Frequency of large crustal earthquakes in Puget Sound–Southern Georgia Strait predicted from geodetic and geological deformation rates

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Earthquake hazard in the Cascadia subduction zone forearc comes from three sources: great subduction earthquakes, Wadati–Benioff slab events, and earthquakes in the forearc crust. This study deals with a concentration of forearc crustal earthquakes in the Puget–Georgia region of southwestern Canada and northwestern United States. These earthquakes are due to margin-parallel shortening, rather than compression in the direction of plate convergence. The frequency of large earthquakes has previously been based mainly on extrapolation of the statistics of smaller events from the short instrumental record. In this study, an independent estimate has been obtained through the seismic moment rate required to accommodate current rates of deformation from GPS and geological data. The catalogue statistics to Mx = 7.5 give a moment rate of 4.0 × 10^{17} Nm/yr and a shortening rate of 2.9 mm/yr if all deformation is seismic. GPS data indicate local current N–S shortening of 3 ± 1 mm/yr. For a seismogenic thickness of 12 km, this deformation represents a moment rate of 4.1 × 10^{17} Nm/yr. The predicted occurrences are 0.022/yr (45 years) for M > 6, and 0.0025/yr (400 years) for M > 7, within the uncertainties of the catalogue statistics. Large characteristic earthquakes are not required.

Forearc long-term average motion based upon larger-scale GPS, paleomagnetic, and geological data ranges from 5 to 7 mm/yr. The estimated long-term occurrence of large events thus is approximately double the current rate, and we infer that the extra earthquakes could follow abrupt N–S shortening in this part of the forearc associated with the oblique rupture motion of great subduction events.

Index Terms: 1208 Geodesy and Gravity: Crustal movements—intraplate (8110); 7230 Seismology: Seismicity and seismotectonics; 7223 Seismology: Seismic hazard assessment and prediction; 8109 Tectonophysics: Continental tectonics—extensional (0905); KEYWORDS: GPS, earthquakes, Cascadia, Puget Sound, geodynamics, hazard


1. Introduction

The earthquake hazard in southwestern British Columbia, western Washington, and western Oregon comes from three sources: great earthquakes on the Cascadia subduction thrust, Benioff–Wadati earthquakes in the downgoing oceanic plate, and earthquakes in the forearc crust (Figure 1). This article focuses on the frequency of occurrence of large forearc crustal earthquakes that are mainly concentrated in the Puget Sound–Southern Georgia Strait region (Puget–Georgia Basin). These earthquakes represent a significant hazard, especially for the larger cities of Seattle, Vancouver, and Victoria (Figures 2 and 3). Most estimates of the frequency of occurrence of large potentially damaging crustal earthquakes have been based on using the Gutenberg–Richter recurrence relation to extrapolate the statistics of mainly smaller events contained in the short instrumental record. This method suffers from the limitations of the earthquake data and that large events may occur where there are few small earthquakes, i.e., characteristic earthquakes.

In this study we examine another approach; we estimate the frequency of large earthquakes required to accommodate current rates of deformation, assuming that most deformation occurs seismically. Rates of current deformation are available from GPS and other geodetic data, from models of forearc motion based on paleomagnetic and geological data, and from paleoseismic geological estimates of fault motion. In a related study, Mazzotti et al. [2002] have determined the tectonic shortening rate of this seismically active forearc region based upon analysis of GPS data. Although the area experienced three damaging Benioff–Wadati deep earthquakes in 1949, 1965, and 2001,
there have been no very large crustal events in the short historical record. The extrapolation of the statistics of smaller crustal events suggests that large events occur at intervals of \( \sim 50 \) years for \( M > 6 \) and \( \sim 400 \) years for \( M \geq 7 \), i.e., similar to the average for much of California in the same size area. Geological studies indicate recently active structures in the region such as the Seattle fault [e.g., Johnson et al., 1999] on which paleoseismic data show that there has been large displacement probably in \( M \sim 7 \) earthquakes. Surprisingly, the three largest historical crustal earthquakes in the general region (\( M \sim 7 \) events in 1872 in north-central Washington, and in 1918 and 1946 on Vancouver Island) occurred outside the area of greatest concentration of smaller events. Also, there is no clear concentration of smaller magnitude seismicity on the Seattle fault and on most similar structures that are inferred to produce the largest earthquakes. Only one identified fault structure in Georgia Strait has clear recent seismic activity [Cassidy et al., 2000; Mosher et al., 2000]. The lack of large historical crustal earthquakes within the present concentration of small earthquakes in the Puget–Georgia Basin region gives concern that the magnitude–frequency of occurrence statistics used to characterize the hazard may not correctly estimate the frequency of occurrence of larger damaging events. There could be “characteristic earthquakes” that are not captured by the statistics. The adjacent Cascadia subduction thrust is an extreme case of such characteristic earthquakes; there is very little current seismicity on this thrust, but there is good evidence that it exhibits infrequent \( M \sim 9 \) giant earthquakes [e.g., Clague, 1997]. Current deformation rates in the forearc region give an independent estimator for the frequency of occurrence of large potentially damaging crustal earthquakes.

[4] In the Cascadia subduction zone between northern California and Vancouver Island (Figure 2), the Juan de Fuca plate subducts beneath the North America plate in a northeasterly direction at about 40 mm/yr. Many subduction zones with oblique subduction have margin-parallel migration of their forearc [e.g., McCaffrey, 1992] and long-term shortening is common at the leading edge of such forearc slivers [e.g., Jarrard, 1986]. The Cascadia forearc repre-

Figure 1. The tectonic sources of large earthquakes on the Cascadia margin.
several comprehensive studies have been carried out relating seismicity to deformation, notably in southern California. Mazzotti and Hyndman [2002] also found agreement between the shortening rate from GPS geodetic data and seismicity in the Mackenzie mountain front of the northern Canadian Cordillera.

[6] To calculate the frequency of occurrence of large earthquakes from deformation rates, it is necessary to know the distribution of deformation (and seismic moment rate) over the different earthquake magnitudes, i.e., the ratio of contribution from large versus small earthquakes. However, within the normal range of magnitude versus frequency of occurrence slopes (“b” values), most of the seismic moment and thus deformation is contributed by the largest magnitude earthquakes. The rate of large earthquakes that we calculate from deformation rates is thus not very sensitive to the slope of the assumed recurrence relation. In our calculation we take the recurrence relation slope to be that obtained from the catalogue statistics.

### 2. Southern California Example

[7] Comprehensive studies have been carried out recently in southern California to compare the rates of earthquake moment release (mainly in large earthquakes) from the historical record, with the moment release predicted from deformation rates determined from GPS, from plate tectonic models, and from the paleoseismic slip rates on active faults [Working Group on California Earthquake Probabilities (WGCEP), 1995]. In that region, there is the strong constraint to the overall total fault slip rate from the Pacific–America relative plate motion. There also are geological data that constrain average motion on most significant faults. These studies show approximate agreement between the observed seismicity data and the frequency of large earthquakes inferred from geodetic and geological data. However, in a comprehensive study [WGCEP, 1995], a discrepancy of up to a factor of two was found, with the deformation data suggesting greater frequency of large earthquakes than estimated from the historical earthquake record. If correct, this difference could be interpreted as indicating the occurrence of infrequent huge earthquakes (characteristic earthquakes), not accounted for in the earthquake statistics model. Alternatively, the seismicity rate may have increased over the past 150 years.

[8] In contrast to the conclusion of the WGCEP study, earlier studies by Anderson [1979] and Hyndman and Weichert [1983] found good agreement between regional seismicity rates and plate model deformation rates for S. California. An important difference in the analyses was that Anderson [1979] and Hyndman and Weichert [1983] used different completeness periods for different earthquake magnitudes to reduce the uncertainty in seismicity rates. The rate of large earthquakes inferred from the catalogue statistics. In our calculation we take the recurrence relation slope to be that obtained from the catalogue statistics.
seismicity rates from deformation data could readily be a factor of two. Field et al. [1999] and Petersen et al. [2000] carefully reviewed and revised all of the input data and assumptions in the WGCEP [1995] calculations. They also found that, within the uncertainties, there was agreement between the frequency of large earthquakes from extrapolation of the historical seismicity catalogue and those predicted from geodetic deformation and geological fault slip rate data. No very large earthquakes are required in addition to those predicted by the catalogue statistics, and no significant aseismic slip is required within the assumed brittle depth range. This effort in southern California gives us confidence that a similar exercise will give useful and important new constraints on the frequency of large potentially damaging earthquakes in the Puget–Georgia Basin region.

[9] We note two differences between the California and Puget–Georgia areas: (1) southern California is a broad transform plate boundary region with seismicity that is primarily due to strike-slip faulting and much smaller amounts of thrust and extensional faulting, whereas the Puget–Georgia Basin is a region of crustal shortening with seismicity that is a combination of thrust and strike-slip faulting; (2) the Puget–Georgia region is much smaller, probably containing only a few structures that generate the largest magnitude earthquakes, and only a few of the larger structures have been determined. The stress migration and redistribution with successive large earthquakes may limit accurate application of the method to large regions, such as the whole plate boundary region of southern California. It may not be applicable to individual fault zones in a broad deforming zone. In our study of the Puget–Georgia region, we include the whole deforming forearc cross-section from the trench to the volcanic arc.

3. Relation Between Deformation Rates and Seismicity

[10] Deformation rates may be related to earthquake rates through seismic moment if most deformation occurs seismically. The seismic moment rate constrains the frequency of occurrence of large events, recognizing that most of the seismic moment normally comes from the larger earthquakes. In this study we focus on the frequency of occurrence of large earthquakes for application to seismic hazard. Therefore, we first estimate the seismic moment rate from Gutenberg–Richer recurrence extrapolation of the catalogue seismicity, and then for comparison estimate the seismic moment rate from the deformation data.

3.1. Gutenberg–Richer Recurrence Relation and Catalogue Seismicity Moment Rate

[11] The instrumental record usually is too short for the number of large events to provide long-term average statistics to useful accuracy. This is especially true for the Puget–Georgia area where the depth constraint required to separate crustal from Wadati–Benioff events is only quite recent. For comparison with seismicity estimates from deformation we follow Anderson [1979], and Hyndman and Weichert [1983], and stabilize the frequency of occurrence estimate for large events by applying the additional assumption that the seismicity follows the Gutenberg–Richer recurrence relation up to an assumed maximum magnitude. This is the method commonly used in seismic hazard characterization. The approach requires that the recurrence relation be a valid representation of the observed seismicity and that the seismicity statistics within the area do not change with time. It also requires that estimates can be made of both the maximum magnitude and the form of the recurrence statistics curve as the maximum magnitude is approached. The main moment contributions usually come from the largest magnitudes, so the form of the recurrence relation near the maximum magnitude is important. We take the maximum earthquake magnitude limit to be an abrupt truncation in the density function, which results in a smooth fall off toward the maximum for the cumulative recurrence function (Figure 4) (see discussion by Hyndman and Weichert [1983]). Alternatives are given by Anderson and Luco [1983]; they discuss the alternative that some fraction of the seismic moment occurs in characteristic earthquakes of size near the maximum magnitude. Good agreement was found for southern California between the seismicity catalogue and moment rate estimates from geological and geodetic deformation rates, assuming the truncated cumulative exponential function and small or no concentration of characteristic earthquakes at the maximum magnitude [Field et al., 1999]. We thus take that distribution function, but recognize that this function is a source of uncertainty in the final results.

[12] The density function is:

\[ n(M) = \alpha \exp(-\beta M) \quad M \leq M_x \]
\[ n(M) = 0 \quad M > M_x \]
where \( \alpha \) and \( \beta \) are the density recurrence coefficients and \( M_x \) is the maximum magnitude. The common Gutenberg–Richer cumulative recurrence relation is given by \( b = 3 / \ln 10 \). The moment–magnitude relation is assumed to be deterministic with moment \( M_o \) being given by the relation \( \log M_o = c + dM \), where \( c \) is of the order 9.0 and \( d \) the order of 1.5 in S.I. units. The total moment rate of the incremental or density recurrence function is then given by:

\[
M_o' = b 10^{4s/[d - b]} M_x + a + c/[d - b]
\]

### 3.2. Seismic Moment Rate From Deformation Rate

[13] For an individual earthquake, the average slip displacement \( d \) and the fault area over which it occurred define the seismic moment, \( M_o = \mu Ad \), where \( \mu \) is the shear modulus. The shear modulus is taken to be \( 3.3 \times 10^{10} \text{ N m}^{-2} \) [e.g., Brune, 1968] for continental crustal rocks. For the whole area of a single fault, the slip rate \( s \) on the fault is proportional to the rate of moment release per unit time \( M_o' \) and per unit fault area, \( s = M_o'/(\mu A) \). The overall deformation rate in a broadly deforming converging region also may be estimated [e.g., Kostrov, 1974; Anderson, 1979]. We employ the simple scalar relation:

\[
s' = CMo'/(2\mu A')
\]

where \( A' = WL \) is the total cross-sectional area of the convergence zone in the plane perpendicular to the convergence direction. It is noted that in thrust faulting, the effective fault area is greater than the cross-section, and the effective rate of fault slip is greater than the N–S shortening rate because of the average fault angle relative to the convergence direction. \( C \) depends on the orientation of the faulting with respect to the regional motion. The fault motions may include both strike-slip and thrust events. The angle depends upon the slip properties of the faults. Molnar [1979] assumed 45° so that \( C = 1.0 \). Anderson [1979] estimated \( C \) empirically to be 0.75 using the data of Chen and Molnar [1977]. The Seattle and Tacoma faults have been interpreted variously to have dips between 25–30° and 55–65° [Brocher et al., 2001; Pratt et al., 1997; Wells and Weaver, 1993]. The Georgia fault is estimated to dip at about 50° [Cassidy et al., 2000; Mosher et al., 2000]. The orientations of mapped strike-slip faults are highly variable, but also average approximately 45° to the convergence direction, i.e., Devils Mountain fault, Southern Whitby Island fault, Coast Range Boundary fault [e.g., Johnson et al., 1999]. We have used \( C = 1.0 \) but include the possibility of 0.75 in our uncertainty analysis. Thus, the moment rate is:

\[
M_o' = 2\mu A'
\]

### 3.3. Maximum Magnitude

[14] The maximum magnitude is an important factor in estimating seismicity from deformation rates. A change in \( M_x \) of 0.5 will result in a change in earthquake rate estimated from deformation by a factor of about two. There are not enough earthquakes at the larger magnitudes to resolve the maximum magnitude in the earthquake statistics. Thus, the maximum must be estimated from the maximum fault area and empirical relations between fault area and magnitude or moment and from paleoseismicity studies.

Taking the relation \( \log A = -4.07 + 1.0 M \) from Wells and Coppersmith [1994], gives a maximum magnitude of about \( M = 7.5 \) if most of the cross-section of the seismically active region was to rupture in one ~ 3000 km² thrust event. The fault structure is very complex, and the largest active structure so far mapped in the area, the Seattle fault is about one quarter of this area. The suggested magnitude of the most recent event on this fault about 1000 years ago was estimated to be approximately \( M = 7.3 \) based upon comparison of the displacement with similar more recent events elsewhere [e.g., Bucknam et al., 1992]. We take \( M_x = 7.5 \) but also have calculated the deformation for assumed maximum magnitudes, \( M_x \), of 7.3 and 7.7. For offshore transform faults and for the San Andreas fault, this approach to estimating deformation rates from seismicity gave results that corresponded well to those from plate tectonic models, generally within 30% [Hyndman and Weichert, 1983; Anderson and Luco, 1983]. However, we note that much of the crustal seismicity of this study represents diffuse deformation rather than the focused slip on large transform faults of that study.

### 3.4. Magnitude–Frequency of Occurrence and Moment Rate

[15] Calculation of the seismicity rate from the deformation rate requires, first, determining the moment release rate for a particular deformation rate. Second, the relative distribution of moment release for each magnitude must be specified, i.e., recurrence relation. Since seismic moments are available for only a few events in the catalogue, we have chosen to work directly with magnitude rather than moment. Therefore, we have calculated the magnitude versus frequency of recurrence from the earthquake catalogue. For comparison with the seismicity predicted from deformation data, we have then employed an empirical magnitude–moment relation.

#### 3.4.1. Magnitude–Frequency of Occurrence

[16] An accurate recurrence relation for the Puget–Georgia area is not as yet available, and warrants future detailed review and analysis. The catalogue data have two important problems: (1) only for the past ~ 20 years has there been sufficient seismograph coverage to allow accurate event depths. For earlier larger events (before about 1980), attempts have been made to separate occurrence within the downgoing Juan de Fuca plate from occurrence in the forearc crust (depth phases, number of aftershocks, etc.), but considerable uncertainty remains. Restricting the data to post ~ 1980, the instrumental seismic record is too short for the number of large events to provide long-term average statistics to useful accuracy. Using the more numerous small events requires extrapolation by about 3 magnitude units to estimate the rates for the largest earthquakes. (2) The catalogue used comes from merging data from three different agencies that employ somewhat different data acquisition and processing. Data for the north part of the area are from the Geological Survey of Canada earthquake database (GSC-PGC); older data for the south are from the U.S. Geol. Survey (USGS), and more recent data are from the Univ. Washington (UW). Most GSC-PGC data have \( M_L \) magnitudes, whereas UW data are mainly \( M_C \) (coda or duration) magnitudes. The conversion of the latter to seismic moment is less well calibrated. For most of the
larger events since 1980 in the southern part of our study region. M_L magnitudes from the GSC-PGC database have been combined with UW locations. We have tried two approaches to define the recurrence, first, extrapolation of the statistics of post-1980 well-located earthquakes and, second, statistics including older earthquakes. We show the results for the latter approach here because the required extrapolation is over a smaller magnitude range. The recurrence from both approaches proved to be very similar, within about 20% in the estimated moment rate.

For the latter approach, the times for completeness of the catalogue varies with magnitude. There are two choices, (1) the data may be limited to the time interval for which all events above a specified magnitude are complete. This is the approach used by some investigators in California. However, as noted above, this approach results in poorly constrained statistics since only the larger magnitudes can be used to define the recurrence relation. Also, there is a tendency to extend the time interval back to when the completeness is not certain. Use of this approach and lack of completeness appears to have been one of the major sources of the apparent discrepancy between deformation and seismicity data for the WGCEP [1995] study, i.e., Stein and Hanks [1998]. (2) A better alternative is to apply the additional assumption that the seismicity statistics do not change with time. This allows us to use longer time intervals for the statistics of the large earthquakes that have a low occurrence rate, and use shorter more recent time intervals for small events that occur with greater frequency. Weichert [1980] presented the statistical treatment required for this approach. The completeness periods we have used are: for magnitudes larger than M > 5.75, 1899; M > 5.25, 1917; M > 4.75, 1940; M > 3.75, 1956; M > 3.5, 1970. There are only a few events prior to about 1940 and they have only a small effect on the recurrence. The uncertainties shown for each magnitude assume a Poisson distribution; an example is shown in Figure 4. Weichert and Milne, [1979] discuss optimizing the best fit and uncertainties to the magnitude–frequency relation. They note that the recurrence relation using this method is mainly constrained by the small to mid magnitudes that have better constrained statistics.

We have taken the minimum magnitude to be as large as possible because our interest is mainly in the large earthquakes. Also, the depths of older smaller events are not well determined, and the magnitude–magnitude relation has been defined using larger events. We find that a minimum magnitude of M = 3.5 provides just enough events such that the statistical uncertainty in the recurrence fits is not a significant error compared to other sources of error in the calculated moment rates. However, including data to M = 3.0 results in a negligible difference in the recurrence. The recurrence relation parameters depend only to a small degree on the assumed maximum magnitude (Table 1).

Two main areas were used to define the earthquake statistics in the Puget–Georgia Basin area to examine the sensitivity of the deformation estimates to the area chosen (Figure 3). The smaller area covers the most intense small earthquake concentration; the larger area includes the surrounding region and is about four times larger. We also have further subdivided the small area into east and west halves to assess the effect of the Puget Sound Tertiary basins and the increase in heat flow to the east. The east half has earthquakes nearly to the surface but has a shallow maximum depth because of the high heat flow approaching the volcanic arc. The west half has a greater maximum depth because of low heat flow, but the upper part is aseismic because of the presence of sedimentary basins. The rate of occurrence M ≥ 7 from the recurrence relations are: for the small box, 0.0013/yr (770 years) and, for the large box, 0.0025/yr (400 years).

### 3.4.2. Seismic Moment From Magnitude

The magnitude rates must be converted to moment rates using an empirical relation. Most of the magnitudes we have used are local magnitude M_L. Earlier we have used log Mo = 9.0 + 1.5 M_L (see Hyndman and Weichert [1983] for a summary of earlier data and references), as did WGCEP [1995]. However, Field et al. [1999] emphasized that there is significant sensitivity of the moment rate to the conversion parameters, and that better agreement is found between moment rate from deformation and from the earthquake catalogue using a more precise value from the relationship of Hanks and Kanamori [1979], i.e., log Mo = 9.05 + 1.5 M_L. We have used their relationship. It gives a ~12% higher moment rate for our data compared to the earlier relation. Additional earthquake moment data for continental western Canada have been provided from moment tensor solutions by Ristau et al. [2002]. We have combined their moment data (4 < M < 6) and earlier data from Oregon State University [Braunmiller, 1999] with the M_L values from the catalogues in Figure 5. There is excellent agreement with the Hanks and Kanamori [1979] relation.

A small correction to the computed moment rates from the magnitudes is required because of the logarithmic asymmetry of the stochastic moment–magnitude relation [e.g., Hyndman and Weichert, 1983]. The average log-moment leads to an average moment that is greater than just the antilog, by approximately a factor exp(s^2/2), where s^2 is the variance of the relation. This factor is exact if the log-moment is normally distributed. We estimate an RMS scatter of about a factor of 2 in Mo, or 0.3 in log Mo, which requires a correction of +27% in moment rate. The required

<table>
<thead>
<tr>
<th>Mx</th>
<th>Small Region</th>
<th>a</th>
<th>b</th>
<th>Shortening Rate</th>
<th>M ≥ 7 Recurrences</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.5</td>
<td>All</td>
<td>2.63</td>
<td>0.75</td>
<td>2.4</td>
<td>2.7</td>
</tr>
<tr>
<td>7.5</td>
<td>West half</td>
<td>2.48</td>
<td>0.78</td>
<td>1.2</td>
<td>2.6</td>
</tr>
<tr>
<td>7.5</td>
<td>East half</td>
<td>2.14</td>
<td>0.72</td>
<td>1.3</td>
<td>2.9</td>
</tr>
<tr>
<td>7.0</td>
<td>Mx = 7.7</td>
<td>2.67</td>
<td>0.73</td>
<td>5.7 ± 1.1</td>
<td>4.1 ± 0.9</td>
</tr>
<tr>
<td>7.0</td>
<td>Mx = 7.5</td>
<td>2.66</td>
<td>0.72</td>
<td>4.0 ± 0.8</td>
<td>2.9 ± 0.6</td>
</tr>
<tr>
<td>7.0</td>
<td>Mx = 7.3</td>
<td>2.64</td>
<td>0.72</td>
<td>2.9 ± 0.6</td>
<td>2.1 ± 0.5a</td>
</tr>
</tbody>
</table>

Table 1. Gutenberg–Richer Recurrence Coefficients, Seismic Moment Rates, and Crustal Shortening Rates for Two Regions and a Range of Maximum Magnitudes, Mx
3.5. Deformation Length and Cross-Section

correction becomes large for larger scatter, so this effect could be a significant source of uncertainty.

3.5. Deformation Length and Cross-Section

[22] Predicting the seismic moment rate from the margin-parallel shortening requires the length and the cross-sectional area that deforms seismically. The margin-parallel length of the small study region was chosen to correspond to the main shortening extent define by GPS data [Mazzotti et al., 2002]. There is much less seismicity to the north and to the south so there is little increase in seismic moment for increasing the margin-parallel length to that of the larger study region.

3.5.1. Margin-Perpendicular Seismogenic Width

[23] The moment rate scales directly with the area so careful consideration is required of both the vertical and horizontal limits of the cross-sectional area that deforms mainly in earthquakes. The map width of the two earthquake areas used in the analysis are 110 and 175 km (Figure 3). To establish the seismogenic depth limits we have used an updated earthquake catalogue provided by T. Mulder (personal communication, 2001), mainly using events since 1985 for which the depth is well controlled. The earthquake depth extent in the study area is found to vary east–west. The western horizontal limit is an abrupt decline in seismicity approximately at the eastern limit of the accreted sediments of the Olympic Peninsula (Core Rocks). The Core Rocks appear to deform mainly aseismically. The sediments probably dip landward following the base of the Crescent volcanic terrane under westernmost Puget Sound [e.g., Brocher et al., 2001] and may lie at depth beneath the larger of our two computational areas. The western boundary of the seismicity concentration varies somewhat to the north and south. On the east side, the region of the volcanic arc has limited seismicity, presumably because of the associated high temperatures. These limits to the earthquake concentration suggest that the small study area deforms mainly seismically, whereas the large study area may have significant aseismic deformation on its west and east edges.

3.5.2. Seismogenic Depth Extent

[24] The effective depth extent of the seismogenic zone is more difficult to establish. Examination of east–west cross-sections of earthquakes (example in Figure 6) shows that most of the earthquakes in the western half of our areas are between 10 and 30 km depth, whereas in the eastern half, they are between 0 and 20 km. The base of seismicity is probably temperature controlled. The lack of seismicity in the upper zone of the western part appears to be determined by aseismic Tertiary sedimentary basins. An east–west cross-section through Seattle (Figure 3) is shown in Figure 6. This is close to the critical maximum temperature for the seismogenic layer of 350°C in the area [Hyndman and Wang, 1993, 1995]. However, the effective seismogenic thickness is probably somewhat less. There is a tapering of seismicity below a depth of about 20 km that probably represents an increasing fraction of aseismic deformation as the thermal limit is approached. We therefore take the base of an equivalent fully seismic layer to be the middle of this taper at 25 km. In the eastern part of our study area the base of seismicity rises to about 15 km; this probably still represents the critical temperature of about 350°C, reached at a shallower depth in the high heat flow zone approaching the arc. There is a thermal boundary between low heat flow in the forearc to high heat flow in the arc-back arc located about 20 km seaward of the volcanic arc [Blackwell et al., 1990; Lewis et al., 1988].

[25] In the western part of our area there are few earthquakes in the upper 10 km. The lack of earthquakes in the upper few kilometers is frequently observed elsewhere and is sometimes ascribed to stable-sliding clays gouge [e.g., Marone and Scholz, 1988]. Also, the western half is underlain by a series of Tertiary sedimentary basins extending to depths of ~10 km [e.g., Brocher et al., 2001; Zelt et al., 2001]. These basins appear to have relatively few earthquakes. They may accommodate much of the upper crust shortening through deformation by folding and by aseismic fault slip in the young sediments. In the eastern part of our areas where there are no thick young sediments, seismicity extends to the surface but the higher heat flow results in a shallower maximum depth. From the above analysis, the study area deforms mainly seismically, whereas the large study area may have significant aseismic deformation on its west and east edges.

Figure 6. Seismicity cross-section across northern Puget Sound region showing the main 12 km thick crustal seismic zone in the lower crust beneath Puget Sound rising to the upper crust in the east as the volcanic arc is approached.
seismogenic thickness is from 10–25 km depth in the west and 0–15 km in the east.

For a second approach to the seismogenic thickness, we have assumed that the seismicity rate at the depth of maximum earthquake rate (~20 km in the west, ~10 km in the east) represents fully seismic deformation (Figure 7). At other depths, the fraction of deformation that is seismic is taken to be the earthquake frequency at that depth divided by the frequency at the depth of maximum seismicity. Since the shortening rate is not expected to vary with depth, the earthquake frequency shallower and deeper represent varying amounts of aseismic deformation. The result is an equivalent fully seismic or effective thickness of 12 km, similar but slightly smaller than the 15 km estimate above. Ideally we would estimate the seismicity frequency and thus shortening rate for the depth interval that we conclude to be fully seismic. However, the number of recent events with adequate depth control is too small for a result with useful accuracy. The 12 km is similar to the 11 km used by WGCEP in S. California. Stein and Hanks [1998] discuss the arguments for an effective thickness of 10–12 km for S. California. In the latter area the heat flow is generally higher than the Cascadia forearc away from the arc, so the thermal maximum seismogenic depth may be less. However, for most of the S. California region there are few deep aseismic Tertiary basins such as in Puget–Georgia Basin. With our effective seismogenic thickness of 12 km, and the narrow computation region with width of 110 km, the total cross-sectional area is 1320 km². For the larger computational region width 175 km, the area is 2100 km². We consider the uncertainty in seismogenic thickness and thus uncertainty in calculated deformation rate from this source to be at least ±25%.

An important complication for large events is that the rupture depth extent may be greater for the largest earthquakes compared to the seismogenic zone for small events. This is of particular concern for earthquake hazard. Large events, greater than about \( M = 6 \), may initiate in the seismogenic zone concentrated at depth of about 20 km, below the sedimentary basins, but rupture up (and perhaps down) to depths that are aseismic for small magnitudes. The rupture for the largest earthquakes may extend to the surface, even through Tertiary sediments that are aseismic at small magnitudes. Thus there may be more of the largest events than predicted by extrapolating the statistics of small events.

3.6. Moment and Deformation Rates From Seismicity

Table 1 provides, first, the catalogue seismicity recurrence relation parameters \( a \) and \( b \) for the small and large study regions and for different maximum magnitudes. Second, the table presents the seismic moment rates and estimated shortening rates, using the effective seismogenic thickness, the moment–magnitude relation, and other computation parameters discussed above. Surprisingly, the large study region has a slightly larger calculated moment rate per cross-sectional area than the smaller study region, in spite of the evidently lower average seismicity. This difference is only in part because of the greater margin-parallel length of the large area. The difference emphasizes the sensitivity of the seismic moment rate and computed deformation rate to the slope \( b \) of the recurrence. The large region data has a \( b \) of 0.72 whereas it is 0.75 for the small region.

4. Deformation Rate and Earthquake Recurrence From GPS and Geological Data

4.1. GPS Shortening Rates

Geodetic data show clearly that the central Cascadia forearc is moving to the north relative to North America (see Figure 2). This motion was first inferred geodetically from very long baseline interferometry (VLBI) that showed the Sierra Nevada block moving to the northwest at about 11 mm/yr [Angus and Gordon, 1991]. From a local campaign network, Savage et al. [2000] showed that southern Oregon is rotating clockwise at a rate consistent with longer-term paleomagnetic data, and with a northward component of motion. Khazaradze et al. [1999], McCaffrey et al. [2000], and Miller et al. [2001] used more data to show that the northward motion of the Oregon forearc block relative to North America decreases to the north from about 15 to 7 mm/yr, and that the motion must be taken up mainly by shortening in Washington and southernmost British Columbia. In work associated with our study, Mazzotti et al. [2002] specifically addressed the north–south shortening across Puget Sound and Southern Georgia Strait in the region of concentrated crustal seismicity. They used the data from the continuous GPS stations in southwestern British Columbia (WCDa), and in Washington and Oregon (PANGA) as well as lower accuracy campaign data from McCaffrey et al. [2000] and Henton [2000]. The present differential motion between southern Washington and southern Vancouver Island is 3.0 ± 0.7 mm/yr in the inner part of the forearc (Figure 8) and 3.5 ± 0.8 mm/yr in the outer part, with the shortening concentrated in Puget Sound–Southern Georgia Strait.

In the forearc the measured velocities consist of a short-term great earthquake cycle interseismic loading part, and a long-term north–south forearc shortening part. Mazzotti et al. [2002] calculated and removed the subduction
loading using the Flück et al. [1997] 3-D dislocation model with an updated convergence rate and direction. Because of the slab geometry and oblique convergence, the subduction loading results in a north–south extension between Oregon and Vancouver Island of 1–2 mm/yr for the inner forearc and 3–4 mm/yr for the outer forearc. These velocities may be subtracted from the GPS data to give estimates of long-term deformation rates.

4.2. Deformation Rate From Models of Forearc Motion and From Geological Studies of Fault Offsets

[31] Cenozoic paleomagnetic data indicate that the Cascadia forearc is rotating clockwise with respect to North America [Magill et al., 1982; Sheriff, 1984; England and Wells, 1991; Beck et al., 1986; Wells, 1990]. This tectonic motion in Cascadia is linked to Pacific–North America shear and northwest migration of the Sierra Nevada block [Walcott, 1993; Pezzopane and Weldon, 1993; Wells et al., 1998]. Recent models for this forearc motion relative to stable North America show clockwise rotation of an Oregon coastal block about a pole in the general area of central Oregon–Washington border [Wells et al., 1998; Wells and Simpson, 2001]. The motion is about 11–15 mm/yr west–northwest in the southernmost forearc and northward at 4–8 mm/yr in the northern forearc. Because of the proximity of the model pole for Oregon block–North America relative motion, the northward motion of the forearc increases seaward from about 4 mm/yr near the arc to about 8 mm/yr near the coast of the western Olympic Peninsula. The model northerly motion in the Southern Puget Sound region is about 6 mm/yr, in agreement with the long-term shortening in the Puget–Georgia region based on the GPS results.

[32] Quantitative fault motion rates from paleoseismicity studies are difficult to obtain in the region because of the effects of glaciation and, on land, because of the common thick vegetation and steep topography. Most fault data have come from marine seismic reflection data. The most striking fault appears to be seismically active [Wells et al., 1993]. Further north in Georgia Strait the Georgia thrust fault and the Devils Mountain fault are a number of other similar structures but most probably have smaller deformation rates. In the Puget Sound region they include the Tacoma fault, Kingston and Sequim faults, southern Whidbey Island fault, and the Devils Mountain fault [Wells and Weaver, 1993; Johnson et al., 1996; Pratt et al., 1997]. This east–west striking broad fault zone is a southward dipping structure with an estimated slip rate of 0.7–1.1 mm/yr [Johnson et al., 1999].

The fault produced relative vertical displacement of marine terraces of 5–7 m in an event about 1000 years ago. There are a number of other similar structures but most probably have smaller deformation rates. In the Puget Sound region they include the Tacoma fault, Kingston and Sequim faults, southern Whidbey Island fault, and the Devils Mountain fault [Wells and Weaver, 1993; Johnson et al., 1996; Pratt et al., 1997]. Further north in Georgia Strait the Georgia thrust fault appears to be seismically active [Cassidy et al., 2000]. The total north–south convergence rate accommodated by all of these faults is poorly determined as yet, but is about 3–6 mm/yr [e.g., Wells and Simpson, 2001], in approximate agreement with the GPS estimates.

4.3. Earthquake Recurrence Rates

[33] The recurrence of large crustal earthquakes from deformation rates requires an assumption on the distribution of seismic moment rate earthquake magnitudes. The two extreme models are that the deformation (and moment) is accommodated by earthquakes with the Gutenberg–Richter distribution over magnitude. The second model is that the deformation is primarily accommodated by large characteristic earthquakes, with few if any small events. For the first assumption we have used a recurrence slope (“b” value) equal to that for the catalogue statistics. The short-term current GPS determined deformation then gives the earthquake rates that agree within the uncertainties of the rates from extrapolating the catalogue statistics based mainly on
small magnitudes. Large characteristic earthquakes are not required, although a small number are allowed by the uncertainties. However, the north–south shortening associated with great subduction earthquakes may be accommodated by a higher rate of crustal seismicity with the current recurrence slope (dashed line) or by characteristic events, i.e., about 6.5 M7 earthquakes or one M7.5 earthquake.

5. Discussion and Conclusions

In this study we have used geodetic and geological estimates of deformation rates in the Puget–Georgia basin area to constrain the frequency of occurrence of large earthquakes. GPS and geological data show that the northward motion of the Oregon forearc block is mainly accommodated by north–south shortening in this region, in good correspondence with the main concentration of crustal seismicity. The Seattle fault has an estimated rate of 1000–2000 years for M > 7 earthquakes. The most recent such event occurred about 1000 years ago [e.g., Johnson et al., 1999]. Several other structures occur to the north and south that are capable of producing earthquakes nearly as large [e.g., Johnson et al., 1999; Mosher et al., 2000].

That the geodetic deformation predicts a frequency of large earthquakes similar to that for the earthquake catalogue statistics, indicates that large characteristic earthquakes are not required. However, we emphasize that there are substantial uncertainties in both the earthquake catalogue and in the deformation rates, as well as in the calculation parameters. Thus, a moderate number of such events are allowed by the data. We also note two other possibilities, (1) that very large earthquakes may rupture seismically further toward the surface, i.e., into the sedimentary basins than small events, so have a larger magnitude than predicted by the recurrence relation statistics, (2) that more crustal earthquakes are expected to follow a megathrust event, to accommodate the predicted ~0.7 m abrupt N–S shortening [Mazzotti et al., 2002].

The large earthquakes of central Vancouver Island do not appear to be directly generated by the same margin-parallel shortening; GPS data indicate that there is at most minor margin-parallel forearc shortening extending that far north [Henton, 2000]. These latter earthquakes could be generated by shear tractions on the base of the forearc of the oceanic Nootka transform fault that subducts beneath this area, or by the differential subduction rates beneath the margin to the north and south of the Nootka transform fault.
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References


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