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Earthquake Hazards in the Pacific Northwest of the United States

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THE POTENTIAL FOR GREAT EARTHQUAKES IN OREGON AND WASHINGTON: AN OVERVIEW OF RECENT COASTAL GEOLOGIC STUDIES AND THEIR BEARING ON SEGMENTATION OF HOLOCENE RUPTURES, CENTRAL CASCADIA SUBDUCTION ZONE

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Foreword

This paper is one of a series dealing with earthquake hazards of the Pacific Northwest, primarily in western Oregon and western Washington. This research represents the efforts of U.S. Geological Survey, university, and industry scientists in response to the Survey initiatives under the National Earthquake Hazards Reduction Program. Subject to Director's approval, these papers will appear collectively as U.S. Geological Survey Professional Paper 1560, tentatively titled "Assessing Earthquake Hazards and Reducing Risk in the Pacific Northwest." The U.S. Geological Survey Open-File series will serve as a preprint for the Professional Paper chapters that the editors and authors believe require early release. A single Open-File will also be published that includes only the abstracts of those papers not included in the pre-release. The papers to be included in the Professional Paper are:

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ABSTRACT

Fundamental questions in earthquake hazards research in the Pacific Northwest are concerned with the magnitude and recurrence of great earthquakes in the Cascadia subduction zone (CSZ). Geologic work of the last few years has produced convincing evidence of coseismic subsidence along the Washington and Oregon coasts. Regional subsidence recorded by estuarine deposits suggests that plate-interface earthquakes of at least M_w 8 (>100-km-long ruptures) occurred during the late Holocene in northern Oregon and southern Washington. Differences in the types of coastal marsh sequences between northern and south-central Oregon, however, suggest that the Oregon coastline did not subside due to regional deformation south of about 44° N. North of this latitude, the coast may intersect the seaward edge of a zone of coseismic subsidence that may continue southeastward onshore. Alternatively, the CSZ may be segmented near 44-45° N; a segment boundary at this location would indicate that plate-interface events near M_w 8 along the central CSZ are more frequent than larger (M_w 9) events. South of this boundary in the Coos Bay region, the tectonic framework developed through mapping and dating of marine and fluvial terraces indicates that episodes of abrupt marsh burial in south-central Oregon are best interpreted as the product of deformation on local structures. Some of the local deformation could be associated with moderate earthquakes (M_s<6). At most sites in south-central Oregon, however, it is still unclear whether the observed deformation is related to local faulting and folding during moderate-magnitude earthquakes, to regional deformation during great plate-interface earthquakes, or to both.

INTRODUCTION

One of the fundamental questions in earthquake hazards research in the Pacific Northwest is the potential for great earthquakes in the Cascadia subduction zone (CSZ) (Heaton and Hartzell, 1987; Atwater, 1987) (fig. 1). The present seismic quiescence in the region, initially attributed to aseismic subduction (Ando and Balazs, 1979; Acharya, 1981; Reilinger and Adams, 1982), is increasingly regarded as evidence for a locked plate interface where accumulating stress will result in a great earthquake (Savage and others, 1981; Heaton and Kanamori, 1984; Adams, 1984; Heaton and Hartzell, 1986; Rogers, 1988; Spence, 1989). Recent geologic field studies support the locked plate-interface model (Atwater, 1987; Kelsey and Carver, 1988; Kulm, 1989; Clarke and Carver, 1989; Adams, 1990; Darienzo and Peterson, 1990; Peterson and Darienzo, this volume), but have thus far provided only rough estimates of great earthquake recurrence and few constraints on earthquake magnitude.

Resolution of seismic-safety issues in the region requires reasonable estimates of the magnitude, extent, and recurrence of prehistoric plate-interface events during the late Holocene (last 5000 years; Shedlock and Weaver, 1989). A probable locked plate interface, a convergence rate of 30-40 mm/yr, and the young age (<10 Ma) of the subducting crust suggest plate-interface events of M_W 8-9 are plausible (Heaton and Kanamori, 1984). But the recurrence and magnitude of these events probably differs from place to place along the CSZ (Spence, 1988; Weaver and Shedlock, 1989). The temporal and spacial pattern of events needs to be determined for the last few seismic cycles (for example, Thatcher, 1989) to assess accurately the hazard from great earthquakes (Nishenko, 1989).

The degree to which changes in the characteristics of the plate interface along the CSZ control the extent of Holocene ruptures, that is, the extent to which the CSZ is segmented, is critical in determining whether events near magnitude M_W 8 or near M_W 9 have occurred more often. Subduction continues at a higher rate beneath Washington than beneath Oregon due to rotation of the Juan de Fuca plate about a pole in northern California (Riddihough, 1984; Nishimura and others, 1984; Wilson, 1986). The properties of the subducting plate and the width of the locked zone also differ from north to south (Crosson and Owens, 1987; Weaver and Shedlock, 1989; Rasmussen and Humphreys, 1989). Furthermore, the stress regime in the crust throughout Cascadia is dominated by northerly compression due to the interaction of the Pacific, Gorda, Juan de Fuca, and North American plates (fig. 1; Spence, 1989; Magee and Zoback, 1989). The complex interactions of these plates and limited seismicity data suggest that the CSZ is composed of 4-7 segments (Hughes and others, 1980; Weaver and Michaelson, 1985; Weaver and Baker, 1988; Heaton and Hartzell, 1986; Rogers, 1988; Spence, 1989). Although plate-interface properties within regions of high seismic moment release may provide the primary control on rupture initiation and extent (Thatcher, 1990), the boundaries between these segments probably reflect the endpoints of many plate-interface ruptures during the Quaternary.

Here we review some of the preliminary conclusions of geologic field studies that are underway in the coastal areas of Oregon and Washington (as of December 1989) and speculate on some of their implications for the hazard posed by plate-interface earthquakes in the region. Because the record of paleoseismic events is fragmentary, a variety of methods is being used to piece together the late Quaternary tectonic and paleoseismic history of the region. Work in progress falls into two groups--paleoseismology studies of the late Holocene coastal record, and studies of cumulative middle and late Quaternary tectonic deformation as expressed by marine and fluvial terraces.

HOLOCENE PALEOSEISMOLOGY

THE RECOGNITION OF COSEISMIC COASTAL DEFORMATION IN WASHINGTON AND OREGON

A linear pattern of coseismic deformation has been observed along subduction zone coasts during great plate-interface earthquakes (Plafker, 1965; Ando, 1975; reviewed in Lajoie, 1986). Plafker (1969; 1972) developed a model of coseismic deformation with an elongated zone of uplift and a parallel zone of subsidence based on mapping of the coseismic deformation in the coastal areas of southern Alaska and south-central Chile following the two largest earthquakes of this century. Uncoupling of the leading edge of the overiding plate during megathrust events can produce up to 2 m of regional coseismic uplift in a zone 80-160 km wide. Locally in this zone, up to 10 m of coseismic uplift can occur by thrusting on small, steep faults in the toe of the upper plate (Plafker, 1969; Plafker and Rubin, 1978; Yonekura and Shimazaki, 1980). The zone of subsidence, which is arcward of the zone of uplift, is often wider than the zone of uplift and can be hundreds of kilometers long; maximum regional subsidence can reach 2-3 m during the largest events (for example, Plafker and Savage, 1970).

INITIAL SEARCH FOR EVIDENCE OF HOLOCENE UPLIFT

Most previous paleoseismology studies along subduction zone coasts have focused on shoreline features inferred to have been raised by sudden coseismic uplift (reviewed in Lajoie, 1986; Berryman, 1987). Although large storms can deposit terrace-like beach berms, flights of Holocene marine terraces along active convergent margins are usually interpreted to be of coseismic origin. Examples of probable Holocene coseismic terraces from other subduction zones include those in Chile (Kaizuka and others, 1973), Japan (Matsuda and others, 1978), New Zealand (Berryman and others, 1989), the New Hebrides Islands (Taylor and others, 1990), and Alaska (Plafker and Rubin, 1978; Winslow, 1988).

Initial paleoseismic work on the CSZ also focused on a search for features produced by coseismic uplift, but no extensive Holocene terraces were found, even along the southern part of the Oregon coast that lies closest to the trench (distance of the coast from base of continental slope, 60-105 km; fig. 2; Golder Associates, Inc., 1986). West and McCrumb (1988a) interpreted the broad, modern wave-cut platforms and the lack of Holocene marine terraces along the Washington and Oregon coasts as evidence that repeated, great, plate-interface earthquakes probably had not occurred (at least in central and southern Oregon). They suggested that if plate-interface earthquakes had occurred during the late Holocene, then sea level rise must have equaled or exceeded the rate of coseismic plus interseismic uplift, or that plate-interface events were too small to produce significant amounts of coseismic uplift.

In the southernmost part of the CSZ, Holocene marine terraces suggesting uplift rates of 3-4 mm/yr have been reported from the Cape Medocino, California area near the plate triple junction (fig. 1; Lajoie and others, 1982; Carver and Burke, 1987; Carver and others, 1989). In addition, Kelsey (1989; 1991) found a single example of a Holocene beach berm of probable coseismic origin near the Elk River 6 km south of Cape Blanco on the southern Oregon coast. Cape Blanco is only about 60 km east of the Cascadia trench and is inferred to lie within a zone of coseismic uplift during plate-interface events (fig. 3). Rates of late Pleistocene uplift for Cape Blanco are higher than rates to the north and south (West and McCrumb, 1988a; Muhs and others, 1990; Kelsey, 1990). The proximity of Cape Blanco to the trench may thus explain the location of this raised berm along a coastline otherwise devoid of evidence for Holocene uplift.

EVIDENCE FOR COSEISMIC SUBSIDENCE

Heaton and Hartzell (1986), Atwater and Grant (1986), and Atwater (1987) changed the focus of paleoseismicity studies in the Pacific Northwest by proposing that much of the coastline in Oregon and Washington is in a zone of coseismic subsidence rather than uplift during Holocene great earthquakes.

Atwater (1987) based his interpretation of repeated coseismic subsidence on the stratigraphy of estuarine deposits in southwestern Washington. He inferred that sudden jerks of subsidence had repeatedly submerged wetland soils on the margins of estuaries during great earthquakes. Plafker (1972, p. 917) had previously suggested the paleoseismic potential of this type of stratigraphic record, and Ovenshine and others (1976), Bartsch-Winkler and others (1983), Combellick (1986), and Bartsch-Winkler (1988) had investigated records of this type in Alaska. However, Atwater (1987) was the first to recognize evidence for coseismic subsidence in the Pacific Northwest and to use stratigraphic methods to show that such subsidence had occurred repeatedly over a large area. Darienzo (1987) and Darienzo and Peterson (1987; Peterson and Darienzo, 1988a; Darienzo and Peterson, 1990), working independently in the Netarts Bay estuary in northern Oregon, adopted the coseismic subsidence model to explain similar sequences of interbedded peat and mud. They and other investigators have subsequently found similar evidence for coseismic subsidence in many estuaries along the Oregon and Washington coasts (Grant and McLaren, 1987; Hull, 1987; Nelson and others, 1987; Nelson, 1987; Grant, 1988; Peterson and Darienzo, 1988b; 1989a; 1989b; McInelly and others, 1989), as well as in the Humboldt Bay area of California (Vick, 1988; Carver and others, 1989; Grant and others, 1989).

The repetitive stratigraphy identified by Atwater (1987; 1988a) in cores and tidal outcrops consists of peaty soils representing densely vegetated lowlands abruptly overlain by intertidal muds. Typically, the peats are thin (0.05-0.2 m), have abrupt upper contacts and gradational lower contacts, are separated by thick (0.5-1.0 m) intervals of intertidal muds, and often have spruce or western red cedar stumps rooted in them. Atwater's (1987) evidence that these peaty soils were submerged suddenly due to coseismic subsidence, in some cases accompanied by tsunamis, included: (1) the abruptness of the upper peat contact, (2) the amount of submergence (0.5-2.0 m) suggested by plant macrofossils above and below the abrupt contacts, (3) the similar character of the peat-mud couplets both laterally at different sites and in sequence, (4) the presence of continuous peat-mud couplets where these sequences lap up on older, Pleistocene sediments at valley sides, (5) the fining-upward, landward-thinning sand beds capping some peats at sites where no sand is presently being deposited, and (6) the similarity of these sequences to those reported to have been deposited following great earthquakes in Alaska and Chile (Ovenshine and others, 1976; Wright and Mella, 1963). In southwestern Washington Atwater (1987; 1988b) identified at least 8 separate buried peats; the 5 most widespread peaty soils are dated at <3 ka (kilo-annum) and all are <5 ka. Atwater (1987) also used the coseismic-subsidence model to reconcile the relatively slow regional uplift of the coast expressed by raised Pleistocene marine terraces in southwestern Washington (0.1-0.4 mm/yr) with the much higher, regional rate of uplift obtained from tide gauge records (2-3 mm/yr). Periods of jerky subsidence (0.5-2 m), superimposed on a low interseismic uplift rate throughout the late Pleistocene, were hypothesized to yield an even lower net uplift rate.

Because interbedded peat-mud couplets can also be produced by non-tectonic processes (discussed below), the most convincing studies of coseismic subsidence focused on evidence suggesting *sudden, significant rises* in relative sea level (submergence). Rooted stems of herbaceous plants are found locally entombed by overlying estuarine mud at the upper contact of some peaty soils in southwestern Washington (Atwater, 1988c). These fossils provide compelling evidence for sudden submergence because they are generally not preserved at the marsh surface for more than a few years. Tree-ring studies by Yamaguchi and others (1989a) have also shown that cedar trees rooted in the youngest buried soil at 4 sites in southwestern Washington probably died suddenly within a few years of each other. In-progress microfossil studies of some peat-mud couplets in south-central Oregon (A.R. Nelson, A.E. Jennings, Univ. of Colorado; K. Kashima, Kyushu Univ., unpub. data) northern Oregon (Darienzo, 1987; Peterson, 1989; Darienzo and Peterson, 1990), and southwestern Washington (Hemphill-Haley, 1989) support Atwater's (1987) deduction that the transitions across the upper contacts of the peat-mud couplets represent 0.5-2.0 m of sudden submergence.

Atwater (1987; 1988a) suggested that evidence of sediment deposition by tsunamis and sediment liquefaction directly overlying the abruptly submerged lowland soils, would be compelling evidence of regional coseismic subsidence. Initial mapping of the sand beds that cap some peaty soils in some parts of the marshes bordering tidal streams at Willapa Bay, Washington (Atwater, 1987), showed that the sand pinches out landward--the expected result of tsunami deposition. The sand beds were only found capping peaty soils and not within the thick estuarine muds, suggesting that all sand beds were deposited immediately following sudden submergence of the marsh surface. Similar sand beds capping buried peaty soils at Grays Harbor, Washington (Reinhart and Bourgeois, 1988), Netarts Bay, Oregon (Peterson and Darienzo, 1988a; Darienzo and Peterson, 1990), along the Salmon River, Oregon (Grant and McLaren, 1987), at Siletz Bay, Oregon (Peterson, 1989), and at Alsea Bay, Oregon (Darienzo, 1989; Peterson and Darienzo, 1989b), show similar relations and have also been attributed to tsunamis. Field and modeling studies of the sand beds at Willapa Bay (Reinhart and Bourgeois, 1987; 1988; 1989; Bourgeois and Reinhart, 1988) indicate that large, landward-directed surges of sandy water that accompanied some of the subsidence events were locally-generated tsunamis rather than storm surges, seiches, or distant tsunamis. Microfossil studies also lend support to the tsunami hypothesis, especially where abrupt peat contacts lack capping sand beds. In estuarine mud located directly above an abrupt peat contact in Willapa Bay, Eileen Hemphill-Haley (USGS, Menlo Park, unpub. data) has found marine diatoms, which are not characteristic of estuarine environments; possibly these diatoms were transported from the continental shelf by a tsunami.

So far, only one estuarine site in southwest Washington has yielded definitive evidence of prehistoric sediment liquefaction, but the eruption of a sand blow at this site was apparently not concurrent with a subsidence event (Atwater, 1988b). However, liquefaction features usually develop in well-sorted fine sand or silt beds that are saturated and confined by overlying finer-grained units (for example, Obermeier and others, 1989). Furthermore, adequate (1-2 m high) vertical exposures where the water table is now below the base of the exposure are required to study such features. No systematic search for liquefaction features has yet been made in sandy areas along the Oregon or Washington coasts, but the continuing rise of late Holocene sea level in the region ensures that any exposures of liquefiable sediments will be difficult to find. Thus, the present lack of well-documented paleoliquefaction sites does not preclude great earthquakes along the CSZ during the late Holocene.

ALTERNATIVES TO THE COSEISMIC-SUBSIDENCE HYPOTHESIS

Much of the initial debate about the applicability of the coseismic-subsidence model to Oregon and Washington has centered on whether observed changes in the sequences of estuarine sediment are distinctive enough to be due to coseismic movements or whether they might have been produced by non-tectonic processes. Early investigations (Darienzo and Peterson, 1987; Darienzo, 1987; Reinhart and Bourgeois, 1987; and others) questioned whether at least some of the peat-mud couplets might not result from (1) storm surges, distant tsunamis, or floods coupled with relatively rapid rises in regional sea level, (2) complex patterns of tidal channel cuts and fills, or (3) sudden changes in tidal range at a site caused by migration of bars near the mouths of tidal inlets. Atwater (1987) argued convincingly that storm surges, distant tsunamis, and floods can only raise sea level very briefly (less than a few days), and that these processes could not permanently change woody lowlands into intertidal mud flats as suggested by the lithologies and plant macrofossils in the sequences in southwest Washington. Marsh vegetation should readily incorporate any sediment deposited on the marsh from floods or storms without permanent submergence of the marsh. In support of these arguments, Peterson and Darienzo (1988b; 1988c; 1989b) found no evidence of significant sediment deposition in estuarine marsh sequences during either major historic floods or large storm surges in fluvially-dominated estuaries in Oregon. Most investigators (for example, Atwater, 1987; 1988c; Peterson and Darienzo, 1989b; Darienzo, 1989; Darienzo and Peterson, 1990) also found that the lateral continuity of most of the buried peaty soils (as demonstrated by transects of cores or laterally-extensive outcrops) eliminated tidal channel cut-and-fill processes as a possible cause at most sites. The regular sequence of lithologic changes within most peat-mud couplets and the changing environments indicated by microfossil assemblages (Hemphill-Haley, 1989; A.R. Nelson, A.E. Jennings, K. Kashima, unpub. data) are also incompatible with a channel origin.

Sudden changes in tidal range are more difficult to dismiss as a possible cause of interbedded peat-mud sequences (Darienzo, 1987). In fact, this mechanism is often cited to explain sequences in northwestern Europe (Shennan, 1986) that are similar to the peat-mud couplets in the Pacific Northwest. On the Atlantic coast of the U.S., Thomas and others (1989) also suggested this explanation for the sudden changes in lithology and foraminiferal assemblages in cores from a Connecticut salt marsh. Sudden changes in tidal range could explain the burial of some peats in some inlets in the Pacific Northwest, but correlation of similar sequences between sites in inlets of different morphology and tidal characteristics shows that local changes in tidal range cannot have produced all peat-mud couplets.

The weakest part of Atwater's (1987) case for coseismic subsidence was his suggestion that the sequences of interbedded peat and mud so typical of late Holocene estuarine sequences in the Pacific Northwest should be characteristic of tectonically active coasts. He argued that on non-tectonically-active coasts, the slow, uniform rise of sea level during the late Holocene should have produced thick, uniform layers of peat like those found in Puget Sound, San Francisco Bay, and Cape Cod marshes (Fig. 4A). Using this line of reasoning, the widespread peat-mud couplets in southwestern Washington were thought to be good evidence for the jerky rise of Holocene sea level. Each jerk was inferred to result from an episode of sudden relative sea level rise due to coseismic subsidence (Fig. 4B). But many tens of studies on other non-tectonic coasts, and particularly those from northwestern Europe, show sequences that are, at least superficially, similar to those in the Pacific Northwest (for example, Devoy, 1987; Tooley and Shennan, 1987). Many of these long-term studies are much more extensive than any work so far attempted in the Northwest (for example, Jelgersma, 1961; Tooley, 1978; Plassche, 1982; Ters, 1987) and are supported by thousands of ^{14}C ages (for example, Berendsen, 1984; Shennan, 1987).

The stratigraphy at many individual sites on non-tectonic coasts suggests that local rates of sea level rise have varied markedly (Shennan, 1986); a number of processes have been invoked to explain these inferred rate changes. In most cases, the upper and lower contacts of peaty units in sequences from non-tectonic coasts are gradational with overlying and underlying intertidal muds, suggesting transitions between upper intertidal and lower intertidal environments that lasted at least tens of years. Oscillations in late Holocene sea level are the most often cited explanations for the interbedded peats and muds on stable coasts (for example, Rampino and Sanders, 1981; Shennan, 1982; Shennan, 1986; Fairbridge, 1987; Ters, 1987). Despite the limits of resolution of altitudinal and ^{14}C age data in sea level studies, some regional oscillations of sea level have been identified (Shennan, 1987), but these oscillations are not necessarily synchronous from region to region (Kidson, 1982). Other processes that may be involved in sea-level changes that produce peat-mud couplets on non-tectonic coasts include: (1) large storms (Tooley, 1985; Cullingford and others, 1989), (2) long periods with many storm surges (Streif, 1980), (3) changes in the shapes and gradients of estuarine systems due to river changes and rising Holocene sea level (Plassche, 1980), and (4) gradual regional tectonic warping (Newman and others, 1980; Shennan, 1989). An additional problem is that lithologic changes in the peat-mud sequences may not be direct proxies of small sea level changes (for example, Orson and others, 1985; Stevenson and others, 1986). Such lithologic changes do show, however, that marsh aggradation has not kept up with late Holocene sea level rise on many non-tectonic coasts (for example, Brooks and others, 1979; Kearney and Ward, 1986) and that the lack of thick sequences of peat is not necessarily evidence for tectonic instability.

There have been no detailed studies of Holocene sea level in Oregon and Washington, so the rate and character (smooth vs. oscillatory) of sea level rise in this region is poorly known. In general, the rate of rise is thought to be similar to that for the U.S. east coast, about 1-2 mm/yr (Peterson and others, 1984, fig. 6; Phipps and Peterson, 1989). However, small (<1.0 m), short-term (10-500 yr), non-tectonic oscillations in regional sea level, combined with changes in local sediment influx and irregular local sediment compaction (for example, Kaye and Barghoorn, 1964), cannot be ruled out as the cause of some peat-mud couplets (fig. 4C). Only where there is good evidence of a sudden, substantial change in water depth across the upper contact of uneroded peaty units can coseismic subsidence be inferred.

The inability to rule out small regional oscillations in the rate of sea level rise casts uncertainty on the suggestion (Peterson and Darienzo, 1988b; Peterson, 1989; Darienzo and Peterson, 1990) that some gradual contacts in marsh sequences in northern Oregon were produced by slow accumulation of tectonic strain. Given currently accepted models of seismic strain accumulation, release, and recovery (for example, Savage, 1983; Thatcher, 1984), a tectonic mechanism for the origin of gradual contacts is certainly plausible. However, we see no way to distinguish lithologies in marsh cores or outcrops resulting from these gradual, tectonically-induced changes in regional uplift rate from those resulting from gradual, non-tectonic changes in the rate of sea level rise.

REGIONAL DIFFERENCES IN LATE HOLOCENE ESTUARINE SEQUENCES IN OREGON

Studies of late Holocene coastal-marsh stratigraphy in Oregon and Washington began with Darienzo's (1987) initial fieldwork in 1985 at Netarts Bay. However, most projects are still in progress and only the Netarts Bay work has been published as a full-length paper (Darienzo and Peterson, 1990). Reconnaissance investigations have been made in most of the 27 estuaries along the Oregon and Washington coasts, but detailed work has been completed in fewer than half. Despite the preliminary nature of the interpretations of sequences in many estuaries, there appear to be significant differences in the character of marsh stratigraphic sequences along the Oregon coast.

NORTHERN OREGON

Work at Netarts Bay (Peterson and Darienzo, 1988a; 1988b; Darienzo and Peterson, 1990) and at the Nehalem River and Salmon River estuaries (Grant, 1988; 1989) shows that late Holocene estuarine sequences in northern Oregon contain multiple (3-5), widespread, buried salt-marsh and occasional spruce-swamp soils, abruptly overlain by intertidal muds and sometimes by sand beds. In a reconnaissance at the Nestucca River estuary, Peterson and Darienzo (1988b) also encountered 3-5 buried peaty soils. The peat-mud couplets in northern Oregon are similar to those studied by Atwater in southwestern Washington, but the peaty horizons are somewhat thicker (0.2-1 m) and are generally separated by thinner intervals of mud (0.3-1.5 m). Stratigraphic relationships are more difficult to document in Oregon because the height of tidal outcrops is lower and outcrops are less extensive than in southwestern Washington. Darienzo and Peterson (1990) estimated that the peat-mud couplets at Netarts Bay each represent 1-1.5 m of sudden coseismic subsidence. Thus, similarities in marsh stratigraphy in northern Oregon and Washington suggest recurrent coseismic subsidence of regional (>200 km) extent (Grant, 1989; Darienzo and Peterson, 1990).

CENTRAL OREGON

Few details of marsh stratigraphy studies at estuaries along the central Oregon coast have been published. Peterson and Darienzo (1988c) report 3-5 buried marsh surfaces in the upper 5 m of cores at the Siletz River estuary. More extensive work at the Alsea River estuary showed 3-10 peaty marsh soils in cores 4-7 m deep (Peterson and Darienzo, 1988b; 1989b; Darienzo, 1989). Most of the buried peaty soils at Alsea Bay have sharp upper contacts and gradational lower contacts, but peaty units in the upper 2.5 m of most cores are separated by thin (2-40 cm) intervals of mud (fig. 5C). Buried peaty soils in the lower portions of the cores are separated by greater thicknesses of mud (0.4-1.8 m), similar to the peat-mud couplets in northern Oregon. Low outcrops and transects of cores show that many stratigraphic units can be correlated for hundreds of meters across the marsh. Sandy capping beds are less well developed on the buried peaty soils in the fluvially-influenced estuaries of the Siletz and Alsea Rivers than in the more tidally-influenced estuaries such as Netarts Bay (Peterson and Darienzo, 1988b). However, distinctive landward-thinning, sandy capping beds on the peat units in the western Alsea River marshes, combined with the repetitive, abrupt upper contacts of the peats led Peterson and Darienzo (1989b) to conclude that the contacts were produced by coseismic subsidence accompanied by tsunamis. Based on changes in sediment lithology across the abrupt peat-mud contacts, the amount of subsidence during each event was estimated to be 0.5-1.0 m, less than that estimated for events in northern Oregon (Peterson and Darienzo, 1989b; Darienzo, 1989).

We found a different type of marsh stratigraphic sequence preserved in the Siuslaw River estuary about 50 km to the south of Alsea Bay. About 3.5-4.0 m of continuous, uniform salt-marsh peat overlies tidal sand and mud at an inlet on the south side of the river (fig. 5B); 3 km to the north, at the mouth of the North Fork of the Siuslaw, about 4 m of salt-marsh peat overlies 3 m of brackish or freshwater peat. The thickness of the salt-marsh peats and several ^{14}C ages show that the upward growth of the marsh surface was able to keep up with rising sea level over a time span of 2000 yr (fig. 4A). Reconnaissance coring upstream of the sites with thick peat showed interbedded peats (5-30 cm thick) and muds (10-50 cm thick), but contacts are mostly gradational; these gradational contacts suggest gradual rather than sudden submergence. Thus, there may have been gradual changes in the rate of sea level rise at the Siuslaw estuary, but the thick peats preclude any significant (>0.3-0.5 m) sudden changes in sea level.

Although the types of marsh stratigraphic sequences in the Siuslaw River and Alsea River cores differ significantly, the difference is less distinct than the differences between sequences in northern Oregon and the Siuslaw. Perhaps a few of the most gradual contacts in the upper parts of the Alsea Bay cores correlate with some of the subtle, gradual lithologic changes in the thick peats in our cores from the Siuslaw. If so, we cannot rule out a non-tectonic origin for these lithologic changes because these changes are of smaller magnitude (probably <0.3 m water level change) and much less abrupt than lithologic changes observed in similar sequences on non-tectonically active coasts. Tectonic and non-tectonic processes may be producing identical small couplets along the Oregon coast, and present methods may not allow separation of a small tectonic component from small oscillations in sea level and sediment input.

SOUTHERN OREGON

South of the Siuslaw River, reconnaissance coring at many sites in the Umpqua River, Coos Bay, and Coquille River estuaries by Nelson (1987; 1988) and Peterson and Darienzo (1989a; Darienzo, 1989) shows no consistent regional pattern of multiple, abruptly buried peaty soils like those found in northern Oregon. Many sequences show a single abruptly buried soil (5-50 cm thick) at 0.8-1.5 m depth; the soil at many sites was apparently submerged suddenly and more gradually at others. An outcrop along the Coquille River exposes such a buried soil, which has the trunks of a few small spruce trees rooted in it and is very similar to buried soils in southwestern Washington. This peaty soil is thin (5 cm), but a possibly correlative soil at North Slough, in northern Coos Bay, is much thicker (50 cm) and resembles the thick salt-marsh peat found at the Siuslaw River. The thickness, depth, and organic material content of uncommon, deeper (>1.5 m depth) peaty soils in the Coos Bay region is highly variable; some have abrupt upper contacts and others do not. However, multiple buried peats are rarely found more than 5-10 m from valley side slopes and few marshes appear to have been as extensive in the last 3000 yr as they are today.

A major exception to the lack of widespread, multiple peat-mud couplets in the Coos Bay region is at South Slough, which occupies the axis of an active north-trending syncline in western Coos Bay. Extensive coring in a small marsh at the south end of the slough revealed 4-9 buried salt marsh surfaces in cores 5-8 m long (fig. 5A). At least 4 of the buried surfaces can be traced in cores across the inlet. Based on lithology, unit thickness, and preliminary foraminifer and diatom data (A.R. Nelson, A.E. Jennings, K. Kashima, unpublished data) these couplets represent about 0.5 m of sudden submergence followed by deposition of 0.3-0.5 m of intertidal sediment. A core from Day Creek, near the middle of South Slough, penetrated a similar sequence of 8 buried peaty soils (Peterson and Darienzo, 1989a); these authors also suggested about 0.5 m of submergence during these events. Reconnaissance coring by Nelson at other sites in the western part of South Slough shows 2-4 abruptly buried peats, two of which are capped by thin beds of silty sand.

Nelson (1987), Peterson and Darienzo (1989a; Peterson, 1989), and McInelly and Kelsey (1990; McInelly and others, 1989) suggest that at least some of the submergence events marked by the buried peaty soils at South Slough are the result of coseismic displacement on local structures. Mapping and dating of Pleistocene marine terraces and underlying structures in the Coos Bay region by McInelly and Kelsey (1990) shows that the South Slough syncline is being deformed, but that the style of deformation in this region is more complex than previously assumed by Adams (1984). McInelly and Kelsey (1990) have also dated tree stumps in the intertidal zone adjacent to flexure-slip faults within the South Slough syncline that indicate Holocene displacements on these structures. Thus, most of the abruptly buried marsh surfaces in South Slough are probably the result of coseismic displacements due to folding or faulting of the syncline. These displacements were small (about 0.5 m) and cannot easily be correlated with the few buried marsh surfaces found elsewhere in the Coos Bay area.

South of Cape Blanco, the only sizable estuaries in southernmost Oregon are at the mouths of the high-gradient Rogue and Chetco Rivers. Sediments in these two estuaries are predominantly gravel and sand and only small areas of marsh fringe the rivers. The lack of quiet-water intertidal mud in these estuaries indicates little chance for a well-preserved, unambiguous stratigraphic record of late Holocene sea level changes.

GREAT EARTHQUAKE RECURRENCE AND THE LIMITATIONS OF ^{14}C DATING

Until the advent of recent coastal field studies, estimates of the recurrence of great earthquakes in the CSZ were limited to average values, largely based on analogies with other subduction zones. Heaton and Kanamori (1984) used empirical and theoretical relations to calculate an average event for the Juan de Fuca plate of M_w 8.3±0.5; such an event could recur every 126-420 years depending on the ratio of aseismic to seismic slip (Kanamori and Astiz, 1985). Heaton and Hartzell (1986) compared the Juan de Fuca plate to other subduction zones in south-central Chile, southwest Japan, and southern Alaska. Using these three analogous subduction zones and the convergence rate of the Juan de Fuca plate, they suggested great earthquake recurrences for the Juan de Fuca plate of 250-500 yr, 100-250 yr, and possibly 1000 yr, respectively. In a study of turbidite deposition on the continental shelf, Adams (1989; 1990) inferred a fairly uniform recurrence of 590±170 yr for major earthquakes. However, this study left open questions about the size and location of earthquake sources and whether all turbidites on the shelf are earthquake-generated. For example, Thatcher's (1990) recent analysis of great earthquakes in other circum-Pacific subduction zones suggests reccurence of plate-interface events in the CSZ might be very non-uniform.

A key assumption in initial paleoseismic studies in the Pacific Northwest was that conventional radiocarbon dating could provide fairly accurate estimates of great earthquake recurrence through analysis of organic materials at or just above the upper contact of abruptly buried peaty soils. It also was hoped that the ages of buried peats were separated by enough time to allow the soils to be correlated from one estuary to another on the basis of conventional radiocarbon dating. However, studies with over a hundred conventional ^{14}C ages in Washington (Atwater, 1988b; Hull, 1988) and our comparative dating program in southern Oregon shows that conventional ^{14}C ages from the uppermost parts of buried wetland soils vary widely depending on the type of organic material analyzed and how it is prepared prior to analysis (Grant and others, 1989). For example, mean ages from identical splits of the same peat samples from Coos Bay sent to two different laboratories that use different pretreatment methods varied by 100-700 yr. Detrital wood picked from some of these same peat samples varied from 100 yr older to 1000 yr younger than ages on the bulk samples. More limited data for older layers shows similar variability, so, as ages increase the percentage of uncertainty in each age decreases.

These results indicate that, with the exception of tree trunks well-rooted in peaty soils, individual conventional ages on most materials can only provide a general estimate (±150-300 yr) of the time of peat burial (Grant and others, 1989); most ages from the buried peaty soils are maximum estimates of the time of burial. The wide range in ages from the same peat layers also suggest that many buried peaty soils formed over time intervals of at least 200-1000 yr. Where time intervals between events are >1000 yr, or where multiple ages on different materials above and below the upper contact of peaty soils can be used to constrain the age of the surface of the soil (for example, some of Atwater's (1988b) sites), conventional ages may allow buried soils to be correlated. However, where the mean event recurrence interval is 400-500 yr and where high-quality tree-trunk samples are lacking (central and southern Oregon), conventional ages cannot be used by themselves to demonstrate correlations.

Despite the problems discussed above, samples from most of the soils inferred to have been buried during as many as 10 late Holocene coseismic events in the coastal Pacific Northwest have been dated by conventional ^{14}C methods (Grant and others, 1989). Atwater (1987; 1988b) collected a variety of materials from above and below the upper contacts of five of the youngest of eight buried wetland soils recognized in southwestern Washington. Mean ages for the subsidence events inferred from these samples were about 0.3, 1.6, 1.7, 2.7, and 3.1 ka. An additional subsidence event about 1.0 ka was identified at the Waatch River estuary near the northwest tip of Washington and at an inlet on the north side of the mouth of the Columbia River (Atwater, 1988b). Grant (1989) obtained mean ages of about 0.4, 1.0, and 1.6 ka from three buried soils along the Nehalem River and about 400 yr BP from the single buried soil exposed along the Salmon River. Darienzo and Peterson (1990) used peat ages from six buried soils at Netarts Bay to suggest synchrony of events in northern Oregon and Washington about 0.3-0.5, 1.0-1.3, 1.4-1.8, and 3.0-3.5 ka. At Alsea Bay in central Oregon, Peterson and Darienzo (1989b) also dated peat from six of ten identified peaty soils that were buried within the past 5000 years. Finally, Nelson (unpublished data) used 20 ages on peat and detrital wood from nine buried soils in a core from the south end of South Slough in Coos Bay to infer times of peat burial about 0.3-1.0 ka, 1.0-1.3 ka, 1.1-1.5 ka, 1.7 ka, 2.1-2.5 ka, 2.3-2.5 ka, 2.6-2.9 ka, 2.7-3.0 ka, and 3.6-3.9 ka (fig. 5A). Peterson and Darienzo (Geology Dept., Portland State Univ., unpublished data) obtained similar ages on six of eight buried soils in a core from central South Slough. But few of the ages from Washington and northern Oregon and none from central and southern Oregon are precise enough to allow correlation of most buried soils.

The conventional ages show that recurrence of submergence events at individual sites is non-uniform. Average recurrence values are in the range of 400-700 yr, but the sites with many ages suggest recurrence intervals as small as 100-200 yr and as large as 1000-1600 yr. Most ages on the youngest buried soil suggest the youngest event occurred during the interval 0.1-0.5 ka, but wide fluctuations in ^{14}C flux in the upper atmosphere during this period (Stuiver and Pearson, 1986) and the large errors on ages from peaty soils make it impossible to determine whether or not the youngest soil was buried about the same time everywhere (Grant and others, 1989).

The most accurate method of dating subsidence events from one site to another is through tree-ring studies, such as those by Yamaguchi (1988). Maximum-limiting calendar ages of AD 1618-1687 appear to constrain the timing of the most recent subsidence event at four sites that span 100 km of the southwest Washington coast, but some interpretive problems remain (Yamaguchi and others, 1989b). Unfortunately, this method can only be applied to those few sites with well-preserved fossil cedar trees and such sites have been found only in southwestern Washington. High-precision conventional ^{14}C analysis (Pearson, 1979) of large samples of less than 10 rings from large buttress roots (near the base of the trunk) may result in the dating of one or two earlier events to within a few decades in northernmost Oregon and southwest Washington where spruce trunks rooted in buried peaty soils are common (Atwater, 1988b, table 2). These are probably the best available methods for testing the synchrony of individual subsidence events along substantial portions of the Pacific Northwest coast.

STUDIES OF LATE AND MIDDLE QUATERNARY DEFORMATION

Determination of rates and styles of longer-term Quaternary deformation is critical in interpreting evidence for Holocene seismic or aseismic deformation in active subduction zones (Bloom, 1980). Studies of raised marine terraces are the most traditional type of geologic investigation of Quaternary deformation along tectonically active coastlines. Other types of geologic investigations include examination of fluvial terrace sequences and analysis of regional patterns of river and drainage basin morphology (for example, Wells and others, 1988; Rhea, 1989; Merritts and Vincent, 1989). These investigations can yield information about the cumulative effects of Quaternary deformation, which may help determine whether a coastal region has been affected by earthquakes of regional or more local extent. Quaternary deformation is commonly quantified through analysis of uplift rates, identification of active structures, and geodetic and geologic analysis of regional tilting. The use of marine and fluvial terraces to describe the nature of these types of Quaternary deformation along the CSZ will be discussed below.

UPLIFT RATES

Uplifted marine terraces are found along many active coastlines. Most Pleistocene marine terraces represent paleo-shorelines that were formed during sea level high stands associated with interglacial periods and subsequent tectonic uplift (Lajoie, 1986). For this reason, terrace studies have been conducted along many tectonically active coastlines around the world (Berryman, 1987); of particular interest are studies in regions undergoing active subduction such as Japan (Yonekura and Ota, 1986), the New Hebrides Islands (Taylor and others, 1985), and Chile (Hsu and others, 1989). Analysis of uplift rates in other active areas has helped to define the nature of tectonic deformation associated with the CSZ.

Calculation of uplift rates from marine terrace data requires that the age and highest present elevation of the wave-cut platform be known, as well as the elevation of sea level at the time the terrace was formed. The elevation of paleo-sea level is determined by comparing the age of the terrace to paleo-sea level curves from well studied areas such as New Guinea, California, and Japan (for example, Lajoie, 1986). The present elevation of the platform is then corrected for paleo-sea level and any tilting that may have occurred since the terrace was formed. The resulting value of tectonic uplift can then be used with the age of the terrace to calculate the net rate of uplift since the terrace was formed. Uplift rates are commonly compared from place to place along the coast, and to other active coasts, to identify patterns of regional deformation (for example, Ota and Yoshikawa, 1978).

Pleistocene marine terraces are preserved discontinuously along the Oregon and Washington coasts (fig. 6; Golder Associates, 1986), but are best developed and have been mapped and dated in the most detail near Cape Blanco and Cape Arago in southern Oregon (Griggs, 1945; Kennedy and others, 1982; Adams, 1984; West and McCrumb, 1988a; McInelly and Kelsey, 1990; Kelsey, 1990). A recent detailed study by Muhs and others (1990) used new uranium-series ages, and amino acid and oxygen isotope data to correlate and date the two lowest terraces at Cape Blanco. Their results include uplift rates of about 0.85-1.25 mm/yr for the 80 ka Cape Blanco terrace and about 0.8-1.5 mm/yr for the 105 ka Pioneer terrace. These uplift rates are lower than those reported by earlier studies (West and McCrumb, 1988a), but they are still the highest rates reported from marine terraces along the Oregon and Washington coasts.

Muhs and others (1990) also compared their new uplift rates with rates from well-dated terrace sequences at other active subduction zones, and concluded that uplift rates at Cape Blanco were not unusually high or low. They also found no clear relation between the style of plate convergence and uplift rates of marine terrace sequences in other subduction zones. These conclusions conflict with those of West and McCrumb (1988a), who used such a comparison to infer that large plate-interface earthquakes had probably not occurred on the CSZ in late Holocene time.

In addition to marine terrace studies, we examined fluvial terraces because these features can be useful for measuring incision rates at considerable distances from the coast. In this approach, we consider incision rates to be an indirect measure of uplift rates. Our studies of fluvial terraces along the Siletz, Siuslaw, Smith, and Umpqua Rivers suggest slow (0.2-0.6 mm/yr) rates of regional uplift of the central part of the Oregon coast (fig. 7A). These uplift rates are based on terrace heights measured from modern river level to the stream-cut bedrock bench (strath) underlying each terrace. Terrace ages are derived from radiocarbon dating of Holocene and latest Pleistocene terrace sediments and thermoluminescence (TL) dating of older terrace deposits. The use of TL dating is relatively new, and our dates are the first such analyses of fluvial sediments in the Pacific Northwest. Although the results are not without some inconsistencies (fig. 7A), we believe our TL ages are accurate to within +/- 20% because our uplift rates are consistent with rates from marine terrace studies in the region (West and McCrumb, 1988a; Muhs and others, 1990).

DEFORMATION ON QUATERNARY STRUCTURES

The new marine terrace ages of Muhs and others (1990) were used by McInelly and Kelsey (1990) and Kelsey (1990) to evaluate the amount and style of deformation on specific structures near Cape Arago and Cape Blanco, respectively (fig. 6). Kelsey (1990) mapped a prominent west-trending fold (the Cape Blanco anticline) that deforms late Pleistocene marine terraces at Cape Blanco. The uplift rates determined by Muhs and others (1990) in this area were measured along the axis of this anticline. The Holocene beach berm of probable coseismic origin (discussed above) is located on the south flank of this anticline, which indicates that this fold has been active in the late Holocene. Kelsey (1990) also noted several faults that offset Pleistocene marine terrace surfaces near Cape Blanco, but the apparent association of these structures with major terrane-bounding suture zones may limit their usefulness as indicators of modern stress conditions. Similar studies in the region between Cape Arago and South Slough by McInelly and Kelsey (1990) identified several northwest-trending folds (fig. 6) and flexural-slip and high-angle faults. Stratigraphic evidence from marshes in the South Slough syncline (discussed above) suggests repeated late Holocene coseismic folding and flexural-slip faulting events in this region. The results of these studies indicate that crustal shortening along the southern Oregon coast is the result, at least in part, of growth of west- and northwest-trending folds and flexure-slip and high-angle faults, and not simply a result of regional uplift.

In our study of fluvial terrace sequences along the central Oregon coast, we examined a north-trending anticline (previously described by Adams, 1984) that deformed Quaternary deposits along the Siuslaw River (fig. 7B), but we found no evidence of folds or faults with demonstrated Holocene or latest Pleistocene movement. Although we may have missed a few structures, especially where terraces are poorly preserved, the general lack of deformation on local structures indicates that the seismotectonic setting of the central Oregon coast is different than the setting of the Cape Blanco-Cape Arago-South Slough region in southern Oregon.

The recent studies of marine and fluvial terraces summarized above suggest that the southern part of the CSZ is dominated by deformation on local structures rather than by regional uplift. As suggested by Muhs and others (1990), deformation on local structures may be the most important reason that uplift rates can vary widely along active subduction zones. Variable uplift rates in Oregon and elsewhere may reflect deformation on local folds and faults rather than changes in the style of plate convergence.

The only other places where onshore Quaternary structures have been described in detail are in the Olympic Peninsula (Wilson and others, 1979) and in northern California (Kelsey and Carver, 1988; Carver and others, 1989)) at the southern end of the CSZ near the Mendocino triple junction. Although there are probably other small Quaternary structures along the coastline and in the coast ranges of Oregon and Washington, the active structures in the Cape Blanco-Cape Arago region mark a distinct change in style of Quaternary deformation along the CSZ. This change in structural style, located near 43.5° N, is probably related in part to the central coast's increased distance from the trench (figs. 2 and 6). Latest Quaternary deformation documented in the offshore record (Clarke and others, 1985; Peterson and others, 1986; Snavely, 1988) indicates the presence of a fold and thrust belt that appears to trend onshore north of Coos Bay (Peterson and Darienzo, 1989a; Darienzo, 1989). The presence of a Quaternary anticline along the Siuslaw River suggests that this area may mark the relatively inactive easternmost part of the fold and thrust belt. Another possibility is that the increasing structural complexity and activity near Coos Bay is caused by a change in plate-interface geometry related to segmentation of the Juan de Fuca plate.

REGIONAL TILTING

Landward-tilted terraces have commonly been observed along active subduction zones (for example, Ota, 1986), so geodetic and geologic evidence for regional tilting has been used as evidence for subduction at Cascadia (Reilinger and Adams 1982; Adams, 1984; Vincent and others, 1989). Reilinger and Adams (1982) noted that their geodetically measured short-term rates of east-west landward tilt of the Oregon coast were similar to long-term rates of landward tilt calculated from reconnaissance information on marine terraces along the Oregon coast. They interpreted the agreement between long and short-term rates as evidence of aseismic subduction. Adams (1984) discussed the agreement in rates in more detail and calculated crustal shortening rates for the Oregon and Washington coasts using geodetic and limited (and sometimes conflicting) geologic tilt data. He noted that the shortening rates suggested coupling of the plate interface during the Pleistocene, but concluded that there was insufficient evidence to determine whether present-day subduction was seismic or aseismic. More recent marine terrace mapping by Kelsey (1990) and McInelly and Kelsey (1990) indicates that landward tilting of marine terraces discussed by Adams (1984) in the Cape Blanco and Cape Arago regions is attributable to deformation on local folds rather than to regional deformation (fig. 6). At Cape Blanco, only the three lowest terraces have landward tilts; two higher terraces have seaward tilts (Kelsey, 1990). Similarly, our work on fluvial terraces in the Oregon Coast Range has uncovered no evidence of a regional pattern of either landward or seaward tilting as suggested by Adams (1984), but rather slow, uniform uplift of the Coast Range complicated by movement on local structures, such as the anticline on the Siuslaw River (fig 7). Therefore, shortening across structures in the North American plate in southern Oregon indicates continuing subduction along the CSZ, but terrace tilting and the agreement between geologically and geodetically determined east-west tilt rates in Oregon are not evidence for either seismic or aseismic slip on the plate interface. More recently, Vincent and others (1989) interpreted new geodetic evidence of long wavelength, north-south deformation along the Oregon coast. Repeated leveling surveys between 1941 and 1988 show a broad trough centered between the Yaquina and Tillamook estuaries, Oregon, and upward tilts southward to the Coquille River and northward to the Columbia River. Vincent and others (1989) interpreted these elevation changes as evidence of interseismic deformation due to strong interplate coupling beneath the Yaquina-Tillamook region.

POTENTIAL MAGNITUDE AND RECURRENCE OF CASCADIA PLATE-INTERFACE EARTHQUAKES

Although the interpretation and implications of the emerging paleoseismic record in the CSZ continues to be debated, the seismic histories of subduction zones that have some characteristics in common with Cascadia limit the probable magnitude and recurrence of future events in the zone. The results of recent geologic studies suggest some additional constraints on postulated rupture histories, but several different hypotheses of probable paleoearthquake magnitude and rupture extent are permitted by the available preliminary data.

CONSTRAINTS FROM OTHER MODERN SUBDUCTION ZONES

Several recent geophysical studies (Heaton and Hartzell, 1987; Rogers, 1988; Spence, 1989) conclude that portions of the plate interface in the CSZ are locked and, thus, there is a significant potential for great earthquakes (M_w 7 3/4-9 1/2). Because no plate-interface events have been recorded instrumentally on the central CSZ, the magnitude of postulated future great earthquakes seems to depend mostly on the modern analog subduction zone most favored by a particular study. Some subduction zones similar to Cascadia, such as southwest Mexico, seem to be characterized by events of M_W 7 1/4-8 1/4 that rupture 60-250 km of the zone (for example, Singh and others, 1981; Nishenko and Singh, 1987). Other zones, however, have experienced much larger events, such as the 1960 south-central Chile earthquake of M_W 9.5, which ruptured 900 km of the plate interface (Plafker and Savage, 1970; Nishenko, 1985; Cifuentes, 1989). Studies that propose south-central Chile as a good analogy for the CSZ argue that a M_W 9+ earthquake should be seriously considered (Heaton and Hartzell, 1986; Rogers, 1988). Heaton and Hartzell (1986) also suggested that the largest events in the Nankai Trough, Japan, or the Colombia subduction zone (M_W 8.5-8.8) may be the best analogies for Cascadia events. In contrast, Kanamori (1986), Rogers (1988), and Spence (1989) suggest that the 1932 Jalisco, Mexico, event (M_s 8.2) on the Rivera plate interface may be the closest analog for plate-interface events in the CSZ. But Heaton and Hartzell (1986) stated that the low level of modern seismicity at Cascadia ruled out an analogy with the southwest Mexican and other similar subduction zones. All studies of empirical correlations of subduction zone parameters and modern analogs are limited, however, because there are significant differences between the subduction setting at Cascadia and all suggested analog source zones (Kanamori and Astiz, 1985; Heaton and Hartzell, 1986; Spence, 1989).

In fact, the historic record of subduction zone events shows that, although the maximum magnitude of thrust events depends on plate parameters (Ruff, 1989), events of variable magnitude typify most subduction zones (Thatcher, 1990). For example, in the Colombian subduction zone, the M_W 8.8 earthquake of 1906 was more recently followed by three M_W 7.6-8.2 events about 20 years apart (Kanamori and McNally, 1982). Other examples of variable magnitudes are compiled in Nishenko (1989). Thatcher (1989; 1990) argues that great plate-interface earthquakes along segments of subduction zones occur in cycles and that although the magnitude of events differs from cycle to cycle, the duration of each seismic cycle is similar. Of more importance for evaluating earthquake potential, the events within each cycle seem clustered near the end of the cycle and tend to occur with increasing rupture length and magnitude in the latter part of the cycle. Reviews by Sykes and Quitmeyer (1981), Kanamori (1986), and Ruff (1989) and others show, as well, that the majority of subduction zones are characterized by events closer to M_W 8 than to M_W 9. If these generalizations developed from other subducting circum-Pacific margins apply to Cascadia, events near M_W 8 and near M_W 9 might both be possible, but events near M_W 8 would be much more common than M_W 9 events and events would tend to cluster near the end of the seismic cycle.

Although the magnitude of plate-interface earthquakes may vary considerably, the location of ruptures along a subduction zone is not random (Thatcher, 1990). Historical (for example, Ando, 1975) and geologic data (for example, Wells and others, 1988; Berryman and others, 1989; Taylor and others, 1990), as well as instrumental records (Sykes and Quitmeyer, 1981), show that subduction zones are segmented by ruptures that tend to fill seismic gaps (Ruff, 1989). Several seismic cycles can be required to fill some gaps, which suggests that asperities (zones of high shear strength on the plate interface) may control the size, timing, and starting points of ruptures (Kanamori, 1986; Thatcher, 1989; Ruff, 1989). Where asperities influence the extent of ruptures through several seismic cycles, the subduction zone can be divided into segments. Identification of potential segment boundaries along a subduction zone is of fundamental importance because the length of segments, their relationship to the structural geometry of the convergence zone, and their relationship to the recent history of ruptures along the zone set limits on the possible lateral and down-dip extent of the most probable plate-interface events.

Although segment boundaries are often associated with weak zones between asperities, major zones of faulting or other deformation that divide the subducting plate into subplates also tend to limit the extent of ruptures on the plate-interface (Ruff, 1989). At Cascadia, deformation zones dividing the north and south ends of the CSZ into subplates (fig. 1) are likely locations for segment boundaries (Spence, 1989). Changes in lower plate dip or the position of the bend within a subducting plate must also have a major effect on the lateral and down-dip extent of plate-interface ruptures (Kanamori, 1986; Spence, 1987); such changes at Cascadia have been identified at 47° and 49°N (Riddihough, 1984; Michaelson and Weaver, 1986). Hughes and others (1980) noted changes in the alignment and type of volcanoes in the Cascade range that can be projected to most of the segment boundaries in figure 1. Characteristic seismicity patterns may also serve to identify segments with differing states of stress.

Finally, the instrumental record of events in subduction zones shows that the rupture process is very complex (for example, Kanamori, 1989; Savage, 1989; Thatcher, 1990) and that simple models of segmentation and recurrence may be of limited use in accurately forecasting great earthquakes. Because the physical characteristics of the plate interface and the mechanical properties of asperities within fault segments differ, magnitude and recurrence characteristics of events in earthquake cycles differ from segment to segment (for example, Taylor and others, 1990). However, segments may persist on widely varying timescales. A major fault zone between two subplates might always halt propagating ruptures, while a diffuse zone of weakness between asperities or an asperity of moderate size might limit rupture extent only *one fourth* of the time (for example, Schwartz and Sibson, 1989). For this reason, even for subduction zones that have been characterized by historical events of relatively moderate magnitude, the rupture of 2-5 segments in a M_w 9 earthquake cannot be precluded (Heaton and Hartzell, 1986).

CONSTRAINTS FROM RECENT GEOLOGIC STUDIES IN OREGON

Although most coastal investigators now agree that M_w 8+ plate-interface events have occurred during the late Holocene in the northern half of the Juan de Fuca plate (Atwater, 1987; Grant, 1989; Peterson and Darienzo, 1989b; Peterson, 1989; Darienzo and Peterson, 1990), discussions about future events on the CSZ tend to focus on one of two end-member hypotheses for the recurrence and magnitude of late Holocene plate-interface events. The first hypothesis argues that M_w 9+ events characterize the CSZ; those events were accompanied by ruptures that extended for more than 600 km along the Juan de Fuca plate interface and parts of adjacent subplates (fig. 3A). Arguments based on possible Cascadia analogs and the rate of plate convergence indicate that recurrence of these M_w 9+ events would be many hundreds of years. The second hypothesis proposes that the CSZ is segmented in a manner similar to figure 1 and that ruptures of 100-250 km, associated with M_w 7 3/4-8 1/4 earthquakes (fig. 3C), would occur much more frequently than larger events. Only rarely, perhaps near the end of a long earthquake cycle (for example, Thatcher, 1989), would several segments rupture in an event near M_w 9 (fig. 3D). If the second hypothesis is correct, earthquake recurrence at Cascadia is apparently non-uniform because there have been no great earthquakes on the plate interface in the past 150-200 years. As discussed above, there are distinct differences in the estuarine sequences attributed to coseismic deformation along the Oregon coast (Peterson, 1989; Darienzo, 1989). We think these differences are best explained by a segmented subduction zone (most events $<M_w$ 8 1/2), but present data do not rule out larger events.

The segmented zone hypothesis requires at least 4 segment boundaries. Likely segment boundaries for the Cascadia subduction zone were suggested by Hughes and others (1980) using the distribution and alignment of different types of volcanoes in the Cascade Range and by Spence (1989) using the work of Riddihough (1984), Michaelson and Weaver (1986), and Weaver and Baker (1988) (fig. 1). Lateral offsets of some north-south trending structures in the accretionary wedge on the continental shelf (Snavely, 1988) might also reflect segment boundaries. The independent motions of the Explorer subplate and the Gorda block show that their plate contacts mark segment boundaries near the northern and southern ends of the CSZ (Spence, 1988). Less certain are differences in seismicity rates within the subducting plates and differences in plate densities and dips in the north-central part of the zone near 49°, 48.5°, and 47° N that may mark segment boundaries. The very low seismicity rate in the south-central part of the zone, our primary area of interest, makes identification of the properties of the subducting Juan de Fuca plate very difficult (for example, Keach and others, 1989), but Spence (1989) suggested that the low seismicity rate between 45°N and the Blanco fracture zone (43°N) might also characterize a separate segment of the CSZ.

A segment boundary in central Oregon (between 44-45° N; fig. 1) that would tend to limit the propagation of plate interface ruptures is one possible explanation for changes in the types of marsh stratigraphic sequences that occur between northern and south-central Oregon. For example, the thick peat sequences in the Siuslaw River estuary may mark the southern extent of late Holocene plate-interface ruptures (figs. 3B and 3C). The coseismic movements inferred from marsh sequences decrease from as much as 2.7 m of subsidence estimated for some sites in southwest Washington (Eileen Hemphill-Haley, USGS, written comm., 1989) to events accompanied by 0.5-1 m subsidence estimated by Peterson and Darienzo (1989b) in Alsea Bay. Thus, the Alsea Bay sequences may record the southernmost extent of plate-interface ruptures that are hundreds of kilometers long (Darienzo, 1989; fig. 3B). Alternatively, the central Oregon coast may have experienced only relatively short ruptures (most <100-200 km), produced by events of small slip (fig. 3C), perhaps with a cycle-ending event of 250-300 km (for example, Thatcher, 1990). In this case, the size and recurrence of events in central Oregon would differ from those in northern Oregon, but the Siuslaw sequences would still mark the southern limit of the long, large-slip ruptures (>0.5 m coseismic subsidence; fig. 3C).

An explanation for the change in the types of marsh sequences that is consistent with the M_w 9-event hypothesis discussed above is related to the width and location of the zone of subsidence inferred to have been produced during repeated great (M_w 9+) earthquakes (see Atwater, 1988d). If the western edge of the zone of subsidence for all these postulated great earthquakes was located between Alsea Bay (107 km east of the trench) and the Siuslaw estuary (104 km from trench; figs. 2 and 3A) only the Alsea Bay sequences would be expected to show evidence of repeated coseismic subsidence. Of course the length, width, and location of the zone of subsidence varies from event to event, but it must always have been east of the Siuslaw estuary, which shows no evidence of major coseismic subsidence events.

The differences in the types of marsh records from the Siuslaw and Alsea sites suggest that these sites are probably on separate late Holocene segments of the CSZ. Both sites are about the same distance from the Cascadia trench (fig. 2) and the ages of sediments in cores from both estuaries probably span at least the last 2-5 seismic cycles (2,000-3,000 yr). Thus, most events recorded at Alsea Bay should have been recorded at the Siuslaw site if the ruptures on the plate interface during these events extended throughout this area. Possibly some long ruptures extended far south of the Siuslaw (for example, fig. 3D), but if so, the zone of coseismic subsidence was inland of the coast and most of these long-rupture events were not recorded at Alsea Bay. Unfortunately, the width of the zone of subsidence and its distance from the trench during historic great earthquakes in other subduction zones are sufficiently variable (Atwater, 1988d; West and McCrumb, 1988b) that none of the possible rupture modes in figure 3 can be ruled out using arguments based on analogies with other subduction zones.

It also is unfortunate that stratigraphic studies of the marsh record may not be capable of conclusively distinguishing between the two end-member hypotheses (figs. 3A and 3C). Marshes of sufficient size to allow extensive coring along transects are needed to document coseismic changes in sea level (for example, Atwater, 1987), but in central Oregon, estuaries with sufficient areas of marsh are spaced 28-50 km apart. Furthermore, in significant portions of these estuaries, preserved sequences consist mostly of sandy tidal channel, eolian, and fluvial sediments in which the record of small sea level changes is obscure. The size of these river-dominated estuaries (Peterson and others, 1984) and the distribution of marshes along them also limits to <8 km the length of the east-west trending zone over which clear evidence of sea level changes can be identified. Thus, only 4-5 estuaries in central Oregon are likely to yield definitive data on coseismic changes in sea level. The distances of potential core sites in these estuaries from the Cascadia trench do not differ enough to distinguish between the two end-member hypotheses discussed above, and there are not enough widely-spaced sites within each estuary to determine differences in the amount of coseismic subsidence along east-west transects. Such differences might be used to define better the position of the zone of subsidence for individual events.

In contrast to the northern and central Oregon coast, the earthquake record of the Coos Bay region appears to be a composite of regional and local events. Here the active fold and thrust belt of the accretionary wedge is well-documented by onshore (McInelly and Kelsey, 1990; Kelsey, 1990) and offshore studies (Clarke and others, 1985). Within this tectonic framework we ascribe the multiple, abruptly buried marsh peats in the South Slough syncline to coseismic subsidence due to local faulting and folding of the syncline or adjacent anticlines (McInelly and others, 1989; Peterson, 1989). The scattered occurrence of other abruptly buried peaty soils elsewhere in the Coos Bay area may also be due to local displacements on other structures. Lesser amounts of local coseismic deformation could extend as far north as the Siuslaw River (Peterson and Darienzo, 1989a), although we have as yet found no evidence of Holocene coseismic deformation there. Although deformation in the Coos Bay region is probably controlled by local structures, plate-interface events are not precluded because movements on local structures often occur during plate-interface events (for example, Plafker, 1969; Sykes, 1989) and the local coseismic deformation could be overprinting regional deformation (Nelson, 1987; McInelly and others, 1989; Peterson and Darienzo, 1989a).

A segment boundary north of the Siuslaw River in central Oregon would indicate that the 180 to 230-km-long part of the central CSZ from that boundary south to Cape Blanco consists of one or possibly two segments (see, Peterson and Darienzo, 1989a). For example, between Coos Bay and the Siuslaw River, the submerged mouths of river valleys and the relatively low elevation of Pleistocene marine terraces suggests that this part of the coast has been uplifted at a lower rate than areas to the north and south (Adams, 1984). If this zone of relative submergence is reflecting north-south changes in the properties of the subducting plate, then it could mark a segment boundary. However, because this zone of submergence coincides with the eastern edge of the fold and thrust belt, it is more likely to be related to an east-west structural transition in the overiding North American plate (Peterson and Darienzo, 1989a). The change in structural style shown by the marine and fluvial terrace studies probably reflects the same transition.

Determining paleoearthquake recurrence in the fold-and-thrust-belt region of southern Oregon is complicated by the probable composite nature of the paleoseismic record. Analogies with similar areas (Berryman and others, 1989; Carver and others, 1989) suggest some local events in the Coos Bay region may have been of small magnitudes (M_s <7) and were not necessarily triggered by larger plate-interface events (see however, Sykes, 1989). As argued by Yeats and others (1981) in southern California, events like those recorded in the South Slough syncline may be very shallow, quite possibly the result of M_s <6 events on flexural-slip faults during coseismic folding. Unfortunately, without a means of precisely dating or directly correlating individual events, the possibility seems remote that marsh stratigraphic methods can be used consistently to distinguish local from regional events on a segment or segments of the CSZ between the Siuslaw River and Cape Blanco.

CONCLUSIONS

Field efforts of the last few years have produced convincing evidence of coseismic changes in land level along the Washington and Oregon coasts. Studies of coastal deposits suggest that regional plate-interface events of at least M_W 8 (>100-km-long ruptures) have occurred in northern Oregon and southwestern Washington.

Differences in the types of marsh sequences between north and south-central Oregon are consistent with several hypotheses for the recurrence and magnitude of plate-interface events. We suspect a segment boundary exists along the CSZ near 44-45° N; such a boundary would suggest that the CSZ is segmented and, therefore, that plate-interface events near M_W 8 along the central CSZ are more frequent than larger events. Alternatively, the western edge of the zone of coseismic subsidence during larger, more infrequent (M_W 9) great plate-interface earthquakes may intersect the coast in central Oregon and could produce the observed stratigraphic changes in marsh sequences.

Geophysical and structural studies and mapping of marine and fluvial terraces indicate that many episodes of abrupt marsh burial in south-central Oregon are best interpreted as the product of deformation on local structures. Some of the local deformation could be associated with earthquakes of very moderate magnitude (M_s<6). At most sites in this area, however, it is still unclear whether coseismic events were responses to local faulting or folding, to regional deformation during great plate-interface earthquakes, or to both. Advances in dating technology and extensive mapping of Pleistocene structures will be needed to distinguish regional and local events.

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REFERENCES CITED

Acharya, H., 1981, Juan de Fuca plate--Aseismic subduction at 1.8 cm/yr: Geophysical Research Letters, v. 8, p. 1123-1125.

Adams, John, 1984, Active deformation of the Pacific northwest continental margin: Tectonics, v. 3, n. 4, p. 449-472.

Adams, John, 1989, Turbidites off the Oregon-Washington margin record paleo-earthquakes on the Cascadia subduction zone [abs.]: Seismological Research Letters, v. 60, no. 1, p. 1.

Adams, John, 1990, Paleoseismicity of the Cascadia subduction zone--Evidence from turbidites off the Oregon-Washington margin: Tectonics, v. 9, p. 569-583.

Ando, Masataka, 1975, Source mechanisms and tectonic significance of historical earthquakes along the Nankai Trough, Japan: Tectonophysics, v. 27, p. 119-140.

Ando, Masataka, and Balazs, E. I., 1979, Geodetic evidence for aseismic subduction of the Juan de Fuca plate: Journal of Geophysical Research, v. 84, p. 3023-3027.

Atwater, B. F., 1987, Evidence for great Holocene earthquakes along the outer coast of Washington state: Science, v. 236, p. 942-944.

Atwater, B.F., 1988a, Probable local precedent for earthquakes of magnitude 8 or 9 in the Pacific Northwest, *in*, Workshop on "Evaluation of earthquake hazards and risk in the Puget Sound and Portland areas", Proceedings of Conference XLII: U.S. Geological Survey Open-File Report 88-541, p. 62-68.

Atwater, B.F., 1988b, Geologic studies for seismic zonation for the Puget lowland, *in*, Jacobson, M.L., and Rodriquez, T.R., eds., National Earthquake Hazards Reduction Program, Summaries of Technical Reports XXV, U.S. Geological Survey Open-File Report 88-16, p. 120-133.

Atwater, B.F., 1988c, Subduction-earthquake telltales beneath coastal lowlands: *in*, Crone, A.J., and Omdahl, E.M., eds., Directions in paleoseismology, Proceedings of Conference XXXIX, U.S. Geological Survey Open-File Report 87-673, p. 157-162.
Atwater, B.F., 1988d, Comment on "Coastline uplift in Oregon and Washington and the nature of Cascadia subduction-zone tectonics": Geology, v. 16, p. 952-953.
Atwater, B.F., 1988e, Buried Holocene wetlands along the Johns River, southwest Washington [abs.]: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 4.
Atwater, B. F., and Grant, W. C., 1986, Holocene subduction earthquakes in coastal Washington: EOS, Transactions of American Geophysical Union, v. 67, no. 44, p. 906.
Bartsch-Winkler, S.B., 1988, Cycle of earthquake-induced aggradation and related tidal channel shifting, upper Turnagain Arm, Alaska, USA: Sedimentology, v. 35, p. 621-628.
Bartsch-Winkler, S.B., Ovenshine, A. T., and Kachadoorian, Reuben, 1983, Holocene history of the estuarine area surrounding Portage, Alaska, as recorded in a 93-meter core: Canadian Journal of Earth Sciences, v. 20, p. 802-820.
Berendsen, H.J.A., 1984, Quantitative analysis of radiocarbon dates of the perimarine area in the Netherlands: Geologie en Mijnbouw, v. 63, p. 343-350.
Berryman, K.R., 1987, Tectonic processes and their impact on the recording of relative sea-level changes: *in*, Devoy, R.J.N., eds., Sea surface studies--A global view, London, Croom Helm, p. 127-161.
Berryman, K.R., Ota, Yoko, and Hull, A.G., 1989, Holocene paleoseismicity in the fold and thrust belt of the Hikurangi subduction zone, eastern North Island, New Zealand: Tectonophysics, v. 163, p. 185-195.
Bloom, A.L., 1980, Late Quaternary sea level change on South Pacific coast--A study in tectonic diversity: Earth Rheology, Isostasty and Eustasy, *in*, Morner, N.-A., ed., New York, John Wiley & Sons, p. 505-516.
Bourgeois, Joanne, and Reinhart, M.A., 1988, Potentially damaging waves associated with earthquakes, coastal Washington, *in*, Hayes, W.W., ed., Workshop on "Evaluation of earthquake hazards and risk in the Puget Sound and Portland areas", Proceedings of Conference XLII: U.S. Geological Survey Open-File Report 88-541, p. 96-99.
Brooks, M.J., Colquhoun, D.J., Pardi, R.R., Newman, W.S., and Abbott, W.H., 1979, Preliminary archeological and geological evidence for Holocene sea level fluctuations in the lower Cooper River valley, South Carolina: The Florida Anthropologist, v. 32, p. 85-103.
Carver, G.A., and Burke, R.M., 1987, Late Holocene paleoseismicity of the southern end of the Cascadia subduction zone [abs.]: EOS, v. 68, p. 1240.
Carver, G.A., Vick, G.S., and Burke, R.M., 1989, Late Holocene paleoseismicity of the Gorda segment of the Cascadia subduction zone: Geological Society of America, Abstracts with Programs, v. 21, no. 5, p. 64.
Cifuentes, I.L., 1989, The 1960 Chilean earthquakes: Journal of Geophysical Research, v. 94, p. 665-680.
Clarke, S.H., Field, M.E., and Hirozawa, C.A., 1985, Reconnaissance geology and geologic hazards of the offshore Coos Bay basin, Oregon: U.S. Geological Survey Bulletin 1645, 41 p.
Clarke, Jr., S.H., and Carver, G.A., 1989, Late Cenozoic structure and seismic potential of the southern Cascadia subduction zone: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331-1332.
Combellick, R.A., 1986, Chronology of late-Holocene earthquakes in south-central Alaska: Evidence from buried organic soils in Turnagain Arm: Geological Society of America Abstracts with Programs, v. 18, no. 6, p. 569.
Crosson, R.S., and Owens, T.J., 1987, Slab geometry of the Cascadia subduction zone beneath Washington from earthquake hypocenters and teleseismic converted waves: Geophysical Research Letters, v. 14, p. 824-827.
Cullingford, R.A., Caseldine, C.J., Gotts, P.E., 1989, Evidence of early Flandrian tidal surges in Lower Strathearn, Scotland: Journal of Quaternary Science, v. 4, p. 51-60.
Darienzo, M.E., 1987, Late Holocene geologic history of a Netarts Bay salt marsh, northwest Oregon coast, and its relationship to relative sea level changes: unpub. M.S. thesis, Interdisciplinary Studies Program, University of Oregon, Eugene, 94 p.
Darienzo, M.E., 1989, Evidence for and implications of small-scale (<1 m) tecctonic subsidence in salt marshes of Alsea Bay, Oregon, central Cascadia margin: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331.
Darienzo, M.E., and Peterson, C.D.., 1987, Episodic tectonic subsidence recorded in late-Holocene salt marshes, northwest Oregon: EOS, Transactions, American Geophysical Union, v. 68, no. 44, p. 1469.
Darienzo, M.E., and Peterson, C.D.., 1990, Episodic tectonic subsidence of late Holocene salt marshes, northern Oregon coast, central Cascadia margin, U.S.A.: Tectonics, v. 9, p. 1-22.
Devoy, R.J.N., ed., 1987, Sea surface studies--A global view: London, Croom Helm, 649 p..

Fairbridge, R.W., 1987, The spectra of sea level in a Holocene time frame, *in*, Rampino, M.R., Sanders, J.E., Newman, W.S., and Konigsson, L.K., eds., Climate--History, periodicity, and predictability: New York, Van Nostrand Reinhold, 127-141 p.
Golder Associates, 1986, WNP-3 Geologic support services, coastal terrace study: unpublished report prepared for Washington Public Power Supply System, Richland, Washington, 106 p.
Grant, W.C., 1988, Evidence for late Holocene subduction earthquakes recorded in tidal marsh deposits along the Nehalem and Salmon Rivers, northern Oregon [abs.]: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 10.
Grant, W.C., 1989, More evidence from tidal-marsh stratigraphy for multiple late Holocene subduction earthquakes along the northern Oregon coast: Geological Society of America Abstracts with Programs, v. 21, n. 5, p. 86.
Grant, W.C., and McLaren, D.D., 1987, Evidence for Holocene subduction earthquakes along the northern Oregon coast [abs.]: EOS, Transactions, American Geophysical Union, v. 68, no. 44, p. 1239.
Grant, W.C., Atwater, B.F., Carver, G.A., Darienzo, M.E., Nelson, A.R., Peterson, C.D., and Vick, G.S., 1989, Radiocarbon dating of late Holocene coastal subsidence above the Cascadia subduction zone--Compilation for Washington, Oregon, and northern California: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331.
Griggs, A.B., 1945, Chromite-bearing sands of the southern part of the coast of Oregon: U.S. Geological Survey Bulletin 945-E, p. 113-150.
Heaton, T. H., and Kanamori, Hiroo, 1984, Seismic potential associated with subduction in the northwestern United States: Bulletin of the Seismological Society of America, v. 74, no. 3, p. 933-941.
Heaton, T. H., and Hartzell, S. H., 1986, Source characteristics of hypothetical subduction earthquakes in the northwestern United States: Bulletin of the Seismological Society of America, v. 76, no. 3, p. 675-708.
Heaton, T. H., and Hartzell, S. H., 1987, Earthquake hazards on the Cascadia subduction zone: Science, v. 236, p. 162-168.
Hemphill-Haley, Eileen, 1989, Investigation into the use of estuarine microfossil zones for determining amounts of coseismic subsidence in southwestern Washington: Geological Society of America Abstracts with Programs, v. 21, n. 5, p. 92.
Hsu, G.J.T., Leonard, E.M., and Wehmiller, J.F., 1989, Aminostratigraphy of Peruvian and Chilean Quaternary marine terraces: Quaternary Science Reviews, v. 8, 255-262.
Hughes, J.M., Stoiber, R.E., and Carr, M.J., 1980, Segmentation of the Cascade volcanic chain: Geology, v. 8, p. 15-17.
Hull, A.G., 1987, Buried lowland soils from Willapa Bay, southwest Washington: Further evidence for recurrence of large earthquakes during the last 5000 years: EOS, Transactions, American Geophysical Union, v. 68, no. 44, p. 1468-1469.
Hull, A.G., 1988, Radiocarbon age of probable coseismic buried soil layers from Washington State: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 7.
Jelgersma, S., 1961, Holocene sea-level changes in The Netherlands: Meded. Geol. Sticht. Serie C, v. VI, 7, p. 1-100.
Kaizuka, S., Matsuda, T, Nogomi, M., and Yonekura, N., 1973, Quaternary tectonic and recent seismic crustal movements in the Arauco Peninsula and its environs, central Chile: Geographical Reports of Tokyo Metropolitan University, v. 8, p. 1-49.
Kanamori, Hiroo, 1986, Rupture process of subduction-zone earthquakes: Annual Reviews of Earth and Planetary Sciences, v. 14, p. 293-322.
Kanamori, Hiroo, 1989, Variable rupture mode of subduction zones [abs.]: U.S. Geological Survey Workshop on Fault Segmentation and Controls of Rupture Initiation and Termination, Abstracts of meeting, March 7-9, 1989.
Kanamori, Hiroo, and Astiz, L., 1985, The 1983 Akita-Oki earthquake (M_w 7.8) and its implications for systematics of subduction earthquakes: Earthquake Prediction Research, v. 3, p. 305-317.
Kanamori, Hiroo, and McNally, K.C., 1982, Variable rupture mode of the subduction zone along the Ecuador-Colombia coast: Bulletin of the Seismological Society of America, v. 72, p. 1241-1253.
Kaye, C.A., and Barghoorn, E.S., 1964, Late Quaternary sea-level change and crustal rise at Boston, Massachusetts with note on the autocompaction of peat: Geological Society of America Bulletin, v. 75, p. 63-80.
Keach, R.W., Oliver, J.E., Brown, L.D., and Kaufman, S., 1989, Cenozoic active margin and shallow Cascades structure--COCORP results from western Oregon: Geological Society of America Bulletin, v. 101, p. 783-794.
Kearney, M.S., and Ward, L.G., 1986, Accretion rates in brackish marshes of a Chesapeake Bay estuarine tributary: Geo-Marine Letters, v. 6, p. 41-49.
Kelsey, H.M., 1989, 3000-year-old elevated beach berm south of Cape Blanco, Oregon--Probable evidence of late Holocene tectonic uplift: Geological Society of America Abstracts with Programs, v. 21, n. 5, p. 101.

Kelsey, H.M., 1990, Late Quaternary deformation of marine terraces on the Cascadia subduction zone near Cape Blanco, Oregon: Tectonics, v. 9, p.983-1014.
Kelsey, H.M., and Carver, G.A., 1988, Late Neogene and Quaternary tectonics associated with northward growth of the San Andreas transform fault, northern California: Journal of Geophysical Research, v. 93, no. B5, p. 4797-4819.
Kennedy, G.L., Lajoie, K.R., and Wehmiller, J.F., 1982, Aminostratigraphy and faunal correlations of late Quaternary marine terraces, Pacific coast, USA: Nature, v. 299, p. 545-547.
Kidson, C., 1982, Sea level changes in the Holocene: Quaternary Science Reviews, v. 1, p. 121-151.
Kulm, L.D., 1989, Cascadia subduction zone--Structure, tectonics, and fluid processes of the accretionary wedge and adjacent abyssal plain: *in*, Hays, W.W., ed., 3rd annual workshop on "Earthquake hazards in the Puget Sound, Portland area", Proceedings of Conference XLVIII, U.S. Geological Survey Open-File Report 89-465, p. 38-39.
Lajoie, K.R., 1986, Coastal tectonics, *in* Active Tectonics: Washington D.C., National Academy Press, p. 95-124.
Lajoie, K.R., Sarna-Wojcicki, A.M., and Ota, Yoko, 1982, Emergent Holocene marine terraces at Ventura and Cape Mendocino, California--Indicators of high tectonic uplift rates: Geological Society of America Abstracts with Programs, v.14, p. 178.
Magee, M.E., and Zoback, M.L., 1989, Present state of stress in the Pacific Northwest: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1332-1333.
Matsuda, T., Ota, Yoko, Ando, M., and Yonekura, N., 1978, Fault mechanism and recurrence time of major earthquakes in southern Kanto district, Japan, as deduced from coastal terrace data: Geological Society of America Bulletin, v. 89, p. 1610-1618.
McInelly, G.W., and Kelsey, H.M., 1990, Late Quaternary tectonic deformation in the Cape Arago-Bandon region of coastal Oregon as deduced from wave-cut platforms: Journal of Geophysical Research, v. 95, no. B5, p. 6699-6713.
McInelly, G.W., Kelsey, H.M., and Nelson, A.R., 1989, Late Pleistocene and Holocene tectonic deformation of the Whiskey Run wave-cut platform in the Cape Arago-Coos Bay area, coastal Oregon: Geological Society of America Abstracts with Programs, v. 21, n. 5, p. 115.
Michaelson, C.A., and Weaver, C.S., 1986, Upper mantle structure from teleseismic P wave arrivals in Washington and northern Oregon: Journal of Geophysical Research, v. 91, p. 2077-2094.
Merritts, D., and Vincent, K.R., 1989, Geomorphic response of coastal streams to low, intermediate, and high rates of uplift, Medocino triple junction region, northern California: Geological Society of America Bulletin, v. 101, p. 1373-1388.
Muhs, D.R., Kelsey, H.M., Miller, G.H., Kennedy, G.L., Whelan, J.F., and McInelly, G.W., 1990,Age estimates and uplift rates for late Pleistocene marine terraces--Southern Oregon portion of the Cascadia forearc: Journal of Geophysical Research, v. 95, no. B5, p. 6685-6698.
Nelson, A.R., 1987, Apparent gradual rise in relative sea level on the south-central Oregon coast during the late Holocene--Implications for the great Cascadia earthquake hypothesis [abs.]: EOS, v. 68, p. 1240.
Nelson, A.R., 1988, Implications of late Holocene salt-marsh stratigraphy for earthquake recurrence along the coast of south-central Oregon, *in*, Hayes, W.W., ed., Workshop on "Evaluation of Earthquake Hazards and Risk in the Puget Sound and Portland Areas", Proceedings of Conference XLII: U.S. Geological Survey Open-File Report 88-541, p. 149-151.
Nelson, A. R., Atwater, B. F., and Grant, W. C., 1987, Estuarine record of Holocene subduction earthquakes in coastal Oregon and Washington, USA: International Union for Quaternary Research, 12th International Congress, Ottawa, Canada, July 31-August 9, 1987, Programme with Abstracts, p. 231.
Newman, W.S., Cinquemani, L.J., Pardi, R.P., and Marcus, L.F., 1980, Holocene deleveling of the United States east coast: *in*, Morner, N.-A., ed., Earth rheology, isostasy, and eustasy, New York, John Wiley, p. 449-463.
Nishenko, S. P., 1985, Seismic potential for large and great interplate earthquakes along the Chilean and southern Peruvian margins of South America: A quantitative reappraisal: Journal of Geophysical Research, v. 90, no. B5, p. 3589-3615.
Nishenko, S.P., 1989, Circum-Pacific seismic potential 1989-1999: U.S. Geological Survey Open File Report 89-86, v. 89-86, 126 p.
Nishenko, S.P., and Singh, S.K., 1987, Conditional probabilities for the recurrence of large and great interplate earthquakes along the Mexican subduction zone: Bulletin of the Seismological Society of America, v. 77, p. 2095-2114.
Nishimura, C., Wilson, D.S., Hey, R.N., 1984, Pole rotation analysis of present-day Juan de Fuca plate motion: Journal of Geophysical Research, v. 89, no. B10, p. 10,283-10,290.
Obermeier, S.F., Weems, R.E., Jacobson, R.B., and Gohn, G.S., 1989, Liquefaction evidence for repeated Holocene earthquakes in the coastal region of South Carolina: Annals New York Academy of Science, v. 558, p. 183-195.

Orson, Richard, Panageotou, William, and Leatherman, S.P., 1985, Response of tidal salt marshes of the U.S. Atlantic and Gulf coasts to rising sea levels: Journal of Coastal Research, v. 1, p. 29-37.
Ota, Yoko, 1986, Marine terraces as reference surfaces in late Quaternary tectonics studies: Examples from the Pacific Rim: *in*, Reilly, W.I., and Harford, B.E., eds., Recent crustal movements of the Pacific region, Private Bag, New Zealand, The Royal Society of New Zealand Bulletin 24, Publication no. 0102, International Lithosphere Program, p. 357-375.
Ota, Yoko, and Yoshikawa, T., 1978, Regional characteristics and their geodynamic implications of late Quaternary tectonic movement deduced from deformed former shorelines in Japan: Journal of Physics of the Earth, v. 26, Supplement, p. s379-s389.
Ovenshine, A.T., Lawson, D.E., and Bartsch-Winkler, S.R., 1976, The Placer River Silt-an intertidal deposit caused by the 1964 Alaska Earthquake: U.S. Geological Survey Journal of Research, v. 4, p. 151-162.
Pearson, G.W., 1979, Precise ^{14}C measurement by liquid scintillation counting: Radiocarbon, v. 21, p. 1-21.
Peterson, C.D., 1989, Potential evidence of subduction zone tectonics from stacked peat horizons in late Pleistocene coastal terraces of the northern Cascadia margin: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331.
Peterson, C.D., and Darienzo, M.E., 1988a, Coastal neotectonic field trip guide for Netarts Bay, Oregon: Oregon Geology, v. 50, no. 9/10, p. 99-106.
Peterson, C.D., and Darienzo, M.E., 1988b, Episodic tectonic subsidence of late-Holocene salt marshes in Oregon--Clear evidence of abrupt strain release and gradual strain accumulation in the southern Cascadia margin during the last 3,500 years, *in*, Hayes, W.W., ed., Workshop on "Evaluation of earthquake hazards and risk in the Puget Sound and Portland areas", Proceedings of Conference XLII: U.S. Geological Survey Open-File Report 88-541, p. 110-113.
Peterson, C.D., and Darienzo, M.E., 1988c, Discrimination of flood, storm, and tectonic events in coastal marsh records of the southern Cascadia margin [abs.]: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 11.
Peterson, C.D., and Darienzo, M.E., 1989a, Episodic abrupt tectonic subsidence recorded in late Holocene deposits of the South Slough syncline--An on-land expression of shelf fold belt deformation from the southern Cascadia margin: Geological Society of America Abstracts with Programs, v. 21, n. 5, p. 129.
Peterson, C.D., and Darienzo, M.E., 1989b, Discrimination of climatic, oceanic, and tectonic forcing of marsh burial events from Alsea Bay, Oregon, U.S.A.: final [unpub.] report (Part B) to the National Earthquake Hazards Reduction Program, U.S. Geological Survey, 33 p. (Peterson and Darienzo, this volume, is an expanded version of this work.).
Peterson, C.D., Scheidegger, K.F., and Schrader, K.J., 1984, Holocene depositional evolution of a small active margin estuary of the Northwestern United States: Marine Geology, v. 59, p. 51-83.
Peterson, C.L., Kulm, L.D., and Gray, J.J., 1986, Geologic map of the ocean floor off Oregon and the adjacent continental margin: Oregon Department of Geology and Mineral Industries, Geologic Map Series, Map GMS-42.
Phipps, J.B., and Peterson, C.D., 1989, General submergence of Grays Harbor, Washington during the Holocene: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1332.
Plafker, George, 1965, Tectonic deformation associated with the 1964 Alaska earthquake: Science, 148, p. 1675-1687.
Plafker, George, 1969, Surface faults on Montague Island associated with the 1964 Alaska earthquake: U.S. Geological Survey Professional Paper 543-G, p. G1-G42.
Plafker, George, 1972, Alaskan earthquake of 1964 and Chilean earthquake of 1960--Implications for arc tectonics: Journal of Geophysical Research, v. 77, p. 901-924.
Plafker, George, and Savage, J. C., 1970, Mechanism of the Chilean earthquakes of May 21 and 22, 1960: Geological Society of America Bulletin, v. 81, p. 1001-1030.
Plafker, George, and Rubin, M., 1978, Uplift history and earthquake recurrence as deduced from marine terraces on Middleton Island, Alaska, *in* Methodology for identifying seismic gaps and soon-to-break gaps: U.S. Geological Survey Open-File Report 78-943, p. 687-721.
Plassche, O., van de, 1980, Holocene water-level changes in the Rhine-Meuse Delta as a function of changes in relative sea level, local tidal range, and river gradient: Geologie en Mijnbouw, v. 59, p. 343-351.
Plassche, O., van de, 1982, Sea-level change and water-level movements in the Netherlands during the Holocene: Mededelingen rijks geologische dienst, v. 36-1, 93 p.
Rampino, M.R., and Sanders, J.E., 1981, Episodic growth of Holocene tidal marshes in the northeastern United States--A possible indicator of eustatic sea-level fluctuations: Geology, v. 9, p. 63-67.
Rasmussen, J.R., and Humphreys, Eugene, 1989, Tomographic image of the upper mantle P-wave velocity structure beneath Washington and western Oregon: EOS, Transactions, American Geophysical Union, v. 70, no. 43, p. 1329.

Reilinger, R. E., and Adams, John, 1982, Geodetic evidence for active landward tilting of the Oregon and Washington coastal ranges: Geophysical Research Letters, v. 9, p. 401-403.
Reinhart, M.A., and Bourgeois, Joanne, 1987, Distribution of anomalous sand at Willapa Bay, Washington: Evidence for large-scale landward-directed processes [abs.]: EOS, Transactions, American Geophysical Union, v. 68, no. 44, p. 1469.
Reinhart, M.A., and Bourgeois, Joanne, 1988, Testing the tsunamis hypothesis at Willapa Bay, Washington--Evidence for large-scale landward-directed processes [abs.]: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 9.
Reinhart, M.A., and Bourgeois, Joanne, 1989, Tsunami favored over storm or seiche for sand deposit overlying buried Holocene peat, Willapa Bay, WA: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331.
Rhea, Susan, 1989, Evidence of uplift near Charleston, South Carolina: Geology, v. 17, p. 311-315.
Riddihough, R.P., 1984, Recent movements of the Juan de Fuca plate system: Journal of Geophysical Research, v. 89, no. B8, p. 6980-6994.
Rogers, G.C., 1988, An assessment of the megathrust earthquake potential of the Cascadia subduction zone: Canadian Journal of Earth Sciences, v. 25, p. 844-852.
Ruff, L.J., 1989, Do trench sediments affect great earthquake occurrence in subduction zones?: Pageoph, v. 129, p. 263-282.
Savage, J.C., 1983, A dislocation model of strain accumulation and release at a subduction zone: Journal of Geophysical Research, v. 88, p. 4984-4996.
Savage, J.C., 1989, Aseismic slip in seismic gaps at subduction zones: Seismological Research Letters, v. 60, no. 1, p. 2.
Savage, J.C., Lisowski, M., and Prescott, W.H., 1981, Geodetic strain measurements in Washington: Journal of Geophysical Research, v. 86, p. 4929-4940.
Schwartz, D.P., and Sibson, R.H., eds., 1989, Fault segmentation and controls of rupture initiation and termination, Proceedings of Conference XLV, National Earthquake Hazards Reduction Program: U.S. Geological Survey Open-File Report 89-315, 447 p.
Shedlock, K.M., and Weaver, C.S., 1989, Rationale and outline of a program for earthquake hazards assessment in the Pacific Northwest: *in*, Hays, W.W., ed., 3rd annual workshop on "Earthquake hazards in the Puget Sound, Portland area", Proceedings of Conference XLVIII, U.S. Geological Survey Open-File Report 89-465, p. 2-9.
Shennan, Ian, 1982, Problems of correlating Flandrian sea-level changes and climate, *in*, Harding, A.F., ed., Climatic change in later prehistory: Edinburgh, Edinburgh University Press, p. 52-67.
Shennan, Ian, 1986, Flandrian sea-level changes in the Fenland. II--Tendencies of sea-level movement, altitudinal changes, and local and regional factors: Journal of Quaternary Science, v. 1, no. 2, p. 155-180.
Shennan, Ian, 1987, Global analysis and correlation of sea-level data, *in*, Devoy, R.J.N., ed., Sea surface studies--A global view: London, Croom Helm, p. 198-230.
Shennan, Ian, 1989, Holocene sea-level changes and crustal movements in the North Sea region--An experiment with regional eustasy: *in*, Scott, D.B., Pirazzoli, P.A., and Honig, C.A., eds., Late Quaternary sea-level correlation and applications, Dordrecht, Kluwer Academic Publishers, p. 1-25.
Singh, S.K., Astiz, L., Havskov, J., 1981, Seismic gaps and recurrence periods of large earthquakes along the Mexican subduction zone: a reexamination: Bulletin of the Seismological Society of America, v. 71, p. 827-843.
Snavely, P. D., Jr., 1988, Tertiary geologic framework, neotectonics, and petroleum potential of the Oregon-Washington continental margin: *in*, Scholl, D.S., Grantz, A., and Vedder, J.G., eds., Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, v. 6, p. 305-335.
Spence, William, 1987, Slab pull and the seismotectonics of subducting lithosphere: Reviews of Geophysics, v. 25, p. 55-69.
Spence, William, 1988, Anomalous subduction and the origins of stresses at Cascadia--A review, *in*, Hayes, W.W., ed., Workshop on "Evaluation of earthquake hazards and risk in the Puget Sound and Portland areas", Proceedings of Conference XLII: U.S. Geological Survey Open-File Report 88-541, p. 114-148.
Spence, William, 1989, Stress origins and earthquake potentials in Cascadia: Journal of Geophysical Research, v. 94, p. 3076-3088.
Stevenson, J.C., Ward, L.G., and Kearney, M.S., 1986, Vertical accretion in marshes with varying rates of sea level rise, *in*, Wolfe, D.A., ed., Estuarine variability: Academic Press, New York, p. 241-259.
Streif, Hansjorg, 1980, Cyclic formation of coastal deposits and their indications of vertical sea-level changes: Oceanis, v. 5, p. 303-306.
Stuiver, Minze, and Pearson, G. W., 1986, High-precision calibration of the radiocarbon time scale, AD 1950-500 BC: Radiocarbon, v. 28, p. 805-838.

Sykes, L.R., 1989, Great earthquakes of 1855 and 1931 in New Zealand--Evidence for seismic slip along downgoing plate boundary and implications for seismic potential of Cascadia subduction zone: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1331.
Sykes, L. R., and Quitmeyer, R. C., 1981, Repeat times of great earthquakes along simple plate boundaries: *in*, Simpson, D.W., and Richards, P.G., eds., Earthquake prediction: An international review, Maurice Ewing Series, v. 4, Washington, D. C., American Geophysical Union, p. 217-247.
Taylor, F.W., Jouannic, C., and Bloom, A.L., 1985, Quaternary uplift of the Torres Islands, northern New Hebrides frontal arc--Comparison with Santo and Malekula Islands, central New Hebrides frontal arc: Journal of Geology, v. 93, p. 419-438.
Taylor, F.W., Edwards, R.L., Wasserberg, G.J., and Frohlich, Cliff, 1990, Seismic recurrence intervals and timing of aseismic subduction inferred from emerged corals and reefs of the central Vanuatu (New Hebrides) frontal arc: Journal of Geophysical Research, v. 95, no. B1, p. 393-408.
Ters, Mireille, 1987, Variations in Holocene sea level on the French Atlantic coast and their climatic significance, *in*, Rampino, M.R., Sanders, J.E., Newman, W.S., and Konigsson, L.K., eds., Climate--History, periodicity, and predictability: New York, Van Nostrand Reinhold, p. 204-237.
Thatcher, Wayne, 1984, The earthquake deformation cycle, recurrence, and the time-predictable model: Journal of Geophysical Research, v. 89, no. B7, p. 5674-5680.
Thatcher, Wayne, 1989, Earthquake recurrence and risk assessment in circum-Pacific seismic gaps: Nature, v. 341, p. 432-434.
Thatcher, Wayne, 1990, Order and diversity in the modes of circum-Pacific earthquake recurrence: Journal of Geophysical Research, v. 95, no. B3, p. 2609-2623.
Thomas, Ellen, Varekamp, J.C., and Plassche, O. van de, 1989, Late Holocene fluctuations in sea level rise (Clinton, CT)--Marsh foraminifera and geochemical evidence: Geological Society of America Abstracts with Programs, v. 21, no. 2, p. 70.
Tooley, M.J., ed., 1978, Sea-level Changes in North-west England during the Flandrian Stage: Oxford, U.K., Clarendon Press, 235 p.
Tooley, M.J., 1985, Climate, sea-level and coastal changes, *in*, Tooley, M.J., and Sheail, G.M., eds., The climate scene: London, George Allen and Unwin, p. 206-234.
Tooley, M.J., and Shennan, Ian, eds., 1987, Sea-level changes--Institute of British Geographers Special Publication 20: London, Basil Blackwell, 397 p.
Vick, G.S., 1988, Late Holocene paleoseismicity and relative sea level changes of the Mad River Slough, northern Humboldt Bay, California: unpub. M.S. thesis, Humboldt State University, Dept. of Geology, Arcata, CA, 87 p.
Vincent, Paul, Richards, M.A., and Weldon, II, R.J., 1989, Vertical deformation of the Oregon continental margin: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1332.
Weaver, C.S., and Baker, G.E., 1988, Geometry of the Juan de Fuca plate beneath Washington and northern Oregon from seismicity: Bulletin of the Seismological Society of America, v. 78, p. 264-275.
Weaver, C.S., and Michaelson, C.A., 1985, Seismicity and volcanism in the Pacific Northwest--Evidence for the segmentation of the Juan de Fuca plate: Geophysical Research Letters, v. 12, p. 215-218.
Weaver, C.S., and Shedlock, K.M., 1989, Potential subduction, probable intraplate, and known crustal earthquake source areas in the Cascadia subduction zone: *in*, Hays, W.W., eds., 3rd annual workshop on "Earthquake hazards in the Puget Sound, Portland area", Proceedings of Conference XLVIII, U.S. Geological Survey Open-File Report 89-465, p. 10-26.
Wells, S.G., Bullard, T.F., Menges, C.M., Drake, P.G., Karas, P.A., Kelson, K.I., Ritter, J.B., and Wesling, J.R., 1988, Regional variations in tectonic geomorphology along a segmented convergent plate boundary, Pacific coast of Costa Rica: Geomorphology, v. 1, p. 239-265.
West, D.O., and McCrumb, D.R., 1988a, Coastline uplift in Oregon and Washington and the nature of Cascadia subduction-zone tectonics: Geology, v. 16, p. 169-172.
West, D.O., and McCrumb, D.R., 1988b, Reply to: Comment on "Coastline uplift in Oregon and Washington and the nature of Cascadia subduction-zone tectonics": Geology, v. 16, p. 952-953.
Wilson, D.S., 1986, A kinematic model for the Gorda deformation zone as a diffuse southern boundary of the Juan de Fuca plate: Journal of Geophysical Research, v. 91, no. B10, p. 10,259-10,269.
Wilson, D.S., 1989, Deformation of the so-called Gorda plate: Journal of Geophysical Research, v. 94, no. B3, p. 3065-3075.
Winslow, M.A., 1988, Holocene deformation patterns and paleoseismicity in the eastern Shumagin seismic gap--southwestern Alaska: Geological Society of America, Abstracts with Programs, v. 20, no. 7, p. A382.

Yamaguchi, D.K., 1988, Preliminary tree-ring dating of late-Holocene subsidence along the Washington coast [abs.]: Holocene Subduction in the Pacific Northwest, A symposium sponsored by the Quaternary Research Center, University of Washington, Seattle, WA, p. 8.
Yamaguchi, D.K., Woodhouse, C.A., and Reid, M.S., 1989a, Tree-ring dating of late Holocene subsidence along the Washington and Oregon coasts: A progress report: National Earthquake Hazards Reduction Program, Summaries of Technical Reports Volume XXVIII, U.S. Geological Survey Open-File Report 89-453, p. 162-169.
Yamaguchi, D.K., Woodhouse, C.A., and Reid, M.S., 1989b, Tree-ring evidence for synchronous rapid submergence of the southwestern Washington coast 300 years B.P.: EOS, Transactions of the American Geophysical Union, v. 70, no. 43, p. 1332.
Yeats, R.S., Clark, M.N., Keller, E.A., and Rockwell, T.K., 1981, Active fault hazard in southern California: ground rupture versus seismic shaking: Geological Society of America Bulletin, v. 92, p. 189-196.
Yonekura, N., and Ota, Yoko, 1986, Sea-level changes and tectonics in the late Quaternary, Recent Progress of Natural Sciences in Japan, v. 11 - Quaternary Research: Science Council of Japan, p. 17-34.
Yonekura, N., and Shimazaki, K., 1980, Uplifted marine terraces and seismic crustal deformation in arc-trench systems: A role of imbricated thrust faulting [abs.]: EOS, Transactions of American Geophysical Union, v. 61, p. 1111.

FIGURE CAPTIONS

FIGURE 1.--Major features of the Cascadia subduction zone in the northwestern United States and southwestern Canada modified from Rogers (1988), Spence (1989), and Wilson (1989). The large arrows mark generalized areas along the coast that coincide with boundaries between tectonic subplates, projections of boundaries between volcanic segments in the Cascade Range (Hughes and others, 1980), or other areas where seismicity and subducting plate parameters may change, and so could correspond with the boundaries between segments of the subducting-plate as outlined by Spence (1989). No query is shown at the Mendocino fracture zone boundary because the location of this feature is accurately known. Distances between boundaries are shown only to suggest a range of possible segment lengths. The range of distances shown for segments north and south of 44.5o N reflects several locations for a possible boundary along this part of the coast. We do not know whether most Holocene ruptures along the CSZ have been influenced by these postulated boundaries. Small black triangles mark volcanoes in the Cascade Range.

FIGURE 2.--Map showing the location of estuaries in Oregon and southern Washington. The distance of the Alsea Bay and Siuslaw River estuaries from the base of the continental slope (east edge of the Cascadia trench) are also shown.

FIGURE 3.--Four possible models of the location of zones of coseismic subsidence during plate-interface ruptures in the central Cascadia subduction zone. A. A rupture of more than 700 km of the plate boundary from an earthquake of at least M_w 9 could produce a zone of coseismic uplift extending eastward (arcward) from the trench and a zone of coseismic subsidence east of the zone of uplift. Estuaries north of the Siuslaw River estuary contain stratigraphic sequences with evidence of coseisimic subsidence, while the Siuslaw estuary contains evidence of a slow, uniform rise in relative sea level (fig. 4). Thus, if the edge of the zone of coseismic subsidence during M_w 9+ events trends between the Siuslaw and Alsea Bay estuaries, as shown, events of this magnitude may have occurred. The eastward extent of the zone of coseismic subsidence for such an event is unknown. B. Ruptures of 300-600 km of the plate boundary during M_w 8 1/2-9 events would also produce extensive zones of coseismic subsidence. If these zones were located farther west than in (A) they could not include the Siuslaw estuary and the ruptures could not extend north or south of the Siuslaw River. C. If only single plate-boundary segments or portions of segments ruptured in events of less than M_w 81/4, then zones of coseismic subsidence less than 250 km long could be produced. These smaller events might not produce any significant (>0.3 m) permanent subsidence and so could include the area of the Siuslaw estuary. Slip along some parts of the plate boundary could also be aseismic (Savage, 1989). D. A variation of models A and C is that events of M_w 7 3/4-8 1/4 could be most common, with rare events of M_w 8 3/4 that rupture two segments of the plate boundary. However, in this model the zone of coseismic subsidence for the rare, largest events would have to be east of the Siuslaw estuary.

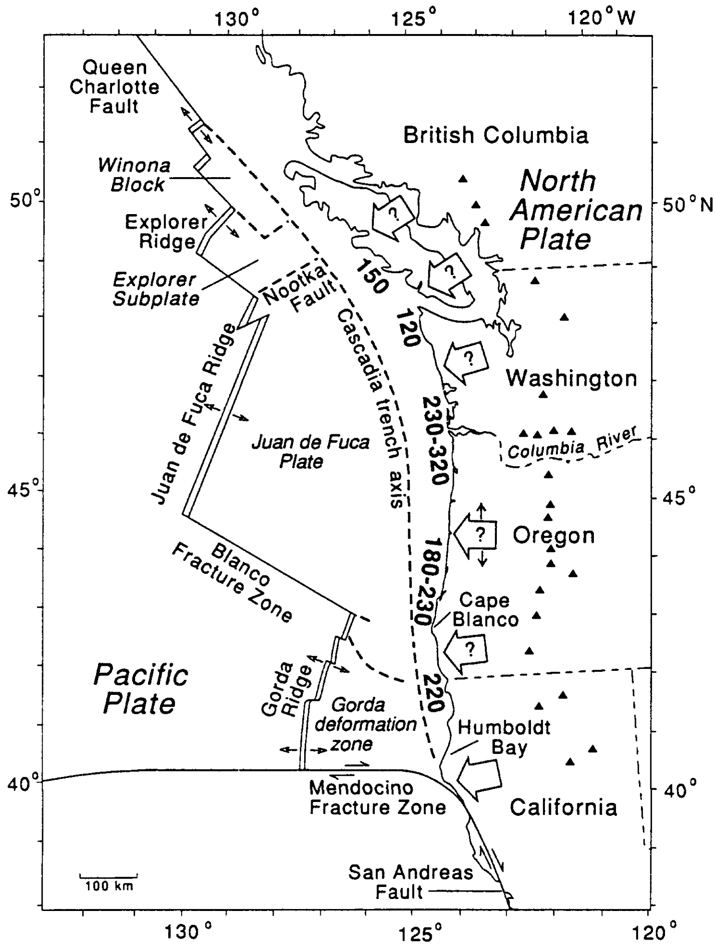
FIGURE 4.--Models of regional sea level rise, land level movements, resulting relative sea level changes, and sediment depostion rates in tidal inlets in the Pacific Northwest during the late Holocene. The curves illustrate how different, but plausible, histories of sea level rise, coastal uplift, and sedimentation rates could produce different types of stratigraphic sequences in these inlets. Sequence lithologies and contacts as in figure 5. Sequence A could only be produced in an inlet that had not experienced major (>0.3 m) sudden changes in

relative sea level. The abrupt upper contacts on the peat units in sequence B suggest sudden (coseismic) subsidence. Gradational contacts characterize sequence C, which is produced by fluctuating regional sea level and rapidly changing sedimentation rates. Sequence B could easily be confused with sequence C, which was produced without any sudden changes in relative sea level.

FIGURE 5.--Comparison of typical cores from estuarine sequences at three sites along the south-central Oregon coast. Age estimates for different levels in cores are based on unpublished ^{14}C ages from these or nearby cores. Core WC-12 contains 9 abruptly-buried peaty soils, but coseismic subsidence events in this area may result from movements on local structures. The thick peat in core SI-09 from the Siuslaw River suggests that no sudden changes in relative sea level have occurred since 2.0 ka. The abrupt contacts at the top of peaty soils in core AB8 have been interpreted as marking regional coseismic subsidence events (0.5-1 m subsidence) by Peterson and Darienzo (1989b; this volume).

FIGURE 6.--Relation between height of 80 ka marine terrace and distance from the Oregon coast to the base of the continental slope. Upper curve shows distance from the coast to the base of the continental slope (east edge of the Cascadia trench), as measured from Peterson and others (1986). Lower curve shows the present height of lowest (80 ka) marine terrace along the Oregon coast between Port Orford and the Columbia River. Solid line elevations in the Cape Blanco-Cape Arago region are from the detailed mapping of Kelsey (1990) and McInelly and Kelsey (1990). Dashed lines between South Slough and the Columbia River are from West and McCrumb (1988a). Note that terraces are poorly preserved along much of the Oregon Coast, so terrace correlations north of the South Slough area are tentative.

FIGURE 7.--Longitudinal profiles of terraces along the Umpqua and Siuslaw Rivers, central Oregon Coast Range A. Longitudinal profiles of terraces along the Umpqua River. Age estimates and analytical errors from thermoluminescence (TL) analyses are marked by filled circles and labeled "ka" (thousands of years ago); a single radiocarbon age is marked by a triangle and labeled "yr BP" (radiocarbon years before 1950). For comparison, the lowest marine terrace present near the mouth of the river (labelled "WR" and an arrow; height from Golder Associates, 1986) has been tentatively correlated with the Whiskey Run terrace near Bandon, Oregon, which has been dated by Muhs and others (1990) at about 80 ka. Preliminary uplift (incision) rates calculated from the age determinations and the height of the terrace surface or the bedrock benches underlying the terraces are shown in parentheses; rates from fluvial terraces are uncorrected for variations in sea level. Close examination of the nearly continuous early Holocene terrace revealed no evidence of latest Pleistocene folding or faulting along the Umpqua River. B. Longitudinal profile of terraces along the lower part of the Siuslaw River. Although no age data has been obtained on the pre-Holocene terraces, the position of the lowest marine terrace (tentatively correlated with the 80 ka Whiskey Run terrace) suggests that uplift rates near the mouth of the Siuslaw River are similar to those along the Umpqua River. The presence of very well developed soils on the terrace deposits suggest they were probably formed well over 100 ka. The terrace remnants are clearly modified by an anticline (double arrow symbol) that is present in the underlying Eocene bedrock. We cannot determine if terraces below an elevation of about 50 m have been folded because terraces are so poorly preserved below this height. The nearly continuous early Holocene terrace appears to be undeformed. The large area affected by the folding (over 20 km wide) indicates that these types of structures should be easy to detect on terrace profiles from other Coast Range Rivers.



130°
125°
120°W
Queen
Charlotte
Fault
British Columbia
Winona
Block
North
American
Plate
50°
50°N
Explorer
Ridge
Explorer
Subplate
Nootka
Fault
150
120
Cascadia trench axis
Juan de Fuca Ridge
Washington
Juan de Fuca
Plate
230-320
Columbia River
45°
45°
Oregon
Blanco
Fracture Zone
180-230
Cape
Blanco
Pacific
Plate
Gorda
Ridge
220
Gorda
deformation
zone
Humboldt
Bay
40°
40°
Mendocino
Fracture Zone
California
San Andreas
Fault
100 km
130°
125°
120°

Fig. 1

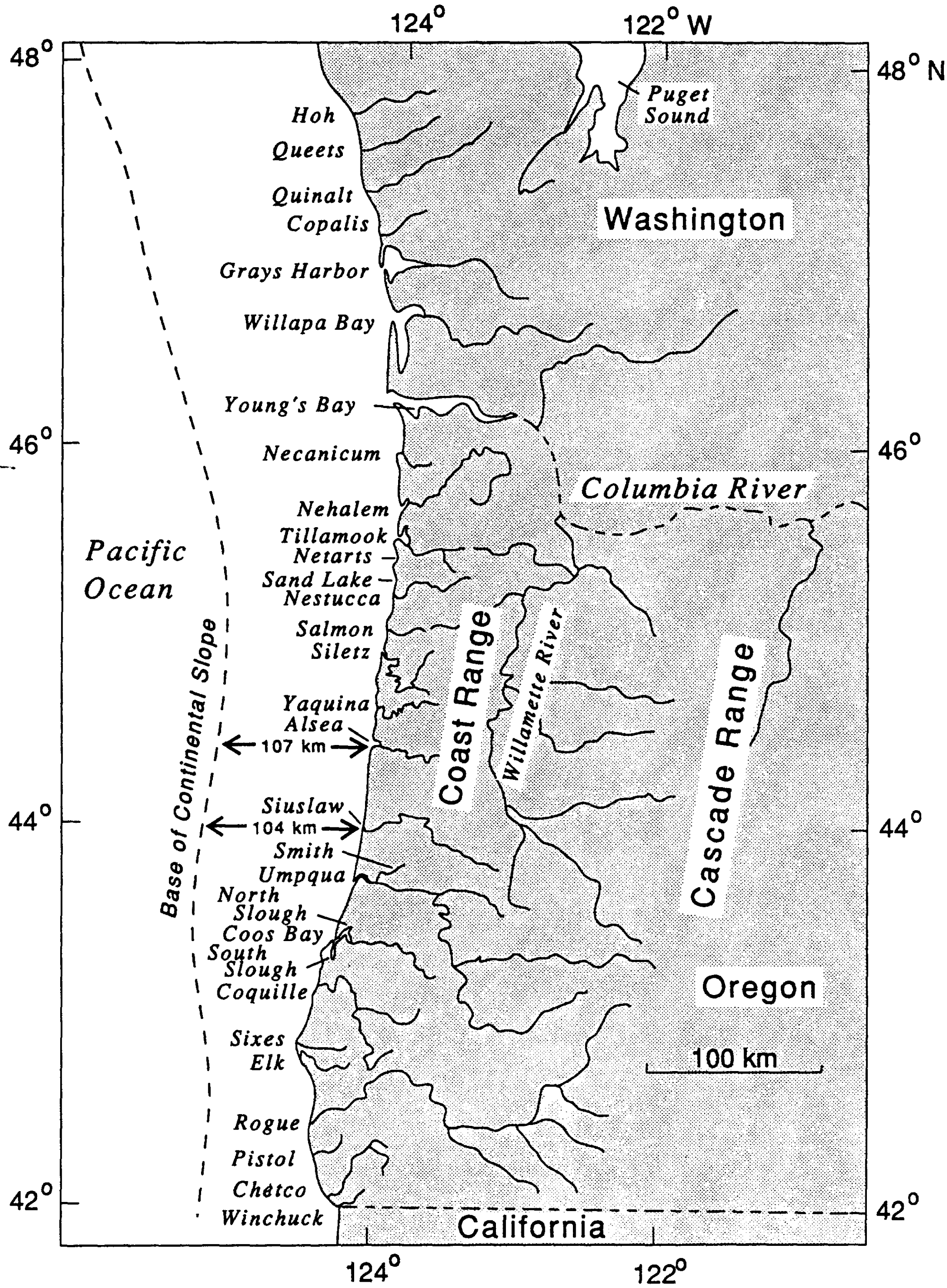
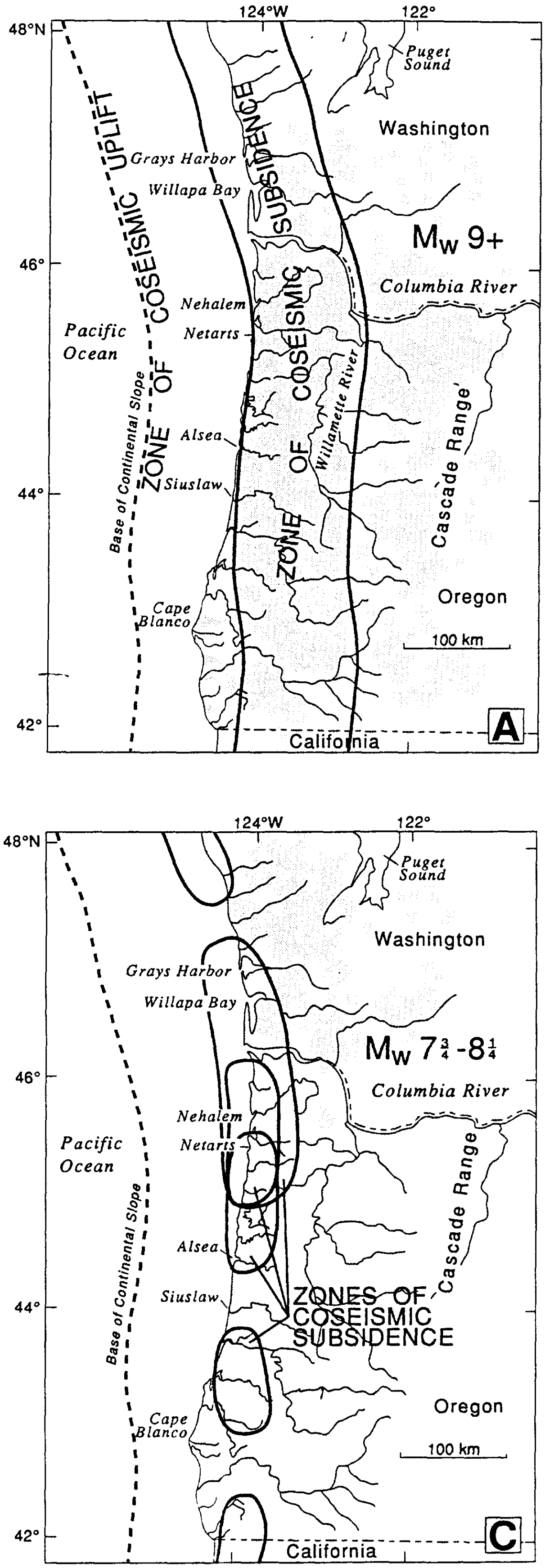

124°
122° W
48°
48° N
Puget Sound
Hoh
Queets
Quinalt
Copalis
Washington
Grays Harbor
Willapa Bay
Young's Bay
46°
46°
Necanicum
Columbia River
Nehalem
Tillamook
Netarts
Pacific Ocean
Sand Lake
Nestucca
Salmon
Siletz
Coast Range
Willamette River
Cascade Range
Base of Continental Slope
Yaquina
Alsea
107 km
Siuslaw
104 km
44°
44°
Smith
Umpqua
North Slough
Coos Bay
South Slough
Coquille
Oregon
Sixes
Elk
100 km
Rogue
Pistol
Chetco
Winchuck
42°
42°
California
124°
122°

Fig. 2



A
Mw 9+
ZONE OF COSEISMIC UPLIFT
ZONE OF COSEISMIC SUBSIDENCE
B
Mw 8½-9
ZONES OF COSEISMIC SUBSIDENCE
C
Mw 7¾-8¼
ZONES OF COSEISMIC SUBSIDENCE
D
Mw 7¾-8¾
ZONES OF COSEISMIC SUBSIDENCE
48°N
46°
44°
42°
124°W
122°
Puget Sound
Washington
Grays Harbor
Willapa Bay
Columbia River
Nehalem
Netarts
Pacific Ocean
Willamette River
Cascade Range
Alsea
Siuslaw
Base of Continental Slope
Cape Blanco
Oregon
California
100 km

Fig. 3

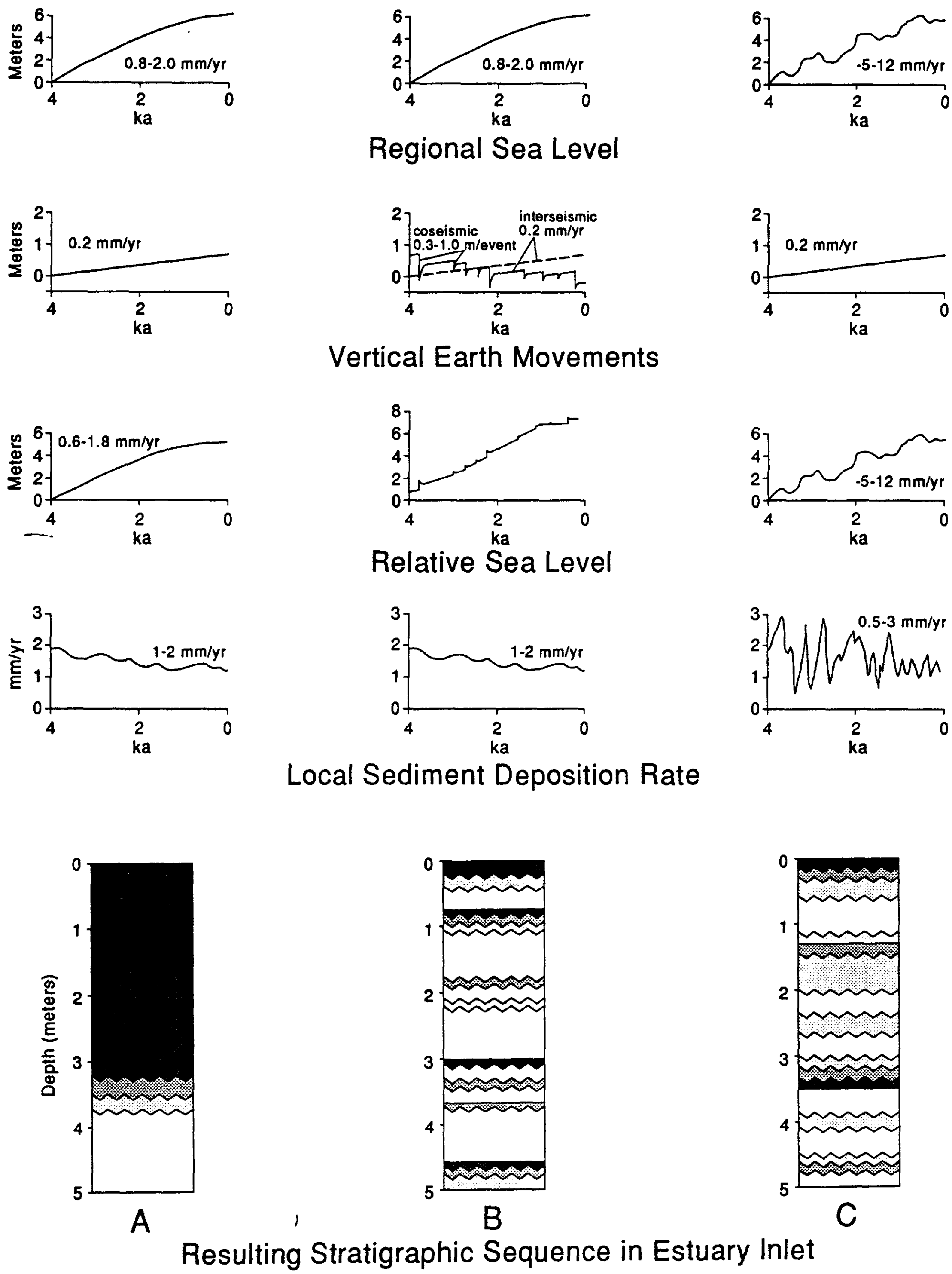

Meters
0.8-2.0 mm/yr
ka
0.8-2.0 mm/yr
ka
-5-12 mm/yr
ka
Regional Sea Level
Meters
0.2 mm/yr
ka
coseismic
0.3-1.0 m/event
interseismic
0.2 mm/yr
ka
0.2 mm/yr
ka
Vertical Earth Movements
Meters
0.6-1.8 mm/yr
ka
ka
-5-12 mm/yr
ka
Relative Sea Level
mm/yr
1-2 mm/yr
ka
1-2 mm/yr
ka
0.5-3 mm/yr
ka
Local Sediment Deposition Rate
Depth (meters)
A
B
C
Resulting Stratigraphic Sequence in Estuary Inlet

Fig. 4

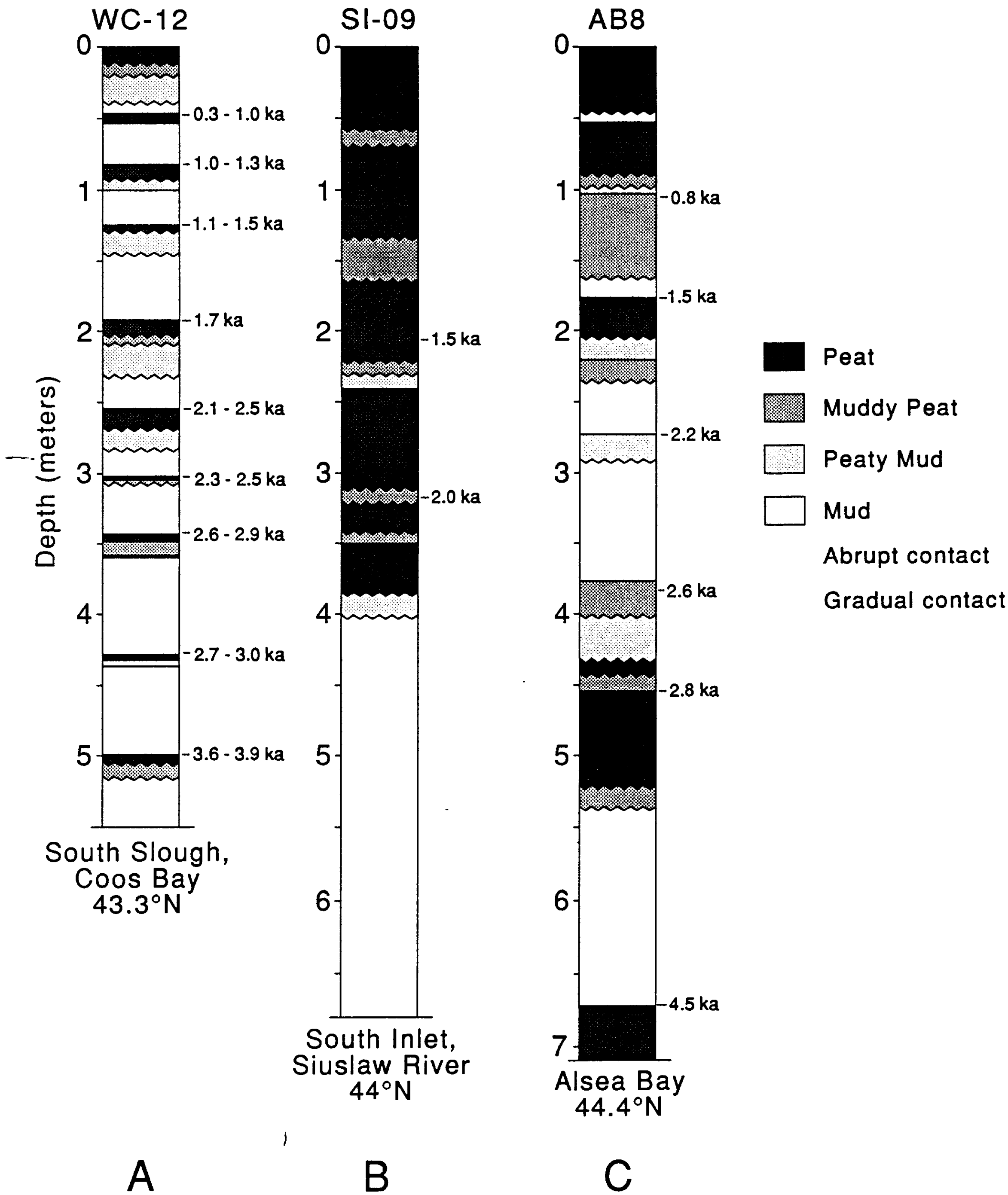

WC-12
SI-09
AB8
Depth (meters)
0
1
2
3
4
5
6
7
0.3 - 1.0 ka
1.0 - 1.3 ka
1.1 - 1.5 ka
1.7 ka
2.1 - 2.5 ka
2.3 - 2.5 ka
2.6 - 2.9 ka
2.7 - 3.0 ka
3.6 - 3.9 ka
1.5 ka
2.0 ka
0.8 ka
1.5 ka
2.2 ka
2.6 ka
2.8 ka
4.5 ka
Peat
Muddy Peat
Peaty Mud
Mud
Abrupt contact
Gradual contact
South Slough,
Coos Bay
43.3°N
South Inlet,
Siuslaw River
44°N
Alsea Bay
44.4°N
A
B
C

Fig. 5

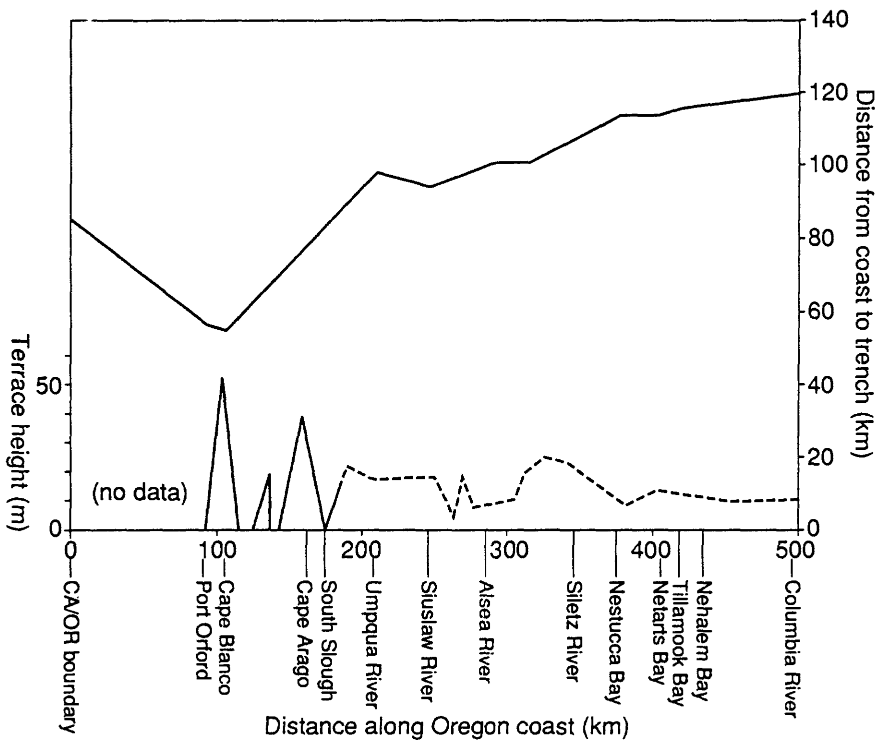
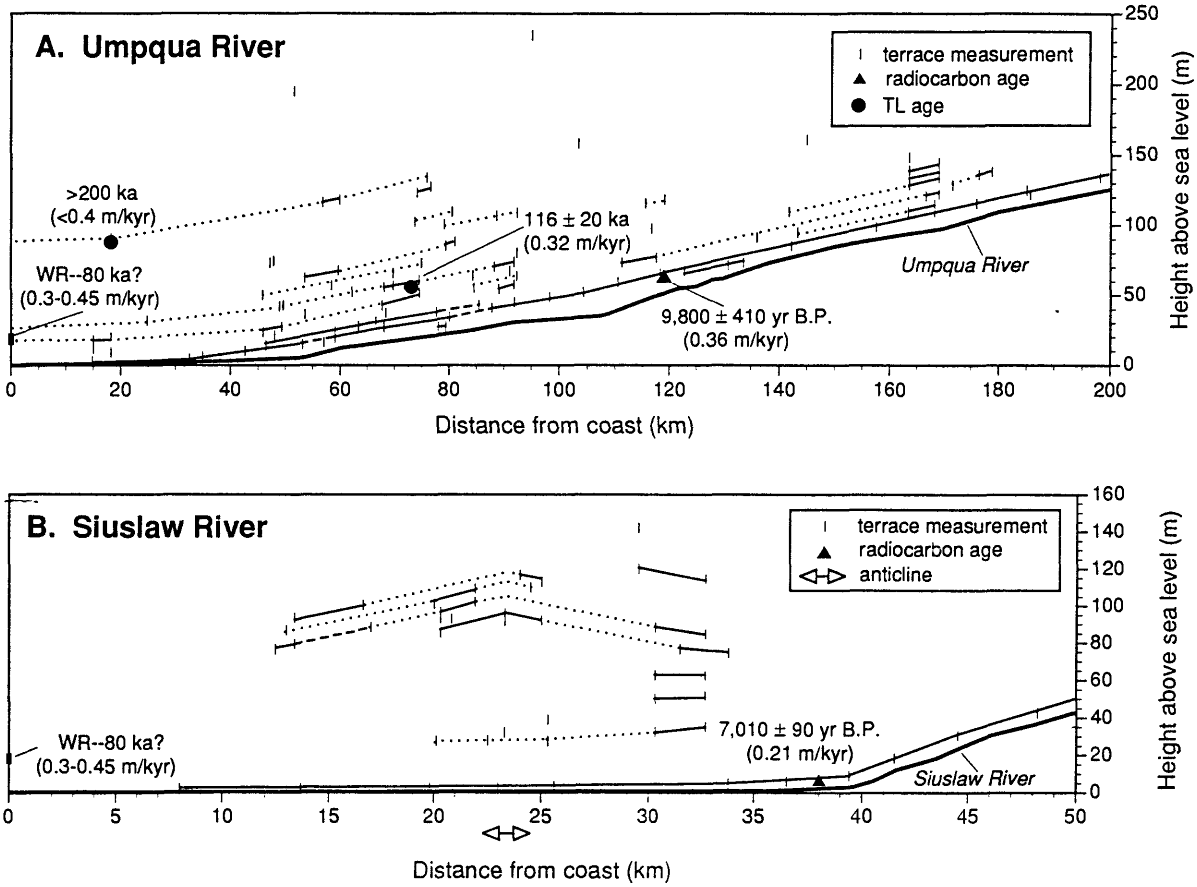

Terrace height (m)
Distance from coast to trench (km)
Distance along Oregon coast (km)
(no data)
0
50
0
20
40
60
80
100
120
140
0
100
200
300
400
500
CA/OR boundary
Port Orford
Cape Blanco
Cape Arago
South Slough
Umpqua River
Siuslaw River
Alsea River
Siletz River
Nestucca Bay
Netarts Bay
Tillamook Bay
Nehalem Bay
Columbia River

Fig. 6



A. Umpqua River
terrace measurement
radiocarbon age
TL age
>200 ka
(<0.4 m/kyr)
WR--80 ka?
(0.3-0.45 m/kyr)
116 ± 20 ka
(0.32 m/kyr)
9,800 ± 410 yr B.P.
(0.36 m/kyr)
Umpqua River
Distance from coast (km)
Height above sea level (m)
B. Siuslaw River
terrace measurement
radiocarbon age
anticline
WR--80 ka?
(0.3-0.45 m/kyr)
7,010 ± 90 yr B.P.
(0.21 m/kyr)
Siuslaw River
Distance from coast (km)
Height above sea level (m)

Fig. 7