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Crustal Deformation at the Leading Edge of the Oregon Coast Range Block, Offshore Washington (Columbia River to Hoh River)

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Earthquake Hazards of the Pacific Northwest Coastal and Marine Regions

Robert Kayen, Editor

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By Patricia A. McCrory, David S. Foster, William W. Danforth, and Michael R. Hamer

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Crustal-block motions complicate the slip rates of seismotectonic elements of the Cascadia subduction boundary

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Crustal Deformation at the Leading Edge of the Oregon Coast Range Block, Offshore Washington (Columbia River to Hoh River)

By Patricia A. McCrory[1], David S. Foster[2], William W. Danforth[2], and Michael R. Hamer[1]

Abstract

Determination of slip rates for the seismotectonic elements that make up the Cascadia subduction boundary is complicated by crustal-block motion within the forearc. In particular, geodetic measurements (McCaffrey and others, 2000) indicate that the Oregon Coast Range block is translating north-northwest at 6-10 mm/yr relative to stable North America. Differential motion of the Oregon Coast Range block with respect to other crustal blocks that compose the forearc results in a complex pattern of faults and folds with orientations highly oblique to those expected from subduction-driven contraction.

The leading (northwestern) edge of the Oregon Coast Range block traverses coastal Washington, where it abuts subduction-complex rocks of the Olympic Mountains to the north. Block kinematics predicts north-northwest-directed contraction where the boundary trends east-northeast near Grays Harbor, Washington. Crustal deformation observed near Grays Harbor is consistent with north-northwestward motion of the Oregon Coast Range block. Deformation is localized within the more ductile subduction-complex rocks of the Olympic coast rather than the more rigid basaltic rocks that underlie the Oregon Coast Range block. The boundary near Grays Harbor is relatively discrete. Seismic-reflection and sidescan-sonar data image several faults and folds that trend east-northeast on the inner continental shelf between Grays Harbor and Cape Elizabeth, Washington, across a zone 40 km wide from south to north. In contrast, in Puget Sound to the east, where the Oregon Coast Range block abuts the relatively rigid pre-Tertiary rocks of Vancouver Island, relative motion is distributed across a zone more than 200 km wide from south to north.

An estimate of the contraction rate across the transverse coastal zone can be derived from the degree of folding and faulting of Quaternary strata. Constraints on the contraction rate come from two datums: (1) the older datum (Raft datum) is an unconformity at the base of a widespread, glacially derived deposit, and (2) the younger datum (Langley datum) is a widespread unconformity marking the base of upper Pleistocene sediments (<150 ka). Shortening of the older, ca. 900-600-ka Raft datum is ~2.0 km, corresponding to a shortening rate of 2.2-3.3 mm/yr.

Quaternary folds range in length from 3 to 15 km and have amplitudes as great as 160 m. Of the seven primary faults and associated folds imaged in this coastal zone, the southernmost structure appears to be the most important. This structure marks the boundary between the Oregon Coast Range block and the Olympic Mountains block, vertically offsets the late Pleistocene Langley datum by more than 40 m, and displaces the sea floor several meters. Other faults in this zone displace the late Pleistocene datum 2-10 m.

Block kinematics predicts transpressional shear where the block boundary trends north-northwest through Willapa Bay, Washington. Crustal deformation observed in Willapa Bay documents a component of contraction in an east-northeast direction. Lateral slip, if any, on mapped faults remains undetected. In contrast to the kinematic model, this portion of the boundary exhibits the geometry of an east-dipping thrust fault.

Faults mapped with both new and earlier seismic-reflection data are consistent with a thrust geometry. These data depict a 3-km-wide, 30-km-long fault zone trending north-northwest through Willapa Bay. However, the main strand shows a sense of vergence opposite to that of the inferred boundary thrust. The main fault strand vertically displaces a ca. 20-ka erosional surface beneath the bay 10-12 m, west side up. A vertical displacement rate of 0.5±0.2 mm/yr has been calculated across the entire fault zone. This "Willapa Bay" fault zone may include a backthrust above the block boundary. Alternatively, the fault geometry may reflect lateral juxtaposition of bedrock slivers. No shallow trace (<90 m) has been identified for the postulated block boundary fault. Nonetheless, the recency and magnitude of movement on the "Willapa Bay" fault zone implies activity on the postulated main thrust to the west.

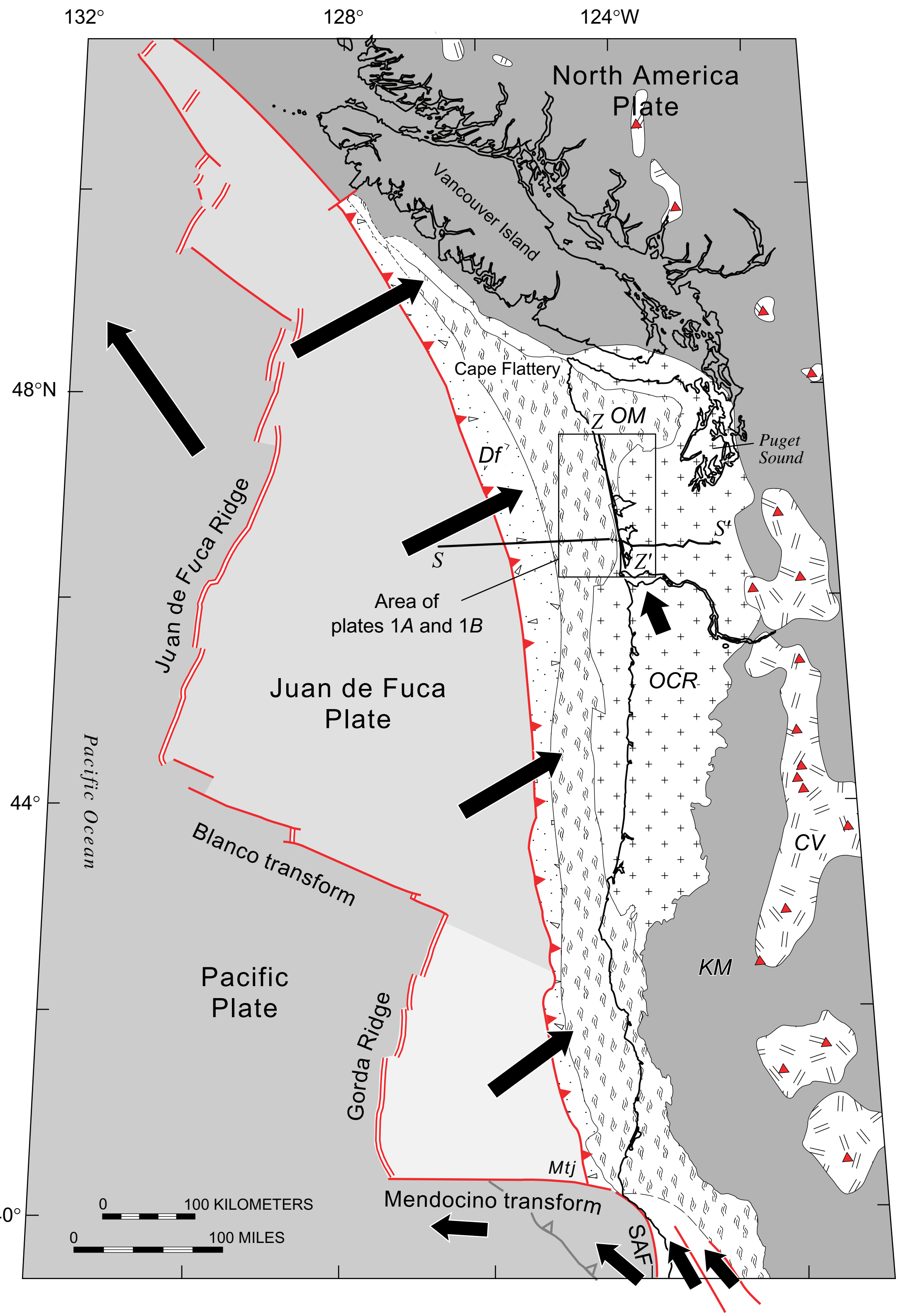
These observations provide the first constraints on rates of Quaternary structural activity in coastal Washington. Furthermore, they suggest that this region may accommodate as much as half of the differential motion between the Oregon Coast Range block and stable North America. The remaining differential motion may be accommodated in the Olympic Mountains, which are uplifting as much as ~3.2 mm/yr as indicated by geodetic measurements (Savage and others, 1991), or as broad-scale buckling of the Oregon Coast Range basement beneath Grays Harbor and Willapa Bay.

Introduction

The Washington Continental Shelf lies within the Cascadia subduction margin and continues to be shaped by plate-boundary processes far from the deformation front. A major

[1]Menlo Park, California
[2]Woods Hole, Massachusetts

crustal block boundary in the subduction margin traverses the study area and continues onshore to mark the southern edge of the Olympic Mountains (fig. 1). Structural evidence for differential motion across this block boundary has been documented both onshore (McCrory, 1996, 1997) and offshore (Wolf and others, 1997, 1998). Sidescan-sonar and high-resolution seismic-reflection data were collected offshore in 1997 and 1998 on the RV *McArthur* (Foster and others, 1999a, b; McCrory and others, in press a, b) and RV *Corliss* (Cross and others, 1998, 1999; Twichell and others, 2000). These data image a structural high that extends 40 kilometers from Grays Harbor northward to Cape Elizabeth, Washington, and is at least 35 kilometers wide from the coast to the mid shelf. This high encompasses a zone of faults and tight folds,



which accommodates north-northwest-directed crustal shortening along the block boundary. The geophysical data reveal numerous sites where fault scarps, anticlinal folds, or diapirs breach the sea floor. Little was previously known about the age of most recent movement on these structures or the rate at which they deform (Grim and Bennett, 1969; Wagner and others, 1986; Palmer and Lingley, 1989).

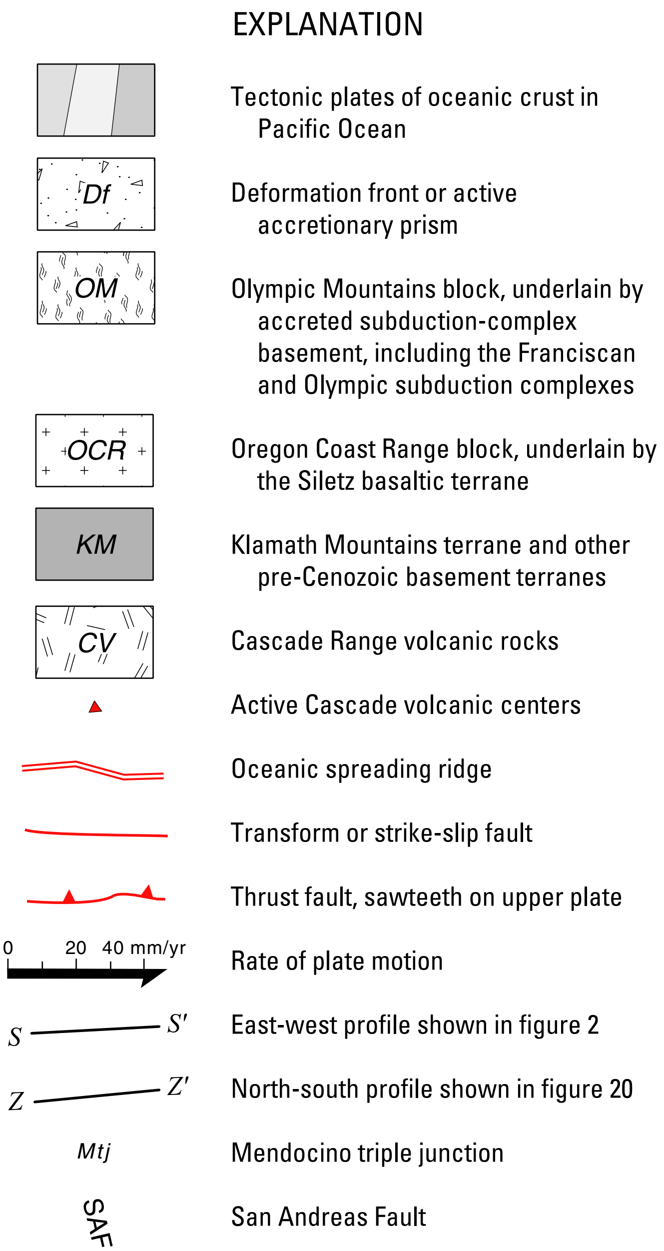


Figure 1. Generalized map of the Cascadia subduction system showing tectonic plates and major basement units (modified from Jennings and others, 1977; Tipper and others, 1981; Hyndman, 1995; Snavely and Wells, 1996; McCrory, 1996; Wells and others, 1998). The Cascadia subduction margin extends from the thrust fault along the trench axis to the volcanic arc, approximately 300 km to the east. Black arrows denote modern motion vectors of the oceanic plates, and of the Oregon Coast Range (OCR) forearc block, with respect to a fixed North America Plate, in mm/yr (Wilson, 1993; D. S. Wilson, written commun., 2001). North-northwest motion of northern California relative to a rigid North America requires a component of shortening along the Oregon-Washington coast, in part concentrated along the northern boundary of the OCR block. Cross section S-S′ is shown in figure 2; cross section Z-Z′ is shown in figure 20.

Active faults in the nearshore area have been understudied because of the difficulty of collecting marine seismic and sidescan data in shallow water and the difficulty of quantifying fault-slip histories. However, nearshore faults can represent a significant seismic hazard owing to their proximity to populated areas and their shallow rupture depths as compared to the subduction plate boundary. In southern Washington, many nearshore faults extend onshore, so the full length of these faults needs to be determined in order to calculate expected earthquake magnitudes. In fact, the ability to link structures across the shoreline makes the Olympic coast an ideal area in which to recognize and quantify recent structural activity in the nearshore zone and to develop methodologies suited to this challenging zone.

Tectonic Setting

The 1,100-km-long Cascadia subduction boundary accommodates relative motion between the subducting Juan de Fuca oceanic plate and the North America continental plate across a 300- km-wide region from the trench to the volcanic arc (fig. 1). Relative motion is directed northeastward and varies from about 23 mm/yr at the southern end of the Cascadia subduction zone to 40 mm/yr at the northern end (Wilson, 1993). Along the southern half of the plate boundary, relative motion is oblique to the trend of the trench axis. This obliquity may drive translation of crustal blocks in the forearc region (Wells and others, 1998). Along the northern half of the boundary, where the study area is located, relative motion is fairly orthogonal.

The Cascadia Continental Slope is primarily composed of weakly consolidated material that has been scraped off the subducting Juan de Fuca Plate and accreted to the margin over the past 2 m.y. (Barnard, 1978). The Continental Shelf and onshore forearc regions are underlain by an amalgamation of terranes that were accreted during Cenozoic subduction. The Siletz terrane, a major forearc terrane, is composed of Paleogene oceanic basalt and overlying marine strata (Wells and others, 1984, Tréhu and others, 1994; Snavely and Wells, 1996). This terrane underpins the relatively rigid Oregon Coast Range (OCR) block, which is more than 500 km long and 200 km wide (Wells and others, 1998) and occupies the central portion of the forearc (fig. 1). Differential motion between the OCR block and adjacent forearc blocks creates regional structural complexities within the overall subduction regime.

Ongoing north-south shortening in western Washington is attributed to northward translation of the Oregon Coast Range block relative to stable North America, nominally represented by pre-Tertiary basement rocks of southern British Columbia (Snavely and Wells, 1996). Beneath Puget Sound this shortening occurs within the relatively thin (15-20 km thick) Siletz basement (Tréhu and others, 1994; Stanley and others, 1999) and overlying strata where a series of rhomboidal basins and intervening basement highs provide striking aeromagnetic images (Blakely and others, 1999). Here, approximately 4 mm/yr of contraction is distributed across at least six faults

(Gower and others, 1985; Johnson and others, 1996, 1999; Wells and Johnson, 2000). In coastal Washington the Siletz terrane is thicker (~30 km thick, Parsons and others, 1999) than under Puget Sound and abuts the significantly more ductile subduction-complex terrane that underpins the Olympic Peninsula (fig. 2) rather than pre-Tertiary basement. Velocity models based on wide-angle refraction data suggest a rhomboidal structural configuration for Siletz rocks at mid-crustal depths in coastal Washington (Parsons and others, 1999), with lows under Willapa Bay and Grays Harbor and an intervening high. This configuration is similar to that observed in the Puget Sound region, but with significantly less apparent relief and no direct evidence of recent structural movement associated with basement irregularities.

Kinematic Model of Coast Range Block Motion

Cenozoic reconstruction of the western continental margin of North America based on the kinematic model of Wilson and others (in press) assigns relative motion to several crustal blocks that compose the continental margin. Specifically, this model refines the late Cenozoic Pacific-Juan de Fuca-North America velocity vectors by considering Oregon Coast Range motion relative to North America. The kinematic model indicates that the Oregon Coast Range (OCR) block has moved about 150 km northward in concert with the Sierra Nevada-California Coast Ranges block relative to stable North America over the past 15 m.y. at an average rate of 10 mm/yr, similar to calculations by Atwater and Stock (1998).

We presume that northward translation of the block over the past 15 m.y. has been converted to convergence in the upper crust at the northern or leading edge of the Coast Range block. In coastal Washington, we propose that this convergence is primarily accommodated by a combination of east-northeast-striking thrust faults and north-northwest-striking strike-slip faults.

The kinematic models of Wells and others (1998) and Wells and Simpson (2001) for modern block motion within the Cascadia forearc defines a pole of rotation for the OCR block in central Washington based on very long baseline interferometry (VLBI) and paleomagnetic data. This model predicts northward motion of about 8 mm/yr in coastal Washington. Recent global positioning system (GPS) measurements indicate a rate of about 6-10 mm/yr (McCaffrey and others, 2000) for the same region. Northward translation of the OCR block during the Quaternary results in convergence directed 340° between this block and the Olympic Mountains block in southwestern Washington. We attribute crustal deformation observed between the Columbia River and the Queets River (46° to 47°30'N.) (fig. 3) primarily to this differential motion.

Location of Block Boundary in Coastal Washington

The OCR block boundary in coastal Washington is characterized by juxtaposition of basaltic crust of the Siletz terrane against marine sedimentary and volcanic rocks and mélange of the Olympic subduction complex. Aeromagnetic data depict a sharp contrast in the upper crust where magnetic basaltic basement is juxtaposed against the mélange basement (Finn and others, 1998; R. J. Blakely, written commun., 2000). The boundary inferred from aeromagnetic data trends north-south in southwestern Washington, paralleling the coastline between the Columbia River and Grays Harbor, and then swings eastward around the Olympic Mountains. A similar pattern is seen in Bouger gravity data, where the strong gradient tracks the boundary between the Olympic subduction complex and the more dense basalt basement of the Crescent Formation (Finn and others, 1998; R. J. Blakely, written commun., 2000).

The *P*-wave velocity model calculated by Parsons and others (1999) suggests that the boundary is an eastward-dipping thrust fault beneath the Willapa Bay-Grays Harbor area. Subsurface lithostratigraphies from petroleum exploration wells (Rau and McFarland, 1982; Snavely and Wagner, 1982) also suggest a dipping contact in the upper 2 km beneath Willapa Bay and Grays Harbor—OCR basaltic rocks are there underlain by fractured and sheared sedimentary rock interpreted to be subduction-complex mélange (fig. 3). In contrast, only subduction-complex rocks are found beneath overlap strata in wells north of the boundary, and only OCR basaltic rocks are present in wells south of the boundary.

Evidence for Contraction at Block Boundary

The 6-10 mm/yr of relative motion between the OCR and North America provides a contraction budget of about 8 mm/yr in coastal Washington. The distribution and orientation of faults and folds onshore to the east of the study area suggest that direct convergence occurs between the OCR and Olympic Mountains (OM) and is expressed in part as permanent crustal contraction within the subduction-complex rocks. In the Grays Harbor-Moclips River area (47° to 47°15'N.), Quaternary thrust faults and folds trend east-northeast (fig. 3), indicating northward-directed crustal shortening (McCrory, 1996, 1997).

This deformation pattern is consistent with the contemporary north-south compressional stress field observed in petroleum exploration wells in the area. Magee and Zoback (1992) infer a N.12°W. principal compressive stress direction in the Shell Sampson Johns No. 1-15 well beneath Saddle Hill on the basis of borehole-breakout data (fig. 3, site N). Magee and Zoback (1992) infer stress directions of N.05°W. and N.14°E. in the Shell Luse No. 1-23 well (fig. 3, site M) and the Shell Grays Harbor No. 1-15 well (figure 3, site O), respectively.

The few recorded earthquakes in the study area are typically quite deep (>30 kilometers, fig. 2), at depths corresponding to the subducted Juan de Fuca slab. Crustal seismicity is sparse in the subduction complex in the coastal region (fig. 2), even though *P*-wave (compressional) velocities of ~4.5-6.0 km/s at 5-15 kilometers depth (Parsons and others, 1999) are within the range where crustal seismicity is observed elsewhere along the subduction margin (McCrory, 2000). A small cluster

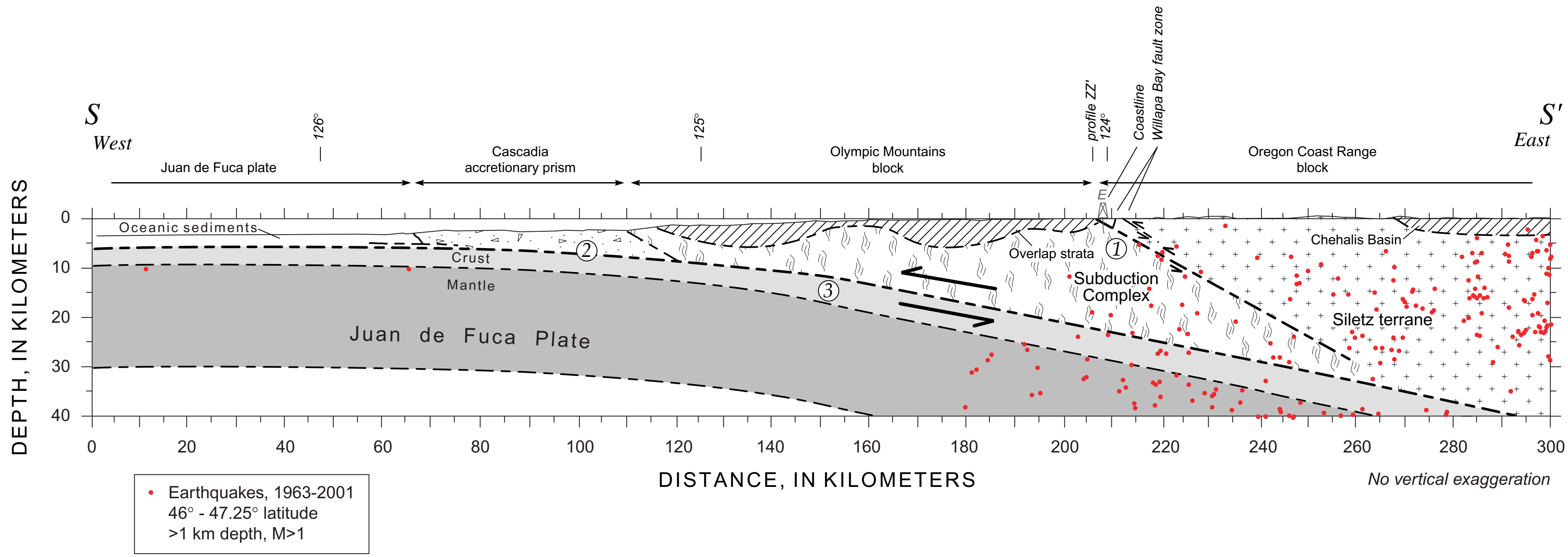
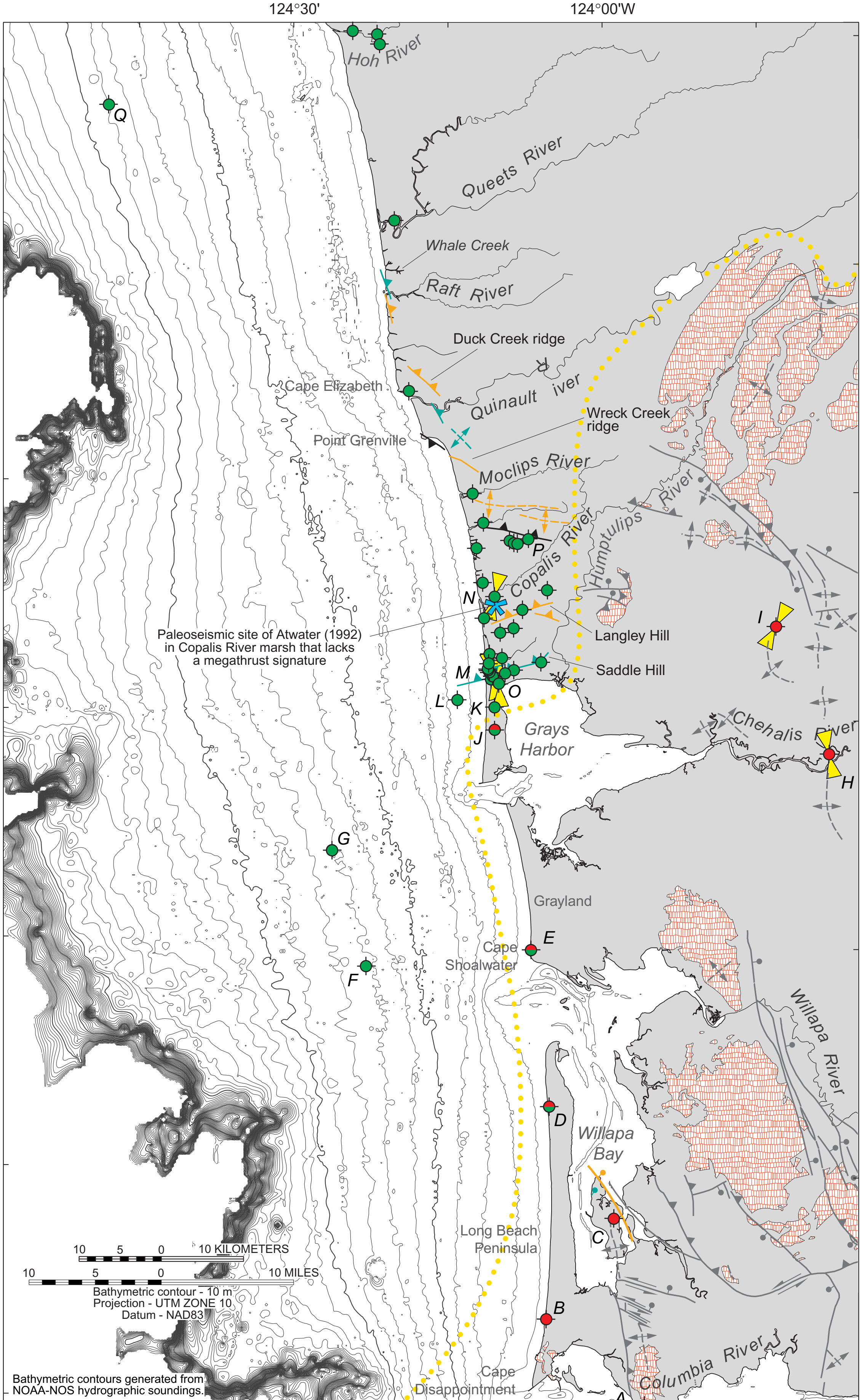


Figure 2. Generalized east-west cross section across the central Cascadia subduction margin showing inferred configuration of major basement units (modified from Parsons and others, 1999) at no vertical exaggeration. See figure 1 for location and explanation of patterns. The location of the Juan de Fuca Plate beneath the margin is constrained by a velocity model of Parsons and others (1999). Paired arrows denote relative movement on Cascadia megathrust and relative movement on block boundary. Circled numbers identify potential seismic sources: (1) upper plate, western Washington crustal sources: (2) interplate, megathrust sources; (3) lower plate, Juan de Fuca sources. Red dots show ambient seismicity 1963-2001 in a 140-km-wide swath centered on the profile line (data derived from a search of the Council of the National Seismic System Earthquake Catalog, online at http://quake.geo.berkeley.edu/cnss/cnss-catalog.html). Much of the ambient seismicity occurs within the subducted oceanic slab, but a small cluster occurs along the boundary fault.


124°30'
124°00'W
47°30'N
47°00'
46°30'
Hoh River
Q
Queets River
Whale Creek
Raft River
Duck Creek ridge
Cape Elizabeth
Quinault River
Point Grenville
Wreck Creek ridge
Moclips River
Humptulips River
Copalis River
P
N
Paleoseismic site of Atwater (1992) in Copalis River marsh that lacks a megathrust signature
Langley Hill
Saddle Hill
I
M
O
L
K
J
Grays Harbor
Chehalis River
H
G
Grayland
E
Cape Shoalwater
F
Willapa River
D
Willapa Bay
C
Long Beach Peninsula
10 5 0 10 KILOMETERS
10 5 0 10 MILES
Bathymetric contour - 10 m
Projection - UTM ZONE 10
Datum - NAD83
B
Bathymetric contours generated from NOAA-NOS hydrographic soundings.
Cape Disappointment
Columbia River
A

of earthquakes does occur along the block boundary beneath Willapa Bay. Given the low to moderate rate of slip postulated for individual faults (<1 mm/yr) in the region, this sparsity of ambient seismicity is not particularly significant.

Present-day displacement inferred from geodetic measurements indicates that the coastal region deviates from the general model of interseismic contraction expected from a locked megathrust. The velocity measured at the Point Grenville site, north of the block boundary, is more northerly than that at coastal sites south of the boundary (Svarc and others, in press). Also, the east-west Grays Harbor transect indicates greater than expected uplift in coastal Washington (Savage and others, 1991; Svarc and others, 1999). These anomalies may result from superimposed effects of deep and shallow strain fields from the megathrust and upper plate structures, respectively.

Temporal Constraints on Quaternary Structural Activity

Cenozoic Stratigraphic Sequence of Coastal Washington

Two major basement units, the Siletz terrane and the Olympic subduction complex, underlie the study area. Both units are separated from overlying Miocene to Pliocene marine strata (Rau, 1970; Wagner and others, 1986; Palmer and Lingley, 1989) by a regional unconformity of late middle Miocene age (fig. 4) (Rau and McFarland, 1982; Kulm and Fowler, 1974; Snavely, 1987; McNeill and others, 2000). Locally,

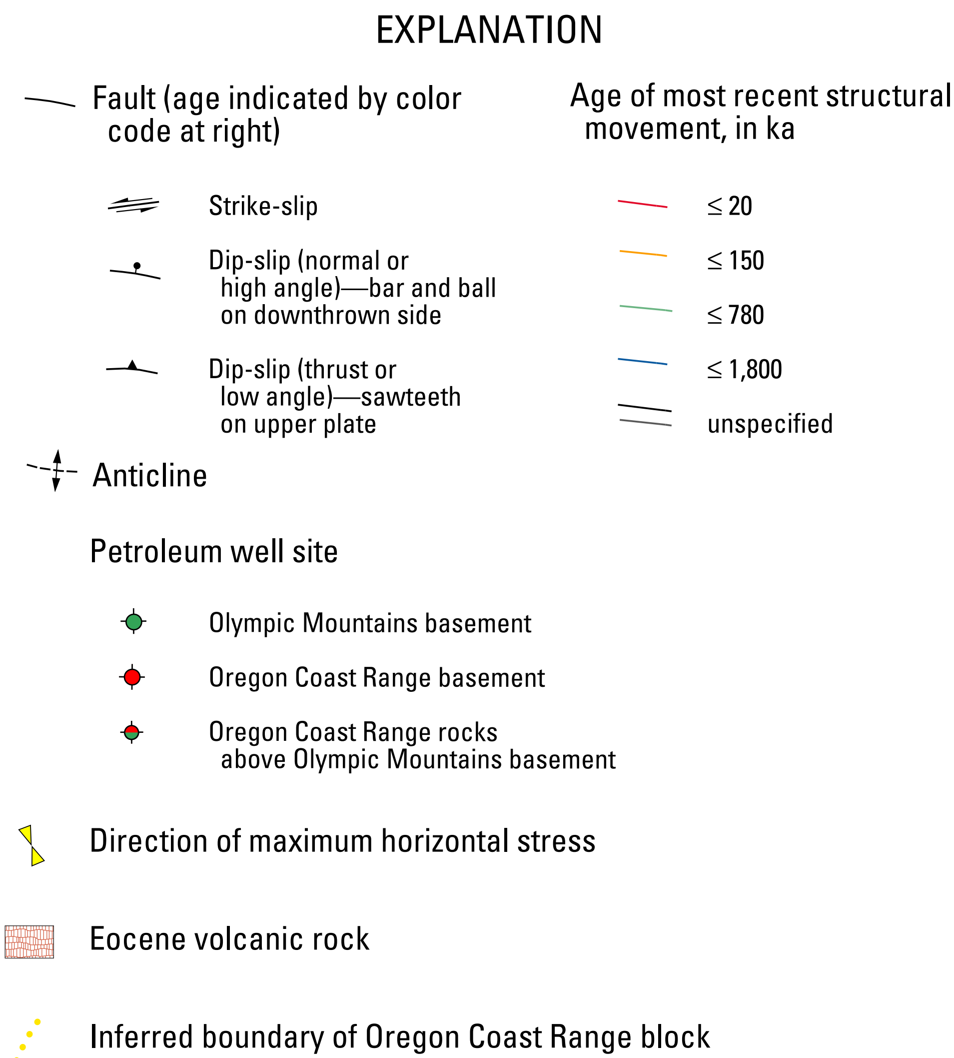


Figure 3. Map showing location of onshore faults and anticlinal folds in coastal Washington (from Walsh and others, 1987; Wells, 1989; McCrory, 1996, 1997) that correspond to those mapped offshore. Location of exploratory petroleum wells from Rau and McFarland (1982), Palmer and Lingley (1989), and Magee and Zoback (1992); stress orientations from Magee and Zoback (1992). Location for Shell Luse No. 1-23 well (site M) has been corrected (moved south) as compared with Magee and Zoback (1992) and McCrory (1996). Site N corresponds to three wells, from south to north: Shell Sampson Johns No. 1-15, Shell Sampson Johns No. 2-15, and Union State No. 3. Note proximity of the Atwater paleoseismic site to the thrust fault just to south.

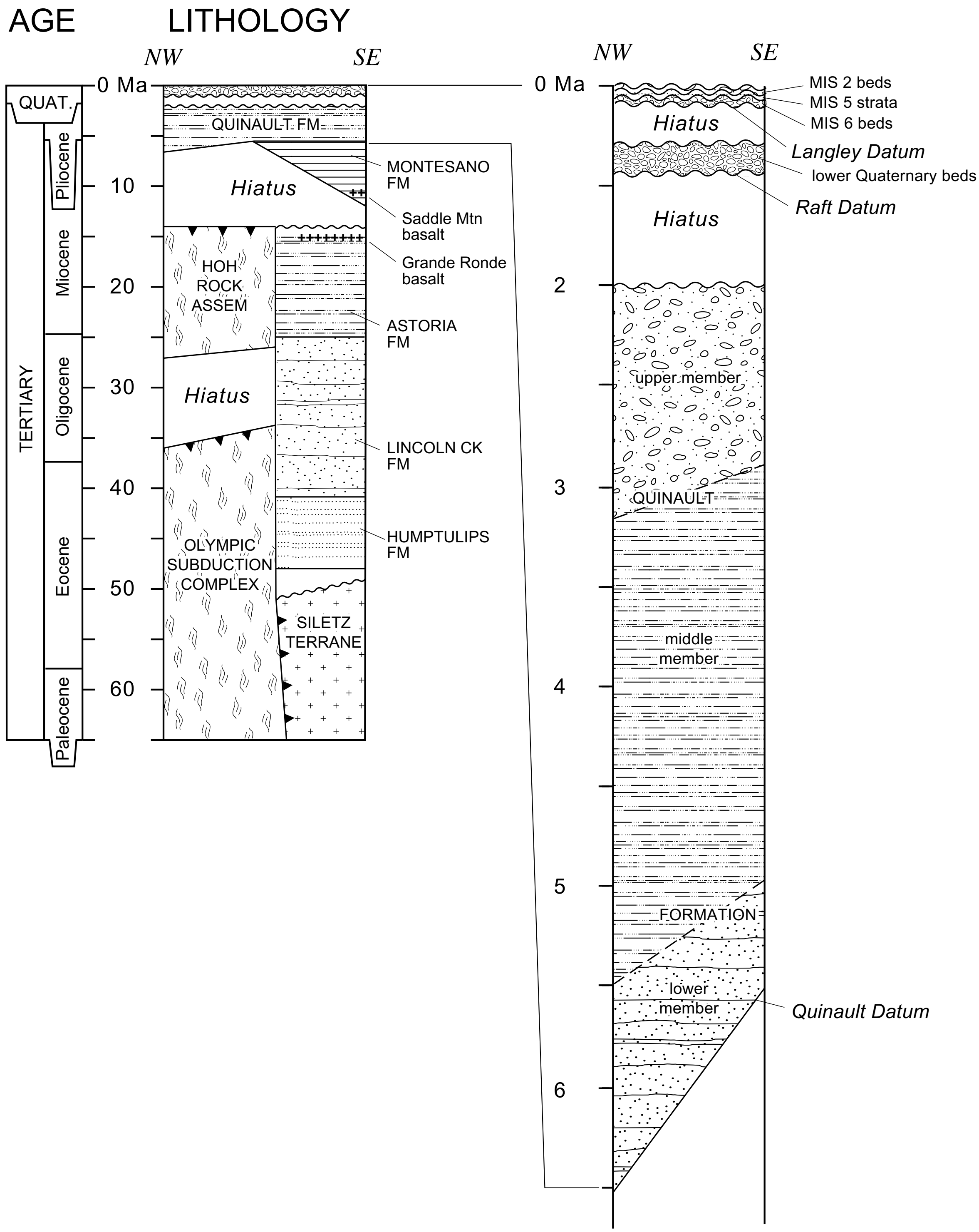


Figure 4. Generalized stratigraphic column showing tectonostratigraphic units in the study area (modified from Palmer and Lingley, 1989), with expanded column of upper part (Pliocene-Quaternary). Northwest edge of column is in the vicinity of the Queets River. Southeast edge is in the vicinity of the Columbia River. Siletz terrane is the principal basement rock for the Oregon Coast Range block. Olympic subduction complex and Hoh rock assemblage are the principal basement rocks for the Olympic Mountains block. Age for the base of the Quinault Formation (ca. 6.5 Ma) is based on biostratigraphy of Rau (1970). Age for the top of the Quinault Formation (ca. 2 Ma) is based on biostratigraphy of Ingle (1967). Dashed division of the Quinault Formation into informal members is schematic and based on lithostratigraphic data of McCrory (unpub. data, 1989). The Quinault datum that separates the Quinault Formation from the underlying Hoh rock assemblage is well exposed at several locations along the coast. The Raft datum that separates the Quinault Formation from overlying lower Pleistocene gravels is poorly exposed at the coast just north of the Raft River. The Langley datum that separates lower Pleistocene gravels from upper Pleistocene gravel or sand deposits is exposed at a number of locations along the topographic ridges discussed in text. See text for descriptions of the stratigraphic units.

these basement rocks and overlap strata are unconformably overlain by shallow marine or nonmarine Quaternary sediments (Grim and Bennett, 1969; Palmer and Lingley, 1989).

Siletz Terrane

Basement rocks in much of coastal and nearshore Oregon are assigned to the Siletz terrane (Wells and others, 1984, Tréhu and others, 1994; Snavely and Wells, 1996), a terrane that accreted to the North America subduction margin by Eocene time, ca. 50 Ma. The Siletz terrane consists of Paleogene oceanic basalt (fig. 5) and overlying forearc strata. The basaltic terrane is quite thick in Oregon and southernmost Washington, where it behaves as a rigid block. The terrane thins northward in the Puget Sound region of Washington, where it is warped and broken into a series of subblocks (Brocher and others, 2001; ten Brink and others, 1999). Overlying Miocene and younger strata show little evidence of deformation away from the subblock boundaries (Johnson and others, 1996, 1999). In coastal Washington, the upper surface of the Siletz basement has relief; however, available data are not sufficient to discern whether this relief is derived from structural activity or, in turn, whether any postulated structures might be active.

Olympic Subduction Complex and Hoh Rock Assemblage

Basement rocks in much of coastal and nearshore Washington are composed of Oligocene to Miocene sedimentary and volcanic rocks scraped off the subducting Juan de Fuca Plate and accreted to the subduction margin by late Miocene time, ca. 10 Ma (Rau, 1973, 1975; Tabor and Cady, 1978; Snavely, 1987). Pervasively deformed rocks of the subduction complex (fig. 6) are divided into two major units onshore, a middle Eocene to lower Oligocene Olympic subduction complex (Tabor and Cady, 1978; Brandon and Vance, 1992) that underlies the Olympic Mountains and the middle Miocene Hoh rock assemblage (fig. 6) (Rau, 1975; Orange and others, 1993) that is exposed in coastal Washington.

The subduction complex is poorly exposed onshore adjacent to the study area, but petroleum wells drilled along the coast north of Grays Harbor penetrated subduction-complex basement beneath middle Miocene to upper Pliocene overlap strata (fig. 3) (Rau and McFarland 1982; Palmer and Lingley, 1989). Equivalent rocks occur on the shelf as inferred from offshore well stratigraphies (Snavely and Wagner, 1982; Rau and McFarland, 1982; Palmer and Lingley, 1989) and correlation with onshore rocks. The subduction complex is inferred to generally young seaward, with the rocks underlying the

Figure 5. Exposure of Siletz basaltic rocks onshore near Grays Harbor, adjacent to the study area.

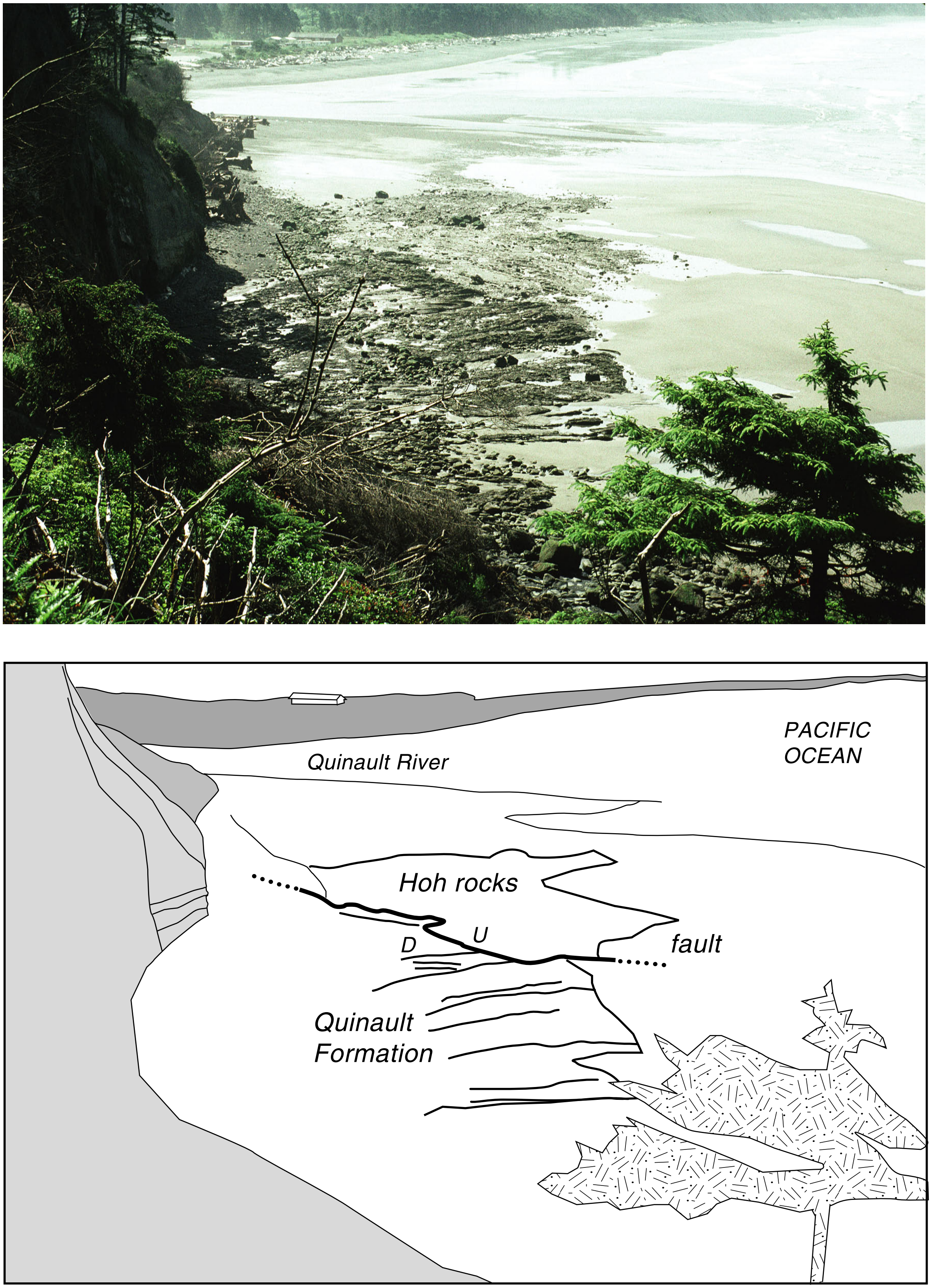


Figure 6. Subduction-complex (Hoh mélange) rocks in fault contact with Quinault strata at the mouth of the Quinault River, near Cape Elizabeth. See plate 1*A* for location and direction of view.

Continental Slope having been accreted during the past 2 m.y. (Barnard, 1978).

Onshore, the older Olympic subduction-complex rocks (Brandon and Vance, 1992) override the younger Hoh rock assemblage along a major thrust system (Snavely and Kvenvolden, 1989). This configuration is observed in Rayonier No. 1-A well in the Grays Harbor area (fig. 3, site P) (Rau and McFarland, 1982), where mélange rocks of late Eocene to Oligocene age overlie mélange rocks dated as early Miocene (on the basis of benthic foraminiferal fauna) at a depth of 1.85 km.

Neogene Overlap Strata

Moderately deformed marine strata of middle Miocene to Pliocene age are divided into two units onshore adjacent to the study area, the middle-upper Miocene Montesano Formation and the upper Miocene-Pliocene Quinault Formation (fig. 4) (Rau and McFarland, 1982; Palmer and Lingley, 1989). These two units are separated by a 15°-20° angular unconformity in the Grays Harbor area (Palmer and Lingley, 1989) based on well data and seismic-reflection profiles. The age of the Montesano Formation is constrained by radiometrically dated volcanic rocks, the Pomona Member of the Saddle Mountain Basalt (ca. 12 Ma), which intrude basal Montesano strata in the Grays Harbor area (Walsh and others, 1987).

The slope to inner shelf lithofacies of the Montesano Formation are poorly exposed onshore (Fowler, 1965). Petroleum well data indicate that the Montesano Formation is as much as 775 m thick (Rau and McFarland, 1982; Palmer and Lingley, 1989). Well and nearshore seismic-reflection profiles indicate that Montesano strata thin markedly northward from Grays Harbor (Rau and McFarland, 1982; Palmer and Lingley, 1989). These strata are not observed in the subsurface north of the Copalis River (fig. 3). The Montesano Formation thins rapidly to the south as well; Montesano strata are not observed south of Shell Minard No. 1-34 well (fig. 3, site K) on the northern spit of Grays Harbor. Strata equivalent to the Montesano Formation are present offshore in the Tidelands State No. 2 well, 3 kilometers southwest of Saddle Hill (fig. 3, site L) (Rau and McFarland, 1982; Palmer and Lingley, 1989). The lithostratigraphy in this well is used to identify and map Montesano-equivalent strata on the shelf (Palmer and Lingley, 1989).

The upper slope to deltaic lithofacies of the Quinault Formation are exposed onshore from Point Grenville north to the Raft River (Horn, 1969; Rau, 1970, 1975) and penetrated in petroleum wells from Long Beach Peninsula north to Cape Elizabeth (Rau and McFarland, 1982). Well data indicate that the Quinault Formation reaches at least 620 m in thickness. Planktonic foraminiferal fauna (Ingle, 1967, 1976) constrain the upper Quinault Formation (Horn, 1969; Rau, 1970), exposed in the sea cliffs between Point Grenville and Cape Elizabeth, to Pliocene age (ca. 5-2 Ma). The Quinault Formation in the Cape Elizabeth area is composed of interbedded conglomerate and sandstone (Horn, 1969; Rau, 1970). The conglomerate beds are highly resistant to erosion and display a distinctive ribbed pattern where they form the shore plat-

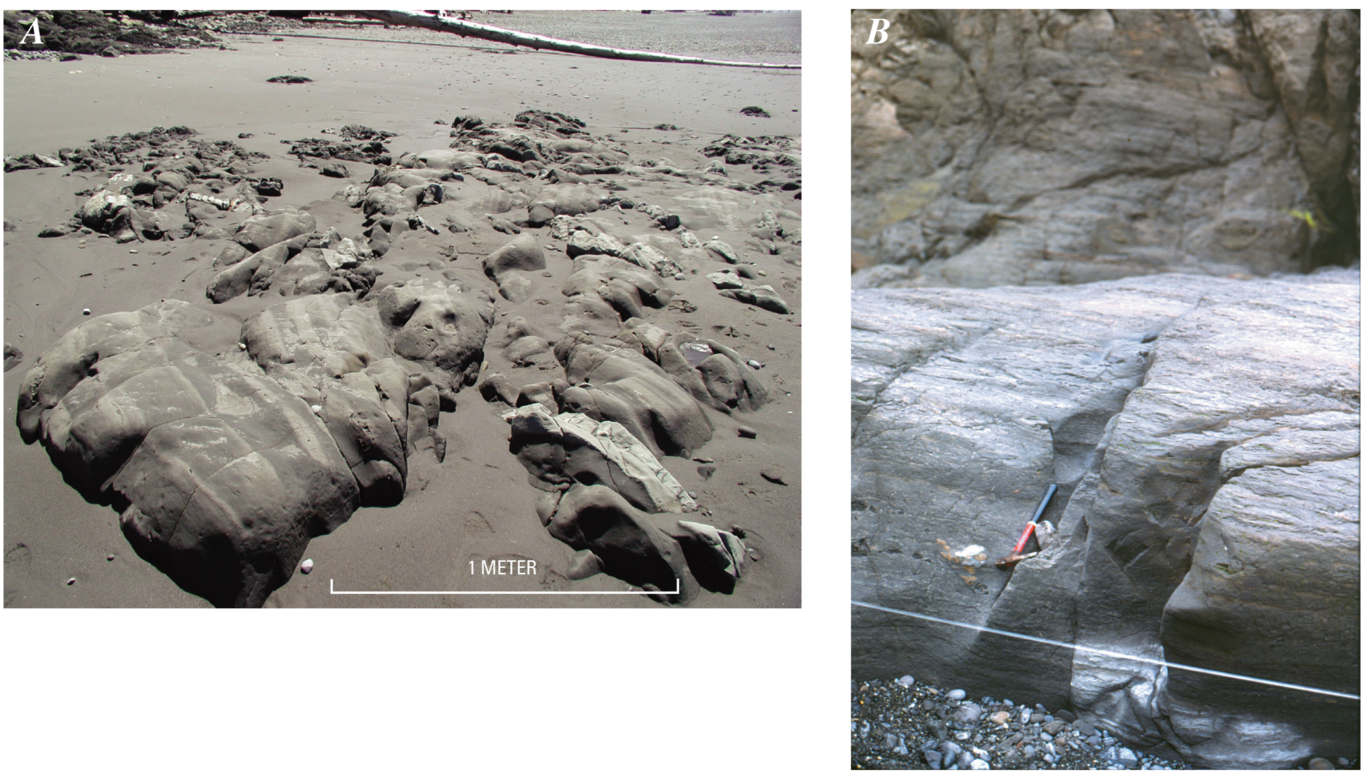
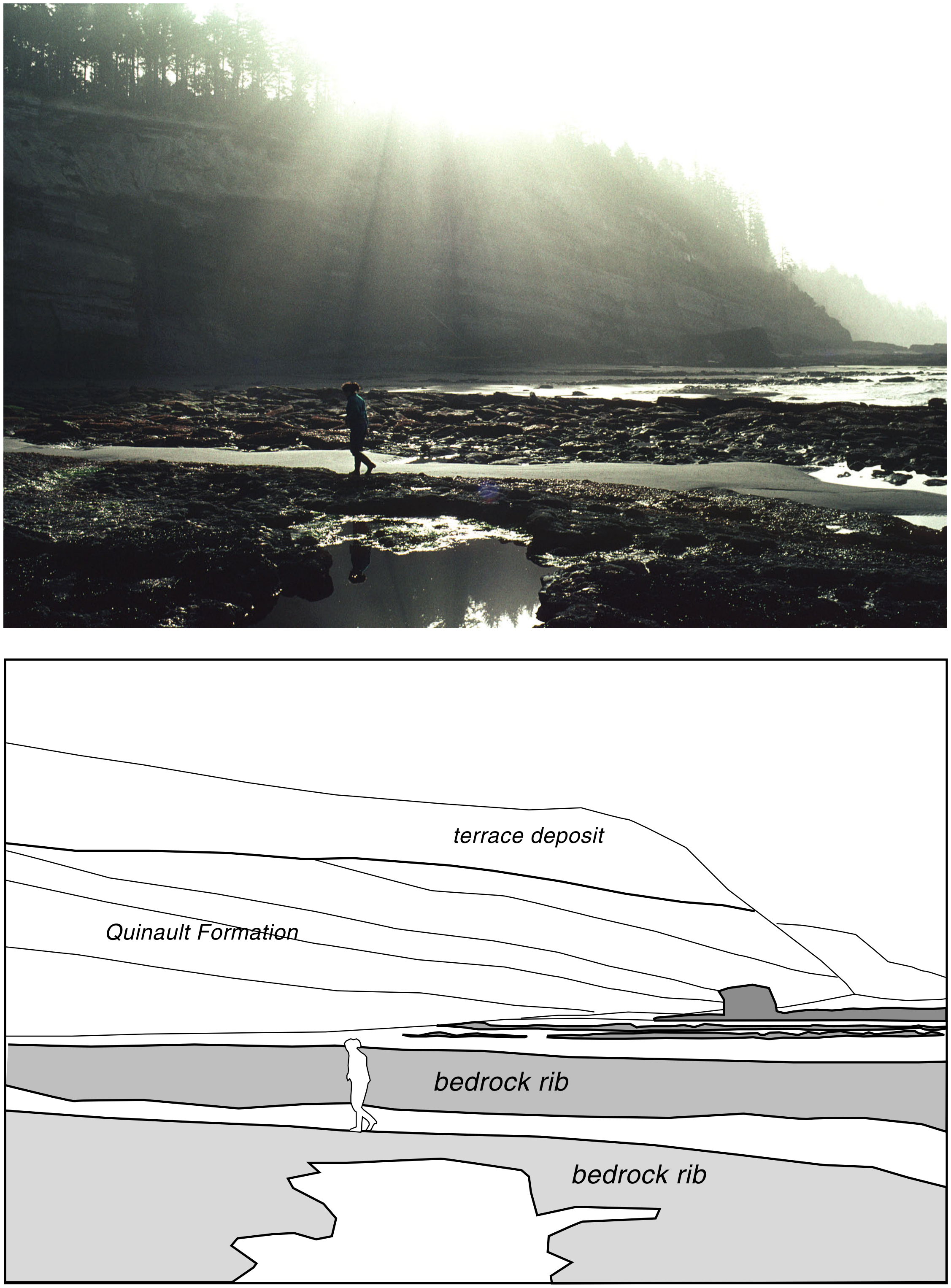


Figure 7. Marine onlap strata onshore adjacent to the study area. See plate 1*A* for locations and directions of views. *A*, Quinault Formation, lower turbidite member, near the Raft River. *B*, Quinault Formation, middle siltstone member, near Duck Creek. *C*, Quinault Formation, upper conglomerate member, near Cape Elizabeth. *D*, Quinault Formation, upper conglomerate member, near Duck Creek. Note the bedrock ribs extending seaward.

C



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D

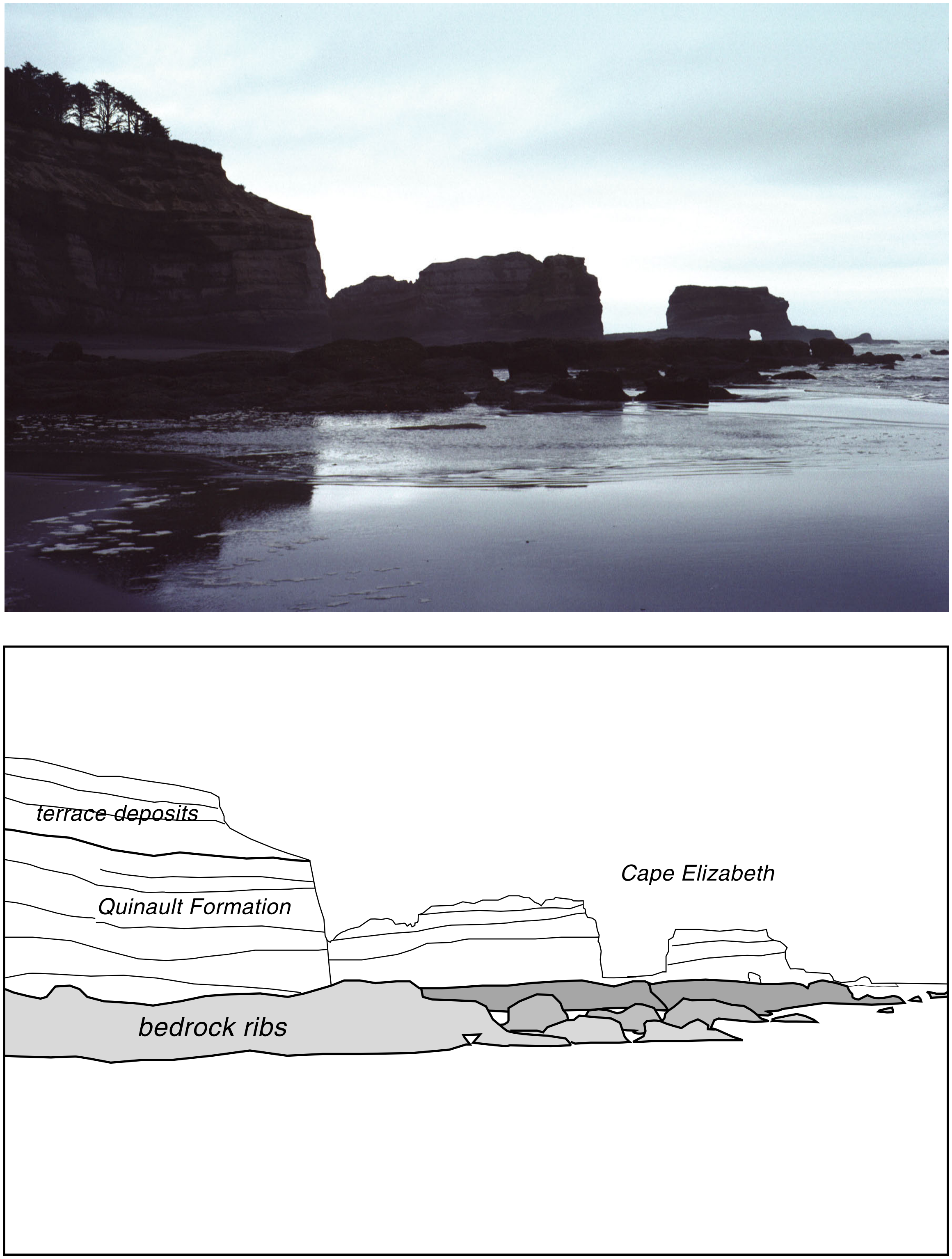


Figure 7.—Continued

form (fig. 7). Folded strata on the inner shelf form ribs that jut from the sea floor in similar fashion and may represent equivalent facies offshore.

Mud diapirs intrude the overlap strata both onshore (Rau and Grocock, 1974; Rau, 1987; Orange and Campbell, 1997) and offshore (Grim and Bennett, 1969; Wagner and others, 1986). The diapirs are commonly found along fault traces that apparently provide conduits for overpressured material in the underlying subduction complex to rise through the overburden. Some diapirs breach the sea floor and appear to be venting fluids into the water column; one forms a small mud volcano on the sea floor (see plate 1*A*).

Quaternary Sediments

Quaternary deposits can be divided into four main units, three glacially derived deposits and one marine-terrace deposit, each bounded by erosional unconformities. A coarse-grained sedimentary unit overlies Quinault strata and underlies upper Pleistocene marine-terrace and glacial deposits onshore between Grays Harbor and Whale Creek. A similar unit, in terms of stratigraphic position, clast composition, and degree of weathering, occurs in the Willapa Bay area. The contact between the Quinault Formation and this gravel unit is poorly exposed in the sea cliffs near the Raft River. Well logs and nearshore seismic-reflection profiles indicate that an angular unconformity (Palmer and Lingley, 1989) separates equivalent strata offshore. Deformation of this unit ranges from mild to strong; strata are flat-lying to severely tilted (fig. 8). This coarse-grained unit has not been dated directly and is tentatively assigned an early Pleistocene age (Walsh and others, 1987; McCrory, 1997) on the basis of its relative position above Pliocene Quinault strata and beneath upper Pleistocene glacial sediments (Moore, 1965).

Onshore exposures of lower Pleistocene strata reveal a range of fluvial and alluvial lithofacies, including poorly sorted gravel lenses, cross-bedded sands and gravels, and minor clay beds (fig. 8) that are interpreted to be glacially derived (Moore, 1965; Pringle, 1986). The distribution of mapped soils that correspond to this unit (Pringle, 1986) suggests that lower Pleistocene deposits are widely distributed between Willapa Bay and Point Grenville. Gravels containing clasts derived from the Olympic Mountains are observed as far south as Long Island in Willapa Bay. Because of its widespread distribution, this unit is tentatively correlated with the major glacio-eustatic lowstand associated with either MIS-22 or MIS-16 (marine-isotope stage), ca. 0.9 or 0.6 Ma (fig. 9), respectively (Shackleton and others, 1990; Williams and others, 1997). This correlation is based on the assumption that a strong, long-lived glacial period was required to develop and preserve the widespread deposit.

Subsurface well data (Rau and McFarland, 1982) and field investigations (McCrory, 1997) indicate that the lower Pleistocene (?) sands and gravels vary widely in thickness onshore, with the unit ranging up to about 200 m thick. An equivalent gravel unit is inferred offshore between Grays Harbor and Cape Elizabeth on the basis of bottom samples (Venkatarathnam, 1969; this paper) and sidescan-sonar data (Twichell and others, 2000; this paper). However, these gravels may in fact represent an amalgamation of several glacio-fluvial units of differing ages deposited during a series of lowstands. In this case, reworking during subsequent sea-level transgressions likely has obscured distinct units on the inner shelf. Nearshore seismic-reflection profiles and well data indicate that lower Pleistocene strata together with overlying upper Pleistocene sediments reach 300 m in thickness (Palmer and Lingley, 1989), with the upper Pleistocene sediments never more than about 40 m thick (Herb, 2000). Offshore, the strata thin and pinch out over structural highs (Palmer and Lingley, 1989).

Two distinct depositional units derived from episodes of alpine glaciation overlie the lower Pleistocene strata with an angular unconformity onshore. These two glacial units mainly consist of till lithofacies bounded by moraines and flanked on the west by outwash lacustrine, fluvial, and alluvial lithofacies (fig. 10) (Moore, 1965) and are locally separated by a marine sand and cobble lithofacies (fig. 11). This intervening marine unit accumulated during late Pleistocene interglacial conditions (MIS-5, ca. 125 to 83 ka; Thackray and Pazzaglia, 1994; McCrory, 1996). An equivalent unit likely occurs offshore between a widespread late Pleistocene erosional surface and a Holocene shell lag found in gravity cores (R. L. Phillips, unpub. data, 1998).

The younger, more restricted glacial unit correlates with the major MIS-2 glacial advance (ca. 20 to 15 ka) on the basis of radiocarbon dates (Moore, 1965). Similar radiocarbon ages have been reported for young glacial deposits in the Queets River area to the north (Thackray and Pazzaglia, 1994). The older, more widespread glacial unit is tentatively assigned to the penultimate major glacial advance (MIS 6, ca. 155 to 150 ka) on the basis of correlation with the global sea-level curve (fig. 9) (Chappell, 1983; Shackleton and others, 1990; Williams and others, 1997). This correlation is again based on the assumption that a strong, long-lived glacial period was required to develop and preserve the widespread deposit.

The outwash facies appear truncated at the coastline, suggesting that they originally extended further west across an emergent shelf. Minor intervening glacial advances are recorded as discontinuous sets of glacially derived deposits along the Quinault River (Moore, 1965).

General Constraints on Offshore Contraction Estimates

Constraints on the location of Quaternary structures and on the recency and magnitude of activity come from four types of geologic and geophysical data. (1) Seismic-reflection profiles allow the mapping of sea-floor offset or warping and the correlation of stratigraphic packages between individual tracklines. The profiles also are used to detect growth faulting or folding and to map the depth and extent of subsurface unconformities. (2) Sidescan-sonar swaths are used to detect sea-floor offsets and to map the orientation of faults and folds expressed on the sea floor. When sea-floor reflective character is matched with bottom samples to determine sea-floor sedi-

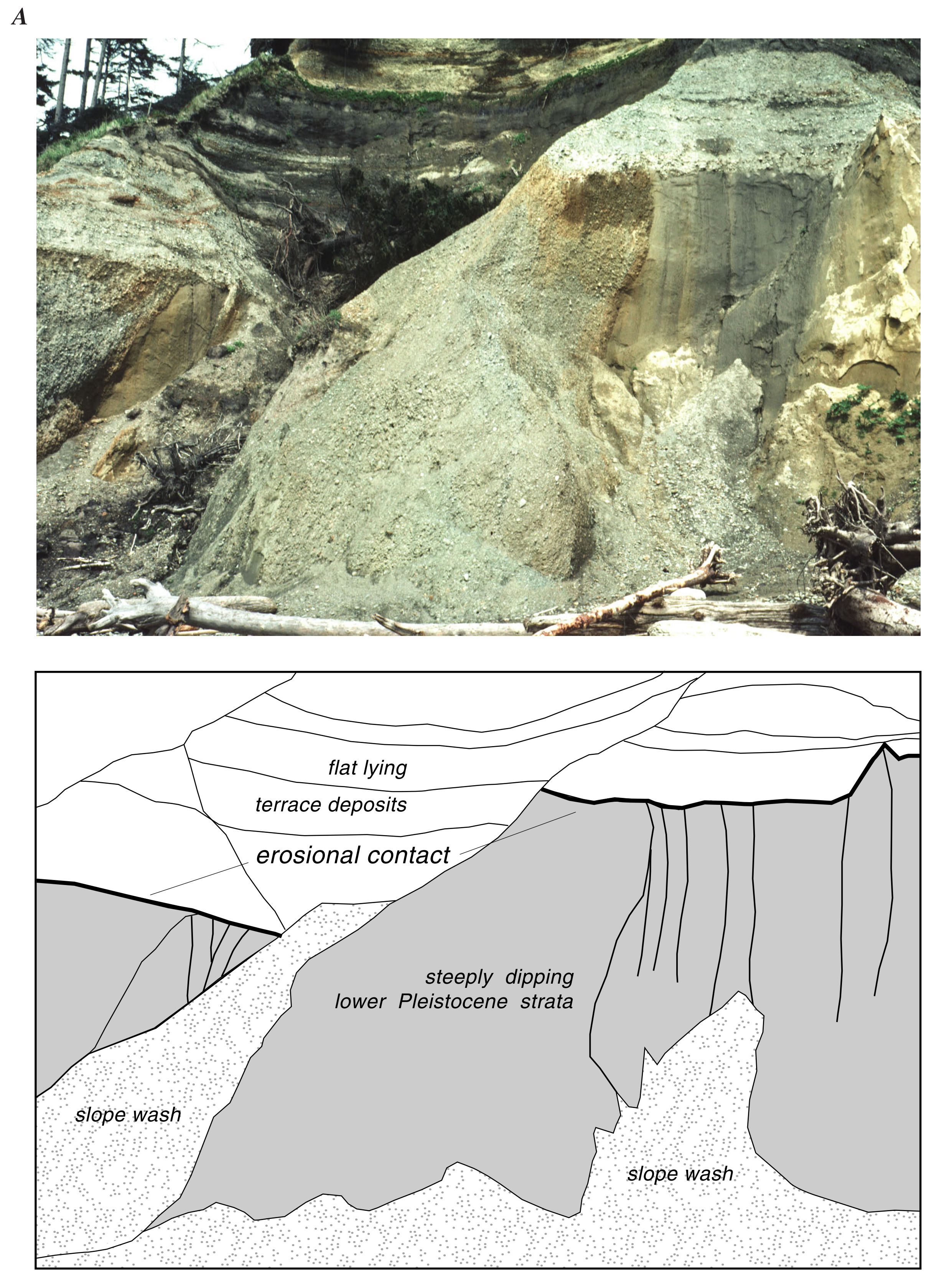


Figure 8. Lower Pleistocene sediments onshore adjacent to the study area. See plate 1A for locations and directions of view. *A*, Tilted lower Pleistocene gravels near Whale Creek overlain by flat-lying MIS-2? gravels. *B*, Tilted lower Pleistocene sand and gravel deposits, near Wreck Creek; exposure is about 3 m high. *C*, Faulted lower Pleistocene gravels on Langley Hill.

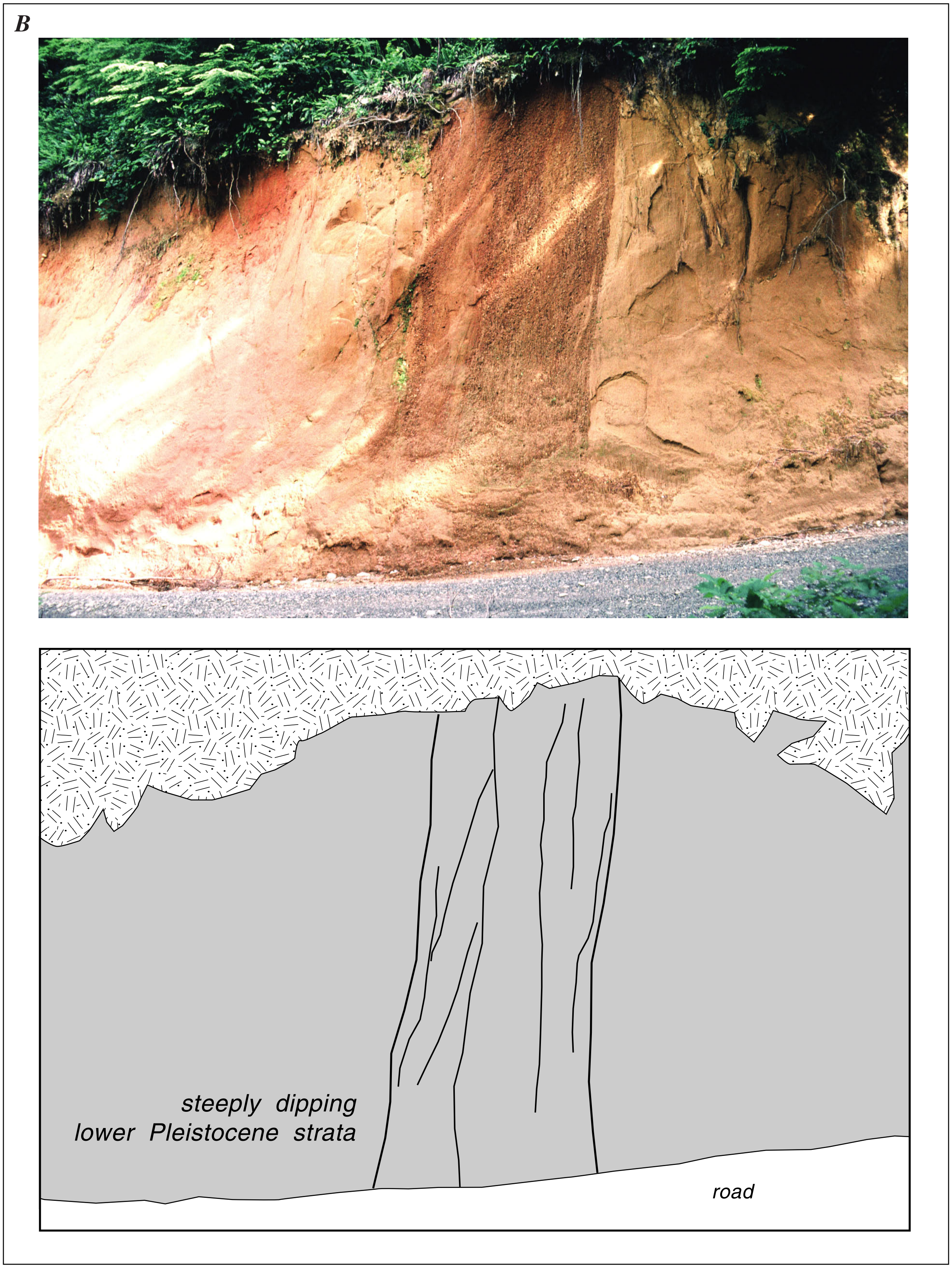


Figure 8. Lower Pleistocene sediments onshore adjacent to the study area. See plate 1A for locations and directions of view. *A*, Tilted lower Pleistocene gravels near Whale Creek overlain by flat-lying MIS-2? gravels. *B*, Tilted lower Pleistocene sand and gravel deposits, near Wreck Creek; exposure is about 3 m high. *C*, Faulted lower Pleistocene gravels on Langley Hill—Continued.

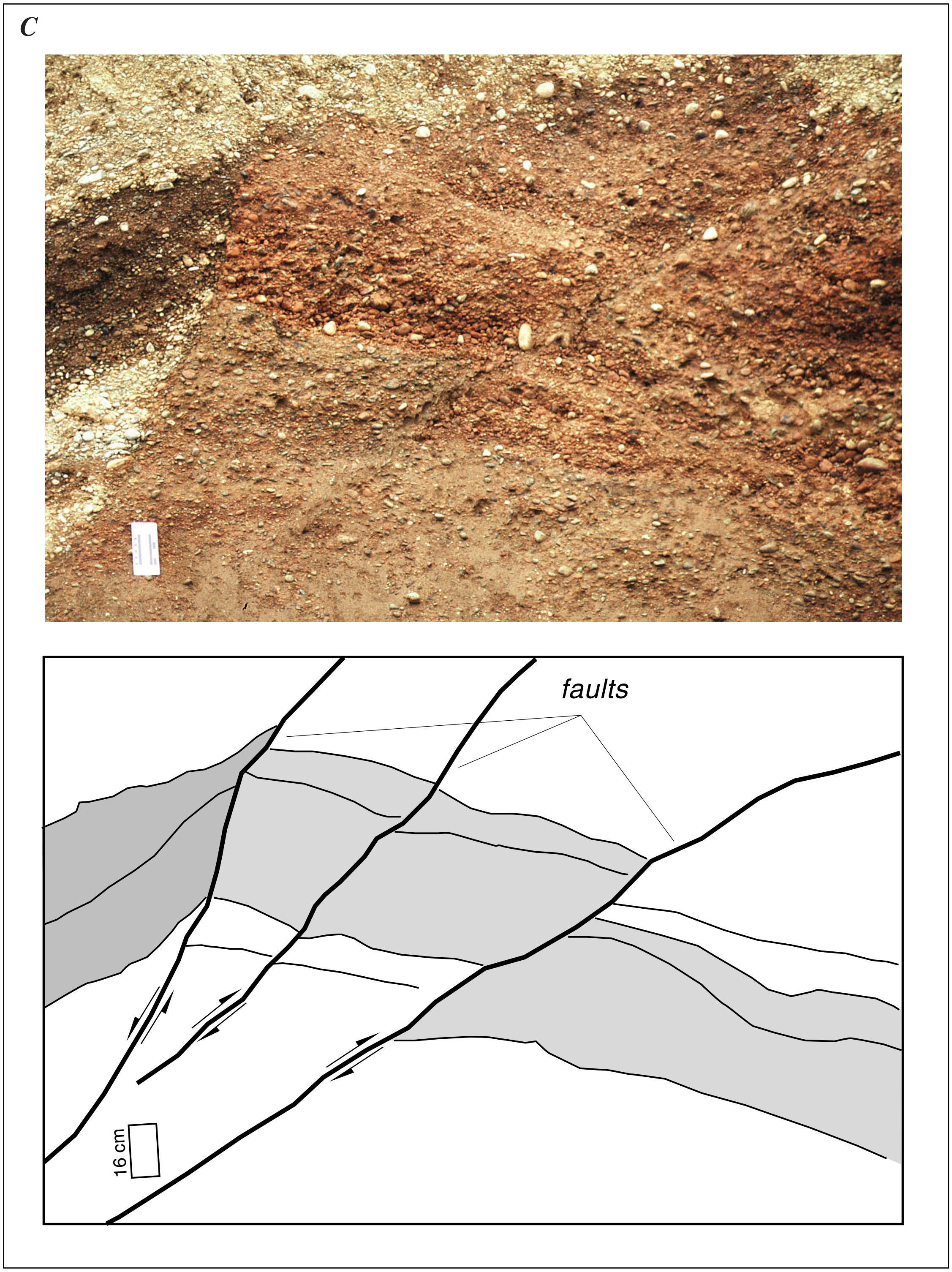


Figure 8.—Continued

ment or rock type, the age of deformed strata can be estimated as well. (3) Well and borehole data are used to determine the age and thickness of stratigraphic units depicted in the seismic-reflection profiles. (4) Dated fossil or wood material and ashes from sediment cores and bottom samples are used to estimate rate of sediment accumulation and age of unconformities depicted in the seismic-reflection profiles. Together these data sets allow the mapping of structures and datums and provide age constraints on the deformed stratigraphic units.

Data Acquisition

The locations of geologic structures mapped on the inner continental shelf between the Columbia River and the Hoh River represent a significant revision to previous mapping efforts (for example, Grim and Bennett, 1969; Wagner and others, 1986; Wolf and others, 1997). This revision is primarily based on new data collected during research cruises in 1997 and 1998 aboard the National Oceanic and Atmospheric Administration ship RV *McArthur* and the Washington Department of Fish and Wildlife vessel RV *Corliss* (fig. 12). The 1997 *McArthur* cruise (M197WO; see appendix 1) collected approximately 400 km of high-resolution seismic-reflection profiles (Foster and others, 2001) and sidescan-sonar swaths (McCrory and others, in press b) in a grid pattern with a 5-km spacing. These data were collected on the Continental Shelf between Grays Harbor and the Raft River in water depths ranging from 20 to 100 m. In addition, the cruise collected seven sea-floor sediment samples in water depths ranging from 28 to 70 m (table 1) along an east-west transect offshore Cape Elizabeth using a Shipek grab sampler. The companion 1997 *Corliss* cruise collected about 500 km of high-resolution seismic-reflection profiles (Cross and others, 1998) and sidescan-sonar swaths (Twichell and others, 2000) in the nearshore area between the Columbia River and Point Grenville, in a similar grid pattern (fig. 12). *Corliss* water depths ranged from 10 to 100 m.

The 1998 *McArthur* cruise (M398WO; see appendix 2) collected about 370 km of high-resolution seismic-reflection profiles (Foster and others, 1999a, b) and sidescan-sonar swaths (McCrory and others, in press a) between Grays Harbor and the Hoh River, in a grid pattern with either a 3-km or 5-km spacing (fig. 12). Water depths ranged from 12 to 78 m. The 1998 *Corliss* cruise collected about 130 km of high-resolution seismic-reflection profiles (Cross and others, 1999) between the Columbia River and Grays Harbor in water depths ranging from 8 to 70 m. The *Corliss* cruise also collected about 160 km of sidescan-sonar swaths (Twichell and others, 2000) in water depths ranging from 8 to 40 m. In addition, 95 bottom grab samples were collected on this cruise along east-west transects in water depths ranging from 8 to 70 m, using a Van Veen grab sampler (fig. 12). A 1998 *Wecoma* cruise (W198WO) collected 15 grab samples (table 2) in water depths ranging from 4 to 172 m (G. Dunhill, unpub. data, 1998) using a Shipek grab sampler (fig. 12) and 10 gravity cores (16-151 cm long; table 3) in water depths ranging from 33 to 132 m (fig. 13). Shell materials from the cores have been radiocarbon dated (R. L. Phillips, unpub. data, 1999) and range in age up to 2.4 ka.

Accuracy in trackline location for the new data sets is excellent, within 10 m. Sidescan-sonar data were collected in 375-m-wide swaths with 19-cm pixel resolution on the *McArthur* and 200-m swaths with 10-cm pixel resolution on the

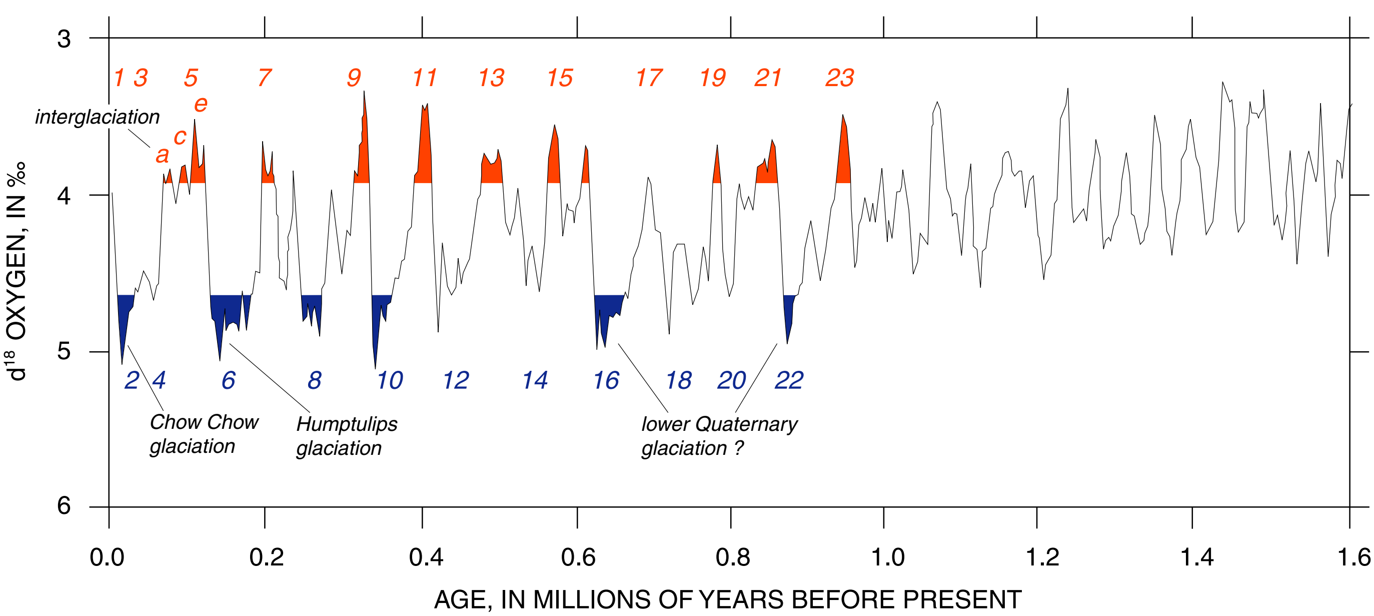


Figure 9. Graph of marine isotopic stages (MIS) (from Shakleton and others, 1990) based on benthic foraminiferal data from ODP Site 677 in the equatorial east Pacific Ocean. Oxygen-isotope data record both ocean temperature and ocean volume and are commonly employed as a proxy for sea-level change. Numbered minima and maxima refer to specific cool and warm marine isotopic stages, respectively; letters refer to specific warm substages during MIS 5. This plot shows the inferred correlation between Pleistocene glacial units and major glacio-eustatic sea-level lowstands (blue). Graph also shows correlation between Pleistocene interglacial marine-terrace units and major MIS-5 sea-level highstands (red).

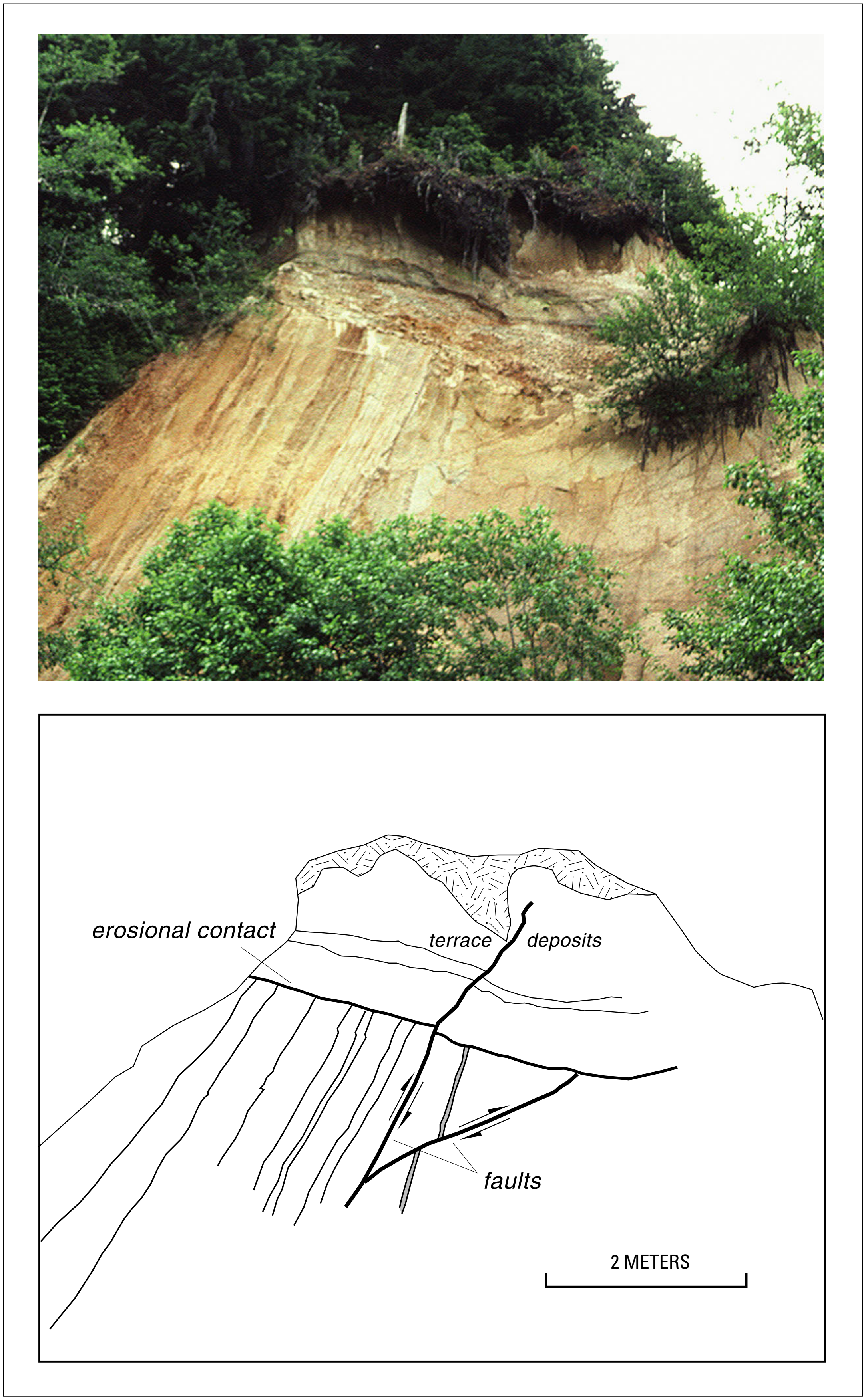


Figure 10. Exposure of MIS-2 glacial gravels overlying tilted lower Pleistocene sands at the Quinault River. Note the fault in lower Pleistocene strata that offsets their upper contact and continues into upper Pleistocene (MIS-2) deposits. See plate 1A for location and direction of view.

A

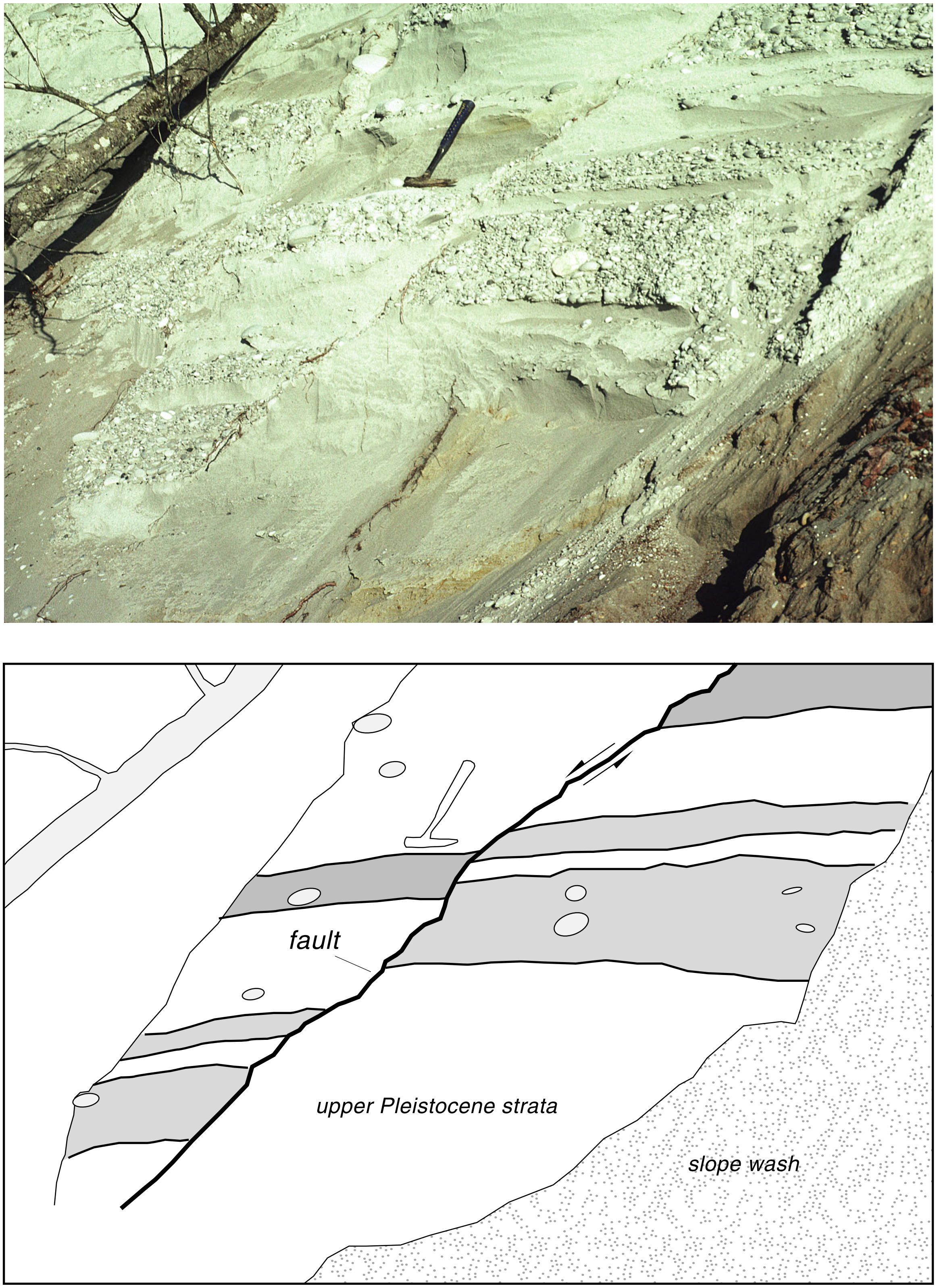
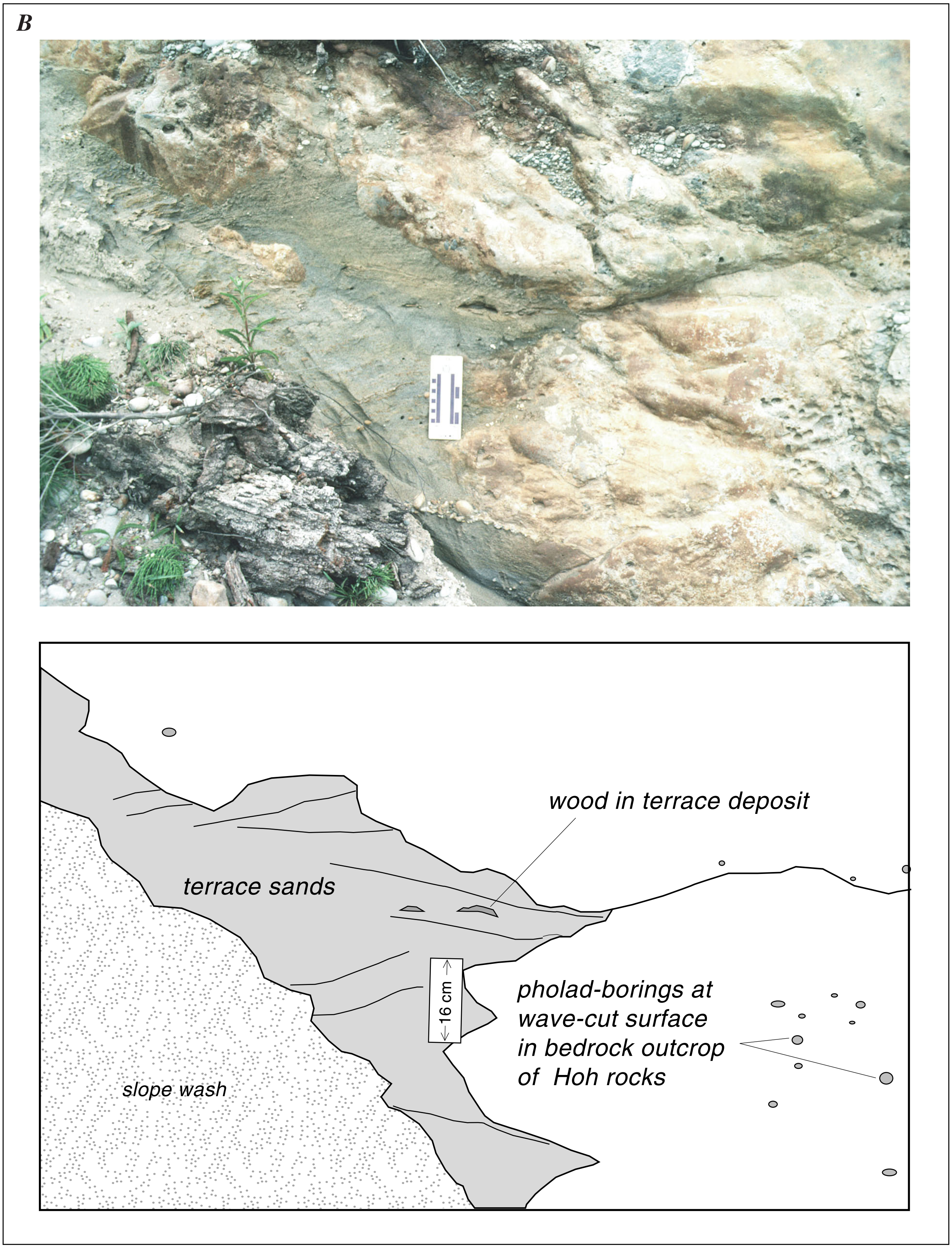


Figure 11. Exposures of MIS-5 marine-terrace sediments onshore adjacent to the study area. See plate 1*A* for locations and directions of view. *A*, Faulted upper Pleistocene marine-terrace deposits near Raft River. *B*, Upper Pleistocene marine-terrace sands and pholad-bored outcrop of Hoh rocks near Hogsback. The borings are diagnostic of an intertidal environment and confirm the marine origin of the terrace platform. *C*, Upper? Pleistocene bioturbated marine muds on Langley ridge. *D*, Faulted upper? Pleistocene estuarine channel sands, near Cape Shoalwater.



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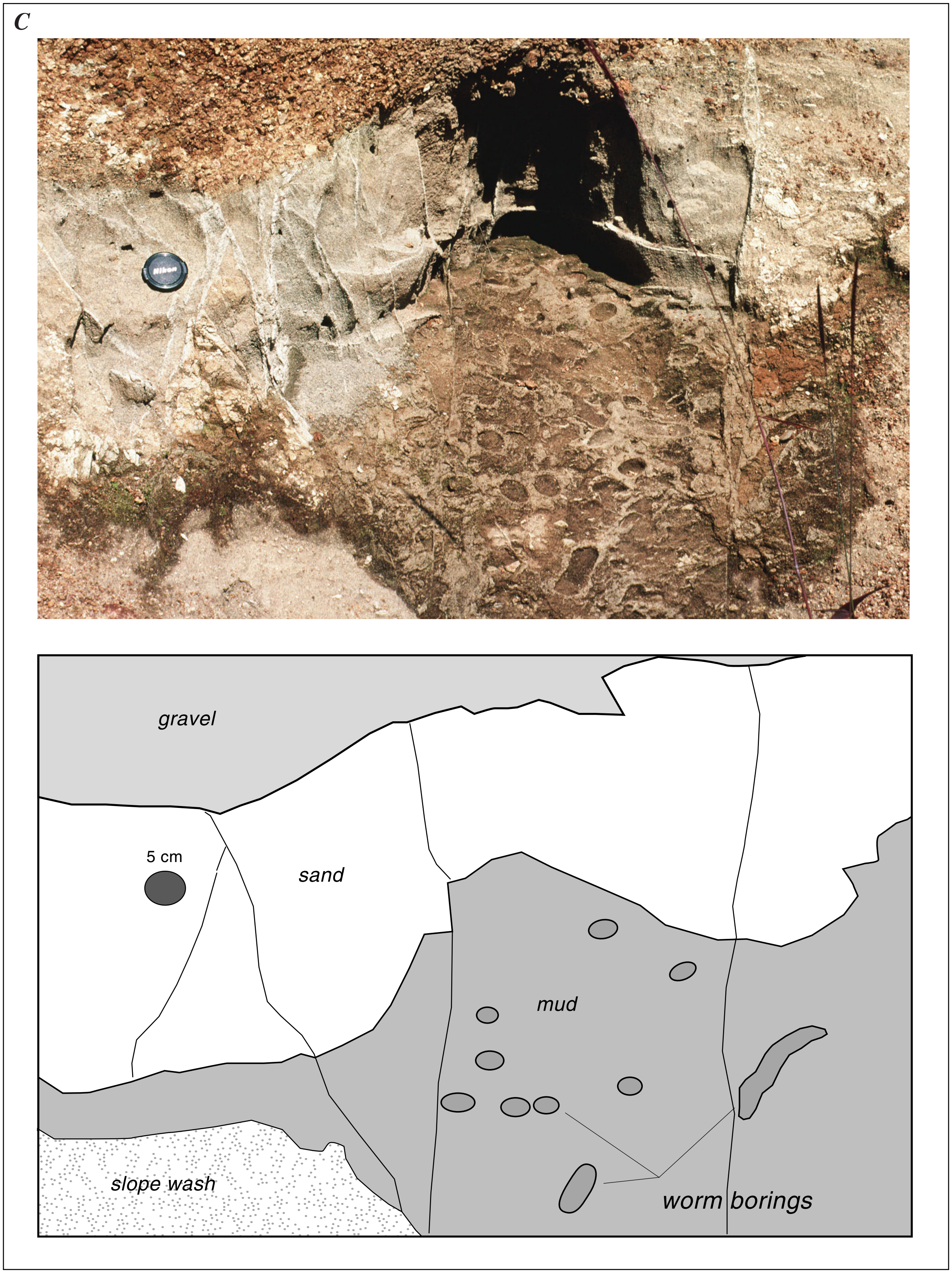
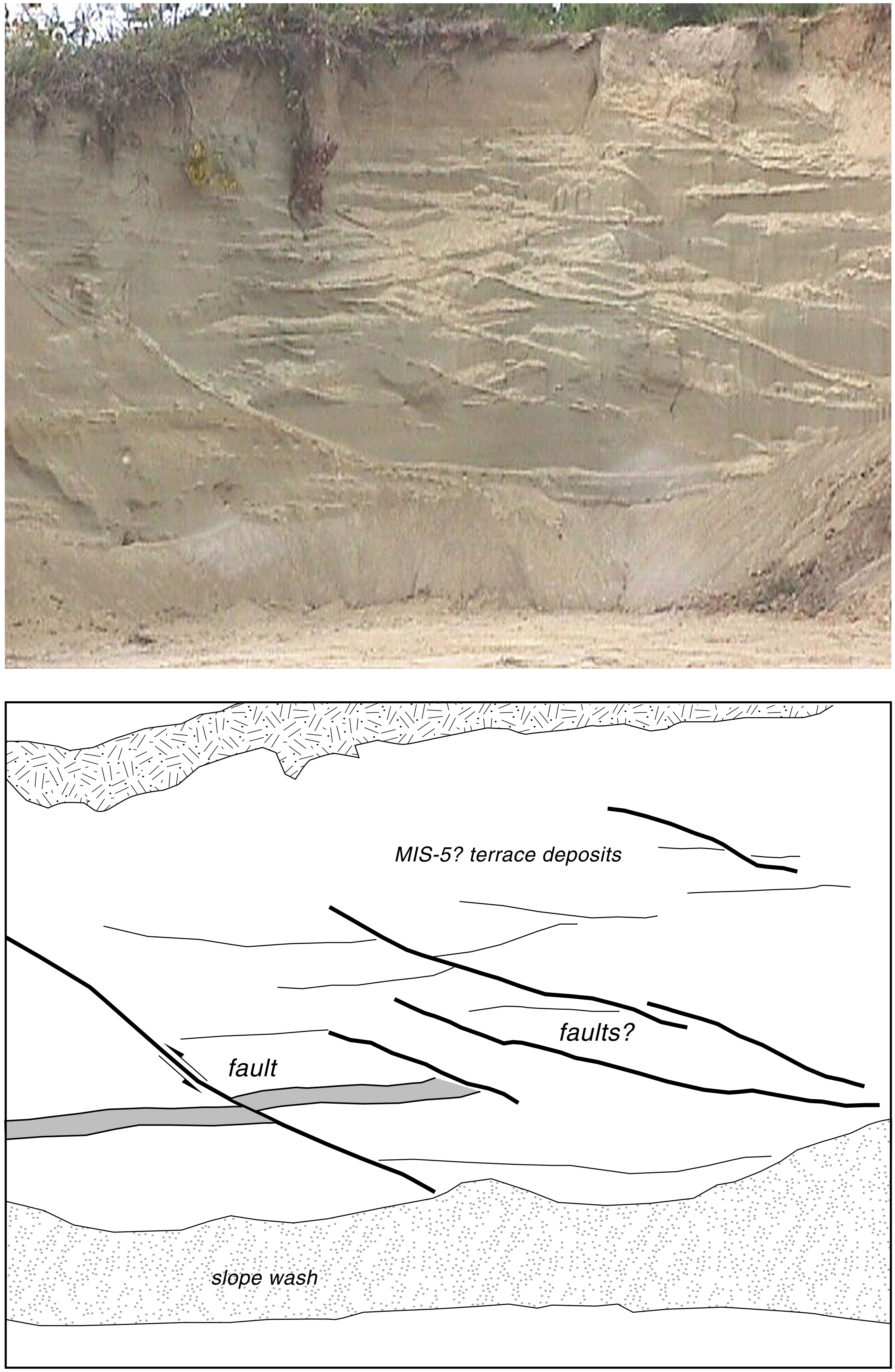


Figure 11. Exposures of MIS-5 marine-terrace sediments onshore adjacent to the study area. See plate 1*A* for locations and directions of view. *A*, Faulted upper Pleistocene marine-terrace deposits near Raft River. *B*, Upper Pleistocene marine-terrace sands and pholad-bored outcrop of Hoh rocks near Hogsback. The borings are diagnostic of an intertidal environment and confirm the marine origin of the terrace platform. *C*, Upper? Pleistocene bioturbated marine muds on Langley ridge. *D*, Faulted upper? Pleistocene estuarine channel sands, near Cape Shoalwater—Continued.

D



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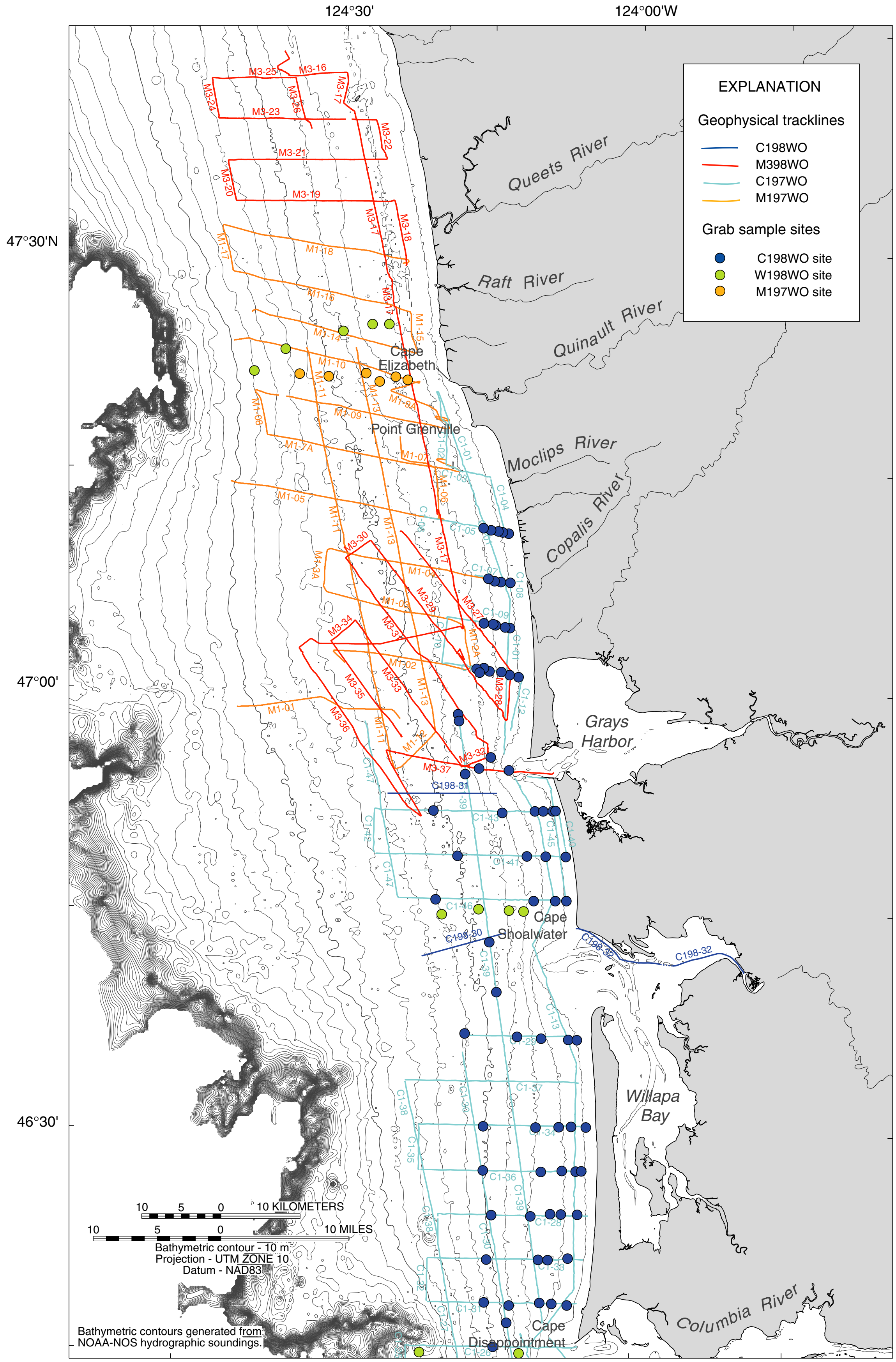


Figure 12. Trackline locations for the 1997 and 1998 RV *McArthur* and 1997 and 1998 RV *Corliss* cruises. Also shown are the locations of grab samples collected during the 1997 *McArthur*, 1998 *Corliss*, and 1998 RV *Wecoma* (W198WO) cruises.

Table 1. Grab-sample log, cruise of RV *McArthur*, June 25-July 4, 1997.
[Sampler: Shipek grab sampler. * means no sample recovered. JD is Julian Day; GMT is Greenwich Mean Time.]

Time JD/GMT	Latitude N.	Longitude W.	Water Depth (m)	Sample Number
191/1115	47° 21.53’	124° 34.01’	70.0	SH-1
191/1149	47° 21.45’	124° 31.18’	51.4	SH-2
191/1218	47° 21.56’	124° 27.36’	39.7	SH-3*
191/1223	47° 21.61’	124° 24.49’	40.2	SH-4
191/1251	47° 21.26’	124° 25.32’	29.3	SH-5*
191/1257	47°’21.27’	124° 25.52’	30.3	SH-6
191/1325	47° 21.44’	124° 23.27’	27.5	SH-7

Table 2. Grab-sample log, cruise of RV *Wecoma*, March 16-30, 1998.
[Sampler: Shipek grab sampler. See URL: walrus.wr.usgs.gov/docs/infobank/ for links to RV *Wecoma* cruise data]

Date/Time (local)	Latitude N.	Longitude W.	Water Depth (m)	Station Number	Sample Number	Description
16 Mar/1930	45° 44.99'	123° 59.95'	40	1	G- 1	Clean sand, with many sand dollars
17 Mar/1249	45° 44.98'	124° 02.97'	65	2	G- 2	Fine- to medium-grained sand; poorly sorted; mafics and quartz
18 Mar/1249	45° 45.02'	124° 08.20'	92	3	G- 3	Fine-grained sand; mafics and quartz
18 Mar/0215	45° 45.01'	124° 13.80'	125	4	G- 4	Silty, fine-grained sand; silt is greenish; subrounded mafics, rounded quartz (more mafics than quartz: 70/30%)
18 Mar/0400	45° 45.10'	124° 28.20'	172	5	G- 5	Clay with >5% quartz
19 Mar/1747	45° 45.01'	124° 07.72'	14	6	G- 6	Medium- to coarse-grained sand; muscovite, biotite, angular olivine, angular quartz, many mafics
19 Mar/1840	46° 15.05'	124° 10.71'	33	7	G- 7	Fine-grained sand; muscovite, biotite, clear mineral?
20 Mar/1244	46° 14.89'	124° 14.06'	65	8	G- 8	Silty sand with abundant organic material (shells, tube worms, etc.)
20 Mar/2153	46° 15.02'	124° 20.66'	133	9	G- 9	No recovery
21 Mar/1030	46° 45.00'	124° 19.62'	62	10	G-10	Silt and fine-grained sand; muscovite, plant material
22 Mar/0351	46° 45.08'	124° 15.24'	50	11	G-11	Silty fine-grained sand; muscovite, many worms
22 Mar/0512	46° 45.02'	124° 12.29'	40	12	G-12	No recovery
22 Mar/0630	46° 45.10'	124° 10.98'	29	13	G-13	Clean fine-grained sand; abundant mica; subrounded mafics, feldspar, quartz
24 Mar/1435	47° 21.67'	124° 42.47'	112	16	G-15	Success on second try; silty clay
24 Mar/1601	47° 23.44'	124° 35.89'	74	17	G-16	100% mud
24 Mar/1803	47° 24.60'	124° 29.96	39	18	G-17	100% gravel
24 Mar/1849	47° 25.21'	124° 27.17'	26	19	G-18	No recovery
24 Mar/1944	47° 25.55'	124° 25.88'	23	20	G-19	No recovery
24 Mar/0954	47° 52.25'	124° 46.09'	58	21	G-20	Medium- to fine-grained sand; quartz, mafics, small clams and organic debris

Table 3. Core log, cruise of RV *Wecoma*, March 16-30, 1998.
[Sampler: gravity corer, weight 973 lbs. See URL: walrus.wr.usgs.gov/docs/infobank/ for links to RV *Wecoma* cruise data]

Date/Time (local)	Latitude N.	Longitude W.	Water Depth (m)	Core Length (m)	Station Number	Sample Number
20 Mar/1600	46° 14.92'	124° 14.02'	65	1.35	8	C-1
20 Mar/2126	46° 15.02'	124° 20.66'	33	0.16	9	C-2
22 Mar/0925	46° 45.00'	124° 25.40'	84	1.51	14	C-3
23 Mar/1458	46° 44.96'	124° 39.90'	132	1.15	15	C-4
29 Mar/0719	46° 59.94'	124° 32.01'	75	0.47	22	C-5
29 Mar/0954	46° 49.98'	124° 35.96'	106	0.96	23	C-6
29 Mar/1208	46° 49.93'	124° 25.89'	75	0.8	24	C-7
29 Mar/1509	46° 35.81'	124° 22.14'	81	0.95	25	C-8
29 Mar/1756	46° 24.01'	124° 17.04'	63	0.61	26	C-9
29 Mar/2047	46° 08.02'	124° 12.99'	85	0.52	27	C-10

Corliss. These data have been integrated as a mosaic at 2-m pixel resolution (P.A. McCrory and W.W. Danforth, unpub. data, 1999). The boomer and sparker data can resolve features as small as 3 m within the upper 200 m beneath the sea floor. Resolution of the accompanying bathymetric or chirp data is less than 1 m with as much as 15 m of penetration.

These new data are supplemented with a sparse set of single-channel seismic-reflection data collected during four USGS and one University of Washington research cruises between 1967 and 1976 (fig. 14; see appendix 3). The quality of these older data range from good to poor owing to variations in the acoustic sources and receivers and the sea state. Accuracy in trackline location for these earlier cruises was within 500 m.

Sea-Floor Exposure of Fault Traces and Fold Axes

The seismic-reflection data provide an image of the subsurface geometry of faults and folds offshore in the upper 100-500 m of strata. The grid spacing allows reconnaissance coverage of shallow features and, in general, is adequate to map structures from one track to the next. *McArthur* line 98-27 crosses within 0.7 km of Tidelands State No. 2 well (fig. 3, site L; fig. 15*A*) and is used to correlate between onshore and offshore strata. *Corliss* line 97-47 crosses within 0.5 kilometer of P-0155-1 well (fig. 3, site G); *Corliss* line 97-43 crosses within 1.7 km and *McArthur* line 98-36 crosses within 3.5 km of this well. All three lines are used to identify the early Pleistocene Raft datum (fig. 4). The sidescan-sonar data depict the orientation of faults and folds exposed on the sea floor (fig. 15*B*). These data proved essential for constraining the strike and length of emergent structures. Faults on the Continental Shelf are generally considered active if they offset the sea floor, as sea-level rise during the past 20 k.y. should have eroded any prior fault scarps flat. However, this designation proves ambiguous in areas where the surficial strata are older than Quaternary and the possibility of relict pre-Quaternary morphology exists.

Sediment recovered from the grab samples and gravity cores is used both to verify the interpretation of sidescan-sonar facies (Twichell and others, 2000) and to provide sedimentation rates for surficial deposits (R. L. Phillips, unpub. data, 2000). The sea floor displays a typical westward progression of beach, nearshore, and mid-shelf facies except for the structural high beneath the inner shelf north of Grays Harbor. This structural high is emergent on the sea floor at its northern end, where Neogene and lower Pleistocene strata are exposed, and is covered by a thin gravel layer at its southern end.

Correlation of Subsurface Folded and Faulted Strata

The orientation of geologic structures mapped in the study area varies considerably (Wagner and others, 1986; this study). These structures deform strata of inferred Pliocene and Pleistocene age (Grim and Bennett, 1969; Nittrouer, 1978; Wagner and others, 1986). The new marine data image a series of faults and associated anticlinal folds separated by broad synclinal folds. Structures in the central portion of the study area trend east-northeast across the shelf and project onshore between Grays Harbor and Cape Elizabeth. In general, strata can be correlated between folds by matching the character and pattern of the seismic reflectors. Palmer and Lingley (1989) used formation thickness logged in exploratory petroleum wells offshore (fig. 3, sites F, G, L, and Q) and several coastal wells onshore (Rau and McFarland, 1982) to correlate with seismic reflectors and thus identify and map offshore stratigraphic units. Many of the strata exposed on the sea floor and in the shallow subsurface in the axes of the anticlines are likely equivalent to the Pliocene Quinault strata exposed onshore between Cape Elizabeth and Point Grenville (figs. 7, 15*B*), and present in the subsurface south of Point Grenville. Strata exposed on the sea floor and in the upper subsurface of synclinal axes include younger units, likely equivalent to the upper Quaternary fill in the stream valleys onshore.

The control points on plate 1*A* represent locations where deformed strata have been identified from seismic-reflection data, largely from high-resolution boomer or sparker profiles, or sidescan-sonar data. These deformation features are principally offset reflectors interpreted to result from shallow faulting or tilted and folded reflectors interpreted to result from crustal tectonic processes. The high vertical exaggeration of the high-

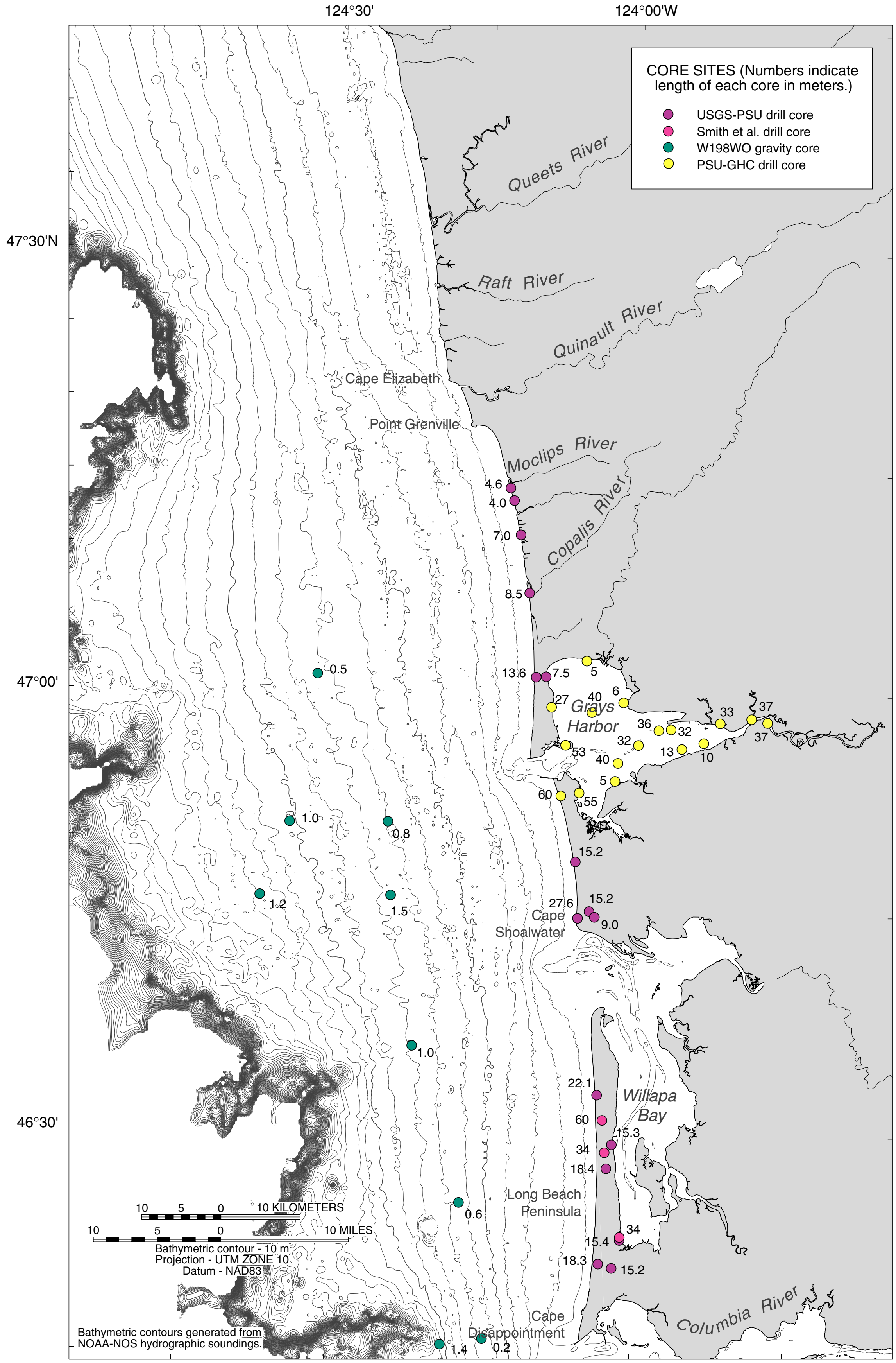


Figure 13. Locations of drill-core sites in the Willapa Bay and Grays Harbor areas (from Peterson and Phipps, 1992; Smith and others, 1999; and Herb, 2000) and 1998 RV *Wecoma* (W198WO) gravity-core sites (G. Dunhill, unpub. data, 1998) on the mid shelf.

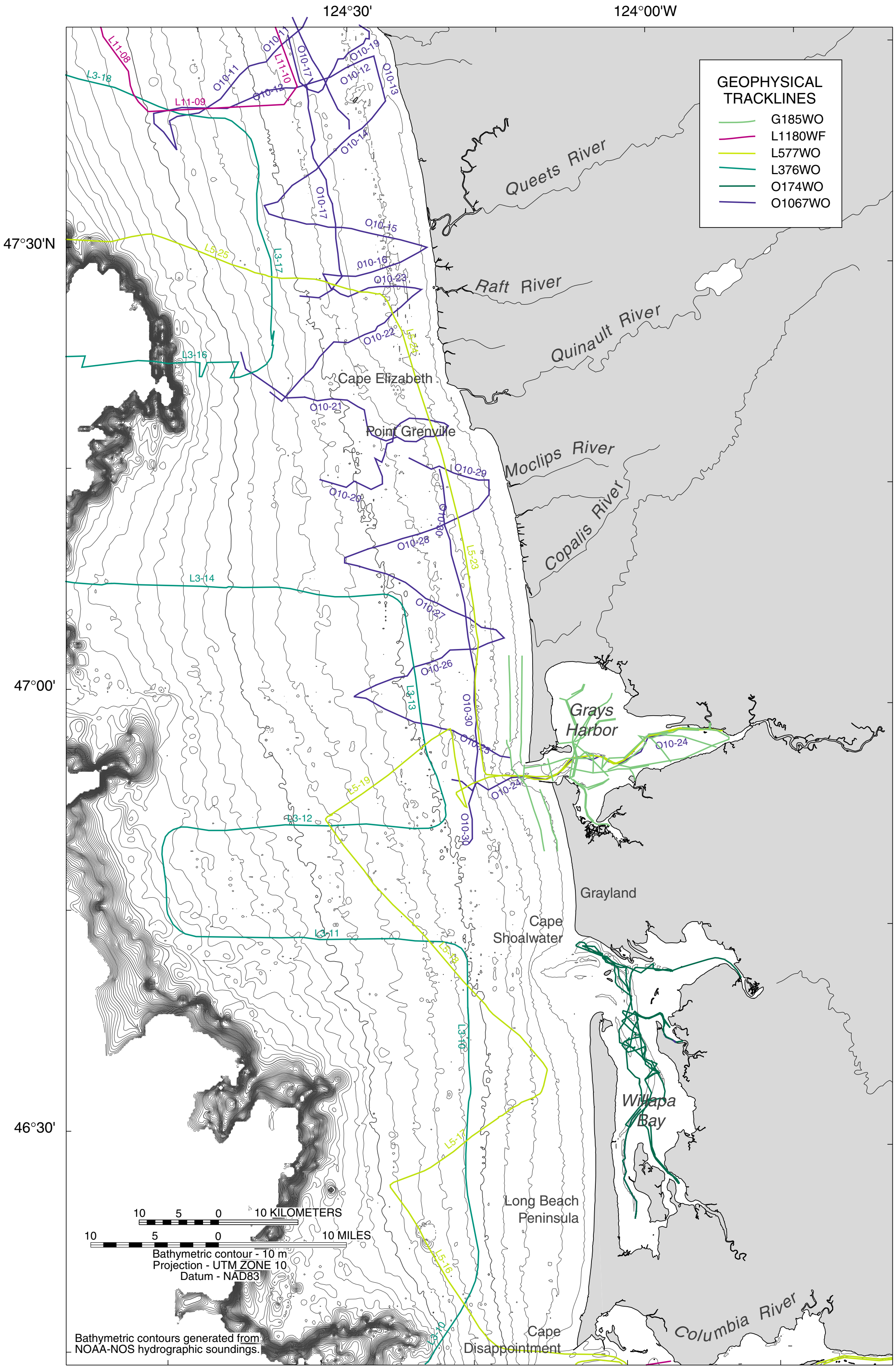
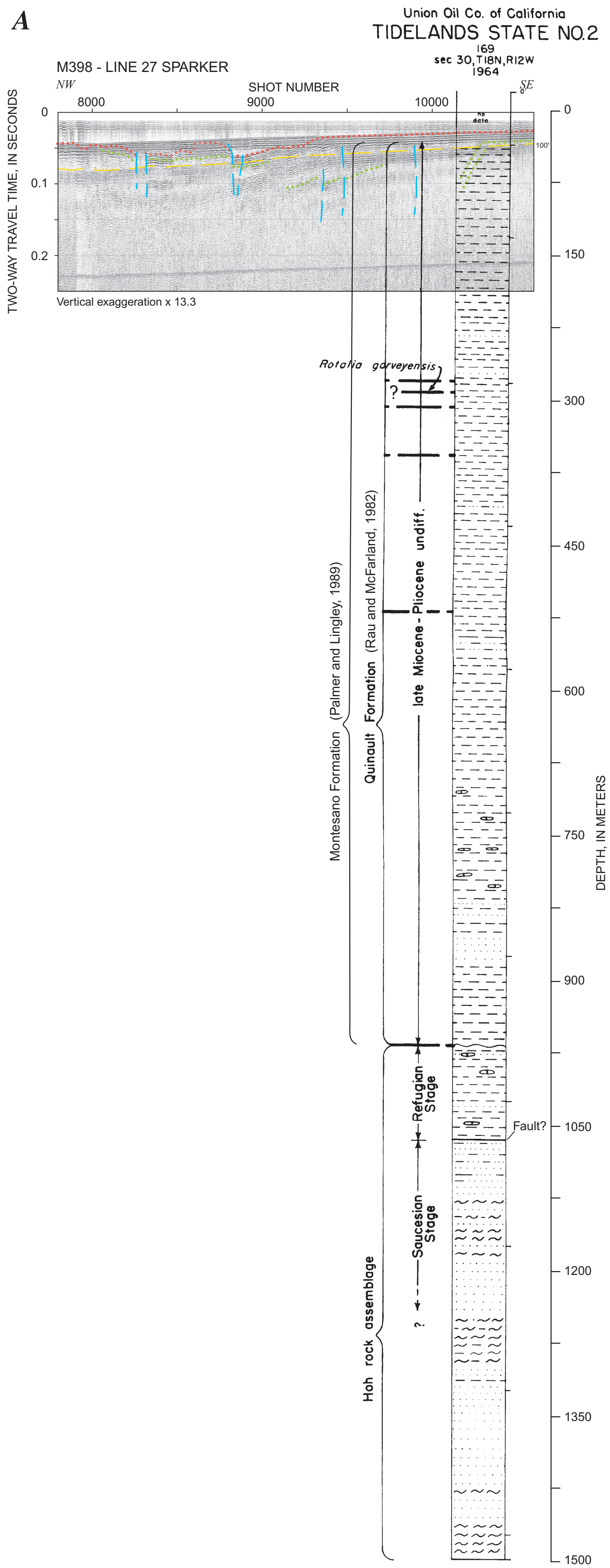


Figure 14. Trackline locations for the 1967 to 1976 research cruises that collected seismic-reflection data used in this study. See appendix 3 for description of cruise data. Subsurface information in Grays Harbor used in this study was derived from the 1995 research cruise G185WO (see Peterson and Phipps, 1992).



resolution profiles makes it difficult to determine fault dips with any accuracy. In general, all fault dips greater than 30° appear as vertical structures. Observation of deformation is also hampered by sea-floor multiples overprinting the seismic reflectors. This problem is most evident in shallow water, such as west of Long Beach Peninsula, or where the sea floor is highly reflective, such as in Willapa River (underlain by the Crescent basalt).

Deformation of Quaternary Unconformities

The seismic-reflection data are adequate for identifying and mapping the erosional surface at the base of inferred upper Pleistocene sediments offshore (fig. 15*C*). We contoured the highest widespread unconformity beneath the sea floor using present sea level as the reference datum. Seismic velocity was converted to depth using an average speed of 1,500 m/s (two-way travel time). The contour map depicts a gently dipping surface beneath the inner shelf (plate 1*B*) that has been modified by erosional and tectonic processes. Between Willapa Bay and Grays Harbor, this surface forms a flat bench beneath the mid-shelf at a depth of 70-85 m. The older data (1976 and 1977 Lee cruises) depict a deeper bench at a depth of 115-125 m beneath the outer shelf to the west, separated from the mid-shelf bench by a linear scarp (plate 1*B*). This morphology is similar to marine terraces onshore that form benches cut during sequential glacio-eustatic highstands. We infer that the offshore benches represent similar erosional platforms cut during glacio-eustatic lowstands.

From Grays Harbor past the northern end of the study area (Hoh River), the erosional unconformity has an irregular surface morphology that extends from near the present coastline out to a depth of about 70 m (Wolf and others, 1997; this study). Available seismic-reflection profiles and sidescan-sonar images depict wide areas of rocky outcrop throughout this region. The lack of a Holocene sediment cover may result from recent uplift (Palmer and Lingley, 1989), as this region has undergone intense folding and faulting compared with adjacent areas to the west and south. Filled paleochannels cross the region of rocky outcrop (plate 1*B*), suggesting that streams have cut through it during sea-level lowstands.

The unconformity changes character both east to west

Figure 15. Examples of seismic-reflection and sidescan-sonar data used to map structures offshore Washington. Green dotted lines denote stratigraphic reflectors, red dotted line denotes late Pleistocene unconformity surface, blue dashed lines denote inferred faults, yellow dashed line denotes first sea-floor multiple. See plate 1*A* for locations of the images. *A*, Tidelands State No. 2 well stratigraphy (see fig. 3, site L) (from Rau and McFarland, 1982) is superimposed on a *McArthur* seismic-reflection offshore profile that crosses within 700 m of the well. Plot depicts correlation between seismic stratigraphy and lithofacies. The Quinault Formation designation of Rau and McFarland (1982) for this well was reinterpreted by Palmer and Lingley (1989), who separate the Quinault Formation into Montesano and Quinault Formations at an observed angular unconformity. Palmer and Lingley (1989) assign the overlap strata in this well to the Montesano Formation only. See figure 18A for a larger version of the profile. *B*, *Corliss* seismic-reflection profile and corresponding sidescan-sonar image (375-m swath width) that depict the rugged relief of strata equivalent to the upper Quinault lithofacies onshore. *C*, *McArthur* seismic-reflection profile that shows inferred lowstand and highstand deposits and bounding unconformities within inferred upper Quaternary strata.

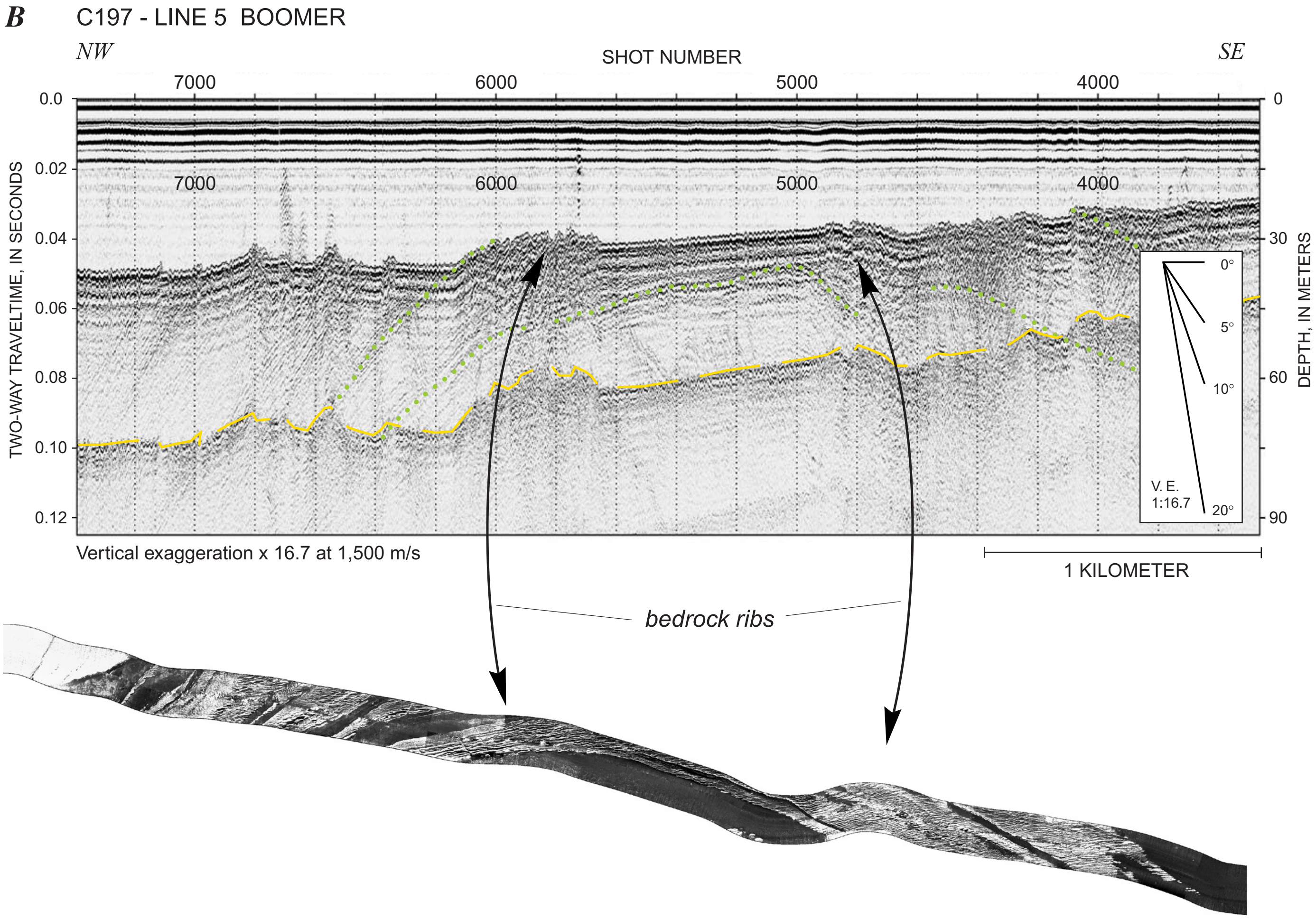


Figure 15. Examples of seismic-reflection and sidescan-sonar data used to map structures offshore Washington. Green dotted lines denote stratigraphic reflectors, red dotted line denotes late Pleistocene unconformity surface, blue dashed lines denote inferred faults, yellow dashed line denotes first sea-floor multiple. See plate 1*A* for locations of the images. *A*, Tidelands State No. 2 well stratigraphy (see fig. 3, site L) (from Rau and McFarland, 1982) is superimposed on a *McArthur* seismic-reflection offshore profile that crosses within 700 m of the well. Plot depicts correlation between seismic stratigraphy and lithofacies. The Quinault Formation designation of Rau and McFarland (1982) for this well was reinterpreted by Palmer and Lingley (1989), who separate the Quinault Formation into Montesano and Quinault Formations at an observed angular unconformity. Palmer and Lingley (1989) assign the overlap strata in this well to the Montesano Formation only. See figure 18A for a larger version of the profile. *B*, *Corliss* seismic-reflection profile and corresponding sidescan-sonar image (375-m swath width) that depict the rugged relief of strata equivalent to the upper Quinault lithofacies onshore. *C*, *McArthur* seismic-reflection profile that shows inferred lowstand and highstand deposits and bounding unconformities within inferred upper Quaternary strata—Continued.

across the shelf and south to north along the coast. Variations along the shelf primarily reflect differing sediment supplies and differing capacities of coastal streams to distribute sediment. Discussion of these variations is beyond the scope of this paper. Variations across the shelf suggest that the surface is a composite of multiple erosional episodes, and thus may vary in age (see discussion below). In general the outer shelf appears to be a region of sediment accumulation and the inner shelf a region of sediment erosion and reworking. This configuration may be explained by tilting of the Continental Shelf, with the outer shelf subsiding and the inner shelf rising.

Contraction Estimates for Offshore Faults and Folds

The study area can be divided into three regions on the basis of the orientation of geologic structures. South of Grays Harbor (46°53’N.), folds and faults trend north-south, roughly parallel to the western margin of the Coast Range block. Between Grays Harbor and Cape Elizabeth (47°21’N.), structures trend east-west, subparallel to an eastward step in the block boundary. The region north of Cape Elizabeth is defined by a shift in structural orientation back to north-south, typical of subduction-related deformation elsewhere in the accretionary complex.

North-South Domain: Columbia River to Grayland

Between the Columbia River and Grayland, the boundary between the Oregon Coast Range (OCR) and Olympic Mountains (OM) blocks is marked by a north-south fault zone. The boundary parallels the coastline and apparently dips about 25° eastward (fig. 2) (Snavely and Wagner, 1982; Parsons and others, 1999). On the basis of block kinematics, the north-south-trending boundary on the west side of the OCR block

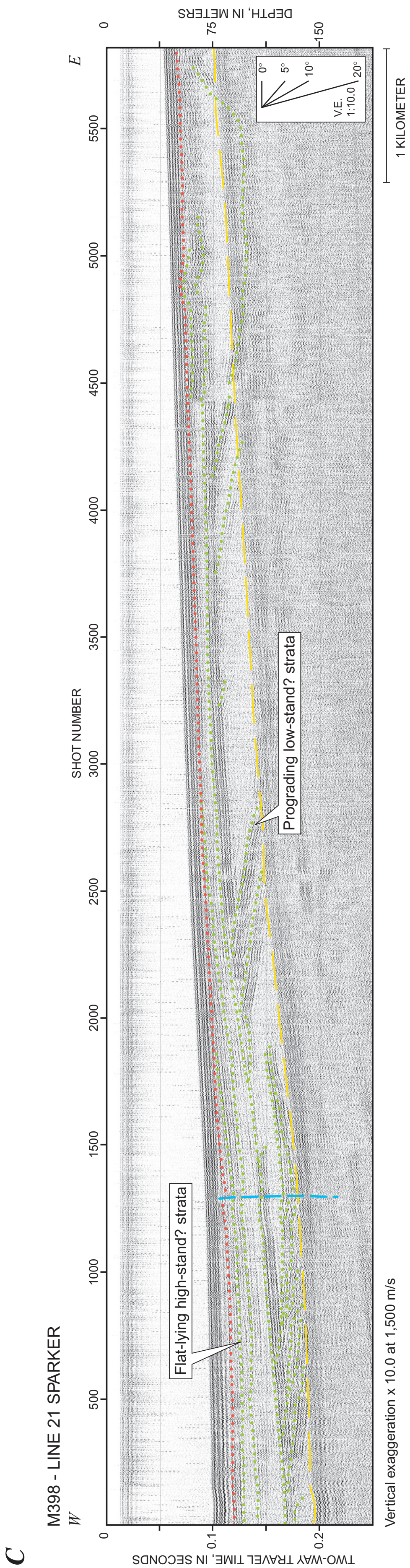


Figure 15—Continued.

is expected to accommodate predominantly dextral shear. However, the subsurface geometry of the boundary suggests thrust movement. From well data and wide-angle seismic data, the boundary can be extrapolated to the surface just west of the coastline. This location has not been confirmed, as no subsurface trace has been identified in the available high-resolution seismic-reflection data. Locating the trace of this fault is complicated by sea-floor multiples overprinting and obscuring the erosional unconformity in water depths < 20 m. In Willapa Bay to the east, we have mapped a 3-km-wide fault zone in the upper plate of the dipping boundary (plate 1*A*). This multi-strand fault zone elevates upper Quaternary strata on its western side and lowers strata on its eastern side to form a horst-graben pair (plate 1*A*). If the horst is elevated from thrust faulting, then the central fault strand would represent a back thrust in the upper plate of the east-dipping block boundary. Alternatively, the fault strands may accommodate transpressional slip and the graben may represent an elongate sag between fault strands. A back thrust would intersect the boundary thrust no deeper than 7 km, implying that such a structure terminates at too shallow a depth to be independently seismogenic. Steeply to gently east-dipping transpressional faults would intersect the boundary thrust much deeper. In either case, the recency and magnitude of movement on these fault strands imply activity on the postulated main thrust to the west.

Single-channel seismic-reflection profiles in Willapa Bay (Hill and others, 1981; Cross and others, 1999) depict a 30-km-long fault zone that trends north-northwest through the bay and may continue another 10+ km northwestward across the inner shelf (plate 1*A*). The fault zone vertically offsets a widespread buried unconformity beneath Willapa Bay (fig. 16). This buried surface has a surprisingly complex morphology that depicts areas of roughness, small paleochannels, and closed depressions (plate 1*B*). Rough paleorelief occurs where older strata have been elevated in a structural ridge to form the buried erosional surface on the western side of the fault zone. The rough patches, which are 25-35 m below sea level and 15 m below the bay floor, together form a linear trend that parallels the fault zone,. To the north, the structural ridge abuts a north-south-trending topographic high near Cape Shoalwater, and to the south it projects toward Long Island, a north-northwest-trending topographic high underpinned by an anticlinal fold (Walsh and others, 1987; Wells, 1989) that juts into Willapa Bay. The closed depressions east of the buried ridge and the minor channels that breach the ridge are reminiscent of geomorphic features associated with young strike-slip faults onshore. A number of north-northwest-trending Neogene structures have been mapped south and east of Willapa Bay (Wells, 1989). Some of these structures project into the bay and closely align with individual fault strands we have mapped beneath the bay. Confirmation of these potential connections will require further work.

The erosional surface shoals eastward, rising from about 25 m below sea level on the shelf west of Willapa Bay to 14 m below sea level on the east side of Willapa Bay (plate 1*B*). Upper Pleistocene strata are exposed in some modern tidal channels and in the bluffs that bound the east side of Willapa Bay (Clifton and others, 1989; Kennedy and others, 1982; R. L.

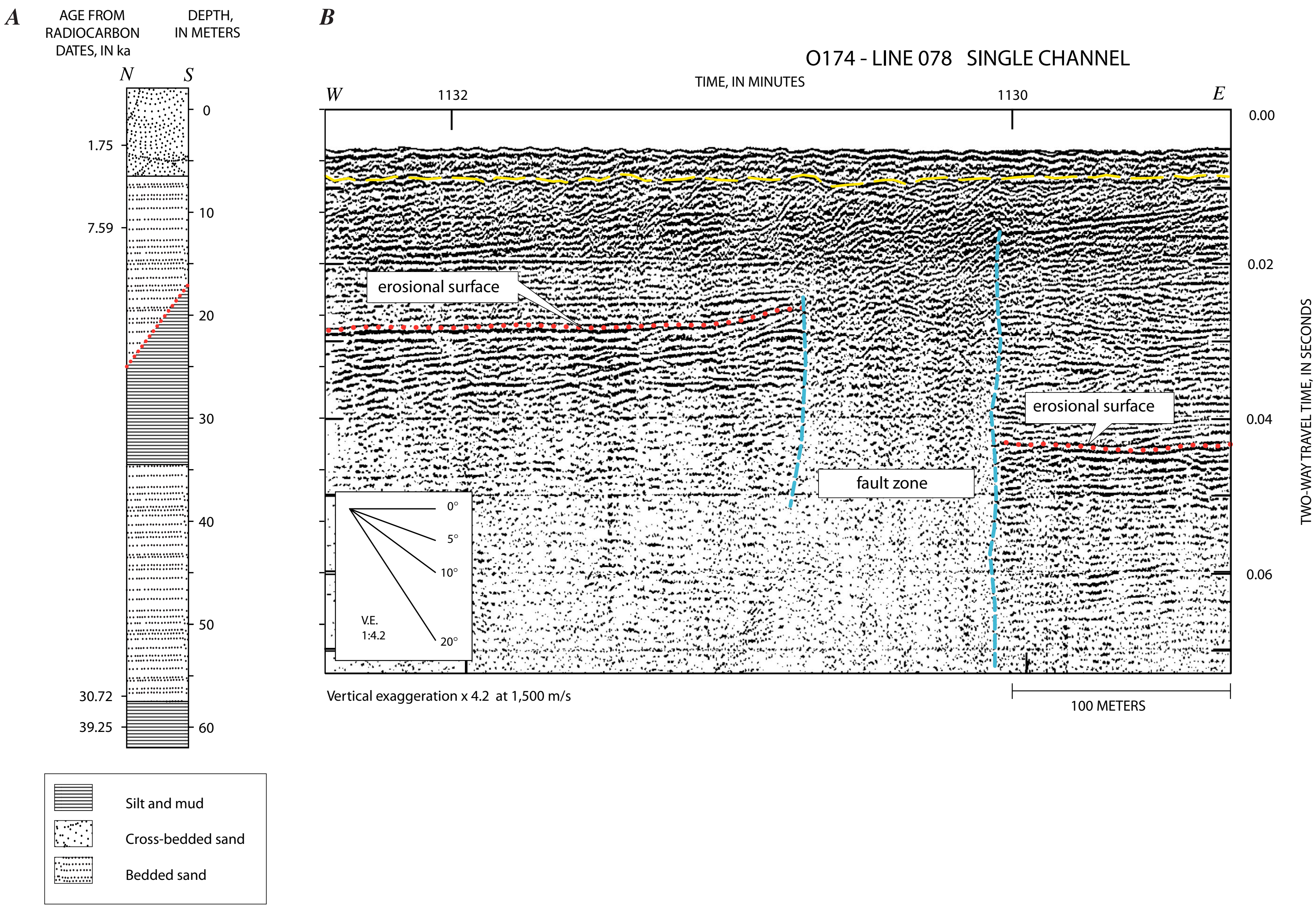


Figure 16. Composite lithostratigraphic column and seismic-reflection profiles in Willapa Bay. See plate 1*A* for locations of the seismic-reflection images. *A*, Composite lithostratigraphic column from borehole studies (Smith and others, 1999; Herb, 2000), with radiocarbon dates indicating that the erosional surface (dotted red line) formed about 20 ka. See figure 13 for locations of borehole sites. *B*, *C*, Seismic-reflection profiles showing offset ca. 20-ka erosional surface across "Willapa Bay" fault zone in central Willapa Bay (*B*) and in northern Willapa Bay (*C*). Red dotted line denotes unconformity; blue dashed lines denote inferred faults, yellow dashed line denotes first sea-floor multiple.

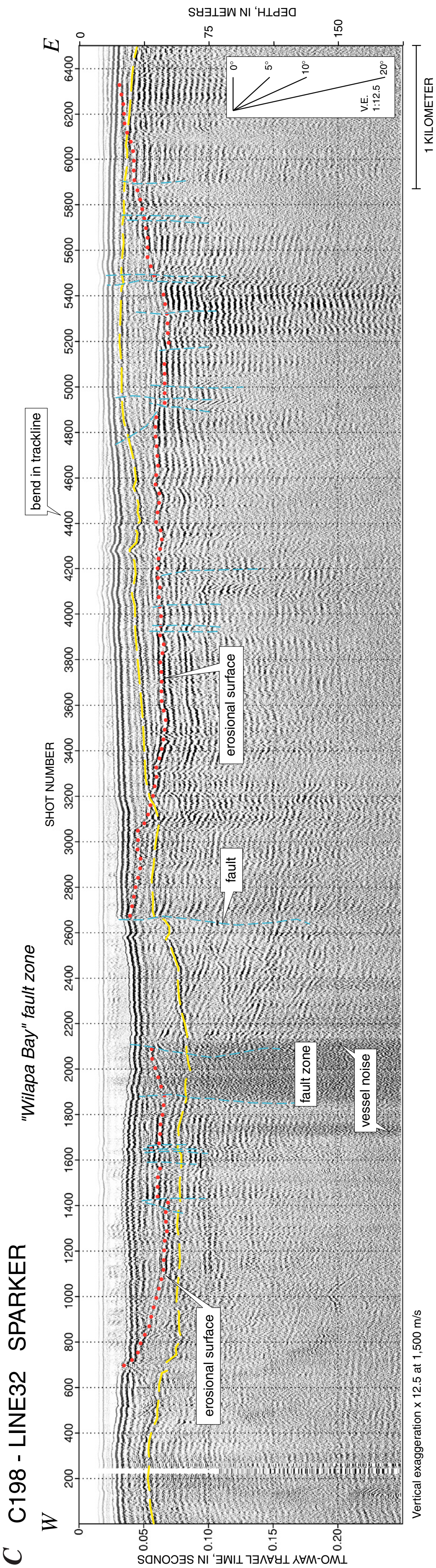


Figure 16—Continued

Phillips, oral commun, 1998). The upper Pleistocene strata may correlate with the unit beneath the eastward-shoaling surface.

Age of Late Pleistocene Datum

The age of the erosional surface beneath Willapa Bay (plate 1*B*) is constrained by radiocarbon dates from shell and wood material recovered from boreholes that penetrate Long Beach Peninsula (fig. 16*A*) (Smith and others, 1999; Herb, 2000). In particular, dates from ~36 m below and ~8 m above the erosional surface indicate that the surface formed between 30.7 and 7.6 ka. The surface is likely a ravinement or lowstand surface, judging from the paleochannels associated with it, which would restrict its age to the MIS-2 glacial period, ca. 20 ka. Alternatively, the surface could have been cut during the subsequent rise in sea level that would have flooded this portion of the shelf starting about 9 ka. Peterson and Phipps (1992) mapped a similar surface at similar depths beneath Grays Harbor, where marine sediment overlying the erosional surface is radiometrically dated at ca. 7.5 ka, suggesting that the two estuarine surfaces are correlative.

The highest widespread unconformity mapped beneath the Continental Shelf (Nittrouer, 1976; Wolf and others, 1997; Twichell and others, 2000) was likely cut during major late Pleistocene sea-level transgressions and modified by stream erosion during sea-level lowstands. The nearshore portion of this surface may be a continuation of the prominent unconformity mapped in Willapa Bay, as well as of the one previously mapped in Grays Harbor (Peterson and Phipps, 1992). The paleoshelf surface projects eastward into both estuaries (plate 1*B*) at elevations within a few meters of those beneath the estuaries (all range from 20 to 30 m below sea level).

Radiocarbon dates from gravity cores recently collected at mid-shelf depths offshore Willapa Bay (fig. 13) place the base of Holocene deposits at 4-7 m below the sea floor in 60-80 m of water (R.L. Phillips, unpub. data, 1999) and 25-40 m above the mapped unconformity. Extrapolating constant rates of sedimentation calculated from the dated cores (0.4-0.7 mm/yr) suggests that the surface formed about 70 ka in the mid-shelf region. If, however, sediment accumulation occurred mainly during submerged conditions, then this surface could be approximately twice as old or have been cut about 140 ka—during the MIS-5 sea-level transgression. If this is correct, then the surface beneath Willapa Bay clearly represents a much younger surface.

The disparity in age estimates for the erosional surface suggests that a simple model of the surface representing a single transgressional event is unlikely. In fact, the surface mapped on the inner shelf onlaps a buried mid-shelf high and then separates into a series of three unconformity-bound stratigraphic sequences beneath the outer shelf. This geometry suggests that the outer shelf is a region of sediment accumulation (owing perhaps to downward tilting or subsidence), where successive cycles of sea-level fluctuations are preserved as a stack of unconformity-bounded strata. Apparently the inner shelf, with a single unconformity, is a region of sediment reworking and removal (owing perhaps to upward tilting or uplift), where subsequent sea-level cycles remove or significantly modify previous ones. In this scenario the mid-shelf serves as a hinge-line where successive transgressions are preserved as lag deposits, without, however, angular discordance, making them difficult to discern. Because of these complexities and the general lack of direct age control, we assign a broad late Pleistocene age (<150 ka) to the surface in the outer and mid-shelf regions and a latest Pleistocene age (<20 ka) to the surface in the inner shelf and estuarine regions.

Slip Rate of Fault Zone in Willapa Bay

The fault zone in Willapa Bay is composed of multiple fault strands with both west-side-up and east-side-up sense of movement. Evidence for significant late Pleistocene movement on this fault zone is observed both onshore adjacent to Willapa Bay and beneath the bay floor. For example, in southern Willapa Bay, the fault zone projects onshore on the eastern side of Long Island, where a thrust fault offsets upper Pleistocene deposits 2 m. On the northwest side of the island, a vertical fault offsets Pleistocene deposits 0.8 m, west side up (fig. 17).

The fault zone aligns with a series of depressions in the erosional surface, reminiscent of sag ponds aligned with strike-slip faults. For example, in central Willapa Bay, a fault strand that vertically offsets the erosional surface about 5 m bounds the southwest side of an elongate depression (Wolf and others, 1998). The parallel alignment of the depression and the fault suggest that fault displacement may have partially controlled development of the depression. Four kilometers to the north, a fault strand, observed on several seismic-reflection profiles, offsets the unconformity between 4 and 8 m, west side up (fig. 16*B*). If the fault zone has a component of lateral shear, then these depressions may reflect differential slip between strands, causing a graben to form.

The largest cumulative offsets and widest zone of faulting are observed in northern Willapa Bay. Here, the buried unconformity and underlying strata are vertically offset 10-12 m, west side up, across a narrow zone (Wolf and others, 1998). However, another fault strand 500 m to the west offsets the unconformity 4-5 m, west side down. A *Corliss* trackline crosses the fault zone at an oblique angle, and the seismic-reflection data depict a multistrand fault zone about 2 km wide (fig. 16*C*). This zone projects onshore at Cape Shoalwater. A quarry near Cape Shoalwater exposes offset estuarine deposits along the northern projection of the fault zone (fig. 11*D*). These are minor thrust faults with a cumulative displacement of about 1 m in cross-bedded sands overlying the marine-terrace platform. Lateral slip, if any, is undetected. The Willapa Bay data yield a maximum vertical offset rate of ~10 m per 20 k.y. or 0.5 mm/yr.

A fault mapped 4 km west of Cape Shoalwater may be an en echelon continuation of the “Willapa Bay” fault zone. A fault mapped 11 km west of Cape Shoalwater vertically displaces the ravinement or transgressive surface about 8 m. This scarp is buried by about 12 m of post-ravinement sediment. If the surface has an age of 20 ka, this age yields an offset rate of 0.4 mm/yr. If this surface is ca. 140 ka, then the rate is 0.06 mm/yr. This structure may correlate with a thrust fault and associated anticline mapped by McNeill and others (1998) from industry multichannel data about 5 km further west.

A number of faults and folds underlie the shelf north of Willapa Bay, adjacent to the Grayland area. In general the structures are buried fairly deep, often beyond reach of the high-resolution seismic-reflection data. New seismic data do not image faults or folds in the upper 90 m of strata adjacent to Willapa Bay. The older, deeper penetration data image faults and folds beneath the shelf that trend north-south, parallel to the block boundary. These structures typically do not deform strata above the late Pleistocene unconformity and thus may not have been active in late Quaternary time.

East-West Domain: Grays Harbor to Cape Elizabeth

Near the entrance to Grays Harbor, structures abruptly shift from a north-south orientation to an east-northeast orientation over a 6-km-wide zone (plate 1*A*). The east-northeast-trending folds and faults are clearly contractional structures both offshore (plate 2) and onshore (McCrory, 1996), and they characterize the deformation between Grays Harbor and Cape Elizabeth. The change in structural orientation is marked by a major fault that cuts across the shelf and projects into northern Grays Harbor (plate 1*A*). Onshore, the multistrand fault appears to mark the boundary between the Oregon Coast Range and Olympic Mountains blocks as judged by basement affinities in petroleum wells drilled on either side of the fault zone (fig. 3). The Sunshine Minard No.1 well (fig. 3, site J), on the south side of the fault, bottoms in Eocene basalt, and the Shell Minard No. 1-34 well (fig. 3, site K), 4 km to the north, bottoms in Hoh mélange (Rau and McFarland, 1982). The projected block boundary offshore coincides with the bounding fault mapped with new seismic-reflection and sidescan-sonar data. In addition, recently collected multibeam data image a trough that extends to within 3 km of the coastline (G. R. Gelfenbaum, unpub. data, 1999) along trend with the fault, suggesting recent disruption of the sea floor.

The “Grays Harbor” fault and associated anticline mark the southern end of a bedrock high that extends 60 km from Grays Harbor northward to the Raft River (plate 1*B*). Nearshore, the high is either barren of sediment, covered with relict, oxide-coated gravels, or crossed by small channels filled with unconsolidated sediment (plate 2*G*). Seaward from about 60-m water depth, the high is commonly onlapped by rippled sediment. The lack of significant sediment cover may result from ongoing uplift associated with underlying structures. Uplift would disrupt modern sediment transport (Wolf and others, 1997; Twichell and others, 2000) by isolating the region from both northward-transported sediment from the Columbia River and westward-transported sediment from local coastal streams.

The bedrock high coincides with the mapped unconformity in the central region (plate 1*B*). This unconformity sur-

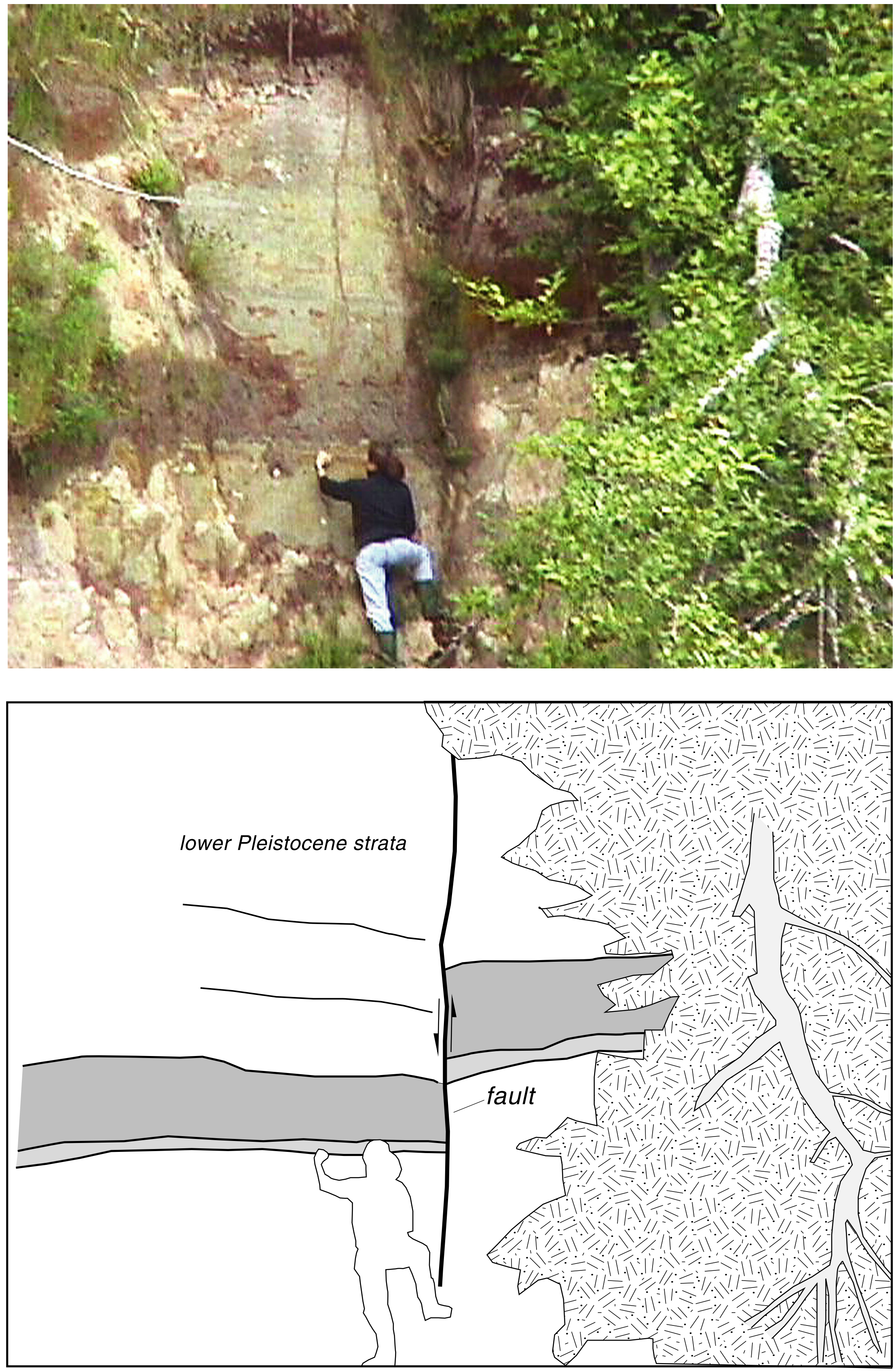


Figure 17. Bluffs on northwestern Long Island in Willapa Bay, showing 0.8-m vertical offset of lower Pleistocene strata. See plate 1A for location and direction of view.

face has been modified by erosion. In particular, three filled channels cut through the bedrock surface: one aligns with the entrance to Grays Harbor, another projects toward Copalis River, and the third projects toward Moclips River. In detail, the morphology of the surface shows some correlation with underlying structures. Anticlinal folds tend to underpin areas of emergent bedrock, and synclinal folds tend to align with paleochannels. This correlation suggests that the complex morphology is in part controlled by Quaternary structural activity.

"Grays Harbor" Fault Zone

Seismic-reflection profiles depict the "Grays Harbor" fault as a 2.5-km-wide zone of reverse faults and back thrusts within a broad anticlinal fold with a wavelength of 15 km. The upward extent of fault strands varies from profile to profile. On some profiles, fault strands appear to die out in lower Pleistocene strata. On other profiles, individual fault strands breach and offset the sea floor. Twenty km west of Grays Harbor, the fault vertically displaces the sea floor 12 m over a 1.7-km-wide zone (plate 2*J*, *K*). Two kilometers west of Grays Harbor, the fault vertically offsets the late Pleistocene erosional surface by more than 40 m, and one of the fault strands displaces the sea floor a few meters. Owing to uncertainties about the age of the unconformity near shore (150-20 ka), a vertical displacement rate is not calculated for this offset.

"Saddle Hill" Fault Zone

About 4 km north of the "Grays Harbor" fault, another multistrand fault crosses the shelf and continues onshore. This fault bounds a filled channel that breaches the bedrock high. Moore (1965) documented thrust faults within Pleistocene gravels beneath Saddle Hill (McCrory 1996), where the fault projects onshore. Petroleum well Tidelands State No. 2 (fig. 3, site L) was drilled into a fault strand about 3 km offshore. Rau and McFarland (1982) document thrust faulting within Olympic subduction rocks penetrated in this well using a biostratigraphic age reversal. *McArthur* lines 98-27 and 98-29 cross near the well location. These seismic-reflection profiles depict a pair of faults about 1 kilometer apart on either side of an anticline (fig. 18).

Montesano sandstone lithofacies (fig. 4) abruptly thin north of Saddle Hill between the Union State No. 3 and Shell Sampson Johns No. 2-15 wells (fig. 3, both site M), suggesting substantial fault displacement (Rau and McFarland, 1982; Palmer and Lingley, 1989). Palmer and Lingley (1989) propose that the "Saddle Hill" fault accommodates translational motion as well as compressional motion, on the basis of this mismatch in sandstone thickness across the fault zone. Given the evidence for thrust faulting discussed above, the facies mismatch may instead result from sediment-dispersal barriers that formed by growth faulting concurrent with sediment accumulation in the Miocene.

Structures in the Point Grenville Area

On the inner shelf between Saddle Hill and Point Grenville, additional faults and folds are well imaged in seismic-reflection profiles and sidescan-sonar swaths. These structures range in length from about 5 to 20 km. Available data are not sufficient to map the longer structures further west than the mid-shelf region, about 25 km west of the coastline. A series of synclinal folds with wavelengths of 3-6 km (plate 2*A-C*) can be mapped from the mid shelf to the coastline. Several synclinal axes project towards stream valleys onshore, suggesting a structural basis for the location of Connor Creek, Copalis River, Boone Creek, Joe Creek, and Moclips River (plate 1*A*). Additionally, several intervening anticlinal folds project towards structural ridges previously mapped onshore (McCrory, 1996, 1997). McCrory (1996, 1997) documented thrust faults and folds in Quaternary strata that underpin the onshore ridges just south of Copalis River, Joe Creek, and Moclips River. One large anticlinal structure, imaged on *Corliss* line 97-01 between Copalis River and Boone Creek (fig. 19), is not emergent onshore.

The structural pattern changes adjacent to Point Grenville. Here structures are more closely spaced, with wavelengths of 0.5-1 kilometer, and appear to curve around Point Grenville (plate 1*A*). These curved folds and faults are bounded on the north by a large anticlinal fold (plate 2*E*) that is emergent on the sea floor. Wagner and others (1986) mapped the anticline as diapiric because of the lack of internal reflectors on seismic-reflection profiles. This interpretation is supported by additional seismic-reflection profiles collected in this study and by the incoherent mélange material exposed on the sea floor in sidescan-sonar images of the fold core. The fault-bounded diapir exposed in the adjacent sea cliffs at Point Grenville (Rau and Grocock, 1974; Rau, 1975) may be an onshore extension of this structure.

Faults and folds to the northwest of the tightly folded area trend northeast-southwest, consistent with structural orientations to the south. One of these faults displaces the sea floor (Wagner and others, 1986; this study) and projects onshore near Cape Elizabeth where the MIS-5a terrace datum jumps from 24 to 51 m in elevation (West and McCrumb, 1988, McCrory, 1996), reaching there its highest elevation along the Washington coast. A 26-m offset of the ca. 83-ka marine terrace platform (MIS-5a) yields a rate of vertical displacement of 0.3 mm/yr.

Contraction Estimate

The onshore-offshore structural correlations are here used as a basis for estimating Quaternary deformation. To obtain a first-order approximation of the rate of shortening across this 40-km-wide region, we assume that the topographic ridges onshore reflect Quaternary shortening that correlates with the erosionally truncated anticlines offshore. This assumption ignores the effects of erosion that would tend to subdue relief. All the topographic relief is assigned a structural origin analogous to the anticlinal folds preserved offshore. The presence of estuarine deposits on the crest of Langely Hill at 67-m

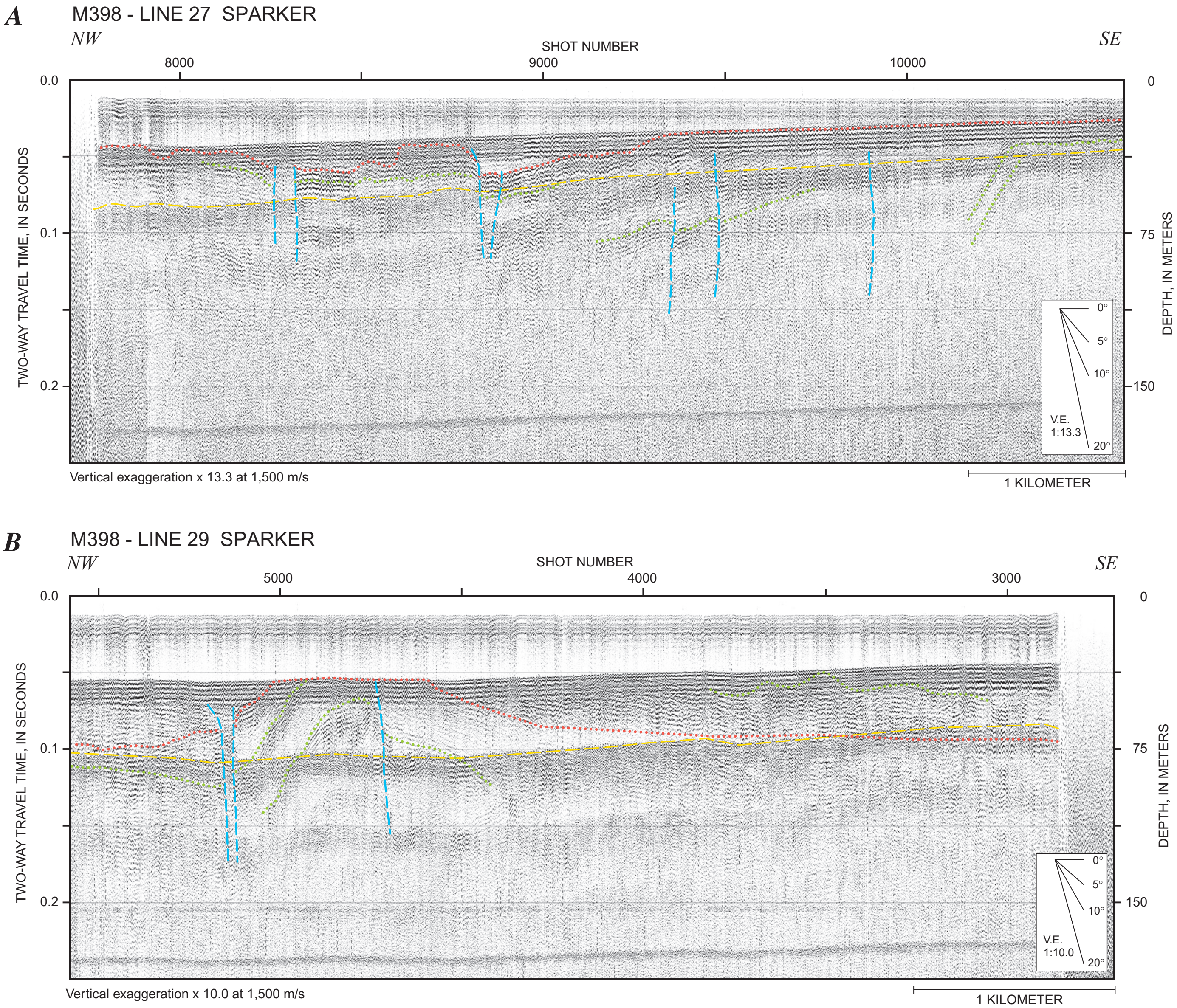


Figure 18. Seismic-reflection profiles of the "Saddle Hill" fault zone at 4 km (*A*) and 8 km (*B*) west of the coastline, showing displacement of upper Quaternary strata. See plate 1*A* for locations of the images. Green dotted lines denote stratigraphic reflectors, red dotted line denotes late Pleistocene unconformity surface, yellow dashed line denotes first sea-floor multiple, blue dashed lines denote inferred faults. Profile *A* crosses within 700 m of Tidelands State No. 2 well.

elevation (fig. 11*C*) lends support to this assumption. The ridges are cored by lower Pleistocene deposits as indicated by well stratigraphies (fig. 19; Rau and McFarland, 1982), and a maximum early Pleistocene age is assigned to the deformation. We also assume that the unconformity surface preserved in synclinal folds offshore is equivalent to the surface buried in the stream valleys onshore. The offshore surface is younger than the onshore one, and at the south end of the profile likely represents the late Pleistocene datum mapped in plate 1*A*.

The combined amplitude of the folds ranges up to 160 m, and their wavelengths vary from 15 km in the south to 3 km in the north. We have measured the length of the geomorphic surface onshore and of the synclinal folds offshore (fig. 19) from Grays Harbor to Point Grenville following the shape of the folds. The offshore measurement only includes synclinal folds imaged on *Corliss* line 97-01. A 3.5-km gap across the strike of the "Saddle Hill" fault zone was caused by equipment malfunction during the cruise. In addition, the shortening estimate does not include the offset terrace north of Point Grenville, which is not completely imaged on the *Corliss* line.

The composite datum suggests a decrease in line length of about 2.0 km (fig. 19). If the age of the surface is 0.9 Ma, the shortening rate is about 2.2 mm/yr; if its age is 0.6 Ma, then the rate is 3.3 mm/yr. If the folding reflects movement on buried thrust faults offshore analogous to those mapped in the core of the anticlinal ridges onshore, then the cumulative slip rates assigned to such faults would be considerably higher. As a first approximation, this region adjacent to the block boundary appears to accommodate as much as half the available 6-10 mm/yr convergence at the surface.

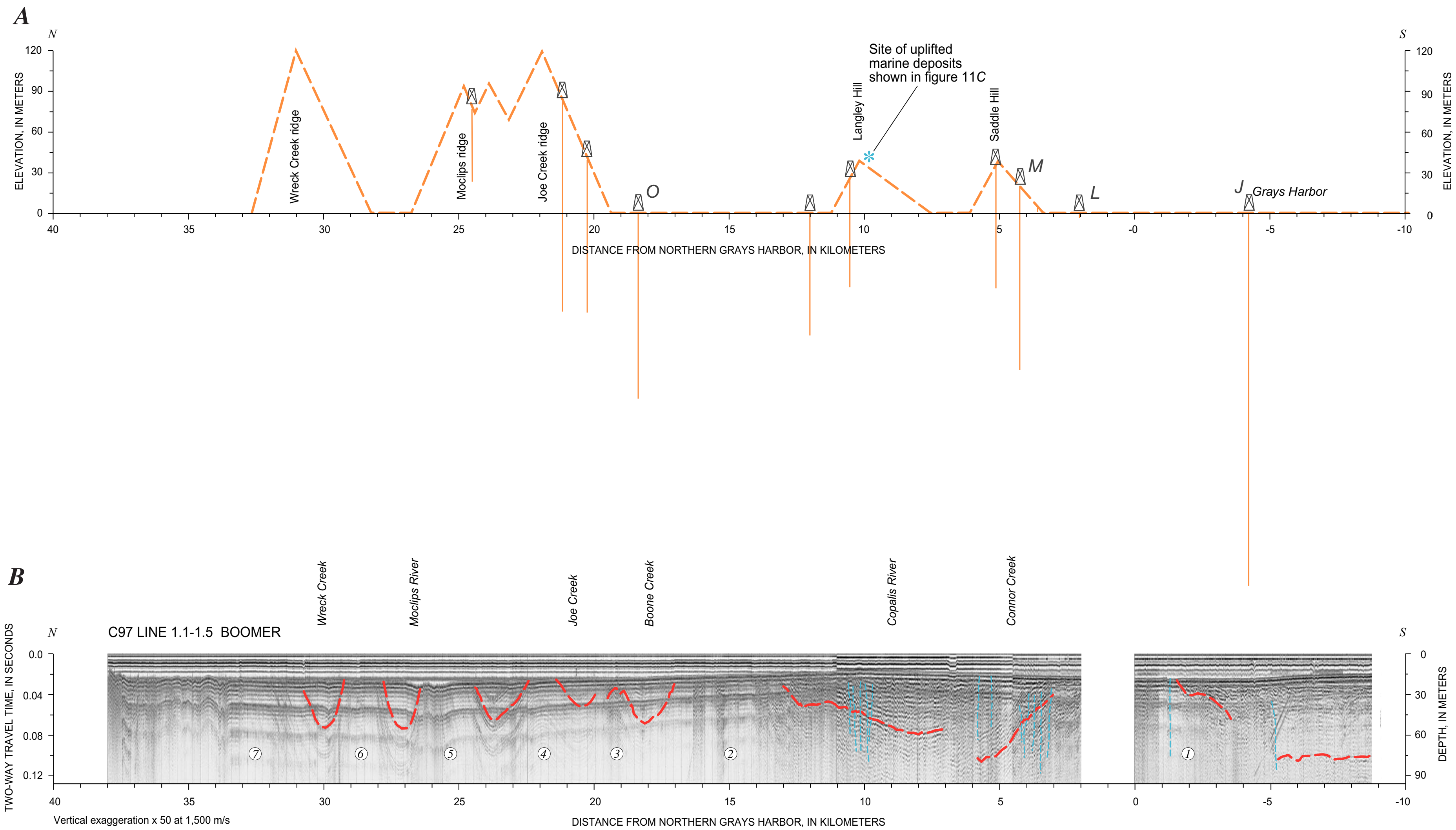


Figure 19. Profiles parallel to coastline used to calculate shortening of Quaternary strata. See plate 1*A* for locations of the profiles. *A*, Onshore geomorphic profile used to calculate Quaternary shortening (Raft datum) accommodated in anticlinal folds. Labeled well symbols denote locations of coastal wells with respect to topography. See figure 3 for map locations of wells. Vertical lines beneath well symbols denote thickness of Quaternary strata in these wells (from Rau and McFarland, 1982). *B*, Offshore seismic-reflection profile used to calculate shortening (red line is approximately equivalent to Langley datum) accommodated in synclinal folds (a portion of this profile is also shown in plate 2*A*). Circled numbers identify major anticlinal folds and associated faults depicted in profile. Note the decreasing wavelength of folds from south to north.

Seven primary anticlinal folds and associated faults are imaged on *Corliss* line 97-01 (fig. 19). The western extent of these transverse structures is not well documented. Some are more than 35 km long and others appear to die out after 10-15 km. The morphology of the unconformity—with contours jutting to the west— suggests that the fault zone may extend another 10 km westward. The wavelength of the folds decreases northward: the fold associated with the bounding fault is about 15 km wide, whereas further north the fold wavelength is as short as 3 km. This configuration suggests the presence of a northward-shoaling detachment fault in the subsurface below the folds on the basis of fault-bend fold models. Detailed consideration of this inference is beyond the scope of this paper. Interestingly, the onshore ridges are higher in elevation in the north, where the fold wavelength is shorter, suggesting that these northern structures are growing in size along trend from offshore to onshore. The lack of surface relief in the south may result from stream erosion or from the structures diminishing rapidly eastward. Available data are not sufficient to assign slip rates to specific structures. For preliminary hazard assessments, therefore, the shortening rate of 2.2-3.3 mm/yr is spread evenly across this zone of seven primary structures.

Northern Domain: Cape Elizabeth and North

North of Cape Elizabeth, structures shift progressively from an east-northeast orientation to a north-northwest orientation (plate 1*A*) over a 6-km-wide zone. The erosional surface in this region forms a gently westward-dipping surface down to a depth of about 85 m (plate 1*B*). Below 85 m, the surface dips more steeply and is buried by a thickening wedge of sediment. A narrow filled channel cuts across the surface and projects eastward toward the mouth of Queets River. A large apron of prograded strata flanks the channel (fig. 15*C*) and likely represents glacial outwash deposits derived from the Queets River drainage. In general, the structures underlying the northern region are not expressed in the morphology of the erosional surface. However, isolated patches of emergent bedrock do occur along some structures.

Many faults mapped in the northern region do not offset or deform the late Pleistocene unconformity. However, some faults do show evidence of ongoing contraction consistent with subduction-driven processes. A nearshore fault just north of Cape Elizabeth projects onshore to join with the late Pleistocene structure beneath Duck Creek ridge (McCrory, 1996). A fault and associated anticline that deform the sea floor offshore from Whale Creek (plate 1*A*) project toward faulted upper Pleistocene deposits (McCrory and others, 1996) exposed in the sea cliffs south of the Raft River (fig. 11*A*).

Faults on the Continental Shelf north of the Raft River do not generally project onshore; rather, they tend to subparallel the coastline. Some of these faults disrupt the late Pleistocene erosional unconformity and offset the sea floor. For example, a fault about 20 km west of the Queets River (plate 2*I*, shot point 9000) disrupts the sea floor and underpins localized bedrock outcrops (plate 1*A*, *B*). This evidence of late Quaternary structural activity in the northern region implies continued subduction-related contraction far from the deformation front, perhaps driven by interplate coupling. Slip-rate estimates, however, are not possible with the available data.

Tectonic Implications of Crustal Shortening

Differential motion across the central Cascadia forearc in coastal Washington results in a complex pattern of folds and faults within the continental-shelf and onshore portions of the accretionary margin. Northward translation of the Oregon Coast Range (OCR) block is driven either by a margin-parallel component of oblique convergence between the Juan de Fuca and North America Plates or by broad northwest-southeast shearing between the Pacific and North America Plates across the Juan de Fuca Plate (fig. 20) (see, for example, England and Wells, 1991; Walcott, 1993). In either scenario, the OCR block acts as an indentor against which the relatively weak accretionary rocks of coastal Washington deform.

On a more regional scale, the pre-Cenozoic terrane composing southern British Columbia acts as a buttress at the northern boundary of the Olympic Mountains (OM) block. Uplift rates in the northern Olympic Mountains, ~3.2 mm/yr (Savage and others, 1991), are the highest along the Cascadia subduction margin, except for the region of the Mendocino triple junction. Brandon and Vance (1992) document ongoing unroofing of the subduction complex in the Olympic Mountains on the basis of fission-track studies, and attribute this unroofing to uplift driven by underplating of subducted material. Savage and others (1991) attribute the uplift to elastic strain accumulation above a locked megathrust. Models that account for elastic strain accumulation leave significant positive residuals (greater than predicted uplift) in the Grays Harbor area (Svarc and others, 1999). These residuals may represent shortening in the upper plate of the megathrust and, in particular, may represent shortening from forearc block convergence. Alternatively, the residuals may indicate a zone where the subduction fault is locked that is locally wider than previously modeled.

Strain Partitioning near the Block Boundary

Subduction-complex rocks beneath the study area were accreted to the Washington continental margin during Cenozoic time. Their primary structural fabric is expected to reflect east-northeast-directed convergence associated with subduction processes. Folds that trend north-northwest in the northern portion of the study area are attributed to this regime. Too few shallow structures are mapped in the southern portion of the study area to adequately assess their kinematic origins. We attribute the faults beneath Willapa Bay to forearc-block motion. However, in the absence of observations of inferred strike-slip motion, a subduction regime cannot be ruled out.

Quaternary structural orientations in the central portion of the study area denote north-south shortening. These structural orientations are skewed relative to mapped structures in the surrounding region (Wagner and others, 1986; this study) that reflect the subduction regime (sensu stricto). In detail, the skewed structures show a progressive clockwise shift in ori-

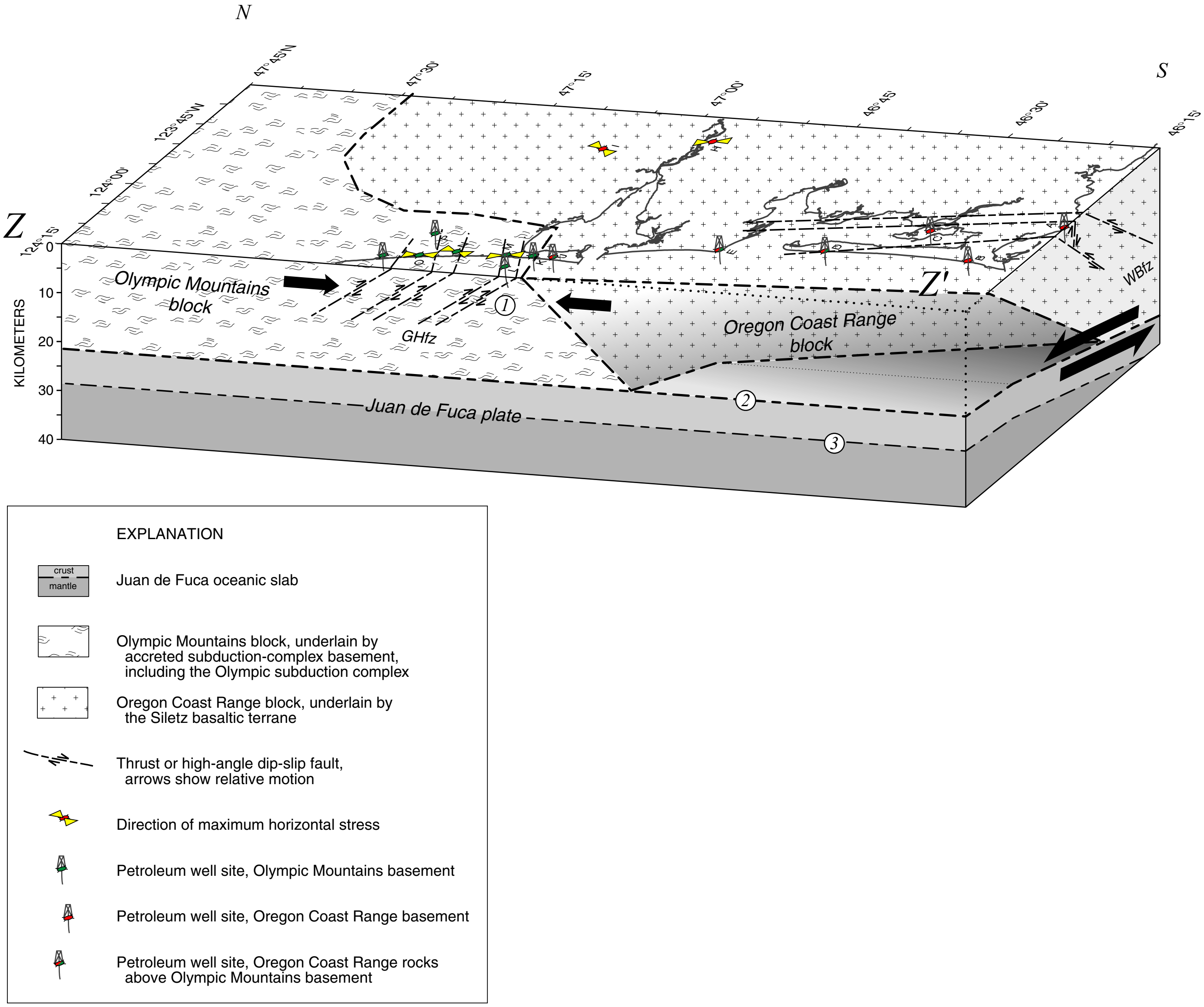


Figure 20. Generalized north-south block diagram at no vertical exaggeration showing inferred configuration of Oregon Coast Range and Olympic Mountains blocks. See figure 1 for location of profile *Z-Z'*. Circled numbers identify potential seismic sources: (1) upper plate, interblock sources; (2) interplate, megathrust sources; (3) lower plate, Juan de Fuca sources. The location of the Juan de Fuca Plate beneath the margin is constrained by a velocity model of Parsons and others (1999). Heavy paired arrows denote relative movement on Cascadia megathrust between the crustal blocks and the Juan de Fuca slab; heavy opposed arrows denote relative movement between crustal blocks. WBfz indicates "Willapa Bay" fault zone, and GHfz indicates "Grays Harbor" fault zone.

entation from south to north. The southernmost structures—in the Copalis River area—trend east-northeast, and the northern ones—in the Wreck Creek area—trend west-northwest (plate 1*A*). Farther north, thrust fault and fold trends abruptly shift to a north-northwest orientation (McCrory, 1996).

The progressive shift in structural orientation suggests that the study area serves as a transition zone between north-south-directed compression in the south and east-west-directed compression in the north. This shift occurs adjacent to a major boundary between crustal blocks of widely differing rheology. Paleomagnetic and geodetic data indicate that the relatively rigid southern block, the OCR block, is actively translating northward along the Cascadia margin (Wells and others, 1998; McCaffrey and others, 2000; Wells and Simpson, 2001). Observed geologic shortening discussed above occurs primarily in the relatively ductile subduction complex just north and west of the block boundary (fig. 1). The proximity of the shortening to the block boundary suggests that these structures accommodate ongoing convergence between the OCR and OM blocks.

If a third to a half of available contraction is consumed across this zone of transverse structures, the question remains as to where the remainder is consumed. The broad-scale warping of Siletz basement on the southern side of the block boundary is one likely sink. Wide-angle seismic data indicate relative basement lows beneath Grays Harbor and Willapa Bay, with an intervening basement high (Parsons and others, 1999). This may be inherited relief or could represent broad-scale buckling with a 25-km wavelength. Geodetically derived uplift rates (Savage and others, 1991) indicate slow uplift beneath Willapa Bay (0.4 mm/yr) in support of continued downwarping relative to the surrounding area. Some contraction may be distributed as general uplift in the Olympic Mountains and some may account for the high geodetically derived uplift rates near the northern block boundary in the Clallam area (fig. 21).

Seismic Hazard Implications of Quaternary Tectonism

The boundary between the OCR and OM blocks is characterized by northeast-trending thrust faults and associated anticlines developed primarily in subduction-complex basement and overlying strata. The location and orientation of these structures indicate ongoing north-south contraction and suggest relative convergence between these basement blocks in coastal and nearshore Washington. A historic record of felt earthquakes (1884, April 1887, 1891, 1909) and tsunamis (May 1887, 1920) exists for the Grays Harbor area (L.J. Workman, written commun., 1994), although the record is too sparse to locate the source of the older events. Some of the earthquakes likely occurred in the subducted Juan de Fuca slab (for example, the 1909 and 1920 events), as did the 1999 M_w 5.9 Satsop earthquake. The 1887 tsunami is intriguing

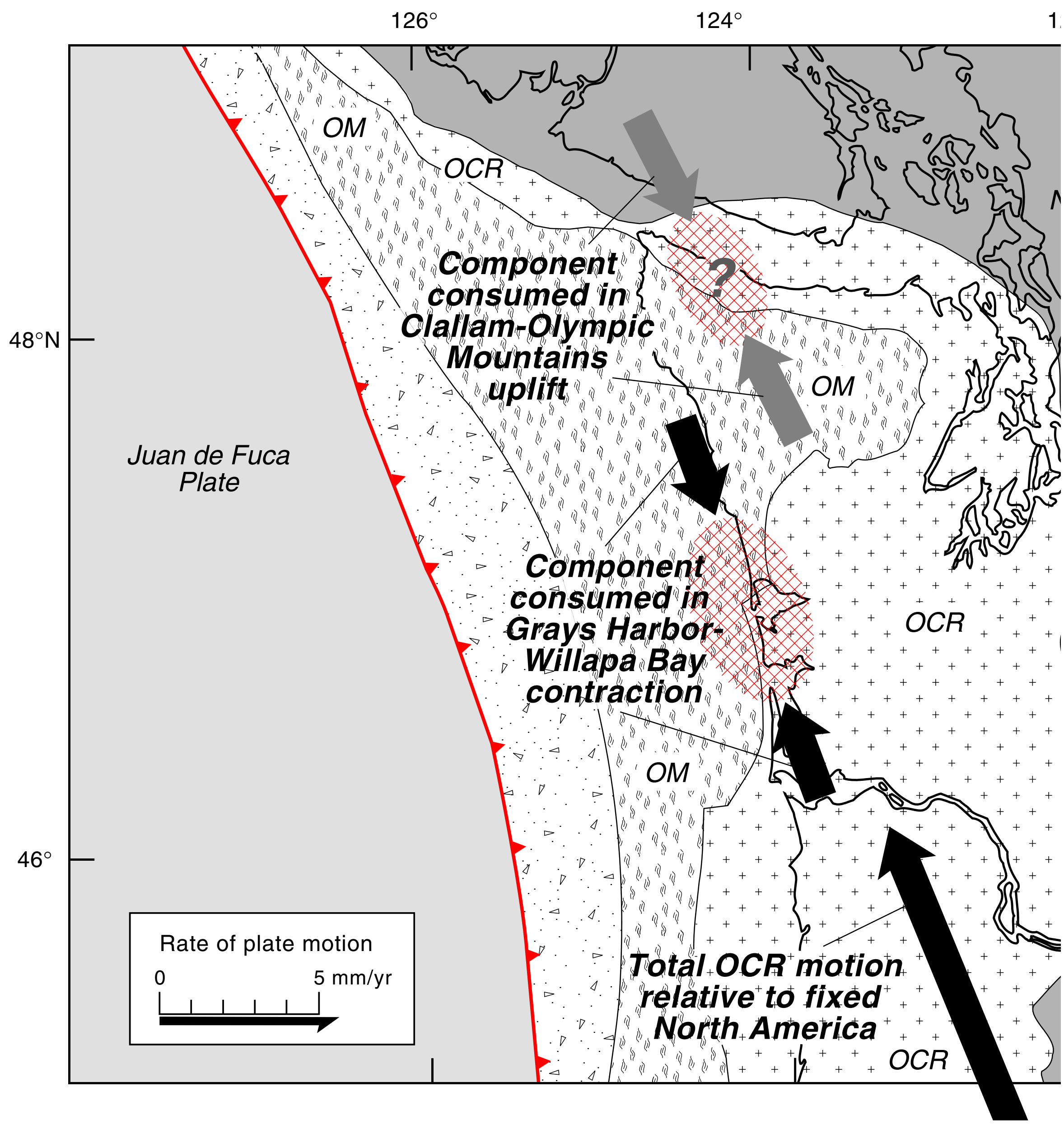


Figure 21. Schematic kinematic diagram of block movements and stratal shortening in the study area. See figure 1 for explanation of patterns and symbols. Long black arrow shows velocity vector for Oregon Coast Range block relative to fixed North America. Opposed black arrows indicate consumption of translational motion as geologic contraction in the Grays Harbor-Willapa Bay area and the Olympic Peninsula. Opposed gray arrows in the Clallam Bay-Olympic Mountains area indicate where remaining strain may be consumed in geologic contraction.

because it occurred two weeks after a locally felt earthquake and may represent delayed slumping triggered by a local, shallow offshore source of crustal seismicity.

If the northeast-trending structures in the study area are seismogenic, they contribute to the regional earthquake hazard and may be represented in the paleoseismic record. In fact, paleoseismic evidence of fault slip beneath the structural ridge south of the Copalis River may be present in Holocene marsh strata exposed along the river. Copalis marsh strata contain three stratigraphic reversals that Atwater (1992) infers to reflect coseismic subsidence during megathrust earthquakes. Atwater (1992) identifies an additional seismic event, ca. 1.1 ka (fig. 3), from liquefaction evidence alone—this event lacks a stratigraphic reversal to indicate subsidence. The structure beneath Langley Hill, 5 km south of the Copalis River, could be the source for this seismic event. In this scenario, subsidence would not be expected on the Copalis River because it is located on the upper plate of the thrust fault.

Holocene marshes in southern Washington estuaries may contain a record of both megathrust and crustal seismic events, as has been proposed for the southern and central Oregon coast (Goldfinger and others, 1992; Nelson and Personius, 1996). The marshes that fringe the estuaries typically preserve only the last 3-4 k.y. of the Holocene stratigraphic record (Atwater and others, 1995; Atwater and Hemphill-Haley, 1997). Even though recurrence intervals for faults with slip rates less than 1 mm/yr are typically measured in thousands of years, such an event could fall within the available stratigraphic record.

The estimated 2.2-3.3 mm/yr of contraction in coastal Washington could indicate as much as 5-7 mm/yr of cumulative fault slip spread across seven thrust faults (each dipping 30°). In this scenario, each fault would have an average slip rate between 0.7 and 1.0 mm/yr depending on the age of the early Pleistocene Raft datum. If the 2.5-m fault slip measured at the Langley quarry (McCrory, 1996) is a typical surface rupture, then a fault with a slip rate of 1.0 mm/yr would require ~2.5 k.y. to accumulate enough elastic strain to trigger a comparable event. If the slip rate is 0.7 mm/yr, ~3.6 k.y. would be the average interval between rupture events. In the scenario described above, the tidal marsh stratigraphies in Copalis River, Grays Harbor, and Willapa Bay would likely record no more than one such crustal seismic event.

The 10-12 m fault offset of the ca. 20-ka surface beneath Willapa Bay likely required multiple rupture events. Such slip events would be accompanied by uplift of western Willapa Bay and subsidence of the eastern bay. The main fault strand may terminate in the boundary fault, inferred to be about 7 km beneath Willapa Bay (fig. 20). In this case, the rupture area would be small and generally above the seismogenic zone. Nonetheless, recent movement on this fault and ambient seismicity (fig.2) point to the potential for recent movement on the boundary fault as well. Borehole studies map out a late Pleistocene bedrock high beneath Long Beach Peninsula (Herb, 2000) (plate 1), consistent with its location on the upper plate of the boundary thrust. Alternatively, this bedrock high may delineate a northern continuation of the Chinook Fault (Wells, 1989) south of Willapa Bay (fig. 3) and of the anticlinal structure that underpins Long Island.

Distinguishing crustal or upper-plate fault slip events from megathrust slip events in the paleoseismic record requires documentation of differential movement. A crustal slip event should produce local areas of uplift juxtaposed against areas of subsidence or no detectable change. Ideally, radiometrically dated sites would be sufficient to document uplift of the upper plate of a crustal fault and subsidence of the lower plate. However, documenting uplift alone, or a pattern of subsidence inconsistent with a megathrust signature may be adequate to demonstrate local slip events.

Conclusions

The Cascadia subduction boundary comprises several seismotectonic elements. The primary element is defined by subduction of the Juan de Fuca Plate beneath the continental margin. In the central portion of the subduction zone, relative convergence is about 40 mm/yr and is expected to generate great earthquakes with recurrence intervals measured in hundreds of years. A secondary seismotectonic element results from relative motion within the Cascadia forearc. The Oregon Coast Range block in the Cascadia forearc translates northward with rates that range from about 4 mm/yr in the Puget Sound region to about 8 mm/yr in coastal Washington. Faults that accommodate relative motion of this block with respect to neighboring blocks are expected to generate moderate to large earthquakes with recurrence intervals measured in thousands of years. Even with the much smaller expected magnitudes and the much longer recurrence intervals as compared to expected megathrust earthquakes, these faults pose significant seismic hazard owing to their close proximity to populated regions.

Marine investigations along the southern Washington coast document intense Quaternary tectonic activity at the leading (northwestern) edge of the Oregon Coast Range block. Permanent shortening is primarily localized within the relatively ductile rocks of the Olympic Mountains block, on the north side of the block boundary. Here folds range in length from 3 to 15 km and have amplitudes as great as 160 m. The wavelength of inferred buckling of the Oregon Coast Range block adjacent to its leading edge is about 25 km; the amplitude is not discernible at the surface except for the general correlation with topographic relief.

We have mapped a zone of faults and folds, 40 km wide from south to north, that are oriented transverse to the Cascadia subduction fabric. This zone appears to convert 2.2-3.3 mm/yr (as much as half) of the Oregon Coast Range translation into permanent crustal shortening. We correlate several faults mapped offshore with previously mapped onshore faults that displace Quaternary deposits. Some of the offshore faults displace the sea floor as much as 12 m, indicating Holocene movement. A late Quaternary fault bisects Willapa Bay above the Oregon Coast Range boundary. This fault offsets a ca. 20-ka datum 10-12 m, yielding a vertical offset rate of 0.5 mm/yr. Fault geometry suggests that some strands of the "Willapa Bay" fault zone may terminate at shallow depths above the main thrust boundary whereas others may extend far deeper.

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Appendix 1. M197WO—The July 7-14, 1997, Cruise on NOAA Ship RV *McArthur*

See URL: walrus.wr.usgs.gov/docs/infobank/ for links to cruise data. See Foster and others (2001) for seismic-reflection data archives and McCrory and others (in press a) for sidescan-sonar data archive.

USGS Scientific and Technical Personnel

Patricia McCrory, Menlo Park, California
Thomas O'Brien, Woods Hole, Massachusetts
David Foster, Woods Hole, Massachusetts
Scott Harrison, Saint Petersburg, Florida

Scientific Equipment

Navigation—USGS differential GPS navigation was logged using an Ashtech system with USGS YoNav navigation software. *McArthur* GPS navigation was logged on a Magnovox system.

Bathymetry—The pipe-mounted bathymetric system had a 3.5-kHz acoustic source (transducer). Seismic data were received on an ORE system and recorded digitally on a MudSeis acquisition system. Analog data were recorded on an EPC graphic recording system.

Sidescan sonar—The EdgeTech sidescan system had dual 100- and 500-kHz acoustic sources. Only the 100-kHz source was recorded on this cruise. Slant range from the ship to the sonar towfish was provided by a Benthos 7000 system. Reflectivity data were recorded with a Triton-Elics ISIS digital acquisition system and processed onboard using USGS XSonar software.

Boomer—The dual-plate Geopulse system with a 1-kHz EG&G acoustic source was towed on a surface catamaran. Seismic data were received on an ORE system and recorded digitally on a MudSeis acquisition system using a Benthos hydrophone streamer. The seismic data were processed onboard using Seismic-UNIX software. Analog data were recorded on an EPC graphic recording system.

Sediment sampler—A Shipek grab sampler was used to collect bottom samples.

Cruise Instructions

The cruise instructions called for eight east-west survey lines and three north-south cross lines in the Grays Harbor-Queets River, Washington, area. In general, the nearshore portions of tracklines were truncated at 30-m water depth to avoid tangling the sidescan tow fish in crab-pot strings. The sidescan fish was pulled onboard during nearshore north-south transects. During the last trackline (Line 18) on the final night of data collection, the sidescan fish did become tangled in a crab-pot line and was rendered inoperable. Because of the shortened tracklines and favorable weather conditions, three additional lines were completed. In total, geophysical data were collected along 400 km of trackline (fig. 12). In addition, seven sea-floor sediment samples (table 1) were collected during an east-west Conductivity-Temperature-Depth (CTD) transect offshore Cape Elizabeth.

YoNav, the differential global positioning system (DGPS) navigation software used in the geophysical lab, had a bug that prevented real-time display of the ship's location. However, the DGPS navigation was logged with the sidescan-sonar data and used to plot these sidescan swaths. The *McArthur* GPS navigation system was used for boomer trackline locations. The DGPS navigation located tracklines with an accuracy of 5-10 m; the GPS system had an accuracy of ~20 m. The DGPS was used for the sidescan navigation, and the NOAA GPS was merged with the boomer shotpoint files by correlating time between data sets. The bathymetric system was pipe-mounted on the port side of the ship and, even in moderate sea conditions, yielded noisy data owing to motion of the ship. The sidescan system was towed about 20 m above the sea floor using a USGS winch from the aft A-frame. This configuration yielded a 375-m swath width and a resolution of 19 cm. The seismic-reflection (boomer) system consisted of two plates towed on a catamaran from the starboard *McArthur* winch and A-frame. Bottom grab samples were collected in 30 to 70 m of water using the starboard *McArthur* winch and A-frame.

Appendix 2. M398WO—The June 25 - July 4, 1998, Cruise on NOAA Ship RV *McArthur*.

See URL: walrus.wr.usgs.gov/docs/infobank/ for links to cruise data. See Foster and others (1999a, b) for bathymetry and seismic-reflection data archives and McCrory and others (in press b) for sidescan-sonar data archive.

USGS Scientific and Technical Personnel

Patricia McCrory, Menlo Park, California
Thomas O'Brien, Woods Hole, Massachusetts
David Foster, Woods Hole, Massachusetts
William Danforth, Woods Hole, Massachusetts

Scientific Equipment

Navigation—USGS differential GPS navigation was logged using an Ashtech system with Capt'n navigation software.

Bathymetry—The chirp subbottom profiling system, with a 2-7-kHz acoustic source, was mounted on a DataSonics sonar tow fish. The chirp data were recorded with the sidescan-sonar data on a Triton-Elics ISIS digital data acquisition station and processed onboard using Seismic-UNIX software.

Sidescan sonar—The DataSonics SIS-1000 sidescan-sonar system had a swept FM (chirp) acoustic source in the 100-kHz band. Slant-range distances from the ship to towfish were provided by a Benthos 7000 system. The reflectivity data were recorded using a Triton-Elics ISIS digital acquisition system and processed onboard using USGS XSonar software.

Sparker seismic reflection—The sparker system, with a 0.8-kHz acoustic source, was constructed by K. F. Parolski (USGS, Woods Hole) and towed from the fantail of the ship. The seismic data were recorded digitally on a MudSeis acquisition system using a Benthos hydrophone streamer and processed onboard using ProcessSEGY software.

Boomer seismic reflection—A single-plate Seismitec system with a 1-kHz acoustic source and a 500-J Applied Acoustics power supply were towed on a surface catamaran. The seismic data were received on a Geopulse system, digitally recorded on a MudSeis acquisition system using a Benthos hydrophone streamer, and processed onboard using Seismic-UNIX software.

Cruise Instructions

Cruise instructions called for six east-west survey lines and three north-south cross lines in the Cape Flattery, Washington, area. A few hours of sonar and chirp data are missing from the first north-south trackline (Line 5), owing to repair of an intermittent malfunction in the towfish. On the second day of data collection, we switched from the boomer seismic-reflection system to the sparker system to increase subbottom penetration from less than 50 m to greater than 100 m, and the sparker system was deployed for the remainder of the cruise. The east-west lines were shortened to reduce time, and we were able to complete all but the westernmost north-south line. Cruise instructions called for four east-west survey lines and two north-south cross lines in the Queets River, Washington, area as a northward continuation of tracklines collected during the 1997 *McArthur* cruise. All planned tracks were completed except the westernmost north-south trackline. Cruise instruction called for eight northwest-trending survey lines in the Copalis River area as a more detailed site survey of an extensive region of bedrock outcrop observed during the 1997 *McArthur* cruise. These lines were in a region of dense crab-pot strings, and care was taken to maneuver the ship through these obstacles during daylight hours. Nonetheless, the towfish became entangled with crab-pot lines four times during our first day in the Copalis study area. We quickly learned how to avoid the pots. To reduce time, only six northwest-trending lines were run in this area and an additional east-west line was added—from 20 km offshore Grays Harbor into the harbor. The sonar was not deployed on this final trackline (Line 37). In total, geophysical data were collected along 545 km of trackline in water depths ranging from 15 to 170 m (fig. 12).

The DGPS (differential global positioning system) navigation located tracklines with an accuracy of 5-10 m. The sidescan-sonar system was towed about 20 m above the sea floor using a USGS winch from the aft A-frame. This configuration yielded a 375-m swath width and a resolution of 19 cm. The chirp system recorded seismic-reflection data with a resolution of 10-20 cm in the upper 5-15 m of strata beneath the sea floor. The sparker system consisted of 2 electrodes in a wire cage towed from the starboard corner of the fantail. The lower frequency sparker system recorded subbottom data with a resolution of 2-4 m in the upper 150-200 m below the sea floor.

Appendix 3. Previous Cruises in the Area Studied

See URL: walrus.wr.usgs.gov/docs/infobank/ for links to cruise data.

O1067WO—1967 Cruise on UW Ship RV *Oceaneer*

Oceaneer seismic data (fig. 13) were recorded at a 1/2-second scan rate (200-fathom or 365.8-m scale) on an analog (paper) system. Penetration depth for the *Oceaneer* sparker system was about 150 m, with a nominal resolution of 3-4 m at the sea floor and 2-m resolution in the subsurface. Data quality is good.

O174WO—1974 Cruise on Vessel *Oomiak*

Oomiak seismic data (fig. 13) were recorded at a 1/4-second scan rate (187.5-m scale) on an analog (paper) system (Hill and others, 1981). Penetration depth for the *Oomiak* Uniboom system was about 50 m, with nominal resolution of 1-2 m at the bay floor and 1 m or less resolution in the subsurface. *Oomiak* data quality ranged from average to poor.

L376WO; L577WO; L1180WF—1976, 1977, and 1980 Cruises on RV *S.P. Lee*

S.P. Lee seismic data (fig. 13) were recorded at 1/2-second scan rate (375-m scale) on both analog (paper) and digital systems. Penetration depth for the *S.P. Lee* Uniboom system was about 75 m, with a resolution of 1-2 m at the sea floor and 1 m or less in the subsurface. Data quality is average.

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