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Notes
Stonewall anticline: An active fold on the Oregon continental shelf

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ABSTRACT
Stonewall Bank is the site of a growing west-verging anticline striking north-northwest on the continental shelf at 44.5° N, southwest of Newport, Oregon. To the east are Pliocene-Pleistocene strata of the Newport syncline, onlapping eastward against gently west-dipping late Miocene and older rocks of the Oregon Coast Range. Folding of Stonewall anticline results in fine-grained middle Miocene strata being exposed at the sea floor along the anticlinal crest; rocks as old as Eocene are encountered in the Unocal P-0093-1 Grohe well, drilled at the anticlinal crest. Rates of folding are based on deformation of an unconformity between Pliocene and Miocene strata (PM unconformity) and of a stream channel that crossed Stonewall Bank during the last glacial maximum. The bed length of the unconformity is shortened across the Stonewall anticline and adjacent folds by about 400 m between structures west of Stonewall Bank and the Newport syncline. The PM unconformity has a vertical separation of about 1000 m between the anticline and the first syncline to the west. The horizontal shortening and vertical separation imply that Stonewall anticline is underlain by a blind reverse fault. Retrodeforming the PM unconformity shows that this fault dips 65°–70° E. A vertical separation of 1000 m on a fault with this dip yields a slip of 1070–1080 m along the fault. If folding of the PM unconformity is assumed to have begun 2–3 Ma, this would give a long-term slip rate of 0.4–0.6 mm/yr. If most folding began after deposition of the entire Pliocene-Pleistocene sequence, the slip rate would be 1.0–1.1 mm/yr. Stonewall anticline has arched the late Pleistocene lowstand wave abrasion platform since sea level underwent a rapid rise from 14.5 to 8 ka. This arch is crossed by an antecedent stream channel that is 275–550 m wide and is marked by side drainages and cut banks up to 12 m high. Warming of the platform on the eastern limb of the anticline has back-tilted the stream channel eastward toward its present onshore continuation, the Yaquina River. The platform slopes downward 10–13 m westward from the crest of Stonewall anticline. We estimate that the platform stopped abrading when sea level reached about –40 m at 11–12 ka. Assuming that the west slope of the platform is controlled by the same blind fault that produced the west dip of the PM unconformity, the Holocene slip rate on this fault would be 0.9–1.3 mm/yr, comparable to the long-term rates based on a folded unconformity. Here we compare rates of deformation of a warped Holocene stream channel with longer-term rates based on a folded unconformity. Stonewall Bank is a growing anticline that is crossed by a stream channel cut while the bank was subaerially exposed during the latest Pleistocene sea level lowstand about 21 ka. Warming of the thalweg of the stream channel allows us to determine the Holocene uplift rate of the abrasion platform, and the slip rate of an underlying blind reverse fault that generates the anticline. These rates are consistent with long-term rates based on folding of the unconformity between Miocene and Pliocene strata on the continental shelf (PM unconformity). The slip rate of the blind reverse fault can be used to assess the earthquake hazard to adjacent coastal communities.

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Fig. 1. Tectonic map of the Cascadia deformation zone off central Oregon, modified from Tréhu et al. (1995). Onshore geology modified from Walker and MacLeod (1991). Continental-shelf formation contacts shown in Figure 3. Continental-slope and abyssal-plain structures from Goldfinger et al. (1997a, 1997c). Because folds on continental slope are so closely spaced, only anticlines are shown (thin line with cross bar); thrust faults have teeth on hanging-wall side. Symbols: Tt—Siletz River Volcanics; Ty—Tyee Formation; Tb—Eocene basalt; Tn-Ta—Nestucca through Astoria formations, undifferentiated. P-0093 and P-0103 locate two exploratory wells. Inset: Plate-tectonic setting of Figure 1, and location of Cascade volcanoes.

METHODS

We use a modified raster-vector geographic information system (GIS) for geologic mapping (Goldfinger et al., 1997b). The ERDAS IMAGINE system emphasizes raster data manipulation and analysis, with a vector data model compatible with other systems such as ARC/INFO. For the Stonewall Bank area, this system integrates AMS (Alpha Marine Systems) 150 kHz sidescan sonar imagery and bathymetry, digitized shelf bathymetry from a map published by the U. S. Coast & Geodetic Survey (1968), tracks of seismic reflection profiles, locations of bedrock and sediment samples, offshore oil-exploratory well locations, gravity and magnetic anomaly data, seismicity, and interpretation of structural geologic data. Sidescan and submersible dive data were acquired in September, 1993, as part of a National Oceanic and Atmospheric Administration (NOAA) Undersea Research Program cruise aboard the support vessel Cavalier. During this cruise, AMS 150 SI (150 kHz) sidescan sonar imagery was acquired at night to identify targets for a two-person submersible, Delia, to investigate the following day.

The AMS 150 system maps a swath 1 km wide, yielding an across-track pixel size of about 0.5 m. Initial processing of the sidescan data carried out during acquisition consisted of slant-range correction and a time-varying gain correction for spherical spreading of the signal (cf. Johnson and Helferty, 1990). Position of the sidescan towfish was calculated using the Sonar processing package developed by Goldfinger et al. (1997b). Geologic interpretations of the sidescan imagery and the submersible dives were plotted at sea using the Intergraph Microstation Computer-Assisted Design (CAD) system and imported into the GIS. This allowed the observer in the submersible to use a geologic map based on the sidescan sonar image. The ship’s position was monitored by the Global Positioning System (GPS), and the submersible position with respect to the ship was monitored by a transponder on the bridge to plot successive positions of the submersible on the geologic map during the dive. A video camera recorded the view from the submersible during the entire time the submersible was in view of the sea floor; the video was supplemented by still photographs taken from within and outside the submersible. Formation attitudes were taken with a hand-held gyro compass calibrated with the ship’s gyro compass before and after each dive. Rock samples were taken using a sampling claw mounted on the hull of the submersible. The sidescan swath position and geological interpretations were later imported into a GIS vector layer referenced to the position of the ship and towfish.

Petroleum-industry multichannel seismic-reflection profiles were used to map the underlying bedrock structure. Time-depth conversions were based on a synthetic seismogram constructed for the Chevron Nautilus offshore well (Cranswick and Piper, 1992), together with sonic velocity data from the Unocal Grebe well at Stonewall Bank (located in Fig. 1).

TECTONIC SETTING

The Cascadia subduction zone, including the Oregon offshore region, is formed by oblique subduction of the young, warm, heavily-sedimented Juan de Fuca plate beneath the North America plate (Fig. 1) at a convergence rate of 40 mm/yr (DeMets et al., 1990). East of the Juan de Fuca plate, the continental slope is underlain by an accretionary wedge which is actively deforming by folding, thrusting, and strike-slip faulting (MacKay et al., 1992; MacKay, 1995; Goldfinger et al., 1997a). At latitude 45° N, the accretionary wedge is in abrupt contact with the Siletzia terrane, which consists of lower to middle Eocene, highly magnetic, high-velocity oceanic basalt (Siletz River Volcanics; cf. Snavely, 1987) overlain by Eocene and younger marine strata (Tréhu et al., 1994; 1995; Fleming, 1996). The Siletzia terrane, which underlies the continental shelf, Coast Range, and Willamette Valley (Yeats et al.,...
Figure 2. Fence diagram showing stratigraphy of the central Oregon coast and the regions around the Unocal Grebe (P-0093) and Chevron Nautilus (P-0103) wells, located in Figure 1. Heavy lines mark sequence penetrated by wells; other stratigraphy inferred from multichannel seismic lines and sea-floor samples. Onshore stratigraphy from Snavely et al. (1969). Coarse-grained clastic deposits are shaded. CRBG—Columbia River Basalt Group. Basalt at deeper stratigraphic level may be CRBG sills or older flows, such as Yachats Basalt at the bottom of the Nautilus well.

Figure 3. (A) Geologic map of central Oregon continental shelf on a bathymetric base constructed from U.S. Coast and Geodetic Survey (1968), together with geology of adjacent coastal region. Onshore faults of Kelsey et al. (1996) not included. Contour interval 10 m shallower than 200 m below sea level, 50 m deeper than 200 m. Tsr—Siletz River Volcanics; Tt—Tyee Formation; Tn—Nestucca and Yamhill Formations; Tal—Alsea Formation; Tyn—Yaquina Formation; Tm—middle Miocene strata; Tcr—Columbia River Basalt Group (CRBG); Tci—CRBG intrusive phase; QTp—Pliocene-Pleistocene strata, with structure contours on base in kilometers below sea level (dashed lines). (B) Location of multichannel seismic lines (solid line), seafloor sample traverses (heavy solid line) and sidescan track line at Stonewall bank, together with Nautilus and Grebe wells (filled circles with crosses).
1996), was accreted to North America by middle Eocene time (Duncan, 1982; Wells et al., 1984). The Coast Range was upwarped and the Willamette Valley downwarped following emplacement of the middle Miocene Columbia River Basalt Group.

The actively deforming accretionary wedge and easternmost abyssal plain are cut by left-lateral faults striking oblique to the wedge with slip rates of 5–7 mm/yr (Goldfinger et al., 1997a). This suggests that the accretionary wedge is deforming not only by east-west shortening but also by north-south extension and clockwise rotation (McCaffrey and Goldfinger, 1995). In contrast, deformation rates in the Oregon Coast Range and Willamette Valley are very low, even though the Willamette Valley has been struck by historical earthquakes such as the destructive 1993 Scotts Mills earthquake (ML 5.6). Geomorphic evidence of active deformation consists of deformed marine terraces on the southern and central Oregon coast (Kelsey, 1990; McInelly and Kelsey, 1990; Kelsey et al., 1996) and different rates of stream incision, indicating varying rates of uplift in different parts of the central and northern Coast Range (Personius, 1995). In the last 70 years, repeated highway leveling indicates that coastal Oregon has been uplifted at variable rates, from 0 at Newport to 1–2 mm/yr near the Columbia River and 2–4 mm/yr in southern coastal Oregon. This short-term uplift has been interpreted as a response to elastic strain accumulation in the interseismic period between great subduction-zone earthquakes.

Figure 4. East-west multichannel seismic-reflection profile including the Chevron Nautilus well. Top: uninterpreted; bottom: interpreted. Note the unconformity between Pliocene-Pleistocene and Miocene strata (PM unconformity). Strong reflector below the unconformity at the Nautilus well is Columbia River Basalt Group (CRBG); note normal faults offsetting this reflector and others to the east. Siletz River Volcanics after Snavely et al. (1980b) and Tréhu et al. (1995), TWTT—two-way traveltime.
earthquakes (Mitchell et al., 1994; Drager et al., 1994; Hyndman and Wang, 1995).

The Oregon continental shelf is characterized by low relief. Its surface is in large part a broad wave-averaged surficial feature that cuts across strata as young as Quaternary in the Newport syncline. At the maximum Wisconsinan sea-level lowstand of 121±5 m below present sea level (Fairbanks, 1989), marine abrasion was effective at depths as much as 20 m deeper (see discussion below). Sea level rose rapidly during the latest Pleistocene and Holocene, preserving subaerial landforms of the abrasion platform. Tilting, warping, and flexural-slip faulting of the lowstand platform (Goldfinger, 1994) are evidence of Holocene deformation, just as tilted, warped, and faulted marine terraces are evidence of deformation during the late Pleistocene in southern coastal Oregon (Kelley, 1990; Mcintyre and Kelley, 1990) and elsewhere.

**STRATIGRAPHIC SETTING**

Our summary of the stratigraphy of the Oregon Coast Range in the vicinity of Newport is based on Snavely et al., 1964 (Fig. 2). Stratigraphy beneath the continental shelf to the west (Fig. 2) is based on two oil-exploratory wells (Figs. 1 and 3; cf. Snavely et al., 1982; well logs on file at the Oregon Department of Geology and Mineral Industries), multichannel seismic-reflection profiles from the petroleum industry and from Snavely et al. (1980b) and Snavely (1987), and cored offshore strata taken by Shell Oil Company and by one of us (Kuhl). Offshore macrofaunal and lithologic correlations are based on drill cuttings examined by the U. S. Minerals Management Service (unpublished reports) and the petroleum industry. Eocene benthic foraminiferal stages (Ulatisian, Narizian) are those of Mallory (1959), and Oligocene and Miocene stages (Refugian, Zemorian, Saucesian, Relizian, Lusian, Mohrian) are those of Kleinpenn (1938). Study of coccolith assemblages (Bukry and Snavely, 1988) shows that these foraminiferal stages are time-transgressive and document water depth more accurately than age; accordingly, they are indicative of age only in a very general way.

Lower Eocene oceamic basaltic basement, the Siletz River Volcanics of Snavely and Baldwin (1948), is exposed in the Coast Range northeast of Newport (Snavely et al., 1976). It is assumed to extend offshore based on high seismic-wave velocity in a wide-angle seismic-reflection profile (Tribus et al., 1994, 1995) and on aeromagnetic evidence of high magnetic susceptibility beneath the continental shelf (U. S. Geological Survey, 1970; Snavely et al., 1980b, b: Fleming, 1996).

In the Coast Range, the Siletz River Volcanics are overlain by the middle Eocene Tyee Formation, in which deep-marine turbidite sandstone is the dominant lithology (Newport section, Fig. 2). The Tyee is overlain by and interfingers with the middle and early late Eocene Yamhill Formation, which is predominantly siltstone with a few interbeds of sandstone. The Yamhill contains the uppermost Ulatisian and lower Narizian stages (Snavely et al., 1969, 1976, 1980a). The Yamhill is overlain by tuffaceous, deep-marine siltstone of the latest Eocene Nestucca Formation; this locally overlies older strata to rest directly on the Siletz River Volcanics. The Nestucca contains benthic foraminifers from the upper Narizian and lower Refugian stages. Subaerial and submarine basalt flows and breccia interbedded with the Nestucca are correlated to the Yachats Basalt.

The Nestucca Formation is overlain conformably by tuffaceous, shallow-marine siltstone of the Alsea Formation, with Oligocene foraminifers from the upper Refugian and Zemorian stages. The Alsea Formation is overlain by a deltaic sandstone unit of late Oligocene age named the Yaquina Formation (Goodwin, 1972). The Yaquina is overlain by and locally interbedded with the Nye Mudstone, predominantly a deep-marine, massive, organic-rich mudstone and siltstone with interbeds of concretionary dolostone. Foraminifers are correlated to the Saucesian stage, indicating an early Miocene age. The Nye is overlain with slight angular unconformity by the early to middle Miocene Astoria Formation, a predominantly shallow-marine sandstone and dark-gray carbonaceous siltstone (Newport member of Cooper, 1981) with Saucesian (early Miocene) foraminifers.

The Astoria Formation is overlain by basalt flows of the Columbia River Basalt Group (CRBG) forming a strike ridge on the seafloor just west of the coastline (Fig. 3; Snavely et al., 1984). Basalts of the Columbia River Basalt Group, occurring as shallow intrusions and basaltic breccias in lava deltas at Yaquina Head and Seal Rock and as flows farther north at Cape Foulweather, are the youngest rocks exposed onshore in this part of the Coast Range. Two flow units are distal parts of flows in the Columbia Plateau; (1) the basalt of Depoe Bay, correlated to the low-Mg N2 Grande Ronde Basalt, and (2) the overlying basalt of Cape Foulweather, correlated to the Ginkgo Flows of the Frenchman Springs Member of the Wanapum Basalt (Wells et al., 1989). These basalt flows are separated by a poorly-dated massive to thick-bedded shallow-marine arkosic sandstone (sandstone of Whale Cove, Snavely et al., 1969). Offshore, sandstone overlies basalt west of the entrance to Yaquina Bay (Snavely et al., 1980a).

The offshore stratigraphy west of Newport is based on the Unocal Grebe P-0093-1 well at Stonewall Bank (total depth 3051m) and the Chevron Nautilus P-0103-1 well (total depth 3849m) west-northwest of Cape Foulweather (Fig. 1). The oldest strata penetrated by the Grebe well are the Yamhill and Nestucca formations of late middle and late Eocene age (Ulatisian and Narizian stages), with the boundary between these formations uncertainly placed (Fig. 2). The strata that are probably Yamhill Formation are highly tuffaceous, whereas the Nestucca Formation is mainly claystone, lacking the tuffaceous interbeds that are found in the Yaquina Bay section. Basalt at the bottom of the Nautilus well, below the Alsea Formation, is correlated to the Yachats Basalt on the Oregon coast. The Alsea Formation in both wells is tuffaceous claystone and siltstone, as is the Nye Mudstone, and it contains foraminifers of latest Eocene and Oligocene (Refugian and Zemorian) age. The Alsea is overlain by the Nye Mudstone of early Miocene (Saucesian) age. The deliše sandstone of the Yaquina Formation and the shallow-marine sandstone of the Astoria Formation pinch out offshore toward the two wells (Goodwin, 1972; Cooper, 1981). Benthic foraminifers show that the entire Eocene to early Miocene sequence was deposited in bathyal water depths.

Basalt of the Grand Ronde Member of the Columbia River Basalt Group (Snavely et al., 1980b) was encountered in the Chevron Nautilus well, where a vertical synthetic seismogram from Cranwike and Piper (1992) indicates correlation with a prominent reflector in a seismic profile through the well site (Fig. 4). This same reflector is overlapped unconformably by Pliocene-Pleistocene strata east of the Grebe well (Fig. 5). In the Nautilus well, the basalt is both underlain and overlain by dark gray siltstone with scattered turbidite sandstone interbeds, abundant foraminifers indicate a middle Miocene age (Relizian stage) and upper bathyal water depths. Refugian strata were also dredged in the core of an anticline at Stonewall Bank, but were not encountered in the Grebe well. In a syncline east of the Nautilus well, the Relizian strata are overlain by more than 800 m of late Miocene strata (Mohrian stage) of similar lithology. These predominantly fine-grained strata have no counterparts onshore. Middle Miocene fine-grained strata cropping out on the seafloor near the coastline were deposited in 45–90 m water depths in comparison with 180–600 m water depths in the Nautilus well and north of Stonewall Bank, suggesting that the location of the Miocene shoreline was close to (probably slightly east of) that of today.

Unconformably overlaid by Miocene sequence are claystone and siltstone with shell fragments and fine-grained to very fine-grained sandstone deposited in upper bathyal water depths. Benthic foraminiferal assemblages have been correlated to the southern California Repetian and Venetian stages of Nathan (1952), considered to be Pliocene, but these stages primarily reflect wa-
Figure 5. East-west multichannel seismic-reflection profile across Stonewall anticline and Unocal Grebe well. Miocene-Pliocene contact (PM unconformity) based on bottom samples. East-dipping reflectors below 3 s two-way traveltime (TWTT) are interpreted as Alsea-Nestucca contact; these reflectors are offset by a fault, down to the west, or by a steep fold. Top: uninterpreted; bottom: interpreted. CRBG—Columbia River Basalt Group.
Figure 6. Geologic cross section from Stonewall anticline to the Oregon coast along the seismic line of Figure 5. Blind reverse fault is assumed to cut the Alsea-Nestucca contact but not the PM unconformity. YAQ—Yaquina Formation; CRBG—basalt flow of Columbia River Basalt Group; P/M—PM unconformity. Benthic microfaunal determinations from dart core samples: MM—middle Miocene; Plio—Pliocene; Quat.—Quaternary; ON—outer neritic water depths of deposition; UB—upper bathyal water depths. No vertical exaggeration.

Figure 7. Simplified cross section of Figure 6 with calculation of displacement on blind fault as expressed by deformation of the PM unconformity. Seismic data permit the Alsea-Nestucca contact to be deformed along the steep limb of a fold or along a fault. See text for calculations. A, B, C, D, E, F discussed in text.
ter depth and have little age significance (cf. Blake, 1991). A similar sequence cored at DSDP Site 176 in 193 m water depth, southwest of the mouth of the Columbia River, consisted of Pleistocene greenish-gray clayey silt overlying, with angular unconformity, an indurated olive-gray shale (Kulm et al., 1973). The shale contains a Pliocene benthic foraminiferal assemblage indicating water depths ≤200 m. However, data from the same core were assigned by Schrader (1973) to North Pacific Diatom Zone III of the lower Pleistocene. Their dating problems preclude an age designation more precise than post-Miocene for these strata in the Newport syncline. Like the underlying Miocene strata, this sequence has no onshore equivalent at this latitude, but in lithology and fossils, it resembles the Rio Dell Formation of the Eel River basin in northwestern California (Piper et al., 1976). Water depths determined from foraminiferal assemblages in the Namutsu well and at Stonewall Bank are 180–600 m, shallowing up-section to 45–90 m, indicating that rates of sedimentation became greater than rates of subsidence. Like the Miocene sequence, the eastward shallowing based on benthic foraminifers suggests a Pliocene coastline close to that of the present day. However, the continental shelf was accumulating sediments during the Pliocene and early Pleistocene, whereas sediments today are bypassing the shelf and are accumulating in large part on the continental slope and abyssal plain. Data from piston cores and gravity cores show that the Pliocene-Pleistocene strata onlap eastward against the west-sloping PM unconformity, although they also may crop out on the seafloor. Piston cores from the continental slope and observations from Delius show that stiff, light gray Pleistocene clay is overlain by a thin veneer of greenish-brown Holocene mud, with the color change occurring approximately at the base of the Holocene (Barnard and McMannis, 1973).

STRUCTURE OF STONEWALL ANTICLINE AND NEWPORT SYNCLINE

West of the coastline, strata are folded in the Newport syncline and Stonewall anticline. Miocene and older strata were broadly folded and offset by normal faults prior to deposition of the Pliocene-Pleistocene sequence (Fig. 4). The largest of these normal faults, east of the syncline, displaces the Columbia River Basalt Group about 500 m, down to the east. East of the Stonewall anticline, the Pliocene-Pleistocene sequence overlaps the CRBG and older middle Miocene strata, indicating a gentle east dip of Eocene to Miocene strata prior to Pliocene deposition (Figs. 5, 6). Two shallow synclines west of Stonewall Bank show evidence of Miocene growth, and an abrupt westward increase in the thickness of post-Nes-tucca strata beneath Stonewall Bank is evidence for pre-Pliocene displacement on the Stonewall fault, discussed further below. These relations require a westward change, prior to Pliocene deposition, from normal faulting near the coast to folding and reverse faulting farther offshore.

Faulted and folded Miocene and older strata are overlain by Pliocene-Pleistocene strata along an unconformity (PM unconformity) with relief of up to 60 m due to Miocene folds that were not completely beveled by erosion. Eastward onlap of the Pliocene-Pleistocene strata against Miocene rocks indicates that this erosion surface sloped west about 1°, and locally as much as 3°, prior to Pliocene deposition. Strata on both sides of the unconformity were deposited in water depths greater than 90 m; however, the low relief and gentle westward slope of this unconformity suggest that it may have been cut by wave abrasion.

The Pliocene-Pleistocene strata are broadly folded into the Newport syncline (Figs. 5, 6). Structure contours based on a grid of seismic-reflection profiles (located in Fig. 3B) show that this syncline plunges gently about N10°W, and it may have been eroded from the crest of the anticline. East of the coastline, strata are folded in the Newport syncline, with horizontal shortening of about 400 m (A–C) and uplift of 1500 m, as shown by seismic-reflection profiles through the site of the Green Bank well (Fig. 5). The Stonewall anticline is no more than 25 km long and trends north-northwest (Fig. 3A).

A seismic-reflection profile shows that the east flank of the Stonewall anticline dips 15–18°, and the west flank dips 25°, evidence that this anticline is seaward-vergent. A multichannel seismic-reflection profile through the site of the Grebe well shows that 1500 ± 500 m of strata, including about 400 m of Miocene strata, have been eroded from the crest of the anticline. North of the Grebe well about 14 km, where the middle-Miocene outcrop on the seafloor is broad, 4000 ± 1000 m of strata were removed by erosion after deposition of the Pliocene-Pleistocene sequence. Uncertainty in the amount of strata eroded from the anticlinal crest allows the possibility of some uplift during deposition, and the possibility that the youngest strata in the Newport syncline may have been overlain by still younger sediments, now removed by erosion.

LONG-TERM SLIP RATE ACROSS THE STONEWALL ANTICLINE

Can we retrodeform the PM unconformity to work out long-term deformation rates across the Stonewall anticline? The seismic profiles and outcrop pattern of the PM unconformity show that the Stonewall anticline is not faulted at the surface. We interpret the east flank of the Newport syncline as part of the broad warp of the Coast Range (Fig. 7) and attribute the west flank of the syncline, Stonewall anticline, and structures farther west as folds generated by east-dipping blind reverse faults. Thus the folding of the PM unconformity can be used to estimate the long-term slip rate on the blind fault beneath Stonewall anticline. Because we are not sure of the age of the post-Miocene strata other than being of “Pliocene–Pleistocene” age, the long-term slip rate on the blind fault is uncertain. The folded strata could be entirely Pliocene, or their age could extend well into the Pleistocene. We assume an age range from 4 Ma, or early Pliocene (late Repetto of southern California), to 1 Ma, or early Pleistocene.

The anticline began to form prior to development of the PM unconformity, as shown by pre-Pliocene dips in Figures 5 and 6. However, the absence of onlap of Pliocene strata against the PM unconformity indicates that the unconformity had no relief across the anticline. Furthermore, multichannel seismic profiles show that the Stonewall anticline was renewed after one-third to one-half of the Pliocene-Pleistocene section does not thin, and actually thickens locally toward the anticline from the east. We interpret these relations as evidence that growth of the Stonewall anticline was renewed after one third to one-half of the Pliocene-Pleistocene strata had been deposited. Based on an age range of 1 to 4 Ma for the Pliocene-Pleistocene sequence, the more recent folding probably began about 2 to 3 Ma.

To work out a slip rate on the blind fault, we first retrodeform the PM unconformity to a planar surface, sloping about 1° west to accommodate the bathymetry of Pliocene strata above the unconformity (Fig. 7). Point A on this unconformity prior to folding (Fig. 7) would be displaced by folding to point B relative to the center of the Newport syncline, with horizontal shortening of about 400 m (A–C) and uplift of about 300 m (B–C). Nearly all of the horizontal shortening takes place across Stonewall anticline. To determine the dip of the blind reverse fault generating the Stonewall anticline, we use the horizontal shortening, 400 m (E–F), and the difference in altitude, 1000 m, of the PM unconformity between the eroded crest of the anticline and the preserved trough of the first syncline to the west (D–E, Fig.
The age of initiation of pre-Pliocene deformation of Stonewall anticline is too poorly constrained to determine a slip rate based on the offset of the Alsea-Nestucca contact. The similar amplitudes of the large synclines east and west of Stonewall anticline raise the possibility that the blind reverse faults generating the folds could flatten into a low-dipping thrust in line-to-medium-grained strata below the Alsea-Nestucca contact but above the top of Siletz River Volcanics. This low-dipping decollement could form, even the upper part, where there is evidence of minor thinning toward Stonewall anticline. If we assume that all displacement on this blind fault would be about 1.0–1.1 mm/yr.

The difference in amplitude between the forelimb and backlimb of Stonewall anticline may be due to the presence of another blind reverse fault west of the two small synclines at the west end of the cross section shown in Figure 6. We do not work out the slip rate of the fault generating this structure because we are unable to compare long-term deformation rates with Holocene rates, as we can at Stonewall Bank.

Another prominent reflector, probably the Alsea-Nestucca contact, dips 35°E near the Grebe well and about 15°E on the west flank of the surface anticline (Figs. 5, 6). This reflector has a vertical separation of about 2000 m at the anticline, which is twice that of the amplitude of the folded PM unconformity at the Stonewall Bank anticline. The vertical separation of 2000 m is about twice that of the amplitudes of the folding and faulting prior to Pliocene deposition. The age of initiation of pre-Pliocene deformation of Stonewall anticline is too poorly constrained to determine a slip rate based on the offset of the Alsea-Nestucca contact. The similar amplitudes of the large synclines east and west of Stonewall anticline raise the possibility that the blind reverse faults generating the folds could flatten into a low-dipping thrust in line-to-medium-grained strata below the Alsea-Nestucca contact but above the top of Siletz River Volcanics. This low-dipping decollement could form, even the upper part, where there is evidence of minor thinning toward Stonewall anticline. If we assume that all displacement on this blind fault would be about 1.0–1.1 mm/yr.

The difference in amplitude between the forelimb and backlimb of Stonewall anticline may be due to the presence of another blind reverse fault west of the two small synclines at the west end of the cross section shown in Figure 6. We do not work out the slip rate of the fault generating this structure because we are unable to compare long-term deformation rates with Holocene rates, as we can at Stonewall Bank. Another prominent reflector, probably the Alsea-Nestucca contact, dips 35°E near the Grebe well and about 15°E on the west flank of the surface anticline (Figs. 5, 6). This reflector has a vertical separation of about 2000 m at the anticline, which is twice that of the amplitude of the folded PM unconformity at the Stonewall Bank anticline. The vertical separation of 2000 m is about twice that of the amplitudes of the folding and faulting prior to Pliocene deposition.
wave motion of long-period storm waves may form ripples in water depths of at least 150 m (Komar et al. 1972; Kulm et al., 1975) and transport sand across the shelf by unidirectional currents ranging up to 70 cm/sec (Smith and Hopkins, 1972). However, most sand is too fine grained to abrade underlying consolidated sedimentary bedrock, as would the coarser-grained sands which occur on the outer shelf as relict deposits from the LGM lowstand (Scheidegger, et al., 1971; Kulm et al., 1975). However, bottom currents laden with sand transported from Coast Range rivers and localized by topographic lows on the shelf may have eroded the shelf in the initial stages of submarine-canyon formation.

We suggest that deformation of the abrasion platform at the LGM lowstand at 121 m water depth began to be preserved at about 14.5 ka, during (or slightly prior to) mwp IA, when sea level had already risen about 20 m. A channel on Stonewall Bank, discussed below, now at 60-70 m water depth, would have been removed from wave action when sea level was 50-60 m higher about 11-12 ka (see inset, Figure 8). This assumes that no differential tilting occurred from 14.5 to 11-12 ka between Stonewall Bank and the LGM shoreline. However, we argue below that this surface has been uplifted about 15 m in the last 11–12 k.y. Because sea level was rising very rapidly when the shoreline crossed Stonewall Bank, adding Holocene uplift makes the age of final preservation of the platform at Stonewall Bank only slightly older than assuming no uplift.

For purposes of slip-rate determination, we assume that Stonewall Bank was no longer affected by marine abrasion after 11–0.5 ka, using the Barbados sea level curve (inset, Fig. 8). For the last 6–8 k.y., global sea level has risen from ~10 m to the present level at a rate of about 1–2 mm/yr (Fairbanks, 1989; inset, Fig. 8). This timing is consistent with that derived from a Holocene transgressive sequence in Alsea Bay, east-southeast of Stonewall Bank (Peterson et al., 1990, 1993), and Peltier (1994). mwp-IA marks a time of extremely rapid sea-level rise.

We turn now to the suitability of the abrasion platform for measuring deformation on the continental shelf. East-west profiles (Fig. 8) across the Stonewall anticline at 44°30′ and 44°38′N, constructed from the bathymetric map along tracklines of multichannel seismic-reflection profiles (Fig. 3), show that the platform rises westward as much as 30 m between the trough of Newport syncline and the crest of Stonewall anticline, in part controlled by a strike ridge of Pliocene–Pleistocene strata east of the axis of Stonewall anticline. The anticlinal crest shows no evidence of upwarping along the profile at 44°38′N, probably because it crops out at a depth shallower than the LGM lowstand, and a tectonic signal cannot be distinguished from the differential erosion of the strike ridge. However, in the profile at 44°44′N, the anticlinal crest crops out at ~150 m below the LGM lowstand, and is expressed topographically as a ridge trending north-northwest (Fig. 8). Further northwest, the...
submarine slope is dissected by erosion (Fig. 3) due to bottom currents laden with sediment derived probably from the Coast Range via the Yaquina and Alsea rivers. Still farther north, the anticline cannot be distinguished among several folds of relatively short wavelength in a multi-channel profile at 44°52' (Tréhu et al., 1995).

In summary, the abrasion platform is not a smooth, beveled surface. It contains rocky, formerly subaerial highlands like the strike ridges east of the Stonewall anticlinal axis, reaching heights of –7 m (7 m below sea level). These highlands were dominated by differential subaerial erosion. In addition, the platform appears to have undergone further scouring by sediment-charged bottom currents following the axis of the Newport syncline and dissecting the upper continental slope at the north end of Stonewall anticline. Finally, the abrasion platform is a composite of many Pleistocene lowstands. Earlier Pleistocene lowstand platforms could have been downdropped enough to be removed from subsequent wave abrasion, so that only the shallower part of the platform was abraded during the LGM lowstand. Accordingly, it is possible to document deformation of the abrasion platform in a general way, but not quantitative enough to determine deformation rates.

Fortunately, Stonewall Bank is crossed by an antecedent stream (Figs. 9, 10), originally subaerial, at 44°5°N. We traced the stream channel with AMS 150 kHz sidescan sonar for a distance of 5.7 km, and expression on the –60 and –70 m contours on the 1:250 000-scale map (Fig. 3) permits the channel to be mapped for an additional 2.3 km, for a total distance of 8 km. The location and orientation of the channel, S65°W, suggests that it is the latest Pleistocene lowstand continuation of the Yaquina River (Figs. 1, 3; cf. Kulm, 1965). The stream channel cuts across Pliocene strike ridges, and the floodplain increases in width downstream from 275 m to 550 m. There is at least one low terrace above the main channel, and side drainages enter the main channel from both sides. swath bathymetry acquired with the AMS 150 sidescan sonar shows that the terrace riser north of the channel is about 12 m high, with a riser slope of 18° to 42°. Dohu submersible dives showed that the channel is now partly covered by Holocene mud (Fig. 9). Bathymetry of the channel accompanying the AMS 150 sidescan sonar shows a slope from about –65 m at the crest of Stonewall anticline eastward to –72 m. Bathymetry from the 1:250 000 map shows the channel expressed farther east by the –80 m contour, although the map contours shown on Figure 3 are less accurate; these show the channel as shallow as –60 m at the crest of the anticline, 5 m shallower than the AMS 150 bathymetry. In addition, the sidescan sonar image (Fig. 9) confirms observations from submersible dives that mud (dark, caused by low reflectance) is ponding behind the crest of the anticline, where the channel shows higher reflectance.

The observation that the stream channel at Stonewall Bank is the downstream continuation of the Yaquina River led us to construct a longitudinal profile from the upper reaches of the Yaquina River across Stonewall Bank (Fig. 11). The profile of the Yaquina River surface from Niem (1976) shows that the river is graded to present sea level, as are other west-flowing streams in the Oregon Coast Range (Niem, 1976; Personius, 1995). The thalweg of the river is at a water depth of nearly 4 m where the river is tidal (Kulm, 1965). To this must be added an unknown thick-
ness of sediments deposited during late Holocene sea level rise, as documented by geotechnical boreholes drilled in the 1930s for the Yaquina Bay Bridge (logs stored at Oregon Department of Transportation) and high-resolution seismic reflection profiles obtained by the U.S. Army Corps of Engineers immediately west of the coastline (Sylwester et al., 1996). For the lower Umpqua River, for example, these sediments are more than 40 m thick (Curtiss et al., 1984), and the thalweg of the Alsea River, buried by Holocene transgressive sediments, is at a depth of –55 m (Peterson et al., 1984). We assume in Figure 11 that the buried thalweg of the Yaquina River is at the same depth as that of the Alsea River.

The bathymetric contours from 0 to –50 m are relatively straight, parallel to the coastline, and show no evidence of the channels of west-flowing rivers (Fig. 3). This indicates that the nearshore bathymetry is strongly affected by late Holocene wave abrasion. Wave abrasion at shallow depths and downcutting by sediment-charged bottom currents in the trough of the Newport syncline prevent the direct mapping of the Yaquina River thalweg west to the thalweg of the Stonewall Bank channel. We estimate that the difference in elevation of the thalweg between the crest of the anticline and trough of the Newport syncline was no more than 15 m prior to Holocene erosion.

West of Stonewall anticline, the topography slopes westward to –72 m in the AMS 150 bathymetry. We estimate the depth to be about –75 m and no more than –78 m in the synclinal trough to the west, based on the depth recorder accompanying the east-west multichannel seismic line crossing Stonewall Bank and on the 1:250,000 bathymetric map. Although the depth of the channel in the synclinal trough to the west was not mapped directly, the difference of 10–13 m between the thalweg on the syncline to the west may be used as the vertical component of Holocene deformation on the Stonewall blind reverse fault. Adding the tidal gradient of Coast Range rivers of 0.1–0.2 m/km (Personius, 1995) would reduce this figure by less than one meter.

If the 10–13 m difference were due to a blind reverse fault dipping 65–70°E, with the fault dip based on retrodeforming the PM unconformity, the slip rate would be 0.9–1.3 mm/yr. The range in slip rate values is due to uncertainty in the time (11–12 ka) and depth that wave action ceased planing off the abrasion platform and in uncertainty in the depth of the thalweg in the syncline to the west.

Why is the estimated depth of the thalweg of the stream shallower west of the Stonewall anticline than it is in the Newport syncline? As discussed above for the PM unconformity, it is probably related to another east-dipping blind-reverse fault west of Stonewall Bank, west of which is a syncline on the upper continental slope containing a thickness of Pliocene–Pleistocene strata almost as great as the thickness in the Newport syncline (Fig. 5). We did not determine the shortening rate across these structures because we cannot compare this rate with a deformed Holocene feature, such as a stream channel, as we can at Stonewall Bank.

In summary, displacement rates of the blind
reverse fault deforming the PM unconformity are 0.4–0.6 mm/yr if folding began during sedimentation and 1.0–1.1 mm/yr if folding began after sedimentation. Displacement rates on the fault deforming the thalweg of the stream channel crossing Stonewall Bank are 0.9–1.3 mm/yr, if the fault dips 65°–70° and the channel is 11–12 ka in age. Considering the uncertainties in each estimate, we conclude that the long-term and short-term slip rates are similar.

DISCUSSION

Origin of the Stonewall Anticline

The Stonewall anticline, considered here as controlled by a blind fault with a slip rate close to 1 mm/yr, is only 25 km long; it is not found on multichannel seismic lines at latitude 44°22′ N and 44°52′ N. Its low length-to-width ratio (aspect ratio) and the steep dip of the blind fault based on deformation of the PM unconformity suggest that this is not part of a thin-skinned fold-thrust belt. The fold is east of the probable termination of the Daisy Bank left-lateral fault (Fig. 1), with a slip rate of 5.7–2.0 mm/yr, and southeast of the known termination of the Wecoma fault, with a left-slip rate, also measured at the deformation front, of 8.5 ± 2.0 mm/yr (Goldfinger et al., 1997c).

The continental slope with the left-slip faults is characterized by folds of relatively short wavelength, in contrast to the long-wavelength folds of the continental shelf (Fig. 1). The long-wavelength folds of the shelf may reflect the underlying rigid basement of Siletz River Volcanics, in contrast to the continental slope, where this basement is absent (Tréhu et al., 1994; 1995; Fleming, 1996), although the syncline on the upper continental slope west of Stonewall Bank (Fig. 5) has approximately the same wavelength as the Newport syncline. The western boundary of Siletzia should be close to the Stonewall anticline, based on criteria for Siletz River basement outlined above (Fleming, 1996). The change in structural style may be expressed in the short-wavelength folds immediately west of the Stonewall anticline (Fig. 1).

AMS 150 sidescan sonar at Stonewall Bank shows a pronounced joint pattern striking N65°–70° W, approximately parallel to the Daisy Bank and Wecoma faults, but oblique to the north-northwest strike of bedding (Fig. 10). However, there is no evidence of strike-slip offset of strike ridges along any of these joint surfaces. The joints could be a reflection of simple shear distributed across the entire Stonewall structure, possibly because so much of the underlying section consists of fine-grained strata.

Seismic Hazard to Coastal Communities

The Stonewall anticline is about 30 km west-southwest of the City of Newport. If the entire fault (25 km long × 20 km downdip width) ruptured in an earthquake with average slip of 1 m, regression curves of Wells and Coppersmith (1994) suggest that an earthquake of Mw = 6.8 ± 0.25 would be generated. An average recurrence interval of about 1000 yr is based on slip of 1 m divided by the slip rate of 1 mm/yr, with at least a 50 percent uncertainty in slip rate and slip per event. Using attenuation relations for rock sites affected by crustal earthquakes in western North America (Geomatic Consultants, 1995) an earthquake of this magnitude would produce peak ground accelerations up to 0.15–0.2 g at 30 km source-to-site distance. This hazard assessment assumes that all slip is released by earthquakes, although there is no paleoseismological evidence to indicate
whether all slip is released seismically or whether part (or all) is released aseismically. However, if Stonewall anticline overlies thick Eocene basaltic basement of the Siletz River Volcanics (Tebba et al., 1994; 1995; Stuevly et al., 1980; Fleming, 1996), it would seem likely that this basement could store elastic strain energy and generate earthquakes.

THE CONTINENTAL SHELF ABRASION PLATFORM AS A STRAIN MARKER

The back-tailing of the thalweg of the antecedent stream channel at Stonewall Bank toward its onshore continuation, the Yaquina River, is evidence that the thalweg of this stream channel can be used as a strain marker. To a lesser extent, the erosion surface of the continental shelf itself can also be used as an indication of Holocene deformation, but only qualitatively unless differential subaerial erosion, Holocene scouring by sediment-charged bottom currents that could eventually carve submarine canyons. The potential of mapping lowstand features may be even greater because of the episodic nature of sea-level rise, including the near-synoptically occurring 14.5 ka. Well-dated pulses in sea-level rise permit the measurement of deformation rates after the removal from wave action of the abrasion platform and the thalwegs of rivers crossing it.

We have determined the slip rate on a blind reverse fault beneath the continental shelf based on deformation of a horizon several million years old and a formerly-subaerial Holocene river channel now submerged in 65 m of water. The new techniques of determining slip rates at sea have widespread application on continental shelves around the world, indicating that the continental shelves themselves may serve as a monitor of crustal deformation.

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