Empirical Study of the
Ground Motion Produced by Earthquakes
Originating in the Subduction Zones of
Japan and Cascadia

By

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A thesis submitted to the
Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Masters of Science

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Chair, Department of Earth Sciences

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Abstract

Response spectra are calculated for approximately 26000 ground motion observations of 350 earthquakes in Japan and Cascadia during 1993-2001. Data range for **M** is 4.5 - 7.0 (including Geiyo and Nisqually, **M** 6.8) and hypocentral distance 10 - 1000 km. Focal mechanism and depth classify the subduction earthquakes into three types: crustal, inslab and interface.

The study shows that Japanese interface earthquakes exhibit lower mean ground motion and less attenuation than inslab earthquakes. In Cascadia, larger **M** inslab events have similar ground motions to those in Japan, however attenuation increases as **M** decreases. For moderate magnitudes, inslab earthquakes in Cascadia attenuate rapidly, following the Atkinson & Boore (2002) ground motion relations. In contrast, attenuation for Japanese inslab events is less pronounced, following the Youngs et al (1997) relations, resulting in larger ground motions than Cascadia. Further data are needed to confirm these apparent regional differences.

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“Dyfal donc a dyrr y garreg” (traditional)

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1 Seismological Background

1.1 Study Purpose

“Buildings kill people, not earthquakes.” This commonly used phrase emphasises that magnitude alone does not control an earthquake’s human and economic impact. Population density and building construction in the affected area determine the extent of damage and injury. To estimate the impact of future earthquakes in an inhabited locale and anticipate the emergency measures needed it is vital to know the amplitudes, frequency content, and duration of shaking, and how that shaking will affect buildings and other structures. The latter is the concern of civil engineers and will not be discussed here. The characteristics of shaking, here referred to as “ground motion,” are the focus of this study.

During the last few decades, researchers have shown that the Cascadia region of northwestern United States and southwestern Canada has experienced large earthquakes in the past and is very likely to do so again. Magnitude estimates vary, with general agreement in the order of $8 < M < 9$, i.e. a great earthquake. By analogy with other very large global earthquakes, movements on the thrusting interface between the subducting and the overlying plate are considered to have caused these events in Cascadia. Section 2.4 presents a summary of the evidence for these conclusions.

A great earthquake in Cascadia has the potential to be felt throughout the densely populated areas of Seattle, Portland, Vancouver, and Victoria. Local and national governments are actively involved in preparing these communities by reviewing building codes, retrofitting older structures, adding safety and redundancy features to the utility supplies, and practising emergency procedures. These preparations require a seismological estimate of the shaking that may be experienced across the region.

Providing an estimate of the shaking for Cascadia is challenging since the local seismic records include neither an event this large nor any occurring on the subduction interface. Present ground motion estimates being used in Cascadia (Adams et al, 1996) are based upon relationships developed for crustal earthquakes in California, and global subduction zone events. It is not known whether these global relationships accurately represent ground motions that would result from subduction interface thrust earthquakes in Cascadia.

The goal of this study is to evaluate the similarity between ground motions in Cascadia and other subduction zones. The strategy is to consider Japan, a region similar tectonically to Cascadia that has abundant records of interface, inslab and crustal earthquakes. The pattern of ground motion is first established for events of all three types in Japan. If ground motions from inslab and crustal events behave similarly in Japan and Cascadia, it would be reasonable to assume the levels of shaking for large interface events will also be similar. This

would improve the confidence in predicting ground motions for a Cascadia interface earthquake from estimates that have been derived in other subduction zones.

1.2 Seismic Hazard Analysis

Defining the expected ground motion for an area is a complex task. What measure is desired? One might be the maximum ground motion observed to date, but this would exclude the possibility of a larger or closer earthquake occurring in the future. Alternatively, using the absolute maximum considered physically reasonable may reassure some people, but result in over-engineering of structures. However, it is very difficult to reliably define the maximum ground motion possible in a region.

Seismologists and engineers have developed a middle-ground approach based on probabilistic seismic hazard assessment. This defines regional seismogenic source zones and calculates the probability that any source zone will produce ground motion exceeding a certain amplitude level at the site of interest (Cornell, 1968, McGuire, 1977). A site specific hazard curve is developed using magnitude recurrence relations for the source zones, and attenuation relationships that describe ground motion amplitude as a function of earthquake magnitude and distance. Details of seismic hazard methods are given in Anderson et al (2000), and an introductory summary may be found in British Geological Survey (1998).

Fundamental to creating a seismic hazard curve is the knowledge of how ground motion attenuates with magnitude and distance. Various authors have simulated the ground motion expected from a Cascadia interface earthquake (Atkinson & Beresnev, 1998; Atkinson & Boore, 1997; Cohee et al, 1991) using seismological parameters relevant to the region. Alternatively, empirical subduction data from other regions is used to estimate ground motion for Cascadia (Youngs et al, 1997; Atkinson & Boore, 2002). This study uses empirical data to investigate whether earthquakes in Cascadia exhibit ground motion behaviour consistent with events in other geographic regions.

1.3 Ground Motion Variations

The strength of ground motion felt by an observer is dependent upon many variables: distance from the earthquake source, earthquake magnitude, seismic wave path through the Earth from the source, and conditions where the person is standing, e.g. rock or soil, consolidated or loose ground materials. Seismic energy released during an earthquake is transmitted by waves, which may be reflected, refracted, scattered or attenuated by the material through which they pass. This results in great variation of the wave amplitude at any point. Ground motion is often quantified for engineering purposes by pseudo acceleration (PSA) as a function of frequency (see section 1.4).

The attenuation of the waves along their travel path is affected by the subsurface geological structure, the seismic velocity layers present, and large scale discontinuities between the tectonic plates, as found in subduction zones.

Previous work exists showing that the observed ground motion and rate of attenuation also varies with the depth of the hypocentre (Crouse, 1991, Atkinson, 1995, Youngs et al, 1997, Molas and Yamazaki, 1995). In the context of a subduction zone, 'depth' may be interpreted as differentiating between the different tectonic origins of earthquakes.

Atkinson (1995) subdivided data observed in Cascadia into shallow crustal events (depth $H < 10$ km), intermediate crustal & plate events ($20 < H < 30$ km), and deep inslab events ($H > 30$ km) rather than by subduction structure.

Cascadia data from small and moderate earthquakes, and deep inslab events ($H > 30$ km) were combined in Atkinson (1997) with larger events from global subduction zones to derive ground motion relations. Source depth rather than mechanism was investigated as a variable. The results indicated a magnitude and depth dependence of the attenuation rate.

Youngs et al (1997) developed subduction zone ground motion relations using a global database. They defined a tectonic classification using focal mechanism to distinguish inslab and interface events but resorted to depth in most cases due to lack of data. They made comparisons between crustal and subduction events

using ground motion relations from previous published work to predict peak ground acceleration (PGA) from crustal events and model PGA data for large magnitude interface events. Their analysis indicated that the PGA for subduction zone events decreases more slowly away from the source than that for crustal events. Secondly, the PGA from an inslab event may be expected to be greater than an interface event at the same distance. Finally, they conclude that both depth and tectonic type are significant to the PGA relationships.

Ambiguities in the definition of earthquake type, combined with limitations of the database have made these previous studies inconclusive with regards to the effects of event type on observed ground motion.

This study investigates further the possibility that the earthquake source location within the subduction zone produces differences in ground motion. Subduction zones contain spatial variations and changes in seismic velocity (see section 2.1), and waves from a hypocentre located on the subduction interface may be expected to propagate differently than waves from events whose hypocentres are within the slab or crustal layer. The subduction zone structure may offer preferential channels of propagation for seismic signals resulting in larger than anticipated ground motions at a distance from the source.

Beginning with a tectonic model of the source processes occurring during subduction, earthquakes will be classified as crustal, inslab and interface and the

observed ground motions compared. Accounting for all other major contributions to observed ground motion differences (earthquake magnitude, hypocentral distance, and recording site) any gross effects of the seismic source type or wave travel path will be revealed.

Most earthquakes are caused by shearing motion along a fault plane, and can be best represented as double couple force systems. This motion produces a four-lobed pattern of varying ground motion amplitude. This radiation pattern may be distorted in the region close to the hypocentre due to the influence of fault size (near-field effects). Directivity effects may also stretch the pattern in the direction of rupture propagation. Somerville et al (1997), Frankel et al (1999), and Anderson et al (2000) discuss manifestations of these radiation and directivity effects in real data. Very few, if any of the data in this study would be considered “near-field” and no specific correction is made for directivity effects. Both horizontal components are included in the data set, and there is a good distribution of stations in a range of azimuths. The spread of maximum and minimum values incorporates directivity and radiation pattern effects.

1.4 Response Spectra

An action plan of earthquake preparedness needs to know how this expected ground motion will affect buildings. When the ground vibrates at a frequency close to the natural frequency of a building, larger movements are produced in

the structure and there is more likelihood of failure. Earthquakes produce seismic waves at a wide range of frequencies, and each frequency is attenuated to different degrees as the wave travels through the earth. In very general terms, the earth acts as a low-pass filter, with greater attenuation of high frequency waves. Low frequency seismic waves from distant large earthquakes often damage tall structures, whereas short, rigid structures are damaged by high frequency waves from smaller, closer events.

One measure of ground motion is insufficient for earthquake resistant design and engineering. Instead of the single value of peak ground acceleration or velocity, the response spectrum of the ground motion is calculated (Newmark and Hall, 1982). The record of ground motion with time observed at a site during an earthquake is applied to a series of damped, single degree of freedom oscillators. Typical damping is 5% critical damping. The natural frequencies of the oscillators are chosen within the range for typical structures. For example, ten storey buildings typically have a natural frequency near 1 Hz, whereas for a one-storey building it may be 10 Hz. The maximum displacement of each oscillator plotted against the oscillator frequency becomes the relative displacement response spectrum (SD). The calculated response spectrum may be considered as representing the movement of a simple structure of natural frequency equal to those plotted if subject to this earthquake. In most cases, the relative displacement is converted to acceleration, termed “pseudo acceleration” (PSA) where $PSA = (2 \pi f)^2 SD$.

Response spectra are preferred as a measure of ground motion for seismic hazard studies since they can better represent the effects of ground motion on structures and are used in many structural response analysis methods. The most recent edition of the Geological Survey of Canada seismic hazard maps (Adams et al, 1996) pointedly moved to a response spectra approach, instead of peak ground velocity or acceleration as in previous publications. This method requires a frequency dependent ground motion attenuation model for each region.

Some previous works on ground motion relationships for Japan and Cascadia largely focus on peak ground motions (Youngs et al, 1997, Molas and Yamazaki, 1995, Fukushima and Tanaka, 1990). In contrast, Atkinson (1995, 1997) used a spectral approach for Cascadia that will be continued here.

1.5 Study Design Principles

This study has been designed to use a large database from one seismic region dominated by a subduction zone (Japan) and containing all three types of earthquakes to investigate the effect of event type on ground motion. Ground motion response spectra are compared for events of similar types and magnitudes occurring in Japan and Cascadia. The recording network in Japan has dense station coverage and produces digital strong motion data. All earthquakes recorded in Japan are classified into the three types of subduction zone events based upon reported earthquake parameters that are considered

reliable. The classification scheme can be applied rigorously and quickly to a large data set. A minimal number of institutional sources are used to provide ground motion and classification parameters to avoid methodological and regional differences. (Note: Tichelaar & Ruff (1993) place great importance on using only one method of depth determination for all earthquakes in their data set. They indicate the possibility of errors in previous works when multiple databases were used since each source of earthquake parameters will use a unique data set or method to determine the reported value.)

2 Tectonic Framework

2.1 Subduction Zones

The largest recorded earthquake to date occurred on 22 May 1960 in Chile (38.2S, 72.6W) with moment magnitude (**M**) = 9.5. There have been three other earthquakes with moment magnitude greater than 9 this century:

M = 9.2, 28 March 1964, Aleutian Islands, Alaska (61.1N, 147.5W)

M = 9.1, 9 March 1957, Alaska (51.3N 175.8W)

M = 9.0, 4 November 1952 Kamchatka (52.75N, 159.5E)

All of these are thrust events and concluded to be caused by displacement on the interface between the subducting oceanic plate and the crust of the overriding continental plate. Nine of the ten largest earthquakes worldwide are subduction interface events (NEIC). Figure 2.1 shows a simplified cross section view of a subduction zone.

Earthquakes are also produced within the subducting slab (inslab events) and the overriding continental crust (crustal events). Inslab events may occur at depths from 20 to 500 - 600 km, with magnitudes typically below **M** = 7.5 (Ludwin, 2001). They are often referred to as Wadati-Benioff zone earthquakes. In some regions, hypocentres from inslab earthquakes outline a double layered Wadati-Benioff zone as the slab descends into the mantle, possibly due to

changing stress distribution or mineral phases within the subducting slab.

Although not reaching magnitudes as large as the great interface earthquakes, inslab events occur more frequently and so have a significant impact on the seismic hazard for a region. Crustal events tend to include greater numbers of smaller magnitude earthquakes and a range of focal mechanisms, caused by a variety of stress-strain patterns in the continental plate.

Earthquakes do not occur everywhere along the interface. The degree of coupling between the subducting and overriding plates determines the aseismic or seismogenic interface segments. Aftershock location and actual rupture areas of interface events can delineate the dimensions of the seismogenic zone (Tichelaar and Ruff, 1993). Typically, the temperature and physical properties needed for interface earthquakes to occur exist at depths between 10 km and 40 km (Satake & Tanioka, 1999).

Mathematical models of the interface can simulate observable characteristics related to subduction, coupling depth being one. Oleskevich, Hyndman, and Wang (1999) apply this method to determine the controlling limits of the seismogenic zone. By entering known parameters of heat flow, trench temperature, and elastic deformation, they calculated thermal and structural models of the subduction zone. The model parameters that best fit the observed seismogenic regions in all subduction zones are considered reliable. They conclude that the dominant controls on the extent of the seismogenic zone are

the isotherms. Earthquakes do not occur until the interface temperature reaches 100-150 °C, and cease when the temperature rises above 350-450 °C. These temperatures correlate to depths of approximately 5 - 10 km and 20 -30 km for the interface in Nankai and Cascadia. The exception to this is when the interface intersects with the continental Moho, the boundary between the crust and the mantle, before this temperature is reached. In this case, hydrated forearc mantle rocks permit stable sliding on the interface, and the Moho depth provides the maximum down dip extent of the seismogenic zone.

Seismic surveying and tomography can establish the subduction zone geometry since the different constituents have temperature and density variations, creating velocity contrasts. Typically the subducting slab is cooler and denser than the surrounding mantle, and therefore has higher seismic velocities. Definite boundaries are imaged at the top and bottom, giving a global average for slab thickness in the order of 90 km (Zhao et al, 1994, Zhao et al, 1997, Deal and Nolet, 1999). Double Wadati-Benioff zones within the slab may be seen extending from 60-200 km deep as a narrow upper layer of velocity increase, and lower layer of velocity decrease (Zhao et al, 1994, Hasegawa et al, 1994). Immediately on top and parallel to the subducting slab is a low velocity zone that continues to the order of 200 km depth that may be associated with dehydration of the slab and reciprocal hydration of the above mantle (Zhao et al, 1994). The forearc mantle wedge is generally a zone of intermediate velocities. Extremely low velocities are associated with the higher temperatures seen beneath many

forearc volcanoes, and in the mantle below the subducting slab (Hasegawa et al, 1994, Deal and Nolet, 1999). Finally, the overriding continental crust has low seismic velocities.

Generally, seismic velocity is inversely related to wave attenuation: high seismic velocity will result in low attenuation. The velocity structure present in a subduction zone supports the premise that the slab would be a preferential propagator of seismic waves since it has lower attenuation than the surrounding mantle. Figure 2.2 shows a simplified view of the velocity and attenuation distribution within a subduction zone.

2.2 Cascadia Subduction Zone

The Cascadia subduction zone parallels Vancouver Island, British Columbia, Canada, and the coastal regions of the states of Washington and Oregon, U.S.A. Here the Juan de Fuca oceanic plate is subducting under the continental North American plate. The subduction trench where the Juan de Fuca plate submerges lies between 100 and 200 km offshore in the Pacific Ocean. Further west is a series of spreading and transform plate boundaries that frame the subduction zone from its northern limits at the tip of Vancouver Island to the Oregon-California state line (see figure 2.3).

Cascadia experiences frequent crustal earthquakes, which sometimes reach large magnitudes. In 1993 two crustal events in Klamath Falls of $M = 5.9$ and 6.0

caused two deaths and significant damage. The 1872 magnitude 7.3 earthquake in Washington State and the 1946 magnitude 7.3 earthquake on Vancouver Island are classified as crustal events (Source: PGC website).

There are regular inslab earthquakes, with damaging inslab events occurring every few decades. The most recent was the 2001 Nisqually $M = 6.8$ event that caused major interruptions to business, transportation and everyday life in Seattle. Unlike other subduction zones, there have been no interface earthquakes along the Cascadia subduction zone since formal seismic recording began in 1790.

The Juan de Fuca plate is young and hot relative to other global subducting plates, with an age of 6-8 million years (Ma) and temperatures between 220 and 260 °C beneath the overlying sediments at the subducting trench (see table 2.1). These sediments overlying the oceanic crust add a 3200 m thick insulating layer that along with the young age contribute to the very high temperatures (Oleskevich et al, 1999).

Global positioning satellite (GPS) data has measured the converging movement of the Juan de Fuca plate as 40 mm/yr in the direction N68°E. Convergence is therefore slightly oblique to the subduction trench (Hyndman et al, 1998, Parsons et al, 1998).

The Juan de Fuca plate's dip varies across the subduction zone, reported at the trench as 5-6° downwards to the east, increasing to 12-16° under the coastline and further east (Gerdom et al, 2000, Parsons et al, 1998). The dip is shallower (8°) under the Olympic peninsula west of Seattle but increases both north and south (Ludwin, 2001, Hyndman et al, 1998). Contours of depth to the top of the subducting plate describe a slightly arched surface with the bulge where the dip is shallower, i.e. the Olympic peninsula. The top of the subducting plate is 28 km deep on west Vancouver Island, increasing to 44 km in the centre of the island (Cassidy et al, 1998) and 30 km deep at the coastline west of Seattle (Ludwin, 2001). An important marker for the seismogenic zone is where the subducting plate intersects with the continental Moho, which here is seen at 35-40 km deep, 145-260 km east of the trench (Oleskevich et al, 1999).

The geometry of the subduction interface seismogenic zone is more speculative. The only diagnostic method available in Cascadia is to use modelling techniques, since there are no recorded interface events. Using their thermal control limits, Oleskevich et al (1999) determine that the interface seismogenic zone stretches from 5 km east of the trench (100-150°C isotherm) to 50-90 km from the trench at a depth of 12 km (350°C isotherm). They also define a transitional seismogenic zone as temperature increases with depth, which cannot initiate earthquake rupture but would be part of the fault plane in a great earthquake scenario. For Cascadia this transitional zone extends to 100-170 km from the trench, 25 km deep (450°C isotherm). This downdip limit follows the coastline, except in the

Olympic peninsula. The shallower subduction dip there extends the isotherms and hence the seismogenic zone calculated by Oleskevich et al (1999), putting the downdip limit of transitional rupture just under the land (see figure 2.3).

The maximum compressive continental stress in Cascadia is in a direction parallel to the subduction trench, i.e. northwest, as determined by earthquake focal mechanisms, in situ stress measurements, and quaternary neotectonic evidence (Wells et al, 1998, Wang, 2000, Geist, 1995). Wang (2000) attributes the origin of this margin parallel compression to the existence of a forearc sliver being transported along the subduction zone strike by complex relative plate movements. Along-strike resistance to this motion causes compressive stress within the sliver. A continental shear zone parallel to the margin, characteristic of obliquely converging subduction zones defines the inland extent of the sliver and bounds its motion.

In contrast, laser ranging and GPS geodetic measurements record crustal shortening perpendicular to the trench, i.e. in the same northeast direction as plate convergence (Wang, 2000). Wang (2000) explains that this “stress-strain paradox” is not necessarily a contradiction since in an elastic process, observed strain changes are not always representative of the absolute stress regime. The geodetic measurements record margin normal elastic shortening during an earthquake cycle due to the locked subduction fault. This pattern of crustal shortening would be seen irrespective of the regional stress.

2.3 Nankai Trough Subduction Zone

To better estimate the likely ground motions in Cascadia from future large earthquakes, a subduction zone with more ground motion data and similar tectonic properties can be used for comparison. Table 2.1 lists the basic properties of a selection of the world's major oceanic subduction zones, all of which have produced great thrust events. Cascadia may be summarised as young, hot, and slowly converging with a shallow downdip limit to the interface seismogenic zone. The best match to these parameters would be southwest Japan, the Nankai trough subduction zone where the Philippine Sea oceanic plate is converging with the Eurasian continental plate.

Table 2.1: Subduction Zone Properties

	Age at trench (Ma)	Convergence rate (mm/yr)	Dip angle (degrees)	Temp at trench °C	Downdip seismogenic limit (km)	Sediment layer (m)
Cascadia ¹	6-8	40-45	8-13	225-260	21-29 ^a	3200
SW Japan ¹	0-15	40	11-20	90	32-38 ^a	1500
Chile (south) ¹	23	90	14	40	50 ^a	135
Chile (north) ¹	48	84	12-19	15-30	40-50 ^b	175
Alaska ¹	48	60	11	50	40-50 ^b	2400
NE Japan ²	125	94	13	-	37-55	-
Mexico ²	15	60	15	-	20-30	-

Notes:

¹Oleskevich et al, 1999, ²Tichelaar and Ruff, 1993

^a Downdip limit controlled by 450° isotherm

^b Downdip limit controlled by intersection with Moho

The match is not perfect: there is a significant difference in the temperature of the interface at the trench. Nankai is hotter than the other subduction zones listed, but cooler than Cascadia, attributed to thinner sediments in Nankai. The temperature difference moves the thermally controlled limits to the interface seismogenic zone in Nankai slightly further landward from the trench and deeper than Cascadia. In Nankai, the updip limit for interface earthquakes is 12-14 km deep, at 30 km from the trench. The downdip limit is still reached above Moho-interface intersection, for which both the depth and distance from the trench is very similar between the two regions (33-35 km deep, 165-220 km from the trench for Nankai) (Oleskevich et al, 1999).

Instead of being bounded by active transform and spreading margins as in Cascadia, the Nankai trough is in close proximity to the Izu-Bonin trench, another active subduction zone, and an extinct, small spreading ridge. This relic spreading ridge causes the Philippine plate to range in age from 0 - 15 Ma along the trench (Oleskevich et al, 1999). Figure 2.4 summarises the tectonic setting for Japan.

The subduction stress regime in Nankai is remarkably similar to Cascadia. The subduction interface is weakly coupled, with oblique convergence and a margin-parallel continental shear zone similar to Cascadia (Wells et al, 1998). Nankai also presents the same “stress-strain paradox” as seen in Cascadia. Specifically,

the direction of crustal shortening is perpendicular to the plate margin, whereas the maximum compressive stress is parallel to the margin (Wang, 2000).

Nankai is suitable as a source of data for this study using the design criteria given in section 1.5. The seismic records available are extensive, including all three types of subduction zone earthquakes (crustal, inslab and interface). The recurrence of large interface earthquakes is 100 years or so and events of moderate magnitudes occur more frequently.

Tectonically, southwest Japan (Nankai) is distinct from northeast Japan, with differences in age, convergence rate, and temperature of the subducting plate. However, seismically the regions are very alike. Crouse (1991) observed similar variations of peak ground motion with hypocentral distance and decided to combine data from all of Japan into one data set. When studying the data set accumulated for this study, earthquakes originating in the Nankai Trough region produce the same distribution of ground motion as those from the Japan Trench. A small sample comparison is shown in Figure 2.5. In the interests of maximising the data available, the same decision as Crouse (1991) is made here and there is no subdivision of Japan into separate subduction zones.

2.4 Past Cascadia Subduction Thrust Earthquakes

In the two centuries of written or instrumental records in Cascadia, there is no mention of a great earthquake like those that occur on the plate interface in other

subduction zones. The oral histories of the coastal peoples include cataclysmic events that may relate to a giant earthquake, but these cannot be used to reliably denote a time, place or magnitude. When formulating the seismic hazard assessment for the region, it was very necessary to determine if the Cascadia subduction zone had ever generated an earthquake of magnitude similar to the Alaska or Chile great events.

The answers came not from seismology, but field geology along the shores and inlets of the Pacific Northwest. Four main categories of evidence were unearthed (Atwater et al, 1995, Hyndman et al, 1998). First, tidal muds are found on top of marsh and forest soils, containing relics of trees and plants that grow above high tide. This indicates the coast experienced sudden subsidence, as opposed to a gradual rising of water level that would produce peat layers. Second, sand layers are seen covering buried soils, consistent with deposition following a tsunami. Third, strong shaking has produced liquefaction features such as sand volcanoes extruded through and over soil layers. Finally, ocean shelf deposits contain periodic turbidite layers of coarse material brought in by landslides, interspersed with fine mud from long term sedimentation. All four groups of features may be attributed to large interface earthquakes occurring in the Cascadia subduction zone.

Radiocarbon dating of trees killed during subsidence indicates the most recent of these major earthquakes occurred between 1680 - 1720 AD (Atwater et al, 1995)

and similar although less precise dates are obtained from isotope dating of plants and tree ring studies. The ocean turbidites provide a longer recurrence pattern. Since a distinctive volcanic ash marker layer was deposited when Crater Lake was formed, thirteen landslide episodes have occurred with time intervals of 300 - 900 years, the most recent being close to 1700 AD (Hyndman et al, 1998). These general dates have been narrowed down by Satake et al (1996). Using detailed reports of wave height and damage in Japan, they investigate possible global earthquake sources for a tsunami event in 1700 AD. Accounting for directivity effects and correlating with seismic records where available, they conclude that Cascadia is the most likely tsunamigenic source zone. To match the observed wave heights and arrivals in Japan, the earthquake would have occurred on the Cascadia subduction thrust at 05.00 (GMT) 27th January 1700 with a moment magnitude approximately 9.

Obermeier and Dickenson (2000) attempted to quantify a magnitude from the paleoseismic evidence. They estimated the peak surface acceleration necessary to produce the observed properties of liquefaction features using in situ geotechnical measurements. Their results ranged from 20% to 42% g (198 - 415 cm/s²) which are less than predicted for **M** = 9 by the Youngs et al (1997) ground motion relations, and approximately equal to that predicted for **M** = 8. The authors stress the limits of their methods and need for refinement, however the results present interesting issues in general terms relevant to seismic hazard assessment in Cascadia. It may be that the magnitude estimate from tsunami

evidence of Satake et al (1996) is overly large. It may be that existing relations are not appropriate for describing how ground motion from the 1700 AD great thrust earthquake scales with distance.

With the advent of modern seismology and recording network improvements, the surprise is that no earthquakes have been recorded whose source may be attributed to movement on the subduction interface. Normally the average number of earthquakes occurring on a fault decreases log-linearly to a maximum magnitude (Gutenberg-Richter relation). The Cascadia thrust interface is atypical since it appears to produce no small earthquakes, only intermittent very large events (Hyndman et al, 1998).

2.5 Future Cascadia Subduction Thrust Earthquakes

Paleoseismic evidence reveals the potential for a great interface earthquake in Cascadia, but this is tempered by the absence of any modern events on the thrust interface. GPS data and levelling surveys consistently confirm landward convergence of the Juan de Fuca plate. Stable aseismic sliding on the interface is discounted by uplift observed throughout the coastal regions. Part of this uplift may be due to rotational relative motion between segments of the Cascadia forearc plate (Wells et al, 1998) although not all the deformation can be accommodated this way. Finally the significant convergence rate would produce

large coastal mountains, which are absent in Cascadia. Therefore observed deformation must be elastic and has been released during earthquakes.

This evidence best supports a model for Cascadia where subduction is ongoing but the interface between plates is “locked”, i.e. no motion currently occurs. A great thrust earthquake on the interface will most likely release the accumulating strain some time in the future.

Estimates of magnitude for the great thrust event ultimately rely on estimates of the potential rupture zone of the interface. Heaton and Hartzell (1987) equate the maximum dimensions of the Cascadia subduction interface with the 1000 km long rupture zone of the giant 1960 Chilean earthquake. They conclude that the maximum moment magnitude for a Cascadia event, representing total rupture of the interface would be $M = 9.5$. Moment magnitude is closely dependent on the area of rupture. Atwater et al (1995) estimate that for the Cascadia subduction zone a rupture plane 50 km by 200 km would produce an earthquake of $M = 8$. $M = 9$ would require a rupture of 100 km wide along 1000 km of the Cascadia interface. Oleskevich, Hyndman and Wang (1999) model a seismogenic interface for Cascadia with maximum widths of 100 km in the north, 180 km in the centre, and 140 km in the south, and a length of approximately 600 km. These widths include the transition zone, which is only considered seismogenic when incorporated into a rupture that begins in the main seismogenic interface. In

terms of simple rupture dimensions, the Cascadia interface therefore seems capable of producing a great earthquake of $M = 9$.

While investigating conditions that affect the degree of interface coupling, Ruff and Kanamori (1980) (reported in Lay & Wallace, 1995) developed an empirical relationship between the age of the subducting plate (T) in millions of years, convergence rate (V) in cm/yr, and maximum moment magnitude produced at global subduction zones:

$$M = -0.00886 T + 0.134 V + 7.96$$

Substituting the plate parameters for Cascadia (see table 2.1) predicts a maximum moment magnitude $8.4 < M < 9.1$, i.e. a great earthquake occurring in Cascadia would be in accordance with observations at other subduction zones.

Fig 2.1: Subduction Zone Earthquakes

Modified from: Tichelaar & Ruff (1993), Satake & Tanioka (1999),
Hyndman et al (1998), Parsons et al (1998)

Fig 2.2: Seismic Velocities and Attenuation in a Subduction Zone

Modified from: Hasegawa et al (1994), Zhao et al (1994)

Fig 2.3: Cascadia Tectonic Setting

Large arrow = direction of convergence
 Hatched line = subduction trench

(Hyndman et al, 1998)

Fig 2.4: Japan Tectonic Setting

Fig 2.5: Comparison of Nankai Trough to Japan Trench

3 Data

3.1 Data Sources

Background seismicity for both areas of study was established using the National Earthquake Information Centre (NEIC) of the United States Geological Survey (USGS): <http://neic.usgs.gov/neis/epic/epic.html>

All ground motion data collected were strong motion accelerograph or broad band seismographic data. Only the horizontal channels were used since this study is interested in S-wave motion, which is predominantly recorded in the horizontal components.

3.1.1 Japan

Hypocentre parameters (latitude, longitude, depth, time, moment magnitude) were obtained from the National Research Institute for Earth Science and Disaster Prevention, Science and Technology Agency (NIED) event listings, both directly from the NIED: <http://www.bosai.go.jp/index.html> and the NIED parameters reported by KNET (see next paragraph).

Ground motion data for earthquakes of $M > 4.5$ were supplied by the Japanese Kyoshin-Net (KNET) network of NIED. KNET data were available on CD-ROM for 1996, 1997, and 1998 and online for 1999, 2000, and 2001:

<http://www.k-net.bosai.go.jp>

Focal mechanism parameters (strike, dip, rake or fault plane solution) necessary to classify the events were collected from F-Net, the Full Range Seismograph Network of Japan (formerly known as Fundamental Research on Earthquakes and Earth's Interior Anomaly (FREESIA)):

<http://argent.geo.bosai.go.jp/freesia/index.html>.

F-Net automatically calculates centroid moment tensor (CMT) solutions using waveform inversion of long-period surface waves for events detected by NIED networks. F-Net data were online for 1997 onwards; therefore a majority of the KNET 1996 ground motion data are not used, as no classification is possible. Some large $M > 5.5$ events absent from F-net have mechanism parameters available in CMT solutions published by Harvard University Seismology:

<http://www.seismology.harvard.edu>

For the special event study of the Geiyo earthquake (see section 7.1), more detailed files including hypocentre and mechanism parameters were available from the Earthquake Research Institute (ERI) of the University of Tokyo:

<http://www.eri.u-tokyo.ac.jp/>

Spatially, NIED, ERI and F-Net reports all have a northern limit around 45° latitude, covering epicentres occurring within the main islands of Japan.

3.1.2 Cascadia

Ground motion data were collected for events of $M > 4.0$ from four sources. Each contributed data for 1999 and 2000, plus some archived events of special interest dating back to 1993.

1) Canadian National Seismic Network (CNSN) station records from Natural Resources Canada autodrm: <http://www.seismo.nrcan.gc.ca>

2) United States National Seismic Network (USNSN) station records from the United States Geological Services (USGS) via autodrm:
<http://neic.usgs.gov/neis/AUTODRM/autodrm.html>

3) Global Digital Network stations from the Incorporated Research Institutions for Seismology (IRIS) Consortium online data source. Local area network (University of Washington) station records were also made available from the IRIS Consortium as special event files: <http://www.iris.edu>

4) For the special study of the Nisqually earthquake (see section 7.1), additional strong motion data were obtained from the Consortium of Organisations for Strong Motion Observation Systems (COSMOS) Virtual Data Centre:
<http://db.cosmos-eq.org/>

Focal mechanism parameters to be used in the classification scheme were extracted from the CMT solutions published by three sources:

1) Pacific Geoscience Centre, Natural Resources Canada (PGC NRCan) has published CMT solutions for significant Cascadia events from 1999 onwards:

<http://www.pgc.nrcan.gc.ca/seismo/MTS/index.html>

2) Oregon State University, College of Oceanic and Atmospheric Sciences has archived CMT solutions for pre-1998: *<http://quakes.oce.orst.edu/moment-tensor>*

3) Harvard University Seismology CMT archives catalogues events of **M** > 5.5. Data are sparse for Cascadia but used where available:

<http://www.seismology.harvard.edu>

3.2 Kyoshin Net (KNET)

Using KNET data from Japan to study differences between event types fulfils a number of the design principles as described in section 1.5. Stations are densely and evenly distributed throughout Japan (see figure 3.1). A high number of stations report data for each event. Most events of **M** > 5.5 have in excess of one hundred station files. For smaller magnitude events, the number of stations per event is quite variable, some even in single digits. KNET routinely provides site information for its stations, which allows classification and correction for site effects (see chapter 6).

KNET stations are equipped with JMA87 accelerometers. These instruments have a flat response to ground motion in the frequency range 0.05 to 10 Hz

(Molas & Yamazaki, 1995). This is advantageous since required instrument corrections are minimal if the spectral response is kept within these frequency limits (see section 5.1.2). Molas & Yamazaki (1995) also describe the JMA87 accelerometer as measuring accelerations from 30×10^{-3} to 980 cm s^{-2} for periods 1 sec to 10 minutes, and having a resolution of $\pm 0.03 \text{ cm s}^{-2}$. They consider this resolution to mean peak motions less than 1 cm s^{-2} in one horizontal component are "unreliable," and therefore omit them. Recorded PGA values for KNET in this study are all above this value. PSA values are less than this at some frequencies since they are the calculated spectral response levels rather than the input ground motions.

The plotted response spectra (see section 1.4) from KNET for small to moderate events from stations at distances greater than 500 km from the earthquake show a large increase in the spread of PSA values. This "ballooning" is magnitude and frequency dependent and would not be expected in normal observations.

Likewise there is a flattening off of PSA amplitudes at low PSA levels and large distances. These are clear indications that PSA amplitudes for small events at large distances are below the reliable sensitivity range of the instruments. These observations are analysed in detail with explanatory plots in section 7.2.1. The result is to limit the magnitude and distance range where results may be considered valid for each spectral frequency.

3.3 Earthquake Parameters and Limitations for Japan

3.3.1 Event Identifiers

Comparing the institutional sources of earthquake data reveals inconsistencies in the events included in each catalogue. No one database has been determined as a definitive source of events or parameters. When cross-referencing between databases, it is assumed that the reported date, time, latitude and longitude are suitable identifiers. One parameter alone is insufficient to uniquely identify any earthquake. A small degree of variability in time and position is allowed when matching event parameters, ground motion data, and focal mechanisms obtained from different sources. The merged data sets have been repeatedly checked manually for discrepancies and incorrectly matched data. A preferred hierarchy is used as to which data source provides the most reliable value of each event parameter.

3.3.2 Hypocentre Location

The preferred values of hypocentre latitude, longitude, and depth are the NIED data as reported by KNET in the ground motion data files. However see section 3.3.5 for more discussion of depth uncertainties. In keeping basically to one source for both hypocentre and ground motion data, difficulties in event identification were minimised. In addition, NIED reports position errors on their web site. Typically the errors in latitude or longitude are in the order of <0.01

degrees, although in a few instances the reported position error is up to 1.9 degrees. Applying these generally small errors to the data does not alter the anticipated classification by position (see section 4.1.1).

Comparing NIED hypocentre location data with that from NEIC, the source of the background seismicity, reveals a systematic east/west offset between the two databases through the latitudes under study. This contributes to the decision to use only one database as a source of hypocentre parameters.

3.3.3 Moment Magnitude

Reviewing the data sources clearly emphasises the variability between the magnitude scales in use, for example, Richter magnitude (M_L), surface wave or body wave magnitudes (M_S and M_b), the local scale used in Japan (M_{JMA}), and moment magnitude (**M**). Also clear is that different data sources often produce different values of the same magnitude type. Magnitude of any type is not suitable as an event identifier.

The goals of the study are to make comparisons of ground motion at a wide range of magnitudes between geographic regions, so a consistent value of magnitude for each event is essential. The use of moment magnitude (**M**) is preferred, as it has been proven to be more consistent globally, at all hypocentral distances, and at large magnitudes (Lay & Wallace, 1995; Wells & Coppersmith,

1994). It is the only magnitude scale that provides a physical measure of the size of the earthquake rupture.

The preferred source for **M** is that reported in the F-Net database, calculated from their waveform inversion moment tensor solution. The data are supplemented by the Harvard CMT solution value when **M** is not available from F-Net.

3.3.4 Hypocentral Distance

The distance measure between a station and the earthquake source used in this study is hypocentral distance (R). It is easily calculated from station position and hypocentre, both widely reported parameters. Other studies use the closest distance to the fault surface when comparing ground motion at different stations. This is harder to determine, even if a simplified formula is used. Molas & Yamazaki (1995) compared the use of both distance measures and concluded that their results in comparing ground motion trends were unaffected. This issue is of limited importance here due to the moderate magnitudes of the study events. Large earthquakes typically have a large rupture area, and so the relative difference between hypocentral distance and distance to the fault becomes more noticeable at large magnitudes.

3.3.5 Depth Uncertainties

NIED reports large uncertainties for the depths assigned to earthquakes, often greater than 50%, and even higher than 100% in some circumstances. Figure 3.2 shows hypocentres for one latitude range, with error bars. Hasegawa et al (1994) comment that for their study in the northeastern area of Japan the accuracy of reported focal depths was quite low. They attribute the difficulty in hypocentre determination to the complex tectonic structure under Japan, the associated complex seismic velocity zones, and to the geographic fact that the majority of seismic stations in Japan are on land, whereas a substantial number of earthquakes occur under the ocean. This latter fact is especially true for subduction earthquakes, which are the focus of this study.

In light of the above, hypocentral depth is not used as a primary identifier or classifier for earthquakes and is given minimal weight during the analysis. See section 4.2 for how depth is applied to the classification process. Where used, the preferred value of depth is that reported by KNET in the ground motion data header file.

3.4 Database for Study

3.4.1 Geographic Distribution

The initial data collection for Japan was in excess of 1000 events during the period 1996 - early 2001. This list is filtered for only those of $M > 4.5$ and cross-referenced to the sources of focal mechanisms, i.e. F-Net and Harvard University. The resulting data set contains 331 events for which both ground motion records and focal mechanisms are available. From these 331 events are 25 851 individual horizontal components, i.e. one station recording one component for one event at one hypocentral distance. Figure 3.1 shows the KNET station locations and epicentres for events in the Japan data set.

For Cascadia, the data quantity is much smaller due to the tectonic and instrumentation differences (see sections 2 and 3). Only events associated with the subduction zone are included, and those from the transform boundaries in the Pacific are excluded. The final data set represents 381 horizontal components from 17 earthquakes. Included are some earthquakes of special interest in the region dating back to 1993, as listed in table 3.1. Figure 3.3 maps the 65 recording stations and 17 epicentres for Cascadia events.

Table 3.1: Cascadia Earthquakes

Name	Date	Time (UT)	Lat	Lon	M	Depth (km)	Station Records	Type ¹
Klamath Falls	1993-09-21	03:29	42.32	-122.03	6	9	2	C
	1993-12-04	22:16	42.29	-122.01	5.48	6	4	C
	1994-01-03	01:26	49.72	-126.77	5.58	25	4	S
Robinson Point	1995-01-29	03:11	47.39	-122.36	4.85	18	8	C
	1995-05-20	12:48	46.88	-121.94	3.84	15	8	C
	1996-02-22	20:06	49.9	-123.90	3.94	9	6	C
	1996-03-18	08:02	49.9	-126.88	4.92	15	4	C
Duvall	1996-05-03	04:03	47.76	-121.88	5.22	10	12	C
Vancouver Island (offshore)	1996-10-06	20:13	49.05	-127.88	6.22	4	12	C
Bremerton	1997-06-23	19:13	47.60	-122.57	4.53	4	8	C
Okanagan	1997-06-24	14:23	48.36	-119.89	4.06	12	8	C
	1997-06-24	14:41	49.24	-123.62	4.3	12	8	C
Satsop	1999-07-03	01:42	47.1	-123.50	5.8	45	18	S
	1999-12-11	12:53	48.53	-123.26	5	54	34	S
	1999-12-25	06:49	48.56	-125	5	27	1	S
	2000-08-01	21:33	49.48	-126.14	5.2	40	40	S
Nisqually	2001-05-28	18:54	47.17	-122.73	6.8	52	200	S

Note: ¹ C = crustal earthquake, S = Inslab earthquake

3.4.2 Magnitude and Distance Distribution

During this study, comparisons will be made between data for each event type grouped by hypocentral distance (**R**) and moment magnitude **M**. The choice of comparison ranges takes into account the data distribution within the three main classification subsets (see section 4.2.1). Figure 3.4 (a) shows that the best data

coverage for Japan is between $4.5 < \mathbf{M} < 6$ and $80 \text{ km} < R < 300 \text{ km}$. At extremes of both $\mathbf{M}$ and R , the density of coverage is inconsistent. Included for reference in figure 3.4 (b) is the same data distribution plot for Cascadia, which as anticipated is extremely sparse. Only inslab and crustal earthquakes have been recorded in Cascadia (see section 2.2).

3.4.3 Frequency Distribution

The classification scheme (see section 4.2.1) can be considered as three main or eight subsets of data. It is important to appreciate how both the events and station records are distributed within these classes. Figure 3.5 (a) shows the distribution for Japan. With the exception of the crustal classification there are typically 5 - 10 events, or 500 - 1000 station records in each 0.5 unit $\mathbf{M}$ bin for Japan. For $\mathbf{M} > 6.0$, the number of events decreases significantly. In general the number of data points is in proportion to the number of events.

The situation in Cascadia is more variable. As figure 3.5 (b) shows, some individual events have a very large number of data points and others very few. For the purposes of comparison the same $\mathbf{M}$ intervals are used as for Japan. The scarcity of Cascadia earthquakes means that not every interval includes data. This is not as restricting to the study as it might seem since relationships between event types are investigated in Japan, with good data coverage and then applied to Cascadia.

Fig 3.1 Japan: Location of Epicentres and KNET Seismograph Stations

Earthquakes occurring during 1996 - 2001 and used in this study.

Fig 3.2: Depth Errors for Japan Hypocentres between 30° and 40° N
Reported by NIED

Fig 3.3 Cascadia: Location of Epicentres and Seismograph Stations

Figure 3.4 (a): Japan: Moment Magnitude (M) and Hypocentral Distance (R)

Figure 3.4 (b): Cascadia: Moment Magnitude (M) and Hypocentral Distance (R)

Figure 3.5 (a) Frequency Analysis for Japan Data 1996 - 2001

Figure 3.5 (b) Frequency Analysis for Cascadia Data 1993 - 2001

4 Earthquake Classification Scheme

4.1 Classification Groups and Determinants

The tectonic background for the subduction zones being studied (see section 2.1) identifies three types of earthquakes: crustal events occurring in the continental crust above the subducting plate, interface events occurring where the two plates are in contact, and inslab events occurring within the subducting plate. Youngs et al (1997), Tichelaar & Ruff (1993), and Hasegawa et al (1994) classify events into these three groups using a variety of methods, which are described below and form the basis of the classification scheme for this study.

4.1.1 Location of Hypocentres

Within a subduction zone, tectonic structure broadly defines the seismogenic areas. The outline of both the overriding and the subducting plate can be established by plotting the distribution of earthquake hypocentres in the region. A single event can then be classified by plotting its hypocentre within this three dimensional framework of background seismicity. Based on the geometry determined by seismicity patterns, earthquakes with depths less than 10 km will be crustal events, and any events deeper than 50 km will be inslab. Between these limits though it is possible for all three types of subduction earthquakes to occur, at different lateral positions.

4.1.2 Focal Mechanism

General principles dictate that two types of subduction earthquakes, interface and inslab will have contrasting focal mechanisms. Thrust mechanisms will dominate for interface events caused by one plate overriding the other, whereas the environment of plate flexure or extension within which inslab events occur will result in a wide range of mechanisms. The third class of subduction zone earthquakes, crustal events may also have any type of focal mechanism.

Fault mechanism solutions assume the earthquake can be represented as a double-couple force system, and map quadrants of compressional or dilatational first motion. Due to symmetry in the double-couple, two orthogonal fault plane solutions exist. Both solutions will be reported and either one may be the actual value. Determination of the actual fault plane requires surface observation of the fault, directivity analysis, aftershock studies, or detailed analysis of the source mechanism (Lay & Wallace 1995).

4.1.2.1 Rake Angle

Rake angle is a measure of the direction of movement on the fault plane, in relation to the strike of the fault. It is most simply determined by a first motion stereographic plot of earthquake arrivals. The rake angle can identify the focal mechanism. Motion that is entirely in the horizontal plane has a rake angle (λ) of 0° or 180° , representing strike-slip fault mechanisms. Pure thrust would have $\lambda =$

90°, and for pure normal $\lambda = 270^\circ$ (or -90°). Figure 4.1 shows the full range of subdivisions and classifications. Two possible rake angles are calculated from a fault mechanism solution for each event due to symmetry as described in section 4.1.2.

4.1.2.2 *Compressive Stress Axes*

The axes of maximum (P), minimum (T), and intermediate (B) compressive stress may also be determined when completing a stereographic plot of first motion arrivals (Lay & Wallace, 1995). Alternative nomenclature terms B as the 'null axis' and T as the 'tension axis.' The projections of these axes on the stereonet give singular angular values. Thrust focal mechanisms have $45^\circ < T < 90^\circ$ and $0^\circ < B < 90^\circ$ whereas normal mechanisms have $0^\circ < T < 45^\circ$ and $0^\circ < B < 45^\circ$. Strike slip mechanisms have $0^\circ < T < 90^\circ$ and $45^\circ < B < 90^\circ$ (Harvard University Seismology <http://www.seismology.harvard.edu>).

4.1.2.3 *Slip Ratio*

Wells & Coppersmith (1994) classify events by the ratio of horizontal to vertical slip as observed at surface displacement. Where the ratio is greater than 1:1, the event is classified as strike slip. If the ratio is less than 1:1, the event is either a normal or thrust (reverse) depending on the orientation of the vertical displacement. They find that their slip ratio classification is in good agreement with that from the rake angle of earthquake focal mechanisms.

4.1.2.4 *Focal Mechanism Solutions*

Whether derived from first motion analysis or central moment tensor solutions (CMT), focal mechanism diagrams, or “beach balls” provide a broad classification of earthquake type. Figure 4.2 illustrates five basic focal mechanism diagrams: strike slip, normal, thrust, oblique normal and oblique thrust. The “beach ball” represents a lower hemisphere stereographic projection of the direction of P wave first motion around the earthquake’s hypocentre onto a flat plane. The curves between dark (compressional motion) and light (dilatational motion) regions are projections of the orthogonal fault plane and auxiliary plane (Lay & Wallace, 1995). The distribution pattern of light and dark regions provides the possible orientation and parameters of the fault planes, hence the fault type.

Published numerical CMT solutions give parameters such as the orientation of compressive stress axes and fault plane properties. Wells and Coppersmith (1994) prefer to use fault parameters from focal mechanisms derived by waveform inversion of long period seismic waves, rather than those from short period first motion analysis since the difficulties of identifying phase arrivals are avoided.

4.1.3 Subduction Interface Events

Youngs et al (1997) followed established convention in defining subduction interface events as having “shallow angle thrust” focal mechanisms. This relates

to prior knowledge about the geometry of the subducting plate (see section 2.1), specifically the typical down-going dip angle. It is assumed that events due to movement on the interface will have fault planes with a similar dip.

The fault plane dip angle is obtained from a stereographic first motion plot or focal mechanism solution. Due to symmetry, two possible dip angles are reported for an event. When delimiting ranges for classification as “shallow” or “steep” it is necessary to consider both reported orthogonal dip angles, relative to the horizontal and vertical planes. For example, for a plate dipping at 10° to horizontal, “shallow angle” thrust earthquakes may be classified as interface if their dip angle (δ) is in the range either $\delta < 20^\circ$ or $\delta > 70^\circ$. A “high” or “steep” angle thrust event for the same plate would be when $20^\circ < \delta < 70^\circ$.

Subduction interface events can only initiate at depths up to that of the Moho in the over-riding plate. Interface events are most likely to occur in a depth range of 10 to 40 km. Global averages for plate geometry indicate that the depth of interface events will be less than 50 km. Any event at depths greater than this must be classified as inslab (Tichelaar & Ruff, 1993). Along the subduction interface, the seismogenic zone correlates to the zone of high coupling between the plates. This is necessary for the interface to be capable of generating a thrust earthquake (Tichelaar & Ruff, 1993). Typical dimensions of the seismogenic interface have been discussed in sections 2.1, 2.2, and 2.3. Youngs et al (1997) classified events with hypocentral depths between 20 and 50 km as interface,

and deeper as inslab. Using 50 km as a maximum depth for interface events is acceptable, but making it a minimum depth for inslab events allows the possibility of erroneously classifying an inslab event as interface.

The strike of an interface event must be parallel to the strike of the associated subduction trench. This is applied in the work of Reyners & McGinty (1999) when studying earthquakes in northeastern New Zealand and to this data set in section 4.2.3.

4.2 Progression of Classification Scheme

Initially, classification by hypocentre location was thought to be the most acceptable route. However, due to limitations of both location and depth (see sections 3.3.2 and 3.3.5), it has been deemed impossible to reliably identify an interface event from inslab on this basis alone. Hypocentre location and specifically depth are rejected as primary parameters for classification.

In moving to use focal mechanism as a classifier, the first approach is to use published CMT solutions. Two problems arise here: first, not all events have a published solution, and second, the need for a visual interpretation of the “beach ball” is cumbersome. Even for Japan, using sources such as Harvard Seismology, the Earthquake Research Institute (ERI), and the quarterly catalogue from Physics of the Earth and Planetary Interiors (PEPI) covers a very

small proportion of the earthquake catalogue and does not always offer a numerical solution.

Whilst accessing the Harvard Seismology database, classification is possible by using their facility to search by the compressive stress axes parameters. Results still have only limited coverage of the Japanese database and cross-referencing events to the KNET ground motion catalogue has proved challenging.

With the use of searchable, numerical parameters being identified as a positive factor, an alternate method of determining focal mechanism was required. The F-Net database publishes numerical focal mechanism parameters (strike, dip, and rake) for an extensive portion of the Japan seismicity catalogue. At this fourth and final stage, there are fewer problems cross referencing to the KNET ground motion catalogue (the two sources both use NIED seismicity parameters), more mechanism parameters are available for each earthquake, and the data are searchable by computer. Where an event is not reported by F-Net, the focal mechanism may be obtained elsewhere (see section 3.1.1).

4.2.1 Classification Categories

Using the fault plane rake and dip to define the focal mechanism (see sections 4.1.2.1 and 4.1.3) and the principles of event types in a subduction zone (see section 2.1), the three main types of subduction zone earthquakes are subdivided into eight classification categories. Table 4.1 identifies the categories

and lists classification criteria. The eight categories may be grouped into the three main event types as follows: categories 1 and 2 are interface events, categories 3, 4, 5, 6, and 7 are inslab events, and category 8 represents crustal events. This grouping is supported by the tectonic principles described in sections 2 and 4.1.3, and in part by the use of hypocentre locations as a secondary classifier, as described in section 4.2.2.

The need for categories termed “ambiguous” arose where the event’s two possible solutions for rake angle do not both lie within the same fault mechanism range. In the case of categories 2, 4, and 7, one possible rake angle would have classified the event as strike slip.

Table 4.1: Classification Categories

			Rake angle	Dip angle	Depth (km)
Interface	1	Low angle thrust	$45 < \delta < 135$ (both angles)	$* < 20, * > 70$	>20
	2	Low angle ambiguous thrust	$45 < \delta < 135$ (one angle)	$* < 20, * > 70$	>20
Inslab	3	High angle thrust	$45 < \delta < 135$ (both angles)	$20 < * < 70$	>20
	4	High angle ambiguous thrust	$45 < \delta < 135$ (one angle)	$20 < * < 70$	>20
	5	Normal	$-135 < \delta < -45$ (both angles)	-	>20
	6	Strike Slip	$-45 < \delta < 45$ or $\delta < -135$ or $\delta > 135$	-	>20
	7	Ambiguous Normal	$-135 < \delta < -45$ (one angle)	-	>20
Crustal	8	Shallow	-	-	<20

To separate events associated with the seismogenic interface or subducting slab from those occurring in the crust on either side of the subduction trench (crustal events) it is necessary to use depth as the classifier. The criteria for identifying category 8, the shallow or crustal events must be approached with caution when

considering the limitations on reported depth parameters for earthquakes (see section 3.3.5). The depth criterion dividing crustal events from interface is taken as equivalent to the top of the interface seismogenic zone. In section 2.3, previous studies are reported (Oleskevich et al, 1999, Satake & Tanioka, 1999) that place the upper limit of the interface zone in Japan at between 8 and 20 km. These values result from seismic surveys, thermal modelling, and seismicity studies. Taking the deeper limit minimises the number of crustal events that will be misplaced into interface or inslab categories.

Due to the unreliable nature of reported hypocentral depth, no maximum depth is imposed for interface events. Although there are physical constraints on depth for interface and inslab events (see sections 2.1 and 4.1.3), in this study focal mechanism is used as the only reliable classification criterion. Approximately 25% of data points from events classified as interface have reported depths greater than that suggested “allowable” by tectonics.

4.2.2 Location Transects

After the primary classification scheme using focal mechanism parameters has been applied, the hypocentres are plotted on transects across 5° latitude ranges. When plotted against the background seismicity as reported in the NIED catalogue, these transects help to confirm the grouping of eight categories into three event types. As shown in figure 4.3 (a), low angle thrust events and low

angle ambiguous thrust events (categories 1 and 2) are both reasonably well defined along the dipping seismogenic zone. Shallow events (category 8 on figure 4.3 (c)) occur to the continental side of the subducting slab, by definition restricted to the upper 20 km. The remaining five categories (high angle thrust and ambiguous thrust, normal, ambiguous normal, and strike slip) have a wider distribution throughout and away from the seismogenic zone (see figure 4.3 (b)). Thus the grouping of categories into three event types is supported, especially when considering the results for low and high angle thrust events. The reminder that the depths used are those given by KNET must weight these comments. The reported errors and considered uncertainties in these values are in some cases very large. The location plots are used only as confirmation and a secondary classifier. They will not change the primary classification using the scheme described in section 4.2.1.

4.2.3 Strike Orientation

Interface earthquakes will have a strike angle oriented parallel to the strike of the subducting plate, whereas inslab events have no such constraint (see section 4.1.3). Figure 4.4 shows the data from Japan classified as interface and inslab according to the scheme described in section 4.2.1. Interface events are constrained within strike angles in the order of 15 - 60 and 210 - 240. This is the same range as the strike of both the Nankai Trough and Japan Trench. As explained in section 4.1.2, there are two possible strike angles for each

mechanism solution. The dip angle for interface events is fixed under the classification scheme, hence the clear radial distribution in the plot. Inslab events show no such constraint in the strike angles, as is expected by theory. This result gives a high degree of confidence in the interface earthquake category. There is still some ambiguity in the inslab classification group.

Fig 4.1: Focal Mechanism Rake Angle

Fig 4.2: Example Focal Mechanism Solutions for Japan Events

1999-10-22 **M** = 5.9 Thrust

1998-06-14 **M** = 5.6 Normal

1997-05-13 **M** = 6.0 Strike slip

1998-10-27 **M** = 5.5 Oblique thrust

1998-05-03 **M** = 7.0 Oblique normal

Dark regions = P wave compressional first motion

(Earthquake Research Institute, Japan)

Fig 4.3 (a): Location of KNET Interface Events

Fig 4.3 (b): Location of KNET Inslab Events

Fig 4.3 (c): Location of KNET Crustal Events

Fig 4.4: Strike and Dip for Inslab and Interface Earthquakes

5 Analysis Methods

5.1 Spectral Analysis Software

For each earthquake, institutional sources (see section 3.1) provide a number of digital records from individual stations, giving the recorded ground motion at fixed time intervals (typically 40 to 200 samples per second). If the data are provided as velocity or displacement, a conversion to acceleration is applied during processing. Response spectra are produced from the instrument corrected acceleration time series by AGRAM, a Fortran software program written within the Carleton University seismology research group and developed using methods described by Boore (1996). The various formats of downloaded data (see section 5.2) necessitated data specific modifications to AGRAM, although the methods of analysis were unchanged. The processing steps of AGRAM are described in sections 5.1.1 and 5.1.2.

5.1.1 Windowing

Data are collected in a time interval that includes the strongest earthquake shaking, but the digital record may not begin at the exact time of earthquake onset. Viewing software is used to identify the time of the main P and S wave arrivals relative to the start of the recorded signal. These times define two “windows” as shown on figure 5.1: the pre-signal noise, from the beginning to 10

seconds before the P wave arrival, and the main body of the earthquake record that follows the S wave arrival. AGRAM calculates the spectral response for this S-window, during which the majority of the earthquake energy is recorded. When manually selecting the S wave arrival it is important that all of the S wave is included in this window. It is desirable to err on the early rather than the late side. When selecting the P wave arrival, choosing a time too early has no effect. If late, there is a possibility of increasing the calculated noise because part of the P wave is included in the initial window.

5.1.2 Data Processing Corrections

AGRAM applies a range of filters and corrections, controlled by the user through a program option file: low pass, high pass and taper filters, baseline trend removal and instrument response corrections. The signal to noise ratio may also be calculated by comparing the spectrum of the noise window, normalised to the same length as the signal window, to the spectrum of the signal window. The instrument response correction requires the station and instrument response functions for each channel in the form of poles and zeros. These are included with the IRIS data files and extracted via RDSEED (see section 5.2.1), or obtained from the CNSN and USGS autodrm services. Instrument response correction is minimal for the KNET data since the frequencies studied are within the flat response range of the instruments. It is only necessary to correct for the constant response level and any offset in the baseline trend.

When calculating the response spectra, AGRAM reports when glitches or jumps are present in the record. Single glitches are removed. Signals with multiple glitches are not included in the database.

5.2 Download Formats

One of the challenges of seismic processing is the multitude of seismic formats, despite a long recognised international effort to make digital data available and compatible. The data sources used result in three categories of download format:

5.2.1 Incorporated Research Institutions for Seismology (IRIS)

IRIS consortium files are obtained in the Standard for Exchange of Earthquake Data (SEED) format. The RDSEED reader program, available through the same download portal, extracts the data into separate station records of ground motion, which may be chosen by the user to be formatted in Seismic Analysis Code (SAC) format. SAC 2000 software is then used to view each station record, “window” the P and S wave arrivals, then make a final conversion to SAC ASCII which can be input into the spectral analysis software being used (AGRAM or a modified version).

5.2.2 CNSN and USGS

CNSN and USGS data obtained via the autodrm service (see section 3.1.2) arrive as compressed files which are opened into separate station and channel records using DECOMPRESS, software obtained from the Geological Survey of Canada. RSEC, a program written in the research group displays the time series for windowing. With the necessary parameter files, these ground motion records are then input into AGRAM.

5.2.3 KNET and Special Event Files

KNET data, and the special event files obtained from COSMOS and the University of Washington records via IRIS are processed by uniquely modified versions of AGRAM taking into account each occurrence of format, and corrections applied or needed. These categories of data come in files processed by the institutions to contain only the event signal and its coda. Windowing is not necessary, and either the complete signal is used or, if too long, only the central portion which will include the strongest ground motion.

Fig 5.1 Acceleration Time Series after "Windowing"

Nisqually **M** = 6.8

Recorded at station WA, 19 km from epicentre

Source: USGS

6 Site Effects

6.1 Effect of Recording Site

Ground motion recorded at a seismograph station is the result of source, path and site effects (see section 1.3). This study is looking to establish gross differences in source and path effects between different types of subduction earthquakes. Any amplification or attenuation attributable to local site conditions should be accounted for prior to comparing the ground motions. Typically stations on hard rock, such as most seismograph stations in the Cascadia region, will record smaller ground motion than stations on soil, which are the norm for KNET stations in Japan. One solution is to group station records into similar soil or rock types. Alternatively, the estimated effect of the general site conditions at each recording station may be removed to directly compare all data. The latter approach is used in this study.

6.2 Site Classification

A simple classification scheme using the letters A through E is now used by most seismologists to describe the site conditions at a recording station. NEHRP (National Earthquake Hazards Reduction Program) has devised a consensus approach to site classification that is based on the average shear wave velocity (V_s) in the top 30 metres of the site (FEMA, 2000). Table 6.1 presents the

NEHRP site classifications. The NEHRP scheme is one of a number of such classifications in use and caution is needed before equating one author's "site type" with another.

Table 6.1: NEHRP Site Classification (FEMA, 2000)

NEHRP Site Classification	Description	V_s (ms^{-1}) (Average over upper 30 m)
A	Hard rock	> 1500
B	Rock	760 - 1500
C	Very dense soil and soft rock	360 - 760
D	Stiff soil	180 - 360
E	Soil	< 180

All stations reporting data from Japan and Cascadia are classified using the NEHRP scheme as presented in table 6.1 (Atkinson and Boore, 2002). Sites on soft rock and soil (NEHRP C, D, and E) dominate Japanese data. In contrast, Cascadia data are mostly from stations on soft rock or harder (see figure 6.1).

6.3 Determining Site Amplification

The most accurate method to determine the amplifications of seismic recording sites is to measure them directly using borehole seismometers. Unfortunately data from this technique are not widely available.

Site amplification can be calculated theoretically using knowledge of local geology and seismic velocity structure. The variation with depth of shear wave velocity (V_s) and density (Φ) must be either known or assumed beneath the recording station. Effective amplification at the surface is then represented by the square root of the ratio of the average seismic impedance ($= V_s \Phi$) over a depth equal to one quarter of the seismic wavelength, to the seismic impedance at the depth of the source. The frequency at which this amplification is appropriate is found by basic wave equations to be $1/[4 \times \text{seismic travel time to the quarter wavelength depth}]$. Hence this method is termed the “quarter wavelength approximation” (Boore & Joyner, 1997).

Recent work has shown that the ratio of horizontal (H) to vertical (V) components in a station record is a reasonable estimate of the site amplification (Lermo & Chavez-Garcia, 1993, Atkinson & Cassidy, 2000, Chen, 2000), both measured and theoretical. These authors report the H/V ratio to be a stable site parameter when averaged over many events, but it is variable between events at one site.

Site amplification may also be derived empirically, using regression of observed data to a predetermined functional form that includes site effects. A correction based on this approach is implemented here since there is insufficient information available to apply any of the other methods to the complete data set. The correction is based on ground motion relationships developed by Atkinson and Boore (2002) by regression of a data set consisting of over 5000 records

from subduction zone earthquakes. The functional form used by Atkinson and Boore (2002) is:

$$\log \text{PSA} = c_1 + c_2 \mathbf{M} + c_3 h + c_4 R - g \log R \\ + c_5 nl / S_C + c_6 nl / S_D + c_7 nl / S_E$$

where:

PSA is 5% damped pseudo-acceleration in cm^{-2} , random horizontal component

$\mathbf{M}$ = moment magnitude with the condition that $\mathbf{M} \leq 8.5$

h = focal depth in km

$R = \sqrt{(D_{\text{fault}}^2 + \Delta^2)}$ with: D_{fault} = closest distance to fault surface, in km

$$\Delta = 0.00724 \cdot 10^{0.507 \mathbf{M}}$$

$S_C = 1$ for NEHRP C soils ($360 < V_s < 760 \text{ ms}^{-1}$), otherwise = 0

$S_D = 1$ for NEHRP D soils ($180 < V_s < 360 \text{ ms}^{-1}$), otherwise = 0

$S_E = 1$ for NEHRP E soils ($V_s < 180 \text{ ms}^{-1}$), otherwise = 0

$g = 10^{(1.2 - 0.18 \mathbf{M})}$ for interface events

$g = 10^{(0.301 - 0.01 \mathbf{M})}$ for inslab events

$nl = 1$ for $\text{PGArx} \leq 100 \text{ cms}^{-2}$ OR frequencies $< 2 \text{ Hz}$

$nl = 1 - (\text{PGArx} - 100)/400$ for $100 < \text{PGArx} < 500 \text{ cms}^{-2}$ ($f \geq 2 \text{ Hz}$)

$nl = 0$ for $\text{PGArx} \geq 500 \text{ cms}^{-2}$ ($f \geq 2 \text{ Hz}$)

PGArx = predicted PGA on rock (NEHRP A/B), in cms^{-2}

The coefficients c_1 to c_7 are listed in table 6.2 and have been determined in the regression for a range of frequencies and for the peak ground acceleration (PGA). The first six terms of the functional form model the expected ground motion for NEHRP site types A and B, which were combined for the purposes of Atkinson and Boore's work. Adding the c_5 , c_6 , or c_7 terms provides the expected observed ground motion, including local amplification effects, at NEHRP site types C, D, or E respectively. In this study, subtracting the appropriate site response term (i.e. either c_5 , c_6 , or c_7) from the observed ground motion at NEHRP site type C, D, or E corrects the data to what would be observed if the site were NEHRP A or B. Data from all site types may then be directly compared with the predominant influence of different site amplifications having been removed.

Table 6.2: Coefficients for Atkinson and Boore (2002) Ground Motion Relation

6.2 (a): Interface Events

Freq	C ₁	C ₂	C ₃	C ₄	C ₅	C ₆	C ₇
0.33	2.301	0.02237	0.00012	0.000	0.10	0.25	0.36
0.5	2.1907	0.07148	0.00224	0.000	0.10	0.25	0.40
1	2.1442	0.1345	0.00521	-0.00110	0.10	0.30	0.55
2.5	2.5249	0.1477	0.00728	-0.00235	0.13	0.37	0.38
5	2.6638	0.12386	0.00884	-0.00280	0.15	0.27	0.25
10	2.7789	0.09841	0.00974	-0.00287	0.15	0.23	0.20
25	2.8753	0.07052	0.01004	-0.00278	0.15	0.20	0.20
PGA	2.991	0.03525	0.00759	-0.00206	0.19	0.24	0.29

6.2 (b): Inslab Events

Freq	C ₁	C ₂	C ₃	C ₄	C ₅	C ₆	C ₇
0.33	-3.70012	1.1169	0.00615	-0.00045	0.10	0.25	0.36
0.5	-2.39234	0.9964	0.00364	-0.00118	0.10	0.25	0.40
1	-1.02133	0.8789	0.00130	-0.00173	0.10	0.30	0.55
2.5	0.005445	0.7727	0.00173	-0.00178	0.13	0.37	0.38
5	0.51589	0.69186	0.00572	-0.00192	0.15	0.27	0.25
10	0.43928	0.66675	0.01080	-0.00219	0.15	0.23	0.20
25	0.50697	0.63273	0.01275	-0.00234	0.15	0.20	0.20
PGA	-0.04713	0.6909	0.01130	-0.00202	0.19	0.24	0.29

6.4 Applying Site Effect Correction

Site amplification factors as described in section 6.3 are removed from all data from Japan and Cascadia. The broad result of removing site effects is to reduce the amplitude of ground motions, and does not dramatically affect the distribution pattern or spread of data. Figure 6.2 shows a small selection of events with different moment magnitudes and event type, before and after the site effect correction is applied. Fig 6.2 (a) presents data from Japan, and 6.2 (b) Cascadia. In Cascadia a majority of sites are classified as NEHRP A / B. Ground motions from these sites are unchanged after removing site effects.

In working with amplification coefficients derived for each station in their study of northeast Japan, Molas and Yamazaki (1995) comment that the values vary within soil type categories. The presence of soft, resonant near-surface layers defines four soil type categories in their study area. They argue that assigning mean coefficients to each soil type poorly represents the amplification at any one station.

Research carried out in California (Rodriguez-Marek et al, 1999) concludes that it is advantageous to subdivide the broad rock/soil type categories such as those in the NEHRP scheme, and place more importance on the overburden depth than the shear wave velocity in order to more accurately described the site amplification.

The NEHRP general site-type classification scheme used in this study may be an over simplification and source of error that could be improved in future studies by applying station specific coefficients.

Fig 6.1: Data Frequency by Site Type

Fig 6.2 (a): Japan Sample Events with Site Effect Correction

Fig 6.2 (b): Cascadia Sample Events with Site Effect Correction

7 Results

7.1 Special Study: Geiyo and Nisqually

7.1.1 Event Parameters

In the early months of 2001, nature provided an interesting opportunity. Two earthquakes occurred, easily identified as seismological twins. First, on Wednesday 28 February 2001 at 10.55 a.m. local time, Seattle experienced its largest earthquake in 40 years. Determined by seismologists to have a moment magnitude = 6.8 and depth of 52 km, it was named “Nisqually” after a local wild life refuge and delta. Nisqually is an inslab event having a normal focal mechanism and being located within the subducting Juan de Fuca plate.

One month later, and an ocean apart, the Geiyo earthquake occurred in southwestern Japan, between the islands of Honshu and Shikoku. Again, a normal faulting inslab event, with the focal mechanism strike in the direction of the subducting Nankai plate, just as Nisqually was aligned with the Juan de Fuca plate. A moment magnitude of 6.8 was determined by F-Net, and all earthquake parameters are remarkably similar to Nisqually (see the summary in table 7.1). Both events were well recorded by regional networks. As model events for a comparison of ground motion between the two regions, one could not ask for

better. (Sources of information from USGS Earthquake Bulletin and ERI Special Event Pages)

Table 7.1 Nisqually and Geiyo: Hypocentre and Mechanism Parameters

	Nisqually	Geiyo
Date/time (UT)	28/02/01, 18.55	24/03/01, 06.28
Moment magnitude	6.8	6.8
Depth	52 km	60 km
Fault mechanism	Normal	Normal
Rake	-89, -90	-56, -133
Dip	18, 72	59, 45
Strike	183, 1	188, 315
Classification	Inslab	Inslab

7.1.2 Data Processing Modifications

Nature delivered two model earthquakes for direct comparison, but the recording networks are not identical. Data from Geiyo are provided by KNET, and are processed using the same methods as all other KNET data, described in sections 5 and 6. The final data set for the Nisqually earthquake includes data from five different sources, each requiring special attention in the data acquisition

and processing. In addition to records from CNSN, USNSN, and IRIS (already discussed in section 5.2.1 and 5.2.2), data from the University of Washington local seismic network were obtained from two sources: direct download of a special event file from the University file transfer site, and a similar designated file from COSMOS. The University of Washington file contains the time series in simple data pairs, time and acceleration for each recording station, corrected for the instrument response. The header, meta data and data structure provided by other institutions are not present, hence the spectral analysis software AGRAM requires modification to accept the alternative input format.

In contrast, the COSMOS file consists of a table giving the ground motion response spectra for four frequencies (0.33, 1, 3.33, and 10 Hz) at each station. Again, the instrument response has been removed, but no details are available as to the spectral analysis routine applied. Where stations appear in the COSMOS and the University of Washington file, the values are not identical, although similarity is considered acceptable for data quality.

Spectral response values obtained from each method are added to the Nisqually data set and the site correction method is applied as detailed in section 6.3.

7.1.3 Comparison of Ground Motion for Geiyo and Nisqually

Figure 7.1 presents the ground motion response as a function of distance for data from Geiyo and Nisqually at frequencies of 1 Hz to 10 Hz. Figure 7.1 (a)

plots the individual data points and 7.1 (b) shows the geometric mean values in 0.3 log unit distance bins. Geiyo data are not used at distances beyond 250 km for the 5 Hz response spectra, or beyond 150 km for 10 Hz. Data from these distances are considered unreliable due to limitations of the KNET instruments. This is explained further in section 7.2.1.

There is good agreement between the amplitudes of ground motion from the two events for the 1 Hz and 2.5 Hz response spectra. At 5 Hz and 10 Hz, the separation between the mean ground motion for the two data sets increases, but is still within the same order of magnitude. The mean values show that attenuation is very similar for both events at 1 Hz. At higher frequencies, attenuation is constant for Geiyo in the distance range 40 - 300 km. At hypocentral distances less than 100 km, attenuation is higher for Geiyo than Nisqually. For distances greater than 200 km attenuation is lower for ground motion from Geiyo, i.e. attenuation of ground motion from Nisqually is increasing with distance.

These observations reflect the behaviour of single events and should not be generalised to the two seismic regions without further study. It may also be biased by potential problems with the distant Geiyo data at high frequencies, or by the soil response. A recent study (Casey, 2002) found that apparent differences in ground motion amplitudes between the two events could be explained by regional differences in site amplification.

The broad conclusion here is that two events, occurring in different geographic regions but with similar source parameters and tectonic setting produce ground motions that follow very similar relationships with distance and frequency. This is important in the context of this study since it supports the idea that ground motion in Japan could be used to supplement and extend existing models of ground motion in Cascadia.

7.2 Japan

7.2.1 Overall Data Trends

The primary purpose of this study is to investigate the variation of ground motion (pseudo acceleration) with moment magnitude and hypocentral distance. In general descriptive terms, it would be expected that ground motion increases with moment magnitude, but has an inverse relationship with hypocentral distance. Moving further away from the source would result in smaller ground motions. Figure 7.2 shows the overall trends for all records obtained from earthquakes in Japan.

The ground motion levels at 1 Hz (figure 7.2 (a)) and 2.5 Hz (figure 7.2 (b)) display the relationships with moment magnitude and hypocentral distance as described above, with a range of pseudo acceleration from 0.01 cm/s^2 to 1000 cm/s^2 (approximately 1 g). The data appear to exhibit a normal distribution, being

denser in the middle of the range, and more sparse at both maximum and minimum values. At 2.5 Hz, the absolute value of ground motion is larger by about a factor of 4 than that for 1 Hz.

At the higher frequencies, 5 Hz (figure 7.2 (c)) and 10 Hz (figure 7.2 (d)) the data show an apparent low-amplitude cut-off in the order of 1 cm/s^2 ($\log \text{PSA} = 0$) that is indicative of instrument limitations. Instead of the normal distribution visible at lower frequencies, the data is truncated quite abruptly at this cut-off value, which does not change with magnitude or distance. This suggests that the instrument's minimum sensitivity is frequency dependent, and capabilities are reduced at higher frequencies.

The study plans to use statistical analysis to accomplish comparison between large data sets. Therefore it is important to establish ranges of magnitude, hypocentral distance, and frequency within which the data are reliable. Figure 7.3 expands the plots showing ground motion scaling with magnitude and distance at hypocentral distance intervals of 1.3 log units. Maximum distance values, within which the data distribution is valid, for each minimum magnitude and frequency are established from these graphs as being where the slope changes between upward trending (expected) and horizontal (the cut-off value). Table 7.2 summarises the results giving the maximum hypocentral distance (in kilometres) in order to ensure reliable data at each frequency.

Table 7.2 (a) Maximum Distance for Reliable KNET Data (log R)

Frequency	1 Hz	2.5 Hz	5 Hz	10 Hz
Minimum M				
7	3	3	2.4	2.1
6.5	3	3	2.4	2.1
6	3	3	2.1	2.1
5.5	3	2.7	2.1	1.8
5	2.7	2.7	1.8	1.8
4.5	2.4	2.4	1.5	-

Table 7.2 (b) Maximum Distance for Reliable KNET Data (km)

Frequency	1 Hz	2.5 Hz	5 Hz	10 Hz
Minimum M				
7	1000	1000	250	150
6.5	1000	1000	250	150
6	1000	1000	150	150
5.5	1000	500	150	65
5	500	500	65	65
4.5	250	250	35	-

7.2.2 Event Type Comparisons

The ground motion data set is classified into three main event types: crustal, inslab, and interface, including all sub classes of event types and ambiguous

focal mechanisms. The comparison of ground motion attenuation with distance between event classes is shown in figures 7.4, 7.5, and 7.6. Each figure divides the data into 0.5 unit moment magnitude intervals, and includes graphs for 1 Hz (figure (a)), 2.5 Hz (figure (b)), 5 Hz (figure (c)), and 10 Hz (figure (d)). Data beyond the hypocentral distance limits given in table 7.2 have been deleted.

The immediate observation is of significant overlap between the three data sets. One data set easily obscures another in this style of graph. All three classification groups show similar attenuation of ground motion with distance from the hypocentre. At the extremes of the range for both distance and PSA there are irregularities: points lying far from the bulk of the data concentration. This is an expected result in observational seismology. As the frequency increases from 1 Hz to 10 Hz, the amplitude of ground motion also increases, by approximately one power of ten.

There are some apparent distinctions between the classification groups, and the visible differences are more prominent at higher frequencies. Interface events are sometimes producing lower ground motion than inslab; this is most pronounced in the magnitude interval $6.5 < \mathbf{M} < 7.0$, especially at 5 Hz (figure 7.4 (c)). Ground motion amplitudes from crustal events are generally lower than from inslab at high frequencies (figures 7.6 (c) and 7.6 (d)). The comparison between interface and crustal in figure 7.5 is puzzling. At low magnitudes, crustal events have lower ground motion amplitudes, whereas above $\mathbf{M} = 6.5$, the reverse is true.

In seeking a more distinct pattern or one that is consistent across all magnitude and frequency ranges, a series of comparisons is made for subdivisions of the event types using the eight categories as defined by the classification scheme explained in section 4.2.1. The ground motions are shown in figures 7.7, 7.8, 7.9, and 7.10 for each magnitude range and at three frequencies: 1, 5 and 10 Hz. Interface events are those with clear low angle thrust mechanisms (category 1), excluding from the data set ambiguous thrust events (category 2). Inslab events are subdivided into those with either normal (category 5) or strike slip (category 6) mechanisms, excluding those with ambiguous normal (category 7) or steep angle thrust (category 3 and 4) mechanisms.

Of interest in figure 7.7 (a) is the $6.5 < M < 7.0$ magnitude interval, where amplitudes for interface events are significantly lower than inslab amplitudes. In the same figure this relationship is reversed in the magnitude range $5.5 < M < 6.0$, although the difference in ground motion amplitude is less obvious.

The comparisons between event classes seen in figures 7.7, 7.8, 7.9, and 7.10, which include just those events which are unambiguously classified, are in agreement with those from the complete data set in figures 7.4, 7.5, and 7.6. Subdivision of the three main classification groups does not reveal any more details of the ground motion relationships. Tentatively, it is concluded that both interface and crustal events produce ground motion lower than inslab events of the same magnitude, with anomalies in individual event type and magnitude

comparisons. It is difficult to ascertain from this data set whether these are part of a more complex pattern of ground motion, or simply reflect the underlying variation in the behaviour of seismic events. In light of this ambiguity, the continuing analysis applies only to the main classification scheme, and disregards the sub categories of focal mechanisms within each event type.

7.2.3 Summary Statistical Analysis

The observational data in section 7.2.2 provide an overview but in some instances are difficult to interpret. Summary statistical analysis simplifies the data plots and may offer further insight into the ground motion relations.

The data are subdivided into bins for statistical analyses by hypocentral distance (0.3 log unit ranges) and moment magnitude (half unit range). The mean PSA and the standard deviation of observations within each subdivision are calculated for each of the three primary event classification groups. Figure 7.11 plots the results, with error bars of $\pm$ one standard deviation of the observations, at frequencies 1 Hz (figure (a)), 2.5 Hz (figure (b)), 5 Hz (figure (c)), and 10 Hz (figure (d)). The data are plotted at $\pm$ 0.05 log units from the midpoint of the distance interval, to make the error bars visible. Adjacent data for each event type are calculated across the same distance interval. The standard error of the mean value (standard deviation of N observations / $\sqrt{N}$) is plotted on the graphs when it is greater than 0.1 log PSA (factor of 1.25). In most instances, the error of

the mean is significantly less than this value and may be considered adequately represented by the vertical dimensions of the marker symbols used on the graphs.

The graphs have been edited by stringently applying the reliability limits presented in table 7.2, resulting in the mostly empty plots for 10 Hz. There is only one event classified as crustal in the magnitude interval $6.5 < \mathbf{M} < 7.0$, so the mean values have not been included.

The following statements and exceptions may be made from reviewing the distribution of the mean ground motion data:

The range of ground motion significantly overlaps for interface, inslab, and crustal events, considering the standard deviation error bars that reflect random variability. The exception is the magnitude interval $6.5 < \mathbf{M} < 7$ where the difference in ground motion between interface and inslab events increases with frequency, with inslab ground motions being a factor of 10 larger than interface at 10 Hz.

The mean ground motion values are lower for interface events than inslab events. The exception is in the magnitude interval $5.5 < \mathbf{M} < 6.0$ at a frequency of 2.5 Hz and 5 Hz (figures 7.11 (b) and (c)). The values for interface and inslab events are very close and marginally higher for interface events at close distances. This reversal is also visible in figure 7.7, which considers only a

subset of the inslab data. At 1 Hz for the magnitude interval $6.5 < \mathbf{M} < 7$, the trend in ground motion amplitudes cross due to differing attenuation (see next paragraph), being higher for interface events at large distances.

The rate of ground motion attenuation differs for each class of event, especially at 1 Hz. Ground motion from interface events attenuates less rapidly than that from inslab events. This causes the situation in the magnitude interval $6.5 < \mathbf{M} < 7.0$ where at distances below 100 km, inslab events produce larger ground motion, whereas further than 100 km, inslab events are at lower amplitude levels than interface events. The attenuation of crustal events is harder to establish but does appear to be very rapid at close distances, lessening as distance increases.

There is no consistent pattern in the comparison of crustal ground motion amplitudes relative to those from interface or inslab events, perhaps due to the contrasting attenuation patterns described in the preceding paragraph. In various magnitude distance combinations, crustal events plot both higher and lower than interface or inslab events.

7.2.4 Investigation of Interface and Inslab Events for $6.5 < \mathbf{M} < 7$

The statistical analysis reveals a strong contrast between the ground motion from interface and inslab events in the magnitude interval $6.5 < \mathbf{M} < 7.0$, not observed in other magnitude ranges. The possibility of inter-event variability being responsible for the elevated mean ground motion for inslab events and lowered

interface event ground motion in this magnitude interval is investigated in figures 7.12 and 7.13.

This magnitude range includes data from four inslab events, including Geiyo (see section 7.1). Figure 7.12 (a) displays the individual data points for the inslab events, with the mean ground motion values for inslab and interface events at 5 Hz frequency response. The hypocentral distance has been extended past the reliability limit of 250 km ($\log R = 2.4$) (see table 7.2) to allow the inclusion of data from the 22/12/96 event. It clearly demonstrates how the data quality deteriorates past the hypocentral distance maximum limit as determined in section 7.2.1. This region should not be considered in any conclusion. All the inslab events produce ground motion significantly higher than the mean from interface events. At hypocentral distances below 300 km ($\log R = 2.5$), the inter-event variability between the inslab earthquakes is low compared to the contrast with interface events.

Further confirmation is provided by figure 7.12 (b). The statistical analysis was repeated, first excluding Geiyo and then using only Geiyo. The mean values of ground motion for inslab events fairly represent the motion from individual events.

The same technique was applied to interface events in this magnitude interval. In figure 7.13 (a), the interface event of 19/10/96 appears to be closer to the mean

values for inslab events than those of interface. However, combining this with the results of repeating the statistical analysis for each event, displayed in figure 7.13 (b) it is more correct to say that this event has a greater variability of ground motion values. The mean ground motion data for the event of 19/10/96 are higher than the other interface events, but closer to the interface mean values than those for inslab.

The conclusion is that the pattern of observed ground motion displayed in figure 7.11 for the magnitude interval $6.5 < \mathbf{M} < 7.0$ accurately represents a contrast in the behaviour of inslab and interface events. In this magnitude range, inslab events produce greater ground motion than interface at similar hypocentral distances.

7.2.5 Ground Motion Scaling with Moment Magnitude

The results in section 7.2.4 raise the interesting question of whether the ground motion from interface events in the magnitude interval $6.5 < \mathbf{M} < 7.0$ is unusually low, relative to interface events at other magnitudes, or whether that from inslab events is unusually high. Figure 7.14 focuses the analysis by using mean ground motion data for the three event types, displayed at four frequencies for three hypocentral distance intervals. The mean data for the magnitude interval under question is plotted at the midpoint of the range, for example $\mathbf{M} = 6.75$.

The scaling of ground motion with magnitude for interface events does not follow a simple trend. At hypocentral distances less than 100 km (figures 7.14 (a) and (b)), the interface ground motion in the interval $5.0 < M < 6.5$ increases very slowly with magnitude. Between $M = 6.5$ and $M = 7.0$, the increase in mean value is similar to that for inslab events within each 0.5 unit magnitude interval, but the absolute values for interface events are now significantly lower than that for inslab. This results in the large differences in ground motion for this magnitude interval, seen here and on figure 7.11 (b), (c), and (d).

An unusual aspect of figure 7.11 (b) and (c) is the reversed relationship between interface and inslab events in the magnitude interval $5.5 < M < 6.0$ at close distances. Reviewing the magnitude scaling pattern in figure 7.14 (a) indicates that at $M = 5.75$ and 6.25 , the ground motion from inslab events at 2.5 Hz and 5 Hz is low relative to the general trend. This would result in inslab ground motion being lower than interface in this magnitude interval, and the reversal seen on figure 7.11 may be considered anomalous rather than indicative of inslab events.

The overall pattern revealed by figure 7.14 is that ground motions from interface earthquakes increase with M more slowly than that from inslab or crustal events. As already observed in section 7.2.3, ground motion from interface events attenuates more slowly with distance than that from inslab. These two properties result in the varying relationships observed between ground motion amplitudes for the event types in different magnitude and distance intervals.

These results are important when extrapolating ground motion relations determined for intermediate magnitude events to large magnitudes, a primary motivation for this study. It is a fundamental point to consider in any extrapolation that ground motions from interface and inslab events are dissimilar in how they scale with magnitude.

7.3 Cascadia

7.3.1 Overall Data Trends

The data collected from Cascadia includes events with moment magnitude between 3.8 and 6.8, and a range of hypocentral distance from 20 to 2000 km. Figure 7.15 shows the ground motion as it varies with event magnitude, at frequencies of 1 Hz, 2.5 Hz, 5 Hz, and 10 Hz. Hypocentral distances below 500 km are plotted in figure 7.15 (a), and those above 500 km in figure 7.15 (b). As would be expected, there is a clear upward trend to each plot showing increasing ground motion with increasing magnitude. These relationships are continued even at the extremes of the data range. Contrast this to the overall data pattern for Japan (figures 7.2 and 7.3) where limits of instrument sensitivity are indicated.

One observation pertinent to the following analysis is that in the Cascadia ground motion data set, crustal events predominate at low magnitudes and larger distances. There are no recorded interface events in Cascadia (see section 2.2).

7.3.2 Event Type Comparisons

Although the Cascadia data set is much smaller than that from Japan, ground motion attenuation must still be investigated for events of similar magnitudes. Figure 7.16 presents the data for 2.5 Hz (figure 7.16 (a)) and 5 Hz (figure 7.16 (b)) in the same magnitude intervals as used for Japan.

The ground motion from Cascadia data presented in figure 7.16 exhibit a consistent rate of attenuation with distance for both crustal and inslab events between $M = 5$ and $M = 6.5$. Above $M = 6.5$, which represents only data from the Nisqually inslab event, the attenuation is still well defined, but less steep than in lower magnitude intervals. Nisqually data are recorded at closer hypocentral distances than the remainder of the Cascadia data.

There is limited overlap of crustal and inslab event in one magnitude interval so the variation between event types within the overall trend for Cascadia is not determined.

7.3.3 Comparison between Cascadia and Japan

To compare the earthquake ground motion in the two regions, Cascadia data have been statistically analysed using the same parameters as described in section 7.2.3 for Japan. The mean, error bars of $\pm$ one standard deviation, and the standard error of the mean are again calculated in bins of hypocentral

distance (0.3 log unit ranges) and moment magnitude (half unit range). The smaller data set for Cascadia means that not every statistical bin is represented.

As explained in section 7.2.1 and table 7.2, Japan ground motion observations are considered reliable to a frequency and magnitude dependent maximum limit of hypocentral distance, ranging from 35 to 1000 km ($\log R = 3$). Data from Cascadia, with the exception of Nisqually, are predominately at distances greater than 300 km ($\log R = 2.5$), resulting in limited data overlap. Japanese ground motion at 2.5 Hz frequency has the best combination of reliability over a large distance range plus observable differences between event classes. Detailed analysis being carried out elsewhere of data from Geiyo and Nisqually (Casey, 2002) anticipates that at 2.5 Hz the regional variation in amplification factors within each soil type is at a minimum. Using the ground motion response at a frequency of 2.5 Hz for both regions optimises the validity of the comparison.

7.3.3.1 Inslab Events

Ground motion from inslab events for both regions are compared on figure 7.17 (a). In the magnitude range $5.0 < \mathbf{M} < 6.0$ the Cascadia amplitudes are lower than those for Japan, but at distances less than 100 km there is the suggestion that the ground motions from the two regions may converge. At these intermediate magnitudes, Cascadia ground motion attenuates more rapidly than that from Japan. As magnitude increases, the attenuation rate for Cascadia data appears to decrease until in the magnitude interval $6.5 < \mathbf{M} < 7.0$ ground motion

amplitudes are similar at all recorded distances for both regions. Conversely, the attenuation of Japanese data is consistent through all magnitude intervals. For large magnitudes and large distances, e.g. $M > 6.0$ and $R > 400$ km, the “levelling” of the Japanese mean data may be due to approaching the reliability limits rather than indicating ground motion variations.

The $6.5 < M < 7.0$ magnitude interval shown on the final plot in figure 7.17 (a) is the only range where the mean ground motion values coincide in both level and attenuation for Cascadia and Japan. For Cascadia this represents the Nisqually ($M = 6.8$) earthquake. In section 7.1, Nisqually has already been reported to produce ground motions of similar amplitudes to Geiyo, an inslab earthquake occurring in Japan. The ground motion data from Nisqually are unique for Cascadia. Recorded at close hypocentral distances and the only event at this magnitude ($M = 6.8$), the ground motion initially appears significantly higher than other Cascadia data.

Was the Nisqually event a typical Cascadia earthquake? Confirmation of this is limited by the lack of events with this magnitude in the Cascadia region. One route of investigating the behaviour of the Nisqually event is to revisit the pattern for the scaling of ground motion with magnitude, as was carried out in section 7.2.5 for Japan. Figure 7.15 (a) and (b) present the Cascadia ground motion observations varying with moment magnitude, in two intervals of hypocentral distance. Data from Nisqually ($M = 6.8$) is only represented in the closer

hypocentral distance range ($R < 500$ km) and observations follow the general trend of ground motion values. Using the mean ground motion statistics in figure 7.18 confirms that the values from the Nisqually event are as expected from the slope of PSA increase with magnitude and not anomalous when compared to other observations in Cascadia.

Figure 7.18 presents the same statistical data now showing the increase of ground motion with magnitude for three hypocentral distance intervals. Closer inspection of the inslab data (left column) reveals that in Cascadia, the rate of ground motion scaling with magnitude decreases as hypocentral distance increases, i.e. the slope of the Cascadia trend line is less at larger distances. In contrast, the Japanese data maintains a very similar increase of ground motion with magnitude at all three distance intervals, even when $R = 1000$ km, which is reaching the limits of data reliability. Figure 7.18 also demonstrates the different rate of scaling with magnitude in Japan and Cascadia. Ground motions in Cascadia from intermediate and low magnitude earthquakes are smaller than from similar magnitude events in Japan.

7.3.3.2 *Crustal Events*

Observations come from several crustal events in Cascadia in the data set, however at limited ranges of hypocentral distance. On figure 7.17 (b) the mean values for Japan and Cascadia have minimal overlap in distance intervals. As is concluded for inslab events, the Japanese data is considered to lose some

integrity as hypocentral distance increases past $R = 400$ km. When comparing the Cascadia data at large distances with Japanese data at small distances, the general trend is very similar. It is possible to imagine a single relationship that would describe all the mean values, except the Japanese data at large distances. The attenuation or slope of this line would be constant as magnitude increases.

An alternative description is obtained by considering the distance interval where data exists from both regions, $250 < R < 500$, plotted at the midpoint $R = 375$ km on figure 7.17 (b). Here the absolute value of ground motion in Japan is significantly higher than that for Cascadia in all observed magnitude ranges. The limited overlap in hypocentral distance makes it difficult to compare the ground motion scaling with magnitude, as displayed on figure 7.18. The only plot including crustal events (right column) for both regions represents the distance interval where Japan ground motion may be unreliable, but is greater than that for Cascadia.

It is not possible to conclude from the available data which interpretation should be given more weight: that ground motion from crustal events is similar in both regions or that it is greater in Japan than Cascadia.

7.3.3.3 Ground Motion Relationships

The results described in section 7.3.3.1 and 7.3.3.2 indicate contrasts between Japan and Cascadia in the pattern of ground motion observations for inslab and

crustal earthquakes. Comparing the observed data from both Japan and Cascadia to sample ground motion relationships developed elsewhere will assist in determining the relevance of these contrasts, and validating the use of global ground motion relationships in each region. It is important to use relationships providing spectral response values in contrast to those dealing only with peak ground motion. It is also important to distinguish between different types of subduction earthquakes.

The work of Atkinson and Boore (2002) is described in section 6.3 where it is used to calculate the site response corrections. Using the full relationship models the spectral response for inslab events, and is shown on the left column of figure 7.19 as AB2002.

Youngs, Chou, Silva and Humphrey (1997) developed spectral ground motion relations for subduction zones that are frequently used by other authors. Their relationship for inslab events is included in figure 7.19 as YCSH1997. The classification scheme was based largely on depth instead of the focal mechanisms used in this study. This may result in inslab events shallower than 50 km being classified as interface, which should not affect the comparison here. The functional form of the YCSH1997 relationship is:

$$\ln(\text{PSA}) = 0.2418 + 1.414 \mathbf{M} + c_1 + c_2 (10 - \mathbf{M})^3 + \\ c_3 \ln (r + 1.7818 e^{0.554 \mathbf{M}}) + 0.00607 h + 0.3846 Z$$

where:

$\ln(\text{PSA})$ is acceleration in units of g

r = closest distance to rupture (km)

h = hypocentral depth (km)

$Z = 0$ for interface events, 1 for inslab events

c_1 to c_3 are constants determined at specific frequencies (-0.4, -0.0048, and -2.36 respectively for rock sites at 2.5 Hz)

For both AB2002 and YCSH1997, the input parameters to produce the relationship shown on figure 7.19 are hypocentral depth of 40 km, equating nearest distance to fault as the hypocentral distance, and using a magnitude value at the midpoint of the statistical analysis magnitude bins, e.g. $M = 5.25$ for the interval $5.0 < M < 5.5$.

Atkinson and Silva (2000) provide an equation (AS2000) to predict earthquake response spectra for shallow crustal events in California. The functional form for soft rock type California sites (NEHRP C) is given by:

$$\log \text{PSA} = c_1 + c_2 (M - 6) + c_3 (M - 6)^2 - \log R - c_4 R$$

where:

c_1 to c_4 are frequency dependent constants (3.6576, 0.476, -0.0669, 0.0019 respectively, interpolated at 2.5 Hz from AS2000 values at 2 Hz and 3.2 Hz)

$$R = \sqrt{d^2 + h^2}$$

$$\log h = -0.5 + 0.5 M$$

d = closest distance to the fault, approximated here by the hypocentral distance

This simple equation was determined from simulations using a generic stochastic method with region specific parameters to predict ground motions. The simulations were shown to agree well with data from 43 California events. These events of moment magnitude 5 to 7.5, at distances from 5 to 150 km are described in Atkinson and Silva (1997) and would all be considered crustal events in the classification scheme used in this study. This relation is shown as AS2000 on the right column of figure 7.19. The functional form is calculated using magnitude values in the midpoint of the statistical analysis bins, and has been corrected to NEHRP site type A/B using the Atkinson & Boore (2002) coefficients described in section 6.3.

Figure 7.19 superimposes the three ground motion relations over the mean data for inslab and crustal events from Japan and Cascadia. The left column represents inslab subduction earthquakes. At large magnitudes ($6.5 < M < 7.0$) there is good agreement between both models and the observed data. Ground motion data from Japan in this magnitude interval are more closely defined by YCSH1997, and would be slightly under-predicted at distances greater than 300 km by AB2002. The AB2002 relation more closely defines Cascadia data,

including the increase of attenuation with distance seen in this magnitude interval.

For the magnitude interval $5.0 < M < 6.0$, there is a clear distinction in which model best represents each region. YCSH1997 shows a good fit to Japanese data, with a slight under prediction of observations at distances greater than 400 km and over prediction when $R < 100$ km. Neither the attenuation nor amplitude of ground motion in Cascadia is adequately represented by YCSH1997.

Conversely, AB2002 well describes the Cascadia data in this magnitude interval but significantly under predicts ground motion in Japan.

When the Atkinson and Silva relationship (AS2000) is compared to crustal data from Japan and Cascadia on the right column of figure 7.19, the overall shape adequately describes the observed ground motion attenuation in both regions. The relationship is very similar to the amplitude of the mean ground motions calculated from Japanese crustal events, with a slight over-prediction. However, it greatly over-predicts ground motions observed in Cascadia for crustal events of magnitudes less than 6.5, which are at larger hypocentral distances than Japan data. Atkinson and Silva (2000) report that when using the relationship in California ground motions for smaller and distant events were over predicted. They explain this as arising from the simple relationship being biased towards fitting data at large magnitudes and near distances, which have the greatest seismic hazard contribution.

Fig 7.1 (a): Geiyo $M=6.8$ and Nisqually $M=6.8$
Ground Motion with Distance

Fig 7.1 (b): Geiyo $M=6.8$ and Nisqually $M=6.8$
Mean Ground Motion with Distance

Fig 7.2 (a) i : Japan Ground Motion Scaling with Magnitude at 1 Hz

Fig 7.2 (a) ii : Japan Ground Motion Scaling with Distance at 1 Hz

Fig 7.2 (b) i : Japan Ground Motion Scaling with Magnitude at 2.5 Hz

Fig 7.2 (b) ii : Japan Ground Motion Scaling with Distance at 2.5 Hz

Fig 7.2 (c) i : Japan Ground Motion Scaling with Magnitude at 5 Hz

Fig 7.2 (c) ii : Japan Ground Motion Scaling with Distance at 5 Hz

Fig 7.2 (d) i : Japan Ground Motion Scaling with Magnitude at 10 Hz

Fig 7.2 (d) ii : Japan Ground Motion Scaling with Distance at 10 Hz

Fig 7.3 (a): Japan Ground Motion Scaling with Magnitude at 1 Hz

Fig 7.3 (b): Japan Ground Motion Scaling with Magnitude at 2.5 Hz

Fig 7.3 (c): Japan Ground Motion Scaling with Magnitude at 5 Hz

Fig 7.3 (d): Japan Ground Motion Scaling with Magnitude at 10 Hz

Fig 7.4 (a): Ground Motion for Japan
Interface and Inslab Events at 1 Hz

Fig 7.4 (b): Ground Motion for Japan
Interface and Inslab Events at 2.5 Hz

Fig 7.4 (c): Ground Motion for Japan
Interface and Inslab Events at 5 Hz

Fig 7.4 (d): Ground Motion for Japan
Interface and Inslab Events at 10 Hz

Fig 7.5 (a): Ground Motion for Japan
Interface and Crustal Events at 1 Hz

Fig 7.5 (b): Ground Motion for Japan
Interface and Crustal Events at 2.5 Hz

Fig 7.5 (c): Ground Motion for Japan
Interface and Crustal Events at 5 Hz

Fig 7.5 (d): Ground Motion for Japan
Interface and Crustal Events at 10 Hz

Fig 7.6 (a): Ground Motion for Japan
Inslab and Crustal Events at 1 Hz

Fig 7.6 (b): Ground Motion for Japan
Inslab and Crustal Events at 2.5 Hz

Fig 7.6 (c): Ground Motion for Japan
Inslab and Crustal Events at 5 Hz

Fig 7.6 (d): Ground Motion for Japan
Inslab and Crustal Events at 10 Hz

Fig 7.7 (a): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Normal Mechanism Events at 1 Hz

Fig 7.7 (b): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Normal Mechanism Events at 5 Hz

Fig 7.7 (c): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Normal Mechanism Events at 10 Hz

Fig 7.8 (a): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Strike Slip Mechanism Events at 1 Hz

Fig 7.8 (b): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Strike Slip Mechanism Events at 5 Hz

Fig 7.8 (c): Ground Motion for Japan
Interface: Low Angle Thrust and Inslab: Strike Slip Mechanism Events at 10 Hz

Fig 7.9 (a): Ground Motion for Japan
Interface: Low Angle Thrust and Crustal Events at 1 Hz

Fig 7.9 (b): Ground Motion for Japan
Interface: Low Angle Thrust and Crustal Events at 5 Hz

Fig 7.9 (c): Ground Motion for Japan
Interface: Low Angle Thrust and Crustal Events at 10 Hz

Fig 7.10 (a): Ground Motion for Japan
 Inslab: Normal and Crustal Events at 1 Hz

Fig 7.10 (b): Ground Motion for Japan
 Inslab: Normal and Crustal Events at 5 Hz

Fig 7.10 (c): Ground Motion for Japan
 Inslab: Normal and Crustal Events at 10 Hz

Fig 7.11 (a): Japan Mean Log PSA with Hypocentral Distance, Frequency 1 Hz

Fig 7.11 (b): Japan Mean Log PSA with Hypocentral Distance, Frequency 2.5 Hz

Fig 7.11 (c): Japan Mean Log PSA with Hypocentral Distance, Frequency 5 Hz

Fig 7.11 (d): Japan Mean Log PSA with Hypocentral Distance, Frequency 10 Hz

Fig 7.12 (a): Japan Inslab Events $6.5 < M < 7$
Ground Motion with Distance at 5 Hz

Fig 7.12 (b): Japan Inslab Events $6.5 < M < 7.0$
Mean Ground Motion with Distance at Frequency 5 Hz

Fig 7.13 (a); Japan Interface Events $6.5 < M < 7$
Ground Motion with Distance at 5 Hz

Fig 7.13 (b): Japan Interface Events $6.5 < M < 7$
Mean Ground Motion with Distance at 5 Hz

Fig 7.14 (a): Japan Ground Motion Scaling with Magnitude
Hypocentral Distance $R = 35$ km

Fig 7.14 (b): Mean Ground Motion Scaling with Magnitude
Hypocentral Distance $R = 65$ km

Fig 7.14 (c): Japan Ground Motion Scaling with Magnitude
Hypocentral Distance $R = 150$ km

Fig 7.15 (a): Cascadia Ground Motion Scaling with Magnitude
Hypocentral Distance $R < 500$ km

Fig 7.15 (b): Cascadia Ground Motion Scaling with Magnitude
Hypocentral Distance $R > 500$ km

Fig 7.16 (a): Cascadia ground motion with event type at 2.5 Hz

Fig 7.16 (b): Cascadia ground motion with event type at 5 Hz

Fig 7.17 (a): Comparison of Ground Motion for Inslab Events at 2.5 Hz

Fig 7.17 (b): Comparison of Ground Motion for Crustal Events at 2.5 Hz

Fig 7.18 Comparison of ground motion scaling with
Moment magnitude at 2.5 Hz

Fig 7.19 Comparison with ground motion relations at 2.5 Hz

8 Conclusions

8.1 Summary

8.1.1 Japan: Ground Motion Amplitude

Response spectra of ground motion observed in Japan show differences between the three types of subduction earthquakes. There is significant overlap between the three data sets, but the mean ground motion amplitudes from interface and crustal earthquakes are smaller than from inslab events of the same magnitude. The three types trend towards similar ground motion values at large distances and small magnitudes. The differences are greatest at larger magnitudes and higher frequencies (see figures 7.11 (a), (b), (c), and (d)). The largest magnitudes for comparison of inslab and interface earthquakes are in the interval $6.5 < M < 7.0$ and include the Geiyo event. In this interval, the difference in ground motion amplitude is significant and inter event variability has been deemed not responsible for the contrast.

8.1.2 Japan: Ground Motion Attenuation

In Japan, the rate of attenuation of mean ground motion with hypocentral distance is constant within the distance range ($10 < R < 1000$ km) and magnitudes ($4.5 < M < 7.0$) observed, and is greater for inslab earthquakes than

interface. Ground motion from crustal earthquakes attenuates very rapidly at small hypocentral distances, then more gradually as distance increases (see figures 7.11 (a), (b), (c), and (d)).

8.1.3 Japan: Ground Motion Scaling with Magnitude

The mean ground motion from interface earthquakes in Japan appears to scale more slowly with moment magnitude than that for inslab events. Scaling comparisons of interface or inslab to crustal earthquakes are ambiguous (see figures 7.14 (a), (b), and (c)).

8.1.4 Comparing Cascadia and Japan

For data from Cascadia the response spectra amplitude levels from crustal and inslab earthquakes are very similar in the limited overlapping distance range (see figure 7.16 (a) and (b)). There are no interface events in the Cascadia data set for comparison.

When comparing ground motions observed in Japan and Cascadia, there is very little overlap in magnitude and hypocentral distance ranges. At small magnitudes and hypocentral distances for inslab events, the ground motion amplitudes may trend to similar values. When comparing inslab events of $M < 6.0$, attenuation is more rapid in Cascadia than Japan so at most distances under study, ground motion amplitudes in Cascadia are lower. As moment magnitude increases,

ground motion attenuates more slowly in Cascadia whereas attenuation is constant with magnitude in Japan. For larger events in the magnitude range $6.5 < M < 7.0$, which includes Geiyo and Nisqually, ground motion amplitudes are similar at all distances recorded (see figures 7.17 (a) and (b)).

The rate of attenuation for ground motion from inslab earthquakes in Cascadia increases with greater hypocentral distance, in contrast to Japan where the rate is unchanged with distance (see figures 7.17 (a) and (b)).

Attenuation of crustal earthquakes in Cascadia is not determined from this data set.

The increase of ground motion with moment magnitude is more rapid in Cascadia than Japan, for both crustal and inslab earthquakes (see figure 7.18).

8.1.5 Application of Ground Motion Relationships

The Atkinson and Silva (2000) ground motion relation quite satisfactorily predicts the amplitudes from crustal earthquakes in Japan but overestimates distant (> 200 km) observations in Cascadia (see figure 7.19). However the relation was derived from data recorded in California at distances less than 200 km so definite conclusions cannot be made.

For large magnitude ($6.5 < M < 7$) inslab earthquakes, both the Atkinson and Boore (2002) and the Youngs et al (1997) ground motion relations well describe the ground motion observed in Japan and Cascadia. This is encouraging for validating the use of globally derived ground motion models in specific regions. At smaller magnitudes ($M < 6.5$) the observed ground motions in the two regions and the two models increasingly diverge with greater hypocentral distance. The Atkinson and Boore (2002) relation stays closely aligned with the Cascadia data, whereas the Youngs et al (1997) continues to match the Japan data.

8.2 Limitations and Improvements

Classification of earthquakes into the three types of subduction events is principally determined with information from automatically calculated focal mechanism solutions. These solutions are often recalculated, both automatically and through specific seismological study. Mechanism parameters reported in the institutional sources used change as work is updated. The classification scheme itself is based upon generalised criteria applied to a large data set. Under these two conditions, misclassification of some earthquakes is unavoidable. A more reliable classification may be achieved by using only data from earthquakes that have been uniquely located within the subduction zone and whose focal mechanisms have been confirmed.

Data used in this study have been collected from stations located on a wide variety of soil types. The different amplification effects have been removed using generic responses for the five NEHRP categories. As discussed in section 6.4, this may insufficiently model the range of site amplifications. Station specific amplification corrections would increase the confidence that this effect has been completely removed.

This study revealed frequency dependent reliability of the KNET data. The magnitude and distance reliability ranges determined here might inadequately represent the limitations of KNET data. It is possible that the mean ground motions for small magnitude events at large distances are incorrectly high. Lower values would reduce the divergence of the Japan ground motions from the amplitudes of Cascadia ground motions. This instrument response issue warrants further study to determine if the regional differences portrayed here are actual or due to data limitations.

8.3 Application of Results

This study concludes that in Japan differences exist between ground motions from the three types of subduction earthquakes. Interface earthquakes have been shown to attenuate and scale with magnitude differently to inslab events. It is reasonable to assume that ground motions from a great thrust earthquake on the subduction interface in Cascadia would not follow the same pattern as from

inslab events in the same region. It has also been shown that regional differences exist between Japan and Cascadia in the attenuation and magnitude scaling of inslab events, at least at smaller magnitudes. To conclude that any differences between interface and inslab events observed in Japan may be carried over to Cascadia would be premature based on the observed data. However it may be argued that the study areas were chosen for their tectonic similarity. In this context and with large magnitude inslab events showing good agreement of ground motion parameters, it may be acceptable and indeed inevitable to use Japanese data to predict interface ground motion in Cascadia. At the magnitude and distance ranges with a significant contribution to regional seismic hazard, the results here show that globally derived relations are applicable to model Cascadia ground motion.

Data Sources and Online References

Canadian National Seismic Network (CNSN), Natural Resources Canada

autodrm <http://www.seismo.nrcan.gc.ca>

Consortium of Organisations for Strong Motion Observation Systems (COSMOS)

Virtual Data Centre <http://db.cosmos-eq.org/>

Earthquake Research Institute (ERI) of the University of Tokyo

<http://www.eri.u-tokyo.ac.jp/>

Full Range Seismograph Network of Japan (F-Net), formerly known as

Fundamental Research on Earthquakes and Earth's Interior Anomaly (FREESIA)

<http://argent.geo.bosai.go.jp/freesia/index.html>

Harvard University Seismology <http://www.seismology.harvard.edu>

IRIS Consortium: The facilities of the IRIS Data Management System, and specifically the IRIS Data Management Center, were used for access to waveform and metadata required in this study. The IRIS DMS is funded through the National Science Foundation and specifically the GEO Directorate through the Instrumentation and Facilities Program of the National Science Foundation under Cooperative Agreement EAR-0004370. <http://www.iris.edu>

Kyoshin-Net (KNET) of NIED <http://www.k-net.bosai.go.jp>

National Earthquake Information Centre (NEIC) of the United States Geological Survey (USGS) <http://www.neic.cr.usgs.gov/neis/epic/epic.html>

National Research Institute for Earth Science and Disaster Prevention, Science and Technology Agency (NIED) <http://www.bosai.go.jp/index.html>

Oregon State University, College of Oceanic and Atmospheric Sciences
<http://quakes.oce.orst.edu/moment-tensor>

Pacific Geoscience Centre, Natural Resources Canada (PGC NRCan)
<http://www.pgc.nrcan.gc.ca/seismo/MTS/index.html>

United States National Seismic Network (USNSN), United States Geological Services (USGS) autodrm
<http://wwwneic.cr.usgs.gov/neis/AUTODRM/autodrm.html>

Data analysis and graphs in this document were prepared using:

CoHort Software (2002) CoStat Monterey, California www.cohort.com

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Appendices

See attached CD-ROM

Appendix A Ground Motion Data from Japan (KNET)

File names:	Japan.dt	CoStat format	11 MB
	Japan.xls	Microsoft Excel format	17 MB
	Japan.txt	Tab delimited text file	9 MB
	Japan.ssv	Space delimited ASCII file	12 MB

Data file contents are explained in the table below:

Column	Title	Details
1	Date (JT)	Date in local Japan time format
2	Time (JT)	Time in local Japan time format
3	UT	Time in universal time format, Day -1 = previous day in UT
4	lat	Latitude
5	lon	Longitude
6	depth (km)	Hypocentral depth in kilometres
7	M (JMA)	Japan Meteorological Agency magnitude
8	M _w	Moment magnitude
9	M ₀ (Nm)	Seismic moment
10	Strike 1	Focal mechanism: first reported strike angle
11	Strike 2	Focal mechanism: second reported strike angle
12	Dip 1	Focal mechanism: first reported dip angle
13	Dip 2	Focal mechanism: second reported dip angle
14	Dip Class	Classification of dip angle, 1 = shallow, 2 = steep (see section 4.2.1)
15	Rake 1	Focal mechanism: first reported rake angle
16	Rake 2	Focal mechanism: second reported rake angle
17	Rake Class	Classification of rake angle: 1 = thrust, 2 = normal, 3 = strike slip, 4 = ambiguous thrust, 5 = ambiguous normal (see section 4.2.1)
18	Type	Type of Subduction earthquake: C = Crustal, S = Inslab, I = Interface (see section 2.1)
19	stn	Recording station
20	Site Class	NEHRP site type (A, B, C, D, E) (see section 6.2)
21	R (km)	Hypocentral distance of station

22	0.33	Pseudo acceleration response at 0.33 Hz
23	0.5	Pseudo acceleration response at 0.5 Hz
24	1	Pseudo acceleration response at 1 Hz
25	2.5	Pseudo acceleration response at 2.5 Hz
26	5	Pseudo acceleration response at 5 Hz
27	10	Pseudo acceleration response at 10 Hz
28	25	Pseudo acceleration response at 25 Hz
29	log R	log hypocentral distance
30	log0.33	log pseudo acceleration response at 0.33 Hz
31	log0.5	log pseudo acceleration response at 0.5 Hz
32	log1	log pseudo acceleration response at 1 Hz
33	log2.5	log pseudo acceleration response at 2.5 Hz
34	log5	log pseudo acceleration response at 5 Hz
35	log10	log pseudo acceleration response at 10 Hz
36	R	Distance to fault calculated from AB2002 (see section 6.3)
37	g	Geometric spreading factor calculated from AB2002 (see section 6.3)
38	PGArx	log peak ground acceleration on rock as calculated from AB2002 (see section 6.3)
39	Abs PGArx	Absolute value of peak ground acceleration on rock as calculated from AB2002 (see section 6.3)
40	0.33SC	Pseudo acceleration response at 0.33 Hz corrected for site type
41	0.5SC	Pseudo acceleration response at 0.5 Hz corrected for site type
42	1SC	Pseudo acceleration response at 1 Hz corrected for site type
43	2.5SC	Pseudo acceleration response at 2.5 Hz corrected for site type
44	5SC	Pseudo acceleration response at 5 Hz corrected for site type
45	10SC	Pseudo acceleration response at 10 Hz corrected for site type
46	log0.33cor	log pseudo acceleration response at 0.33 Hz corrected for site type
47	log0.5cor	log pseudo acceleration response at 0.5 Hz corrected for site type
48	log1cor	log pseudo acceleration response at 1 Hz corrected for site type
49	log2.5cor	log pseudo acceleration response at 2.5 Hz corrected for site type
50	log5cor	log pseudo acceleration response at 5 Hz corrected for site type
51	log10cor	log pseudo acceleration response at 10 Hz corrected for site type

Appendix B Ground Motion Data from Cascadia

File name:	Cascadia.dt	CoStat format	116 KB
	Cascadia.xls	Microsoft Excel format	213 KB
	Cascadia.txt	Tab delimited text file	107 KB
	Cascadia.ssv	Space delimited ASCII file	133 KB

Data file contents are explained in the table below:

Column	Title	Details
1	Special event	Name given to earthquakes of particular regional importance
2	Date (UT)	Date in Universal Time format
3	Time (UT)	Time in Universal Time format
4	lat	Latitude
5	lon	Longitude
6	Depth (km)	Hypocentral depth in kilometres
7	ML	Richter magnitude
8	Mw	Moment magnitude
9	Strike 1	Focal mechanism: first reported strike angle
10	Strike 2	Focal mechanism: second reported strike angle
11	Dip 1	Focal mechanism: first reported dip angle
12	Dip 2	Focal mechanism: second reported dip angle
13	Dip Class	Classification of dip angle: 1 = shallow, 2 = steep (see section 4.2.1)
14	Rake 1	Focal mechanism: first reported rake angle
15	Rake 2	Focal mechanism: second reported rake angle
16	Rake Class	Classification of rake angle: 1 = thrust, 2 = normal, 3 = strike slip, 4 = ambiguous thrust, 5 = ambiguous normal (see section 4.2.1)
17	Type	Type of Subduction earthquake: C = Crustal, S = Inslab, I = Interface (see section 2.1)
18	stn	Recording station
19	site type	NEHRP site type (A, B, C, D, E)
20	R (km)	Hypocentral distance of station
21	log R	log hypocentral distance
22	0.33	Pseudo acceleration response at 0.33 Hz
23	0.5	Pseudo acceleration response at 0.5 Hz
24	1	Pseudo acceleration response at 1 Hz
25	2.5	Pseudo acceleration response at 2.5 Hz
26	5	Pseudo acceleration response at 5 Hz
27	10	Pseudo acceleration response at 10 Hz
28	20	Pseudo acceleration response at 20 Hz

29	log 1	log pseudo acceleration response at 1 Hz
30	log 2.5	log pseudo acceleration response at 2.5 Hz
31	log 5	log pseudo acceleration response at 5 Hz
32	log 10	log pseudo acceleration response at 10 Hz
33	R	Distance to fault calculated from AB2002 (see section 6.3)
34	g	Geometric spreading factor calculated from AB2002 (see section 6.3)
35	PGArx	log peak ground acceleration on rock as calculated from AB2002 (see section 6.3)
36	absPGAx	Absolute value of peak ground acceleration on rock as calculated from AB2002 (see section 6.3)
37	1SC	Pseudo acceleration response at 1 Hz corrected for site type
38	2.5SC	Pseudo acceleration response at 2.5 Hz corrected for site type
39	5SC	Pseudo acceleration response at 5 Hz corrected for site type
40	10SC	Pseudo acceleration response at 10 Hz corrected for site type
41	log1cor	log pseudo acceleration response at 1 Hz corrected for site type
42	log2.5cor	log pseudo acceleration response at 2.5 Hz corrected for site type
43	log5cor	log pseudo acceleration response at 5 Hz corrected for site type
44	log10cor	log pseudo acceleration response at 10 Hz corrected for site type