Array measurements of deep tremor signals in the Cascadia subduction zone

Mario La Rocca,[1] Wendy McCausland,[2] Danilo Galluzzo,[1] Steve Malone,[2] Gilberto Saccorotti,[1] and Edoardo Del Pezzo[1]

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[1] Preliminary analysis of deep tremor recorded during July, 2004, in the Cascadia Subduction zone shows that small aperture arrays can resolve the slowness and back azimuth of seismic waves with a useful resolution. Data were collected by three dense arrays of short-period seismometers specifically deployed in the Puget Sound area under an US-Italy-Canada cooperative effort. Slowness analyses at the three arrays indicate that the 2–4 Hz tremor wave-field is composed of waves propagating with apparent velocities higher than 4 km/s. Combining this with polarisation analysis show these waves to be transverse (SH) waves. However, P-waves, though smaller in amplitude, can be detected by different slowness values obtained for the radial and transverse components. The intersection of wave vectors determined by the back azimuth and slowness values measured at the three arrays provides a preliminary estimate of source location for a sample of the recorded deep tremor. **Citation:** La Rocca, M., W. McCausland, D. Galluzzo, S. Malone, G. Saccorotti, and E. Del Pezzo (2005), Array measurements of deep tremor signals in the Cascadia subduction zone, *Geophys. Res. Lett.*, *32*, LXXXXX, doi:10.1029/2005GL023974.

1. Introduction

[2] Tremor-like seismic signals were observed a few years ago in southwest Japan and interpreted as being generated in the zone of subduction of the Philippine Sea plate beneath the Japan plate [*Obara*, 2002]. These seismic signals show the same spectral characteristics of volcanic tremor, but are recorded far from volcanoes. The signal amplitude is only slightly greater than that of background noise. The onset is emergent. The frequency content is between 1 and 5 Hz, and the duration variable from minutes to hours. Rough tremor source locations spread in a broad area, often with a clear migration along subduction zone strike with time. Calculated depths are in the range of 20–40 km [*Obara*, 2002]. Similar seismic signals have been recently observed in the Cascadia subduction zone. Here periods of deep tremor are clearly correlated both in space and time with the slip episodes observed every 14 ± 2 months by continuous GPS measurement on Vancouver Island and northern Washington [*Dragert et al.*, 2002; *Rogers and Dragert*, 2003; *McCausland and Malone*, 2003]. Data from the widely-spaced stations of the Pacific Northwest Seismic Network (PNSN) have been used to infer rough estimates about the source location, using either the signal's envelopes [*McCausland and Malone*, 2004] or a modified beam-forming technique [*Kao and Shan*, 2004]. In both cases, the results are analogous to those obtained in Japan, indicating a similar depth range and epicenter migration rate along the subduction strike.

[3] The absence of coherent sharp pulses clearly recognizable at regional seismic stations makes the accurate determination of source locations using classical techniques based on inversion of picking phase arrivals nearly impossible. The many successful experiences in volcanic tremor studies [e.g., *Konstantinou and Schlindwein*, 2002; *Chouet*, 2003], suggested that seismic arrays could provide a powerful tool for investigating the complex wavefields of Cascadia deep tremor. The recurrence period of 14 ± 2 months observed in northern Washington and British Columbia [*Rogers and Dragert*, 2003] suggested that the next tremor episode should have occurred between May and July 2004. For this reason a field survey using small aperture seismic arrays was installed during this period by the University of Washington in cooperation with the Istituto Nazionale di Geofisica e Vulcanologia, Italy (INGV) and the Pacific Geoscience Centre of the Geological Survey of Canada (PGC). Three seismic arrays were set up during the spring 2004 in the northern Puget Sound region (Figure 1). A deep tremor episode started on July 8 and lasted for about two weeks in the vicinity of the arrays. Preliminary, rough locations of several selected strong tremor bursts were determined by the analysis of waveform envelopes observed at the regional stations [*McCausland and Malone*, 2004; W. McCausland et al., Temporal and spatial occurrence of deep non-volcanic tremor: From Washington to Northern California, submitted to *Geophysical Research Letters*, 2005, hereinafter referred to as McCausland et al., submitted manuscript, 2005]. Array data were then used to investigate the kinematic properties of these tremor bursts with the aim to track the source using estimates of apparent velocity and back azimuth. In this paper we describe the first results obtained by array and polarization analysis of some tremor bursts.

2. Instruments and Data

[4] The three arrays were located near Sequim (SEQ) in the northern Olympic Peninsula, on Lopez Island (LOP), and Southern Vancouver Island (SOK) (Figure 1). Each array consisted of six or seven three-component, short-period seismic stations with spacing of 150–300 meters. All arrays were set to record in continuous mode at

[1]Osservatorio Vesuviano, Istituto Nazionale di Geofisica e Vulcanologia, Naples, Italy.
[2]Department of Earth and Space Sciences, University of Washington, Seattle, Washington, USA.

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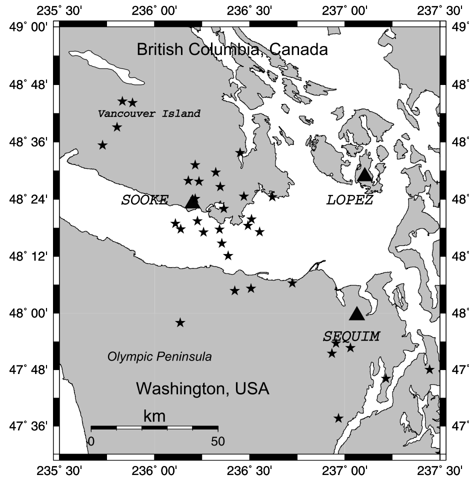

Figure 1. Location of the three arrays deployed during the summer, 2004 deep tremor experiment. For the three array configurations see Figure 4b. Stars represent the epicenters of some strong tremor bursts located by relative timing of waveform envelopes at PNSN stations. See color version of this figure in the HTML.

sampling frequencies of 125 Hz (Lopez and Sequim) and 100 Hz (Sooke). Service runs for data recovery were made every two weeks to one month.

[5] During the Eposodic Tremor and Slip (ETS) event many bursts of deep tremor were recorded with considerable amplitude at the three arrays, whereas for others it was recognizable only at one or two arrays. An example of tremor signals is shown in Figure 2, as recorded at stations LOP4, SEQ4 and SOK4. Bottom plots in Figure 2 show the spectra averaged over the array stations for each components. Most of the energy is concentrated in the 2–6 Hz frequency band. The high-amplitude, broad peaks at frequencies below 1.5 Hz at Lopez and Sequim are attributed to oceanic microseismic noise. The same peak is not observed at Sooke because of the instrument response of the 2 Hz seismometers used at this array.

[6] Harmonic components, often observed in volcanic tremor and often interpreted as due to an oscillating source process, are not present in deep tremor. The horizontal components of the three arrays all show a broad spectral peak at about 3 Hz. The persistence of this peak at such widely-spaced sites suggest that this energy represents a contribution from the source.

3. Kinematic and Polarization Properties

[7] Array processing techniques can characterize the details of arriving seismic waves for wave type and direction of approach. We use the Zero Lag Cross-Correlation method in the time domain (ZLCC [*Frankel et al.*, 1991; *Del Pezzo et al.*, 1997]) and polarisation analysis [*Jurkevics*, 1988] to resolve the back azimuth, slowness and particle motion of tremor waves. From a preliminary analysis we estimate the main propagation direction, then we rotate the horizontal components along the radial and transverse directions. Detailed analysis is then applied separately to the three component seismograms filtered in a 2–4 Hz or 3–6 Hz frequency band, using one second long sliding windows with 80% overlap. Figure 3 shows an example of the correlation, back azimuth and slowness for a 2-minute-long section of deep tremor recorded at the Lopez array. The results of the ZLCC analysis of a ten minute segment of tremor is shown in Figure 4. The azimuthal distribution of back azimuth and the slowness distributions are for signals with high correlation (>0.8) and high rectilinearity (>0.7).

[8] In addition to propagation parameters, we also estimate the polarisation attributes of the incident wave-field from application of the covariance matrix method [*Kanasewich*, 1981; *Jurkevics*, 1988]. The stacking of individual array station's covariance matrices delayed according to the slowness estimated for that particular window allows for a consistent reduction in the variance of polarisation estimates. The resulting eigenvector associated with the largest eigenvalue of the stacked covariance matrix is the polarisation vector. Under the convention positive is upward the polarization azimuth of a P-wave is coincident with the propagation azimuth, whereas a SV-wave propagation and polarisation azimuths will differ by 180 degrees. The rectilinearity of particle motion, computed from the eigenvalues and shown at the bottom of Figure 3, gives an estimate of the body wave quality.

[9] Only short segments of tremor have been analyzed at the three arrays thus far. At the Sooke array the tremor amplitude is generally higher than the other two arrays because it is close to the sources and on very hard competent rock. The slowness distributions at this array is characterized by mean values smaller that those observed at Lopez and Sequim indicating a steeper incidence angle. The tremor wave-field at both Sooke and Sequim is often fairly complex with respect to the Lopez arrays due to simultaneous activity of more than one source near the arrays. For this reason the back azimuth distributions are spread over a wide angle (Figure 4a). The Lopez array almost always shows very high correlation and stable values of back azimuth for each tremor burst. We attribute a lower corre-

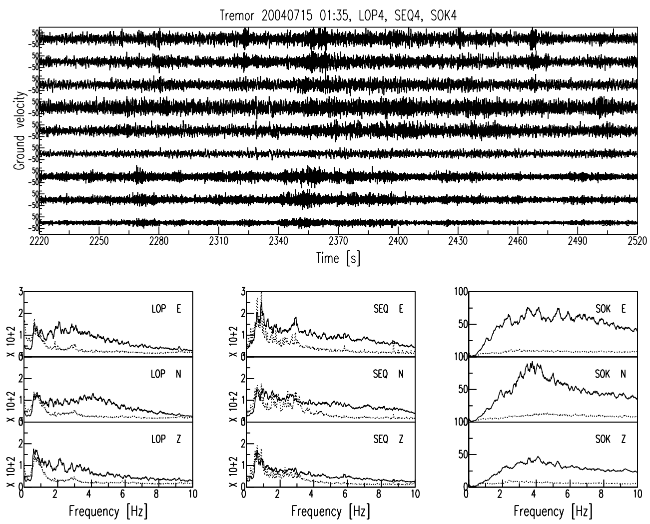

Figure 2. Example of deep tremor bursts recorded at the three arrays. Top plot shows the three component unfiltered seismograms at stations LOP4, SEQ4 and SOK4. Bottom plots show array-averaged spectra of the same tremor signals. Dotted line is from a sample of noise before the tremor burst.

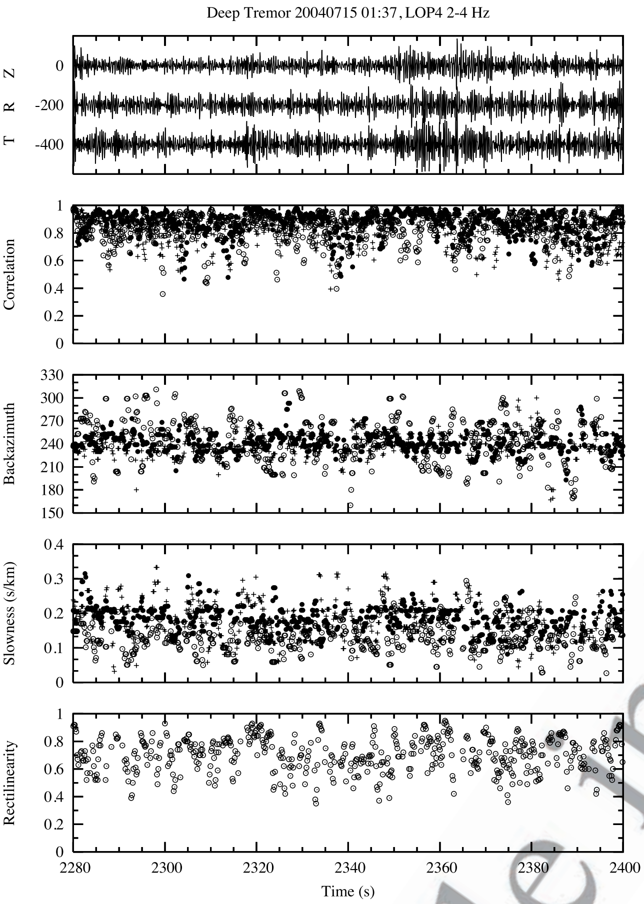

Figure 3. Results from Zero Lag Cross-Correlation analysis at Lopez array for two minutes of strong tremor. Seismogram amplitudes are in microns/sec. Full circles, empty circles and crosses represent results for radial, transverse and vertical components respectively. Bottom plot depicts the particle motion rectilinearity obtained by the polarization analysis. Only the results of windows with correlation >0.8 are used in the spectral averages.

lation level at the Sequim array as due to a higher background noise level and more complex geology under the array stations. At all three arrays, both the array-averaged signal correlation and signal amplitude on horizontal components are larger than those measured on the vertical component. Horizontal slownesses are typically lower than 0.25 s/km indicating a dominance of body waves in the tremor wave-field. It is noteworthy to observe that average ray parameters associated with the radial component are generally lower than those observed for the transverse components. This is a clear indication that the deep tremor wave field contains a small but measurable contribution of compressional waves, since they can be recorded only on the radial component and not on the transverse.

[10] Results shown in Figure 4, obtained by the analysis of ten minutes of tremor, represent the most common signal characteristics observed at the three arrays typical of the analyses of several strong tremor bursts. The back azimuth distributions at the three arrays indicate that the tremor burst described in Figures 3 and 4 must be produced by at least two sources. One gives the predominant contribution to the signals recorded at Lopez and Sequim, and can be located by the intersection of the two maxima of back azimuth distributions. Other sources must be very close to the Sooke array with much stronger energy than at the other two arrays, a variety of back azimuths and lower slownesses indicating steeper arrival angles. The occurrence of more than one source at nearly the same time has been inferred by tremor amplitudes at PNSN stations [*McCausland and Malone*, 2004].

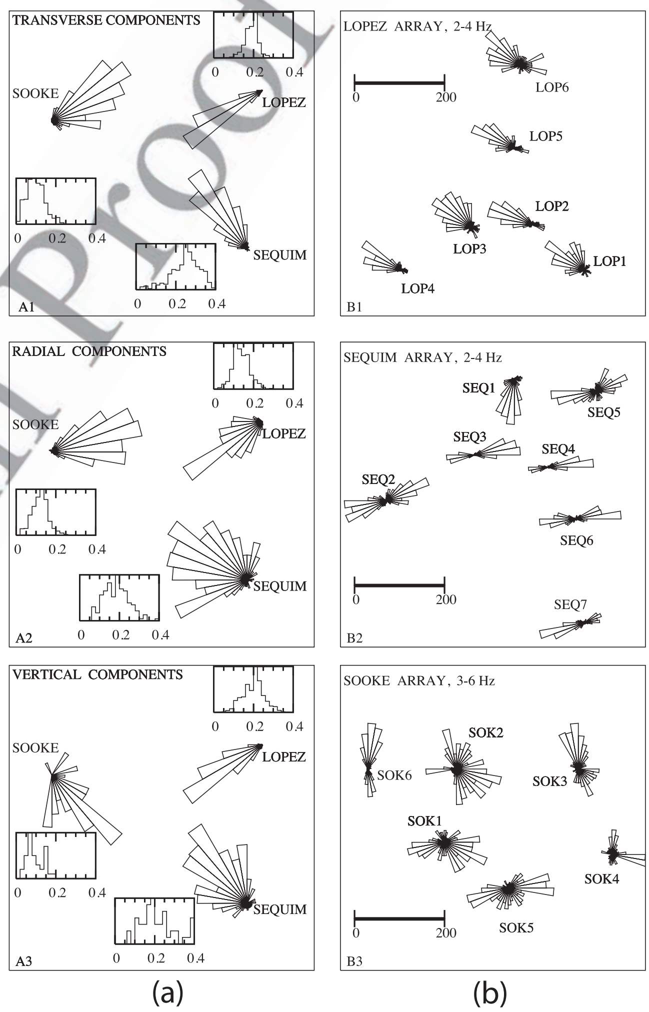

Figure 4. Analysis of ten minutes of deep tremor filtered in the 2–4 Hz at Lopez and Sequim and 3–6 Hz at Sooke. Only signals with correlation >0.8 and rectilinearity >0.7 have been selected for these plots. (a) Normalized distributions of back azimuth (rose diagrams) and slowness (histograms in sec/km) at the three arrays obtained by ZLCC distributions separately for the transverse, radial and vertical components. (b) Stacked normalized projections on the horizontal plane of the polarization azimuth obtained by single station polarization analysis for each station of the three arrays. Only signals with rms >0.125 micron/s at Lopez and Sequim arrays and rms >0.3 micron/s at Sooke, and rectilinearity >0.7 have been selected for these distributions. A comparison with the back azimuth distributions shown in Figure 4a gives an idea about the predominance of shear waves, particularly at Lopez and Sequim.

[11] We can estimate the epicentral area of some tremor energy bursts by the simple intersection of the backazimuth directions estimated at at least two of the three arrays. The source depth inferred by the slowness values in this example using Lopez and Sequim is around 30 km.

[12] As an example of local detail we also measure the polarization parameters at each station of the three arrays separately, setting a high amplitude and rectilinearity threshold to consider only well defined body waves. Figure 4b shows the polarization azimuth distributions computed at all stations for the same 10 minutes of tremor described in Figure 4a. Comparing the distributions in Figure 4B with the back-azimuth distributions plotted in Figure 4a, we can deduce that most of the high amplitude signals are composed of SH waves, particularly at Lopez and Sequim, while at Sooke the complexity of the wave-field indicated by the back azimuth distribution is confirmed by the polarisation azimuth distributions. At the Sequim array the anomalous behaviour of station SEQ1 with respect to all the others of the same array is evidence of a strong local site effect at this station.

4. Discussion and Conclusions

[13] Joint array-polarization analysis of data recorded by small aperture arrays is shown to be a useful tool for detailed investigation of the wave-field properties of deep tremor. Polarization results confirm that tremor signals are composed mostly of shear waves, though a small contribution of P waves is evident at many stations for a number of tremor episodes. However, P-wave phases cannot be easily associated with specific S-waves from a common source. The evidence of P-waves in the deep tremor signals comes only from the distribution of polarization azimuths and from the lower value of slowness measured on the radial components with respect to the transverse components. We cannot establish whether these P-waves come directly from the source or are produced by SV- to P-wave conversions. Further analyses are necessary to understand if there is a correspondence between P- and S-wave packets. If this correspondence could be established, the hypothesis of small earthquakes occurring semi-continuously at depth in a small volume would be supported.

[14] It seems clear that the occurrence of multiple tremor sources at nearly the same time as determined by network locations (McCausland et al., submitted manuscript, 2005) is consistent with the array analysis in at least a few cases thus far analyzed. Thus tremor can take place simultaneously over an extended region, either as individual isolated sources or a distribution of sources. The details of the individual source dynamics may be much harder to determine than for isolated earthquakes. We feel that further refinements of array analysis techniques applied to many more samples of data will be needed to gain additional insight into the deep tremor source.

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E. Del Pezzo, D. Galluzzo, M. La Rocca, and G. Saccorotti, Osservatorio Vesuviano, Istituto Nazionale di Geofisica e Vulcanologia, I-80124 Napoli, Italy.
S. Malone and W. McCausland, Department of Earth and Space Sciences, University of Washington, Seattle, WA 98195, USA. (steve@ess.washington.edu)