Spatial Correlation of Subduction Interplate Coupling and Forearc Morpho-Tectonics

Abstract approved:

Andrew J. Meigs

The two largest earthquakes ever recorded, the 1964 M\text{w} 9.2 Alaskan and 1960 M\text{w} 9.5 Chilean, occurred on seismogenic plate interfaces at subduction zones. It has been theorized that the catastrophic failure of a locked zone along the contact between the downgoing slab and the upper plate causes these earthquakes, although determinations of the position, attitude, and extent of this locked zone vary from model to model. Four methods used to constrain the positions of the locked zones are: (1) historical great earthquake rupture extents, (2) heat flow/thermal profiles along the seismogenic plate interface, (3) patterns of surface deformation across the subduction zone forearc, and (4) spatial patterns of upper plate seismicity. Secondary parameters, such as subducted sediment thickness, upper plate lithology, and dip angle of the subducting slab likely play a role in locked zone location as well. In addition to a locked zone, the upper plate of most subduction zones is marked by paired inner and outer forearc highs and basins between the deformation front (trench) and the volcanic arc. Although such surface morphological features are easy to recognize, their spatial and geometric relationships to the locked zone have not been investigated systematically. This thesis investigates correlations between the spatial position of these morpho-tectonic features and the underlying locked zone at the Aleutian, Alaskan, Cascadia, Costa Rican, Javanese,
Sumatran, Nankai, and Southern Chilean subduction zones. For all subduction zones other than Cascadia, which has yet to experience a great earthquake in historical times, the applied means of determining the position of the locked zones place them on plate interface regions between the inner and outer forearc highs. A strong correlation exists between dip of the downgoing plate and the width of both the locked zone and the spacing of the forearc morphologic elements for each of the subduction zones examined. The concept of comparative subductology is updated and enhanced in this study by creating GIS databases incorporating geological, seismological, geodetic, and geophysical observations. Correlations between surface morphological features and geologic and geophysical observations provide insight into controls on the position of the locked zone responsible for great earthquakes within the eight subduction zones examined, indicating that forearc morphology and interplate coupling are related via basic subduction parameters and the structural-tectonic regime of the forearc region.
Testing Spatial Correlation of Subduction Interplate Coupling and Forearc Morpho-Tectonics

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A THESIS

submitted to
Oregon State University

in partial fulfillment of
the requirements for the
degree of

Master of Science

Presented October 9 2003
Commencement June 2004
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Introduction

Great ($M_w > 8.0$) earthquakes occurring along plate interface megathrusts in subduction zones (Figure 1) are the most powerful and devastating seismic events on earth, and are responsible for the vast majority of seismic energy released worldwide.

Great earthquakes spawned from subduction megathrusts result from the catastrophic failure of part of the interface between the two plates that is coupled during interseismic periods of strain accumulation, or what is commonly conceptually modeled as a "locked zone." Although there remains some debate (which will be addressed in the "Discussion" chapter) as to the most accurate spatio-mechanical definition of the locked zone and the best means to conceptually and physically characterize it, this wide portion of the plate interface has significant surface area, and lies at some depth between an updip region of the plate interface where stable sliding occurs and a deeper downdip transition zone below which the plate interface exhibits plastic behavior (Scholz, 2002).
Figure 2 shows a cartoon block diagram through a hypothetical subduction zone, demonstrating a typical locked zone geometry along the shallowly dipping plate interface. Great earthquake ruptures are characterized by shallowly dipping moment tensor solutions with a reverse sense of motion (Scholz, 2002). The intricacies of the exact definition of the locked zone and how it behaves through great earthquake cycles are hotly debated in the present literature, particularly concerning the best proxies for determining its location (e.g. Hyndman and Wang, 1995, Oleskevich et al., 1999, Fluck et al., 1997, others).

Because of the large energy release and potential for far-reaching damage during great subduction zone earthquakes, extensive research has been undertaken on subduction zones and their seismicity. Although the fundamental mechanisms of great earthquake generation and the subduction system are well understood, much debate remains as to the intricacies of locked zone locations, and little work has been done on the direct relationships between the locked zones that give rise to these damaging events and morphologic elements of the overlying subduction zone forearc regions.
Subduction zones that have experienced great earthquakes share a number of common geomorphic and tectonic features. Landward from the trench these features include an accretionary prism, outer-arc high, outer forearc high (coast range or island arc), outer forearc basin, and eventually, a volcanic arc, idealized in Figure 3. Despite the recognition of this common pattern of topographic/geomorphic elements, very little is known about the relationship between their spatial and geometric positions and the location, attitude, and extent of the locked zone of the subduction interface.

**Goals of This Study**

This study strives to investigate the relationships between forearc morphologic elements (FAME) and the locked zone by updating the concept of comparative subductology first put forth in 1982 by Seiya Uyeda. Comparative subductology involves gathering and then comparing and contrasting data describing subduction zone parameters such as dip angle, age of downgoing lithosphere, rate of plate convergence, locked zone position and area, historical great earthquake rupture extent, and others. Observations and models from better-constrained subduction zones such as the Nankai in Japan can be used to understand lesser-constrained convergent boundaries, such as Cascadia or southern Chile.
Comparative subductology is updated in this study through integration of many kinds of geophysical data atop a base of digital elevation and bathymetric data in Geographic Information Systems (GIS) databases to explore correlations between surface morphologic features and geophysical observations related to the subsurface structure of subduction zones. Eight different subduction zones are examined (Figure 1): Alaska, Aleutian, Cascadia, Costa Rica, Java, Nankai, Southern Chile, and Sumatra. Historical and modern seismicity, geodetic data, and extent of great earthquake surface rupture are gathered and compiled in GISs to facilitate comparative subductology as a means of answering fundamental first-order questions about the relationships between subduction locked zones and forearc morphologic elements.

Specifically, this study strives to address the following questions:

1. Where are the regions of interplate coupling and earthquake generation in relation to the principal forearc morphologic elements?
2. Are those first-order forearc morphologic elements linked genetically to the position of the seismogenic zone?
3. Does morphologic analysis provide a predictive framework for understanding downdip changes in the mechanical behavior of the plate interface?

These questions, considered within the framework of the general eight-subduction zone model constructed in this study, have implications ranging from future scientific study of subduction zones, great earthquakes, to understanding potential seismic hazards.
Background

Subduction of one lithospheric plate beneath another was first theorized many years ago to explain the constancy of the surface area of the earth in light of the recognition of sea-floor spreading (e.g. Melville, 1966; Sykes, 1966; Issacks et al., 1969; White et al., 1970; Uyeda, 1982, and many others) and the existence of sub-planar regions of elevated seismicity descending deep into the earth’s mantle beneath island arcs (Wadati-Benioff zones; Wadati, 1928, see below). The advent of global seismic networks during the Cold War provided the physical seismological evidence to confirm the suspicions of scientists developing the early theories of plate tectonics. Further advances, including quantification of heat flow, finite element modeling, GPS technology, and multi-channel broadband local seismic networks have allowed researchers to explore the various processes at work in subduction zones around the planet.

Large shallow earthquakes arising from subduction megathrusts (Figure 4, Pacheco and Sykes, 1992) account for almost 90% of the seismic moment released globally. The tremendous amount of energy released during subduction zone great earthquakes is related to the larger surface area of the plate interface that accumulates

![Figure 4 - Global Seismic Moment. After Pacheco and Sykes, 1992. Subduction zones examined in this study are highlighted in color.](image-url)
strain and undergoes deformation during interseismic periods, also known as the coupled, or locked zone. Researchers have long recognized the existence and importance of understanding the locked zone and its implications for great earthquake generation within subduction zones (Kelleher, 1972, Kelleher et al., 1974, Kanamori, 1977; Singh et al., 1992, Tichelaar and Ruff, 1993, Hyndman and Wang, 1995, Hyndman et al, 1995, Ruff and Tichelaar, 1996, Oleskovich et al., 1999, Ruff and Kanamori, 1983, Mazotti et al., 2000, etc.). Much of this research has focused on defining the updip and downdip position, attitude, and extent of the locked zone, as these parameters are directly proportional to great earthquake energy release, and therefore seismic hazards.

Means of Locating the Locked Zone

The locked portion of the subduction plate interface has two boundaries—an updip and a downdip limit (question marks, Figure 5). Locating these boundaries carefully is of utmost importance to understanding the full seismic potential of each subduction zone, as the strength of any earthquake is directly related to the area of rupture of the fault that generates it (Pacheco 1993). Both the upper and lower locked zone boundaries are located at transitional areas between stable sliding and stick/slip regions of the

Figure 5 - Block diagram through the Sunda Island area of the Java subduction zone, demonstrating the variable position of the locked zone with relation to the elements of the forearc morphology (colored lines, see Figure 3 for description of ideal FAMEs). Scale varies, but 1 cm ~35 km.
subduction plate interface (Tichelaar and Ruff, 1993; Furukawa, 1993; Ruff and Kanamori, 1983; Pacheco et al, 1993; many others). Above the updip limit, the presence of porous, unconsolidated marine sediments infused with fluids inhibits the development of rock strength needed for earthquake-inducing brittle-slip behavior (Ruff, 1991).

Figure 6 – Temperature/depth/strength profile for the San Andreas fault, modified for comparison of changes in rock strength with depth along a subduction zone plate interface, and consequently the locked and transition zones of that plate contact. After Scholz, 2002.

Similarly, at depths below the downdip limit of the locked zone, temperatures and pressures are such that the plate interface transitions back to stable sliding behavior.

Between the updip and downdip limits, conditions are suitable for the rocks of the downgoing plate to become mechanically impinged by the rocks of the upper plate, creating the locked portion of the plate interface. The lithology of the weaker upper plate, however, controls seismogenic behavior because the downdip limit to seismicity occurs where either plate behaves plastically.

Scholz (1988) presented a key diagram depicting the upper and lower limits to the seismogenic zone shown in Figure 6. This diagram was intended to be applied to the San Andreas fault, which cuts continental crust that is quartzo-feldspathic in nature; however, the fundamental concepts are quite applicable to oceanic crust whose major phase is
olivine. The basalt-eclogite phase transition, along with the dewatering of hydrous igneous minerals commonly found in subducting oceanic crust (such as biotite and muscovite) are parallels to the quartz plasticity onset depth and alteration shown in Scholz’s diagram (Ruff and Kanamori, 1983). $T_4$ represents the updip limit of the seismogenic zone, below which rocks behave in a brittle fashion until temperatures and pressures build and the downdip seismogenic limit is reached at $T_1$. There, quartz plasticity becomes a factor as the temperature rises past 300°C, fault rocks transition from cataclastites to mylonites, and adhesion replaces abrasion as the major means of wear (Scholz, 2002). A particularly important facet of Scholz’s theory on fault mechanics as it applies to subduction megathrust great earthquakes is the concept of friction rate behavior (middle of figure 6). Above the updip limit to the seismogenic zone in Scholz’s model, rocks experience velocity strengthening, and are incapable of nucleating earthquakes. Below this point, in the locked zone, velocity weakening in the faulted rocks allows earthquakes to nucleate as stored strain is released during a great earthquake. The downdip limit to the seismogenic zone lies in the region below which the rocks once again experience velocity strengthening, and cannot accommodate brittle rupture, or earthquake nucleation. It is important to note that even though great earthquake rupture cannot nucleate above or below the seismogenic zone, it can extend above and beyond it during earthquakes. This important distinction between the coupled, or locked zone, and great earthquake rupture extents will be re-examined below in the “Discussion” chapter.

Finite element thermal modeling of many of the subduction zones examined in this study (e.g. Hyndman and Wang, 1995; Oleskevich et al., 1999; Fluck et al., 1997) agrees with the depths and temperatures depicted in Figure 6, placing the downdip limit to seismicity at or near the 350°C isotherm, underlain by a transition zone between the
350°C and 450°C isotherms, although this modeling often neglects to account for the influence of elevated pore-fluid pressure on the placement of the updip limit to the locked zone.

**Locked Zone Determination Methods**

Although most researchers agree on the existence of the locked zone and concur that its upper and lower limits are subject to geologic controls, there remain several methods by which locked zones’ locations and extents are defined. In many cases, the upper limit to seismicity on a subduction megathrust is depicted by researchers as extending updip from an earthquake’s source region all the way seaward, nearly to the subduction zone’s trench. Great earthquakes often nucleate at or on the downdip limit of the locked zone (Kelleher et al., 1973; Scholz, 2002), propagating upwards and outwards to the updip limit of the seismogenic zone (Kato and Seno, 2003). Observations suggest that for the largest earthquakes (Alaska 1964, Chile, 1960), the updip limit for the locked zone extended almost to the associated trenches, where rupture mapping indicated that the earthquakes ruptured the plate boundary to very shallow depths within a few kilometers of the seafloor (e.g. Chile, 1960, Barrientos and Ward, 1990). Some great earthquakes, however, began near the updip limit to the seismogenic zones and ruptured downwards, such as the Mw 7.7 1994 Sanriku-oki and Mw 8.3 1968 Tokachi-oki events (Kato and Seno, 2003) in Japan.

The majority of modern research on subduction locked zones focuses on the location of the downdip limit to the seismogenic zone, which is less well defined. Four methodologies are typically called upon by researchers to define the downdip extent of the locked zone:

1. **Great Earthquake Rupture Zones**
The first and most concrete method by which to define the locked zone involves

Figure 7 - Great Earthquake ruptures from (A) the Alaska/Aleutian subduction zone from Nishenko and McCann (1979, upper) and Sumatran subduction zone, (B) from Zachariassen (1994 – left). Shaded areas in (A) represent Aleutian terrace (crosshatched) versus the trench and trench slope (stippled). Figures are shows as examples of types of great earthquake rupture data gathered from the literature for digitizing and inclusion in GIS databases in this study.

mapping of the rupture zones of previous great earthquakes (Figure 7). This method relies on the principle that the ruptured portion of a plate interface that gives rise to a great earthquake represents the section of the megathrust that was coupled so strongly that it stored sufficient energy to cause such an earthquake. As strain accumulates and is periodically released in a great earthquake, different along-strike regions of a subduction megathrust assume the role of the locked zone. In a subduction zone that has experienced many great earthquakes in historic time, the soundest means of delineating the locked zone is to identify the area between the volcanic arc and the deformation front that has ruptured during past great earthquakes (i.e Figure 7, Java subduction great earthquakes of 1883 and 1881). Such an area is most easily mapped by locating the mainshock and its associated aftershocks; although some disagreement exists about which temporal window of aftershocks should be considered to be directly tied to the
mainshock. Most researchers define subduction zone great earthquake rupture zones according to a time-period grouping of geophysically related aftershocks following a large shallow underthrusting mainshock, such as 12-hour, 24-hour, 36-hour, or 72-hour aftershocks. Literature sources for great earthquake rupture areas typically include documentation of how the rupture areas were defined, as there is no accepted uniform time period with which these ruptures are delineated.

Historical great earthquake rupture extent mapping does not usually distinguish between mainshock rupture area and overall great earthquake rupture extent (aftershocks included), which may or may not extend beyond the region that was locked or coupled prior to the event. In this study, rupture areas are taken to define the locked zone on a first-order approximation of its position and extent.

2 - Thermal Constraints (Heat Flow Models)

A second means of delineating the locked zone is via thermal modeling of temperature conditions along the plate interface. Following basic principles of rock mechanics, such as those of Scholz (1988 and 2002) discussed above, there exists a temperature/depth along a shallowly dipping subduction megathrust below which either the upper or downgoing plate behaves in a manner more plastic than brittle. This temperature/depth constitutes the downdip limit to the seismogenic zone, as earthquakes could only nucleate above it where the rocks on both sides of the fault could store strain until they fail (Tichelaar and Ruff, 1993; Savage et al., 1991; Hyndman and Wang, 1993). The “heat-flow” or thermal method of defining the locked zone limits is typically applied via numerical modeling of temperature conditions along the megathrust. Models often incorporate several real subduction parameters, such as incoming plate age and
convergence rate, rock type, regional sedimentation influx into the megathrust, pore-fluid pressure along the plate interface, and regional geothermal gradient. Building on the work of Hyndman and Wang (1993 and 1995) and Tichelaar and Ruff (1993), Oleskovich et al. (1999) identify temperature as the most significant factor in determining the updip and downdip limits of the coupled region of the plate interface (Figure 8). Oleskovich et al. used finite element numerical models to estimate the temperatures along the top surface of the downgoing slab in Cascadia, south Alaska, SW Japan, and Chile, and concluded that the updip edge of the locked zone begins where the temperature reaches 100°C, and the downdip limit of the coupled seismogenic zone lies at ~40km depth, or 350°C (Figure 8). Using temperature alone to delineate the updip limit of the locked zone can be problematic, however, as sedimentation, pore-fluid pressure along the fault, and other factors can work in tandem or override temperature as the main control over rock behavior along the megathrust. This will be discussed in greater detail in the Discussion Chapter; here it is important to recognize that many studies of subduction interplate coupling have based their locked zone determinations on thermal models using the criteria established by Tichelaar and Ruff (1993) and Hyndman and Wang (1995). Only after future $M_{w} > 8.0$ subduction earthquakes occurs will the ability of thermal models to predict the position of the locked zone be tested.

**Figure 8** - Limits to seismogenic zone. From Oleskovich et al., 1999.
3 – Upper Plate Seismicity

The locked zone’s downdip limit can be constrained by upper plate seismicity. The bottom depth of seismicity in the upper plate represents the lower limit of brittle behavior in the upper plate half of the subduction plate interface (lower dashed white line, Figure 9). Ruff and Kanamori (1996) focused on the geology of the two plates, ascribing the downdip end of the locked zone to the basalt to eclogite phase transition at 30-35 km depth, below which the plates become uncoupled due to the onset of superplastic deformation. Following this concept, there should be some depth in the upper plate of a subduction zone below which earthquakes cannot nucleate on the megathrust between the two plates. Although this technique is somewhat speculative, and no examples of its use exist in the published literature (although it technically is the application of Scholz (2002) model to the upper plate), it nevertheless makes sense following the simple geologic principles on which it is based.

Figure 9 – Cartoon cross section through an ocean-continent subduction zone showing the correlation of the lower limit to seismicity in the upper plate with the downdip extent of the locked zone. After Fluck et al. (1997) and Byrne and Hibbard (1988).
4 - Surface Deformation

Careful examination of forearc ground deformation above a subduction zone measured by GPS and other geodetic measurement techniques presents a fourth way of defining the locked zone. In interseismic periods as the downgoing plate subducts and strain accumulates across the soon-to-be seismogenic plate interface, the upper plate behaves like an elastic beam, flexing and generating uplift across the forearc (Savage, 1983) as shown in Figure 10a. In coseismic periods (during earthquakes) the strain is released, the upper plate returns to an unflexed or normal state, and the forearc experiences subsidence (Figure 10b). Savage's (1983) description of the importance of ground deformation across a subduction zone's forearc in terms of revealing the state of coupling on the megathrust below paved the way for studies in the coming decade that would utilize the new technology of GPS to enhance the ability to measure ground deformation above subduction zones. Dragert et al. (1990), Murray and Lisowski (1994) in Cascadia along with Shimada et al. (1990) and many others in Japan were amongst the first users of GPS to measure deformation across a subduction zone forearc. Since then, numerous studies have shown the ability of GPS measurements to be used in constructing a model describing plate coupling processes at subduction zones. The locked zone can be defined by dislocation modeling of current deformation data in an attempt to fit that data to plate coupling models which are sensitive to width, depth, and geometries.
of the locked and transition zones. Savage et al. (1998) presented deformation data collected over a period of a few years across the rupture zone of the 1964 M\textsubscript{w} 9.2 great earthquake in southern Alaska, and constructed four models with differing locked zone geometries as an example of how GPS data can be used to locate the locked portion of a subduction zone plate interface. Figure 11 provides a similar comparison of inter- and co-seismic deformation data from the 1946 great earthquake in the Nankai subduction zone of southwestern Japan, based upon the work of Thatcher (1984), which demonstrated conclusively that interseismic deformation is approximately the inverse of coseismic deformation across subduction forearcs (Figure 11).

Except for the M\textsubscript{w} 8.3 great earthquake of September 25, 2003 in the northern Japanese subduction zone (which is presently being studied in detail), there have been no great earthquakes since the advent of GPS technology. The ability of GPS data and coupling models of subduction zones through complete seismic cycles remains to be seen.
Forearc Morphologic Elements

Great earthquake-generating subduction zones around the planet share a number of common geomorphic and tectonic features distributed across the forearc region between the trench and the volcanic arc (idealized in Figure 3, and in an actual topographic/bathymetric model in Figure 5). These features consist of a pair of topographic highs and lows, known as the outer and inner forearc high (OFAH and IFAH, respectively), and interspersed between them lie two topographic lows, the inner and outer forearc basins (IFAB and OFAB, respectively). These features are thought to be structurally controlled by either landward- or ocean-verging shallow thrust sheets within the accretionary prisms of subduction zones (Dickinson and Seely, 1979) (See Figure 26 below, seismic profile across Java subduction zone, and Discussion Chapter).

Past work on Subduction Zone Forearc Morphology

The first detailed publications on topographic features formed by subduction tectonic processes came from Seely (1977) and Dickinson and Seely (1979). Dickinson and Seely (1979) categorized the structural features seen across accretionary wedges in petroleum-exploration seismic profiles into groups based upon the structural vergences and basin structure. Seely (1979) examined the vergence of structures in subduction forearcs, and a few years later, Melosh and Raefsky (1980) recreated the overall (trench-high-basin-arc) forearc morphology common to many subduction zones via finite element numerical modeling, and found that the overall forearc morphology of subduction forearcs arises from viscous stresses generated in the lower lithosphere during subduction. In 1992, Wdowinski constructed viscous flow models of subduction zones to model the deformation near trenches, ignoring the more subtle morphologic features further landward. Ruff and Tichelaar (1996) pointed out an oft-overlooked correlation
between the coastline of many subduction zones and the trace of the downdip edge of the seismogenic zone as defined by thermal modeling. They suggested that the upper-plate Moho intersects the downgoing slab at the downdip limit to the locked zone, and that rocks of the mantle wedge are aseismic. Cattin and Lyon-Caen (1997) built a 2-dimensional finite element model to examine the long-term effects of coupling for three subduction zones: northern Chile, northern Japan, and Tonga, noting that the observed topography at those three subduction zones’ forearcs are inconsistent with a megathrust whose effective friction coefficient ($\mu$) is smaller than $\mu = 0.2$.

None of the above studies, however, made quantitative comparisons of subduction forearc morphology and the seismogenic coupled zone. The work of Song and Simons (2003) marks the first published effort of researchers to quantitatively and simultaneously examine forearc morphology in relation to great earthquakes in multiple subduction zones. Great earthquake ruptures, they concluded, are associated with negative topographic anomalies (areas of low mean topography) and positively correlated to areas with negative free-air gravity measurements. Song and Simons, however, only considered onshore regions of the forearc morphology, eliminating analysis of the OFAH and OFAB in most subduction zones. Other related recent work by Wells et al. (2002) presented preliminary work examining the connection between large, sediment filled OFABs and great earthquakes, and are presently working on this relationship in more detail.

A cursory examination of the bathymetry and topography of the eight subduction zones examined in this thesis reveals both the presence of these forearc morphologic elements and a variety of their arrangement, spacing, and distribution (see figures describing topography and bathymetry of each subduction zone below). The examination
of these features and how they relate to each other within each subduction zone and from one zone to another forms the central focus of this thesis. Further discourse on the geology, structure, and morphology of these forearc morphologic elements is deferred to the “Overview of Plate Margins” and “Discussion” chapters. Even with aberrations in the myriad properties of subduction zones (degree of dip of the downgoing slab, plate age, convergence rate, plate thickness, sediment input, etc.), the fundamental mechanism of subduction does not vary drastically amongst subduction zones, and neither do the basic geologic, tectonic, structural, seismic, geophysical, and morphologic features of each subduction zone. Comparing forearc morphologic features of one subduction zone to another along with geologic, tectonic, and geophysical information about the locked zone will help relate the locked zone to the forearc morphology, and constrain the potential for great earthquake generation at each subduction zone examined.
Overview of Plate Margins Included in this Study

Eight plate margins are examined in this study to illustrate the geologic, tectonic, and great earthquake history of each subduction zone. Table 1 highlights common geophysical, seismological, and structural features that relate to each region’s locked zone and overlying forearc structure and morphology.

Table 1 – Overview of Subduction Zones

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Synopsis of Subduction Zones

Alaskan Subduction Zone

The Pacific plate subducts beneath the North American plate at ~57 mm/yr (Zweck et al., 2002) along the Alaskan subduction zone (Figures 12 and 13). In 1964, the southern Alaskan subduction zone generated the second most powerful earthquake in recorded history, a Mw 9.2 shock whose epicenter was located in Prince William Sound between the inner forearc basin and the trench (Figure 12). Significant deformation occurred across the forearc as a result of this event (Plafker, 1965, 1972; Savage et al., 1998, Zweck et al., 2002; Santini et al., 2003), which has been extensively studied.

Modeling of GPS data in recent years suggests the existence of two separate asperity-dominated locked patches exist in the Alaskan subduction zone (Zweck et al., 2002, Figures 13 and 14). Santini et al. (2003) further confirmed this via modeling of the slip distribution in the 1964 great earthquake rupture area via Monte Carlo simulation.

Figure 12 – Diagram outlining large earthquakes in the Aleutian (top) and Alaskan (bottom) subduction zones, highlighting the rupture areas from the earthquakes (stipple). From Tichelaar and Ruff, 1993.
Figure 13 – GPS station velocities across the forearc of the Alaskan subduction zone. From Zweck et al, 2002. South oriented stations on the northern side of the Kenai peninsula indicate shortening across the forearc resulting from strong plate coupling in that area, which recommenced shortly after the 1964 great earthquake.

Figure 14 – Plate locking model of Zweck et al (2002) which calls upon two separate locked asperities, (A) near Kodiak Island, and (B) in the vicinity of Prince William Sound and the rupture area of the 1964 Mw 9.2 great earthquake.
Plate convergence to the east of the Anchorage region is complicated by the subduction of the Yakutat terrane, a composite oceanic-continental block (Mazzotti and Hyndman, 2002) that is subducting along with the Pacific plate beneath North America and the Prince William Sound southeast of Anchorage. Seismic reflection surveys across Prince William Sound (Brocher et al., 1994) imaged the ramp megathrust along which the Yakutat Terrane and the Pacific plate subducts beneath North America at ~20 km depth beneath the sound, dipping to the NW at ~3°. Further to the SE in the vicinity of the Wrangell-St. Elias and Chugach Mountains, the Yakutat Terrane is engaged in oblique convergence with North America (Plafker, 1987).
Aleutian Subduction Zone

Several great earthquakes have occurred in the Aleutian subduction zone, notably the Andreanof Island 1957 $M_w$ 9.1 and 1986 $M_w$ 7.9 events, the Shumagin Island 1938 $M_w$ 8.2, and the Rat Island 1965 $M_w$ 8.7 earthquake (Zheng et al., 1996, and Kato and Seno 2003, see also Figure 12). In addition, there have been numerous smaller $M_w$ 6-7 events (Figure 12). Subduction of Pacific plate oceanic crust throughout the Aleutians is oblique, as much as 30° from orthogonal since at least Miocene time, at a rate of 80-90 mm/yr (Ryan and Scholl, 1989). The structure of the Aleutian forearc region was imaged by Ryan and Scholl (1989) and consists of mainly seaward-verging thrust sheets piled atop a seaward verging backstop (Figure 16). Throughout the Central Aleutian subduction zone, right-lateral shearing of actively deforming forearc structures driven by the oblique convergence dominates the

Figure 16 - Seismic reflection profile across the Amlia area of the Aleutian Islands, from Ryan and Scholl (1989). Note seaward vergence of forearc structure throughout accretionary prism and seaward dip of backstop (purple dashed line).
accretionary prism region (Ryan and Scholl, 1989). GPS-based geodesy and modeling of the Shumagin Island region (Zheng et al., 1996) indicates that the megathrust beneath the eastern part of the Aleutian subduction zone has strong enough coupling capable of generating a large earthquake. Due in part to the more oblique convergence in the western region of the Aleutians, coupling is not as strong, and smaller earthquakes occur. The average rupture length of great earthquakes in the Aleutians exceeds 400 km (Sykes, 1971; Mogi, 1968), and although the population density in the area remains low, tsunamis generated from Aleutian earthquakes can affect localities around the Pacific.

**Cascadia Subduction Zone**

The subduction of the Juan de Fuca plate beneath the North American continent has been well studied. Although there has not been a great earthquake in the recorded history of this subduction zone, there is ample evidence of coastal subsidence (Atwater, 1987, 1995, 2003, Jacoby 1997, Kelsey, 1998) and offshore turbidites (Goldfinger et al., 2002) that the Cascadia megathrust has produced large earthquakes throughout the Holocene.

The southern boundaries of this active margin is characterized by a
complicated menagerie of structural and tectonic elements due to the transition from the strike-slip plate boundary in northern California northward through the Mendocino triple junction to subduction of the Juan de Fuca plate beneath North America in Cascadia (Figure 18). Notably, a $M_w$ 7.1 earthquake occurred in 1992 at (Oppenheimer et al., 1993).

Some confusion remains as to whether the earthquake occurred in the accretionary prism or along the interplate megathrust between the Juan de Fuca and the North American plates (Hagerty and Schwartz, 1996).

Figure 18 – Topography, bathymetry, and seismicity of the Cascadia subduction zone. Seismicity from NEIC catalogue (See Appendix).

According to Goldfinger et al. (1994 and 1997), strain partitioning induced by oblique convergence of the two plates has led to the development of strike-slip faulting in the forearc region of the Juan de Fuca plate. Vergence of forearc structures in Cascadia varies with latitude: in northern Oregon and southern Washington vergence is landward, whereas central and southern Oregon have seaward vergence, and
southernmost Oregon and California have landward verging structures (Trehu et al., 1994; Trehu et al., 1995; McNeill et al., 2001; Goldfinger, 1997).

The location and physical dimensions of the locked zone in the Cascadia subduction zone has been the subject of much recent debate (Hyndman and Wang, 1995; McCaffrey et al., 2000; Oleskevich et al., 1999, etc.), and different schools of thought exist as to the most accurate means to constrain the locked zone (Fluck et al., 1997). Heat-flow based modeling (Figure 17) places the locked zone lying offshore under the accretionary prism, whereas GPS-based modeling generally calls for the locked zone to lie farther onshore (Figure 19).

Recent work by Rogers and Dragert (2003) presented evidence of silent-slip events occurring on the deeper part of the northern Cascadia megathrust, capable of triggering a large subduction earthquake. Plate coupling and locked zone modeling in Cascadia will be discussed in more detail below in the “Discussion” chapter.
Costa Rican Subduction Zone

Subduction in Costa Rica (Figure 20) takes place as a result of the thrusting of the Cocos plate beneath the Caribbean Plate at a rate of 70-94 mm/yr (Protti et al., 1995). Protti et al. (1994) note that coupling along this subduction zone changes along strike in a manner correlated with the roughness of the bathymetry on the seafloor being subducted. The high-relief Cocos Ridge subducts beneath the Osa Peninsula in southern Costa Rica, and according to Protti et al. (1994), causes varying degrees of plate coupling along the Costa Rican trench. There have been several large earthquakes in the Costa Rican subduction zone, most recently the Nicoya 1990 Mw 7.0 event, postulated by Protti et al. (1995) to have been the result of the rupture of a seamount asperity. Graefe et al. (2002) identify the subduction angle as 60° in central

Figure 20 - Topography and bathymetry of the Central American subduction zone. Active volcanoes are indicated by red triangles.
Figure 21 - Idealized cross section through the Costa Rican accretionary prism, from high-resolution broadband seismic profiles. From Ye et al., 1996.

Costa Rica, and 30° in southern Costa Rica for the first 60 km of the subducting Cocos plate, although Guzman and Cardenas (1998) attribute an angle of 45° to the subduction along the length of the subduction zone. High resolution seismic profiling of the Costa Rican subduction forearc (Ye et al., 1996; Hinz et al., 1996; Ranero and von Huene, 2000) identifies mainly seaward verging structures in the accretionary prism (Figure 21), and although these profiles do not extend landward far enough across the forearc to provide information about the structures beneath the IFAH and OFAH, these FAMEs are clearly visible from bathymetric and topographic data (Figure 20).

Sumatran Subduction Zone

The Sumatran subduction zone (Figure 22) results from the underthrusting of the Indian and Australian plate beneath Eurasia at a rate of 67 +/- 7 mm/year (Zachariasen et al., 1999) at dips of ~7° (Kopp et al., 2002) to ~15° (Newcomb and McCann, 1987) beneath the forearc, and ~50° (Fauzi et al., 1996) beneath the volcanic arc. Subduction in the Sumatran convergent zone is 40° oblique from orthogonal (Prawirodirdjo et al., 1999). Most of this oblique motion is accommodated along the Great Sumatra fault (shown in Figure 23). The Sumatran subduction zone ruptured in 1833, generating an earthquake with Mw ~8.8-9.2 (Zachariasen et al.)
Evidence for uplift and trenchward migration of deformation as a result of this great earthquake, in addition to rapid subsidence prior to the event are seen in the uplift and downdropping of coral microatolls (Zacahariasen, 1999). Convergence in this subduction zone is 67 +/- 7 mm/yr, at a vector of N11°E, +/- 4° (Tregoning et al., 1994). Although Figure 23 displays the inferred rupture zones from the two largest great earthquakes in this subduction zone, they are not outlined with the certainty and detail with which aftershock or tsunami-based rupture inversions would be.

Newcombe and McCann (1987) reported that the rupture in the 1833 earthquake was 550 km long, spanning from the trench to below the outer-arc islands, and accommodated 4-8 m of slip along the plate interface. Seismic profiles by Kopp et al. (2003) showed seaward verging structures through the Sumatran forearc accretionary prism, as well as a landward verging backstop and outer forearc basins (OFABs).
Javanese Subduction Zone

Plate convergence in this subduction zone is much more orthogonal than the Sumatran subduction zone to the northwest. Abercrombie et al. (2001) studied the 1994 thrust earthquake off the southwest coast of Java, finding it to be the solitary shallow thrust event in the Harvard CMT catalogue. Interestingly, all of the aftershock mechanisms studied and relocated by Abercrombie et al. were normal fault events, in contrast to the mainshock. Abercrombie et al. use evidence from seismic waveform analyses, along with bathymetric data to infer that the relatively slow earthquake (∼12 seconds in duration) resulted from the subduction of a seamount. The overall interpretation of Abercrombie et al. is that this earthquake occurred from subduction of a seamount in the only locked region of the Java subduction zone, which is decoupled everywhere else. As in the Sumatran

Figure 23 - Sumatra subduction zone, showing inferred rupture zones from the 1861 and 1833 great earthquakes. From Zachariasen, 1999.

Figure 24 - GPS geodesy of the northwestern end of the Sumatran subduction zone, where convergence is the most oblique. Prawirodirjo et al., 1999.
subduction zone, seismic profiles of Kopp et al (2003) show seaward prism structures buttresses against a seaward dipping backstop in northwestern Java, as well as perched OFABs.

Nankai Subduction Zone

Of all the eight subduction zones examined in this study, the Nankai trough convergent plate boundary region of southwest Japan (Figure 25) where the Philippine sea plate subducts under Eurasia at 460 mm/yr (Seno et al., 1993) is by far the most well constrained margin. According to Thatcher (1984), data describing interseismic deformation within the Nankai subduction zone is the most complete for any subduction zone on the planet. Numerous studies of seismicity, geodesy,

![Figure 25 - Topography and bathymetry of the Nankai trough, southern Japan. Showing segmented forearc basins (idealized as colored lines, A – Enshu, yellow, B – Kumano, red, C – Muroto, green, d – Tosa, orange, E – Hyuga blue). Segments B and C generated the 1944 and 1946 great earthquakes, respectively.](image-url)
tsunamis, seismic refraction, heat flow/thermal modeling, and coupling models have been done on Nankai, and all have benefited from an extraordinarily long historical and instrumental record of geological events. In the world of great earthquakes, the Nankai subduction zone of southwestern Japan is well known because of two $M_\text{w}$ 8.0 and 8.1 events in 1944 and 1946, the epicenters of which were located at the northeastern end of the subduction zone (Figure 25). In the many decades since these events, researchers (notably Awata and Sugiyama, 1989; Sugiyama, 1989 and 1994) have determined that the Nankai subduction zone is segmented (Figure 25), with ruptures occurring in distinct segments of the margin in consecutive temporal increments (Sugiyama, 1994, and references therein).

The 1946 $M_\text{w}$ 8.2 event, the larger of the two, resulted from rupture of the fully coupled plate interface in what is now known as Tosa segment of the Nankai trough. Slip estimates range from 6.2 m (Tanioka and Satake, 2001) from inverted

![Figure 26](image_url)

**Figure 26** - Cross-section through southern portion of the Nankai subduction zone, showing interpretation of seismic profiles taken through the Muroto Basin, the segment of the Nankai trough that ruptured in the 1946 $M_\text{w}$ 8.1 earthquake. From Nakanishi et al., 2002.
Figure 27 - Topography and bathymetry of the southern Chile subduction zone. Star marks hypocenter of 1960 $M_w$ 9.5 great earthquake.

tsunami data) to 11.1 m (Sagiya and Thatcher, 1999, inverted geodetic data). Rupture in 1946 extended laterally for $\approx 500$ km (Savage 1995) over an area of several hundred km$^2$.

**Southern Chile Subduction Zone**

Subduction of the Nazca plate beneath the South American continent has been continuous since the mid-Tertiary (Lavenu and Cembrano, 1999) at the Chilean subduction zone (Figure 28). Plate convergence in this region has a velocity of 85 mm/yr (DeMets et al., 1990), and becomes more and more margin-oblique to the south and southeast. The largest historic earthquake ever recorded on earth, a $M_w$ 9.5 event, occurred on the Chilean subduction zone in 1960. A detailed examination of the rupture area of this event by Barrientos and Ward (1990) inverted surface deformation to find the rupture area.
over which slip occurred from the mega-quake, some 850 km long and 130 km wide. Although occurring in northern Chile, the Mw 8.0 earthquake that occurred near Antofagasto ruptured a seismic gap and ruptured \( \sim 185 \times 90 \text{ km}^2 \) of the plate interface (DeLois et al., 1997). De Lois et al (1997) also conclude that the down-dip extent of the main rupture corresponds to the deepest edge of the locked zone, at a depth of 50 km. Klotz et al. (2001) estimate that the depth of coupling is heterogeneously \( \sim 33 \text{ km} \) deep north of 30° S, and \( \sim 50 \text{ km} \) deep south of 35° S. Kelleher (1972) examined the rupture zones of large \( (M_L > 7.7) \) earthquakes all along the Chilean margin, and used their spatial positions to infer the locations of seismic gaps to facilitate predictions of future shocks. Kelleher noted that in southern Chile, larger \( (M_R > 8.0) \) earthquakes occur in an orderly N-S temporal progression about every 100 years. Rubio et al. (2000) examined the crustal structure of the southernmost Chilean margin, and found that the high obliquity of the subduction in that region produces similarities between the forearc structures seen there and those found on the central and western Aleutian subduction zone. Rubio et al. (2000) found predominantly landward verging structures in the Patagonian accretionary prism, along with a seaward verging backstop.
Methods

In order to examine the spatial correlation between subduction zone earthquakes and forearc morphologic elements, databases containing primary geophysical observations were constructed within Geographic Information Systems (GISs). Each database was constructed with two goals, the first being the depiction of the spatial extent and position of the locked zone of each subduction zone via the four major methods by which locked zones are identified: (1) historical great earthquake rupture, (2) GPS-based deformation measured across the forearc, heat flow/temperature, (3) heat flow/temperature measurements and the resulting models of the plate interface, and (4) upper plate seismicity. The second goal was to outline in a spatially quantifiable way the trench/deformation front, the inner and outer forearc highs (IFAHs and OFAHs) and intermediary basins, and the volcanic arc for each subduction zone.

![Figure 28 - Location and extent of GIS databases constructed for each subduction zone.](image)

Separate GIS databases for each of the eight subduction zones examined in this study were constructed for the regions delineated in Figure 28, with each database characterized by uniform data type and resolution.
GIS Database Construction

Due to variable coverage and data availability, some databases contained more information than others. Spatial extents of each subduction zone GIS model are found in Table 2. Sources of all data included in the GIS databases are given in Table 3. Each type of data and notable aspects are discussed below.

*Topography / Bathymetry*

**ETOP02 Data - Bathymetry**

Global two-minute (~3.7 km) resolution ETOPO2 topographic and bathymetric data were used as the base layer for each subduction zone GIS model. ETOPO2 seafloor bathymetric data were derived from the work of Smith and Sandwell (1997), which used satellite altimetry observations along with field-checked shipboard echo-sounding measurements. The raw-ASCII elevation data provided by the National Geophysical Data Center and the National Oceanic and Atmospheric Association were converted into ESRI grid format datasets and imported into ArcGIS, and then clipped according to the earthquake search extents described above, and projected to a UTM projection appropriate for each subduction zone. See “Appendix” for details on data sources and GIS methods.

**GTOPO30 - Topography**

Regional 30-arc second (~0.93 km) resolution SRTM30 topographic data were added to each subduction zone’s GIS database in order to identify finer details in features.

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**Table 2 – Spatial Extent of GIS Databases**

<table>
<thead>
<tr>
<th>Subduction Zone</th>
<th>N</th>
<th>S</th>
<th>E</th>
<th>W</th>
</tr>
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<td>-145</td>
<td>-180</td>
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<td>23</td>
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<td>-25</td>
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<td>Japan</td>
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<td>15</td>
<td>120</td>
<td>147</td>
</tr>
<tr>
<td>Java / Sumatra</td>
<td>7</td>
<td>-15</td>
<td>95</td>
<td>125</td>
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<td>Region</td>
<td>Bathymetry</td>
<td>Topography</td>
<td>Earthquakes</td>
<td>GPS Data</td>
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<td>-------------</td>
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<td>Zueck et al. 2002</td>
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<td></td>
<td>ASGDC 300 m</td>
<td>Harvard CMT</td>
<td>Ratchkovsky, 1999</td>
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<td>Pacheco and Sykes, 1992</td>
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<td></td>
<td></td>
<td></td>
<td>Sykes, 1992</td>
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<td>GTOPO30</td>
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<td>GTOPO30</td>
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<td>ANSS</td>
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<td></td>
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<td>LePichon et al, 1998</td>
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<td>Linde and Sacks, 2002</td>
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<tr>
<td>S. Chile</td>
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<td>GTOPO30</td>
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<td>Tichelaar and Ruff, 1993</td>
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<td>Pacheco and Sykes, 1992</td>
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<td>Location</td>
<td>Great Earthquake Rupture</td>
<td>Plate Depths</td>
<td>Seismic Cross-Sec.</td>
<td>Plate Dip</td>
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<td>----------</td>
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<td>Ratnayake, 2002</td>
<td>Loughran et al., 2002</td>
<td>Elnakash et al., 1999</td>
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<td>Nishenko and McCann, 1979</td>
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<td></td>
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<td></td>
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<td>Ranero and Von Huene, 2000</td>
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<td>Kopp et al., 2002</td>
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<td>Rubio et al., 2000</td>
<td>Clekmovich et al., 1999</td>
<td>Laurens, 2002</td>
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<td></td>
<td>Horimoto and Ward, 1990</td>
<td></td>
<td></td>
<td>Laurens et al., 2002</td>
</tr>
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</table>
of the forearc morphology located above sea level, as the SRTM30 data only covers topography and contains no bathymetric information. SRTM30 data are the most recently updated version of the GTOPO30 dataset, which was derived from eight sources, described in full detail in the Appendix. The majority of DEM data in the GTOPO30 dataset used in this study were derived from the Digital Chart of the World (commonly known as DCW) and the US Military’s Digital Terrain Elevation Dataset, (commonly referred to as DTED). The SRTM data were derived from NASA’s 1999 Shuttle Radar Topography Mapping (SRTM) mission, also described in the Appendix.

Locked Zone Determinations

Four main proxies are commonly used by researchers to delineate the updip and downdip extents of subduction locked zones. Methods specific to each locked zone determination are described below.

(1) Historic Great Earthquake Rupture Area

Great earthquake rupture area was digitized from figures published in the literature (Table 3 and Table 4, below). Numerous studies have been undertaken on the extent of great earthquake surface rupture occurring from events at the eight subduction zones examined in this study. These data are typically presented in graphical/map form in peer-reviewed journal articles. The rupture figures were scanned, and georegistered according to the projection, datum, and geographic coordinate systems of the original published figures. Efforts were made to gather information about the sources of the data in their figures beyond that provided in the papers themselves in order to maintain the high accuracy of spatial great earthquake extent. Table 4 gives the sources of great
Table 4 – Sources of Great Earthquake Rupture Areas

<table>
<thead>
<tr>
<th>Zone</th>
<th>Date</th>
<th>$M_w$</th>
<th>Rupture Source</th>
<th>Downdip Area (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sumatra</td>
<td>1833</td>
<td>9.2$'$</td>
<td>Shaking, Tsunami$^i$</td>
<td></td>
</tr>
<tr>
<td>Sumatra</td>
<td>1861</td>
<td>8.5$'$</td>
<td>Shaking, Tsunami$^i$</td>
<td></td>
</tr>
<tr>
<td>S. Chile</td>
<td>1928</td>
<td>8.3$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
<td></td>
</tr>
<tr>
<td>Alaska</td>
<td>1938</td>
<td>8.2</td>
<td>Aftershocks (months)$^{bc}$</td>
<td></td>
</tr>
<tr>
<td>Costa Rica</td>
<td>1939</td>
<td>7.3$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
<td></td>
</tr>
<tr>
<td>S. Chile</td>
<td>1939</td>
<td>8.3$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
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<tr>
<td>Costa Rica</td>
<td>1941</td>
<td>7.5$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
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<tr>
<td>Nankai</td>
<td>1944</td>
<td>7.9$^b$</td>
<td>Geodesy$^k$</td>
<td></td>
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<tr>
<td>Nankai</td>
<td>1946</td>
<td>8.2$^h$</td>
<td>Geodesy$^k$</td>
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<tr>
<td>Costa Rica</td>
<td>1950</td>
<td>7.7$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
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<tr>
<td>Costa Rica</td>
<td>1956</td>
<td>7.3$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
<td></td>
</tr>
<tr>
<td>Aleutian</td>
<td>1957</td>
<td>9.1$^h$</td>
<td>Aftershocks (months)$^{bc}$</td>
<td></td>
</tr>
<tr>
<td>Chile</td>
<td>1960</td>
<td>9.5$^b$</td>
<td>Surface Def.$^d$</td>
<td>110,500$^d$</td>
</tr>
<tr>
<td>Costa Rica</td>
<td>1962</td>
<td>7.4$'$</td>
<td>Aftershocks (unspec.)$^i$</td>
<td></td>
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<tr>
<td>Alaska</td>
<td>1964</td>
<td>9.2$^a$</td>
<td>Aftershocks (months)$^a$</td>
<td>150,000$^c$</td>
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<tr>
<td>Alaska</td>
<td>1965</td>
<td>6.9</td>
<td>Aftershocks (months)$^{bc}$</td>
<td></td>
</tr>
<tr>
<td>Alaska</td>
<td>1986</td>
<td>7.9</td>
<td>Aftershocks$^b$</td>
<td></td>
</tr>
<tr>
<td>Costa Rica</td>
<td>1990</td>
<td>7.0$^e$</td>
<td>1000 days (50%-100 days)$^k$</td>
<td>4000$^g$</td>
</tr>
</tbody>
</table>


earthquake rupture extents used in this study, while Figures 12 and 23 show examples of typical great earthquake rupture figures.

(2) Geodetic Surface Deformation / GPS Data

Surface deformational data from across the forearc regions of the eight subduction zones examined in this study were gathered from the literature and from databases published on the internet, such as the PANGA database containing geodetic station motion information across the Cascadia subduction zone (Miller et al., 2003).

When possible, data were also gathered from published studies of deformation across
subduction zone forearcs, such as that of Zweck et al. (2002) for southern Alaska, or e.g. Savage (1995) and Tabei et al. (1996) for Nankai. More details, along with a comprehensive list of GPS data included in each GIS database, can be found in Table 3.

(3) Heat Flow

Heat flow data pertinent to each of the subduction zones examined in this study were gathered from the literature. Sources are given in Table 3. Heat flow measurements included in the GIS database by digitizing the thermal contours reported in the literature, or by gathering tabular data describing temperature/depth relationships across subduction zone forearcs. Refer to the Discussion section, below, for a discourse on the benefits and caveats of using heat flow data.

(4) Upper Plate Seismicity / Earthquakes

Earthquake data from several earthquake catalogs were added to each GIS database. Table 3 lists the spatial extent of the model for each subduction zone. Figure 23 shows the global extent of searches made from the various earthquake catalogs described below. Each GIS database contained all available earthquakes from the global Harvard CMT catalogue (http://www.seismology.harvard.edu/projects/CMT/, Dziewonski and Wodehouse, 1983) as well as the global ANSS catalogue (http://www.anss.org/, Benz et al., 2001) and published global earthquake datasets, such as Tichelaar and Ruff (1993) and Pacheco and Sykes (1990). In addition, wherever possible, high-accuracy, locally-relocated earthquake data were collected from the literature or imported from datasets generously provided in electronic format by the authors of papers whose study areas matched those of this study.
F.A.M.E. Delineations

Identification and delineation of the different forearc morphologic elements (FAME) in each of the eight subduction zones are central to this analysis. The primary means by which each FAME was identified are described below.

*Shaded Relief Maps of DEMs - Inner and Outer Forearc Highs and Basins*

These paired topographic highs and lows are easily visible on shaded relief maps generated from the DEM data described above. Shaded relief maps (e.g. Figure 29) were generated with ArcGIS’s Spatial Analyst extension, which simulates illumination of a DEM grid and assigns a greyscale value between 1 and 254 to cells in the resulting image. Vertical exaggeration of elevation values from 1.0 to 3.0 was used variably to accentuate shading as the two grids were overlain atop one another and the top color-shaded relief grid made 40% transparent to facilitate identification of areas of high relief such as the

*Figure 29 - Perspective view to the east along the strike of the Java subduction zone, illustrating how the FAME were delineated from color tinted shaded relief images of DEMs.*
inner and outer forearc highs. When delineating the deformation front, care was taken to ensure that only the deepest cells were selected regardless of the apparent color-depth imparted by the relief shading process.

**Volcanic Arcs**

The axes of the volcanic arcs were delineated by drawing smoothed arcs through each volcano closest to the trench identified by the Smithsonian Global Volcanism Program in the Volcanoes of the World database. A spline arc was drawn through the selected volcanoes and eruptive centers of each subduction zone in AutoCAD and then imported back into ArcGIS as an arc coverage.

**Mathematical Derivations of Elevation Grids as a Test of FAME Delineations**

In order to test the accuracy of the FAME delineations as traced from the DEMs, lower-resolution DEMs were derived from the ~3.7 km ETOPO2 DEMs to overly...
accentuate the topographic and bathymetric features of the forearc. Figure 30 shows a 
~92.5 km resolution grid generated by resampling the ETOPO2 grid by standard
deviation in 25-cell increments of the Alaskan and Aleutian subduction zone. These
values provide roughly the same locations for the FAME, providing a validating test of
the hypothesis that they are first-order morphologic features of subduction zones and
that estimating the FAMES from shaded relief DEMs is a sound methodology.

Trench-Perpendicular and Seismic Profiles

A secondary test of the FAME delineations was made by comparing subduction
zone strike-perpendicular seismic profiles from the literature to profiles constructed
across the subduction zones using the ETOPO2 and GTOPO30 topographic and
bathymetric data. Ten profiles were constructed across each subduction zone at roughly
90° to the trench (See Appendix). In order to construct the profiles, the ETOPO2
elevation data were projected to UTM coordinate system in order to have meters as a unit
of horizontal spatial measurement as opposed to decimal degrees used as a unit of spatial
measurement in the geographic coordinate system native to the data. Profiles were
constructed with no vertical exaggeration in order to accentuate elements of the FAME.
The method by which the profiles were generated (the Krugh method) is described in
detail in the Appendix. Figure 31 shows the seismic profile across the west end of Java,
along with Java Profile 6, the 1:1 profile constructed from the DEMs, and a cross-section
across the subduction zone showing a model of the plate interface and locked zone.

F.A.M.E. Spatial Arrangement Measurements

The spatial arrangement of the FAME was measured directly in GIS. Each of the
FAME were modeled as a curved polyline using the methods described above, and then
measured via the following methods.
Profile Across the Javanese Subduction Zone

Volcanic Arc

Seamount (IFAB)

40 km

dots represent ANSS earthquake foci
Distances between the IFAH, OFAH, Trench, and Arc

Prior to measurement, the lines depicting the FAME were clipped so that they were approximately the same length. Since the FAMEs are modeled as sub-parallel non-orthogonal lines, ordinary techniques to measure linear offset were inapplicable. Therefore, the “Nearest Features” GIS tool (Jenness, 1996) was implemented to measure the average offset of the FAME from one another. The Nearest Features ArcView extension (see Appendix) calculates the distance and azimuth to the nearest feature of a line theme from a designated point on another line theme. This extension was used in conjunction with another extension that places points along a line feature (Lead, 2003) at either regular intervals or divides the entire length of the line into a specified number of points. Table 5 describes the spacing of points used to measure the FAME in each subduction zone. Once the points were generated (dots along FAME in Fig. 32), measurements of shortest proximity were made from the trench to the OFAH, the trench to the IFAH, the trench to the arc, the IFAH to the OFAH, the arc to the IFAH, and the arc to the trench.

Figure 32 - Aleutian subduction zone FAME spacing measurements, spaced at 200km along each FAME.
arc to the OFAH. Distances were then calculated from each of the points on the outer FAME to the nearest point on the inner FAME of the pair being measured (lines from trench to OFAH, IFAH to OFAH, etc. in Figure 32). The resulting measurements were then averaged to give the offset or spacing of the FAME to one another.

### Table 5 – Spacing of FAME Measurements

<table>
<thead>
<tr>
<th>Zone</th>
<th>Point Spacing</th>
<th>Number of Points/Feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alaska</td>
<td>200 km</td>
<td>8</td>
</tr>
<tr>
<td>Aleutian</td>
<td>200 km</td>
<td>5</td>
</tr>
<tr>
<td>Cascadia</td>
<td>Varies, ~48-55 km</td>
<td>20</td>
</tr>
<tr>
<td>Costa Rica</td>
<td>Varies, ~26-28 km</td>
<td>20</td>
</tr>
<tr>
<td>Java</td>
<td>200 km</td>
<td>7</td>
</tr>
<tr>
<td>Nankai</td>
<td>Varies, ~38-44 km</td>
<td>20</td>
</tr>
<tr>
<td>S. Chile</td>
<td>100 km</td>
<td>23</td>
</tr>
<tr>
<td>Sumatra</td>
<td>200 km</td>
<td>9</td>
</tr>
</tbody>
</table>
Results

The principal results of this study are twofold:

1. The creation of GIS databases containing geographic, geologic, and geophysical data describing the bathymetry, topography, great earthquake history, modern and past ground deformation, and therefore the best-estimate location and extent of subduction interplate coupling for eight subduction zones.

2. Measurement of the length and spatial arrangement of elements of forearc morphology in each of the subduction zones contained in the databases mentioned above.

GIS databases are by nature customizable; they contain information which can be displayed in an unlimited number of configurations in order to draw attention to certain spatial relationships and quantitative measurements. Because of the inability to carry this feature over to printed medium, the databases created in this study are found on DVD-ROMs in the Appendix of this document. Sources are found in Table 3.

Below are maps (Figures 33-41) derived from the GIS databases compiled for each subduction zone being examined in this study. The Appendix contains the data used to compile the databases, which can be loaded into a GIS database and displayed in combinations other than those of Figures 33-41. Furthermore, Figure 42 provides a side-by-side comparison of the FAMEs of the eight subduction zones, and Tables 6 and 7 show the results of the FAME measurements. The implications of each map along with the quality of the methodologies with which they were generated are discussed below in the following Discussion chapter.
Figure 33 - The locked zone of the Aleutian and Alaskan subduction zones, as defined in the Aleutian by the rupture extents of the historical large and great earthquakes, and in the Alaskan by rupture extent and tsunami models from the 1964 Mw 9.2 great earthquake, and heat flow measurements of Hyndman and Wang (1995). Sources are given in Table 4. Errors in rupture extent are on the order of +/- 10 km. Earthquake rupture and magnitude, and therefore the locked zone in the Aleutian subduction zone seems to continue almost to the volcanic arc, unlike the Alaskan subduction zone to the northwest, where rupture and the locked zone lies primarily between the IFAH and OFAH. This could be due to several geologic differences between the two zones, including shallower downgoing plate dip in Alaska, faster convergence in the Aleutians, less sediment influx in the Aleutians, or possibly the thinner and therefore weaker oceanic crust backstop in the Aleutians.
Figure 34 - (left) The Cascadia subduction zone, showing the locked zone as proposed by Hyndman and Wang (1995) and Fluck et al. (1997). The absence of great earthquakes in Cascadia makes it difficult to determine the location of the locked zone, although modelling of other geophysical data such as GPS ground deformation measured across the forearc, or heat flow data collected across the subduction zone aids in estimating its location. Much debate remains as to the location of the locked zone in Cascadia, as the many different available datasets do not agree in symphony as to where the locked zone should be located. See Discussion chapter for more thorough treatment of this enigmatic subduction zone. Right figure shows heat-flow model of Hyndman and Wang (1995).
Figure 35 - The Costa Rican subduction zone, showing the locked zone as historical great earthquake ruptures, most of which fall astride the IFAH and OFAH. Note the correlated segmentation of the surface morphology of the incoming Cocos Plate (smoother to the NW, rough to the SW near the Cocos Ridge) and the topographic variations along-strike in the subduction zone (Nicoya Peninsula in the NW, lack of prominent structural high in the central region, and the Osa Peninsula to the SE). Great earthquake rupture sources given in Table 4.
Figure 36 - The locked zone of the Java and Sumatra subduction zones, as defined in Sumatra by the rupture extents of the 1833 and 1881 great earthquakes (Zachariasen et al., 1999), and in Java loosely by the contours to the downgoing plate top, which identify the depth of 100 km of the downgoing plate to lie approximately under the volcanic arc. The downdip limit of the locked zone (dashed brown line) can be estimated from the historical earthquake ruptures as well as the depth and heat flow of the downgoing plate, which are estimated to be similar to those measured by Hyndman and Wang (1995) in Alaska, Nankai, and Cascadia. Note how the majority of the zone of locking falls between the IFAH and OFAH.
Figure 37 - The locked zone of the Nankai subduction zone, as defined by the GPS-measured deformation of the forearc, as modeled by Hyndman and Wang (1995). Colored lines indicate FAMEs as modeled in this study. Note how the majority of the zone of locking falls between the IFAH and OFAH, as when the locked zone is defined by other means.
Figure 38 - The locked zone of the Nankai subduction zone, as defined by the heat flow measurements of Hyndman and Wang (1995). Colored lines indicate FAMEs as modeled in this study. Note how the majority of the zone of locking falls between the IFAH and OFAH, as when the locked zone is defined by other means.
Figure 39 - The locked zone of the Nankai trough subduction zone, as defined by the rupture zones of the 1944 Mw 8.0 and the 1946 Mw 8.1 great earthquakes presented by Savage, 1995. Colored lines indicate FAMEs as modeled in this study. Note how the majority of the zone of rupture (and therefore implied locking) falls between the IFAH and OFAH, as when the locked zone is defined by other means.
Figure 40 - The southern Chile subduction zone, showing the locked zone as historical great earthquake ruptures, which all fall astride the IFAH and OFAH. A notable exception is the 1960 Mw 9.5 earthquake (red dashed line - slip inversion of Barrientos and Ward (1990) that ruptured the entire width of the Chilean forearc, past the trench to the seaward and just beyond the volcanic arc to the landward, for over 1000 km along strike. Tichelaar and Ruff (1991) describe coupling along sparts of the Chilean margin to the north as extending from depths of 48-53 km.
Figure 41 and Tables 6 and 7 - 1:18,000,000 Maps of Subduction Zones

Measurements of FAME

<table>
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Table 6 - FAME Length

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Discussion

Models of eight subduction zones have been compiled in GIS integrating first-order model-based primary geophysical observations of subduction interplate coupling with digital topographic and bathymetric data describing the surface morphology of the subduction zones' forearcs. These subduction zones were selected from a larger potential dataset of subduction zones because (1) they have experienced great earthquakes in historical time, or geological data suggests a pattern of great earthquakes over the past several thousand years, (2) they are for the most part well-constrained by a variety of geologic and geophysical data, and (3) they have characteristic forearc morphology.

Locked Zone Determinations

1) Historical great earthquake rupture

Of the four major methods for locked zone determination, historical great earthquake rupture extents are the least speculative, as they coincide with the locked zone location to the degree that the rupture patch for any given great earthquake represents the maximum possible extent of the locked zone. Whereas this method is robust, certain caveats must be considered with respect to determining the locked zone via this method. Published zones of great earthquake rupture are only as accurate as the methods of the studies which produced them, and therefore the accuracy of each subduction zone's historical great earthquake rupture extent depends upon the studies from which they were obtained. Estimates of error are given in Table 8.

Most researchers define subduction zone great earthquake rupture zones on the basis of a time-period grouping of geophysically related aftershocks following a large shallow underthrusting mainshock, such as during 12-hour, 24-hour, or 36-hour windows.
of aftershocks. While some of the sources used in this study stated the time-period of aftershocks, not all gave the source-time period of their rupture extent. Wherever possible, sources of great earthquake rupture are given in Table 3 and 4, and the reader is referred to these sources for detailed information about the earthquake rupture areas contained therein.

Following the fault model of Scholz (2002), it is likely that the locked zone of a subduction megathrust which generates a great earthquake is smaller in area than the rupture zone of that great earthquake. This is due to the fact that rupture often extends beyond the coupled zone of the plate interface, and propagates co-seismic strain through adjoining areas (transition zones) of the plate interface where rock mechanical behavior transitions from brittle to ductile above the updip and below the downdip limits to seismicity. Within the locked zone, rocks which experience velocity weakening accommodate great earthquake nucleation, which can lead to rupture extending beyond the bounds of this previously coupled area. Consequently, mapped great earthquake historical rupture areas represent the maximum extent of the pre-mainshock coupled zone, which is likely to be substantially smaller. The 1960 Chilean great earthquake provides an example of how the coupled zone of a megathrust was much smaller than the overall rupture area as mapped by aftershocks and ground deformation (Barrientos and Ward, 1990, Figure 56).

Tsunami waveform modeling in conjunction with tide gauge and inundation height data are often used to determine historical great earthquake rupture areas. This technique is quite useful in examining older great earthquakes, such as the Tonankai and Nankaido events of 1944 and 1946, which occurred prior to the era of high-resolution seismometers. Tsunami rupture model results, however, should be approached with some
caution because the rupture extent is determined from proxy data not directly related to
the actual mainshock.

Positional error on the order of ~10 km was introduced by digitizing rupture data
into the GIS databases. Published rupture zones themselves have errors from the
earthquake, tsunami, and deformational data upon which they are based, and errors in
digitizing are likely less than or on par with errors in the locations of the rupture zones
themselves. The quality with which historical great earthquake ruptures define the spatial
position and extent of the locked zone is somewhat variable. Because historical great
earthquake rupture areas define former locked zone extents within the limits of rock
mechanical behavior, they are useful in identifying previously locked patches of
subduction megathrusts. Utilizing them to define potential future rupture areas, while
possible, should be undertaken with caution and should be viewed in the light of other
means of locked zone determination.

(2) Heat Flow / Thermal

Temperatures of crustal materials within the subduction system are often inverted
from measurements of heat flow taken at or near the earth’s surface in transects across
the forearc region. Some measurements are made from wells which penetrate a few
hundreds of meters into the uppermost crust. Others are extrapolated from laboratory
analyses of samples from rock outcrops assuming that exposed rocks are similar to those
found at depth. Seismic refraction surveys and sporadic deep-sea drill core samples
provide some evidence as to the rock types found above and somewhat close to the plate
interface, as do offshore marine geologic samples, but these are second-order estimates of
actual geologic conditions on the plate interface. Heat-flow measurements are often
applied to subduction zones without due consideration of the impact of fluid-flow
conditions along the plate interface, particularly in the fluid-saturated accretionary prism where some researchers place the updip limit to the seismogenic zone. As early as 1979, Seeley noted the water-retention ability of young subducted sediments examined in the accretionary prism of the Cascadia forearc. The elevated pore-fluid pressures found in this part of the subduction system can drastically alter the thermo-mechanical behavior of the rocks at depth along the megathrust. Heat transport via pore-dwelling groundwater alters the mechanical behavior of the rocks there, and therefore casts some uncertainty on heat-flow based locked zone determinations, particularly of the updip limit to the seismogenic zone.

Heat flow measurements done at the surface are a measure of the total flux out of the crust and can only be used as supporting evidence for locked zone location by making assumptions about plate and fault zone rheology. Furthermore, heat flow calculations are usually made along a few select transects across the deformation front and forearc of subduction zones, and then interpolated along the strike-length of the subduction zone. The heat flow data used in this study should therefore be approached with caution, and each particular dataset (e.g. Hyndman and Wang, 1995; Wang et al., 1995; Nakanishi et al., 2002, etc.) should be closely examined when considering the involved errors in measurements and applied as a proxy method of delineating the locked zone.

(3) Geodesy-based Measurement of Forearc Deformation

Time-series measurements of crustal geodetic station velocities and subsequently, determinations of locked zone position and extent are only as accurate as the deformational data themselves. The quality of geodetic data is subject to a variety of errors, including but not limited to satellite position, signal strength, quality of differential correction, and post-processing procedures, all integral parts of GPS-geodesy. Older,
more traditional techniques often occur less frequently, and although highly accurate, are more costly and time-consuming to acquire. Geodetic constraint of the surface velocity field and locked zone position is modeled from these data and fault models. To date, there has not been a great earthquake in a region with an existing GPS network that would serve to test the ability of GPS to accurately model plate coupling in subduction zones. (The M₉ 8.3 earthquake which occurred ~13 kilometers deep along the northern Japanese subduction zone on September 25th, 2003, occurred in a region with excellent GPS coverage, and will likely be the first such earthquake to examine the ability of GPS-based models to describe plate coupling in detail). Numerous studies use velocity field data to identify coupled portions of megathrusts below subduction forearcs (Svarc et al., 2002; McCaffrey et al., 2000; Savage, 1995; Zweck et al., 2002; Fluck et al., 1997; McCaffrey et al., 2001, and many others). Some of these studies do so by comparing interseismic non-GPS geodetic measurements to co- and post-seismic ground motion response (Figure 10) in regions which have experienced larger earthquakes.

Determinations of the locked zone via geodetic measurements, then, are taken with some confidence in locations with high-density recording of ground motion and ample amounts of corroborating geophysical data describing the status of relative plate motion and coupling (i.e. Nankai).

Perhaps the most glaring drawback to using geodetic data to model the locked zone stems from the inability of the data to constrain velocity of offshore regions. As hard as modelers try to postulate about the status of plate coupling across an entire subduction zone via velocity-field geodesy, there remains a disconnect between offshore upper-plate response to subduction conditions and onshore plate motions. A full spectrum of model locked zone geometries can be called upon to successfully explain the
same set of observed GPS data, highlighting the inadequacy of exclusively onshore
géodetic data to model plate coupling on and off shore. Only through the integration and
inclusion of seafloor geologic structures and strain accommodation (e.g. Goldfinger et al.,
1992 and 1997) can conditions on the offshore portions of convergent margins be
examined thoroughly via forearc deformational data.

(4) Upper Plate Seismicity

Using the lowest depth of earthquakes purportedly located in the backstop region
of the upper plate (Figure 9) as a proxy for the location of the downdip limit to the
seismogenic zone along the adjacent megathrust is inexact for several reasons. First, the
locations of the earthquakes found in publicly available earthquake catalogs are subject to
the quality of the methodology with which they are generated, particularly due to
uncertainties in the earth model used to locate the earthquakes. Therefore, each dataset
(see Table 3) has its own quality and propensity for error. Secondly, in a fashion similar to
heat flow-based locked zone determinations, upper plate seismicity data are removed
from actual conditions along the plate interface in that they behave in a manner subject to
conditions only on the upper half of the fault. Therefore, this method is perhaps the most
indirect of the four locked zone determinations, as it rests wholly on the location of
earthquakes whose focal mechanisms are controlled by the transition to stable sliding-
type behavior assumed to be at a similar depth as the downdip end of the locked zone.
The depths of such earthquakes are often hard to measure accurately, given uncertainties
in knowledge about the lithologies at those depths along with lateral heterogeneities in
lithology along the upper plate contact with the subducting slab at the depth of the
downdip edge of the seismogenic zone.
Accurate relocations of numerous upper-plate earthquakes along with knowledge of the focal mechanisms of those earthquakes and their precise depth location within the subduction system is required to utilize this methodology. Furthermore, data would have to be carefully selected so as to differentiate upper plate convergence-related tectonic seismicity from magmatic and plutonic earthquakes. Unfortunately, seismic data of sufficient quality and uniformity was not available for all of the eight subduction zones examined, and therefore this method was not used in this study as a primary means of locating the locked zones.

**Locked Zone Definition**

Some debate remains as to the precise meaning of the term "locked zone." Essentially, the locked zone concept is a conceptual model to explain plate coupling, strain accumulation and earthquake generation. Most researchers define the locked zone as the part of the plate interface that accumulates strain during interseismic periods and releases that strain via a great earthquake. Aftershocks, tsunami models, or historical earthquake ruptures are commonly used to locate the locked zone; no attempt to differentiate between rupture, earthquake nucleation, and coupling is typically made, despite the fact that each may occur in spatially different locations. It remains difficult to differentiate, for example, between a locked zone comprised of a collection of asperities, or of an extensive fault plane with low friction.

Modeling of plate coupling from GPS data remains limited by the lack of data in offshore areas, which overly a substantial width of the coupled portion of subduction plate interfaces, and from heat flow data by the accuracy and model-dependence of their interpretation. Some facts remain uncontested, however: not all of the megathrust stores strain, and not all of it ruptures during earthquakes. There must be, then, a region of the
plate interface that (1) elastically stores strain, (2) releases that strain as an earthquake, including (3) an area of earthquake nucleation. Locked zone, seismogenic zone, and zone of earthquake nucleation are not synonyms and adequate definition of the "locked zone" requires better understanding of the physical differences between what part of the plate interface is strong (locked) versus the region of earthquake nucleation in subduction zones.

Delineation and Measurement of Forearc Morphologic Elements (FAMEs)

FAME Delineations

Delineations of the FAME in this study are estimates based upon large-scale subduction zone-length topographic and bathymetric morphological features. In reality, subduction zone FAMEs such as the IFAH consist of an entire mountain range such as

Figure 42 - Northern Oregon segment of Cascadia subduction zone, showing discrepancy between trace of Willamette Basin (derived from a high resolution DEM) and estimated axis of inner forearc high (Coast Range). Note how watershed boundary is highly irregular as a result of being controlled by geomorphologic processes as opposed to "average" inner forearc high trace.
the coast range of western Oregon (Figure 42, below). To facilitate spatial quantification, FAMEs are presented in this study as curvilinear arcs of a uniform length; this simplification facilitates quantification of their spatial relationships along a chosen part of each subduction zone.

Some error on the order of a few kilometers is intrinsic to the delineation of mountain range-scale structural highs by curvilinear arcs due to the inexact nature of depicting a mountain range with a line. A more systematic estimate of this error would begin with consideration of the resolution of the digital topographic and bathymetric data from which the FAMEs were delineated. The sub-oceanic portion of the forearc was modeled with DEM data having a resolution on the order of several kilometers, whereas the onshore DEM resolution is ~1 km. Therefore, error in the placement of the arcs used to delineate the FAMEs on the order of ~10 km are within reason, with the error in tracing the volcanic arcs and the deformation front (trenches) less than that involved in drawing the trace of the IFAH and OFAH because of the nature of these FAMEs.

Furthermore, offshore sediment deposition has been known to obscure forearc morphologic features, such as the 7 km thick sedimentary basins in Cascadia (Trehu et al., 1995) or the five accretionary basins of the Nankai trough (Sugiyama, 1994). In the case of Cascadia, the break-in-slope between the continental margin and the slope leading downwards to the deformation front was used as a rough marker of the axis of the OFAH. Wherever possible, seismic profiles were used to delineate FAMEs as a double check of the topographic and bathymetric interpretations. Wells et al. (2003) are currently investigating possible relationships between these basins and great earthquakes in the Nankai trough subduction zone following the work of Sugiyama (1994), which will be discussed below.
Quantitative Means of Delineating the FAMEs – Why They Were Not Used

An alternative to estimating or tracing the approximate location of the axes of FAMEs such as the IFAH and OFAH would be to use spatio-analytical techniques of GIS to delineate the forearc highs as arcs following the drainage divides between the inner and outer forearc basins (Figure 42) as shown by the digital topographic and bathymetric data. This method was not utilized, however, as the drainage divides were often far from linear and are not nearly parallel to the trench axis in the way the overall forearc high trends. Despite resulting indirectly from subduction-related tectonic uplift, the jagged nature of the watershed boundary lines results mostly from basin-scale erosion and other geomorphologic processes too minute to consider at the scale of this analysis.

In addition to basin modeling, numerical derivations of DEM data were considered to more quantitatively identify linear topographic and bathymetric features across the subduction zones’ forearcs (See Figure 30). These methods lead to grid-coarsening and the eventual delineation of the FAMEs in almost exactly the same locations as when they were estimated from the subduction-zone scale topographic and bathymetric data, and so were not utilized.

Magnitudes of Error

A summary of the errors inherent in each of the methodologies used in this analysis follows in Table 8, and errors intrinsic to specific sources used in this study are found in Figures 32-39 and Table 4 (great earthquake rupture areas).
Table 8 – Spatial Scale of Subduction Zone Features and Errors in Measurement

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<td>GPS Deformation Grids (tabular)</td>
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<td>Upper Plate Earthquakes</td>
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<td>Plate Depths</td>
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Discussion of Questions Posed at the Beginning of This Study

1. Where are the regions of interplate coupling and earthquake generation in relation to the principal forearc topographic elements?

Evidence from the Alaskan, Aleutian, Costa Rican, Javanese, Nankai, Sumatran, and South Chilean subduction zones suggests that the majority of subduction interplate coupling regions lie primarily below the inner and outer structural highs (IFAH and OFAH, respectively) of the subduction zones’ forearcs, within +/- ~20 km (See Figures 33-42 in the Results chapter). Of all of the subduction zones examined, perhaps the best demonstration of this correlation comes from the Nankai trough of Southwestern Japan. Figures 36 through 38 depict the Nankai trough locked zone as modeled via historical great earthquake rupture, heat flow, and GPS-based forearc deformation, demonstrating the methods’ agreement in placing the locked zone mostly underneath the inner and outer forearc highs. Not all examined subduction zones, however, clearly demonstrated the spatial correlation between their locked zone and their FAMEs. Because the locked zone location is still debated in the Cascadia convergent margin, the relationship between its FAMEs and its seismogenic zone remains nebulous.
Cascadia – Where is the Locked Zone?

The Cascadia subduction zone (Figure 33) remains enigmatic in that it has yet to experience a historical great earthquake, which would reveal the previously locked portion of its megathrust. Because of this lack of historical seismicity, secondary observations must be used to predict the position of the locked zone in Cascadia.

Existing Plate Coupling Models of Cascadia

In most recent literature, heat flow and GPS-geodesy remain the most prevalent techniques utilized to model the locked zone of most subduction zones, including Cascadia. Heat flow based estimates of Cascadia plate coupling (Figure 43) place the locked zone mostly offshore, beginning at the deformation front and extending landward transitionally to the coastline. GPS-based modeling of Cascadia’s locked zone (Figure 19) delineates the locked zone in a similar fashion, although some models (Stanley and Villasenor, 2001; McCaffrey et al., 2000) allude to the presence of a double locked zone lying both offshore below the accretionary prism, and also under the
eastern forearc basin (Figures 19, 44). The locked zone placement suggested by these models differs significantly in its correlation with the FAMEs from the other seven subduction zones examined.

2. Are those first-order forearc topographic elements linked genetically to the seismogenic zone?

The linkage between FAMEs (first-order topographic elements) and the underlying locked zone appears to be genetic in two ways: first, subduction parameters such as plate dip and convergence rate control the width of the plate interface (downdip widening). Secondly, FAMEs are linked via the structural and tectonic components of the subduction forearc, which provide a direct geologic link between conditions along the plate interface and the morphology of the upper plate.

Relationship between FAMEs and Seismogenic Zone via Basic Subduction Parameters

The spatial arrangement of the FAME are systematically related to the fundamental parameters of subduction. Figure 45 shows three of the FAME spatial measurements done at eight subduction zones plotted against published estimates of...
Figure 45 - FAME Spatial Relationships to Plate Dip

Figure 45 (a) - Arc-Trench, or forearc width (m) vs. plate dip, showing excellent correlation between steeper subduction and narrower forearcs.

Figure 45 (b) - Trench-Outer forearc high distance (m) vs. plate dip, showing reasonable correlation between steeper subduction and narrower outer forearc basins.

Figure 45 (c) - Inner to outer forearc high spacing (m) vs. plate dip, showing weak correlation between subduction angle and width of inner and outer forearc basins to one another.
subducting plate dip. As expected, wider forearcs, such as Cascadia, Costa Rica, and Alaska show a strong correlation with shallower dip. Figure 45a shows the excellent correlation between arc-trench distance and plate dip: wide arc-trench subduction zones such as Alaska and Nankai have shallower dips, whereas narrower forearcs such as the Aleutians or Costa Rica have steeper plate dips. Structurally, this phenomenon can be explained by the propensity for the shallower subducting plate to have more mechanical contact with the upper plate, propagating strain and therefore deformation across a wider geographic area of the overlying forearc region of the upper plate. This pattern also has implications for seismogenic zone width as seen in Alaska, which has the widest forearc examined in this study.

The correlation between the trench and the OFAH in the eight subduction zones (Figure 45b) also suggests a genetic relationship between FAMEs and the seismogenic zone. While the correlation is not as strong as the arc-trench spacing, it nevertheless shows that downgoing plate dip exerts a direct influence on the trench slope region that marks the seaward edge of the outer forearc basin.

**Figure 46** - Block cartoon of outer forearc basin developing from seaward verging thrust sheets of an accretionary prism, showing OFAH (orange) and IFAH (blue). From Sugiyama, 1994.

Outer forearc basin width (OFAH-IFAH spacing, Figure 45c) shows a weak correlation with plate dip. This is probably due to the fact that outer forearc basin width has
more to do with the structural regime in that particular forearc accretionary region and sediment influx than with downgoing plate dip. The correlation is strong enough ($R^2 = 0.4$) to suggest, however, that although not as straightforward as arc-trench spacing, the links between OFAB wideness and plate dip are still somewhat genetic. Wells et al., (2003) built on the work of Sugiyama (1994) in suggesting that the outer forearc basins mark large asperities on the plate interface that can be used to infer areas of strong coupling. Figure 46 shows a schematic block diagram through the accretionary prism of the Nankai trough in the region of the 1946 $M_w 8.1$ Nankaido great earthquake.

**Relationships between Forearc Structure and Seismogenic Zone**

The FAMEs of the eight subduction zones examined are linked genetically to the seismogenic zone beneath them through the structures found in the subduction zones’ forearcs. As plate convergence and subsequently subduction continue, coupling between the two plates propagates strain upwards into the upper plate, which manifests itself in deformation of the upper plate. In their study of the Central Aleutian forearc in 1989, Ryan and Scholl note that the formation of outer-arc structural highs and forearc basins depends on the manner in which the accretionary wedge is attached to the older bedrock framework of the arc. They go on to emphasize how the backstop geometry controls not only the structural style of the outer-arc high but also the structural geologic behavior of the accretionary prism, as did Davis (1988) and Byre and Hibbard (1978). The vergence of structures in the forearc, as well as the orientation of the upper plate backstop have a significant influence on the mechanics of the accretionary prism, and particularly on how it interacts with the upper surface of the downgoing plate in the seismogenic zone.
**Cascadia**

In Cascadia, there are three different areas of vergence – the northern region of the forearc verges seaward, whereas the central segment in southern Washington and northern Oregon verges landward, and the vergence returns to seaward in central and southern Oregon, and northern California. Landward verging forearc structures in the presence of a seaward verging backstop consisting of dense, mafic, rigid material buttressed against the craton of North America (Figure 44) acts much like a bulldozer blade, scraping the sedimentary input into thrust sheets, and creating a decollement overlain by kilometers of fluid rich sediment which likely has virtually no basal shear stress between its lowermost regions and the downgoing plate sliding smoothly below. Only the interaction between the Siletz terrane (IFAH and backstop in central Oregon) or a similarly dense, package of rocks and the downgoing plate, then, could generate enough brittle-rock interaction to generate a great earthquake. If this scenario is true, then the locked zone in Cascadia as inferred from thermal modeling (Hyndman and Wang, 1995) is located beneath a portion of the forearc that is relatively weak in comparison with that beneath the IFAH.

Mapping of seafloor structural geology off central Oregon suggests that the outer accretionary wedge in Cascadia is decoupled from the downgoing plate due to high fluid pressures along the interface (Goldfinger, 1992; 1997; pers. comm. 2003; Trehu et al., 1997). Structural studies of seafloor folds and faults in the upper accretionary prism in Cascadia (Goldfinger et al. 1992) suggest a broad zone of coupling extending from the mid-slope region of the Cascadia subduction zone landward at least to the coastline. A lack of measured seismicity on any of the structures suggests that motion along the structures and growth of folds must occur concurrently with large earthquakes which
rupture large portions or the entire margin. More detailed examination of the folds and faults by Goldfinger et al. (1997) found that the Cascadia accretionary wedge is rotating and translating northward in response to the component of Juan de Fuca/North America plate convergence that is tangential to the overall convergence. Nevertheless, proponents of the offshore locked zone in Cascadia have argued that the high sediment influx into the accretionary prism along with the young, hot downgoing plate justify placing the locked zone far offshore. These observations argue strongly against the heat-flow based locked zone models for Cascadia which place regions of strong coupling far seaward into the accretionary prism.

Other subduction zones

Seismic profiles across the Javanese subduction forearc (Figure 31, Kopp et al., 2002) and Sumatran forearc (Kopp et al., 2003) show the outer forearc high as an actively folding and faulting landward-verging sedimentary mélangé compressed against the inner forearc high. The present seaward-verging backstop is a fossil accretionary prism against which the active prism is pinned (Kopp et al., 2003). Seismic profiles of the Central Chilean subduction zone (Laursen et al., 2002) also show similar thrust faults adjacent to the deepwater Valparaiso outer-forearc basin. Deformation in these regions occurs predominantly along imbricate thrust faults whose vergence varies, propagating the positions of the forearc highs seaward over time as the arc migrates landward, widening the overall forearc. Seismic profiles across the Alaskan and Aleutian (Figure 16) subduction zones show seaward verging structures pinned against a seaward-verging backstop, whereas profiles across the Costa Rican margin (Figure 21) show similar structures.
The genetic relationship between FAMEs in subduction zones and the underlying seismogenic locked is manifested via the structural features that develop in the accretionary prism and upper plate. This relationship accounts for the good correlation of the OFAB with the locked zone observed at seven of the eight subduction zones, and possibly even Cascadia. Furthermore, this observation is corroborated by the ongoing work of Wells et al., (2002) which points out the correlation of IFAB subsidence with seismogenic portions of the Nankai subduction zone.

3. Does morphologic analysis provide a predictive framework for understanding downdip changes in the mechanical behavior of the plate interface?

Making predictions about the mechanical behavior of the plate interface based solely on the surface morphology of the upper plate above the seismogenic zone is possible, but should be done in conjunction with other supporting geophysical observations. Downdip changes in the mechanical behavior of the plate interface are largely dependent upon a variety of conditions along the interface, such as temperature, rheology, asperity and barrier distribution, stress regimes imparted from previous earthquakes, sediment input and the fluid content of that sediment, as well as radiogenic heat production, and simpler properties of the interface, like the dip angle of the downgoing plate, or the rate at which the plate convergence occurs.

Because of the genetic structural links that appear to exist between FAMEs in subduction zones and the seismogenic plate interface beneath them, analysis of the morphology of subduction zone forearcs can provide information about plate coupling and thus the condition of the locked zone. Song and Simons (2003) used gravity and morphologic analysis of onshore subduction forearcs at the same convergent margins examined in this study, and found that morphologic analysis can be used to constrain
along-strike variations in forearc topography are significant in controlling seismogenic behavior.

Clear links exist between the roughness of downgoing plates and both the seismic coupling and the spatial arrangement of the FAMEs above those regions. Documentation exists for several subducted seamounts having a profound influence on coupling (Ruff, 1992), earthquake generation (Abercrombie et al., 2001), and FAME character (Graefe et al., 2002). Some researchers (Ruff, 1992) believe that asperities play a strong role in coupling, whereas others (Wang, 1995) call upon small amounts of basal shear stress distributed across a wide area as mainly responsible for low coupling yet elevated seismogenic potential.

Possibility of Very Low Coupling Stress – Wang et al. (1995)

Wang et al. (1995) presented evidence for very low coupling stress along the megathrust beneath Cascadia, noting that evidence exists which points to high pore-fluid pressures, and therefore very low shear stress along the megathrust in northern Cascadia. Furthermore, they note that the notion of a weak thrust fault does not necessarily stand in disagreement with the idea that the plates are storing strain, building up to a future great earthquake. One of the major tenets upon which Wang et al. (1995) base their argument for weak coupling along the Cascadia megathrust is the location of the locked portion of the plate interface beginning near the continental slope and shelf and terminating near the coast. Wang et al. (1995) call upon near-lithostatic pore-fluid pressure in conjunction with a stress regime in Cascadia where convergence-direction stress is similar in magnitude to vertical stress, and the majority of strain is arc-parallel to support the fact that the coupling stress on the fault is very low.
Regardless of the true position of the locked zone in Cascadia, morphologic analysis of subduction zones is a powerful tool that, if used in conjunction with corroborating geophysical data, can provide information about the status of coupling mechanics on the megathrust.
Conclusions

(1) Evidence from seven of the eight subduction zones examined in this study suggests that the regions of subduction interplate coupling lie below the inner and outer structural highs (IFAH and OFAH, respectively) of the subduction zones' forearcs, regardless of the means by which the locked zone location is determined.

Cascadia is anomalous in this regard in that its locked zone, when modeled with heat flow data (Hyndman and Wang, 1995; subsequent studies based upon the same models), lies offshore between the deformation front and the outer forearc high. GPS forearc deformational data, on the other hand, models a double locked zone both between the trench and the OFAH, and appearing again beneath the western portion of the Willamette Valley (inner forearc basin) in Oregon, and the Seattle basin in Washington (Stanley and Villasenor, 2001; McCaffrey et al, 2000; etc). GPS data, however, cannot model coupling conditions for offshore areas.

Furthermore, heat flow data do not take into account the possibility of low basal shear stresses throughout the Cascadia accretionary prism as suggested by the vergence of prism structures and the backstop, as well as elevated pore-fluid pressures. At best, heat flow data are a proxy for plate locking several orders of magnitude removed in spatial scale from actual conditions on the plate interface, and should accordingly be viewed with some skepticism unless substantiated by more direct measurements. Although morphologic analysis of the FAMEs and locked zones at the other seven subduction zones would suggest that the locked zone in Cascadia lies underneath its OFAB, more work is needed to pin down the location of the locked zone in Cascadia.
(2) Forearc morphologic element (FAME) spacing and basic subduction parameters appear to be genetically linked. Steeper downgoing plate dip, for example, occurs in subduction zones whose forearc are the least widely spaced, whereas shallower plate interface dips are found in the widest subduction zones. While less strongly correlated, OFAB width (IFAH-OFAH spacing) also relates to plate dip positively. The primary means by which FAME and the seismogenic zone of the subduction plate interface are linked are the structural and tectonic faults and features of the forearc region. The conspicuous package of paired highs and basins that appears in all eight subduction zones owe their existence to the transmission of strain from the underlying subduction megathrust in both inter- and coseismic periods.

(3) Making predictions of the mechanical behavior of the plate interface based solely on the surface morphology of the upper plate above the seismogenic zone is possible, but should be done in conjunction with other supporting geophysical observations. Nevertheless, this represents an oft-overlooked tool to predict coupling behavior, and if methodologies can be refined to isolate true seafloor structure from overlying sediment cover (the ETOPO2 Bathymetric data used in this study cannot do this) then this may prove to be a powerful tool of prediction.

Locked zone, seismogenic zone, great earthquake rupture area, coupled zone, and region of earthquake nucleation are fundamentally based on different mechanical concepts, and do not necessarily refer to the same spatial/physical region of a subduction megathrust. Care must be taken to avoid using these terms interchangeably.
In summary, clear correlations exist between the morphology of the forearc regions of subduction zones and the conditions on the underlying plate interface with respect to seismic coupling and great earthquake behavior.

Advancement of this research might involve application of the upper-plate seismicity method to delineate the locked zone. Although this would include significant efforts in gathering a high-resolution catalogue of earthquakes, it would provide additional control on the locked zone location. Inclusion of published data on the response of the northern Japanese forearc to the great earthquake of September 24\textsuperscript{th}, 2003 would provide an excellent first-hand look into the ability of GPS data to model plate coupling in subduction zones before, during, and after great earthquakes.
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Appendix

List of Acronyms

1. FAME – Forearc morphologic element
2. FAMEs – Forearc morphologic elements
3. IFAH – Inner Forearc High
4. OFAH – Outer Forearc High
5. EQ – Earthquake
6. GEQ – Great Earthquake (M_w > 8.0)
7. ETOPO2 – Global Topography and Bathymetry 2 minute elevation data
8. GTOPO30 – Global topography 30 arc second elevation data
9. SRTM30 – Shuttle Radar Topography Mission 30 arc second elevation data
10. DCW – Digital Chart of the World

Earthquake Catalogues

- NEIC (National Earthquake Information Center) http://neic.usgs.gov/
- ANSS (Advanced National Seismic System) http://www.anss.org/
- Harvard CMT (Central Moment Tensor) Catalogue
  http://www.seismology.harvard.edu/projects/CMT/

Topography and Bathymetry Data

- ETOPO2 – http://www.ngdc.noaa.gov/mgg/fliers/01mrg04.html

A thorough description of the procedures used in generating this data are found at the following URL:

http://www.ngdc.noaa.gov/mgg/bathymetry/predicted/explore.HTML

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• SRTM30 - http://www.jpl.nasa.gov/srtm/dataprod.htm

GIS Methods
• Nearest Features Extension -
  http://www.jennessent.com/arcview/nearest_features.htm
• Trench-Perpendicular Seismic Profiles

These profiles were constructed by a variety of methods. First, the trenches were delineated via the methodology described in the “Methods Chapter.” Next, a line was drawn perpendicular to the trench axis. Approximately ten lines per trench were drawn. Then, a custom Visual Basic button action script was generated within ArcGIS which queried the underlying ETOPO2 DEM for the UTM northing, easting, and elevation value of each cell underlying the line. These values were exported to tables in Microsoft Excel, and then graphed in Surfer.

• Krugh Method – This method is a custom spreadsheet written in Microsoft Excel by William C. Krugh (william.krugh@erdw.ethz.ch) which takes a collection of data points with X,Y, and Z values, and queries the dataset to create a spatial subset of data lying within a specified distance on either side of a specified line. In this study, this method was used to add earthquake epicenters (as seen on Figure 31) lying within some distance (usually 20 km on each side) of each profile line. The finished product including the profile lines and the earthquake data subsets were graphed in Surfer and manipulated in Adobe Illustrator so as to include seismic profile raster images. Feel free to contact me for further details (kayeg@geo.orst.edu).