ES106 Earth System Science
Summer 2014

Field Trip – Coastal Processes and Geology of the Lincoln City Area

Compiled by Dr. Taylor
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August 26, 2014

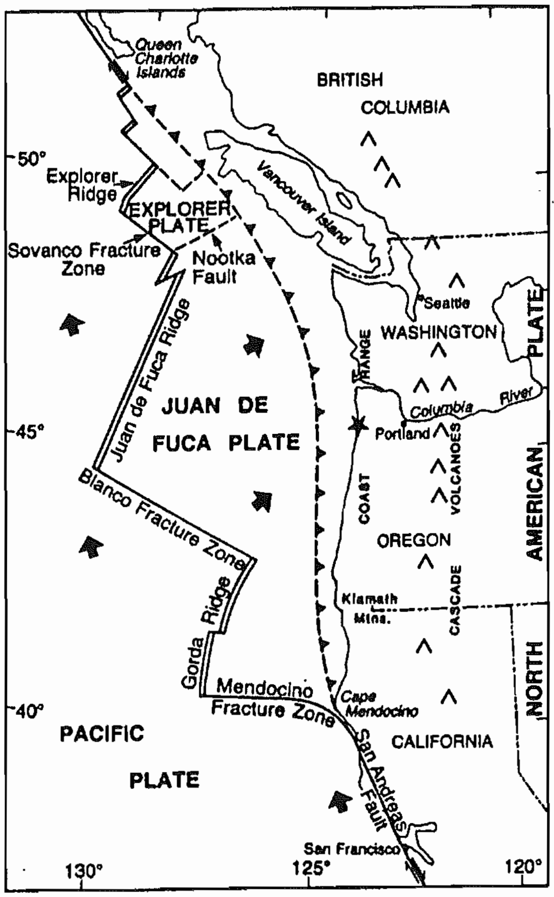

Fig. 1. Tectonic map of the Cascadia Margin. Star marks the study site. The approximate location of the trench axis, marking the boundary between the North American plate and the subducting Juan de Fuca Plate, is shown by a thrust boundary (dashed line).

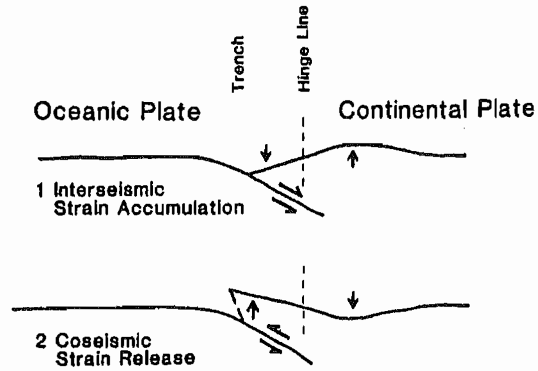

Fig. 2. Diagram of vertical coastal tectonics associated with (1) coupled strain accumulation and (2) coseismic shear dislocation between a subducting oceanic plate and an overriding continental plate [after Ando and Balazs, 1979].

margin, north of 46° latitude, indicate tensional forces associated with trench normal subduction of the northern Juan de Fuca plate [Taber and Smith, 1985; Weaver and Baker, 1988].

One test of active subduction tectonics in the Cascadia margin would be the evidence of cyclic uplift and abrupt subsidence of coastal areas. Such vertical tectonics are forced by the alternation of interseismic coupling (strain accumulation) and coseismic shear dislocation (strain release) of the subducting oceanic plate and overlying continental plate [Fitch and Scholtz, 1971]. When the subducting plate and overlying continental plate are coupled, the leading edge of the continental plate is dragged downward, producing an associated uplift on the opposite (landward) side of a flexure hinge line (Figure 2). When interseismic stresses overcome the frictional coupling, the coseismic strain release results in abrupt tectonic uplift (seaward of the hinge line) and corresponding abrupt subsidence (landward of the hinge line). Significantly, such abrupt vertical displacements of the sea floor produce large tsunamis [Heaton and Hartzell, 1986]. The alternation of coastal uplift and abrupt coastal subsidence together with tsunami deposition provides a potentially unique record of interplate paleoseismicity in strongly coupled subduction zones.

Such events of abrupt coastal subsidence (1-3 m) on the landward side of trench parallel hinge lines were observed in association with the 1946 Nankaido earthquake [Fitch and Scholz, 1971], the 1960 Chile earthquake [Plafker and Savage, 1970] and the 1964 Alaska earthquake [Savage and Hastie, 1966; Plafker, 1972; Ovenshine et al., 1976]. Multiple events of abrupt submergence of late Holocene wetland surfaces have also recently been reported for the coast of southwest Washington [Atwater, 1987] and for northwestern Oregon (W. Grant, personal communication, 1988). These records of relative sea level change are interpreted as positive evidence for active subduction tectonics (including coseismic subsidence) along the central Cascadia margin. Similar field evidence of abrupt changes in relative sea level have been reported for Netarts Bay in northern Oregon [Peterson et al., 1988].

In this paper we describe in detail the stratigraphy and tectonic implications of buried marsh deposits from Netarts Bay, a coastal lagoon in northern Oregon (Figure 3). The small marsh of Netarts Bay was chosen for a detailed study of late Holocene records of relative sea level change on the basis of its protection from ocean storm waves and its negligible fluvial influence. These conditions are important to insure as complete and uncomplicated a record of relative sea level change as possible. In addition, its central position in the Cascadia margin (45°N) allows for the comparison of neotectonic processes in northwestern Oregon with reported marsh burial events in larger estuaries of southwestern Washington

Figure 1. Cascadia Subduction Zone and northern Oregon estuaries. Distance between the Necanicum and Alsea estuaries is approximately 175 km.

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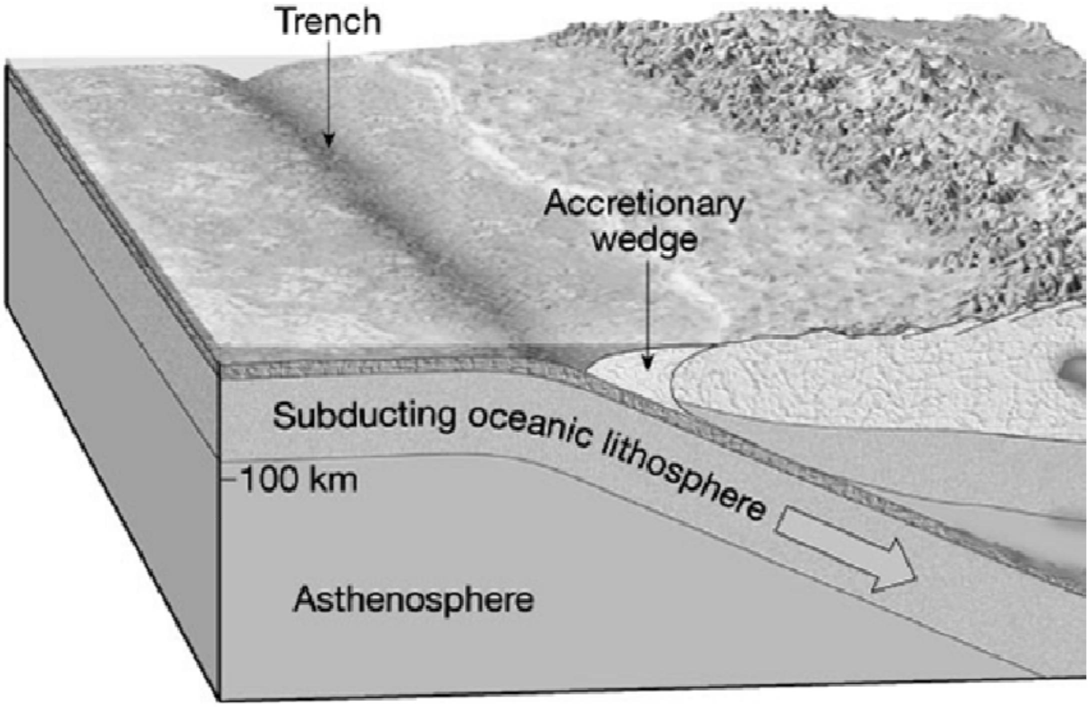
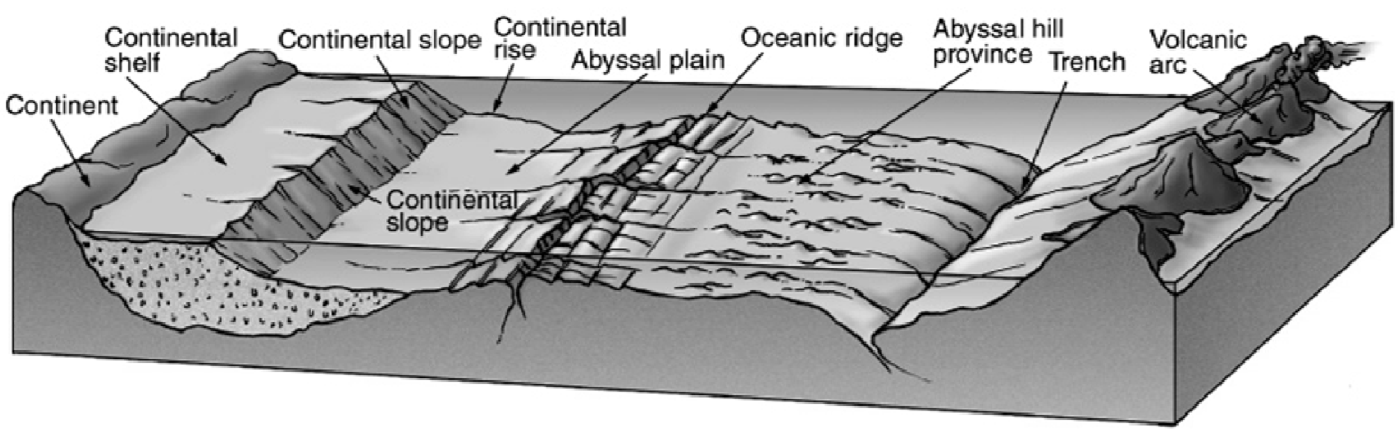
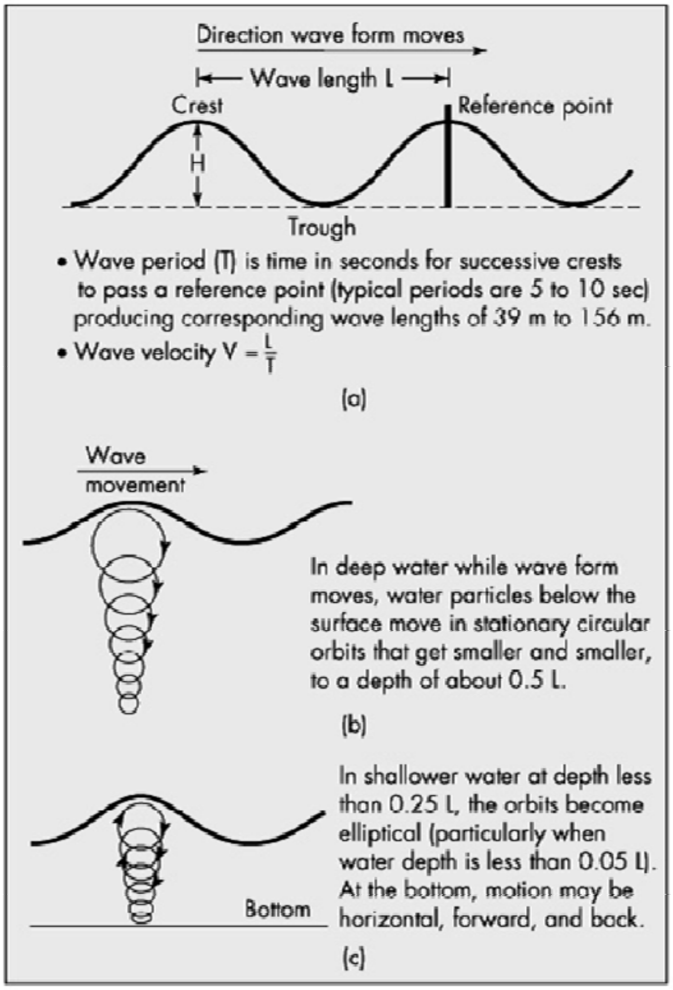
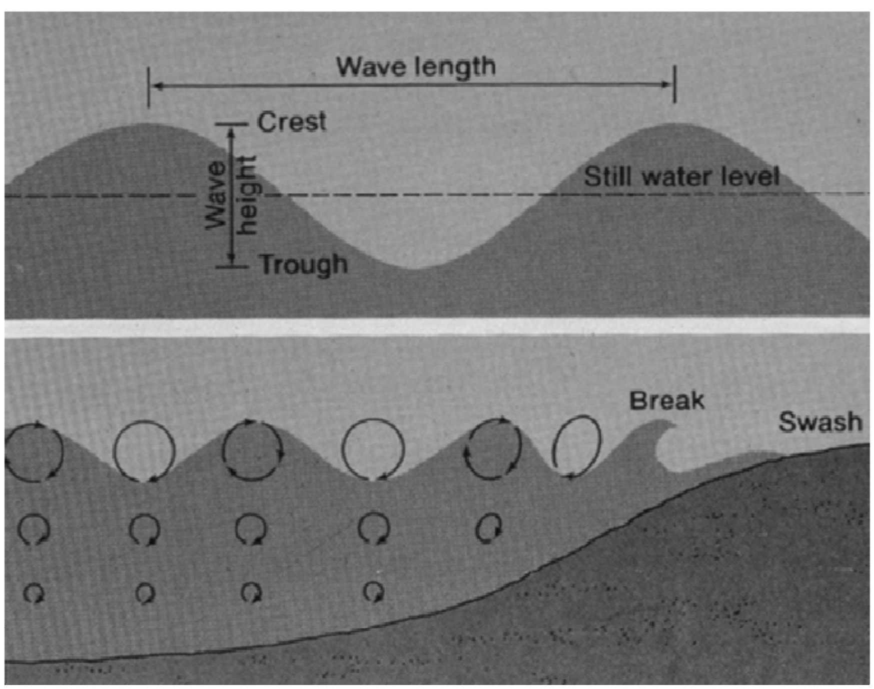
Plate Tectonic Setting





Wave Processes – Wind Driven





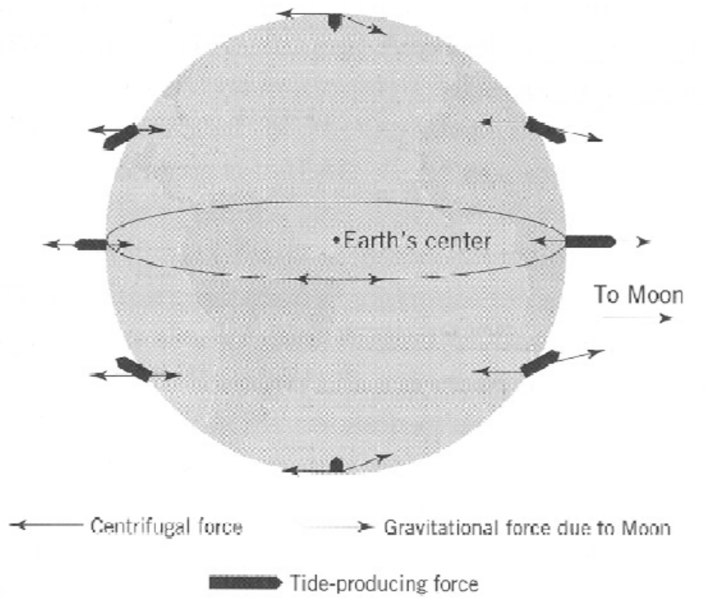
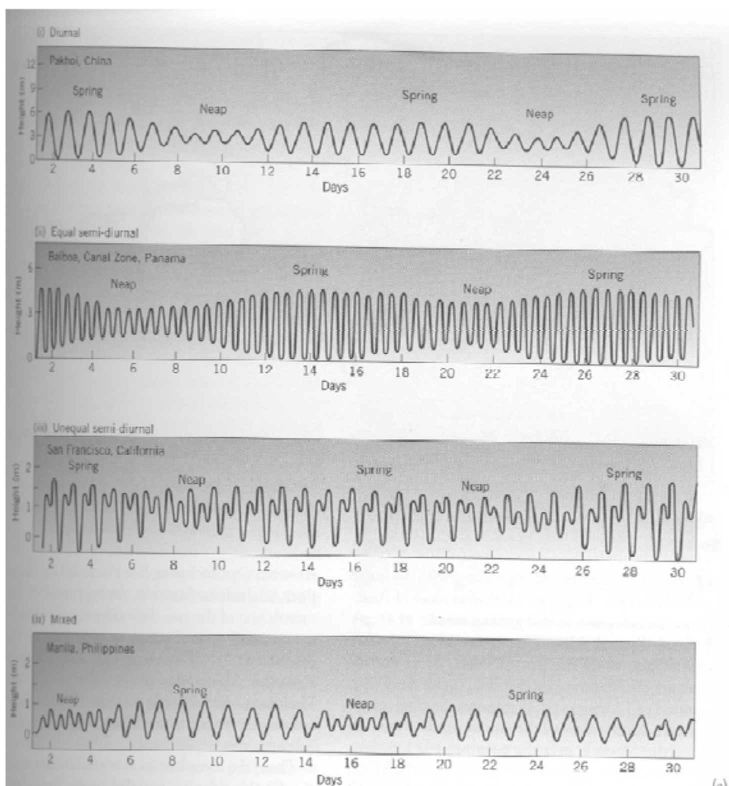
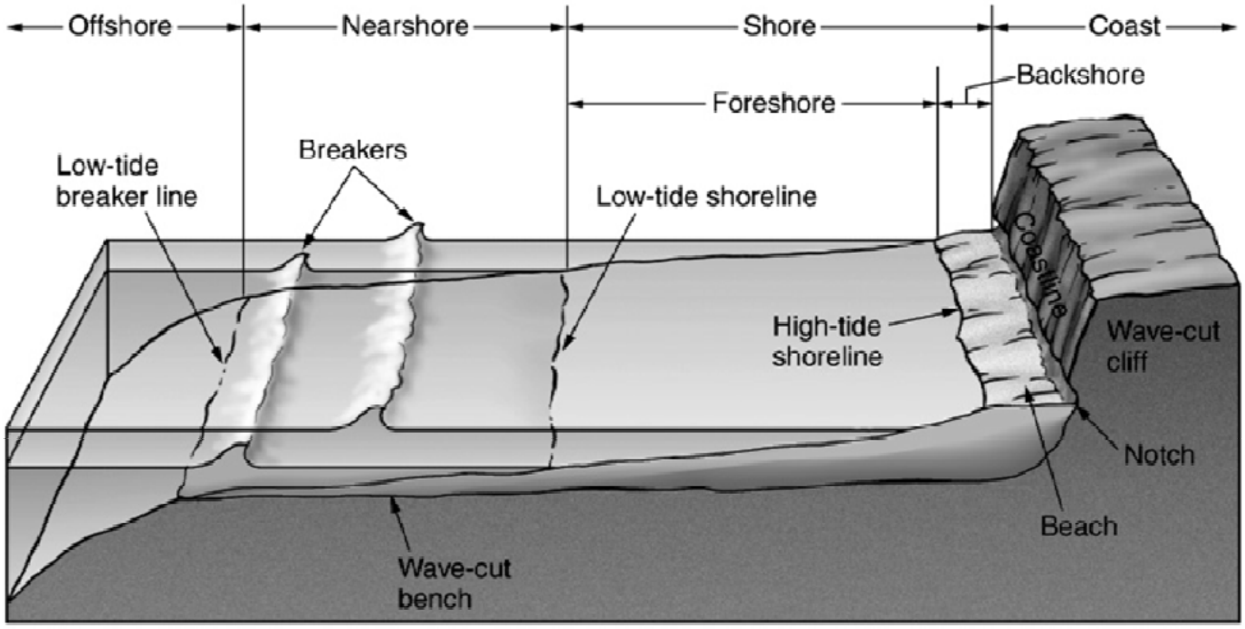
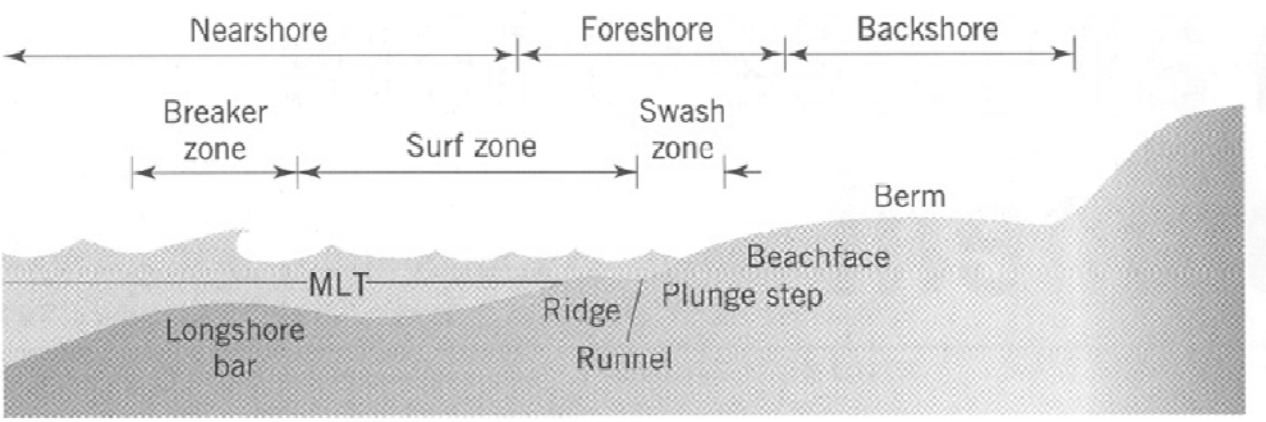


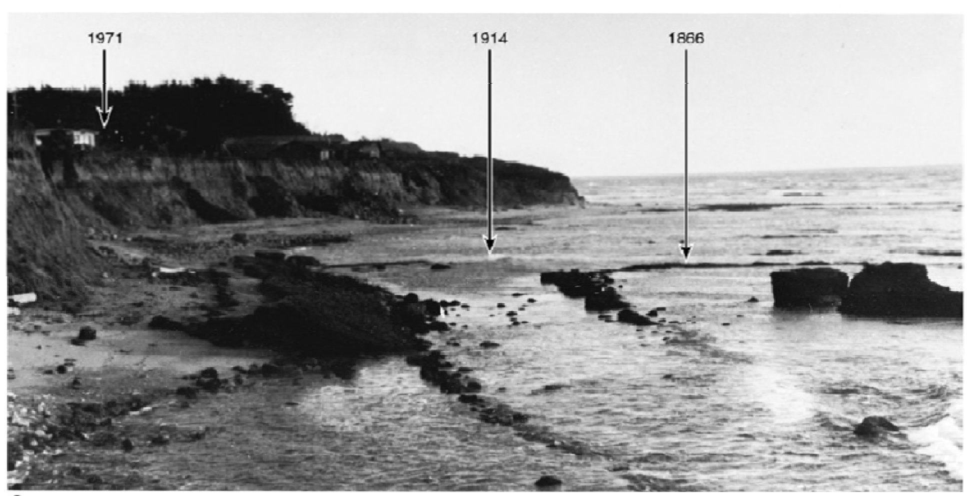
Fig. 11.4 Tides are generated in the Earth's hydrosphere due to differences between the centrifugal force and the Moon's attractive force.

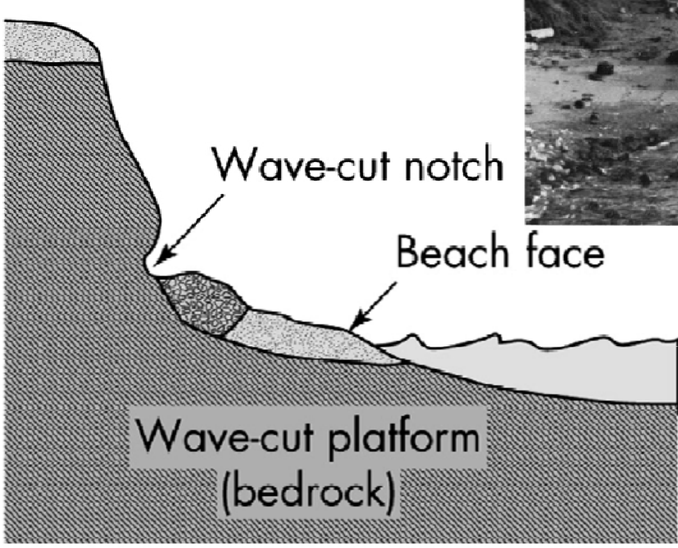


11.14 Coastlines throughout the world experience a variety of tidal signatures. (a) The major types include: (I) diurnal tides, one daily; (II) semi-diurnal tides, two tides daily; (III) semi-diurnal tides with strong inequality; (IV) mixed tides, combination of diurnal semi-diurnal tides. (From R. A. Davis, 1977, *Principles of Oceanography*, 2nd edn. Reading, MA: Addison-Wesley, p. 155, Fig. 8.9.)

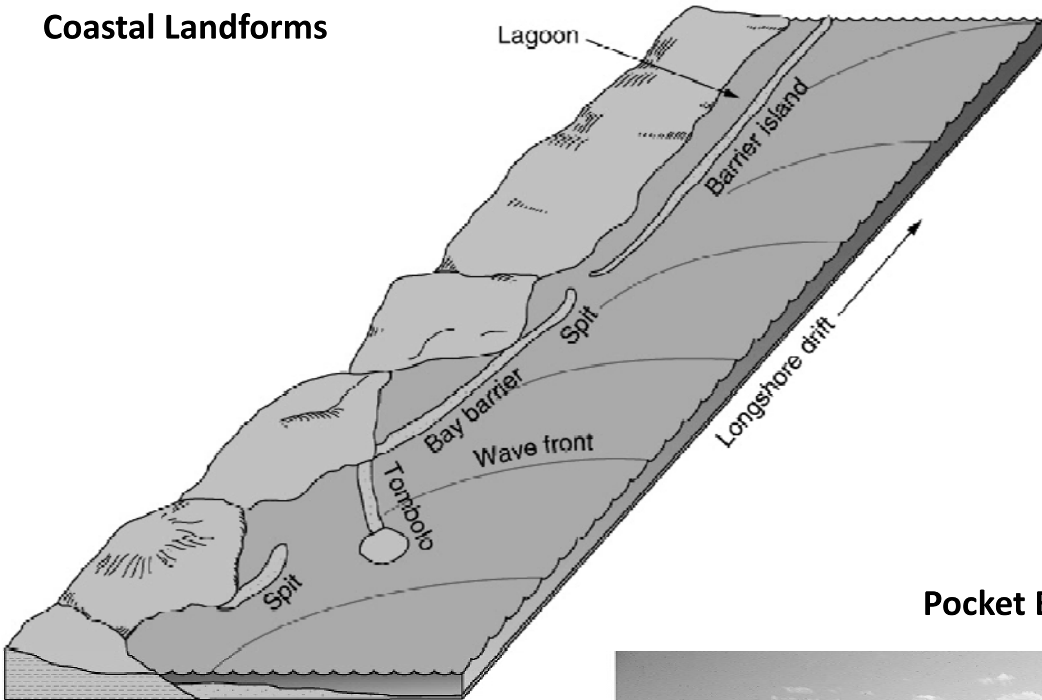

Offshore
Nearshore
Shore
Coast
Backshore
Foreshore
Breakers
Low-tide breaker line
Low-tide shoreline
Coastline
High-tide shoreline
Wave-cut cliff
Notch
Beach
Wave-cut bench


Nearshore
Foreshore
Backshore
Breaker zone
Surf zone
Swash zone
Berm
Beachface
MLT
Ridge
Plunge step
Longshore bar
Runnel

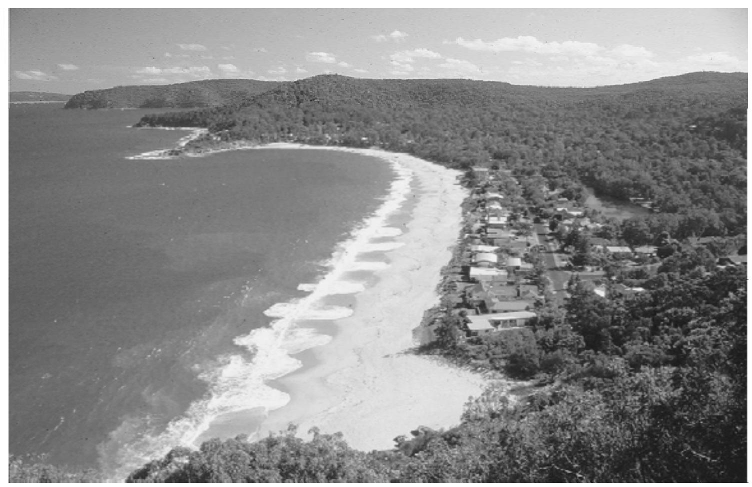

1971
1914
1866


Wave-cut notch
Beach face
Wave-cut platform
(bedrock)

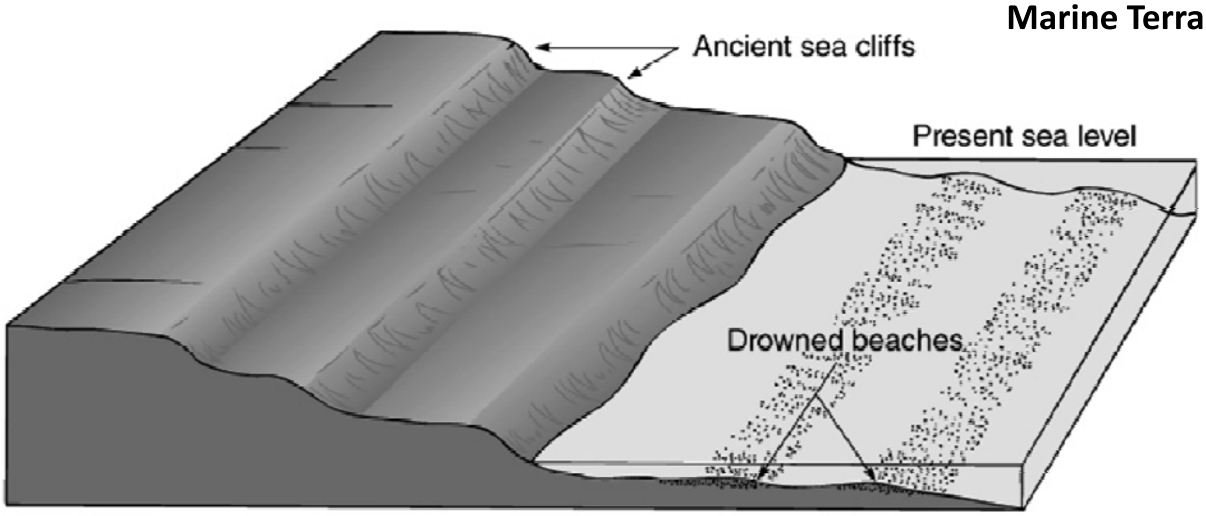
Coastal Landforms



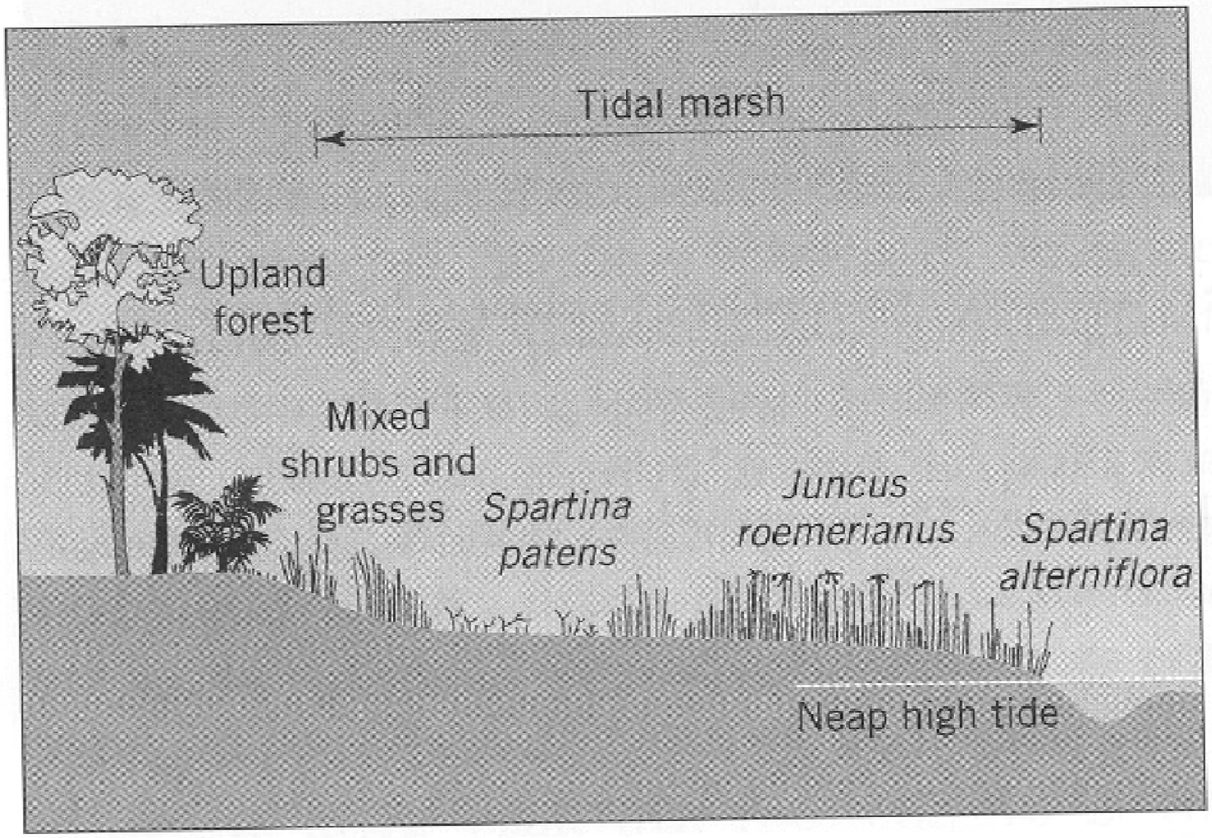
Pocket Beaches



Marine Terraces



Vegetation Patterns – Coastal Bays and Marshes



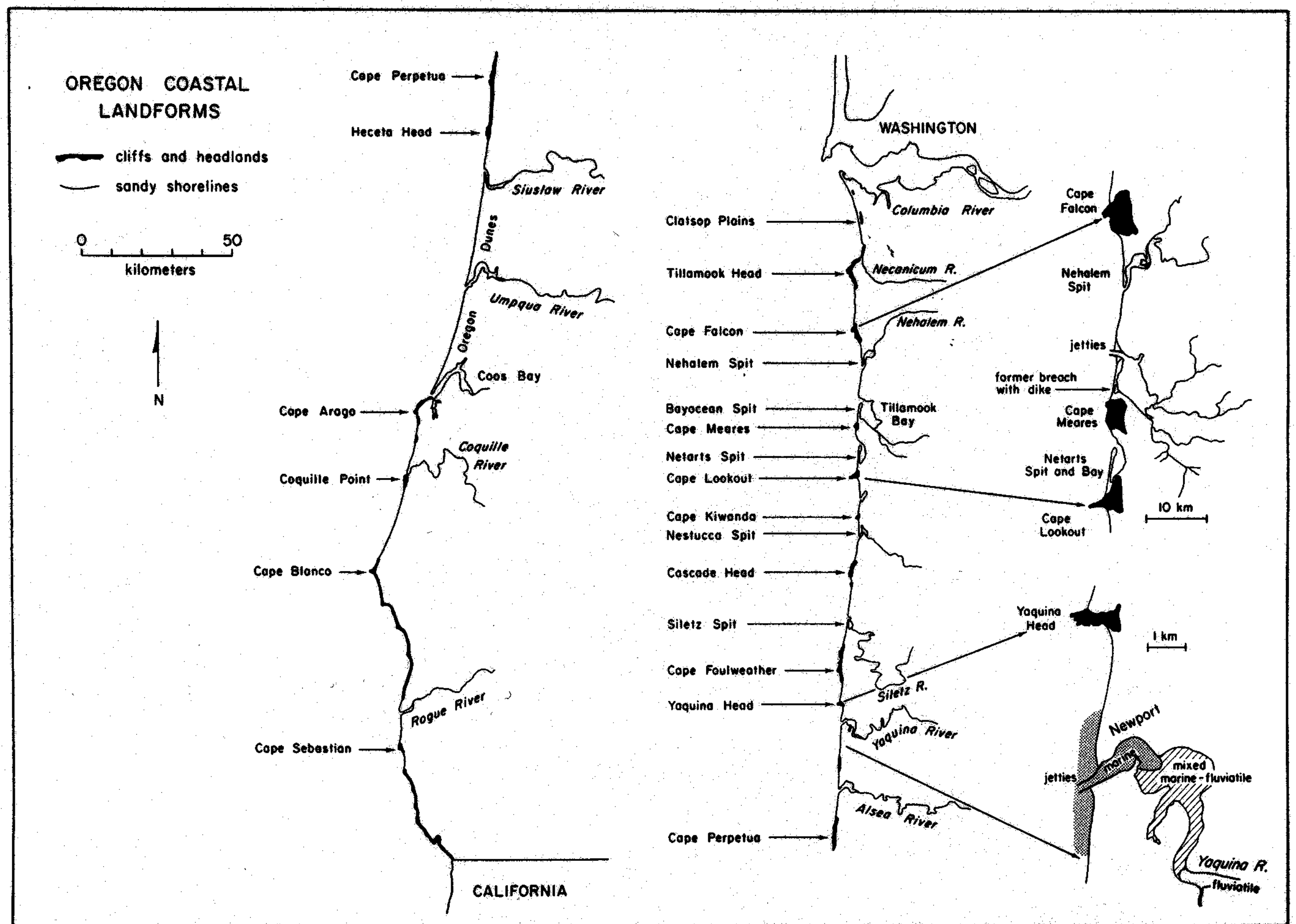
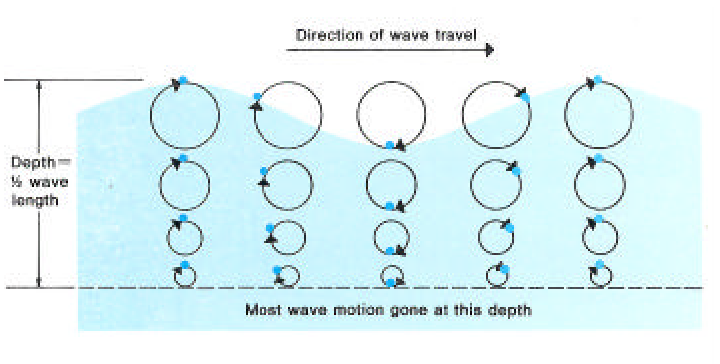
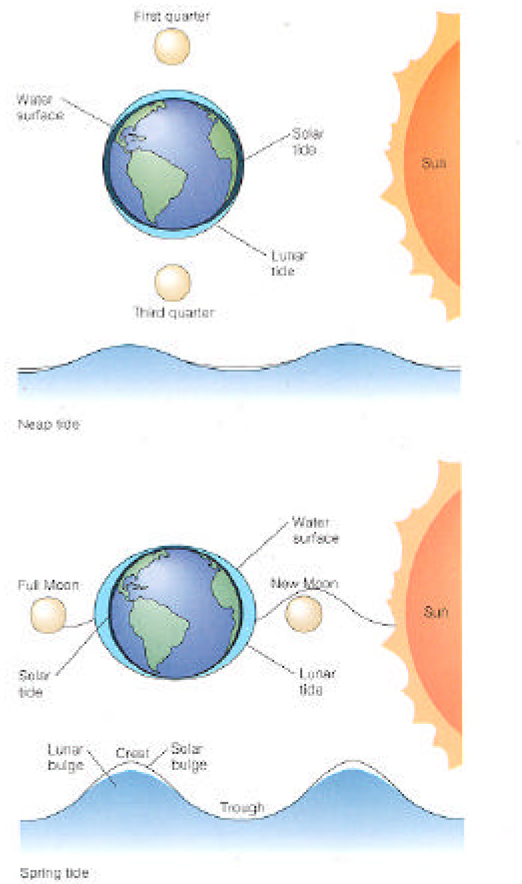
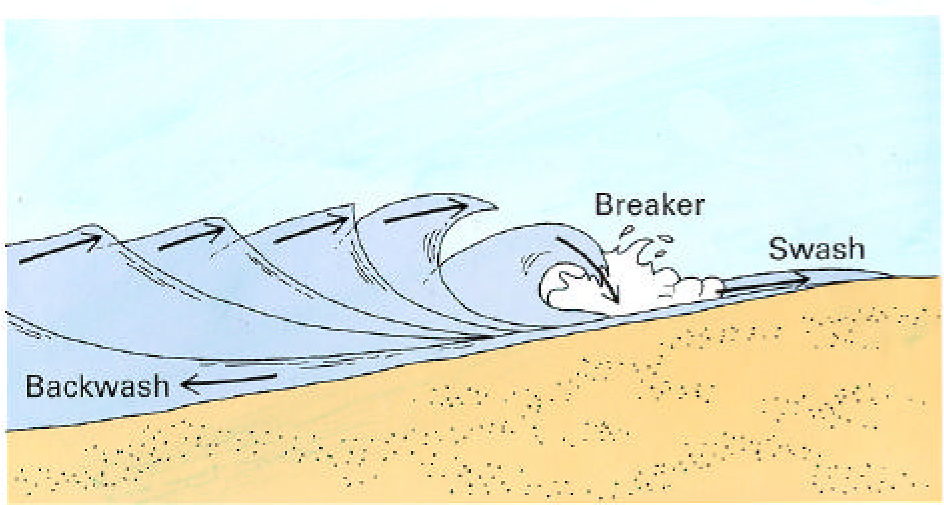


Figure 1. Coastal landforms of Oregon, consisting of stretches of rocky shorelines and headlands, separating pockets of sandy beaches. From Komar (1985).

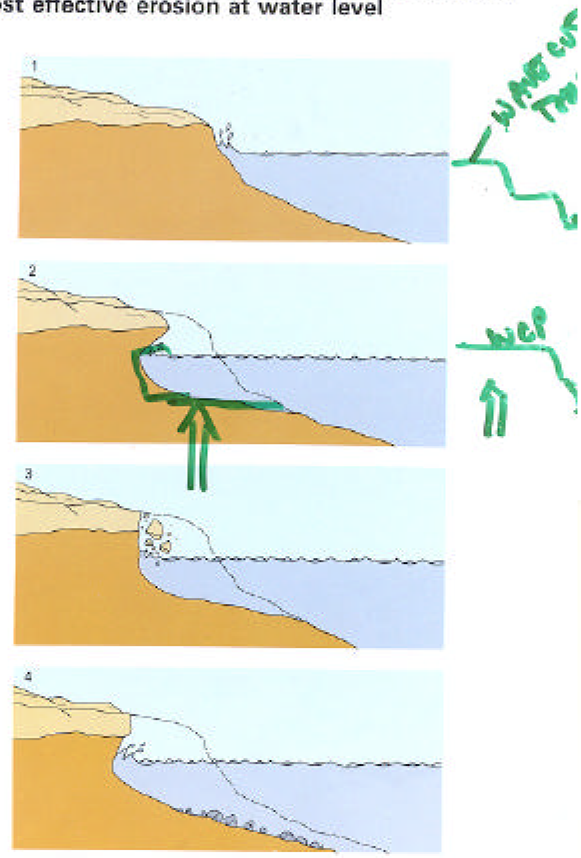


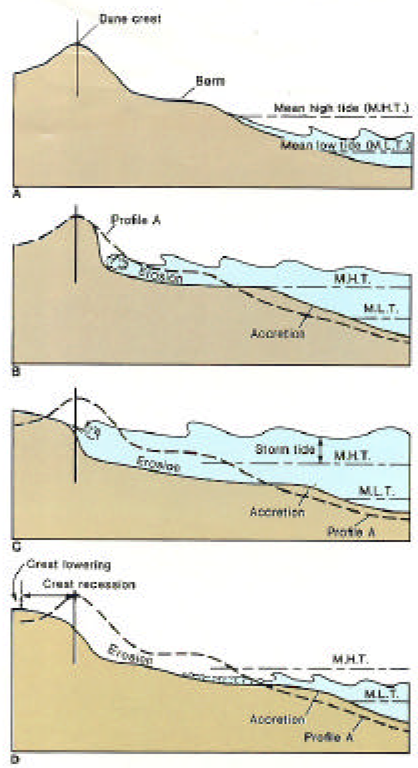


A breaking wave

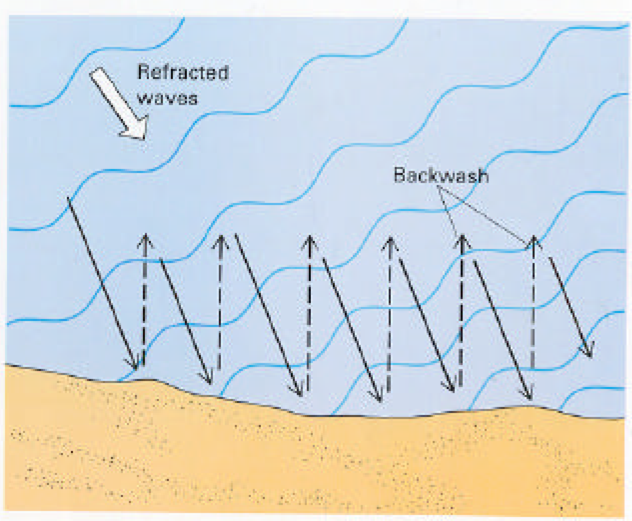


Waves pounding on exposed headlands accomplish the most effective erosion at water level T70





Beach drifting involves a zigzag movement in a general downwind direction along the coast



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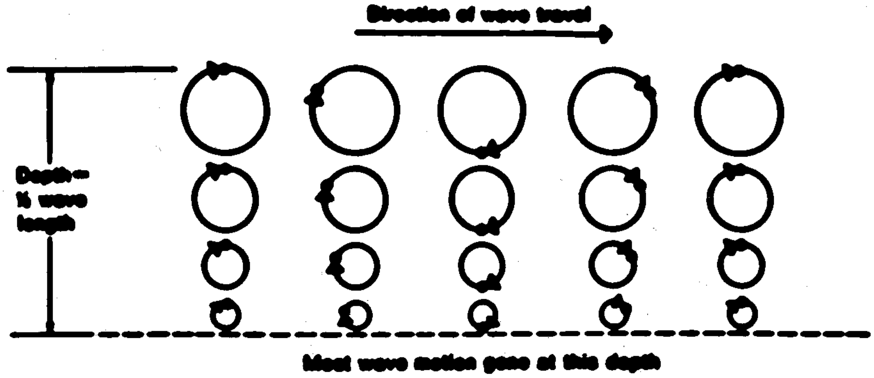


Figure 14.3
Orbital motion of water in waves dies out with depth. At the surface the diameter of the orbits is equal to the wave height.

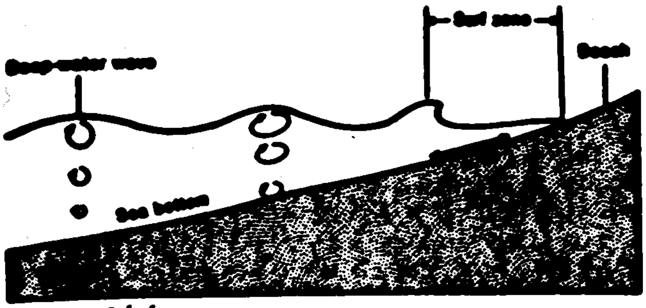


Figure 14.4
As a deep-water wave approaches shore, it begins to "feel" the sea bottom and slow down. Circular water orbits flatten and the wave peaks and breaks. In the foamy surf zone, water moves back and forth rather than in orbits.

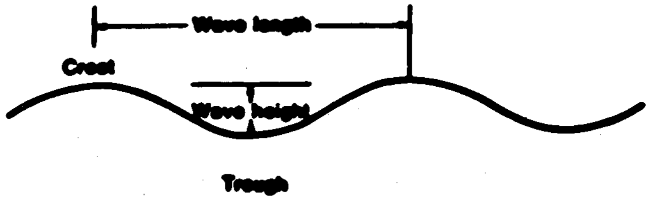


Figure 14.2
Wave height is the vertical distance between the wave crest and the wave trough. Wave length is the horizontal distance between two crests.

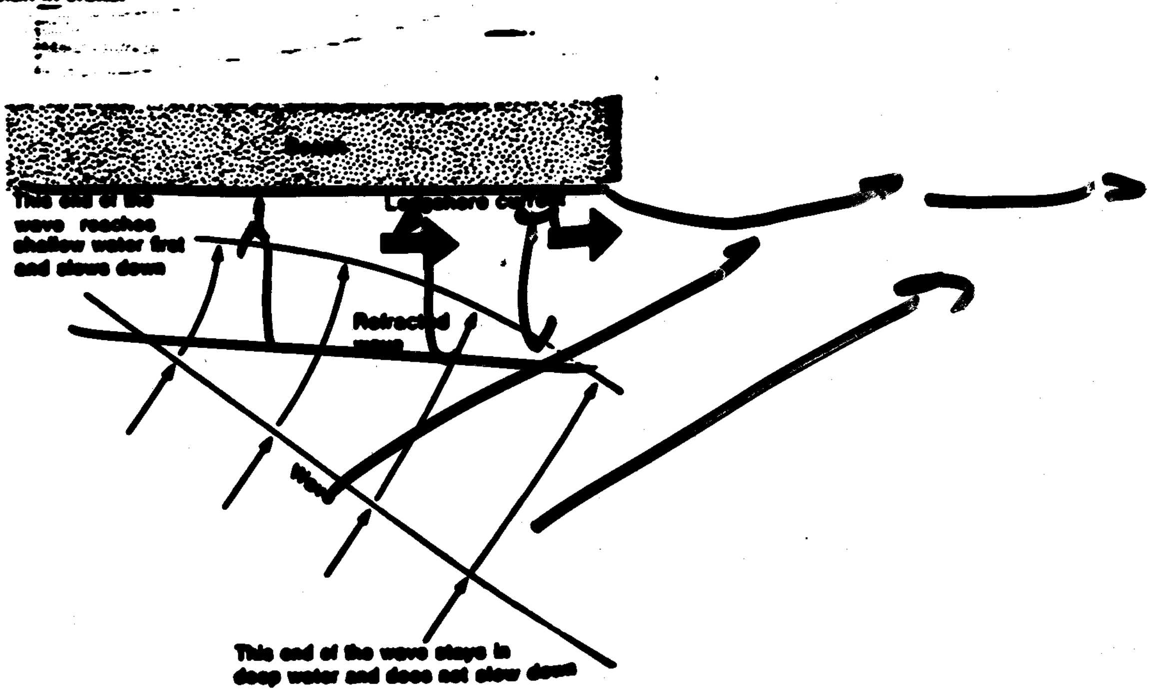
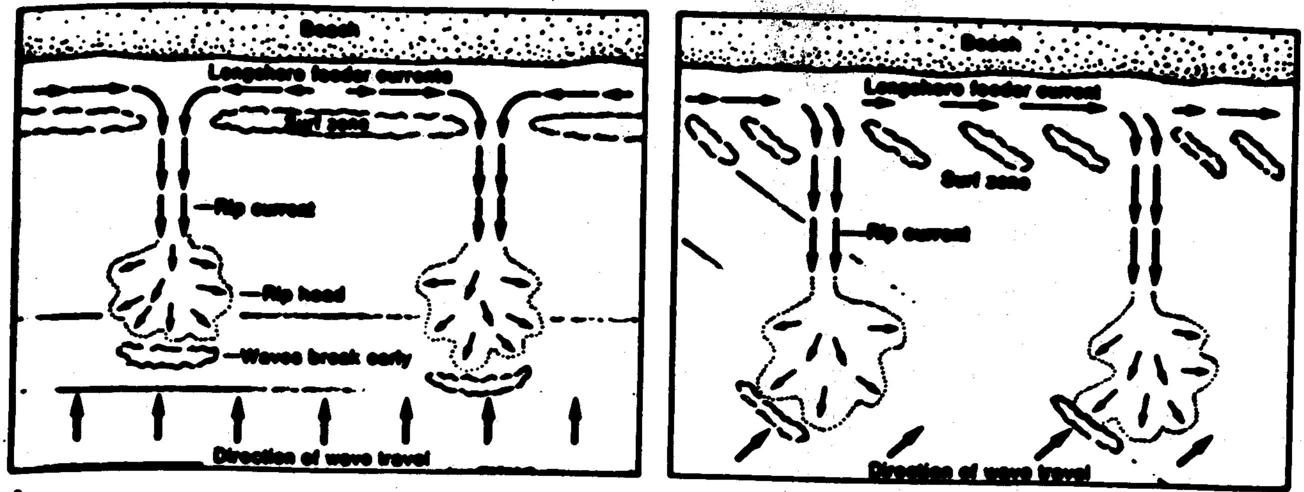
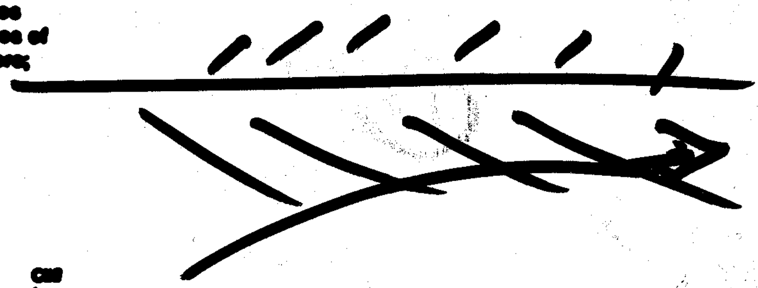


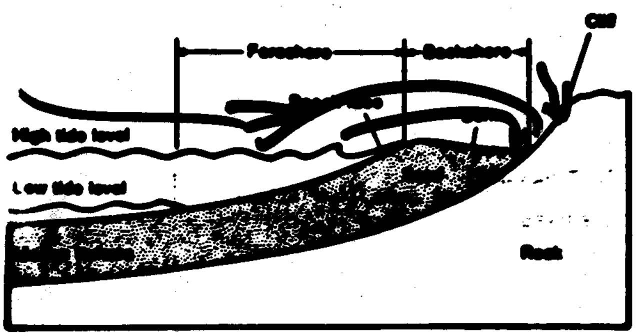
Figure 14.6
Wave refraction changes the wave direction, bending the wave so it becomes more parallel to shore. The angled approach of waves to shore sets up a longshore current parallel to the shoreline.



A B

Box 14.1 Figure 1
Rip currents and their feeder currents can develop regardless of the angle of approach of waves. (*A*) Waves approach parallel to shore; feeder currents on both sides of rip currents. (*B*) Waves approach at an angle to the shore; feeder current on only one side of rip current.





A

Figure 14.7
(*A*) Parts of a beach. (*B*) The beach face (on the left) and berm (on the right) on a northern California beach.
Photo by K. Lajoie, U.S. Geological Survey.

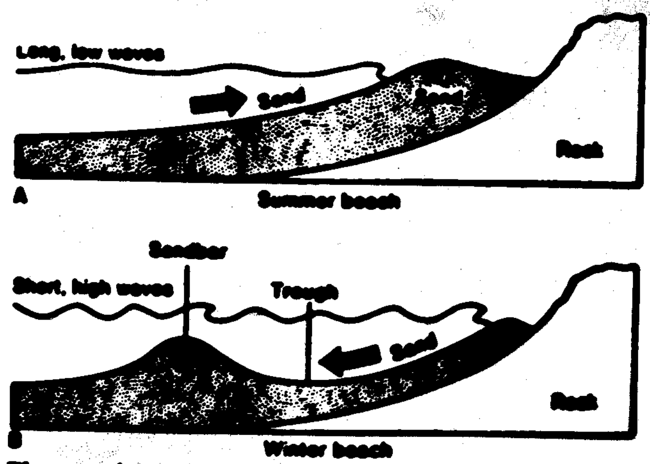
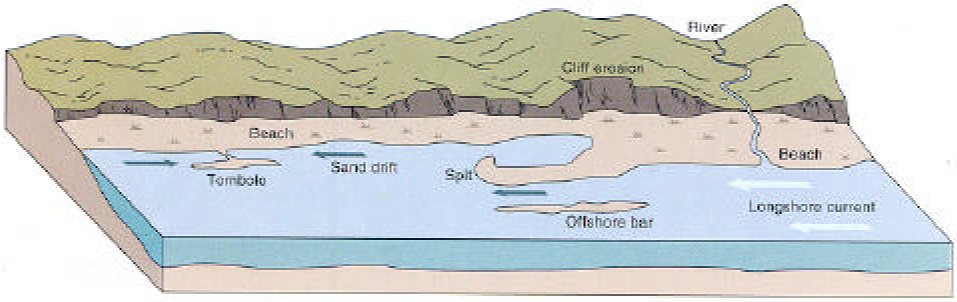
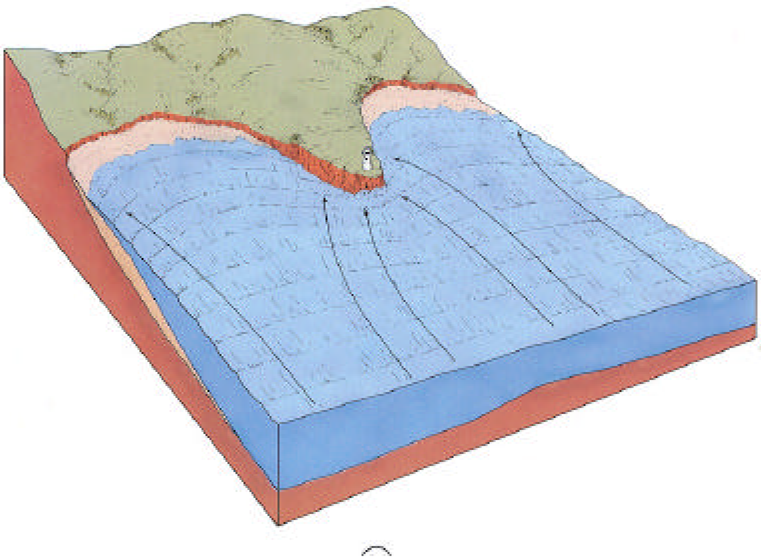
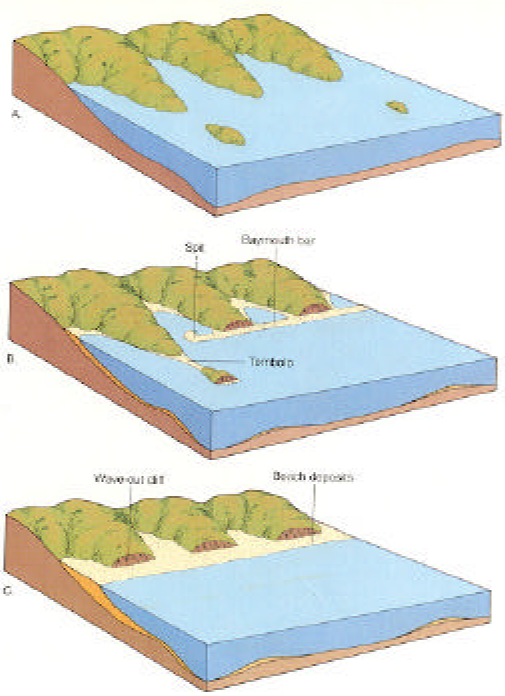
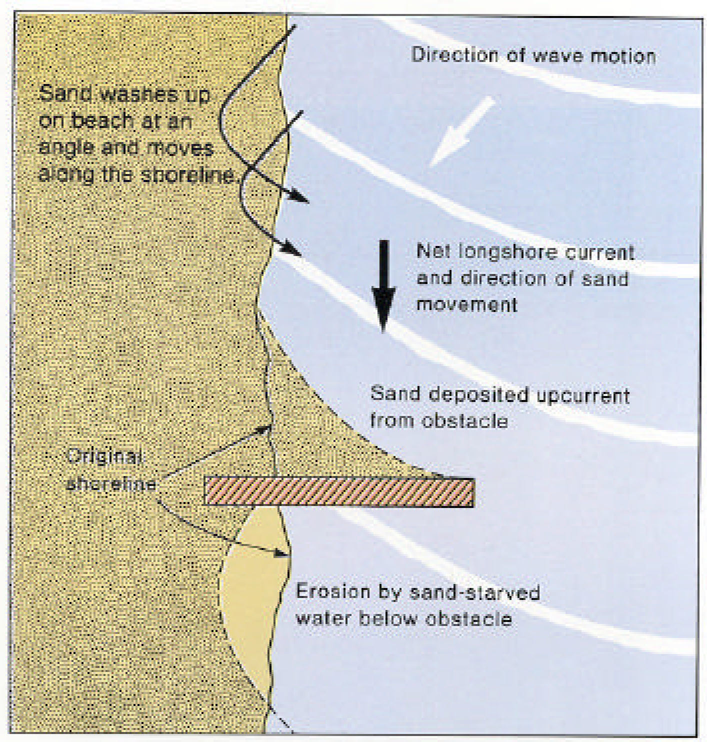


Figure 14.8
Seasonal cycle of a beach caused by differing wave types. (*A*) Summer beach. (*B*) Winter beach. Waves may break once on the winter sandbar, then re-form and break again on the beach face.


River
Cliff erosion
Beach
Sand drift
Tombolo
Spit
Beach
Offshore bar
Longshore current




A.
Spit
Baymouth bar
B.
Tombolo
Wave-cut cliff
Beach deposits
C.


Direction of wave motion
Sand washes up on beach at an angle and moves along the shoreline
Net longshore current and direction of sand movement
Sand deposited upcurrent from obstacle
Original shoreline
Erosion by sand-starved water below obstacle

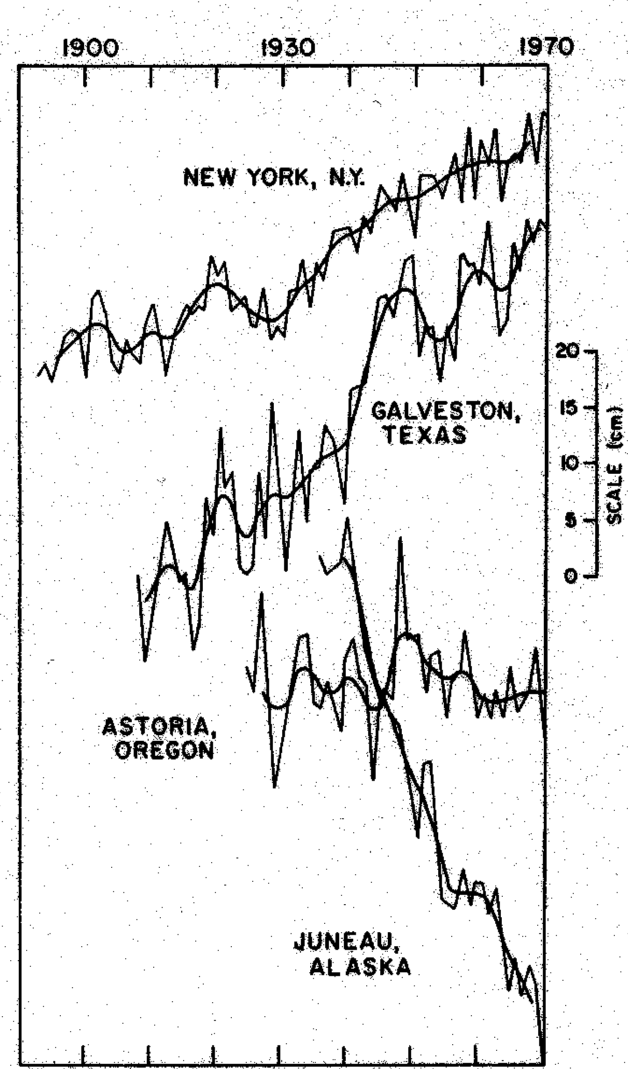


Figure 2. Yearly changes in sea levels determined from tide gauges at various coastal stations. After Hicks (1972).

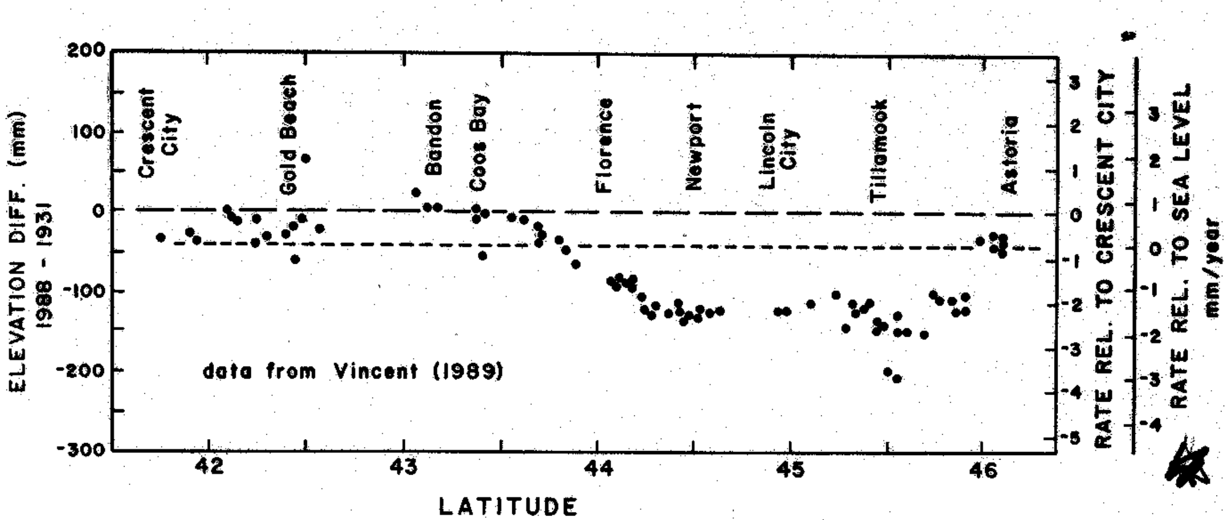
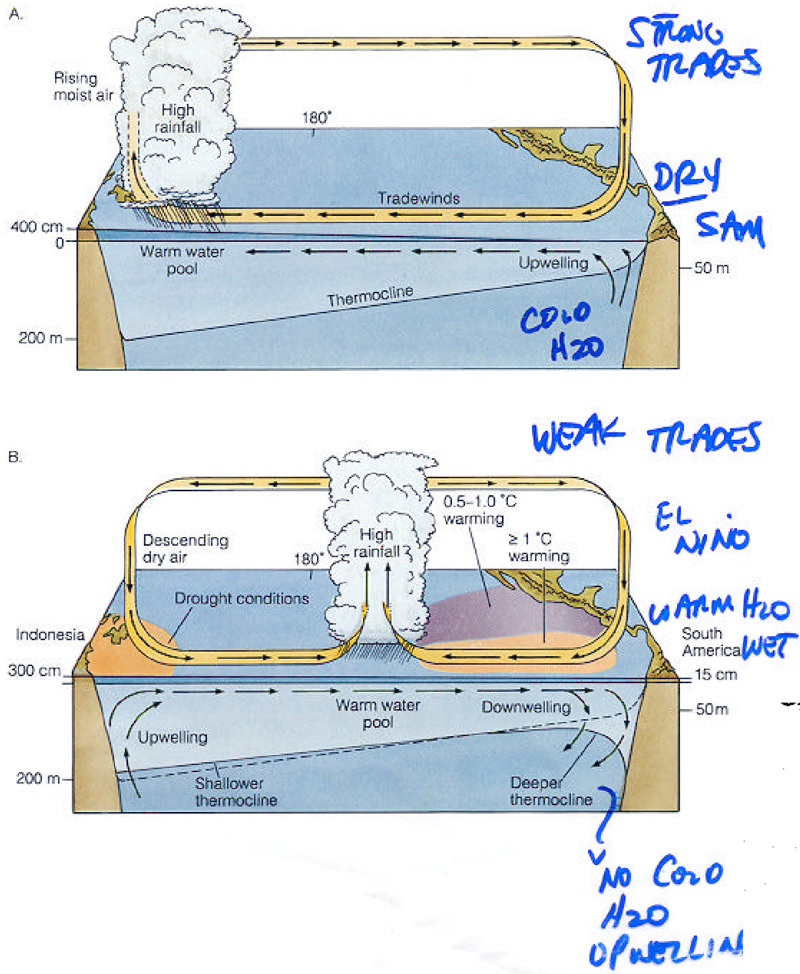


Figure 3. Elevation changes and the relationship to sea-level rise along the length of the Oregon coast from Crescent City in California north to Astoria on the Columbia River, based on repeated geodetic surveys along the coast. After Vincent (1989).



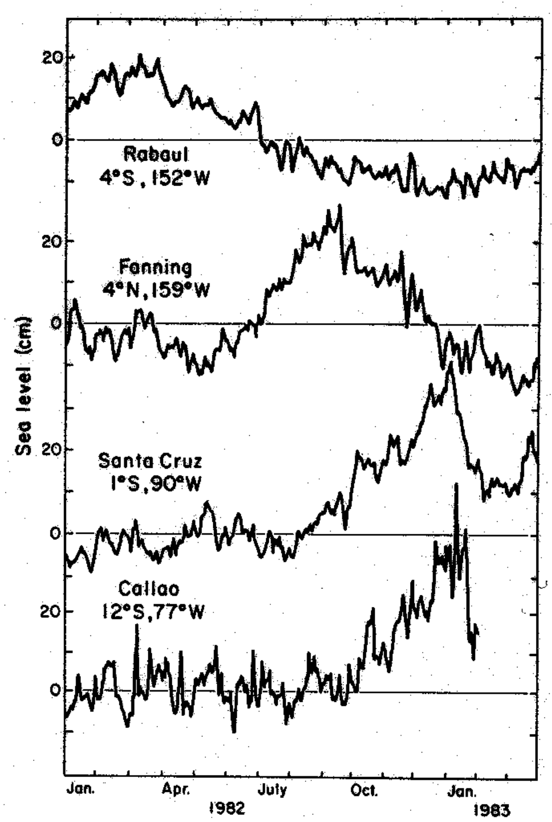


Figure 20. Sea-level "wave" during the 1982-83 El Niño measured at a sequence of islands from west to east near the equator, and finally at Callao on the coast of Peru. After Wyrtki (1984).

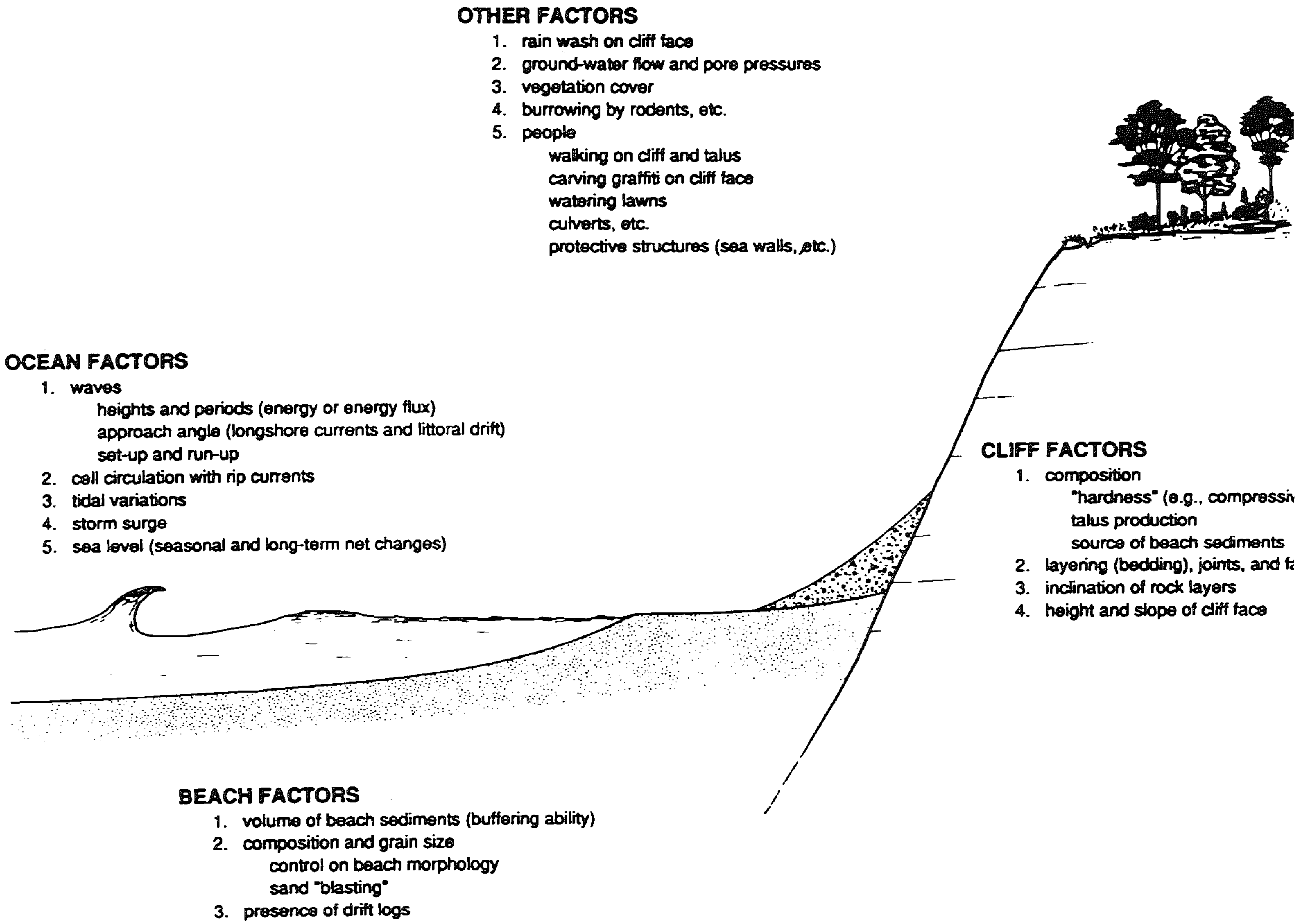


Figure 29. Schematic diagram illustrating the many factors and processes in

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ROCK UNITS AND COASTAL LANDFORMS BETWEEN NEWPORT AND LINCOLN CITY, OREGON

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The coast between Newport and Lincoln City is composed of sedimentary rock punctuated locally by volcanic rock. Where the shore is on sedimentary rock, it is characterized by a long, sandy beach in front of a sea cliff. Perched above the sea cliff is the sandy marine terrace that fringes most of this segment of the Oregon coast. Where volcanic rock interrupts the continuity of sedimentary rock, the regularity of the coastline is broken by jagged promontories, rocky shores, and sea stacks (Figure 1).

Figure 1. Yaquina Head composed of volcanic rock; Moloch Beach is on sedimentary rock. (State Highway Division photo by Kinney)

Figure 2. Landslide at Jumpoff Joe in Newport destroyed about 15 acres of coastal frontage and involved a number of houses.

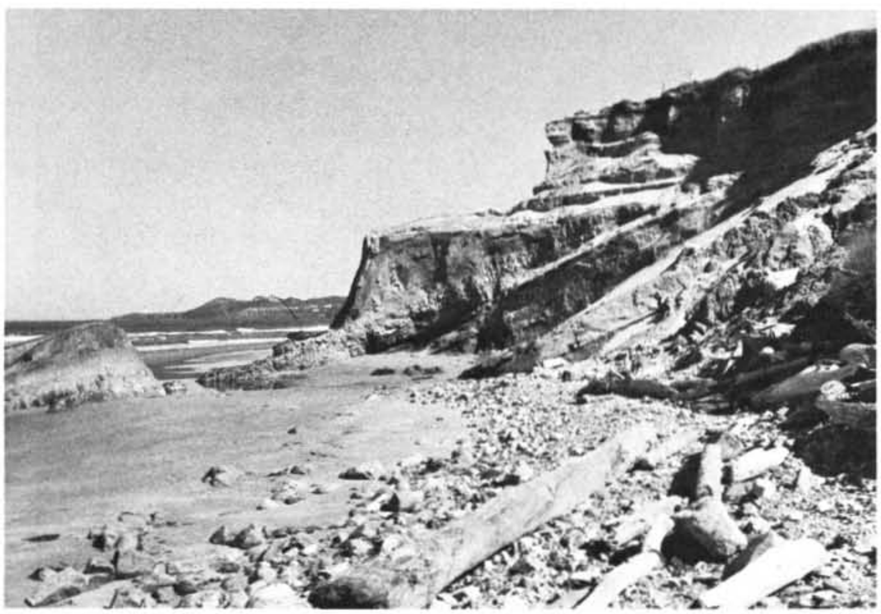


Figure 3. Seaward dipping beds of Astoria Formation are exposed at Jumpoff Joe.

A number of different rock units (geologic formations) are exposed along this part of the Oregon Coast (see accompanying geologic map). Each unit is characterized by a kind of rock or suite of rocks that distinguishes it from other units. For this reason, each unit plays a different role in shaping the coastal landforms.

In the following text, the geologic formations are discussed from oldest to youngest beginning with those of Eocene age, dating back 40 or 50 million years ago, to those of Quaternary age, some of which are forming today. The distinctive landforms associated with each unit are described and also illustrated in the photographs.

Eocene formations

Siletz River Volcanics: This formation, consisting of basalt flows and pyroclastic rocks, is not exposed along the shore but crops out east of Devils Lake and is responsible for the rugged topography in the Coast Range to the east.

Nestucca Formation: These late Eocene siltstones and interbedded sandstones are the bedrock beneath the marine terrace north of Siletz Bay. They underlie the beach sand at Lincoln City and are often exposed during the winter and early spring after storm waves have swept away the sand. The Nestucca Formation is described in more detail by Snavely and others (1969).

Oligocene formations

Siltstone of Alsea: This formation, named informally by Snavely and others (1969), consists of massive to thick-bedded tuffaceous siltstone and very fine-grained sandstone. It is the bedrock along the east side of Siletz Bay and is exposed in a roadcut along U. S. Highway 101 just south of Schooner Creek at the north end of Siletz Bay.

Yaquina Formation: The Yaquina Formation is of diverse composition, consisting mainly of sandstone but also of conglomerate and siltstone (Snavely and others, 1969). It is the bedrock beneath the marine terrace south of Siletz Bay and is exposed in a roadcut at the south end of the bay, but neither it nor the siltstone of Alsea is exposed along the shore in this segment of the coast.

Miocene formations

Nye Mudstone: The Nye Mudstone, of lower Miocene age, is described by Snavely and others (1964) as a "...mixture of clay, silt, and very fine sand-size particles in varying proportions; the lithologic designation differs from place to place, but commonly is mudstone and siltstone, and less commonly silty, very fine-grained sandstone." The Nye crops out along

Figure 4. View south from Cape Foulweather toward terrace promontories on Astoria Formation; Yaquina Head in distance. (State Highway Division photo)

Figure 5. Devils Punchbowl in Astoria Formation is the collapsed roof of two sea caves. (State Highway Div. photo by Kinney)

the sea cliff at Newport beneath the Astoria Formation and is exposed in contact with the Astoria Formation at Jumpoff Joe, $1\frac{1}{2}$ miles north of Yaquina Bay. It does not crop out along the shore north of Yaquina Head but underlies the terrace south of Gleneden Beach.

Beds of the Nye Mudstone have a high clay content and therefore have low shearing strength, especially when wet; areas underlain by this formation are subject to landslides along cliffs and in steep terrain. Along the shore between Yaquina Bay and Yaquina Head, the beds have a seaward inclination of as much as 30 degrees, which increases the threat of landslides in this formation. Much of the slumping along this part of the coast is attributed to yielding of weak beds along inclined bedding planes, and bedrock slumps are numerous. The largest slumped area is at the Jumpoff Joe locality in Newport; it involves about 15 acres along a quarter mile of the shore (Figure 2).

Astoria Formation: The middle Miocene Astoria Formation, which overlies the Nye Mudstone, is described by Snavely and others (1969) as consisting "...principally of olive-gray, fine- to medium-grained micaceous, arkosic sandstone and dark carbonaceous siltstone. The sandstone beds range from massive to thin-bedded and generally are thicker in the upper part of the sequence." The Astoria Formation contains many fossils.

Between Yaquina Bay and Yaquina Head, the Astoria Formation is exposed above the Nye Mudstone in the sea cliff (Figure 3). From Yaquina Head to Cape Foulweather it is the bedrock below the terrace sediments, and it crops out extensively in the sea cliff and on the wave-cut platform between these points. It is exposed at inner Depoe Bay, in the sea cliffs at Boiler Bay, and at the mouth of Fogarty Creek. It underlies the terrace at Lincoln Beach but is not exposed along the beach.

From Yaquina Head to Otter Rock, a continuous, gently sloping beach lies in front of the sea cliff on the Astoria Formation and overlying terrace sandstone (Figure 4). The northern part is Beverly Beach and the southern part Moloch Beach.

Two promontories (Figure 4) south of Cape Foulweather are segments of a marine terrace developed on the Astoria Formation. The terrace is capped by a layer of sandstone, which gives the flat surface to these points of land. The Astoria Formation is exposed along the outer edges of the promontories where wave erosion has removed the terrace sandstone. It forms the present-day wave-cut platform, where marine gardens with many tidal pools support a luxuriant and varied assemblage of plants and animals.

The southernmost of the two promontories, at Otter Rock community, is the site of the famed Devils Punchbowl (Figure 5). This hole in the terrace was formed by collapse of the roof where two sea caves met, one from the north and the other from the west. Water enters the bowl at high tide, and during storms it churns and foams as in a boiling pot.

One of the most remarkable features in the Astoria Formation is the inner Depoe Bay (Figure 6). This very small and secure harbor lies behind

Figure 6. Basaltic wall shelters inner Depoe Bay from storm waves and erosion. (State Highway Division photo by Kinney)

Figure 7. Government point is a wave-cut platform on volcanic rock; Boiler Bay to north is eroded into sedimentary rock. (State Highway Division photo by Kinney)

a wall of basalt lava rock, and its access to the sea is a narrow chasm through the basalt wall. The basin is a low area in the topography formed principally by stream erosion along North and South Depoe Creeks. Much of the shaping of this small basin was done during the most recent glacial stage, the Wisconsin, when sea level stood several hundred feet lower than it does now. At that time the passage through the wall was a gorge through which the ancestral Depoe Creek flowed. With melting of the glacial ice, the sea rose to a level that caused flooding at the lower end of the drainage basin of Depoe Creek. In this respect, the origin of this small bay has something in common with large water bodies such as Yaquina Bay, where rise in sea level has "drowned" their lower parts and formed estuaries.

Boiler Bay (Figure 7), about 1½ miles north of Depoe Bay, is eroded in the Astoria Formation, which is exposed beneath a thick terrace deposit along the inner cliffs of the bay. Boiler Bay was formed where wave erosion breached a layer of the same basalt as that at Depoe Bay. A fault and other fractures trending at nearly right angles to the shore helped direct the erosion.

A short distance north of Boiler Bay there is a cove in the Astoria Formation at the mouth of Fogarty Creek (Figure 8). This indentation is defended on both north and south sides by remnants of the basalt.

Depoe Bay Basalt: This middle Miocene basalt flow was described by Snavely and others (1965) and named by the same writers (1973). The flow is exposed along the shore at Depoe Bay and forms the wall that separates the inner and outer bays (Figure 6). Here the rock is a pillow basalt breccia consisting of more or less dense ellipsoidal masses (pillows) of basalt enclosed in a matrix of angular basaltic glass fragments formed by the sudden chilling of hot lava coming in contact with sea water (Figure 9). Snavely and co-workers state that the 75-foot thick breccia at Depoe Bay grades into a 50-foot subaerial (deposited on land) flow to the south. Numerous dikes and sills of this type of basalt cut the Astoria Formation just east of Depoe Bay, and it is believed that lava erupted from local fissures and flowed into the ancient sea (Snavely and MacLeod, 1971).

At Depoe Bay, the lava flow forms a jagged, steeply sloping surface along the water's edge, and in places trenches and small caves have been eroded along fractures. One of the caves has an opening at its landward end through which the force of waves causes water to erupt, at times in columns several tens of feet high. These features, called spouting horns, are common along parts of the Oregon coast, and the one at Depoe Bay is one of the most spectacular (Figure 10).

The Depoe Bay Basalt extends northward from Depoe Bay but lies inland from the shore until it appears again at Boiler Bay, where small isolated masses, through their superior resistance to erosion, impart an irregularity to the inner edge of the bay. It lies along the shore north of Boiler Bay as far as Lincoln Beach. The rock knob at the mouth of Fogarty Creek (Figure 8) and the sea cliff northward to Fishing Rock (Figure 11) are of this basalt, and an isolated remnant of the flow forms a reef opposite Lincoln

Figure 8. Fogarty Creek enters a small cove protected by basalt shoulders. (State Highway Div. photo by Kinney)

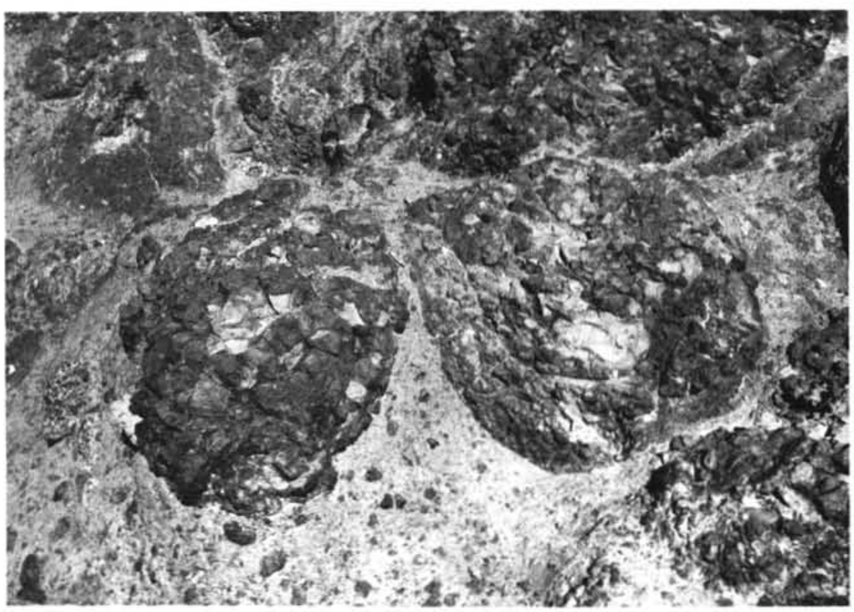


Figure 9. Pillow structure formed when hot lava of Depoe Bay Basalt poured into seawater.

Figure 10. Spouting Horn at Depoe Bay shoots up through a hole in the roof of a sea cave. (State Highway Division photo)

Beach Wayside, just north of Fishing Rock. Off Neptune Beach, a reef and two small islands, Otter Rock and Whaleback, are remnants of subaerial flows of this formation.

Sandstone of Whale Cove: An unnamed middle Miocene unit of sandstone 200 to 300 feet thick lies between the Depoe Bay Basalt and the slightly younger Cape Foulweather Basalt in the Whale Cove-Depoe Bay locality. This unit, informally referred to as the sandstone of Whale Cove, is described by Snavely and others (1969), who state, "Massive to thick-bedded medium- to fine-grained arkosic sandstone and thin-bedded micaceous carbonaceous siltstone to fine-grained sandstone constitute the bulk of the unit."

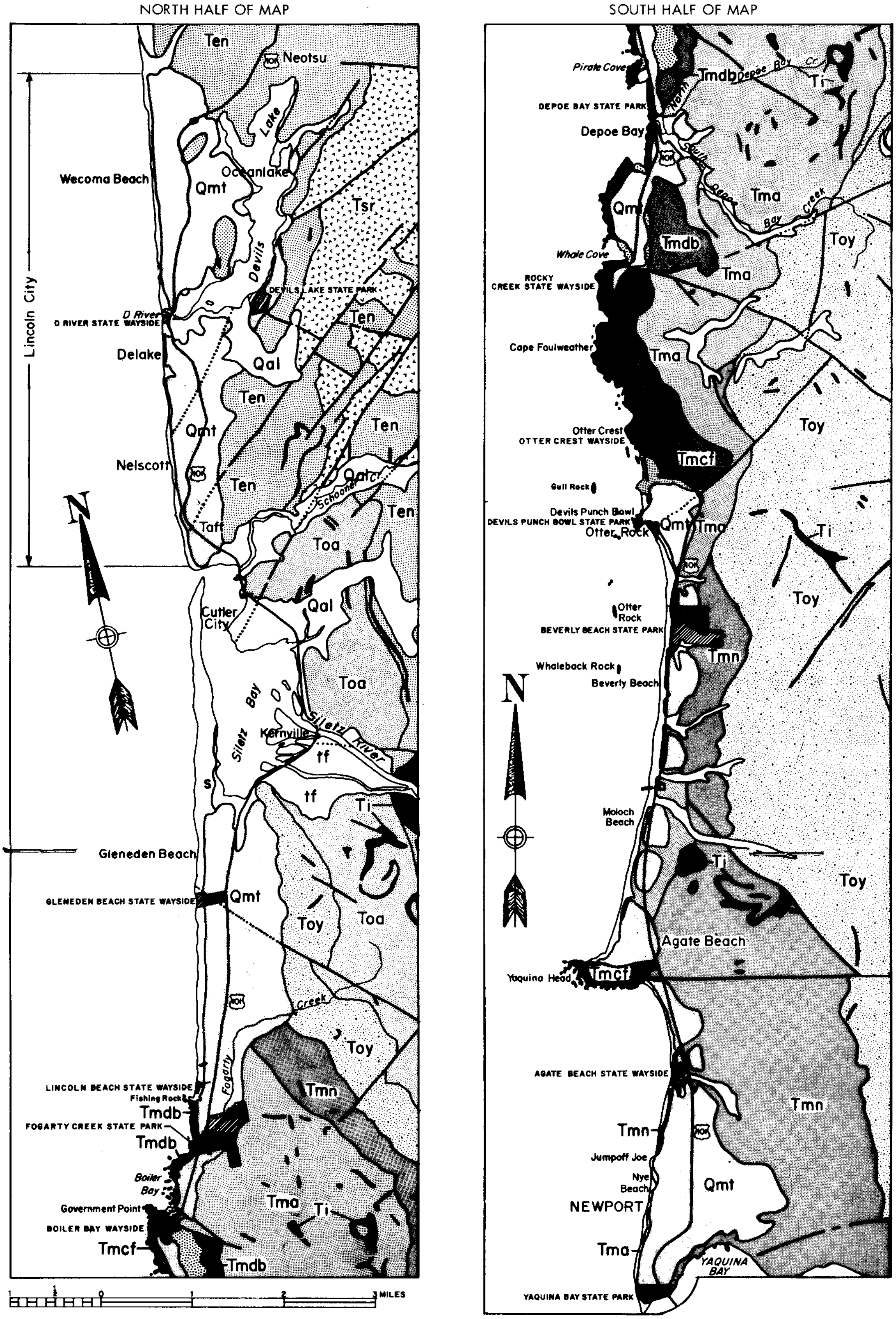
Whale Cove (Figure 12) was eroded in this sandstone, which forms the sea cliffs along the east and north edges of the cove. The erosion was probably directed initially along an east-west fault that cuts across the sandstone and the basalt that lies between it and the ocean. Once erosion breached

Figure 11. Fishing Rock at south end of Lincoln Beach is a remnant of Depoe Bay Basalt. (State Highway Div. photo by Kinney)

Figure 12. Whale Cove in the foreground and outer Depoe Bay in the distance. (State Highway Div. photo by Kinney)

Figure 13. North end of outer Depoe Bay (lower left of photo) and Pirate Cove (center left of photo) carved into sandstone and basalt. (State Highway Div. photo by Kinney)

Figure 14. Otter Crest on Cape Foulweather rises abruptly out of sea and provides spectacular views from its top. (State Highway Div. photo)



EXPLANATION

- Qal tf — Alluvium and estuary deposits
- s — Beach, sandspits, and dunes
- Qmt — Marine terrace deposits
- Tmcf — Cape Foulweather Basalt
- Tmwc — Sandstone of Whale Cove
- Tmdb — Depoe Bay Basalt
- Tma — Astoria Formation
- Tmn — Nye Mudstone
- Toy — Yaquina Formation
- Toa — Siltstone of Alsea
- Ten — Nestucca Formation
- Tsr — Siletz River Volcanic Series
- Ti — Intrusive rocks

Geologic map of coastal Lincoln County between Newport and Lincoln City. Adapted from maps by Snavely and MacLeod, published in Schlicker and others (1973).

the wall of basalt, the cove was enlarged in the softer sandstone. North-trending faults and fractures in the rock helped promote the cove's enlargement, and sea caves and small re-entrants formed along some of the fractures.

From Whale Cove, the sandstone formation extends northward beneath the terrace sandstone and is the rock in which the wide outer Depoe Bay was formed (Figures 12 and 13). It forms the sea cliffs at the north and south ends of this bay, and, as at Whale Cove, the sea cliffs are irregular and penetrated by small coves and sea caves.

Pirate Cove (Figure 13), a short distance north of Depoe Bay, is partly in the sandstone of Whale Cove, and the sea cliffs along the inner edges of the cove have caves, chasms, and other features characteristic of this sandstone. The sandstone ends just south of Boiler Bay.

Cape Foulweather Basalt: The Cape Foulweather Basalt, named by Snavely and others (1973), is the younger of the two middle Miocene basalt units along this part of the Oregon Coast. It is more extensive than Depoe Bay Basalt and is the major contributor to the shore's irregularity. The unit includes several textural varieties of volcanic basalt but consists predominantly of breccia and water-laid fragmental debris (Snavely and others, 1969). It also contains massive subaerial flows and submarine pillow lavas. Volcanic necks and feeder dikes that cut through the layers of erupted rock indicate that the eruptions were from local centers. Particles that went into the ocean were reworked into poorly sorted and crudely stratified deposits that fringe the main body of erupted rock.

The main center of eruption of this basalt was at Cape Foulweather, where layers of basalt breccia and dense lava accumulated to a great thickness. The Cape, with cliffs and precipitous seaward slopes that rise to a height of 500 feet at Otter Crest (Figure 14) is one of the most ruggedly beautiful headlands on the Oregon Coast. From the viewpoint at Otter Crest, one can see miles of coastline to the south (Figure 4). North of Otter Crest the shoreline around the Cape is very irregular, with numerous small coves and smaller promontories (Figure 15).

The Cape Foulweather Basalt extends along the shore as far as Boiler Bay, and north of the Cape the rock consists mostly of fragmental types. It is the bedrock beneath the terrace along the shore between the Cape and Government Point (Figure 7). The basalt helps to enclose and protect Whale Cove, outer Depoe Bay, and Pirate Cove.

At Rocky Creek Wayside the Cape Foulweather Basalt is an unsorted and crudely stratified breccia (Figure 16) overlying dense flow basalt. The strata are fairly steeply inclined and are the accumulations of volcanic fragments and flows on the flank of a volcanic cone. North-trending fractures in the rock at this locality have directed the wave erosion that has shaped a picturesque setting of a chasm, sea caves, trenches, and isolated wave-washed rocks (Figure 17).

At Government Point the fragmental rock is roughly size-sorted and

Figure 15. Rugged shore of Cape Foulweather north of Otter Crest.

Figure 16. Basalt breccia of Cape Foulweather Basalt at Rocky Creek Wayside was erupted from a volcanic cone in Miocene time.

Figure 17. Rocky Creek Wayside is situated on a small marine terrace composed of basalt. (State Highway Div. photo by Kinney)

Figure 18. Waves and turbulent water at Government Point, a marine terrace eroded on fragmental volcanic rock. (State Highway Div. photo)

stratified, which suggests a submarine site of deposition. Much of the material is about the size of pebbles. Volcanic fragments of this size are called lapilli, and the rock is designated a lapilli tuff; the coarser textured rock is classed as breccia. These rocks are part of the apron of water-deposited fragmental volcanic debris that formed a fringe around the seaward edge of the main center of eruption at Cape Foulweather.

Government Point is a favorite viewpoint for wave-watchers. During storms, waves strike the rocks with eruptive violence (Figure 18) or rush plunging into adjacent Boiler Bay (Figure 7), at times causing the surface of the bay to turn white with foam and seething water. Though the turbulence of the water in Boiler Bay during storms is not the basis for the name, the condition does lend appropriateness to it. The bay gets its name from the boiler of a wrecked ship which was visible for many years, until it finally rusted away.

Yaquina Head (Figure 1) is composed of the same kind of basalt as that at Cape Foulweather, and the breccias and dense flows are similar to those at the Cape. Dikes and sills cutting the extrusive layers indicate a local center of eruption. The flows at Yaquina Head overlie the Astoria Formation, and in places the lava became intermixed with the sandstone. The contact is clearly exposed in the cliffs at the ends of the beaches on both sides of Yaquina Head (Figure 19).

This point of land was once much more extensive and is slowly but steadily being worn away by the waves. Because of more rapid erosion in areas of weakest rock, owing to either hardness difference or fracturing, the outline of Yaquina Head is very irregular. As erosion has progressed, more resistant parts have become separated from the mainland and are now small stacks and rock knobs.

The steep-sided island (Figure 20) lying offshore about half a mile southwest of Otter Crest is an isolated remnant of the Cape Foulweather Basalt. This island is referred to as both Gull Rock and Otter Rock.

Dikes cropping out on the wave-cut platform between Otter Crest and Otter Rock headland are related to the Cape Foulweather Basalt and were probably part of the feeder system for the flows. Dikes and sills in the hills and on the beach north of Cape Foulweather are of middle Miocene age but have not been correlated with either the Depoe Bay Basalt or the Cape Foulweather Basalt.

A row of rock knobs at the north end of Siletz Bay near the mouth of Schooner Creek is along a dike that is also exposed in the nearby roadcut. Another dike about a quarter of a mile to the south forms a rock knob on a point of land. The Inn at Spanish Head is built on a basalt sill which has two seaward projections that form natural breakwaters. Remnants of the sill are awash in the surf zone a few hundred yards to the north. A dike complex about half a mile north of Nelscott community forms a reef that projects seaward nearly at right angles to the beach.

Figure 19. Intermixed sedimentary rock and basalt on south side of Yaquina Head.

Figure 20. Gull Rock south of Otter Crest is a remnant of the former headland.

Quaternary deposits

Terrace sediments: At times of higher sea level during the Pleistocene Ice Age, wave erosion cut platforms and benches on the bedrock along the shore the same way it does today, but at a higher elevation. After the platform was cut, sediments were deposited over it as the water level rose. When the sea level again lowered, a terrace was left behind.

The terrace sediments are mostly weakly consolidated sand, but in many places there is a gravel layer directly above the bedrock. Less commonly, gravel is interlayered with the sand. These gravel beds are a source of agates and other interesting rocks on the beaches. Agate hunting is best during the winter, when much of the sand has been removed by storm waves that leave exposed gravel. The source of the agates at Agate Beach is cavity fillings in the basalt at Yaquina Head and gravel layers in the terrace sediments. As the sea cliff is eroded, a new supply of agates is added to the beach.

Terraces are widest and longest where the bedrock is sedimentary, but they are well formed on the fragmental basalts north of Cape Foulweather and on the outer end of Yaquina Head. Pleistocene sediments along the shore range in thickness from a few feet to more than a hundred feet. Sea cliffs at Lincoln City reach heights of more than 100 feet and are mostly of terrace sandstone with a capping of dune sand.

River and estuary deposits: There are two estuaries in this stretch of the coast, Yaquina Bay (Figure 21) and Siletz Bay. Estuaries are formed where the sea has encroached on the lower ends of rivers because of a rise in sea level. In some of the larger rivers, they are affected by tides and the inflow of salt water for many miles upstream.

The flat plains along the estuaries are of sediments, mostly silt, sand, and clay, that were carried by the rivers and deposited in the quieter water of the bays. Most of the deposition took place at times when the sea level was higher and the bays were enlarged, but alluvium continues to be added to the plains whenever the rivers spill over their banks during floods.

Beaches and sandspits: The sand and gravel on the beach form a thin veneer over a wave-eroded surface on bedrock (the wave-cut platform). The beach is sometimes referred to as a river of sand because the sand moves along the shore with the longshore currents. Along the Oregon Coast, winter winds from the south and southwest give the currents a predominantly northward motion; summer winds from northerly directions drive the currents southward. Sand accumulates on the beach during summer, but most of it is swept away during the winter storms when the wave energy is high. In some places, the bedrock is swept clear, and in others the beach deposit becomes mostly gravel.

Sandspits are extensions of the beach that project into and terminate in open water (Figure 22). They are located at the mouths of bays, some projecting from the north and others from the south. Some bays may have

Figure 21. Entrance to Yaquina Bay, an estuary that formed by drowning of the river mouth with rising sea level. (State Highway Div. photo by Kinney)

Figure 22. Sandspits at entrance to Siletz Bay. Sandspit at right is more than 2 miles long. (State Highway Div. photo by Kinney)

both a north and a south projecting spit, as at Siletz Bay, but one will usually be longer than the other. A sandspit is in delicate balance with the waves, currents, tides, and sand supply; fluctuations in any one or a combination of these factors have an immediate and sometimes devastating effect, such as breaching during storms. A given sandspit may survive as a landform, but its position, size, outline, and contours will change.

Figure 23. Low dunes on Siletz Spit stabilized by grass. Basalt riprap provides temporary protection for houses until next assault by high tides and storm waves. (State Highway Div. photo by Kinney)

Sand dunes: Because most of the shore between Newport and Lincoln City has a sea cliff along it, preventing sand from blowing inland from the beach, there are few recent dunes. Old Pleistocene-age dunes composed of brownish-yellow sand are situated on the marine terrace at a number of places between Newport and northern Lincoln City. The old dunes are stabilized by soil and vegetation, but where exposed they are subject to erosion. They are visible in the higher sea cliffs, in highway cuts, and in excavations.

Dunes along the Siletz sandspit are the most recently active, but over most of the sandspit they have been stabilized with European Beach grass (Figure 23). On the seaward side at the end of the spit, sand is still being moved by the wind. Erosion by waves during storms and high tides of winter are constantly wearing away the outer margin of this spit, endangering houses built too near the edge and requiring the placement of rock (riprap) to retard the erosive undermining of the structures.

Devils Lake, the only large lake on this part of the Oregon coast, owes its origin to drowning of the lower part of a stream valley and blocking of the stream's mouth by dunes and beach deposits. Since the volume of water flowing in D River is not sufficient to keep the bottom of the river channel below sea level, a fresh-water lake results rather than a salt-water estuary.

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Ocean processes and hazards along the Oregon coast

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INTRODUCTION

Early explorers of the Oregon coast (Figure 1) were impressed by the tremendous variety of its scenery. Today, visitors can still appreciate those same qualities. 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 mi. 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 whose lengths are governed by the spacings between the headlands. 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 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, a significance that continues to the present.

In contrast, today the most important "commodity" for the Northwest coast economy is the vacation visitor. Vacationers arrive in thousands during the summer months, but in spite of their numbers it is still possible to leave coastal 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,

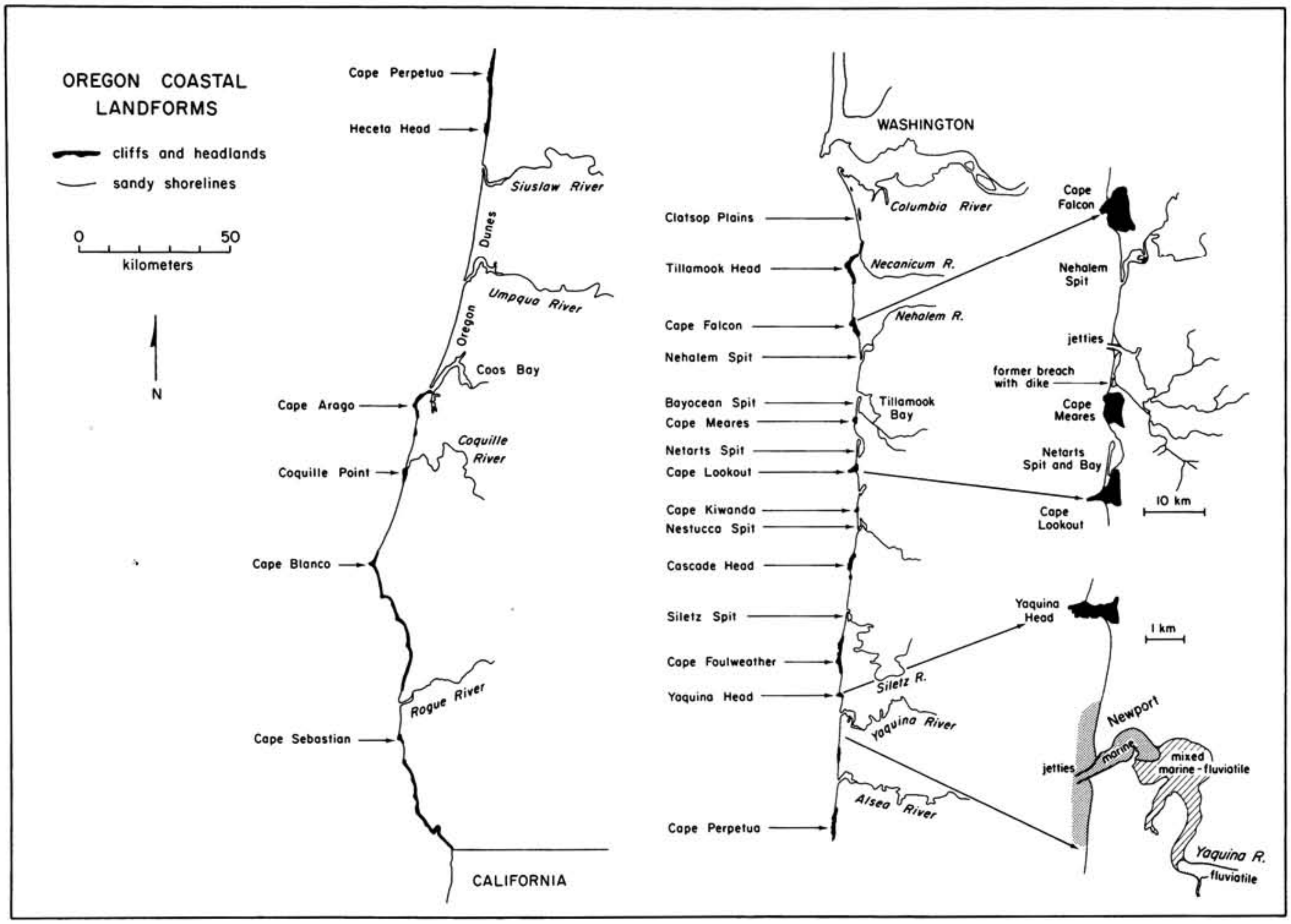


Figure 1. Coastal landforms of Oregon, consisting of stretches of rocky shorelines and headlands, separating pockets of sandy beaches. From Komar (1985).

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 an evening sunset, as well as easy access to a beach. These desires are not always compatible with nature, and as a result there are increasing problems with homes that are being threatened and sometimes lost to beach erosion and cliff landsliding.

Such problems can usually be avoided if one recognizes that the coastal zone is fundamentally different from inland areas because of its instability. This requires some knowledge of ocean waves and currents and how they shape beaches and attack coastal properties, and it requires an understanding and recognition of the land's potential instabilities that might cause disasters such as sudden landslides. A familiarity with the processes and types of problems experienced in the past can aid in the selection of a safe location for one's home. It can also enhance enjoyment of the coast and, it 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. Especially significant is the presence of active sea-floor spreading beneath the ocean to the immediate west. New ocean crust is formed at the Juan de Fuca and Gorda Ridges, and the movement of the resulting plates is generally eastward toward the continent. These ocean plates collide with the North American plate (which includes the continental land mass). That collision zone lies along the margin of the coasts of Washington, Oregon, and northern California. There is also evidence that the oceanic plates have been undergoing subduction beneath the continental North American plate, evidence that 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 are accreted 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 epochs, some 40-60 million years ago. These accreted marine sediments, mainly gray mudstones and siltstones, can be seen in many sea cliffs along the coast (see cover photo, lower left). As will be discussed in a later section, the presence of these mudstones is important to the erosion of sea cliffs and particularly the occurrence of landslides.

In addition to Tertiary mudstones, many sea cliffs contain an upper layer of clean sand (cover photo, lower left). These Pleistocene marine terrace deposits 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 the photo is the lowermost and youngest of a series of terraces that in some places form a stairway up the flank of the Coast Range. Their presence documents that the Oregon coast has been tectonically rising for hundreds of thousands of years, while at the same time the level of the sea has oscillated due to 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 and others, 1983). Examples obtained by yearly averaging and covering up to 80 years are shown in Figure 2.

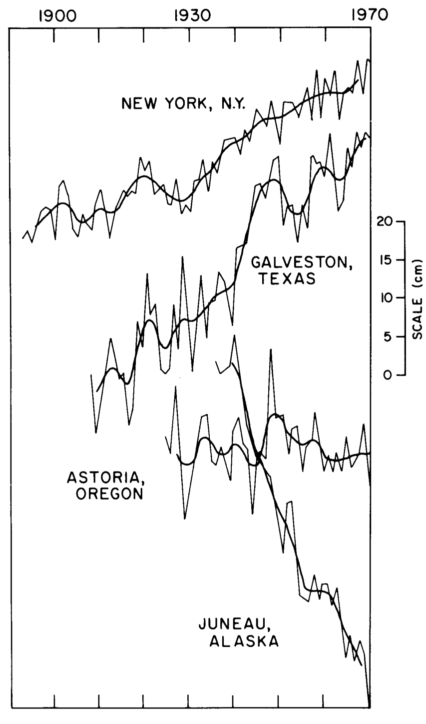


Figure 2. Yearly changes in sea levels determined from tide gauges at various coastal stations. After Hicks (1972).

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 that produce these irregular fluctuations. In spite of such irregularities, most tide-gauge records reveal a long-term rise in the sea that can in part be attributed to the melting of glaciers. The record from New York City is typical of such observations: in that example, the long-term average rise is 3.0 mm/year, about 12 in. per century (1 in. = 25 mm). The record from Galveston, Texas, also shows a rise, but the average rate is much higher at 6.0 mm/year (24 in. per 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 land that affect the record obtained at a specific tide-gauge site. It is known that the Galveston area is subsiding, so the 6.0-mm/year record from that tide gauge represents the combined effects of the local land subsidence plus the actual rise in sea level. An extreme case of this is Juneau, Alaska, which is tectonically rising at a rate that is faster than the rise in sea level. The Juneau tide-gauge record, therefore, indicates a net fall in the water level relative to the land. According to the record from the tide gauge at Astoria, Oregon, as included in Figure 2, 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 tide gauge indicates that the land is actually rising slightly faster than the water, the net increase in the land elevation relative to the sea being 0.1 to 0.2 mm/year. This change is small, amounting to 10 to 20 mm (<1 in.) of land elevation increase if continued for 100 years. Greater rates of uplift of the land must be occurring at Neah Bay on the north coast of Washington, the net rate there being 1.3 mm/year (5 in. per century) in excess of the global sea-level rise, and at Crescent City in northern California with 0.7 mm/year or 2.8 in. per century of net land emergence (Hicks and others, 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 and others (1991) have analyzed the geodetic data along a north-south line extending the full length of the Oregon coast. Surveys made in 1931 and 1988 were compared to establish elevation changes; the values are graphed in Figure 3. 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. Positive values for other locations represent an increase in elevation relative to Crescent City, and negative values indicate reduced elevation relative to Crescent City (but could still involve tectonic uplift).

The overall pattern seen in Figure 3 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 north-

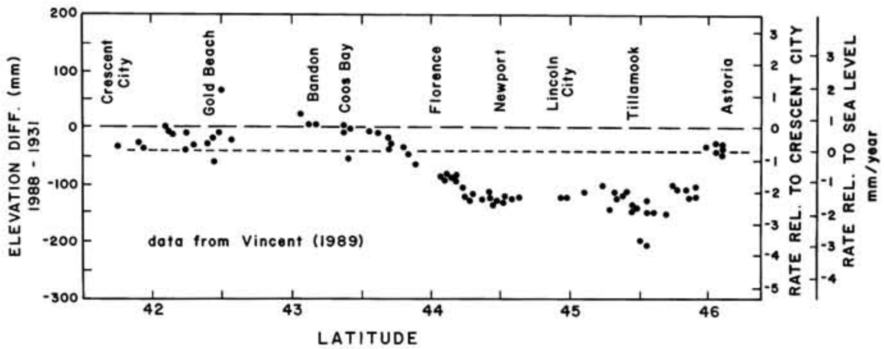


Figure 3. Elevation changes and the relationship to sea-level rise along the length of the Oregon coast from Crescent City in California north to Astoria on the Columbia River, based on repeated geodetic surveys along the coast. After Vincent (1989).

ernmost portion of the coast toward Astoria and the Columbia River. The first scale on the right of Figure 3 indicates the equivalent rates, the elevation changes divided by the lapsed time between surveys, 1988-1931 = 57 years. The differential rates are significant, for example amounting to 2-3 mm/year when comparing 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 as they were determined by Vincent (1989) into rates compared with the global change in sea level. This is done simply by shifting the first scale on the right of Figure 3, the one that is relative to the Crescent City bench mark, by an amount of 0.7 mm/year as determined from the tide gauge at that location. This shift yields the rate scale furthest to the right in Figure 3, 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 scale by 0.7 mm/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 0.1-0.2 mm/year, just as found by the tide gauge at that location—confirming the validity of Vincent's analysis of geodetic data to determine elevation changes and of the analyses undertaken to convert those data into a rate of change compared with the increasing level of the sea.

According to the results graphed in Figure 3, the southern half of the Oregon coast and the far north coast near Astoria are presently rising faster than the global sea level, while the central stretch between Newport and Tillamook is being submerged by the rising sea. The submergence rates are on the order of 1-2 mm/year (4-8 in. per century) and therefore are small, compared with submergence rates experienced on most coastlines: Rates of 4-6 mm/year (16-24 in. per century) are common along the east and Gulf coasts of the United States (Figure 2). The global rise in sea level has been estimated by various workers to be on the order of 1-3 mm/year (4-12 in. per century). The large range is due to the difficulty of separating that worldwide component from local tectonic and isostatic effects included in records from tide gauges. Assuming that the eustatic rise in sea level is on the order of 2 mm/year (8 in. per century), the results from Figure 3 indicate that the south coast of Oregon is tectonically rising at a rate of about 2-3 mm/year (8-12 in. per century), while the stretch between Newport and Tillamook is approximately stable, neither rising nor falling tectonically.

It is apparent that the along-coast differences in tectonic uplift versus changing levels of the sea deduced from Figure 3 will be relevant to spatial patterns of coastal erosion. However, there also appears to be a temporal change in the tectonics that would be important to erosion. Earthquake activity is generally associated with subduction zones such as the one in the Northwest—seismic events formed by the plates' scraping together as the oceanic plate slides beneath the continental plate. The Northwest coast is anomalous in that respect in that there have been no historic earthquakes that 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 limited to establish whether earthquakes do accompany subduction. This evidence has come from investigations of estuarine marsh sediments buried by sand layers, deposits suggesting that during prehistoric times portions of the coast have abruptly subsided, generating an extreme tsunami that swept over the area to deposit the sand (Atwater, 1987; Darienzo and Peterson, 1990; Atwater and Yamaguchi, 1991).

Based on the number of such layers found in Willapa Bay, Washington, and Netarts Bay, Oregon, 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 indeed occur along the Northwest coast—but with long periods of inactivity between events. An earthquake releases strain built up by subduction, and the result is that some areas of the coast drop by 1-2 m (3-6 ft) during the release, whereas other areas undergo minimal subsidence. Between earthquake events the strain is accumulating, and this produces a general uplift of the coast as recorded by the tide gauges and geodetic surveys within historic times (Figures 2 and 3).

Another potential change in the present-day pattern of sea-level rise versus coastal uplift is associated with predictions for an accelerated rise in sea level associated with future greenhouse warming. Global temperatures have been predicted to increase from 1.5° to 4.5° by the year 2050 (National Research Council, 1983). Those 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 National Research Council (1987) predicts that by the year 2025 the global sea level will rise by 10-21 cm (4-8 in.). Although this may seem insignificant, the effects on sandy shorelines may be magnified 100 times in the horizontal direction, resulting in shoreline erosion of 10-21 m (33-70 ft).

There are many uncertainties in the analyses of sea-level rise resulting from greenhouse warming, and therefore the resulting predictions have been controversial among scientists. Different investigators who studied 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, 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 world's most dynamic environments. 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 give way to waves and currents, retreating back toward the land. At times, this beach loss continues until the erosion threatens homes and motels and cuts away at public parklands.

Ocean waves

The extreme seasonality of the Oregon climate results in parallel variations in ocean processes and exerts the primary control on natural cycles observed on beaches. The varying energy of ocean waves parallels the seasonally varying storm winds, because the strength of those winds is the primary factor in causing the growth of waves. In general, the greater the wind velocities blowing over the ocean's surface, 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's 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 first 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, a storm need not be in the immediate coastal zone. Waves reach the shores of Oregon from storms all over the Pacific Ocean, even from storms in the southern hemisphere near Antarctica. Our largest waves are derived, however, from winter storm systems moving 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 that are usually employed in measuring 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 they formerly obtained by visual determination. A microseismometer system is also in operation at the Oregon State University Hatfield Marine Science Center in Newport, one that 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 in the Marine Science Center can provide records of ocean waves—after all, the Center is nearly 2 mi from the ocean. However, even more impressive is the fact that the waves can be detected on the seismometer at Oregon State University in Corvallis, 60 mi inland. When the surf is high on the coast, its effects can be seen as small jiggles in the seismometer recordings.

The microseismometer in the Marine Science Center differs from normal seismometers in that it is so tuned as to amplify small tremors, whether they are caused by earthquakes too minor to be felt or generated by ocean waves along the coast. In order to use the recordings from the microseismometer to measure ocean waves, it was necessary to first calibrate the system (Zopf and others, 1976; Creech, 1981). This calibration 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 taken with a pressure transducer, an instrument that rests on the ocean bottom and records wave pressures that are directly proportional to the heights of the waves passing over the transducer.

The pressure transducer, the most commonly used instrument for measuring ocean waves directly, would be preferable to the microseismometer. However, winter storms experienced along the Northwest coast are so intense that 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 in 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 were continued for 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 the daily wave measurements obtained from the microseismometer is shown in Figure 4, covering the period from mid-December 1972 through January 1973. Most apparent in this series are the storm waves that struck the Oregon coast during Christmas. The breaker heights at that time reached 7 m, about 23 ft, 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.

Thus, the significant wave height can be evaluated from measurements obtained with 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

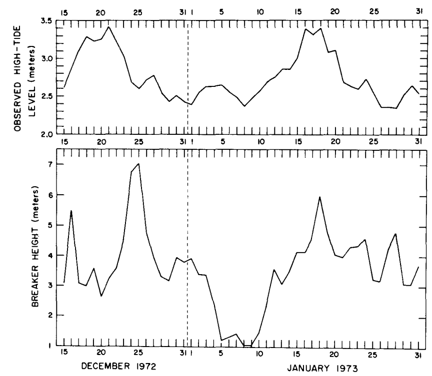


Figure 4. An example of daily variations in wave conditions measured by the microseismometer at Newport, covering the interval from December 1972 through January 1973. From McKinney (1977).

is because an observer normally tends to weight the 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 time interval will be a factor of about 1.8 times the significant wave height (Komar, 1976). Therefore, when the graph of Figure 4 indicates the occurrence of a significant wave height of 7 m during Christmas 1972, there must have been individual waves having heights of about 1.8 x 7 m = 12.6 m (>41 ft)! As might be expected, there was considerable erosion along the coast during that storm, the severest impact occurring at Siletz Spit on the mid-Oregon coast.

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

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

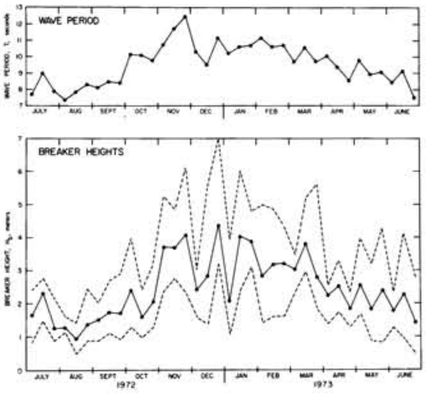


Figure 5. The monthly variations of wave breaker heights and periods at Newport, illustrating the occurrence of higher wave conditions during the winter months. Solid line is for mean heights (significant wave heights) for one-third-month intervals; dashed lines are for largest and smallest breakers for those intervals. From Komar, Quinn, and others (1976).

rigs exploring for oil measured an individual wave at a height of 29 m (95 ft) (Rogers, 1966; Watts and Faulkner, 1968). This is close to the 112-ft height of the largest wave ever reliably measured in the ocean, 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 is similar on most coastlines and is illustrated schematically in Figure 6. 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 profile that is above the high-tide line). This eroded sand moves to deeper water, where it then accumulates in offshore bars, approximately in 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.

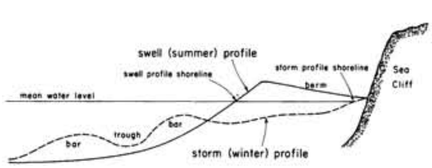


Figure 6. General pattern of seasonal changes in beach profiles associated with parallel variations in wave energies. From Komar (1976).

This cycle has been demonstrated to occur on Oregon beaches, just as it has been observed 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 the one at Gleneden Beach south of Lincoln City (Aguilar-Tunon and Komar, 1978). These two beaches were selected because of their contrasting sand sizes that 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 increases as grain size increases. Gravel beaches are the steepest, their slopes sometimes reaching 25°-30°, whereas the overall slope of a fine-sand beach may be only 1°-2°. This is seen in the comparison of the beach profiles at Gleneden Beach and Otter Rock (Figure 7), the beach at Gleneden being coarser and hence steeper.

The month-by-month changes in the profiles at Gleneden Beach are shown in Figure 8. These profiles were obtained with standard surveying gear and by wading into the water. They do not show the offshore bars, which were too deep to reach. However, these profiles do illustrate the rapid retreat of the beach as the winter season develops. 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 change was much smaller than at Gleneden Beach. Sand elevations at Gleneden changed by as much as 2-3 m (8 ft) (Figure 8), while the changes at Otter Rock amounted to less than 1 m (3 ft). This 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.

The greater response of coarser grained beaches to storm waves is of importance to coastal-erosion processes, since the waves are able to rapidly cut through the beach to reach homes and other structures. This 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. So, 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 that 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 that property erosion does not occur. It is only when the beach berm completely disappears and the waves can wash against the 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 that 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

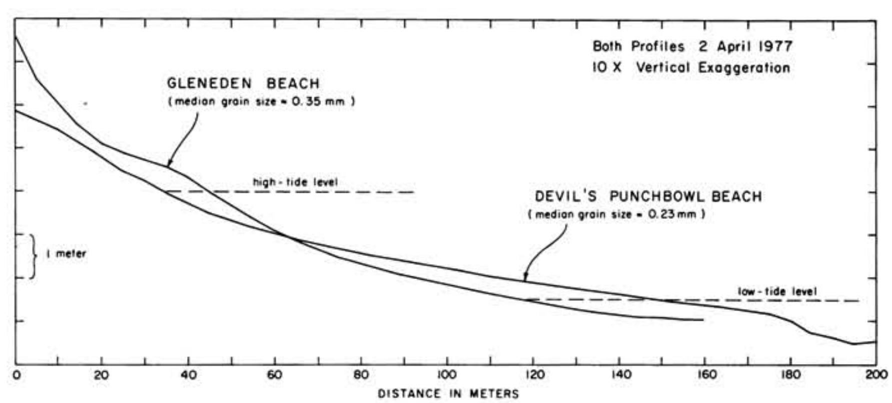


Figure 7. Beach profiles from Gleneden Beach and Devil's Puchbowl Beach (Otter Rock), Oregon, illustrating that coarser-sand beach (Gleneden) is steeper. From Aguilar and Komar (1978).

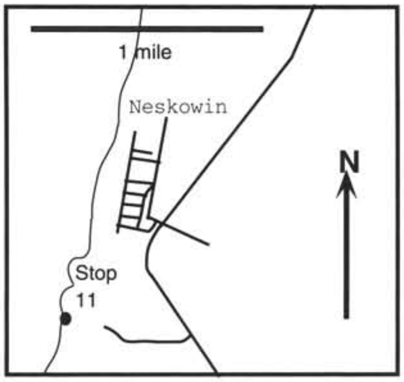


Figure 40. Location of Stop 11.

Figure 41. Photograph of large stumps developed on late Holocene wave-cut platform now in the surf zone. Age of one tree is about 2,000 RCYBP.

one cell to another. Progradational beaches north of Neskowin contrast sharply with widespread shoreline retreat in the Lincoln City cell to the south. The two cells differ in sand supply, even though the adjacent cells are similar in size, river drainage-basin area, and predicted rate of uplift (Goldfinger and others, 1992). We hypothesize that sand is excavated from bluffs in the Lincoln City cell and bypassed around Cascade Head to supply the Neskowin-Pacific City cell. Such a process could be accelerated during periods of beach and sea cliff erosion following coseismic subsidence.

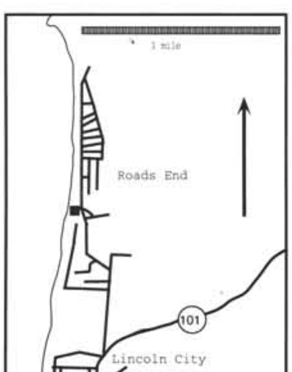


Figure 42. Location of Stop 12.

Stop 12. Roads End Beach Wayside

Location (Figure 42): At the north end of Lincoln City find Roads End intersection with Highway 101 (McDonalds Restaurant on opposite side of intersection). Take road north to the Roads End Beach Wayside parking lot. Walk down the path to the beach and head south 300 ft to view the late Pleistocene sea cliffs.

Features: Coseismic liquefaction features (convolute bedding) in late Pleistocene marine-terrace deposits.

Site description: The late Pleistocene terraces (assumed 80 ka in age) contain highly convoluted heavy-mineral layers that were originally deposited in planar foreshore beds (Figure 43). These convoluted beds are associated with clastic dikes, sills, and vent-collapse structures in many of the Oregon marine terrace deposits (Peterson and Madin, 1992). The tops of the convoluted beds at this site are eroded, proving the liquefaction to have occurred during the period of marine terrace deposition and not during the subsequent 80,000 years (Figure 44). Higher up in the section, some transitional backshore to eolian dune sands show a second liquefaction event. Some sites in the Lincoln City terraces show as many as three distinct liquefaction events that occurred during the period of deposition, near the end of that marine transgression.

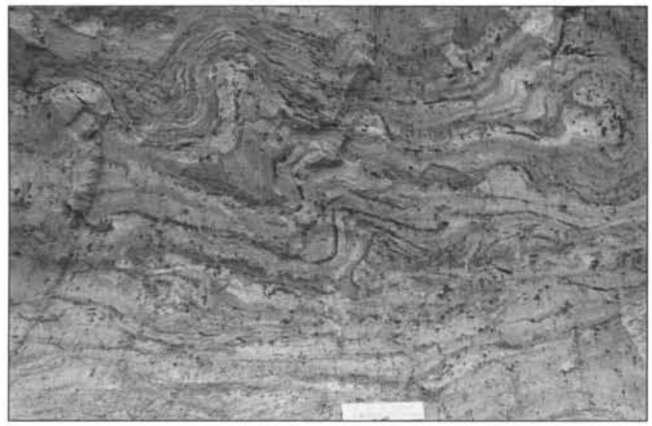


Figure 43. Photograph of highly convoluted beds in marine terrace deposits.

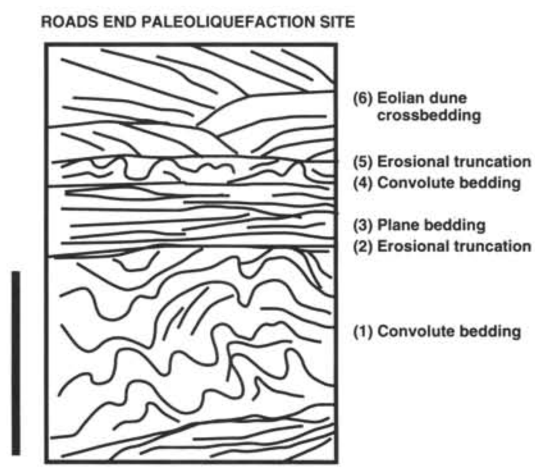


Figure 44. Convoluted foreshore deposits (liquefaction) are truncated by overlying planar beds. A second liquefaction event is represented by smaller scale convolute beds in overlying dune deposits. Liquefaction in dune deposits generally corresponds to permeability caps under interdune pond sediments. High pore pressure allowed localized liquefaction above the water table.

Stop 13. Salishan Spit, Siletz Bay

Location (Figure 45): Stop at the office of Salishan Lease Holders, Inc., on the west side of Highway 101, opposite Salishan Lodge. With permission, drive out to the Siletz Bay spit. Park at the north end of the Golf course and walk east on a dirt access road to about half the length of the fairway. Leave the road and walk north 300 ft through trees and/or brush to wetland areas.

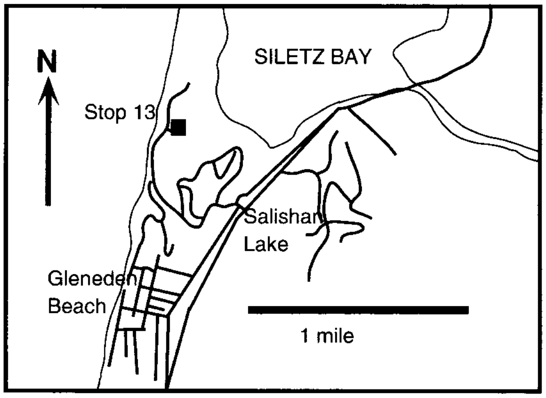


Figure 45. Location of Stop 13.

Features: Coseismically buried peats and tsunami-deposited sands.

Site description: Siletz Bay is separated from the Pacific Ocean by a spit that now contains an artificially stabilized foredune and numerous residences (Figures 1 and 46). Six coseismically buried peats with overlying tsunami deposits up to 25 cm (10 in.) thick are identified in marsh deposits along the spit and across the bay in the Millport Slough area (Darienzo, 1991) (Figures 46 and 47). The Salishan House site at the southern end of the spit contains particularly well-developed buried peats and overlying tsunami sands.

This is an example of one of our tsunami sensitive sites. The tsunami sands on the spit are derived directly from the beach (spit overtopping) except for the fifth buried peat, which shows a significant river sand component. Since the bay deposits adjacent to the spit are predominantly of river sand mineralogy (Peterson and others, 1984), a tsunami that propagated up the bay would deposit a mixed (beach/river) sand mineralogy. The overtopping of the spit by three of the last four paleotsunamis substantiates estimates of tsunami runup (+ 6 m [20 ft] MSL) predicted from the Cannon Beach sites discussed above. Additional work is needed here to establish the age of the topmost tsunami layer, which is contaminated by young roots descending from the modern wetlands.

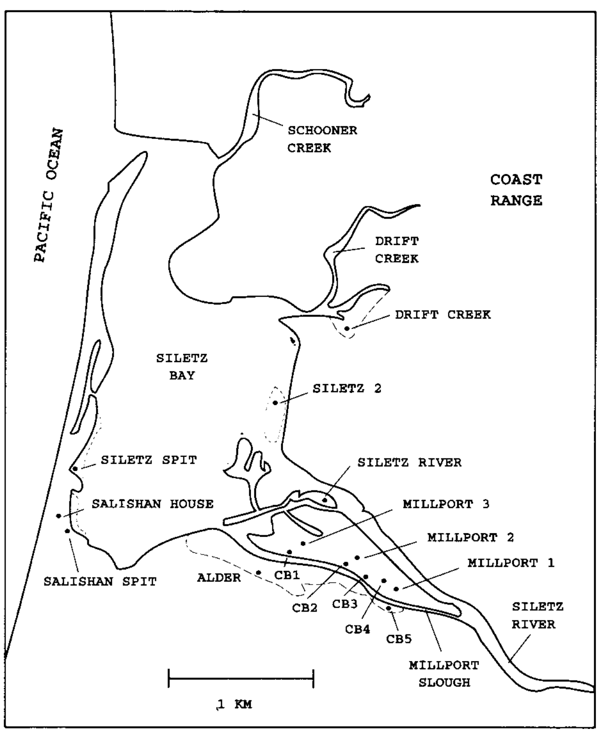


Figure 46. General map of Siletz Bay showing core site locations. Dashed lines show cored marsh areas.

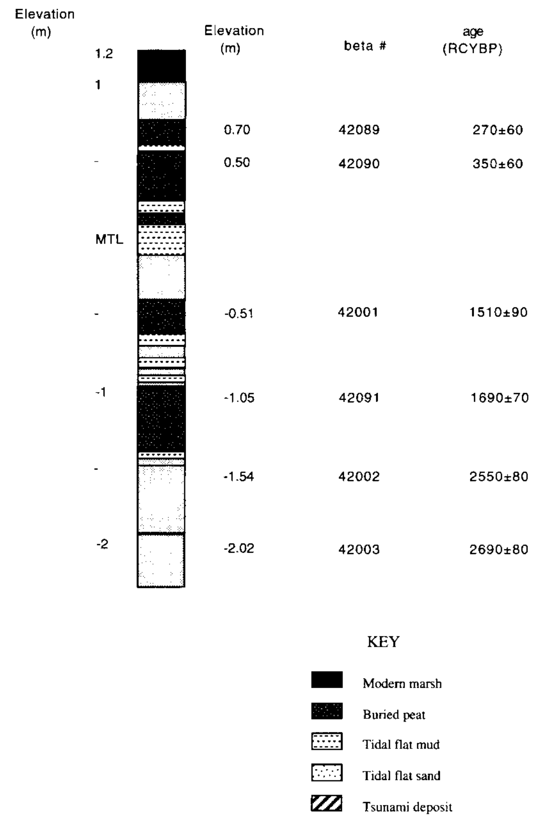


Figure 47. Radiocarbon ages of buried peats from Salishan House. Beta number is radiocarbon laboratory sample number.

Stop 14. Millport Slough, Siltez Bay

Location (Figure 48): Turn east off Highway 101 at Alder Road, immediately north of Salishan Lodge. Take the road to a small bridge over a tidal channel (Millport Slough). Explore cutbanks southeast of the Millport Slough bridge to find one to three buried wetland horizons. **Requires low tide.**

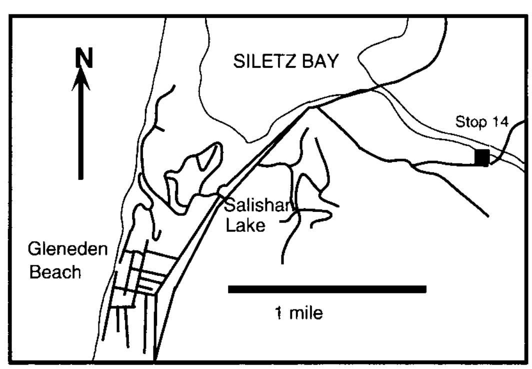


Figure 48. Location of Stop 14.

Question 3-9. For the sake of simplicity, let's assume that the Oregon coast has been tectonically stable for the late Pleistocene (which, we know is false, but let's make it simple). According to the graph in Fig. 5D, what was the position of relative sea level, compared to modern, along the Oregon coast 120,000 years ago? How about 50,000 years ago?

Task 5C. Figure 5E is a topographic profile of the Oregon coast at the latitude of Newport. Note the position of modern sea level, the continental shelf, and continental slope. Using colored pencils, draw and label the position of sea level along the Oregon coast at 120,000 years ago and 50,000 years ago (refer back to your graph in Fig. 5D).

Question 3-10. Comment on what the topography of the Oregon coast looked like 120,000 and 50,000 years ago, compared to that of today. Were there rocky shorelines and sea cliffs? Was there more or less "beach" area. What did the coastal rivers (e.g. Yaquina, Nestucca, Columbia) look like at those times, compared to today? Draw diagrams as necessary to support your answer.

Part 4. Rates of Coastal Erosion in Oregon.

Coastal Oregon is very dynamic with respect to geomorphic process. Cliff erosion, beach erosion, landsliding, and flooding are common occurrences, particularly during the stormy winter months. Each year, millions of dollars in damage occurs to property along the Oregon coast. The understanding of geomorphic process is critical for the appropriate design of land-use regulations and housing plans.

Figure 6 is a map of the Nye Beach area of Newport, OR. The Jumpoff Joe sea cliff area is a classic example of very active historic erosion along the Oregon coast. By examining historic aerial photographs, it is possible to map out the position of the sea cliffs, and the rate of erosion over time. The map shows sea cliff positions for the years 1868, 1939, and 1967. The black squares represent the position of buildings and houses during this time frame.

Task 6. Using the map data in Fig. 6, determine the historic rates of sea cliff erosion at the Nye Beach area. Fill in the table below for cross-sections A-A', B-B', and C-C'.

Section I.D.	Record Period (date range)	Retreat Distance (meters)	Time Interval (years)	Retreat Rate (m/yr)
A-A'	1868-1939	________	________	_______
A-A'	1939-1967	________	________	_______
A-A'	1868-1967	________	________	_______
B-B'	1868-1939	________	________	_______
B-B'	1939-1967	________	________	_______
B-B'	1868-1967	________	________	_______
C-C'	1868-1939	________	________	_______
C-C'	1939-1967	________	________	_______
C-C'	1868-1967	________	________	_______

Question 4-1. How does the rate of sea cliff erosion vary spatially from section A-A', south to section C-C' along Nye Beach?

Question 4-2. How has the rate of sea cliff erosion varied temporally between 1868 and 1967, at each of these localities? Which show the highest rates, the lowest rates, how have the rates changed over time?

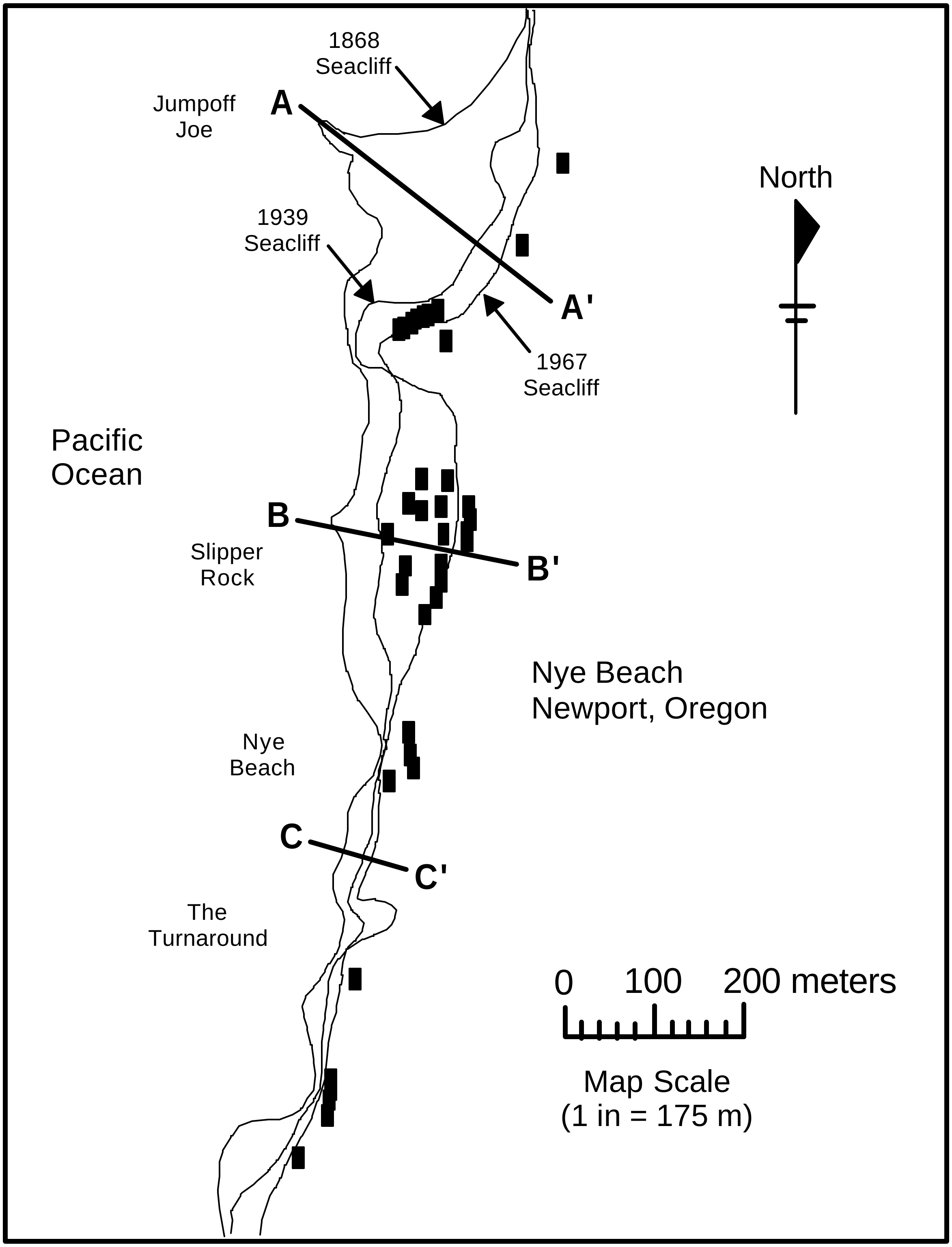
Question 4-3. Hypothesize why the rates of sea cliff erosion have slowed in the Jumpoff Joe area, historically.

Question 4-4. Look at a geologic map of Oregon (e.g. on the lab room wall), what geologic unit(s) underlies the Nye Beach area? Hypothesize how bedrock geology may be controlling the spatial patterns you observed in Question 4-1.

Question 4-5. What has happened to the large cluster of houses between points B and B'? When did this happen? List the processes that were likely involved.

Question 4-6. Given the 1939-1967 rates of erosion in the Jumpoff Joe area (line A-A'), how long did it take for the three houses southeast of the 1967 shoreline to bite the dust? Do you think they are still there? Why or why not?

Figure 6. Map showing sea cliff retreat in the Jumpoff Joe area of Newport, Oregon (from Komar, 1992; as documented by Stembridge, 1976).



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Reference Tables

APPENDIX 7
Table for length conversion

Unit	mm	cm	m	km	in	ft	yd	mi
1 millimeter	1	0.1	0.001	10^{-6}	0.0397	0.00328	0.00109	6.21×10^{-7}
1 centimeter	10	1	0.01	0.0001	0.3937	0.0328	0.0109	6.21×10^{-6}
1 meter	1000	100	1	0.001	39.37	3.281	1.094	6.21×10^{-4}
1 kilometer	10^{6}	10^{5}	1000	1	39,370	3281	1093.6	0.621
1 inch	25.4	2.54	0.0254	2.54×10^{-5}	1	0.0833	0.0278	1.58×10^{-5}
1 foot	304.8	30.48	0.3048	3.05×10^{-4}	12	1	0.333	1.89×10^{-4}
1 yard	914.4	91.44	0.9144	9.14×10^{-4}	36	3	1	5.68×10^{-4}
1 mile	1.61×10^{6}	1.01×10^{5}	1.61×10^{3}	1.6093	63,360	5280	1760	1

APPENDIX 8
Table for area conversion

Unit	cm^2	m^2	km^2	ha	in^2	ft^2	yd^2	mi^2	ac
1 sq. centimeter	1	0.0001	10^{-10}	10^{-8}	0.155	1.08×10^{-3}	1.2×10^{-4}	3.86×10^{-11}	2.47×10^{-8}
1 sq. meter	10^{4}	1	10^{-6}	10^{-4}	1550	10.76	1.196	3.86×10^{-7}	2.47×10^{-4}
1 sq. kilometer	10^{10}	10^{6}	1	100	1.55×10^{9}	1.076×10^{7}	1.196×10^{6}	0.3861	247.1
1 hectare	10^{8}	10^{4}	0.01	1	1.55×10^{7}	1.076×10^{5}	1.196×10^{4}	3.861×10^{-3}	2.471
1 sq. inch	6.452	6.45×10^{-4}	6.45×10^{10}	6.45×10^{-8}	1	6.94×10^{-3}	7.7×10^{-4}	2.49×10^{-10}	1.574×10^{-7}
1 sq. foot	929	0.0929	9.29×10^{-8}	9.29×10^{-6}	144	1	0.111	3.587×10^{-8}	2.3×10^{-5}
1 sq. yard	8361	0.8361	8.36×10^{-7}	8.36×10^{-5}	1296	9	1	3.23×10^{-7}	2.07×10^{-4}
1 sq. mile	2.59×10^{10}	2.59×10^{6}	2.59	259	4.01×10^{9}	2.79×10^{7}	3.098×10^{6}	1	640
1 acre	4.04×10^{7}	4047	4.047×10^{-3}	0.4047	6.27×10^{6}	43,560	4840	1.562×10^{-3}	1

APPENDIX 9
Table for volume conversion

Unit	mL	liters	m^3	in^3	ft^3	gal	ac-ft	million gal
1 milliliter	1	0.001	10^{-6}	0.06102	3.53×10^{-5}	2.64×10^{4}	8.1×10^{-10}	2.64×10^{-10}
1 liter	10^{3}	1	0.001	61.02	0.0353	0.264	8.1×10^{-7}	2.64×10^{-7}
1 cu. meter	10^{6}	1000	1	61,023	35.31	264.17	8.1×10^{-4}	2.64×10^{-4}
1 cu. inch	16.39	1.64×10^{-2}	1.64×10^{-5}	1	5.79×10^{-4}	4.33×10^{-3}	1.218×10^{-8}	4.329×10^{-9}
1 cu. foot	28,317	28.317	0.02832	1728	1	7.48	2.296×10^{-5}	7.48×10^{6}
1 U.S. gallon	3785.4	3.785	3.78×10^{-3}	231	0.134	1	3.069×10^{-6}	10^{6}
1 acre-foot	1.233×10^{9}	1.233×10^{6}	1233.5	75.27×10^{6}	43,560	3.26×10^{5}	1	0.3260
1 million gallons	3.785×10^{9}	3.785×10^{6}	3785	2.31×10^{8}	1.338×10^{5}	10^{6}	3.0684	1

APPENDIX 10
Table for time conversion

Unit	sec	min	hours	days	years
1 second	1	1.67×10^{-2}	2.77×10^{-4}	1.157×10^{-5}	3.17×10^{-8}
1 minute	60	1	1.67×10^{-2}	6.94×10^{-4}	1.90×10^{-6}
1 hour	360	60	1	4.17×10^{-2}	1.14×10^{-4}
1 day	8.64×10^{4}	1440	24	1	2.74×10^{-3}
1 year	3.15×10^{7}	5.256×10^{5}	8760	365	1