COASTAL NATURAL HAZARDS
Science, Engineering, and Public Policy

Edited by James W. Good and Sandra S. Ridlington
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In early October 1991, more than 160 coastal geologists, oceanographers, engineers, planners, resource managers, and citizens gathered in Newport, Oregon, to learn about recent research on coastal natural hazards and discuss the implications for coastal development and management. At that conference, "Coastal Natural Hazards: Science, Engineering, and Public Policy," distinguished scientists, engineers, and policy analysts reviewed the state of knowledge in their specialties. We learned about the effects of periodic El Niños on beach and shore erosion and about recent research on factors that control sea cliff erosion. Scientists presented evidence for periodic great subduction zone earthquakes that have occurred along the Pacific Northwest coast and speculated on when the next quake might strike. We were introduced to planning and engineering approaches to hazard mitigation on the West Coast and learned about the successes and shortcomings of public policies designed to deal with development in hazardous areas.

This book is a collection of the principal papers delivered at that conference, along with critiques and supplementary remarks of panelists. For the most part, the papers are written in nontechnical language, with ample illustrations. As such, they serve as useful primers for the newcomer to the subject, whether a local official, property owner, realtor, or coastal visitor. Together, the papers should also be a useful reference for the policymaker, emergency manager, professional planner, beach and coastal manager, academic, and student. And for long-time observers of the coastal scene, the papers will confirm many of their hunches about the workings of our dynamic Pacific Northwest coastline.
SCIENCE
SEISMIC HAZARDS ON THE OREGON COAST

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Seismic hazards have been considered a relatively minor threat in Oregon for most of our recorded history. Recent advances in the geological and seismological understanding of earthquakes in Oregon changed this perception during the 1980s, and there is now fairly widespread acceptance among the scientific community that Oregon, particularly coastal Oregon, faces significant seismic hazards. In this paper I explain the changes in scientific understanding that led to this conclusion and describe the many types of hazards associated with earthquakes. In addition, I illustrate examples of the evaluation of hazard-prone areas, using the coastal geologic hazard maps published by the Oregon Department of Geology and Mineral Industries (DOGAMI).

This paper is intended for a lay audience. Thus, in the interest of clarity, I have omitted many arguments and details of the scientific data. Although I cite many sources, the paper is not a complete review of the existing literature.

Plate Tectonics: The Driving Force

The theory of plate tectonics explains the large-scale structure of the surface of the earth and major earth movements. The theory is based on the assumption that the rigid outer rock shell of the earth, called the crust, is essentially floating on a plastic or semiliquid layer 100-150 kilometers deep in the earth’s mantle (figure 1). Over hundreds of millions of years, circulation in the body of the earth has broken the crust into fragments the size of continents. These fragments are called plates, and as they move slowly across the face of the earth, they interact with each other along their edges, producing earthquake and volcanic activity. The boundaries between plates take one of three forms: divergent boundaries, where plates pull apart; convergent boundaries, where plates come together; and transform boundaries, where plates slide horizontally past one another.

Figure 1. Three types of plate boundaries. A spreading boundary (a) marks the divergence of two plates. A convergent boundary (b) occurs where one plate moves toward another. A transform boundary (c) occurs where relative plate motion is parallel to the plate edges. After Nisbet and others, 1988.
Around the world, the majority of earthquake and volcanic activity is concentrated along the plate boundaries. Spreading centers produce huge, but relatively quiet, eruptions of basalt. Subduction zone volcanic chains create smaller, but often explosive, eruptions of lava and ash. Spreading centers produce normal fault earthquakes, caused by the pulling apart of the crust, which are typically no larger than magnitude 6 or 7. Transform boundaries create earthquakes up to magnitude 8 along horizontal slip faults, where the opposite sides of the fault move horizontally past each other. Subduction zones produce thrust earthquakes, where one side of the fault is shoved beneath the other. These subduction earthquakes are the largest recorded, with magnitudes commonly greater than 8. Subduction zones also produce intraplate earthquakes up to magnitude 7 or 8 in the subducting plate, as it buckles on its way down into the body of the earth.

Cascadia: The Faults under Our Feet

The Pacific Northwest is endowed with examples of all three types of plate boundaries, as three plates interact in the region. Oregon is situated on the North American Plate (figure 2), which stretches from the Pacific coast of the U.S. to the middle of the Atlantic Ocean. To the west of the North American Plate is the Pacific Plate, the largest on the planet, which extends to Alaska, Japan, and Antarctica. Last and least, sandwiched between these two giant plates is the Juan de Fuca Plate, which forms the deep ocean floor just off the coast of Oregon and Washington. The Pacific and North American plates share a transform boundary in California (San Andreas Fault) and northern British Columbia (Queen Charlotte Fault), and the Pacific Plate moves inexorably north past North America along these two great horizontal slip faults. Dozens of major historical earthquakes on these transform faults clearly indicate that these are active plate boundaries. The Juan de Fuca and Pacific plates are separated by a spreading center, which is very seismically active and which has experienced undersea volcanic eruptions in the last few years. Finally, there is a subduction zone plate boundary between the Juan de Fuca and North American Plates. The Juan de Fuca Plate slides beneath...

At divergent boundaries, spreading centers form where lava erupts along the length of the boundary, congealing to form new crust. As the plates continue to pull apart, the newly formed crust splits, half with each plate, and this process creates tens to hundreds of kilometers of new crust over millions of years. The crust formed by this process is composed of dense basalt rock, which floats low in the mantle and therefore makes up the floors of the earth's oceans.

Where two plates collide in a convergent boundary, one will typically duck beneath the edge of the other and be pushed or pulled several hundred kilometers into the depths of the earth. This process is called subduction. When the subducted plate is sufficiently deep, it melts; the resultant magma rises to feed a chain of volcanoes parallel to the convergent boundary. This kind of plate boundary, called a subduction zone, consumes the crust produced at spreading centers.

At a transform boundary, two plates simply slide past each other horizontally, and crust is neither produced nor consumed.
the North American Plate along a great fault that extends from Cape Mendocino in California to Vancouver Island in British Columbia. This great fault is called the Cascadia subduction zone (CSZ). The CSZ originates (figure 3) at the base of the continental slope off Oregon and Washington, and angles gently beneath the North American Plate. It reaches a depth of 100 to 150 kilometers beneath the high Cascades, where the Juan de Fuca Plate melts to feed the Cascade volcanoes. As such, this great fault underlies virtually all of Oregon, and along the coast it may be as little as 30 or 40 kilometers down. Because all of the other plate boundaries in the area and the Cascade volcanoes are active, we conclude that the CSZ is also active. The Juan de Fuca Plate is probably subducting along the CSZ at 3.8 to 4.8 centimeters per year (Riddihough 1984), a rate quite similar to the 3.3 to 4.8 centimeters per year measured and estimated on the San Andreas Fault (Harbert 1991). The clear conclusion is that Oregon sits on top of the CSZ, a major active plate boundary fault.

Earthquake Sources: The Triple Threat

From our understanding of the plate tectonic setting of the Pacific Northwest, we can identify three possible earthquake types (figure 4): crustal, intraplate, and subduction. Each of the three types occurs in geographically discrete source zones. Although the three types have distinct characteristics, they are all driven by the convergence of the North American and Juan de Fuca plates across the CSZ. Crustal earthquakes occur within the North American Plate at depths of 10 to 20 kilometers. Intraplate earthquakes occur within the descending Juan de Fuca Plate at depths of 40 kilometers. Subduction earthquakes are hypothetical, as none have been observed, but they are believed to occur in the upper portion of the CSZ, along the great fault which separates the two plates.

Crustal Earthquakes: Close to Home

In Oregon, the majority of historical earthquakes have probably been crustal events. Most of these earthquakes have occurred in the Portland area, the Willamette Valley, the northern Oregon Cascades, and eastern Oregon. Coastal Oregon has been almost completely devoid of earthquakes, with the exception of a cluster of small events near Newport, and the 1863 Port Orford earthquake (Jacobson 1986; Johnson and Scofield 1991), both of which occurred before the establishment in 1970 of modern seismic networks in the Pacific Northwest. As a result, it is not known whether these earthquakes are crustal or intraplate. The history of seismicity along the
Oregon coast may suggest that there is little threat from crustal earthquakes. However, the record of historical seismicity extends only to 1841, and instrumental measurement of earthquakes in Oregon began only in the late 1950s.

The geologic record suggests that crustal earthquakes may pose some hazard at a few sites along the coast. McInelly and Kelsey (1990) reported numerous faults in the South Slough-Charleston region of Coos Bay that may represent a seismic hazard (figure 5). The various faults have broken and offset marine terrace deposits that are probably only 80,000 to 120,000 years old and hence may have some potential for future movement. The mapped extent of these faults is short, which may suggest that they are not capable of generating earthquakes greater than magnitude 5 to 6.

Work in progress (Harvey Kelsey, personal communication, 1991) suggests faults near Alsea Bay which offset marine terrace deposits, also a few hundred thousand years old. Finally, detailed offshore geologic mapping (Goldfinger and others 1990) has identified dozens of major offshore crustal faults that appear to have moved in at least the last 1.6 million years (Pleistocene time), possibly as recently as the last 10,000 years (Holocene time). These faults pose a potential threat, particularly if they extend onshore. Similar offshore crustal faults have been responsible for significant historical earthquakes, including the magnitude 6.6 earthquake of July 12, 1991, which occurred 110 kilometers west of Brookings. If that earthquake had been 50 kilometers closer, damage could have been widespread. Where potentially active crustal faults occur beneath urban areas, the possibility exists for damaging earthquakes.

From what is now known, most of the Oregon coast is probably not greatly at risk from crustal earthquakes. Detailed fault mapping of the coast has been in progress for only a few years, and seismic monitoring capabilities on the coast have
always lagged behind the rest of the state. Improved seismic monitoring by the University of Oregon and University of Washington should help to define potential crustal faults along the coast. Ongoing coastal fault studies by Western Washington State University (Harvey Kelsey), University of Oregon (Ray Weldon), the U.S. Geological Survey (Ray Wells, Parke Snively) Oregon State University (Vem Kulm, Chris Goldfinger, John Dittrle), and DOGAMI (Ian Madin) should also provide a more reliable estimate of crustal earthquake hazards.

**Intraplate Earthquakes: Danger in the Depths**

In western Washington, the majority of damaging historical earthquakes have been intraplate earthquakes, which occur in the descending Juan de Fuca Plate (figure 4). The largest of these earthquakes was the magnitude 7.1 Olympia earthquake of 1949. Along the Oregon coast, a small number of earthquakes have been positively identified as intraplate events. The largest of them was a magnitude 2.8 event that occurred at a depth of 41 kilometers near Newport in June 1981 (Weaver and Baker 1988). This suggests that many of the other earthquakes located in the Newport area before 1970 may have been intraplate events. The largest intraplate event in Oregon may have been the 1873 magnitude 6.7 Port Orford earthquake. This event was felt along the southern Oregon and northern California coasts and had no aftershocks. The absence of aftershocks has led to speculation that it was an intraplate earthquake: intraplate earthquakes typically do not have aftershocks (Ludwin and others 1989). Weaver and Shedlock (1989) have proposed that much of the Oregon Coast from Astoria to Waldport and from Cape Blanco to the California border is susceptible to intraplate earthquakes as large as magnitude 7 (figure 6).

No amount of surface geological investigation will improve our understanding of intraplate earthquakes, which occur 45 to 60 kilometers beneath the surface. Improved seismic monitoring capabilities being installed by the Universities of Washington and Oregon will provide a more reliable estimate of the hazard of intraplate earthquakes. It is clear that a major source of potential earthquakes as large as magnitude 7 underlies the entire Oregon coast, but it is not clear whether these earthquakes will happen sufficiently often to present a significant hazard.

**Subduction Earthquakes: The Big One**

No large earthquakes have been reported from the CSZ during the 150 years of recorded history in the Pacific Northwest, and modern seismic networks detect essentially no earthquakes in the zone. This has led seismologists to speculate that subduction on the CSZ, although almost certainly active, is aseismic and never produces large earthquakes (Ando and Balazs 1979). However, Heaton and Kanamori (1984) discussed the seismic potential of the CSZ and noted that it shared many characteristics with other subduction zones which had great earthquakes. They concluded that the Juan de Fuca Plate was similar to other subduction zones in which active subduction was accompanied by a great earthquake of magnitude.
8 or larger. Adams (1984) studied modern deformation of the CSZ using leveling, tide gauge, and geomorphic data and concluded that it was possible that subduction was accomplished during great subduction earthquakes every 200 to 500 years. Adams also noted that it might be possible to search for evidence of prehistoric great earthquakes by looking for disturbed layers in lake sediments, landslides triggered by earthquakes, periodic submarine landslide deposits, and uplifted or subsided coastal features. Other researchers (Byrne and others 1988) contended that the rocks in the CSZ are sufficiently weak and hot that they act in effect as a lubricant, allowing subduction to proceed without any great earthquakes. The picture is further complicated by the example of the San Andreas fault, which has "aseismically" creeping segments, which produce constant microearthquakes, and an almost completely aseismic segment, which moved in 1906 to produce the great San Francisco earthquake. Without direct evidence, the earlier debate was largely academic, as there was no way to prove or disprove the hypothesis of great earthquakes on the CSZ.

**Buried Marshes: The Smoking Gun**

The theoretical arguments about whether or not the CSZ moved in periodic great earthquakes were overshadowed by Brian Atwater's (1987) discovery of direct geologic evidence for prehistoric great earthquakes. Atwater's study was the first to find direct evidence of great CSZ earthquakes and was based on looking for the geologic footprint of a great earthquake. Other great subduction earthquakes around the world—Alaska, 1964, and Southern Chile, 1960 (Plafker 1972)—produced distinct and gigantic footprints on the land. Typically, the upper plate in the subduction zone undergoes immediate and permanent land level changes during a great subduction earthquake with a pattern as shown in figure 7. The leading edge of the upper plate is uplifted, with subsidence farther inland and less pronounced uplift farther inland yet. The simple mechanical explanation for this pattern is that during the hundreds of years between earthquakes, the two plates are locked together but still converging. This steady convergence causes the upper plate to flex slowly, as shown in figure 8. When the earthquake occurs, the flex is released and the land rises or subsides accordingly. The earthquake cycle produces a distinctive pattern of land level changes, with slow steady uplift or subsidence between earthquakes that instantaneously reverses during the earthquake. This phenomenon can be used in effect as a natural seismograph to record prehistoric earthquakes, because the sea leaves a "ring around the bathtub" on the land. As the land moves up and down with respect to sea level, coastal processes leave geologic features and deposits that form at very specific elevations. Where the land is uplifted, wave-cut platforms or beach ridges formed at or below mean tide level are often stranded high above the highest tides. Where the land subsides, freshwater marshes or lowland forest lands may sink below the level of the tides and be converted to intertidal mudflats.

Atwater (1987) studied Willapa Bay in southwestern Washington, where he noted a distinctive pattern of sediment in the banks of tidal channels in modern marshes. Typically, the modern vegetation would be found growing on a modern
observed sand layers directly above several of the buried marsh peats, which he speculated might have been deposited by tsunamis (popularly known as tidal waves) generated by the same earthquake that caused the subsidence.

Atwater's discovery provided the first geologic evidence that great megathrust earthquakes might have occurred before the arrival of Europeans in the Pacific Northwest, but there were still many skeptics, many unanswered questions. Perhaps the burial of the marshes was due to floods, storm surges, breaches of spits, distantly generated tsunamis, or periodic great forest fires that choked streams with silt and filled in bays. Alternatively, it might be possible that the land had indeed subsided in an earthquake, but in a minor earthquake on a local fault instead of a great earthquake stretching from Vancouver Island to California.

Subsequent to Atwater's original research in Willapa Bay, other researchers began to explore Oregon estuaries for similar evidence. They found it in almost every significant estuary along the northern and central coast (figure 10). Grant and McLaren (1987) found evidence for several episodes of abrupt marsh subsidence and burial at the Salmon and Nehalem River estuaries.


Clearly, the phenomenon of abruptly buried marshes is not due solely to local faults in Washington. All along the Cascadia subduction zone, repeated cycles of slow uplift followed by rapid submergence of the land have occurred, with many submergence events accompanied by tsunamis. The simplest explanation for these deposits is the periodic occurrence of great subduction earthquakes that involve hundreds of
kilometers of the coast all at once. If true, the implications for Oregon coastal communities are awesome, because such an earthquake would cause simultaneous strong shaking and coastal subsidence, which would be followed quickly by a local tsunami.

Corroborating Evidence: More Pieces of the Puzzle

Although the evidence from buried marshes is fairly persuasive, it is vital to look for other evidence to prove the great earthquake hypothesis.
The adverse consequences of spending money and restricting coastal development unnecessarily in response to a false subduction earthquake threat are probably outweighed only by the consequences of preparing inadequately for a true threat. Although earthquake-related subsidence remains the only satisfactory explanation for the buried marshes, it is important to look for other types of evidence. To date, this has come from undersea landslides, modern geodetic measurements, Indian legends, and archaeological sites.

Adams (1990) has proposed a completely independent line of evidence for great subduction earthquakes based on submarine landslide deposits. Sand, silt, and clay flushed into the coastal waters of Oregon and Washington by rivers accumulate in thick deposits offshore on the continental shelf and slope. Periodically, these piles become unstable and slump in a submarine landslide, causing a slurry of sediments and water (called a turbidity current) to flow down submarine channels onto the deep abyssal plain. Each turbidity current leaves a distinctive layer of sediment, called a turbidite, and it is possible to count the number of turbidity currents that have passed any given site by counting the turbidite layers. Griggs and Kulm (1970) first noted that sediment cores from a number of submarine channels off the coast of Oregon and Washington could be used to count the number of turbidity currents that had occurred since the eruption of Mt. Mazama (now Crater Lake) about 7,000 years ago. They determined this by counting the number of turbidites above the first layer which contained the distinctive ash from Mt. Mazama. In his analysis, Adams (1990) noted that there were similar numbers of post-Mazama turbidites in the upper reaches of many channels along the coast. Most important, he noted that even where two channels came together, there were the same number of turbidites below the confluence as above. This requires the turbidity currents in each channel to have been triggered simultaneously. Adams (1990) argues that the only plausible explanation for simultaneous triggering of turbidity currents at sites tens to thousands of kilometers apart is a great subduction earthquake.

Geodetic techniques compare very precise measurements of the position and elevation of a network of stations over time to determine how the land is currently expanding or contracting, rising or falling. The first attempt to use geodetic data to constrain the behavior of the CSZ was by Ando and Balazs (1979), who used historical leveling data to show that the Oregon Coast Ranges were tilting to the east. They concluded that the Juan de Fuca Plate was subducting aseismically and would not have great earthquakes. Adams (1984) looked at historical data as well as geologic data to determine long-term deformation rates all along the CSZ. He concluded that the modern deformation did not require aseismic subduction and suggested that great earthquakes might occur. Vincent (1989), and later Weldon (1991), used historic leveling data and tidal records along the Oregon coast and across the Coast Ranges to show that parts of the coast are clearly rising at a significant rate. This result is very important because it shows clearly that the Juan de Fuca and North American plates are in fact locked together, and not slipping aseismically.
past one another on some layer of sedimentary “grease.” Both studies note that the amount of geodetically measured uplift is dramatically less along the north-central Oregon coast than areas farther north or south (figure 11), which suggests that the subduction zone is broken into small independent segments.

![Figure 11. Schematic representation of geodetically measured deformation in the Pacific Northwest. Vertical data in Oregon from Weldon, 1991. Horizontal data in Washington from Savage and Lisowski, 1991.](image)

In Washington, Savage and Lisowski (1991) measured the ongoing deformation of the Olympic Range with precision laser instruments. They concluded that the Olympics are currently being shortened horizontally in a direction essentially parallel to the direction of subduction of the Juan de Fuca plate (figure 11), and this shortening is consistent with the accumulation of strain energy on a locked subduction zone.

These preliminary results from geodetic studies still leave questions about the shape of the locked portion of the subduction zone and about our current position in the strain cycle, but they are inconsistent with the notion of subduction without great earthquakes.

Indian legends of great earthquakes and tsunamis are known from the Pacific Northwest. Heaton and Snaively (1985) report several legends from the region. One legend of the Makah Indians at Neah Bay recorded by James Swan states that the waters of the bay receded dramatically for four days, then returned to flood the land for another four days before receding. The same legend described a permanent land level change at the same time, with an island being converted to a peninsula, although it also noted that the water that flooded the community was hot. Woodward (1990) reports a similar tsunami legend from the Tillamook area. Unfortunately, Indian legends are somewhat ambiguous about the timing of events, and contain enough references to clearly supernatural occurrences that they provide only weak corroborating evidence to the great earthquake hypothesis.

More concretely, Woodward (1990) noted archaeological evidence for significant changes in the lifestyles of Indians along the coast of Oregon which have occurred at times coincident (see discussion below) with hypothetical prehistoric subduction earthquakes. At Nehalem Bay, Woodward reports an Indian campsite dated to 380 years before the present (BP) that is now permanently below tidal levels. In Tillamook Bay, changes in species of shellfish deposited in middens suggest a change from a bay environment to open shore at 1070 years BP. At Netarts Bay, shell middens at an Indian campsite formed 1,400 years ago have now subsided below the level of high tides. The results from these sites and others are intriguing, but they provide only circumstantial evidence of major, perhaps catastrophic changes in coastal Indian settlements that may have accompanied great earthquakes.

The evidence listed above is consistent with a history of great megathrust earthquakes in the Pacific Northwest, and a majority of geoscientists working in the region now accept that these events have occurred. There are, however, problems with the theory of great subduction events, which are reviewed in the following section.

**Conflicting Evidence: It’s Not a Done Deal**

One of the most fundamental problems with the great earthquake story is the assumption that the buried marsh layers are in fact due exclusively to abrupt land subsidence during earthquakes. Alternating layers of peat and intertidal
mud are known from coastal regions without subduction zones (Nelson and Personius, in press). Atwater (1987) and Atwater and Yamaguchi (1991) cite a variety of evidence from Washington marshes that seem to require earthquakes to explain buried marshes. Peterson and Darienzo (in press) have shown that in Alsea Bay, abrupt land subsidence is the only likely cause for the buried marshes observed there. However, the origin of buried layers in other bays may still be questioned.

If we accept that the marshes do subside during earthquakes, we must assess the possibility that each estuary is responding to independent movements on local faults rather than great subduction earthquakes that cause many estuaries to subside at the same time. Goldfinger and others (1990) have studied faults on the continental shelf and slope of Oregon and have identified dozens of major faults which may have moved in geologically recent times. Many of the estuaries where buried marshes occur appear to lie on these faults, raising the possibility of numerous local subsidence events. Further investigation is necessary to determine whether these faults are independently responsible for marsh burial, but several general observations suggest that they are not. First, at least a dozen estuaries between central Oregon and central Washington subsided about 300 years ago (see below). If each subsidence event was the result of an independent earthquake, the implication is that over a dozen occurred in the late 1600s, but none have occurred since the 1840s. There are so many estuaries with relatively recent and frequent marsh burials that we should have historical records of marsh burial events if they are due to random earthquakes on a dozen independent faults. In addition, geologic mapping onshore, in some cases quite detailed, has yet to uncover evidence that any of the offshore faults associated with estuaries has moved in the last few thousand years.

Finally, almost every estuary has evidence of tsunamis associated with one or more of the buried marsh layers. Peterson and Darienzo (in press) have pointed out that if each estuary has an independent earthquake which generates a local tsunami, there will be a tsunami deposit directly above the subsided marsh in that estuary, and tsunami deposits at a variety of levels in adjacent estuaries that did not subside. This implies that tsunami sands should be distributed throughout the peat and intertidal mud layers if there are numerous independent events. On the Oregon coast, Darienzo and Peterson (1988, 1990), Peterson and others (1991), and Peterson (personal communication, 1991) find that the vast majority of tsunami deposits occur directly above buried marshes.

Another unresolved problem with the great subduction earthquake hypothesis is the common occurrence of uplifted marine terraces adjacent to estuaries which contain buried marshes. Sea level has changed dramatically during the last few hundred thousand years, falling during ice ages when water is tied up in glaciers, and rising between ice ages as glaciers melt. During each high stand of sea level, wave action cuts a platform across coastal bedrock, which is then covered by marine sediments to form a distinct, flat marine terrace. The most recent high stand was about 80,000 years ago, and at many sites along the Oregon and Washington coast this terrace is now several meters to tens of meters above modern sea level. If sea level now is about the same as it was 80,000 years ago, these terraces must have been uplifted by earth movements. However, the uplifted terraces are often adjacent to estuaries in which there is clear evidence of several meters of submergence in the last few thousand years. It is necessary to resolve the contradictory evidence for net uplift over the last 80,000 years and net submergence over the last 5,000 to 10,000 years.

A final unresolved problem with the great subduction earthquake hypothesis is the apparent lack of widespread evidence of liquefaction. Liquefaction occurs when loose, water-saturated sand deposits are shaken strongly in an earthquake. The sand becomes fluid, and a mixture of sand and water often erupts onto the ground surface through fissures. These sand fissures and erupted sand piles are commonly observed in many other areas of the world that have been shaken by strong earthquakes. The presence of such features in association with buried marsh horizons would strongly support the great earthquake hypothesis. The widespread absence of liquefaction features along the Oregon and Washington coast could suggest that whatever caused the marshes to subside did not involve
strong shaking. Widespread liquefaction features have not been reported from the Oregon coast to date; however, no systematic effort has been made to locate them. In Washington, Atwater (personal communication, 1991) has found liquefaction features associated with buried marshes at sites on the Copalis River. Peterson (personal communication, 1991) has observed widespread liquefaction on the Oregon coast in marine terrace sediments which are 80,000 years or more old. I have observed similar features in old marine terrace sediments in the Coos Bay area. The critical problem is to find liquefaction features in sediments that are only a few thousand years old. Clearly, a concerted effort must be made to establish whether or not liquefaction features are widespread along the Oregon coast, and if they are not, the great earthquake hypothesis must be carefully re-examined.

When is the Next Big One? The Big Question

If we accept for the time being that buried marsh deposits in Oregon and Washington are natural seismographic records, then the next step is to determine how often, on average, the prehistoric earthquakes occurred. If it is possible to calculate a reliable average time between events, then it is possible to calculate the probability that the next event will occur in some given time frame. This technique has been widely applied in other areas where there is a reasonably well-dated geologic record of prehistoric earthquakes.

The time of burial of marshes in Oregon has been dated by two techniques, each of which has significant drawbacks. Radiocarbon dating can be used to date plant material preserved in the buried marsh or forest peats. The technique is relatively fast and inexpensive, and dateable plant material is abundant. Analytical errors inherent in the technique are typically plus or minus 50 to 100 years, which is not significant for materials that are several thousand years old, but is very significant for materials that are only a few hundred years old. Calibrations for prehistoric variations in radioactive carbon production introduce additional uncertainty, and many relatively young samples correspond to several calendar dates when calibrated. The second source of error is even more of a problem. Radiocarbon ages date the time of death of the plant material, and samples taken from peats may have been dead for a long time. This error can be greatly reduced by dating material from trees rooted in the buried marsh that were presumably killed by the subsidence, but such trees are far less common than peats. In general, at any site, it may not be possible to date the time of marsh subsidence any closer than plus or minus 100 to 200 years. This means that we cannot necessarily distinguish between events that occurred a day apart and events that occurred a few hundred years apart, and it may well be that the average time between earthquakes is similar to or smaller than the best resolution of radiocarbon dating.

The second dating technique is tree-ring dating, which is accomplished by comparing the patterns of annual growth rings in trees killed by subsidence to those in living trees on adjacent uplands. This technique allows dating of the time of death of the trees to within a decade, or often within a few years (Atwater and Yamaguchi 1991). However, well-preserved trees are not present in many sites, and living trees are not old enough to compare with buried marshes that are more than 1,000 years old. This technique is most useful for looking at the most recent events.

A final problem in calculating the average time between earthquakes is the possibility that due to conditions of sedimentation, timing, local climate, sea level fluctuations, and so on, not all earthquakes will make unambiguous buried marsh horizons at all sites. This means that recurrence intervals estimated for any one site will be based on a minimum number of events thought to have occurred. If one or two events were not clearly recorded, then the resultant estimate of recurrence interval will underestimate the probability of the next earthquake.

The uncertainties associated with dating marsh subsidence mean that a credible calculation of the probability of the next earthquake is still not possible, even assuming that buried marshes represent past earthquakes. The best we can do with the radiocarbon numbers at this point is to take the reported ages at face value and treat the resulting estimates of recurrence intervals with a great deal of skepticism. An important result we
can derive from this kind of analysis is not so much which day to be out of town in order to avoid the Big One, but a sense of how short an interval is possible between great earthquakes, and a reasonable estimate of when the last one occurred.

The most recent event is probably the best dated, because it is best exposed and because locally the radiocarbon dating can be checked with tree-ring dating of cedar and spruce trees killed by marsh subsidence. Atwater and Yamaguchi (1991) find that in southwest Washington, radiocarbon and tree-ring dating suggest that the most recent subsidence occurred about 300 years ago. Peterson and others (1991) report a range of ages for the most recent event in Oregon bays, with the youngest at 270, plus or minus 60, and the oldest at 550, plus or minus 70 years BP. Grant (written communication, 1991) reports the most recent subsidence in the Salmon River of 247, plus or minus 25 years BP, and in the Nehalem River, 225, plus or minus 19 years BP. Adams (1990) estimated the age of the most recent turbidite offshore at 300 years BP by studying the thickness of sediment layers on top of the turbidite. Most of these dates are consistent with the more precise tree-ring data indicating that the last great event or set of events occurred in the late 1600s, but it is not possible to distinguish between one great simultaneous event and several smaller events scattered over decades.

The average intervals between earthquakes calculated from this data must be treated skeptically. Atwater (personal communication, 1991) is not sure that a significant return time can be calculated, but points out that there have been either 6 or 7 events in the last 3,500 years. This suggests a nominal recurrence of 500 to 580 years.

Peterson and others (1991) report average intervals of 370 years for 4 events at Netarts Bay, 340 years for 3 intervals in Alsea Bay, and a regional average over 11 events in Northern Oregon of 330 to 340 years. Adams calculated an average of 590 years for 13 events, using the turbidite data. There is wide variability in this data, but two things are clear. If all of these events were due to independent earthquakes on local structures, then there have been tens of earthquakes in the last few thousand years. The return interval between subsidence-causing earthquakes somewhere along the coast then becomes so short that we would expect to have a historical record of one. The other important fact to note is that recurrence intervals from many sites are at least as short as the time since the last event, within the limits of radiocarbon error.

We have a long way to go before we can quantify the likelihood of the next great earthquake, but this event is not necessarily going to occur in some remote future. In fact, it is quite possible that the next big shake will happen in the near future. This possibility should be sufficient to cause emergency managers, land-use planners, and public officials of coastal communities to start looking at where they are vulnerable.

Where and How Big: What Can We Expect?

Estimates of the size and potential location of future great subduction earthquakes also vary widely and are based on a limited understanding of the structure of the CSZ. The size of future earthquakes will depend on the area of the locked fault between the plates that moves. The location of the earthquake will similarly depend on the portion of the fault that moves.

The area of the fault that moves depends on the width of the locked portion of the fault and the length of fault along the coast that fails. The total length of the CSZ is fairly well known, but few researchers think that the entire 1,000 km will fail all at once. Instead, the CSZ is likely to break in a series of relatively short segments. Geoscientists can guess at the location of segment boundaries but still cannot demonstrate where they lie. Segments may be as short as 100 kilometers or the full 1,000 kilometers. Similarly, the width of the locked portion of the fault strongly influences the possible size of an earthquake. The location of the locked zone also controls where the earthquakes can occur. There is little agreement on the likely width of the locked zone. In southern Oregon, Clark and Carver (1991) proposed that the locked zone might be as wide as 75 to 100 kilometers in southern Oregon. Peterson and others (1991) present a model of the locked zone constrained by marsh subsidence data that is best fit by a 90-kilometer-wide locked zone. Blackwell (1991) proposes a locked zone as narrow as 20
kilometers based on thermal modelling. According to Pezzopane and others (1991), geodetic data suggests that it may vary widely in width. A pair of potential locked zones is shown in figure 12.

These maps can be used by trained professionals to make a first-order assessment of potential earthquake hazards. For this report, the maps are out of Bulletin 81, Environmental Geology of Lincoln County (Schlicker and others 1973).

**Ground Shaking and Amplification**

The most widely experienced effect of an earthquake is ground shaking, which is also typically responsible for the majority of earthquake damage. The strength of shaking at any site during an earthquake will depend on the size of the earthquake, the distance of the site from the epicenter, and the nature of the geologic materials under the site. Larger earthquakes produce stronger ground shaking, but the strength of shaking dies off rapidly with distance from the epicenter. To predict the strength of shaking at a given site, we need to know how large the earthquake will be and where it will be centered, both currently impossible to know. A few general models of the strength of ground shaking have been made for the Oregon coast. The strength of ground shaking is usually expressed as a fraction of the force of gravity. Levels above .2 acceleration of gravity (g) are significant, and modern buildings in Oregon are designed for .2 g. Pezzopane and others
Figure 13.
Geologic map of the Newport area, Lincoln County, Oregon. After Schlicker and others, 1973.
Figure 14
Environmental
Geology map of the
Newport area, Lincoln
County, Oregon. After
Schlicker and others,
(1991) suggest that peak horizontal accelerations of .2 g to .4 g can occur along the coast. Coohee and others (1991) model a magnitude 8.1 subduction zone earthquake and suggest that coastal Oregon might experience .14 g to .41 g of peak horizontal acceleration. An additional threat unique to CSZ earthquakes is the unusually long duration of shaking. The magnitude (Mw) 8.1 earthquake modeled by Coohee and others (1991) would cause strong shaking for over 45 seconds. Damage increases dramatically as the duration of shaking increases.

The ground motion levels discussed above are for bedrock sites. The presence of thick soils, alluvial deposits, or soft rock over the bedrock can greatly amplify the ground shaking, often by factors as high as six. In general, young (Quaternary) deposits of sand, silt, and clay are most likely to amplify ground shaking, although less frequently they may actually reduce ground shaking. Figure 15 is derived from figure 13, the geologic map from DOGAMI Bulletin 91, and shows the areas covered by the geologic units labelled Qmt (Quaternary Miocene terrace) and Qal (Quaternary alluvium). The Qmt deposits are young marine terrace sand deposits, and the Qal deposits are young sand, silt, and gravel deposits lining the bays and river valleys. These units are most likely to amplify shaking, in contrast to the bedrock deposits present in the rest of the area. Therefore, for a preliminary assessment, these areas would be considered more potentially hazardous, and more refined hazard assessments would be focused there. The actual threat of amplification can be modeled by computer techniques for a given site, a procedure that might be appropriate for large structures or critical facilities like hospitals.

To illustrate the importance of soil amplification, we can look at the Mexico City earthquake of 1985. This earthquake, a magnitude (Mw) 8.1 subduction zone megathrust event, was centered 300 kilometers from Mexico City. Soft alluvium in the old lake beds on which the city is built amplified the shaking sufficiently to cause complete collapse of numerous modern structures engineered to withstand earthquakes. Similarly, the portion of the Cypress Freeway structure that collapsed in the 1989 Loma Prieta earthquake was only that part built on soft bay mud.

Coseismic Subsidence

As we saw earlier, the footprint that a great subduction earthquake makes on the land is a pattern of rapid subsidence or uplift of the land. This movement, which takes place during the earthquake, is called coseismic movement. It is the occurrence of coseismic subsidence along the Oregon coast that is thought to be responsible for the repeated burial of marshes, and a future great subduction earthquake would be likely to produce similar effects. It is possible to estimate the amount of coseismic subsidence at a marsh site by identifying the ecological zones represented by the successive layers and measuring the difference in elevation between modern representatives of those zones. Peterson and others (1991) have made such estimates of the average coseismic subsidence at three bays for the last four burial events. They found 1.0 to 1.5 meters of subsidence at Netarts Bay, 0.5 meter to 1.5 meters at Alsea Bay, and 0 to 0.5 meter at the Siuslaw River. These are not dramatic amounts of subsidence and are unlikely to cause large-scale flooding of coastal communities. However, this subsidence adds to the flooding by the subsequent tsunami and causes increased flooding during storms and accelerated coastal erosion.

Fault Rupture

As discussed in the section on crustal earthquakes, we know of few young faults on the coast of Oregon. However, there are numerous offshore faults. These offshore faults appear to cut the seafloor and are therefore likely to have moved in geologically recent times. Ground rupture caused by movement of an offshore fault is not a great problem because there is no development offshore. Figure 16, derived from the geologic map in figure 13, shows several major west-northwest trending faults passing south of Yaquina Bay. These faults are very similar in trend to the geologically young offshore faults, and there remains a possibility that they may move during a great subduction earthquake or independently in a smaller crustal earthquake. The likelihood is probably remote, so again, this hazard might be of concern only in the siting and construction of critical structures. It is very expensive to engineer structures to tolerate fault rupture beneath their
foundations, but it is relatively easy to site structures well away from the potential rupture zone.

**Liquefaction and Settlement**

Many geologically young sand and silt deposits are relatively loose, meaning that the sand particles are not tightly packed together and there are significant spaces between grains. When shaken by an earthquake, loose sand or silt can become more compact, just as flour settles when shaken in a measuring cup. If the sand is dry, ground settlement occurs, which may locally be sufficient to damage structures. An even more destructive situation exists when the sand is saturated with water before the earthquake. The settlement of the sand pressurizes the water in the spaces between grains, and the pressurized water causes the sediment to liquefy. Because liquefied sediment has very little strength, it is common for structures to tilt, sink or settle dramatically when the underlying soil liquefies. Even more devastating is the tendency for liquefied soil to flow towards free faces (such as river or bay banks) and down very gentle slopes. Mass movement of liquefied or partly liquefied soils results in the most spectacular of earthquake damage and is particularly devastating to coastal areas, damaging bridges, docks, and port facilities. Liquefaction also causes widespread failure of buried pipes and cables, affecting fire fighting and emergency communications after the event.

As with amplification, the tendency of any site to liquefy in an earthquake can be estimated accurately only with a detailed site-specific study. The Qmt and Qal deposits are the only geologic materials in this area with any significant potential for liquefaction. Although they are widespread, these materials pose a threat only where they are
saturated with groundwater. Again, we can use the geology and environmental hazard maps for the Newport area to roughly estimate the areas most susceptible to liquefaction, and thus narrow down the area where more specific studies are needed. Figure 17 shows areas likely to be susceptible to liquefaction. It is derived by overlaying areas of shallow ground water (depicted on the environmental geology map, figure 14) on areas of Qm or Qa sands and silts (depicted on the geologic map, figure 13).

Landslides

One of the most common secondary hazards associated with earthquakes is earthquake-induced landslides. Slopes which are stable under ordinary conditions may be destabilized by the strong shaking of an earthquake and begin to move. Wilson and Keefer (1985) note that earthquake-induced landslides can occur up to 200 kilometers from the epicenter of a magnitude 8 earthquake. As with the amplification and liquefaction hazards, detailed site studies are required to determine how likely a slope is to slide in the event of a given earthquake. Again, it is possible to use the information available in the DOGAMI environmental hazard maps to outline areas most likely to experience this hazard. Figure 18 shows two types of landslide data derived from the maps. Areas of existing landslides or landslide topography are taken directly from the environmental geology map (figure 14). These areas may be reactivated in future earthquakes, particularly where they have been developed, cut by roads, or logged. Landslide-prone areas are derived by overlaying areas of mudstone bedrock from the geologic map (figure 13) on areas with slopes over 25% from the environmental geology.
map (Figure 14). These areas are the most likely to have new landslides in an earthquake. In addition, areas of rapid sea cliff erosion or riverbank erosion may be susceptible to earthquake-induced landsliding. In all cases, extensive development, logging, forest fires, or road building may increase the likelihood of earthquake-induced landslides because of changes in drainage and stability of the slopes.

**Tsunami and Seiche**

The final class of secondary earthquake hazard is mass movements of water which may inundate shoreline areas. In a seiche, the water in a relatively small body of water, like a lake or bay, sloshes from bank to bank, just like a full coffee cup on a bumped table. A tsunami occurs when a large area of the seafloor moves, displacing a huge amount of water in the ocean. Both of these hazards are likely to occur in the event of a subduction zone earthquake, but only seiches are likely to occur in a crustal or intraplate earthquake.

The extent of inundation caused by a seiche in any body of water will depend on the strength of ground shaking at the site. It will also depend on the degree of similarity between the natural period of oscillation of the body of water and the period of shaking of the earthquake. This makes estimation of seiche hazards extremely difficult, because the periods of shaking of earthquakes are quite variable. Sophisticated computer modelling can put rough limits on the maximum seiche run-up, but this technique is relatively expensive.

Tsunamis are great waves produced by vertical motion of large portions of the seafloor. The waves travel at speeds of several hundred kilometers per hour in the open ocean, where they may
be only a fraction of a meter high. When a tsunami wave approaches shore, it begins to slow down and get higher, and what began as a wave only a half a meter high on the open ocean may be several meters high when it reaches shore. The maximum elevation above sea level that the tsunami reaches is called the run-up. The area covered by the tsunami is the inundation. Tsunamis are not likely to be generated by crustal or intraplate earthquakes, because these types of earthquakes are relatively small and do not involve vertical movements of the seafloor. Subduction zone earthquakes, on the other hand, are very large, cause large vertical movements of the seafloor, and usually cause tsunamis. There is currently a warning system in place to alert residents of the Oregon coast to the approach of tsunamis generated in Alaskan, Chilean, or Japanese subduction zones, but the tsunami generated by an earthquake on the CSZ would arrive without any warning other than the earthquake itself.

Without knowing the exact size and location of future subduction zone earthquakes, it is difficult to predict tsunami run-up heights for the Oregon coast. There are, however, several crude approaches available to get a general feel for the possible magnitude of locally generated tsunamis. The first approach is to look at the “tsunami” sand deposits associated with buried marshes along the coast. This has been done by Peterson and others (1991a), who produced maps of the areas thought to have been inundated by the tsunamis that followed past subduction earthquakes. Unfortunately, all the tsunami deposits are preserved in the modern estuaries, so these maps show only the minimum area covered by the tsunami. Tsunami sands are not preserved if they are deposited on slopes above the bay, so we cannot
use this technique to determine the maximum water level, only the minimum. Peterson and others (1991a) found prehistoric tsunami sands at least 2 kilometers (and possibly 18 kilometers) up Yaquina Bay.

The other approaches to tsunami height is computer modeling. The modeling of waves traveling in water is fairly straightforward, but it is extremely complex to model how the wave behaves when it enters shallow water (less than 50 meters) and interacts with the irregular floor of the shallow sea. It is even more complicated to model how the wave behaves in estuaries. Two attempts have been made to model a locally generated tsunami caused by a subduction zone earthquake. Hebendt (1988) modeled the tsunami likely to accompany a magnitude 9.1 (Mw) earthquake (figure 19). His model shows expected wave height along the Oregon coast at points a few kilometers offshore, thereby sidestepping the shallow-water problem. Clearly, these wave heights, locally as much as 12 meters, represent a serious threat. Baptista and others (1991) have produced a simple model as a prelude to a more complete model. Their initial model is designed to test the sensitivity of tsunami height to various factors and only estimates tsunami height at the latitude of Astoria. Again, this model gives wave height only at a water depth of 50 meters and does not carry the wave onshore. The Baptista and others model suggests that a wave about 7 meters high would be likely from an average subduction zone earthquake. The wave height in this model is very dependent on variables that are still poorly known, so the wave height may not be reliable. The arrival time of the tsunami is much less variable, however, and underscores the unique threat associated with locally generated tsunamis. The tsunami crest in the model reaches the coast 20 to 30 minutes after the earthquake. This is not enough time for an official warning to be issued, so all coastal residents should consider strong ground shaking as a natural tsunami warning and should seek high ground immediately.

The actual height above sea level reached by any tsunami will depend on many local factors, including the offshore wave height, the shape of the shore or estuary, the normal tidal stage at the time, and the amount of coseismic subsidence. It is not unreasonable for many parts of the Oregon coast to expect tsunami run-up of 5 to 10 meters, with inundation extending several kilometers up many estuaries.

**Figure 19.** Computer model of local tsunami in the Pacific Northwest from a hypothetical Mw 9.1 subduction earthquake. Right hand figure shows the pattern of wave elevation for all recording points; the solid line is the average for all points. Wave heights are for points offshore; they cannot be used to estimate coastal run-up or inundation. From Hebendt, 1988.
Conclusions: Should We All Move to Nebraska?

Where does all of this uncertain science leave the residents and decision makers of Oregon's coastal communities? Some may think that we must evacuate the coast forever; others will think we can continue to develop without regard to seismic hazards. The truth, of course, lies in between. Let's look at a few key facts.

- In 150 years or so of our history, there has been no earthquake damage on the coast, yet there has been abundant damage caused by mundane hazards like storms, coastal erosion, and landslides.

- The best geologic data now available strongly suggests, but cannot prove, that most of the coast is susceptible to large damaging earthquakes. These events are certainly rare on human time scales, but could occur at any time.

- The natural geologic makeup of the coast makes it prone to a variety of earthquake hazards, and any large earthquake is likely to cause a large amount of damage.

- It is possible now to make a broad assessment of hazard zones in which individual sites need to be investigated in more detail.

- Lifelines in Oregon coastal communities are likely to be severely impaired in the event of large earthquakes, affecting emergency response operations.

- The long-term economic impact of a large earthquake may destroy communities more thoroughly than the ground shaking.

- No community can afford to "earthquake proof" all of its lifelines and economic infrastructure in the short run.

What should be done, given these facts? Certainly we need more research to answer many of the uncertainties about the earthquake threat, but we know enough to begin to act. Earthquake hazards can be reduced in communities by increasing public awareness of the hazard and by protecting lifelines and structures. The first is relatively inexpensive, and can save many lives. Community groups, the Red Cross, and others can help to educate the community about earthquake and general disaster preparedness. Protecting the infrastructure is economical over the long run, as long as it is integrated into long-range building and land-use plans. Hazardous buildings will probably not get fixed, but they should be replaced by earthquake-resistant structures when their natural life is over. Similarly, facilities sited in hazard zones probably won't get moved, but their replacements should be sited properly. Planning carefully, identifying hazard zones, and considering potential earthquake safety as an element in any development project will lead in the long run to a much more earthquake-resistant Oregon coast. Odds are that we have decades to prepare. We should not squander that opportunity.

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SEISMIC HAZARDS ON THE OREGON COAST—
A RESPONSE

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I am going to limit my discussion to things that I actually have knowledge of, namely landsliding.

In my opinion, landsliding holds the most potential for liability and is the most visible hazard along the Oregon coast, especially between Newport and Lincoln City. This is not to say that landsliding is confined to this portion of the coast; rather, it is one of the most populated areas and subjected to more human activity than most other areas.

Madin has noted that the most slide-prone areas are mudstone bedrock and slopes over 25% and areas of rapid sea cliff erosion or riverbank erosion. I would add terrace deposits overlying seaward-dipping mudstone with slopes as flat as 10 degrees. Typically, the landsliding occurs within a few hundred feet of the beachline and during or after heavy or prolonged rainfall. Severe storms that result in pounding and erosion of the sea cliff compound the land movement.

My area of concern is landsliding connected to the subduction, or severe crustal quake. From my observations of the morphology of the marine terrace deposits up and down the Oregon coast, abnormal drainage patterns appear to be common. Erosion of the Coast Range and nearshore sediments should result in drainage ways perpendicular to the coast. Seemingly more often than not, the drainages are deflected at the margins of, or within, the terrace deposits, and for variable distances they parallel the shorelines, as shown on the contour map example used for figure 1.

Figure 2 is the same map as figure 1 with geologic units delineated from the mapping for DOGAMI Bulletin 81 (admittedly very broad and general). Assuming that the terrace deposits are more erodible than the underlying mudstone bedrock units, one would think that the erosional channels would continue straight toward the beach. An argument could be made that the upper (eastward) margin of the terrace deposits has pulled away from the underlying bedrock, creating a new drainage path. One could also imply from figure 2 that the Astoria and Nye mudstone formations could have undergone similar movements.

These terrace deposits were apparently once uniform sand or poorly indurated sandstone that rested on seaward-sloping or dipping mudstones. From my experience, when excavating the terrace deposits one finds that they are highly fractured and contain large volumes of water. Normal coastal erosion and saturation by heavy rainstorms can cause, and has caused, sections to break off and slide onto the beach. The active sliding is usually within one or two hundred yards of the beach. My concern is that this pattern of fracturing (figure 1) continues many hundreds of yards inland. Observations also show that the fractures farther from the shoreline do not appear to show any recent movement.

Figure 3 depicts a possible sequence of events without specific ages or intervals.

This phenomenon could possibly contain a geologic record in the form of Carbon-14 from buried organics or tree rings (if any old enough still exist) in the base of the ravines. Assuming that all of the fractures did not occur simultaneously, different ages may be established for different events. At the very worst, a most recent event may be isolated.

In summary, I feel that the research is moving steadily forward. This is serious business. I urge the researchers to avoid searching for data to fit preconceived notions (one set of errors can mean hundreds of years for recurrence intervals). Coastal governments should not panic; the probability for disaster was the same in the last decade as it will be in the next.
Figure 3.
A. Uplifted terrace deposits in equilibrium. No disturbance.
B. Subduction quake. Terrace deposits move along bedrock surface, creating fractures parallel to the shoreline. Note movement into zone of maximum erosion potential and parallel to the shoreline. Note movement into zone of maximum erosion potential and downwarping.
C. Long period of quiescence (perhaps today?). Note that beach erosion has moved terrace deposits back to sea level/bedrock contact. Nearshore landsliding is continual as the result of wave undercutting. Ravine slopes reaching natural angle of repose.
D. Subduction quake (tomorrow?). Terrace deposits again move into zone of maximum erosion. Destruction of structures on marine terrace deposits. Ravines open up again.
How do we plan for a catastrophic event that has a low probability of occurring at any given time but that, when it does occur, will have enormous consequences? At the conclusion of his paper, Ian Madin suggests a number of steps various parties should initiate in light of our knowledge about earthquakes in subduction zones. I agree with their general direction and offer the following additional comments.

Emergency Planning

The first step in emergency planning is to increase the level of public awareness. Most Californians know about the San Andreas fault. But how many Oregonians are aware of the potential for a devastating earthquake in their state?

We can learn from public information campaigns in California and perhaps those in the south, where officials are used to dealing with hurricanes. This is an area in which the Federal Emergency Management Act (FEMA) should be doing a lot more.

Any public information campaign will be complicated by the large number of tourists and visitors in coastal communities. How can we reach this group effectively?

Buildings

1. Reinforcing Public Buildings

Ideally, public facilities should be retrofitted to withstand earthquakes. I agree with Madin’s conclusion that little will occur. With budgets limited, such improvements are likely to be a very low priority. Cannon Beach had some experience with this last year. The city hall is of masonry and would not be safe in an earthquake. For this reason, a consultant had recommended extensive repairs. However, after lengthy discussions of the situation, the city council voted to make only minor repairs.

2. Building Codes/FEMA

There is a conflict between FEMA flood regulations, which require the construction of pile-supported buildings in coastal high-hazard areas, and the poor performance of such structures in an earthquake. Is there some way to reconcile this conflict?

The same conflict exists where pile-supported structures have been built in filled estuaries and flood plains. Much of Cannon Beach’s downtown is located in a filled wetland, and I suspect this is not uncommon for other coastal towns located on estuaries.

Land Use Planning

1. Relocation of Threatened Structures

It will be difficult to relocate a public facility that is currently in an area at high risk from tsunamis until that facility is totally worn out. An example of such a structure is the Cannon Beach grade school, which is located on the Ecola Creek estuary, an area extremely susceptible to tsunami hazard.

2. Planning for Tsunami Hazard

Present FEMA mapping and regulations do not consider tsunami hazards, either from a distant earthquake or from one in the subduction zone. Should they? Is it technically feasible to prepare for a tsunami? If so, what might be the implications of incorporating tsunami planning into the regulations, including its effect on insurance rates?

The fact that a tsunami wave could reach 10 meters or more does not leave much room for land use planning in many communities. For example, in Cannon Beach, the elevation of downtown is 12 feet mean sea level (MSL). The area is protected by a dike with a height of 20 to 25 feet MSL. Many of the city’s oceanfront areas have a height of less than 30 feet MSL.
CATASTROPHIC COASTAL HAZARDS IN THE CASCADIA MARGIN U.S. PACIFIC NORTHWEST

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After decades of debate, scientists now believe that the Cascadia subduction zone, encompassing the Pacific Northwest (PNW) coastal zone, is coseismic, that is, predisposed to earthquakes. Prehistoric earthquakes of potentially very large magnitude (>8.5 Mw) are implied by past episodes of abrupt coastal subsidence, tsunami inundation, and sediment liquefaction (table 1; Atwater 1987; Reinhart and Bourgeois 1989; Darienzo and Peterson 1990; Vick 1988; Peterson et al. 1991a; Carver, pers. comm.). The prehistoric subduction zone earthquakes are estimated to have taken place at intervals of between 300 and 600 years, with the last event occurring about 300 years ago.

While earthquake sources, magnitudes, and recurrence intervals in the Cascadia margin are currently being investigated (Shedlock and Weaver 1991) little is being done to establish site-specific risks from the collateral earthquake effects. Locally, these effects can include unconsolidated sediment liquefaction, coastal landslides, tsunami inundation, and persistent shoreline subsidence and related flooding. The magnitude of coastal subsidence (zero to two meters relative sea level rise) could vary regionally, producing extensive beach erosion and severe seasonal flooding in bays and tidal-river flood plains. Beach retreat might shift some shorelines landward by as much as 100 meters. We estimate that as much as 90 percent of the present wetlands and low pastures in some bays will be submerged following the next subsidence event. For the most part, PNW coastal planners at present have little or no site-specific data with which to address concerns about these collateral seismic hazards.

In addition to earthquake hazards, the catastrophic responses of some PNW beaches to the anomalous storm conditions of the 1982-83 El Niño event (Komar 1986; Tuttle 1987) have clearly shown the susceptibility of the beaches to extreme interannual climatic events. Sustained beach erosion, sand dune accretion, or coastal flooding were experienced in many PNW coastal zone beaches following the longshore redistribution of beach sands during the 1982-83 winter period. Some beaches experienced northward sand displacements of 5 to 10 million cubic meters, over multikilometer distances, for a duration of several years (Peterson et al. 1990). The northward shift in beach sand resulted from an unusually oblique approach of winter storm waves associated with anomalously low latitudes of North Pacific storm centers in 1982-83. The delayed return of beach sand to the south (1986 and 1987) followed a two-year period of high-latitude winter storms (1984 and 1985) that were unable to mobilize the northward displaced sand (Peterson et al. 1992). The several years following the 1982-83 El Niño appear to be the most widespread erosional period in the PNW coastal zone during the last several decades.

Locally, the multiyear redistribution of littoral sand (1) stripped beaches to underlying bedrock, (2) exposed sea cliffs and foredunes to direct wave attack, or (3) caused the rapid growth of eolian dune fields (dunes caused by wind). The presence of jetties, for example those at Humboldt Bay and at the mouths of the Siuslaw, Yaquina, and Columbia rivers, might have contributed to the post-El Niño effects of local beach erosion. Furthermore, the long-term effects of sea walls, dune stabilization, and offshore dredge
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Published and unpublished data from PSU Geology Department and other sources listed below:

*Pers. Comm., B. Atwater, USGS and J. Bourgeois, UW
**Pers. Comm., W. Grant, USGS
***Pers. Comm., A. Nelson, USGS
****Pers. Comm., G. Carver, HSU

? Features tentatively identified.

Disposal on littoral sand supply in the PNW coastal zone have not been quantitatively evaluated. Of particular concern are the additive impacts of (1) extreme changes in storm wave climate, (2) physical restrictions to longshore transport, and (3) diminished sand supply on existing beach sand buffers. Because coastal managers have not had much experience with such unusual erosional events, they generally have not considered the potential impacts of interannual redistributions of beach sands during shoreline planning or permitting processes.

In addressing these newly identified coastal hazards, it is important to recognize the diversity of shoreline conditions and associated hazard susceptibilities in the PNW coastal zone. For example, the open ocean shoreline from the Juan de Fuca Straits, Washington, to Cape Mendocino, California (1,000 kilometers in distance), contains some 42 separate beach segments. These segments possibly represent proxies for independent littoral cells (2 to 165 kilometers long) totaling some 770 kilometers, or about 77 percent of the coast (Peterson et al. 1991b). Catastrophic
shoreline erosion could differ between and within these beach segments as a function of the local distribution of beach sand buffers. For example, measured sand volumes in selected beaches range from 15 to 3,400 cubic meters per meter of shoreline (Peterson et al. 1991c). As yet, no quantitative relations between pre-existing sand volume and susceptibility to catastrophic erosion have been established in the PNW coastal zone. Some 38 of the beach segment boundaries, that is, about 45 percent of the cell-bounding headlands, project less than 500 meters seaward of adjacent shoreline embayments. Assuming 0.01 to 0.02 nearshore gradients (slope), these small headlands can be expected to terminate in less than 10 meters of water, well within reported water depths of active sand suspension and transport (U.S. Army Corps of Engineers 1973). However, no field experiments have been conducted to test the effects of these small headlands in restricting longshore transport under highly variable conditions of directional wave climate. For example, chronic beach erosion or dune sand accretion in some cells might result from infrequent events of sand bypassing around small headlands during extreme climatic events. Finally, there have been no studies of the potential long-term flux of beach sand between inshore, offshore, or longshore sand reservoirs following sustained coastal subsidence (decades) associated with earthquake subsidence or uplift.

An increasing concern of many PNW coastal communities is their susceptibility to near-source tsunami hazards. In the event of a megathrust earthquake in the central Cascadia margin, as few as 20 minutes might elapse between the termination of seismic shaking and the advance of the corresponding tsunami (Baptista, pers. comm.). Although evidence of prehistoric tsunami inundation is now established in more than a dozen PNW bays (table 1), the geologic records do not provide accurate estimates of the heights of tsunami run-ups. Preliminary computer numeric models of tsunami generation and shoreward propagation have been developed for the Cascadia margin (Hebenstreit 1988; Baptista, pers. comm.). However, a great deal of work is needed to refine the models for accurate prediction of tsunami onshore run-up (land elevations swept by the tsunami wave) and inshore attenuation (landward distance reached by the tsunami). In addition to the uncertainty of tsunami run-up, the lack of detailed coastal topography (land elevations) severely limits the prediction of site-specific tsunami hazard needed by planners and emergency managers.

Of the beach-fronted PNW coastline, approximately 460 kilometers (60 percent of the total) are backed by unconsolidated dune or bay deposits. The remainder (40 percent of the total) are backed directly by sea cliffs. Unconsolidated beach, dune, and bay sediments within reach of perched water tables are likely to be the foundation soils most susceptible to liquefaction from seismic shaking. Ironically, the flat topography and close proximity of these deposits to modern shorelines make them very appealing to private and commercial developers. Although liquefiable deposits have been mapped in the Portland and Seattle metropolitan areas, they have not been regionally mapped or systematically tested for liquefaction potential anywhere in the PNW coastal zone.

Seasonal and interannual variations in eolian dune sand supply are major complicating factors in coastal planning for shoreline development, jetty maintenance, harbor mouth dredging, and dune habitat ecology. Surprisingly little information exists regarding the site-specific rates of beach sand transport by eolian processes in the PNW coastal zone. It has been suggested that sand supplies to dune fields are alternately terminated and reactivated following periods of coseismic cycles of subsidence and uplift, respectively (Hunter, pers. comm; Carver, pers. comm.). Unfortunately, there have been few geologic studies of the origin of the major dune fields, their timing of formation, or their long-term growth dynamics since Cooper’s pioneering work (Cooper 1958 and 1967). Finally, there have been no quantitative, site-specific studies on the long-term effects of the “locking up” of beach sand in artificially stabilized dune fields, for example, foredunes stabilized by dune grass plantings or shore protection structures.

Most of the beach-fronted sea cliffs contain poorly consolidated Pleistocene deposits overlying wave-cut marine terraces, tectonically
upwarped between 0 and 120 meters above present sea level. The longshore distribution of modern sea cliff failures appears to vary widely in northern Oregon (Galster 1987; Komar and Shih 1991) as well as throughout the PNW. Although some 90 percent of the observed sea cliffs in the PNW coastal zone are oversteepened, less than 10 percent of modern sea cliff shoreline (pre-1982-83 El Niño) shows evidence of catastrophic slope failure (Peterson et al. 1992). In addition, we find no regional correlations between reported modern uplift rates (Mitchell et al. 1991) and apparent sea cliff retreat in the Cascadia margin. We speculate that periods of rapid sea cliff retreat immediately follow coseismic subsidence events or anomalous conditions of beach sand redistribution. The susceptibility of existing sea cliffs to future erosion and retreat, due either to coseismic tectonic subsidence (abrupt sea-level rise) or interannual events of sand redistribution by anomalous wave climate, have yet to be evaluated in the Cascadia margin.

In conclusion, the PNW coastal zone is particularly susceptible to Cascadia margin earthquakes from the multiple threats of (1) relative proximity to earthquake epicenters, (2) near source tsunami run-up, (3) abundance of liquefiable foundation soils, and (4) persistent coastal subsidence and flooding. The less dramatic, but potentially more frequent, events of unusual wave climate make “apparently stable” shorelines in the PNW coastal zone far more dynamic than previously assumed. Finally, increasing development pressures on shoreline properties are certain to yield increasing land-use conflicts between people who want to build artificial structures and the natural dynamics of shoreline erosion or accretion. Coastal planners, emergency managers, and the general public need comprehensive assessments of potential, catastrophic shoreline hazards resulting from earthquakes and extreme climatic conditions in the Cascadia margin. Focused research efforts are now needed to provide site-specific information for catastrophic hazard mitigation in the Pacific Northwest coastal zone.

Acknowledgments

Work on natural coastal hazards performed by the Geology Department at Portland State University has been recently supported by the National Coastal Research Institute grants no. 2-5632- 03 and CZ17.90-5635-01, the USGS National Earthquake Preparedness Program grants 14-08-0001-G1512 and 14-08-0001-G2120, the Oregon Department of Geology and Mineral Industries interagency agreements 1989-1991, and the National Science Foundation grant EAR-8903903.

References


Introduction

Visitors to the Oregon coast are impressed by the tremendous variety of its scenery. The low rolling mountains of the Coast Range serve as a backdrop for most of the length of its ocean shore. In the south the Klamath Mountains extend to the coast, and the edge of the land is characterized by high cliffs being slowly cut away by ocean waves. The most resistant rocks persist as sea stacks scattered in the offshore. Sand and gravel are able to accumulate only in sheltered areas, where they form small pocket beaches within the otherwise rocky landscape.

The more extensive stretches of beach are found in the lower-lying parts of the coast. The longest continuous beach extends from Coos Bay northward to Heceta Head near Florence, a total shoreline length of some 60 miles. This beach is backed by the impressive Oregon Dunes, the largest complex of coastal dunes in the United States. Along the northern half of the coast there is an interplay between sandy beaches and rocky shores. Massive headlands jut out into deep water, their black volcanic rocks resisting the onslaught of even the largest storm waves. Between these headlands are stretches of sandy shoreline.

Figure 1: Coastal landforms of Oregon, consisting of stretches of rocky shorelines and headlands, separating pockets of sandy beaches. (From Komar [1985])
whose lengths are governed by the spacings between the headlands (figure 1). Portions of these beaches form the ocean shores of sand spits such as Siletz, Netarts, Nehalem, and Bayocean. Landward from the spits are bays or estuaries of rivers that drain the Coast Range.

The first western explorers and settlers were attracted to the Oregon coast by the potential richness of its natural resources. Earliest were the traders, who obtained pelts of ocean otter and beaver from the Indians. Later came prospectors, who sought gold in the beach sands and coastal mountains, but who in many cases were content to settle down and “mine” the fertile farm lands found along the river margins. Others turned to fishing, supporting themselves by harvesting the abundant Dungeness crab, salmon, and other fish in the coastal waters. Also important to the early economy of the coast were the vast tracts of cedar and sitka spruce. Their significance continues to the present. However, today the most important “commodity” for the Northwest coastal economy is the vacation visitor. Vacationers arrive by the thousands during the summer months.

It is still possible, in spite of the number of tourists who visit the state, to leave Highway 101 and find the seclusion of a lonely beach or the stillness of a trail through the forest. However, there is cause for concern that the qualities of the Oregon coast we cherish are being lost. Like most coastal areas, Oregon is experiencing developmental pressures. Homes and condominiums are being constructed immediately behind the beaches, within the dunes, and atop cliffs overlooking the ocean. Everyone wants a view of the waves, passing whales, and the evening sunset, as well as easy access to a beach, but these desires are not always compatible with nature. As a result, increasingly homes are being threatened and sometimes lost to beach erosion and cliff landslides. Such problems can usually be avoided if builders recognize that the coastal zone is fundamentally different from inland areas because of its instability. Builders need some knowledge of ocean waves and currents and how they shape beaches and attack coastal properties. In addition, they need to understand and recognize potential instabilities of the land that might cause it to suddenly slide away. A familiarity with the processes and types of problems experienced in the past can aid in the selection of a safe location for a home. It can also enhance people's enjoyment of the coast, and is hoped, lead to an appreciation of the qualities of the Oregon coast that must be preserved.

**Tectonic Setting and Geomorphology**

The tectonic setting of the Oregon coast is extremely important to the occurrence and patterns of erosion. Significant is the presence of active sea-floor spreading beneath the ocean to the immediate west. New ocean crust forms at the Juan de Fuca and Gorda ridges, adding to the Juan de Fuca and Gorda South plates. These oceanic plates, which are moving generally eastward toward the continent, collide with the North American plate, which includes the continental land mass. The collision zone lies along the margin of the coasts of Washington, Oregon, and northern California. There is evidence that the oceanic plates have been undergoing subduction beneath the continental North American plate, evidence which includes the still-active volcanoes of the Cascades, the existence of marine sedimentary rocks accreted to the continent, and the occurrence of vertical land movements along the coast.

Most of the marine sediments deposited on the oceanic plates are scraped off during the subduction process and accrete to the continental plate. The addition of ocean sediments to the continent has led to the long-term westward growth of the Pacific Northwest. The oldest rocks found in the Coast Range date back to the Paleocene and Eocene periods, some 40 to 60 million years ago. These accreted marine sediments, mainly gray mudstones and siltstones, can be seen in many sea cliffs along the coast (figure 2). As will be discussed in a later section, the presence of these mudstones is important to the erosion of sea cliffs and particularly to the occurrence of landslides.

In addition to the Tertiary mudstones, many sea cliffs contain an upper layer of clean sand (figure 2). These are Pleistocene marine terrace deposits and consist of uplifted beach and dune sands. In some areas the Pleistocene sands form the entire sea cliff, with no outcrop of Tertiary mudstones beneath. The flat marine terrace seen
in figure 2 is the lowermost and youngest terrace of a series that in some places form a stairway up the flank of the Coast Range. The presence of this stairway documents that the Oregon coast has been tectonically rising for hundreds of thousands of years, while at the same time the sea level has oscillated because of the growth and retreat of glaciers.

The general uplift of the Northwest coast is also demonstrated by records from tide gauges where the hourly measurements are averaged for the entire year, removing the tidal fluctuations and leaving the mean sea level for that year (Hicks et al., 1983). Examples of up to 80 years in length obtained by yearly averaging are shown in figure 3. Each record reveals considerable fluctuations in the level of the sea from year to year, with many small ups and downs. The sea level in any given year is affected by many oceanic and atmospheric processes. These processes cause the irregular fluctuations.

In spite of such irregularities, most tide-gauge records reveal a long-term rise in the sea that can be attributed in part to the melting of glaciers. The record from New York City in figure 3 is typical of such analyses. In that example the long-term average rise is 3.0 millimeters a year, about 12 inches a century (1 inch = 25 millimeters). The record from Galveston, Texas, also shows a rise, but the average rate is much higher at 6.0 millimeters a year (24 inches a century). The actual level of the sea cannot be going up faster at Galveston than at New York City—the discrepancy results from changing levels of the

Figure 3: Yearly changes in sea level determined from tide gauges at various coastal stations. (After Hicks [1972])
combined effects of the local land subsidence and the actual rise in sea level. An extreme case of this is Juneau, Alaska, figure 3, which is tectonically rising at a rate that is faster than the rise in sea level. Its tide-gauge record, therefore, indicates a net fall in the water level relative to the land.

The record from the tide gauge at Astoria, Oregon, is included in figure 3—the level of the sea there has remained relatively constant with respect to the land. This must indicate that during at least the last half century, Astoria has been rising at just about the same rate as the sea. A detailed analysis of the measurements from the Astoria gauge indicates that the land is actually rising slightly faster than the water, the net increase in the land relative to the sea being 0.1 to 0.2 millimeters a year. This change is small, amounting to a 10- to 20-millimeter (less than an inch) increase in land elevation if it continued for 100 years. The land must be rising at a faster rate at Neah Bay on the north coast of Washington, where the net rate is 1.3 millimeters a year (5 inches a century) in excess of the global sea-level rise, and at Crescent City in northern California, with 0.7 millimeter a year, or 2.8 inches a century, of net land emergence (Hicks et al. 1983).

Data from geodetic surveys collected by the National Geodetic Survey permit us to infer the movement of the land relative to the sea along the remainder of the Oregon coast. Vincent (1989) and Mitchell et al. (1991) have analyzed the geodetic data along a north-south line extending the full length of the Oregon coast. To establish elevation changes, they compared surveys made in 1931 and 1988; the values are graphed in figure 4. The movement so determined is relative rather than absolute, so the elevation changes have been normalized to the bench mark in Crescent City. Accordingly, the elevation change scale on the left of the diagram gives 0 for Crescent City, while positive values for other locations represent an increase in elevation relative to Crescent City and negative values indicate reduced elevation relative to Crescent City. (However, the elevation could still involve tectonic uplift.) The overall pattern seen in figure 4 indicates that the smallest uplift has occurred along the north-central coast between Newport and Tillamook, with progressively higher uplift further south and along the very northernmost portion of the coast toward Astoria and the Columbia River. The first scale on the right of figure 4 indicates the equivalent rates, calculated as the elevation changes divided by the lapsed time between the surveys (1988-1931 = 57 years). The differential rates are significant; for example, they amount to 2 to 3 millimeters a year when we compare Astoria and the south coast with the Newport and Lincoln City areas. It is possible to use the tide-gauge data to convert the elevation changes relative to Crescent City determined by Vincent (1989) into rates relative to the annual change in the global level of the sea. This is done simply by shifting the first scale on the right of figure 4, that relative to the Crescent City bench mark, by an amount 0.7 millimeter a year determined from the tide gauge at that location. This shift yields the rate scale farthest to the right in figure 4, the rate of land-level change relative to the changing global sea level. A positive value again indicates that the elevation of the land is increasing relative to the sea, while a negative value corresponds to inundation of the land by the rising sea. This coast-wide shift of the

Figure 4: Elevation changes and their relationship to sea-level rise along the length of the Oregon coast from Crescent City in California north to Astoria on the Columbia River, based on repeated geodetic surveys along the coast. (After Vincent [1989])
scale by 0.7 millimeters a year, based on the tide
gauge at Crescent City, indicates that Astoria at
the far north is rising faster than the sea by an
amount on the order of 0.1 to 0.2 millimeter a
year, the same measurement recorded by the tide
gauge at that location. These matching data con-
firm (1) the validity of the geodetic data analyzed
by Vincent to determine elevation changes and
(2) the analyses undertaken to convert that data
into a rate of change that can be compared with
the increasing level of the sea.

According to the results graphed in figure 4,
the southern half of the Oregon coast is currently
rising faster than the global sea level, as is the far
north coast near Astoria. Conversely, the central
stretch between Newport and Tillamook is being
submerged by the rising sea. The latter rates are
on the order of 1 to 2 millimeters a year (4 to 8
inches a century), and therefore are small com-
pared with submergence rates experienced on
most coastlines: rates of 4 to 6 millimeters a year
(16 to 24 inches a century) are common along the
east and Gulf coasts of the United States (figure
3). The global rise in sea level has been estimated
by various workers to be on the order of 1 to 3
millimeters a year (4 to 12 inches a century), the
large range being due to the difficulty of separat-
ing that worldwide component from local tec-
monic and isostatic effects included in records
from tide gauges. Assuming that the eustatic rise
in sea level is on the order of 2 millimeters a year
(8 inches a century), the results from figure 4 in-
dicate that the south coast of Oregon is tectoni-
cally rising at about 2 to 3 millimeters a year (8 to
12 inches a century) whereas the stretch between
Newport and Tillamook is approximately stable,
neither rising nor falling tectonically.

It is apparent that the along-coast differences
between tectonic uplift and changing levels of the
sea deduced from figure 4 will be relevant to spa-
tial patterns of coastal erosion. However, there
also appears to be a temporal change in the tec-
nonics that is important to erosion. Earthquake
activity is generally associated with a subduction
zone such as that in the Northwest, where seismic
events are triggered by the plates scraping to-
gether as the oceanic plate slides beneath the con-
tinental plate. The Northwest coast is anomalous
in that respect in that there have been no historic
earthquakes which can be attributed to plate
subduction. However, recent evidence suggests
that the plates are temporarily locked together and
that the 200-year historical record from the
Northwest is too short to establish whether earth-
quakes do accompany subduction. This evidence
has come from investigations of estuarine marsh
sediments buried by sand layers, deposits which
suggest that during prehistoric times portions of
the coast have abruptly subsided, generating an
extreme tsunami that swept over the area to de-
posit the sand (Atwater 1987; Atwater and
Based on the number of such layers found in
Willapa Bay, Washington, and Netarts Bay, Or-
egon, it has been estimated that catastrophic
earthquakes have occurred at least six times in the
past 4,000 years, at intervals ranging from 300 to
1,000 years. The last recorded event took place
about 300 years ago. Therefore, there is strong
evidence that major subduction earthquakes do
occur along the Northwest coast, but with long
periods of inactivity between events.

An earthquake releases strain built up by sub-
duction. This results in some areas of the coast
dropping by 1 to 2 meters (3 to 6 feet) during the
release, while other areas undergo minimal sub-
sidence. Between earthquake events the strain
accumulates; this produces a general uplift of the
coast as recorded by the tide gauges and geodetic
surveys within historic times (figures 3 and 4).

Another potential change in the present-day
pattern of sea-level rise versus coastal uplift is
associated with predictions that future greenhouse
warming will accelerate the rise in sea level.
Some scientists have predicted that global tem-
peratures will increase from 1.5° to 4.5° by the
year 2050 (National Research Council 1983).
These predictions in turn have led to a variety of
estimates for accelerated sea-level rise caused by
increased glacial melting and thermal expansion
of seawater. For example, a report by the Na-
tional Research Council (1987) predicts that by
the year 2050, the global sea level will have risen
10 to 21 centimeters (4 to 8 inches). Although
this may seem insignificant, the effects on sandy
shorelines may be magnified 100 times in the
horizontal direction, resulting in shoreline erosion
of 10 to 21 meters (33 to 70 feet). There are many
uncertainties in these analyses of sea-level rise
caused by greenhouse warming, and the resulting
predictions have been controversial among scientists. Different investigators studying sea-level curves derived from tide gauges have reached conflicting results, some concluding that they see an increase in the rate of rise in recent decades and others concluding that they do not. Despite the uncertainties, there is a growing consensus that some increased rate of sea-level rise can be expected in the next century. This recognition has led to recommendations that future sea levels be given more serious consideration in coastal management decisions.

Ocean Processes as Agents of Erosion

The Northwest coast is one of the most dynamic environments in the world. Ocean waves and currents continuously reshape the shoreline. Portions of the beach are cut away while others are built out. Severe storms strike the coast during the winter, generating strong winds that drive rain against sea cliffs and homes and form huge ocean waves that crash against the shore. Beaches, giving way to waves and currents, retreat toward the land. At times this beach loss continues until the erosion threatens structures and cuts away at public parklands.

Ocean Waves

The extreme seasonality of the Oregon climate results in parallel variations in ocean processes that exert the primary control on natural cycles observed on beaches. The energy of ocean waves parallels the seasonality of storm winds because the strength of those winds is the primary factor in causing the growth of waves. In general, the greater the wind velocity blowing over the surface of the ocean, the higher the resulting waves. Other factors are involved in addition to the wind speed. One is the duration of the storm—the longer the winds blow, the more energy they are able to transfer to the waves. The third factor is the fetch, the area or ocean expanse over which the storm winds are effective. Fetch operates much like storm duration in that the area of the storm governs the length of time the winds are able to act directly on the waves. As the waves are forming they move across the ocean surface and may eventually pass beyond the area of the storm so they no longer acquire energy from the winds. The importance of fetch is apparent when one contrasts wave generation on the ocean with that on an inland lake. The fetch on the lake can be no greater than its length, so the waves can acquire only a small amount of energy from winds before they cross the entire lake and break on the shore.

Wind-generated waves are important as energy-transfer agents. They obtain their energy from the winds, transfer it across the expanse of the ocean, and finally deliver it to the coastal zone when they break on the shoreline. Therefore, the storm need not be in the immediate coastal zone. Waves reach the shores of Oregon from storms all over the Pacific, even from the southern hemisphere near Antarctica. However, the largest waves reaching Oregon derive from winter storm systems that move down from the north Pacific and Gulf of Alaska.

Ocean waves reaching the shores of Oregon are measured daily by a unique system, a microseismometer like those usually employed to measure small earth tremors. In this application the microseismometer senses ground movements produced by ocean waves as they reach the shore and break. Many Coast Guard stations in the Northwest now use this system to obtain better estimates of wave conditions than were formerly determined visually. A microseismometer system is also in operation at OSU Hatfield Marine Science Center in Newport; it is connected to a recorder to obtain a permanent record of the waves. This system has been in operation since November 1971 and has yielded the longest continuous record of wave conditions on the west coast of the United States. These measurements have been valuable in research examining the causes of beach erosion along the Oregon coast.

It might come as a surprise that a microseismometer at the Marine Science Center can provide records of ocean waves—after all, the center is nearly two miles from the ocean. However, even more impressive is that the waves can be detected on the seismometer at Oregon State University in Corvallis, 60 miles inland. When the surf is high on the coast, its effects can be seen as small jiggles in the seismometer recordings.

The microseismometer at the Marine Science Center differs from normal seismometers in that
it is tuned to amplify small tremors, whether they are caused by earthquakes too minor to be felt or by ocean waves along the coast. To use the record from the microseismometer to measure ocean waves, it was necessary to first calibrate the system (Creech 1981; Zopf et al. 1976). This was accomplished by obtaining direct measurements of waves in the ocean at the same time their tremors were measured with the microseismometer. The direct measurements of waves were collected with a pressure transducer, an instrument that rests on the ocean bottom and records pressures that are directly proportional to the heights of the waves passing over the transducer. This is the most common method for directly measuring ocean waves, and it would be preferable to use such an instrument rather than a microseismometer. However, winter storms experienced along the Northwest coast are so intense they usually destroy pressure transducers or other wave-measuring instruments that must be placed in the water. On this coast we need a microseismometer that can remain at the Marine Science Center, safe from the reach of waves. Although the direct comparisons between the pressure-transducer records and those obtained with the microseismometer lasted only a few months, the results showed that the motions on the microseismometer are directly proportional to the heights of the offshore waves. Now only the microseismometer is needed to monitor daily ocean-wave conditions.

An example of daily wave measurements obtained from the microseismometer is shown in figure 5, covering the period from mid-December 1972 to mid-January 1973. Most apparent in this series are the storm waves that struck the coast during Christmas. The breaker heights at that time reached 7 meters, about 23 feet, roughly the height of a three-story building. This reported height represents what is termed a “significant wave height,” defined as the average of the highest one-third of the waves. The significant wave height can be evaluated from measurements of the waves obtained using wave-sensing instruments. However, it turns out that the significant wave height also roughly corresponds to a visual estimate of a representative wave height. This is because observers normally tend to weight their observations toward the larger waves, ignoring the smallest. There will of course be many individual waves that are still higher than this reported significant wave height, which remains something of an average. Measurements have shown that the largest wave height during any 20-minute interval will be a factor of about 1.8 times.

![Figure 5: An example of daily variations in wave conditions measured by the microseismometer at Newport, covering the interval from December 1972 through January 1973. (From McKinney [1977])](image-url)
the significant wave height (Komar 1976). Therefore, when the graph of figure 5 indicates the occurrence of a significant wave height of 23 feet during Christmas 1972, there must have been individual waves of about 1.8 x 23 feet—41 feet high! As might be expected, there was considerable erosion along the coast during that storm, the severest impact having been at Siletz Spit on the mid-Oregon coast.

Figure 6 gives an example of annual changes in wave-breaker heights measured by the microseismometer. The measurements were obtained from July 1972 through June 1973 but are typical of annual variations (Komar et al. 1976a). These data again represent significant wave heights. The solid line gives the average of the significant breaker heights measured during each one-third month interval. It shows that the breakers are on the order of 2 meters high (7 feet) during the summer months and nearly double to about 4 meters (13 feet) in the winter. The dashed lines are the maximum and minimum wave breaker heights that occurred during those one-third month intervals; these extremes provide a better impression of the effects of individual winter storms. The largest waves recorded within this 1972-73 period (the storm waves that are shown on a daily basis in figure 5) reached the coast during the final third of December 1972.

Although extremely high, the waves during that December 1972 storm are well below the largest that have been measured off the Northwest coast. In the early 1960s, a wave-monitoring program on offshore rigs exploring for oil measured an individual wave having a height of 95 feet (Rogers 1966; Watts and Faulkner 1968). This is close to the 112-foot height of the largest wave ever reliably measured in the ocean. It was observed from a naval tanker traveling from Manila to San Diego in 1933 (Komar 1976). All of the measurements on the Oregon coast confirm that it has one of the highest wave-energy climates in the world.

**Beach Cycles on the Oregon Coast**

Beaches respond directly to the seasonal changes in wave conditions. The resulting cycle (illustrated schematically in figure 7) is similar on most coastlines. The beach is cut back during the winter months of high waves when sand is eroded from the shallow underwater and from the beach berm (the nearly horizontal part of the beach.

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**Figure 6:** The monthly variations of wave breaker heights and periods at Newport, illustrating the occurrence of higher wave conditions during the winter months. The solid line is for the mean heights (significant wave heights) for one-third month intervals; and the dashed lines are for the largest and smallest breakers for those intervals. (From Komar et al. 1976a)
profile which is above the high-tide line). This eroded sand moves to deeper water where it accumulates in offshore bars, approximately the zone where the waves first break as they reach the coast. Sand movements reverse during the summer months of low waves, moving back onshore from the bars to accumulate in the berm. Although this cycle between two beach-profile types is approximately seasonal due to changing ocean waves, the response is really one to high storm waves versus low regular swell waves. At times, low waves can prevail during the winter and the beach berm may actually build out, although not generally to the extent of the summer berm. Similarly, should a storm occur during the summer, the beach erodes.

This cycle has been demonstrated to occur on Oregon beaches, just as along other coasts. In one study, profiles were obtained monthly during the winter of 1976-77 from two beaches, that to the south of Devil's Punchbowl at Otter Rock and that at Gleneden Beach south of Lincoln City (Aguilar and Komar 1978). These two beaches were selected because of their contrasting sand sizes, which produce marked differences in overall slopes of the profiles. The sediment grain size is the primary factor that governs the slope of a beach, the slope increasing with increasing grain size. Gravel beaches are the steepest, their slopes sometimes reaching 25 to 30 degrees, whereas the overall slope of a fine-sand beach may be only 1 to 2 degrees. This is seen in the comparison of beach profiles of Otter Rock and Gleneden Beach, figure 8, the latter being coarser and hence steeper.

The month-by-month changes in the profiles at Gleneden Beach are shown in figure 9. These profiles were obtained by using standard...
surveying gear and by wading into the water. They do not show the offshore bars that were too deep to reach. However, these profiles do illustrate the rapid retreat of the beach as the winter season develops. The erosion began as early as October and continued through the spring. The return of sand to the berm and the buildup of the beach did not take place until April through June. The cycle of profiles at the Otter Rock beach was basically the same, at least in its timing. However, the magnitude of the change was much smaller than at Gleneden Beach. Sand elevations at Gleneden changed by as much as 2 to 3 meters (8 feet) (figure 9), while the changes at Otter Rock amounted to less than 1 meter (3 feet). This difference again can be attributed to differences in grain sizes between these two beaches. In general, the coarser the grain size of the beach sand, the larger the changes in its profile in response to varying wave conditions. The response to storms is also much faster for the coarser-grained beach: the storm waves not only cut back the coarser beach to a greater degree but also erode it at a much faster rate. Here nature goes counter to what might intuitively have been expected.

This greater response of coarser-grained beaches to storm waves is important to coastal erosion processes since the waves are able to cut rapidly through the beach to reach homes and other structures. This fact points to the general role of the beach as a buffer between the ocean waves and coastal properties. During the summer when the beach berm is wide, the waves cannot reach the properties. Erosion is not a problem, thanks to the buffer protection offered by the beach. However, when the beach is cut back during the fall and early winter, it progressively loses its buffering ability and property erosion is more likely. If a storm strikes the coast in October, there may be enough beach to serve as a buffer so

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**Figure 9:** A series of beach profiles obtained at Gleneden Beach, Oregon, illustrating the seasonal variations for Oregon coast beaches as shown schematically in figure 7. (From Aguilar and Komar [1978])
that property erosion does not occur. It is only when the beach berms completely disappear and the waves can wash against cliffs and foredunes that the potential for property losses is great. This is often the condition from about November through March, but in fact the extent of the remnant berm is extremely variable along the coast, as is the parallel threat of property erosion. This longshore variability results from the patterns of nearshore currents which assist the waves in cutting back the beach.

Nearshore Currents and Sediment Transport

Waves reaching the coast generate currents in the nearshore zone that are important to sand movements on the beach and thus to erosion processes. These wave-generated currents are independent of ocean currents that exist farther offshore since those deep-ocean flows do not extend into the very shallow waters of the nearshore.

Most of the time waves along the Oregon coast approach the beaches with their crests nearly parallel to the shoreline. Under such circumstances the nearshore currents take the form of a cell circulation, the most prominent part of which is the seaward-flowing rip currents (figure 10). The rip currents are fed by longshore currents flowing roughly parallel to shore, but extending along only a short stretch of beach. The currents of this cell circulation are able to move sediments and thus to affect the morphology of the beach. The longshore currents hollow out troughs into the beach, generally increasing in width and depth as a rip current is approached. Rip currents can be very strong, cutting through the offshore bars to produce deeper water and a steeper but more uniform beach slope. The rips move sand offshore and thereby tend to erode crescent-shaped embayments into the beach berm. Aerial views of the coast typically show beaches that are extremely irregular, consisting of a series of rip embayments of various sizes together with troughs cut by the longshore currents and rip currents (figure 11). At times these rip-current embayments extend across the entire width of the beach and begin to cut into foredunes and sea cliffs. Such rip embayments have played a major role in property losses due to erosion. Although rip embayments seldom produce much property erosion on their own, they have the effect of eliminating the buffer protection of the beach berm. When a storm occurs, the waves are able to pass through the deep water of the rip embayment, not breaking until they reach the properties. Thus, rip embayments can control the center of attack by storm waves. The resulting erosion is commonly limited in longshore extent to only one or two hundred yards; this is the longshore span of a rip embayment that reaches the foredunes or sea cliff (figure 12).

When waves break at an angle to the beach, they generate a current that primarily flows parallel to the shoreline. However, even then seaward-flowing rips may be present. This longshore current, together with the waves, produces a transport of sand along the beach, a sand movement that is known as "littoral drift." This is more than a local rearrangement of the beach sand with accompanying topographical changes as produced by rip currents and the cell circulation. Instead, the littoral drift may involve along-coast movements that displace sand by many miles.

On Oregon beaches the waves tend to arrive from the southwest during the winter and from
the northwest during the summer (corresponding to changes in wind directions). As a result, there is a seasonal reversal in the direction of littoral drift—north in the winter, south during the summer. The net littoral drift is the difference between these north and southward sand movements. Along most of the Oregon coast this net drift is essentially zero, at least if averaged over a number of years. This is demonstrated by the absence of continuous accumulations of sand on one side of jetties or rocky headlands, with

Figure 12: A schematic diagram illustrating how rip currents erode embayments that can cut through the beach and locally threaten properties.

erosion on what would be the downdrift side (Komar et al., 1976b). Patterns of sand accumulation and erosion on opposite sides of jetties, figure 13A, are found on many coasts where there is a net littoral drift. For example, along the shores of southern California and most of the east coast of the United States, erosion in the downdrift directions from jetties has caused major problems and considerable loss of property (Komar, 1976, 1983b). In contrast, when jetties have been built on the Oregon coast, sand has accumulated on both their north and south sides. This pattern is diagramed schematically in figure 13B and is illustrated specifically by the Yaquina Bay jetties in figure 14. In the case of the Yaquina Bay jetties, more sand accumulated to the south than to the north, but this was due to the oblique orientation of the jetties to the overall trend of the coastline and because the pre-jetty shoreline curved significantly inward toward the bay. More significant is that sand accumulated both north and south of the jetties until the embayments between the jetties and the pre-jetty shoreline filled and an equilibrium shoreline developed. Subsequent to achieving equilibrium, there has been almost no change in the shoreline configuration. The sand that accumulated adjacent to the jetties derived from erosion of the beaches more distant from the jetties, and so an overall symmetrical pattern emerged, one that is significantly different from the asymmetrical pattern found on coasts where there is a large net littoral drift (compare figure 13A with figure 13B). This reduces the potential for major erosional and property losses due to the construction of jetties on the Oregon coast, at least compared with other coasts where there is a large net littoral drift. However, one severe erosion problem did occur on the Oregon coast in direct response to jetty construction, that which led to the destruction of the town of Bayocean (discussed below).

The Pocket-Beach Nature of the Oregon Coast and Sources of Nearshore Sands

The ultimate cause of the zero net littoral drift of sand along the Oregon coast is that the beaches are contained between rocky headlands, in effect forming pocket beaches (figure 1). The headlands are large and extend to sufficiently deep water to prevent beach sand from passing around them. Therefore, the sand within each