Bulletin of the Seismological Society of America, Vol. 95, No. 5, pp. 1880–1902, October 2005, doi: 10.1785/0120040132

# Low Potential for Large Intraslab Earthquakes in the Central Cascadia Subduction Zone

by Ivan G. Wong

**Abstract** The Cascadia subduction zone (CSZ) can be divided into three distinct sections based on the characteristics of intraslab seismicity. Based on a 150-year historical record, no moderate-to-large intraslab earthquakes of moment magnitude (**M**) 5.5 or greater have occurred within the subducting Juan de Fuca plate of the central CSZ from south of the Puget Sound in northwestern Washington to the Oregon–California border. Also very few intraslab earthquakes as small as **M** 3 have been instrumentally located within the central CSZ since 1960, and a Wadati–Benioff zone is not apparent. In the southern CSZ beneath northwestern California, a Wadati–Benioff zone is present to a depth of about 40 km, but no large Gorda block earthquakes have been observed in the downgoing slab, although large events have occurred near the trench axis. In contrast, the Puget Sound region within the northern CSZ has been repeatedly shaken by large intraslab earthquakes of **M** ≥6.5 in the depth range of 40 to 60 km, such as the recent 2001 **M** 6.8 Nisqually event. A critical question addressed in this article is what is the potential for such large, shallow intraslab earthquakes in the central CSZ beneath western Oregon and southwestern Washington?

I have evaluated the available information on the thermal and physical properties, geometry, and historical and contemporary seismicity of the central CSZ, and performed thermal modeling. Based on these analyses and comparisons with other subduction zones worldwide, the lack of shallow intraslab earthquakes in the central CSZ is not unusual. The hot temperatures (>500°C) within the Juan de Fuca plate, particularly below a depth of 40 km where large events are expected, are not conducive to earthquake generation, resulting in either the complete absence of **M** ≥6.5 shallow intraslab earthquakes or long recurrence intervals (hundreds of years) between such events. Temperatures appear to be sufficiently high in the central CSZ so that no Wadati–Benioff zone can exist even at shallow depths (<40 km). The young plate age, slower convergence rate, and the insulating effect of the Siletz terrane above the plate are factors that probably lead to the hot temperatures in this portion of the slab. The variability in the maximum depth of the Wadati–Benioff zone along the CSZ, 60 km beneath the Puget Sound, 40 km within the subducting Gorda block, and essentially zero in the central CSZ, reflect the differing temperature conditions, that is, the cutoff temperature varies with depth and rock composition, and also the potential for large shallow intraslab earthquakes. In addition to the effects of temperature, the level of tectonic stresses, which vary along the length of the CSZ, must also be a factor in controlling the occurrence of large intraslab earthquakes. Large events can occur in the Puget Sound region, probably because of cooler intraslab temperatures and possibly because of a stress concentration or zone of weakness along the pronounced arch in the Juan de Fuca plate.

A previous study has suggested an intraslab subduction zone origin for a **M** 7.3 earthquake that occurred in 1873 near the town of Brookings, in southernmost Oregon. However, analysis of its seismotectonic setting and comparison with other historical earthquakes in northernmost California suggest that the event probably had a very shallow origin within the Gorda block (southern CSZ) and was not a deep intraslab earthquake in the central CSZ.

## Introduction

An examination of the relatively brief historical record of the Pacific Northwest dating back to about 1850 indicates that no moderate-to-large (moment magnitude [**M**] ≥5.5) intraslab earthquakes have occurred within the subducting Juan de Fuca plate of the central Cascadia subduction zone (CSZ) defined herein as extending from south of the Puget Sound region, Washington, to the Oregon–California border (Figs. 1 and 2). A possible exception is a large earthquake that shook much of western Oregon and Washington and northern California in 1873, centered near Brookings, Oregon (Ludwin *et al.*, 1991) (Fig. 2). The more complete instrumental record, which dates back to about 1960, indicates that very few intraslab earthquakes, down to a location threshold of Richter local magnitude ($M_L$) 3, have occurred within the central CSZ (Fig. 2).

In contrast, the Puget Sound region within the northern CSZ has been shaken by moderate-to-large intraslab earthquakes (depths between 40 to 60 km), including events in 1939 (unspecified magnitude *M* 5¾), 1946 (*M* 6.3), 1949 (*M* 7.1), 1965 (body-wave magnitude $m_b$ 6.5), and most recently in 2001 near the town of Nisqually (**M** 6.8) (Ludwin *et al.*, 1991; Pacific Northwest Seismograph Network [PNSN] Staff, 2001). Although none of these events resulted in a heavy loss of life or numerous injuries, structural damage and economic losses were significant. No large intraslab earthquakes have been observed beneath Vancouver Island in the northern CSZ (Ludwin *et al.*, 1991).

In the southern CSZ, large earthquakes have occurred near the trench axis in the Gorda block offshore of northern California. However, slab seismicity terminates at a depth of about 40 km, so no large intraslab earthquakes at greater depths have been observed in the southern CSZ. Thus, the CSZ can be divided into three distinct sections based on intraslab seismicity.

A critical question is what is the potential beneath western Oregon and southwestern Washington for large (**M** ≥6.5) intraslab earthquakes such as those that have occurred historically in the Puget Sound region (Ludwin *et al.*, 1991; Rogers *et al.*, 1996; Weaver and Shedlock, 1996; Wong, 1997; Kirby *et al.*, 2002; Petersen *et al.*, 2002)? No large historical earthquakes have struck the populated urban corridor in the Willamette Valley in western Oregon, which includes the Portland metropolitan area and the cities of Salem and Eugene (Ludwin *et al.*, 1991; Wong and Bott, 1995) (Fig. 2).

Weaver and Shedlock (1996) proposed that there is a potential for large intraslab earthquakes at shallow depths of 45 to 60 km extending throughout western Washington and as far south as west-central Oregon to about latitude 44.5° (Fig. 1). For example, the 2001 Nisqually earthquake occurred near the source of the 1949 Olympia event at a depth of 54 km (PNSN Staff, 2001; K. Creager, University of Washington, written comm., 2005). Weaver and Shedlock's (1996) shallow depth range is based on the observation that all recorded intraslab events of $M_L$ 4 or greater have occurred deeper than 45 km. They were unsure of the intraslab potential along the central Oregon Coast because of the unknown plate geometry and lack of historical seismicity. They did, however, include the 1873 epicentral area as an area of large intraslab earthquake potential as well as the Gorda block further south along the Mendocino coast (Fig. 1).

The focus of this study is on large shallow intraslab earthquakes occurring at depths of 40 to 70 km. Because all large Puget Sound intraslab events have been in this depth range and hence shallow, similar events located beneath the most populated portions of the Pacific Northwest are of greatest importance in terms of seismic hazards. The term "intraslab" is used synonymously with "intraplate" and includes earthquakes conventionally defining the Wadati–Benioff zone.

In this article, I suggest that the potential for large intraslab earthquakes in the central CSZ is very low because the internal temperatures within the Juan de Fuca plate at depths greater than 40 km, where such events occur, are too high to sustain the stresses necessary for seismogenesis. I review what is known about the current seismotectonic setting and seismicity of the central CSZ and compare its attributes with those of other subduction zones. I perform thermal modeling of the central CSZ to quantify these intraslab temperatures and compare them with those in other subduction zones worldwide. Although temperature is probably the most important factor controlling intraslab seismicity, other factors such as the level of tectonic stresses within the slab at shallow depths (40 to 70 km) are also significant, and I describe these in this article also. The origin of the 1873 earthquake has direct implications for the potential of large intraslab earthquakes in the region, and I evaluate and discuss this issue.

## Intraslab Earthquakes

Gutenberg and Richter (1954) classified earthquakes at depths between 70 and 300 km as intermediate earthquakes and those deeper than 300 km as very deep earthquakes. Thus, to maintain consistency with this terminology, I use the term "shallow" for events whose focal depths are less than 70 km. Note some investigators consider intraslab earthquakes as shallow as 40 km deep to be intermediate-depth events (e.g., Astiz *et al.*, 1988). An important question, which I do not address in this study, is whether shallow intraslab earthquakes are caused by a different mechanism than deeper intraslab events. Regardless of the mechanisms (e.g., brittle failure, dehydration embrittlement), temperature is a major factor controlling the occurrence of intraslab seismicity (Isacks *et al.*, 1968; Vlaar and Wortel, 1976; Molnar *et al.*, 1979; Kirby *et al.*, 1996; Stein and Stein, 1996) and is a focus of this study.

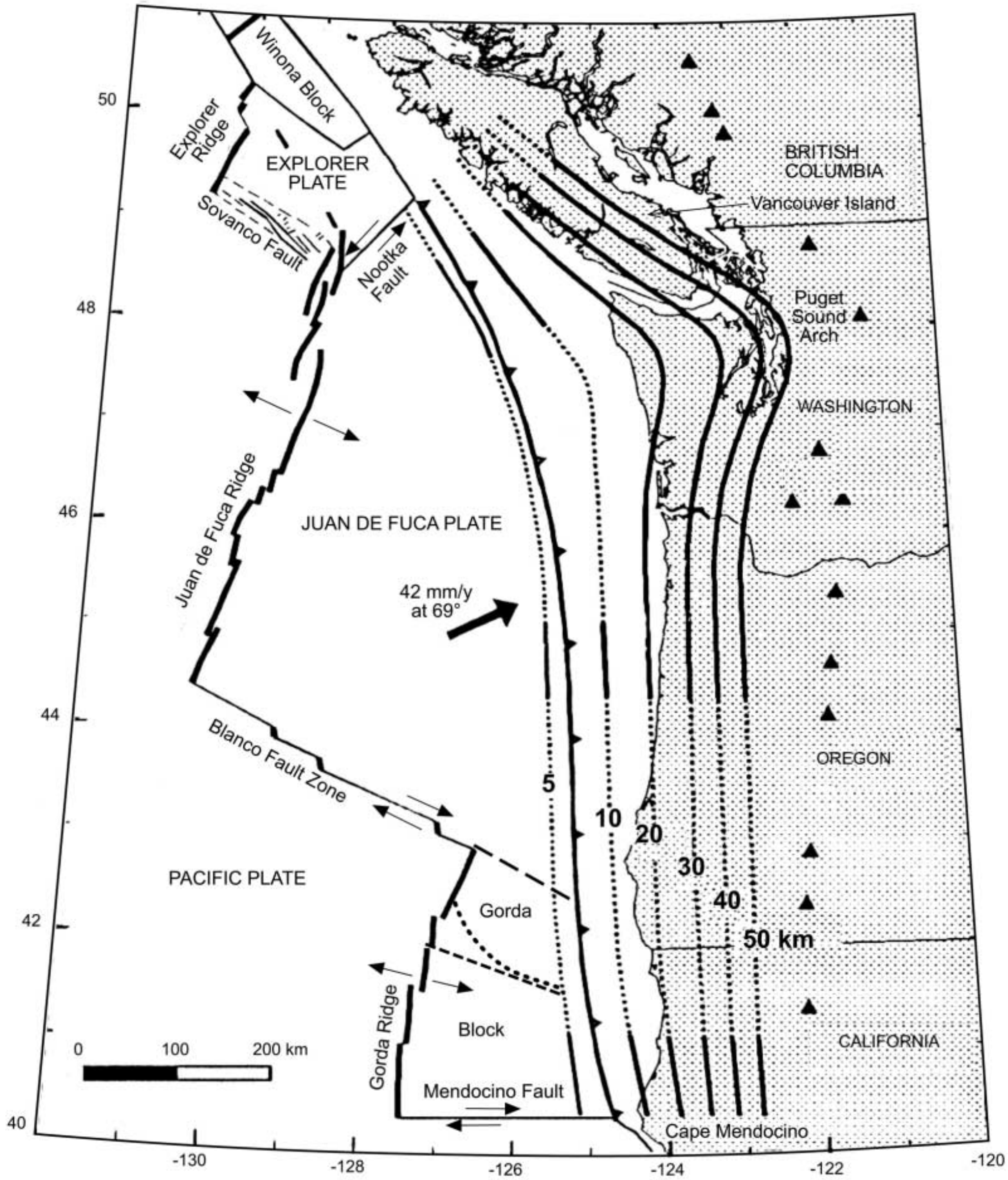


Figure 1. Geometry of the CSZ (modified from Flück *et al.*, 1997). The northern CSZ extends from Vancouver Island to south of Puget Sound; the central CSZ from the latter to the northern boundary of the Gorda block; and the southern CSZ is composed of the subducting Gorda block. The short-dashed line is the northern boundary of the southern Gorda plate, from Riddihough (1984). The dotted line represents the northern boundary of the Gorda deformation zone as defined by Wilson (1989). The long-dashed line marks the boundary of the oceanic plate being generated by the Gorda ridge and of the Gorda block, inferred in this study from Smith *et al.* (1993). The deformation front is indicated by the barbed line. Solid and dotted depth contours (observed and inferred, respectively) are to the top of the Juan de Fuca plate. Triangles represent Cascade volcanoes.

To aid in the discussions presented in this article, I adopt the following intraslab classification, which is similar to the scheme of Astiz *et al.* (1988), based principally on the location of earthquakes relative to the subduction zone (Fig. 3): (A) earthquakes within the subducting plate at depths greater than 40 km; (B) shallow events near the trench axis or in the outer-rise region; and (C) events well removed from subduction zones and in the interior of the oceanic plates. All the damaging intraslab earthquakes in the Puget Sound region have been type A events. Type A and B events make up Wadati–Benioff zones, are associated with the colder portions of the subducting plate, and are a result of the tectonic

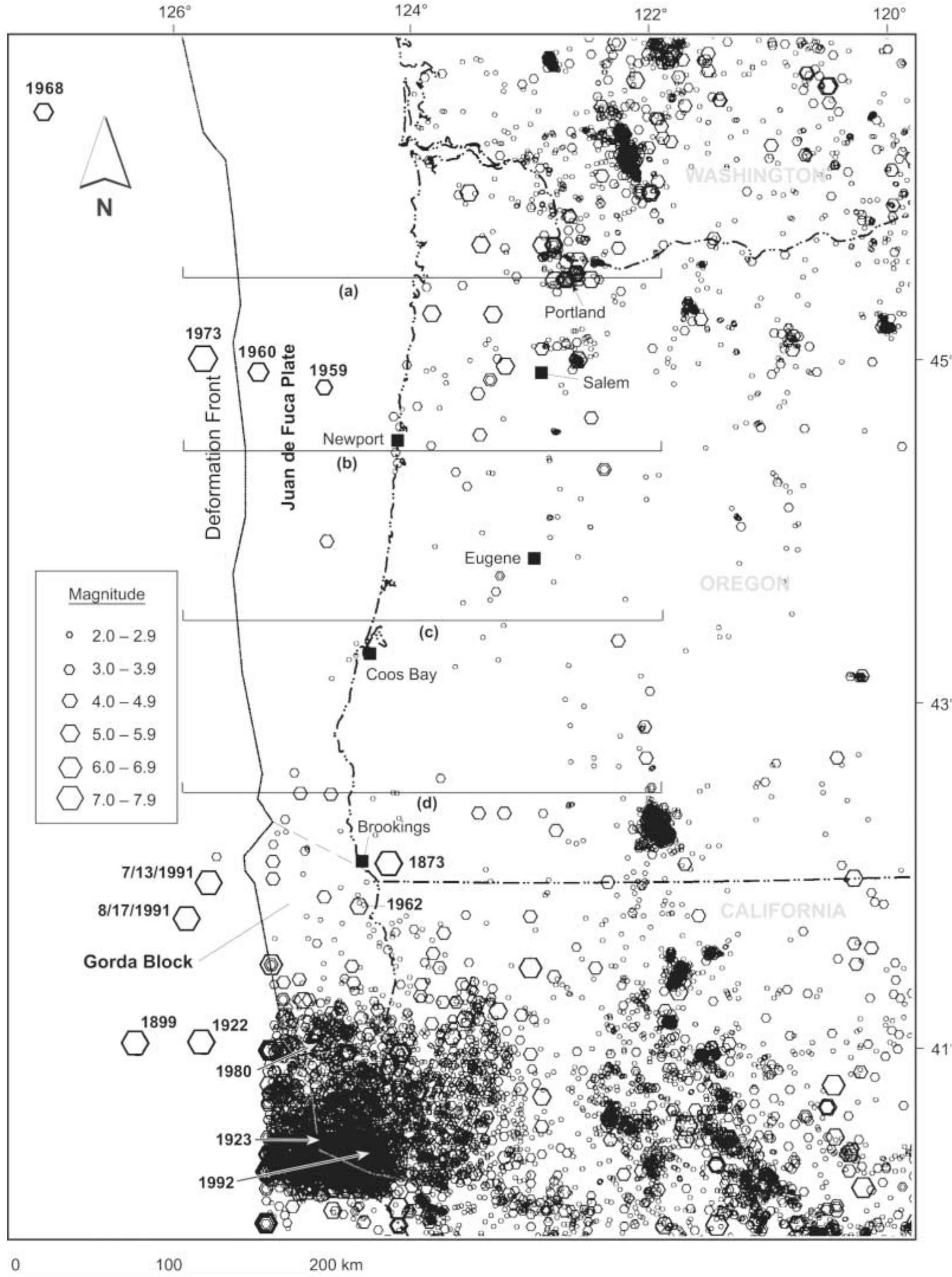


Figure 2. Historical seismicity, 1850 to 2000, of the central and southern CSZ (**M** ≥2.0). Significant Gorda block earthquakes are labeled. Only significant events are shown west of the deformation front. Dashed line is the approximate boundary between the Juan de Fuca plate and the Gorda block. The 1968 and 1973 type B events west of the deformation front are also shown. Seismicity data compiled from the University of Washington, U.S. Geological Survey, Northern California Earthquake Data Center, and Bakun (2000). Locations of cross sections shown on Figure 4 are also indicated.

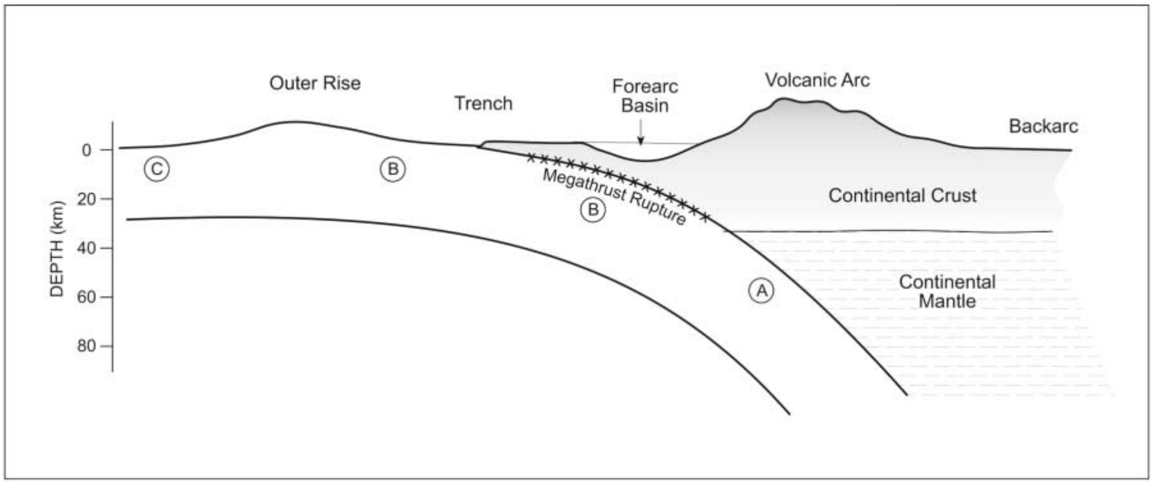


Figure 3. Classification of intraslab earthquakes modified from Astiz *et al.* (1988): (A) intraslab earthquakes deeper than 40 km; (B) intraslab events near the trench axis or outer-rise region; and (C) events in the interior of the oceanic plate beyond the outer rise.

stresses being induced within the downgoing plate (Isacks *et al.*, 1968). Type B earthquakes result from bending of the plate at the trench axis and outer rise (Fig. 3). Interplate coupling at shallow depths may be important in influencing the distribution and timing of type A and B intraslab seismicity (Astiz *et al.*, 1988).

A possibly significant aspect of type A events is that they generally occur beneath the contact where the plates are coupled, generally at depths below 40 km (Fig. 3). Bends in the subducting plates are often observed at this depth (Astiz *et al.*, 1988), and this is also the case for the CSZ (see further discussion). At depths of about 40 km, basaltic and gabbroic rocks undergo a phase change to ecologite, resulting in a denser plate and hence steeper subduction (Ruff and Kanamori, 1983). Furthermore, all subduction zones are generally uncoupled below a depth of 40 km possibly due to this phase change in the down-going plate (Ruff and Kanamori, 1983).

Astiz *et al.* (1988) noted that the most active regions of large shallow and intermediate-depth earthquakes (e.g., Altiplano, Timor, Sulawesi) have significant distortions (e.g., lateral bends, tears, buckles) in the subducting plate. Several investigators have suggested that shallow and intermediate-depth earthquakes are probably the result of rupture along reactivated pre-existing zones of weakness in the oceanic plate in response to regional tectonic stresses (e.g., Kirby *et al.*, 1996; Jiao *et al.*, 2000). Kirby *et al.* (1996) further suggest that the reactivation of these normal faults down to depths of about 325 km is due to dehydration embrittlement. Recent analyses of six large (**M** 7.1 to 7.7) intermediate-depth earthquakes by Tibi *et al.* (2002) support this model. Kirby *et al.* (1996) also suggested that most shallow and intermediate-depth earthquakes occur in the topside of the subducting plates, possibly in the oceanic crust. This region is the coldest portion of the slab as it enters the trench, where brittle normal faults are expected to be concentrated and where hydrous phases are expected to be initially stable.

## Seismotectonic Setting

The CSZ marks the convergent plate boundary between the subducting Juan de Fuca plate, and Explorer and Gorda subplates and the overlying North America plate. The subduction zone extends for a distance of about 1100 km along the Pacific coast from Cape Mendocino in the south to Vancouver Island in the north (Fig. 1). Prior to the 1990s, it was thought that the megathrust along the CSZ was aseismic, owing principally to an absence of observed seismicity (see Rogers *et al.*, 1996 for summary). However, based largely on coastal paleoseismic evidence (e.g., Atwater *et al.*, 1995) and a tsunami recorded in Japan in January 1700 (Satake *et al.*, 1996), prehistoric megathrust earthquakes as large as **M** 9 are believed to have ruptured the CSZ every 400 to 600 years on average.

The Juan de Fuca plate was created at the Juan de Fuca ridge (Fig. 1). Subduction appears to have been occurring for at least the past 40 m.y. (Riddihough, 1984). The current rates of convergence range from 29 mm/yr at the Oregon-Washington border increasing to the north to 40 mm/yr near Vancouver Island and to the south where a maximum of 35 mm/yr is reached at the Mendocino Coast (Wells *et al.*, 2002). These rates are significantly lower than values previously estimated for the CSZ (e.g., Riddihough, 1984). The central CSZ has convergence rates of 29 to 33 mm/yr with an average rate of 31 mm/yr (Wells *et al.*, 2002). The northern CSZ has the highest convergence rates, ranging from 30 to 40 mm/yr (average $\sim$35 mm/yr), and the rate for the Gorda block in the southern CSZ is 33 to 35 mm/yr (Wells *et al.*, 2002).

From southern British Columbia to central Oregon, the inferred age pattern is very simple, with the Juan de Fuca plate being about 12 m.y. old below the coastline (Wilson, 2002). From central Oregon southward, the age pattern of the subducting plate is more complex. Beneath the Puget

Sound, the plate age is 14 to 15 m.y. while in contrast, the plate is only 11 to 12 m.y. old beneath the Willamette valley (Wilson, 2002). The Gorda block is about 10 to 11 m.y. old beneath inland northwest California. As will be discussed later, plate age is an important parameter with regard to the thermal state of the CSZ.

A commonly accepted model for the geometry of the CSZ was developed by Flück *et al.* (1997), which is a minor revision of previous models by Hyndman and Wang (1993, 1995) (Fig. 1). The geometry is based on seismic reflection, refraction and tomography data, Wadati–Benioff seismicity, and teleseismic waveform analysis. The amount of data varies along the length of the subduction zone, as demonstrated by the absence of a Wadati–Benioff zone in Oregon. The Flück *et al.* (1997) model indicates a gently dipping plate in Oregon to a depth of at least 50 km (Fig. 1). Based on more recent teleseismic body-wave analysis, however, the Juan de Fuca plate dips shallowly at about 12° beneath the central Oregon coast and more steeply (~27°) below the Willamette valley (Rondenay *et al.*, 2001). This change in dip occurs at a depth of approximately 40 km, as observed worldwide. At a depth of about 70 km, the plate steepens significantly (Rondenay *et al.*, 2001).

The region east of the Gorda ridge between the Blanco and Mendocino fracture zones has been considered a separate plate or a more deformed subplate or block of the Juan de Fuca plate (Wilson, 1989) (Fig. 1). The location of the northern boundary of the Gorda block appears to be controversial, as illustrated by various depictions in the literature. As will be discussed later with regard to the origins of the 1873 earthquake, an important question is whether the Juan de Fuca plate or the Gorda block is being subducted at the latitude of the Oregon–California border (Fig. 1). Riddihough (1984) defined northern and southern Gorda plates based on differences in spreading rates between the North and South Gorda ridges (Fig. 1). Some have referred to this boundary as the division between the Gorda and Juan de Fuca plates. Smith *et al.* (1993) suggested that the boundary (long-dashed line in Fig. 1) may be farther north than the Wilson (1989) boundary, based on seismicity possibly colinear with the Blanco fault zone.

## Historical and Contemporary Seismicity

In the following, I review and evaluate the historical and contemporary seismicity of the CSZ relevant to large shallow intraslab earthquake potential. Based on the characteristics of intraslab seismicity, the CSZ can be divided into three sections: northern, including the Puget Sound region and Vancouver Island; central; and southern (Gorda block).

### Central Cascadia Subduction Zone

The historical record indicates that very few earthquakes appear to be associated with the subducting Juan de Fuca plate in western Oregon (Fig. 2). This contrasts sharply with the Puget Sound region and the Gorda block (e.g., Ludwin *et al.*, 1991). The preinstrumental portion of the historical record since 1850 is probably complete for **M** >5.0 events in western Oregon (Wong and Bott, 1995). The seismographic coverage of coastal Oregon has been very poor. The instrumental record is probably only complete for events down to $M_L$ 4 since the early 1960s, when the Worldwide Standardized Seismograph Network came into existence. Until recently, no seismographic stations operated within 100 km of Brookings, the epicentral area of the 1873 earthquake (Wong and Bott, 1995).

Figure 4 shows east–west cross sections of seismicity recorded by the University of Washington's Pacific Northwest Seismographic Network (PNSN) from 1980 to 2000 along the central CSZ between the latitudes of 42° and 46°N. Also shown is the top of the subducting plate, as estimated from the model of Flück *et al.* (1997). Very few events appear to be located in the subducting plate, except at the boundary with the Gorda block, and hence a well-defined Wadati–Benioff zone is not apparent. Fewer than 35 earthquakes deeper than 30 km have been observed beneath western Oregon since 1963 (Fig. 2). The largest earthquakes of intraslab origin (or at least at suspected intraslab depths) observed in western Oregon are two events of $m_b$ 4.9 and 4.7, which occurred recently, on 12 July and 19 August 2004, respectively, just offshore of Newport, Oregon. West of the deformation front, seismicity (type B) is generally at a low level within the Juan de Fuca plate. A $m_b$ 4.6 event in 1968 and a $m_b$ 5.8 event in 1973 are the largest events observed to date off the Oregon coast (Fig. 2). Two possible intraslab events, in 1959 and 1960, $M_L$ 4.6 and 5.0, respectively, occurred offshore of Oregon (Fig. 2), but their focal depths are not well determined.

In addition to the lack of a Wadati–Benioff zone beneath western Oregon, a very low level of crustal seismicity has been observed all along the Oregon coast since at least 1980, when seismographic coverage was initiated in Oregon by the PNSN (Fig. 2). Few historical earthquakes have reportedly occurred in the proposed 1873 epicentral region (Fig. 2).

### Gorda Block

South of the Oregon–California border in the southern CSZ, the Wadati–Benioff zone is well imaged by contemporary seismicity down to a depth of only about 40 km within the subducting Gorda block (Smith *et al.*, 1993). Occasional small events have been observed as deep as 80 km (Walter, 1986).

Large type B earthquakes of **M** 7.0 and greater occur frequently in the Gorda block and are followed most often by aftershocks (Table 1). However, no large (**M** ≥6.5) type A events have been observed historically in the Gorda block (Table 1), which is consistent with the absence of a well-defined Wadati–Benioff zone below a depth of 40 km. Since

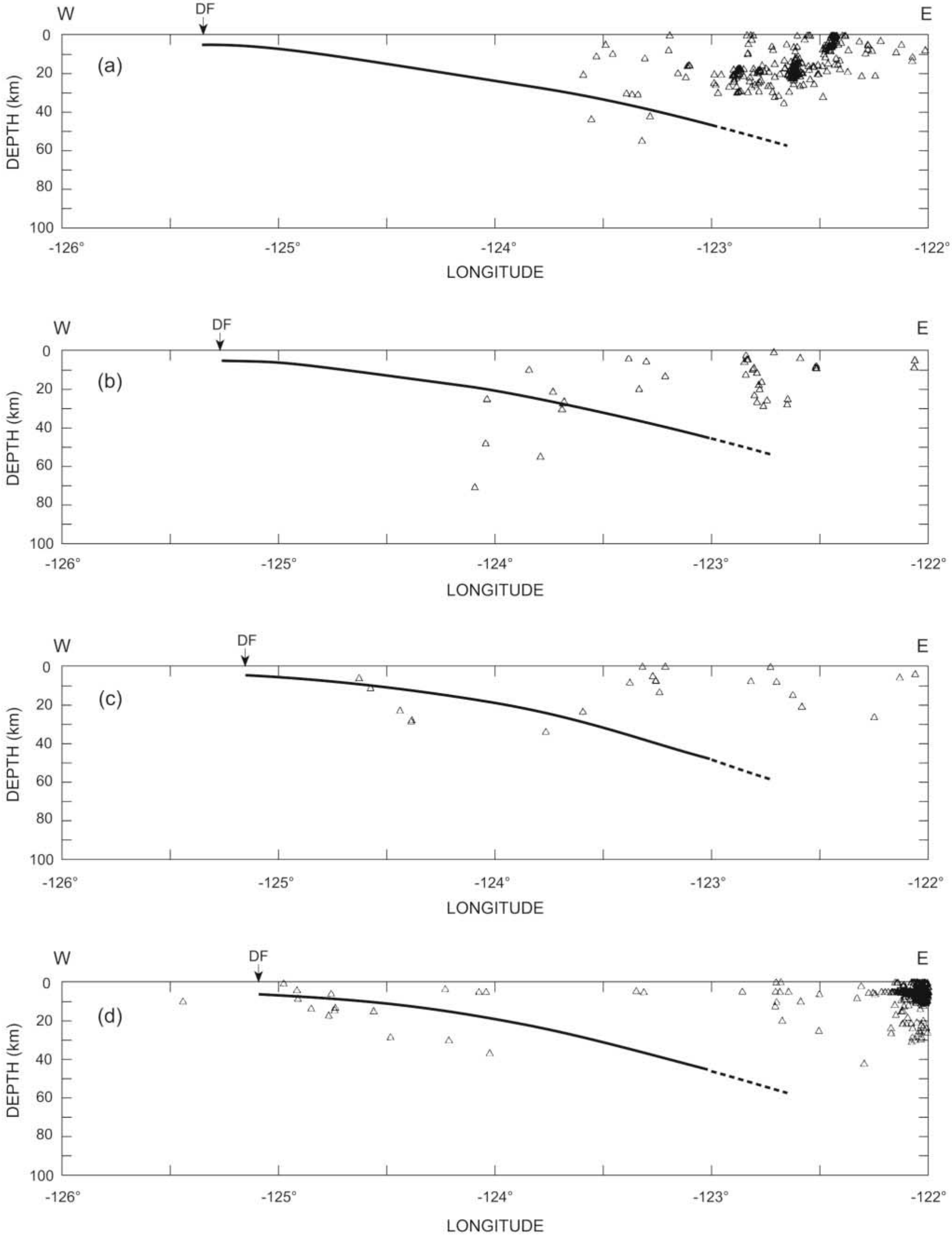


Figure 4. East–west seismicity cross sections (1980–2000) beneath western Oregon centered along one-degree latitude intervals: (a) 45°–46°N; (b) 44°–45°N; (c) 43°–44°N; (d) 42°–43°N. Data principally from the University of Washington. DF is the deformation front.

modern seismographic monitoring began, one of the most notable Gorda block events was the 8 November 1980 **M** 7.4 Eureka, California, earthquake, located about 120 km to the south–southwest of the 1873 event (Fig. 2). The mainshock occurred at a shallow depth (<20 km) and was followed by numerous aftershocks (largest $M_L$ 5.4), which defined a 14-km-long northeast-striking zone (Smith *et al.*, 1993). The sequence appeared to have been the result of rupture of a left-lateral, strike-slip fault, based on focal mechanisms (Smith *et al.*, 1993). A surface-wave magnitude ($M_S$) 7.3 earthquake also occurred just offshore of Cape Mendocino in 1923 (Toppozada and Parke, 1982). Other significant Gorda block earthquakes include events in 1899 (**M** 7.0, Bakun, 2000), 1922 ($M_S$ 7.4), and August 1991 (**M** 7.0). All occurred to the west of the deformation front or trench axis (Bakun, 2000) (Fig. 2).

On 23 August 1962, a $M_S$ 5.6 earthquake occurred offshore just to the southwest of the Oregon–California border (Fig. 5). The focal mechanism of this type B event (Bolt *et al.*, 1968) is very similar to the mechanism of the 1980 earthquake. Smith *et al.* (1993) suggested that earthquakes such as the 1980 event indicate internal deformation of the Gorda block, which is undergoing north–south shortening concentrated on northeast-trending, left-lateral strike-slip faults. The strike-slip focal mechanism of the 1962 earthquake suggests that the northern boundary of the Gorda block is at least as far north as the Oregon–California border (Fig. 2). The relatively large size ($M_S$ 5.6) of the 1962 earthquake and the sparse seismicity of the Juan de Fuca plate west of Oregon (Fig. 2) also suggest a Gorda block origin.

### Puget Sound

Abundant small-magnitude seismicity recorded by the PNSN images quite well the geometry of the seismogenic portion of the subducting Juan de Fuca plate beneath the Puget Sound (Weaver and Baker, 1988; Ludwin *et al.*, 1991). The active zone extends to a depth of 60 km with some events as deep as 100 km (Ludwin *et al.*, 1991; McCrory *et al.*, 2003).

In addressing the question of whether there is a potential for large intraslab earthquakes in the central CSZ, understanding why such large events have apparently been confined to the Puget Sound region, at least in historical times, is important. Note no large intraslab earthquakes have been observed beneath Vancouver Island, also within the northern CSZ (Ludwin *et al.*, 1991). Based on an evaluation of the historical seismicity, Weaver and Baker (1988) suggested that the subducting Juan de Fuca plate changes dip relatively rapidly from a shallow angle of about 11° to about 25° over a distance of a few tens of kilometers beneath the Puget Sound. They suggest that this bend in the plate induces downdip extensional bending stresses that may explain the presence of large-magnitude intraslab earthquakes. Weaver and Baker (1988) concluded that the 1949 earthquake, at a depth of 54 km (Baker and Langston, 1987), was the result of these downdip tensional stresses, based on its strike-slip focal mechanism. The focal mechanisms of the 1965 $m_b$ 6.5 Seattle–Tacoma earthquake displayed normal faulting consistent with bending stresses in the slab. The 2001 Nisqually earthquake focal mechanism is very similar to that of the 1949 event (PNSN Staff, 2001).

Based on an evaluation of the intraslab seismicity, Crosson and Owens (1987) and Weaver and Baker (1988) suggested that an upward arch also occurs in the subducting Juan de Fuca plate beneath the central and southern Puget Sound (Fig. 1). *T* axes from focal mechanism data also support the existence of the arch (Ma *et al.*, 1996). North of the arch, the subducting plate dips to the northeast beneath northwestern Washington, changing to east–southeast on the south side of the arch. No such structure is observed elsewhere along the CSZ (Fig. 1), although there is only limited data between central Oregon and the Mendocino triple junction.

Overall, the pattern of intraslab seismicity in the Puget Sound correlates with the structure of the arch; major clusters of earthquakes may result from bending stresses due to flexure of both the plate and arch (McCrory *et al.*, 2003). McCrory *et al.* (2003) also noted that all the large intraslab earthquakes, including the 1949, 1965, and 2001 events, oc-

Table 1
Significant Gorda Block Mainshocks ($M \geq 6.8$) and Aftershocks

| Date | Location | Mainshock | Source | Largest Aftershock | Number of Aftershocks $M \geq 5.0$ | References |
|---|---|---|---|---|---|---|
| 16 Apr 1899 | Offshore | $M_S$ 7.0 | Intraslab B | MM IV (?) in Eureka | At least 3 events felt in Eureka | Townley and Allen, 1939; Toppozada *et al.*, 1981 |
| 31 Jan 1922 | Offshore | $M_S$ 7.4 | Intraslab B | MM IV in Eureka | 1 + ? | Toppozada and Parke, 1982 |
| 22 Jan 1923 | Just Offshore | $M_S$ 7.3 | Intraslab B | $M_L$ 5.0 | 1 + ? | Toppozada and Parke, 1982 |
| 8 Nov 1980 | Eureka | **M** 7.4 | Intraslab B | $M_L$ 5.4 | 2 | Velasco *et al.*, 1994 |
| 13 July 1991 | Offshore | **M** 6.8 | Intraslab B | $M_L$ 4.9 | 0 | Velasco *et al.*, 1994 |
| 17 Aug 1991 | Offshore* | **M** 7.1 | Intraslab B | $M_L$ 4.2 | 0 | Velasco *et al.*, 1994 |
| 25 Apr 1992 | Cape Mendocino | **M** 7.2 | Megathrust ? | **M** 6.5 and 6.6** | 2 | Oppenheimer *et al.*, 1993 |
| 15 June 2005 | Offshore | **M** 7.2 | Intraslab B | **M** 6.7 | 1 | USGS PDEs |

*Foreshock of $M_S$ 6.3 on 16 August 1991.

**The two largest aftershocks are type B events based on Oppenheimer *et al.* (1993).

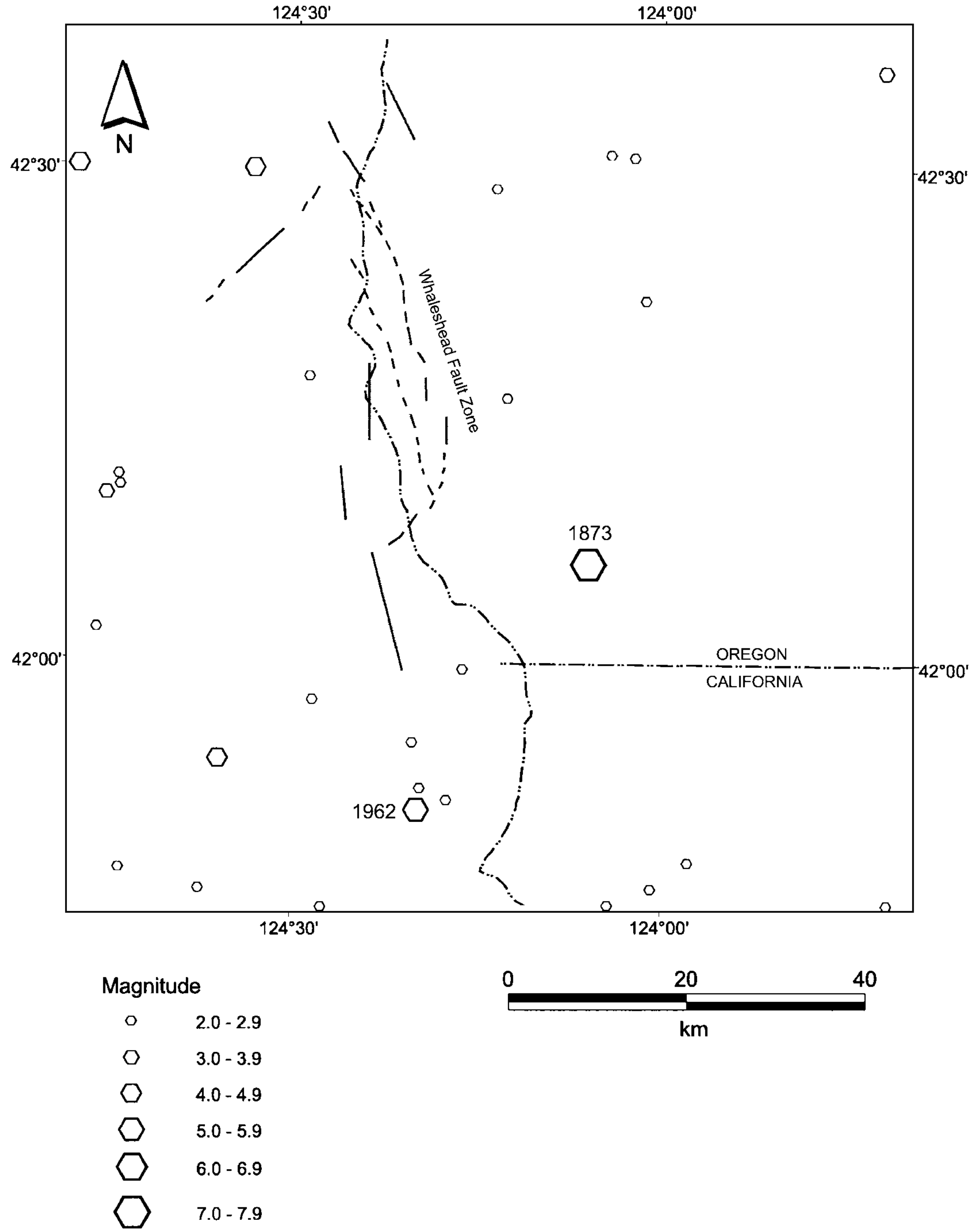


Figure 5. Historical seismicity (1873 to 2000) and crustal faults in the vicinity of the 1873 earthquake. Portrayal of the Whaleshead fault zone from Geomatrix Consultants (1995).

curred along an east–northeast-striking linear trend in the southern Puget Sound that extends from a depth of 40 km to nearly 100 km. They speculate that this trend may represent a tear in the plate along the southern flank of the arch, where stresses are concentrated. These observations are consistent with the global observation that most active subduction zones with large shallow intraslab events have distortions in the downgoing slabs (Astiz *et al.*, 1988).

Intraslab events appear to be concentrated in the oceanic crust and uppermost mantle in the northern CSZ (Preston *et al.*, 2003), and the 2001 Nisqually earthquake probably occurred within the Juan de Fuca plate crust (K. Creager, University of Washington, written comm., 2004). Furthermore, Preston *et al.* (2003) suggest that intraslab earthquakes beneath the Puget Sound can be separated into two groups: those updip of the Moho 45-km depth contour, which occur in the oceanic mantle, and those downdip, which are within the subducting oceanic crust. The former events may be the result of dehydration of serpentinite and the latter of dehydration embrittlement.

## 1873 Brookings Earthquake

On 22 November 1873, at about 9:00 p.m. local time, a large earthquake with a maximum reported intensity of Modified Mercalli (MM) VIII occurred near the Oregon–California border, possibly near the present town of Brookings (Toppozada *et al.*, 1981) (Figs. 5 and 6). Toppozada *et al.* (1981) placed the epicenter of the 1873 earthquake at the center of the MM VI isoseismal contour along the Oregon–California border (Fig. 6); its location uncertainty is large. They also estimated that the size of the event was $M_L$ 6¾, based on the size of its felt area. None of the evaluated accounts report aftershocks occurring after the mainshock (T. Toppozada, California Geological Survey, personal comm., 2002) although such events may have occurred if they were of small size or distant from the coast. The apparent lack of aftershocks led Ludwin *et al.* (1991) to suggest either (1) that the earthquake occurred far enough offshore such that no aftershocks were felt or (2) that it was deep, perhaps within the subducting Gorda plate.

With regard to (2), Astiz *et al.* (1988) noted that nearly half of the $\mathbf{M} \geq 6.5$ shallow and intermediate-depth intraslab earthquakes in their worldwide survey had no aftershocks ($\mathbf{M} > 3.5$), and 85% had five or fewer aftershocks within 1 week of the mainshock. No significant aftershock activity followed the 1949 or 1965 earthquakes, and this appears to be characteristic of Puget Sound intraslab earthquakes (Algermissen and Harding, 1965). The 2001 Nisqually earthquake was followed by only a few aftershocks, with the largest event of $M_L$ 4.3 (PNSN Staff, 2001). Thus, a lack of aftershocks in 1873 would be consistent with a subducting plate origin.

Bakun (2000) re-evaluated the 1873 earthquake and estimated a magnitude of **M** 7.3 (**M** 6.9 to 7.6 at 95% confidence level), considerably higher than the $M_L$ 6¾ estimated by Toppozada *et al.* (1981). This magnitude increase has significant implications of hazard in all of the Pacific Northwest if the 1873 earthquake is an intraslab earthquake because if would be the largest historical event in the northern and central CSZ, exceeding the 1949 Olympia earthquake.

Bakun (2000) relocated the 1873 epicenter onshore, about 15 km north–northeast of Brookings, Oregon (Fig. 5 and 6). This location would place the 1873 earthquake within the central CSZ. At a $1\sigma$ (67%) confidence level, however, the actual location of the event could lie within a northwest–southeast-elongated region 175 km long and 60 km wide and mainly onshore (Bakun, 2000) (Fig. 6). There is no known available information to characterize the ground shaking in the vicinity of Brookings or to indicate whether any small aftershocks were felt.

Based on the Flück *et al.* (1997) model, the top of the downgoing plate lies at a depth of 15 to 20 km beneath the vicinity of the 1873 epicenter (Fig. 7). If the 1873 earthquake was an intraslab earthquake, it would have likely occurred within the oceanic crust or uppermost mantle at a depth range of 15 to 30 km (type B) (Fig. 7), considerably shallower than the large type A events in the Puget Sound region. This shallower depth also places the 1873 earthquake in a portion of the subducted plate overlain by North American crust (Fig. 7) rather than North American mantle, as is probably the case in the Puget Sound. However, there are no definitive data to constrain the focal depth of the earthquake. Obviously, if the earthquake had a crustal origin, its focal depth would be less than 15 to 20 km (Fig. 7).

A significant characteristic of the 1873 earthquake is the elongated nature of its isoseismal contours (Fig. 6). Although this elongation may be an artifact of the population being concentrated west of the Cascades, the pattern is consistent with an earthquake occurring on a generally north-striking fault. Earthquakes occurring on the strike-slip faults of the San Andreas fault system exhibit an isoseismal pattern elongated along the strike of the faults (see isoseismal maps in Toppozada *et al.*, 1981). The isoseismal pattern for the 1980 Eureka earthquake, though offshore and affected by observer bias, also exhibits a somewhat elongated north–south trend (Stover and von Hake, 1982). The earthquake resulted from the rupture of a northeast-striking, strike-slip fault (Smith *et al.*, 1993), which probably accounts for its elongated isoseismal pattern. An elongated isoseismal pattern suggests an extended source that ruptures in a strike-slip manner. The intensity patterns for the Puget Sound intraslab events do not show this elongated nature.

If the 1873 earthquake was as large as suggested by Bakun (2000), aftershocks of at least **M** 5 would typically be expected. If Bakun's (2000) onshore location is close to the actual location, it is likely that the 1873 earthquake was deficient in moderate- to large-sized aftershocks, because they should have been felt. Despite the sparse settlement in the epicentral area, aftershocks of **M** 6.0 and greater would most certainly have been reported, and most likely those of **M** 5.0 and greater as well. Ludwin *et al.*'s (1991) alternative explanation that the mainshock may have been well offshore is still possible, although the analysis by Bakun (2000) indicates that most of the area within the 67% confidence level is onshore (Fig. 6). It is very possible that aftershocks smaller than **M** 5.0 followed the 1873 aftershock.

### 1873 Source

Could the 1873 earthquake have been the result of rupture along a crustal fault, a megathrust, or is it indeed an intraslab event? Assuming a crustal origin, it is likely that it occurred on a fault expressed at the surface rather than on a blind or buried fault given its large magnitude. The Whaleshead fault zone is a north–northwest-striking, left-lateral, strike-slip fault aligned along the coast north of Brookings

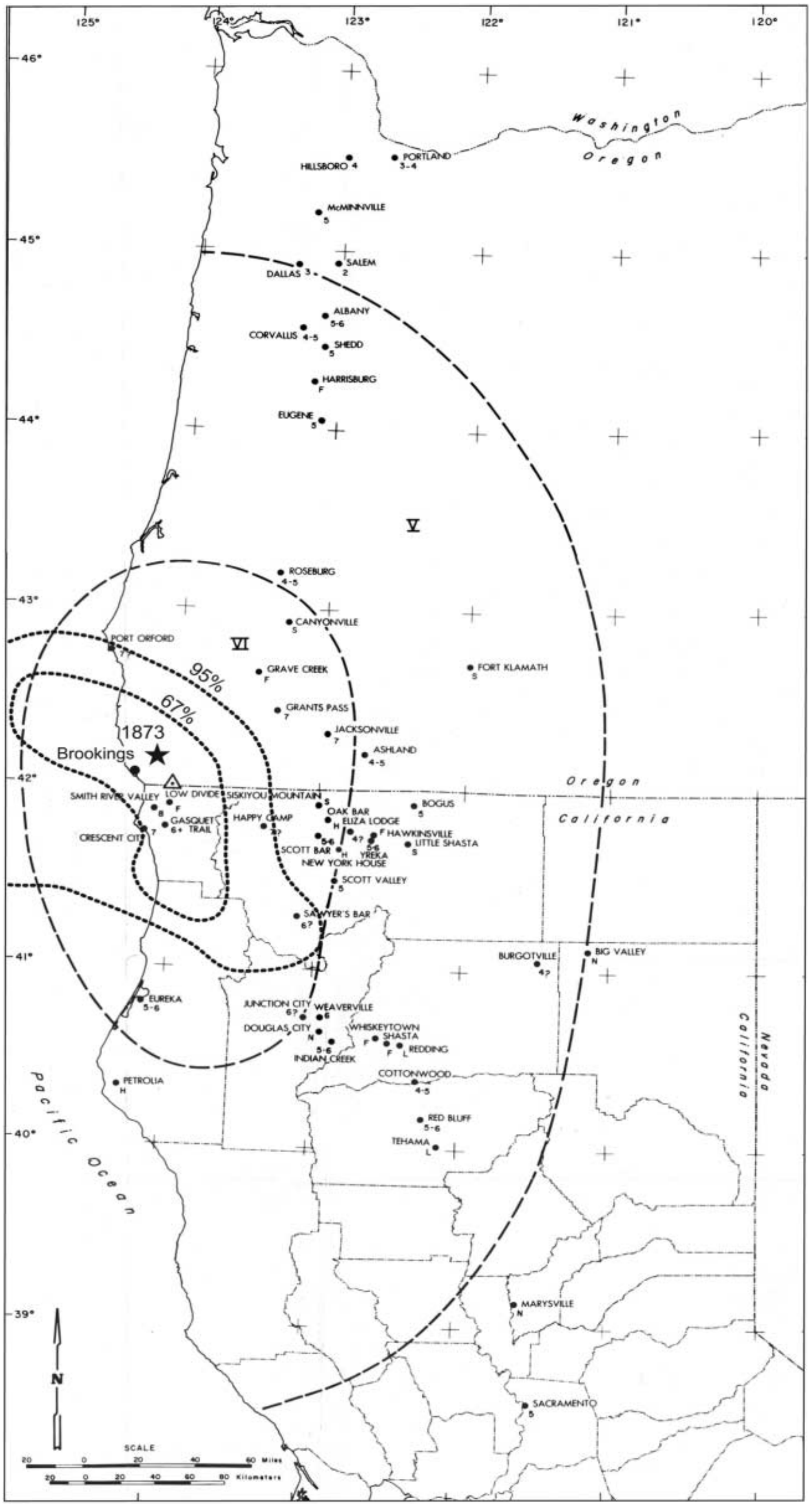


Figure 6. Isoseismal (dashed lines) map of the 1873 earthquake from Toppozada *et al.* (1981). Numbers indicate MM intensities. The epicenter estimated by Toppozada *et al.* (1981) is indicated by the triangle. Also shown are the epicenter (star) estimated by Bakun (2000) and his 67% and 95% confidence level contours regarding the location (dotted lines).

(Pezzopane, 1993; Fig. 5). Differential uplift of marine terraces in Whaleshead Cove indicates that the Whaleshead fault zone has been active in the late Pleistocene (Kelsey and Bockheim, 1994). There are no clear Holocene scarps along this trace of the fault zone, although it is unlikely one could be identified because of the steep and undulating topography (H. Kelsey, Humboldt State University, written comm., 2002). Geomatrix Consultants (1995) suggested that, given its setting within the active accretionary wedge above the CSZ, the Whaleshead fault zone most likely ruptures only together with the CSZ megathrust and is not an independent earthquake source.

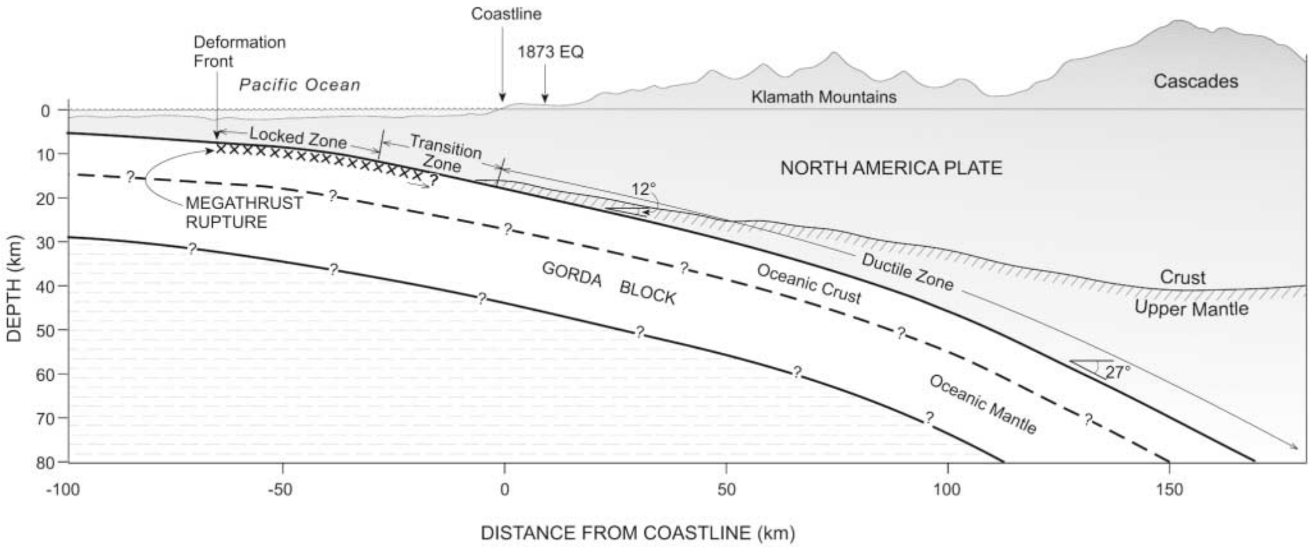


Figure 7. Cross section through the central CSZ at the latitude of the Oregon–California border. Geometry of the subduction zone is based on the model of Flück *et al.* (1997).

Moderate-to-large crustal earthquakes in the Pacific Northwest are generally followed by significant aftershocks (Ludwin *et al.*, 1991). The 1872 **M** 6.5 to 7.0 North Cascades, Washington, earthquake was followed by numerous felt aftershocks (Ludwin *et al.*, 1991). Based on the lack of an apparent crustal fault source and lack of aftershocks, it is unlikely that the 1873 earthquake had a crustal origin within the North America plate.

It seems also unlikely that the 1873 earthquake had a megathrust origin, because interplate earthquakes, almost without exception, have robust aftershock sequences. The 1992 Cape Mendocino earthquake, which may have occurred on the megathrust (Oppenheimer *et al.*, 1993), had numerous large aftershocks (Table 1). Large interplate earthquakes generally have aftershocks as large as one magnitude smaller than the mainshock (L. Ruff, University of Michigan, written comm., 2002).

Given these observations, an intraslab origin for the 1873 earthquake appears to be the most probable scenario. However, I believe the evidence, albeit circumstantial, suggests that the event was a shallow type B intraslab earthquake that occurred within the seismically active Gorda block rather than a type A event in the quiescent Juan de Fuca plate in Oregon. The uncertainties in the location of the Juan de Fuca plate–Gorda block boundary would certainly allow the 1873 earthquake to have occurred in the Gorda block and not the Juan de Fuca plate (Fig. 1). The absence of aftershocks, which is typical of type A intraslab earthquakes, is probably not applicable to the shallower 15- to 30-km-deep 1873 event. Like the 1923 and 1980 Gorda block events, the 1873 earthquake may have had aftershocks in the **M** 4 to 5 range that went unreported (Table 1). The elongated isoseismal pattern of the 1873 earthquake suggests a northerly-trending strike-slip mechanism similar to the 1980 earthquake. A Gorda block origin is consistent with the historical record of multiple large type B earthquakes, in contrast to the probable complete absence of such events in the central CSZ (Table 1). The 1873 earthquake probably occurred in a more shallow portion of the subducting plate (Fig. 7), which is not being subjected to the same stresses (due to plate flexure), temperature, or pressure conditions as those causing large intraslab type A earthquakes in the Puget Sound region.

## Comparison with Other Young Subduction Zones

Three subduction zones, the Nankai trough, northern Mexico, and southernmost Chile, are similar to the CSZ. All three zones are subducting young, hence warm, oceanic lithosphere (<20 m.y.) and have convergence rates slower than 50 mm/yr.

### Nankai Trough

The Nankai and Kyushu troughs comprise the southwestern Japan subduction zone, where the Philippine Sea plate is subducting beneath the Eurasian plate (Wang *et al.*, 2004). The convergence rate is about 45 mm/yr The age of the subducting plate in the Nankai trough is estimated to be from 15 to 17 m.y. old (Hyndman *et al.*, 1995). Along the margin of the Nankai trough, the geometry of the subduction zone is complex, with the downgoing plate being warped or contorted or having a tear between the Nankaido and Tonankai areas (Hyndman *et al.*, 1995).

A well-defined Wadati–Benioff zone is observed to a depth of about 200 km southwest of Nankai along the older (45 m.y. old) and colder Kyushu trough (Wang *et al.*, 2004). In contrast, typical of young warm slabs, the Wadati–Benioff zone extends to a shallow depth of about 50 km in the Nankai trough (Wang *et al.*, 2004). Only two large shallow intraslab earthquakes are known to have occurred in the Nankai trough, and both were confined to the Geiyo area: the 1905 $M$ 6.8 and 2001 **M** 6.7 events. Both events occurred at a depth of about 50 km at the southwestern end of the Nankai trough (Cummins *et al.*, 2002; Zhao *et al.*, 2002) and were located in an area where the plate changes dip (Zhao *et al.*, 2002) and where a fossil spreading ridge appears to be subducting. Seven smaller **M** 6 to 6.5 events have also occurred in the Geiyo area in the past 100 years (D. Zhao, Ehime University, written comm., 2002). All these shallow earthquakes appear to be concentrated in this area of complex plate geometry similar to the Puget Sound region.

As in the CSZ, no other large (**M** $\geq$ 6.5) intraslab earthquakes have been observed in the Nankai trough, based on the United States Geological Survey (USGS) Preliminary Determination of Epicenter (PDE) database and the National Oceanic and Atmospheric Administration's (NOAA) catalog of significant worldwide earthquakes (2150 B.C. to 1994) (http://neic.usgs.gov/neis/epic/epic_global.html), although Cummins *et al.* (2002) report possibly a third large intraslab earthquake, the $M$ 6.8 Yoshino event in 1952, which occurred at a depth of 60 km in the central portion of the Nankai trough.

### Northern Mexico

The northern Mexico subduction zone may be most comparable to the CSZ because it is a small plate containing only young lithosphere (Rogers, 1988). Although not as well understood as the subduction of the Cocos plate, the Rivera plate is subducting beneath the North America plate at a rate that ranges from 12 to 30 mm/yr (average 23 mm/yr) (Minster and Jordan, 1978). The age of the seafloor near the trench is 10 m.y. (Rogers, 1988). Examination of the historical seismicity from 1973 to 2003 based on the USGS's PDEs indicates that only four earthquakes of **M** $\geq$4.0 have occurred within the subduction zone with assigned focal depths greater than 33 km. The largest known earthquake since 1900 was a type B event in 1948 of **M** 6.9 well offshore of the coastline. In contrast, Singh *et al.* (1981) identified 13 probable shallow and intermediate-depth intraslab earthquakes of $M_S$ $\geq$7.0 from 1806 to 1979 in the Mexican subduction zone to the southeast.

### Southernmost Chile

South of about latitude 46°S to about 55°S, the Antarctica plate is subducting beneath South America. In the past 25 m.y., the convergence rate has been fairly uniform at approximately 20 mm/yr (Cande and Leslie, 1986). Given the remoteness of the region from global seismographic coverage, it is difficult to assess whether any shallow intraslab earthquakes have occurred in the southernmost Chile subduction zone. Examination of the Regional Center of Seismology for South America (CERESIS) catalog (1471–1981) and USGS PDEs reveals that nine earthquakes of **M** $\geq$6.5 have occurred within the subduction zone from 1879 to present. Only one event has a focal depth estimate: the 1949 $M_S$ 7.8 earthquake at the very southern end of the subduction zone had an estimated depth of 60 km, although this value must be poorly constrained given the seismographic coverage at the time. It is possible that the 1949 event was actually a megathrust earthquake, given that its size would be at the upper end of intraslab earthquakes. A **M** 6.3 earthquake in 1963 had an assigned depth of 48 km.

### Large Shallow Intraslab Seismicity Rates

To place the rates of large shallow intraslab earthquakes in the CSZ in the context of other subduction zones worldwide, I have estimated the rates of shallow **M** 6.5 to 8.0 intraslab events by examining the NOAA worldwide historical catalog, the USGS PDEs, the CERESIS catalog, and other available sources (Table 2). To allow a direct comparison between subduction zones, the rates have been normalized per year and for a subduction zone length of 1000 km, similar to the length of the CSZ. Table 2 also lists the maximum normalized average recurrence interval (the inverse of the normalized minimum rate).

These rates are minimum values because many events have no computed focal depths and thus were not included in the rate calculations. A problem with several subduction zones is the common practice of fixing the earthquake's depth at 33 km to help constrain the location solution. Such fixed-depth events may well have occurred at depths greater than 40 km. Because of the large uncertainties in computed focal depths, at least $\pm$10 km, from teleseismic locations, I have included events as deep as 80 km in the rate calculations. This inclusion may also help adjust the rate for events assigned a 33-km depth, which may have been deeper. Some events with depths of 40 to 50 km have been assigned magnitudes greater than $M_S$ 8.0. I assume that these events were likely earthquakes occurring on the megathrust, and thus I do not count them in the intraslab rates. It is interesting to note the inconsistencies among the catalogs in terms of both location and focal depth. Events as large as **M** 7.5 sometimes occur in one catalog but not in others, even as recently as 1985.

For the reasons given above, the computed rates listed in Table 2 should be viewed only as first-order estimates. They are probably most reliable for North and Central America and Japan and least certain for the western Pacific (e.g., Izu-Bonin) and southwest Pacific (e.g., Sumatra). Probably the most complete record of observation is from 1973 to the present, corresponding to the USGS PDEs. I computed rates for the periods both back to 1973 and to the beginning of the historical record for each subduction zone and generally selected the highest rate. The rates for Kamchatka, Kuriles, Philippines, Kermadec, Scotia, Solomons, Sumatra, and Java are too low because a large number of events have fixed

Table 2
Worldwide Rates of Shallow (40–80 km Deep) Intraslab Earthquakes ($M$ 6.5 to 8.0)

| Subduction Zone | Search Area | | Approx. Length (km) | Start of Historical Period | Minimum No. $M \geq 6.5$ | Minimum Rate No. / 1000 km/yr | Maximum Normalized Average Recurrence Interval (years) |
|---|---|---|---|---|---|---|---|
| Northern Cascadia (Juan de Fuca) | 49.0° N | 46.5° N | 260 | 1833 | 3 | 0.068 | 15 |
| Central Cascadia (Juan de Fuca) | 46.5° N | 42.0° N | 500 | 1850 | 0 | 0.000 | > 150 |
| Southern Cascadia (Gorda) | 42.0° N | 40.0° N | 240 | 1850 | 0 | 0.000 | > 150 |
| Alaska (Pacific) | 160.0° W | 140.0° W | 1350 | 1900 | 17 | 0.221 | 5 |
| Aleutians (Pacific) | 180.0° W | 160.0° W | 1600 | 1911 | 19 | 0.330 | 3 |
| Caribbean (North America) | 19.0° N | 11.0° N | 1000 | 1965 | 1 | 0.026 | 38 |
| Northern Mexico (Rivera) | 21.0° N | 19.0° N | 250 | 1900 | 0 | 0.000 | > 100 |
| Mexico (Northern Cocos) | 19.0° N | 15.0° N | 775 | 1806 | 27 | 0.177 | 6 |
| Central America (Southern Cocos) | 15.0° N | 7.0° N | 1500 | 1973 | 10 | 0.222 | 5 |
| Colombia/Ecuador (Northern Nazca) | 7.0° N | 0.0° | 850 | 1904 | 9 | 0.113 | 9 |
| Ecuador/Peru (Northern Nazca) | 0.0° | 16.0° S | 2150 | 1897 | 16 | 0.071 | 14 |
| Northern Chile (Central Nazca) | 16.0° S | 23.0° S | 800 | 1906 | 18 | 0.234 | 4 |
| Central Chile (Central Nazca) | 23.0° S | 34.0° S | 1200 | 1909 | 35 | 0.310 | 3 |
| Southern Chile (Southern Nazca) | 34.0° S | 46.0° S | 1300 | 1914 | 7 | 0.061 | 17 |
| Southernmost Chile (Antarctica Plate) | 46.0° S | 55.0° S | 1000 | 1910 | 1 | 0.011 | 93 |
| Kamchatka (Pacific) | 56.0° N | 50.0° N | 800 | 1973 | 7 | 0.292 | 3 |
| Kuriles (Pacific) | 50.0° N | 42.0° N | 1200 | 1973 | 15 | 0.417 | 2 |
| NE Japan (Pacific) | 42.0° N | 35.0° N | 900 | 1897 | 19 | 0.199 | 5 |
| Nankai (Philippine) | 35.0° N | 32.0° N | 630 | 1899 | 3 | 0.045 | 22 |
| Kyushu (Philippine) | 32.0° N | 30.0° N | 330 | 1899 | 7 | 0.204 | 5 |
| Ryukyus/Taiwan (Philippine) | 30.0° N | 23.0° N | 1200 | 1901 | 12 | 0.098 | 10 |
| Izu-Bonin (Pacific) | 35.0° N | 22.0° N | 1550 | 1916 | 5 | 0.037 | 27 |
| Philippines (Philippine) | 15.0° N | 1.0° S | 1600 | 1973 | 11 | 0.229 | 4 |
| Marianas (Pacific) | 22.0° N | 10.0° N | 1800 | 1973 | 1 | 0.019 | 54 |
| Tonga (Pacific) | 14.0° S | 30.0° S | 1900 | 1948 | 13 | 0.124 | 8 |
| Kermadec (Pacific) | 30.0° S | 37.0° S | 850 | 1968 | 5 | 0.168 | 6 |
| New Zealand (Pacific) | 37.0° S | 42.0° S | 625 | 1921 | 6 | 0.117 | 9 |
| Solomons (Indo-Australian) | 5.0° S | 13.0° S | 2150 | 1973 | 32 | 0.496 | 2 |
| New Hebrides (Indo-Australian) | 13.0° S | 24.0° S | 1600 | 1973 | 8 | 0.167 | 6 |
| Sumatra (Indo-Australian) | 92.0° E | 104.0° E | 1900 | 1973 | 3 | 0.053 | 19 |
| Java (Indo-Australian) | 104.0° E | 120.0° E | 1800 | 1921 | 3 | 0.020 | 49 |
| Scotia (Antarctic) | 54.0° S | 61.0° S | 800 | 1973 | 5 | 0.208 | 5 |

1. To incorporate the estimated minimum uncertainty in focal depths of intraslab earthquakes of $\pm$ 10 km, particularly those occurring prior to 1960, I have included in the compilation those events which have assigned depths up to 80 km.
2. Subducting plate in parentheses.
3. $m_b$ values assumed to be equal to **M**.
4. Historical seismicity from NOAA catalog, USGS PDEs, and CERESIS catalog.

focal depths of 33 km, and I believe that most of these events were shallow intraslab earthquakes. For example, if I include the 33-km deep events in the count for Sumatra, the rate would be five times as high, at 0.263 (Table 2).

For subduction zones where the normalized rates are zero (no **M** $\geq$6.5 earthquakes) such as the central CSZ, the associated maximum normalized average recurrence intervals are minimums estimated from the length of the historical record (Table 2).

Figure 8 is a plot of convergence rates (mm/yr) versus plate age (m.y.) for the most significant subduction zones or sections of subduction zones worldwide in a manner first illustrated by Ruff and Kanamori (1980). I have indicated by different symbols the maximum normalized average recurrence intervals for shallow intraslab (**M** 6.5 to 8.0) events from Table 2. Of the 32 subduction zones analyzed, 19 have maximum normalized average recurrence intervals of 10 years or less (Table 2; Fig. 8). Four more subduction zones have relatively short recurrence intervals of 11 to 20 years, including the northern CSZ. With the exception of the northern CSZ and the Colombia/Ecuador subduction zone, the zones with the smallest recurrence intervals ($\leq$10 years) are older than 20 m.y. and have a convergence rate of more than 50 mm/yr (Fig. 8).

The four subduction zones with very long recurrence intervals (approximately more than 100 years), the central and southern CSZ, northern Mexico, and southernmost Chile, are very young (< 20 m.y.) and slow moving (< 40 mm/yr). Four old (>100 m.y.) subduction zones, the Caribbean, Marianas, Izu-Bonin, and Java, have unexpectedly long recurrence intervals (Table 2; Fig. 8). As mentioned, the intervals for Java and possibly for the other three subduction zones are too long because of poor focal depth control.

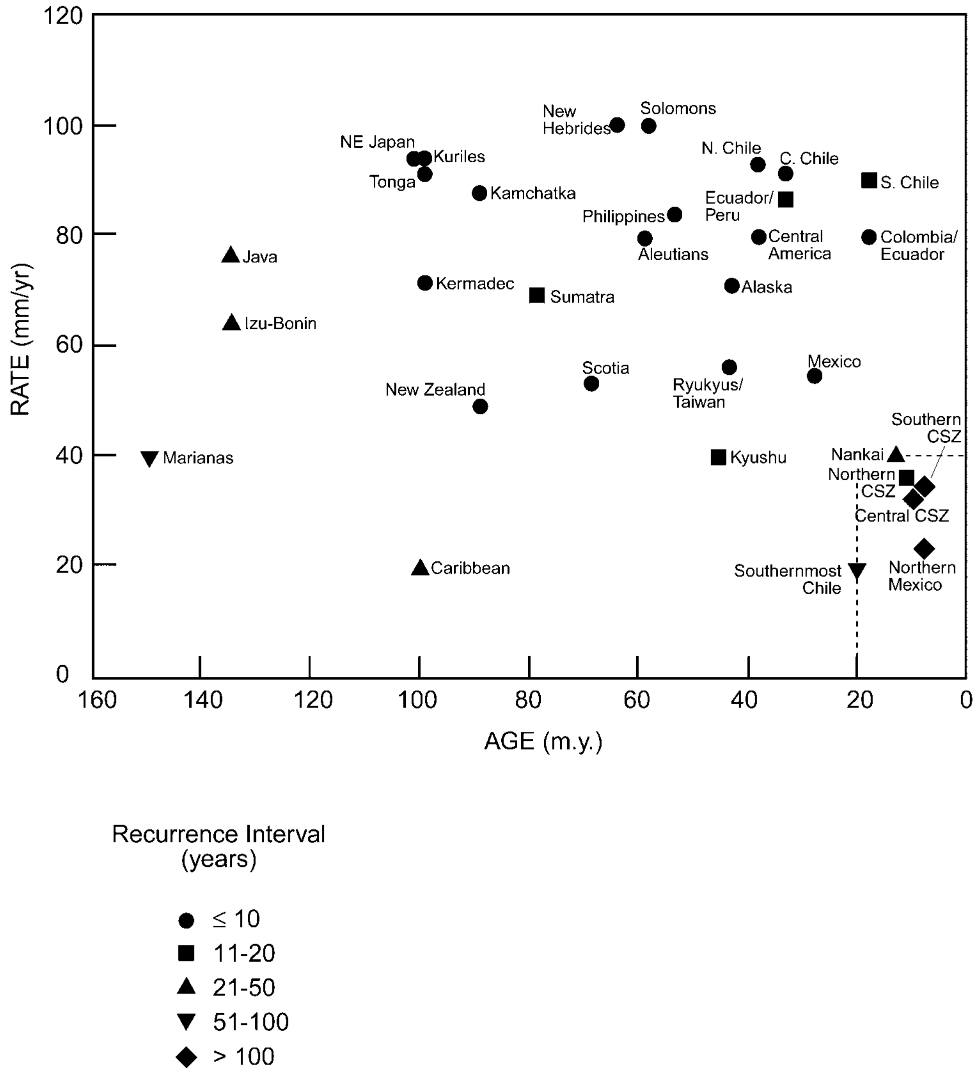


Figure 8. Normalized maximum recurrence intervals for large (**M** 6.5–8.0) shallow intraslab earthquakes of some significant subduction zones worldwide as a function of convergence rates and plate ages. Convergence rates and plate ages are generally from Astiz *et al.* (1988). Plate age and convergence rates for Cascadia are from Wilson (2002) and Wells *et al.* (2002), respectively. The intraslab recurrence intervals are minimum values and may be too high by a factor of 2 or more (e.g., Sumatra, Java, Kermadec).

Based on this comparison (Fig. 8), the central CSZ's lack of large shallow intraslab earthquakes is not uncommon among subduction zones worldwide that are similarly young and slowly subducting and, as I will discuss, probably hot.

## Factors Controlling CSZ Intraslab Seismicity

Similarities in plate age and convergence rate between the central and southern CSZ and northern Mexico, and the southernmost Chile subduction zones suggest that large intraslab earthquakes (**M** ≥6.5) are generally absent in these zones because the subducting plates are youthful and slowly subducting. All are less than 20 m.y. old, and convergence rates are less than 40 mm/yr. Their common characteristics of youth and low convergence rate are consistent with a number of hypotheses, as described subsequently. The Nankai and northern CSZ exhibit a similar characteristic in that the large intraslab earthquakes that occur are confined to a distorted portion of the downgoing plate.

### Effects of Temperature

Isacks *et al.* (1968) were the first to suggest that the maximum depth of intraslab earthquakes corresponds to a critical temperature where strain accumulation or strain release can no longer occur because of thermal weakening of the slab. Vlaar and Wortel (1976) attributed the termination of intraslab seismicity at different depths in different regions to a depth-dependent critical temperature. Molnar *et al.* (1979) inferred that intermediate and deep earthquakes probably occur in the coldest portion of the slab. Kirby *et al.* (1996) showed that the maximum depth of intraslab earthquakes varies as a function of a thermal parameter, the product of the vertical descent rate and plate age.

The thermal structure of a subduction zone depends primarily on the temperature of the plate when it enters the trench and how it warms up as it descends (Stein and Stein, 1996). Peacock (1996) specified six factors that control the temperatures within subducting plates: the age of the incoming lithosphere; convergence rate; geometry of subduction; radioactive heating; induced convection in the overlying mantle wedge; and the rate of shear heating along the interface.

Oceanic plates generally remain colder, denser, and mechanically stronger than the surrounding mantle because they subduct faster than the time it takes for them to warm up (Stein and Stein, 1996). Although plates are warm when they are generated at mid-ocean ridges, they lose heat and cool along the seafloor as they move toward the trench. As the slab descends, it warms primarily by conduction through its upper and lower surfaces (Molnar *et al.*, 1979).

The crust in young subducting plates (<30 m.y.) entering trenches may be as much as 300°C. warmer than older plates (Kirby *et al.*, 1996). This observation is consistent with the maximum depth of intraslab seismicity in young plates being limited to 50 to 100 km (Kirby *et al.*, 1996). The northern CSZ, Mexico, and Nankai subduction zones exhibit some of the shallowest seismicity worldwide, reflecting the relatively hot thermal conditions of their oceanic crust and uppermost mantle (Kirby *et al.*, 1996). The maximum intraslab seismicity depths of 40 km in the Gorda slab indicate an even warmer environment than the Puget Sound region (60 km), or, conversely, the Juan de Fuca plate in the northern CSZ appears to be cooler than both the Gorda block and the central CSZ (see next section).

Molnar *et al.* (1979) first noted that the absence of shallow intraslab earthquakes beneath Oregon and Washington was not surprising owing to high temperatures. They defined a "cutoff temperature" as the temperature above which rock should be too warm, and thus too weak, to support stresses needed for seismogenesis. They calculated the cutoff temperature for intraslab earthquakes as about 600° ± 100°C at a depth of 200 km, increasing to 830° ± 50°C at 650 km depth. The observation that the cutoff temperature increases with depth suggests that pressure may also be important in constraining seismogenesis with depth (Molnar *et al.*, 1979). Because shallow earthquakes in oceanic plates (type C, Fig. 3) also occur only when temperatures are less than about 800°C, Stein and Stein (1996) suggested similar temperatures may limit earthquake occurrence throughout the descending plate. Because of the transition from basalt to eclogite in the downgoing Juan de Fuca plate at a depth of 40 km, the cutoff temperature may differ above and below the bend in the slab even without considering the effect of pressure, which would increase the cutoff temperature.

For the shallow depth range of 40 to 70 km in the Juan de Fuca plate, the cutoff temperature should be less than 600°C, possibly on the order of 500° ± 100°C based on a simple extrapolation of the temperatures estimated by Molnar *et al.* (1979). This latter temperature range is in the range appropriate for crustal rock in continental settings (Chen and Molnar, 1983; Wong and Chapman, 1990). In the analysis by Hyndman and Wang (1995), modeled temperatures on the CSZ megathrust are lowest on a profile across northwestern Washington and are increasingly higher in order along the Columbia River, central Oregon, and southern Vancouver Island profiles. Temperatures are high along the megathrust under southern Vancouver Island because of the younger plate age (Hyndman and Wang, 1993). This pattern is reflected in the surface heat flow along the margin, which is primarily a function of the dip of the plate and the thermal conductivity of the overlying crust (Hyndman and Wang, 1995). Heat flow is lowest across northwestern Washington, where the plate dip is only about 5°.

In contrast, low thermal conductivity and low heat-generating Siletz/Crescent basalts in the crust overlying the downgoing slab (dip assumed at 11° at 60 km depth) along the central Oregon profile increase the temperatures along the megathrust (Hyndman and Wang, 1995). At a distance of 150 km from the deformation front (Fig. 1), the temperature difference along the megathrust between northwestern Washington and central Oregon is about 50°C (Hyndman and Wang, 1995). These observations of the megathrust temperatures indicate that the underlying intraslab temperatures of the Juan de Fuca plate are higher beneath western Oregon and southern Vancouver Island than beneath northwestern Washington.

## Thermal Modeling of the Central Cascadia Subduction Zone

Hyndman and Wang (1993, 1995) and Oleskevich *et al.* (1999) performed thermal modeling of the CSZ partially constrained by heat flow measurements to estimate the downdip temperatures on the megathrust in an effort to characterize the extent of the seismogenic zone. In the following, I follow the approach of Wang *et al.* (1995) and Wong and Harris (2003) to estimate the intraslab temperatures at depths of 40 to 100 km beneath the central CSZ at the latitude of 44.5°N.

Wang *et al.* (1995) solve the thermal equation for a steady-state underthrusting slab using a finite-element method. Within each element, the thermal properties are uniform but the temperature may vary quadrantically. The principal inputs into the modeling are the plate convergence rate, age of the subducting plate, thickness of the insulating sediments on the incoming crust, and the downdip geometry of the slab (Wang *et al.*, 1995). Several assumptions are made in the modeling, and I refer the reader to Hyndman *et al.* (1995) for a detailed discussion.

In the modeling, I adopt the downgoing slab geometry proposed by Rondenay *et al.* (2001) as illustrated in Figure 9a. The 27° dip of the subducting plate below a depth of 35 km is significantly greater than the approximate 13° dip assumed by Oleskevich *et al.* (1999) and the 11° previously adopted by Hyndman and Wang (1995) at the downdip end of the megathrust. A plate age of 8 m.y., incoming sediment thickness of 3.2 km, and a temperature at the top of the oceanic plate at the deformation front of 240°C are adopted from Oleskevich *et al.* (1999) (Table 3). An average convergence rate of 31 mm/yr for the central CSZ, significantly lower then the value of 40 mm/yr previously used in thermal modeling (e.g., Oleskevich *et al.*, 1999) is adopted from Wells *et al.* (2002). The model is divided into seven thermostratigraphic units (Fig. 9a): the accreted prism (subduction zone complex) and a thin sediment layer blanketing the downgoing slab, Crescent–Siletz basaltic crust, oceanic and continental crust, oceanic mantle, and the continental mantle wedge. Thermal model parameters for each unit, thermal conductivity, radioactive heat generation, and heat capacity (for the subducting plate only), are adopted from Oleskevich *et al.* (1999) (Table 3). The finite-element mesh is shown in Figure 9b. Elements are 20 km long and vary in thickness. Owing to limitations of the modeling, the bend in the plate at a depth of 70 km could not be simulated, but this should not influence the computed temperatures at shallower depths.

The modeling results (Fig. 10) indicate that (1) the downgoing slab maintains a temperature warmer than 500°C beneath the layer of accreted sediments at depths greater than 10 km and (2) it has a temperature range of 600° to 1200°C at depths between 40 and 70 km, beneath the bend in the plate. This temperature range is above the possible cutoff temperature of 500° ± 100°C. Thus the hot temperatures of the downgoing Juan de Fuca plate in the central CSZ may be the principal reason for the absence of seismicity and, in particular, the lack of large shallow intraslab earthquakes in historical times.

The hot temperatures of the subducting Juan de Fuca plate in the central CSZ are probably due to its young age, the relatively slow convergence rate, and the insulating effect of the overlying crust. Because the downgoing plate subducts at a slower rate beneath western Oregon (~31 mm/yr) than along the northern CSZ (~35 mm/yr), the slab has a longer time to be warmed up by the mantle (Riddihough, 1984; Spence, 1989). The central CSZ is warmer than the rest of the CSZ because the age of the slab beneath western Oregon is younger (11 to 12 m.y. old) than for example, beneath western Washington (~14 to 15 m.y.) (Wilson, 2002). As stated earlier, because of the thicker and more extensive low conductivity Siletzia crust in coastal Oregon, the plate is not allowed to cool as fast as beneath the Puget Sound (Hyndman and Wang, 1995). The Siletzia terrane is about 30 km thick beneath western Oregon, thinning to the north, where it is less than 10 km thick beneath southern Vancouver Island (Trehu *et al.*, 1994). The terrane also extends much further to the east beneath western Oregon than elsewhere along the CSZ (Hyndman and Wang, 1995).

Two previous thermal modeling studies by Wang *et al.* (1995) and Peacock *et al.* (2002) that yield information on intraslab temperatures were examined as part of this analysis. It should be noted, however, that more current input (e.g., convergence rates) would change the results of their modeling. The intraslab isotherms near Vancouver Island (Wang *et al.*, 1995; Fig. 11) suggest that temperatures range from about 550° to 1100°C at depths of 40 to 70 km in the subducting oceanic plate. Active intraslab seismicity is observed to a depth of 40 km, where the basalt to eclogite transition occurs (Fig. 11).

Peacock *et al.* (2002) calculated the thermal structure of the slab at the boundary of the central and southern CSZ (latitude 42°N) based on a preliminary finite fault model. At the depth range of 40 km in the downgoing slab, they predict a temperature of about 600°C. The Gorda block's Wadati–Benioff zone terminates at a depth of 40 km and hence possibly at a temperature of about 600°C, based on preliminary modeling. The 40-km maximum depth of intraslab seismicity in the Gorda block also coincides with the possible basalt/gabbro to eclogite transition in the slab.

### Mechanical Coupling/Stress State

In addition to the thermal state, other factors such as the tectonic stress state may be significant in dictating the occurrence and character of intraslab seismicity. Heaton and Kanamori (1984) first noted that the CSZ is similar to other strongly coupled subduction zones. Because the Juan de Fuca plate is young and warm, it is highly buoyant, resulting in strong mechanical coupling with the North America plate. The inverse relationship between age of the subducting plate and interplate mechanical coupling was first recognized by Ruff and Kanamori (1980). The degree of seismic coupling has played strongly in the arguments of whether the CSZ could produce large-magnitude megathrust earthquakes (e.g., Heaton and Kanamori, 1984). Ruff and Kanamori (1980) suggested that the degree of seismic coupling could be represented by the largest megathrust earthquake that has occurred in that zone. The occurrence of the 1700 **M** 9 earthquake suggests that the CSZ is strongly coupled. Astiz *et al.* (1988) also suggested that the CSZ is strongly coupled, based on the focal mechanisms of the 1949 and 1965 Puget Sound earthquakes. They suggested that in strongly coupled subduction zones, downdip tensional stresses are concentrated in the slab near the lower edge of the megathrust.

Strong interplate coupling due to the buoyancy of the plate can lead to slowed subduction, resulting in reduced stresses in the overlying crust (Riddihough, 1984; Rogers *et al.*, 1996) and possibly in the interior of the underlying subducting slab. The absence of a Wadati–Benioff zone in the central CSZ would be consistent with a lower level of tectonic stresses south of Puget Sound.

Rogers *et al.* (1996) also offered two contrasting hy-

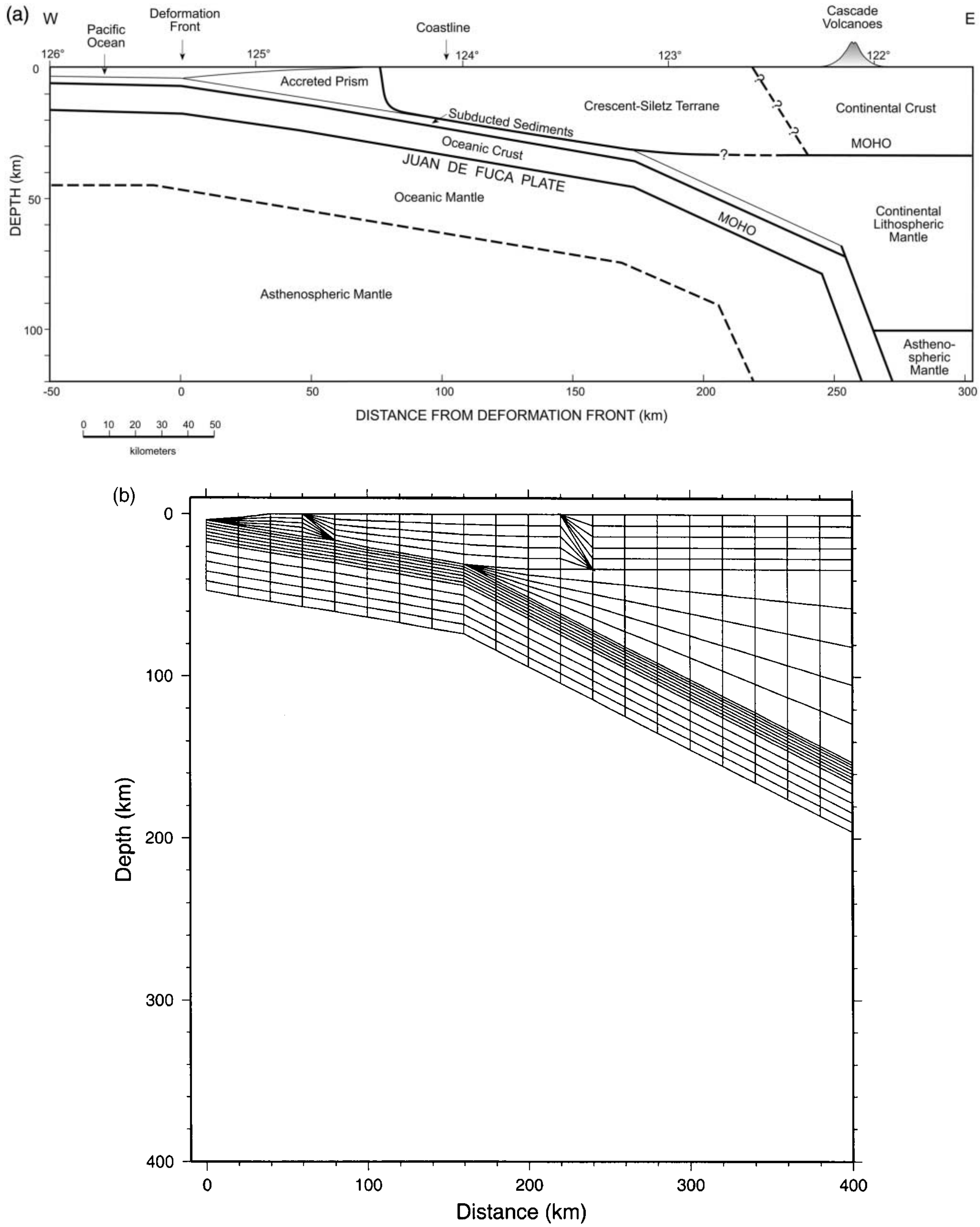


Figure 9. (a) Simplified structural model across central Oregon at latitude 44.5° N used in thermal modeling and (b) finite-element mesh.

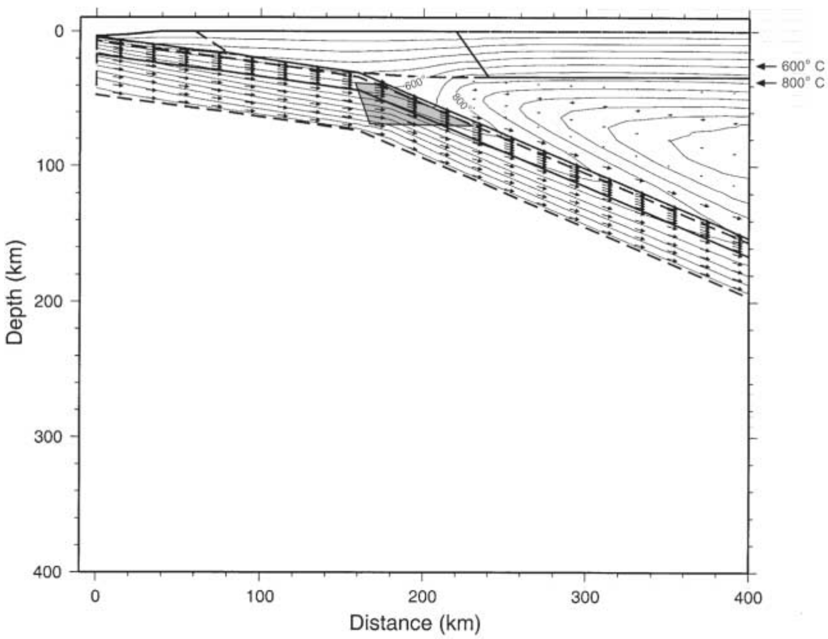


Figure 10. Thermal model of the central Cascadia subduction zone calculated in this study. Isotherms from the 2D model calculations are at 100°C intervals. The black- and red-dotted isotherms indicate the 600°C and 800°C isotherms. Arrows indicate the flow vectors. The results show that at depths of 40 to 70 km, the temperatures within the Juan de Fuca plate range from 600° to 1200°C.

Table 3
Thermal Model Parameters

| | |
|---|---|
| Age of Juan de Fuca Plate | 8 m.y. |
| Convergence Rate | 31 mm/yr |
| Incoming Sediment Thickness | 3200 m |
| Megathrust Temperature at Deformation Front | 240°C |

| Thermostratigraphic Unit | Thermal Conductivity W/m/K | Radioactive Heat Generation $\mu$W/m$^3$ | Heat Capacity MJ/m$^3$/K |
|---|---|---|---|
| Accreted Prism | 1.5–2.5* | 0.6 | — |
| Crescent-Siletz | 1.59 | 0.05 | — |
| Continental Crust | 2.5–3.0 | 0.2–0.6 | — |
| Subducted Sediments | 1.2 | 0.6 | — |
| Oceanic Crust | 2.9 | 0.01 | 3.3 |
| Oceanic Mantle | 2.9 | 0.0 | 3.3 |
| Continental Mantle Wedge | 2.9 | 0.0 | — |

Model parameters adopted from Hyndman and Wang (1995) and Oleskevich *et al.* (1999).

*Increases linearly with distance landward and with depth from toe of the prism to the backstop.

potheses to explain the low seismicity rates in both the crust and Wadati–Benioff zone beneath western Oregon: (1) the megathrust is unlocked and the continental crust and slab are in a low stress state or (2) the megathrust and overlying crustal faults are locked and in a stage of the seismic cycle where few events are produced. Given the now accepted fact that the CSZ megathrust last ruptured in the 1700 **M** 9 earthquake (Satake *et al.*, 1996), it is likely that the megathrust is locked and accumulating strain to be released in a future event.

Several investigators (e.g., Frohlich, 1989) attribute slab-pull force as a source of stress in intraslab seismicity, and thus a reduction in such forces may result in a reduction in intraslab seismicity. Spence (1989) stated that the increased buoyancy of the young Juan de Fuca plate under Oregon caused by slowed subduction will decrease the slab-pull forces of the subducted plate. Rogers *et al.* (1996) and Wells *et al.* (2002) hypothesize that intraslab seismicity may be absent beneath Oregon because of reduced slab-pull forces caused by a lack of a deep slab. Peacock *et al.* (2002) suggested three reasons for the absence of intraslab seismicity beneath western Oregon: (1) the subduction zone is anhydrous such that dehydration embrittlement cannot occur; (2) water is released by metamorphic dehydration, but the hydrogeology of the slab prevents fluid pressures from attaining levels required for faulting; or (3) the slab may be unusually warm because of interaction with the Newberry hot spot resulting in conditions too warm for seismogenesis.

## Discussion

At this time, it is difficult to assess the validity of other hypotheses such as interplate coupling, reduced slab-pull, or hydrogeologic conditions without further investigations outside the scope of this study. The point of this study is that the thermal state of the slab is the most important control on intraslab seismicity. Overall, it appears that most of the CSZ and Nankai, and the northern Mexico and southern Chile subduction zones are generally too warm for the occurrence of large shallow intraslab earthquakes. For all four subduction zones, their youthfulness suggests that the subducting plate is warmer than average as it enters the trench. Because of slower convergence (<40 mm/yr; Fig. 8), these slabs also have more time to be heated up by the surrounding lithosphere as they subduct. The Siletz terrane overlying the subducting plate in the central CSZ also acts as an insulator in keeping the slab hot. The thermal modeling of the central

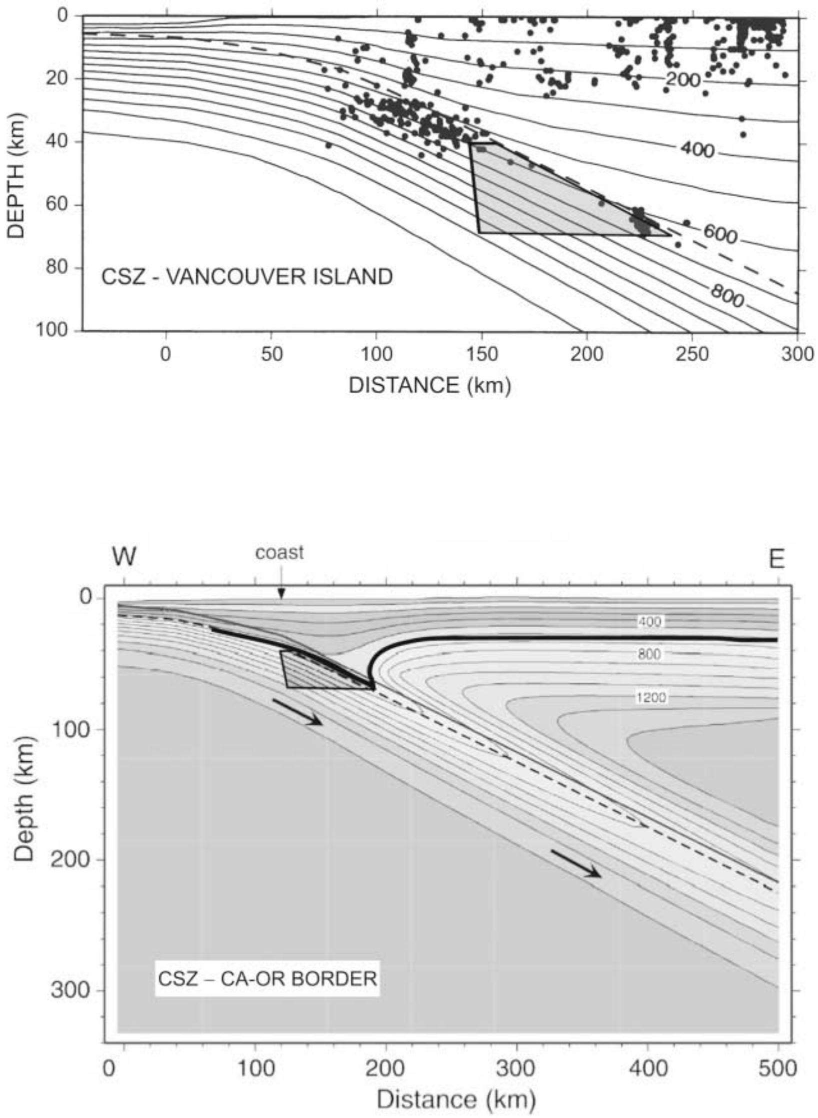


Figure 11. Thermal models of the Cascadia subduction zone near the southern end of Vancouver Island from Hyndman and Wang (1995) and at the Oregon–California border (Peacock *et al.*, 2002). Seismicity is also shown for the former. The tops of the slabs are indicated by the dashed lines. The depth range of 40 to 70 km of large intraslab earthquakes observed in the Puget Sound region is shown by the shaded area.

CSZ indicates that the downgoing slab beneath western Oregon is hot, which likely precludes intraslab seismicity, at least at depths greater than 40 km. At depths shallower than 40 km, temperatures still appear to be warm enough, above a possible cutoff temperature (>500°C), that intraslab seismicity could be almost completely absent as observed. There probably is, however, one or more other factors such as the level of tectonic stress, though not as significant as temperature, that influence the occurrence of intraslab seismicity.

Still the question remains why such events occur in the Puget Sound region. It appears that large-magnitude intraslab earthquakes there may be associated with subducting plate conditions or features that are not observed elsewhere along the central or southern CSZ. Temperatures appear to be cooler in the northern CSZ than other portions of Cascadia. This observation is supported by the fact that seismicity beneath the Puget Sound occurs to depths of 60 km. Hence, only in this portion of the northern CSZ is there a Wadati–Benioff zone at depths greater than 40 km, where large intraslab events are thought to be generated. Stresses in the Puget Sound might also be higher possibly because of the arch in the Juan de Fuca plate. The presence of the arch in the Puget Sound, in contrast to the relatively smoothly bending surface of the Juan de Fuca plate farther to the south, may act as a locus of increased tectonic stresses. As observed globally by Astiz *et al.* (1988), distortions in downgoing slabs appear to localize large shallow earthquakes in the most active subduction zones, so the arch in the plate must be a region of enhanced levels of tectonic stresses, or the tear along the arch, as proposed by McCrory *et al.* (2003), acts as a zone of weakness where large intraslab earthquakes can occur.

Similarly, despite the youthfulness of the Nankai subduction zone, the Geiyo area in the southern region has had two large shallow intraslab earthquakes in historical time, possibly because of a distortion in the Philippine Sea plate due to the subduction of a fossil ridge. Oceanic intraslab seismicity occurs preferentially in active tectonic features such as fossil spreading ridges, suggesting that these zones are weaker than the surrounding lithosphere (Stein and Stein, 1996; Kirby *et al.*, 1996).

The presence of a shallow Wadati–Benioff zone in the Gorda block would be consistent with a high level of tectonic stresses and evidence of internal deformation (Smith *et al.*, 1993) due to the deformation occurring at the Mendocino triple junction to the south. Future stress modeling along the length of the CSZ will aid in the assessment of stresses as a factor in intraslab seismicity.

## Conclusions

Although not complete down to **M** 5.0, there have been no known earthquakes of this size or greater observed within the central CSZ in the past 150 years. Adopting a Gorda block origin for the 1873 earthquake, I believe the potential for large (**M** $\geq$6.5) intraslab events within the central Juan de Fuca plate is very low. Such events are believed to occur in the depth range of 40 to 70 km, coinciding with a bend and associated stresses in the Juan de Fuca plate. This is not to say that such events cannot ever occur in the central CSZ. Their average recurrence interval conceivably could be very large ($>$500–1000 years), although examination of other subduction zones indicates typical recurrence intervals of less than tens of years (Table 2). Earthquake recurrence in subduction zones follows the Gutenberg–Richter relationship and thus small- to moderate-magnitude seismicity should be observed beneath western Oregon if larger events are possible. The apparent absence of a Wadati–Benioff zone based on decades of seismic monitoring and the historical absence of even moderate-magnitude intraslab earthquakes within the Juan de Fuca plate outside of the Puget Sound region, and particularly beneath western Oregon, is a compelling argument that large events are unlikely to occur unless the 150-year historical record is anomalous. However, based on a comparison with other subduction zones, I see no valid reason to doubt that the historical record is representative of the intraslab seismogenic potential of the Juan de Fuca plate.

Comparison of the CSZ with other young, warm subduction zones suggests that the absence of large shallow intraslab earthquakes is not unusual. The variability in the maximum depth of the Wadati–Benioff zone along the CSZ, 60 km in the northern CSZ beneath Puget Sound, 40 km within the southern CSZ (Gorda block), and essentially zero in the central CSZ, reflect differing temperature conditions and hence potential for large shallow intraslab earthquakes. The Juan de Fuca plate, at least beneath western Oregon, is probably too hot ($>$500°C) to sustain the stresses necessary for seismogenesis throughout its depth range ($>$10 km). The Gorda block has a Wadati–Benioff zone to a depth of 40 km, where a temperature of about 600°C is reached. The maximum depth of the Wadati–Benioff zone suggests that the northern CSZ is the coolest section of the subduction zone. Temperature estimates along the megathrust (Hyndman and Wang, 1995) are consistent with this observation. Relatively fast convergence rates, the older age of the plate beneath western Washington, and the lack of an insulating crust may be factors in cooling the plate in the northern CSZ.

In addition to temperature, the level of tectonic stresses must play an essential role in whether large intraslab earthquakes can occur. The Puget Sound region appears to be unusual in its capability to generate large shallow intraslab earthquakes, because the slab is not only cooler than the rest of the CSZ but also because stresses may be concentrated there due to the arch in the subducting Juan de Fuca plate or because of the existence of a pre-existing zone of weakness, as proposed by McCrory *et al.* (2003).

Despite the absence of a Wadati–Benioff zone beneath western Oregon, seismic hazard analyses (e.g., Geomatrix Consultants, 1995; Wong *et al.*, 2000; Petersen *et al.*, 2002) have traditionally included the potential for large intraslab earthquakes in their evaluations. If indeed this potential has been overestimated, then the implications for seismic hazard assessment in the Pacific Northwest are significant.

## Acknowledgments

My thanks to Larry Ruff, Bob Crosson, Chuck Langston, Harvey Kelsey, Dapeng Zhao, Ken Creager, Ray Wells, and Bill Bakun for generously sharing their time and knowledge. Special thanks to Rob Harris for assisting in the thermal modeling and to Kelin Wang for the use of his thermal modeling computer program. Thanks also to Melinda Lee, Mark Dober, Fabia Terra, and Fumiko Goss for their assistance in the preparation of this article. The article benefited greatly from thoughtful reviews by Walter Arabasz, Jim Pechmann, Kris Pankow, two anonymous reviewers, and Associate Editor Fred Pollitz.

## References

Algermissen, S. T., and S. T. Harding (1965). Preliminary seismological report, the Puget Sound, Washington, earthquake of April 28, 1965, U.S. Coast and Geodetic Survey.

Astiz, L., T. Lay, and H. Kanamori (1988). Large intermediate-depth earthquakes and the subduction process, *Phys. Earth Planet. Interiors* **53,** 80–166.

Atwater, B. F., A. R. Nelson, J. J. Clague, G. A. Carver, D. K. Yamaguchi, P. T. Bobrowsky, J. Bourgeois, M. E. Darienzo, W. C. Grant, E. Hemphill-Haley, H. M. Kelsey, G. C. Jacoby, S. P. Nishenko, S. P. Palmer, C. D. Peterson, and M. A. Reinhart (1995). Summary of coastal geologic evidence for past great earthquakes at the Cascadia subduction zone, *Earthquake Spectra* **11,** 1–18.

Baker, G. E., and C. A. Langston (1987). Source parameters of the 1949 magnitude 7.1 south Puget Sound, Washington, earthquake as determined from long-period body waves and strong ground motions, *Bull. Seism. Soc. Am.* **77,** 1530–1557.

Bakun, W. H. (2000). Seismicity of California's north coast, *Bull. Seism. Soc. Am.* **90,** 797–812.

Bolt, B. A., C. Lomnitz, and T. V. McEvilly (1968). Seismological evidence on the tectonics of central and northern California and the Mendocino escarpment, *Bull. Seism. Soc. Am.* **58,** 1725–1767.

Cande, S. C., and R. B. Leslie (1986). Late Cenozoic tectonics of the Southern Chile trench, *J. Geophys. Res.* **91,** 471–496.

Chen, W., and P. Molnar (1983). Focal depths of intracontinental and intraslab earthquakes and its implications for the thermal and mechanical properties of the lithosphere, *J. Geophys. Res.* **88,** 4183–4214.

Crosson, R. S., and T. J. Owens (1987). Slab geometry of the Cascadia subduction zone beneath Washington from earthquake hypocenters and teleseismic converted waves, *Geophys. Res. Lett.* **14,** 824–827.

Cummins, P. R., Y. Kaneda, and T. Hori (2002). Large intermediate-depth earthquakes in southwest Japan, in *The Cascadia Subduction Zone and Related Subduction Systems—Seismic Structure, Intraslab Earthquakes and Processes, and Earthquake Hazards*, S. Kirby, K. Wang, and S. Dunlop (Editors), *U.S. Geol. Surv. Open-File Rept 02-328* and *Geological Survey of Canada Open-File Rept 4350*, 99–101.

Flück, P., R. D. Hyndman, and K. Wang (1997). Three-dimensional dislocation model for great earthquakes of the Cascadia subduction zone, *J. Geophys. Res.* **102,** 20,539–20,550.

Frohlich, C. (1989). The nature of deep focus earthquakes, *Annu. Rev. Earth Planet. Sci.* **17,** 227–254.

Geomatrix Consultants (1995). Seismic design mapping for the State of Oregon, unpublished final report prepared for the Oregon Department of Transportation, variously paginated.

Gutenberg, B., and C. F. Richter (1954). *Seismicity of the Earth and Associated Phenomena*, Princeton University Press, Princeton, New Jersey, 310 pp.

Heaton, T. H., and H. Kanamori (1984). Seismic potential associated with subduction in the northwestern United States, *Bull. Seism. Soc. Am.* **78,** 933–941.

Hyndman, R. D., and K. Wang (1993). Thermal constraints on the zone of major thrust earthquake failure: the Cascadia subduction zone, *J. Geophys. Res.* **98,** 2039–2060.

Hyndman, R. D., and K. Wang (1995). The rupture zone of Cascadia great earthquakes from current deformation and the thermal regime, *J. Geophys. Res.* **100,** 22,133–22,154.

Hyndman, R. D., K. Wang, and M. Yamano (1995). Thermal constraints on the seismogenic portion of the southwestern Japan subduction thrust, *J. Geophys. Res.* **100,** 15,373–15,392.

Isacks, B., J. Oliver, and L. R. Sykes (1968). Seismology and the new global tectonics, *J. Geophys. Res.* **73,** 5855–5899.

Jiao, W., P. G. Silver, Y. Fei, and C. T. Prewitt (2000). Do intermediate- and deep-focus earthquakes occur on pre-existing weak zones? An examination of the Tonga subduction zone, *J. Geophys. Res.* **105,** 28,125–28,138.

Kelsey, H. M., and J. G. Bockheim (1994). Coastal landscape evolution as a function of eustasy and surface uplift rate, Cascadia margin, southern Oregon, *Geol. Soc. Am. Bull.* **106,** 840–854.

Kirby, S., E. R. Engdahl, and R. Denlinger (1996). Intermediate-depth intraslab earthquakes and arc volcanism as physical expressions of crustal and uppermost mantle metamorphism in subducting slabs, in *Subduction: Top to Bottom, American Geophysical Monograph 96*, G. E. Bebout, D. Scholl, S. Kirby, and J. P. Platt (Editors), 195–214.

Kirby, S., K. Wang, and S. Dunlop (Editors) (2002). *The Cascadia Subduction Zone and Related Subduction Systems—Seismic Structure, Intraslab Earthquakes and Processes, and Earthquake Hazards*, *U.S. Geol. Surv. Open-File Rept. 02-328* and *Geological Survey of Canada Open-File Rept. 4350*, 169 pp.

Ludwin, R. S., C. S. Weaver, and R. S. Crosson (1991). Seismicity of Washington and Oregon, in *Neotectonics and North America*, D. B. Slemmons, E. R. Engdahl, M. D. Zoback, and D. D. Blackwell (Editors), Geological Society of America Decade Map 1, 77–98.

Ma, L., R. Crosson, and R. Ludwin (1996). Western Washington earthquake focal mechanisms and their relationship to regional tectonic stress in the Pacific Northwest, in *Assessing Earthquake Hazards and Reducing Risk in the Pacific Northwest*, A. M. Rogers, T. J. Walsh, W. J. Kockelman, and G. R. Priest (Editors), *U.S. Geol. Surv. Profess. Pap. 1560*, Vol. 1, 257–283.

McCrory, P. A., F. F. Pollitz, and J. L. Blair (2003). Evidence for a tear in the Juan de Fuca plate in the vicinity of the 2001 Nisqually earthquake (abstract), *EOS* **84,** F1109.

Minster, J. B., and T. H. Jordan (1978). Present day plate motions, *J. Geophys. Res.* **83,** 5331–5354.

Molnar, P., D. Freedman, and J. S. F. Shih (1979). Lengths of intermediate and deep seismic zones and temperatures in downgoing slabs of lithosphere, *Geophys. J. R. Astr. Soc.* **56,** 41–54.

Oleskevich, D. A., R. D. Hyndman, and K. Wang (1999). The updip and downdip limits to great subduction earthquakes: thermal and structural models of Cascadia, south Alaska, SW Japan, and Chile, *J. Geophys. Res.* **104,** 14,965–14,991.

Oppenheimer, D., G. Beroza, G. Carver, L. Dengler, J. Eaton, L. Gee, F. Gonzalez, A. Jayko, W. H. Li, M. Lisowski, M. Magee, G. Marshall, M. Murray, R. McPherson, B. Romanowicz, K. Satake, R. Simpson, P. Somerville, R. Stein, and D. Valentine (1993). The Cape Mendocino, California, earthquake sequence of April 1992: subduction at the triple junction, *Science* **261,** 433–438.

Pacific Northwest Seismograph Network (PNSN) Staff (2001). Preliminary report on the $M_w = 6.8$ Nisqually, Washington, earthquake of 28 February 2001, *Seism. Res. Lett.* **72,** 352–361.

Peacock, S. M. (1996). Thermal and petrologic structure of subduction zones, in *Subduction: Top to Bottom*, G. E. Bebout, D. Scholl, S. Kirby, and J. P. Platt (Editors), American Geophysical Monograph 96, 119–133.

Peacock, S. M., K. Wang, and A. M. McMahon (2002). Thermal structure and metamorphism of subducting oceanic crust: insight into Cascadia intraslab earthquakes, in *The Cascadia Subduction Zone and Related Subduction Systems—Seismic Structure, Intraslab Earthquakes and Processes, and Earthquake Hazards*, S. Kirby, K. Wang, and S. Dunlop (Editors), *U.S. Geol. Surv. Open-File Rept. 02-328* and *Geol. Surv. of Canada Open-File Rept. 4350*, 123–126.

Petersen, M. D., C. H. Cramer, and A. D. Frankel (2002). Simulations of seismic hazard for the Pacific Northwest of the United States from earthquakes associated with the Cascadia subduction zone, *Pageoph* **159,** 2147–2168.

Pezzopane, S. K. (1993). Active faults and earthquake ground motions in Oregon, *Ph.D. Thesis*, University of Oregon, Eugene, 209 pp.

Preston, L. A., K. C. Creager, R. S. Crosson, T. M. Brocher, and A. M. Trehu (2003). Intraslab earthquakes: dehydration of the Cascadia slab, *Science* **302,** 1197–1200.

Riddihough, R. (1984). Recent movements of the Juan de Fuca plate system, *J. Geophys. Res.* **89,** 6980–6994.

Rogers, A. M., T. J. Walsh, W. J. Kockelman, and G. R. Priest (1996). Earthquake hazards in the Pacific Northwest—an overview, in *Assessing Earthquake Hazards and Reducing Risk in the Pacific Northwest, U.S.*, A. M. Rogers, T. J. Walsh, W. J. Kockelman, and G. R. Priest (Editors), *Geol. Surv. Profess. Pap. 1560*, Vol. 1, 1–54.

Rogers, G. C. (1988). An assessment of the megathrust earthquake potential of the Cascadia subduction zone, *Can. J. Earth Sci.* **25,** 844–852.

Rondenay, S., M. G. Bostock, and J. Shragge (2001). Multiparameter two-dimensional inversion of scattered teleseismic body waves 3. Application to the Cascadia 1993 data set, *J. Geophys. Res.* **106,** 30,795–30,807.

Ruff, L., and H. Kanamori (1980). Seismicity and the subduction process, *Phys. Earth Planet.* Interiors, **23,** 240–252.

Ruff, L., and H. Kanamori (1983). Seismic coupling and uncoupling at subduction zones, *Tectonophysics* **99,** 99–117.

Satake, K., K. Shimazaki, Y. Tsuji, and K. Ueda (1996). Time and size of a giant earthquake in Cascadia inferred from Japanese tsunami records of January 1700, *Nature* **379,** 246–249.

Singh, S. K., L. Astiz, and J. Havskov (1981). Seismic gaps and recurrence periods of large earthquakes along the Mexican subduction zone: a reexamination, *Bull. Seism. Soc. Am.* **71,** 827–843.

Smith, S. W., J. S. Knapp, and R. C. McPherson (1993). Seismicity of the Gorda plate, structure of the continental margin, and an eastward jump of the Mendocino triple junction, *J. Geophys. Res.* **98,** 8153–8171.

Spence, W. (1989). Stress origins and earthquake potentials in Cascadia, *J. Geophys. Res.* **94,** 3076–3088.

Stein, S., and C. A. Stein (1996). Thermomechanical evolution of oceanic lithosphere: implications for the subduction process and deep earthquakes in *Subduction Top to Bottom*, G. E. Bebout, D. Scholl, S. Kirby, and J. P. Platt (Editors), Geophysical Monograph 96, 1–17.

Stover, C. W., and C. A. von Hake (Editors) (1982). *United States Earthquakes, 1980*, U.S. Geological Survey and National Oceanic and Atmospheric Administration, 132 pp.

Tibi, R., G. Bock, and C. H. Estabrook (2002). Seismic body wave constraint on mechanisms of intermediate-depth earthquakes, *J. Geophys. Res.* **107,** ESE 1-1–1-23.

Toppozada, T. R., and D. L. Parke (1982). Areas damaged by California earthquakes 1900–1949, California Division of Mines and Geology Open-File Report 93-02, 45 pp.

Toppozada, T. R., C. R. Real, and D. L. Parke (1981). Preparation of isoseismal maps and summaries of reported effects for pre-1900 California earthquakes, California Division of Mines and Geology Open-File Report 81-11, 181 pp.

Townley, S. D., and M. W. Allen (1939). Descriptive catalog of earthquakes of the Pacific Coast of the United States: 1769 to 1928, *Bull. Seism. Soc. Am.* **29,** 1–297.

Trehu, A. M., I. Asudeh, T. M. Brocher, J. H. Luetgert, W. D. Mooney, J. L. Nabelek, and Y. Nakamura (1994). Crustal architecture of the Cascadia forearc, *Science* **265,** 237–243.

Velasco, A. V., C. J. Ammon, and T. Lay (1994). Recent large earthquakes near Cape Mendocino and in the Gorda plate: broadband source time functions, fault orientations, and rupture complexities, *J. Geophys. Res.* **99,** 711–728.

Vlaar, N. J., and M. J. R. Wortel (1976). Lithospheric aging, instability, and subduction, *Tectonophysics* **32,** 331–351.

Walter, S. R. (1986). Intermediate-focus earthquakes associated with Gorda plate subduction in northern California, *Bull. Seism. Soc. Am.* **76,** 583–588.

Wang, K., T. Mulder, G. C. Rogers, and R. D. Hyndman (1995). Case for very low coupling stress on the Cascadia subduction fault, *J. Geophys. Res.* **100,** 12,907–12,918.

Wang, K., I. Wada, and Y. Ishikawa (2004). Stresses in the subducting slab beneath southwestern Japan and relation with plate geometry, tectonic forces, slab dehydration, and damping earthquakes, *J. Geophys. Res.* **109,** B08304.

Weaver, C. S., and G. E. Baker (1988). Geometry of the Juan de Fuca plate beneath Washington and northern Oregon from seismicity, *Bull. Seism. Soc. Am.* **78,** 264–275.

Weaver, C. S., and K. M. Shedlock (1996). Estimates of seismic source regions from earthquake distribution and regional tectonics in the Pacific Northwest, in *Assessing Earthquake Hazards and Reducing Risk in the Pacific Northwest*, A. M. Rogers, T. J. Walsh, W. J. Kockelman, and G. R. Priest (Editors), *U.S. Geol. Surv. Profess. Pap. 1560*, Vol. 1, 285–306.

Wells, R. E., R. J. Blakely, and C. S. Weaver (2002). Cascadia microplate models and within-slab earthquakes, in *The Cascadia Subduction Zone and Related Subduction Systems—Seismic Structure, Intraslab Earthquakes and Processes, and Earthquake Hazards*, S. Kirby, K. Wang, and S. Dunlop (Editors), *U.S. Geol. Surv. Open-File Rept. 02-328* and *Geol. Surv. Canada Open-File Rept. 4350*, 17–23.

Wilson, D. S. (1989). Deformation of the so-called Gorda plate, *J. Geophys. Res.* **94,** 3065–3075.

Wilson, D. S. (2002). The Juan de Fuca plate and slab: Isochron structure and Cenozoic plate motions, in *The Cascadia Subduction Zone and Related Subduction Systems—Seismic Structure, Intraslab Earthquakes and Processes, and Earthquake Hazards*, S. Kirby, K. Wang, and S. Dunlop (Editors), *U.S. Geol. Surv. Open-File Rept. 02-328* and *Geol. Surv. Canada Open-File Rept. 4350*, 9–12.

Wong, I. G. (1997). The historical earthquake record in the Pacific Northwest: applications and implications to seismic hazard assessment, in *Earthquakes—Converging at Cascadia, Symposium Proceedings*, Y. Wang and K. K. Neuendorf (Editors), Association of Engineering Geologists Special Publication 10 and Oregon Department of Geology and Mineral Industries Special Paper 28, 19–36.

Wong, I. G., and J. D. J. Bott (1995). A look back at Oregon's earthquake history, 1841–1994, *Oregon Geol.* **57,** 125–139.

Wong, I. G., and D. S. Chapman (1990). Deep intraslab earthquakes in the western U.S. and their relationship to lithospheric temperatures, *Bull. Seism. Soc. Am.* **80,** 589–599.

Wong, I. G., and R. N. Harris (2003). Thermal control of shallow intraslab seismicity: implications to the central and southern Cascadia subduction zone (abstract), *Seism. Res. Lett.* **74,** 223.

Wong, I., W. Silva, J. Bott, D. Wright, P. Thomas, N. Gregor, S. Li, M. Mabey, A. Sojourner, and Y. Wang (2000). Earthquake scenario and probabilistic ground shaking maps for the Portland, Oregon, metropolitan area, Oregon Department of Geology and Mineral Industries Interpretive Map Series IMS-16, scale 1:62,500, 11 sheets with 16 p. text.

Zhao, D., N. Hirata, and K. Obara (2002). The 2001 Geiyo earthquake and the structure of the warm Philippine Sea slab under southeast Japan (abstract), *Seism. Res. Lett.* **73,** 219.

Department of Geology and Geophysics
University of Utah
135 South 1460 East
Salt Lake City, Utah 84112

Seismic Hazards Group, URS Corporation
1333 Broadway, Suite 800
Oakland, California 94612

Manuscript received 13 July 2004.