A regional study of shear wave splitting above the Cascadia subduction zone: Margin-parallel crustal stress

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Abstract. Recordings of local earthquakes from 16 three-component broadband seismic stations in southwestern British Columbia, Washington, and northern Oregon are used to study regional variations of shear wave anisotropy in the North American plate above the subducting Juan de Fuca plate. There is evidence for shear wave splitting at all sites, with good agreement of fast polarization directions and travel time delays at adjacent stations. Most stations exhibit fast directions parallel to the strike of the margin, with anisotropy of 1-2%. These fast polarization directions are consistent with earthquake focal mechanisms and borehole stress studies, indicating that the observed anisotropy is likely due to crustal stresses (i.e., extensive dilatancy anisotropy theory). The margin-parallel stresses may be due to oblique subduction of the Juan de Fuca plate. However, at the station closest to the coast (OZB), the fast direction shows a more margin-normal orientation that may be associated with the proximity of the locked portion of the underlying subduction thrust fault.

Introduction

Regional variations in the direction and magnitude of crustal stresses above a subducting plate can provide important information about dynamics of the subduction process and stress interaction between the subducting and overlying plates. Many tools can be used to study stress within the crust, including borehole breakout studies and earthquake focal mechanisms. In recent years, shear wave splitting has become an important technique for constraining crustal stress [e.g., Crampin and Lovell, 1991].

Differential regional stresses can induce seismic anisotropy in the crust through the alignment of pervasive fluid-filled cracks [e.g., Crampin, 1985]. In an anisotropic crust, vertically propagating shear waves will be polarized into two orthogonal components, one parallel to the direction of regional compressive stress which travels with a higher velocity. Such anisotropy can be studied through the analysis of shear waves generated by local earthquakes.

The Cascadia subduction zone marks the convergence between the subducting Juan de Fuca plate and continental North American plate (figure 1a). In this region, the North American continental crust is approximately 30-35 km thick [e.g., Zelt et al., 1993]. The Juan de Fuca plate is at a depth of ~30 km beneath the westernmost stations used in this study (e.g., OZB) and is at 60-70 km depth beneath the easternmost stations (e.g., TTW, LON) [e.g., Ma et al., 1991].

Shear wave splitting within both the mantle and continental crust of the Cascadia subduction zone has been noted in previous studies that were limited to two locations. Bostock and Cassidy [1995] reported SKS splitting beneath station PGC on southern Vancouver Island with a fast direction striking N74°E. In central Washington State, Silver and Chan [1991] noted SKS splitting beneath station LON with a fast direction striking N84°E. The observed splitting, with a margin-normal fast orientation, was attributed to mantle flow associated with the subduction of the Juan de Fuca plate. In the only previous crustal study, Cassidy and Bostock [1996] looked at crustal shear wave splitting beneath station PGC using local earthquakes and determined a fast direction striking approximately ESE, i.e., margin-parallel.

The current study provides an extension of this work to much of the Cascadia forearc. Local earthquakes recorded at 16 three-component broadband seismometers, including

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Figure 1. a) Seismograph stations used in this study (triangles). Earthquakes locations are shown as circles, and those used by Cassidy and Bostock [1996] are squares. b) Earthquakes projected onto cross-section oriented perpendicular to the subduction margin. The Juan de Fuca plate profile (solid line) is the average structure beneath region 2.
station PGC, are used to constrain crustal shear wave anisotropy beneath each station. Comparisons among the stations provide information on regional variations in shear wave splitting parameters.

Data and Method

The stations used in this study (figure 1a) are operated by the Geological Survey of Canada, the University of Washington, and the U.S. Geological Survey. The stations ALB, LAS, and EGM were temporary seismographs operated from 1987 to 1989 [Cassidy and Ellis, 1993]. All stations have three-component broadband instruments, recorded at sampling rates between 20 and 50 Hz. In shear wave splitting studies, it is desirable to only use energy that has an angle of incidence of less than 40° at the station. Energy at larger incidence angles may undergo distortion due to post-critical reflections at the Earth’s surface [Booth and Crampin, 1985]. We note that events with incidence angles of up to 50° have been used in previous shear wave splitting studies [e.g., Gledhill, 1991]. In the current study, several stations had few earthquakes within the 40° "shear wave window". Thus, we examined events with incidence angles of up to 48°, but only used those with incidence angles less than 40° for the final calculations.

The selected data set consists of 172 earthquakes (magnitudes, $M_3 = 0.6$ to 3.7). Almost all events occurred within the continental crust or in the upper oceanic plate, and thus their raypaths have only sampled the continental crust (figure 1b). Approximately 5% of events occurred deep within the downdropping plate and sample the mantle wedge as well as the crust above the subducting plate. The number of events analyzed at each station varies from 1 to 49, due to regional variations in seismicity and variable recording periods. Earthquake locations for the Canadian sites were obtained using the Canadian National Seismograph Network operated by the Geological Survey of Canada. The U.S. locations were obtained from the Pacific Northwest Seismic Network earthquake catalogue maintained by the University of Washington. Typical uncertainties are ±2–4 km in latitude, longitude, and depth [e.g., Cassidy and Bostock, 1996].

Three-component recordings of each earthquake were integrated to produce displacement waveforms which were then filtered to 1-10 Hz to remove microseismic noise. All earthquakes had small magnitudes. Therefore, the bulk of the energy is in this frequency range. The horizontal components were rotated into the radial (earthquake-to-station azimuth) and transverse directions. A particle motion plot of the horizontal components was used for shear wave splitting analysis. Figure 2 shows the filtered displacement waveforms and resulting particle motion plots for two earthquakes, one at station OZB and one at GNW.

Shear wave splitting was observed for nearly all events. The particle motion plots (e.g., figure 2) show clear first motion directions at the arrival of the first shear wave, followed by a change in particle motion direction several milliseconds later. The polarization direction of the fast shear wave was measured, following the procedure of Chen et al. [1987] and Cassidy and Bostock [1996]. The direction relative to the radial direction was added to the azimuth of the station relative to the earthquake to obtain the geographical direction. This procedure reduces subjective bias in measuring the direction. Fast polarization directions could be measured for most events with uncertainties less than 10°.

Measurements of the slow direction and the time delay between the fast and slow arrivals were more difficult to make in some cases, as noted in previous studies [e.g., Chen et al., 1987]. The slow arrival was signified by a change in polarization to motion approximately perpendicular to the initial arrival (e.g., figure 2). The direction of this arrival was measured following the above procedure. The time delay between the fast and slow shear waves was the time difference between the two arrivals. The slow direction could be confidently measured on ~70% of the waveforms, with average uncertainties of ±15°. The delay time could be determined on ~80% of the waveforms, with average uncertainties of ±0.03 s. For some waveforms for which a slow direction could not confidently be picked, the particle motion showed an abrupt change, indicating another arrival. The time difference between the fast arrival and this change was taken as the minimum time delay for these waveforms.

Observations

Fast Polarization Directions

At all stations, the fast polarization directions are very consistent for energy arriving at a wide range of azimuths and angles of incidence. Figure 3 shows equal area plots of the fast directions measured at stations OZB, GNW, TTW, and LON. Earthquake source effects may contribute to the observed waveforms. However, the fast directions are not believed to be significantly affected by source effects due to the observed consistency. In addition, a variety of focal mechanisms are found in this region, including strike-slip, thrust, and normal faulting [e.g., Ma et al., 1991; Mulder, 1995; Cassidy and Bostock, 1996].

Table 1 provides the mean fast directions for each station. For these means, only events with incidence angles less than 40° and clear arrivals were used. In Washington and Oregon,

![Figure 2](image_url)

Figure 2. Filtered displacement seismograms with relative amplitudes (left), close-up of the shear waves (centre) and horizontal particle motion plots (right) for the time window denoted by the vertical lines on the centre plot. The black dots are at 0.025 s intervals, and the solid line is the fast S-wave. The arrow is the arrival time of the slow S-wave (dotted line). a) earthquake 20 km northeast of station OZB ($M_3 = 1.7$, 15 km depth) - fast azimuth is 92°, delay time is 0.18 s. b) earthquake 28 km northeast of station GNW ($M_3 = 3.0$, 30 km depth) - fast azimuth is 159°, delay time is 0.35 s.)
inland stations exhibit a north-south fast polarization direction (figure 4a), parallel to the Cascadia margin. In southwest British Columbia, the fast direction for most stations is approximately northwest-southeast. Due to the change in orientation of the Cascadia subduction zone, this direction is also margin-parallel. Closer to the coast, station OZB displays a well-constrained fast direction oriented approximately east-west, oblique to the NNW margin orientation. Two earthquake focal mechanisms have been determined for this area [Mulder, 1995]. Both mechanisms show a margin-normal compressive axis (figure 4a).

**Travel Time Delays**

The delay times are generally on the order of 10 milliseconds per kilometre of distance travelled. For all stations, there are no apparent variations in the delay times (normalized by hypocentral distance) with event azimuth or incidence angle. This indicates that the degree of crustal anisotropy is fairly uniform beneath each station.

Figure 4b shows the observed delay time as a function of earthquake focal depth for all stations, except PGC (this is given by Cassidy and Bostock [1996]). At all sites, significant delay times of ~0.1 seconds are observed for events with a shallow focal depth (10-20 km). This was also noted by Cassidy and Bostock [1996] for station PGC. This suggests that the anisotropy must increase in the upper 10-20 km of the crust in order to produce the observed delay times.

A least-squares straight line fit to a delay time versus hypocentral distance plot was computed at stations for which there are sufficient good quality delay time picks for events at a large range of hypocentral distances. The slope of this line gives an estimate of the path-averaged anisotropy. An average crustal shear wave velocity of 3.6 km/s was assumed beneath all stations. This velocity is derived from P-wave refraction studies [Zelt et al., 1993], combined with $V_p/V_S$ values from earthquake studies [Mulder, 1995]. All stations exhibit anisotropy on the order of 1-2% (table 1), which is consistent with that expected for continental crust [e.g., Crampin and Lovell, 1991]. There are no obvious regional variations in the magnitude of anisotropy.

**Table 1. Mean Fast Directions and % Anisotropy (±1σ)**

<table>
<thead>
<tr>
<th>Station</th>
<th>Number of events</th>
<th>Mean Fast Direction</th>
<th>% Anisotropy</th>
</tr>
</thead>
<tbody>
<tr>
<td>PHC</td>
<td>4</td>
<td>160° ± 13°</td>
<td></td>
</tr>
<tr>
<td>CBB</td>
<td>1</td>
<td>150°</td>
<td></td>
</tr>
<tr>
<td>OZB</td>
<td>44</td>
<td>95° ± 21°</td>
<td>1.1 ± 0.4%</td>
</tr>
<tr>
<td>ALB</td>
<td>4</td>
<td>148° ± 16°</td>
<td></td>
</tr>
<tr>
<td>LAS</td>
<td>2</td>
<td>139° ± 47°</td>
<td></td>
</tr>
<tr>
<td>EGM</td>
<td>2</td>
<td>101° ± 16°</td>
<td></td>
</tr>
<tr>
<td>PGC</td>
<td>38</td>
<td>123° ± 23°</td>
<td>2.2 ± 0.5%</td>
</tr>
<tr>
<td>ERW</td>
<td>2</td>
<td>115° ± 19°</td>
<td></td>
</tr>
<tr>
<td>OCWA</td>
<td>5</td>
<td>132° ± 37</td>
<td></td>
</tr>
<tr>
<td>GNW</td>
<td>7</td>
<td>146° ± 23°</td>
<td>1.3 ± 0.5%</td>
</tr>
<tr>
<td>SPW</td>
<td>4</td>
<td>199° ± 41°</td>
<td></td>
</tr>
<tr>
<td>TTW</td>
<td>7</td>
<td>185° ± 7°</td>
<td>1.2 ± 0.3%</td>
</tr>
<tr>
<td>RWW</td>
<td>3</td>
<td>162° ± 28°</td>
<td></td>
</tr>
<tr>
<td>LON</td>
<td>3</td>
<td>180° ± 13°</td>
<td>0.6 ± 0.3%</td>
</tr>
<tr>
<td>RAIO</td>
<td>5</td>
<td>187° ± 19°</td>
<td>1.6 ± 0.7%</td>
</tr>
<tr>
<td>COR</td>
<td>2</td>
<td>153° ± 54°</td>
<td></td>
</tr>
</tbody>
</table>

Data from Cassidy and Bostock [1996]

![Figure 3. Equal area plots covering incidence angles from 0° (centre) to 40° (outer circle) showing the horizontal projection of the leading S-wave polarization direction for a) OZB, b) GNW, c) TTW, and d) LON. The line length indicates the measurement certainty (short lines are lower quality measurements). Note the consistency over all azimuths and incidence angles. Large arrows show the mean fast direction.](image)

![Figure 4. a) Mean fast directions (solid bar indicates stations with ≥5 events). Also shown are the maximum compressive stress directions determined from borehole stress studies [Mueller et al., 1997] and from focal mechanism studies of crustal earthquakes throughout southwestern BC [Mulder, 1995] and Washington [Ma et al., 1991; Mueller et al., 1997; Giamppiccolo et al., 1999]. b) Delay times versus focal depth for OZB (crosses) and all other stations (circles). The dotted line is the best-fit line for the PGC delay times (data not shown) [Cassidy and Bostock, 1996].](image)
Discussion

Clear evidence of shear wave splitting was observed in almost all waveforms recorded at 16 Cascadia forearc stations. The fast polarization directions are approximately north-south for stations in Washington and Oregon, with a rotation to northwest-southeast in British Columbia. The fast directions thus strike parallel to the Cascadia subduction margin over most of the region. An important exception is the station closest to the coast (OZB), where the fast polarization direction is oblique to the margin.

We do not believe that the observed splitting is due to reverberations and scattering of energy. The data were restricted to energy within the "shear wave window" for each station, only good quality events with clear arrivals and linear particle motion were used, and the observed slow arrivals were approximately perpendicular to the initial arrival. There are several possible causes for shear wave splitting: 1) fractures associated with active faults; 2) anisotropy due to lithology or geological structures; and 3) alignment of fluid-filled fractures by the regional stress field (extensive dilatancy anisotropy (EDA) [e.g., Crampin, 1985]). Local geological effects (1 and 2) are not believed to be the primary cause because of the observed consistency in fast directions at each station for events at a wide variety of azimuths and incidence angles. In addition, local effects would not be expected to produce the observed depth and lateral extent of anisotropy.

It is believed that stress-aligned fluid-filled fractures are the dominant source of crustal anisotropy for this area. Large-scale regional stresses produce an anisotropic crust by preferentially closing cracks that strike perpendicular to the direction of maximum compressive stress. In this model, the fast polarization direction will lie parallel to the direction of maximum compressive stress.

The results presented here indicate that the direction of maximum horizontal compressive stress is margin-parallel throughout most of the Cascadia forearc. This is supported by several other stress observations within this region (figure 4a). In southwestern British Columbia, Mulder [1995] reports a maximum compressive stress oriented 152.2°±7° based on earthquake focal mechanisms. Both Ma et al. [1991] and Giampiccolo et al. [1999] report focal mechanisms in Washington showing a north-south maximum compressive stress. A variety of borehole stress measurements (borehole breakouts and hydrofractures) show similar compressive stress directions [Mueller et al., 1997]. Theoretical studies also support a dominant margin-parallel compression [Wang et al., 1995]. Wang [1996] attributes this compression to a buttressed forearc sliver and oblique subduction of the Juan de Fuca plate.

The crustal shortening observed over most of the region through geodetic studies [e.g., Hentong et al., 1999] is margin-normal, in the direction of plate convergence, consistent with a currently locked subduction thrust fault. As pointed out by Wang et al. [1995], the resulting margin-normal compressive stresses are transients associated with the great earthquake cycle on a larger margin-parallel steady state stress. Thus, for most of the region, fluid-filled fractures are aligned with the dominant margin-parallel compressive stress. However, closer to the coast (approaching the locked portion of the subduction fault), the magnitude of the margin-normal stress increases, resulting in a more margin-normal orientation of crustal fractures and anisotropy. The obliquity of the OZB fast directions to the margin may be indicative of this. Based on these results, analysis of shear wave splitting within the continental crust above the Cascadia subduction zone appears to be an effective tool for mapping the maximum regional horizontal compressive stress direction.

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References


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