Probable low-angle thrust earthquakes on the Juan de Fuca–North America plate boundary

Anne M. Tréhu*
Jochen Braunmiller
John L. Nabelek

College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, Oregon 97331-5503, USA

ABSTRACT

In 2004, two clusters of earthquakes occurred in the central part of the Cascadia forearc, which displays several characteristics indicative of along-strike and downdip variations in plate coupling. Moment tensor analysis for the main shock in each cluster indicates that both events, which had magnitudes of 4.9 and 4.8, were compatible with low-angle thrust motion on fault planes dipping 6°–15° to the east, consistent with the plate boundary dip of ~12°. By tracing rays through a high-resolution two-dimensional crustal velocity model to match arrival times of secondary arrivals (pP and PnP), we estimate that the source depth was 9–11 km for the M4.9 event and 15–17 km for the M4.8 event. We conclude that these earthquakes probably represent seismogenic thrust motion on the nominally locked or transitional part of the Cascadia megathrust, updip of where episodic tremor and slip (ETS) have been documented. Continued high-resolution observations of seismicity and improved crustal models are needed to confirm whether apparent temporal correlations between ETS and continued seismic activity in these clusters indicate stress transfer along the megathrust.

Keywords: Cascadia, subduction thrust earthquake, subduction zone, episodic tremor, slip.

INTRODUCTION

The Cascadia subduction zone represents a global end member in which a large accretionary complex is developing as a young hot plate is subducting. Compared to other subduction zones, it has been relatively aseismic since the advent of modern seismic instrumentation. While earthquakes occur frequently in the upper and lower plates (McCrory et al., 2003), including several damaging events in the past two decades, no instrumentally recorded low-angle thrust earthquakes have been documented on the plate boundary during historic times, with the possible exception of an earthquake near the Mendocino Triple Junction in 1992 (Oppenheimer et al., 1993). There is, however, considerable paleoseismic evidence for very large interplate earthquakes (Goldfinger et al., 2003; Kelsey et al., 2005), including a magnitude 9 plate boundary event on 26 January 1700, that appears to have ruptured the entire subduction zone (Satake et al., 1996). On 12 July 2004, an earthquake with moment magnitude 4.9 occurred offshore Newport, Oregon, followed on 19 August by an event with moment magnitude 4.8 (see GSA Data Repository Table DR1). Although these events did not cause any significant damage onshore, they were widely felt. The epicenters of these events (Fig. 1A) indicate that they occurred in a part of the forearc where the predicted temperature at the plate boundary is between 350 and 450 °C (Fig. 1A). Based on studies of the frictional properties of metamorphosed sediments, this temperature range is considered to be transitional between where the plate boundary is locked and where plate motion is accommodated aseismically (Hyndman and Wang, 1995). Several other earthquakes have occurred in this region since 1989 (Figs. 1B, 1C), whereas very few earthquakes have occurred elsewhere within the nominally locked or transitional zone. The adjacent Juan de Fuca plate west of the Cascadia deformation front has also been anomalously active. This region of relatively elevated seismic activity coincides with the northern end of a buried ridge on the subducted plate (Fleming and Tréhu, 1999) and an apparent widening of the partially coupled zone, as determined from geodetic observations (McCaffrey et al., 2007), suggesting that structures on the subducted plate may affect plate coupling.

The 2004 events occurred in two well-defined clusters (Fig. 1D) with several aftershocks in the days following the events (Fig. 1E). Several small events also occurred near both main shocks in the years preceding the 2004 events and again in 2007 (Table DR1). Seismograms recorded at seismic station COR for a foreshock and three aftershocks are nearly identical to those of the 19 August earthquake in spite of nearly three orders of magnitude variation in amplitude (Fig. 2), indicating that they all had the same mechanism and hypocenter. Similar repeating earthquake sequences have been reported from many fault zones around the world, including in fault segments that are characterized by a large amount of aseismic creep (Nadeau and McEvilly, 1999; Uchida et al., 2003). Waveforms from events in the July cluster show more variability, although P-to-S time differences indicate that the earthquakes within this cluster also originated within 1 km of each other. Much of the scatter in the reported locations of events within these clusters may be due to location uncertainty.

MOMENT TENSOR INVERSION

The fault mechanisms and depths of the two 2004 main shocks were determined through moment tensor inversion of long-period regional 3-component seismograms in the 25–50 s pass band (Fig. 3; Nabelek and Xia, 1995). For each event, the source mechanism thus derived is well constrained and indicates a thrust fault with an eastward dip of 6°–15°. The slip vector of 66°–73° for the 19 August event is very similar to the expected slip vector for Juan de Fuca–Oregon block motion at this latitude (34.8 mm/yr with an azimuth of 75°; McCaffrey et al., 2007); for the July 12 event, the slip vector is rotated ~25° clockwise. Uncertainties in the moment tensor inversion and depth estimates from teleseismic P arrivals are discussed in Figures DR1–DR7 in the GSA Data Repository.

HYPOCENTER RELOCATION USING 2D RAYTRACING

To obtain an independent, high-resolution estimate of source depth for the 19 August earthquake, we modeled traveltimes picked from short-period data (0.1–1 s) by tracing rays through the crustal model of Figure 4A. The dip of the subducting Juan de Fuca plate (12° at km 185) is well constrained by controlled-source seismic data (Fig. 1D). In the upper plate, the Siletz terrane, a thick, basaltic terrane that underlies the Coast Range and Willamette Valley, forms the backstop for the Cascadia accretionary complex (Snayvel and Wells, 1996; Fleming and Tréhu, 1999). It is underlain by a low-velocity wedge that thickens toward the east (Li and Nabelek, 1999) and is interpreted to indicate subducted oceanic crust and sediments and/or serpentinitized upper-plate mantle (Tréhu et al., 1994;
Bostock et al., 2002). A narrow band of strong reflectivity beneath the continental shelf is interpreted to be the plate boundary on this profile and on a seismic profile located ~40 km to the north (Tréhu et al., 1995). The strong reflectivity likely results from high fluid pressure. Similar strong reflectivity has been observed offshore Vancouver Island (Nedimovic et al., 2003) and in the Nankai Trough (Kodaira et al., 2004). Whether such observations of high reflectivity correspond to regions where the plates are locked or are slipping aseismically remains controversial; both end-member models have been proposed.

Key to our approach to decreasing the uncertainty in hypocentral parameters is a secondary arrival that follows the direct P wave by ~0.85 s.

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Figure 1. A: Seismicity of northeastern Pacific Ocean. All earthquakes in the Advanced National Seismic System (ANSS) composite catalog with magnitudes >3.5 since 1989 are included. Dark gray shading shows where thermal models indicate that temperature at megathrust is below 350 °C, and light gray indicates megathrust temperatures between 350 and 450 °C (Hyndman and Wang, 1995): these have been interpreted to indicate regions where the plate boundary is fully locked or transitional between being locked and freely slipping, respectively. Megathrust earthquakes are expected to rupture shaded portion of plate boundary. Note absence of seismic activity within this region, with the exception of central Cascadia and northern California. The two earthquakes that are the focus of this paper are labeled by their month and year of occurrence. Bathymetric and topographic contour intervals are 1000 m. Box outlines the region shown in B. B: All earthquakes from ANSS composite catalog since 1989 with magnitude >1. Symbol size is proportional to magnitude. Clusters of seismic activity within subduction zone and in adjacent oceanic plate are apparent. Bathymetric contour interval is 250 m. Box outlines region included in D. C: Time history of seismicity since 1989 from ANSS composite catalog. Color coding in C and D shows events before (green dots) and after (blue dots) the summer 2004 earthquakes; violet indicates events in these clusters in 2007. D: Earthquakes since 2003 from ANSS composite catalog. Color coding shows clustering of events before (green dots) and after (blue dots) the earthquake on 12 July 2004. Violet indicates events in these clusters in 2007. Red and blue lines show locations of published two-dimensional crustal velocity models based on controlled-source seismic data. Orange segments show extent of strong lower-crustal reflection interpreted to indicate the presence of fluids at plate boundary. Dotted red line labeled CTZ (crustal transition zone) indicates abrupt north-south increase in seismic velocity beneath continental shelf and Coast Range. Dashed red lines indicate contours of slip deficit (in mm/yr) relative to expected local plate motion rate of 34.8 mm/yr (McCaffrey et al., 2007); these show a transition from a narrow zone of intermediate interplate coupling north of 44.8°N to a wide zone south of 44.5°N. Background shading offshore represents aeromagnetic data illuminated from the west. Subtle ridge in magnetic field is interpreted as buried basement ridge (Fleming and Tréhu, 1999). Bathymetric contour interval is 250 m.

Figure 2. Velocity seismograms for foreshock, main shock, and three aftershocks of the 19 August 2004 earthquake at seismic station COR (source-receiver distance ~80 km) and for main shock at TOLO (~30 km). Waveforms for the 5 earthquakes at COR are remarkably similar in spite of nearly 3 orders of magnitude difference in amplitude (scale shown in counts). Predicted relative arrival times for Pg, PmP, Sg, and SmS are for a source at x = 185 km and z = 16 km and a Poisson’s ratio of 0.29 for the model of Figure 4A. Broadband seismograms (20 samples/s) from the Incorporated Research Institutions for Seismology (IRIS) Data Management Center are shown with no additional filtering. Coda magnitudes (Mc) from the Pacific Northwest Seismic Network (PNSN) catalog are given for the smaller events.
at COR (Fig. 2). Based on the controlled-source seismic data, which indicate that the amplitude of reflections from the Moho of the subducting plate (PmP) is similar to that of the direct P-wave arrival (Pg) at source-receiver offsets of 60–110 km for sources near the epicenter of the August earthquake (Gerdom et al., 2000), we interpret the secondary arrival observed at COR as PmP and used this observation to constrain hypocentral depth. Direct S-wave (Sg) arrivals observed at seismic stations COR and TOLO were used to decrease uncertainty in the earthquake epicenter, which can be quite large in the direction perpendicular to the coastline for offshore earthquakes (Braumiller et al., 1997).

Observed (PmP – Pg) and (Sg – Pg) times at seismic stations COR and TOLO were compared to predicted times for a wide range of possible source positions and Poisson’s ratios, as shown in Figures 4B–4D. This analysis indicates that the earthquake occurred within 1 km of the interpreted plate boundary at a depth of ~16 km. It also confirms that the

![Image](image1.png)

Figure 3. Source mechanisms of the summer 2004 main shocks from full waveform inversion of 25–50 s data. Azimuthal coverage, shown by triangles around the edge of the focal sphere plots, is excellent for an offshore earthquake.

![Image](image2.png)

Figure 4. A: P-wave velocity model for modeling arrivals identified in short-period data (Gerdom et al., 2000). Model was developed from multichannel seismic reflection data and large-aperture data from offshore airgun shots recorded on 17 ocean bottom and 22 onshore seismometers using a hybrid inverse and forward-modeling approach based on RAYINVR software (Zelt and Smith, 1992). White dashed line delimits region imaged by controlled-source data. Base of the Juan de Fuca crust is well constrained everywhere within this region. Red dashed line shows plate boundary, which is observed within ~20 km of deformation front and at strong pre-critical reflection (thick orange line), and inferred to be ~7 km above base of Juan de Fuca plate crust elsewhere. Siletz terrane is indicated by diagonal lines. B: Predicted time difference between Pg and PmP as a function of source depth for a source at x = 185 km, which corresponds to the projection of the 19 August 2004 epicenter onto seismic profile. Blue dot shows time difference for shots near sea surface observed for the same source-receiver offset. Time difference (Δt) of 0.85 ± 0.05 s observed in seismograms for seismic station COR in Figure 2 corresponds to a depth of 16 ± 0.5 km. C: Contours of predicted Δt between PmP and Pg. Source position was varied in 5 km intervals horizontally and in 2 km intervals in depth. Solid red line shows interpreted position of plate boundary. Contour for Δt = 0.85 s has its closest approach to plate boundary at km 185 and moves away gradually as source moves east. To constrain the horizontal position, time differences between Sg and Pg at stations COR and TOLO were modeled for range of Poisson’s ratio (see D); dashed red line shows predicted contours that correspond to the observation at TOLO of 5.25 ± 0.05 s assuming the preferred Poisson’s ratio of 0.29. D: Predicted S-P time differences at COR and TOLO as a function of Poisson’s ratio for three different source positions along the 0.85 s contour in C: (x = 180 km, z = 14 km), dotted lines; (185, 16), solid lines; (190, 16), dashed lines. Observed time differences of 5.25 ± 0.05 s at TOLO and 10.90 ± 0.15 at COR are shown as solid horizontal lines. Near-vertical dashed lines join solutions for same source position and indicate best-fitting Poisson’s ratio for each position. Observed S-P times at both stations are the same as predicted times for source at (185,16) if Poisson’s ratio is 0.285–0.29, which is consistent with laboratory measurements of Poisson’s ratio in mafic rocks. Sources at km 180 or 190 are consistent with observed S-P times only if Poisson’s ratio is different for paths to the two stations, and in both cases required Poisson’s ratio is either too low or too high for mafic rocks. E: Ray diagram for a source at x = 185 km, z = 16 km. Ray paths for Pg, Sg, PmP, and SmS to seismic stations COR and TOLO are highlighted. Star indicates critical angle, defined by Snell’s law, for this source depth and velocity structure; beyond critical angle, all energy is reflected, consistent with large PmP amplitude observed at COR (reflection coefficient 1.0) and with the absence of a clear PmP phase at TOLO (reflection coefficient ~0.1). Radiation pattern for moment tensor solution of Figure 2 is also shown and indicates that Pg and PmP originated in quadrants with opposite polarity, consistent with observed phase shift of ~180°.
epicentral parameters published by the Pacific Northwest Seismic Network (PNSN) for the 19 August event are accurate to within 5 km, even though the PNSN depth was overestimated by ~12 km. The Poisson’s ratio of 0.285–0.29 obtained for the path between the earthquake and stations TOLO and COR is consistent with the mafic nature of the Siletz terrane (Christensen, 1996). The ray-tracing analysis is also consistent with the time difference between Sg and SmS at COR (Fig. 2) and with the observation that Pg and PmP have opposite polarity (Figs. 2 and 4E). Ray tracing through this velocity model to match traveltimes of teleseismic Pp arrivals supports this depth estimate (Fig. DR5).

No PmP arrival is observed for the 12 July 2004 earthquake. The range of source-receiver offsets within which the amplitude of PmP rises above the background noise level is small, and observations of PmP arrivals on temporary stations deployed onshore to record shots along L8 and L9 (Fig. 1D) suggest that stations TOLO and COR were not near the critical distance for PmP arrivals for this event. By modifying the model of Figure 4A to incorporate along-strike changes in the density and velocity of the accretionary complex (Flemings and Trèhu, 1999; Gerdom et al., 2000) and modeling Pp arrivals for this event (Fig. DR6), we conclude that this event occurred at a depth of 9–12 km, also close to the interpreted plate boundary. Proximity to where the subducted ridge abuts the Siletz backstop may explain the apparent rotation of the slip vector for this event.

DISCUSSION

The close correspondence between the preferred hypocentral depths and the plate boundary as interpreted from the high-resolution crustal model suggests that the two main shocks (and other events with similar waveforms) resulted from slip on the plate boundary. This is the first evidence for seismogenic slip in the locked or transitional zone (Fig. 1A) since the great Cascadia earthquake of A.D. 1700 (Satake et al., 1996). A time delay of ~6 months between these earthquakes and the previous episode of episodic tremor and slip (ETS) along this segment of Cascadia (Brudzinski and Allen, 2007) suggests that slip downwarp of the transition zone, where ETS originates, may trigger slip at shallower depth (Mazzotti and Adams, 2004). That seismic activity occurred in both clusters in 2004 and again in 2007 (Figs. 1C, 1D) suggests external triggering by the same process. While it may be premature to conclude whether these earthquakes represent repeated slip on small asperities surrounded by a region of stable sliding or slip on a weak portion of the plate boundary embedded within a locked fault, confirmation of apparent correlations between ETS episodes and clustered seismicity by future observations would support the latter model. This study also underlines the importance of having an accurate model for the crustal velocity structure in order to accurately locate earthquakes and understand their tectonic implications.

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