APPENDIX 2.5N

LATE CENOZOIC TECTONICS OF THE PACIFIC NORTHWEST WITH SPECIAL REFERENCE TO THE COLUMBIA PLATEAU

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1.1 INTRODUCTION

This discussion of the late Cenozoic tectonics of the Pacific Northwest is primarily concerned with Miocene and younger events in the Columbia Plateau and adjacent geologic provinces, unlike an earlier review (Washington Public Power Supply System, 1977a) which treated evolution of the entire Pacific Northwest from Precambrian time to the present. The paper relies heavily on data and concepts developed by earth scientists since 1977, particularly on the structural analyses of Columbia Plateau deformation by Laubscher (1977; 1981). Specific topics to be treated include: (1) late Cenozoic kinematic relationships between the Great Basin and the Pacific Northwest; (2) the timing of plateau deformation; (3) the nature of the Olympic-Wallowa lineament; and (4) a tectonic model for plateau deformation.

2.1 PLATE TECTONIC SETTING OF THE PACIFIC NORTHWEST

The Miocene and younger tectonics of the Pacific Northwest are the consequences of incompletely understood interactions between the Pacific, North American, and Juan de Fuca plates. The latter, a vestigial remnant of the once extensive Farallon plate, lies offshore of Oregon, Washington, and Vancouver Island. It is separated from the Pacific Plate to the north, west and south by the Gorda, Juan de Fuca, and Explorer ridges (from south to north) and the transform faults (fracture zones) which bound them. To the east, the Juan de Fuca-North American plate boundary is one of past convergence and, in the opinion of most recent workers, one along which subduction still continues at rates of several centimeters per year.

Washington Public Power Supply System (1977a), in agreement with Riddihough and Hyndman (1976) and Carlson (1976), concluded that despite the lack of an inclined seismic (Benioff) zone beneath western Washington and Oregon, eastward subduction of the Juan de Fuca plate beneath North America is probably an ongoing process. More recently, Crosson (1980) has presented seismic focal data that indicates that a shallow east-dipping Benioff zone does exist beneath western and central parts of the Puget Trough at the latitude of the Olympic Peninsula. Other features supportive of Quaternary subduction beneath the continent include the compressive deformation of late Pleistocene sediments along the base of the Oregon-Washington

continental slope, the linear array of Quaternary volcanoes of the High Cascades, the present gravity field of the continental margin (see below), and the nature of the deep (60-70km) Puget Sound-Gulf Islands earthquakes. Based on focal mechanism solutions, the deep earthquakes of 1965 (April 29, Puget Sound) and 1976 (May 16, Gulf Islands) were produced by normal faults with north-south strikes, a type of faulting attributed by Washington Public Power Supply System (1977a) to extensional strain in the outer (upper) part of the subducting oceanic plate where it abruptly bends to steepen eastward. That plate, therefore, must dip only 10° to 15° eastward from the base of the offshore continental slope to a 60 km depth below Puget Sound. It must then steepen to 30° to 50°, however, if the Quaternary Cascade volcanoes east of the Puget Trough lie 100 km to 120 km above the subducting plate as many petrologists believe (cf. Washington Public Power Supply System, 1977a, Figure 2R C-7). McKenzie and Julian (1971) had earlier concluded from a study of travel-time residuals of the 1965 Seattle earthquake that a slab of high-velocity material, interpreted by them as subducted oceanic crust, dips eastward ($50^{\circ} \pm 10^{\circ}$) beneath the Cascades. The results of their studies have recently been confirmed by A. Rohay (oral communication, Cascadia Conference, May 1980).

On the basis of a recent analysis of the regional gravity field in western Oregon, Washington, and southwestern British Columbia, Riddihough (1979, p. 351) concludes that the field, which shows an outer "low" and an inner "high", is "similar to that over other subduction zones." A crustal section drawn by Riddihough (his Figure 5, Profile 3) through the Olympic Peninsula, Puget Sound, and Mt. Baker, has a structural interpretation from a gravity profile closely approximating the interpretation drawn by Washington Public Power Supply System (1977a) from other evidence. The descending Juan de Fuca plate is shown to dip 5° eastward to a position beneath Puget Sound, then to dip steeply and abruptly to 40°.

Ando and Balazs (1979) have used geodetic data to conclude that the Juan de Fuca plate is aseismically underthrusting the North American plate at the present time. Leveling surveys across western Washington over the past 70 years indicate a regional down-to-the-east crustal tilting, which Ando and Balazs relate to continuous creep on the interface between the two converging plates. They state that pre-seismic deformation along active subduction boundaries characterized by thrust-type earthquakes (for which there is no evidence in the Pacific Northwest) should be characterized in western Washington by tilting of the coastal area toward the offshore trench (i.e., to the west).

The relative direction of convergence between the North American and Juan de Fuca plates, assuming that such convergence continues today, is imperfectly known. Recent estimates for the vector of convergence between North America and that segment of oceanic crust generated along the Juan de Fuca Ridge range from N 39° E to N 57° E, all at about 4 cm per year (Washington Public Power Supply System, 1977a); the most recent estimate, by Riddihough (1979), is approximately N 55 ° E. A northeastward direction of convergence is at a high angle to the Washington segment of the plate boundary, but oblique enough to conceivably impose a dextral component of strain in the upper, North American plate.

Rogers (1979) suggests that the north-south compressive stress responsible for earthquakes in the Vancouver Island region may be the direct consequence of north-south convergence between the small northerly Explorer plate and North America. This possibility cannot be discounted, but the southerly continuation of that stress field into areas east of the Juan de Fuca-North American plate boundary (with its different vector of convergence) and into northern and western California (astridge the Pacific-North American transform plate boundary) favors a more regional pattern of plate interaction as its cause. Specifically, the stress field may be the combined consequence of well-documented dextral transform motion between the Pacific and American plates and oblique, northeast-southwest convergence between the Juan de Fuca and American plates. The two major segments of the Pacific-American plate boundary, extending northward from the mouth of the Gulf of California to the Mendocino fracture zone (San Andreas fault), and from offshore Vancouver Island to the Gulf of Alaska (Queen Charlotte Islands fault), are sharply defined, but strain effects related to the boundary appear to extend far inland, particularly in areas adjacent to the San Andreas segment (e.g., the western Great Basin).

The idea that the late Cenozoic tectonics of the western United States are related to a broad zone of dextral shear along the western continental margin was popularized by Atwater (1970). Her concept of a broad "transform zone" of plate interaction was essentially inherited from similar, pre-plate tectonics hypotheses of Carey (1958), Eardley (1962), Wise (1963), and Hamilton and Myers (1966). More recently it has been variably developed by Smith (1977, 1979), Eaton (1979), and Christiansen and McKee (1979), all of whom emphasize the extensional phenomena of the Basin and Range province as the critical aspect of late Cenozoic plate margin deformation.

3.1 KINEMATIC INTERRELATIONS BETWEEN THE GREAT BASIN AND THE BLUE MOUNTAINS AND COLUMBIA PLATEAU

3.1.1 INTRODUCTION

A number of recent workers (e.g. Washington Public Power Supply System, 1977a, 1980; Zoback and Thompson, 1978; Eaton, 1979) have called attention to the fact that essentially synchronous ENE-WSW crustal extension occurred in the Great Basin (by normal faults and igneous dikes) and in the central and eastern Blue Mountains and Columbia Plateau (basaltic dikes) beginning about 17 m.y. ago--although the amount of extension greatly diminished northwards into the latter areas. Because plateau basalts continued to be erupted from northerly striking dikes until at least 8.5 m.y. ago (Ice Harbor dikes near Pasco, with strikes of N 20° - 30° W; Swanson and others, 1979), the view has been widely held (e.g. Washington Public Power Supply System, 1977a; Eaton, 1979; Laubscher, 1981) that the eastern Columbia Plateau and the Great Basin shared a common extensional strain history until then (recognizing that plateau extension was much less than that of the Great Basin). That view, in light of new information about the structural evolution of the Columbia Plateau (see below) must now be regarded as overly simplistic, and probably misleading to those who build tectonic syntheses upon it.

3.1.2 OLIGOCENE AND EARLY MIOCENE EVENTS

As a prelude to further discussion of the point just made, a comparison of the Oligocene and Early Miocene histories of the Great Basin, Blue Mountains, and Columbia Plateau prior to the inception of extensional deformation 17 m.y. ago is in order. Between 40 and 20 million years ago volcanic arc activity related to eastward subduction of oceanic lithosphere beneath western North America was widespread. Such activity affected the central longitudinal third of Washington, most of Oregon with the exception of coastal and northeasternmost areas, and all of what is now the Great Basin (Snyder and others, 1976; Armstrong, 1979). Coney (1979) describes this time period, particularly in the Great Basin area, as one of (1) great ash-flow or ignimbrite eruptions from caldera centers, and (2) the development of amphibolite-grade metamorphic-mylonitic complexes at relatively shallow crustal levels (i.e. into Paleozoic and Mesozoic strata; cf. Compton and others, 1977; some of these complexes are now dated as Mesozoic in age, e.g., the Whipple Mountains of southeastern California; this author, unpublished studies). Loring (1976) has documented widespread examples of Oligocene normal faulting of variable

trend (north-south to east-west) within Utah and Nevada, but two lines of evidence suggest that topography and, by inference, tectonic activity were generally subdued. A paucity of Oligocene and Early Miocene sedimentary rocks in the Great Basin suggests that there were few major basins for sediment entrapment (Christiansen and McKee, 1979), and sheetlike Late Oligocene-Early Miocene ignimbrites mantled a topography of low relief over regionally extensive areas 25 to 20 million years ago.

The Oligocene to Early Miocene history of the Columbia Plateau is obscured by its cover of Columbia River basalts and younger units, but the history of the Blue Mountains during this time period is at least partly recoverable. The province was deformed at some time after Eocene arc volcanism (Clarno Fm.), but prior to Columbia River basaltic volcanism. Clarno strata were folded locally as steeply as 60^{o} before deposition overlying Picture Gorge basalts (Thayer and Brown, 1966) on the northern limb of the major east-west Aldrich Mountain anticline, the structural axis of the western Blue Mountains. Robyn (1977) suggests that this deformation occurred between 25 and 20 m.y. ago, but his older age limit is not well-constrained. If folding was earlier, deposition of the Oligocene (less than 43 m.y. ago) to Early Miocene John Day Formation may have been limited to the south by the high-standing Blue Mountains block (Thayer and Brown, 1966), and partly controlled by synchronous folding in the basin of deposition. According to McKee (1972, p. 236-237), "The John Day Formation appears to be thickest near the centers of northwest-trending synclines. The folds are, in part, younger than the Columbia River Basalt, as that unit has been warped by the fold. The thickness data from the John Day strata suggest that the same folds existed in the Oligocene and that deformation along northwest-trending lines has been going on for at least 30 million years." Supporting evidence for early folding (pre-Columbia River Basalt) comes from the Blue Mountains anticline along the northern edge of the John Day basin. This major fold constituted a topographic barrier in Middle Miocene time that separated, except locally, flows of the Lower Yakima and Picture Gorge Basalts (Nathan and Fruchter, 1974). McKee's suggestion, that folding with trends and geometries characteristic of the Columbia Plateau began in Oligocene time, is an extremely important one when the genesis of the younger folds that deform the Columbia River basalts is considered.

3.1.3 POSSIBLE KINEMATIC LINKAGE BETWEEN GREAT BASIN AND THE PACIFIC NORTHWEST

As reviewed below, compressional deformation in the Blue Mountains and Columbia Plateau may have peaked in Late Miocene and Pliocene time, roughly between 10.5 and 3 to 4 m.y. ago. This time period was significant for the Great Basin area as well, since it includes the kinematic transition between east-west extension within the northern and western Great Basin to a west-northwest to northwest direction of extension. According to Eaton (1979) this change in "spreading" direction occurred between 7 and 4 m.y. ago and accompanied the progressive northward development in California of the transform boundary between Pacific and North American plates.

The general coincidence in timing between (1) the onset of west-northwest to northwest extension in the western Great Basin and (2) an acceleration in similarly oriented (although not necessarily aligned) compression in the provinces to the north, may be interpreted to indicate that a kinematic linkage exists between them. Laubscher (1977), for example, referred in general terms to the Columbia Plateau as a compressional "sink" for extensional strain in the Great Basin to the south (the "source"). He does not believe, however, that the large amounts of extension in the southern region are equally compensated by compression in northern areas. Such a direct kinematic linkage is geologically unrealistic, considering the relatively small compressional strain of the Columbia Plateau and Blue Mountains areas.

As outlined above, folding in north-central Oregon along northeast trends began <u>before</u> Columbia River basalt volcanism and the initiation of east-northeast-west-southwest extension in the Great Basin area. Plateau folding continued from 14 to 7 million years ago--within which time interval, according to Eaton (1979) the direction of extension in the Great Basin changed to east-west. Thus to this writer, an explanation for subsequent, but similar plateau deformation that is tied to a radical reorientation of Great Basin extension from east-west to a northwest-southeast direction (ca. 7 m.y. ago) seems forced. Nevertheless, the entire question of kinematic linkage between extension in the Great Basin and the tectonics of the Pacific Northwest has become enormously complicated by the just-published paper of Magill and Cox (1981). These writers conclude that the entire Cascade Range south of southern Washington has rotated about 27° in a clockwise sense since 25 m.y. ago. They attribute this

impressive rotation, which may have affected the Klamath and Sierra Nevada provinces as well, to post-early Miocene extension in the Basin and Range province. The implications of their paper are beyond the scope of this report but may be profound.

3.1.4 NORTHERN MARGIN OF BASIN AND RANGE PROVINCE IN CENTRAL OREGON

A correct understanding of the kinematic interrelationships between the Basin and Range province and the Blue-Mountains/Columbia Plateau must focus on the tectonic boundary between them in central Oregon. There is, however, no unanimity of opinion on exactly how this intraplate provincial boundary is defined. Some writers (e.g. Smith, 1979; Christiansen and McKee, 1979) speak of the High Lava Plains of central and eastern Oregon, with its youthful volcanism and high heat flow, as a geologic transitional zone between the two provinces. Others (e.g. Lawrence, 1976; Eaton, 1979) emphasize structural elements, most importantly the northwest-striking Brothers fault zone, as defining the boundary. Laubscher (1981), most recently has considered the Brothers fault zone as a portion of the southern margin of a broad zone of dextral transform strain extending from western Idaho to the Olympic Mountains (IDOL). However defined, the boundary has generally been accepted as one characterized by dextral transform motion since the interesting paper by Lawrence (1976).

Reasons for postulating dextral slip along a tectonic boundary between the Basin and Range and Blue Mountains provinces include the following: (1) an en echelon pattern of fractures within the Brothers fault zone, reported by Lawrence (1976) and interpreted by him as Riedel shears in a youthful zone of limited dextral shear; (2) eastward termination of the Brothers zone at the throughgoing, north-south Steens normal fault, a geometry necessitating a westward dilation of the normal faulted terrain south of the Brothers zone with respect to the northern, more stable block (Lawrence, 1976); (3) Lawrence's contention that the northwestern end of the Brothers zone merges with a more northerly striking system of normal faults along the eastern edge of the High Cascades; and (4) the northwesterly trend of the Brothers zone--an orientation within the dextrally strained margin of the North American plate that should be characterized by components of right-slip (cf. Atwater, 1970).

Lawrence (1976, p. 849) states that "relatively little displacement has taken place along the Brothers fault zone",

a conclusion compatible with the absence within the zone of a single, throughgoing strike-slip fault. However, his interpretation of right-slip along the Brothers zone--based on an inferred en echelon pattern of fractures within it--deserves scrutiny. Lawrence claims that the N 60° W trend of the zone contrasts significantly with a N 40° W strike of discontinuous, en echelon fractures within it, and that the acute angle between these structural elements indicates right-lateral displacement along the zone.

The most strongly en echelon structural elements of the Brothers zone are a series of northwest (N 50° W ± 10°)-trending ridges that lie north of Oregon State Highway 20 between Brothers and Horse Ridge Summit west of Millikan, and continue south and west of the summit as components of Horse Ridge (Figure 1). Lawrence interpreted these ridges as "fractures" on the basis of his analysis of Landsat imagery (they are plotted within the zone in his Figure 2). However, at least two of the westernmost component ridges of Horse Ridge appear to be elongated, doubly plunging anticlines based on the configuration of lava flows within them. The morphology of the easternmost two ridges of Horse Ridge and those ridges farther east (north of Highway 20) is also indicative of folding. The ridges rise as much as 700 feet above the surrounding lava plains and lose elevation both northwestward and southeastward along their axes. Most are asymmetric, with steeper southwestern flanks. Their grand isolation above the largely unfaulted lava plains would seem to rule out an origin for them by block faulting of Basin-and-Range type (i.e. normal faulting). Scarplike slopes on the southwestern margin of most of the ridges appear to be fault controlled, possibly with reverse or thrust components of slip in light of their association with probable anticlinal folds. This writer's tentative interpretation of the ridges as compressional features (made in the summer of 1979) has subsequently received independent support from an unpublished photogeologic study of the region made by a major oil company in the mid-1950's. In that study, the ridge directly east of Horse Ridge Summit was mapped as a northerly-plunging anticline. If the dozen or so ridges in question are indeed anticlinal, their pronounced en echelon orientation (Figure 1) suggests strongly that they are brachyanticlines formed within, or bordering on the north, a zone of sinistral (left-lateral) shear with a strike in this area of approximately east-west. En echelon ridges of similar trend and morphology are also present south of Highway 20 in the same general area (e.g. Pine Ridge), and because they too rise abruptly above the level of the surrounding lava plains, they are also favored as compressional features. The ridges

of the Millikan-Brothers area thus appear to complicate Lawrence's interpretation of dextral slip within the Brothers fault zone, based on his erroneous assumption that the ridges are expressions of Riedel shears with transcurrent displacment.

Another aspect of Lawrence's tectonic analysis also needs reconsideration. He (1976, p. 850) believes that the northwest-trending Brothers zone separates the Basin and Range province from the Blue Mountains of central Oregon. "Extension virtually ceases at the northern end of the province along the Vale and Brothers fault zones." However, the Brothers zone as he has defined it from Landsat imagery (his Figure 2) appears much too selective when compared with patterns of faulting mapped by Green and others (1972) in the Burns 2° map sheet (see also Walker, 1977). Lawrence shows the Brothers zone as a linear zone that extends southeastwards from Bend and passes south of Harney Lake. Yet Greene and others map scores of minor normal faults north of Lawrence's zone between Hampton (on Highway 20) and Harney Lake that are identical in field appearance and trend to those of the zone as Lawrence defines it. In other words, the eastern linearity and location of Lawrence's zone can be questioned. Certainly, the northern limit of northwest-striking normal faults of the Basin and Range province does not lie along the Brothers zone in this region, but in a more northerly position--roughly along an irregular east-west line connecting Hampton and Burns. Inspection of the Burns map sheet (Greene and others, 1972) suggests that the deformation of the High Lava Plains west and southwest of Burns is more one of crustal extension in a northeast-southwest direction across scores of normal faults than dextral slip across northwest-striking Riedel shears. The northwest orientation of the normal faults is, of course, compatible with right-lateral displacement of the High Lava Plains relative to the northern, Blue Mountains block if the faults are extensional fractures in a zone of distributed dextral strain.

4.1 TEMPORAL AND SPATIAL RELATIONS BETWEEN EXTRUSION AND DEFORMATION OF COLUMBIA RIVER BASALTS

Tectonic models for the origin of Columbia Plateau structures must be cognizant of the relations in space and time of the two major geologic events of the plateau: (1) the extrusion of basaltic lavas from north-northwest trending dikes; and (2) the deformation of these lavas and coeval and younger sedimentary rocks along northeast, east, and northwest trends. The former is known to have occurred from approximately 16.5 to 6 m.y. ago, although Swanson and

Wright (1978) estimate that 99% of the lavas were extruded between 16 and 14 m.y. ago. As amplified below, deformation of the basalts began about 14 m.y. ago and continues to the present.

Figure 2 draws attention to an important spatial relationship between the two plateau phenomena. That is, that with the exception of basalts among the oldest (Picture Gorge) and youngest (Ice Harbor), the feeder dikes for the Columbia River flows were in eastern plateau areas unaffected by plateau folding. To state this relation differently, the Columbia Plateau can be divided into two areas with differing late Cenozoic strain histories--an eastern area, the Palouse subprovince of Rockwell (1979), characterized by east-northeast extension totalling more than 1 km (Taubeneck, 1970), and a western area, the Yakima fold belt subprovince, characterized by north-south compression. Bentley (1980) has recently estimated that north-south shortening across the subprovince decreases from a value of 15 km, west of 120° W longitude, to 4 km, east of 120° W. Although maximum development of structures within the two strain fields did not overlap (dikes, 16 to 14 m.y. ago; folds, 10.5 to 3 or 4 m.y. ago), the fields did coexist for approximately 8 million years (14 to 6 m.y. ago).

Because total strains within each field are small, the structural boundary between them is not clearly expressed. Laubscher (1977, 1981) refers to the eastern, north-south trending boundary of the Yakima subprovince as the Wallula Gap-Moses Lake Belt. It connects the eastern topographically-expressed terminations of the Saddle Mountain and Frenchman Hills structures. He believes the diffuse belt is characterized by dextral transcurrent strain since it separates a north-south-shortened block to the west (Yakima) from the more stable Palouse block to the east.

5.1 TIMING OF PLATEAU FOLDING AND ASSOCIATED FAULTING

5.1.1 MIOCENE DEFORMATION

It has become clear in the last several years that plateau folds and parallel faults, usually with thrust or reverse fault geometries, were developing during the extrusion of some Miocene Columbia River basalts from fissures farther east. The best evidence for the earliest recognized deformation, that of late Grande Ronde time after nearly 85% of the basalts had been extruded, comes from the recent work of Beeson and Moran (1979) in the northern Cascades of Oregon. They report that folds trending N 40° - 65° E began to develop during youngest Grande Ronde time. At some

localities pillow lavas and sedimentary interbeds of this age are restricted to synclinal troughs and anticlinal volcanic sections are abbreviated. Similarly at the northern edge of the Columbia Plateau, in the Wenatchee Mountains of Washington, the Vantage Member of the Ellensburg Formation (post-Grande Ronde Basalt, pre-Frenchman Spring Member of Wanapum Basalt) thins toward the crest of Taneum Ridge anticline (Meyers and Price, 1979, p. lV-16).

Evidence for deformation during Wanapum time, approximately 14.5 to 13.6 m.y. ago, is geographically more widespread. Again, in the northern Oregon Cascades, Beeson and Moran (1979) found evidence for uplift, northeast-southwest faulting, and erosion of the Frenchman Springs member of the Wanapum Basalt prior to deposition of Priest Rapids flows (also Wanapum Basalt). In the Pasco Basin area, stratigraphic and geochemical studies by geologists of the Rockwell Hanford group (Reidel and others, 1980; Rockwell, 1979) indicate that the Saddle Mountains structures (Figure. 3) began to form in post-Grande Ronde time, ca. 14.5 m.y. ago. Thinning of Frenchman Springs flows has been noted across the northwest-trending Smyrna anticline on Saddle Mountain. The distribution of overlying flows of the Roza Member of the Wanapum Basalt on Saddle Mountain suggests east-west folding or arching during their deposition.

Evidence for post-Wanapum deformation is still more widespread. Bentley and others (1980, p. 59) state that "broad, structurally controlled basins had become noticeable at the onset of Saddle Mountain time, about 13 - 13.5 million years ago." They report that anticlinal ridges were locally high enough to confine flows of the oldest member of the Saddle Mountain Basalt, the Umatilla Member. These flows thin over the present site of the Rattlesnake Hills, suggesting that this fold structure began to form at the close of Wanapum time (Rockwell, 1979). The thinning of flow units across Saddle Mountain noted above continued to occur until Elephant Mountain time (ca. 10.5 m.y. ago), enabling Reidel and others (1980; oral presentation, 1980) to conclude that the rate of uplift of this structure between 14.5 and 10.5 m.y. ago was approximately 39 meters/million years.

Some of the clearest evidence for intra-plate Miocene deformation on the Columbia Plateau comes from the Yakima and Umtanum anticlinal structures (Figure 3). Bentley (1977, p. 339) describes "locally substantial deformation...14 to 12 m.y. ago" near Priest Rapids (Umtanum Ridge) and Yakima (Yakima Ridge). In the latter area, just

east of the Yakima River, steeply-dipping Wanapum basalts (as young as the Priest Rapids Member) are overlain with angular unconformity by flat-lying Selah conglomerates of the Saddle Mountains Basalt (Bentley, 1977, p. 364); the conglomerates are older than the 12 m.y. old Pomona Member of the Saddle Mountains Basalt.

Somewhat similar relations are reported by Bentley (1977, p. 377 and summarized in Rockwell, 1979, p. III-166) from the Filey Road area of eastern Umtanum Ridge. Here, 3 km west of Priest Rapids Dam, Grande Ronde and Wanapum basalts, including the Priest Rapids Member of the latter, were tightly folded, overturned, reverse faulted, and eroded prior to the deposition of fanglomerates across them. The fanglomerates intertongue northward with fluvial sediments of Selah age that underlie the Pomona Member. The Pomona basalt in this area is described as relatively undeformed" (Rockwell, 1979, p. III-166; see also their Figure III-73).

Farther east, Goff and Myers (1978) also conclude that most deformation of Umtanum Ridge occurred prior to Saddle Mountain time, but the distribution of the 10.5 m.y. old Elephant Mountain basalts around the eastern end of the structure suggests to them that some folding continued through Elephant Mountains time. Bentley (1977, p. 374) states that more than 400 m of structural relief has developed at the eastern end of Umtaum Ridge since the extrusion of Elephant Mountain lavas.

Despite evidence for geographically widespread deformation of the Columbia Plateau prior to extrusion of the 12 m.y. old Pomona lavas, most deformation affecting the Columbia Plateau and the adjacent Blue Mountains province is post-Pomona in age. In the Blue Mountains, strong north-south compressional deformation postdated the extrusion of Columbia River basalts, but preceded eruption 6.6. m.y. ago of the wideapread tuff member of the Rattlesnake Volcanics (Thayer and Brown, 1966; Robyn and others, 1977; Robyn, 1977). Robyn (1977) dates this compressional event as occurring between 10 and 7 m.y. ago. Deformation was expressed by renewed folding and rupturing of the north flank of the Aldrich Mountain anticline to form the east-west John Day reverse fault. Northeast and northwest-striking fractures and strike-slip faults formed contemporaneously in the northern footwall of the fault as a conjugate response to north-south shortening. Only minor compression and igneous activity (silicic intrusions and basaltic volcanism) have occurred in the eastern Blue Mountains since 6.6 m.y. ago (Robyn, 1977).

In Washington, the widespread folding of the Elephant Mountain Member of the Saddle Mountains Basalt in the Pasco Basin area demonstrates major plateau deformation younger than the 10.5 m.y. age of the member. Rockwell (1979, p. lV-17, 20-21) conclude that most deformation in the Pasco Basin area occurred between 10.5 and approximately 5. m.y. ago, although the younger age limit is not well-controlled (as discussed below).

5.1.2 PLIOCENE DEFORMATION

A problem in establishing an upper (younger) limit for most plateau folding and associated faulting is the incomplete Pliocene and Quaternary stratigraphic record of the plateau area. In many areas, late Quaternary deposits rest directly on deformed Miocene basalts, with a resultant hiatus in the geologic record of the region of up to 16 million years.

Pliocene stratigraphic units (ca. 5 to 3 m.y. old) are preserved in the northern and central Pasco Basin area (Ringold Formation), the Yakima-Ellensburg area (upper Ellensburg Formation), and in southwestern portions of the Washington plateau (Simcoe Volcanics). Unfortunately, plateau workers have differed on the extent of plateau deformation affecting these units, largely because of the lack of precise age controls on them. For example, in the northern Pasco Basin dips of beds in the upper part of the Pliocene Ringold Formation (5.1 - 3.3 m.y. ago, Rockwell, 1979) are very gentle. This relation may indicate waning deformation (Rockwell, 1979, lV-21) since lower Ringold strata on the flanks of Saddle Mountain to the north dip as steeply as 40° (Washington Public Power Supply System, 1977d, p. 2RH 8-6). However, since basalts beneath the gently dipping Ringold sediments also dip gently, the areal differences in Ringold dips may reflect spatial rather than temporal factors.

Unpublished subsurface studies by Golder and Associates (D. Caldwell, personal communication, 1981) of Ringold sediments that lie across the buried southeastern end of the Gable Mountain anticline (Figure 3) indicate that older Ringold sediments on the flanks of the structure dip more steeply than younger. Golder's data suggest that broad arching of Ringold sediments continued through the Pliocene.

Pliocene fanglomeratic deposits in areas between Yakima and the Kittitas Valley, e.g. the Thorp Gravel, have recently been assigned to the upper Ellensburg "Formation" (Washington Public Power Supply System, 1977d) or "group" (Bentley, 1977). Statements made in the two 1977 papers

about the significance of these deposits to the dating of plateau deformation are, however, somewhat at variance. Bentley states (1977, p. 355) that "major deformation" occurred along Manastash Ridge "after some coarse basaltic fanglomerates of the Thorp (?) Gravel were deposited", despite a reference elsewhere (p. 352) that Thorp (?) conglomerates on the north flank of the Manastash structure dip only 3 - 5°. The inclined conglomerates are truncated by an early (?) Pleistocene pediment surface capped by gravels. The age of the Thorp (?) unit is not known, although tephra layers in upper Thorp Gravels in Kittitas Valley have yielded fission trace and K-Ar ages of 3.7 to 4.8 m.y. ago (Rigby and Othberg, 1979, p. 18). From such relations, Bentley (op. cit., p. 339) derives the tenuous conclusion that "the majority of the 'ridges' rose in Pliocene-Pleistocene times" (6 to 1.5 m.y. ago).

Washington Public Power Supply System (1977d, p. 2RH 8-3), however, although favoring local deposition of upper Ellensburg deposits "during or after" ridge uplift take a more conservative view of the significance of these deposits: "The regional significance and age of these deposits is poorly understood. Most deposits are <u>undeformed</u>; only on west end of Smyrna Bench (Saddle Mountain) and in the Kittitas Valley do small dip slip faults cut these deposits" (italics by this writer). Elsewhere, (p. 2RH 8-12) Washington Public Power Supply System concludes that the

> "age of deformation in the area of investigation is dominantly post-Elephant Mountain and pre-Thorp Gravel. Some deformation along the cores of the anticlines may be slightly older and minor tilting of pediment surfaces may be younger, but the majority of the deformation (faulting and folding) must have occurred between these two dates."

Their conclusion is, in turn, challenged by Rigby and Othberg, (1979) who report that Ellensburg sedimentary rocks younger than Columbia River basalts (their "supra-basalt Ellensburg Formation") are deformed widely in the Yakima area:

> "The deformation exhibited by the supra-basalt Ellensburg Formation indicates that the large-scale uplift and deformation of the basalt ridges in the western Columbia Basin occurred after most, if not all, of these sediments had been deposited, or late Miocene to early (?) Pliocene in age."

Basalt terrace gravels preserved in valleys in the Yakima area are interlayered with and overlie supra-basalt Ellensburg sediments. The gravels, which according the Rigby and Othberg resemble the Thorp Gravels of the Kittitas Valley, exhibit near vertical to overturned dips at localities along Ahtanum Ridge, in the foothills of Cowiche Mountain, and in Yakima. Unfortunately, the age of these gravels is not known, although Rigby and Othberg (1979) suspect they may be broadly coeval with the Thorp gravels. They believe (p. 23) that some of the terrace gravels "were deposited before deformation began, while others were deposited for some time after deposition occurred. Deformation of this part of the Columbia Basin, at least, apparently came to an end sometime during deposition of this gravel unit."

Pliocene and Pleistocene lavas are widespread in the southwestern corner of the Washington portion of the Columbia Plateau. The relations of these lavas, the Simcoe volcanics, to plateau fold and fault structures are instructive. According to Bentley and others (1980, p. 60):

> "Eruptions of basalt and related lava began in the Simcoe volcanic field probably during early Pliocene, about 4-5 million years ago. Much of the deformation of the area had been completed before these eruptions, although some flows appear to have been tilted by later folding. The eruptions, which may have taken place over a 2-4 million year period, produced a broad continuous basalt field dotted with cinder cones".

Disagreement exists in the literature concerning the relative age of east-to-northeast-trending anticlines and northwest-striking, high angle faults in this part of the plateau. One such fault appears to cut a 4.5 m.y. old Simcoe flow, but is overlapped by a flow dated at 3.5 m.y. (Shannon and Wilson, 1973). Shannon and Wilson (1973) interpret the 4.5 to 3.5 m.y. old fault as being synkinematic with formation of the Columbia Hills and Horse Heaven anticlines, but studies by Anderson (1980) and Bentley and others (1980) indicate that the northwest-striking cross faults are younger than the folds (see also Rockwell, 1979, p. II-84, 85).

5.1.3 PLEISTOCENE DEFORMATION

Although upper units in the Simcoe Volcanics and the Ringold and Ellensburg Formations may be of Pleistocene age, the Pleistocene stratigraphic record on the plateau is generally represented by poorly dated glacial deposits, flood gravels

and associated sediments, and loess (Rigby and Othberg, 1979). The extent of involvement of these units in plateau folding is uncertain, in part because they tend to be best preserved in basins between major folds, and in part because of their youthfulness and the likelihood of low plateau strain rates.

All workers appear to agree that the uplift of Yakima Ridge ended prior to one million years ago, since the undeformed Tieton Andesite (K/AR age of 1.0 ± 0.1 m.y.) lies in an erosional reentrant across the truncated northern flank of the anticline (Rockwell, 1979, p. 11-77). Bentley (1977, p. 354) concluded that most of the deformation along nearby Manastash Ridge occurred prior to the development of the one m.y. old Thrall pediment surface on its north flank.

6.1 HOLOCENE AND CONTEMPORARY TECTONICS

6.1.1 INTRODUCTION

Information on the geometry and kinematics of active structural elements in the Pacific Northwest comes primarily from two sources -- field recognition and study of such structures, and fault plane solutions from single or composited seismic events. From a practical standpoint, relatively little is known about Quaternary tectonics in Washington from the direct study of active structural elements. None of the seven states west of the Rocky Mountains have fewer known or inferred examples of active faults than the state of Washington (Howard and others, 1978). This situation may reflect a genuine diminishment of youthful faulting in areas of the United States north and west of the Basin-and-Range province or a lack of bedrock exposures in the western, more seismically active half of the state -- or both.

6.1.2 DATA FROM SEISMOLOGY

Washington Public Power Supply System (1977a) summarized evidence, largely from earthquake focal mechanism studies, that the upper lithosphere (less than 25 km depth) of Washington and the northern half and western third of Oregon presently lies within a regional strain field characterized by north-south shortening. Individual and composite focal mechanisms for seismic events within this region typically yield orientations for P, the compressional axis, that have shallow plunges and trends that vary from NNW to NNE. T, the axis of maximum extensional strain, has a more variable orientation, ranging from vertical to east-west and horizontal. The two principal combinations of P and T that

result from these orientations lead to two characteristic types of fault plane solutions for Pacific Northwest earthquakes: 1) east-west striking thrusts or reverse faults (north-south P, vertical to subvertical T); and 2) either steep northwest-striking faults with dextral slip or northeast-striking faults with sinistral slip (north-south P, horizontal to subhorizontal and east-west T). Rogers (1979) has recently documented that the field of north-south shortening extends into the Vancouver Island region of southwestern Canada, where solutions for both strike-slip faulting (predominant) and thrust faulting (subordinate) are also obtained.

However, results of recent seismic studies from the Columbia Plateau indicate that the orientation of P there is considerably more variable than the simple picture of north-south orientation that was presented in WPPSS PSAR Amendment 23 by Washington Public Power Supply System (1977a), Washington Public Power Supply System (1977e), and Washington Public Power Supply System (1977f). In a draft report (Woodward-Clyde Consultants, 1980a) to the Washington Public Power Supply System, orientations of P have been found to vary markedly in different parts of the southeastern Plateau region. For example, composite focal mechanism solutions from microearthquakes with depths of 9 to 24 km yield N-S orientations of P for a more-or-less longitudinal belt (ca. 119'30'W to 119'45'W) that extends southward from east of Priest Rapids Dam to within 15 km of the Columbia River west of McNary Dam. All composite solutions show predominant reverse, dip-slip displacements.

In contrast, composite focal mechanism solutions for areas both east (near Mesa, Washington) and west (near Sunnyside) of the longitudinal belt yield P orientations that are regionally anomalous (N 67° W, 43° and N 74° W, 60°, respectively, for the two areas). Orientations of T for the two areas are not as consistent as those for P, being N 32° E, 9° and N 106° E, 30°, for eastern and western areas respectively. These solutions suggest pure dip-slip normal displacement along north-striking normal faults in the latter case, and faults with oblique-slip (both strike-slip and normal dip-slip compontents) in the former. The solution for normal faulting is regarded by Woodward-Clyde as inconclusive, although that for oblique faulting is described as "fairly well constrained." Similarly, two of the largest historic earthquakes of the southeastern Plateau region, those of July 16, 1936 ("Milton-Freewater", M =6.1, Woodward-Clyde Consultants, 1980b), and April 8, 1979 (College Place, M = 4.1, ibid.), have also yielded fault plane solutions with marked

departures of P from a regional N-S orientation, i.e. N 56° E, 22° and N 91° W, 1°, respectively. The solution for the 1936 event is poorly constrained, but that for the 1979 event appears quite reliable.

Reasons for the characteristic north-south orientation of P in the Pacific Northwest are unresolved, as are reasons for the local or areally restricted departures from that orientation in southeastern Washington. The general consistency of focal mechanism solutions for Washington, northern and western Oregon, and southeastern British Columbia (Washington Public Power Supply System, 1977a; Rogers, 1979) is, nevertheless, important in documenting the regionality of north-south crustal shortening (shallower than 25km) in these areas. Nowhere in the Pacific Northwest are mappable structural features related to north-south shortening better developed than in the Columbia Plateau, although the geologic record of this area (reviewed above) indicates that most of its east-trending folds and parallel thrust and reverse faults are pre-Quaternary in age.

6.1.3 LATE QUATERNARY FAULTING

The low-level seismicity of portions of the Columbia Plateau attests to ongoing deformation of the region, despite the fact that in most areas, seismicity is not associated with surface manifestations of faulting. Surface ruptures of Quaternary age, essentially unknown in 1977 at the time of submittal of PSAR Amendment 23, have since been recognized in three areas of the Plateau: 1) Toppenish Ridge, approximately 80 km west of Richland (Campbell and Bently, 1980); 2) about 25 km southeast of Richland, from the vicinity of Wallula Gap on the Columbia River southeastward to the Walla Walla/Milton-Freewater area (Shannon & Wilson, 1980); and approximately 40 km north of Richland on the eastern end of Gable Mountain.

6.1.3.1 Toppenish Ridge

Campbell and Bentley (1980) report that the summit, north flank, and alluvial fans at the base of the north flank of Toppenish Ridge (Figure 3) are broken by nearly 95 surface ruptures up to 9 km in length. Most of the ruptures are less than 1 km long and only six have lengths in excess of 3 km. The faulted zone varies in width from 0.5 to 2.2 km and has a length, more-or-less parallel to the ridge, of 32 km. Most flank and summit ruptures are sub-vertical faults for which no strike-slip displacement is evident. Faults at the base of the north flank are generally lobate in form and are interpreted as comprising a thrust zone coincident with the

older south-dipping Toppenish fault. Some cut glaciofluvial slackwater deposits of Touchet type and others displace post- "Touchet" alluvial fans. Mt. St. Helens ash (12,800 y.b.p.) is present in some of the slackwater deposits, but has not yet been shown to be faulted. Campbell and Bentley (1980) attribute the Toppenish faults to north-south compression, with thrusting at the northern base of the anticlinal structure and extension across its hinge (Figure 4). Bentley (1980, personal communication) believes that the 30 km-long faulted segment of Toppenish Ridge may be terminated at its western and eastern ends by northwest-striking strike-slip faults, which may be responsible in some way for localizing Quaternary rupture along the anticline.

Although the surface ruptures on and adjacent to Toppenish Ridge may represent Quaternary tectonic activity, an alternative explanation appears viable at this time. Specifically, the combination of apparent low-angle thrusting along the northern base of the ridge and normal (extensional) faulting at higher elevations raises the possibility that the two are expressions of gravitationally-induced slope failure (Figure 4). Several lines of evidence support the interpretation that the Toppenish structures represent either an aborted phase or the incipient development of massive slope failure along the north flank of Toppenish Ridge. Included among them are the dramatic north flank landslides on the eastern end of the ridge and around Ortney Lake, immediately to the west. These slides demonstrate in spectacular fashion the instability of the northern flank. The close spatial relations between apparent low-angle faults at the base of the ridge and high-angle faults at higher elevations are suggestive of an interrelated landslide toe and headscarp geometry. An active youthful landslide at the south end of the White Bluffs, on the Columbia River opposite Richland, is characterized by just such prominent headscarps and a subhorizontal plane of movement near the base of the slope (Kiel, personal communication, 1980). Bentley and others (1980, p. 51) argue that a landslide origin of the Toppenish ground ruptures is "unlikely, because such an interpretation does not explain the abundance of south-side-down faults on the north slope of the ridge." However, such faults can be reasonably interpreted as antithetic faults to the main north-dipping sole fault. The White Bluffs landslide referred to above displays just such an antithetic rupture which defines, with the headscarp fault to the east, the edges of a shallow graben. Until the Toppenish ground ruptures receive further study, it is premature to conclude unequivocally that they represent tectonic activity.

6.1.3.2 Wallula Gap and Milton-Freewater Areas

Youthful faulting in the vicinity of Wallula Gap is apparently related to a zone of dextral transcurrent faulting (Shannon & Wilson, 1979a) that includes the topographically prominent Wallula fault zone (Figure 3). Bingham and others (1970) first noted features within the zone that suggested to them the possibility of Quaternary displacement. Recent studies (Shannon & Wilson, 1979a; Shannon & Wilson, 1979b; Shannon & Wilson, 1980) have documented or indicated the involvement in faulting of Quaternary units (including undated colluvium, Palouse Formation, Touchet Formation, and younger loess) at seven separate localities, the most western of which is at Finley Quarry on the northern end of the Butte (west of Wallula Gap). The three westermost localities (Figure 5) lie within a 20 km-long segment of the Wallula fault zone. Colluvium of probable Quaternary age is faulted at two localities (Finley Quarry and east of Warm Springs Canyon, Shannon & Wilson, 1979a, loc. C), and horizontal fault striae have been observed within a clastic dike of Touchet (?) silt and clay in basalts near Wallula Gap (Shannon & Wilson, 1979a, loc. A).

Bingham and others (1970, p. 77) believed that Touchet beds in an area east of Vansycle Canyon (east of Wallula Gap and between Shannon & Wilson's localities A and C) had apparently "been ruptured by relatively recent, minor fault movement." A curved linear topographic feature (referred to as "Bingham's linear" in subsequent reports) with a trend at nearly right angles to the drainage of the area was the basis for the conclusion that 12,000 year-old Touchet beds had been displaced by faulting. Preliminary results of trenching across "Bingham's linear" by Woodward-Clyde Consultants in early 1981, do not confirm the existence of a throughgoing fault across Touchet beds exposed in two trenches (D. Hitchcock, personal communication, 1981) although trench logging continues at the time of this writing.

Farther east, two possible fault localities in the Milton-Freewater area are in general alignment with the Wallula zone to the west (Figure 5). South of Umapine, north-dipping (30^{o} to 60^{o}) gravity slump surfaces or normal faults cut Touchet beds and cross-cutting clastic dikes with a maximum offset of 0.5 meters (Shannon & Wilson, 1979b). Youthful faulting 10 km farther to the southeast of the Umapine locality is suspected by Shannon & Wilson, but not demonstrated. Shannon & Wilson (1979b) found angular basaltic debris in loess along the trace of an inferred

bedrock fault. They suggest that the basalt fragments may have been derived from a fault scarp and were subsequently mixed with surficial loess deposits.

The other two eastern fault localities are the Buroker thrust fault east of Walla Walla, and the Little Dry Creek fault south of Milton-Freewater (Figure 5). The base of the Pleistocene Palouse Formation is offset approximately one meter along the former fault, a west-dipping (26°) reverse fault that strikes north-south. Higher loess deposits (Holocene?) appear to be unfaulted (Shannon & Wilson, 1980, p. 17) near Little Dry Creek, basalt and Palouse beds are downdropped along a steep (75°) north-east dipping fault about 0.5 meters. This fault lies south of an east-projected trace of the Wallula fault zone, and is not in alignment with it.

6.1.3.3 Gable Mountain

Faulting on the eastern end of Gable Mountain, 40 km north of Richland was studied by Bingham and others in 1970. They concluded 1) that south-dipping low-angle thrust faults on the northern and southern flanks of the structure were connected beneath an intervening cover of glacial flood deposits, and 2) that glaciofluvial deposits exposed in trenches above the faults had not been disturbed by faulting. Both conclusions have been shown to be in error on the basis of detailed trenching and borehole studies conducted since the summer of 1980 by Golder and Associates (D. Caldwell, personal communication, 1981). The two south-dipping thrust faults are separate faults, although the possibility that they merge into a single fault in areas to the southwest is still open. Flood gravels believed by Golder to be of Missoula age (latest Pleistocene - earliest Holocene) can be seen to have been thrust faulted in several trenches across the northern fault. Furthermore, clastic dikes which cut these glaciofluvial deposits and Miocene bedrock units, and which have been injected along both the northern and southern thrust faults, are sheared and striated at a number of localities. In partial support of the conclusions of Bingham and others (1970), glaciofluvial deposits which lie across southwestern portions of the northern fault and the eastern portion of the southern fault have not been disturbed by faulting. Golder's studies of the two faults are currently in progress. It is this writers opinion that both faults have experienced Quarternary displacement of tectonic origin.

7.1 OLYMPIC-WALLOWA LINEAMENT

The Olympic-Wallowa lineament (OWL), originally postulated by Raisz (1945) as a northwest-trending alignment of topographic features between the Olympic Peninsula, Washington, and the Wallowa Mountains, Oregon, is a cryptic feature of Pacific Northwest geology that may have bearing on the tectonic history of the Columbia Plateau. Raisz believed that the lineament was probably fault-controlled, but he stated (1945, p. 483) "that in most places the lineament is rather a zone than a line, with many parallel ridges and splinters...(p. 484) it appears to be a more complex structural line than a simple fault. It may have started as a transcurrent fault, but the line of weakness thus created probably suffered further dislocation." His reference to transcurrent faulting alludes to his perception from physiographic relations that both the crests of the Cascades and the Blue Mountains have been offset along the lineament for about six miles -- in a left-lateral sense.

Skehan (1965) suggested that the lineament may mark a fundamental boundary in the continent between former oceanic crust to the south and older continental crust to the north. Washington Public Power Supply System (1977a) proposed on geologic grounds that the basement for much of the Columbia Plateau on both sides of the lineament is Mesozoic oceanic crust, and associated sedimentary rocks, accreted to the continent prior to the Cenozoic era. His conclusion is generally supported by Hill's 1972 interpretations that the crust beneath the Plateau is 1) thinner than that of the granitic-metamorphic terrane of northern Washington by as much as 12 km, or 2) that it has an average P-wave velocity as much as 0.8 km/sec. higher than that terrane, or 3) that some combination of 1) and 2) prevails. In 1979, however, Hill concluded that the crust beneath the Pasco Basin is indeed thin (ca. 25 km minimum), but that it has a low P-wave velocity (ca. 6.1 km/sec.). This low velocity, if valid, is difficult to reconcile with an accreted basement of oceanic character. As an alternative, Laubscher (1981) has proposed that the basement is genuinely continental, but was thinned during early Tertiary regional doming -- evidence for which are the Eocene grabens of the northern Cascades and Okanogan terranes. He ascribed thinning of the crust to the combined consequences of east-west stretching, subareal erosion of the crest of the dome, and subcrustal "erosion" by processes unknown. This writer still believes that exposures of ophiolite rocks and associated marine sediments north (Ingalls ophiolite of Washington Cascades) and west of the plateau (Rimrock Lake) argue for an accreted oceanic basement beneath central areas of the plateau.

Whatever kind of crust underlies the Pasco Basin, recent geophysical studies support both Washington Public Power Supply System's (1977a) and Laubscher's (1981) contentions that this crust does not change across OWL. No evidence is seen in a recently compiled total Bouguer gravity anomaly map of the Columbia Plateau (1"500,000; c.i. = 4 mgal) that a change in basement rocks or crustal character occurs along the lineament (Weston Geophysical Research, 1981). Furthermore, a strong gravity gradient that separates the Yakima and Pasco Basins trends north-south across OWL. The gradient is so linear that the Weston report states (p. 23). "One edge of the causative body extends north-south with a density contrast that is positive with respect to the rocks toward the west. Because the contours are relatively straight, any faults striking between N 45° W and S 45° W that cross the gradient would have horizontal displacements less than two-three km." This conclusion supports Laubscher's contention (1977, 1981) that any strike-slip displacement along Raisz's lineament (and Laubscher's Cle Elum-Wallula lineament, see below) must be less than 2 km. Incidentally, the north-south gravity gradient referred to above, is interpreted by Laubscher as delineating the western edge of a master north-south trending graben of Eocene age that developed longitudinally along the crest of the regional dome he has postulated.

A recent study (Rodi and others, 1980) to model the three-dimensional structure of crust and upper mantle beneath the Columbia Plateau utilized joint inversions of regional Bouguer gravity data and P-wave travel-time residuals for teleseismic events recorded at stations in eastern Washington. The joint inversion model resulting from the study revealed no changes in crustal or mantle structure at depths greater than 10 km coincident with the surface trace of OWL.

It thus seems unlikely that the Olympic-Wallowa lineament is a fundamental or profound crustal break, or that diffuse transcurrent displacement along its inferred Plateau segment has been greater than a few kilometers since extrusion of the Miocene Columbia River basalts. Is it a throughgoing feature from the Cape Flattery area of the Olympic Peninsula to the linear northeastern margin of the Wallowa Mountains, Oregon (near Enterprise and Wallowa Lake)? Almost certainly not, for the reasons discussed below.

The northwest-trending, northern margin of the Olympic Peninsula, the western end of the Olympic-Wallowa lineament, is controlled by the strike and steep dip of Eocene units on the northern flank of a major, Miocene or younger antiform

which plunges steeply eastward beneath Puget Sound. The Eocene Crescent Volcanics define this antiformal structure, with their prominent horseshoe-shaped outcrop pattern around the northern, eastern, and southern margins of the Olympic Mountains (Tabor and Cady, 1978). A recent U.S.G.S. report on the geology of the Olympic Peninsula (Tabor and Cady, 1978) does not refer to Raisz's speculations about an Olympic-Wallowa lineament, but does show high-angle faults along most of the valleys believed by Raisz to define his lineament. The faults generally parallel steeply inclined bedding within Eocene units and much of their traces are mapped as concealed beneath Quaternary glacial deposits. Although they are thus difficult to evaluate in terms of their displacement history, they appear to this writer to be unlikely representatives of a hypothesized 650 km-long fault zone - a zone postulated in no small measure on their existence. There are no a priori reasons why structures related to the Olympic antiform should extend to the east of Puget Sound, and at present, no evidence that they do so.

Raisz believed that the easternmost segment of the lineament extended up the troughlike valley of the South Fork of the Walla Walla River and across the Blue Mountains to the Wallowa Mountains. Evidence has since accumulated to the contrary. Mapping by D. Swanson, USGS (unpublished), by Shannon & Wilson (1979b), and by R. Dale and J. Kendall (Kendall and others, 1981) have all demonstrated the continuity across the South Fork of the Walla Walla River of vertical major faults (including the Hite fault) and west-dipping dikes of Frenchman Springs basalts. It is, therefore, clear that this segment of the topographically defined Olympic-Wallowa lineament cannot be related to faulting in rocks of Miocene age.

In an earlier report (Washington Public Power Supply System, 1977a. p. 2RC-34) this writer stated that the Olympic-Wallowa lineament is "as originally defined a fictional structural element of the Pacific Northwest." I am still inclined to that view when the entire lineament postulated by Raisz is considered, but the existence of a disturbed plateau structural zone (including the Wallula fault zone) coincident with the central third of Raisz's lineament cannot be questioned. The nature and tectonic significance of this disturbed structural zone ("CLEW" of Laubscher, 1977) is discussed below.

8.1 ORIGIN OF COLUMBIA PLATEAU STRUCTURES

8.1.1 INTRODUCTION

Much has been written about the origin of the faulted fold structures of the Columbia Plateau with their enigmatic, more-or-less east-west orientations. The plateau folds have long attracted the attention of geologists because of their dramatic topographic rise above a generally subdued topography and because their trends are so atypical of the Cordillera as a whole.

The relationship of plateau fold (and fault) structures to deformation in underlying basement rocks has been a matter of particular contention among geologists since at least the writing of Laval (1956). He proposed that plateau folds could be caused by either 1) draping of basalt strata over reactivated basement faults ("thick-skinned" deformation), or 2) by folding of strata independent of the basement above a shallow-dipping surface of detachment ("thin-skinned" deformation).

Perhaps the most recent proponent of "thick-skinned" deformation is Bentley (1977, p. 343 and Figs. 14 and 20), who contended that the anticlinal folds of the plateau are localized above "basement weakness zones" (Figure 6). Such zones, according to Bentley (p. 343), have "localized the horizontal stresses and 'caused' the vertical uplift of the ridges at successive times. In gross character, the anticlines are 'drape' folds caused by vertical breakup of basement blocks..."1

1/ A similar model has recently been proposed by G. H. Davis (1978, not this author) for the widely spaced monoclines and folds of the Colorado Plateau. Davis concludes that the Colorado Plateau structures formed above steep, inactive basement fracture zones or faults that became reactivated in Laramide time when the plateau was subjected to regional compression. As with the Columbia Plateau folds, the impressive uplifts of the Colorado Plateau represent very little lateral shortening of the crust (1% or less according to G. H. Davis as compared with an analysis of Laubscher, 1977, of roughly 2% for the Columbia Plateau between the Columbia River and Kittitas Valley Washington; Bentley's more recent analysis, 1980, concludes approximately 7% shortening in western plateau areas and 2% in eastern).

8.1.2 STRUCTURAL ANALYSES OF LAUBSCHER

In the opinion of the writer, the most innovative analyses of plateau structure in the past several years have been those of Laubscher (1977, 1981). Laubscher (1981) relates internal deformation of the Columbia Plateau to its position within a broad belt of late Cenozoic regional dextral shear that separates the distending Basin and Range province from stable North America to the north. This inferred belt is several hundred kilometers wide and extends northwestward from Idaho to the Olympic Peninsula, hence the acronym IDOL given to it by Laubscher. IDOL is viewed as a complex mosaic of nearly two dozen tectonic blocks of variable extent. The most important from the standpoint of Columbia Plateau structure is the Yakima block, that IDOL component that contains all of the plateau folds south of a Cle Elum-Wallula lineament (CLEW) recognized earlier by Laubscher (1977).

A difficulty in testing the IDOL concept is the relatively small translational and rotational strains proposed by Laubscher to have occurred within IDOL. For example, he postulates only about 7 km of dextral shear across the entire 200 to 400 km-wide belt, and no more than 2 km of shear across CLEW itself (1981). Small strains have produced impressive structures in the Columbia Plateau because of the general horizontality of strata and flatness of topography there before latest Cenozoic deformation. Elsewhere, in geologically complex or topographically varied areas, comparable strains might produce structural features so diffuse that they would be difficult or impossible to recognize.

Laubscher's CLEW (1977), which forms a portion of the northern boundary of IDOL and is coincident with the central third of the Olympic-Wallowa lineament as defined by Raisz (1945), is a northwest-trending zone of structural disturbance that extends between Cle Elum to the northwest and Wallula Gap to the southeast (Figure 3). It is now recognized by most plateau workers as a fundamental structural component of the Washington portion of the Columbia Plateau. The zone is characterized by a "system of mostly faulted east-southeast-trending folds, or fold segments deflected in that direction where the long east to east-northeast-trending folds enter CLEW. This pattern is that of a right-lateral en echelon belt" (Laubscher, 1977, p. 11).

In a refinement of his 1977 draft report, Laubscher (1981) has concluded that CLEW developed prior to the plateau

folds, since they either terminate against it or are deflected or otherwise influenced by it. He interprets CLEW as the upper crustal consequence of deep-seated (15 to 20 km) right-lateral shear along a break in the lithosphere of the Columbia Plateau.

Folding within the Yakima Block is attributed by Laubscher to compressive stresses directed from the south as a result of block interactions with IDOL. These stresses are deemed responsible for a northward-migrating detachment (decollement) of the 20 km thick plateau crust from its serpentinized (?) peridotite mantle. The detachment is believed to root southward in or at the base of the thicker Blue Mountains crust. Broad fold structures spaced at roughly 25 to 30 km intervals lie above south-dipping reverse or thrust faults that progressively left the decollement at depth to ramp upwards (Figure 7). The periodicity of these folds is related by Laubscher to the initial periodic spacing of an echelon anticlinal folds (his "brachyanticlines") in CLEW, possible as the consequence of some kind of interference phenomenon between CLEW and the subcrustal detachment. "It is, therefore, proposed that as decollement instability at the base of the crust spread to the north, the pre-existing regularly-spaced brachyanticlines on CLEW, or rather the corresponding deep faults, were stress concentrators on the decollement surface from which thrusting spread sideways" (Laubscher, 1981). Laubscher believes that the sharply-defined ridges of the Columbia Plateau, which sit atop the much broader fold structures referred to above, are similar to kink bands and are due to "localized decollement at depth of 1-3 km, probably at the base of the Yakima sequence which is a mechanical discontinuity" (1981).

Several features do suggest regional northward translation of the Columbia Plateau and, hence, support some variety of detachment model for the origin of the compressional structures within it. Among these features are the generally asymmetric geometry of the plateau folds (steepest or over-turned flanks to the north -- with some prominent exceptions), and a southwestward deflection of the western ends of several major folds (Columbia Hills, Horse Heaven-Simcoe, Toppenish, Ahtanum?) as they enter the foothills of the Washington Cascades from the east. Laubscher defined the western edge of the Yakima block of the plateau on this basis, and he considered this rather diffuse boundary a zone of sinistral strain (HOOK, Laubscher, 1981). Similarly, the eastern ends of the Saddle Mountains and Umtanum-Gable Mountain structural trends may be deflected southeastward in a dextral sense. If so, and

this latter geometric relationship has not yet been adequately confirmed, then Laubscher's concept of the plateau folded region as a block or slab which has indented more northerly areas gains credence.

A major modification of Laubscher's crustal detachment model seems necessary to this writer, however, because of two important geometric aspects of the plateau folds: 1) north-south shortening across individual fold structures is small (0.5 to 2.5 km according to Bentley, 1980); and 2) the direction of fold vergence is southward in some structures (e.g. Cleman Mountain anticline) and alternatively both southward and northward along trend in others (Manastash, Umtanum, Gable Mountain, Toppenish?).

Both characteristics are compatible with local detachment of strata in the vicinities of the fold structures, much as Laubscher invisioned, but without the associated north-vergent, regional thrust faults which Laubscher postulates as extending to depth for 40 to 50 km across the entire crust (Figure 7). The alternating northward/southward vergence along trend of the Umtanum and Gable Mountain anticlines, with requisite tear faults separating the divergent fold structures is indicative of a history of localized folding, followed by later development of associated faults. Hypothetical stages in the generalized kinematic development of such doubly-vergent structures are presented in Figure 8. As illustrated, variable directions of overturning along the axial trace of the anticlinal fold -- possibly the consequence of pre-existing structural or stratigraphic anisotropies in the rocks being folded -- lead to the development of secondary cross-structure tear faults and, if deformation continues, to still younger flanking thrust (or reverse) faults. Price (1980), in a detailed structural analysis of the Umtanum Ridge anticline near Priest Rapids Dam, has concluded that concentric folding with a kink-like" geometry occurred during north-south compression (a geometry similar to that envisioned by Laubscher, 1981). "When shortening reached its maximum amount achievable by folding, it continued by overthrusting along thrust faults" (Price, 1980, p. 21). In other words, folding was not the consequence of the thrusts which now flank the structure.

Theoretical modeling of the plateau folds has yet to be undertaken, but to this writer the existence of a single, regional detachment surface beneath the Yakima fold province seems unlikely in the light of the small compressive strains (2 to 7%, Bentley, 1980), within the province. The distances between individual fold structures and the broad

uplifts they surmount are more likely, in my opinion, to be related to a buckling geometry for the plateau crust than to thrust fault risers ascending from a deep-seated plane of detachment at the base of the crust. Thrust-fault displacements of a kilometer or less for individual fold structures would, in all likelihood, be dissipated within the stratigraphic section well before the next fold-thrust structure some tens of kilometers distant is encountered.

Finally, Laubscher's hypothesis of northward-migrating subcrustal decollement as an explanation for plateau folding appears to be contradicted, as reviewed above, by the irregular development in both time and space of the plateau folds and their associated thrust faults. There is clearly no simple geographic progression of fold development discernible from the geologic record. Stratigraphic studies in the Pasco Basin area, for example, indicate that one of the most northerly fold structures, the Saddle Mountains anticline, began to form prior to some major folds now located to the south. In Laubscher's defense, much of the data bearing on the chronology of plateau fold development was unavailable to him at the time of his most recent writing.

This writer believes that one of Laubscher's principal contributions is his recognition that plateau structures fall into two major categories: (1) the prominent east-trending folds and their associated faults; and (2) CLEW, the zone of dextral strain that traverses the fold belt from northwest to southeast. The concept of CLEW as a structurally disturbed zone impressed upon the more regional plateau fold structures has its origins in Raisz's Olympic-Wallows lineament (1945), but Laubscher appears to have been the first to recognize the kinematic significance of the zone and its characteristics of dextral transcurrent strain.

The origin of CLEW and the timing of its development with respect to the plateau folds are critical aspects of any tectonic model for plateau deformation. In the following section this writer discusses these topics in a context somewhat different than that proposed by Laubscher (1977, 1981). Nevertheless, my indebtedness to his structural analyses will be obvious and is strongly acknowledged here.

8.1.3 GEOMETRY AND KINEMATICS OF CLEW

The plateau disturbed zone named CLEW by Laubscher (1977) can be conveniently divided from northwest to southeast into three structural domains (Figure 3), all of which contain

features characteristic of dextral transcurrent strain: (I) a broad zone of deflected or anomalous fold and fault trends (south from Cle Elum to Rattlesnake Mountain); (II) a narrower aligned belt of topographically expressed domes and doubly-plunging anticlines (Red Mountain to Wallula Gap) and, to the southwest, the northwest-trending eastern end of the Horse Heaven anticline; and (III) the Wallula fault zone (Wallula Gap to vicinity of Milton-Freewater, Oregon). As illustrated in Figure 3, CLEW narrows southeastward. East of Wallula Gap it is primarily represented by the transcurrent Wallula fault zone.

The Wallula fault zone was studied by Bingham and others (1970) in reconnaissance and by Shannon & Wilson (1979a) in greater detail. Strike-slip displacement within the zone is evidenced by the occurrence of subhorizontal striae at fault localities near Wallula Gap (loc. A, Shannon & Wilson, 1979a) and east of Warm Springs Canyon (loc. C, Shannon & Wilson, 1979a). Dextral (right-lateral) slip is suggested by the en echelon pattern of presumably normal faults directly south of the Wallula fault zone (Figure 5), by possible right-lateral stream deflections along the fault-line scarp east of Wallula Junction, and by the very slight en echelon alignment of the long axes of the domes and doubly-plunging anticlines west of Wallula Gap.

Field studies by Shannon & Wilson (1979b) and J. Kendall and R. Dale (in progress, University of Southern California) lead this writer to the conclusion that the Wallula fault terminates southeastward in the vicinity of Milton-Freewater. It is proposed here that dextral displacements along the Wallula fault can be geometrically explained as the consequence of differential crustal extension in the northern and southern walls of the zone. Horizontal (extensional) components of slip across normal faults south of the Wallula zone between Wallula Gap and Milton-Freewater would produce a northwestward displacement of the southern wall with respect to the northern. As illustrated in Figure 9, this displacement increases northwestward across each successive normal fault. A similar geometry of normal faulting is seen in the southwestern wall of the dextral Furnace Creek fault zone in Death Valley near its area of termination (Wright and Troxel, 1973).

Near Wallula Gap, the dextral strain which is largely concentrated along the Wallula fault to the east, becomes more diffuse. It does so in part by a westward splaying of the fault into two major branches. The southern branch (Wallula Gap fault of Jones and Deacon, 1966) dies out 4 to

6 km west-northwest of Yellepit according to Shannon & Wilson (1979a), although Foundation Sciences (1980) infers that it extends several tens of kilometers farther west. The northern, presumably main branch of the zone is inferred to underlie the next structural domain (II) of CLEW, the slightly en echelon doubly-plunging anticlines and domes that lie between Wallula Gap and Rattlesnake Mountain; this structural alignment has been called the Rattlesnake Hills-Wallula lineament by some previous workers. Here, dextral strain is apparently manifested not by a single throughgoing fault, i.e. the Wallula fault, but by the discontinuous fold and fault structures of the domain (see additional discussion of this point below) and by a southeastward deflection of the eastern end of the Horse Heaven anticline (Figure 3).

Northwest of Red Mountain and the Yakima River, the zone of CLEW (III) in which dextral strain effects can be observed becomes still broader and, presumably, still more diffuse. Laubscher contends on the basis of his geometric analyses (1977, 1981) that cumulative dextral strain across CLEW is less than 2 km, an estimate supported by Weston Geophysical's interpretation (1981) of the linear pattern of north-south-trending gravity anomalies along the western edge of the Pasco Basin. If right-slip within CLEW is geometrically related to crustal extension by limited normal faulting south of the Wallula fault, then estimates of total right-slip (or strain) of less than several kilometers appear reasonable.

The isolated fold structures of the Red Mountain-Wallula Gap domain of CLEW themselves attest to limited strike-slip displacement within the domain. Tchalenko (1970) and Wilcox and others (1973) have demonstrated from field and clay model studies that isolated fold structures such as those of the Red Mountain-Wallula Gap domain are the consequence of limited displacement across diffuse zones of transcurrent strain. As stated by the latter authors:

> "...<u>en echelon</u> folds in wrench zones form early. As the amount of displacement on the wrench zone increases, the initial folds are broken first by fractures and then by faults (p. 77). The development of the main, throughgoing wrench fault is the last state in the early phase of wrench-zone deformation (p. 82). After a short interval of concurrent folding and conjugate faulting, the rocks (or clay) fracture in a relatively narrow zone within the overall deformational swath, and the master wrench fault is created (p. 87)."

Because the bedrock-cored folds of the Red Mountain-Wallula Gap structural domain are separated by low-lying areas of Holocene deposits, the existence of a throughgoing wrench fault between them is difficult to establish. Bingham and others (1970) concluded that such linkage does exist in the form of a zone of shearing, the Rattlesnake-Wallula fault, which they interpreted as joining 15 widely spaced outcrops of breccia along the Rattlesnake Hills-Wallula structural trend. Continuity of the fault-breccia zone, according to Bingham and others (1970, p. 74) "can usually be inferred by the presence of subdued topographic features, such as straight scarps, saddles, and gully alinements (sic), along which the breccia is usually concealed by loess." They regarded the hypothesized fault zone as younger than the domal uplifts, but parallel to and almost congruent with the aligned fold structures.

Washington Public Power Supply System (1977b, p. 2R F-17) describes a Wallula Gap-Rattlesnake Hills topographic lineament as being "very gently curved" and appearing as a "northeast-facing break in slope." Washington Public Power Supply System (1977c, p. 2R H.5-8) supports the existence of a throughgoing topographic lineament ("the Rattlesnake Hills-Wallula lineament is formed by a slight en-echelon alignment of discontinuous, tightly-folded, plunging anticlines separated by interfold segments with a north homoclinal dip off of the Horse Heaven anticline"), but he concluded (p. 2R H. 509) that "there is no field evidence that this lineament is fault caused, either continuouly or en echelon."

Two other lines of data bear on the question of continuity of fault structure within the Red Mountain-Wallula Gap structural domain. J. Doherty of Weston Geophysical (personal communication, 1980) reports that several aeromagnetic profiles that transect the domain at high angles--across both anticlinal folds and the topographic lows between them--have anomalies at the Rattlesnake Hills-Wallula lineament that consistently model as a steep fault, north side down. However, if a throughgoing fault does exist, it does not appear to lie within the bedrock exposures of the various anticlinal folds. For example, faults exposed in the Butte, south of Finley, and in the next two hills to the northwest are not parallel and have slickenside striae that set each apart from the others in a kinematic sense (Table 1).

The question of fault continuity between the isolated anticlines and domes of CLEW's Red Mountain-Wallula Gap domain has not yet been satisfactorily resolved. A strong

case, based on geometric analogies with wrench fault zones elsewhere, can be made for the existence of a throughgoing transcurrent zone of movement at some depth beneath the fold alignment, but the upward extent of that zone has yet to be defined. Accordingly, additional geological and geophysical studies are planned along the structural trend in early 1981.

The tectonic model proposed here for the geometric and kinematic interrelationships between the three domains of CLEW draws support from geometrically analogous relations along the Newport-Inglewood fault zone in the western Los Angeles Basin. Recent studies of that zone (Harding, 1973; Yeats, 1973; Barrows, 1974) describe it as a wrench fault-controlled zone, consisting of a series of rather evenly spaced and complexly faulted anticlines with an en echelon pattern that requires dextral wrenching. The folds are topographically expresed (Figure 10) along a northwest-southeast trend approximately 65 km in length. Individual folds are cut obliquely by synthetic, right-lateral strike slip faults. Offsets increase southeastward where the disturbed zone is characterized by longer, more throughgoing faults (Figure 10), but total dextral displacement for the zone is no more than 3 km (Yeats, 1973). Despite geometric similarities between the two zones, they differ pronouncedly in present seismicity and regional tectonic setting. No historic earthquakes have conclusively been tied to the Rattlesnake Hills-Wallula structural alignment (see concluding section of this report), but the Newport-Inglewood zone has experiences scores of significant events since 1920 (Barrows, 1974), including earthquakes of magnitude 6.3 (March, 1933, with 78 aftershocks between 3.9 and 5.2), 5.4 (October, 1933),4.9 (1941), "near" 4.9 (1920),and 4.5 to 4.6 (1939, 1944, 1961, 1969). Unlike CLEW, which lies within an intraplate region of low strain, the Newport-Inglewood zone lies within the San Andreas fault system, the active boundary between the Pacific and North American plates in California. Furthermore, geologic studies summarized elsewhere indicate that most of the development of southern portions of CLEW occurred prior to Pleistocene time, unlike the pronounced contemporary strain along the Newport-Inglewood zone.

8.1.4 ORIGIN OF CLEW

Laubscher (1981) proposes that CLEW initially developed across the Columbia Plateau lithosphere as a zone of "broad and gentle dextral <u>en echelon</u> brachyanticlines" above a deeper wrench fault of very small displacement (<u>much</u> less than 2 km). The early brachyanticlines of CLEW, or the deep fault(s) inferred to lie elow them, are viewed by Laubscher

as acting as stress concentrators and nucleating in some way the Yakima-type plateau folds during slightly younger crustal decollement (detachment).

This writer's view differs somewhat from Laubscher's in terms of the timing of the development of CLEW. The components of CLEW are viewed here as development synchronously with plateau fold and fault structures of more easterly trend. According to this view, north-south shortening of the Columbia Plateau has produced the two major plateau structural elements-- (1) the east-trending fold-fault structures of higher plateau strata by buckling and local detachment ("thin-skinned") mechanisms, and (2) CLEW, by dextral movement along a deeper fault or zone of anisotropy, with the synchronous distortion of developing fold structures at stratigraphic and structural levels directly above the deeper structure (Figure 11). Changes in trend of folds (e.g. Horse Heaven Hills anticline) as they enter CLEW are not interpreted as rotations (drag) of older structures, but rather as expressions of the complex local stress field within CLEW that resulted from interaction of the two levels of differing deformational behavior. The zone of structural interference is very broad in northern parts of CLEW, but narrows progressively southeastward to become the Wallula fault zone east of Wallula Gap. It is not clear whether the end of the Wallula fault near Milton-Freewater coincides with the terminus of the deeper strike-slip structure, or merely represents the south-eastward limit of Miocene and younger reactivation along a pre-existing structure.

8.1.5 NATURE OF THE BASEMENT CONTROL OF CLEW

An important question is whether the location of CLEW was predetermined by a pre-exiting fault or structural flaw in the crust of the plateau, and reactivated in Miocene time, or whether it is a pristine zone formed during late Miocene deformation. Laubscher (1981) adopts the latter view, pointing out that neither regional gravity patterns nor isopachs of crustal thickness change across CLEW (in this writer's opinion, his isopach observation is inadequately controlled by data). Alternatively, Shannon & Wilson (1978, p. 23), in an evaluation of CLEW based on Laubscher's 1977 manuscript, favored pre-existing controls on the location of CLEW and the nature of deformation within it: "In our opinion, most individual structures in the CLEW are consistent with a tectonic model of local compression that occurred along trends controlled by regional zones of inherited weakness in a stress system oriented north-south or north-northeast south-southwest." Shannon & Wilson

(1978) did not, however, accept the regional wrench tectonics model proposed by Laubscher for CLEW and endorsed here.

Evidence for the existence along CLEW of an older fault below the Miocene plateau basalts comes from northwestern areas in the Washington Cascades. Both Tabor and Frizzell (1979) and Vance and Miller (1981) report that early Eocene structures of the Straight Creek fault zone turn southeastward, south of Kachess Lake and the Yakima River, into a position coincident with the Olympic-Wallowa lineament (and, therefore, CLEW). According to Vance and Miller the Goat Peak segment of the north-south Straight Creek fault becomes the northwest-striking Taneum Lake fault 15 km south of the Yakima River (Figure 3). They view the Taneum Lake fault as a late dip-slip splay of the Straight Creek fault, -- not the main transcurrent structure. Because the Taneum Lake reverse fault cuts only pre-Miocene units (Bentley, 1977), its age relationship to Miocene plateau basalts is unclear. However, Tabor and Frizzell (1977) conclude that Miocene or younger movement along either the southern Straight Creek fault or the Olympic-Wallowa lineament "must be minimal or absent because the Miocene Snoqualmie batholith and its satellite stocks cut faults in the Straight Creek zone and are unmarked by structures paralleling and on strike with the lineament." The Snoqualmie batholith, dated at 17-18 m.y., is essentially coeval with the earliest Columbia River basalts. Other intrusive and stratigraphic relationships discussed by Vance and Miller (1981) and Tabor and Frizzell (1977) support the conclusion that both strike-slip and dip-slip displacements along southern parts of the Straight Creek fault zone had terminated by mid-Oligocene time, ca. 33 m.y. ago.

Hammond (1977) has described a broad northwest-striking zone of faulting, coincident with OWL, that is exposed along the South Fork of the Snoqualmie River on the west flank of the Cascade Range. This fault zone is presumably intruded by the Snoqualmie batholith (Tabor and Frizzell, 1979), but it too projects towards CLEW. It's geometric and kinematic relationships to the Taneum Lake fault are not known. Collectively, the fault relations described above strongly imply that faulting parallel to CLEW and of an age older than the Columbia River basalts was present in at least northern portions of what was to become CLEW in late Miocene time. Whether such faulting contributed to CLEW's development is not known.

The interpretation of CLEW as a diffuse zone of wrench-fault deformation presupposes the existence at depth of a narrower zone of horizontal strain. In Laubscher's words (1981) CLEW is the consequence of "deep-seated right-lateral shear that drives a deformable sequence of partially decoupled more superficial layers." Drawing upon the studies of Emmons (1969) and Harding and Lowell (1979), Laubscher concludes that the depth to the driving structure is about half the width of the deformed belt that develops above it. Accordingly, he estimates that a throughgoing zone of shear would lie at a 15-20 km depth below what has been called domain I of CLEW in this paper (Figure 3), but less than 2 km below the narrower brachyanticlinal domain II. The zone presumably extends to surface levels in domain III (the Wallula fault zone). Another way of stating relations as perceived by Laubscher is that the depth to the controlling transcurrent (wrench) structure responsible for CLEW increases northwestward from domain III to I. If Laubscher's analysis is correct, the driving structure beneath domain I (Rattlesnake Mountain to Cle Elum), which makes up most of CLEW, lies within the lower plateau crust and/or upper mantle. An alternative explanation for the greater width of CLEW in domain I is not that the controlling wrench structure is deeper there, but that the zone of "basement" wrenching simply becomes broader and more diffuse to the northwest.

8.1.6 AGE OF CLEW AND THE QUESTION OF ITS CAPABILITY

The Wallula fault zone within domain III of CLEW can be inferred to extend from the vicinity of Wallula Gap to near Milton-Freewater. Although the topographically-defined trace of the fault zone becomes indistinct several kilometers east of Warm Springs Canyon, the continued presence of northwest-striking normal (?) faults in the hills to the south as far east as Milton-Freewater argues for an extension of the Wallula zone into that region (Figure 3). Shannon & Wilson (1979a) has concluded that the 45-50 km-long Wallula fault zone thus defined is a "capable fault" from the standpoint of nuclear power plant siting criteria.

In light of preliminary findings from the Woodward-Clyde trenches across "Bingham's linear" that Touchet beds are not offset along this part of the Wallula zone, the evidence for capability of the Wallula structure is diminished. The age of colluvial deposits cut by faulting, both at Finley Quarry and west of Warm Springs, is not known. Fanglomerates dated as 50,000 years old (Woodward-Clyde Consultants, 1978, as referenced in Shannon & Wilson, 1979a) lie unbroken across

the Wallula Gap strand of the fault zone near Yellepit. This relation casts some doubt on a Touchet age (ca. 13,000 y.b.p.) for the sheared clastic dike described by Shannon & Wilson (1979) in the Wallula Gap fault zone 3 km east of Yellepit. Unsheared clastic dikes of probable Touchet origin lie within and across fault zones exposed at Finley Quarry and in quarries on "K" and "L" hills to the northwest (all three localities in CLEW domain II). Touchet beds and associated clastic dikes are definitely offset along north-dipping surfaces in a road cut south of Umapine near Milton-Freewater, but the convex-upwards geometry of the surfaces may indicate an origin for them by gravity slumping rather than by tectonic activity. Even if a tectonic origin for these surfaces is assumed, their normal fault geometry is not obviously compatible with their being part of the strike-slip Wallula zone.

In short, although capability of the Wallula fault zone cannot at present be refuted, its unequivocal designation as "capable" (Shannon & Wilson, 1979a) now appears unwarranted in light of recent trenching along "Bingham's linear." In this regard, the 1979 College Place earthquake (M = 4.1) which occurred along the Oregon-Washington state line just north of Umapine is of interest. Slemmons and O'Malley (1980, p. 34) state that historic tectonic activity on CLEWW (a southeast-extended version of CLEW) is "recorded by the College Place earthquake." That conclusion is suspect. Although near the southeast-projected trace of the Wallula fault zone in the Milton-Freewater area, the focal mechanism solution for this significant plateau seismic event is grossly incompatible with right slip along a fault parallel to the Wallula zone (P in the solution is east-west in orientation; and the N 40° W - striking fault plane of the solution is characterized by <u>left</u> slip; Woodward-Clyde Consultants, 1980a). The earthquake may indicate that the north-south compressive stress field responsible for the development of CLEW is no longer extant in the Milton-Freewater area, having been replaced by east-west shortening. The "Milton-Freewater earthquake of 1936 (M = 6.1) has an instrumentally located epicenter near Waitsburg, Washington, approximately 30 km north-northeast of Milton-Freewater." This conclusion is based on a recent evaluation of historic seismographic records by Woodward-Clyde Consultants (1980b) and is in agreement with the original epicentral determination reported shortly after the event. Reports of an epicenter near Milton-Freewater were based on "felt" reports and intensity data. Thus, this earthquake also did <u>not</u> occur within the Wallula fault zone. It's focal mechanism solutions, although poorly constrained, are not compatible with dextral slip on a fault plane parallel to the Wallula zone.

Evidence is good that the Wallula fault zone terminates southeastward in the vicinity of Milton-Freewater. Its northwestern extent is more problematical. Laubscher (1977, 1981), Shannon & Wilson (1979a), and this writer have suggested that the isolated anticlines of CLEW domain II, which extend like beads in a string northwestward from Wallula Gap, lie above a continuation at depth of the Wallula zone. Although the existence of a throughgoing surficial fault hidden beneath Quaternary deposits and lying directly north of the anticlines cannot be discounted, the geometry of the slightly en echelon folds supports the premise that they lie above a shallow buried zone of limited lateral displacement (cf. Tchalenki, 1970, and Wilcox and others, 1973). As proposed by Laubscher (1981) the greatly increased width of CLEW domain I to the northwest argues for a much deeper position of the controlling wrench fault structure there, or (as proposed here) an increasingly broad zone of transcurrent shear at depth, or both. It is the diffuseness of strain within CLEW domains I and II, coupled with evidence that total dextral strain across CLEW is less than 2 or 3 km, that complicates the question of capability of this zone of wrench tectonics. It is important to point out, however, that no Quaternary fault displacements have been documented northwest of Finley Quarry. If, as discussed above, the structures of CLEW, particularly those of domains I and II, represent the interaction of developing thin-skinned plateau folds above an active deeper wrench system, then the age of the folds gives us the age for the formation of CLEW. For reasons discussed at length elsewhere in this report, most plateau folding appears to be of Pliocene and older age (Washington Public Power Supply System, 1977d; Bentley and others, 1980; Rockwell, 1979) although the possibility of major folding in the Yakima area until one or one and one-half m.y. ago cannot be discounted (Bentley, 1977; Rigby and Othberg, 1979).

Tectonic models for CLEW which link it to synchronous plateau folding, thus lead to the conclusion that strain along CLEW has waned significantly within the past 1 1/2 to 3 or 4 m.y. Unlike the San Andreas fault, a dextral plate boundary with sharp definition and large fault displacements (ca. 300 km) in the past 4 to 6 m.y., CLEW is an intraplate zone of small dextral strain (less than a few kilometers) that is characterized by its diffusiveness and its waning nature in the past several million years. These characteristics must be considered when fault-related design criteria are approved for the Hanford plants.

9.1 REFERENCES

Anderson, J. L., 1980, Deformation and Canyon Cutting in Post-Grande Ronde, pre-Frenchman Springs Time, Grayback Mountain, South-central Washington: Geological Society of America, Abstracts with Programs, V. 12, No. 3, p. 93.

Ando, M., and Balazs, E. I., 1979, Geodetic Evidence for Aseismic Subduction of the Juan de Fuca Plate: Journal of Geophysical Research, V. 84, No. B6, p. 3023-3028.

Armstrong, R. L., 1979, Cenozoic Igneous History of the United States Cordillera from Latitude 42° to 49° N, in Smith, R. B., and Eaton, G. P., Eds., Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, Geological Society of America Memoir 152, p. 265-282.

Atwater, Tanya, 1970, Implications of Plate Tectonics for the Cenozoic Tectonic Evolution of Western North America: Geological Society of America Bulletin, V. 81, No. 12, p. 3513-3536.

Barrows, A. G., 1974, A Review of the Geology and Earthquake History of the Newport-Inglewood Structural Zone, Southern California: California Division of Mines and Geology Special Report 114, 115 p.

Beeson, M. H. and Moran, M. R., 1979, Stratigraphy and Structure of the Columbia River Basalt Group in the Cascade Range, Oregon: U.S. Department of Energy Report RLO-1040-T1, p. 5-77.

Bentley, R. D., 1977, Stratigraphy of the Yakima Basalts and Structural Evolution of the Yakima Basalts and Structural Evolution of the Yakima Ridges in the Western Columbia Plateau, in Geological Excursions in the Pacific Northwest, Department of Geology, Western Washington State University, p. 339-389.

Bentley, R. D., 1980, Magnitude of Neogene Horizontal Shortening in the Western Columbia Plateau, Washington-Oregon: American Geophysical Union Transactions, V. 62, No. 6, p. 60.

Bentley, R. D., Anderson, J. L., Campbell, N. P., and Swanson, D. A., 1980, Stratigraphy and Structure of the Yakima Indian Reservation, with Emphasis on the Columbia River Basalt Group: United States Geological Survey Open-File Report 80-200, 83 p.

Bingham, J. W., Londquist, E. T., and Baltz, E. H., 1970, Geologic Investigation of Faulting in the Hanford Region, Washington (with a Section on the Occurrence of Microearthquakes, by A. M. Pitts); United States Geological Survey Open-File Report 70-27, 104 p.

Campbell, N. and Bentley, R. 1980, Late Quaternary Faulting, Toppenish Ridge-Southcentral Washington, Geological Society of America, Abstracts with Programs, V. 12, No. 3, p. 101.

Carey, S. W., 1958, The Tectonic Approach to Continental Drifting, in Continental Drift, a Symposium: Geology Department, University of Tasmania, Hobart, p. 177-355.

Carlson, R. L., 1976, Cenozoic Plate Convergence in the Vicinity of the Pacific Northwest: A Synthesis and Assessment of Plate Tectonics in the Northeastern Pacific: University of of Washington, Ph.D. dissertation., 129 p.

Christiansen, R. L., and McKee, E. H., 1979, Late Cenozoic Volcanic and Tectonic Evolution of the Great Basin and Columbia Intermontane Regions: in Smith, R. B., and Eaton, G. P., Eds. in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, Geological Society of America Memoir p. 283-322.

Coney, P. J., 1979, Mesozoic-Cenozoic Cordilleran Plate Tectonics, in Smith, R. B., and Eaton, G. P., Eds., Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, Geological Society of America Memoir 152, p. 33-50.

Crosson, R., 1980, Seismicity and Tectonics of the Puget Sound Region: Results from the Regional Seismograph Network: Seismological Society of America Annual Meeting, Seattle, Washington, Earthquake Notes, V. 50, No. 4, p. 58-59.

Davis, G. A., 1980, Problems of Intraplate Extensional Tectonics, Western United States, in Continental Tectonics: National Academy of Sciences, Washington, D.C., p. 84-95.

Davis, G. A., 1978, Monocline Fold Pattern of the Colorado Plateau, in Vincent Matthews, III, Ed., Laramide Folding Associated with Basement Block Faulting in the Western United States, Geological Society of America Memoir 151, p. 215-234.

Eardley, A. J., 1962, Structural Geology of North America (2nd ed.): Harper and Row Publishers., New York, N. Y., 743 p.

Eaton, G. P., 1979, A Plate-Tectonic Model for Late Cenozoic Crustal Spreading in the Western United States, in Riecker, R. E. Ed., Rio Grande Rift: Tectonics and Magmatism, American Geophysical Union, Washington, D. C., p. 7-32.

Emmons, R. C. 1969, Strike-Slip Rupture Patterns in Sand Models: Tectonophysics, V. 7, p. 71-87.

Foundation Sciences, Inc., 1980, Geologic Reconnaissance of Parts of the Walla Walla and Pullman, Washington, and Pendleton, Oregon 1° x 2° AMS Quadrangles: Prepared by Kienle, C. F., 76 p.

Goff, F. E., and Myers, C. W., 1978, Structural Evolution of East Umtanum and Yakima Ridges, South-Central Washington: Geological Society of America, Abstracts with Programs, V. 10, No. 7, p. 408

Green, R. C., Walker, G. W., and Corcoran, R. E., 1972, Geologic Map of the Burns Quadrangle, Oregon: United States Geological Survey Miscellaneous Investigation Map I-680, 1:250,000.

Hamilton, W., and Myers, W. B., 1966, Cenozoic Tectonics of the Western United States: Review of Geophysics and Space Physics, V. 5, p. 509-549.

Hammond, P. E., 1977, Stratigraphy, Structure, and Plutonism along the Upper Yakima and South Fork Snoqualmie River Valleys, Central Cascade Range, Washington, in Geological Excursions in the Pacific Northwest: Department of Geology, Western Washington State University, p. 292-308.

Harding, T. P., 1973, Newport-Inglewood Trend, California - An Example of Wrenching Style of Deformation: American Association of Petroleum Geologists Bulletin, V. 57, No., 1 p. 97-116.

Harding, T. P., and Cowell, J. D., 1979, Structural Styles, Their Plate-Tectonic Habitats, and Hydrocarbon Traps in Petroleum Provinces: American Association of Petroleum Geologists Bulletin, V. 63, No. 1, p.1016-1058.

Harding, T. P. and Lowell, J. D., 1979, Structural Styles, their Plate-Tectonic Habitats, and Hydrocarbon Traps in Petroleum Provinces, American Association of Petroleum Geologists Bulletin, V. 63, No. 7, p. 1016-1058.

Hill, D. P., 1972, Crustal and Upper Mantle Structure of the Columbia Plateau from Long Range Seismic-Refraction Measurements: Geological Society of America Bulletin, V. 83, No. 6, p. 1639-1648.

Hill, D. P., 1980, Seismic Evidence for the Structure and Cenozoic Tectonics of the Pacific Coast States, in Smith, R. B., and Eaton, G. P., Eds., Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, Geological Society of America Memoir 152, p. 145-174.

Howard, K. A., et. al., 1978, Preliminary Map of Young Faults in the United States as a Guide to Possible Fault Activity: United States Geological Survey Miscellaneous Field Studies Map MF-916, 1:5,000,000.

Jones, F. O., and Deacon, R. J., 1966, Geology and Tectonic History of the Hanford Area and its Relation to the Geology and Tectonic History of the State of Washington and the Active Seismic Zones of Western Washington and Western Montana: Report prepared for Douglas United Nuclear, Inc., DUN-1410, Richland, Washington.

Kendall, J. J., Dale, R. C., and Davis, G. A., 1981, The Structural Relationship of the Olympic-Wallula Lineament, Hite Fault System and LaGrande Fault System, the Blue Mountains of Umatilla County Oregon: Geological Society of America Abstracts with Programs V. 13 No. 2, p. 64.

Laubscher, H. P., 1977, Structural Analysis of Post-Yakima Deformation, Columbia Plateau, Washington: Draft report prepared for United Engineers and Constructors, Inc.

Laubscher, H. P., 1981, Models of the Development of Yakima Deformation: Prepared for United Engineers and Constructors, Inc.

Laval, W. N., 1956, Stratigraphy and Structural Geology of Portions of South-Central Washington: University of Washington, Ph.D. dissertation.

Lawrence, R. D., 1976, Strike-slip Faulting Terminates the Basin and Range Province in Oregon: Geological Society of America Bulletin V. 87, No. 6, p. 846-850.

Loring, A. K., 1976, Distribution in Time and Space of Late Phanerozoic Normal Faulting in Nevada and Utah: Utah Geology, V. 3, No. 2, p. 97-109.

Magill, J., and Cox, A., 1981, Post-Oligocene Tectonic Rotation of the Oregon Western Cascade Range and the Klamath Mountains: Geology, V. 9, No. 3, p. 127-131.

McKee, B., 1972, Cascadia, the Geologic Evolution of the Pacific Northwest: McGraw-Hill Book Co., New York, N. Y., 394 p.

McKenzie, D., and Julian, B., 1971, Puget Sound, Washington, Earthquake and the Mantle Structure beneath the Northwestern United States: Geological Society of America Bulletin V. 82, No. 12, p. 3519-3524.

Nathan, S. and Fruchter, J. S., 1974, Geochemical and Paleomagnetic Stratigraphy of the Picture Gorge and Yakima Basalts (Columbia River Group) in Central Oregon: Geological Society of America, Bulletin V. 85, No. 1, p. 63-76.

Price, E. H., 1980, Strain Distribution and Model for Formation of Eastern Umtanum Ridge Anticline, South-Central Washington: Rockwell Hanford Operations, report prepared for U. S. Department of Energy, RHO-BWI-SA-30, 27 p.

Raisz, E., 1945, The Olympic-Wallowa Lineament: American Journal of Science, V. 243-A, p. 479-485.

Reidel, S. P., et. al., 1980, Rate of Deformation in the Pasco Basin during the Miocene as Determined by Distribution of Columbia River Basalt Flows: Rockwell Hanford Operations, report prepared by U. S. Department of Energy, RHO-BWI-SA-29, 43 p.

Riddihough, R. P., 1979, Gravity and Structure of an Active Margin -- British Columbia and Washington: Canadian Journal of Earth Sciences, V. 16, No. 2, P. 350-363.

Riddihough, R. P., and Hyndman, R. D., 1976, Canada's Active Western Margin--The Case for Subduction: Geoscience Canada, V. 3, No. 4, p. 269-278.

Rigby, J. G., and Othberg, K., 1979, Reconnaissance Surficial Geologic Mapping of the Late Cenozoic Sediments of the Columbia Basin, Washington: State of Washington, Department of Natural Resources, Open File Report 79-3.

Robyn, T. L., 1977, Geology and Petrology of the Strawberry Volcanics, NE Oregon: University of of Oregon, Ph.D. dissertation, 197 p.

Robyn, T. L., Hoover, J. D., and Thayer, T. P., 1977, Geology and Geochronology of the Strawberry Volcanics, NE Oregon: Geological Society of America Abstracts with Programs, V. 9, No. 4. p. 488-489.

Rockwell, 1979, Geologic Studies of the Columbia Plateau, a Status Report: Rockwell Hanford Operations, Richland, Washington, Report RHO-BWI-ST-4.

Rogers, G. C., 1979, Earthquake Fault Plane Solutions near Vancouver Island: Canadian Journal of Earth Sciences, V. 16, No. 3, p. 523-531.

Shannon & Wilson, 1973, Geologic Studies of Columbia River Basalt Structures and Age of Deformation, the Dalles-Umatilla Region, Washington and Oregon: Boardman Nuclear Project: Report to General Electric Co., 52 p.

Shannon & Wilson, Inc., 1978, Geologic Reconnaissance of the Cle-Elum-Wallula Lineament and Related Structures: A report prepared for Washington Public Power Supply System, 33 p.

Shannon & Wilson, Inc., 1979a, Evaluation of Faulting in the Warm Springs Canyon Area, Southeast Washington: Report prepared for Washington Public Power Supply System, 18 p.

Shannon & Wilson, Inc., 1979b, Geologic Reconnaissance of the Wallula Gap, Washington - Blue Mountains - LaGrande, Oregon Region: Report prepared for Washington Public Power Supply System, 63 p.

Shannon & Wilson, Inc., 1980, Geologic Evaluation of Selected Faults and Lineaments, Pasco and Walla Walla Basins-Washington: Report prepared for Washington Public Power Supply System, 25 p.

Skehan, J. W., 1965, A Continental-Oceanic Crustal Boundary in the Pacific Northwest: United States Air Force Cambridge Research Laboratory Report No. AFCRL 65-904, 56 p.

Slemmons, D. B., and O'Malley, P., 1980, Fault and Earthquake Hazard Evaluation of Five United States Corps of Engineers Dams in Southeastern Washington: Prepared for United States Corps of Engineers, 60 p.

Smith, R. B., 1977, Intraplate Tectonics of the Western North American Plate: Tectonophysics, V. 37, p. 323-336.

Smith, R. B., 1979, Seismicity, Crustal Structure, and Intraplate Tectonics of the Interior of the Western Cordillera, in Smith, R. B. and Eaton, G. P., EDS., Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, Geological Society of America Memoir 152, p. 111-144.

Snyder, W. S., Dickinson, W. R., and Silberman, M. L., 1976, Tectonic Implications of Space-time Patterns of Cenozoic Magnetism in the Western United States: Earth and Planetary Science Letters, V. 32, p. 91-106.

Swanson, D. A. and Wright, T. L., 1978, Bedrock Geology of the Northern Columbia Plateau and Adjacent Areas: In Baker, V. R. and Nummedal, D., eds., The Channeled Scabland, Washington NASA Planetary Geology Program, A Guide to the Geomorphology of the Columbia Basin, Washington, p. 37-57.

Swanson, D. A., et. al., 1979, Preliminary Reconnaissance Geologic Maps of the Columbia River Basalt Group in Parts of Eastern Washington and Northern Idaho: United States Geological Survey Open-File Report 79-534, 1:250,000.

System, Science, and Software, Determination of Three-Dimensional Structures of Eastern Washington from the Joint Inversion of Gravity and Earthquake Travel Time Data: report prepared for Weston Geophysical Corporation by Rodi, W. L. et al., 143 p.

Tabor, R. W., and Cady, W. M., 1978, Geologic Map of the Olympic Peninsula: United States Geological Survey Miscellaneous Investigation. Map I-994, 1:125,000.

Tabor, R. W., and Frizell, V. A., 1979, Tertiary Movement Along the Southern Segment of the Straight Creek Fault and its Relation to the Olympic Wallowa Lineament in the Central Cascades, Washington: Geological Society of America, Abstracts with Programs, V 11, No. 3, p. 131.

Taubeneck, W. H., 1970, Dikes of Columbia River Basalt in Northeastern Oregon, Western Idaho and Southeastern Washington: in Proceedings of the Second Columbia River Basalt Symposium, Eastern Washington State College Press, Cheney, Washington, p. 73-96.

Thayer, T. P., and Brown, C. E., 1966, Geologic Map of the Aldrich Mountain Quadrangle, Grant County, Oregon: United States Geological Survey Map GQ-438, 1:62,500.

Vance, J. A., and Miller, R. B., 1981, The Movement History of the Straight Creek Fault in Washington State: Unpublished manuscript, 5 p.

Walker, G. W., 1977, Geologic Map of Oregon East of the 121stMeridian: United States Geological Survey Miscellaneous Investigation Map I-902, 1:500,000.

Washington Public Power Supply System, 1977a, Tectonic Evolution of the Pacific Northwest - Precambrian to Present: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2A, Subappendix 2R C.

Washington Public Power Supply System, 1977b, Imagery and Topographic Interpretation of Geologic Structures in Central Washington: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2B, Subappendix 2R F.

Washington Public Power Supply System, 1977c, Geologic Studies - Wallula Gap to Badger Coulee: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2B, Subappendix 2R H, Chapter 5.0.

Washington Public Power Supply System, 1977d, Geologic Studies of the Saddle Mountains - Manastash Ridge and Umtanum Ridge Structures: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2B, Subappendix 2R H, Chapter 8.0.

Washington Public Power Supply System, 1977e, Evaluation of Microearthquake Activity in Eastern Washington: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2B, Subappendix 2R J.

Washington Public Power Supply System, 1977f, Microearthquake Survey and Evaluation of Stress Orientation in Central Washington: WNP 1/4 Preliminary Safety Analysis Report, Amendment 23, Vol. 2B, Subappendix 2R L.

Weston Geophysical Research, Inc. 80, personal communication.

Weston Geophysical Research, Inc. 1981, Compilation and Interpretation of Gravity in Washington, Oregon, and Adjacent Parts of British Columbia and Idaho: Prepared for Washington Public Power Supply System, WNP-2 FSAR, Appendix 2.5L, 28 p.

Wilcox, R. E., Harding, T. P., and Seely, D. R., 1973, Basic Wrench Tectonics: American Association of Petroleum Geologists Bulletin V. 57, No. 1, p. 74-96.

Wise, D. U., 1963, An Outrageous Hypothesis for the Tectonic Pattern of the North American Cordillera: Geological Society of America Bulletin V. 74, No. 3, p. 357-362.

Woodward-Clyde Consultants, 1978, Paleomagnetic Measurements of the Ringold Formation and Loess Units near Hanford, Washington and Evaluation of Age Dating Potential of Quaternary Deposits near Hanford, Washington: Report prepared for Washington Public Power Supply System under the Direction of United Engineers and Constructors Inc., Philadelphia, Pennsylvania.

Woodward-Clyde Consultants, 1980a, Recent Seismicity of the Hanford Region (Draft Report, January): Prepared for United Engineers and Constructors, 28 p.

Woodward-Clyde Consultants, 1980b, Seismological Review of the July 16, 1936 Milton-Freewater Earthquake Source Region: Prepared for Washington Public Power Supply System, 44 p.

Wright, L. A., and Troxel, B. W., 1973, Shallow-fault Interpretation of Basin and Range Structure, Southwestern Great Basin, in DeJong, K. A., and Scholter, Robert, Eds., Gravity and Tectonics, John Wiley & Sons, New York, N. Y., p. 397-408.

Yeats, R. S., 1973, Newport-Inglewood Fault Zone, Los Angeles Basin, California: American Association of Petroleum Geologists Bulletin V. 57, No. 1, p. 117-135.

Zoback, M. L., and Thompson, G. A., 1978, Basin and Range Rifting in Northern Nevada: Clues from a Mid-Miocene Rift and its Subsequent Offsets: Geology, V. 6, No. 2, p. 111-116.