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Great earthquakes of variable magnitude at the Cascadia subduction zone

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Abstract

Comparison of histories of great earthquakes and accompanying tsunamis at eight coastal sites suggests plate-boundary ruptures of varying length, implying great earthquakes of variable magnitude at the Cascadia subduction zone. Inference of rupture length relies on degree of overlap on radiocarbon age ranges for earthquakes and tsunamis, and relative amounts of coseismic subsidence and heights of tsunamis. Written records of a tsunami in Japan provide the most conclusive evidence for rupture of much of the plate boundary during the earthquake of 26 January 1700. Cascadia stratigraphic evidence dating from about 1600 cal yr B.P., similar to that for the 1700 earthquake, implies a similarly long rupture with substantial subsidence and a high tsunami. Correlations are consistent with other long ruptures about 1350 cal yr B.P., 2500 cal yr B.P., 3400 cal yr B.P., 3800 cal yr B.P., 4400 cal yr B.P., and 4900 cal yr B.P. A rupture about 700–1100 cal yr B.P. was limited to the northern and central parts of the subduction zone, and a northern rupture about 2900 cal yr B.P. may have been similarly limited. Times of probable short ruptures in southern Cascadia include about 1100 cal yr B.P., 1700 cal yr B.P., 3200 cal yr B.P., 4200 cal yr B.P., 4600 cal yr B.P., and 4700 cal yr B.P. Rupture patterns suggest that the plate boundary in northern Cascadia usually breaks in long ruptures during the greatest earthquakes. Ruptures in southernmost Cascadia vary in length and recurrence intervals more than ruptures in northern Cascadia.
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Introduction

In the late twentieth century, coastal paleoseismology changed the perception of the earthquake hazard posed by the Cascadia subduction zone in coastal western North America (Fig. 1; Atwater, 1987; Clague, 1997; Yeats, 1998). In the early 1980s, the lack of historical earthquakes on the boundary between the Juan de Fuca and North America plates was attributed to smooth plate subduction, whereas by the late 1990s, a consensus had been reached that a locked plate boundary slips during great (magnitude 8 to 9) earthquakes that recur with a frequency of hundreds of years (Clague et al., 2000a). Subsided tidal wetlands capped by tsunami-laid sand (Darienzo and Peterson, 1990; Jacoby et al., 1995; Nelson et al., 1996a; Shennan et al., 1996; Kelsey et al., 1998, 2002; Clague et al., 2000b; Witter et al., 2003; Atwater et al., 2005) are the most dramatic of an array of geologic and geophysical evidence (Hyndman and Wang, 1995; Wang et al., 2001; Satake et al., 2003) used to argue for repeated great earthquakes at Cascadia. Low-lying coastal lakes and lagoons also archive as much as 7000 yr of stratigraphic evidence of the local tsunamis that accompany great Cascadia earthquakes (Clague et al., 2000b; Garrison-Laney, 1998; Kelsey et al., 2005). Debate now centers on questions, critical for building code design and emergency planning (Charland and Priest, 1995; Wang and Clark, 1999; Peterson et al., 2000; Petersen et al., 2002; Frankel et al., 2002), that might be answered by studies of Holocene earthquake history. Chief among the questions are the extent of past great earthquake ruptures along the subduction zone (a proxy for magnitude), their distribution in time (recurrence), the height and inland extent of accompanying tsunamis, and the strength of inland ground shaking.

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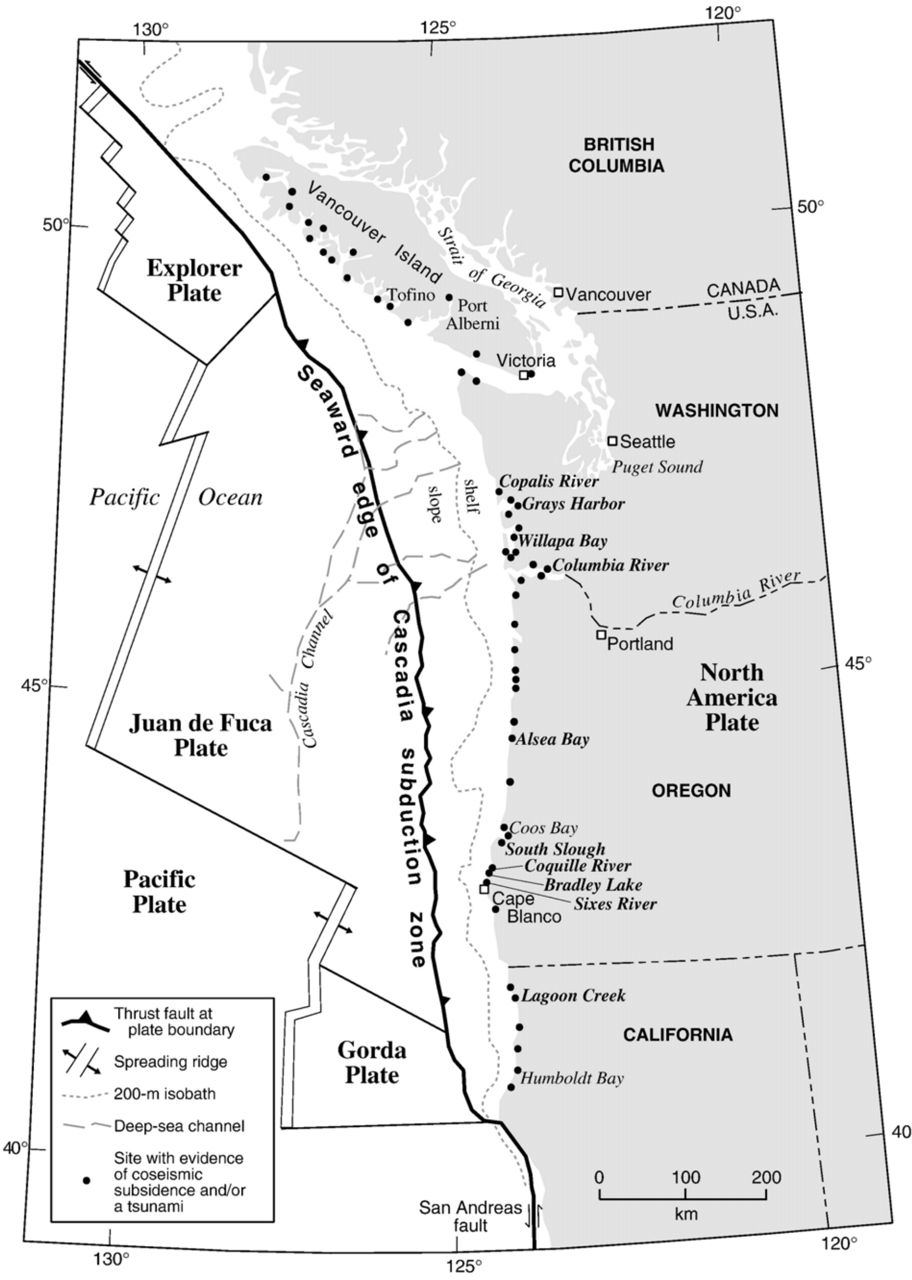
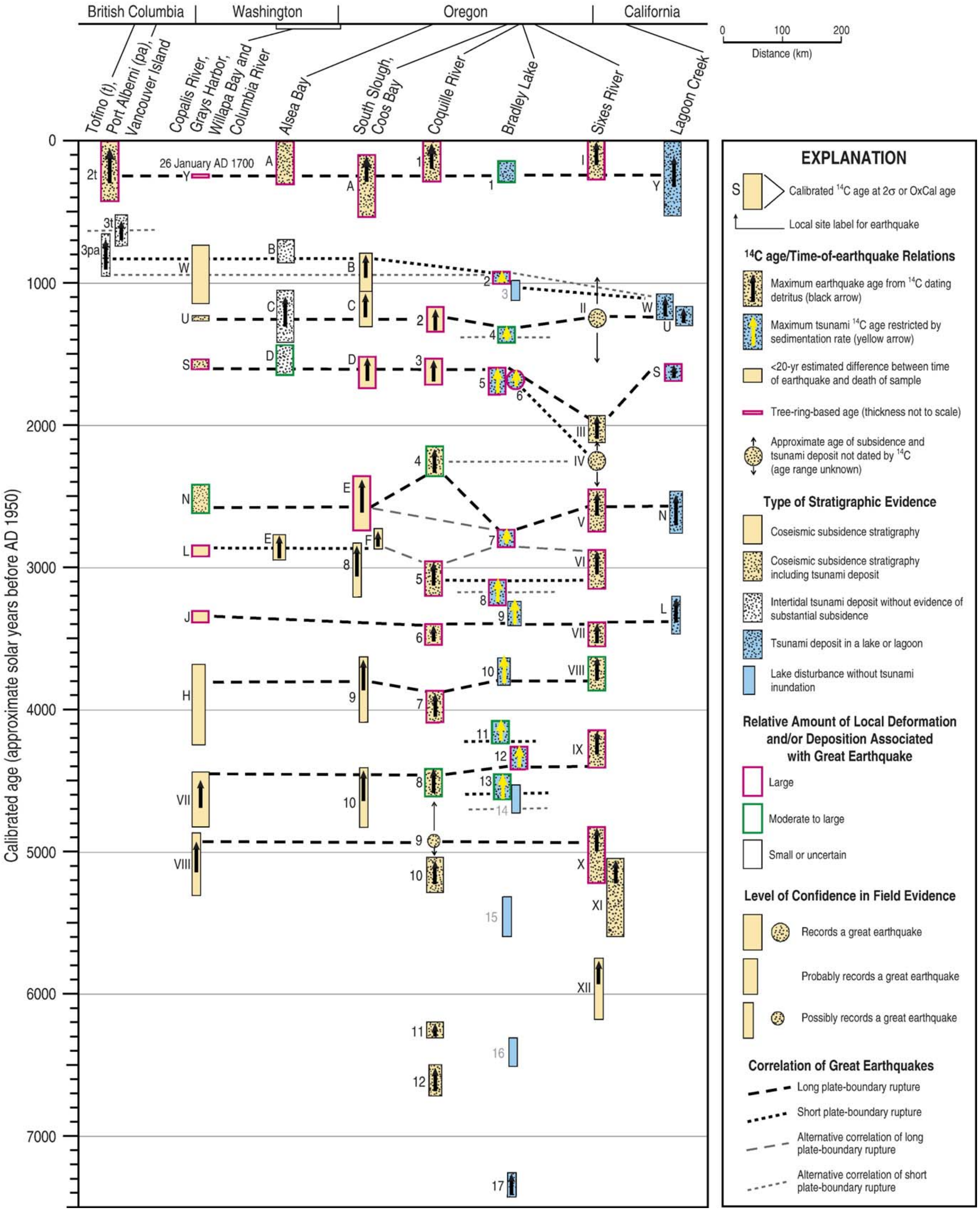


Figure 1. Location of coastal sites along the Cascadia subduction zone with evidence for great Cascadia earthquakes and accompanying tsunamis (after Atwater and Hemphill-Haley, 1997, their Fig. 1). Sites of Figure 2 with long earthquake histories, and with ages more precise than those from most sites, are labeled in bold italics.

But developing an earthquake history for Cascadia has proved difficult because the plate boundary lies 60–130 km offshore and all stratigraphic archives of great earthquakes are, at best, indirect measures of plate-boundary slip (Fig. 1; Nelson et al., 1996a; Leonard et al., 2004). Inferring even the relative amount of fault slip during individual earthquakes, and therefore earthquake magnitude, at Cascadia paleoseismic sites is problematic. Although quantitative analysis of tidal microfossil assemblages shows promise for measuring the amount of coseismic and interseismic deformation of the surface of the upper plate (e.g., Guilbault et al., 1995; Hughes et al., 2002; Sawai et al., 2004; Shennan and Hamilton, 2006; Edwards and Horton, 2006), even semi-quantitative microfossil analyses have been applied to no more than seven of more than 25 Cascadia estuaries. Only semi-quantitative methods with large uncertainties (e.g., Nelson et al., 1996b; Atwater and Hemphill-Haley, 1997; Kelsey et al., 2002; Witter et al., 2003) have been applied to evidence for more than four earthquakes at any site.

More than a decade of debate has not resolved the question of whether plate-boundary earthquakes at Cascadia are largely limited to the greatest earthquakes of magnitude 9 or include a mixture of magnitude-8 and magnitude-9 earthquakes (Clague, 1997). Magnitude-9 earthquakes rupture much of the plate boundary, whereas the segmented ruptures of magnitude-8 earthquakes are of more limited extent. The lack of magnitude-8 earthquakes during the historical period of the past 200 yr, combined with rates of plate convergence that require strain release during earthquakes on time frames of hundreds of years, would seem to favor the magnitude-9 hypothesis (Atwater and Hemphill-Haley, 1997). Although initially many questioned Cascadia's ability to produce magnitude-9 earthquakes (Heaton and Hartzell, 1987; McCaffrey and Goldfinger, 1995), by the



end of the millennium, researchers were suggesting that most great earthquakes recorded in coastal and offshore deposits were close to magnitude 9. Others argued, however, that because analogous subduction zones have a history of great earthquakes of variable magnitude and rupture length (Thatcher, 1990; Ruff, 1996; Nanayama et al., 2003; Cisternas et al., 2005), Cascadia probably does as well (McCaffrey and Goldfinger, 1995; Nelson and Personius, 1996; Kelsey et al., 2002; Witter et al., 2003).

Such variation in rupture length has been difficult to test at Cascadia, however, because great earthquakes there are too frequent and errors on typical radiocarbon samples too large to use ^{14}C-based correlations to rule out either long or short ruptures. But a reasonable inference from differences in the number of great earthquakes or local Cascadia tsunamis among the most carefully studied and best dated sites is that some great earthquake ruptures broke only parts of the subduction zone (Kelsey et al., 2002; Witter et al., 2003; Kelsey et al., 2005). Although most records show four great earthquakes in the past 2000 yr, the tsunami record at Bradley Lake in southern Oregon shows six (Figs. 1 and 2). Similarly, for the past 5000 yr, 12 tsunamis left a record at Bradley Lake, whereas wetland stratigraphy shows 9 or 10 times of coseismic subsidence at nearby tidal sites.

An alternative interpretation of differing numbers of earthquakes at different sites is that the thresholds for creating and preserving earthquake evidence differ from site to site, particularly among sites preserving different kinds of evidence. For example, if a tsunami accompanying a magnitude-8 earthquake in southern Oregon spread sand across the floor of Bradley Lake, the earthquake might produce too little subsidence for the sand to be preserved in nearby coastal marshes.

In this paper, we correlate the Bradley Lake record of tsunamis generated by great earthquakes with the most detailed tidal records of plate-boundary earthquakes, and with a record of tsunamis in a freshwater lagoon in northern California (Fig. 2). In our correlations, we consider possible differing thresholds for creating and preserving evidence of great earthquakes and accompanying tsunamis at the different sites, along with the errors and degree of overlap of ^{14}C-based age ranges for events at each site. To help choose among correlation alternatives, we infer the relative amount of coseismic subsidence during some earthquakes and the relative height of accompanying tsunamis.

Evidence thresholds

Geomorphic and stratigraphic process thresholds explain why evidence of catastrophic events is preserved at relatively few paleoseismic sites. McCalpin and Nelson (1996, p. 14) use the term “magnitude threshold of formation” for the earthquake magnitude required to create identifiable evidence of an earthquake in a particular stratigraphic or geomorphic setting. We distinguish two types of thresholds for earthquake evidence (such as landforms, stratigraphy, or fossil assemblage changes): creation thresholds and preservation thresholds. To exceed creation thresholds, evidence produced by surface deformation or ground shaking (and the erosional or depositional responses to them) must be distinct from similar evidence that might be produced by nonseismic processes in the same setting (Nelson et al., 1996a). In Figure 3, the greater changes in tidal environments measured in cores B and D would more firmly identify coseismic subsidence during great earthquakes than would the apparently smaller changes in other cores from the same site, which sample only adjacent tidal environments above and below the contacts penetrated by the cores (e.g., Nelson et al., 1996b; Witter et al., 2003). Non-uniform coseismic subsidence, differential sediment compaction, and pre- and (or) post-seismic subsidence or uplift may further complicate interpretations of the amount of coseismic subsidence in tidal wetlands. To exceed preservation thresholds, the balance among erosion, deposition, and other processes (such as bioturbation or soil development) at a site must favor preservation of the distinctive earthquake evidence. For example, along subduction-zone coasts that subside suddenly during great earthquakes, the preservation of sand sheets spread by tsunamis accompanying the earthquakes is ensured by quick burial with tidal mud following subsidence. But in tidal wetlands on non-subsiding coasts impacted by tsunamis, sand sheets are commonly removed during the highest tides or made unrecognizable by root stirring (Nelson et al., 1996a; Atwater and Hemphill-Haley, 1997; Clague et al., 2000b).

Interactions among erosional and depositional processes and site characteristics control the variability of creation and preservation thresholds among sites. Both types of thresholds also vary over time at particular sites because site characteristics change over time. For example, Kelsey et al. (2005) infer that westward shoreline progradation at Bradley Lake has increased the evidence-creation threshold for tsunamis accompanying subduction-zone earthquakes of a given magnitude over the past 2000 yr (e.g., Fig. 4). That is, tsunamis of the past 2000 yr have not produced as dramatic a change in lake stratigraphy as tsunamis of 4000–5000 yr ago, and this change is probably not due to differences in the magnitude of source earthquakes.

Interpretation of evidence thresholds at Cascadia sites where evidence is limited to tsunami deposits is complicated by possible non-plate-boundary sources for tsunamis—large

Figure 2. Correlation of great Cascadia earthquakes inferred from ages and evidence of sudden coastal subsidence and tsunamis. Leaders show position of sites along a line extending 1070 km from northern end of Juan de Fuca plate to southern end of Gorda plate (Fig. 1). Age-range rectangles for the eight youngest earthquakes in southern Washington and northernmost Oregon are based on 40 ^{14}C ages from 14 sites as much as 110 km apart (Atwater et al., 2004). Ages for the two oldest earthquakes in southern Washington are on peat (Shennan et al., 1996, their Table 1). Central and southern Oregon ages come from tidal marshes at Alsea Bay (Nelson et al., 2000), South Slough (Nelson et al., 1996b, 1998), the Coquille River (Witter et al., 2003), and the Sixes River (Kelsey et al., 2002; Fig. 1). Bradley Lake (Kelsey et al., 2005) and Lagoon Creek (Abramson, 1998; Garrison-Laney, 1998) are small coastal lakes in southern Oregon and northern California, respectively. Disturbance events 3, 14, 15, and 16 at Bradley Lake are labeled in gray because the strong ground shaking that they record may not coincide with great earthquakes (Kelsey et al., 2005).

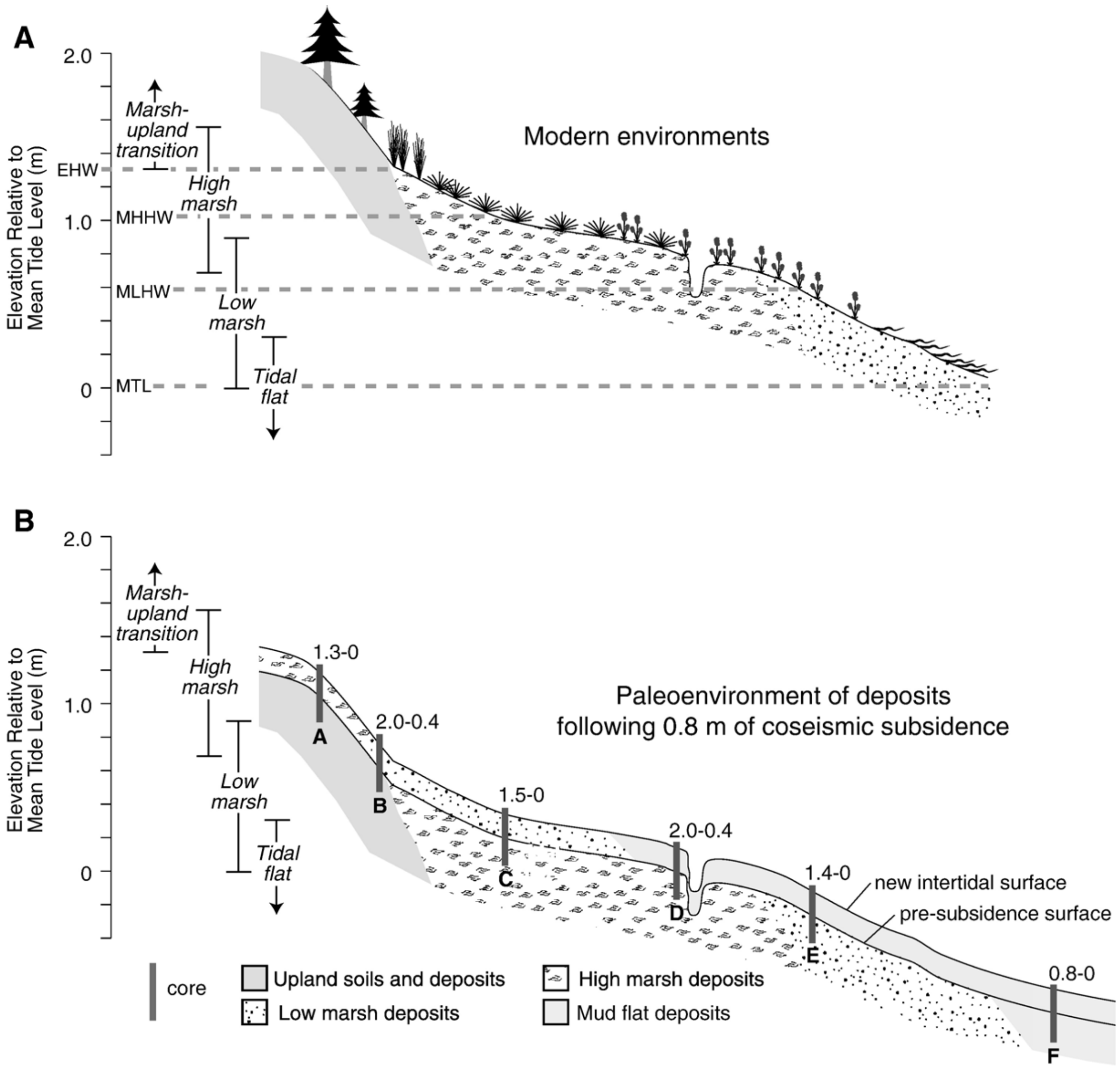


Figure 3. Illustration of uncertainties in estimating the amount of coseismic subsidence recorded in cores from a hypothetical Cascadia tidal marsh. A, elevation ranges of tidal environmental zones and tide levels. B, surface in A has been uniformly lowered 0.8 m as a result of subsidence during a great earthquake. Patterns show litho- and bio-facies for corresponding environmental zones before and after the earthquake. Numbers next to cores show range in amount of subsidence (meters) inferred from sudden change in fossils across the pre-subsidence surface.

submarine landslides and shallow faulting in the upper plate independent of plate-boundary earthquakes (Witter et al., 2003; Kelsey et al., 2005). Offshore landslides may produce very large tsunamis, and mapping along the continental slope shows topography characteristic of large slides (Goldfinger et al., 2000; McAdoo and Watts, 2004). Modeling of tsunamis generated by five types of shallow upper-plate faults at Cascadia shows that slip on seaward-vergent faults, especially those on the continental shelf, could generate tsunamis at least as high as those produced directly by rupture of the plate boundary (Geist and Yoshioka, 1996). Tidal wetland sites are unlikely to preserve evidence of tsunamis independent of coseismic subsidence, and the sources of tsunamis that deposit sand in lakes can only be inferred. For this reason, we follow earlier work (e.g., Atwater and Hemphill-Haley, 1997; Nelson et al., 1998; Kelsey et al., 2002, 2005; Witter et al., 2003) in inferring that offshore slides or upper-plate faulting large enough to produce distinctive tsunami deposits are coincident with great earthquakes.

Additional local and regional factors complicate distinguishing site-to-site differences in creation and preservation thresholds from differences in plate-boundary rupture lengths that are the result of earthquakes of differing magnitude. For example, the amount of coseismic subsidence at a tidal wetland depends on the depth and width of each plate-boundary rupture, distance from the plate boundary, and effect of possible localized upper-plate faulting (Hyndman and Wang, 1995). The height of tsunamis depends on even more factors than the amount of coseismic subsidence—the width and offshore distance of the rupture zone, size and shape of coseismic sea-floor deformation, bathymetry of the shelf and continental slope, tide level at tsunami landfall, wave setup and other meterologic effects,

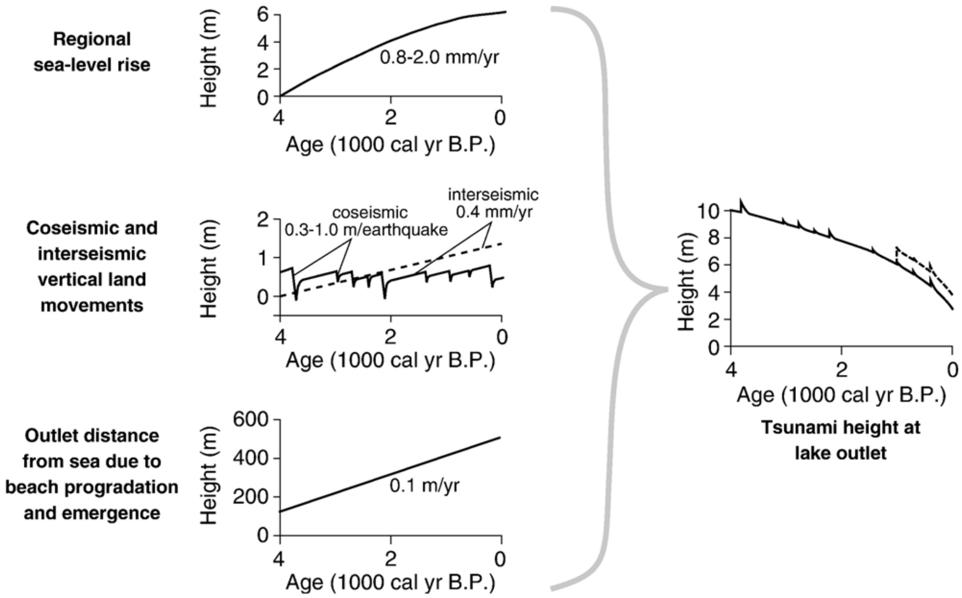


Figure 4. Decrease in tsunami height at seaward outlet of a hypothetical coastal lake in southern Oregon over the past 4000 yr for tsunamis of constant height accompanying Cascadia great earthquakes. Dotted line in right column figure shows the potential increase in height due to erosion of the outlet during inundation by a hypothetical tsunami about 1000 yr ago.

degree of coastal wave refraction, and especially nearshore bathymetry (Myers et al., 1999).

Despite the complicating factors and lack of quantitative data from most sites, we qualitatively infer the relative amount of subsidence or height of tsunamis during earthquakes at each site in Figure 2 to help decide among correlation alternatives. We justify these qualitative inferences about subsidence and tsunamis by assuming that earthquakes of greater magnitude produce greater amounts of coseismic subsidence and higher tsunamis than earthquakes of lesser magnitude. Based on differences in the characteristics of various kinds of earthquake and tsunami evidence (e.g., Atwater et al., 1995; Hemphill-Haley, 1995; Nelson et al., 1996a; Shennan et al., 1996; Peterson and Darienzo, 1996; Atwater and Hemphill-Haley, 1997; Hughes et al., 2002; Kelsey et al., 2002; Witter et al., 2001; Leonard et al., 2004; and Kelsey et al., 2005), we distinguish earthquakes with the greatest subsidence and highest tsunamis from those with moderate to large subsidence and tsunamis, and from events with smaller effects or whose relative size is difficult to infer from the literature. We have no way of accurately scaling the relative size of subsidence and tsunamis among sites.

Limitations of ^{14}C correlation

As in many other fault-parallel correlation figures, we use rectangles to show error ranges for calibrated ^{14}C ages (2σ errors; most are calibrated from means of multiple ages) thought to date great earthquakes inferred from evidence in southern Washington, Oregon, and northern California (Fig. 2). Dashed lines of Figure 2 mark likely correlations and some alternative correlations of rectangles from which we infer a history of both long and short earthquake ruptures along the subduction zone. Evidence of short ruptures is strongest where rectangles overlap least with rectangles to the north or south.

Of course, radiocarbon's inability to distinguish evidence of earthquakes closely spaced in time is a widely recognized limitation of prehistoric earthquake correlation (e.g., Heaton and Hartzell, 1987; Nelson et al., 1995; Atwater and Hemphill-Haley, 1997). Even unusually precise ^{14}C dating could not determine whether or not the series of ruptures that occurred hours to a few years apart along the Nankai coast of southeastern Japan in 1854 were a series of ruptures of different plate-boundary segments or a single through-going rupture like the one that broke the same plate boundary in 1707 (Ando, 1975). Alaska, Colombia, Chile, and most recently the Sumatra–Andaman region offer other historical examples of long ruptures extending along several segments of subduction zones whose segments previously ruptured in separate great earthquakes (e.g., Thatcher, 1990; Cisternas et al., 2005).

Another limitation of all but ten of the ^{14}C-age-range rectangles in Figure 2 is that dated carbon came entirely from detritus older than the earthquake. Age-range rectangles for such detrital samples are marked by upward-pointing arrows to show that the time of an earthquake may postdate the upper end of the rectangle corresponding with its dated evidence. More than a decade of comparisons among ages on Cascadia detrital samples (Atwater, 1992; Nelson et al., 1994; Atwater and Hemphill-Haley, 1997) shows that at least three-quarters of carefully selected samples do not predate an earthquake by more than a few hundred years. But the many examples of age differences of many hundreds of years from the same bed—some greater than a millenium (e.g., Grant, 1989; Nelson, 1992a; Nelson and Personius, 1996; Kelsey et al., 2005)—temper our confidence in the accuracy of even the shortest arrowed rectangles (age ranges) of Figure 2. The six shortest rectangles from southern Washington and northernmost Oregon lack arrows because these most-precise ranges have been limited on both ends through analysis (with the statistical program OxCal; Bronk Ramsey, 2001) of ^{14}C ages of unusual precision on tree rings and other plant parts younger, as well as older, than the earthquakes (Atwater et al., 2004). Other unarrowed rectangles are limited by ages of less precision (H at Columbia River; B and D at Alsea Bay) or are based on the age of a shrub that probably died within 20 yr of the earthquake (W at Columbia River). In a similar analysis, Kelsey et al. (2005) restricted ranges of ages on samples from 12 tsunami deposits at Bradley Lake in southern Oregon with uniform lake sedimentation rates. Although dated materials are detrital, restrictions make the ranges for the lake's ages less open-ended, and therefore more precise, than detrital ages from tidal sites (ranges labeled with yellow arrows in Figure 2; e.g., Nelson et al., 1996b, 1998; Kelsey et al., 2005).

Ages not summarized in Figure 2 are available for similar evidence from many other sites from northern Vancouver Island to Humboldt Bay (Fig. 1; e.g., Atwater et al., 1995; Darienzo and Peterson, 1995; Clague et al., 2000b), but error ranges on ages from these other sites are mostly larger than those of Figure 2, and the materials dated are less certainly associated with earthquake evidence than are the materials used for Figure 2 ages. Like the tallest of the age-range rectangles in the South Slough and Sixes River columns, the large error ranges on ages from these other studies allow many alternative earthquake correlations. Continental slope turbidites produced by regional earthquake ground shaking have been correlated based on turbidite characteristics and ^{14}C-controlled sedimentation-rate ages, but the difficulties of interpreting ^{14}C ages on planktonic foraminfera beneath turbidite unconformities makes turbidite ages – and hence their correlation with coastal records of earthquakes – uncertain (Goldfinger et al., 2003).

Because of the above uncertainties in ^{14}C dating earthquakes and tsunamis, we consider inferences about the relative amount of coseismic subsidence and height of tsunamis, as well as the degree of overlap on age ranges, in correlating evidence from site to site. More quantitative, objective approaches to prehistoric earthquake correlation (e.g., Weldon et al., 2004), while a long-term goal, are difficult to apply in correlating Cascadia earthquakes because of the lack of quantitative—or in many cases even qualitative—information about the amount of coseismic deformation or strength of ground shaking at most sites.

Plate-boundary ruptures of variable length

Long and short ruptures of the past 2000 yr

The most secure of our site-to-site correlations of great earthquakes and tsunamis along the Cascadia subduction zone is of evidence dating from AD 1700 (250 cal yr B.P.). Stratigraphic evidence of substantial coastal subsidence and a high tsunami at sites spanning at least 900 km of the subduction zone (Atwater et al., 1995; Nelson et al., 1995; Clague et al., 2000b; Fig. 2) is consistent with an earthquake near magnitude 9. As with all our correlations, the alternative—of a series of much shorter ruptures during magnitude-8 earthquakes closely spaced in time — cannot be precluded, but unusually precise ^{14}C dating of the AD 1700 earthquake limits such a series to segmented ruptures over a period of less than two decades. A broad plateau in the radiocarbon calibration curve yields tall age-range rectangles (broad error ranges) for most ^{14}C ages near AD 1700; precise ^{14}C dating of older tree-ring wood yields very short ones in southern Washington (Atwater et al., 1991; Nelson et al., 1995). Dendrochronology (Jacoby et al., 1997; Yamaguchi et al., 1997) further limits the time of the earthquake to late 1699 or early 1700, in agreement with assignment of the earthquake to 26 January 1700, probably about 9 PM local time (Satake et al., 2003; Atwater et al., 2005). Changes in site conditions over time have resulted in stratigraphic evidence of this earthquake at some sites, such as Coos Bay and Bradley Lake (Fig. 1), being less distinct than similar evidence for several earlier, probably smaller earthquakes.

The other earthquake of the past two thousand years for which evidence of substantial subsidence and tsunamis is widespread and distinct dates from about 1600 cal yr B.P. Contrasts in lithology and fossils across the upper contact of the wetland soil buried by mud or sand following this earthquake suggest at least as much tidal subsidence during this earthquake as the evidence for any other earthquake in central and northern Oregon and southern Washington (Nelson et al., 1996a). Another reason that stratigraphic evidence for the 1600 cal yr B.P. earthquake is so distinct is that wetland soil horizons buried by the subsidence are commonly thicker and better developed than older or younger buried horizons. The distinctness reflects the long interval prior to the 1600 cal yr B.P. earthquake available for forest expansion and soil development, the longest interseismic interval of the past 5000 yr (Atwater and Hemphill-Haley, 1997; Fig. 2). The highest tsunamis recorded at Bradley Lake (tsunami 5) and Lagoon Creek (tsunami S) also date to about 1800–1600 cal yr B.P. (Abramson, 1998; Kelsey et al., 2005). If these tsunamis record the long rupture implied by our correlation of the 1600 cal yr B.P. earthquake, coseismic subsidence spanned at least 470 km of the subduction zone and a high tsunami inundated sites 620 km apart. Such characteristics are consistent with a magnitude-9 earthquake.

The apparent lack of a dated correlative of the 1600 cal yr B.P. earthquake at the Sixes River is striking (Witter et al., 2003; Fig. 2). Kelsey et al. (2002) attributed the absence of stratigraphic evidence for an earthquake about this time to a plate-boundary rupture of limited extent, arguing that the woody detritus, dated to about 2000 cal yr B.P., in the third oldest buried soil (III) at the Sixes River was probably too old to have been buried about the time of the third oldest earthquake in southern Washington and northernmost Oregon (earthquake U, Fig. 2). But if the soil was buried instead during the fourth oldest earthquake (earthquake S), only 400 yr younger than the dated detritus, it would predate earthquake S by no more than similar materials buried following subsidence from soils widely correlated with earthquake S at Coos Bay (Nelson, 1992b, his Fig. 4), Netarts Bay (Nelson et al., 1996a, their Fig. 5), and Willapa Bay (Atwater and Hemphill-Haley, 1997, p. 92).

Most sites in Figure 2 preserve evidence of at least two other earthquakes and (or) tsunamis between the earthquakes and tsunamis of AD 1700 and 1600 cal yr B.P. Wetland soils dating from about 700 to 1100 cal yr B.P. from Coos Bay into southern Washington are thin with upper contacts that are indistinct compared with similar soils buried following the earthquakes of AD 1700 and 1600 cal yr B.P. Contrasts in fossils and lithology across the upper contact of the soils are less distinct than those across the contacts of the other soils (e.g., Hemphill-Haley, 1995). The indistinct evidence probably reflects (1) less time for soil development since the previous earthquake than for the soil buried following the 1600 cal yr B.P. earthquake, (2) high rates of soil organic matter decomposition in soils at sites in southern Washington where the large tidal range allows greater soil weathering (Atwater and Hemphill-Haley, 1997), and (3) less coseismic subsidence during this earthquake than during earlier and later earthquakes (Nelson et al., 1998). Because the age of tsunami B, produced during the second youngest earthquake at Alsea Bay, is substantially younger than ages for subsidence south of Coos Bay, it probably records a rupture limited to the northern part of the subduction zone (Witter et al., 2003). Ranges for the subsidence of earthquake W at Columbia River and earthquake B at Coos Bay overlap with the range for tsunami B and so may be the same age. Alternatively, the earthquakes could predate tsunami B. In either case, small subsidence during earthquakes W and B is consistent with a short rupture in northern Cascadia. Sand beds from a tsunami accompanying an earthquake about this time (700–1100 cal yr B.P.) may be preserved as far north as southern Vancouver Island (tsunami 3pa) and as far south as Lagoon Creek: although tsunami W's rectangle at Lagoon Creek is at least 200 yr older than tsunami B at Alsea Bay, the twigs used to date tsunami W (Abramson, 1998) may have been reworked.

Field evidence and correlations suggest that plate-boundary rupture during an earthquake about 1350 cal yr B.P. was probably more extensive than during the earthquake about 700–1100 cal yr B.P., but perhaps less extensive than the rupture during the earthquake about 1600 cal yr B.P. Soils and contrasts in fossils and lithology recording the earthquake about 1350 cal yr B.P. are less distinct than the evidence for the earthquake about 1600 cal yr B.P. but commonly more distinct than evidence for the earthquake about 700–1100 cal yr B.P. Because the Lagoon Creek site is not capable of recording coseismic subsidence, we know only that this earthquake's rupture extended at least as far as the 440 km separating the Copalis and Coquille rivers (Fig. 2). If the undated second youngest wetland soil and tsunami sand bed at the Sixes River (labeled II) records subsidence and tsunamis during this earthquake, the rupture extended at least 470 km. Although age ranges for both tsunamis W and U at Lagoon Creek overlap the age range for earthquake U at Willapa Bay, correlation of tsunami U with earthquake U is most likely.

No age-range rectangle dating tsunami inundation at Bradley Lake overlaps with the age ranges for the earthquake and tsunami about 1350 cal yr B.P. in southern Washington and northernmost Oregon, or the potentially correlative tsunami at Lagoon Creek. Overlap would occur, however, if the rectangle for tsunami 4 at Bradley Lake shifted upward by only half a century, as suggested by the age for this event derived from lake sedimentation rates (Kelsey et al., 2005, their Fig. 14). At least some of the dated plant fragments that constrain tsunami ages at Bradley Lake probably were reworked from shallow-water deposits and so could easily date from a century or two prior to the tsunamis that deposited them in the lake (Kelsey et al., 2005). Because the age ranges for adjacent Bradley Lake tsunamis are constrained by sedimentation rates, shifting the tsunami-4 rectangle up, for example by a century, would require shifting age-range rectangles for tsunamis 2, 5 and 6 up by a similar number of years. An upward shift of rectangles for tsunamis 2, 5, and 6 would also increase the overlap of tsunami 5 at Bradley Lake with earthquake S's rectangle in southern Washington and northernmost Oregon as well as create an age overlap of the rectangles for tsunami 2 at Bradley Lake and tsunami B at Alsea Bay. Such upward shifts in Bradley Lake rectangles would strengthen our correlation of these events.

The range of disturbance event 3 at Bradley Lake, dated at about 1100 cal yr B.P., overlaps with ranges for earthquake W at Columbia River, earthquake B at Coos Bay, and tsunami W at Lagoon Creek. Event-3 beds at Bradley Lake probably record ground shaking without tsunami inundation (Kelsey et al., 2005). Thus, event 3 might record earthquake shaking at the time of subsidence in southern Washington and northernmost Oregon (earthquake W), and at Coos Bay (earthquake B), with a tsunami too low to enter Bradley Lake. But a tsunami from the earthquake might have entered Lagoon Creek (tsunami W) because the berm at the seaward end of that site's lagoon is quite low. Whether or not earthquake W in Washington and Oregon and tsunami W at Lagoon Creek correlate, an earthquake recorded by such a limited amount of evidence at Bradley Lake is unlikely to have ruptured the southern part of the subduction zone.

Ruptures 2000–5000 yr ago

Correlation of evidence for great earthquakes and accompanying tsunamis dating from 2000 to 5000 cal yr B.P. is less certain than for younger events because stratigraphic evidence of subsidence and tsunamis is less accessible, less well preserved, preserved at fewer sites, or dated with fewer ^{14}C ages. A further complication is that events during this period are so frequent that the lengths of recurrence intervals between events approach the lengths of earthquake age ranges (Fig. 2).

The rupture extent of a great plate-boundary earthquake about 2500 cal yr B.P. is uncertain. We prefer a long-rupture correlation of earthquake N in southern Washington and northernmost Oregon with tsunami N at Lagoon Creek as shown on Figure 2, but alternative correlations are equally plausible. Although the age-range rectangle for subsidence and tsunami about 2300 cal yr B.P. (event 4) at the Coquille River lies above the rectangle for earthquake N, the Coquille rectangle is based on a single age and so is less reliable than most of the other Coquille rectangles, which are based on multiple ages (Witter et al., 2003).

Choosing among alternative correlations for tsunamis 7 and 8 at Bradley Lake is difficult because the evidence for earthquakes and tsunamis about this time does not suggest any significant differences in the amounts of subsidence or height of tsunamis. Shifting the age range for tsunami 7 upward (to support our correlation with earthquake N) requires shifting the age ranges for tsunamis 8 and 9 by a similar amount because the ranges are linked by lake sedimentation rates (Kelsey et al., 2005). Such a shift would provide almost complete overlap of tsunami 8's rectangle with those of tsunami and subsidence events 5 at the Coquille River and VI at the Sixes River, but no overlap with earthquake L's short rectangle in southern Washington and northernmost Oregon. Thus, it seems unlikely that tsunami 8 at Bradley Lake occurred at the same time as earthquake L; a more likely alternative is that tsunami 8 and events 5 and VI at the Coquille and Sixes rivers, respectively, record a plate-boundary rupture in southern Cascadia that did not reach southern Washington (Witter et al., 2003).

Age-range rectangles for subsidence and tsunami events 6 at Coquille River and VII at Sixes River do not quite overlap with earthquake J in southern Washington. Because these were large events with evidence as distinct as that for younger, long plate-boundary ruptures, we correlate the events with earthquake J in southern Washington, tsunami 9 at Bradley Lake, and tsunami L at Lagoon Creek. This correlation implies both that the rupture during an earthquake about 3400 cal yr B.P. was long, and that the rupture that produced tsunami 8 at Bradley Lake was short. The short rupture may have resulted from an earthquake in southern Cascadia, either one that produced only a tsunami at Bradley Lake, or a somewhat longer rupture that produced subsidence at the Coquille River (event 5) and Sixes River (VI) and tsunamis at all three sites.

Correlation of evidence for great earthquakes older than 3500 cal yr B.P. relies mostly on the degree of overlap of event age ranges because information about the relative amount of subsidence, distinctness of soils, or heights of tsunamis of that age is available only from three sites in southern Oregon. The tall rectangles showing large age ranges for subsidence about 3800 cal yr B.P. in southern Washington and at Coos Bay overlap the rectangles for probable correlative events at the three sites in southern Oregon with stratigraphic records extending that far back in time. Although the rectangle for event 7 at the Coquille River does not overlap tsunami 10's rectangle at Bradley Lake, we follow Witter et al. (2003) in assuming that Coquille event 7's age range is probably a maximum age range by at least half a century. Such a correlation suggests a rupture at least 470 km long. We make a similar correlation of rectangles at the same sites for evidence of an earthquake and tsunami about 4400 cal yr B.P. Substantial subsidence and high tsunamis at the Sixes River and Bradley Lake about this time are consistent with the extensive plate-boundary rupture implied by the correlation. Tsunamis 11, 13, and disturbance event 14 at Bradley Lake lack obvious correlatives at other sites and so may record tsunamis and shaking from earthquakes below the threshold of evidence creation and/or preservation at other sites. The amount of subsidence, distinctness of the buried soil, and thickness of tsunami sand for event X at the Sixes River is similar to those for later large events. For this reason, we suggest a possible correlation with events at the Coquille River and in southern Washington whose rectangles overlap with event X's. Beyond 5000 yr B.P., too little evidence has been identified and dated to infer much about the extent of plate-boundary ruptures.

Summary of great earthquake history

Eight of the 10 great earthquakes inferred from evidence in southern Washington and northernmost Oregon by Atwater (1992), Atwater and Hemphill-Haley (1997), Shennan et al. (1996), and Atwater et al. (2004) can be correlated with other evidence of great earthquakes and accompanying tsunamis along at least 460 km of the subduction zone. Coseismic subsidence during five of the 10 earthquakes can be correlated for at least 470 km and their high tsunamis for at least 620 km—distances consistent with plate-boundary earthquakes near magnitude 9. Evidence of subsidence during an earthquake about 700–1100 cal yr B.P. has only been correlated over a distance of 420 km. A plate-boundary rupture about that time was apparently limited to the northern and central parts of the subduction zone, but its tsunami may have inundated Bradley Lake and Lagoon Creek.

Although alternative long-rupture or short-rupture correlations are plausible for at least three of the earthquakes recorded in southern Cascadia, seven tsunamis, subsidence, or lake disturbance events are recorded in southern Oregon in the past 5000 yr that lack clear correlatives in southern Washington and northernmost Oregon. The greater number of tsunamis at Bradley Lake (12) compared with the maximum number of events at other sites (10) implies that two Bradley Lake tsunamis lack correlatives. Some of the 12 events probably record short plate-boundary ruptures during magnitude-8 earthquakes that did not rupture into Washington (Kelsey et al., 2002; Witter et al., 2003), whereas others are recorded at sites, such as Bradley Lake, with apparently lower thresholds for the creation and preservation of great earthquake evidence than other sites.

Perhaps the strongest evidence for short, closely spaced ruptures, like those we infer to be more common along the southern part of the subduction zone, are lake deposits between the deposits of tsunamis 5 and 6 at Bradley Lake (Fig. 5; Kelsey et al., 2005). Photographs taken shortly after splitting core M show at least 22 poorly preserved, light–dark laminae couplets, which Kelsey et al. (2005) infer to be annual varves, between the distinctive lithofacies of tsunamis 5 and 6. Even allowing for modest (<10 mm) unrecognized erosion of the couplets in this most protected part of the lake, ^{14}C-controlled sedimentation rates (0.5–0.8 mm/yr) suggest that the tsunami deposits were laid down less than 40 yr apart. Such close timing of tsunamis might be explained by two plate-boundary ruptures located largely north and south of Bradley Lake, respectively. Peterson and Darienzo (1996, their Fig. 57) and others argue that the coincidence of tsunami beds and subsided wetland soils at most tidal sites is evidence for extensive rather than segmented plate-boundary ruptures, but a few decades is probably too short a time for deposition of tidal mud to separate tsunami beds from underlying subsided soils at many estuarine sites (e.g., Atwater

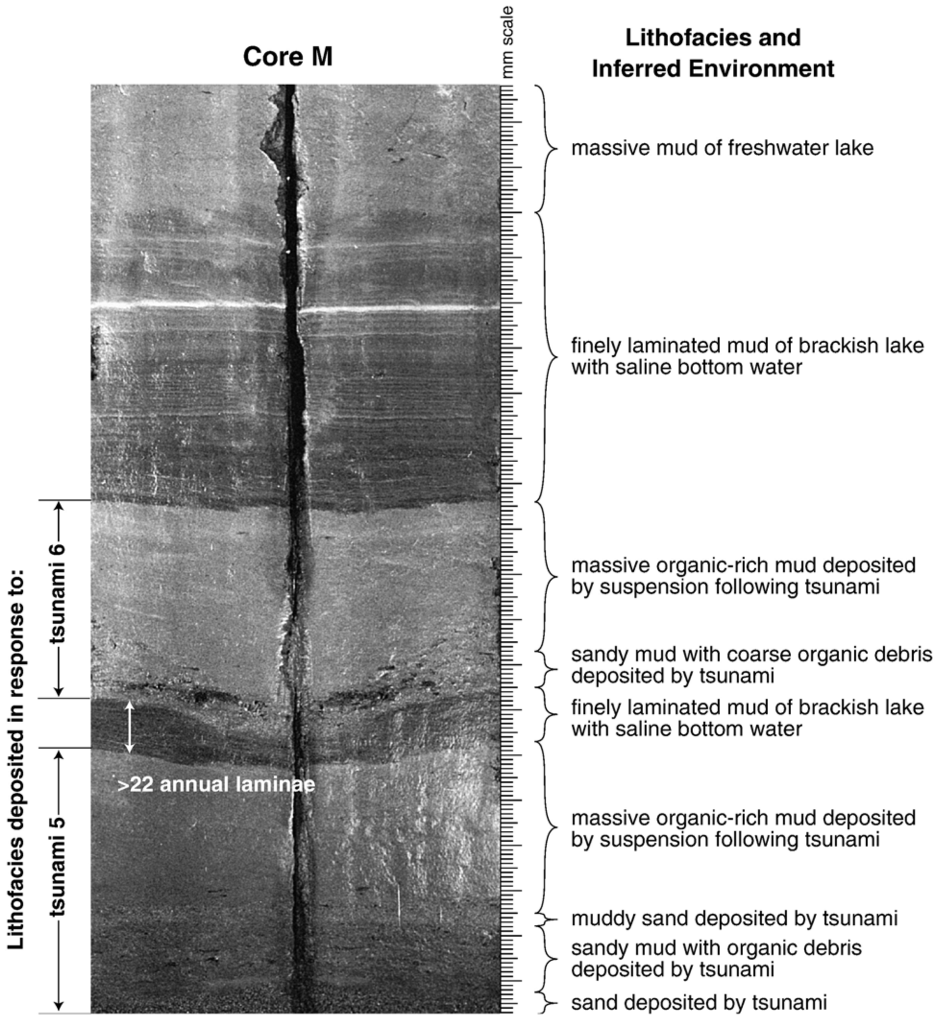


Figure 5. At least 22 light–dark laminae couplets between tsunami deposits 5 and 6 at Bradley Lake suggest great earthquake ruptures spaced decades apart off the southern Oregon coast (Figs. 1 and 2). Core M is a piston core from the landward end of the lake about 1 km east of the sea (Kelsey et al., 2005).

and Hemphill-Haley, 1997; Atwater et al., 2001; Witter et al., 2003).

Conclusions

Age ranges for evidence of great plate-boundary earthquakes and accompanying tsunamis at eight Cascadia paleoseismic sites, combined with estimates of the relative amounts of subsidence and heights of tsunamis, suggest a 5000-yr history of earthquakes of variable magnitude. Written records of a tsunami in Japan provide the most conclusive evidence for rupture of much of the plate boundary during the earthquake of 26 January 1700. Stratigraphic evidence dated at about 1600 cal yr B.P., similar to that for the 1700 earthquake, implies a similarly long rupture with substantial subsidence and a high tsunami. Correlations are consistent with other long ruptures of at least 460 km about 1350 cal yr B.P., 2500 cal yr B.P., 3400 cal yr B.P., 3800 cal yr B.P., 4400 cal yr B.P., and 4900 cal yr B.P. One or two other earthquakes may have ruptured much of the plate boundary in the past 5000 yr. A rupture about 700–1100 cal yr B.P. was limited to the northern and central parts of the subduction zone, and a northern rupture about 2900 cal yr B.P. may have been similarly limited. Times of probable short ruptures in southern Cascadia include about 1100 cal yr B.P., 1700 cal yr B.P., 3200 cal yr B.P., 4200 cal yr B.P., 4600 cal yr B.P., and 4700 cal yr B.P.

Although evidence is insufficient to distinguish short from long ruptures for most of the earthquakes and tsunamis identified between 2000 and 5000 yr ago, rupture patterns of Figure 2 suggest that the plate boundary in northern Cascadia commonly breaks in long ruptures during the largest earthquakes. In contrast, southernmost Cascadia is typified by short as well as long ruptures during great earthquakes of variable magnitude. The variable magnitudes may result in a shorter and more variable great earthquake recurrence in southern versus northern Cascadia.

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