Geophysical Studies of the Northern Cascadia Subduction Zone off Western Canada and Their Implications for Great Earthquake Seismotectonics: A Review

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Abstract. The northern Cascadia subduction zone of western Canada lies between the underthrusting oceanic Juan de Fuca plate and the continental North American plate. Large damaging earthquakes have occurred within the overriding crust and within the downdropping plate. Although there have been no historical great subduction earthquakes, coastal marsh and deep sea turbidite paleoseismicity data indicate that such events do occur, the last 300 years ago. Marine and land-based seismic reflection and refraction studies, combined with other geophysical and geological data, delineate the regional structure and tectonic regime that control earthquake occurrence. The principal structural elements are: the underthrusting oceanic plate, the overlying accretionary sedimentary wedge, and two small terranes accreted along the margin. Underplated oceanic material is present in the lower crust beneath the forearc Wrangellia terrane and adjacent Coast Plutonic Complex. An important use of structural and tectonic data is the precise location and nature of the subduction thrust plane that are needed for modelling the strain buildup of great earthquakes and the temperatures that control earthquake occurrence. Geodetic data from the coast across Vancouver Island show that there is strain buildup toward future great earthquakes. Geodetic and thermal models indicate that the thrust is locked and great earthquakes can occur over a downdip zone 50 to 100 km wide. These parameters allow for magnitude 9 subduction thrust earthquakes.

INTRODUCTION

The Cascadia subduction zone is part of the modern plate tectonic regime of the continental margin off southwestern Canada and northwestern United States. The region is dominated by the interactions of three main lithospheric plates: the large Pacific and North American plates and the intervening Juan de Fuca plate system, which includes the Explorer, Juan de Fuca and Gorda (off northern California) plates (Fig. 1). The Cascadia subduction zone has experienced damaging earthquakes in the forearc crust and in the downdropping oceanic plate. The margin is unusual in that no large thrust earthquakes have occurred during the 200-year written historical period (e.g., Rogers, 1988). However, paleoseismicity studies involving: (1) buried coastal marshes and dead trees in bays from salt water
Fig. 1. Schematic illustration of the northern Cascadia subduction zone. The tectonic regimes for major historical earthquakes and the location of the Lithoprobe corridor (cross-section in Fig. 7) are indicated. The inset map shows the location of the map coverage with respect to northern North America.

Ingress following abrupt subsidence of the land, (2) coastal deposits of tsunami-generated sand layers, and (3) deep-sea turbidite sediment layers inferred to be the result of strong shaking of the continental margin, have established that major thrust earthquakes have occurred in the region with recurrence intervals of 300 to 1000 years (Clague, 1997 and references therein). Based on west coast First Nations oral history, tree ring dating, and the record of the resulting tsunami on the east coast of Japan, the most recent Cascadia subduction thrust earthquake occurred on 26th January, 1700 (references in Clague, 1997).

To learn more about the structure and tectonic history of the northern Cascadia subduction zone, an extensive geological and geophysical data set has been generated off southwestern Canada during the past two decades, making the area one of the most comprehensively studied subduction zones in the world. Regional coverage is provided by gravity and magnetic data and geological mapping, while local measurements include heat flow, borehole data, magnetotellurics, and seismicity. Seismic refraction, shallow seismic profiling, and deep multichannel seismic reflection lines provide profile coverage across and along the study area (Fig. 2). Hyndman et al. (1990) summarize some of these data and their interpretations, with an emphasis on the seismic reflection data. Hyndman (1995a) provides an overview of studies of the northern Cascadia subduction zone. Clowes et al. (1997) present some additional deep reflection data and develop density/magnetic susceptibility structural models for the offshore margin, most of Vancouver Island and the adjacent western mainland.
A critical constraint on great earthquake shaking and thus hazard at coastal cities is which part of the subduction thrust may rupture in great earthquakes. The structural and tectonic studies have been used in analyses and modelling of the pattern of elastic strain accumulation and modelling of the thermal regime that constrain the zone of likely major thrust earthquake failure. Vertical and horizontal strain rates have been determined through long-term trends in tide gauge data, changes in repeated accurate leveling surveys, changes in repeated high-accuracy gravity profiles and horizontal shortening observed in GPS and other precise positioning surveys (Savage et al., 1991; Dragert et al., 1994; Henton, 2000). The locations and widths of the main seismogenic zone and the transition zone into which the rupture can propagate have been estimated from dislocation modelling of the pattern and rates of current deformation in the region (e.g., Hyndman and Wang, 1995; Flück et al., 1997). The thermal regime also can constrain the seismogenic zone, assuming that the downdip limit of seismic behaviour on the fault is thermally controlled (e.g., Hyndman and Wang, 1993).

In this paper, we summarize the tectonic setting and regional geology of the northern Cascadia subduction zone, illustrate some of the geophysical data, show
an interpretation of the structure of the zone, review the current deformation data
and thermal regime, and present the current estimate of the Cascadia subduction
thrust seismogenic and transition zones. We also provide some comparisons with
the southwest Japan subduction zone.

TECTONIC SETTING AND REGIONAL GEOLOGY

Plate interaction along the western margin of North America has been
dominated by convergence for at least the past 150 million years (Riddihough,
1982; Engebretson et al., 1992). Alternating episodes of northeast-directed
convergence and northwest-directed transform motion or oblique convergence
allowed exotic material to be brought in, accreted to the margin, and then sheared
northward (e.g., Riddihough, 1982). The last major collisional episode, around
mid-Cretaceous time, emplaced the Insular superterrane against the existing
continental boundary represented by the Intermontane superterrane, and generated
the mid-Cretaceous to early Tertiary intrusive rocks of the Coast Belt over the
region of the suture; details of this tectonic evolution vary (e.g., Coney et al.,
1980; Monger et al., 1982; van der Heyden, 1992; Monger et al., 1994). During
the latest Cretaceous or early Tertiary, the small Pacific Rim terrane was
emplaced south and west of the Insular superterrane (Johnson, 1984; Brandon,
1989). Subsequently during middle to late Eocene time, the Crescent terrane,
which formed as new oceanic crust in a marginal basin setting, accreted against
the Pacific Rim terrane (Massey, 1986). These terranes formed the backstop for
the modern accretionary wedge. Figure 2 shows the current configuration of the
tectonic units.

The Juan de Fuca plate is presently converging with the North American
margin at a relative rate of about 40 mm/a directed N56°E (e.g., Riddihough and
Hyndman, 1991; Fig. 1). To the north the Explorer plate, a young fragment of the
Juan de Fuca plate that has been moving independently for at least 4 Ma
(Riddihough, 1984), is converging at variable rates along the margin; the
estimated convergence near Brooks Peninsula (Fig. 2) is 21 mm/a at N50°E by
Riddihough and Hyndman (1991). The Nootka fault zone, which separates the
two oceanic plates, is delineated by a northeasterly-trending broad band of
epicentres (Hyndman et al., 1979; Wahlström and Rogers, 1992) and has an
estimated left-lateral motion of approximately 25 mm/a directed N50°E
(Riddihough and Hyndman, 1991). The triple junction formed by the Pacific,
North America and Juan de Fuca (Explorer) plates is presently located near 51°N
and 131°W (Fig. 1). The junction migrated northward to this position about 1 Ma
ago from its previous position near Brooks Peninsula, where it was probably
stable for at least 5 Ma (Riddihough, 1977). Based on new SeaBeam bathymetry
and other data, Rohr and Furlong (1995) proposed that for the most recent
tectonics, the Explorer subduction zone is no longer active. The system is inferred
to be evolving to a single triple junction at the northern end of Juan de Fuca ridge
where it meets the Nootka fault zone. In this model, the western third of Explorer
plate is becoming attached to the Pacific plate while the eastern two thirds
becomes part of North America.
Onshore, Vancouver Island is dominated by the Wrangellia terrane, which is part of the Insular superrterane (Jones et al., 1977; Fig. 2), an accreted package of Devonian through Lower Jurassic volcanic and plutonic rocks (Muller, 1977). Along the west coast and southern part of the island, outcrops of the small Pacific Rim terrane are found. This terrane is a metamorphosed sediment-rich melange unit in contact with Wrangellian rocks along the San Juan-Survey Mountain fault system on southern Vancouver Island (Leech River complex of Fairchild and Cowan, 1982). The Crescent terrane, consisting primarily of tholeiitic basalts, outcrops on the southern tip of Vancouver Island and northwestern Washington state, forming the footwall of the Leech River fault and hanging wall of the Hurricane Ridge fault (Muller, 1977; Tabor and Cady, 1978; Massey, 1986). The Coast Belt lies to the east of Wrangellia and straddles the mid-Cretaceous suture zone between the Insular and Intermontane superrteraneas. The surface contact between the Coast Belt and the Wrangellia terrane occurs near the eastern edges of the Strait of Georgia and Johnstone Strait (Fig. 2), although isolated outcrops of Wrangellian rocks are found in the western Coast Belt (Journeay and Friedman, 1993). On the basis of interpretations of seismic refraction and gravity data, Zelt et al. (1996) suggest that, east of southernmost Vancouver Island, Wrangellia extends in the subsurface to the eastern Coast Belt.

SEISMIC REFLECTION STUDIES

Due to the subsurface resolution provided by the multichannel seismic reflection (MCS) technique, this method forms the “spearhead” of geophysical procedures applied within the Lithoprobe program. In 1984, a series of profiles were recorded on southern Vancouver Island as part of Lithoprobe Phase I (Fig. 2; Yorath et al., 1985a, b; Green et al., 1986; Clowes et al., 1987a). This was followed in 1985 by an offshore continuation of the land lines through a series of marine profiles provided for Lithoprobe through the Geological Survey of Canada’s (GSC) Frontier Geoscience Program (Yorath et al., 1987; Clowes et al., 1987b; Davis and Hyndman, 1989; Hasselgren and Clowes, 1995; Clowes et al., 1997). As part of Lithoprobe’s Southern Cordillera transect (Cook, 1995), a series of vibroseis MCS profiles was acquired in 1988 across the southern Cordillera of mainland British Columbia, including a number of lines in the Coast Belt (Fig. 2; Cook et al., 1991; Varsek et al., 1993; Hammer and Clowes, 1996). In 1989, a second series of marine MCS profiles was recorded primarily over the continental shelf and slope as part of site surveys for the Ocean Drilling Program (Spence et al., 1991a, b; Hyndman et al., 1994). Digital data for the MCS surveys are archived and available from the Lithoprobe Seismic Processing Facility at the University of Calgary (internet address: http://www.lith.ualberta.ca).

The main features of the seismic reflection sections over the Cascadia subduction zone, illustrated by Fig. 3 but determined from many profiles, are: (i) the top of Juan de Fuca plate oceanic crust is imaged continuously below the deep sea sediments and intermittently below the continental slope, shelf and Vancouver Island; (ii) the frontal thrusts, deformation front and large accretionary wedge which form as the thick sedimentary section on the incoming oceanic crust
Fig. 3. Coherency-filtered seismic reflection data with interpretation for (a) offshore line 85-01 (migrated) and (b) onshore line 84-01 (unmigrated); line locations shown in Fig. 2. The main components of the Cascadia subduction zone are illustrated by the two sections. Abbreviations are: AW, accretionary wedge sediments; C, a prominent band of reflectors in the mid-crust, perhaps involving a deformation zone of pre-Tertiary imbricated sediments; CR, Crescent terrane; DF, deformation front; E, a prominent band of reflectors in the lower crust, possibly representing a deformation zone of Tertiary imbricated sediments and/or a zone including free fluids; FT, frontal thrust; HS, hemipelagic sediments; OC, oceanic crust; OM, oceanic mantle; PR, Pacific Rim terrane; TB, Tofino basin; TF, Tofino fault; TS, turbidite sediments; WCF, West Coast fault; WR, Wrangellia terrane. Dashed box outlines region of enlargement in Fig. 4(b).

is scraped off while the crust itself subducts toward the island (Fig. 4; also see Davis and Hyndman, 1989); (iii) the Eocene to recent marine clastic sediments of the Tofino basin which overlies the accretionary wedge and accreted terranes to the east; (iv) the Eocene volcanic Crescent terrane which outcrops on southern Vancouver Island and northwestern Washington state; (v) the Pacific Rim terrane, a narrow accreted section of regionally metamorphosed Mesozoic marine sediments and basalt, which with the Crescent terrane, forms the backstop to the accretionary wedge; (vi) the Wrangellian accreted terrane, a thick Paleozoic to Mesozoic sequence, dominated by volcanic and plutonic rocks, of which the lower crust and sub-crustal lithosphere have been removed; and (vii) two prominent bands, “C” and “E”, of strong reflectivity which have been interpreted as imbricated sediments and volcanics associated with pre-Tertiary (“C”) and Tertiary (“E”) underplating (Clowes et al., 1987a; Hyndman et al., 1990; Calvert and Clowes, 1990) or, for “E”, a non-structural interpretation that involves the trapping of fluids at a metamorphic boundary (Hyndman, 1988; Hyndman et al.,
Fig. 4. (a) SeaMarc plan-view acoustic image of the frontal thrusts on the seafloor near the base of the continental slope off Vancouver Island. The track of the ship is indicated. The lower half of the image, which shows the flat topography of the deep ocean turbidite sediments, has been removed. (b) Enlargement of migrated structural stack along Line 85-01 showing a cross-sectional image of the frontal thrusts in the same region as (a). Note that deformation of the layered oceanic sediments begins 5–10 km seaward of the first visible deformation of the sea floor and almost 20 km seaward of the deformation front (Fig. 3).

Further landward, seismic reflection sections show characteristics of the Coast Belt (Varsek et al., 1993; Hammer and Clowes, 1996) and accreted terranes east of it.

The reflection data provide the details of the third dimension, depth, for the subduction zone and regions to the east. They enable ties with surface geology and thus the extension of the geology to depth, characteristics that are required for tectonic interpretations.

**SEISMIC REFRACTION AND OTHER SEISMIC STUDIES**

An extensive network of seismic refraction profiles has been acquired in the region of the northern Cascadia subduction zone and the southern Cordillera on the mainland; Clowes et al. (1995) provide a summary. The 1980 Vancouver Island Seismic Project (VISP) included a NW-SE marine line in the deep ocean
(R. M. Clowes and W. R. H. White, unpublished interpretation), a line along
Vancouver Island (McMechan and Spence, 1983; Drew and Clowes, 1990) and
an offshore-onshore line from the deep ocean to the mainland of British Columbia
(Spence et al., 1985; Drew and Clowes, 1990). The marine component of the
latter provided information on the deep velocity structure below the continental
slope (Waldron et al., 1990). Airgun shots from the 1989 MCS lines were
recorded by seismographs deployed at a number of land stations to provide deep
velocity information below the shelf (Wang, 1997). In 1989, Lithoprobe carried
out a refraction experiment including profiles extending from east-central
Vancouver Island to the east (Zelt et al., 1993) and northeast (Spence and
McLean, 1998) plus regional 3-d coverage (Zelt et al., 1996).

The marine refraction line combined with MCS lines in the deep ocean
(Hasselgren and Clowes, 1995) clearly establish the structure and characteristics
of the oceanic crust and upper mantle of the Juan de Fuca plate (Fig. 5). The
subducting oceanic crust is defined with smoothly increasing landward dip below
the slope, shelf and Vancouver Island. Above the subducting crust offshore, a
thick layer of low velocity material is associated with the accretionary wedge.
Below Vancouver Island, a thick high velocity layer, bounded by thinner layers
of lower velocity at depths of about 18 and 30 km is defined (Fig. 5). The latter
correspond well with the "C" and "E" reflectors (Fig. 3); the former corresponds
with the intervening zone of low reflectivity (Fig. 3) and can extend no farther
landward than eastern Vancouver Island. Even with the long line along Vancouver
Island, no clear definition of a continental Moho can be interpreted; this is the
region where the subducting plate probably intersects the continental Moho.
Further landward, characteristics of Wrangellia and the Coast Belt are defined by
the refraction data.

Tomographic inversion and related studies of teleseismic earthquake phases
have provided deep velocity structure and defined the downgoing slab in the

![Fig. 5. Generalized velocity structural model from the VISP experiment (Fig. 2) for the Lithoprobe
corridor (Fig. 1). Heavy lines indicate the boundaries constrained by MCS reflection data (e.g.,
Fig. 3). Numbers assigned to each layer/block indicate the range of velocities (km/s) at the
shallowest and deepest points within that layer/block (adapted from Drew and Clowes (1990)).](image-url)
forearc region (Bostock and VanDecar, 1995; Crosson and Owens, 1987), these results being consistent with the slab position and depth derived from geochemical studies of rocks from the Cascadia-Garibaldi volcanic centers (Dickinson and Seeley, 1979). From receiver-function analyses, which involve modelling the wave forms of teleseismic arrivals, the low velocity “C” and “E” layers are well resolved beneath Vancouver Island and, based on the shear-wave velocities, are interpreted to have high fluid-filled porosity (Cassidy and Ellis, 1990; Cassidy, 1995). The same analyses also clearly define the subducting Juan de Fuca plate.

OTHER GEOPHYSICAL STUDIES

A comprehensive set of magnetic and gravity data has been acquired for the Cascadia subduction zone by the Geological Survey of Canada. Recent presentations of the data in map form and 2-d and 3-d interpretations of these data are given by Dehler and Clowes (1992) and Clowes et al. (1997). Trends in both potential field maps parallel the strike of Vancouver Island and the coastal mainland. Highs and lows on the magnetic map clearly define the plan positions of the Crescent and Pacific Rim terranes, respectively. Two-dimensional modeling across the subduction zone also distinguishes the Crescent terrane (Fig. 6) and, with the map data, shows that it does not extend northwest beyond the Nootka fault zone (Fig. 2).

The gravity field shows the parallel bands of low and high gravity characteristic of most subduction zones (e.g., Riddihough, 1979). The primary low is associated with the sediment-filled trough at the base of the slope; the main high is over Vancouver Island. A subsidiary low off the southwest coast of the island is due to a sediment-filled fossil trench associated with the boundary between the Crescent and Pacific Rim terranes. A strong low off the northwest coast of the island is due to a deep basin filled with Pleistocene sediments (Davis and Clowes, 1986). Two-dimensional gravity modeling confirms all of these characteristics; an example is illustrated in Figs. 6(a) and (b) with the interpretation of the density model shown in Fig. 6(c) (Clowes et al., 1997). An important requirement from all modeling studies is the presence of a high-density block beneath Vancouver Island and above the subducting plate (Riddihough, 1979; Spence et al., 1985; Dehler and Clowes, 1992; Clowes et al., 1997). The position of this high density block is similar to that of the thick high velocity layer interpreted from the refraction data, furthering interpretation of wide-scale underplating of Wrangellia. The modeling also shows that the shape of the boundary region between Wrangellia and the Coast Belt to the east varies along strike and may be controlled by properties of pre-existing plutonic rocks (Clowes et al., 1997).

Magnetotelluric data have been collected across Vancouver Island; their interpretation indicates a highly conductive layer deepening landward from about 20 km near the west coast to about 35 km beneath the east coast (Kurtz et al., 1986, 1990). This layer coincides in position with the strongly reflective layer “E” (Fig. 3), which also has a low seismic velocity. A variety of explanations for the
Fig. 6. (a) Comparison of observed magnetic anomaly data (×) and gravity anomaly data (+) with values calculated from the density/susceptibility model of (b). The magnetic modeling did not include undulations at the southwest end of the line, due to seafloor magnetic anomalies, or the large, high frequency variations at the northeastern end which are most likely due to local, near-surface features that do not relate to the regional structure. (b) Density/susceptibility model representative of the Lithoprobe corridor (Fig. 1). Densities are in kg/m³; magnetic susceptibilities (second number, usually following solidus) are in SI units. Geometry from the seismic reflection and refraction interpretations are used as constraints in the modeling. (c) Generalized interpretation of the major blocks in (b) (adapted from Clowes et al. (1997)).
properties of layer “E” have been made: imbricated sediments above a block of Tertiary oceanic crust trapped when the subduction zone jumped seaward (Clowes et al., 1987a); a shear zone localized at the brittle-ductile rheology boundary and/or associated with the earlier phase of subduction (Calvert and Clowes, 1990; Hyndman, 1995a); fluids driven off the subducting oceanic plate that are trapped at a metamorphic front (Hyndman, 1988). Note that the explanations are not exclusive of each other.

An extensive set of heat flow data has been acquired for the Cascadia subduction zone and shows large variations, typical of most subduction zones. Data sources include ocean bottom heat flow measurements in the deep basin and continental slope (Davis et al., 1990; Hyndman et al., 1994); values estimated from bottom simulating reflectors (BSRs) associated with the base of a gas hydrate layer along the slope (Davis et al., 1990; Hyndman et al., 1993); results from ODP borehole measurements on the slope (Westbrook et al., 1994); heat fluxes determined from petroleum exploration wells (Lewis et al., 1988, 1991) on the continental shelf; and numerous high quality borehole measurements on southern Vancouver Island (Lewis et al., 1988, 1991). Results show a steady landward decrease in heat flow values from the ocean basin to Vancouver Island. The average is about 120 mW/m² in Cascadia basin near the base of the continental slope, close to that expected for oceanic lithosphere of 6–7 Ma. On the lower slope, the average is about 90 mW/m²; it decreases landward to 50 mW/m² on the shelf and down to values of 30–35 mW/m² about 20 km seaward of the volcanic centers (Fig. 2). The landward decrease is a consequence of the heat sink provided by the cool, underthrusting oceanic crust. In the region of the volcanic arc, there is an abrupt increase to very high heat flow values (Lewis and Jessop, 1981; Lewis et al., 1991).

LITHOSPHERIC CROSS SECTION OF THE NORTHERN CASCADIA SUBDUCTION ZONE

Based on the seismic reflection, seismic refraction, other geophysical results and geological information, a lithospheric cross section from the deep ocean Cascadia basin across the subduction zone and the southern Cordillera to the fold-and-thrust belt of the Rocky Mountains has been compiled along the Lithoprobe Southern Cordillera Transect (Clowes et al., 1995, 1998, 1999). The western half of this section, across the Cascadia subduction zone to east of the forearc volcanic centers is shown in Fig. 7. All of the components of the zone discussed in the preceding sections are illustrated in this cross section.

SEISMICITY DATA

The region of the northern Cascadia subduction zone generates the greatest seismicity and seismic hazard in Canada (e.g., Rogers, 1994). Figure 8 shows a seismicity map of the larger well-located earthquakes. The concentration of earthquakes offshore between 49°–50° is associated primarily with the Nootka fault zone, the boundary between the Juan de Fuca and Explorer plates (Fig. 2).
Fig. 7. Interpreted lithospheric cross section from the Juan de Fuca plate to the east side of the Coast Belt. Short, thin lines between 30 and 55 km depth just east of Insular-Coast Belt boundary schematically represent high-amplitude, near-vertical-incidence reflections. In the mantle, open arrows show material flow. No vertical or horizontal exaggeration. AW, accreted wedge; BR, Bridge River terrane; CD, Cadwallader terrane; CT, Crescent terrane; DF, deformation front; FF, Fraser fault; gr, granites; GVB, Garibaldi volcanic belt; HA, Harrison terrane; Hi V,ρ, high velocity and density; JdF, Juan de Fuca plate; MR, mantle reflector; MT, Methow terrane; PRT, Pacific Rim terrane; SH, Shuksan terrane.

Earthquakes in the near-offshore and onshore represent two types. The majority are shallow earthquakes in the continental crust; deeper earthquakes occur in the subducting slab. The former are limited to depths of about 30 km by the maximum temperature for crustal earthquake failure of about 350°C (Hyndman and Wang, 1993). The slab earthquakes are limited to maximum depths of about 70 km because the young Juan de Fuca plate exceeds the maximum temperature for mantle earthquakes (~750°C) at greater depths (Riddihough and Hyndman, 1976).

During historical times, the northern Cascadia subduction zone has not experienced a great thrust earthquake, in contrast to most comparable margins around the world. However, recent studies (summarized in a later section) have demonstrated that subduction thrust earthquakes have occurred in the Cascadia subduction zone during the past few thousand years. The most recent one took place off the coast of southern Vancouver Island and northwestern Washington state on 26 January 1700 (e.g., Satake et al., 1996). This earthquake is indicated in Fig. 8 by the ellipse.
Fig. 8. Seismicity map of southwestern Canada and northwestern Washington state. Earthquakes with magnitude M>6 are identified by their year of occurrence. The dotted ellipse labeled 1700 schematically indicates the subduction thrust earthquake of 26 January 1700.

THE GREAT EARTHQUAKE CYCLE

Earthquakes are complex natural events, and the great earthquakes of subduction zones are no different. However, the basic physical process is relatively simple and can be approximated with an elastic rebound model (Fig. 9). For the Cascadia subduction zone, convergence of the oceanic Juan de Fuca and continental North American plates, with the fault between them locked, results in elastic bending and buckling of the continental crust and accumulation of elastic stress in the region of the locked fault. The seaward edge of the continent is dragged down beneath the continental shelf, producing uplift farther inland. This situation continues for hundreds of years until the accumulated stress exceeds the sliding strength of the fault. Then the fault ruptures, releasing the stored elastic energy, a fraction of which radiates as earthquake waves. At the time of the earthquake, the seaward edge of continent springs back and upward, and the uplift gives way to rapid subsidence. The fault then relocks and the cycle resumes.

SUBDUCTION ZONE EARTHQUAKES

According to paleoseismicity studies along the coasts of British Columbia, Washington and Oregon, subduction thrust earthquakes have occurred at intervals ranging from a few hundred to more than a thousand years. Clague (1997) provides a review of the evidence and an extensive reference list; Hyndman (1995b) and Hyndman et al. (1996) provide overviews. The coastal geological
evidence derives from four basic types of effects that would be associated with a great earthquake: (i) subsidence of a marsh into the intertidal zone during an earthquake and subsequent burial by mud; (ii) deposition of sand on a co-seismically subsided surface due to generation of a tsunami; (iii) movement of liquefied sand, caused by the shaking, upward through cohesive sediment and spilling onto a co-seismically subsided surface; and (iv) deep sea deposits with alternating thin layers of mud and thick layers of sandier turbidites.

Marsh vegetation develops at a level between low and high tides in a number of sheltered inlets and bays along the west coast. Excavations below the marshes reveal that, at depth intervals of about one third of a meter, peat layers consisting of vegetation identical to that of the present marsh surface are buried. They are interpreted to be former intertidal marsh vegetation that was submerged by episodic sudden subsidence during a great earthquake. Following the earthquake, mud accumulated on the drowned marsh, raising the level back to intertidal whereupon marsh vegetation was re-established. The procedure repeats with each great earthquake. Atwater (1987) was the first to discover this phenomenon on the west coast and its relation to earthquakes. Atwater et al. (1995) summarize the data.

Other evidence enhances this interpretation. Many of the buried marsh surfaces are covered by sand layers (e.g., Clague and Bobrowsky, 1994). The sand is inferred to have been carried in by the large waves of the tsunami which accompanied the earthquake; i.e., tsunami deposits. Radiocarbon dating and ring studies of drowned trees in the inlets and bays provide a chronology for the earthquakes. These indicate that the intervals between successive subduction thrust earthquakes are irregular, but average about 500 years over the past 4000 years. The most recent such earthquake occurred about 300 years ago (e.g., Atwater et al., 1995).

Ground shaking from earthquakes is inferred to have produced liquefaction features less than 3000 years old at a number of sites on the west coast (e.g., Clague et al., 1992). Such features include dikes and sills of sediment that cross-cut and parallel the normal horizontal sedimentary layers, and blows that extruded mound-like masses of sediment.

On the deep ocean floor offshore of the bays and inlets, further evidence for past great earthquakes is found. Cores collected from deep sea channels in Cascadia basin display repeated layering of thin pelagic (deep-sea) fine clay deposits and thick, coarser, sandy deposits, these being emplaced by turbidity currents generated by submarine landslides triggered by the earthquake shaking. There is evidence for 13 large turbidity currents in Cascadia during the past 7500 years, or an average of about one every 600 years (Adams, 1990, 1996). The argument is based on synchronous turbidite deposits in different channels separated by as much as 580 km.

**THE CASCADE SUBDUCTION SEISMIC SOURCE ZONE**

The locked zone along the subduction thrust fault (Fig. 9), plus a transition zone as described below, determines the seismic source zone. This zone is limited
Fig. 9. Simplified cross section of a subduction zone showing the long-term deformation between earthquakes (top) and the short-term deformation during the earthquake (bottom) (modified from Hyndman et al. (1996)).

in both the updip and downdip directions. Seaward, the seismic zone is bounded by a region that does not generate earthquakes. In the relatively unconsolidated sediments in the region of the subduction zone deformation front, stable sliding behaviour is expected to occur. This is consistent with Byrne et al.'s (1988) global observation that earthquakes usually do not extend up dip to the trench axis. The seismic behaviour is partially controlled by the type of material contrast across the subduction thrust (Hyndman and Wang, 1993). Aseismic slip occurs with unconsolidated or semi-consolidated sediments as the overlying material, while seismic slip could occur where crystalline continental or arc crust, or very consolidated sediments, comprise the overlying material. Another important controlling factor on the region of free slip may be stable-sliding clay minerals common in the region of subduction zone faults. The clays dehydrate and become stronger with increasing temperature. At ~150°C, the clay minerals change into sedimentary rocks strong enough to sustain large elastic strains (Hyndman and Wang, 1993). A further factor that may limit stick-slip behaviour near the seaward end of the subduction thrust is high pore pressures. Davis and Hyndman (1989) have estimated that beneath the lower slope pore pressure is close to lithostatic, whereas below the upper slope it is only slightly above hydrostatic.

The downdip limit of the seismic source zone also may be thermally controlled. At some depth, a temperature is reached at which the rocks behave in a plastic, rather than brittle, manner. Laboratory measurements on continental
crustal rocks indicate that the critical temperature at which rocks revert to stable sliding is ~350°C (Hyndman and Wang, 1993, and references therein). This temperature corresponds well with that estimated for the maximum depth of earthquakes in many continental areas, including crustal ones landward of the subduction trench in Cascadia. Earthquakes may rupture downdip to depths where the temperature reaches ~450°C, the onset of feldspar plasticity. At depths with greater temperatures, the rocks deform plastically (Hyndman and Wang, 1993). The region along the subduction thrust corresponding to temperatures between 350°C and 450°C is called the transition zone and plays an important role in numerical modeling of observed thermal and deformational data.

An important constraint for thermal modeling is the observed surface heat flow. Lewis et al. (1988), Hyndman and Wang (1993) and Wang et al. (1995) have summarized the extensive marine and land data for the margin of the northern Cascadia subduction zone (Fig. 10(a)). For the numerical evaluation of thermal models and frictional heating to compare with the observations, Hyndman and Wang (1993) describe a steady-state two-dimensional finite element method; Wang et al. (1995) incorporate two important refinements to the original model. For the modeling, they take the seaward limit of seismogenic behavior to be located below the lower slope where the frontal thrusts (Fig. 4) reach the main sub-horizontal detachment, about 5 km landward of the deformation front. The stick-slip seismogenic zone is predicted to extend downward to about 350°C and the transitional, stable sliding zone to about 450°C. Results for a number of varying parameters were generated. The best fit of calculated heat flow compared with observations for the Lithoprobe corridor is shown in Fig. 10(a); the thermal field generated by the numerical modeling is shown in Fig. 10(b). The thermally estimated locked and transition zones are each about 50 km long. The model heat flow results provide an excellent fit with the observations.

On a present-day observational time scale, the rates of long-term deformation between thrust earthquakes are very slow and require very precise measurements to resolve. Vertical and horizontal strain rates across the southern Vancouver Island region have been determined through five types of geodetic data: (i) long term trends in tide gaug data; (ii) changes in repeated accurate levelling surveys; (iii) changes in repeated high accuracy gravity profiles; (iv) horizontal shortening observed in repeated precise positioning survey networks; and (v) continuously recording global positioning networks (Dragert et al., 1994). A summary of results indicates that the outer coast is moving upward at a rate of a few millimeters per year with a landward decrease in this uplift (Fig. 11(a)); and shortening perpendicular to the coast is occurring at a rate of about 0.1 mm per kilometer per year. Given the on-land coastal zone is about 100 km wide, this shortening represents only one quarter of the 40 mm per year plate convergence rate; the remainder is taken up as shortening further seaward across the continental shelf and slope. On a geological time scale, these rates are very fast and would produce high coastal mountains in a short time, if maintained. The mountains are not present, leading to the conclusion that the measured deformation is primarily elastic and will be released in the rebound accompanying the next great earthquake.
In the simplest models for the thrust earthquake cycle, the deformation between earthquakes is assumed to be entirely elastic. For the Cascadia subduction zone, Dragert et al. (1994) and Hyndman and Wang (1995) use a simple dislocation model of a dipping thrust fault in an elastic half space. They provide arguments that more complete descriptions of the earthquake cycle, involving a viscoelastic model that describes variations with time in the rate of deformation between earthquakes, do not change the primary results. In the models, a transition zone, as described previously, is included between the fully locked and downdip free slip portions of the fault since an abrupt discontinuity is not physically realistic. Neither the downdip distribution of slip nor the width of the transition zone can be accurately determined from the available deformation data. A linear variation of slip rate from completely locked to completely free is used to model the transition zone. Its width is selected on the basis of results from the thermal modeling (Fig. 11). In general, the wider the locked zone, the farther landward are the flexural uplift and the hingeline marking the transition between interseismic subsidence and uplift.

In Fig. 11(a), uplift rates from leveling data for the northern Cascadia and southwest Japan subduction zones are compared with dislocation modeling results for the temperature-controlled locked and transition zones illustrated in Fig. 11(b). Note that the maximum uplift rates for southwestern Japan are much further landward from the coast than those for northern Cascadia. This results in
widths for the locked and transition zones for Cascadia of 60 km and 60 km, respectively, compared with those for southwest Japan of 150 km and 45 km, respectively (Fig. 11). Such analyses can be carried out along the Cascadia margin wherever sufficient deformational data exist.

Based on results generated from 2-d thermal and deformational modeling compared with available observational data, Hyndman and Wang (1995) estimated the width of the locked and transition zones along the entire coast of the Cascadia subduction zone. Subsequently, Flück et al. (1997) developed a general 3-d dislocation model for thrust faults that accommodates curved fault geometry and non-uniform interseismic locking or co-seismic rupture. For the Cascadia subduction zone, the 3-d model includes the bend in the margin trend and the end effects of the subducting slab. The surface deformation is calculated for a locked zone or a rupture zone of variable width along the margin and compared with observational data. The map projections of the locked and transition zones for the
best fit from these calculations are shown in Fig. 12. The corresponding model has the thrust locked along the whole margin with an average locked zone width of 60 km and transition zone width of 60 km. The two zones lie mainly offshore below the continental shelf and slope. They vary smoothly along the margin, being greater off northern Washington where the thrust dip is shallower and narrower off central Oregon. Great earthquakes may rupture all or some portion of the locked and transition zones. However, under the assumption that the locked plus transition zones along the entire margin represent the maximum co-seismic rupture area, a great Cascadia subduction earthquake of moment magnitude $M_w = 9.2$ is possible.

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