia to the Juan de Fuca plate. Geology, 13, 759-762.

Yorath, C.J. *et al*., 1987. Marine multichannel seismic reflection, gravity and magnetic profiles – Vancouver Island continental margin and Juan de Fuca ridge. Geol. Surv. Can. Open File Rep. 1661, 26 pp.

FIGURE CAPTIONS

Fig 1. Map of the northern Cascadia subduction zone showing the locations of the two composite seismic sections presented in this paper. A dip section, whose orientation is indicated by the dashed green line, has been simulated by projecting the red profile onto an azimuth of 063^{o}. The yellow profile represents a near-strike section along the margin. The location of other seismic sections presented in later figures are shown in white. Earthquakes within the subducting Juan de Fuca subducting plate are marked by filled black circles.

Fig. 2. Unmigrated seismic line 85-01 over the continental shelf and slope near Vancouver Island showing the top of the subducting plate (JdF), the E reflection band, and the F reflector. Beneath the deformation front, the top of the subducting igneous crust is marked by the strong, landward-dipping reflection at around 5.5 s that lies just above several diffraction events. Beneath the continental shelf, the inter-plate boundary is indicated by the intermittent reflections marked JdF. At the landward end of the line the 2 s thick band of E reflections lies above the thinner F reflection.

Fig. 3. Line 89-01, just to the south of line 85-01, showing that the deeper zone of E reflections merge westward with the top of the inter-plate boundary, JdF (lower white line). The upper part of the E reflections merge into shallower structures of the accretionary wedge (upper white line). The lower part of the package of JdF reflections appears to continue into the dipping F reflection,

indicated by the arrow. F is interpreted to be the top of the subducting oceanic plate. The section is unmigrated.

Fig. 4. Composite, unmigrated seismic section near the west coast of Vancouver Island showing that the shallowest of the inslab earthquakes follow the geometry of the top of the subducting Juan de Fuca plate (yellow line). Earthquake hypocentres within 15 km of the seismic profiles are shown by black filled circles. The seismic data are unmigrated, but the position of the top of the plate is derived from a migrated section, indicating that the position of the interface does not move a great distance after migration due to its relatively low dip. The orientation of the original line segments used to construct the composite section are shown at the top. Oceanic Moho assuming subducting crust is 6 km thick – dashed blue line. Vertical exaggeration is 1.5:1.

Fig. 5. Composite dip reflection section (unmigrated) across Vancouver Island. Near the west coast of Vancouver Island the top of the subducting plate is marked by the JdF and F reflections. The F reflection is mostly continuous beneath the western half of Vancouver Island when viewed at high gain, but appears as discrete segments beneath the eastern half of the island, and probably merges into the deepest of the E reflections.

Fig. 6. Interpreted section normal to Cascadia margin. The migrated dip section is superimposed on interval velocities extracted from the 3-D P wave model of Ramachandran (2001) along the seismic line. The interpreted top of the subducting plate dips at approximately 15^o near the west coast of Vancouver Island, dips shallowly at around 4^o beneath Vancouver Island, and then increases in dip to 15^o once more near the east coast. Also shown are earthquake hypocentres (filled black circles), and the locations of tremors (coloured black circles) that occurred during the 2003 episodic tremor and slip event. Focal mechanisms indicate that inslab earthquakes near the west coast of Vancouver Island exhibit a variety of mechanisms, many with a component strike-slip, while those near the east coast are dominated by downdip tension. Red – Leech River fault (LRF), brown – top of E reflectivity, green – base of E in imbricated region, yellow – top of subducting plate, blue – oceanic Moho, dashed where inferred.

Fig. 7. Interpreted strike section along Cascadia margin through Puget Sound and Georgia Strait. The unmigrated strike section is superimposed on interval velocities extracted from the 3-D velocity model of Ramachandran (2001) along the seismic line. Also shown are earthquake hypocentres (filled black circles), and the locations of tremors (white black circles) that occurred during the 2003 episodic tremor and slip event. Focal mechanisms indicate that inslab earthquakes are dominated by downdip tension. Only tremors for February 27th are shown, as subsequent tremor activity had migrated to the northwest away from the seismic line. Brown – top of deep reflectivity, dashed green – edges of deep unreflective region or discontinuities in reflectivity due to merged shear zones, yellow – top of subducting plate.

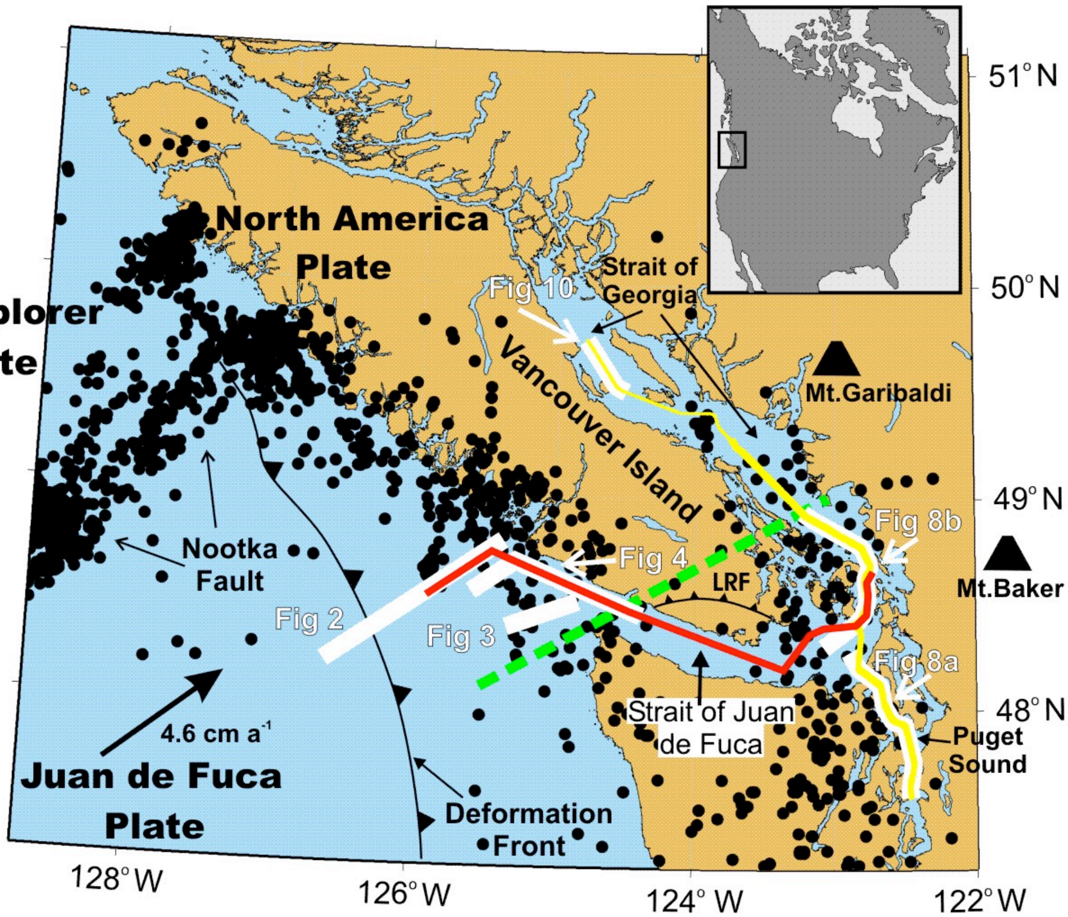
Fig. 8. Anastamosing reflection packages at 35-50 km depth near the southeast corner of Vancouver Island. (a) Line PS-2 (unmigrated) through Puget Sound shows the 2 s thick E reflections, with the F reflection at their base, increasing in depth landward. Another approximately 2 s thick band of reflections, denoted D, overlies the E reflections and merges into them to create a broad zone of reflectivity at 12-16 s. (b) Line SG-1 (unmigrated), which lies to the north of PS-2, shows a similar anastomosing reflector geometry above the subducting plate. South and north are reversed between the two plots to show the similar geometry of the two anastomosing reflection packages. Brown – upper limit of deep reflectivity, dashed green – edges of unreflective region, yellow – F reflector at top of subducting plate

Fig. 9. Reflections from the lowermost crust and upper mantle at north end of line SG-1 (unmigrated) through the Strait of Georgia. The sedimentary cover is less than 1 s thick, and the underlying upper and mid crust is largely unreflective. However, laterally continuous reflections can be observed from 9 s to the base of the section at 16 s, which lies well within the upper mantle.

Fig. 10. Schematic cross-section normal to the Cascadia margin at the location of southern Vancouver Island. The E and F reflectors define the roof and floor thrusts respectively of a 100 km wide duplex structure, beneath which the Juan de Fuca plate subducts. The D reflectors may

also be part of the roof thrust, but have not yet been shown to be continuous with the E reflectors, as indicated by the grey region. Seismicity in the subducting slab occurs primarily where the top of the plate is inferred to steepen.

Fig. 11. Cartoons illustrating two alternative models for formation of thickened forearc crust: underplating of the continental crust by out of sequence thrusting or erosion of the lower forearc crust. At depths greater than 20km, some thrusting may be distributed over shear zones that are several km thick, and crustal blocks are likely to deform in a ductile fashion.. a) Subduction thrust cuts down into the crust of the oceanic plate (dashed line). Oceanic crust is transferred to the overriding plate, but if there is some relative motion between this block and the continent, then the block will be carried downward relative to the continent. b) Subsequently, the subduction thrust cuts down again into the oceanic crust (dashed line), adding another block to the continent, seaward of the first. c) If some motion also continues on the former subduction thrust, then the second block will be carried downward, underthrusting the first. d) In the case of subduction erosion, a thrust cuts up into the forearc crust (dashed line), perhaps controlled by a zone of weakness in the forearc crust due to fluids rising from the subduction plate (thick grey line). If this thrust is the subduction megathrust, a block of lower forearc crust will be transferred to the oceanic plate and carried downward. If the this thrust is a new, second thrust, then the enclosed block will move downward relative to the overriding continent, and the oceanic plate will also subduct beneath the block. e) If the locus of thrusting moves back to the underside of the block when it reaches the corner of the forearc mantle, then subduction can proceed beneath a thickened forearc crust. If a new fault cuts up into the forearc crust farther seaward (dashed line), then a second block will be transported downward in the same fashion. f) Thrusting along the top of the second block will cause it to be transported downward, underthrusting the first. The location of the reflectors identified in the seismic data is identified; the presentday subduction megathrust comprises JdF, F, and the merged E and F reflectors beneath the eastern edge of Vancouver Island.



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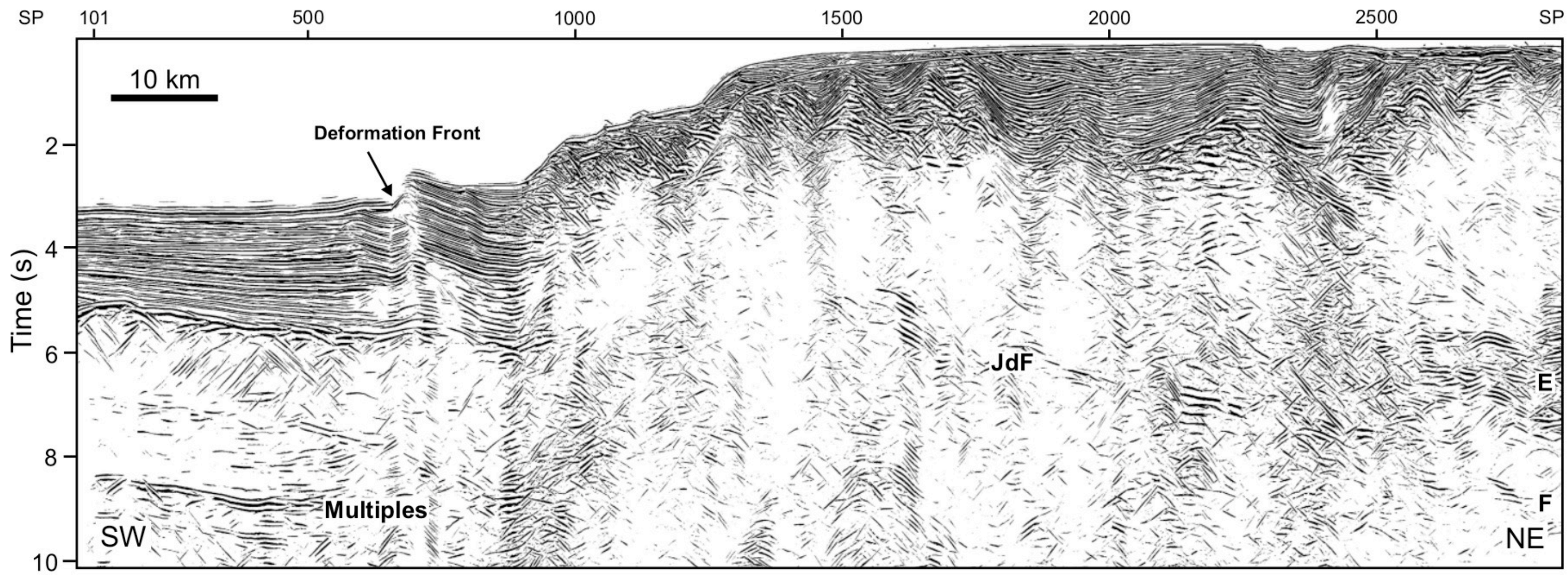

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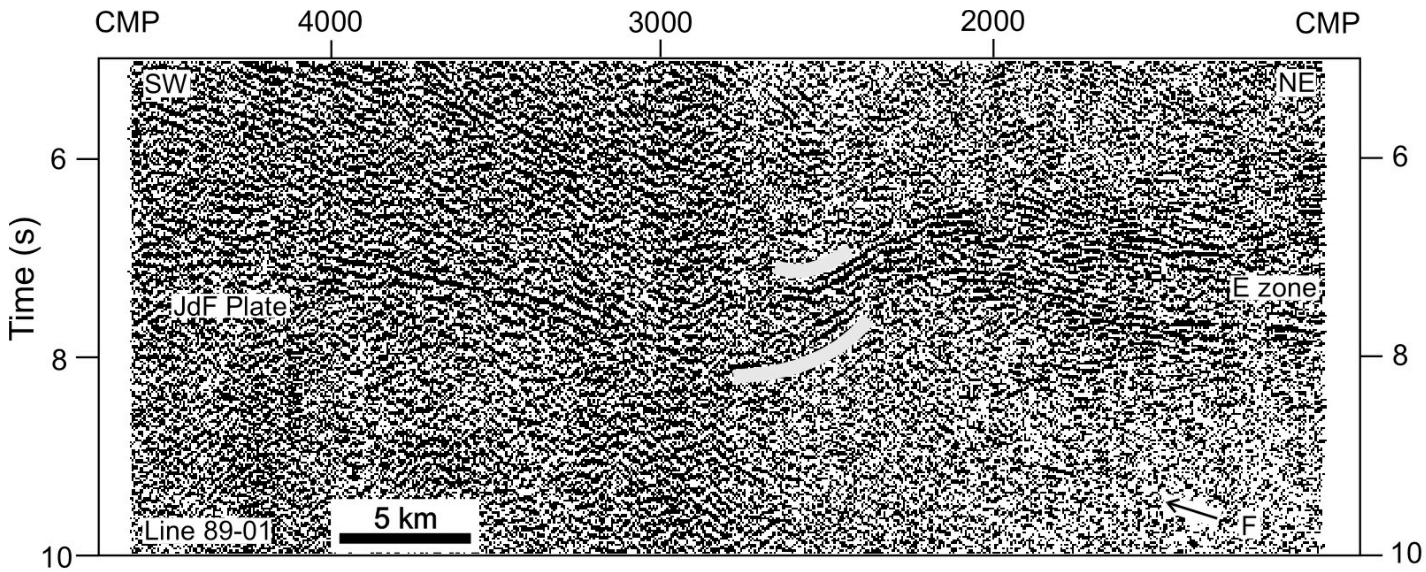

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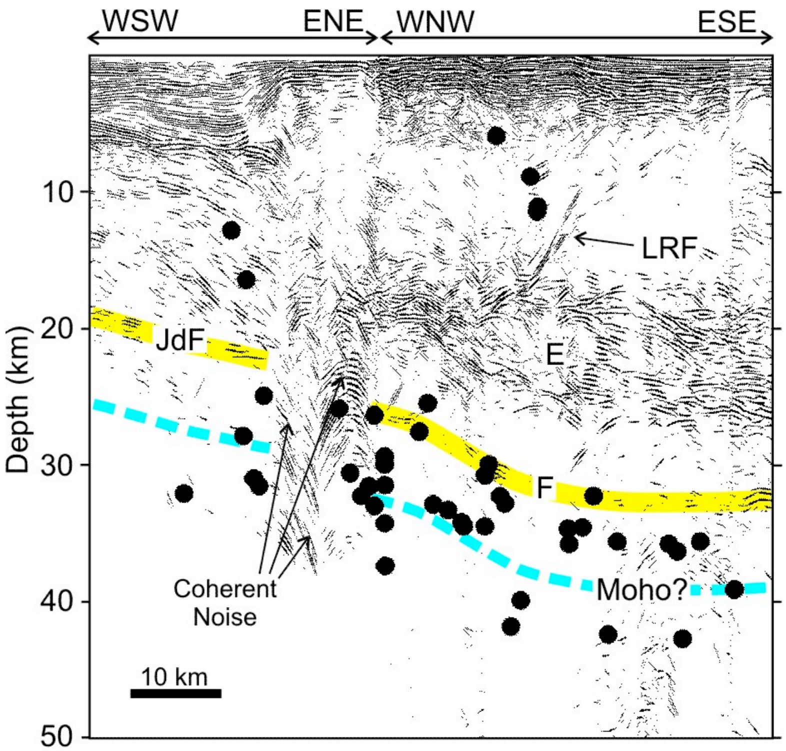

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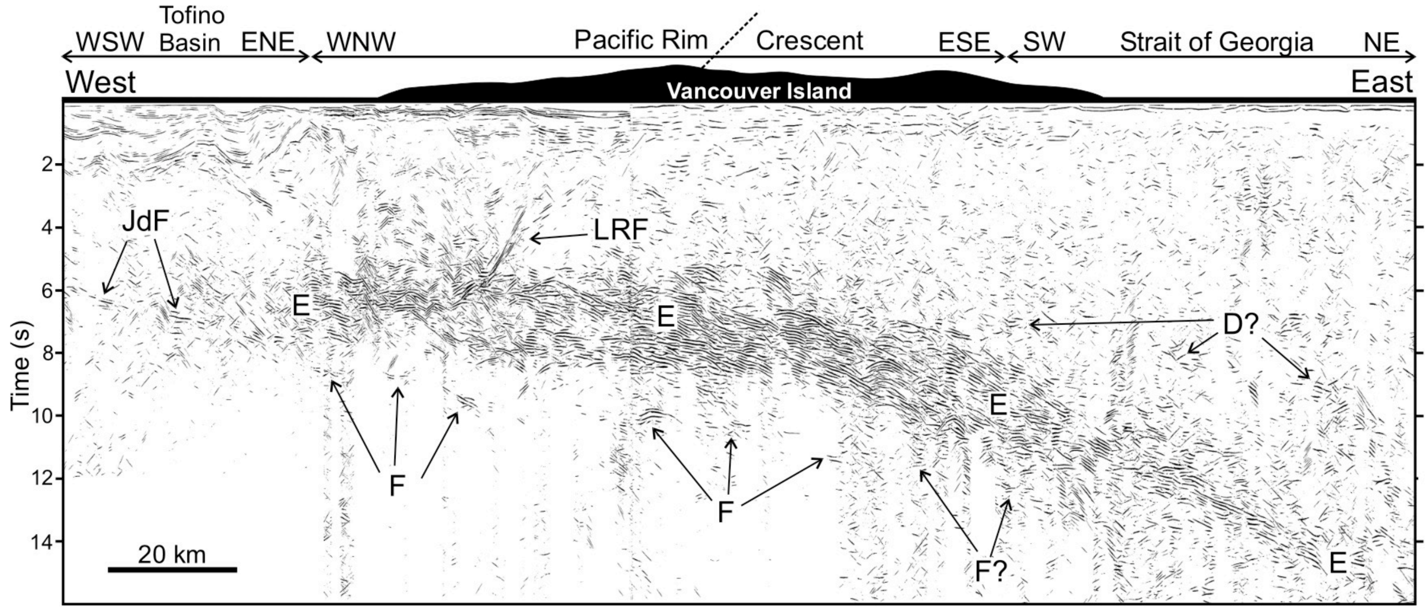

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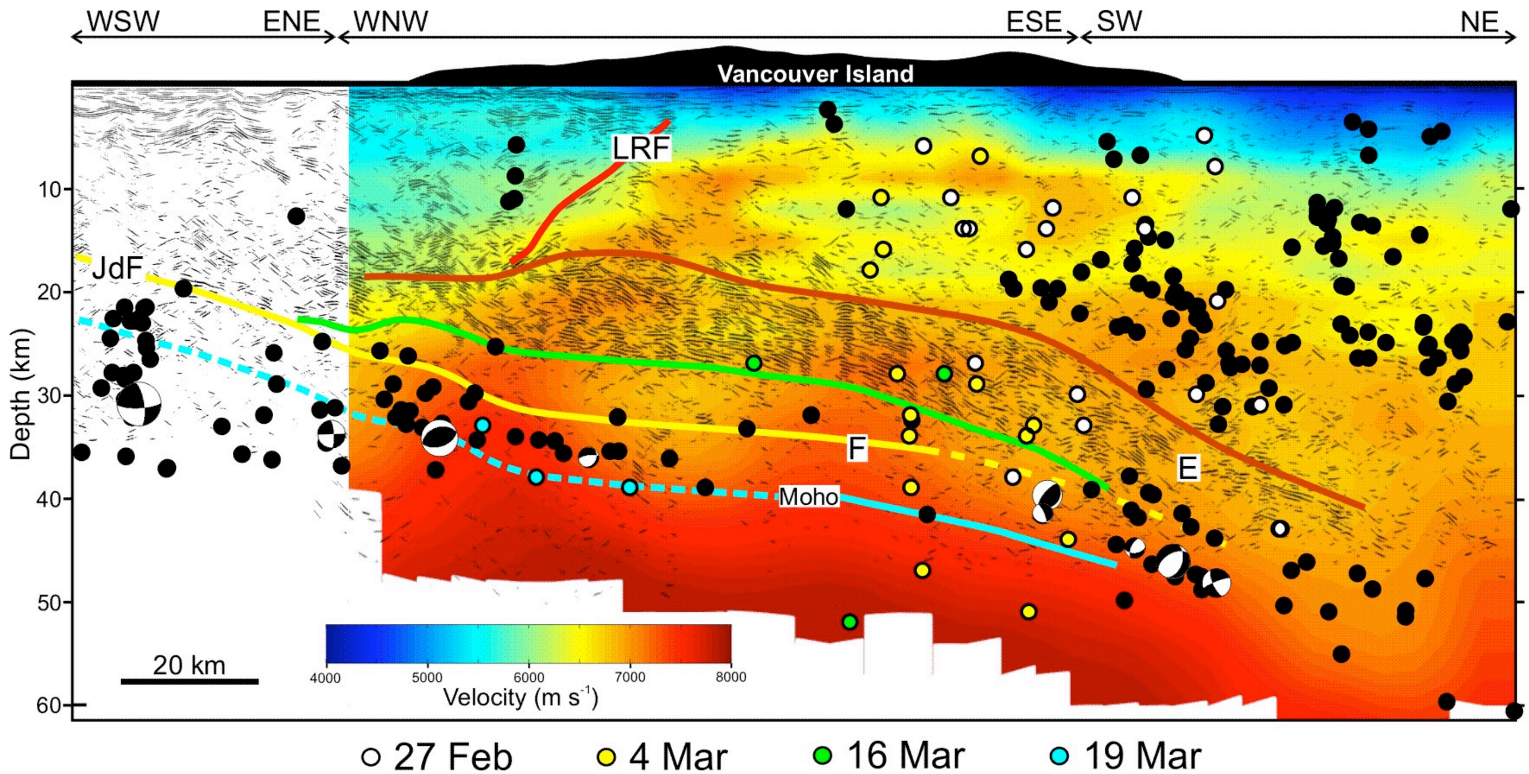

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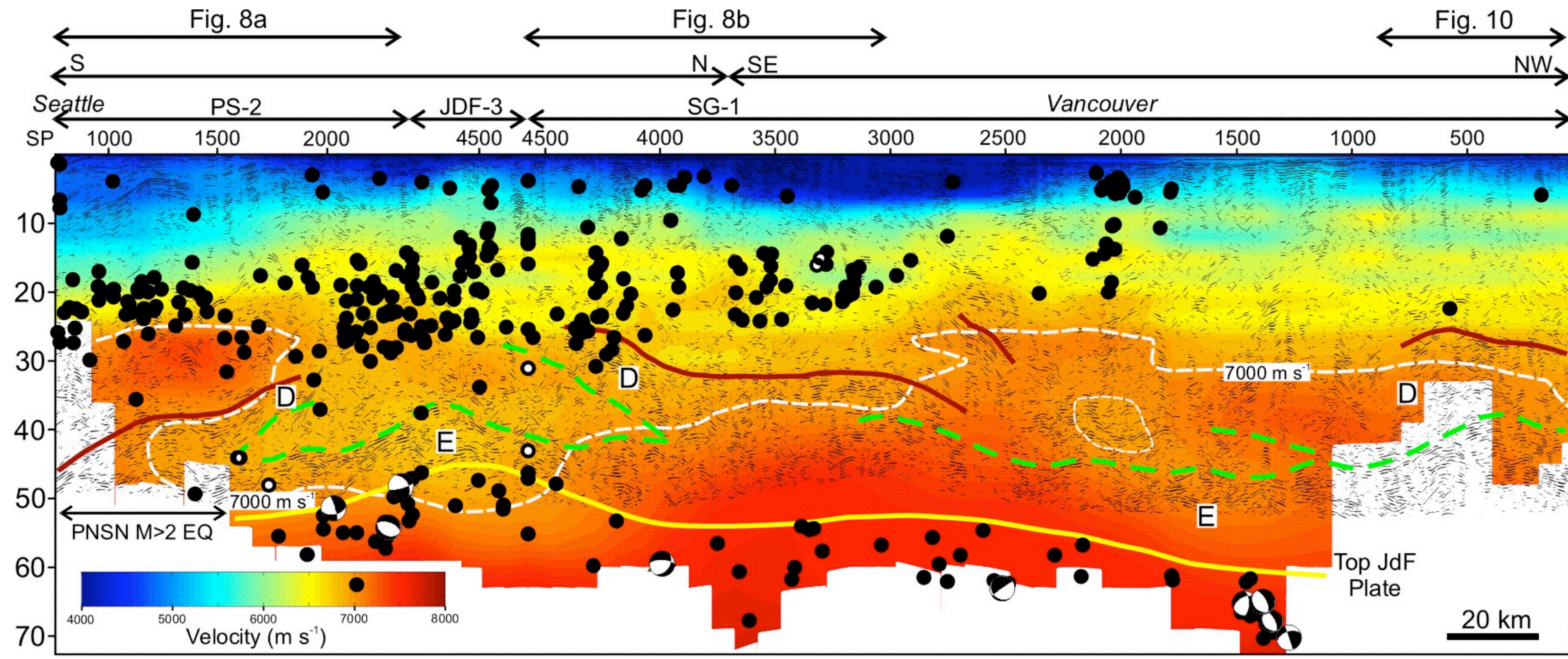

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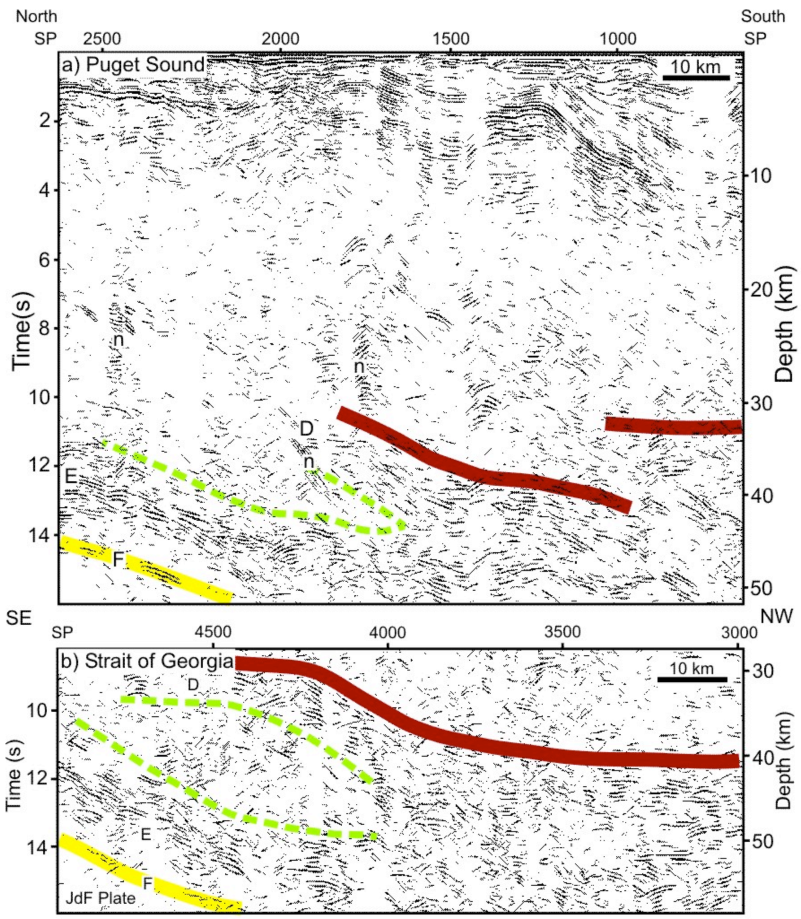

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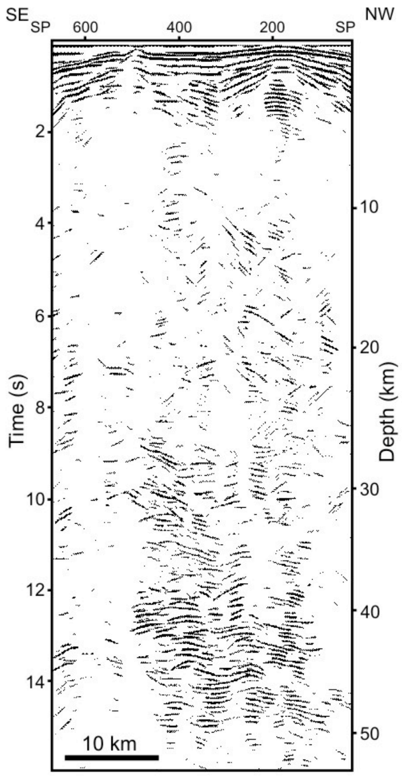

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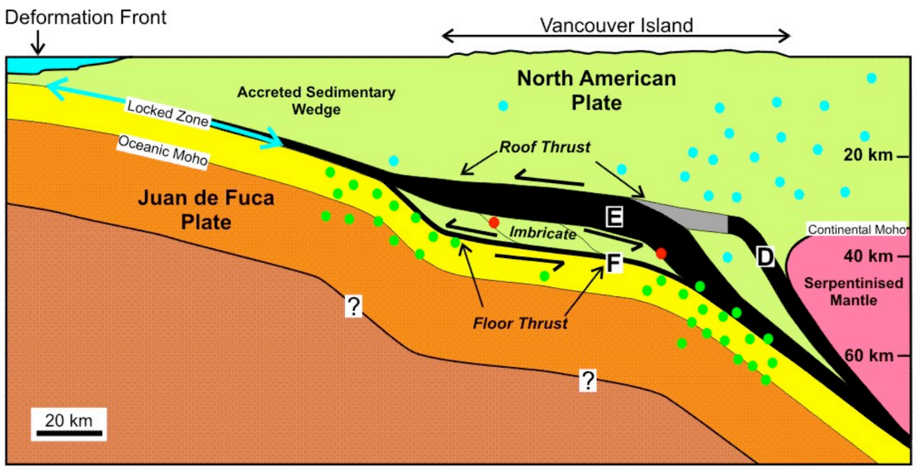

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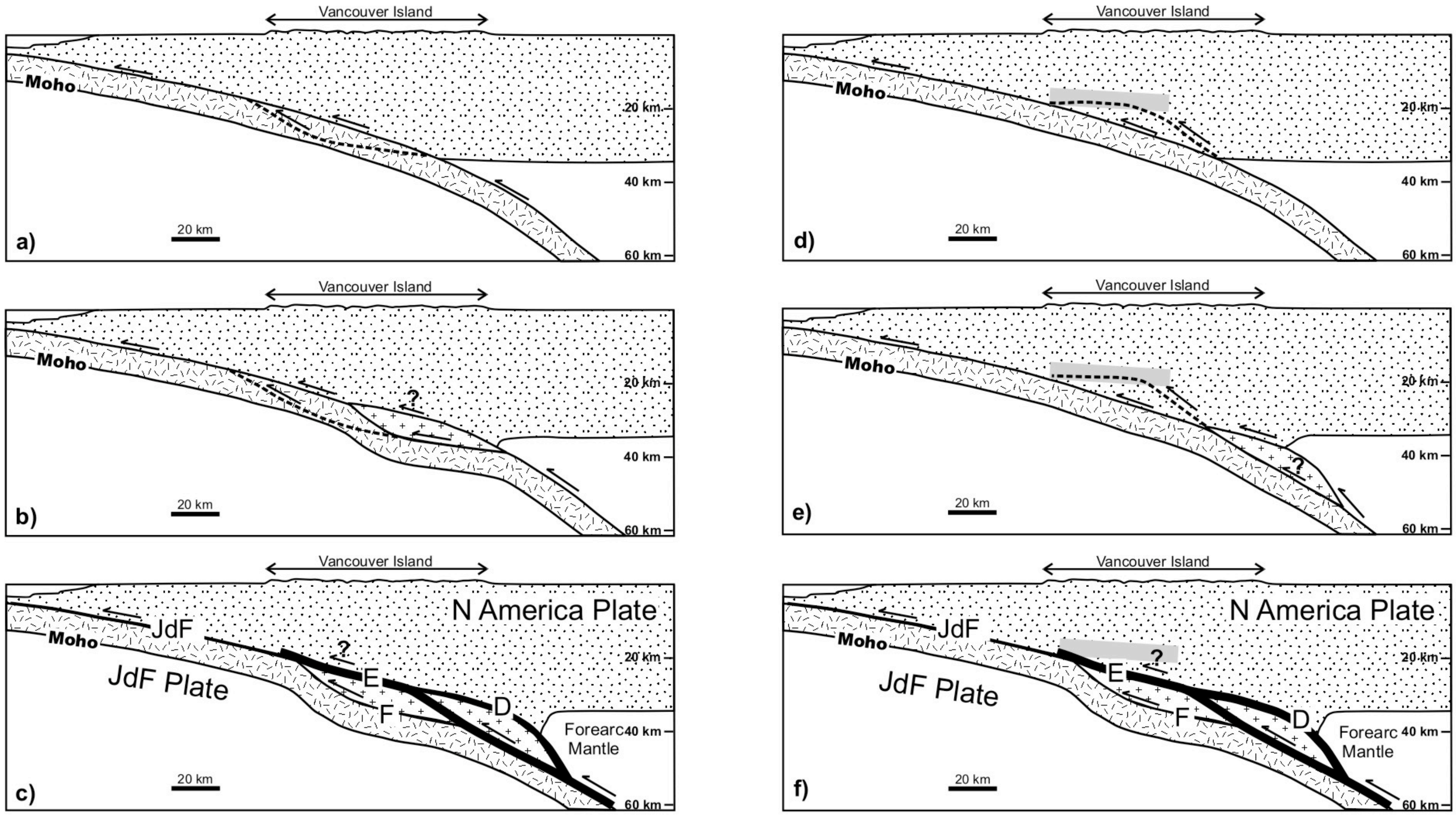

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