**Stochastic Finite-Fault Modeling of the September 26, 2003
M 8.1 Tokachi-Oki Earthquake. Implications for Expected
Ground Motions for Megathrust Events in the Cascadia
Subduction Zone**

Submitted by

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Abstract

In this study, a simulated-based ground-motion prediction model for the Cascadia region was developed. The model allowed prediction of average ground-motion from large interface earthquakes and estimation of the variability in expected ground motion due to possible ranges of variability in input parameters.

This research work was achieved in two main parts. In the first part, linear regression was applied to the Fourier spectra data of the 2003 **M** 8.1 Tokachi-Oki earthquake and some of its aftershocks (**M**: 7.3, 6.4, 5.9, 5.5), in order to determine source, path and site parameters for these earthquakes and, within the context of the stochastic finite-fault modeling technique, calibrate a simulation model for this earthquake. From the empirical analyses, the ground-motion attenuation for all the events can be modeled using a geometric spreading coefficient $b_1 = -1.0$ with an associated inelastic attenuation model given by $Q = 135 f^{0.76}$; typical site amplifications were obtained for NEHRP D and E sites and the possible nonlinear site amplification was not significant. Empirically derived stress drop values agree with those of other studies, ranging from 100 to 200 bars with no apparent dependence on magnitude but possible dependence on type of earthquake; about 100 bars for interface and near 200 bars for in-slab events. These empirical parameters were used as input for the stochastic simulation of PSA, from which a calibrated model was derived.

In the second part, the calibrated model was used first to estimate an average stress drop for large interface earthquakes and then, adapted to the Cascadia region to estimate average PSA for scenario earthquakes from **M** 7.5 to **M** 9.0, for firm site

conditions near the cities of Vancouver, Victoria and Seattle. Results for these conditions suggest mean values of 5%-damped PSA from 100 to 200 cm/s^2 , at frequencies from 1 to 4 Hz; uncertainty in stress drop causes uncertainty in simulated PSA of about $\pm 50\%$. Uncertainties in the attenuation model produce larger uncertainties in PSA, a factor of about two at 100 km. These results were generalized as ground-motion prediction equations for interface earthquakes at the Cascadia region.

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Chapter 1. Introduction

Subduction zones in many regions of the world pose a major seismic hazard to modern urban communities. A significant contribution to this seismic hazard comes from mega-thrust earthquakes occurring along the subduction interface regions; several examples include the 1960 $M \geq 9$ Chilean earthquake (Housner, 1963; Dunber et al., 1992), the 1964 $M \geq 9$ Alaskan earthquake (Lander, J., 1989), the 1985 M 8 Valparaiso, Chile (Choy and Dewey, 1988), and Michoacan, Mexico (UNAM seismology group, 1986) earthquakes; more recently the 2003 M 8.1 Tokachi-Oki, Japan earthquake (Koketsu et al., 2005) and the 2004 $M \geq 9$ Sumatra earthquake (Bilham et al. 2005). In all these events, major damage to structures and / or loss of lives was reported.

Among the most seismically active and densely instrumented regions is Japan, where more than one subduction zone contributes to the high seismic hazard affecting the region (Stauder and Maulchin, 1976; Kinoshita, 1998). On the other hand, the Cascadia subduction zone (SW British Columbia and NW Washington) is a region characterized by relatively low seismic activity. Nevertheless, great earthquakes ($M \geq 9$) have occurred (Atwater et al., 1995; Satake et al., 1996) and are expected in the future (Heaton and Kanamori, 1984; Heaton and Hartzell, 1986; Atwater and Hemphill-Haley, 1996; Satake et al., 2003).

The study of the M 8.1 Tokachi-Oki earthquake, a mega-thrust seism that occurred on September 26, 2003 at the Kurile Trench, NE Japan, and some of its aftershocks (M : 7.3, 6.4, 5.9, 5.5), given the large spatial coverage and density of its records, represents a special opportunity to gain knowledge about ground-motion

attenuation and source scaling properties with magnitude for subduction earthquakes. The information gained may increase knowledge not just of ground-motion processes for Japan but by analogy for similar events in other regions, such as Cascadia.

1.1 Background

Ground-motion attenuation models are necessary input for either deterministic or probabilistic seismic hazard assessments studies (Reiter, L., 1990). These models, commonly known as ground-motion prediction equations (GMPEs), are used to estimate ground-motion measures for a specific site or region as a function of magnitude and distance. Sometimes other parameters are involved, e.g., site class, type of faulting, seismogenic zone, etc. These equations can be obtained empirically (from actual ground-motion records) or theoretically (from seismological models of source, path and site conditions), however, overlap between these two approaches is common practice.

An important feature of observed ground-motion is the tectonic environment where it is generated; as a consequence, GMPEs are often grouped in three different categories: a) shallow crustal earthquakes in active tectonic regions, b) shallow crustal earthquakes in stable continental regions, and c) subduction zone earthquakes (Abrahamson and Shedlock, 1997), subdivided into interface and in-slab events. Ground-motion models for interface events in subduction zones are the focus of this research.

Empirical GMPEs for subduction zones have been derived from global databases, by Crouse, C. B. (1991) (CR91), Youngs et al. (1997) (YS97), and Atkinson and Boore (2003) (AB03). Specific region empirical studies that represent high seismicity zones,

have been derived by Fukushima and Tanaka (1990), Kanno et al. (2006), and Zhao et al. (2006) (ZH06) for Japan; Anderson and Lei (1994) for Mexico.

GMPEs based on global data may not adequately reproduce ground-motion in particular regions (e.g., Cascadia subduction zone). Moreover, the heterogeneity of global subduction databases may represent a potential problem and consequently the results from global ground-motion models may be biased (Atkinson and Boore, 2003). Figure 1.1 shows a comparison between global GMPEs (YS97, AB03) and a region specific attenuation model (ZH06). It plots 5% response spectral amplitudes (PSA) versus distance (measured as the closest distance to fault plane, R_{cd}) for interface earthquakes and for two magnitudes at 0.5 Hz. Figure 1.2 shows this comparison for 5.0 Hz. It is remarkable, from these figures, the significant differences between global attenuation models (AB03 and YS97), and between these models and the regional attenuation model for Japan (ZH06). Particularly, the near-source estimates from the AB03 model are systematically lower than the other models estimates for moderate or large magnitudes. It is also noticeable the very flat attenuation behavior of the AB03 model at low frequencies mainly for large magnitudes. These curves represent, for some frequencies and / or distances, a difference of a factor of about 2.5 times between PSA estimates.

For the Cascadia subduction zone (NW United States, SW Canada), where the number of earthquakes and / or strong ground-motion records is low compared to other subduction regions such as Japan, Mexico, Central and South America, it has not been possible to develop region-specific GMPEs. Thus, the seismic hazard assessment for this zone has relied on global attenuation models (YS97, AB03) or on theoretically derived

ground-motion relationships for which input parameters are highly uncertain, or may not necessarily represent ground-motion properties of the Cascadia region (Gregor et al. 2002). In particular, there are no ground-motion data from large interface earthquakes in this region; however, this possible scenario ($M \geq 7.5$, mega-thrust event; Satake et al., 1996) represents a significant contribution to the seismic hazard affecting the region (Frankel et al., 1999; Halchuk and Adams, 2008; Petersen et al., 2008). Additionally, current building codes provisions for the region and, more specifically, current seismic hazard maps on which the codes are based, use global attenuation models such as YS97 for the Canadian side (SW British Columbia) (Halchuk and Adams, 2008) and regional (ZH06) and global (YS97) models for the United States side (Petersen et al., 2008).

Under this context, a plausible and interesting option is to develop a simulated-based ground-motion prediction model for the region that is first calibrated using data from other regions, then adjusted to represent conditions in Cascadia. This allows prediction of ground-motion from large interface earthquakes and estimation of the variability in expected ground motion due to possible ranges of variability in input parameters. This idea constitutes the principal motivation of this research.

1.2 Purpose of the Study

The main objectives in this research are to study the ground-motion source, path and site effects for the 2003 Tokachi-Oki earthquake sequence, and generalize the lessons learned to the problem of predicting ground-motions for future great interface

earthquakes in the Cascadia subduction zone. I pursue these objectives through several linked studies:

1. Empirical analyses on ground-motion acceleration records and the corresponding Fourier acceleration spectra, for the M 8.1 Tokachi-Oki event and the selected aftershocks, are performed to define source, path and site parameters for this seismic sequence.
2. Stochastic finite-fault modeling techniques are used to explore ground-motion scaling with magnitude, distance and site conditions, and calibrate the simulation model for the 2003, M 8.1 Tokachi-Oki earthquake sequence.
3. The calibrated ground-motion model is used to infer source levels (stress drop) for large interface earthquakes in other regions, and its variability.
4. Simulations of expected ground-motions from large interface earthquakes in the Cascadia subduction zone are made, considering appropriate source, path and site conditions. Results are generalized as ground-motion prediction equations for this region.

1.3 Organization of Thesis

The research work presented in this thesis is organized in six chapters. The introductory chapter outlines the problem addressed and specific objectives of this work. Chapter 2 provides a general description of the tectonic setting for the 2003 Tokachi-Oki earthquake, the regional seismicity, and tectonic information that are relevant for this research work. In Chapter 3, I describe the database for the Tokachi-Oki earthquake

sequence, the empirical analyses (linear regression on Fourier acceleration data) needed to estimate source, path and site characteristics of these earthquakes. Chapter 4 describes the procedure and results for the simulations of response spectral amplitudes for this seismic sequence. In Chapter 5, the calibrated model (from Chapter 4) is used to simulate response spectral ordinates for scenario interface earthquakes of **M** 7.5 to **M** 9.0 in the Cascadia subduction zone, and for the cities of Vancouver, Victoria and Seattle. I generalize simulation results for various magnitudes and distances as ground-motion prediction equations for the Cascadia region. Chapter 6 discusses those aspects of the research work that I think should be explored in more detail and / or improved in order to reduce uncertainties in ground-motion estimation for the Cascadia region and I also summarize results and draw conclusions.

Figure 1.1 - Response spectral amplitudes (PSA) at 0.5 Hz, versus closest distance to fault (R_{cd}), from different ground-motion attenuation models for subduction interface earthquakes. Light gray lines correspond to a M 6.5 earthquake and black lines correspond to a M 8.0 earthquake. Represented models are: Youngs et al., 1997 (YS97); Atkinson and Boore, 2003 (AB03); Zhao et al., 2006 (ZH06).

Figure 1.2 - Response spectral amplitudes (PSA) at 5.0 Hz, versus closest distance to fault (R_{cd}), from different ground-motion attenuation models for subduction interface earthquakes. Light gray lines correspond to a M 6.5 earthquake and black lines correspond to a M 8.0 earthquake. Represented models are: Youngs et al., 1997 (YS97); Atkinson and Boore, 2003 (AB03); Zhao et al., 2006 (ZH06).

Chapter 2. Tectonic setting

On September 26, 2003 at 04:50 (local time) a very large thrust earthquake occurred off Tokachi (Tokachi-Oki), Hokkaido, NE Japan. Preliminary data from the U S Geological Survey (USGS) provides the following parameters: epicenter = 41.78°N , 144.86°E ; depth = 27 km; moment magnitude $M = 8.1$. At least 15 aftershocks above M 5.0 were generated in a 3-day period; the largest reported magnitudes among them were M : 7.3, 6.4, 6.0, 5.9, and 5.5. At the epicenter zone, the Pacific plate subducts towards $\text{N}60^{\circ}\text{W}$ beneath the Hokkaido region from the Kuril trench at a rate of about 80 mm/year (De Mets et al., 1990). Important previous events along the SW Kuril trench are: in 1952 (M 8.1), 1973 (M 7.4), 1969 (M 8.2) and 1968 (M 8.1). Seismological analyses of the 2003 Tokachi-Oki earthquake indicate re-rupture of the 1952 episode, suggesting that it was a characteristic event in this region; for example, Yamanaka and Kikuchi (2003) analyzed the source process of this event from teleseismic body waves but also did an inversion of available near-field records of the 1952 earthquake. They found that the 2003 asperity was also ruptured in 1952. Figure 2.1 shows the epicenters of the main shock and aftershocks with $M > 5.0$, along with a schematic representation of the rupture areas for some of these historical large earthquakes and some tectonic features (plates and trenches) involved in the region.

2.1 Seismicity of the Hokkaido region

The Kurile trench predominantly determines the seismic activity of the Hokkaido region. Stauder and Mualchin (1976) solved 120 focal mechanisms for large earthquakes

($m_b \geq 5.7$) of the Kurile-Kamchatka arc and concluded that for the region, the characteristic fault motion in normal-depth earthquakes is an under-thrust of the Pacific plate under the continental lithosphere in the direction N55°W, along a 20° dipping plane at depths 20-60 km. For the vicinity of the Hokkaido corner, these authors characterize the seismicity with east-west normal faults, dipping northward for intermediate to deep-focus earthquakes, implying hinge faulting and contortion of the Pacific plate between the Japanese segment of the plate (dipping more moderately) and the steeper dipping Kurile Island segment. Later studies for the region (Sadaomi et al., 1983; Kao and Chen, 1995) pointed out that, beneath the Hokkaido Island, a double seismic zone is present between 80 and 170 km depth; this zone consists of two layers of seismicity: an upper layer of down-dip compression events and a lower layer of down-dip extension.

2.2 Tectonic features of NE Japan

This section overviews some tectonic characteristic of NE Japan, which are known to be relevant for the analysis of ground-motion observations in the region (Hashida, 1989; Kanno et al., 2006; Takai et al., 2006). It is not intended to describe, in a detailed way, the tectonic evolution of the region, a subject beyond the scope of this research work. The tectonic features of interest are the basin spatial distribution, relative age of deposits, and geographical description of the volcanic front to separate back-arc from fore-arc region.

Japan can be divided geologically in NE Japan, presently overriding the subducting Pacific plate, and SW Japan, overriding the subducting Philippines Sea plate.

Both subduction zones are associated with volcanic arcs. A triple junction, formed by the Japan trench, Nankai trough, and Izu-Bonin trench, delineates tectonic boundaries (ISTL and KTL) that separates NE from SW Japan tectonic regions (Figure 2.2). The Quaternary volcanic front, in the Hokkaido region, strikes to the East and becomes the Kurile arc (Jolivet et al., 1994; Sato, H., 1994).

In NE Japan, a wide distribution of late Cenozoic back-arc sediments fills Miocene extensional basins (Figure 2.2). More than five km of late Cenozoic deposits have accumulated on the back-arc side of northern Honshu (Sato, H., 1994). The late Cenozoic tectonic evolution of the region has been divided in four regimes: i) unknown (32-25 Ma); ii) extensional deformation (25-13 Ma), featured by a localized thick sedimentation by coarse clastics and volcanoclastics filling half grabens and pull-apart basins; iii) a transitional stage (13-3.5 Ma) characterized by a low rate argillaceous sedimentation; and iv) shortening deformation (3.5 Ma to Recent), marked by a thick basin fill and a high rate of sedimentation. From the middle of Miocene to the present stage (crustal shortening), central Hokkaido was affected by regional compression and a duplex structure developed in the Ishikari Basin, forming a flexural basin in which Quaternary sediments were deposited (Yanagimoto and Iijima, 2003) (denoted in Figure 3.1.7).

The delineation of different origin basins for the region (above) allows definition of geological variables, useful in this study (Chapter 3) to better interpret special features of ground-motion (regional amplification and basin effects). These effects have been previously studied by Cardenas-Soto and Chavez-Garcia (2003) and Yamanaka et al.

(2004), among others. Other factors of relevance to this research are the findings from several experiments concerning the velocity structure beneath the continental and oceanic Hokkaido regions. Iwasaki et al., (1989) observe a Pn velocity change from 7.6 to 8.1 km/s within a narrow zone near the coast of north-eastern Japan, values which were related with low and high Q (quality factor) zones. Hashida (1989) also found correlation for SW and NE Japan between low Q and high Q zones with geological features: type and age of rocks and volcanic activity. The velocity structure is important for determining wave propagation and regional amplifications, whereas thermal conditions and / or rock age are related to seismic attenuation (Q) (Hashida, 1989; Artemieva et al., 2004; Karato, 1993; Roth and Wiens, 2000). Thus these findings could be useful for a better interpretation of the ground-motion attenuation behavior and consequently a more accurate attenuation model. A more detailed description of these factors and the way they are used for the analysis is given in next chapter.

Figure 2.1 - Epicenter of the September 26, 2003, **M** 8.1 Tokachi-Oki earthquake (red – green star), and aftershocks with $M > 5.0$ (yellow open circles). Closed shapes, delimited by continuous white lines, are a schematic representation of the rupture areas for some historical large earthquakes in the region. The number inside a rupture area is the year of occurrence for each earthquake; some estimated magnitudes are **M** 8.2 (1952), **M** 7.9 (1968), **M** 7.4 (1973) (Yamanaka and Kikuchi, 2003). Thick white dashed line represents the trench zones involved at the epicentral region and small red triangles represent volcanoes.

Figure 2.2 - Location map showing the modern tectonic features of Japan. KTL, Kanto tectonic line; HAZ, Hokkaido axial zone; ISTL, Itoigawa-Shizuoka tectonic line, and the spatial distribution of late Cenozoic back-arc sediments for NE Japan (taken from Sato, 1994).

Chapter 3. Empirical modeling of the 2003, M 8.1 Tokachi-Oki earthquake sequence

I apply empirical regression techniques to strong-motion data to determine source, attenuation and site effects for the M 8.1 Tokachi-Oki main shock, and several of its largest aftershocks. The empirical data analyses are mostly based on linear regression techniques applied to the Fourier spectra of each earthquake record. The regressions determine the best-fitting attenuation model and the near-source spectrum for each event. Under the same regression scheme, site amplification is determined for NEHRP D and NEHRP E sites relative to an assigned reference function for NEHRP C sites (NEHRP stands for National Earthquake Hazard Reduction Program). The NEHRP site classification scheme (BSSC, 2001) is based on the shear-wave velocity time-averaged over the top 30m of the soil profile (V_{S30}). The V_{S30} interval for each site class and an explanation of the V_{S30} definition is covered in section 3.1.2. I also determine relative amplification effects attributable to several other geologic variables, in particular whether the site location is in a fore-arc versus back-arc region, the relative age of deposits, and whether the site lies within a sedimentary basin. Different empirical analyses are used to estimate duration of motion (Raoof et al., 1999; Atkinson and Boore, 1995) and attenuation effects (Anderson and Hough, 1984; Atkinson and Silva, 1997; Boore and Joyner, 1997) for these earthquakes. Determination of these parameters is useful, as they are needed as input for the stochastic finite-fault model (Motazedian and Atkinson, 2005) used to model the Tokachi-Oki earthquake sequence, a topic that is covered in Chapter 4.

3.1 Data

In this section I describe the acceleration time series that are available for the 2003 Tokachi-Oki sequence, exploratory analyses that were conducted to characterize salient data features, and the classification of the recording sites based on shear wave velocity information available for each station.

3.1.1 Database for the 2003 Tokachi-Oki main shock and selected aftershocks

The **M** 8.1 Tokachi-Oki main shock and at least 15 aftershocks, with magnitudes from **M** 5.0 to **M** 7.3, were recorded at about 340 stations of the K-NET, NIED network (Kinoshita, 1998), at distances ranging from 40 km to >500 km; three component (NS, EW, Z) acceleration data from this earthquake sequence were downloaded from www.knet.bosai.go.jp, along with shear wave velocity and density profiles at the recording sites. The soil profiles extend to 20 m depth in most cases.

The closest distance to the fault plane (R_{cd}) is used to characterize the distance from each event to each recording site, as it is the most relevant measure for the larger events (Campbell, 1981; Anderson and Lei, 1994). For the main shock, I used the fault plane geometry given by Yagi (2004) (strike = 250°; dip = 20°; Latitude = 41.78°N; Longitude = 144.08°E; depth = 17 km; length = 130 km; width = 170 km); the geometry of Yagi is representative of those reported for this event (e.g. Ito et al., 2004). Figure 3.1.1 shows the main shock fault plane and the location of recording stations (by NEHRP site class); aftershock locations are also indicated. In order to assign the fault plane geometry for the aftershocks, I used the relations of Wells and Coppersmith (1994) to

assign fault lengths and widths based on event magnitude. I used the hypocentral locations given either by USGS or by F-NET (Fukuyama et al., 1996) and assumed that the hypocenter was located in the center of the fault plane, and that the fault strike and dip matches that of the main shock. For the smaller events, the difference between R_{cd} and hypocentral distance is not significant, but it is important for the main shock, and to some extent the largest aftershocks (**M**: 7.3, 6.4). More detailed information about fault plane geometry for each event is given in Table 3.4.

Five events were selected from the earthquake sequence for detailed analysis, including the main shock and its largest aftershocks; the selected events have moment magnitude **M**: 8.1, 7.3, 6.4, 5.9 and 5.5. The acceleration records were visually examined, windowed from the S-wave arrival up to a cut-off time equivalent to 90% of the signal energy (calculated as the sum of the squared amplitudes), and baseline corrected. Fourier acceleration spectra ($Faccn$) were calculated for all records, and used to calculate average $Faccn$ for frequencies from 0.1 to 20 Hz, in log-spaced frequency bins 0.1 log units in width. No systematic differences between the two horizontal components (EW, NS) were observed, as is shown in Figure 3.1.2 for **M** 8.1 data, where I plotted for low and high frequencies, differences of $Faccn$ (between EW and NS components) versus distance; Figure 3.1.5 also shows the same behavior but for PGA data. Thus, I took the geometric mean of the two horizontal components as the ground-motion variable for analysis.

3.1.2 Site classification

Shear wave velocity profiles available at each K-NET recording site were used to classify the sites according to their NEHRP class. The NEHRP site class is based on the shear-wave velocity time-averaged over the top 30m of the soil profile (V_{S30}), where each class, from A to F is defined as:

Class	Soil type	Range of V_{S30} (m/s)
A	Hard rock	$V_{S30} > 1500$
B	Rock	$760 < V_{S30} \leq 1500$
C	Very dense soil and soft rock	$360 < V_{S30} \leq 760$
D	Stiff soil	$180 \leq V_{S30} \leq 360$
E	Soil profile with:	$V_{S30} < 180$
F	Soil requiring site-specific evaluations	

For the K-NET sites, soil profiles are generally only available for the top 10 or 20 m. I used two different approaches, explained in Boore (2004), to extrapolate the velocity profiles from 10 or 20 m down to 30 m in order to obtain the appropriate average velocity over 30 m. Both approaches are based on the time-averaged velocity $\bar{V}_s(d)$ to a depth d , where d is generally the depth to the bottom of the velocity model and is given by,

$$\bar{V}_s(d) = d / \tau(d), \quad (3.1)$$

where $\tau(d)$ is the travel time to depth d defined as,

$$\tau(d) = \int_0^d \frac{dz}{V_s(z)}, \quad (3.2)$$

$V_s(z)$ in expression 3.2 is the velocity model as a function of depth. The first approach (“constant” approach) assumes that the velocity obtained at the deepest available level of

the profile (10 or 20 m) continues as a constant value to a depth of 30 m, therefore, V_{S30} under this assumption is given by,

$$V_{S30} = 30 / (t(d) + (30 - d) / V_{eff}), \quad (3.3)$$

where V_{eff} is the assumed (effective) velocity from d to 30 m. The second approach is based on empirical correlation between the average velocity over the top 20 m (V_{S20}) and the corresponding average over the top 30 m (V_{S30}). Boore (2004) provided such an empirical correlation based on calculations derived from soil profile data in California. The correlation can be used to estimate V_{S30} based on V_{S20} . It was found that for most sites these two methods resulted in the same NEHRP site classification. However, for 10% to 14% of the sites (depending on the analyzed magnitude), the NEHRP site class varied by one unit depending on the approach taken. These few cases (sites) had information available only to 10 m so that the “constant” approach resulted in a softer classification (Figure 3.1.3). Because most of these 10 m profiles reported rock conditions from the first 3 or 4 m down to 10 m with high and constant velocities, I adopted the “constant” approach result as more representative for these cases; A further consideration was that the correlation approach uses results from California soil conditions, which are not necessarily representative of Japanese soil conditions. Moreover, from preliminary results of the regression analysis, the few differences in classification do not significantly modify the regression results.

3.1.3 Distance constraints

Figure 3.1.4 shows the overall behavior of the Fourier spectral acceleration ($Faccn$) versus distance. It plots log-averaged $Faccn$ values and their standard deviation for selected frequencies in distance bins that are roughly 0.25 log units in width. The attenuation plots illustrate a flattening of the spectral decay curves at larger distances and higher frequencies. Careful examination of this phenomenon based on many plots like that of Figure 3.1.4 shows that this effect is dependent on both magnitude and frequency, occurring at progressively closer distances as magnitude decreases and frequency increases. Figure 3.1.5 shows the problem in more detail for peak ground acceleration (PGA), for illustration purposes. It is a consequence of amplitude limitations imposed by the K-NET instrumentation (Strasser and Boomer, 2005, a; Kanno et al., 2006): even for acceptable signal to noise ratio values (~ 3), there may be ground-motion amplitudes that are lower than the instrument's resolution capabilities. This problem imposes limitations on the K-NET data that can be reliably used. Consequently, the determination of what cut-off distances to use for the data analysis was based on detailed inspections of plots such as Figure 3.1.4 and 3.1.5 for each event. I also checked the sensitivity of preliminary regression results to the choice of cut-off distances. I then adopted the following distance cut-offs for each event:

- **M 8.1** $R_{cut} = 325$ km
- **M 7.3** $R_{cut} = 350$ km
- **M 6.4** $R_{cut} = 250$ km
- **M 5.9** $R_{cut} = 150$ km
- **M 5.5** $R_{cut} = 200$ km

Data beyond these distances are not used. Furthermore, data at frequencies higher than 10 Hz are not used.

3.1.4 Anomalous ground-motion intensities for the Hokkaido region, NE Japan

Another important feature that can be observed by visual inspection of the Fourier amplitude data is that there is a remarkable difference in amplitudes between stations sited in different geologic conditions, for the same NEHRP site class, at similar distances. This is illustrated in Figure 3.1.6 by comparing *Faccn* for several pairs of stations located at similar distances on the same NEHRP site class, but occurring in either the fore-arc (FAR) or back-arc (BAR) regions of Japan. Figure 3.1.7 shows the division of NE Japan into fore-arc and back-arc regions (USGS, 2001). Fore-arc stations have significantly greater high-frequency content than do back-arc stations, suggesting that an additional site amplification (or attenuation) factor is required to accommodate this effect.

Two other geologic variables are also defined on Figure 3.1.7; data exploration and previous studies (Hashida, 1989) revealed they might also be significant factors in observed amplitudes. These factors are: (i) the relative age of the materials – Cenozoic (CZC) or Paleozoic- Mesozoic (PMZ); and (ii) whether the site is located in a sedimentary basin province (BAS) or more generally within the volcanic island arc province (VOL). Thus for each site I compile the NEHRP site class, and three additional geological dummy variables to represent fore-arc vs. back-arc, Cenozoic vs. Paleozoic- Mesozoic, and sedimentary basin vs. volcanic island arc province. This information is used in a second stage of the regression analysis to define the additional amplification

and / or attenuation effects that are attributable to these factors, a topic that is covered in section 3.2.4.

Figure 3.1.1 - K-NET, NIED seismic stations sites classified according to NEHRP (BSSC, 2001) criterion, epicenter of the September 26, 2003, **M** 8.1 Tokachi-Oki earthquake (main shock) and some of its aftershocks; black solid squares represent the analyzed aftershocks. A graphical representation of the fault plane for the main shock, modified from Yagi (2004), is also shown.

Figure 3.1.2 – Differences between Fourier acceleration spectral amplitudes, plotted versus closest distance to fault, obtained from EW and NS components of acceleration records for the 2003, **M** 8.1 Tokachi-Oki earthquake. Plots are for frequencies from 0.3 Hz to 10 Hz, text boxes contains mean and standard deviation values for these differences in logarithmic units.

Figure 3.1.3 - Histograms of the site classification for the K-NET stations included in this work. Results are shown for the main shock and three aftershocks; classification is based on NEHRP code and two methods, named (see text) “constant” approach (left column), and correlation approach (right column), described in Boore (2004).

Figure 3.1.4 - Geometric mean of the horizontal Fourier spectral acceleration (F_{acn}) as a function of closest distance to fault (R_{cd}). Averages were taken in distance bins logarithmically spaced and plotted versus the average distance of the corresponding bin (line markers). Three frequencies are plotted for each of the four magnitudes. The number below the square marker represents the number of data in each distance bin. Error bars on the 1.25 Hz line are $\pm$ one standard deviation.

Figure 3.1.5 - Peak ground acceleration (PGA) as a function of closest distance to fault (R_{cd}). Data from E-W and N-S components are plotted for four magnitudes. Visual estimation of the cut off distance, from where the flattening effect starts, corresponds to (in km): 325 (**M 8.1**), 350 (**M 7.3**), 250 (**M 6.4**), 150 (**M 5.9**) and 175 (**M 5.5**).

Figure 3.1.6 - Fourier spectral acceleration ($Faccn$) as a function of frequency (f) observed on stations located at similar distances, on same NEHRP site class, but in different geological regions (fore-arc vs. back-arc).

Figure 3.1.7 - Geological information (USGS, 2001) used to derive the geological variables: *FAR* (fore-arc region), *BAR* (back-arc region), both separated by the volcanic front represented by the black dotted line; *PMZ* (Paleozoic-Mesozoic rocks region), denoted by black plus symbols, *CZC* (Cenozoic rocks region) defined by the area which is not *PMZ*; *BAS* (sedimentary basin region), denoted by grey dot symbols) and *VOL* (volcanic island arc region), defined by the area which is not *BAS*. Each site was assigned a triplet of variables according to its location and this geological information.

3.2 Empirical data analysis

Linear regression techniques (Davis, J., 1986; Draper and Smith, 1998) are used to determine the source, attenuation, and site operators that characterize the observed *Faccn* data. Because the data are so abundant, I analyze each event independently, to obtain the maximum information on its attenuation and source parameters, free from trade-offs in regression parameters between events. It is important to emphasize that data are limited to distances beyond 40 km, and thus it is not possible to study near-source attenuation properties; this is a general condition at subduction zones for interface earthquakes due to their offshore locations. Nevertheless, results applicable to $R_{cd} > 40$ km are obtained, which are crucial in relation to seismic hazard from large subduction earthquakes, not only in Japan but also in other regions such as Cascadia, Mexico, or South America.

3.2.1 Regression analysis

The general representation of observed spectral amplitudes can be written as:

$$A(f, R) = A_0(f) \cdot D(f, R) \cdot S(f) \quad (3.4)$$

where $A(f, R)$ is the observed spectrum, $A_0(f)$ represents source spectral amplitude, $D(f, R)$ is a diminution function that accounts for geometric spreading and inelastic attenuation, and $S(f)$ represents local or regional site amplification functions. I use the following specific form to model the attenuation and site effects by a linear equation:

$$\log_{10} A_j(f, R) = \log_{10} A_0(f) + b_1 \log_{10} R_j + b_2 R_j + b_3 S_j^D + b_4 S_j^E \quad (3.5)$$

where $A_j(f, R)$ is the observed *Facen* at the j^{th} station (geometric mean of two horizontal components), $A_0(f)$ is source spectral amplitude, R_j is the closest distance to the fault (R_{cd}) for the j^{th} station, $s_j^D = \begin{cases} 1, & j = D \\ 0, & j \neq D \end{cases}$ and $s_j^E = \begin{cases} 1, & j = E \\ 0, & j \neq E \end{cases}$ are dummy variables (Draper and Smith, 1998) assigned to D and E site conditions respectively. The regression coefficients b_1 and b_2 describe the geometric spreading and inelastic attenuation respectively, while b_3 and b_4 provide the amplification for NEHRP D sites and E sites respectively, both of which are relative to the reference site class of NEHRP C. NEHRP C was chosen as the reference site condition as it is the firmest site condition for which there are abundant observations. It is noted that Equation 3.5 is a simple linear model, as opposed to a more complicated shape such as a hinged bilinear or trilinear model (Atkinson and Mereu, 1992). It was determined that a more complex form is not required by the data; this could be due in part to the fact that observations are restricted to distances beyond 40 km.

A complication in applying Equation 3.5 is posed by the possibility of nonlinear site amplification for the larger events and closer distances. As seen on Figure 3.1.5, PGA for the main shock generally exceeds 10%g at distances < 100 km, and thus a nonlinear response would be expected (Boore and Atkinson, 2007). To accommodate nonlinearity, I use a correction to the observed data amplitudes to bring them up to the level that it would be expected based on an entirely linear response (consistent with Equation 3.5). This correction is based on the formulation of Boore and Atkinson (2007). These authors provide an estimate of the nonlinear amplification effect based on PGA

and average shear wave velocity, V_{S30} , as derived from the empirical analyses of Choi and Stewart (2005). In this work, I use a slightly simplified version of their site factors; this version (Boore and Atkinson, 2006) is not significantly different from their final formulation adopted in 2007, which simply smoothes the transition from linear to non-linear factors that occurs near 60 cm/s^2 . The additive soil amplification factors in $\ln$ units are:

$$S = S_{lin} + S_{nl} = b_{lin} \ln\left(\frac{V_{30}}{V_{ref}}\right) + b_{nl} \ln\left(\frac{60}{100}\right) \quad \text{for } PGA_{B/C} \leq 60 \text{ [cm/s}^2\text{]} \quad (3.6a)$$

$$S = S_{lin} + S_{nl} = b_{lin} \ln\left(\frac{V_{30}}{V_{ref}}\right) + b_{nl} \ln\left(\frac{PGA_{B/C}}{100}\right) \quad \text{for } PGA_{B/C} > 60 \text{ [cm/s}^2\text{]} \quad (3.6b)$$

where S_{lin} denotes the linear term, S_{nl} the non linear term, $V_{ref} = 760 \text{ m/s}$ (B/C boundary) and slopes b_{lin} and b_{nl} are coefficients that are dependent on frequency and shear wave velocity (given in Boore and Atkinson, 2007). $PGA_{B/C}$ is the predicted PGA for B/C boundary site conditions.

These equations can be used to estimate the nonlinear part of the site amplification, provided we know V_{S30} and $PGA_{B/C}$. I obtain an estimate of $PGA_{B/C}$ for this purpose as follows. First, I fit the PGA data for the main shock as recorded on NEHRP C sites to: $\log_{10} PGA = c_0 + c_1 \log_{10} R + c_2 R$. Then I use the linear factor suggested by Equation 3.6a to convert the predictive PGA equation for C to the equivalent curve for B/C conditions. This is illustrated in Figure 3.2.1; note that the predictive equation implies nonlinear behavior for soil sites for the main shock at

distances inside 125 km ($PGA_{B/C} \approx 60 \text{ cm/s}^2$ at $R = 125 \text{ km}$). The predictive equation for $PGA_{B/C}$ is:

$$\log_{10} PGA_{B/C} = 3.84 - 0.7 \log R - 0.0043R \quad \text{for } 40 \leq R \leq 350 \text{ [km]} \quad (3.7)$$

Based on the predicted $PGA_{B/C}$ from Equation 3.7 and the soil response equations above (3.6b), I calculate the nonlinear response factors S_{nl} . These factors are added to each soil site inside 125 km (by multiplying the observed amplitudes by $\exp(S_{nl})$); this acts to increase the $Faccn$ values, compensating for the amount by which nonlinearity had presumably reduced them. Note that amplitudes are not adjusted for the linear part of the site response, as the linear site amplification factors were determined for each site class, relative to C, in the regression (Equation 3.5). Figure 3.2.2 shows examples of the nonlinear correction factors at $f = 1$ and 5 Hz for different levels of V_{s30} , as a function of $PGA_{B/C}$ and /or closest distance R_{cd} . Nonlinear corrections are only applied for the main shock, as predicted amplitudes for the aftershocks do not reach into the nonlinear range except for a few points.

The database for regression analysis is comprised of the nonlinear-corrected $Faccn$ amplitude data within the cut-off distance for data validity (independent variable), and the predictor variables of Equation 3.5. Regressions are performed separately for each event. I evaluate the goodness-of-fit by analyzing residuals, where the residual is defined as the ratio of observed to predicted $Faccn$.

There is a trade-off in regression analyses between the geometric spreading coefficient, b_1 , and the inelastic coefficient, b_2 . The former controls the slope of attenuation (in log-log space), while the latter controls its curvature. Based on

preliminary regressions for all events, I determined that best-fit values of b_1 are near -1.0; therefore this coefficient was constrained to $b_1 = -1.0$ for all regressions to improve their stability (and repeatability of interpretation regarding attenuation from one event to another). This is also in agreement with previously reported results for subduction zones (Crouse et al., 1988; Youngs et al., 1997; Atkinson, 1997). Thus the regression analysis determines for each event the apparent inelastic attenuation parameter, the apparent source spectrum (on C sites), and the linear site amplification terms relative to NEHRP C.

3.2.2 Attenuation model

The apparent inelastic attenuation of each event, relative to the fixed geometric spreading coefficient of -1.0, is given by the regression coefficient b_2 . This parameter is often linked to the regional Quality factor (Q), to which it is inversely proportional:

$$b_2 = (-\pi f) / (\ln(10) \beta Q(f)), \quad (3.8)$$

where I assume $\beta=3.6$ km/s. I plot the inferred Q values on Figure 3.2.3 for each event, as a function of frequency, along with a simple average trend that is based most heavily on the M 8.1 results (as they cover the largest distance range), but which matches the data for all events reasonably well:

$$Q = 135 f^{0.76}, \quad (3.9)$$

Although the inelastic attenuation has been expressed in terms of equivalent Q , it should be acknowledged that this function is entirely linked to the assumed geometric spreading of -1.0, and would not match the Q function that would be obtained from the more traditional choice of spreading factor of -0.5 at large distances (for surface waves).

Thus I use this apparent Q model only in conjunction with the assumed geometric spreading of -1, and not as a general Q model for NE Japan. This single Q model is assumed for all events for all remaining analyses in this study. It provides a reasonable fit to most events, but it was noted in examining the inelastic attenuation that the **M** 7.3 event has noticeably higher apparent Q than other events, of both higher and lower magnitude (Figure 3.2.3). Based on the 53 km depth of this event and its focal mechanism (discussed in the source parameter section), it is quite likely that this earthquake occurred within the subducting slab, and is thus an in-slab rather than an interface event (Ito et al., 2004). It could be that this is the reason for the higher apparent Q .

3.2.3 Site Amplification for NEHRP classes

For each event, the regression provides the apparent site amplifications for D and E sites, relative to the reference of NEHRP C. On Figure 3.2.4, I plot the amplification for each NEHRP site class as a function of frequency, for each event. Of course, these amplifications should ideally be the same, or nearly so, for all events (i.e.: the site amplification should be independent of the source). They are indeed similar, as shown on Figure 3.2.4. The apparent differences in amplification for any specific site class for different events could be related to differences in data spatial coverage for each case and/or random variability.

The regression cannot determine the amplification function for NEHRP C, as this is the reference condition. In order to make the overall amplification model more

realistic, and allow the inference of source parameters, I have estimated a model for the amplification of NEHRP C (as shown on Figure 3.2.4). This model was developed by using an average velocity profile for the analyzed KNET-NEHRP C sites, which was extended from 20 m to mid-crustal depths based on the velocity model of Iwasaki et al. (1989). These authors provided P wave velocity profiles for the crustal structure of the Kurile trench (off Hokkaido region) based on OBS (ocean bottom seismograph) data. Their interpretation for the continental side covers most of the continental slope (close to the SW Hokkaido coast) and is thus a reasonable representation for the crustal conditions for the fore-arc area. I estimate the shear wave velocity model assuming $V_p/V_s = \sqrt{3}$ (Katsumata et al., 2003; Nishizawa and Suyehiro, 1986). The assumed velocity model for the reference C condition is given in Table 3.1. Quarter-wavelength calculations, based on the method introduced by Joyner et al. (1981), were used to estimate the amplification that would be expected for C sites for this profile. Under the quarter-wavelength approach (Boore and Joyner, 1997), the site amplification as a function of frequency is given by

$$A[f(z)] = \sqrt{\rho_s \beta_s / \bar{\rho}(z) \bar{\beta}(z)}, \quad (3.10)$$

where the product $\rho_s \beta_s$ represents the seismic impedance at source depth, and $\bar{\rho}(z) \bar{\beta}(z)$ is the seismic impedance averaged over a depth z , the average velocity is given by $\bar{\beta}(z) = z / S_u(z)$ where $S_u(z)$ is the shear wave velocity travel time to a depth z . The average velocity and travel time to depth z correspond, for the soil profile case, to equations 3.1 and 3.2 respectively. The frequency corresponding to depth z is

$f(z) = 1/4S_{\mu}(z)$ and the density $\bar{\rho}(z)$ is the travel-time weighted average density over a depth z .

Specifically, I used the computer program SITE_AMP (Boore, 2003) with parameters: $\kappa_0 = 0.045$, as the value to account for the high frequency attenuation effect (Anderson and Hough, 1984); $i = 45^\circ$, for the angle of incidence; ρ (density) = 2.8 g/cm^3 and β (shear wave velocity) = 3.6 km/s , as the source conditions.

The attenuation parameter, κ , describes high-frequency spectral decay. It increases with distance, due to inelastic attenuation effects, but has a residual value κ_0 at short distances. This κ_0 value, an important parameter in high-frequency modeling, was calculated from the linear fit of κ measurements as a function of distance for the **M** 8.1 *Faccn* data for C sites (a topic that is covered in section 3.2.6). The choice of $i = 45^\circ$ is largely arbitrary, but it was noted that the amplification function determined for $i = 45^\circ$ matches the average amplification function that would be inferred from analysis of the H/V ratios (Lermo and Chávez-García, 1993), and I chose it for this reason. The plotted reference amplification curve for C sites on Figure 3.2.4 represents the combined effect of amplification and attenuation, based on the quarter-wavelength calculations. Then the curves for D and E were obtained using the regression results relative to C. The heavy lines show the smoothed average functions that I obtained for the amplification of D and E sites, considering all five events. These amplifications are generally consistent with other empirical results for NEHRP C, D and E amplification functions in other regions (Boore et al., 1997; Atkinson and Boore, 2003). In particular, the Atkinson and Boore analysis, from a worldwide data base for subduction earthquakes, shows similar values

for the amplification function for different site classes, though minor differences are observable; these may be attributable to differences in shallow geological conditions (i.e.: soil thickness, depth to the top of rock) between NE Japan and other subduction regions. The overall behavior of the site amplification functions is as expected, in that softer soil amplifies ground-motions at lower frequencies. Note that these represent linear amplification functions. The smoothed average values for NEHRP C, D and E linear amplification of Figure 3.2.4, listed in Table 3.2, are adopted for use in all remaining analyses of this study. The values are given to 2 significant figures in Table 3.2 to produce a smooth function, but it should be noted they are not known to this precision.

3.2.4 Geological factors

Another site response issue is that of regional variability of site response and/or path effects in Japan, such as those noted in recent studies by Kanno et al., (2006), Takai et al. (2006) and Maeda and Sasantani (2006). Ground-motion observations in Japan may be conditioned by Q structure effects (Hashida, 1989); seismic waves originating in the Kurile, Japan or Izu-Bonin trench that are recorded at stations in the fore-arc region will travel shorter distances in a low- Q zone than those that propagate to the back-arc sites. It is interesting to note that the obtained attenuation model: $b_I = -1.0$, $Q = 135 f^{0.76}$ represents a steeper attenuation compared to the subduction zone attenuation models of Ordaz and Singh (1992) and Atkinson (2005). These authors analyzed mainly fore-arc ground-motion data. By contrast, this study model is similar to that of Singh et al. (2006), who found $Q = 120 f^{0.75}$ for the mantle wedge in the Mexican subduction zone,

including both fore-arc and back-arc information. Thus the observed attenuation rates may be quite sensitive to whether back-arc stations are included. Although this is really a path effect, and may thus depend on distance, a simple way to investigate it is by looking at whether there is a difference in residuals depending on whether a station is in the fore-arc or back-arc region of NE Japan (Figure 3.1.7).

Another factor that may influence site ground-motions is the age of the geologic deposits. Two approaches were used to analyze this factor; the first one is based on the relation between thermal conditions and/or rock age and seismic attenuation (Q) (Hashida, 1989; Artemieva et al., 2004; Karato, 1993; Roth and Wiens, 2000). For example, Hashida (1989) concluded that there are variations in Q structure at 30 km depth from values as low as 50 up to as large as 1000 for the hotter/younger versus colder/older zones in NE Japan, respectively. The second approach takes into consideration the tectonic evolution of the region, particularly its basin development. Northeast Japan is characterized by a wide distribution of late Cenozoic back-arc sediments, which fill Miocene extensional basins (Sato, 1994). Given that these basins are part of the volcanic island arc evolution, they are referred to as volcanic basins (Neogene sediments in Figure 2.2). The age serves as a proxy for identifying these volcanic basins.

Preliminary analyses from the first approach showed that the expected results (higher attenuation for younger / hotter rocks, in this case Quaternary volcanic material) were not observed. I concluded that the factor for the rock age (Quaternary volcanic) was not significant. In the second approach, the age factor (Cenozoic material in Figure 3.1.7)

is used to map the numerous volcanic basins (Figure 2.2), to allow assessment of whether these structures have significant regional effects (e.g., as found by Cárdenas-Soto and Chávez-García, 2003).

Finally, Japan contains two significant sedimentary basins within the framework of its overall volcanic island arc composition: The Ishikari-Hidaka and the Tokachi basins (Figure 3.1.7). These basins were developed in a different way than the volcanic basins (Yanagimoto and Iijima, 2003); they are more recent and are best considered as sedimentary basins.

To consider the influence of the factors discussed above, I have classified each site using additional geologic variables to distinguish fore-arc vs. back-arc, Cenozoic vs. Paleozoic- Mesozoic, and sedimentary basin vs. volcanic island arc province. Figure 3.1.7 shows the geographic classification used. I treat each of these as a dummy variable and introduce them in the regression scheme (Equation 3.5) (Draper and Smith, 1998), in the same way that the site class was treated as a dummy variable. Thus I organized these geological variables in three columns of dummy variables such that for the j^{th} station; $S_j^{BAR} = \begin{cases} 1, & j=BAR \\ 0, & j=FAR \end{cases}$, $S_j^{CZC} = \begin{cases} 1, & j=CZC \\ 0, & j=PMZ \end{cases}$ and $S_j^{BAS} = \begin{cases} 1, & j=BAS \\ 0, & j=VOL \end{cases}$. For example, a site located in the fore-arc region, on Paleozoic-Mesozoic material within a sedimentary basin will have the dummy variable array [001]. Additionally, the geological variable associated to a dummy = 0 represents the reference level of the complimentary geological variable (e.g., *FAR* is the reference for the *BAR* result).

For the analysis of the rock age variable, I excluded from the CZC group all the sites (dummy variables) located on sedimentary basins (by assigning them a CZC dummy

variable of 0). In this way, the CZC variable represents strictly the spatial distribution of volcanic basins, without any “double-counting” of sedimentary basin effects.

I repeated the regression in two ways: (i) by simply adding these new variables into the regression; and (ii) by first fixing the site response factors for D and E, as already determined. The results were very similar for both cases, indicating that these factors represent a robust regional characteristic of the site effect. The BAR vs. FAR and the BAS vs. VOL cases are robustly determined in the data; the factor for rock age in the **M** 8.1 case is well determined too, but this factor is ambiguous for the **M** 7.3 case. Figure 3.2.5 shows the coefficient values (amplification factors) for each geologic variable versus frequency, for regression of the two largest events. This can be interpreted as the relative effect (difference in level) of a specific geological condition on the amplitudes of ground-motion, with respect to the complimentary geological condition (reference level of 0). From Figure 3.2.5, it is evident that there is significant additional attenuation experienced at back-arc stations, reducing observed ground-motions for frequencies above 1 Hz; The enhanced ground-motions at low frequencies within the sedimentary basins is also remarkable. Finally, the age factor is also showing an additional amplification effect, perhaps attributable to their volcanic basin origin, at least for the **M** 8.1 event at long periods. The difference in the significance of the age factor between the **M** 8.1 and **M** 7.3 events could be explained by the fact that this type of effect has been related to the presence of surface waves (Cárdenas-Soto and Chávez-García, 2003); the main shock is shallower than the **M** 7.3 earthquake. The differences in relative amplification levels between the BAS and CZC factors could be related to possible

differences in impedance contrast between basement rocks and sedimentary basins (more recent, less consolidated) and between basement rocks and volcanic basins (older and more consolidated).

3.2.5 Source parameters

Based on the adopted common regression model, I corrected all *Faccn* data for each event back to a reference near-source distance and site condition, and then averaged them to obtain a mean apparent source spectrum. The reference site condition used was hard rock; thus I divided each recorded spectrum by the C, D or E amplification function of Table 3.2 according to the site condition.

The acceleration spectrum at any distance can be compared to the corresponding spectra expected for a Brune model (Brune, 1970, 1971) using the following (Boore, 1983):

$$A(f) = CM_0 \left[(2\pi f)^2 / 1 + (f / f_0)^2 \right] \cdot e^{-\pi f R / \beta Q(f)} \cdot R^{b_1} \cdot e^{-\pi f \kappa_0} \quad (3.11)$$

where $C = R_{\theta\phi} FV / 4\pi\rho\beta^3$, $R_{\theta\phi}$ (average radiation pattern for shear waves)(Boore, 1983) = 0.55, F (free-surface amplification) = 2.0, V (partition onto two horizontal components) = 0.71, ρ (density) = 2.8 g/cm³, β (shear wave velocity) = 3.6 km/s, M_0 = seismic moment (dyne-cm) and the corner frequency, f_0 , is related to stress drop in bars ($\Delta\sigma$) by $f_0 = 4.9 \times 10^6 \beta (\Delta\sigma / M_0)^{1/3}$ (Boore, 1983). The stress drop can be inferred from the high-frequency level of the acceleration spectrum (at $f \gg f_0$), while seismic moment can

be determined from the low-frequency level of the displacement spectrum (at $f \ll f_0$, subject to bandwidth limitations) (Boore, 1983).

Any convenient reference distance can be chosen to compare the average source spectra to the Brune model, provided that the same attenuation model is used to perform the attenuation corrections for both the observed data and the Brune model. A common choice is to use a reference distance of 40 km (eg. Malagnini et al, 2000). In this study, I make the comparisons to model spectra at greater distances, so that the model spectra can be compared to both the Brune model and to individual observations. Because of the reference distances used, both the Brune-model spectra and the observations consider the inelastic attenuation correction ($b_2 R$), in addition to the standard geometric spreading correction applied as in Equation 3.11 ($R^{b'}$).

Figures 3.2.6 and 3.2.7 show these comparisons between average observed spectra and Brune-model spectra, from which the seismic moment and Brune stress can be inferred. Plots are made showing the displacement and acceleration spectra of each event at two selected reference distances, in comparison to those for the best-fit Brune model. The long-period end of the displacement spectrum is used to infer the seismic moment, which is given for each event on Table 3.3, along with published values from other sources (U.S. Geological Survey (USGS) at www.neic.usgs.gov/neis/sopar and / or by F-NET at www.fnet.bosai.go.jp/freesia). For the largest event (**M** 8.1) there is not sufficient bandwidth to observe the long-period end of the spectrum, so I simply adopt the published seismic moment values from other studies (Table 3.3); I also adopt a published value for the **M** 7.3 event, as the bandwidth to determine moment for this event

is marginal. With the seismic moment set, I then use Equation 3.11 to estimate the stress parameter from the high-frequency level of the acceleration spectrum; an appropriate value of kappa is selected to match the observed spectral decay of each event at higher frequencies. It is important to emphasize that the selected κ value in this matching procedure is, in fact, a residual κ value because an overall kappa factor ($\kappa_0 = 0.045$) was already considered in the C amplification reference function (Figure 3.2.4); the total kappa is thus the sum of the assumed overall kappa and the residual kappa for each event. The obtained stress parameter and residual kappa values are also given in Table 3.3.

The purpose of plotting the apparent source spectra at two reference distances in the range of the data on Figures 3.2.6 and 3.2.7 is to include direct comparisons of these spectra to individual observations. In each plot, I show not just the model spectra, but observed spectra (corrected just for site conditions) for those records that are at that distance (± 5 km). This shows that the model spectra are describing observations well.

In Table 3.3, I observe that the moment magnitudes that were inferred from the near-source displacement spectra are in good agreement with moment solutions from other sources. The Brune stress parameters inferred for the events range from 100 to 200 bars, with no systematic dependence on magnitude. The stress parameter values for the events are high relative to those typically inferred for large crustal events. For example, the M 7.6 1999 Chi Chi Taiwan event had a stress drop near 50 bars based on stochastic finite-fault modeling (Roumelioti and Beresnev, 2003), while large California events typically have stress drops of about 50 to 80 bars (Atkinson and Silva, 1997). The large

stress parameter values have important implications for ground-motion hazards from great subduction earthquakes.

Another observed feature of the inferred near-source spectra is that they match the shape of the Brune model well except for the largest two events. Both the **M** 7.3 and **M** 8.1 events show significant spectral “sag” at intermediate periods relative to the Brune model. This shape is consistent with empirical observations for large events in other regions such as California (Atkinson and Silva, 1997, 2000). The shape is believed to be a consequence of finite-fault effects (Beresnev and Atkinson, 2002).

3.2.6 Duration of ground-motion

The duration of ground-motion can largely determine the level of damage of a specific earthquake. Large magnitude earthquakes occurring at far distances may have similar energy content than smaller magnitude events occurring at shorter distances but usually have longer duration of ground-motion and potentially larger damage (Kramer, 1996). Several measures of duration of ground-motion have been proposed: Raoof (1999) uses a time window limited by the 5% and 75% values of the integrated square velocity of filtered (band-passed) signals; Atkinson (1993) links, under the context of random vibration theory and from Parseval’s theorem, time-domain and frequency-domain amplitudes to measure duration of motion; Trifunac and Brady (1975) use the time interval that contains usually 90% of the signal energy. In all of these studies, the distance dependence of duration is assumed. In this research I adopt a Trifunac and Brady type measure of duration, where the time window is defined for the 90% of the signal

energy, measured from the S wave arrival time and also adopt the distance dependence of duration modeled as:

$$T = T_0 + dR, \quad (3.12)$$

where T_0 describes the source duration and d is the distance gradient of the model.

In order to estimate the distance term of equation 3.12, I fit a single line (some other researchers have used more complicated forms of distance dependent duration models, e.g.: Atkinson and Boore, 1995) to duration (as defined above) versus closest distance (R_{cd}) data, for the **M** 8.1 earthquake and some of its aftershocks. In Figure 3.2.8 I plot, for the E-W components of the main shock records, duration versus distance data along with the fitted line with their corresponding prediction bounds (95% confidence intervals). Data are shown for each NEHRP class (C, D, and E) and the general fitted model, which represents all data, is also plotted. This model is very similar to the models obtained for each particular site class (not plotted in Figure 3.2.8 but described in its text box). It is clear that practically all data are within the general model prediction bounds and that differences between site classes are not significant. Figure 3.2.9 emphasizes the previous observation (similar trend between different site classes) but for the **M** 8.1 N-S component data. However, because the data are limited to $R_{cd} \geq 50$ km, the single line fit will poorly determine the source term T_0 , thus I use reported values of source duration for the main shock (Yagi, 2004; Yamanaka and Kikuchi, 2003); thus for the final adjustments of the **M** 8.1 distance term estimation, the source term was constrained by $T_0 = 40 \pm 1$ s. Figure 3.2.10 shows the effect of the source term constraint on the distance term estimation (change from $d = 0.14$ to $d = 0.10$). Figure 3.2.11 shows the comparison

between the **M** 8.1 model for $T_0 = 40$ s (adopted value) and the models obtained for aftershocks with **M**: 6.4, 5.5, and 5.2. I did not constrain the source terms for the aftershocks because I wanted to emphasize the different trend in distance term (at least for 50 km and above) between the **M** 8.1 event and lower magnitude events. Even though the data dispersion from the model for each aftershock is large (see legend), a small difference in distance slope between the aftershocks and the main shock is observable. It is also interesting that, if reasonable values for the aftershock source durations (reported or calculated) are considered (much lower values than the ones in legend of Figure 3.2.11, due to their magnitudes), then a single slope for the distance dependence for these earthquakes would not appear to be a reasonable model. Based on the results shown in Figures from 3.2.8 to 3.2.11, I adopt for subsequent analysis, $d = 0.09$ and $d = 0.07$ s/km for the distance-dependence of duration for the **M** 8.1 event and aftershocks respectively.

3.2.7 Kappa factor

The high frequency attenuation shape of Fourier acceleration spectra for the 2003 Tokachi-Oki earthquake sequence is characterized through the kappa parameter (κ) (Anderson and Hough, 1984). Hanks (1982) suggested that the general trend of the shear wave Fourier acceleration spectrum (F_{accn}) is flat above the corner frequency f_0 to a second frequency (f_{max}) above which the spectrum decays rapidly. He concluded that f_{max} showed a strong dependence on site conditions and that small but significant differences of f_{max} correlate with gross geologic features of recording stations. Subsequently, Anderson and Hough (1984) identified that the exponential decay of F_{accn} starts at

frequencies larger than f_0 but smaller than f_{max} , referred as f_E (effective frequency). They describe the shape of $Faccn$ at high frequencies by

$$a(f) = A_0 e^{-\pi \kappa f} \quad f > f_E, \quad (3.13)$$

where A_0 depends on source properties and epicentral distance, and κ , the spectral decay parameter, depends on distance (as it includes inelastic attenuation) and on local site conditions. To describe the behavior of κ , these authors converted slope values of high frequency $Faccn$ to κ values and plotted against epicentral distance to evaluate, by linear fitting, the distance dependence of κ . They found that κ tends to finite values as distance approaches to zero (referred as κ_0), and related this parameter (κ_0) to subsurface geologic conditions.

In this work, I measured (for the main shock and some aftershocks) the decay slopes of $Faccn$ on log-linear plots of $Faccn$ versus frequency (roughly between 5 and 15 Hz) and then converted them to κ values:

$$\kappa = \frac{\Delta \log Faccn}{\Delta f} \quad 5\text{Hz} \leq f \leq 15\text{Hz}, \quad (3.14)$$

where $\Delta \log Faccn$ is the difference (in logarithmic units) between $Faccn$ amplitudes that correspond to the frequency limits of the interval Δf ; examples are given in Figure 3.2.12a. In this figure, numbers are assigned weights that represents quality of data, where $w = 1, 2$, or 3 are poor quality information, and good quality data from incomplete and complete records respectively; suspicious (possible misclassified sites) data were assigned $w = 0$. Figure 3.2.12b shows the linear fit (κ vs R_{cd}) obtained from **M** 8.1, NS

components data when all information ($w = 1 - 4$) is analyzed. It is remarkable the difference in κ_0 values between site classes. Figures 3.2.13a and 3.2.13b show a comparison between the fitted lines, obtained from EW components data on C and D sites respectively, for the main shock and the aftershocks with magnitudes **M**: 8.1, 7.3, 6.4, 5.5. The comparison is intended to emphasize the similar κ models between magnitudes for the same site class.

Because the data are restricted to $R_{cd} > 50$ km, the estimated κ_0 values from these linear fits are poor. It is expected that the values of κ from closer distance records are lower than the ones at 50 km, and that the actual κ_0 values could be lower than those reported in Figure 3.2.12b. In order to have an idea of the range of variability of κ_0 for C sites, I compare, in Figure 3.2.14, the smoothed amplification functions obtained when the $\frac{1}{4}$ wave-length technique was applied to C sites (see section 3.2.3), calculated for angle of incidence i and κ_0 values: [$i = 45^\circ$, $\kappa_0 = 0.045$]; [$i = 0.0^\circ$, $\kappa_0 = 0.045$]; [$i = 45^\circ$, $\kappa_0 = 0.035$], against the smoothed average amplification functions obtained from applying the spectral ratios technique (H/V) (Lermo and Chávez-García, 1993), to **M** 8.1, C sites data, and to data for all magnitudes (main shock and aftershocks). It can be seen that the function calculated for $\kappa_0 = 0.035$ better fits the high frequency shape of the H/V function from **M** 8.1 data, but the functions calculated for $\kappa_0 = 0.045$ are closer to H/V from all magnitudes. Because the function for $\kappa_0 = 0.045$, $i = 45^\circ$ is closer to low and intermediate frequency amplitudes of the average H/V functions, I adopted for further analysis

(simulation procedures) $\kappa_0 = 0.045$ and $i = 45^\circ$, as the representative value for NEHRP C sites.

3.2.8 Assessment of results

Several aspects of the regression were investigated to test the robustness of the results. First, I tested the impact of the nonlinear amplification effects assumed for the observations inside 125 km for the **M** 8.1 event. Figure 3.2.15 shows the derived linear amplification functions from the regression for D and E sites when the nonlinear correction factor is assumed, in comparison to the linear amplifications obtained when nonlinearity is ignored and all sites are treated as having linear response. I observe that there are some differences in the inferred linear amplification factors for the E sites (e.g., between 15 % and 20 % for the factor at 1.5 Hz); the decrease in inferred linear amplification for E sites, when linearity is assumed, is what is expected. The impact of this noticeable change in E sites amplification factors, if linear amplification is assumed, is on the inelastic attenuation parameter, which changed from $Q_0 = 145$ to $Q_0 = 135$. However, the assumption of linear site amplification does not significantly affect the derived D and E amplification functions, nor the inferred stress drops. Thus the assumption of nonlinear site response has minimal impact on the analyses and conclusions of this research work.

Another fundamental aspect for the assessment of the results is the residual analysis. Under the regression model as described above, including the adoption of common amplification and attenuation parameters for all events, residuals (defined as the

difference between the logarithmic values (base 10) of the observed and the simulated F_{accn} ordinates) were calculated for each record, and are shown as a function of distance for selected events and frequencies on Figure 3.2.16. Average residuals in distance bins were also calculated and plotted, to make the average results more apparent. Overall, the regression model describes the data well. Residuals are scattered about zero and show no apparent trends with distance. An exception is that the residuals tend to lie below zero (implying overestimation by the model) for the **M** 7.3 event, at distances less than 100 km. This suggests that the near-source attenuation may have been different for this event (probably steeper and also noted in Figure 3.2.3) in comparison to that observed for the other events of the sequence. It is interesting that this earthquake was classified as in-slab type by Ito et al. (2004). Based on their depth and focal mechanisms (Table 3.3), other possible in-slab events among the aftershocks are the **M** 5.9 and **M** 5.5 events; due to conflicting information on depth, there is no way to reach a reliable conclusion on the classification of these two events.

The stress drop parameter values obtained from the near-source spectra analysis shown an interesting characteristic. For interface events, the best-match values were between 100 and 120 bars, whereas for the suspected in-slab events the stress drops were about 200 bars. This trend is consistent with other studies. For example, Mikumo et al. (1998) investigated the spatial and temporal variations of shear stress from failures of a large segment of the Cocos-North America plate boundary during a sequence of four large interface earthquakes (from 1979 to 1986) ranging in magnitude from **M** = 7.3 to **M** = 8.1. They found maximum stress drop values of about 80 to 100 bars for these events.

Garcia et al. (2004) analyzed 17 in-slab earthquakes for the Mexican subduction zone and compared their stress drops to those of interface earthquakes of similar magnitudes. They concluded that there are significant stress drop differences between in-slab earthquakes (average $\Delta\sigma \sim 300$ bars) and interface earthquakes (average $\Delta\sigma \sim 80$ bars) that may be attributable to source properties. For the Hokkaido region (same region as this study), Takeo et al. (1993) calculated a stress drop of $\Delta\sigma \sim 400$ bars for the **M** 7.6 in-slab, 1993 Kushiro-Oki earthquake. It thus appears reasonable to suppose that higher stress drop events may indeed be associated with in-slab sources.

The last but not the least interesting aspect of the assessment of results is the geological factors analysis. The geological variables are intended to represent what are actually complicated path effects; however, treating them as a simple site correction factor provides an initial way to represent the overall effects of the phenomena. When these additional geological effects are included in the regression analysis, the pattern and variability of the residuals in frequency and distance show significant improvement. This is illustrated in Figure 3.2.17 for the **M** 8.1 event; the standard deviation of residuals (frame a) is reduced at all frequencies by including the significant geologic variables, and the distribution of residuals versus distance is uniform (frames b, c, and d).

Table 3.1, Average shear wave velocity and density profile for NEHRP C sites (K-NET stations), extended from 20 m depth to mid crust depths on the basis of Iwasaki et al (1989).

Thickness [m]	Shear Velocity [m / s]	Density [g/cm ³]
1	144.0	1.54
1	178.7	1.55
1	245.2	1.61
1	333.9	1.68
1	367.0	1.75
1	383.0	1.77
1	437.8	1.78
1	460.4	1.79
1	477.8	1.81
1	484.3	1.83
1	487.8	1.84
1	511.5	1.84
1	512.0	1.83
1	533.0	1.85
1	534.1	1.84
1	569.8	1.86
1	580.6	1.88
1	611.9	1.89
1	612.0	1.90
1	612.0	1.93
130	943.3	2.08
250	1040.6	2.19
341	1242.0	2.15
3259	2190.0	2.30
4000	3180.0	2.60
10000	3600.0	2.80

Table 3.2, Amplification factors from 0.1 to 10 Hz for NEHRP C, D and E sites (K-NET stations). Calculation of C sites factors was based on the velocity-density profile given in Table 3.1 and the quarter-wavelength technique (Boore and Joyner, 1997) for $\kappa_0 = 0.045$. Factors for D and E sites are the smoothed versions of the average functions from regression results for all magnitudes.

Frequency [Hz]	Amplification factors NEHRP C sites	Amplification factors NEHRP D sites	Amplification factors NEHRP E sites
0.10	1.27	1.22	1.60
0.12	1.31	1.27	1.82
0.15	1.36	1.36	2.05
0.18	1.39	1.46	2.27
0.22	1.44	1.58	2.43
0.27	1.49	1.64	2.66
0.33	1.53	1.69	2.93
0.41	1.58	1.79	3.22
0.50	1.60	1.85	3.58
0.61	1.61	1.91	3.98
0.74	1.64	2.00	4.32
0.90	1.66	2.11	4.76
1.11	1.70	2.23	5.54
1.35	1.72	2.39	5.71
1.65	1.73	2.54	5.41
2.02	1.74	2.68	4.95
2.46	1.72	2.93	4.56
3.01	1.68	3.02	4.06
3.67	1.64	2.92	3.3
4.49	1.59	2.55	2.58
5.48	1.48	2.17	2.14
6.70	1.34	1.68	1.62
8.19	1.19	1.36	1.31
10.0	1.03	1.06	1.02

Table 3.3, Reported moment magnitude M and depth h [km] values for the September 26, 2003, M 8.1 Tokachi-Oki earthquake and four aftershocks, from different sources of information (see text). Calculated (this study) magnitude M , residual κ_0 and $\Delta\sigma$ [bars] for the same events, from near-source spectra. * indicates that M could not be determined due to our bandwidth limitations; in this case the quoted M is the assigned value from other studies. Our preferred moment for each event (used in calculation of $\Delta\sigma$) is indicated in bold. Reported focal mechanism parameters from two sources of information; ϕ , δ , and λ are strike, dip and rake angles respectively. Last column is the classification (**class**) given by Ito et al (2004): 1) the thrust type with nodal plane similar to the main shock (TH-type); 2) the thrust-fault type with nodal different to the main shock (OT-type); and 3) the normal fault type (NF-type). The conditions for the TH-type are: $220^\circ \leq \phi \leq 260^\circ$, $5^\circ \leq \delta \leq 35^\circ$, $100^\circ \leq \lambda \leq 140^\circ$.

M / h [km]				M	κ_0 [resid]	$\Delta\sigma$ [bars]	focal mechanism $\phi / \delta / \lambda$ [degrees]		class
JMA K-NET	F-NET	USGS PDE	USGS HRV	This Study			F-NET	USGS HRV	Ito et al. (2004)
8.1	7.91 / 23	8.12 / 13	8.29 / 28	8.12*	0.01	120	249 / 15 / 127	250 / 11 / 132	TH
7.1	7.34 / 53		7.37 / 47	7.34*	0.005	200	199 / 21 / 66	208 / 18 / 86	OT
6.5	6.43 / 41	6.36 / 31	6.47 / 35	6.41	0.005	100	239 / 23 / 116	244 / 17 / 114	TH
6.1	5.83 / 26		5.90 / 55	6.08	0.02	200	250 / 32 / 140	227 / 28 / 104	TH
5.5	5.38 / 41	5.42 / 39	5.40 / 49	5.50	0.005	200	148 / 42 / -35	148 / 48 / -38	NF

Figure 3.2.1 - Peak ground acceleration (*PGA*) data for the **M 8.1** main shock. Geometric mean data on C class sites are denoted with a C, the black line is the fit to these data considering information up to $R_{cd} = 350$ km; light gray line represents the adjusted equation for the B/C boundary ($V_{S30} = 760$ m/s).

Figure 3.2.2 - Non linear amplification factors for the M 8.1 event as a function of the expected peak ground acceleration for B/C site conditions (PGA_{rock}), or as a function of closest distance to fault (R_{cd}). $R_{cd} \approx 125$ km or $PGA_{rock} \approx 60$ cm / s² represents the cut off between non-linear amplification factors (to the right) and linear amplification factors (to the left). Factors are shown for 1 Hz (a) and 5 Hz (b).

Figure 3.2.3 - Q values implied by the inelastic coefficients of Equation 3.8 under the constraints: geometric spreading coefficient $b_1 = -1.0$ and cut off distances as shown in the legend for each magnitude. Solid line represents the fit to the data for the M 8.1 case according to $Q = Q_0 f^\eta = 135 f^{0.76}$.

Figure 3.2.4 - Absolute amplification functions obtained from the site term coefficients. The C amplification is the generic function (C reference), calculated from the quarter-wavelength technique (Boore and Joyner, 1997), the velocity-density profile given in Table 3.1 and for $\kappa_0 = 0.045$. For D and E class sites, results from individual regressions of three events (**M** 6.4, **M** 7.3, **M** 8.1) are shown along with the smoothed average of these results for each site class (thicker lines).

Figure 3.2.5 - Coefficient values (amplification factors) for each geological variable as a function of frequency, obtained from regression of the main shock and the largest aftershock. Lines show the effects for BAR, CZC and BAS variables; the reference conditions are FAR, PMZ and VOL variables respectively.

Figure 3.2.6 - Average near source Fourier spectral acceleration (Ref. Faccn; black continuous line), the Brune source model (Brune; light gray continuous line) modified by the attenuation factor and the average of Fourier spectral acceleration data (Avg. Faccn; dark gray continuous line) observed on F (gray boxes) number of stations located within ± 5 km of the analyzed distance R (gray boxes). Two R cases for each magnitude are shown to illustrate model performance. Magnitude M in the gray boxes is the preferred value (Table 3.3) to calculate stress drop (Brune spectra). Light gray dashed lines represent the attenuated displacement Brune spectra (Brune / ω^2). Black dashed lines are the average near source displacement spectra used to calculate magnitude (M this study in Table 3.3).

Figure 3.2.7 - Average near source Fourier spectral acceleration (Ref. Faccn; black continuous line), the Brune source model (Brune; light gray continuous line) modified by the attenuation factor and the average of Fourier spectral acceleration data (Avg. Faccn; dark gray continuous line) observed on F (gray boxes) number of stations located within ± 5 km of the analyzed distance R (gray boxes). Two R cases for each magnitude are shown to illustrate model performance. Magnitude M in the gray boxes is the preferred value (Table 3.3) to calculate stress drop (Brune spectra). Light gray dashed lines represent the attenuated displacement Brune spectra (Brune / ω^2). Black dashed lines are the average near source displacement spectra used to calculate magnitude (M this study in Table 3.3).

Figure 3.2.8 - Duration of motion (T) versus closest distance to fault (R_{cd}) from EW components records of the M 8.1 event. Symbols represent data for each site class (as explained in legend), continuous line is the fitted model (all data included) and dashed lines represent 95% confidence intervals of the model. Gray text box indicates the obtained models when data from each site class are analyzed separately along with the general model.

Figure 3.2.9 - Duration of motion (T) versus closest distance to fault (R_{cd}) from NS components records of the M 8.1 event. Symbols represent data for each site class (as explained in legend), continuous line is the fitted model (all data included) and dashed lines represent 95% confidence intervals of the model. Gray text box indicates the obtained models when data from each site class are analyzed separately along with the general model.

Figure 3.2.10 - Duration of motion (T) versus closest distance to fault (R_{cd}) from EW components records of the **M** 8.1 event. Symbols represent data for each site class (as explained in legend), black continuous line is the fitted model for no constraints on source duration (T_o var.), black dashed lines are 95% confidence intervals for this model, light gray continuous line is the fitted model when the source duration is fixed to $T_o = 39$ s and light gray dashed lines are the 95% confident intervals for this adjusted model.

Figure 3.2.11 – Comparison between the **M** 8.1 duration of motion model for $T_0 = 40$ s (black continuous line) and duration of motion models for aftershocks when source durations are not constrained; black dotted line, light gray dashed line and light gray continuous lines correspond to **M**: 6.4, 5.5, and 5.2 aftershocks respectively.

Figure 3.2.12 - a) Kappa values as a function of closest distance (R_{cd}) for M 8.1, D sites, NS components data; numbers (weights) represent quality of data: $w = 1, 2, 3$ are poor quality and good quality on incomplete and complete records respectively, $w = 0$ represents suspicious data (possible misclassified sites); black continuous line is the fitted line when all data ($w = 1 - 4$) are analyzed, and b) kappa models obtained as described for panel (a) but from M 8.1, NS components data on C, D and E sites separately.

Figure 3.2.13 – Kappa models (single line fits) obtained from EW components data for **M**: 8.1, 7.3, 6.4, and 5.5 earthquakes on C sites (panel a) and on D sites (pane b).

Figure 3.2.14 – Comparison between the smoothed average amplification functions (thicker lines) obtained when the $\frac{1}{4}$ wave-length technique was applied to shear wave velocity and density data for K-NET C sites (for different angle of incidence and kappa values) and the smoothed average amplification functions calculated from the spectral ratios (H/V) technique, applied to K-NET C sites. H/V amplification functions were calculated from data of all magnitudes (main shock and aftershocks) (light gray continuous line) and from data only for the **M** 8.1 earthquake (black thinner continuous line).

Figure 3.2.15 - Linear amplification functions for D and E class sites from **M 8.1** data. Two cases are shown; the first one under the assumption that non linear amplification effects were present in the data (nonlin functions in the legend), the second one when linear amplification is assumed for the data (linear functions in the legend). The inferred linear amplifications for D sites are practically the same. The generic C amplification function used in this study is also shown (C ref).

Figure 3.2.16 - Residuals of the logarithm of Fourier spectral acceleration (observed – estimated) as a function of distance (R_{cd}). Results are shown for the analyzed aftershocks, for 0.2 Hz and 3 Hz.; magnitude is indicated in the dark gray boxes. Larger symbols represent the geometric mean (log units) of the residuals contained in a specific distance bin. The distance bins were logarithmically spaced.

Figure 3.2.17 - (a) Standard deviations (as a function of frequency) of residuals of the logarithm of Fourier spectral acceleration for the **M 8.1** event. Two cases are considered: (i) with and (ii) without geological factors applied in the regression. Graphs **b**, **c** and **d** show residuals, for different frequencies, as a function of distance (R_{cd}) after geological factors are applied. Larger symbols represent the geometric mean (log units) of the residuals contained in a specific distance bin and error bars are $\pm$ one standard deviation. The distance bins were logarithmically spaced.

Chapter 4. Stochastic finite-fault modeling of the 2003, M 8.1 Tokachi-Oki earthquake sequence

One of the most used tools in the analysis and prediction of observed ground motion as a function of magnitude and distance has been the stochastic model. This method assumes that high frequency ground motions can be represented as band-limited Gaussian noise, with underlying amplitude spectrum defined by a seismological model of source and propagation processes (Hanks and McGuire, 1981; Boore, 1983; Boore and Atkinson, 1987; Atkinson and Boore, 1995). In the simplest implementations of this model, earthquakes are treated as point sources, an assumption that is difficult to sustain if we want to know ground motion amplitudes close to the causative fault for a large earthquake. It has been reported that finite fault effects (rupture propagation, directivity and source-receiver geometry) strongly influence ground motion characteristics (amplitudes, frequency content and duration of motion) in both near-source and far-field regions for large events (Hartzell, 1978; Somerville et al., 1991; Atkinson and Silva, 1997). Several researchers have worked on this problem (finite fault effects) by extending stochastic modeling to consider rupture along a finite fault plane (Schneider et al., 1993; Beresnev and Atkinson, 1997; Motazedian and Atkinson, 2005). The common approach to modeling these effects has been to subdivide the fault into smaller parts, each of which is considered as a point source. Then the ground motions at an observation point are obtained by summing the contributions from all sub-faults with their respective delay times. I use the stochastic finite-fault modeling approach of Motazedian and Atkinson (2005), as implemented in the computer code EXSIM, for the analysis and simulation of

5% damped spectral ordinates (PSA) of the September 26, 2003, **M** 8.1 Tokachi-Oki earthquake and some of its aftershocks. The method models ground motions as a propagating array of Brune point sources (Brune, 1970, 1971), each of which can be simulated using the stochastic point source methodology of Boore (1983, 2003).

4.1 Stochastic finite-fault modeling technique

The stochastic finite-fault modeling technique considers each sub-fault as a point seismic source and sums contributions to the total ground-motion at a site over all sub-sources. Boore and Atkinson (1987) describe the (historical) development of ground-motion stochastic simulation methods, explaining that two approaches have been considered: time domain simulation and random-process theory methods. They emphasize that in both methods, ground-motion is modeled as band-limited finite duration white Gaussian noise in which the radiated energy is distributed over a specific duration. In both cases, specification of a spectral shape, which represents the radiation from the source, is required. This describes the sub-source spectrum as a function of magnitude. Also needed are frequency dependent functions to specify the spectral energy at a given site, that is: attenuation, site response and / or instrument response functions.

The general form of the stochastic model to estimate Fourier spectra of ground-motion at a site is (Boore and Atkinson, 1987):

$$A(f) = C \cdot S(f) \cdot D(f) \cdot I(f), \quad (4.1)$$

where C is a scaling factor, $S(f)$ is the source spectrum, $D(f)$ is the diminution function (path and instruments effects) and $I(f)$ is the type of motion to be estimated (acceleration,

velocity, etc). If body-wave spreading is assumed in a whole space, the scaling factor for a point source is given by:

$$C = \frac{(R_{\theta\phi})FV}{4\pi\rho\beta^3}, \quad (4.2)$$

where $R_{\theta\phi}$ is the radiation pattern, F is a factor for free surface effects, V is the partition into horizontal components, and ρ and β are density and shear wave velocity in the source region. The source function $S(f)$ is modeled as the product of a deterministic seismological source function multiplied by the Fourier spectrum of windowed Gaussian noise. For point sources $S(f)$ is defined as:

$$S(f) = M_0 / [1 + (f / f_0)^2], \quad (4.3)$$

where M_0 is the seismic moment and f_0 is the corner frequency (Boore, 1983). Another similar form has been obtained to represent this source function defined in terms of two corner frequencies (Joyner, 1984). This source spectrum represents the generation of shear (S) waves. The function $D(f)$ is a product of both geometrical spreading and inelastic attenuation effects, and a high cut filter. Its expression is as follows:

$$D(f) = \exp\left[-\frac{\pi R f}{Q(f)\beta}\right] [1/R] \cdot P(f, f_m), \quad (4.4)$$

where Q is the quality factor as a function of f , R is hypocentral distance and P is the high cut filter. There are two common forms for this filter, the Butterworth filter: $P(f) = [1 + (f/f_m)^8]^{-1/2}$ (Boore, 1983), and the exponential form: $P(f) = \exp(-\pi \kappa f)$. This filter is part of the stochastic model to account for rapid spectral decay above a certain frequency f_m (Beresnev and Atkinson, 1997). The last factor in the stochastic model 4.1, the filter

$I(f)$, shapes the spectrum in order to get the particular ground-motion of interest. It has a simple form: $I(f) = (2\pi f)^p$, with $p = 1$ for velocity and $p = 2$ for acceleration.

The above equations describe a point source simulation. As pointed out by Motazedian and Atkinson (2005), when large earthquakes and / or short distances relative to fault dimensions are analyzed, the stochastic finite-fault model is a more suitable approach. In this case the entire fault is divided into N sub-faults. The ground-motion acceleration at a site is obtained as the summation of sub-source contributions:

$$a(t) = \sum_{i=1}^{nl} \sum_{j=1}^{nw} a_{ij}(t + \Delta t_{ij}), \quad (4.5)$$

where nl and nw are the number of sub-sources (cells) along the length and width respectively ($nl \times nw = N$), Δt_{ij} is the delay time from ij^{th} cell to the site. Then, for each cell the acceleration spectrum is modeled as a point source with a ω^2 shape as:

$$A_{ij}(f) = \left\{ CM_{0ij} (2\pi f)^2 / [1 + (f / f_{0ij})^2] \right\} \exp(-\pi f \kappa) \exp(-\pi f R_{ij} / Q(f) \beta) (1 / R_{ij}), \quad (4.6)$$

where M_{0ij} , f_{0ij} and R_{ij} are for each cell: seismic moment, corner frequency and distance from the observation point respectively. The factor C is as given in 4.2. The seismic moment for a cell is obtained by considering the ratio of its area to the area of the main fault:

$$M_{0ij} = M_0 S_{ij} / \left(\sum_{l=1}^{nl} \sum_{k=1}^{mw} S_{kl} \right), \quad (4.7)$$

where S_{ij} is the relative slip weight of the ij^{th} sub-fault. The previous equations describe a kinematic model of earthquake ground-motion process. Refinements to the model to account for a wide range of factors, such as the shift of frequency content with time as the

rupture propagates, and self-healing slip pulses, are described by Motazedian and Atkinson (2005), and incorporated in the computer program EXSIM. These authors found that, for the kinematic model approach (static corner frequency), as the number of sub-faults is incremented, the received energy at low and high frequency is decreased and increased respectively, and the total radiated energy is not conserved. Therefore they proposed a “dynamic corner frequency” approach to finite-fault modeling to overcome this problem. In this approach the corner frequency varies with time, which means that it can also be expressed as function of ruptured area. The dynamic frequency starts from high corner frequencies at the beginning of rupture, progressing to lower corner frequencies, and ending with f_0 , the corner frequency for the entire fault. A better representation of the low frequency part of the signal in the simulated records may be significant in the analysis of large interface earthquakes.

There are several model parameters to define as input to EXSIM. The two main “free” parameters are stress drop and pulsing percentage area, which control the level of spectra at high frequencies and low frequencies, respectively.

4.2 Parameters of the model

The parameters required for the stochastic finite-fault method describe the source, path and site contributions to the simulated ground-motions; in this case the attenuation and amplification models determined in the empirical analysis (section 3.2), which are mostly based on the M 8.1 data, are used for the simulations. For the selected aftershocks, I make minor modifications to the source parameters (i.e.: the stress drop) of

each event, while holding site and path parameters constant for all events. The analysis for the selected aftershocks adds robustness to the model, providing insight into event-to-event variability in source and path effects.

Source: I follow Yagi (2004) to define most of the source parameters for the main shock, i.e.: fault geometry (length, wide, strike and dip angles of the fault plane) and slip distribution. For the aftershocks, the geometry is not known. Therefore, I used the Harvard moment tensor catalogue (HRV) at www.neictest.cr.usgs.gov/neis/sopar to define seismic moment for each aftershock, and based its dimensions on the empirical relationships of Wells and Coppersmith (1994). Table 4.1 contains the source parameter values used in the simulations, along with the adopted path and site parameters (described next).

Path: As it was mentioned above, I used the attenuation parameters obtained in section 3.2 for all simulations in this section; however, for the duration of ground-motion, represented as: $T = T_0 + d \cdot (R)$, I use the slope value $d = 0.09$ (found previously) for the **M** 8.1 case; this value may be a slight overestimate of the distance-dependent duration effect from the point of view of simulations, as it is based on a long estimate of duration, out to 90% of signal energy. For the main shock this effect is not particularly important, as the source duration dominates the signal duration. For the aftershocks I assigned a slightly smaller value of $d = 0.07$, as being more typical for most events (Atkinson, 1995; Beresnev and Atkinson, 1997). The source duration, T_0 , depends on fault dimension and therefore magnitude.

Site: The site amplification functions for each NEHRP site class, described in subchapter 3.2, were used for the simulations. The site reference function (NEHRP C) was calculated for an angle of incidence $i = 45.0^\circ$ and a kappa factor $\kappa_0 = 0.045$. The choice of $i = 45.0^\circ$, constrained in some extent for the observed H/V spectral ratio function for C sites, was intended to reflect the shallow depth of the main shock and the large area coverage of data. This function is plotted in Figure 4.1; I also show in this figure the effect of changing i from 45.0° to 0.0° (vertical incidence) and κ_0 from 0.045 to 0.02 in the NEHRP C amplification function, the corresponding sets of amplification factors are included in Table 4.2. This illustrates that the reference amplification function is actually quite sensitive to these choices, especially at high frequencies. This is a factor that should be kept in mind in interpreting source parameters; for example, the stress drop will trade off against the reference amplification, with a higher site amplification implying a lower stress drop, as shown in the next section.

4.3 PSA simulation results for the main shock and selected aftershocks

The modeling of the Tokachi-Oki main shock and its aftershocks was done using the source, attenuation, and site parameters described in Table 4.1. For each event, I adjust the stress drop parameter to obtain the best match of simulations to data. This match is defined in terms of residuals, defined as the difference between the logarithmic values (base 10) of the observed and the simulated PSA ordinates:

$$res(f, R_{cd}) = \log_{10}(\text{PSA}_{obs}) - \log_{10}(\text{PSA}_{sim}) \quad (4.8)$$

The best value of stress drop ($\Delta\sigma$) is that which minimizes the average residual, **stat1**, defined by:

$$\mathbf{stat1} = \frac{1}{M} \sum_{n=1}^N \sum_{m=1}^M |res_{nm}(f, R_{cd})| \quad (4.9)$$

while the variability of the residuals is expressed by its standard deviation, **stat2**:

$$\mathbf{stat2} = \left(\frac{1}{N} \sum_{n=1}^N \text{var}_n[res_n(f, R_{cd})] \right)^{1/2} \quad (4.10)$$

where N = number of frequencies, M = number of distances and var_n is the variance calculated on a residual vector distributed in distance for the n^{th} frequency.

It is important to clarify that residuals were calculated at the NEHRP C reference level. The observed PSA ordinates on D and E sites were first corrected to this reference level using the amplification functions for D and E sites calculated in section 3.2.3. Then, the corrected data were compared to the model defined for NEHRP C to calculate residuals.

I determined, for each event, the $\Delta\sigma$ value that minimizes the average residual **stat1**. Figure 4.2 shows how the average residuals are distributed in frequency for the M 8.1 event. To illustrate the effect of the amplification factor on the inferred solution, two cases are presented: i) the reference NEHRP C amplification function is based on $\kappa_0 = 0.045$ and $i = 45.0^\circ$; and ii) the reference C amplification is based on the values calculated with $\kappa_0 = 0.02$ and $i = 0.0^\circ$. The inferred stress drop changes only slightly based on the assumed NEHRP C amplification (120 bars vs. 110 bars), but it is remarkable that the latter amplification function provides a better distribution of residuals

across frequency, by accounting for apparent high-frequency energy in the PSA data. The strong signal at high frequencies is consistent with overall site conditions in Japan, which typically feature a shallow soil layer over rock.

Figure 4.3 plots residuals against distance, for two selected frequencies, for the interface events (**M** 8.1 and the **M** 6.4 aftershock) and for two aftershocks (**M** 5.5, **M** 5.9) believed to be in-slab events (Ito et al, 2004). There are no apparent trends of residuals with distance, indicating a good fit of the attenuation model to the data. Interestingly, the best-fit value of $\Delta\sigma$ appears to be about 100 bars for the interface events, compared to values near 200 bars for the in-slab events. These stress drop values are in good agreement with the results reported in section 3.2.5 for these events based on near-source Fourier spectra, as would be expected. For the **M** 8.1 earthquake, the inferred stress drop agrees with the value reported by Yagi (2004), which is 120 bars.

Table 4.1, Source, path and site parameters used in simulations of PSA for the 2003, Tokachi-Oki main shock (**M** 8.1) and some of its aftershocks (**M**: 7.3, 6.4, 5.9, 5.5).

	Parameter	M 8.1	M 7.3	M 6.4	M 5.9	M 5.5
S o u r c e	orientation (deg) [strike, dip]	[250, 17]	[208, 18]	[244, 17]	[227, 28]	[148, 48]
	dimensions (km) [strike, dip]	[132 x 168]	[60 x 24]	[19 x 10]	[10 x 6]	[6 x 4]
	depth range (km)	[6 – 56]	[43 – 51]	[34 – 37]	[53 – 56]	[47 – 50]
	Location (top) (deg) [lat., lon.]	[145.0, 41.2]	[143.8, 41.9]	[144.7, 42.3]	[145.2, 42.1]	[144.0, 42.5]
	number of sub- sources [strike, dip]	[24, 21]	[10, 4]	[3, 2]	[2, 1]	[1, 1]
	pulsing area (%)	50	✓	✓	✓	✓
	slip distribution	Yagi (2004)	random	random	random	random
P a t h	Inelastic attenuation $Q = Q_0 f^\eta$	$135 f^{0.76}$	✓	✓	✓	✓
	geometric spreading	$1 / R$	✓	✓	✓	✓
	Distance duration term [d] (s/km)	0.09	0.07	0.07	0.07	0.07
S i t e	κ factor	0.045	✓	✓	✓	✓
	amplification factors [reference level]	NEHRP C; $i = 45.0^\circ$ [see Table 3.2]	✓	✓	✓	✓
	shear wave velocity [crustal] (km/s)	3.6	✓	✓	✓	✓
	density (g/cm ³) [crustal]	2.8	✓	✓	✓	✓
	damping factor (%) [PSA]	5	✓	✓	✓	✓

Table 4.2, Amplification functions and average shear wave velocity and density profile for NEHRP C sites of K-NET stations. Two sets of amplification factors (for different i and κ values) are included.

Shear velocity (β) and density (ρ) profile			Amplification functions		
Thickness [m]	β [m / s]	ρ [g/cm ³]	Frequency [Hz]	$\kappa= 0.045$, $i = 45.0^\circ$	$\kappa= 0.02$, $i = 0.0^\circ$
1	144.0	1.54	0.10	1.27	1.45
1	178.7	1.55	0.12	1.31	1.52
1	245.2	1.61	0.15	1.36	1.60
1	333.9	1.68	0.18	1.39	1.69
1	367.0	1.75	0.22	1.44	1.78
1	383.0	1.77	0.27	1.49	1.86
1	437.8	1.78	0.33	1.53	1.93
1	460.4	1.79	0.41	1.58	2.01
1	477.8	1.81	0.50	1.60	2.07
1	484.3	1.83	0.61	1.61	2.13
1	487.8	1.84	0.74	1.64	2.19
1	511.5	1.84	0.90	1.66	2.22
1	512.0	1.83	1.11	1.70	2.25
1	533.0	1.85	1.35	1.72	2.29
1	534.1	1.84	1.65	1.73	2.32
1	569.8	1.86	2.02	1.74	2.37
1	580.6	1.88	2.46	1.72	2.44
1	611.9	1.89	3.01	1.68	2.51
1	612.0	1.90	3.67	1.64	2.57
1	612.0	1.93	4.49	1.59	2.61
130	943.3	2.08	5.48	1.48	2.65
250	1040.6	2.19	6.70	1.34	2.64
341	1242.0	2.15	8.19	1.19	2.60
3259	2190.0	2.30	10.0	1.03	2.51
4000	3180.0	2.60			
10000	3600.0	2.80

Figure 4.1 - Amplification functions for NEHRP C sites of the Hokkaido region, NE Japan. All these functions were calculated from a common shear wave and density profile (Table 4.2) but for different angles of incidence and κ factor values (as shown in the legend). Black solid line is the function obtained in section 3.2.3, dark gray dashed line shows the effect of changing the angle of incidence and light gray dashed line shows the combined effect of changes in i and κ simultaneously.

Figure 4.2 - Average residuals as a function of frequency, in log units, of the observed minus simulated PSA values for the M 8.1 Tokachi-Oki earthquake. They correspond to (a) $\Delta\sigma = 120$ bars, $i = 45.0^\circ$, $\kappa = 0.045$ and (b) $\Delta\sigma = 110$ bars, $i = 0.0^\circ$, $\kappa = 0.02$. Light grey bars are $\pm$ one standard deviation.

Figure 4.3 - Residuals of the logarithm of PSA (observed – estimated) as a function of distance (R_{cd}) for four magnitudes, at low and high frequencies. Larger symbols represent the geometric mean (log units) of the residuals contained in a specific distance bin. The distance bins were logarithmically spaced. The stress drop value that minimized residuals, for each case, is indicated in the dark gray box.

Chapter 5. Implications for expected ground-motion from large interface earthquakes in the Cascadia subduction zone

In Chapter 4 it is shown that the stochastic finite-fault model of Motazedian and Atkinson (2005) reasonably reproduces ground-motion data from the **M** 8.1 Tokachi-Oki earthquake sequence. In this chapter I use this calibrated model to simulate ground-motion for scenario earthquakes of **M** 7.5 to 9.0 on the Cascadia subduction interface and to estimate average PSA for firm site conditions near the cities of Victoria (NEHRP B/C boundary), Vancouver (top of the Pleistocene layer, with $V_{s30} = 440$ m/s) and Seattle (B/C boundary). Thus, adjustments to the calibrated model are required to consider average source, attenuation and site parameters for the Cascadia region. This includes an evaluation of the likely variability in stress drop for large interface earthquakes, based on evaluation of data from other regions of the world in comparison to the Tokachi-Oki model predictions, and an assessment of regional attenuation and site effects. In the last part of this chapter I analyze simulated response spectra values versus distance to generalize these results as ground-motion prediction equations for interface earthquakes at the Cascadia subduction zone.

5.1 Estimation of stress drop variability for large interface earthquakes

The stochastic finite-fault modeling technique accurately reproduces, on average, observed ground-motion characteristics (amplitudes and frequency content) from large interface earthquakes, once the regional source, path and site characteristics are known. For regions where there are no data from large interface earthquakes, the attenuation and

site models can be established from data of smaller magnitudes. However, information on source parameters for large events may be lacking. For example, for the Cascadia region there is empirical information on attenuation and site effects (Atkinson, 2005), but there is no information on the stress drop that might be expected for a great interface event. To place some constraints on this key uncertainty, stress drop variability amongst large events on subduction interfaces worldwide is estimated.

The approach is to perform simple comparisons between observed and simulated PSA values for selected interface earthquakes. I focus on model vs. data comparisons of spectral amplitudes at distances of about 60 (± 20) km; in this distance range the observed motions are reduced by the geometric spreading attenuation factor (Crouse et al, 1988), but the inelastic attenuation factor (which may vary regionally) has a minor effect. The base simulation model for the comparisons is the Tokachi-Oki model as defined in Chapter 4 (TK07). Predictions for this model are compared to PSA data from selected interface earthquakes of different subduction zones, having multiple recordings within about 60 (± 20) km of the fault, as compiled by Atkinson and Boore, 2003 (AB03). Table 5.1 lists the selected earthquakes, along with moment magnitude, hypocentral and focal mechanism information from the Harvard Central Moment Tensor catalogue.

For the distance range of $R_{cd} < 100$ km, in which the comparisons are focused, it is assumed that the geometric spreading coefficient is $b_I = -1.0$ for all subduction regions, and that regional differences in inelastic attenuation behaviour are unimportant (Crouse et al, 1988; Youngs et al, 1997; Atkinson, 2005). I make a preliminary estimation of likely regional crustal amplification effects that would apply (as described below). By

comparing simulated amplitudes from the Tokachi-Oki model (adjusted for regional crustal amplification) with various $\Delta\sigma$ values to observed amplitudes, I make an estimate of the $\Delta\sigma$ value for each the selected earthquakes. It is acknowledged that these are rough estimates, as I make no attempt to model attenuation and site processes in detail for each event. The aim is simply to gain insight on the likely variability of $\Delta\sigma$ for large interface events.

The reference site condition for the comparisons is the NEHRP B/C boundary ($V_{S30} = 760$ m/s). The AB03 database provides site classification (from NEHRP A to NEHRP E classes) for each record. I assigned fixed representative values for each class: $V_A = 2000$ m/s; $V_B = 1050$ m/s; $V_C = 560$ m/s; $V_D = 270$ m/s; $V_E = 100$ m/s. The amplification factors by which to correct each record to the corresponding PSA values for B/C site conditions were obtained by using the empirical site amplification relationships given by Boore and Atkinson (2006); for consistency and simplicity, the expected peak ground acceleration for B/C conditions ($PGA_{B/C}$) to input to these relationships was calculated using their prediction equation for $PGA_{B/C}$. It is emphasized that the required value for $PGA_{B/C}$ is only a preliminary estimate, as the amplification factors are not particularly sensitive to this value. To produce simulations for B/C boundary conditions (for Tokachi-Oki), I used the site information provided by K-NET, selecting stations that lie between B and C (stations IWT from 008 to 010 and MYG from 001 to 004) to define a generic shear-wave velocity and density profile for B/C conditions. The amplification function for this profile was calculated using Boore's (2003) SITE AMP program, as

shown in Table 5.2. I assumed vertical incidence and $\kappa_0 = 0.02$ for the calculations (as kappa is expected to be lower for B/C sites than for softer site conditions).

The amplification function for B/C sites may be expected to show significant regional variability due to typical crustal conditions, especially at high frequencies (Crouse et al, 1988; Atkinson and Boore, 2003; Atkinson and Casey, 2003). Consequently, in the comparisons, both the stress drop of the event and the regional crustal amplification may influence the observed spectral ordinates at high frequencies. To account for this effect, we calculated an expected B/C amplification function for each region. I compile, from different sources of information, shear velocity and density crustal models (Alaska: Niazi and Chun, 1989; Brocher et al., 2004. Chile: Mendoza et al., 1994. Japan: Iwasaki et al., 1989; Nishizawa and Suyehiro, 1986. Mexico: Valdes et al., 1986; Furumura and Singh, 2002; Dominguez et al., 2006), and define an average model for each region. These models define the conditions of the crust in each region, but typical conditions in the upper 30 m are not known. I therefore adopt the top 30 m B/C shear velocity profile defined for Japan (Table 5.2) as a generic top 30 m B/C profile for all regions. Figure 5.1.1 shows the B/C amplification functions for each region under this assumption, calculated for $\kappa_0 = 0.02$ and $i = 0.0^\circ$. Tables 5.3 and 5.4 contain the corresponding shear wave velocity and density models and amplification factors respectively. Figure 5.1.1 suggests remarkable differences in expected amplifications at frequencies below 1 Hz between regions. However, at high frequencies the amplifications are similar, which is not surprising given the common site profile for the top 30 m. This commonality of behavior at high frequencies means that the calculated

stress drops will be insensitive to the regional amplification function, as it is the high frequencies that control stress drop. This is an aspect of the study that could be improved in the future if more information on typical site conditions in each region and their effect on high frequency amplitudes become available.

In Figures 5.1.2, 5.1.3, and 5.1.4, I show (for B/C boundary site conditions) the simulated versus observed PSA at 10 Hz, for the selected earthquakes. Based on inspection of these plots, I made a rough estimate of stress drop for each event, as listed in Table 5.5. The plots show some interesting differences and agreements in attenuation behavior between the TK07 model and data from various regions. I minimized the effect of these differences by focusing on the comparisons at distances < 100 km to draw conclusions regarding stress drop. No fitting of the data to the model calculations is performed (this is not warranted given the lack of region-specific information on attenuation and site conditions); the stress drop estimates are a judgment based on inspection.

Based on these results, and considering that unaccounted-for site effects or erroneous information on site classification may be present in the data, I adopted as the lower limit of $\Delta\sigma$ for interface earthquakes a value of **30 bars**. This value is supported by the stable estimates of stress drop from the Alaska region that suggest this value, and by the lower limits for some of the Mexican earthquakes. For the upper limit on stress drop, I set a value of **150 bars**. The selection of this value was not as clear as the lower limit; there may be some higher values, such as the inferred values of up to 170 bars for the Chilean and Japanese earthquakes. On the other hand, there are some events for

which a value of 120 bars appears relatively well defined (including the Tokachi-Oki main shock). On balance, I selected 150 bars as an upper limit of $\Delta\sigma$ for interface earthquakes. This is not intended to be an absolute upper limit, nor is 30 bars intended to be an absolute lower limit. Rather these represent what may be considered as values that are perhaps one to two standard deviations from a median value. A more precise representation of stress drop variability must await more detailed studies of this parameter for large interface events.

Table 5.1 – Selected earthquakes for stress drop comparisons. Source parameters from HRV catalogue. Codes for regions are as in Atkinson and Boore (2003).

Region	M	Date	Lat. [deg]	Lon. [deg]	Depth [km]	Strike-Dip [deg]	
Alaska	6.0	1983 02 14	54.51	-158.98	39.8	256	25
(AL)	6.6	1985 10 09	54.84	-159.40	31.8	246	16
Chile	8.0	1985 03 03	-33.92	-071.71	40.7	11	26
(CC)	7.1	1985 04 09	-34.26	-071.86	46.6	0	21
Japan	7.6	1978 06 12	38.02	142.07	37.7	184	14
(JA)	7.0	1982 07 23	35.98	141.91	27.0	203	14
	7.7	1983 05 26	40.44	138.87	12.6	16	27
Japan	5.8	1996 06 02	27.27	128.57	42.0	228	29
(JK)	5.9	1997 05 11	37.09	140.91	57.9	160	47
	6.1	1997 07 14	43.19	146.47	34.0	236	28
Mexico	8.0	1985 09 19	17.91	-101.99	21.3	301	18
(ME)	7.6	1985 09 21	17.57	-101.42	20.8	296	17
	7.0	1986 04 30	18.25	-102.92	20.7	290	18
	7.4	1995 09 14	16.73	-098.54	21.8	289	15
	5.6	1996 04 23	17.13	-101.84	36.8	121	59
	6.6	1996 07 15	17.50	-101.12	22.4	297	21

Table 5.2 - Amplification function and average shear wave velocity and density profile for NEHRP B/C boundary site conditions of K-NET stations.

Shear velocity (β) and density (ρ) profile			Amplification function	
Thickness [m]	β [m / s]	ρ [g/cm ³]	Frequency [Hz]	$\kappa=0.02, i=0^\circ$
1	207.1	1.77	0.10	1.32
1	334.3	1.87	0.12	1.35
1	367.1	1.91	0.15	1.38
1	367.1	1.87	0.18	1.41
1	501.4	1.89	0.22	1.43
1	501.4	1.96	0.27	1.45
1	712.9	2.01	0.33	1.46
1	770.0	2.05	0.41	1.47
1	770.0	2.08	0.50	1.48
1	770.0	2.11	0.61	1.50
5	1000.0	2.12	0.74	1.53
15	1250.0	2.15	0.90	1.57
120	1300.0	2.20	1.11	1.61
3850	2190.0	2.30	1.35	1.64
4000	3180.0	2.60	1.65	1.67
10000	3340.0	2.70	2.02	1.70
10000	3740.0	2.90	2.46	1.71
5000	4500.0	3.20	3.01	1.73
			3.67	1.73
			4.49	1.73
			5.48	1.72
			6.70	1.69
			8.19	1.66
			10.0	1.59

Table 5.4 - Frequency dependent amplification factors for the models given in Table 5.3

Amplification functions; NEHRP B/C (Regions)				
$i = 0.0^\circ$; $\kappa=0.02$				
Frequency [Hz]	AL	CC	ME	JA
0.10	1.08	1.27	1.09	1.32
0.12	1.09	1.30	1.10	1.35
0.15	1.11	1.33	1.13	1.38
0.18	1.14	1.37	1.15	1.41
0.22	1.17	1.40	1.19	1.43
0.27	1.20	1.44	1.23	1.45
0.33	1.23	1.46	1.27	1.46
0.41	1.26	1.49	1.32	1.47
0.5	1.30	1.51	1.37	1.48
0.61	1.34	1.54	1.41	1.50
0.74	1.39	1.57	1.46	1.53
0.90	1.46	1.60	1.50	1.57
1.11	1.52	1.63	1.54	1.61
1.35	1.58	1.66	1.58	1.64
1.65	1.63	1.69	1.61	1.67
2.02	1.68	1.72	1.65	1.70
2.46	1.71	1.75	1.68	1.71
3.01	1.74	1.77	1.71	1.73
3.67	1.76	1.78	1.74	1.73
4.49	1.76	1.79	1.77	1.73
5.48	1.75	1.79	1.79	1.72
6.70	1.72	1.76	1.78	1.69
8.19	1.68	1.72	1.75	1.66
10.0	1.62	1.65	1.68	1.59

Table 5.5 – Best estimate of stress drop values for each of the selected earthquakes; the range of stress drop values that is judged to be permitted by the data variability for each case is also included.

Region	M	$\Delta\sigma$ (bars)	$\Delta\sigma$ (bars)
		[average]	[range]
Alaska (AL)	6.0	30	20 – 40
	6.6	40	30 – 50
	Region	35	20 – 50
Chile (CC)	8.0	100	30 – 170
	7.1	120	70 – 170
	Region	110	30 – 170
Japan (JA)	7.6	170 [n/a]	160 – 180
	7.9	150 [n/a]	130 – 170
Japan (JK)	5.7	170 (?)	170 - ?
	6.3	90	60 – 120
	5.9	170 (?)	70 - ?
	6.1	30	10 – 50
	8.0	90 [n/a]	60 – 120
Mexico (ME)	7.6	70	30 – 120
	7.0	40	10 – 70
	7.4	40	10 – 70
	6.6	60	40 – 80
	Region	65	10 – 120

Figure 5.1.1 - Amplification functions calculated for NEHRP B/C boundary site conditions, $\kappa = 0.020$, $i = 0.0^\circ$, for four different shear wave velocity and density crustal models, each one related to a different subduction zone (code for regions are as in Table 5.1).

Figure 5.1.2 - PSA for interface earthquakes as a function of closest distance to fault (R_{cd}) for 10 Hz. Light gray open circles represent the data (Atkinson and Boore, 2003) corrected to NEHRP B/C boundary site condition. Lines show fits to simulated PSA ordinates for four $\Delta\sigma$ values. A separate frame is plotted for each earthquake: M 6.0, 6.6 in Alaska (AL) and M 7.1 and 8.0 in Chile (CC).

Figure 5.1.3 - PSA for interface earthquakes as a function of closest distance to fault (R_{cd}) for 10 Hz. Light gray open circles represent the data (Atkinson and Boore, 2003) corrected to NEHRP B/C boundary site condition. Lines show fits to simulated PSA ordinates for four $\Delta\sigma$ values. A separate frame is plotted for each earthquake: M 7.6, 7.9, 5.7, 6.3, 5.9, and 6.1 in Japan (JA, Japanese Meteorological Agency and JK, K-NET).

Figure 5.1.4 - PSA for interface earthquakes as a function of closest distance to fault (R_{cd}) for 10 Hz. Light gray open circles represent the data (Atkinson and Boore, 2003) corrected to NEHRP B/C boundary site condition. Lines show fits to simulated PSA ordinates for four $\Delta\sigma$ values. A separate frame is plotted for each earthquake: M 6.6, 7.0, 7.4, 7.6, and 8.0 in Mexico (ME).

5.2 Ground-motion prediction (PSA) for the Cascadia subduction zone

The calibrated EXSIM model is used to simulate ground-motion for generic sites in Southwestern British Columbia and Northwestern Washington. The simulation scenarios are constrained by the geometry of the anticipated rupture area (Hyndman and Wang, 1995; Flück et al, 1997), the average stress drop for subduction zone earthquakes and its variability (as estimated in the previous section), and by regional estimates of attenuation and site conditions. I first generate simulations for “best estimates” of these parameters, and then I explore the implications of alternative values for the main uncertainties in source and attenuation parameters.

5.2.1 Parameters of the model

Source: For the seismic source geometries I follow Hyndman and Wang (1995), who define the landward limit of the rupture surface based on the geometry and temperature gradient of the slab. Response spectral ordinates (PSA) were generated for the average stress value $\Delta\sigma = 90$ bar, for moment magnitudes M 7.5, 8.0, 8.5, and 9.0, for three specific reference site locations and reference conditions. The locations, and their corresponding generic site conditions, are chosen to correspond to firm sites in the Fraser Delta region of Vancouver (FRA), Victoria (VIC), and Seattle (SEA). To simulate the motions at each location I locate the fault plane symmetrically about a perpendicular line from the trench to the site. Fault plane areas for each magnitude were defined considering the relationships used by Wells and Coppersmith (1994), Kanamori and Anderson (1975), and Beresnev and Atkinson (1997). Once an average area was defined,

the fault length for the M 8.0 and M 8.5 cases was assigned a 90 km fixed fault width (this is the maximum according to Hyndman and Wang). The M 7.5 event was assigned a narrower width, based on Wells and Coppersmith (1994). For the M 9.0 Cascadia scenario (Satake et al, 1996), we considered both the 90 km width, and a wider rupture zone of 150 km, with the fault length adjusted accordingly. Figures 5.2.1 and 5.2.2 show the fault planes that correspond to the FRA, VIC and SEA simulations respectively, while Table 5.6 lists the simulation geometries and parameters. All simulations assume random slip distribution and random hypocenter location on the fault plane.

Path: By using empirical data from earthquakes of small to moderate magnitude, Atkinson (2005) investigated the attenuation of ground-motion in the Cascadia subduction zone. Her findings suggest that the empirical attenuation model used for California by Atkinson and Silva (2000) (AS00) may be appropriate to express attenuation for offshore events in the area of the subduction zone. I adopted the AS00 attenuation model as the “best estimate” attenuation model for the region. The distance-dependent duration of motion term is taken as 0.10 based on typical values shown by Raoof et al. (1999) (from which the AS00 model was derived); this factor is not important as source duration dominates the total duration.

Site: I calculated generic site amplification factors for reference “firm” sites as a function of frequency for Vancouver’s Fraser delta, Victoria and Seattle. Amplifications are constructed separately for each location, as studies suggest there are significant differences in shallow crustal structure. For example, a thinner layer of accreted sediments lies beneath Victoria in comparison to that beneath the Fraser delta or Seattle

(Ramachandran et al, 2006; Graindorge et al, 2003; Ellis et al, 1983; McMechan and Spence, 1983). In the Fraser delta, there is a pervasive layer of Pleistocene deposits that overlies the tertiary bedrock; the Pleistocene layer has a stable shear velocity gradient from 400 to 1000 m/s, while the tertiary bedrock has an average $V_s = 1500$ m/s (Hunter, 1995; Hunter et al., 1997). For the Fraser delta, the top of the Pleistocene layer was selected as the reference site condition ($V_{S30}=414$ m/s), while for Victoria and Seattle I define a generic NEHRP B/C boundary profile for the top 30 m ($V_{S30}=760$ m/s; this is harder than the conditions at the top of the Pleistocene layer that underlies the Fraser delta). For each generic profile, the quarter wavelength approach with an angle of incidence $i = 0.0^\circ$ was used to compute the amplifications, assuming $\kappa_0 = 0.02$ for NEHRP B/C boundary site conditions (VIC and SEA) and $\kappa_0 = 0.03$ for FRA conditions (Pleistocene). Figure 5.2.3 plots the shear wave velocity profiles versus depth, while Table 5.7 provides the profiles and amplification functions for each location. Figure 5.2.4 shows the amplification functions given in Table 5.7.

Each of the reference site conditions is, to a large extent, arbitrary, and does not necessarily represent a specific site in any of the three studied areas. The generic profiles are intended to reflect shallow crustal properties and average local site conditions for the firmest sites that may be available - so that these motions may in turn be input to overlying soil deposits where applicable. For example, in Seattle several studies demonstrate the presence of soft and thick sediment layers which form the sedimentary basin structure on which Seattle is located (Frankel et al., 1999 and 2002; Jones, 1999). However, stiffer site conditions are found to the West and Southeast of Seattle, for

example, at Tertiary sandstones at Alki Point and Seward Park, or recent glacial tills at Central Seattle and the Space Needle (Williams, 1999). These firmer sites were used to define the generic NEHRP B/C boundary site profile for Seattle. For Victoria, it is also possible to find B/C site conditions; according to Finn et al. (2004), nearly 50 % of shallow soil sites in Victoria would be classed between NEHRP A and C. Finally, for Vancouver, even though most of the Fraser delta has been classified as NEHRP D or E, there exist NEHRP C sites with velocities close to our adopted Pleistocene reference value of $V_{S30} = 414$ m/s in the Surrey Uplands, and within the City of Vancouver to the East and North of the Fraser Delta (Hunter et al., 2002). Thus I believe that the adopted reference models represent realizable “firm” reference site conditions in each of the studied areas.

5.2.2 “Best estimate” simulation results

Figure 5.2.5 plots average 5% damped pseudo-acceleration (PSA) and its standard deviation for the FRA, VIC and SEA reference site conditions for the “best estimate” model described above ($\Delta\sigma = 90$ bars, AS00 attenuation), for a scenario event of **M 9**; these sites are at distances of 170, 112 and 130 km, respectively, from the fault plane. The mean PSA and its standard deviation are determined from 100 simulations with random slip distribution and random hypocenter location; they correspond to a random horizontal component. The random variability in simulated PSA values is around 20 % to 30 % of the average PSA values over all frequencies. Note that this represents only variability in these random simulation parameters, and is not a full measure of aleatory

variability (nor epistemic) in expected ground-motions. It is plotted simply to show the effect of these random simulation parameters on results. The variability for magnitudes **M**: 7.5, 8.0, and 8.5 is similar and shown in Figures 5.2.6, 5.2.7, and 5.2.8 respectively. Figure 5.2.9 shows the scaling of average response spectra with magnitude.

5.2.3 Effects of parameters uncertainties

One of the key uncertainties in the source parameters that control ground-motion amplitudes is the sub-event stress drop parameter. The “best estimate” response spectra assumed a value of $\Delta\sigma = 90$ bars. On Figure 5.2.10, I show the effect of stress drop values of 30 and 150 bars (these are the estimates of the lower and upper limits on stress drop parameter based on interface events around the world) for a **M** 9 scenario. Figure 5.2.11, 5.2.12, and 5.2.13 show the same effect for **M**: 7.5, 8.0, and 8.5. These figures suggest that uncertainty in ground-motion spectral amplitudes for Cascadia events due to uncertainty in the appropriate stress drop parameter is about $\pm 50\%$.

Another important uncertainty in source characterization is the source geometry. This is particularly critical for the **M** 9 scenario due to its large extent. I therefore considered three possible geometries (Figure 5.2.2), including an elongated (1000 km x 90 km) fault geometry at two different orientations (geometries a, b) and a broader (600 km x 150 km) fault geometry (c). The two orientations are used for the long, skinny fault plane because EXSIM models the fault plane as a straight line in plan (future enhancements could improve this to allow the fault plane to be a curve in plan); thus the approximation of a curved fault trend by a straight line is an area of modeling

uncertainty. Figure 5.2.14 shows, for the three localities, the effect of these choices on the average PSA for simulated **M** 9 events (the PSA for **M** 8.5 is also plotted for reference). It is interesting to note that the elongated geometries (a and b in Figure 5.2.2) implied by a narrow 90 km fault rupture width generate lesser ground-motions than would a wider fault zone; this is because the longer rupture results in much of the ground-motion being generated at larger distances from the site. Additionally, beyond the fact that the elongated geometries cannot be justified from the models postulated for the possible rupture area in the region (Hyndman and Wang, 1995; Flück et al, 1997), if the **M** 9 event is truly a very long narrow rupture, then the motions it produces at seaboard cities may well be lower than would be produced by an event of **M** 8.5, especially at high frequencies. An exception to this observation could be, from Figure 5.2.14, the vertical elongated geometry (b) for Seattle, likely due to the relative position of Seattle respecting this “possible” geometry.

A key uncertainty in the path model is the regional attenuation model, which includes both the geometric and inelastic attenuation effects. For the best estimates I adopted the Atkinson and Silva (2000) (AS00) attenuation model, based on the results of Atkinson (2005) for Cascadia events. Figure 5.2.15 explores the influence of possible alternative models to AS00 ($Q = 180 f^{0.45}$ and $b_I = [-1.0; -0.5]$ for $[R \leq 40; R > 40 \text{ km}]$), including those of Ordaz and Singh (1992) for Mexico (OS92) ($Q = 273 f^{0.66}$ and $b_I = [-1.0; -0.5]$ for $[R \leq 100; R > 100 \text{ km}]$), and this study model (section 3.2) for NE Japan (TK07) ($Q = 135 f^{0.76}$ and $b_I = -1.0$ for $R \geq 40 \text{ km}$). (In these comparisons the source and site parameters are fixed at the best estimate values.) On Figure 5.2.15 I also plot the

PSA estimates obtained from the empirical ground-motion relationships for subduction zones reported by Atkinson and Boore (2003) (AB03). These authors, in a very recent report (Atkinson and Boore, 2008), provide a slight correction for their predictions of PSA at 2.5 Hz and 5.0 Hz, presumably due to an error in their database; I use this correction for the plots included in Figure 5.2.15. Atkinson and Boore (2003) also provide regional correction factors for their estimates (including Japan and Cascadia regions); I did not use these regional correction factors for the AB03 curves in Figure 5.2.15 so that they correspond to the general average PSA for interface earthquakes at subduction zones.

It is remarkable that the TK07 model represents a steeper attenuation compared to the other models; this is likely due to the strong attenuation observed on back-arc sites of the Hokkaido region (section 3.2). By comparison, the rest of the models basically reflect fore-arc attenuation conditions. There are significant differences in the attenuation effects at distances roughly from 85 km to 145 km (factor of 1.5-2.0) between the AS00 and OS92 models at low and high frequencies (though their predictions are similar at intermediate frequencies); this could well reflect actual attenuation differences between the central Mexico subduction zone and the Cascadia (SW Canada-NW United States) subduction zone. The AB03 results are a “compilation” from all zones, but may be biased by the predominance of records from Mexico for the large interface events. The underlying generic site effects for the AB03 database are unknown, but appear to be different from those inferred for Cascadia, based on differences in spectral shape. These results point to the importance of regional variations in attenuation and site effects in

controlling ground-motion amplitudes and spectral shapes, and suggest that the use of empirical subduction relations based on a mixture of data from around the world, though a traditional expediency, may not be a well-founded approach.

5.2.4 Attenuation of interface motion with distance

The evaluation of parameter uncertainties points to the potential importance of regional attenuation in controlling the expected amplitudes from Cascadia interface events. Figure 5.2.16 shows, for 0.5 Hz and 5.0 Hz (VIC B/C site conditions), the expected decay of amplitudes with distance from the trench, for profiles that run perpendicular to the trench in southwestern B.C., as shown in Figure 5.2.17 with black square symbols. All calculations use the “best estimate” AS00 attenuation model with $\Delta\sigma = 90$ bars; results from 30 simulations were averaged to establish the mean PSA. The effect of magnitude is clearly much more important at lower frequencies, which is to be expected. On Figure 5.2.18, I show the importance of the attenuation model used by plotting the attenuation along the profile for a single magnitude, but for different attenuation models (average of 10 simulations); the empirical attenuation predictions of Atkinson and Boore (2003) are also shown for comparison. The flat attenuation characteristics of the AB03 estimates relative to what is expected for NE Japan, and for Cascadia at lower frequencies, is apparent on this figure; it is also remarkable the similar shape, at high frequencies and for $R_{cd} \geq 100$ km, between AB03 estimates and the Mexican attenuation model. The flat AB03 attenuation shape is likely a consequence of the fact that they are data-driven empirical relationships, derived from a database for

which approximately 40% of the interface earthquakes occurred in the Mexican subduction zone. Studies have shown that earthquakes in the Mexican subduction zone exhibit a very slow ground-motion attenuation with distance; the suggested cause is a regional amplification effect that is predominant for low frequencies at large distances (200 km and above) (Cardenas and Chavez-Garcia, 2003; Singh et al., 1995). The Mexican attenuation model shown in Figure 5.2.18 (Ordaz and Singh, 1992) is steeper than the Cascadia/California attenuation model (Atkinson and Silva, 2000) from 50 to 100 km (as the transition from body-wave to surface-wave spreading rates occurs at greater distance in OS92 compared to AS00), but less steep at greater distances, for which the effects of higher Q in Mexico become more important.

5.2.5 Ground-motion prediction equations for NEHRP B/C site conditions

The response spectral values from the simulation results can be generalized for ease of use by fitting them to a ground-motion prediction equation (GMPE). A two-step linear regression procedure (Joyner and Boore, 1993) was applied to simulated PSA ordinates and PGA values at sites located along the profiles shown in Figure 5.2.17 for all magnitudes, to generate GMPEs for large interface earthquakes in the Cascadia region.

In the first step I determine the distance attenuation term plus a source term for each scenario (M 7.5, 8.0, 8.5, 9.0 (geometry c)) according to the form:

$$\log Y = C_i - C_1 \log R - C_2 R \quad (5.1)$$

where Y is the PSA value at a selected frequency (or PGA) and $R = (R_{cd}^2 + h^2)^{1/2}$. R_{cd} is the closest distance to fault, and h represents a near source saturation term determined

iteratively to provide the best fit. The second regression step determines the magnitude-dependent coefficients from the source terms C_i to:

$$C_i = C_0 + C_3(\mathbf{M} - 8) + C_4(\mathbf{M} - 8)^2 \quad (5.2)$$

Residuals between “observed” and estimated log PSA were assessed, for different h values and for each scenario, to explore the h dependency on magnitude. Figure 5.2.17 shows, along with the perpendicular profiles, fan shape array profiles that were used to simulate more PSA data for different directions and at far and closer source distances. The fan array is useful to assess the sensitivity of PSA spectral levels to sub-source – site geometry and to provide near-source data to constrain the attenuation shape. The fan array covers 180° in 7 directions (30° increment starting from the strike direction for each case) and 7 distances on each direction at 10, 30, 60, 100, 150, 200 and 400 km from the center of each fault plane.

In order to explore the h dependency on magnitude, I decided not to use distances < 20 km, to avoid inconsistency with the general attenuation shape clearly delineated by longer distance data (Figure 5.2.16). Under this data constraint in the first regression step I derived:

$$h = \mathbf{M}^2 - 3.1\mathbf{M} - 14.55 \quad (5.3)$$

In Figures 5.2.19 and 5.2.20 I show, for 0.5 Hz and 5.0 Hz, the average PSA values on the fan array sites (black symbols) along with the average PSA obtained on the perpendicular profiles (lines with marker symbols), both data sets generated for 30 iterations and for $\mathbf{M}$: 7.5, 8.0 (Figure 5.2.19) and $\mathbf{M}$: 8.5, 9.0 (Figure 5.2.20). I observed, from this comparison, that simulated PSA ordinates are on average insensitive to sub-

source – site geometry, the perpendicular profile results are robust and, even though near-source (between 20 and 30 km, depending on magnitude) PSA ordinates shown larger variability, on average they agree with the attenuation shape delineated by longer distance PSA ordinates. Thus, I used the data (PSA and PGA) from the fan shape array (excluding data below 20 km) and the data (PSA and PGA) from perpendicular arrays for the final two-step regression procedure. I plot, for 0.5 Hz and 5.0 Hz, the obtained PSA models and data for **M** 7.5 and **M** 8.0 in Figure 5.2.21 and for **M** 8.5 and **M** 9.0 in Figure 5.2.22; Table 5.8 contains the regression coefficients corresponding to equations 5.1 and 5.2. I compare in Figures from 5.2.23 to 5.2.26, for **M** 8.0, **M** 9.0, at 0.5, 1.0, 2.5, 5.0 Hertz, and for rock reference conditions, this study model (CS08) versus Youngs et al. (1997) (YS97), Gregor et al. (2002) (GR02), Atkinson and Boore (2003; 2008) (AB03/08) and Zhao et al. (2006) (ZH06) models. A similar comparison but for the PGA model is shown in Figure 5.2.27. Practically in all the cases, the minimum and maximum near-source amplitudes are determined by the AB03/08 and GR02 estimates respectively; for the GR02 model, this “overestimation” is extended, for low and high frequencies, to far-field distances. Overall, for **M** 8.0 and **M** 9.0 at a wide range of frequencies, the PSA estimates from the ZH06 model for Japan and from this study model (CS08) agree most closely, suggesting similar ground-motion source and attenuation properties between Japan and Cascadia. It is also noticeable that, for these two magnitudes, the YS97 model underestimates low-frequency ground-motion and overestimates high frequency ground-motion for most distances for both Cascadia and Japan, emphasizing the limitations of

global ground-motions relations to estimate region-specific ground motion amplitudes from subduction zone earthquakes.

From Figure 5.2.27, a similar conclusion can be drawn for the PGA model. I noted however, that for **M** 7.5 (which is not plotted), the model (CS08) predicts higher PGA values than the estimates for the **M** 8.0 scenario for near-source distances ($\sim < 20$ km) - a trend that is similar to the observations from Figure 16. This oversaturation of high frequencies at short distances could be a geometric effect that is related to the sub source – site distance distribution in the EXSIM program; it is not considered to be of any particular significance.

5.2.6 Ground-motion variability for the prediction equations

For empirical GMPE's, the dispersion of the data from the average value predicted by the equations has been described by σ , the standard deviation of the ground-motion residuals (i.e., the difference between the observed and the predicted ground motions). This parameter, usually known as the total variability σ_t , is separated into the inter-event variability σ_e and the intra-event variability σ_a , being related by

$$\sigma_t = \sqrt{\sigma_e^2 + \sigma_a^2} \quad (5.4)$$

The inter-event variability, which is interpreted as the variation of the average offset of ground-motion from the median for a given earthquake, results from random variations in the source process (given an specific distance and site conditions) thus, represents the temporal component of variability. On the other hand, the intra-event variability, which is the variation of ground-motion within a single earthquake about its

own median, represents the spatial component of variability (Strasser and Bommer, 2005, b)

In the context of ground-motion simulations, the overall inter-event variability (σ_{inter}) is the variability, for a given distance and azimuth, of the computed ground-motion across all stress drop levels, all slip distributions, and all hypocenter locations for an event, whereas the intra-event variability (σ_{intra}) is the variability that can be observed across a group of receivers at an specific distance for a given event (Strasser and Bommer, 2005, b). These authors describe a special type of inter-event variability, the pseudo inter-event variability (σ_{pseudo}), which is comparable with the inter-event variability (σ_e) from empirical data. Another approach (Toro et al., 1997) for the aleatory variability from numerical simulations implies two terms: *i*) the parametric variability, based on the range of ground motions due to varying parameters in the model for future earthquakes and *ii*) the modeling variability, based on misfit in calibrations of the simulation method.

According to these definitions, the inter-event variability (for this study model) that could be calculated from the results on the average simulated PSA and its variability due to changes in stress drop values (for $\Delta\sigma = 30$ bars and $\Delta\sigma = 150$ bars; section 5.2.3), would not represent the model inter-event variability because these stress drop values only define a range which does not reflect an actual estimate of the stress drop distribution for future earthquakes. Thus the simulations do not form a reasonable basis for estimating the aleatory variability. In this case, it is better to be guided by empirical evidence as to the appropriate level of aleatory uncertainty. This evidence is indirect, as

there are no suitable empirical PSA data for this region. Moreover, the Tokachi-Oki data are not very useful for estimating the expected variability, because of the large dispersion (mainly at high frequencies) of residuals observed for soil conditions in Japan (Atkinson and Casey, 2003). As pointed out by Atkinson (2006), in the context of seismic hazard calculations it is a common practice to use ground motion prediction equations, derived from many sites to estimate ground-motion for a single site. There may be, in these equations, factors for correcting roughly site effects, but these factors cannot fully account for the actual site response at each site, and thus the aleatory variability that is applicable to a single site will be overestimated. This component of the aleatory variability, due to variability in site-to-site amplification, should not be carried through to hazard analyses.

Under these circumstances, neither the simulations nor the data for the TK03 earthquake are useful to estimate the total variability for the CS08 model. However, a reasonable option for this purpose may be the use of a set of total variability values from a well developed and reliable uncertainty model. In this case, under the assumption that the variability for subduction events is about the same as that for crustal earthquakes in active regions, practically any of the recent PEER-NGA (Pacific Earthquake Engineering Research Center – Next Generation Attenuation of Ground-Motion Project) models can be used.

These models share common important features such as: they are valid for crustal earthquakes in active tectonic regions, for $5 \leq M \leq 8.5$, for $0 \leq R \leq 200$ km, site response is based on shear-wave velocity, and they were developed from a common, very

extensive and comprehensive database (PEER database for crustal earthquakes in active regions). Particularly, I adopted the Campbell and Bozorgnia (2007) model of aleatory variability (CB07) because it features a remarkable reduction of aleatory uncertainty (inter-event and intra-event variability) due to strict data selection criteria.

In Table 5.9 I include, as an estimation of the inter-event, intra-event, and total variability for the CS08 model, the values reported by Campbell and Bozorgnia (2007); specifically, the values that are calculated for a random horizontal component of ground-motion (the intra-event variability is adjusted accordingly), as they may be more suitable for the simulated PSA ordinates used in the development of the CS08 model.

Table 5.6 - Source, path and site parameters used in simulations of PSA for the 12 analyzed seismic scenarios (4 magnitudes at 3 sites) of the Cascadia subduction zone. Parameters for **M 9.0** are related to the geometry (c) in Figure 10.

	Parameter	Location	M 7.5	M 8.0	M 8.5	M 9.0
Source	orientation (deg)	FRA	[315, 10]	[315, 10]	[310, 10]	[330, 06]
	[strike, dip]	VIC	[315, 08]	[315, 08]	[315, 08]	[330, 06]
		SEA	[345, 06]	[345, 06]	[345, 06]	[345, 06]
	dimensions (km)	All	[80 x 30]	[170 x 90]	[380 x 90]	[600 x 150]
	[strike, dip]					
	depth (km)	All	10	✓	✓	✓
	[top of fault plane]					
	Location (top) (deg)	FRA	[-124.8,48.2]	[-125.2,47.7]	[-124.5,47.1]	[-124.2,44.9]
	[lat., lon.]	VIC	[-124.6,47.9]	[-124.7,47.2]	[-124.1,46.5]	[-124.2,44.9]
		SEA	[-124.3,46.9]	[-124.7,46.4]	[-124.7,45.3]	[-124.7,44.2]
Path	# of sub-sources	all	[8, 3]	[17, 9]	[38, 9]	[60, 15]
	[strike, dip]					
	pulsing area (%)	All	50	✓	✓	✓
	slip distribution & hypocentre location	All	random	random	Random	Random
	inelastic attenuation	All	$180 f^{0.45}$	✓	✓	✓
	$Q = Q_0 f^\eta$					
	geometric spreading		$b_I = [-1.0 ; -0.5]$, for $[R \leq 40 ; R > 40]$ (km)			
	duration [d] (s/km)	All	0.10	0.10	0.10	0.10
Site	κ factor	FRA	0.03	✓	✓	✓
		VIC	0.02			
		SEA	0.02			
	area & amplification factors	Vancouver (FRA)	→	Pleistocene		
		Victoria (VIC)	→	NEHRP B/C		
		Seattle (SEA)	→	NEHRP B/C		
	[reference level]					
	shear wave velocity (km/s)	FRA	3.8	✓	✓	✓
		VIC	3.8			
		SEA	3.6			
	density (g/cm ³)	FRA	2.8	✓	✓	✓
		VIC	2.8			
		SEA	2.7			
	damping factor (%)	All	5	✓	✓	✓
	[PSA]					

Table 5.7 - Shear wave velocity and density models and their corresponding amplification functions for VIC (Victoria), FRA (Vancouver) and SEA (Seattle). For each case, the reference site condition and the corresponding V_{S30} value are in bold.

Shear velocity (β) and density (ρ) profiles (Regions)									Amplification functions			
NEHRP B/C (VIC)			Pleistocene (FRA)			NEHRP B/C (SEA)			$i = 0.0^\circ$			
$V_{S30} = 758$ m/s			$V_{S30} = 414$ m/s			$V_{S30} = 758$ m/s						
depth	β	ρ	depth	β	ρ	depth	β	ρ	Freq.	VIC	FRA	SEA
m	m/s	g/cm ³	m	m/s	g/cm ³	m	m/s	g/cm ³	Hz	$\kappa=0.02$	$\kappa=0.03$	$\kappa=0.02$
0	207	1.77	0	400	2.02	0	207	1.77	0.10	1.07	1.65	1.45
1	207	1.77	600	1000	2.12	1	207	1.77	0.13	1.09	1.84	1.52
2	334	1.87	600	1500	2.20	2	334	1.87	0.16	1.11	2.07	1.59
3	367	1.91	2000	2310	2.38	3	367	1.91	0.20	1.14	2.29	1.65
4	367	1.87	4000	2880	2.59	4	367	1.87	0.25	1.18	2.50	1.70
5	501	1.89	5000	3460	2.67	5	501	1.89	0.32	1.23	2.69	1.75
6	501	1.96	5000	3690	2.71	6	501	1.96	0.40	1.30	2.84	1.78
7	713	2.01	7500	3690	2.71	7	713	2.01	0.50	1.37	2.96	1.80
8	770	2.05	7500	3810	2.80	8	770	2.05	0.63	1.45	3.02	1.81
9	770	2.08	10000	3930	2.88	9	770	2.08	0.79	1.54	3.05	1.82
10	770	2.11	16000	3930	2.88	10	770	2.11	1.00	1.61	3.04	1.83
15	1000	2.12	16000	3980	2.90	15	1000	2.12	1.26	1.69	3.02	1.83
30	1250	2.15	32000	4040	2.91	30	1250	2.15	1.58	1.74	2.95	1.82
200	1732	2.30	35000	4270	2.95	1000	1500	2.28	2.00	1.79	2.87	1.82
500	2540	2.50	50000	4500	3.00	2500	2310	2.38	2.51	1.82	2.75	1.81
1000	3350	2.67	60000	4620	3.10	7000	2880	2.59	3.16	1.85	2.59	1.81
2000	3690	2.71				10000	3460	2.67	3.98	1.86	2.41	1.81
9000	3810	2.80				13000	3690	2.71	5.01	1.88	2.20	1.81
16000	3870	2.84				20000	3810	2.8	6.31	1.88	1.96	1.81
16000	4020	2.90				25500	3920	2.88	7.94	1.87	1.69	1.79
20000	4100	2.92				30000	4040	2.91	10.00	1.79	1.43	1.71
27000	4160	2.93				35000	3870	2.85	12.59	1.69	1.13	1.61
33000	3690	2.90				40000	4040	2.91	15.85	1.51	0.90	1.44
38000	4160	2.93				45000	4156	2.95	19.95	1.29	0.62	1.23
						50000	4387	3.00				

Table 5.8 – Regression coefficients for model defined by equation 5.1 and 5.2

Frequency [Hz]	C_0	C_1	C_2	C_3	C_4
0.10	2.338	-0.6311	0.00000	0.5357	-0.0737
0.13	2.489	-0.6412	-0.00003	0.4760	-0.0629
0.16	2.569	-0.6048	-0.00024	0.4324	-0.0641
0.20	2.671	-0.5942	-0.00040	0.3822	-0.0417
0.25	2.814	-0.6108	-0.00046	0.3490	-0.0299
0.32	2.978	-0.6431	-0.00057	0.3258	-0.0103
0.40	3.104	-0.6585	-0.00063	0.2990	-0.0074
0.50	3.241	-0.6741	-0.00081	0.2696	-0.0064
0.63	3.393	-0.7101	-0.00089	0.2483	0.0103
0.79	3.453	-0.6885	-0.00119	0.2417	0.0125
1.00	3.621	-0.7376	-0.00128	0.2116	0.0328
1.26	3.733	-0.7473	-0.00159	0.2035	0.0292
1.58	3.859	-0.7746	-0.00179	0.2010	0.0153
2.00	3.999	-0.8211	-0.00195	0.1870	0.0271
2.50	4.167	-0.8854	-0.00211	0.1802	0.0258
3.16	4.303	-0.9322	-0.00231	0.1713	0.0270
4.00	4.472	-1.0133	-0.00234	0.1713	0.0255
5.00	4.746	-1.1691	-0.00212	0.1593	0.0432
6.30	4.930	-1.2671	-0.00204	0.1645	0.0301
8.00	5.209	-1.4404	-0.00163	0.1788	0.0151
10.00	5.490	-1.6257	-0.00115	0.1736	0.0261
12.60	5.676	-1.7633	-0.00071	0.1784	0.0245
15.85	5.823	-1.8889	-0.00022	0.1845	0.0160
20.00	5.843	-1.9391	0.00000	0.1813	0.0199
PGA	5.006	-1.5573	-0.00034	0.1774	0.0827

Table 5.9 – Inter-event, intra-event, and total sigma values from the ground-motion prediction model of Campbell and Bozorgnia (2007) (CB07), calculated for a random horizontal component of ground-motion. All sigma values are in $\log_{10}$ units and frequencies do not correspond to the original selection given for the CB07 model but for the CS08 model.

Frequency [Hz]	<i>Intra-event</i> σ_a	<i>Inter-event</i> σ_e	<i>Total sigma</i> σ_t
0.10	0.3159	0.2106	0.3797
0.13	0.2989	0.1884	0.3533
0.16	0.2904	0.1739	0.3385
0.20	0.2806	0.1559	0.3210
0.25	0.2705	0.1290	0.2997
0.32	0.2633	0.1396	0.2980
0.40	0.2638	0.1364	0.2970
0.50	0.2667	0.1286	0.2961
0.63	0.2640	0.1286	0.2936
0.79	0.2640	0.1220	0.2908
1.00	0.2653	0.1107	0.2875
1.26	0.2648	0.1013	0.2835
1.58	0.2612	0.0965	0.2785
2.00	0.2554	0.0929	0.2718
2.50	0.2514	0.0942	0.2685
3.16	0.2514	0.0936	0.2683
4.00	0.2463	0.1042	0.2674
5.00	0.2456	0.1081	0.2683
6.30	0.2443	0.1186	0.2716
8.00	0.2432	0.1226	0.2724
10.00	0.2421	0.1242	0.2721
12.60	0.2374	0.1262	0.2688
15.85	0.2347	0.1212	0.2641
20.00	0.2324	0.1120	0.2580
PGA	0.2198	0.0951	0.2395

Figure 5.2.1 - Fault plane representations for scenario simulations of interface events of M 7.5 (black continuous line), 8.0 (light gray dotted line), and 8.5 (dark gray continuous line), as used for Vancouver (top left), Victoria (top right) and Seattle (bottom center); location, geometry, and size of fault planes given in Table 5.6.

Figure 5.2.2 - Fault plane representations for scenario simulations of interface events of M 9 (geometries a, b and c), as used for Vancouver and Victoria (panel a), and for Seattle (panel b). Location and size of fault planes for geometries c are given in Table 5.6.

Figure 5.2.3 - Shear wave velocity models as function of depth for VIC (Victoria), FRA (Vancouver) and SEA (Seattle). Note that for FRA the reference site condition is Pleistocene ($V_{S30} = 414$ m/s), while for VIC and SEA it is NEHRP B/C.

Figure 5.2.4 - Amplification factors as a function of frequency for the reference site conditions and κ values for VIC, FRA and SEA. Shear wave velocity and density models as well as the amplification factor values are given in Table 5.7.

Figure 5.2.5 - Average 5% damped response spectra amplitudes for hypothetical interface events of M 9.0 (geometry c) at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), solid lines are $\pm$ one standard deviation from random hypocenter location and random slip distribution in simulations.

Figure 5.2.6 - Average 5% damped response spectra amplitudes for hypothetical interface events of **M 7.5** at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), solid lines are $\pm$ one standard deviation from random hypocenter location and random slip distribution in simulations.

Figure 5.2.7 - Average 5% damped response spectra amplitudes for hypothetical interface events of **M** 8.0 at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), solid lines are $\pm$ one standard deviation from random hypocenter location and random slip distribution in simulations.

Figure 5.2.8 - Average 5% damped response spectra amplitudes for hypothetical interface events of **M** 8.5 at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), solid lines are $\pm$ one standard deviation from random hypocenter location and random slip distribution in simulations.

Figure 5.2.9 - Scaling of average 5% damped response spectral amplitudes with magnitude **M**: 7.5, 8.0, 8.5, and 9.0 (c).

Figure 5.2.10 - Response spectra for M 9.0 (c) at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), light and dark solid lines are average PSA for $\Delta\sigma = 30$ and $\Delta\sigma = 150$ bars, respectively.

Figure 5.2.11 - Response spectra for **M 7.5** at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), light and dark solid lines are average PSA for $\Delta\sigma = 30$ and $\Delta\sigma = 150$ bars, respectively.

Figure 5.2.12 - Response spectra for **M 8.0** at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), light and dark solid lines are average PSA for $\Delta\sigma = 30$ and $\Delta\sigma = 150$ bars, respectively.

Figure 5.2.13 - Response spectra for **M 8.5** at Vancouver, Victoria, and Seattle. Light gray dashed line is the average PSA for $\Delta\sigma = 90$ bars (100 trials), light and dark solid lines are average PSA for $\Delta\sigma = 30$ and $\Delta\sigma = 150$ bars, respectively.

Figure 5.2.14 – Average 5% damped PSA for M 9.0 scenarios with different geometries (a, b, and c) for Vancouver, Victoria and Seattle. Average PSA for M8.5 is also shown for reference.

Figure 5.2.15 – Average 5% damped response spectra for an **M** 8.0 scenario for different subduction zone attenuation models. AS00 = Atkinson and Silva (2000), adopted for the Cascadia region; OS92 = Ordaz and Singh (1992) for the Mexican subduction zone and TK07 is the Tokachi-Oki model for this study (all for $\Delta\sigma = 90$ bars). AB03 lines are the PSA estimates from Atkinson and Boore (2003) relationships for NEHRP C and rock site conditions at different distances, modified at 2.5 Hz and 5.0 Hz by Atkinson and Boore (2008) correction factors but by Cascadia regional factors given in AB03.

Figure 5.2.16 – Average 5% damped PSA for Cascadia subduction events (“best estimate” source and path) as a function of closest distance to fault (R_{cd}) for NEHRP B/C boundary site conditions (VIC velocity model). Top panel show attenuation for **M** 7.5, 8.0, 8.5, and 9-c for 0.5 Hz, while lower panel are for 5.0 Hz. Estimates are made for perpendicular profiles from the trench as shown in Figure 5.2.17.

Figure 5.2.17 - Scenario seismic sources and station arrays: perpendicular profiles (black square symbols) and fan shape profiles (gray open symbols) for **M 8.0** (left frame) and **M 9.0 (c geometry; right frame)**.

Figure 5.2.18 – Average PSA versus closest distance to fault (R_{cd}) (10 trials) for B/C site conditions (VIC), for a M8.0 event, at 0.5 and 5 Hz. Symbols depict average PSA at R_{cd} values associated with the perpendicular profile sites for models as given in Figure 5.2.15: AS00, TK07, and OS92 (all for B/C site conditions). Lines show predictions of AB03 model for NEHRP B (rock) and C site conditions, corrected by AB08 factors.

Figure 5.2.19 - Average PSA at 0.5 Hz and 5.0 Hz, from simulated amplitudes for the fan array sites (black symbols), along with the average PSA obtained on the perpendicular profiles (lines with marker symbols), both data sets generated for 30 iterations and for **M 7.5** (top panel) and **M 8.0** (bottom panel) on B/C site conditions.

Figure 5.2.20 - Average PSA at 0.5 Hz and 5.0 Hz, from simulated amplitudes for the fan array sites (black symbols), along with the average PSA obtained on the perpendicular profiles (lines with marker symbols), both data sets generated for 30 iterations and for **M 8.5** (top panel) and **M 9.0** (bottom panel) on B/C site conditions.

Figure 5.2.21 - Average PSA values from simulations for the perpendicular and fan shape arrays of sites (symbols), compared to predictions of derived ground-motion predictions equations of Table 5.8 (lines), all for B/C site conditions and for **M 7.5** and **M 8.0** at 0.5 and 5 Hz.

Figure 5.2.22 - Average PSA values from simulations for the perpendicular and fan shape arrays of sites (symbols), compared to predictions of derived ground-motion predictions equations of Table 5.8 (lines), all for B/C site conditions and for **M 8.5** and **M 9.0** (geometry c) at 0.5 and 5 Hz.

Figure 5.2.23 - Average 5% damped response spectral amplitudes, for $M 8.0$ at 0.5 Hz and 1.0 Hz, from different ground-motion models: CS08 (this study), YS97 (Youngs et al., 1997), GR02 (Gregor et al., 2002), AB03 (Atkinson and Boore, 2003, 2008), and ZH06 (Zhao et al., 2006). All estimates are for rock site conditions.

Figure 5.2.24 - Average 5% damped response spectral amplitudes, for $M 8.0$ at 2.5 Hz and 5.0 Hz, from different ground-motion models: CS08 (this study), YS97 (Youngs et al., 1997), GR02 (Gregor et al., 2002), AB03 (Atkinson and Boore, 2003, 2008), and ZH06 (Zhao et al., 2006). All estimates are for rock site conditions.

Figure 5.2.25 - Average 5% damped response spectral amplitudes, for $M 9.0$ at 0.5 Hz and 1.0 Hz, from different ground-motion models: CS08 (this study), YS97 (Youngs et al., 1997), GR02 (Gregor et al., 2002), AB03 (Atkinson and Boore, 2003, 2008), and ZH06 (Zhao et al., 2006). All estimates are for rock site conditions.

Figure 5.2.26 - Average 5% damped response spectral amplitudes, for $M 9.0$ at 2.5 Hz and 5.0 Hz, from different ground-motion models: CS08 (this study), YS97 (Youngs et al., 1997), GR02 (Gregor et al., 2002), AB03 (Atkinson and Boore, 2003, 2008), and ZH06 (Zhao et al., 2006). All estimates are for rock site conditions.

Figure 5.2.27 - Average peak ground acceleration (PGA) for **M 8.0** (top) and **M 9.0** (bottom) earthquakes, from different ground-motion models: CS08 (this study), YS97 (Youngs et al., 1997), GR02 (Gregor et al., 2002), AB03 (Atkinson and Boore, 2003, 2008), and ZH06 (Zhao et al., 2006). All estimates are for rock site conditions.

Chapter 6. Summary and conclusions

This research work was developed in two main parts. In the first part, I studied the 2003 **M** 8.1 Tokachi-Oki earthquake and some of its aftershocks in order to gain knowledge about source, path and site processes for this earthquake and, within the context of the stochastic finite-fault modeling technique, explore ground-motion scaling with magnitude, distance and site conditions from large interface earthquakes. In the second part, the calibrated model was used to infer an average stress drop value, for large interface earthquakes and its variability. For an average stress drop value, and appropriate regional path and site parameters, I then simulated expected ground motion from this type of earthquakes on the Cascadia subduction zone, and explored consequences of uncertainties in input parameters. I generalized these results deriving theoretical ground-motion attenuation relationships for large interface earthquakes occurring at the Cascadia subduction zone.

The main analyses and conclusions, along with some recommendations for future studies, are summarized as follows.

6.1 Empirical analyses of the 2003 Tokachi-Oki sequence

I derived and explored, as a preliminary step for further analyses, Fourier acceleration spectra (*Faccn*) from time series records of the main shock and selected aftershocks with **M**: 7.3, 6.4, 5.9, and 5.5. I observed a flattening of the spectral shape at larger distances and higher frequencies, occurring at progressively closer distances as magnitude decreases and frequency increases; therefore, distance and frequency

constraints were set on *Faccn* data. This effect imposes limitations on the K-NET data and should be considered in *Faccn* analysis (in terms of distance and frequency) for the region.

Some features of spectral values were observed in terms of geological conditions related to the complex tectonic setting of NE Japan. Specifically, fore-arc stations have significantly greater high-frequency content than do back-arc stations; stations sited on sedimentary basin structures show larger low-frequency amplitudes than do stations located on the volcanic arc province of the region. To accommodate these effects, dummy variables (geological variables) were used in the regression analysis. Regional amplifications or attenuation effects can modify the expected shape or behavior of ground-motion attenuation and, therefore, should be accounted for in the derivation of ground-motion relationships.

Site classification (based on the NEHRP classification scheme) was done from available shear-wave velocity and density profile data for the K-NET stations. Two approaches (“constant velocity” and “correlation”) were used (Boore, 2004) to classify sites; no significant differences were observed between these results.

As a second step I performed empirical data analysis on *Faccn* and time series of the 2003 Tokachi-Oki sequence. *Faccn* data for each event were regressed to extract attenuation and amplification models and near-source amplitudes. The possibility of non-linear site amplification effects was considered by applying correction factors (Atkinson and Boore, 2007) to the larger events and closer distances.

The attenuation of *Faccn* data for all events was reasonably well modeled with a geometrical spreading coefficient $b_l = -1.0$ and the inelastic attenuation described by $Q = 135 f^{0.76}$. Even though this model is a good representation of the Tokachi-Oki sequence data, because it reflects the strong attenuation observed at back-arc stations, should not be considered as a general model for NE Japan.

Within the same regression framework, site amplification functions for each magnitude and for NEHRP D and E sites relative to C sites were derived. Good agreement between functions for different magnitudes (for the same site class) was observed. I also derived a generic amplification function for C sites of the K-NET stations, for which I used regional (crust) and local shear-wave velocity profiles and the $\frac{1}{4}$ wave-length approach (Joyner et al., 1981) for i (angle of incidence) = 45° and $\kappa_0 = 0.045$ s. The obtained κ_0 variability for C sites is large, from $\kappa_0 = 0.056$ (from linear fit of kappa measurements versus distance) to $\kappa_0 = 0.035$ (from comparisons with amplification functions obtained from spectral ratio techniques applied to **M** 8.1 data). This variability is an important aspect that should be studied because of the impact of kappa on the high-frequency shape of *Faccn* and the interaction with stress drop estimation in modeling procedures. A possibility of improvement on the knowledge of this factor (at least for the Japanese stations) is by analyzing data recorded on the KIK-NIED network, where sensors / records are available at depth of about 100 m and on the surface for the same site, possibly allowing a clearer analysis for this factor.

Empirical near-source spectra (from the regression) were compared to near-source “Brune” spectra (Brune, 1970) to estimate stress drop values for the analyzed

earthquakes. Good agreement between reported and estimated magnitudes was found (only for the M 8.1 case, due to frequency band limitations, was not possible to estimate magnitude). Stress drop values are around 100 bars for the interface earthquakes (Ito et al., 2004) of the sequence and about 200 bars for the in-slab events; for the M 8.1 event I obtained 120 bars, which is in agreement with reported values (Yagi, 2004). No dependency of $\Delta\sigma$ on magnitude but possibly on event type (origin) was found.

High-frequency decay rates (kappa factor) were measured on log-linear plot of *Faccn* data and used to estimate κ_0 values. No dependency of κ_0 on magnitude but on site condition was found. For C, D, and E sites, I obtained $\kappa_0 = 0.056$, 0.068 , and 0.08 respectively.

Duration of ground-motion was measured on acceleration records and plotted versus distance to estimate distance-dependent slope of duration models for the main shock and aftershocks. Calculated values for this slope were $d \sim 0.1$ and $d \sim 0.7$ for the M 8.1 and for lower magnitudes respectively.

Due to the fact that data are restricted to $R_{cd} \geq 50$ km and that the amount of data for lower magnitudes is low, the uncertainty in κ_0 and d estimation is high mainly for lower magnitudes.

6.2 Stochastic simulation of the 2003 Tokachi-Oki sequence

The empirically-derived parameters were used as input for the stochastic finite-fault simulation of PSA ordinates for the 2003 Tokachi-Oki sequence. The EXSIM code

(Motazedian and Atkinson, 2005), used for the simulations, implements this particular stochastic technique.

Practically all path and site parameters were the same for all events (the exception is a slight difference in distance-dependent slope of duration from $d = 0.09$ to $d = 0.07$ for the main shock and for the aftershocks respectively). Source parameters (slip distribution and fault geometry) were taken from Yagi (2004) for the main shock, while for lower magnitudes a random slip distribution was assumed and fault geometries were based on the Wells and Coopersmith (1994) relationships. Pulsing area was fixed to 50% for all cases; $\Delta\sigma$ was the only “free” parameter in the procedure. The simulated amplitudes matched the observations well, showing a well-behaved distribution of residuals, in frequency and distance. However, in order to improve a small positive trend observed at high frequencies, I modified the C amplification function (reference condition in simulations), by changing i and κ_0 values from 45° to 0° and from 0.045 to 0.02 respectively, to better accommodate the higher amplitudes of PSA at high frequencies. These results provided a calibrated and validated simulation model for the second part of the research.

6.3 Stress drop variability for large interface earthquakes

Stress drop variability amongst large events on subduction interfaces was estimated to place some constraints on stress drop for mega-thrust events in the Cascadia region. The approach was to perform simple comparisons between observed and simulated PSA values, for selected interface earthquakes, at distances of about 60 (± 20

km), assuming that at this distance range the PSA amplitudes are not significantly affected by inelastic attenuation and the dominant geometrical spreading attenuation is described by $b_1 = -1.0$. The calibrated model (part 1) was used to simulate PSA values at different subduction regions; the model was adapted to regional conditions (by considering crustal amplifications) and the reference site condition was the NEHRP B/C boundary. Atkinson and Boore (2007) equations were used to correct observed PSA data to B/C reference level. Thus, for the simulations, I calculate a generic B/C amplification function from shear-wave velocity profiles for a suite of K-NET stations; this local B/C function was common to all simulations (regions) and because of this commonality, the stress drop estimations were insensitive to the regional amplification functions. Therefore, a better description of the local B/C amplifications and κ values for each region is evidently a necessary task to improve the average stress drop estimation and its variability. From these comparisons, an average value $\Delta\sigma = 90$ bars was estimated for large interface earthquakes with a minimum and maximum limits of $\Delta\sigma = 30$ bars and $\Delta\sigma = 150$ bars, describing its variability, perhaps representing one or two standard deviations from the average value.

6.4 Ground-motion predictions for the Cascadia subduction zone

The calibrated EXSIM model was used to simulate ground-motion for generic sites at Vancouver, Victoria and Seattle. The considered scenario earthquakes with magnitude M : 7.5, 8.0, 8.5, and 9.0, were constrained by the geometry of the anticipated rupture area for the region (Hyndman and Wang, 1995; Flück et al, 1997), by the average

$\Delta\sigma$ for interface earthquakes and its variability and by regional estimations of attenuation and site conditions. The generic site conditions for each area correspond to firm sites: B/C boundary for Victoria and Seattle areas and the Pleistocene layer ($\sim$ NEHRP C) for Vancouver. The attenuation model of Atkinson and Silva (2000) (AS00) was considered and adopted as the most representative for the Cascadia region (based on Atkinson, 2005). Different regional amplification factors were also generated from diverse sources of information for the crust velocity structure. Fault planes were placed symmetrically around perpendicular lines from the trench to each area and fault areas were based on Wells and Coppersmith (1994), Kanamori and Anderson (1975), and Beresnev and Atkinson (1997) formulations; a 90 km width was the fixed value for $M \geq 8$ and length was adjusted according to the average calculated fault area. For the $M 9$ fault area, three different geometries were analyzed from which the wider area (150 km width) was chosen as the preferred geometry. These parameters, along with random slip distribution and random hypocenter location, were considered as the “best estimates” for the input for simulations.

The random variability in 5% damped PSA ordinates (after 100 iterations) for any magnitude or area (city) is around 20% to 30% of the average PSA for the “best estimates” of input parameters. This random variability represents only the effect of aleatory slip distribution and aleatory hypocenter location. However, when minimum and maximum $\Delta\sigma$ values are used, the uncertainty in $\Delta\sigma$ is about 50% of the average PSA for mega-thrust earthquakes in the Cascadia region.

The uncertainty in fault geometry was significant for the **M** 9 scenario for Vancouver or Victoria; the elongated geometry (90 km width) would generate lower PSA amplitudes than the **M** 8.5 scenario for these localities, most probably due to the large distances of a significant part of the fault area (sub-sources) with respect to any observation point. The contribution from distant sub-sources to the ground-motion would be low. In fact, this is a limitation of the EXSIM code; it uses rectangular fault geometries, thus the modeling of PSA for curved subduction zones will require an improvement (adaptation) of this code to accommodate curved geometries.

The uncertainty in the attenuation model for simulations has a significant impact. Differences of a factor of 1.5 to 2.0 between PSA ordinates obtained from AS00 model and those simulated with OS92 model (Ordaz and Singh, 1992), representative of the Mexican subduction zone, were observed at distances from 85 km to 145 km.

A database of simulated PSA ordinates was generated for sites located at distances from 20 to 400 km in 7 directions (30° apart each other, covering 180° from the strike direction), on B/C boundary conditions for magnitudes from **M** 7.5 to **M** 9.0. Results were generalized by fitting them to a ground-motion prediction equation for large interface earthquakes in the Cascadia region; an estimate of aleatory variability associated with the equations was provided. This study model (CS08) was compared to global empirical attenuation models, specifically: Atkinson and Boore, 2003 (AB03); and Youngs et al., 1997 (YS97), and also to regional attenuation models: Gregor et al., 2002 (GR02) for Cascadia; Zhao et al., 2006 (ZH06) for Japan. Similar PSA predictions between this study and the ZH06 model, for a wide range of frequencies and distances,

were noted. Global models (AB03 and YS97) tend to underestimate regional amplitudes (CS08 and ZH06) at near-source distances and overestimate regional amplitudes at far-field distances, reflecting a slower attenuation rate with respect to the regional models, the most critical case being the AB03 model at low frequencies. I speculate that the flat attenuation shape for AB03 model for interface earthquakes, and to some extent the YS97 model as well, could be related to the relatively large amount of Mexican subduction zone data in the AB03 database, for which a significant regional amplification effect has been reported (Singh et al., 1995; Cárdenas-Soto and Chávez-García, 2003). This points to the need for a revision of global subduction databases and a reassessment of the global ground-motion prediction equations for subduction earthquakes.

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