Thermal models of flat subduction and the rupture zone of great subduction earthquakes

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[1] In subduction zones, the size of the seismogenic zone that ruptures during a great thrust earthquake may be thermally controlled. We have constructed finite element thermal models of six transects across three different subduction zones in order to determine the temperature distribution along the plate interface and to predict the size of the seismogenic zone. These models incorporate the complex plate geometries (variable dip downsection) necessary to model a flat slab subduction style. We focus on the rupture zones of the great earthquakes of Nankai (SW Japan) 1946 M8.3 and Alaska 1964 M9.2 as well as on the Cascadia margin. Subduction zone segments with moderate to steep dips exhibit rupture zones of 100–150 km downdip width, consistent with earlier elastic dislocation models and thermal models. For shallow dipping flat slab segments, our models predict larger locked zones, of 150–250 km width, in good agreement with aftershock and geodetic studies. The wider seismogenic zone predicted for flat slab segments results from an uncommonly wide, cold, forearc region, as corroborated by surface heat flow observations. A global analysis of great M > 8 interplate earthquakes of the 20th century reveals that more than a third of these events occurred in flat slab segments, whereas these segments represent only 10% of modern convergent margins. This implies substantially higher interplate coupling for flat slab segments, likely due to the increased downdip extent of the seismogenic zone, and suggests that the seismic risk near such regions may be higher than previously thought.

INDEX TERMS: 3040 Marine Geology and Geophysics: Plate tectonics (8150, 8155, 8157, 8158); 7209 Seismology: Earthquake dynamics and mechanics; 7220 Seismology: Oceanic crust; 7223 Seismology: Seismic hazard assessment and prediction


1. Introduction

[2] Great subduction earthquakes have been the subject of intense study and much debate during the past three decades [Kanamori, 1972, 1977; Ruff and Kanamori, 1983; Thatcher and Rundle, 1984; Byrne et al., 1988; Tichelaar and Ruff, 1993; Hyndman and Wang, 1995; Oleskevich et al., 1999]. More importantly, they pose a substantial risk to population centers along active margins, such as most of the Pacific Rim. Among the most cited examples are the recurrent great earthquakes along the Nankai trough of SW Japan as last expressed by the twin Tonankai 1944 M8.1 and Nankaido 1946 M8.3 earthquakes [Sagiya and Thatcher, 1999, and references therein] and the great Alaska earthquake of 1964 M9.2, the second strongest earthquake ever recorded [Johnson et al., 1996, and references therein].

[3] The interplate surface, which ruptures during a great subduction earthquake, is said to be “seismogenic,” capable of generating earthquakes by elastic failure. In cases of high or complete interplate coupling the rupture zone is considered to be locked during the interseismic part of the earthquake cycle and to release the accumulated relative plate motion of one complete seismic cycle within the time span of about 1 min or less (Figure 1). For recurrence periods of several hundred years and plate motions on the order of 5 cm/yr, this can translate to over 10 m of slip. Indeed average slip values reported for the 1964 M9.2 Alaska earthquake are 5–10 m with a local maximum of over 20 m [Johnson et al., 1996]. Seismic moment is proportional to the area of the rupture surface and to the slip [Wells and Coppersmith, 1994]. In addition to generating a larger earthquake, a greater downdip width means the rupture zone may extend landward of the coast, and thus implies closer proximity to population centers. Therefore, accurate estimates of the dimensions of a locked interplate zone are required for an accurate assessment of seismic risk.

[4] This paper investigates the rupture zone of the 1946 Nankai M8.3 and the 1964 Alaska M9.2 events. For these two margins the dip of the subducting plate actually
decreases beneath the forearc before increasing again near the volcanic arc and corresponds to a flat slab plate geometry [Gutscher et al., 2000]. This type of plate geometry has a major impact on the position of the convecting asthenospheric corner (Figure 2) and can be expected to strongly affect the thermal structure of the subduction zone forearc. The thermal models presented here for the first time investigate the influence of a flat slab plate geometry on the width of the seismogenic zone. We construct numerical models of the subduction zone thermal structure in order to calculate the temperature distribution along the plate interface and thereby determine limits of the seismogenic zone. We also examine the Cascadia margin of the northwestern United States where a substantial locked interplate zone exists [Dragert et al., 1994; Hyndman and Wang, 1995; Satake et al., 1996; Clague, 1997] and where the subducting plate has a low angle or flat slab geometry [Crosson and Owens, 1987; Gutscher et al., 2000].

2. The Seismogenic Zone

[5] The rupture zone of an interplate earthquake is most reliably determined using source time studies or by the distribution of aftershocks. As these observations are not always available for a given margin, a variety of complimentary methods have been developed to estimate the areal extent of the seismogenic zone in different subduction zones. These include geodetic studies (of coseismic and postseismic elastic deformation), tsunami waveform modeling, thermal modeling and marine seismic surveys. Comparative studies indicate that on a global average, the downdip limit of the rupture zone appears to be 40 ± 5 km [Tichelaar and Ruff, 1993]. Among the most common methods for determining the dimensions of the locked zone are geodetic and thermal modeling.
3. Numerical Modeling

[5] Geodetic modeling is based on the elastic dislocation model which suggests that stress builds up during the interseismic phase of the earthquake cycle along the locked zone [Savage, 1983]. This results in an interseismic subsidence signal (between the trench and the coast) and an uplift signal (near the downdip limit) for vertical motions with an amplitude of a few millimeters per year. During rupture (for the purely elastic plate model) these signals are reversed and sudden uplift of several meters occurs offshore (sometimes reaching the coast) simultaneously with subsidence inland (Figure 1). Horizontal motions can also be modeled and this approach has become popular with the great expansion in the number of precise GPS (Global Positioning System) stations and the increase in the period of observation. Such studies applied to the Nankai Trough suggest high to complete coupling along the locked zone [Le Pichon et al., 1998; Ozawa et al., 1999; Mazzotti et al., 2000].

[7] The downdip limit of the seismogenic zone marks the transition from frictional, stick-slip behavior to quasi-plastic, aseismic behavior. In continental fault zones, the base of the seismogenic zone occurs at approximately 15 km depth corresponding to temperatures of 300°C–350°C and the onset of quartz plasticity [Chen and Molnar, 1983; Schol, 1984; Schol, 1990]. In many subduction zones, the downdip limit of the seismogenic zone occurs at 40 ± 5 km depth where estimated temperatures range from 250°C (constant shear stresses) to 400°C (shear stresses increasing with depth and a thick overlying crust) or even 550°C (shear stresses increasing with depth and a thin overlying crust) [Tichelaar and Ruff, 1993]. Hyndman and coworkers proposed that large subduction thrust earthquakes are initiated at T < 350°C, with ruptures able to extend downdip to 450°C [Hyndman and Wang, 1993, 1995; Hyndman et al., 1997]. In this paper, we follow Hyndman’s work and use the 350°C isotherm with a transition to 450°C as marking the downdip limit of the seismogenic zone. We note, however, that the downdip limit may be controlled by factors other than temperature, such as the composition of rocks along the plate interface. Specifically, in places where the downdip limit corresponds to the forearc Moho, deeper aseismic behavior may be a result of aseismic serpentinite in the forearc mantle [Hyndman et al., 1997; Peacock and Hyndman, 1999; Oleskevich et al., 1999].

[8] The updip limit of the seismogenic zone typically occurs beneath the accretionary wedge, where high porosity, unconsolidated sediments, with high pore fluid pressure are present [Byrne et al., 1988]. Some researchers have suggested a thermal control for the updip limit caused by the transformation of ductilely deforming clay minerals (smectite) to illite and chlorite which exhibit stick-slip elastic behavior and are considered to rupture seismically [Frorlijk, 1990], though smectite is not abundant in all subduction zones. These transformations occur at a temperature of 100°C–150°C which is thus proposed as marking the updip limit of the seismogenic zone [Hyndman et al., 1995; Oleskevich et al., 1999].

4. The Nankai Great Earthquakes

[12] The Nankai trough is the site of recurrent great earthquakes (1707 Hoei, 1854 Ansei I and II, 1944 Tonankai and 1946 Nankaido) with an unparalleled historical documentation dating back to 684 A.D. [Ando, 1975; Hori and Oike, 1999]. Early work using tsunami and geodetic data suggested rupture zone widths of 70 km for the Tonankai segment and 100 km for the Nankaido segment (Figure 3) [Ando, 1975]. Thermal models incorporating a
Figure 3. (a) Location map of SW Japan and the 1944 Tonankai M8.1 and 1946 Nankaido M8.3 earthquakes. Four day aftershocks (circles and squares) define the rupture zone (shaded gray), which is widest (200 km) for the Nankaido segment where plate dip is shallowest. (b) Plate geometry along Nankai–Shikoku transect based on seismic profiling [Kodaira et al., 2000], regional tomography [Zhao et al., 2000], and hypocenters (January 1964 to December 1995) [Engdahl et al., 1998] (sampling box shown in a). (c) Plate geometry along southern Kyushu transect (details same as b).
constantly increasing plate dip arrived at similar estimates (Figure 3) [Hyndman et al., 1995; Oleskevich et al., 1999].

Two transects from the SW Japan margin were modeled, one across the shallowest dipping portion of the Nankai Trough, beneath Shikoku and southwestern Honshu (Figures 2a, 3b, and 4b) and a second across the southern Kyushu margin (northeasternmost Ryukyu trench), with a steeper mean plate dip and a volcanic arc 240 km from the trench (Figures 2b, 3c, and 4c). For the Shikoku transect, the arc is located 380 km from the trench axis of the Nankai

Figure 4. Modeled thermal structure and heat flow for SW Japan margin and predicted width of the seismogenic zone. (a and b) Nankai/Shikoku transect and (c) Kyushu transect. See text for discussion.
Based on seismic reflection profiles from the Nankai Trough, the sediment thickness at the trench above the decollement is 1 km [Moore et al., 1990] and this value was taken for both transects. Tomographic images obtained from regional earthquake travel time data indicate the presence of a subhorizontal Philippine Sea Plate slab situated below 40 km depth beneath southwestern Honshu [Hirahara, 1981; Zhao et al., 2000; Zhao, 2001] in agreement with available hypocenters from this region (Figure 3b) and indicate flat subduction occurs here [Sacks, 1983; Gutscher, 2001].

For the Shikoku transect the thermal modeling is complicated by the age structure of the Philippine Sea plate. The lithosphere subducting at the Nankai trough (the Shikoku Basin) was formed at a back arc spreading center between 25 and 18 Ma [Okino et al., 1999]. Because the axis of the fossil spreading center is subparallel to the plate motion vector, the age of the subducting lithosphere at the trench has been increasing through time. Given that subduction is believed to have begun near 15 Ma and the age of the lithosphere at the trench today (for the Shikoku transect) is 20 Myr, the thermal structure has evolved from subduction of 5 Myr old lithosphere to subduction of 20 Myr old lithosphere. We used a mean age of 15 Myr for the incoming lithosphere.

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The Nankai segment of the Japanese islands shows the lowest thermal gradients of 25–50 K/km, compared to values of 50–150 K/km for the Japan trench and Kyushu subduction segments [Tanaka et al., 1999]. This is clearly due to the flat slab geometry, since the age of the subducting lithosphere is the youngest, <25 Ma and accordingly the initial slab temperature at the trench is highest for the Nankai segment, compared to >90 Ma lithosphere at the Japan trench and 40 Ma lithosphere for the Kyushu segment. Recently published [Furukawa et al., 1998] heat flow data from Shikoku and western Honshu constrain the thermal structure of the upper plate. The broad forearc heat flow minimum, extending across the Seto Inland Sea (separating Shikoku from Honshu) and >250 km from the trench, indicates an absence of convecting mantle beneath and is in good agreement with our thermal model incorporating the appropriate flat slab geometry. Our calculated heat flow agrees well with the observed values, within the scatter of the data (Figure 4a). The sharp peak in heat flow at 380 km marks the Quaternary volcanic arc with values >120 mW/m².

Our Shikoku model predicts a temperature of 350°C is reached 160 km from the Nankai Trough. Taking the 150°C isotherm as the beginning of the locked zone and considering the transitional zone to be between the 350 and 450°C isotherms, one arrives at a downdip width of 130–210 km for the Nankai–Shikoku transect (Figure 4b). For the Kyushu transect the lithospheric age is substantially older (formed at 40 Ma) and the locked zone is only 90–100 km wide (Figure 4c). Because of the colder initial temperature the seismogenic zone begins further arcward and the 350°C isotherm is at about the same distance from the trench.

The seismogenic zones predicted by our thermal models are in good agreement with the distribution of 4 day aftershocks from the JMA (Japanese Meteorological Agency) Database (Figure 3) for both the Nankaido and Kyushu segments. For the flat slab Shikoku transect our
Figure 6. (a) Southern Alaska location map showing depth contours to the Wadati–Benioff zone, rupture zone based on source time studies (dashed) and 4 day aftershocks M > 4 (circles) [Engdahl et al., 1998]. Note the rupture zone (shaded gray) based on aftershocks is 250 km wide in the flat slab region to the NE where the 1964 earthquake was initiated and diminishes to <150 km to the SW where slab dip increases. (b) Plate geometry along Prince William Sound transect based on hypocenters (January 1964 to December 1995) (sampling box shown in a). (c) Plate geometry along Kodiak Island transect (details same as b).
locked zone is about 50 km greater than that obtained by previous thermal models which used a steadily steepening plate dip [Hyndman et al., 1995; Oleskevich et al., 1999]. This increased width is in good agreement with long term geodetic measurements of coseismic and interseismic vertical deformation, which show coseismic subsidence as far inland as the north coast of Shikoku island (Figure 5) [Thatcher, 1984]. The shape of these subsidence/uplift curves suggests a downdip limit about 190 km from the trench, in excellent agreement with our thermally predicted downdip limit. This agreement is further corroborated by recent joint inversions of GPS and tsunami data using the multiple polygon technique indicating a wider locked zone of up to 200 km for the Nankaido segment extending from close to the trench all the way across Shikoku Island to the coast of the Inland Sea [Sagiya and Thatcher, 1999].

5. The Great M9.2 Alaska Earthquake

[18] The rupture of the 1964 M9.2 earthquake initiated beneath Prince William Sound in the zone of shallowest plate dip (Figure 6). The position of the Pacific Plate at depth is well constrained by abundant Wadati–Benioff zone seismicity (Figure 6) [Page et al., 1989]. For the shallowest 30 km the crustal structure is imaged by wide-angle seismic data from the TACT [Brocher et al., 1994] and EDGE [Moore et al., 1991; Ye et al., 1997] profiles. Sediment thicknesses at the trench are 2 km for the Kodiak transect.
and 3 km for the Prince William Sound transect [Gutscher et al., 1998; Fruehn et al., 1999]. Source time studies have suggested an 800 km long, trench-parallel rupture zone with an almost constant width of roughly 150 km (downdip), an average slip of 10 m along the plate interface and two major asperities, one beneath Prince William Sound and a secondary one east of Kodiak Island [Ruff and Kanamori, 1983; Christensen and Beck, 1994].

Modern studies of the rupture zone describe the plate interface as a three-dimensional surface consisting of several tens of polygons. An inversion of the coseismic geodetic deformation data for 68 square shaped subfaults yielded peak slip values of 25 m for the northeastern (Prince William) segment and 10–15 m for the Kodiak segment [Holdahl and Sauber, 1994]. A joint geodetic and tsunami inversion using 17 quadrilaterals and incorporating a more realistic and extremely shallow 3° dip below the Prince William Sound segment arrived at similar slip values but with a region of maximum slip (25 m) extending somewhat further north [Johnson et al., 1996]. Both results show a wider rupture zone in the NE (200 km) than in the SW (150 km). Previous thermal models for the Alaska subduction zone featured a simplified plate geometry either with a constant plate dip [Ponko and Peacock, 1995] or with an increasing plate dip, but lacking a convecting mantle wedge [Oleskevich et al., 1999].

Two transects were modeled across the portion of the Alaska margin that ruptured during the great 1964 event, one in the region of shallowest plate dip beneath Prince William Sound and one across Kodiak Island where plate dip is steeper. For the former, the volcanic arc is 560 km from the trench, above a slab at 120 km depth. For the latter, the arc is 320 km from the trench, above a slab at 80 km depth. The plate interface from 200 to 400 km from the trench for the Prince William Sound transect is located at a shallow depth of 30–40 km. Thus, the upper plate lithosphere here is limited to only 40 km. Accordingly, a value of 40 km was chosen for the upper plate thickness for both profiles. Attempts to model the thermal structure with a simple mantle wedge of constant dip (both 30° and 45° and lithospheric thicknesses of 40–60 km were tried) proved unsatisfactory, in particular for the Kodiak Island transect where temperatures beneath the arc were insufficient to permit partial melting of the mantle wedge. Thus, a composite mantle wedge was constructed, with a dip of 20° up to roughly the volcanic arc and 100 km depth, and a 45° dip beyond and at greater depth. This more accurately reflects the gently increasing slab curvature observed in tomographic images from the Alaska subduction zone [Zhao et al., 1995].

The results of our thermal modeling for the Prince William Sound transect (Figure 7a) predict a broad locked zone of 240–340 km. For the Kodiak Island transect our model predicts a locked zone of 160–180 km and a temperature of 1100°C beneath the arc (Figure 7b). These results are supported by the distribution of 4 day aftershocks M > 4 from the best available relocated hypocenter catalog [Engdahl et al., 1998] which clearly shows a broader rupture zone to the NE, up to 250 km in width and narrowing to the SW to 100 km south of Kodiak Island, where plate dip steepens (Figure 6). The position of the main shock and the distribution of aftershocks seen in cross section is in good agreement with the coseismic uplift signal (Figure 8) [Savage and Plafker, 1991; Holdahl and Sauber, 1994] and with the limits predicted by our thermal model (Figure 7). These results are further corroborated by joint geodetic–tsunami inversions [Johnson et al., 1996] and with recent GPS results of postseismic deformation from Kenai Peninsula indicating a very broad zone of nearly complete elastic coupling [Freymueller et al., 2000].

6. The Cascadia Margin

Two transects were modeled across the Cascadia margin, one in northern Washington and one in central

Figure 8. Seismogenic zone of the great 1964 Alaska M9.2 earthquake. (a) Coseismic deformation in Prince William Sound region. (b) Rupture zone in Prince William Sound region based on 4 day aftershocks.
Figure 9. (a) Cascadia location map showing depth contours to the Wadati–Benioff zone and interplate locked zone (shaded gray) based on GPS data [Khazaradze et al., 1999]. P = Pacific Plate, G = Gorda Plate, JF = Juan de Fuca Plate, E = Explorer Plate. (b) Plate geometry along the Olympic–Puget transect in northern Washington at 48°N. Between 12 and 30 km depth the plate geometry is poorly constrained and may be shallow dipping or flat slab (sampling box for seismicity shown in a). (c) Plate geometry along the central Oregon transect. Upper 30 km are constrained by seismic data [Gerdom et al., 2000] and at greater depth only by regional tomography [Parsons et al., 1999] and by assuming a mean depth to the slab of 100 km beneath the arc.
Oregon (Figure 9). The northern Washington transect crosses the Olympic Peninsula and Puget Sound, where the Juan de Fuca slab has the shallowest plate dip [Crosson and Owens, 1987]. This been termed an arch [Stanley and Villasenor, 2000], and has been recently proposed to be a flat slab segment [Gutscher et al., 2000]. Plate geometry is well constrained by seismic profiling near the trench [Flueh et al., 1998; Gutscher et al., 2001]. Sediment thicknesses at the trench above the decollement are 2 km for both the Washington and Oregon transects [MacKay, 1995; Flueh et al., 1998].

Figure 10. Modeled thermal structure for the Cascadia margin and the resulting predicted width of the seismogenic zone, including observed and calculated heat flow. (a and b) Olympic–Puget transect. (c and d) Central Oregon transect. See text for discussion.
Hypocenters beneath Puget Sound constrain the plate geometry at 40–70 km depth (Figure 9b). Between 12 and 30 km depth the plate geometry is not well constrained and may be shallow dipping or horizontal (flat slab). In both cases the position of the corner of convecting asthenosphere will be situated far from the trench as indicated by the 360–380 km trench–arc separation, the greatest observed in the Cascades arc.

The second transect across Central Oregon samples a more steeply subducting slab. The geometry is constrained in the uppermost 30 km by seismic refraction data [Gerdom et al., 1999] and thereafter by regional tomography [Parsons et al., 1999] as well as by estimating a mean depth to the slab of 100 km beneath the volcanic arc (Figure 9c). The Cascadia forearc is composed largely of mafic accreted terranes (Crescent and Siletzia terranes) and thus a mafic composition (with lower radiometric heat production) was used for the forearc as modeled in earlier studies [Hyndman and Wang, 1995; Oleskevich et al., 1999].

The Cascadia margin is well constrained by abundant heat flow data (Figures 10a and 10c) [Blackwell et al., 1990], including offshore data from ODP drilling and gas hydrate studies [Hyndman and Wang, 1995]. The trench normal heat flow profiles differ markedly between the two transects. The forearc heat flow minimum, with values of 35–50 mW/m² is narrower and situated along the coast in Oregon (about 150 km from the trench), whereas it is much broader in Washington (200–300 km from the trench) extending all the way across Puget Sound, indicating an absence of hot convecting asthenosphere beneath. This thermal low corresponds to the region of greatest trench arc separation, shallowest plate dip and increased upper

\[23\] The distance between the trench and the arc is not well constrained and may be shallow dipping or horizontal (flat slab). In both cases the position of the corner of convecting asthenosphere will be situated far from the trench as indicated by the 360–380 km trench–arc separation, the greatest observed in the Cascades arc.

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Table 1. Great Interplate Subduction Earthquakes (M8) of the 20th Century

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<thead>
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<th>Lat.</th>
<th>Long.</th>
<th>Region</th>
<th>Magnitude</th>
<th>Subd. style</th>
<th>v (cm/yr)</th>
<th>Lith. age (Myr)</th>
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Most earthquake dates, locations, and magnitudes in the table are from the compilation of Ruff [1996]. The New Guinea earthquakes are from the work of Okal [1995]. Date is in day, month, year format and Magnitude = Mw unless otherwise specified (i.e., s = Ms). A steep subduction style means steadily increasing plate dip (typically ≥30° beneath the forearc), while a flat subduction style is defined by three inflection points in the downgoing plate [Gutscher et al., 2000]. v = orthogonal plate convergence based on global plate kinematic models [DeMets et al., 1990; Seno et al., 1993], except for Tonga/Kermadec where GPS velocities were used [Bevis et al., 1995]. Lith. age = the age of the subducting oceanic lithosphere at the trench, taken from the compilation of Mueller et al. [1997].

Plate seismicity [Crosson and Owens, 1987; Stanley and Villaseñor, 2000]. Based on parallels to heat flow and seismicity patterns observed in other well documented margins [Henry and Pollack, 1988; Gutscher et al., 2000] we suggest here that there is a causal link between the flat subduction style (Figure 9b) of Cascadia between 46° and 49°N and the low heat flow values observed here.

[25] The heat flow values predicted by our model are shown in Figures 10a and 10c compared to the observed data. Our model successfully predicts the overall shape and position of the heat flow minima. However, as our models do not include the effects of heat transport through genesis of arc magmatism, the short-wavelength heat flow peak is not successfully reproduced. In each case this maximum of 120–150 mW/m² coincides exactly with the position of the volcanic arc (respectively some 360 and 280 km from the trench). The back arc regions in general have 10–20 mW/m² higher heat flow than predicted by our ideal continental geotherm yielding 65 mW/m², though this discrepancy decreases with increasing distance from the arc (Figures 10a and 10c).

[26] Our thermal model of the Olympic–Puget Sound transect (Figure 10b) yields a moderate locked zone of 100–170 km width. For the Central Oregon transect we predict a somewhat narrower locked zone of 80–140 km width (Figure 10d). In both cases, reasonable temperatures of about 1160°C are obtained beneath the arc. The broad forearc heat flow minimum associated to the shallow slab dip along the Olympic–Puget transect suggests that previous elastic dislocation models and thermal models based on steeper plate...
dips [Hyndman and Wang, 1995; Oleskevich et al., 1999] may underestimate the width of the seismogenic zone for northern Washington. This is also the conclusion reached by the most recent GPS deformation models [Khazaradze et al., 2000] which have a locked zone of 100 km plus an additional transitional zone of 130 km, bringing the potential rupture plane to a position beneath the edge of Puget Sound (Figure 9a). Thus the maximum size of a potential earthquake as well as the area affected by intense ground shaking may be larger and may pose a substantially higher risk to the densely populated Puget lowlands region than previously assumed.

7. Discussion

[27] On the basis of our thermal models and of available heat flow observations it is clear that for margins with comparable subduction parameters (slab age, plate velocity) the seismogenic zone will be substantially larger (downdip width) for a flat slab subduction style. The distribution of aftershocks shown here (e.g., SW Japan, Figure 3 and Alaska, Figure 6) is in good agreement with these limits. Geodetic observations of coseismic and interseismic vertical deformation from the Nankai and Prince William Sound margins (Figures 5 and 8) also agree well with our predicted limits. For the colder, more steeply dipping subduction zones (Kyushu, Kodiak) the downdip limit of aftershocks extends close to the tip of the hot wedge of convecting asthenosphere, which governs the position of the 350°C and 450°C isotherms. This appears to be consistent with a thermal control of the seismogenic zone. As noted above, convecting mantle is required to obtain sufficiently high temperatures beneath the arc.

[28] Alternatively, a downdip limit in the forearc mantle may be consistent with a compositional control by serpentinite in the mantle wedge [Hyndman et al., 1997; Oleskevich et al., 1999; Peacock and Hyndman, 1999]. A test between the serpentinite and thermal hypothesis requires wide-angle seismic data on the deep crustal structure of the forearc as well as precise aftershock data on the size of the rupture zone. Such a case exists for the NE Japan subduction zone and the tightly constrained aftershock distribution following the 1994 Sanriku-oki earthquake [Hino et al., 2000]. Here the aftershocks, and thus the inferred rupture zone, extend to 50 km depth and the upper plate Moho is known from deep crustal seismic studies to be located at 20 km depth. Thus, for NE Japan, the serpentinite hypothesis fails to explain the position of the downdip limit and we therefore favor a thermal control for the margins in this study as well.

[29] Geodetic modeling of the interseismic and coseismic elastic deformation offers another independent measure of the limits of the locked zone, though this method is sensitive not only to the chosen width of the rupture zone, but also to the plate geometry. The earliest elastic dislocation model of the coseismic slip during the Nankai earthquakes included only a single plane, dipping at a constant angle of 20°–25° (Figures 3 and 11) [Ando, 1975]. This model predicted a width of 100 km for the locked zone (extending from 50 to 150 km from the trench). Subsequent models divided the downdip plane into several segments and modeled the interseismic deformation as well [Savage and Thatcher, 1992]. Later studies subdivided the rupture plane into numerous smaller rectangular subplanes and inverted for slip along each of these [Satake, 1993; Holdahl and Sauber, 1994]. Finally, such models became truly three dimensional, by assigning each subplane its own plate dip in order to reflect all available constraints on plate geometry. These sophisticated models can also be used for joint inversion of coseismic geodetic and tsunami waveform data [Johnson et al., 1996;
Sagiya and Thatcher, 1999]. Modern geodetic studies from the three margins discussed here are in good agreement with the limits predicted by our thermal models (Figure 11).

Finally, we examine the global record of great interplate earthquakes of the 20th century [Ruff, 1996]. Flat slab segments have proportionally more great earthquakes than margins with a “normal” dipping subduction (Table 1 and Figure 12). Most of the flat slab segments indicated in Table 1 were identified and discussed in a previous paper [Gutscher et al., 2000]. A subsequent addition is the Makran subduction zone, on the basis of the tomographic cross section of Bijwaard et al. [1998, Figure 9a, p. 30.070]. The Makran margin experienced an M8.1 event in 1945 [Byrne et al., 1992]. Whereas the flat slab segments represent only 10% of the world’s convergent margins, they account for over 1/3 of instrumentally recorded great M8 subduction earthquakes. This confirms the results of the other data sets discussed above and supports the idea that interplate coupling is higher in flat slab regions and such regions pose a substantially higher seismic risk.

8. Conclusions

[31] A low subduction angle modifies the thermal structure of a subduction zone by displacing the hot convecting asthenosphere wedge away from the trench. Thus flat slab regions have a cold forearc up to 300–400 km from the trench. The increased size of the rupture zone as predicted from the thermal models is in good agreement with the most recent geodetic studies from the SW Japan, Alaska and Cascadia margins. The thermal results together with other observations (geometric considerations, upper plate seismicity) indicate stronger interplate coupling along flat slab segments. Over 1/3 of great M8 interplate earthquakes in the 20th century occurred in regions of flat slab subduction, which represent only 10% of the world’s convergent margins. The greater downdip width of the seismogenic zone (150–250 km) implies a greater maximum size for an interplate earthquake and a hypocenter further from the trench, thus posing an increased seismic risk to populated coastal regions near flat slab segments, including SW Japan, Alaska, Cascadia and southern Mexico.

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