ARTICLES

The Geological Signature of Great Earthquakes off Canada's West Coast

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SUMMARY
Geological and geophysical evidence, gathered in the last 15 years by a number of scientists working in Canada and the United States, leaves little doubt that some of the largest earthquakes on Earth occur at the Cascadia subduction zone on Canada's western doorstep. No such earthquake has occurred since European settlement of the region in the early 1800s, but the entire 900 km length of the thrust fault separating the Juan de Fuca and North America plates apparently ruptured during a magnitude-9 event on 26 January 1700. Evidence for this and older subduction earthquakes has been found at coastal wetlands from Vancouver Island to northern California. The geological evidence includes buried wetland soils produced by sudden, seismically induced subsidence, sheets of sand and gravel deposited by tsunamis, and sand dykes and blows generated by liquefaction during strong ground shaking. Dating of the buried soils and tsunami deposits in Washington and Oregon has shown that great Cascadia earthquakes have an average recurrence of 500 years; however, intervals between the seven most recent events range from less than 200 years to 700-1300 years. The hypothesis that subduction earthquakes occur in Cascadia is independently supported by geodetic measurements and the results of geophysical modelling, which collectively indicate that part of the plate boundary is locked and accumulating elastic strain that will be released during a future earthquake.

RÉSUMÉ
Les données géologiques et géophysiques recueillies par différents chercheurs au Canada et aux États-Unis au cours des quinze dernières années laissent peu de doutes quant à la réalité de séismes, parmi les plus grands sur Terre, qui se seraient produits dans la zone de subduction de Cascadia, sur la côte ouest du Canada. Aucun séisme de cet magnitude ne s'est produit depuis les débuts de la colonisation européenne dans la région, au début du 19^e siècle, mais il semble que toute la zone de chevauchement de 700 km de longueur, entre la plaque de Juan de Fuca et la plaque nord-américaine, ait été secouée par un séisme de magnitude 9, le 26 janvier 1700. On a trouvé des traces de ce séisme et d'autres plus anciens dans les basses-terres côtières à partir de l'Île de Vancouver jusqu'au nord de l'État de Californie. Les indices géologiques comprennent des sols de terres humides enfouis, des traces de subsidence d'origine sismique, des nappes de sable et gravier dues à des tsunamis et, des dykes et des volcans des sable engendrés par la liquéfaction de dépôts pré-existants lors de forts tremblements de terre. La datation des couches de sols enfouis et des dépôts de tsunamis des régions des États de Washington et de l'Oregon ont montré qu'un grand séisme survient, en moyenne, à tous les 500 ans dans la zone de Cascadia; toutefois, les intervalles entre les sept derniers événements fluctuent entre moins de 200 ans et 700-1 300 ans. L'hypothèse de séismes dans la zone de Cascadia causés par la subduction est corroborée indépendamment, et par des mesures géodésiques et par les résultats de la modélisation géophysique, les deux montrant qu'une portion de la bordure de la plaque est coincée et accumule ainsi l'énergie de contraintes élastiques, énergie qui sera libérée lors de séismes à venir.

INTRODUCTION
It struck without warning on 27 January 1700. Tsunami waves up to 3 m high swept a 1000-km length of the Pacific coast of Japan, damaging houses and inundating rice paddies and storehouses. This was not the first time the Japanese had experienced a large tsunami, but this one was different: it affected an unusually long section of the coast, yet did not follow a local earthquake.

Ten hours earlier, the same tsunami surged onto the coasts of California, Oregon, Washington and Vancouver Island. In these areas, however, the

Oral Tradition of the Destruction of Lo<u>h</u>t'a

Native oral traditions of tsunamis provide evidence for recent large earthquakes at the Cascadia subduction zone. An example, which is notable in that it mentions both the tsunami and the ground shaking that immediately preceded it, describes the destruction of Lo<u>h</u>t'a, the principal winter village of the Pachena Bay people on the west coast of Vancouver Island, at a time shortly before European contact:

"This story is about the first !Anaql'a or "Pachena Bay" people. It is said that they were a big band at the time of him whose name was <u>H</u>ayoqwis'is, 'Ten-On-Head-On-Beach.' He was the Chief; he was of the Pachena Bay tribe; he owned the Pachena Bay country. Their village site was Lo<u>h</u>t'a; they of Lo<u>h</u>t'a live there. I think they numbered over a hundred persons...There is no one left alive due to what this land does at times. They had no way or time to try to save themselves. I think it was at nighttime that the land shook...They were at Lo<u>h</u>t'a; and they simply had no time to get hold of canoes, no time to get awake. They sank at once, were all drowned; not one survived...I think a big wave smashed into the beach. The Pachena Bay people were lost...But they on their part who lived at Ma:lts'a:s, 'House-Up-Against-Hill', the wave did not reach because they were on high ground. Right against a cliff were the houses on high ground at M'a:lsit, 'Coldwater Pool'. Because of that they came out alive. They did not drift out to sea along with the others..." (Arima *et al.*, 1991, p. 230-231).

waves were much larger than in Japan, ranging up to 20 m in height in some bays and at the heads of inlets. Many coastal Indian villages were devastated. Oral traditions, handed down through the generations, tell of powerful waves that swept away entire villages on a cold winter night.

We now know that the tsunami that struck both Japan and the Pacific coast of North America in 1700 was triggered by a very large earthquake at the Cascadia subduction zone, and that earthquakes like this occur, on average, once every 500 years (Atwater *et al.*, 1995; Clague, 1997). This paper recounts the scientific detective work that literally unearthed the evidence for these earthquakes and improved our understanding of seismic hazard on the west coast.

SHIFTING GROUND: THE FIRST SUGGESTIONS THAT GREAT EARTHQUAKES HAPPEN IN CASCADIA

Oceanic crust of the Juan de Fuca plate descends eastward beneath low-density continental crust of North America at the Cascadia subduction zone, off Canada's west coast (Fig. 1). Prior to the early 1980s, no one thought that this subduction zone could generate great (magnitude-8 or larger) earthquakes. Geophysicists had earlier established that subduction is indeed occurring off the Pacific coast (Riddihough and Hyndman, 1976; Riddihough, 1984), but it was widely assumed that the North America and Juan de Fuca plates slide smoothly and continuously past one another (Ando and Balazs, 1979). Yet, even then, the record of instrumented earthquakes hinted that this assump-

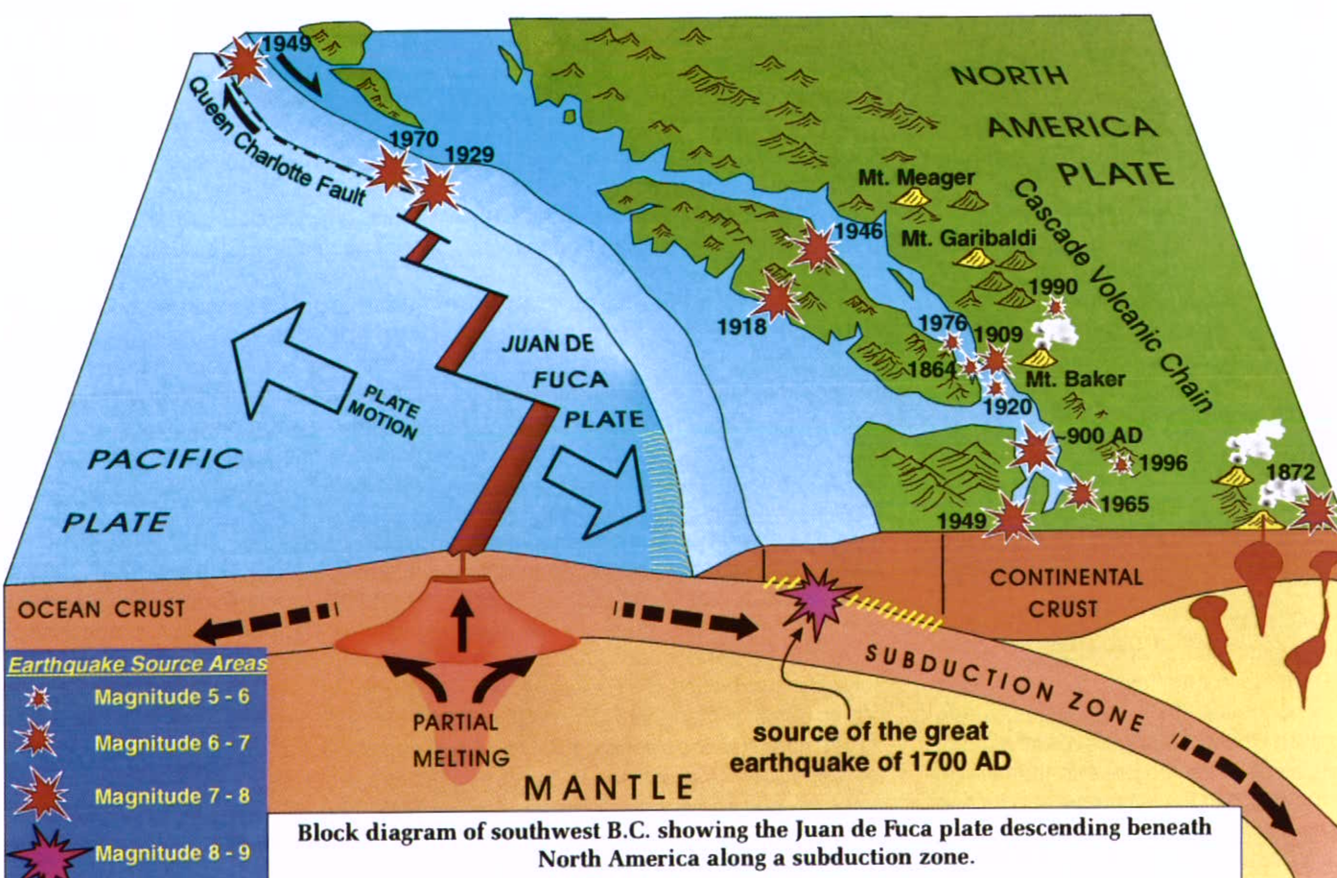

Figure 1 Block diagram of southwest British Columbia, showing the Juan de Fuca plate descending beneath North America at the Cascadia subduction zone. The diagram also shows the locked portion of the thrust fault separating the two plates and epicentres of large historical earthquakes in the region.

tion was incorrect: there should be small earthquakes on the fault that separates the two plates if sliding is occurring continuously, but no such earthquakes have been recorded.

Geophysical thinking on this subject began to shift in 1981 when Jim Savage and colleagues published a paper showing that the pattern of crustal strain in western Washington is consistent with a locked subduction zone (Savage *et al.*, 1981). Soon thereafter, Heaton and Kanamori (1984) compared subduction of the Juan de Fuca plate with subduction at other places where there are great earthquakes. Their ideas were further developed by Heaton and Hartzell (1986, 1987), who suggested that the boundary between the North America and Juan de Fuca plates is locked and is accumulating strain that could ultimately be released in one or more great earthquakes. Garry Rogers made a Canadian contribution to the subject by showing that the Cascadia subduction zone is similar to other Pacific subduction zones that have produced great earthquakes in the last several centuries (Rogers, 1988). He concluded that great earthquakes have also occurred in Cascadia but that the last one predates European exploration of the region.

These studies provided a foundation for more recent geophysical research on the Cascadia subduction zone, which is summarized below. The early work also stimulated the search for geological evidence for past earthquakes in Cascadia. The paradigm that underlies much of the recent geophysical and geological research is the great earthquake cycle.

THE GREAT EARTHQUAKE CYCLE

The process of sudden fault slip during a subduction earthquake may be approximated by an elastic dislocation model. Convergence between the North America and Juan de Fuca plates causes elastic bending and buckling of the crust. The seaward edge of the continent is dragged down beneath the continental shelf, producing a flexural bulge farther inland (Fig. 2). After hundreds of years, the stress exceeds the strength of the fault, causing it to rupture. The rupture releases stored elastic energy, a small fraction of which radiates outward as earthquake waves. The seaward edge of the continent springs back and upward during the earthquake, and the bulge collapses (Fig. 2). The fault then locks, and strain begins to accumulate anew.

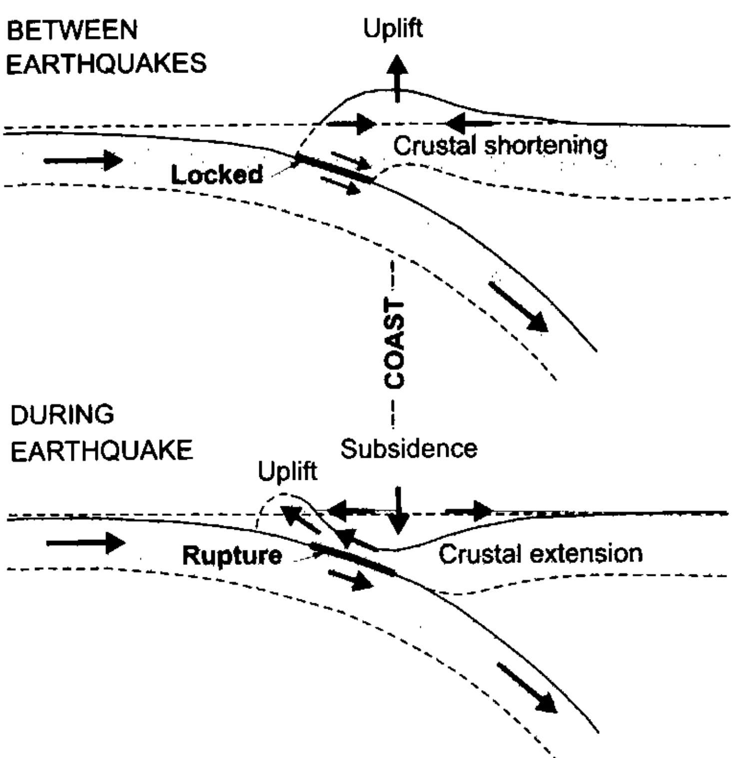

Figure 2 Simplified cross-section of a subduction zone, showing coseismic and interseismic deformation. Between earthquakes (top), the seaward edge of the continent is dragged down, and a flexural bulge forms farther inland. During an earthquake (bottom), the area seaward of the locked zone is uplifted and the flexural bulge collapses. Vertical exaggeration is approximately 6000:1. (Modified from Hyndman and Wang, 1993, fig. 10)

Chile 1960 and Alaska 1964: Analogues for the 1700 Cascadia Earthquake

The elastic dislocation model explains the pattern of coseismic deformation of the giant subduction earthquakes in Chile in 1960 (Plafker and Savage, 1970) and Alaska in 1964 (Plafker, 1969), which are probably good analogues for Cascadia subduction earthquakes. During the two historical earthquakes, an area nearly 1000 km long, directly above the fault rupture, moved seaward and upward (Fig. 3), producing very strong shaking over a large area and triggering devastating tsunamis. A broad area landward of the zone of uplift subsided up to 2 m when the flexural bulge collapsed (Fig. 3).

GEODETIC STUDIES: EVIDENCE MOUNTS FOR GREAT EARTHQUAKES IN CASCADIA

The pattern of deformation in Cascadia, which has been inferred from geodetic measurements, indicates that a portion of the fault separating the North America and Juan de Fuca plates is locked and accumulating elastic strain. The types of geodetic measurements that have provided this information include 1) repeated levelling of benchmarks along survey lines (Reilinger and Adams, 1982; Dragert, 1987; Dragert *et al.*, 1994), 2) tide gauge records (Ando and Balazc, 1979; Reilinger and Adams, 1982; Holdahl *et al.*, 1989; Dragert *et al.*, 1994), 3) repeated gravity surveys (Dragert *et al.*, 1981, 1994), 4) repeated positional survey networks (Savage *et al.*, 1991; Dragert *et al.*, 1994), and 5) continuously recording Global Positioning System (GPS) networks (Lisowski *et al.*, 1989; Dragert and Hyndman, 1995). The first three types of measurements (levelling, tide gauges and gravity) reveal elevation changes over periods of years to decades. All give similar results (Dragert *et al.*, 1994): the coast of Cascadia is rising at rates of 1-4 mm a^{-1}, with uplift decreasing inland to near zero 100 km from the coast (Fig. 4).

Present-day horizontal deformation in the region has been estimated from repeated positional survey and GPS networks. Positional surveys involve precise laser measurement of distances between benchmarks on mountain tops up to 50 km apart. This method is timeconsuming and costly, and has been replaced in recent years by GPS, which uses distance-ranging to satellites. Shortening, based on measurements made by both techniques, is about 0.1 $m \cdot a^{-1}$ across the 100-km-wide coastal zone.

Viewed in total, the geodetic measurements indicate that the edge of North America is being compressed and uplifted. Present-day uplift rates are at least 10 times higher than average late Quaternary rates (Kelsey *et al.*, 1994a), which suggests that the deformation cannot continue indefinitely and that much of the strain is being stored elastically and will be released during future earthquakes.

GEOPHYSICAL MODELS OF THE DEFORMATION

The current pattern of deformation in Cascadia has been modelled mathematically by scientists at the Geological Survey of Canada's Pacific Geoscience Centre (PGC) to determine the position of the inferred locked zone (Dragert *et al.*, 1994; Hyndman and Wang, 1995). Temperature is the main factor controlling the width of the seismic source zone in Cascadia (Hyndman and Wang, 1993). The thrust fault separating the North America and Juan de Fuca plates may become seismogenic at about 150°C, the temperature at which clay minerals dehydrate and unconsolidated sediments change into rocks strong enough to sustain large elastic strains (Hyndman and Wang, 1993). This temperature is reached at a depth of about 10 km, beneath the lower continental slope where oceanic crust is covered by thick sediments. The downdip limit of the seismogenic zone is also thermally controlled. The critical temperature at which there is stable sliding, rather than "stick-slip" movement, is about 350°C (Hyndman and Wang, 1993, and references therein). Earthquakes may rupture farther downdip, in a transition zone, to temperatures of about 450°C, beyond which the rocks behave plastically (Hyndman and Wang, 1993). These temperatures are reached at shallow depths in Cascadia, compared with most other subduction zones, because the incoming oceanic plate is hot and is covered by a thick blanket of insulating sediments.

Detailed thermal modelling has been carried out along a series of profiles across the Cascadia margin to obtain temperatures on the plate-boundary thrust fault (Hyndman and Wang, 1993, 1995). The important conclusion of this work is that the position of the 350° isotherm on the fault agrees with the downdip limit of the locked zone inferred from the geodetic data (Fig. 4).

Roy Hyndman and his PGC colleagues next demonstrated that the inferred locked portion of the Cascadia thrust fault corresponds to the area of great earthquake rupture. They did this by applying the type of analysis outlined above to a subduction zone in southwest Japan that has a history of well documented great earthquakes. They showed that the rupture areas of great Japanese subduction earthquakes in 1944 and 1946 correspond closely to the locked seismogenic zone inferred from analysis of geodetic data and temperature constraints (Fig. 4; Hyndman *et al.*, 1995).

THE SEARCH FOR GEOLOGICAL TRACES OF GREAT CASCADIA EARTHQUAKES

By the late 1980s, arguments based on the above-mentioned geodetic data had convinced many earth scientists that great earthquakes occur in Cascadia. The onus was then on geologists to find evidence for such earthquakes. Brian Atwater of the United States Geological Survey was the first geologist to test the great-earthquake hypothesis; he reasoned that earthquakes of magnitude 8 or larger must leave evidence in the geological record. Historical subduction earthquakes in Japan, Chile and Alaska provided Atwater with clues as to what to look for and where to find it.

Land-level Change

Tens of thousands of square kilometres of land and ocean floor are suddenly raised or lowered during great subduction earthquakes (Fig. 3). An elongate zone of coseismic uplift, with maximum displacements of 5 m or more, is present outboard of the locked zone. This zone can be up to 100 km wide and 1000 km long. Uplift occurs in this area because the edge of the overriding plate springs back and upward at the time of

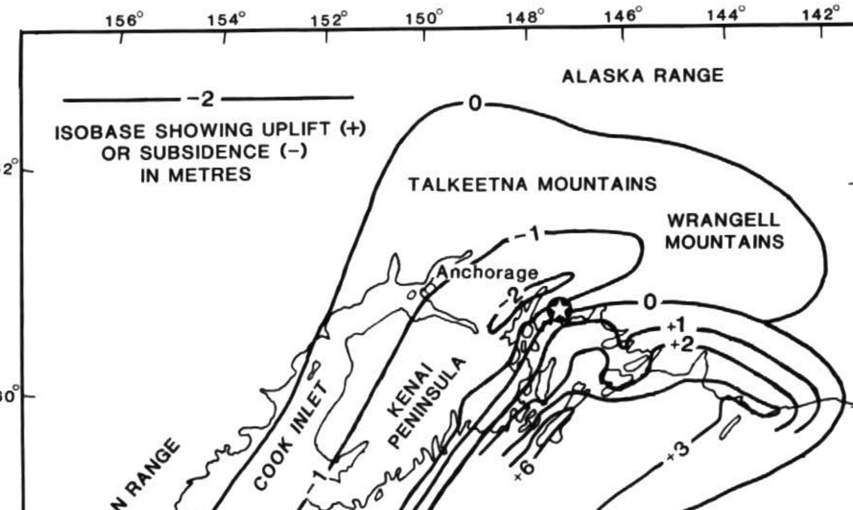

Figure 3 Map showing coseismic uplift and subsidence produced by the 1964 Alaska earthquake. The earthquake is an analogue for great earthquakes at the Cascadia subduction zone. (Modified from Plafker, 1969, fig. 3)

the earthquake (Fig. 2). The uplift zone is bordered landward by an extensive area of subsidence, with displacements of up to 2 m (Fig. 3). Subsidence is greatest just inboard of the ruptured section of the thrust fault.

In Cascadia, the area that would be uplifted during an earthquake lies entirely offshore on the continental shelf. Because there are no islands there, features indicative of coseismic uplift, notably raised beaches and raised intertidal rock platforms, are absent. Much of the Pacific coast from central Vancouver Island to northern California, however, lies within the zone of expected coseismic subsidence. It is in this region that geologists have focussed their search for evidence of the earthquakes. In particular, tidal marshes provided an excellent target because vegetated marsh surfaces might submerge during an earthquake and later be covered by tidal mud (Fig. 5). The expectation of geologists, then, was that a subduction earthquake in Cascadia would leave a distinctive peat-mud couplet in the marsh stratigraphic record: a peat layer, representing the pre-earthquake marsh surface, abruptly overlain by a layer of tidal mud (Fig. 6). Furthermore, trees at the edge of the marsh might subside into the intertidal zone where their roots would be submerged during part of the tidal cycle. The trees would then die, but their roots and stems would remain long afterward as testament to the earthquake.

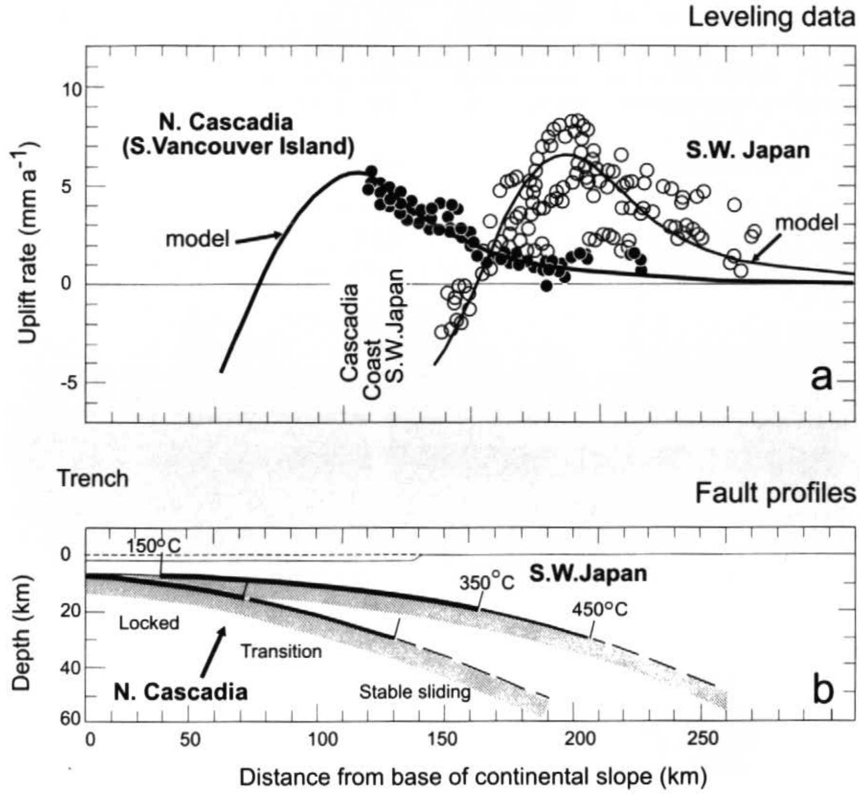

Figure 4 (**a**) Uplift of coastal regions of northern Cascadia and southwest Japan as determined from repeated leveling surveys (symbols) and as predicted for the temperature-controlled zones shown in Figure 4b (curves). (**b**) Profiles showing locked (150°-350°C), transition (350°-450°C), and stable sliding (>450°C) zones, corresponding to model curves in Figure 4a. Note that the flexural bulge and inferred locked zone extend farther landward in Japan than in Cascadia. (Modified from Hyndman and Wang, 1995, fig. 16)

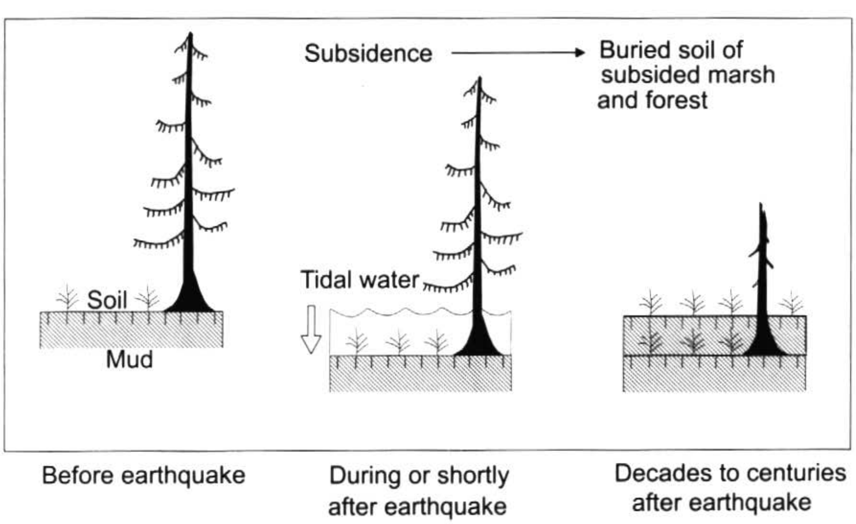

Figure 5 Simplified diagram illustrating how a peat-mud couplet forms in a tidal wetland during a great subduction earthquake. The marsh subsides into the intertidal zone during the earthquake and subsequently is covered by tidal mud. (Modified from Atwater *et al.*, 1995, fig. 3)

Peat-mud couplets were first reported in Cascadia in 1987 at an estuary in southwest Washington (Atwater, 1987). They have since been found at nearly 20 sites on the Pacific coast between Vancouver Island and northern California, and are particularly well exposed in 2 m- to 4 m-high outcrops at several estuaries in southwest Washington and northwest Oregon (references in Atwater *et al.*, 1995; Clague, 1997; and Clague *et al.*, 1998).

Caution is required when interpreting such stratigraphy. Buried peats alone are not proof of coseismic subsidence because they are also found on tectonically passive coasts (*e.g.*, Tooley, 1978; van de Plassche, 1982, 1991; Shennan, 1986, 1987; Devoy, 1987; Fletcher *et al.*, 1993). A number of criteria need to be evaluated to rule out nonseismic causes such as storms, river floods, stream migration, local changes in sedimentation, and sea level rise. The criteria, discussed below, include suddenness of submergence, amount of submergence, lateral extent of buried peats, synchroneity of submergence at widely spaced sites, and coincidence of submergence with tsunami deposits (Nelson *et al.*, 1996c).

Many buried peats in Cascadia are abruptly overlain by tidal mud. Stems and leaves of herbaceous plants rooted in some peats are covered by mud (Atwater and Yamaguchi, 1991), which suggests that submergence was rapid. Rapid submergence is also indicated by abrupt changes in fossil plant and animal assemblages across contacts between peats which formed in forest-margin and high-marsh environments, and overlying muds which are low-marsh and tidal-flat deposits (Atwater and Yamaguchi, 1991; Nelson and Kashima, 1993; Guilbault *et al.*, 1995, 1996; Hemphill-Haley, 1995; Nelson *et al.*, 1996a,c; Atwater and Hemphill-Haley, 1997; Shennan *et al.*, 1997). Amounts of coseismic subsidence have been estimated by determining the elevations of the marsh surface immediately before and after submergence. Pre- and post-earthquake marsh elevations are determined from elevation ranges of living intertidal plants and animals that occur as fossils in the peats and muds (Guilbault *et al.*, 1995, 1996; Hemphill-Haley, 1995; Nelson *et al.*, 1996a). Submergence estimated in this manner is commonly in the range of 0.5 m to 2 m, too large to be explained by nonseismic processes (Nelson *et al.*, 1996c; Peterson *et al.*, 1997). "Ghost" forests, killed by the most recent great Cascadia earthquake in January 1700, still exist in a few areas where submergence was near the upper limit of this range.

Buried peats at many estuaries in Cascadia are continuous over distances of tens to hundreds of metres and have little relief; thus their burial cannot be explained by deposition in stream or tidal channels. Some peats in southwest Washington and northern Oregon have been correlated over distances of 50 km to 200 km on the basis of 1) differences in the thickness of the peats in vertical sequences of sediments at different sites (these differences create a "bar code" pattern that is similar from site to site), 2) differences in the abundance of tree roots in the peats, and 3) radiocarbon dating, which has shown that correlated peats are not statistically different in age (Atwater, 1992; Atwater and Hemphill-Haley, 1997).

Tsunamis

Tsunamis accompany great subduction earthquakes in Cascadia (Fig. 7). As mentioned earlier, the tsunami of the most recent of these earthquakes, in 1700, was recorded in Japan, and devastated Native Indian communities along the west coast of North America.

A large tsunami leaves a distinctive geological footprint in some coastal environments. Turbulent onrushing waters may carry sand, gravel and flotsam far inland over low flat terrain. The sedi-

Figure 6 Buried peat layer (arrow) exposed in a pit dug in a tidal marsh near Tofino, British Columbia. The peat layer delineates a former marsh surface that subsided about one-half metre during the last great Cascadia earthquake 300 years ago. The peat is overlain by a thin layer of sand (tip of knife), deposited by the tsunami that followed the earthquake. The sand, in turn, is overlain by tidal mud (photo by J.J. Clague).

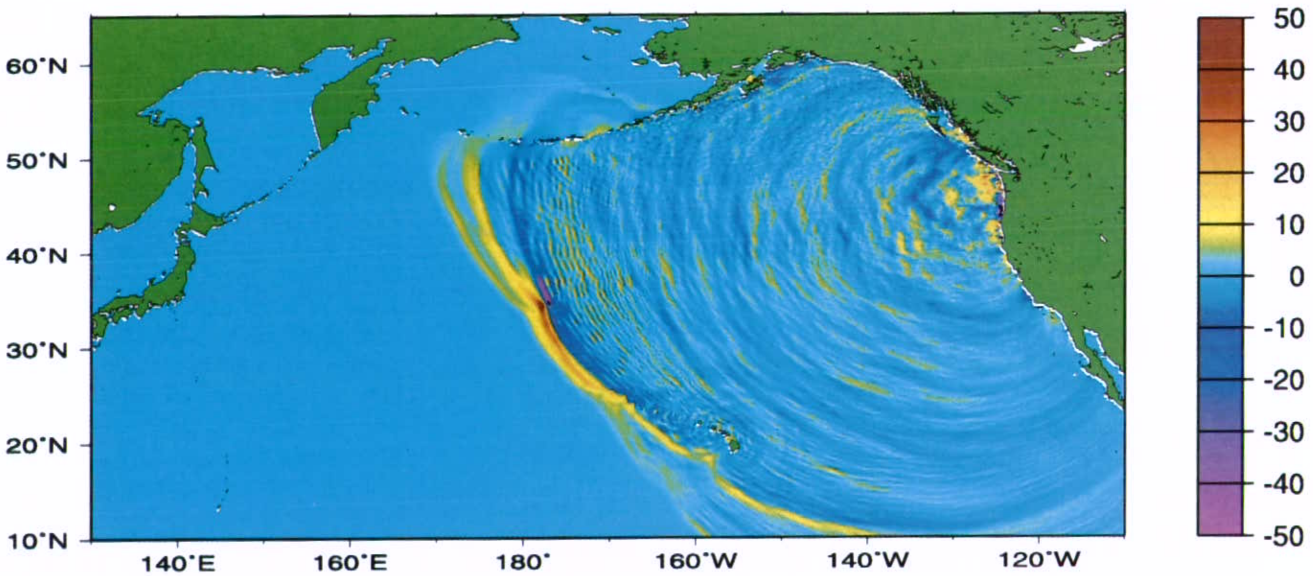

Figure 7 Snapshot of the simulated tsunami of the 1700 Cascadia earthquake, six hours after initiation. The simulation was done using a finite-difference method, using digital bathymetry data. The sea-floor displacement used in the simulation was that produced by an M_w 9 earthquake that ruptured the entire length of the Cascadia subduction zone (image produced by K. Satake).

ment is deposited as the flow slackens and then ebbs. A good place to look for this type of sediment is tidal marshes, the same places where coseismic subsidence is likely to be recorded. At many tidal marshes in Cascadia, thin sheets of sand or gravel rest directly on peaty marsh soils and are overlain by tidal muds (Fig. 8; Atwater, 1987, 1992, 1996; Reinhart and Bourgeois, 1987, 1989; Darienzo and Peterson, 1990, 1995; Atwater and Yamaguchi, 1991; Clague and Bobrowsky, 1994a,b; Darienzo *et al.*, 1994; Peterson and Darienzo, 1996). At some marshes, delicate stems and leaves of herbaceous plants rooted in the underlying peat extend into, or are matted down beneath, the sand (Atwater and Yamaguchi, 1991; Clague and Bobrowsky, 1994b). The association of subsidence and sand deposition argues for a common cause: an earthquake that caused the coast to subside while generating a tsunami offshore (Fig. 9). In contrast, at marshes outside the zone of coseismic subsidence, tsunami deposits are not associated with buried peaty marsh soils, but rather occur within monotonous peat or mud sequences (Benson *et al.*, 1997).

Figure 8 Tsunami sand overlying an eroded peaty soil in a pit dug near Tofino, British Columbia. The tsunami was triggered by the last great Cascadia earthquake 300 years ago. (Photo by J.J. Clague)

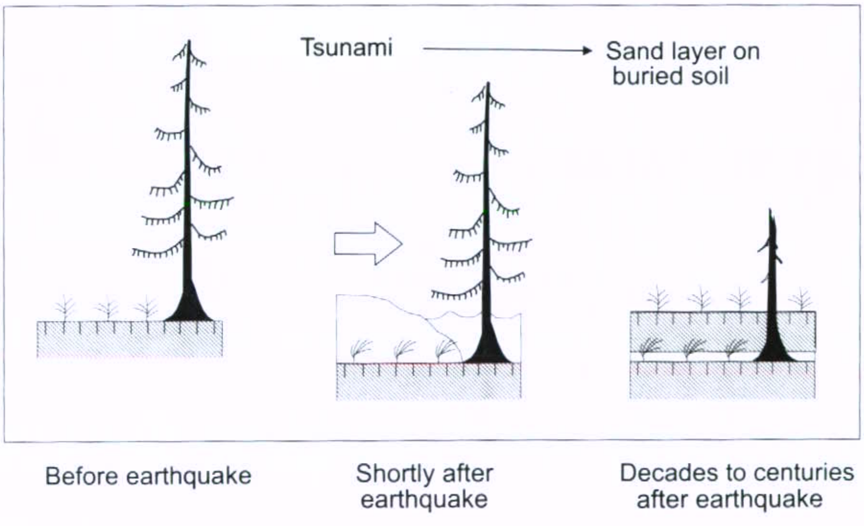

Figure 9 Simplified diagram showing tsunami sand being deposited on a coseismically subsided marsh surface immediately after an earthquake. The sand subsequently is covered by tidal mud. (Modified from Atwater *et al.*, 1995, fig. 3)

Tsunami of the 1964 Alaska Earthquake and Its Impact on British Columbia

The most destructive historical tsunami in British Columbia was triggered by the great Alaska earthquake of 27 March 1964. A series of waves radiated out from the epicentre of the earthquake near the head of Prince William Sound at a velocity of over 800 $km \cdot h^{-1}$ and, within a few hours, reached the outer shores of British Columbia (Wigen and White, 1964; Murty and Boilard, 1970; Thomson, 1981). Although there was no loss of life in British Columbia, most of the 130 deaths attributable to the Alaska earthquake, including some as far away as California, were caused by tsunamis (Hansen *et al.*, 1966).

The hardest hit community in British Columbia was Port Alberni, located at the head of Alberni Inlet (Fig. 10). Three main waves struck Port Alberni between 12:20 a.m. and 3:30 a.m. on 28 March 1964. Most people in the town were asleep when the first wave rolled in. The sea surged up Somass River at a velocity of about 50 $km \cdot h^{-1}$ and spilled onto the land, inundating entire neighbourhoods in chest-deep water. This first wave reached 3.7 m above geodetic datum and knocked out the Port Alberni tide gauge. The second and most destructive wave swept into town less than two hours later, at 2 a.m. The lights of the waterfront mills went out as the water smashed through the facilities. The ground floor of the Barclay Hotel, 1 km inland, was splintered by the surging water; guests had to be plucked from an upper floor by police in boats. Logs and debris crashed into buildings, automobiles were cast about, and houses were swept off their foundations and hurled inland. As the water subsided, some buildings were dragged seaward; two houses drifted into Alberni Inlet and were never seen again. The second wave reached 4.3 m above geodetic datum. The third wave, which arrived at about 3:30 a.m., was the largest of all, but because the tide had fallen it crested at 3.9 m and did little further damage. Other waves oscillated in Alberni Inlet with decreasing strength for another two days. Two hundred and sixty homes in Port Alberni were damaged by this tsunami — sixty extensively — and the total economic losses here and elsewhere on Vancouver Island were estimated at about C$10 million (1964 dollars).

The tsunami left a sheet of sand in marshes at Port Alberni (Clague *et al.*, 1994) and many other sites on Vancouver Island (Benson *et al.*, 1997). At most sites, including Port Alberni, the sand is thinner and finer than the deposit of the AD 1700 tsunami, which suggests that the older tsunami was the larger of the two (Fig. 10).

The interpretation that beds of coarse sediment in peat and mud sequences in coastal wetlands are tsunami deposits should be made with caution, for similar beds may be deposited during river floods or storms or as a result of tidal channel migration. Most inferred tsunami deposits in Cascadia, however, are distant from streams, contain marine or brackish-water microfossils, and/or become thinner and finer landward. These characteristics argue for an offshore, rather than a fluvial, source (Clague and Bobrowsky, 1994b; Hemphill-Haley, 1995; Nelson *et al.*, 1996c; Atwater and Hemphill-Haley, 1997). The deposits of rare large storms are commonly difficult to distinguish from those of tsunamis, but sediment containing marine fossils at elevations well above the reach of storm waves is probably a tsunami deposit (Nelson *et al.*, 1996c). Storm deposits are unlikely to be found far inland from the coast in protected inlets because as water levels rise, current velocities in channels drop below values required to transport sufficient sand to form widespread sheets (Bourgeois and Reinhart, 1989; Peterson and Darienzo, 1996). Also, some sand sheets in Cascadia comprise two or more beds, suggestive of deposition by successive tsunami waves rather than a storm surge (Benson *et al.*, 1997, fig. 6).

Tsunami deposits have also been found in low-lying lakes along the coasts of Vancouver Island (Hutchinson *et al.*, 1997; Clague *et al.*, 1998, 1999) and Oregon (Kelsey *et al.*, 1994b; Nelson *et al.*, 1996b, 1998). Tsunamis enter the lakes by surging up outlet streams or by crossing beaches and dunes. Sediments deposited by the tsunamis are strikingly different from normal, organic-rich lake sediments; they are generally much coarser and may contain marine fossils and abundant plant detritus. Run-up heights of tsunamis can be estimated by comparing sequences of sand layers from two or more lakes at different elevations within the same area. The sources of the earthquakes that produce the tsunamis, however, cannot be determined from lake sediments alone.

The sizes of tsunamis triggered by Cascadia subduction earthquakes have been estimated using numerical models (Ng *et al.*, 1991; Whitmore, 1993). The largest waves may have heights of 5 m or more at some sites on the open coast and could further amplify by a factor of up to three in bays and inlets.

Shaking

Strong seismic shaking causes loose, water-saturated silt, sand and even gravel to transform from a solid to a liquid state, a phenomenon termed liquefaction. The transformation happens when pore pressures increase to the point that sediment grains no longer support one another, but rather become suspended in the pore water. When liquefaction occurs at depth, as often is the case, overlying cohesive sediments may glide on the liquefied layer toward a free face such as a river bank, creat-

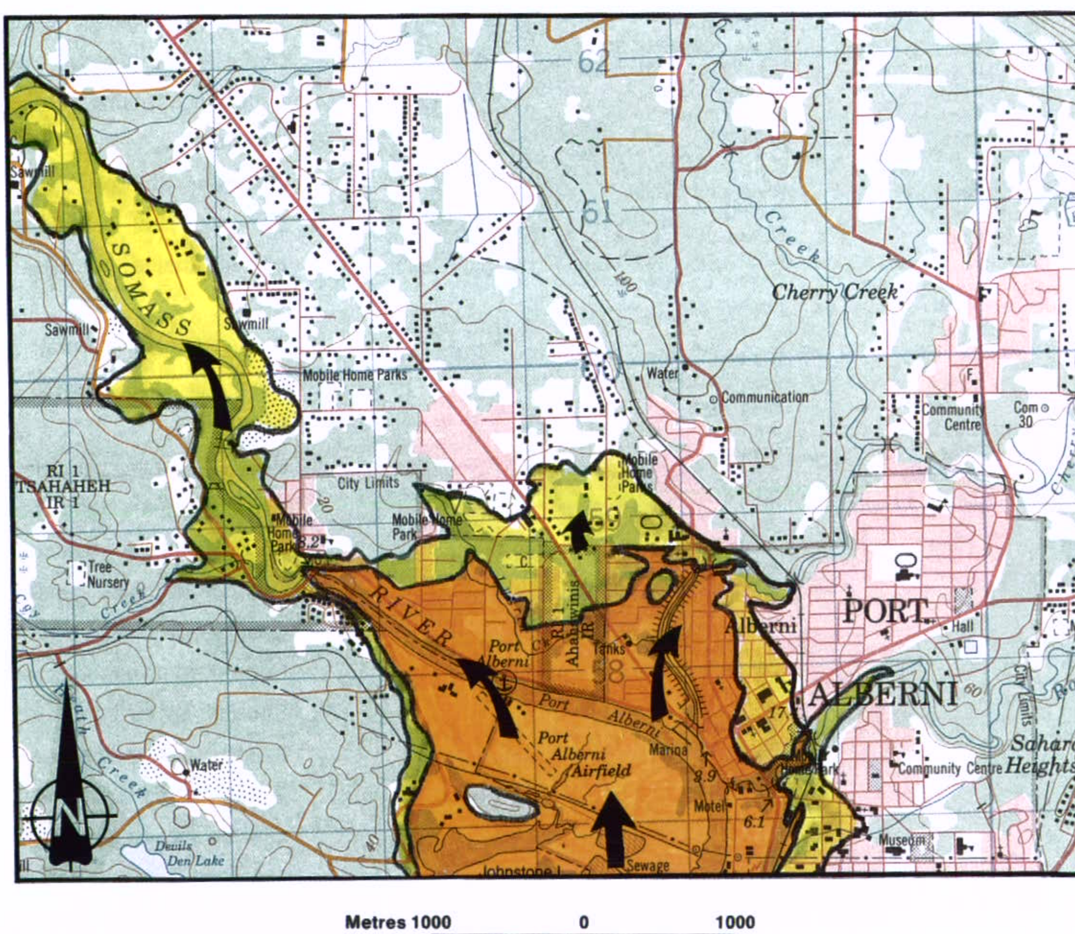

Figure 10 Map showing areas inundated by the 1964 Alaska and 1700 Cascadia tsunamis at Port Alberni on Vancouver Island (1964 area = orange; 1700 area = yellow + orange). Inundation areas are approximate: that of the 1700 tsunami is based on a maximum wave height of 16 m, calculated using a numerical model (Ng *et al.*, 1991).

ing fractures along which the liquefied sediment flows. The sediment stops moving along the fractures after the shaking ends and is left as dykes and sills. Liquefied sediment may also erupt onto the surface to form mound-like accumulations referred to as blows.

Sills, dykes and blows ascribed to earthquake-induced liquefaction have been found in Cascadia near Vancouver (Clague *et al.*, 1992; Mathewes and Clague, 1994), along the lower Columbia River at the Washington-Oregon border (Atwater, 1994; Obermeier, 1995), and at other sites in western Washington and Oregon (Kelsey *et al.*, 1993; Moses *et al.*, 1993; Siskowic *et al.*, 1994; Obermeier, 1995; Peterson and Madin, 1997). Only the liquefaction features along the lower Columbia River, however, can be definitely linked to a subduction earthquake (Atwater *et al.*, 1995). Sand was erupted onto tidal swamps in the Columbia River estuary, 30-60 km inland from the coast, 300 years ago. The sand blows rest directly on a buried peat that itself is cut by dykes connected to the blows. This association suggests that the liquefaction and subsidence occurred simultaneously, which argues for a common cause, specifically a large earthquake (Fig. 11).

SOURCE AND SIZE OF EARTHQUAKES

The evidence for sudden subsidence, tsunamis, and strong ground shaking at numerous sites convincingly demonstrates that large earthquakes have occurred in Cascadia in the recent geological past. Other arguments, however, are required to infer a plate-boundary, as opposed to a crustal, origin for the earthquakes. The arguments, which focus on the location and areal extent of coseismic subsidence and on earthquake frequency, are briefly reviewed below (see Atwater *et al.*, 1995 and Clague, 1997 for details).

Rupture Location and Area

Plate-boundary rupture is required to explain the location and amount of coseismic subsidence in Cascadia. Such rupture should cause up to 2 m of sudden subsidence of coastal areas by elastically thinning and extending the North America plate (Fig. 2; Savage, 1983; Atwater, 1988; Darienzo and Peterson, 1990; Goldfinger *et al.*, 1992). Plant fossils and sediment types show that large areas of the coast from northern California to Vancouver Island subsided 0.5 m to 2 m during past earthquakes (Atwater *et al.*, 1995 and references therein; Guilbault *et al.*, 1995, 1996; Hemphill-Haley, 1995; Atwater and Hemphill-Haley, 1997; Peterson *et al.*, 1997). The lack of evidence for greater amounts of subsidence is consistent with plate-boundary rupture.

Large earthquakes are also required to account for the inferred large area of most or all of the ruptures. Magnitude-8 earthquakes rupture fault areas of about 10,000 km^2 (Wells and Coppersmith, 1994); ruptures in Cascadia are probably larger than this, based on inferences that can be made about their widths and lengths.

Geodetic data and thermal modelling suggest that a 50 km to 100 km width of the North America-Juan de Fuca plate boundary would rupture during a subduction earthquake (Savage *et al.*, 1991; Hyndman and Wang, 1993; Dragert *et al.*, 1994; Mitchell *et al.*, 1994; Dragert and Hyndman, 1995). Such rupture widths are consistent with what is known about the pattern and extent of sudden subsidence inland from the coast along the Cascadia margin (Atwater, 1992, 1994).

Rupture lengths of tens to hundreds of kilometres are implied by the presence of correlative buried peats over such distances in southern Cascadia (Darienzo *et al.*, 1994; Atwater *et al.*, 1995). Stratigraphic correlation and radiocarbon ages of plants killed during the most recent earthquake indicate a single rupture along 900 km of the subduction zone (Nelson *et al.*, 1995). A rupture of this length implies an earthquake of magnitude 9 or larger.

Plate-boundary rupture may also have triggered localized displacements along shallow folds and faults in the North America plate (McInelly and Kelsey, 1990; Clarke and Carver, 1992; Goldfinger *et al.*, 1992, 1996; Nelson, 1992b). However, no single structure in the North America plate can account for all areas that subsided during past Cascadia earthquakes; thus the subsidence cannot be due to crustal faulting alone.

Earthquake Frequency and Displacement

Recurrence intervals in Cascadia appear to require earthquakes at least as large as magnitude 8. Recurrence intervals estimated from coastal geological evidence average about 500 years (see below), and 300 years have elapsed since the most recent great earthquake. Over a period of 500 years, the orthogonal component of plate convergence (McCaffrey and Goldfinger, 1995) would equal at least 17 m. In comparison, slip during historic magnitude-9 earthquakes in Chile and Alaska averaged about 20 m (Plafker, 1969; Plafker and Savage, 1970; Barrientos and Ward, 1990). Viewed differently, if the boundary between the North America and Juan de Fuca plates has a total rupture area of 60,000 km^2 (Hyndman and Wang, 1993), 500 years of plate

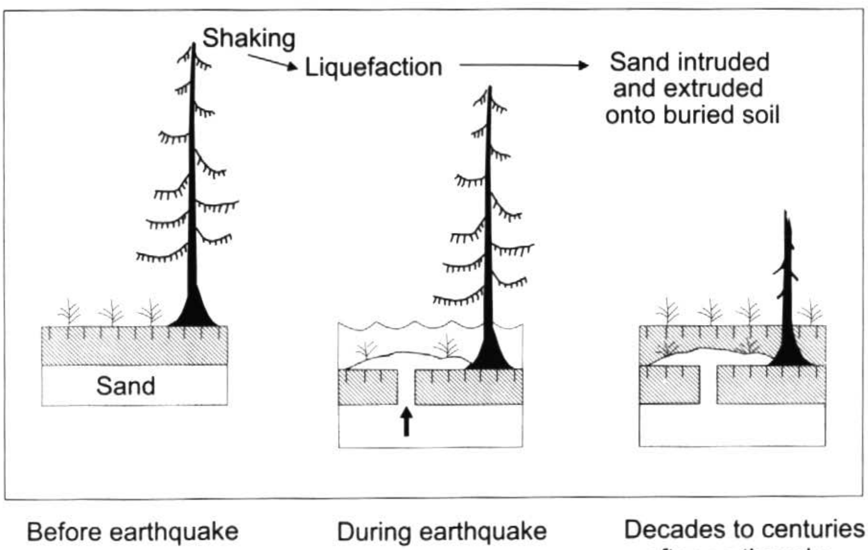

Figure 11 Diagram showing liquefied sand moving upward through cohesive sediment and spilling onto a coseismically subsided marsh surface during or shortly after an earthquake. The vented sand is later covered by tidal mud. (Modified from Atwater *et al.*, 1995, fig. 3).

motion produces enough seismic moment for one earthquake of magnitude 9 or more than 30 earthquakes of magnitude 8 (Hanks and Kanamori, 1979); yet no such earthquake has happened in the last 300 years.

These arguments assume, however, that all of the accumulated strain is elastic, which is not the case; some strain is permanent, as shown by long-term deformation of the North America plate (Goldfinger *et al.*, 1992, 1996). Part of the plate convergence may also be accommodated by aseismic slip at the plate boundary (Kanamori and Astiz, 1985; Pacheco *et al.*, 1993).

EARTHQUAKE RECURRENCE

The ages of the most recent subduction earthquakes in Cascadia have been estimated by dating buried peats and tsunami deposits at many sites in the region (Peterson *et al.*, 1988; Darienzo and Peterson, 1990, 1995; Atwater and Yamaguchi, 1991; Atwater *et al.*, 1991, 1995; Atwater, 1992, 1996; Nelson, 1992a,b; Clague and Bobrowsky, 1994a,b; Darienzo *et al.*, 1994; Nelson *et al.*, 1995, 1996a,c; Nelson and Personius, 1996; Atwater and Hemphill-Haley, 1997; Shennan *et al.*, 1997). Radiocarbon and dendrochronological techniques have been used to date fossil plants that were killed by submergence or tsunamis, or that started to grow soon after the earthquakes.

The best estimate of average recurrence of Cascadia subduction earthquakes, about 500 years, comes from a study of buried peats at estuaries in southwest Washington (Atwater and Hemphill-Haley, 1997). The buried peats in this area record seven earthquakes in the last 3500 years (Fig. 12). Five of the seven events have uncertainties of 20-380 years, based on high-precision (long count time) radiocarbon ages on plants that died, or began to grow, at about the time of the earthquakes. Age uncertainities for the remaining two events are larger.

Intervals between successive earthquakes differ considerably (Fig. 12; Atwater and Hemphill-Haley, 1997). Three earthquakes occurred between 3500-3320 and 2800-2400 years ago; their mean recurrence is thus more than 270 years but less than 550 years. An interval of 700-1300 years separated the youngest of these three events from the next earthquake, which dates to sometime between 1700 and 1500 years ago. Two earthquakes occurred 1350-1130 and 1300-900 years ago; after this, a period of 600-1000 years elapsed before the most recent earthquake 300 years ago.

DATING THE LAST EARTHQUAKE: PERSISTENCE PAYS

An extraordinary effort has been made to date the youngest subduction earthquake in Cascadia, in part to address the question of whether the uppermost peat-mud couplet in tidal marsh deposits is the product of a single, magnitude-9 earthquake, or two or more smaller, magnitude-8 quakes that were minutes, days, months or even years apart. The latter scenario would suggest that the subduction zone is segmented and possibly incapable of rupturing along its full 900-km length during a single earthquake, an important issue for earthquake risk assessment.

Conventional radiocarbon dating gives ages with uncertainties in excess of 40 C^{14} years, and these uncertainties become larger upon calibration. A clever approach that was used to reduce the uncertainty in the age of the last Cascadia earthquake was to date different blocks of annual rings from

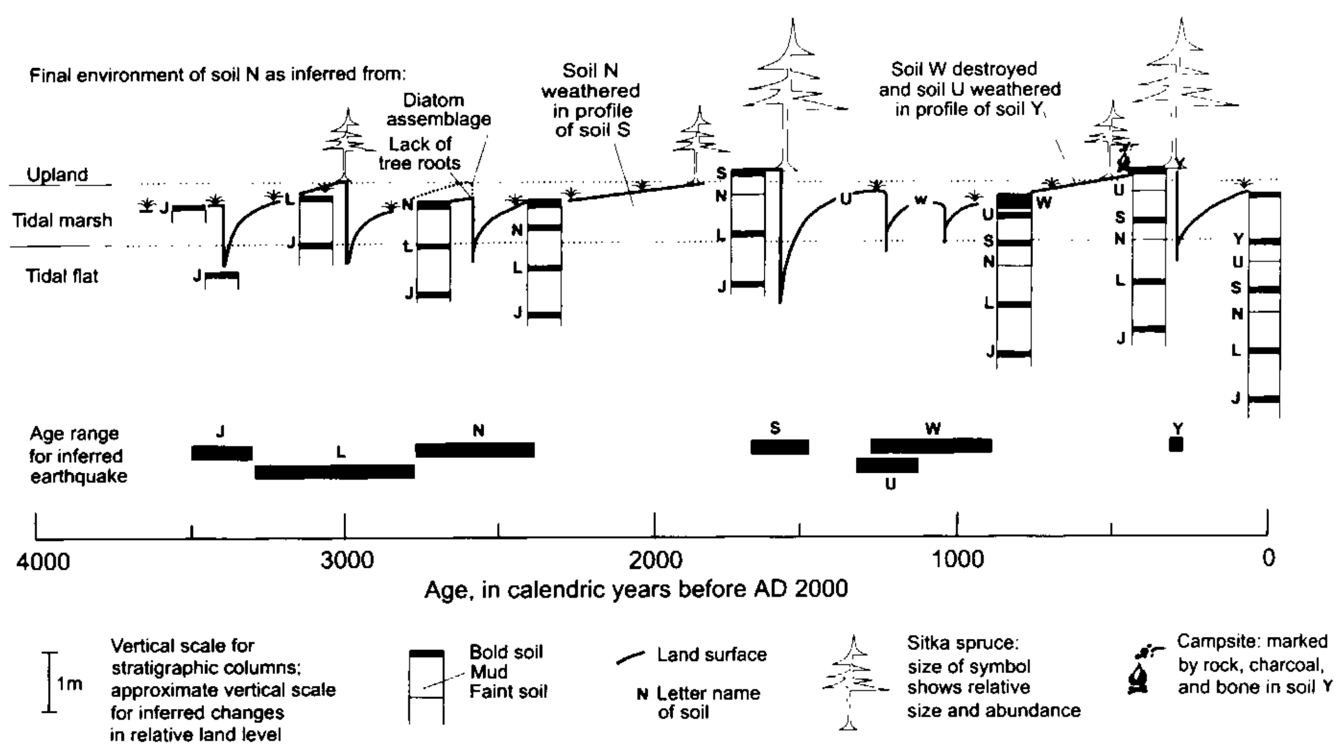

Figure 12 Representation of land-level, vegetation, and soil changes at an estuary in southwest Washington where there is evidence for seven large earthquakes in the last 3500 years. Each buried soil, shown as a lettered horizontal line in the stratigraphic columns, records an earthquake that caused at least 0.5 m of subsidence. Soils are separated by muds deposited between earthquakes. The age ranges of the earthquakes are based on radiocarbon ages. (Modified from Atwater and Hemphill-Haley, 1996, fig. 15).

earthquake-killed trees. High-precision radiocarbon ages on three different blocks of rings from a stump at a site in northern California limit tree death there to about AD 1700-1720 (Nelson *et al.*, 1995); the three radiocarbon ages eliminate other possible dates by matching distinctive steps (wiggles) in the calibration curve relating radiocarbon time to calendar time (Fig. 13). Similar wiggle matching for a stump in Washington limits tree death there to between AD 1695 and 1710 (Atwater *et al.*, 1991). These dates are consistent with age estimates obtained by high-precision radiocarbon dating of large numbers of earthquake-killed herbaceous plants rooted in the uppermost buried peat at widely separated sites in Cascadia (Nelson *et al.*, 1995). They are also consistent with a date of about AD 1690 obtained by dendrochronological techniques, *i.e.*, by matching the outermost preserved rings of standing red cedars that were killed by the earthquake with rings of old living trees (Atwater and Yamaguchi, 1991). The latter date, however, is inexact because the bark and some outer rings of the dead trees have been lost due to weathering and decay since the earthquake.

Efforts to date the last great earthquake in Cascadia came to a successful conclusion 2 years ago when two research teams, using different approaches, both demonstrated that the quake happened in AD 1700. One of the teams, led by Kenji Satake, found written records of damaging waves at several sites along the Japanese coast in January 1700 (Satake *et al.*, 1996). The waves were not the result of a local earthquake or storm, and there are no historical records or paleoseismic evidence for a large earthquake in 1700 in South America, Alaska or Kamchatka. Satake argued, by the process of elimination, that the tsunami was triggered by a giant earthquake at the Cascadia subduction zone. He corrected for the time it would take for the waves to travel to Japan and concluded that the earthquake happened on 26 January 1700, at about 9 p.m. Pacific Standard Time: the precision obviously exceeds that which is possible through radiocarbon dating!

A quite different approach was taken by dendrochronologists David Yamaguchi, Gordon Jacoby, and several of their co-workers. Confronted with the problem that weathering has removed the outermost rings of standing, earthquake-killed trees, Yamaguchi *et al.* (1997a,b) studied the roots of these trees (roots are better preserved than stems because they are entombed in mud). Jacoby *et al.* (1997), on the other hand, examined the rings of "survivor trees," trees that were damaged, but not killed by the earthquake. The dead and survivor trees gave the same result: the trees died or were damaged shortly after the 1699 growing season.

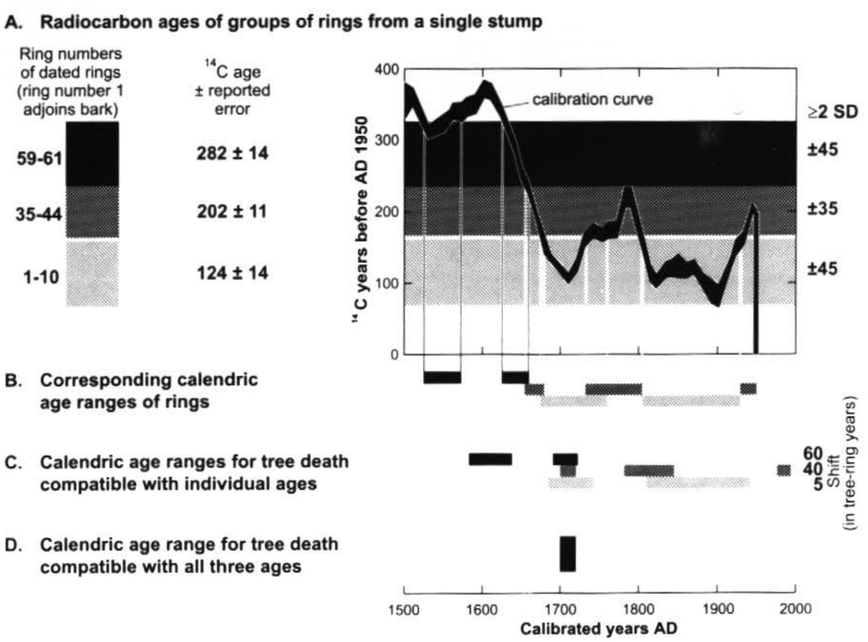

Figure 13 An example of how radiocarbon dating has been used to constrain age estimates for the last Cascadia plate-boundary earthquake. (**A**) High-precision radiocarbon ages were obtained on three different groups of rings cut from a single stump rooted in the youngest buried soil at a tidal wetland in northern California. The radiocarbon ages intersect different parts of the ^{14}C calibration curve, which relates radiocarbon age to calendar time (Stuiver and Becker, 1993). Two-standard-deviation uncertainties for each age are shown at the right edge of the diagram. (**B**) Horizontal bars showing the corresponding calendar age ranges, shaded as in Figure 13a. (**C**) Ranges in Figure 13b have been shifted to the right by the number of years between the dated rings and the bark. (**D**) The time at which the dated tree was killed by submergence during an earthquake is limited to the narrow range AD 1700-1720, where the ranges for all three ages overlap. (Modified from Nelson *et al.*, 1995, fig. 2; reproduced with permission from Nature; copyright 1995 Macmillan Magazines Limited.)

CONTRIBUTION TO SEISMIC RISK ASSESSMENT IN CASCADIA

Geologists and geophysicists have demonstrated that 1) great earthquakes have occurred in Cascadia in the recent geological past (on average once every 500 years during the last 3500 years), 2) the last earthquake took place in AD 1700 and probably was a magnitude-9 event, and 3) strain is presently accumulating that will be released in a future great earthquake.

Prior to the 1980s, United States and Canadian building codes did not consider subduction earthquakes. Seismic design was based on the instrumented record of earthquakes: local, crustal and subcrustal earthquakes smaller than moment magnitude 7.5. However, subduction earthquakes will be considered in the next version of the National Building Code of Canada, scheduled for release in the year 2001. A deterministic analysis done for this purpose suggests that wave velocities and accelerations at major cities in the Pacific Northwest during a great subduction earthquake will be no larger than those of a large, local, crustal earthquake (Adams *et al.*, 1996). Seismic energy will attenuate considerably over the 100-150 km between the source of a subduction earthquake and the largest cities in the region (Vancouver, Seattle and Portland). Strong shaking, however, would last much longer during a subduction earthquake than during a large crustal earthquake: several minutes as opposed to perhaps a few tens of seconds. Finally, subduction and crustal earthquakes would have different wave characteristics at our coastal cities. Relatively long-period, low-amplitude seismic waves

Table 1 Comparison of average recurrence and affected area for different earthquakes in Cascadia.

Type and size of earthquake (moment magnitude)	**Average recurrence**[1] (years)	**Area of damage**[1] (km^2)
Subduction (M_w 8-9+)	500	100,000
Crustal/subcrustal[2] (M_w 7-7.5)	30-40	20,000

[1]Values are approximate.
[2]A crustal earthquake has a hypocentre within the North America plate; a subcrustal earthquake occurs within the subducting Juan de Fuca Plate.

would dominate the energy spectrum of a subduction event at epicentral distances of 100-150 km, whereas much shorter-period, high-frequency waves would characterize large crustal events at short epicentral distances. Low-frequency waves pose a significant hazard to certain tall buildings and linear structures such as bridges and tunnels.

An important contribution of the geological work has been to provide a better understanding of some of the damaging effects of a future great earthquake in Cascadia. One of the consequences of a subduction earthquake is that the outer coasts of Vancouver Island, Washington and Oregon would probably subside up to 2 m. Such subsidence would inundate low-lying areas and could seriously damage development along the shoreline.

The tsunami of a subduction earthquake would damage towns and cities around the North Pacific Ocean; the highest run-ups would be on Vancouver Island and the Pacific coasts of Washington and Oregon. Some of the communities at greatest risk are those at the heads of inlets on Vancouver Island that open westward to the Pacific Ocean, for example Port Alberni and Zeballos (Clague *et al.*, 1994). On the other hand, no tsunami deposits have been found along the shores of the Strait of Georgia. It thus appears that Vancouver, Nanaimo and other cities along this inland sea are not themselves at risk from Cascadia tsunamis.

Liquefaction will occur in granular, water-saturated sediments over a large area during a subduction earthquake, although it is not yet known whether shaking will be strong enough at Vancouver and Seattle to induce liquefaction there. Although sand dykes and blows are present near Vancouver, they may have formed during a strong local earthquake (Clague *et al.*, 1992, 1997).

The phenomena discussed above, and others such as landslides, are not limited to subduction earthquakes: they can accompany large crustal earthquakes and earthquakes within the subducting Juan de Fuca plate. What sets a subduction earthquake apart is the possible large area of damage (of the order of 100,000 km^2 as opposed to perhaps 10,000 km^2 for a magnitude 7 quake) and long duration of shaking. A subduction earthquake might damage cities throughout the Pacific Northwest, from Portland to Vancouver, whereas a large crustal earthquake would affect a much smaller area (Table 1). The total damage caused by a large crustal earthquake would depend greatly on its location. An earthquake centred in a sparsely populated area such as west-central Vancouver Island would cause far less damage than one close to a large city. Although the total economic losses from a magnitude-9 earthquake would probably be huge, the geological record indicates that these events are 25-40 times less common than moderate to large (magnitude 6-7) crustal and subcrustal earthquakes in the region (Table 1). In the worst-case scenario, a magnitude-7 crustal earthquake close to Vancouver, Seattle or Portland might cause damage comparable to that of the 1995 Kobe earthquake ($100 billion, 1998 dollars). In view of this possibility, we would be well advised to be at least as concerned about local, crustal and subcrustal earthquakes as larger, but much rarer subduction events.

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