Morphology of the Explorer–Juan de Fuca slab edge in northern Cascadia: Imaging plate capture at a ridge-trench-transform triple junction

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ABSTRACT

The Explorer plate is a young oceanic microplate that accommodates relative motion between the Pacific, Juan de Fuca, and North America plates near northern Vancouver Island, Canada. The northern limit of Explorer plate–Juan de Fuca subduction and the fate of the slab in northern Cascadia are poorly understood. We use passive teleseismic recordings from an array of POLARIS broadband seismic stations to image crustal and upper mantle structure beneath northern Vancouver Island into the interior of British Columbia. A clear signature of subducted material extends northeast from the Brooks Peninsula at crustal levels, beneath Georgia Strait and the mainland deep into the mantle to 300 km depth. Complexity in slab morphology results from Juan de Fuca ridge subduction and toroidal flow around the slab edge, in agreement with geophysical and geological data. We propose a tectonic model for the Explorer plate in which its separation from the Juan de Fuca plate is caused by the thermomechanical erosion of the slab edge and slab thinning at shallow levels, both of which slow convergence with North America and lead eventually to plate capture.

INTRODUCTION

The northern end of the Cascadia subduction zone is characterized by the interaction of three oceanic plates (Pacific, Juan de Fuca, and Explorer) with North America, where they meet in a set of poorly defined triple junctions (Fig. 1) (Braunmiller and Nábělek, 2002). The Explorer plate is a small oceanic fragment that detached from the subducting Juan de Fuca plate ca. 4 Ma during creation of the left-lateral Nootka transform fault located offshore Nootka Island, south of the Brooks Peninsula (Rohr and Furlong, 1995). Since Explorer inception, the Explorer ridges have been rotating clockwise and spreading centers have jumped northward, slowly reducing convergence with North America (Braunmiller and Nábělek, 2002). The Explorer region is currently undergoing major internal shear deformation as an ephemeral adjustment accommodating relative motion between the Pacific and North America plates (Rohr and Furlong, 1995; Kreemer et al., 1998, Dziak, 2006).

Explorer microplate evolution is an active example of an important process associated with ridge collision and subduction that has occurred throughout geological history, typically leaving few geological records (Stock and Lee, 1994). Given the strong dependence of plate reconstructions on oceanic data, it is critical that we understand processes associated with microplate formation. To gain insight into the causes of plate fragmentation, we need to consider the factors controlling the balance between driving and resistive forces acting at plate boundaries. In a subduction setting the most effective driving force is the pull exerted by the subducted slab (Forsyth and Uyeda, 1975; Govers and Meijer, 2001). Hence much could be learned about Explorer microplate evolution by exploring its structure at crustal and upper mantle levels beneath northern Vancouver Island and landward. Unfortunately, little is known about subducted slab structure in northern Cascadia. Indirect evidence for the location of the slab edge from heat flow, gravity, and geochemical data loosely defines its surficial projection along a NE-trending corridor landward and parallel to the Brooks Peninsula (Lewis et al., 1997). However, those data lack resolution along the third (depth) dimension that is necessary to constrain morphology of the subducting slab.

Cassidy et al. (1998) used teleseismic data recorded at an array of five stations sparsely deployed across northern Vancouver Island to provide the first direct evidence for a dipping low-velocity zone up to the Brooks Peninsula, and the resumption to normal continental-like seismic signature to the north. Here we use seismic data from a recently deployed

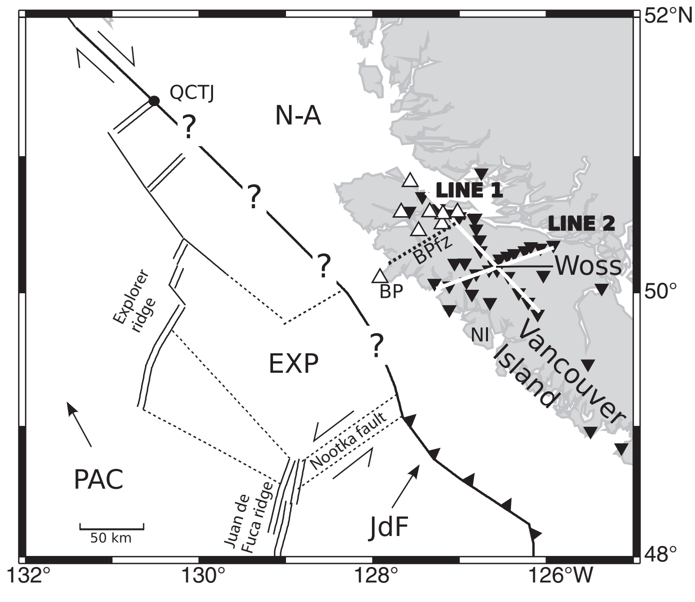

Figure 1. Identification of major tectonic features in western Canada. BP—Brooks Peninsula, BPfz—Brooks Peninsula fault zone, NI—Nootka Island, QCTJ—Queen Charlotte triple junction. Dotted lines delineate extinct boundaries or shear zones. Seismic stations are displayed as inverted black triangles. Station projections along line 1 and line 2 are plotted as thick white lines. White triangles represent Alert Bay volcanic field centers. Center of array locates town of Woss. Plates: N-A—North America; EXP—Explorer; JdF—Juan de Fuca; PAC—Pacific.

POLARIS portable array together with data from permanent stations to map the morphology of the subducting slab in the forearc region beneath northern Vancouver Island and the interior of western British Columbia using receiver functions and P-wave traveltime tomography.

SHALLOW STRUCTURE FROM RECEIVER FUNCTIONS

The POLARIS–British Columbia Northern Vancouver Island (NVI) array comprises 26 broadband seismometers along two mutually perpendicular arms (Fig. 1). One arm trends NW-SE in a direction parallel to strike and straddles the assumed northern end of the subduction zone (line 1), and the second arm trends SW-NE in the dip direction, just north of the extension of the Nootka fault beneath Vancouver Island where convergence is observed (line 2) (Mazzotti et al., 2003). The array has been in operation since June 2005 and to date each station has recorded an average of 50 teleseismic events with high signal-to-noise ratio. The NVI data set is complemented by recordings from a few permanent stations of the Geological Survey of Canada and a previous experiment (Cassidy et al., 1998).

All receiver functions filtered between 0.05 and 0.35 Hz are plotted in raw form along each line according to station position and, for individual stations, sorted by back azimuth (Fig. 2; see the GSA Data Repository[1] for details on receiver function method). These data are most sensitive to structures with scale lengths of 1–10 km and are dominated across both profiles by the signature of a low-velocity zone. This signature comprises three sets of oppositely polarized pulses representing forward scattered P to S (Ps) (3–5 s at station VI57) conversions, and backscattered P to S (Pps) (9–14 s at VI57) and S to S (Pss) (15–20 s at VI57) conversions afforded by reflection of the teleseismic wave field at the Earth's free surface. These signals are interpreted as oceanic crust of the subducting Juan de Fuca–Explorer slab, consistent with its expression in studies farther south beneath Vancouver Island (Nicholson et al., 2005), Oregon (Bostock et al., 2002), and worldwide (e.g., Abers, 2005). Along line 1 the low-velocity zone is evident from VI57 to VI52, and disappears farther north. The seismic response there is more similar to that of a single discontinuity at a typical continental crust-mantle boundary (Moho) with positive Ps and Pps arrivals at ~4 s and ~15 s, and a negative Pss pulse at ~22 s. A low-velocity zone is inferred at station PHC, although it lacks clear Ps signals and its relation to subducted oceanic crust to the south is unclear. Structural signals along line 2 show evidence of a well-defined low-velocity zone dipping gently NE along the entire profile that is most easily seen in oppositely polarized reverberations (Pps, Pss) at ~10–20 s. Oppositely polarized Ps signals are clearly imaged from stations CHM to VW01, whereafter they show polarity crossovers, as seen by the blue-red checkerboard pattern at ~3 s from stations VI09 to VI07. This polarity reversal with respect to the back azimuth of the incident wave field manifests elastic anisotropy. This change in elastic symmetry also coincides with a shallowing of the low-velocity zone signature signaled by earlier arrivals of reverberated phases (Pps, Pss). Dipping low-velocity zone signals resume thereafter to station VI01, where the signature disappears.

The timing of scattered modes relative to P (0 time in Fig. 2) can be used to characterize both thickness and average Vp/Vs (V is velocity) of the overlying column by assuming a dipping planar geometry of the subsurface. This is accomplished by stacking waveforms of the three scattering modes with time delays that correspond to the propagation of plane waves through a range of models (Zhu and Kanamori, 2000). Using this technique we determined the depth to top and bottom

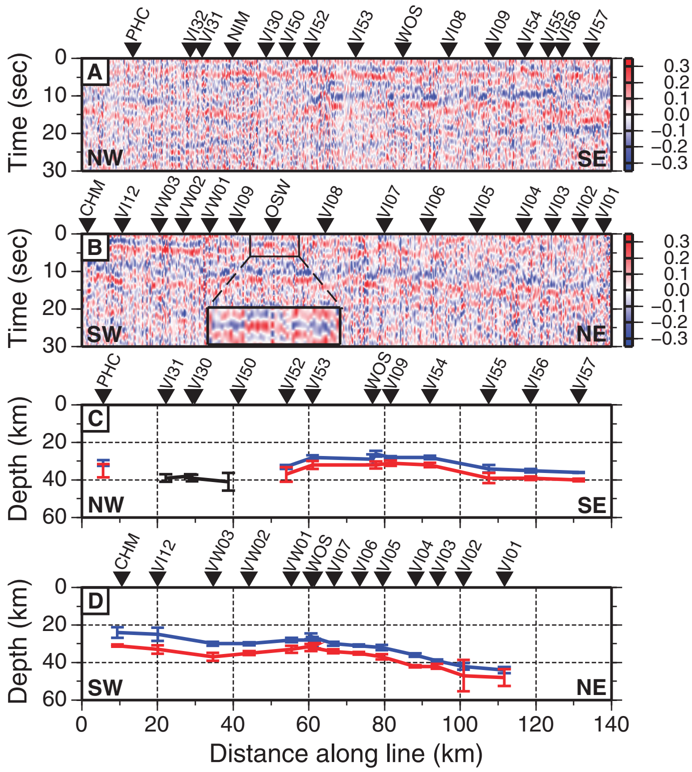

Figure 2. A, B: Raw receiver functions along line 1 (A) and line 2 (B) sorted by station location and, for each station, by back azimuth of incident wave field. Inset in B shows enlargement of first 6 s of Ps conversions beneath Woss (station WOS) to illustrate polarity reversals that are indicative of anisotropy. C, D: Best-fit estimates of depths to top (blue line) and bottom (red line) of subducted oceanic crust and crust-mantle discontinuity (Moho—black line) along both lines. Error bars are calculated as in Zhu and Kanamori (2000). See the GSA Data Repository (see footnote 1) for details.

of the low-velocity zone and mapped the morphology of the subducted oceanic crust along both lines (Fig. 2) and across northern Cascadia using a larger data set (Fig. 3).

DEEP STRUCTURE FROM P-WAVE TOMOGRAPHY

Deep slab structure (50–300 km depth) is imaged using P-wave traveltime tomography of the upper mantle in the Pacific Northwest using the method described in Bostock and VanDecar (1995). For this component of the study, data were collected over a much broader network comprising broadband and short period seismic stations from several Canadian and U.S. portable and permanent arrays (Fig. 3). We derived 13,595 P-wave traveltimes from 738 earthquakes, with good source coverage in back azimuth and epicentral distance. The data set was inverted for upper mantle structure below Washington and western British Columbia, extending the coverage much farther north than a previous study (Bostock and VanDecar, 1995). Resolution tests performed with a synthetic slab model extending along the full margin of northwestern North America indicate that deep structure is well resolved by the data.

The velocity model is characterized by a quasi-planar, high-velocity layer steeply dipping beneath British Columbia to depths of at least 300 km (Fig. 3; see the Data Repository). This material is inferred to represent the thermal and compositional anomaly of the subducted Juan de Fuca plate. The slab signature appears to be continuous with a slight change in strike from Washington to northern Vancouver Island, where the signature ends abruptly.

[1]GSA Data Repository item 2008228, details on analysis, and Figures DR1–DR6 (receiver functions, phase stacking, and cross sections of tomographic model), is available online at www.geosociety.org/pubs/ft2008.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

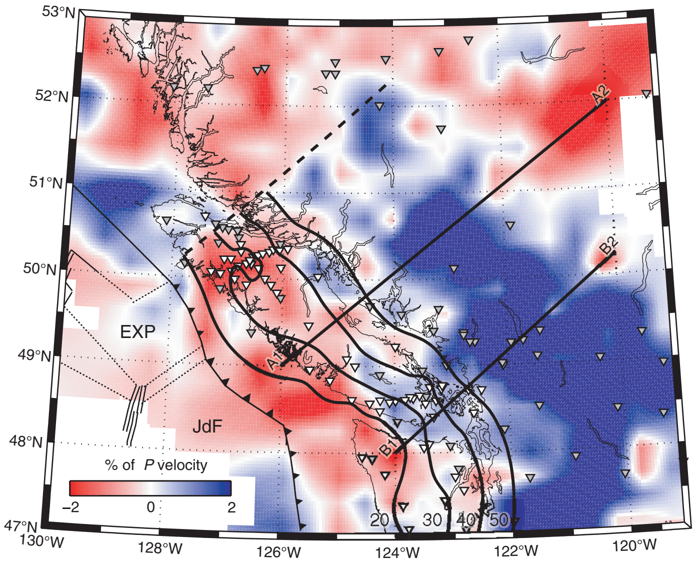

Figure 3. P-wave tomography image of upper mantle structure at 200 km depth. Blue high-velocity body represents thermal and compositional anomaly of subducting Juan de Fuca (JdF) slab. Black dashed line indicates northern limit of subduction. Cross sections of the model along lines A1–A2 and B1–B2 are shown in the GSA Data Repository (see footnote 1). Shallow oceanic crust contours from receiver functions are overlaid as black lines with depth to top of oceanic crust (in km) indicated. Inverted white triangles are broadband seismic stations used in receiver functions. Inverted gray triangles are additional short-period and broadband seismic stations used in traveltime tomography.

DISCUSSION

Both shallow and deep slab results are plotted in Figure 3. We consider five main features: (1) the depth contours of oceanic crust outline a NE-dipping underthrusting plate in northern Vancouver Island at shallower levels than farther south; (2) the top of oceanic crust shallows over a region centered on the town of Woss, where seismic anisotropy is present; (3) the contours deepen at the extrapolated location of the Nootka fault beneath Vancouver Island; (4) the oceanic crustal signature disappears sharply north of the Brooks Peninsula; and (5) the edge of the deep slab is imaged at all depths, indicating continuity in structure from Vancouver Island into the interior of British Columbia.

Further Evidence of a Sharp Slab Edge

The imaged morphology of the slab in northern Vancouver Island is in agreement with all available geological and geophysical data in the region. Heat flux measurements show a transition from low values (~46 mW/m^2) over the subducted portion of the slab in central and southern Vancouver Island to higher values (~67 mW/m^2) to the north of the Brooks Peninsula over a distance of 50 km (Lewis et al., 1997). This shift also coincides with a transition from low to high Bouguer anomalies, indicating hotter but denser material at crustal depths north of the Brooks Peninsula (Lewis et al., 1997). Moreover, this region shows subdued topography, lower mean elevation, and the absence of deep seismicity, suggesting the absence of a slab (Lewis et al., 1997).

Of these constraints, heat flow is the most informative regarding the time evolution of the slab margin due to the ~40 m.y. necessary to establish the surficial heat flow transition through the conductive crustal regime (Lewis et al., 1997). The large heat flow difference in northern Vancouver Island roughly coincides with the imaged slab edge, consistent with a stable location of the Queen Charlotte triple junction at the Brooks Peninsula during that time (Lewis et al., 1997). Stability of the ridge-trench-transform triple junction at the northern end of Cascadia for ~40 m.y. and subduction of the Juan de Fuca ridge likely imply toroidal mantle flow around the slab edge that can cause thermomechanical erosion and slab melting, generating a complex geochemical environment at the surface (Thorkelson and Breisprecher, 2005).

In northern Vancouver Island the imaged edge of the slab coincides with the Alert Bay volcanic field, which has a geochemical signature resembling that of ocean-floor basalts and within-plate basalts and is distinct from volcanic rocks expected within the subduction zone forearc (Armstrong et al., 1985). The slab edge also coincides with a change in fault orientation northwest of the Brooks Peninsula fault zone (Fig. 1), in agreement with periods of extension in the Tertiary, which, assuming triple junction stability, is presumably caused by the viscous coupling of the overriding North America crust with infilling asthenospheric material (Lewis et al., 1997). The location of the deep slab edge beneath the interior of British Columbia also coincides with the disappearance of normal Cascade arc volcanism.

Explorer Tectonics: Slab Stretching?

The current state of Explorer plate kinematics (Explorer is converging with North America [Braunmiller and Nábělek, 2002; Mazzotti et al., 2003] or Explorer is a new transform boundary between Pacific and North America [Rohr and Furlong, 1995; Kreemer et al., 1998; Dziak, 2006]) is a conundrum that we cannot address directly in this study. The creation and fate of the Explorer plate, however, are governed by dynamical processes driven by the balance of driving to resistive forces acting at plate boundaries that are intimately linked with Explorer plate structure.

In a subduction zone, the ratio of negative buoyancy of the subducting slab (slab pull) to resistive forces acting on the plate generally drives plate motion (Forsyth and Uyeda, 1975; Govers and Meijer, 2001). Buoyancy of the subducting slab is controlled mainly by temperature and, to a lesser extent, composition. At a ridge-trench-transform triple junction the subducted slab is young and hot, and its increased buoyancy diminishes slab pull, and hence convergence. However, the net slab pull exerted along the much larger Cascadia margin dominates over diminished pull at the edge and the ridge is forced to plunge down underneath the continent. The ensuing toroidal flow around the subducted slab edge generates hot mantle upwellings and contributes to increased temperature near the edge, and thus further increases its buoyancy. It is then likely that the feather slab edge will remain at shallower levels than farther south of the triple junction, consistent with the observed oceanic crustal depth contours of the Explorer–Juan de Fuca slab beneath Vancouver Island (Fig. 3).

These factors hold the key to the formation of the Explorer microplate. Creation of the Nootka fault implies that Explorer started resisting subduction ca. 4 Ma due to increased slab buoyancy at shallow levels. Elevated temperatures most likely prevented the Nootka fault from tearing through the entire subducted Juan de Fuca slab. Hence a possible cause of the creation of the Explorer microplate is the stretching and/or tearing of the slab by tensile forces that accommodate plate reconfiguration at shallower levels (Fig. 4). Subducted slab stretching is the analogue of crustal thinning during continental rifting, with tensile forces provided by slab pull (ten Brink et al., 1999). Assuming a linear relative velocity between the Explorer and Juan de Fuca slabs with time from 0 to 2 cm/yr, we estimate the separation to cause ~40 km of slab stretching. Ensuing mantle upwelling results in the thermal uplift of the slab at the location of maximum extension, consistent with an increase in Bouguer anomaly and a shallowing of the subducted oceanic crust.

We postulate that the shallow portion of the oceanic crust centered around Woss represents the expression of stretching of the Explorer slab at subcrustal levels, and accounts for the localized strong anisotropic fabric presumably due to shearing (Fig. 4). Note that slab contours deepen near

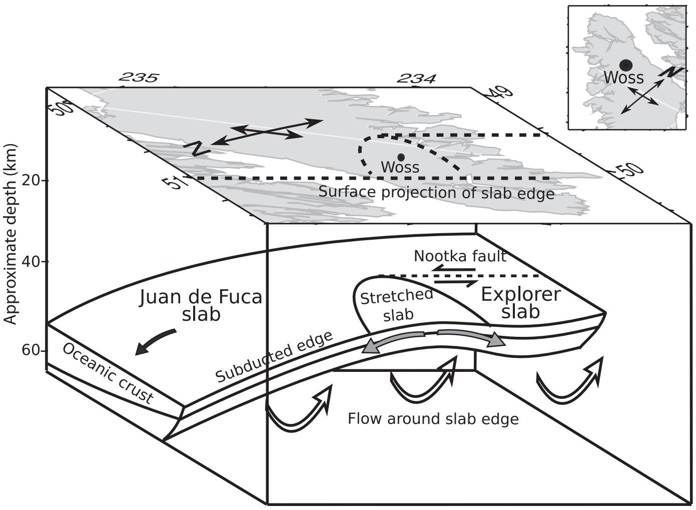

Figure 4. Tectonic interpretation of subducted slab edge morphology in northern Cascadia. Inset shows areal view of region. We postulate that the shallowing of oceanic crust centered on Woss represents the expression of slab stretching. Explorer slab is detached from Juan de Fuca plate along Nootka fault and region of inferred stretching.

the extension of the Nootka fault beneath Vancouver Island at station VI57, indicating a disruption in slab continuity. We interpret the subducted portion of the Explorer plate to be the small underthrust segment bordered by the Nootka fault, the shallow swell near Woss where maximum stretching is inferred, and the Brooks Peninsula fault zone (Fig. 4). This model implies that the Juan de Fuca slab remains unperturbed to the north and east of the Explorer slab and still contributes to slab pull, and is consistent with the reduced convergence of the Explorer plate north of the Nootka fault and eventual Explorer plate capture by North America. Such episodes of plate capture have been reported in similar ridge-trench triple junction settings, e.g., the Rivera plate north of the Cocos plate in Central America (DeMets and Traylen, 2000), and the Monterey and Arguello plates offshore Baja California (Stock and Lee, 1994; Zhang et al., 2007). Ongoing convergence of the Explorer plate with North America is probably due to a combination of viscous coupling between the Explorer and Juan de Fuca slabs downdip and along the creeping Nootka fault beneath Vancouver Island.

CONCLUSION

Resolving the creation and fate of the Explorer microplate in northern Cascadia requires characterization of the structural elements controlling plate dynamics at a triple junction. Based on seismic and geophysical data we propose a tectonic model in which complexity in slab morphology is attributed to thermomechanical erosion of the slab edge caused by ridge subduction and toroidal mantle flow, increased buoyancy, reduced slab pull, slab stretching, and detachment from the Juan de Fuca plate along the Nootka fault. Our model implies that Juan de Fuca subduction is still active north and east of the detached Explorer slab, and that the Explorer plate is being captured by North America.

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