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ACTIVE DEFORMATION OF THE PACIFIC NORTHWEST CONTINENTAL MARGIN

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Abstract. Tilted and uplifted marine terraces in southern Oregon show progressive landward tilting of the coastal ranges at about 5 - 16 x 10^{-8} rad. yr^{-1} for the last 0.25 m.y. Tide gauges in Washington and British Columbia, and ten resurveyed leveling lines running inland from the coast, indicate contemporary landward (down-to-the-east) tilt rates of about 1-12 x 10^{-8} rad. yr^{-1} averaged over periods of from 10 to 50 years. The leveling lines traverse, and the terraces cut across, dipping Cenozoic strata: Pleistocene (dips to 3°), Mio-Pliocene (dips to 30°) and Eocene (dips to 60°). Southern Oregon from Cape Blanco to the Siletz River shows geodetic or terrace tilting in the same direction as the underlying stratal dips. Hence present-day deformation continues past deformation of the coastal ranges and is most likely related to active subduction of the Juan de Fuca plate. The steep stratal dips, lack of major active faults and historic earthquakes, and presence of very young bedding-plane faults suggest that much of the onshore deformation and shortening within the overlying North American plate is taken up by folding rather than thrust faulting. Present shortening rates across north-trending folds near the coast are between 0.02 and 1.9 x 10^{-7} yr^{-1}. The rate of shortening decreases rapidly eastwards from the Juan de Fuca - North American plate boundary. A total of about 25 mm yr^{-1} of permanent shortening could be occurring within the North American plate; most of it in the westernmost 40 km. The landward tilt and shortening rates are similar to those above many other subduction zones that have experienced great thrust earthquakes. While a high strain rate measured near Seattle, Washington, has been interpreted as elastic strain accumulation before a thrust earthquake, the low level of historic seismicity and the similarity of short- and long-term deformation rates suggest alternatively that the subduction beneath Washington is aseismic. The issue has considerable implications for seismic hazard evaluation in the Pacific Northwest and could be resolved by a search for the effects (or lack of effects) of prehistoric great earthquakes.

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INTRODUCTION

The Pacific coast of Oregon and Washington lies close to the boundary between the North American and Juan de Fuca plates. South of Vancouver Island marine magnetic anomalies indicate present convergence at 35 mm yr^{-1} between the two plates [Riddihough and Hyndman, 1976; Riddihough, 1977]. The calc-alkaline volcanic rocks of the Cascade Range, the morphology of the sea floor and the deformation of young offshore sediments, all suggest subduction of the Juan de Fuca plate beneath the North American plate, but present-day underthrusting is unproven.

The case for active subduction has been well presented by Riddihough and Hyndman [1976], who explained the absence of a deep, well-defined Benioff Zone in terms of the low plate-convergence rate and the young age (and hence hot and more plastic nature) of the underthrust sea floor. Recently, Ando and Balazs [1979] showed a simple landward tilt for the Olympic Peninsula of northwest Washington from both repeated geodetic level surveys and tide gauge records. They compared the simple tilt with the well-documented and complex pattern of pre-, co-, and post-seismic movements observed in Japan, and interpreted the Washington data to indicate contemporary, though aseismic, subduction.

An earlier paper [Reilinger and Adams, 1982], closely related to the present one, analyzed leveling and tide gauge data for Oregon and Washington, showed that there was a well-defined landward tilting of the 600-km-long coastal ranges at about 3×10^{-8} rad. yr^{-1} and noted that the tilting agreed with studies in progress on the deformation rate of marine terraces in Oregon (now published in the present paper). Although the agreement between long- and short-term deformation rates suggests that the subduction is occurring by aseismic creep as hypothesized by Ando and Balazs, recent measurements of horizontal deformation on northwest Washington, interpreted as elastic strain accumulation by Savage et al. [1981], seem to be incompatible with aseismic subduction. Further contemporary deformation studies seemed needed before the correct alternative--aseismic subduction, or subduction with large, infrequent thrust earthquakes--might be determined.

In this paper I describe coastal terraces in Oregon and Washington and quantify their deformation. By analysis of east-west geodetic leveling lines and tide gauge records I show that contemporary deformation rates are similar to the rates for the last 0.25 m.y. and by extrapolation of terrace tilts to the dip on the underlying strata that the pattern of deformation in the coastal ranges of the Pacific Northwest has remained substantially the same over the last half million years. Both the rates and the persistance of deformation patterns argue that subduction continues beneath Oregon and Washington despite the lack of shallow thrust earthquakes.

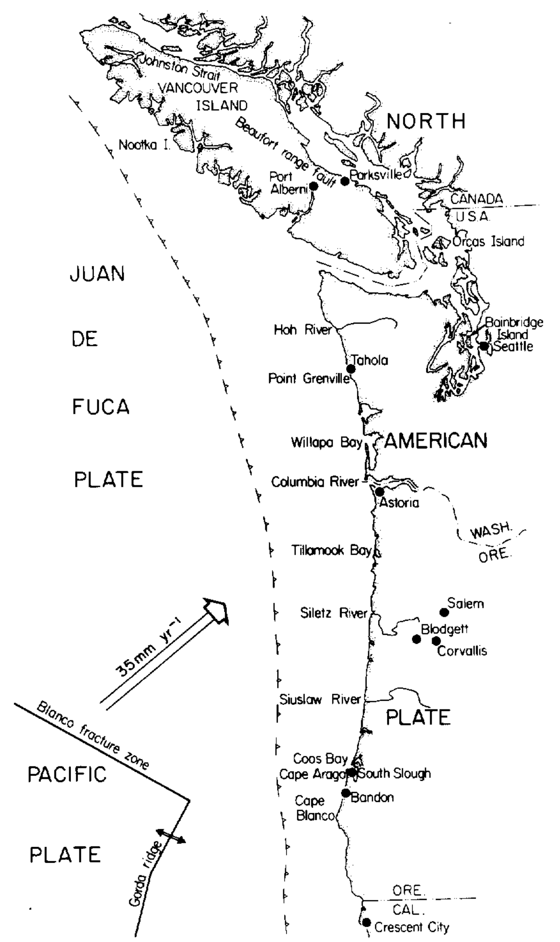

Fig. 1. Map of Pacific Northwest coast showing localities mentioned in the text, inferred trench position (broken barbed line), and compression vector representing motion of the Juan de Fuca plate relative to North America [Riddihough, 1977].

GEOMORPHIC EVIDENCE FOR LANDWARD TILT OF THE COASTAL RANGES

Marine Terraces in Oregon and Washington

Along much of the Oregon coast (Figure 1) there are emerged wave-cut surfaces and terraces that demonstrate coastal uplift. The highest and most spectacular terraces are to the south between Cape Arago and Cape Blanco, and have attracted considerable attention since the 1890's [Griggs, 1945, Baldwin, 1966]. Cape Blanco forms a salient that approaches to within 60 km of the plate boundary (Figure 1). Consequently, onshore rates are highest here and decrease northwards as the coast and boundary diverge.

Griggs [1945, plate 42] correlated and mapped the extent of the three lowest terraces along the coastal strip between Coos Bay and Bandon. These terraces (from youngest to oldest) are Whisky Run, Pioneer, and Seven Devils. Griggs also mapped the extent of higher gravels which belong to an older surface, unnamed by him, but in this paper termed the Metcalf terrace (named for the Metcalf Ranch shown on Griggs' plate 42). Although four distinct terraces are recognized in the coastal strip, they have not been distinguished just 6 km to the east, on the east side of South Slough, probably because there the uplift rate has been too low to preserve them as separate terraces.

At Cape Blanco there is a well developed terrace considered by many authors to correlate with the Pioneer terrace to the north, presumably on the basis of height and continuity along the coast and degree of weathering in the terrace cover beds. Higher terraces are also recognized, but lower terraces are absent at the cape itself [Beaulieu and Hughes, 1976], although they may be represented by fluvial terraces along the Elk and Sixes Rivers [Janda, 1972]. From Cape Blanco south to the California border there are terraces mapped as the Pioneer and higher terraces [Beaulieu and Hughes, 1976].

North of Coos Bay, the coast is low and sandy and may be downwarped. From the Siuslaw River north to Tillamook Bay, geological maps by the State of Oregon show there are discontinuous terraces below 150 m, some as small remnants and others several kilometers wide. The elevation of the best developed terrace (tentatively considered to be 100,000 yr old, see below) decreases from about 50 m at the Siuslaw River to about 20 m at Tillamook Head. This decrease away from the Blanco salient reflects the increasing distance of the coast from the plate boundary and is consistent with the overall landward tilt of the coast ranges documented below.

In Willapa Bay, Washington, 25 km north of the Columbia River, a 13 m terrace is underlain by esturine sediments containing shells that by amino acid racemization

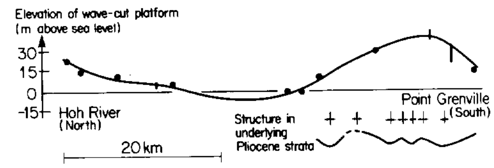

Fig. 2. North-south section along the central part of the Washington coast between Point Grenville and Hoh River, showing varied elevation of the wave-cut surface [after Rau, 1973]. Also shown is schematic structure in the Quinault Formation (Pliocene) at approximately equal vertical and horizontal scales.

dating are 120,000 ± 40,000 yr old [Kvenvolden et al., 1979] and that indicate an uplift rate of less than 0.1 mm yr^{-1} for the coast. Further north, between Point Grenville and Hoh River, a wave-cut surface possibly 82,000 yr old is elevated 35 m above sea level in some places, and at others it is submerged (Figure 2; Rau [1973]) indicating differential uplift along the coast.

In the Puget lowlands and in British Columbia the last glaciation would have removed any emerged terraces, and so the evidence for long-term uplift or subsidence. However, some parts of the lowlands are rapidly emergent, most probably as a consequence of glacio-isostatic rebound [Mathews et al., 1970]. A shell deposit on Orcas Island (48.69°N 122.93°W) has been radiocarbon dated at 12,350 ± 400 years [Wehmiller et al., 1977], and gives a mean uplift rate of 4 mm yr^{-1}. The present rate is perhaps far slower than it was immediately after deglaciation. More significantly, a 3260 ± 80 year radiocarbon date on shells at Bainbridge Island [Wehmiller et al., 1977] indicates 6 m of land emergence (~2 mm yr^{-1}) for a period well after the most immediate deglacial effects. Bainbridge Island is within 20 km of the Seattle tide gauge which indicates present land subsidence (see below), and the difference between these long and short-term rates has not yet been explained.

Age of Terraces in Southern Oregon

Critical to the calculation of tilt and uplift rates is the knowledge of the age of the wave-cut benches beneath the terraces. The benches are cut during periods of relatively high sea level when the rate

TABLE 1. Uplift and Tilt of Marine Terraces Near Coos Bay, Oregon

Terrace	Assigned Age, yr	Elevation,[a] m	Eustatic Correction, m	Uplift[a] Rate, mm yr^{-1}	Tilt, rad.	Average Tilt Rate, 10^{-8} rad. yr^{-1}
A Cape Arago						
Metcalf	230,000	150	0	0.65	0.019	8
Seven Devils	124,000	85	-6	0.64	0.006	5
Pioneer	103,000	55	+15	0.68	0.005	5
Whisky Run	83,000	21	+13	0.41	0.004	5
B Trig 691						
Metcalf	230,000	230	0	1.0	0.037	16
Seven Devils	124,000	123	-6	0.94	0.017	14
Pioneer	103,000	44	+15	0.57	0.0098	10

[a]Relative to sea level at longitude 124.376 degrees west.

of sea level rise approximates that of the land uplift, as is well explained by Bloom [1980]. In regions of slow uplift the youngest emerged feature is the Holocene (~6000 yr) bench, the next, the bench or benches that correspond to high sea levels about 100,000 yr (at 82,000, 103,000, and 124,000 yr ago), and above that a bench 230,000 yr old. In regions of more rapid uplift there may be additional benches cut during periods 28,000, 42,000 and 60,000 yr ago when sea level was relatively high but still lower than the present [Bloom et al., 1974].

Radiocarbon dating of wood and shells from the Whisky Run and Pioneer terraces near Bandon gives ages between 36,000 and 45,000 yr [Janda, 1972, p. 63], but are probably minimum ages. A more definite age for the Whisky Run terrace at Bandon is a Th/U date of 72,200 ± 5300 yr on solitary corals [Kennedy et al., 1982]; G.L. Kennedy (personal communication, 1982) infers the terrace to represent the 82,000-yr sea level maximum. This plausible age of 82,000 yr for the Whisky Run and consideration of the known sea level history leads to the following ages for the terraces near Cape Arago: Whisky Run, 82,000; Pioneer, 103,000; Seven Devils, 124,000; and Metcalf 230,000 years. When these ages are applied to the tilted terraces near Cape Arago, they indicate approximately constant tilt and uplift rates for the last 230,000 years (Table 1), a geologically reasonable result given the tectonic environment.

At Cape Blanco, C^{14} and Th/U dating of shells on the main wave-cut surface (widely considered to correlate with the Pioneer, e.g., Beaulieu and Hughes [1976]), gave ages of 35,000 yr, but in view of carbonate-closure problems supply only a minimum age [Richards and Thurber, 1966]. More recently, Wehmiller et al. [1977] have determined an amino acid racemization date of 50,000 ± 20,000 yr for shells from the same place, considered the terrace to represent either the 42,000 or 60,000 yr sea level maxima of Bloom et al. [1974], and hence calculated an uplift rate of either 2.75 or 1.5 mm yr^{-1}. If the dates for both the Whisky Run and main Cape Blanco terrace are correct, the Blanco terrace cannot correlate with the Pioneer as widely thought, but must instead represent a post-Whiskey Run terrace.

Above the main terrace at Cape Blanco, there are three named terraces: the Silver Butte (about 60 m elevation), the Indian Creek (170 m), and the Poverty Ridge (270 m) [Janda, 1972; Beaulieu and Hughes, 1976]. It is not possible to assign terrace ages unambiguously, but the elevations and Janda's [1972] correlation of the Indian Creek and Seven Devils terraces are more consistent with a 60,000 yr age for the main Blanco terrace rather than 42,000 yr.

The present uncertainty in the absolute ages of the terraces at Cape Arago and at Cape Blanco does not detract from the evidence they provide for progressive tilting and uplift. At worst, the adopted ages could be wrong by a factor of 2 (e.g.,

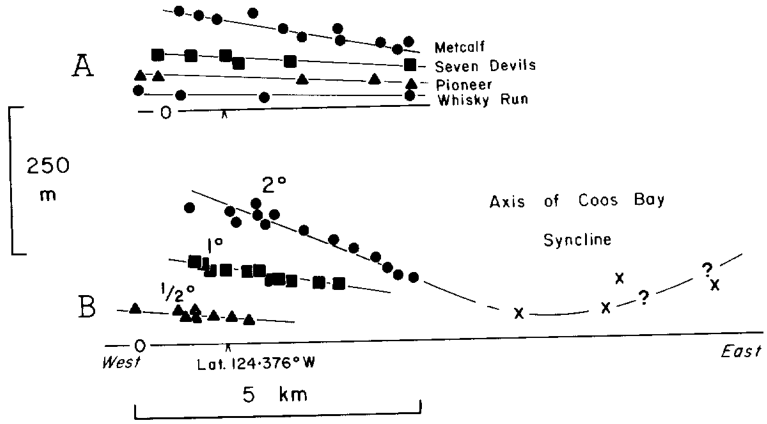

Fig. 3. East-west cross section of terrace elevations north (A) and south (B) of Cape Arago showing the landward tilt of the terraces and demonstrating the progressive nature of the deformation. Elevations are of the terrace surface and not the wave-cut platform and are taken from Griggs [1944, plate 42]. Note that individual terraces cannot be distinguished on the east side of South Slough, but the highest remnants appear to rise above sea level on the east flank of the Coos Bay Syncline.

the youngest plausible age for the Whiskey Run terrace is 42,000 yr) and so the derived rates, even if in error by this factor, still represent rapid deformation.

Landward Tilting of Marine Terraces, Coastal Oregon

Because the Oregon coast is nearly parallel to the subduction trench, there are only a few places where irregularities allow marine terraces to demonstrate tilting normal to the trench.

Wave-cut surfaces that form today in California have an initial seaward slope of 0.3°-1.0° [Bradley and Griggs, 1976], and it is therefore likely that the newly emerged wave-cut surfaces in Oregon had similar seaward slopes. Hence their present landward tilt represents not only tilting from the horizontal but also back-tilting of the initial seaward slope. For an initial seaward slope of 0.3°, surfaces of 42,000, 60,000, ~100,000 and 230,000 years would need to be tilted landward at 12, 8, 5, and 2 x 10^{-8} rad. yr^{-1}, respectively, to be horizontal, and any present tilt in the landward direction increases the total tilt rate. Therefore tilt rates calculated assuming a horizontal surface may be low by a factor of 2 or more, the error being greatest for the youngest terraces. The seaward slope correction applies only to surfaces cut across wide platforms parallel to the tilt direction and is not applicable at Cape Arago because those platforms were cut perpendicular to the tilt direction. At Cape Blanco, however, the main terrace is 6 km wide, probably 60,000 years old, and cut parallel to the tilt direction, so that although it has a landward tilt of 0.10° (see map of Cape Blanco in the work of Beaulieu and Hughes [1976]), and also Janda [1972, p. 48], the total amount of landward tilting, including removal of an initial seaward tilt, could be as much as 0.4°, corresponding to a mean tilt rate of 11 x 10^{-8} rad. yr^{-1}.

The terraces near Cape Arago have a distinct landward (down-to-the-east) dip [Baldwin, 1966]. For example, Janda [1972, p. 28] gives the elevation of the Whisky Run wave-cut surface at Cape Arago as 29 m, but as less than 5 m at Charleston, 5.5 km to the east. Ideally, the elevation of the wave-cut surface for every terrace should be plotted against distance normal to the trench, but, in general, only the elevations of the terrace cover beds are known. The terrace cover beds lie on the wave-cut surface and may be 15-20 m thick; to determine tilt of the terraces from elevation of the cover beds, it is necessary to assume that their thickness is constant. When terrace elevations (taken from Griggs [1945, plate 42]) are plotted as two east-west pro-

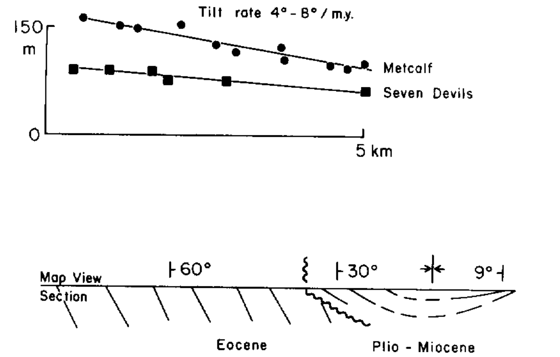

Fig. 4. Relationship between tilted marine terraces (top) and underlying geologic structure (bottom), Coos Bay Syncline, Oregon. The terraces are tilted down toward the axis of the syncline.

files, one north (A) and one south (B) of Cape Arago, they show simple eastward tilt with the amount of tilt being greatest for the oldest and highest surface (Figure 3). The simple pattern for each surface suggests that in each case a single surface has been identified (Table 1).

For the oldest surface, the maximum dip is about 2° on profile B. From an assigned age of 230,000 years, the average landward tilt rate is 16×10^{-8} rad. yr^{-1} (9° per million years). Profile A gives 8×10^{-8} rad. yr^{-1}, and both profiles indicate progressive tilting of surfaces of successive age. Two further profiles (not shown) to the south give rates of 6 and 7×10^{-8} rad. yr^{-1} from the tilt of the Metcalf terrace. A few kilometers to the south of all four profiles a geodetic tilt rate of $(12 \pm 7) \times 10^{-8}$ rad. yr^{-1} between Bandon and Coquille (see below) suggests that tilting in the region has continued at substantially its present rate for 230,000 years.

The easterly tilting of the Cape Arago terraces is related to the tightening of the underlying north trending syncline (Figure 4). Volume constraints require bedding-plane slip during the tightening of folds, the fault throws being opposite in sense to the tilt trend (Figure 5). Hence the apparently smooth landward tilt of the terraces shown in Figure 3 may in fact be interrupted by successive faults.

Youthful Faulting in the Coastal Ranges

Faults with throws of less than 10 m would not show on the 50-foot contour interval map [Griggs, 1945, plate 42] used to construct Figure 3, but are observed in the coastal section east of Cape Arago where there has been an extensive study of the structure and stratigraphy of the Coos Bay Coalfield. Fault displacements probably occur at many places on terraces throughout coastal Oregon and Washington but have been inadequately studied.

At Mussel Reef (Yokam Point), near Cape Arago, a reverse bedding-plane fault which dips east at 70° displaces the Whisky Run terrace surface by 5 m, west side down [Baldwin, 1966, p. 199]. Across Sunset Bay, the Whisky Run Terrace is 6 m lower to the east than the west and may well be offset by a concealed fault [Beaulieu and Hughes, 1975, p. 43]. A further fault has displaced the wave-cut platform by a meter and formed a ridge on the sand [Allen and Baldwin, 1944, p. 40].

At the Seven Devils Mine, on the Seven Devils terrace and 8 km southeast of Cape Arago, a southeast-striking fault in the terrace cover beds displaces the bedrock by more than 3 m, with the northeast side being upthrown [Griggs, 1945]. Along the Whisky Run terrace north of Bandon there are several offsets in the Eocene bedrock, and faulting of the terrace is illustrated by Murphy et al. [1979]. The illustrated fault is one of several faults within the cover beds of the Whisky Run terrace. It strikes N42°W, dips 30°W and has 0.3 m of thrust displacement (P.J. Murphy, Stone and Webster Corp., written communication, 1981).

Further south at Cape Blanco, Dott [1971, p. 52, 56] notes "youthful

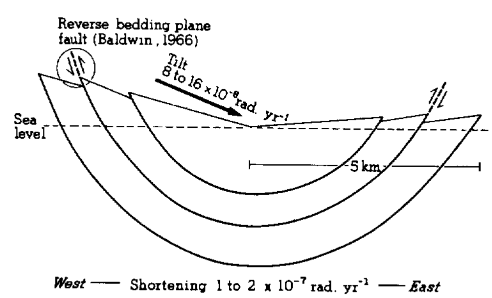

Fig. 5. Schematic cross section across the Coos Bay Syncline showing slip on reverse bedding-plane faults as a consequence of compression across the syncline. Tilt on the flanks is superimposed on a lesser landward tilt (see Figure 9) and produces greater uplift on the western (Cape Arago) flank than on the eastern flank.

faulting" that displaces Pliocene strata, although displacement of the terrace surface is not proven. Finally, 10 km southeast of Cape Blanco the Indian Creek terrace is offset about 5 m (northwest side up) across the Beaver Creek fault [Janda, 1972].

In addition to the above faults in Oregon, three postglacial thrust faults on the Olympic Peninsula, Washington, displace till layers and dam small lakes [Wilson et al., 1979]. These three faults are probably of minor tectonic significance, and like the bedding plane faults near Coos Bay they probably represent a small-scale response to regional compression--flexural slip during folding--rather than being significant tectonic features in themselves. Hence they present a ground-rupture hazard and probably break during earthquakes caused elsewhere rather than causing earthquakes themselves (compare Yeats et al. [1981]).

From tilting and faulting of the marine terraces in a rather restricted part of the coastal ranges, I now turn to more extensive and recent evidence for deformation and landward tilting before attempting to synthesise the long-term pattern of deformation in the coastal ranges.

TIDE GAUGE ANALYSIS

Tide gauges, which can provide short-term elevation changes, exist on both the Oregon-Washington coast [Hicks, 1978] and the Canadian coast [Wigen and Stephenson, 1980]. However, as their recording histories are different and have not yet been compared year for year, I analyze the Canadian and United States data separately.

Oregon-Washington

The land is rising relative to sea level 1.7 mm yr^{-1} at Neah Bay, Washington, and 0.7 mm yr^{-1} at Astoria, Oregon, and is sinking 2.3 mm yr^{-1} at Seattle (Figure 6; values are for the 1940-1975 period common to all gauges and have not been corrected for any eustatic sea level effect). Solving these three results geometrically as a three-point problem gives a landward tilt across the Olympic Peninsula of $(2.8 \pm 0.8) \times 10^{-8}$ rad. yr^{-1} toward N84°E, nearly normal to the coast. A simple linear tilt is suggested by the more detailed geodetic leveling (Figure 7). At Friday Harbour, Washington, the land is sinking at 0.6 mm yr^{-1}, and taken with the Neah Bay value indicates a landward tilt of $(1.9 \pm 0.8) \times 10^{-8}$ rad yr^{-1} toward N82°E, a fortuitous result, but nearly parallel to the tilt direction from the other three tide gauges. A plane fitted to velocities derived from preand post-1940 data from all four tide gauges indicates a landward tilt rate of $(2.7 \pm 0.8) \times 10^{-8}$ rad. yr^{-1} [Reilinger and Adams, 1982].

Canada

The best tide gauge data for the Pacific coast of Canada uses yearly mean differences between gauges to establish trends [Wigen and Stephenson, 1980]. They indicate the land to be sinking relative to sea level 0.64 mm yr^{-1} at Vancouver and 0.27 mm yr^{-1} at Victoria, and rising 1.25 mm yr^{-1} at Tofino and 1.95 mm yr^{-1} at Alert Bay (Figure 6). Rooted tree stumps now below sea level and partly inundated archeological sites indicate that sea level in the Victoria-Vancouver area has risen a few meters relative to the land over the last 5000 yr [Clague et al., 1982], and probably indicates land subsidence like that shown by the tide gauges, because world-wide sea level has changed little over the last 5000 years.

Solving the Tofino, Victoria and Vancouver results geometrically gives a landward tilt of $(1.0 \pm 0.5) \times 10^{-8}$ rad. yr^{-1} toward N75°E, quite consistent in both direction and magnitude with the tide gauge results to the south. However, along Vancouver Island from Alert Bay to Victoria there is a tilt that parallels the coast and is down to the southeast 0.6×10^{-8} rad. yr^{-1}. If this tilt, of unknown origin, is subtracted from the above result, there remains a net tilt of 0.8×10^{-8} rad. yr^{-1} toward N34°E, approximately normal to the coast.

By using additional tide gauge records of somewhat lesser reliability, Riddihough [1982] showed the same pattern of outer coast-up, interior-down, i.e., landward tilting, extends northwards to the Queen Charlotte Islands. From Riddihough's uplift contours, the landward tilt rate may be about 3×10^{-8} rad. yr^{-1} across northern Vancouver Island and 1×10^{-8} rad. yr^{-1} at Queen Charlotte City. Both places are north of the northernmost interaction of the Juan de Fuca plate with North America.

The Canadian tide gauges therefore indicate landward tilting of the coastal range

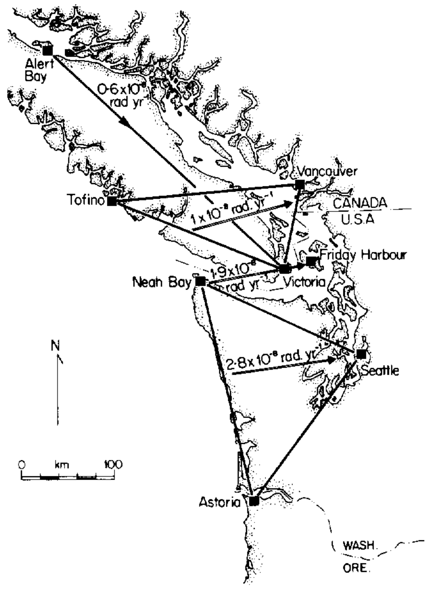

Fig. 6. Location of eight tide gauges in the Pacific Northwest and the tilt vectors obtained by analysing the different sea level change between (or among) the gauges. Land uplift relative to sea level is given for each gauge in the text.

(here Vancouver Island) in the same sense and of similar magnitude as the tilting in the coast ranges of Oregon and Washington.

GEODETIC LEVELING RESULTS

Surveys of five west-east geodetic leveling routes in Oregon and four in Washington provide additional evidence for current landward tilting of the coastal ranges (Figure 7) and have been discussed in detail by Reilinger and Adams [1982]. The two central Oregon lines and the short, Westport to Raymond line in Washington, weakly support, and the remaining lines strongly support landward tilting of the coastal ranges (Table 1 of Reilinger and Adams). Tilt rates for all but the relatively short Bandon-Coquille line are about 3×10^{-8} rad. yr^{-1} (Figure 7). The Bandon-Coquille line runs inland just south of the Cape Arago tilted terraces and gives a landward tilt of $(12 \pm 7) \times 10^{-8}$ rad. yr^{-1} quite consistent with the tilt rate estimated from nearby terraces.

A 35-km west-east profile across the narrowest part of Vancouver Island (Port Alberni to Parksville) was leveled in 1930 and again in 1973 [Rogers and Hasegawa, 1978, Figure 8]. Elevation differences between the levelings reach 80 mm, but for the two end points correspond to an uplift of the west end relative to the east of 0.4 mm yr^{-1} and a landward tilt rate of $(1 \pm 2) \times 10^{-8}$ rad. yr^{-1}. However the profile shows the strong positive correlation with topographic relief that elsewhere suggests systematic surveying errors [Reilinger and Brown, 1981], and so the significance of the simple endpoint difference is not certain.

LONGER TERM DEFORMATION OF THE COASTAL RANGES

On Vancouver Island, no onshore evidence for Neogene or more recent deform-

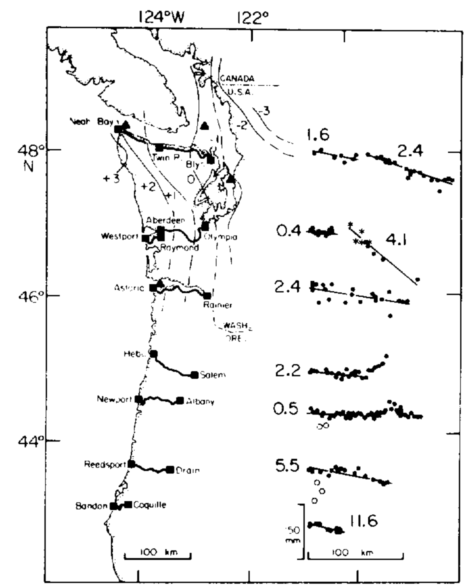

Fig. 7. Location of east-west leveling profiles in coastal Oregon and Washington (left), and elevation change profiles relative to arbitrary zero points for each line (right) [after Reilinger and Adams, 1982]. Stars on the Aberdeen-Olympia profile represent displacements of bench marks in the period 1947-1974; their displacements have been doubled to plot with the 1920-1974 data (dots). Linear regression lines fit to segments of the elevation change profiles give tilt rates shown (units = 10^{-8} rad. yr^{-1}). A few bench marks (open circles) were considered unstable and were not used in the least squares fit. Triangles show location of tide gauges. Contours give uplift (in mm yr^{-1}) in the Olympic Peninsula according to Ando and Balazs [1979].

ation rates has been published, but some indication comes from the deformed sediments on the continental shelf [Tiffin et al., 1972; Yorath, 1980]; these rates are discussed later.

Along the central Washington coast, the wave-cut surface described by Rau [1973] has not been dated directly but, by extrapolation of C^{14} dates in the cover beds, was considered more than 70,000 years old by Florer [1972]. If the surface corresponds to the 82,000-yr sea level maximum, then the most rapid uplift rate is 0.6 mm yr^{-1}, the north-south tilt rate along the coast about 1.5×10^{-8} rad. yr^{-1}, and the wavelength of deformation about 60 km (Figure 2). However, dips in the Pliocene Quinault Formation that underlies part of the wave-cut surface are steeper than 50° in places (Figure 2), and a series of northwest to southwest trending synclines and anticlines at 1 km centers north of Tahola [Rau, 1973] indicates that locally the tilt wavelength is much smaller, and hence the tilt rate higher. The Quinault Formation is less than 7 m.y. old, and dips of 50° in the formation imply an average north-south shortening rate of at least 0.3×10^{-7} yr^{-1} and current tilting at least 4×10^{-8} rad. yr^{-1} by methods discussed in the appendix.

Two major anticlines that parallel the Washington coast are flanked by Quinault Formation beds with dips of 10° to 40° [Glover, 1940]. Another anticline near Wreck Creek (47.28°N 124.40°W) is perpendicular to the coast and plunges east at 6° [Glover, 1940]. If the plunge is entirely due to landward (eastward) tilting at the ~2.5×10^{-8} rad. yr^{-1} rate shown by tide gauge and leveling analysis (Figures 6 and 7), then the anticline could have had no plunge only 4 m.y. ago.

At several places along the Oregon coast, deformed terraces and other features that record short-term tilt rates have the same dip direction as the underlying Cenozoic strata but are naturally less deformed (e.g., Figure 4). The dip of the strata gives the total amount of tilting since deposition, and the tilt of the overlying terraces, the part that occurred in the most recent period of time.

During the steady growth of a sinusoidal fold, the tilt rate on the flanks of the fold is not constant but slows down as the amount of shortening increases. The first 2% of shortening produces 15° flank dips, 6% shortening gives 30° dips, and 22%, 50° dips, with the rate of increase in dip for a given amount of shortening slowing down rapidly as the dip increases (see Figure A1 in the appendix). Hence a given shortening rate will produce very fast angular changes (tilt) on the flanks of very young folds but very slow changes for old folds with dips of 50° or more. For this reason it is physically more reasonable to discuss the tilt rates in terms of the contemporary shortening rates they imply, rather than extrapolating the present tilt rates to account for the stratal dips directly. A relation between shortening rate, flank dip, and tilt rate is newly derived in the appendix (equation A3).

Along the coast north and east of Cape Arago, marine terraces now dip landward at up to 1.1° and have been tilted at 8×10^{-8} rad. yr^{-1} (Figure 3 and Table 1). The terraces are cut across strata on the west side of the Coos Bay Syncline. The syncline contains Eocene and Empire Formation (Pliocene) strata (separated by an unconformity) that have maximum eastward dips of 70 and 20°, respectively [Baldwin et al., 1973]. From the tilt rate of 8×10^{-8} rad. yr^{-1} given by the tilted terraces, the present shortening rate is 1.9×10^{-7} yr^{-1}. The higher tilt rate shown by terraces to the south of Cape Arago (see Table 1 and Figure 3) may be partly explained by more gently dipping (50°) strata; the present shortening rate of 1.3×10^{-7} yr^{-1} there is comparable to that north of the Cape. From the elevation of the terraces, the west flank of the growing syncline is presently being uplifted at a long-term rate of 0.5-1 mm yr^{-1}, and the axis of the syncline is about stable relative to sea level.

The leveling line from Bandon to Coquille crosses the west flank of the Coos Bay Synclinorium and smaller superposed folds [Baldwin et al., 1973], and along the line Eocene strata dip east at an average 35°. From the geodetic tilt rate of $(12 \pm 7) \times 10^{-8}$ rad. yr^{-1}, the current shortening rate is $(0.5 \pm 0.3) \times 10^{-7}$ yr^{-1}.

Similarly, at Cape Blanco, the wave-cut platform of the main terrace dips landward at about 0.10° [Janda, 1972], and if 60,000 years old has been tilted at an average rate of at least 3×10^{-8} rad. yr^{-1}. The actual tilt rate may well be four times more rapid because of the expected seaward tilt of the marine

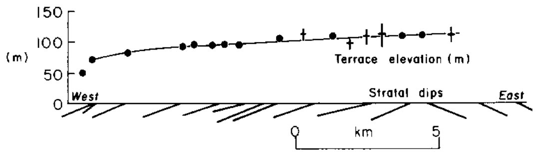

Fig. 8. Profile along Siuslaw River, Oregon showing anomalous seaward tilt of the river terrace and its relation to underlying stratal dips. In the same distance the present river floodplain falls only 8 m.

platform when formed. Elk River beds at Cape Blanco have normal magnetic polarity and an 0.5 ± 0.1 m.y. amino acid age estimate [Wehmiller et al., 1978, Table 1]. They dip at about 3° to the east [Janda, 1972, p. 56] and indicate a landward tilt of 10×10^{-8} rad. yr^{-1}. Also near Cape Blanco, beds of the Empire Formation dip landward at about 7° and Eocene strata dip landward at about 30° [Dott, 1971, p. 52]. The current shortening rate is 0.3×10^{-7} yr^{-1}.

In contrast to Cape Arago and Cape Blanco, a river terrace that extends 14 km up the Siuslaw River dips seaward at 0.13° (Figure 8). The tilt is "more probably due to regional folding and uplift than to a steep ancestral stream gradient" [Schlicker and Deacon, 1974, p. 19]. If the terrace was graded to a 100,000 yr marine bench, then the tilt rate is 2×10^{-8} rad. yr^{-1} seaward. The terrace cuts across Eocene strata that have a mean seaward dip of about 11°, indicating a shortening rate of 0.02×10^{-7} yr^{-1}.

On the Siletz river, upstream of Siletz, Oregon, discontinuous terrace remnants mapped by Schlicker et al. [1973] appear to decrease in height above the present river in a downstream direction, as on the Siuslaw River. If the terrace is 100,000 yr old, it suggests a seaward tilt rate of $(3 \pm 2) \times 10^{-8}$ rad. yr^{-1}. The river cuts across Eocene strata that dip seaward at about 8°, and so the tilt rate suggests a shortening rate of 0.02×10^{-7} yr^{-1}.

The 30° angular unconformity between Miocene and Eocene strata at Cape Arago indicates that deformation on the 10^6-yr time span must be episodic. Nevertheless, the analysis shows that regions of steeply dipping strata are being more rapidly shortened than regions of gently dipping strata, and that at present shortening rates deformation need have continued for only a few million years to form the present geological structures in coastal Oregon.

It is noticeable that the tilt rates on the land (east) facing flanks of the anticlines and synclines of the coastal ranges are faster than on the sea facing flanks. For example, Cape Arago and Cape Blanco give $\geq 9 \times 10^{-8}$ rad. yr^{-1} on landward flanks and Siuslaw River and Siletz River give $\leq 3 \times 10^{-8}$ rad. yr^{-1} on seaward flanks. These observations can be reconciled if the flank tilting is $\sim 6 \times 10^{-8}$ rad. yr^{-1} on the westernmost flank, the flank-tilting rate decreases eastward, and it is superposed on a linear landward tilt of 3×10^{-8} rad. yr^{-1} across the entire coastal ranges, as shown by the leveling data (Figure 9). The linear tilt serves to increase the tilt on landward flanks and decrease it on seaward flanks.

The linear tilt rate of 3×10^{-8} rad. yr^{-1} determined from the releveling, together with stratal dips averaging 10° (away from the Blanco salient; estimated from geological maps produced by the Oregon Department of Geology and Mineral Industries), implies shortening of the order of 0.03×10^{-7} yr^{-1} across the 70-km wide Oregon coastal ranges.

A summary of tilt rates and directions in the Pacific Northwest is given in Figure 10. Although the directions are in

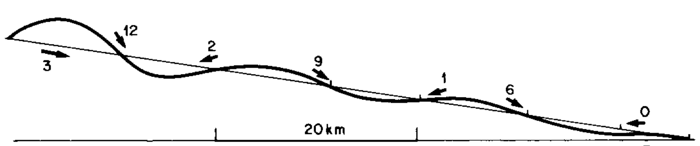

Fig. 9. Schematic interpretation of tilts from terraces and geodetic releveling across the coast ranges. A decaying sinusoidal tilt with maximum flank rates of 6×10^{-8} rad. yr^{-1} and a wavelength of 20 km is superimposed on a linear landward tilt of 3×10^{-8} rad. yr^{-1}.

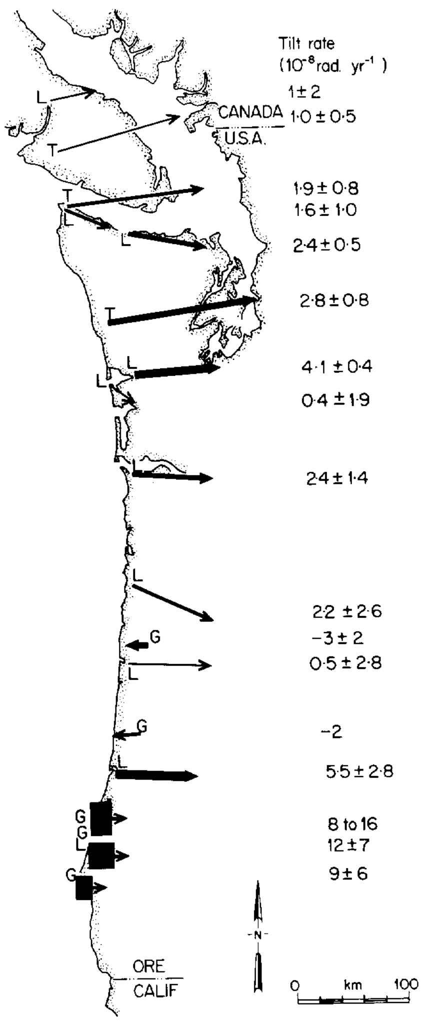

Fig. 10. Summary of tilt rates and directions derived from geodetic, tide gauge, and terrace deformation data. Arrow lengths represent spatial extent of data, arrow widths represent magnitude of tilt rate. Negative rates represent seaward tilting. Letters beside arrows represent nature of evidence for rate: G, geomorphic, L, leveling, T, tide gauge. Note the consistent direction and similar magnitude of tilting in the coast ranges. The exceptions are the short coastal profiles in the south, which are controlled by local structure.

most cases fortuitous and need not represent the true tilt direction, they are uniformly down to the east, and overall suggest an easterly or east-northeasterly tilt. Further, although the rates are determined for periods as short as 10 yr and as long as 230,000 yr, there are no serious differences in magnitude among them. The rates are most rapid in the south where they are determined for the flanks of individual folds, but where they span the entire width of the coast range the rates are consistently about 3 x 10^{-8} rad. yr^{-1} south of Vancouver Island, and on rather scant evidence perhaps half that rate to the north. The general agreement between the rates and their consistency with stratal dips and geologic structure suggest that deformation has continued in the coastal ranges at the contemporary rate for about the last half million years; the rates are similar to those observed near other active subduction zones that are associated with earthquakes.

ONSHORE LANDWARD TILTING AND ITS RELATION TO SUBDUCTION ELSEWHERE

Landward tilting of the coastal ranges is a distinctive component of onshore deformation in the Pacific Northwest, and similar tilting occurs adjacent to subduction zones elsewhere. Away-from-trench tilting is usually easily documented for island arcs, as the islands have shorelines that are normal to the trench. Thus tilting has been established for the Ryukyu Islands [Ota and Yoshikawa, 1978], and for the New Hebrides (Taylor et al. [1980]; rate about 10 x 10^{-8} rad. yr^{-1}).

Landward tilting of coastal ranges is commonly more difficult to quantify as the coasts themselves are parallel to the trench, and measurements must be made at fortuitous places where peninsulas or reentrants occur. Thus in Costa Rica, landward tilting across the Nicoya Peninsula is indicated by drowned shorelines to the east and elevated terraces to the west [Alt et al., 1980] and by bioerosion morphology on coastal platforms [Fischer, 1980]. In Chile, marine terraces on the Arauco Peninsula and adjacent islands are tilted 1°-3° down the east (landward) at inferred rates of 5 to 60 x 10^{-8} rad. yr^{-1} [Kaizuka et al., 1973, Table 3]. The landward tilt rate from the terraces is very nearly the same as the rate of 10-40 x 10^{-8} rad. yr^{-1} deduced from the uplift pattern caused by historic earthquakes and their probable frequency.

The North Island of New Zealand is being obliquely underthrust by the Pacific plate at 50 mm yr^{-1} [Cole and Lewis, 1981] and the complexity of deformation decreases along the east coast

(K.R. Berryman, written communication, 1980), from complex folding and faulting at the south end (Paliser Bay) to gentle folding and no faulting at the north end (East Cape). At Paliser Bay, Ghani [1978] used the differential elevation of Holocene and older marine terraces to deduce deformation rates. Tilt rates on the flanks of folds with 15 km wavelengths are in the range 25 to 65 x 10^{-8} rad. yr^{-1}, and as Oregon, the active tilting is increasing the dip of underlying Tertiary sediments. The complexity of folding and thrust prevents any simple recognition of landward tilting.

However, at East Cape, marine terraces are progressively tilted away from the trench [Yoshikawa et al., 1980] at a mean tilt rate of 5 x 10^{-8} rad. yr^{-1}. Along the same coast the Holocene (6000 yr) bench has been tilted at 5 x 10^{-8} rad. yr^{-1}, indicating substantially constant deformation rates.

In Japan, along the southeast shore of Cape Muroto, Shikoku Island, Yoshikawa et al. [1964] describe tilted marine terraces that indicate landward tilting for the last 100,000 yr at about 6 x 10^{-8} rad. yr^{-1}. In the same area an estimate of the net deformation (preseismic + seismic + postseismic) from geodetic leveling corresponds to a landward tilt rate of 7.5 x 10^{-8} rad. yr^{-1}, and a Holocene rate of 12 x 10^{-8} rad. yr^{-1} to landward [Yoshikawa et al., 1980] shows that the present pattern of deformation can account for the tilt of the terraces.

Cape Muroto is only one of the peninsulas on the south coast of Japan; eight others are uplifted during great earthquakes or have terraces that rise towards the promontory, thus indicating landward tilting [Ota and Yoshikawa, 1978; Matsuda et al., 1978]. In many places the Holocene bench (~6000 yr old) and older terraces have been tilted at a similar rates so providing good long-term evidence that plate underthrusting in southeastern Japan--like the other places cited--causes a consistent landward tilting of the coast at rates similar to those found in Oregon.

TECTONIC DEFORMATION ON THE PACIFIC NORTHWEST CONTINENTAL SHELF AND SLOPE

Off Vancouver Island, Pliocene and younger sediments are deformed into folds, some with active "diapiric" cores, that parallel the coast [Tiffin et al., 1972]. The amount of deformation decreases northwards, and north of the Nootka Fault zone [Hyndman et al., 1979], the sediments are far less deformed. Off central and southern Vancouver Island the folds in Pliocene and younger strata that underlie the shelf have flank dips of 2 to 4°, indicating 0.6% shortening in less than about 7 m.y.

The Apollo structure is a well-described anticline in the Tofino Basin and is southwest of Nootka Island. It is 18 km long, 5 km wide and trends N35°W [Yorath, 1980]. Its recent age--late or post Pleistocene--is shown by deformed Quaternary strata and 20 m of bathymetric relief. Since the Pliocene (5 m.y.), 1.5% of shortening has occurred across the structure and 0.8 km of sediment has accumulated in the broad synclines to the east and west. Yorath believed the anticline represented compression at the toe of a massive submarine slide; if, however, it is tectonic, it represents a shortening rate of at least 3 x 10^{-9} yr^{-1} averaged over the 5-m.y. period.

The continental shelf and slope off Oregon are together 135 km wide off the Columbia River, but they narrow to only 60 km off the Cape Blanco salient. In the vicinity of the salient the folds on the continental shelf are complex and associated with faults and the surface of the shelf is irregular [Kulm and Fowler, 1974]. Elsewhere the surface of the shelf is marked by north to northwest trending ridges 10 to 75 km long that have considerable bathymetric relief. Seismic reflection profiling shows that the ridges (e.g., Heceta Bank off the Siuslaw River mouth) are the uplifted cores of anticlines, and samples dredged from them are Pleistocene to pre-late Miocene in age, far older than the Pleistocene sediments that occupy the broad (30 km wide) synclines between the banks.

Reflection profiling also shows that much deeper (~5 km) and older sedimentary basins underlie the Pleistocene sediments, and that within each syncline the older sediments are the most deformed, showing that the basins and synclines have formed progressively. Miocene and Pliocene siltstones dredged from the uplifted banks contain benthic foraminifera that lived in water depths much greater (~1000 m) than the 100 m depth of the present banks [Byrne et al., 1966]. The siltstones are about the same age as the onshore Empire Formation, and like that formation were probably continuously uplifted and deformed since deposition. They indicate an uplift rate of 0.1-0.3 mm yr^{-1} averaged over millions of years, and indicate that the bathymetric relief of the ridges is mostly due to tectonic uplift. Uplift rates decrease to north and south away from the Cape Blanco salient. Minimum uplift rates (averaged over millions of years) are from 0.1 to 1.0 mm yr^{-1}, although for most of the shelf the rate is about 0.2 mm yr^{-1} [Kulm and Fowler, 1974].

Seismic reflection profiling of the continental shelf and slope off Washington shows that even very young sediments are now deformed into en echelon faulted anticlines that trend parallel to the coast [Carson et al., 1974]. The folds are typically less than 10 km wide, 1 km high, and have flank dips of 15° or less. A time sequence is indicated, with the younger and less deformed folds being the closest to the trench, and deformation migrating outwards as the older structures become tightly folded. The folds are estimated to be between 0.13 and 0.35 m.y. old and so the uplift on the fold crest appears to be 2-5 mm yr^{-1} faster than surrounding areas. Despite the age sequence, all the folds indicate essentially constant shortening rates (about 0.5 x 10^{-7} yr^{-1}), suggesting that compression is still being taken up across the older folds.

Further to the work of Carson et al. [1974], Barnard [1978] concluded that the sediment deformation shown by seismic reflection profiles indicates about 7 mm yr^{-1} (range 4 to 15 mm yr^{-1}) of shortening across the outermost marginal ridge. On his profiles, the outermost ridges are about 7 km wide, so that a shortening rate (including thrusting) across the folds of 10 $\pm$ 5 x 10^{-7} yr^{-1} is indicated for the last 0.2-0.5 m.y. This should be compared with rates of ~0.5 x 10^{-7} yr^{-1} for folding alone [Carson et al., 1974], and as noted by Barnard [1978, p. 90], 90-95% of the shortening may be accommodated by thrusting.

Shortening rates across the marginal ridge are therefore very high. As shown by Carson et al. [1974], the second fold in from the margin (e.g. fold 79-5b) also shows high shortening rates. One syncline 50 km east of the base of the slope (Barnard [1978], Figure 6, profile H, syncline to landward of fold 7S) has an unconformity (presumed to be 0.5 m.y. after Barnard [1978]) that dips at 5.3° (after correction for vertical exageration and profile orientation relative to the margin) toward the axis of the syncline and so is inferred to indicate an average tilt rate on the flank of the fold of 19 x 10^{-8} rad. yr^{-1}. From the same profile the maximum flank dips are 20° so that over the 0.5 m.y. period the flank dip has increased from 14.7° (= 1.7% shortening by equation A2) to 20° (3.1% shortening), a mean shortening rate of 0.3 x 10^{-7} yr^{-1}. At this rate the fold might be only 1.1 m.y. old.

Unfortunately, further profiles that might allow calculation of shortening rates for the folds on the shelf have not been interpreted. Nevertheless, I expect substantial shortening to be occurring across the entire shelf, because the broad synclinal folds beneath the continental shelf show gently folded strata over older, more deformed strata [Kulm and Fowler, 1974], the fold crests are still being uplifted as shown by their bathymetric relief [Kulm and Fowler, 1974], and onshore shortening rates are about 10^{-7} yr^{-1} near Cape Arago.

GEOLOGIC SHORTENING RATES AND DIRECTIONS: SUMMARY

In the previous sections I have deduced shortening rates from the tilting of marine and river terraces onshore, from the geodetic tilt measurements across the coastal ranges, and from the tilt of a warped unconformity on the continental shelf. I have also cited shortening rates determined by others for very young folds close to the Juan de Fuca-North American plate boundary. Figure 11 summarizes the rates and their orientation, and Figure 12 proposes a composite profile of shortening rate eastwards from the plate boundary.

I conclude that shortening rates across the continental margin are ~10 x 10^{-7} yr^{-1} (folding and thrust faulting) at the immediate margin, $\geq$ 0.3 x 10^{-7} yr^{-1} (folding only, unknown additional amount due to thrusting) on the shelf, (0.02 to 1.5) x 10^{-7} yr^{-1} (folding only, thrusting minor) for coastal onshore folds with rates being highest where the coast is nearest the slope (Figure 9), and ~0.03 x 10^{-7} yr^{-1} (from tilting) for the remainder of the coastal ranges. All the above rates represent permanent deformation, and as seems reasonable, the shortening rates are highest near the surface expression of the plate boundary and decrease away from it toward the continent.

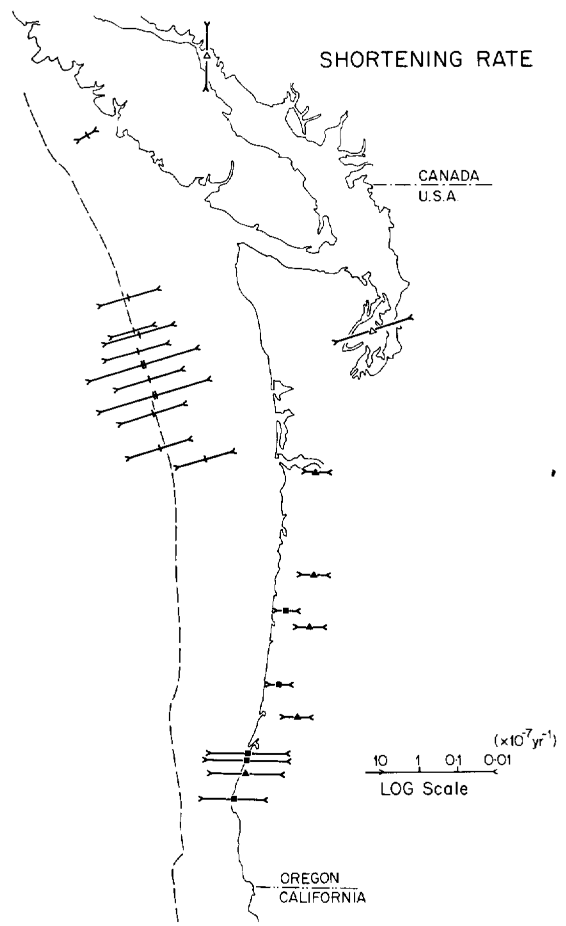

Fig. 11. Summary of contemporary shortening rates and directions derived from trilateration (open triangles), leveling (filled triangles), terrace tilting (squares), stratal folding (single bars), folding and faulting (double bars) as discussed in the text. In order to plot the large range of rates, arrow lengths are proportional to the logarithm of the rate.

The straight line on Figure 12 represents an exponential decay of shortening rate away from the plate boundary. Obviously, the exact decay relationship is poorly defined, but the one on the figure is equivalent to total shortening across the margin of 25 mm yr^{-1}, 80% of which occurs in the most westerly 40 km. The total shortening rate could be a sizeable fraction of the plate convergence rate. For the slope, shelf, and the immediate coast, the young geologic structures indicate contemporary shortening in a direction approximately parallel to the Juan de Fuca-North American plate compression vector.

SHORTENING AND STRESS DIRECTIONS IN THE PACIFIC NORTHWEST

For the central part of the Pacific Northwest continental margin, the analysis of magnetic anomalies indicates relative convergence between the Juan de Fuca and North American plates of 35 mm yr^{-1} along N50°E, or 32 mm yr^{-1} normal to the continental margin [Riddihough, 1977]. The convergence rate has decreased sharply in the last 4.5 m.y. Two fragments of the Juan de Fuca plate appear to be moving independantly of the main plate. North of the Nootka Fault Zone, the Explorer plate is moving N6°E at 19 mm yr^{-1} relative to North America and 14 mm yr^{-1} normal to the Vancouver Island continental margin [Riddihough, 1977]. South of 42°N the Gorda plate is moving north, parallel to

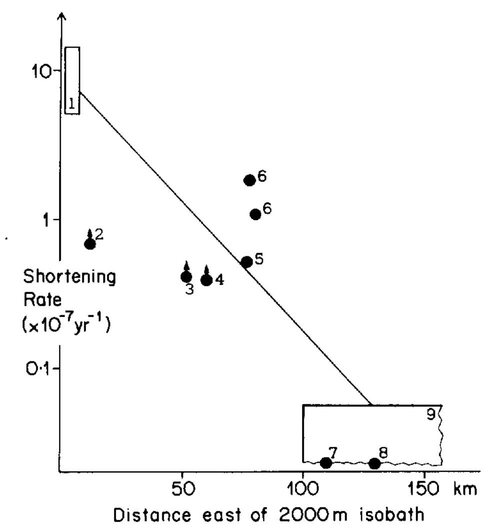

Fig. 12. Composite profile of shortening rate eastwards from the Juan de Fuca-North American plate boundary (represented by the 2000 m isobath). Note logarithmic scale. Straight line (with equation $y = 10^{-6} \exp(-0.04\ x)$) represents suggested rapid decay of shortening rate eastwards. Data are 1, folding and thrusting of marginal folds [Barnard, 1978]. 2, folding of marginal folds [Carson et al., 1974]. 3, folded unconformity on shelf. 4, tilted terrace at Cape Blanco. 5, leveling near Bandon. 6, tilted terraces at Cape Arago. 7, terrace on Siuslaw River. 8, terrace on Siletz River. 9, range of values for leveling across coast ranges.

the North American plate at 40 mm yr^{-1} (Riddihough, 1980) and past convergence between the two plates may have recently ceased.

Geodetic Determinations of Shortening Directions and Rates

The direct determination of earth strain has been made for three networks in British Columbia and two in Washington. From triangulation surveys near the 1946 British Columbia earthquake, Slawson and Savage [1979] calculated shortening of $(22 \pm 12) \times 10^{-6}$ directed towards 179°, and inferred right-lateral normal slip on the Beaufort Range fault during the earthquake. A triangulation network across Johnstone Strait (between Vancouver Island the mainland) gives a secular shortening rate of $(0.6 \pm 0.3) \times 10^{-7}$ yr^{-1} directed toward 178° ± 15° for the period 1914 to 1966 [Savage et al., 1981]. The secular rate is a tenth that across the fault, but both measurements indicate north-south compression in the region, probably released by dextral movement on northwest striking faults [Rogers, 1979]. A recent measurement on the west coast of Vancouver Island [Lisowski and Slawson, 1983] indicates significant compression along 022° and may represent a transition between the compression measured above and compression parallel to the Juan de Fuca-North America plate motion.

Each of the two Washington networks was measured annually for 7 years, so giving results with relatively large errors but showing distinct trends [Savage et al., 1981]. At Seattle a shortening rate of $(1.3 \pm 0.2) \times 10^{-7}$ yr^{-1} directed toward 071° ± 6° is indicated and the strain may have accumulated uniformly with time. At Hanford, south-central Washington and 370 km east of the coast, the shortening rate is $(0.4 \pm 0.1) \times 10^{-7}$ yr^{-1} toward 054° ± 23°, but the accuracy of the measurement is such that no strain need have accumulated at all. Both networks show northeast-southwest compression that is parallel to the convergence between the Juan de Fuca and North American plates.

Geodetically determined strains are well determined for coastal ranges adjacent to subduction zones in Japan and New Zealand. In areas of southern Japan not affected by large earthquakes, the maximum shear strain rate is normal to both coast and plate boundary, and decreases from 4×10^{-7} yr^{-1} to 0.5×10^{-7} yr^{-1} away from the plate boundary [Nakane, 1973a, b]. Geodetic releveling across folds 200 km west of the plate boundary [Mizoue, 1980] shows that permanent deformation occurs even at this distance. In the southeast part of the North Island of New Zealand strain rates of $(3\text{-}5) \times 10^{-7}$ yr^{-1} indicate compression normal to the coast [Walcott, 1978a]. At least some of the strain accumulation is nonelastic, since some shortening is occurring by active folding [Ghani, 1978], as in Oregon. The measured strain rate at Seattle is lower than at the other plate boundaries, perhaps partly because plate convergence is not as rapid.

Seismic Determinations of Shortening Directions and Rates

Large earthquakes have been rare in the coastal Pacific Northwest, no earthquake larger than magnitude 7.5 having occurred since settlement began about 160 years ago. With the chief exception of some deep earthquakes below Puget Sound, many earthquakes appear to relate to crustal deformation and the region lacks the great earthquakes and deep, well-defined Benioff zone associated with subduction zones elsewhere.

From their depth and focal mechanisms, the two largest earthquakes--the 1949 magnitude 7.1 Seattle and the 1965 magnitude 6.5 Puget Sound--appear to lie within a bending plate of subducted lithosphere [Rogers, 1979], perhaps near the point where the dip of the plate steepens [Langston, 1981]. Thus they need not be closely related to crustal stresses.

Weaver and Smith [1983] have recently assessed crustal seismicity and earthquake focal mechanisms for their St. Helens seismic zone and for other earthquakes in western Washington. They show that the zone is a dextral shear that is responding to NE directed compression from the Juan de Fuca plate. Furthermore, focal mechanisms seaward of a line through Seattle have northeast-trending compression axes that parallel the plate convergence vector and the compression direction measured at Seattle. To landward the mechansims indicate north-south compression. The earlier apparent disagreement between focal mechanisms [e.g., Crosson, 1972] and plate convergence was due mainly to scarce data and to the assumption that a single stress province existed in the Pacific Northwest [e.g., Sbar, 1983].

Weaver and Smith (1983) interpret the focal mechanisms as evidence for a locked interface between the Juan de Fuca and North American plates, with the eastern boundary of the locked region causing the change in stress direction. They further suggest that the locked area is bounded to the south, roughly along the Columbia River. A focal mechanism from Oregon [Bolt et al., 1968] and mechanisms from British Columbia [Rogers and Hasegawa, 1978; Rogers, 1979] also indicate north trending compression. Following the argument of Weaver and Smith, these additional measurements may constrain the proposed locked region to beneath western Washington.

Alternatively, a stress gradient across the continental margin overlying the subduction zone may account for the abrupt change in the maximum compression direction. Nakamura and Uyeda [1980] note that in Alaska and southern Japan the maximum horizontal stress is parallel to the plate convergence vector near the trench but roughly normal to it 200-1000 km inland. They conclude that compressional stress at the trench is not transmitted directly to the plate interior but is partly taken up (attenuated) by folding and thrust faulting within the overlying plate.

The Juan de Fuca plate is a small plate between the much larger Pacific and North American plates, which are in north-south compression (or equivalent shear) north and south of the Juan de Fuca plate. Attenuation of subduction stress eastwards by deformation near the continental margin would allow the broader north-south compression to dominate the stress field inland of the coast ranges. To seaward the compression would be roughly northeast-southwest and produce folds parallel to the coast, to landward the compression would be north-south. Such an attenuation of stress could be measured directly by borehole deformation gauges, but to my knowledge such measurements have not been made.

From a recurrence relationship and an assumed largest earthquake, and irrespective of compression direction, Weichert and Hyndman [1983] estimated a total seismic moment release rate of about 0.1×10^{18} Nm yr^{-1} in their Cascade region, equivalent to shortening of 30-km-thick crust at 0.13 mm yr^{-1}. Including a crustal contribution from their Puget region, the overall shortening rate would be 0.16 mm yr^{-1}. The Cascade region lies about 100 km east of the 2000-m isobath, and the shortening-rate decay relationship on Figure 12 would suggest shortening at 0.45 mm yr^{-1} within this region. Although the moment shortening rates could be in error by a factor of 2 or 4 [Weichert and Hyndman, 1983], and the shortening-rate decay relationship is no better defined, the eastwards decrease in shortening rate can explain the extreme disagreement between plate tectonic convergence rates (~35 mm yr^{-1}) and the crustal seismic moment shortening rates (~0.15 mm yr^{-1}) that concerned Weichert and Hyndman.

For comparison, permanent crustal shortening of at least 3 mm yr^{-1} (3% of the plate motion across the subduction zone) has been estimated from the historical record of large intraplate earthquakes in Japan [Wesnowsky et al., 1982]. The Cascade rate of seismic intraplate shortening is an order of magnitude lower than the Japanese rate.

DISCUSSION

The volcanoes of the Cascade Range and the presence of a dipping slab of oceanic crust beneath the Puget Lowlands [Langston, 1981] suggest three broad alternatives to the seismotectonic framework of the Pacific Northwest.

First, after continuing for the preceeding 5-10 m.y., subduction might have ceased in the last few hundred thousand years, the age of the youngest compressive folds at the continental margin [Carson et al., 1974]. A more radical view--that there has been substantially no subduction for the last 9 m.y.--has been argued by Farrar and Dixon [1980]. A recent cessation of subduction would imply that the Juan de Fuca plate is now welded to North America and so explain the apparent dominance of north-south compression over the east-west compression that formed the Pleistocene folds in Oregon and Washington. It could also explain the lack of seismicity in the area, and in particular the absence of large thrust earthquakes and a Benioff Zone. However, the onshore record of marine terrace deformation indicates substantially constant tilt rates for the last 0.06-0.23 m.y. These rates are even faster than the long-term deformation rates indicated by the dip of Eocene and Miocene strata beneath the terraces, and they indicate continued deformation. Releveling and tide gauge data further show that deformation at similar rates has continued into the present and so suggest that the overall tectonic environment has changed little in the past half million years: certainly a dramatic change in the last 10^5 yr seems highly unlikely.

A second alternative accepts the plate motion and deformation studies that indicate subduction continues at about the 35 mm yr^{-1} convergence rate, but argues that the subduction is occurring aseismically at a steady pace through stable sliding or creep. Aseismic subduction might be favoured by the relatively slow convergence rate and young age of the subducted plate which allow it to behave plastically at a shallow depth [Riddihough and Hyndman, 1976]. In addition some shortening occurs by deformation within the overlying sedimentary wedge, perhaps reducing the amount of slip required at seismogenic depths.

The geodetic leveling results which indicate continued landward tilting for about the last 70 yr [Ando and Balazs, 1979; Reilinger and Adams, 1982], are at their simplest consistent with the net pattern of deformation in earthquake zones. Ando and Balazs suggest that the deformation in the Pacific Northwest averages-out the discrete earthquake strain accumulation and release phases commonly seen where great thrust earthquakes occur, and that subduction is occurring aseismically. However, the simple pattern of plastic deformation above a megathrust, proposed by Ando and Balazs [1979] to account for the steady landward tilting in Washington, has been challenged by Yonekura and Shimazaki [1980], who suggest that the uplift that results in landward tilting is restricted in space and time and is directly related to infrequent ruptures on imbricate thrust faults that branch upwards from the megathrust [Fukao, 1979]. It is by no means clear that such imbricate faults or restricted uplift occur in Oregon and Washington, because no major thrust faults have been identified onshore, leveling shows consistent contemporary tilting for the entire coastal range (600 km x 60 km), and at present evidence (either for or against) episodic uplift is lacking. Hence completely aseismic subduction remains a distinct possibility.

A third alternative is that subduction is continuing in the Pacific Northwest, that it is accompanied by great thrust earthquakes, but that the return period of these is such that (coincidentally) none has occurred during historic times. In this interpretation the presently observed movements are interpreted as interseismic deformation, dominantly plastic and permanent near the plate boundary but partly elastic and temporary away from it; the strain now accumulating will be released by some future earthquake. Savage et al. [1981] have interpreted the accumulation of strain near Seattle to represent elastic strain accumulation and have postulated a megathrust dipping east at 15° and extending from the base of the continental slope downdip for 130 km. Modeled strain accumulation on this locked thrust, assuming plate convergence of 32 mm yr^{-1}, matches the measured strain rates at Seattle reasonably well. As the measured strain rate is large compared with permanent shortening rates calculated herein (Figure 12) for places as distant from the plate boundary, it might plausibly represent elastic strain accumulation.

Given the contemporary deformation rates presented in this paper, the problem is not whether subduction continues beneath Oregon and Washington, but whether the subduction zone is characterized by locked regions and occasional large thrust earthquakes or is aseismic. The historic record of the region contains no great shallow earthquakes, either within the overlying North American plate or along its shallow-dipping boundary with the Juan de Fuca plate.

According to Sbar [1983], present crustal earthquakes in the Pacific Northwest represent the release of long-accumulated strain in the north-south direction. The scarcity of earthquakes releasing substantial NE-SW compressional strain suggests that as yet insufficient strain has accumulated in this direction. However, the NE-SW strain is generated by long-term plate subduction. Is then the geodetic strain measured at Seattle episodic, accumulating and being released over a period of a few years as nearly smooth aseismic subduction occurs, as suggested by Sbar [1983]? Or is the accumulated NE-SW compressive strain currently too low to generate many earthquakes, because much previous strain was released in some large prehistoric earthquake?

In the case of episodic strain accumulation, the recently measured compression at Seattle would presumably be released by a period of increased, aseismic slip on the subduction thrust causing a period of relative extension at Seattle. This implies rather a weak coupling between the two plates, despite the evidence for active deformation of the margin presented in this paper, which suggests that the coupling is quite strong. Furthermore, the agreement between geodetic leveling and tide gauge trends in Washington suggests that any change in coupling must occur at time scales of less than about 10 yr, and since the Seattle network has now been analyzed for 7 yr, any such episodic behavior should soon become apparent.

Were infrequent large earthquakes to occur, strain accumulation amounts and rates might vary greatly over periods of hundreds of years. For example, Walcott [1978b] in his analysis of geodetic strain before and after a large thrust earthquake in New Zealand showed that the earthquake was followed by a period of extension normal to the margin. The extension period soon changes to one of compression, and after the extension is recovered the strain builds up for the next large earthquake. Walcott estimated 70-130 yr of strain accumulation at a rate of $\sim 4 \times 10^{-7}$ yr^{-1} (for a total strain of $(30\text{-}50) \times 10^{-6}$) would be needed before the next large earthquake, part of which is required to recover the spreading during the extensional period.

At the 2×10^{-7} yr^{-1} rate of shear strain accumulation at Seattle, earthquakes that need $(40\text{-}100) \times 10^{-6}$ of shear strain (as proposed by Sbar [1983]) would require 200-500 years of accumulation. If great earthquakes had such a long return period it would explain their absence in the short historic record, and if the extensional phase had only recently been completed (say, in 1900 prior to all tide gauge and geodetic measurements) it might explain the contemporary scarcity of thrust earthquakes along the continental margin. In the absence of such earthquakes, those earthquakes representing the release of north-south strain could dominate the recent seismicity.

If the region is prone to great thrust earthquakes, then past earthquakes should have imprinted their effects on the landforms and sediments of the region. I discuss below four lines of evidence for past earthquakes in the region, and suggest that each is worthy of future study.

Possible Evidence for Past Earthquakes

Sims [1975] showed that the San Fernando and previous earthquakes could be deduced from the depth and nature of disturbances in the sediments of Van Norman Lake, California. At a site in the Puget Lowlands of Washington, he considered 14 of 21 zones of deformation in 40,000-year-old varved sediments were caused by earthquakes. At a single site such deformation zones might represent moderate, close earthquakes like those felt in 1949 and 1965, or large, distant earthquakes on the plate boundary that are not known from historic times. Nevertheless, they do record earthquakes with a return period of 70-170 yr and hence place some constraints on the total earthquake history of the area.

One effect of earthquakes in hilly terrain is the triggering of landslides, some of which may dam river valleys to form lakes. The lakes then provide good opportunities for dating the earthquakes [Adams, 1981]. Landslides may also be triggered by other causes, but synchronous age, areal extent, and distribution pattern help distinguish earthquake-caused landslides. Landslide-dammed lakes occur in the Olympic Mountains [Tabor and Cady, 1978; Tabor, 1975] and in Oregon [Baldwin, 1958], and their systematic dating would seem desirable. The dating of drowned trees has already been used to infer movements on some small faults in the Olympic Peninsula [Wilson et al., 1979].

A further effect of large earthquakes is the slumping of shelf-edge sediments, their transport by turbidity flows, and their deposition on the deep-sea floor as turbidites. Off the Columbia River mouth, Griggs and Kulm [1970] found evidence for turbidity flows every 400 to 500 years in the 6600-yr period since the deposition of the Mazama Ash. They note that the thickness of the hemipelagic clay between the turbidites varies by a factor of 2 or less, suggesting similar time intervals between flows, and that the thickness of the turbidites themselves are similar, suggesting a uniform time for sediment accumulation in the source area. While the regularity of the turbidity flows could be due to self-triggering (accumulation of sediment at constant rate until it becomes unstable), it is also consistent with an earthquake cycle. It is very unlikely that great thrust earthquakes beneath the shelf could fail to trigger slumps on the slope.

Therefore while large slumps could probably occur without an earthquake trigger, the turbidite record suggests that great earthquakes occur no more frequently than every 400 yr.

In addition to the further analysis and dating of the high-level marine terraces in coastal Oregon and Washington, as mentioned earlier, much more work is needed on Holocene shoreline features. A Holocene beach ridge formed when sea level attained its present elevation 6000 years ago would now be 9 m above its original level if uplift continued at the 1.5 mm yr^{-1} rate determined for Cape Blanco. The elevation of the highest Holocene beach along the Pacific coast might therefore record differential uplift along the plate boundary. If great thrust earthquakes occur in this region, their coseismic coastal uplift might strand the then-current beach ridge, and so over a period of time form a series of beach ridges below the highest, each recording an earthquake. For example, a series of beach ridges in New Zealand indicates five earthquakes in the past 6500 yr [Wellman, 1967]. To date there appears to have been no systematic study of emerged Holocene beaches on the Pacific Northwest Coast, although the promise of such a study is suggested by a 3000-yr old storm berm at Cape Blanco that is 2.5 m higher than the present berm (Janda [1972]; R.J. Janda, unpublished data, 1970; C^{14} date W2523).

Confirming Contemporary Deformation

Equally important as the study of uplifted features is the search for evidence--e.g. rooted trees below high tide level--that might indicate regional subsidence of the coast and hence a change to preseismic strain buildup before a great earthquake. In this connection it is interesting to note the report of tree stumps exposed at low tide along Sunset Bay near Cape Arago [Baldwin, 1959, p. 40; Janda, 1972, p. 23]. These are undated and possibly old; however, they occur in a region of long-term uplift and would have important implications if they were very young.

There are many reasons why I believe the leveling routes in coastal Oregon should be releveled in the near future. Substantial time has elapsed since the last leveling in 1941, and so a new leveling could provide substantial confirmation (and with much greater precision) of the geodetic results presented. The new leveling would also be of great value to studies of the temporal behaviour of deformation in the coastal ranges and to our understanding of the subduction (aseismic or otherwise) of the Juan de Fuca plate. Finally, should a large earthquake occur, it would be invaluable to have geodetic observations made shortly (and not 40 years) before the event for comparison with earlier and later surveys.

Of great value to complement the releveling program would be the establishment of an array designed to detect tilt of the Oregon coast. The most obvious site is on the extensive flat surface of the main terrace at Cape Blanco, where long-term tilt rates are known from terrace deformation to be about 10×10^{-8} rad. yr^{-1}. In addition to confirming long-term tilting, the array might show irregularities in the landward tilt rate and possibly the reversals in tilt direction that might preceed large earthquakes.

CONCLUSIONS

Although I have derived much new data on rates of contemporary landward tilting and shortening of the Pacific Northwest continental margin, I have been unable to provide definitive evidence as to whether the subduction causing the deformation is steady or episodic, aseismic or seismic.

On the one hand there is the agreement (coincidental or otherwise) between geodetic leveling near Bandon (10-yr period) and the long-term tilt rate of terraces (105-yr period) [Reilinger and Adams, 1982], the steady tilting for the past 50 yr shown by Ando and Balazs [1979] across the Olympic Peninsula, and the low current seismicity of the region, all suggesting steady deformation, and interpretable (in the absence of historic earthquakes) as resulting from aseismic subduction. On the other hand, the aseismic subduction is difficult to reconcile with the observation that deformation rates and style along the Pacific Northwest margin are closely comparable to those associated with strongly seismic subduction elsewhere in the world.

Barring a major change in seismicity in the near future, it is apparent that a definitive assessment of large-earthquake risk in coastal Oregon and Washington will require the search for the effects of past earthquakes as discussed earlier. While such a search could reveal evidence of past large earthquakes, the lack of such evidence would not eliminate entirely the possibility that large earthquakes occur. However, the significance of the problem is such that all methods of investigation must be explored.

APPENDIX: DETERMINATION OF SHORTENING RATES FROM THE TILT RATE OF FOLDED STRATA

Folded beds approximate sine curves in shape [Currie et al., 1962], and so the shortening they represent can be analyzed mathematically. In terms of the length along the curve S, the wavelength L of a sine curve is

$$L = \frac{\pi S}{2} \cdot \frac{\cos\theta}{E(\theta)} \qquad \text{(A1)}$$

where θ is the slope of the curve at a point of inflection and $E(\theta)$ is the complete elliptic integral of the second kind (see, for example, tables in "Handbook of Tables for Mathematics," 3rd edition, CRC Company). For a sinusoidal fold, θ represents the steepest flank dip, and beds of original length S (beds are assumed to conserve their length) now have horizontal extent L and so the total amount of horizontal shortening, s, is

$$s = L - S = S \cdot \left[\frac{\pi}{2} \cdot \frac{\cos\theta}{E(\theta)} - 1\right]$$

and the fractional length change (shortening) with respect to the original length, S, is

$$x = \frac{s}{S} = \frac{\pi}{2} \cdot \frac{\cos\theta}{E(\theta)} - 1 \qquad \text{(A2)}$$

The beds are now 100[1+x]% their original length and have been shortened by 100[-x]%.

As shown by the graph of this relationship (Figure A1), the flank dip increases rapidly at first for a small amount of shortening, and thence increases more slowly. The sine model of folds breaks down at intermediate to large values of shortening because the folds "lock up" and shortening occurs by other mechanisms. Therefore quantitative results from beds dipping more than about 50° should be used with caution.

The rate of horizontal shortening is obtained by differentiating L with respect

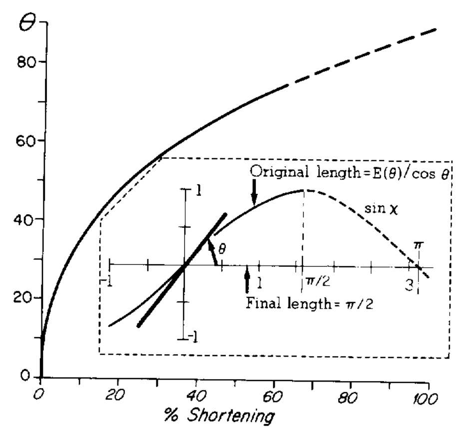

Fig. A1. Diagram showing the dip on the flanks of a sinusoidal fold that is caused by varying amounts of shortening.

to time, dividing by L then gives the strain rate (units = T^{-1}):

$$\frac{dx}{dt} = \left[-\tan\theta - \frac{1}{\tan\theta}\left[1 - \frac{K(\theta)}{E(\theta)}\right]\right]\frac{d\theta}{dt} \qquad \text{(A3)}$$

where $K(\theta)$ and $E(\theta)$ are the complete elliptic integrals of the first and second kind and $d\theta/dt$ is the rate of change of dip on the flank of the fold.

Equation (A3) links the shortening rate to the dip of the fold and the rate of increase of that dip; the latter is also the tilt rate of marine terraces cut across the folded strata and examples of shortening rates determined from tilt rates are given in the paper.

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