SECTION 2.5.1

AMENDMENT TO WNP-2 FSAR

Prepared by

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CHAPTER 2

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2.5 GEOLOGY, SEISMOLOGY, AND GEOTECHNICAL ENGINEERING

The site is located in the Pasco Basin, a physiographic depression of the Columbia Plateau province in southeastern Washington state. The adjacent provinces are the Cascade Mountains on the west, the Blue Mountains on the south, the Northern Rocky Mountain and the Idaho Batholith on the east, and the Okanogan Highland on the north. Underlying the site of the WNP 1-2-4 plant facilities is a thin cover of eolian deposits; an average of 45 feet of Quaternary glaciofluvial sands; and approximately 480 feet of Pliocene Ringold sediments. The upper Ringold sediments consist of 205 feet of predominantly gravel with some interbedded layers and lenses of sand and silt. The lower Ringold sediments consist of 275 feet of interbedded claystone, siltstone and conglomerate. These sediments in turn overlie several thousand feet of Miocene-Pliocene basalt flows and interbeds. All stratigraphic units beneath the site appear to be near horizontal.

The documented geologic history of the Pasco Basin begins in Miocene time with the extrusions of the Columbia River Basalts. Older rocks are not known to crop out in the Pasco Basin area. A 10,000-foot hole drilled on Rattlesnake Mountain encountered Oligocene (?) sediments and Eocene basalts below the Columbia River basalts. After cessation of the basalt extrusions, Ringold sediments were deposited during Pliocene time. The principle development of the anticlinal ridges and associated faulting appears to have preceded the Ringold deposition. In late Pleistocene time the Pasco Basin was inundated by catastrophic glacial flooding. These events resulted in deposition of alluvial deposits that cover most of the Pasco Basin. During Holocene, wind-blown silt and sand deposits have mantled the landscape.

Within the Columbia Plateau and the Pasco Basin, the significant geologic features of the site and surrounding area are west and northwest-trending anticlinal folds. The geologic evidence supports the conclusion that folding is the predominant mode of deformation on the Columbia Plateau and that known faults with trends parallel to fold structures are secondary. Such faults have been identified on Saddle Mountains, Umtanum Ridge, Gable Mountain, and Rattlesnake Mountain.

A northwest-trending zone of structure within the Columbia Plateau between Cle Elum and Wallula Gap is known as CLEW. The Rattlesnake-Wallula alignment lies within the southeast portion of CLEW. CLEW is interpreted as a diffuse zone of

dextral strain, evidenced by the northwest-trending structures, that appears to have developed sychronously with the predominant east-west trending structures. The northwestern portion of CLEW is ill defined by the fold structures. In the southeastern portion, CLEW becomes more distinctly defined by the Wallula fault zone.

The more southerly Columbia Plateau fold structurs (e.g., Toppenish Ridge, Horse Heaven Hills, Columbia Hills) developed before extrusion of the Plio-Pleistocene Simcoe volcanics, that lie essentially undisturbed across the western traces of these structures. This relationship indicates a waning of southern plateau deformation, including CLEW, since the Pliocene. Surface ruptures of probable Quaternary age have been mapped on Toppenish Ridge. Both landslide and tectonic origins for the structure appear to be viable and are under investigation.

The Rattlesnake-Wallula alignment is a fold dominated structural trend west of Wallula Gap. East of Wallula Gap, the trend is primarily expressed by the Wallula fault zone. Geologic mapping has been unable to demonstrate the existence of a through-going surface fault associated with the fold structures west of Wallula Gap. Trenching across the Wallula fault zone, east of Wallula Gap, indicates non-involvement of the late Pleistocene-Holocene Touchet sediments. At Yellepit, west of Wallula Gap, a trench across the southern strand of the Wallula fault zone (Wallula Gap Fault) shows undisturbed fanglomerates of latest Pleistocene age, resting on faulted Miocene basalts. While this data limits the age of last movement, capability, as defined by 10CFR100, Appendix A, of the Rattlesnake-Wallula structural trend, cannot be demonstrated unequivocally on the basis of existing data.

Gable Mountain is the closest structure to the WNP 1-2-4 site, at a distance of approximately 6 miles to the subsurface extension of the anticlinal axis. There is evidence of faulting with probable late Pleistocene displacement of 0.2 - 0.3 feet mapped over a distance of about 1100' along strike. The cause for this displacement is currently under investigation.

No surface faulting within a distance of 5 miles of the WNP 1-2-4 sites is known. Approximately 3 miles to the east of the WNP 1-2-4 site, Pliocene Ringold sediments lie essentially horizontal with no sign of deformation, for a distance of at least five miles both north and south. No evidence was detected for faulting at depth directly underneath the WNP 1-2-4 site on the basis of borehole

geophysics. Within the limits of resolution of the gravity and aeromagnetic data, no linear features interpreted to be faulting have been identified within 5 miles of the WNP 1-2-4 sites.

From detailed geologic and geophysical studies, there appear to be no potential hazards that will adversely affect the plant structures at the site due to natural geologic phenomena of from man's activities.

Historically the seismicity in the vicinity of the site has been low, with infrequent earthquakes of low to moderate intensity or magnitude. Damage from earthquakes in the vicinity of the site has been minor.

The maximum vibratory ground motion affecting the site during historic time is estimated to have been approximately 0.015 g. The maximum vibratory ground motion potential at the site from an earthquake associated either with a tectonic structure or with a tectonic province is estimated to be 0.125 g.

Vibratory acceleration levels of 0.25 g and 0.125 g are assigned as conservative design bases for the Safe Shutdown Earthquake (SSE) and the Operating Basis Earthquake (OBE), respectively.

The existing glaciofluvial sand was excavated down to the underlying, very dense Ringold gravel and replaced in a denser state by compaction (structural backfill). Groundwater was not encountered during excavation operations and appears to be stable at about elevation 380 feet (about 10 to 15 feet below the surface of the Ringold gravel).

Excavated site soils were used for the structural backfill. Close control on compaction procedures was maintained to verify that specified densities (average relative density of 85 percent, with a minimum relative density of 75 percent) and uniformity of compaction were achieved during placement of backfill.

Foundations of all WNP-2 plant structures are supported in structural backfill. The backfill provides safe bearing for the structural foundations, and settlements are estimated to be minimal. Systematic monitoring and analysis of settlements of Seismic Category I foundations at the site have been continuous since the beginning of construction.

The structural backfill and underlying dense Ringold gravels do not appear to be susceptible to loss of strength,

subsidence or liquefaction resulting from potential vibratory ground motions that might be associated with the design earthquakes.

It is concluded that the site is suitable for a nuclear power plant in terms of Appendix A to 10 CFR Part 100, "Seismic and Geologic Siting Criteria for Nuclear Power Plants", and that the design basis for vibratory ground motion (Safe Shutdown Earthquake) should be 0.25g without the need to consider surface faulting at the site. It is also concluded that subsurface materials will adequately provide vertical and lateral stability to all structures under, and dynamic conditions contemplated for, this facility at this site as constructed.

The geological, geophysical and seismological investigations performed in support of Chapter 2.5 include:

Geologic, geophysical, and seismologic studies of the site and region for a distance of 200 miles;

Remote sensing studies of the site and region for a minimum distance of 50 miles;

A 40,000 line mile aeromagnetic survey centered on the site;

Compilation of a gravity maps for the site, site province and Pacific Northwest region;

Compilation of geologic and tectonic maps for the Pacific Northwest region;

Detailed geologic mapping of the site and surrounding area for at least five miles and additional detailed mapping of significant structures;

Subsurface exploration at the site and vicinity involving borings for geologic and soil engineering purposes, geologic trenching, and test pits;

Representative penetration resistance tests and in situ deformation and density tests in boreholes, and percolation tests and in situ density test pits;

Laboratory testing of samples taken from borings, trenches and test pits for engineering and geologic analysis;

Measurement of groundwater levels in boreholes;

Borehole geophysical surveys for stratigraphic correlation among borings;

Uphole/downhole and crosshole seismic geophysical surveys;

Surface seismic refraction to define overburden stratigraphy and to delineate depth to rock in the vicinity of the site.

The preparation of the Geology and Seismology section of the original WNP-2 PSAR was under the direction of Burns and Roe, Hempstead, New York, with assistance of the following:

Dr. Howard A. Coombs - Principal Consultant
Dr. James W. Crosby III - Consultant for Borehole Geophysics
Shannon & Wilson, Inc. - Soil Properties, Foundation Engineering

Dr. R. P. Miller - Project Manager
D. Schwantes - Project Soils Engineer
R. E. Brown - Consultant for Bedrock Stratigraphy

Weston Geophysical Research, Inc. - Seismic Surveys and Borehole Geophysics

T. F. Sexton - Project Manager
E. N. Levine - Field Project Manager
Dr. F. T. Turcotte - Seismologist

The original preparation of the Geology and Seismology section of the PSAR for project WNP-1 and WNP-4, which is utilized in the WNP-2 FSAR, was under the direction of United Engineers and Constructors, Philadelphia, Pa. with the assistance of the following:

Weston Geophysical Research, Inc. - Seismicity, Seismic Surveys, Borehole Geophysics, and Response Spectra

R. J. Holt - Project Director/Seismologist
Dr. F. T. Turcotte - Seismologist
Dr. H. Coombs - Consultant/Geology
Prof. J. W. Crosby III - Consultant for Borehole Geophysics
Rev. D. Linehan, S. J. - Consultant/Seismicity
Prof. R. W. Whitman - Consultant/Response Spectra
T. F. Sexton - Project Manager

E. N. Levine - Project Coordinator for Field Studies

FUGRO, Inc. - Geology

E. A. Danehy - Project Manager
Owen Swanson - Project Geologist
Donald Hitchcock - Assistant Project Geologist

Shannon & Wilson, Inc. -
Soil Properties, Foundation Engineering, Soil Engineering

Dr. R. P. Miller - Project Supervisor
H. H. Druebert - Project Soils Engineer
T. L. Olmsted - Project Engineering Geologist
Dr. I. Arango - Structure-Soil Interaction Evaluations
Prof. A. J. Hendron - Consultant/Soil Dynamics
Prof. H. B. Seed - Consultant/Structure-Soil Interaction
Prof. J. Lysmer - Consultant/Structure-Soil Interaction

Extensive additional geologic, geophysical and seismological data was compiled for the WNP1 and WNP4 PSAR as part of studies related to the December 14, 1872 earthquake. This information was filed in October, 1977 as Amendment 23 to the WNP 1 and WNP 4 PSAR and is included herein by reference. This work was performed under the direction of United Engineers and Constructors with technical guidance provided by the following:

Dr. H. A. Coombs - Principal Consultant
Dr. G. A. Davis - Professor Geology, USC
Dr. Don Tocher - Chief Seismologist, Woodward-Clyde Consultants

The preparation of the Geology and Seismology section presented in the initial submittal of the WNP-2 FSAR was the responsibility of Burns and Roe, Woodbury, New York with assistance of the following:

Dr. H. A. Coombs - Principal Consultant

Shannon & Wilson, Inc. -
Chapter 2.5 preparation, Regional Geology and Seismicity, Site Geology, Soil Properties, Foundation

Dr. R. P. Miller - Project Supervisor
H. H. Druebert - Project Manager and Soils Engineer
H. H. Waldon - Consultant Geologist

Weston Geophysical Research -
Seismicity Review, Site Geophysics, and Borehole Geophysics

V. J. Murphy - Project Manager
T. F. Sexton - Assistant Project Manager
Andrew LaCroix - Project Coordinator/Seismology
Richard Allen - Supervisor Data Processing/Site Geophysics

Amendment 18 to the WNP-2 FSAR integrates all of the new geological and geophysical data accumulated since the FSAR was originally docketed. It is a synthesis through 1980 based on the original FSAR and incorporates new information contained in Amendment 23 (WNP-1/4 PSAR), consultant reports, and Rockwell Hanford Operations' Basalt Waste Isolation Project reports. The amendment was prepared under the direction of the Supply System by Weston Geophysical Corporation with the assistance of the following:

Dr. H. A. Coombs - Professor Emeritus, University of Washington
Dr. G. A. Davis - Professor Geology, University of Southern California
Dr. R. B. Smith - Professor Geophysics, University of Utah
Mr. D. D. Tillson - Consulting Geologist

Weston Geophysical Corporation - Geophysics

Mr. V. J. Murphy - Project Manager
Mr. J. T. Doherty - Assistant Project Manager
Mr. J. Imse - Geophysicist
Prof. G. Simmons - Consultant/Geophysics
Prof. K. E. Glass - Consultant/Remote Sensing
Prof. B. C. Birchfiel - Consultant/Geology
Prof. H. P. Laubscher - Consultant/Geology

Woodward-Clyde Consultants - Geology

Dr. A. Patwardhan - Project Director
Mr. D. Gross - Project Manager
Mr. D. Hitchcock - Project Geologist

Information on Gable Mountain, Gable Butte, and Umtanum Ridge was provided by Northwest Energy Services Company with assistance of the following:

Golder Associates - Geology

D. Caldwell - Project Manager
G. Anttonen - Geologist

2.5.1 BASIC GEOLOGIC AND SEISMIC INFORMATION

2.5.1.1 Regional Geology

The WNP1-2-4 site is located within the Pasco Basin, one of several physiographic depressions occupying the Columbia Plateau (Figure 2.5-1). The Columbia Plateau, a major physiographic, geologic, and tectonic province is surrounded by the Blue Mountains and High Lava Plains provinces on the south, Northern Rocky Mountains and Idaho Batholith provinces on the east, the Okanogan Highlands province on the north, and the Cascade Mountains, Puget-Willamette Trough, and Washington-Oregon Coast Ranges on the west.

Discussions of the regional geology and tectonics of the Pacific Northwest have been presented by McKee (1972) and King (1959, 1969). Other notable treatments related to individual states and provinces have been made by Gunning and White (1966), Livingston (1969), Baldwin (1969), Ross and Savage (1967), the United States Geological Survey (1964, 1966, 1969), Rockwell (1979), Laubscher (Appendix 2.5-O), and Davis in (Washington Public Power Supply System 1977; and Appendix 2.5-N).

The Tertiary rocks of the Columbia Plateau are bordered on the northwest, north, east, and south by pre-Cenozoic rocks. Only to the west, in the Middle Cascades, Puget-Willamette Trough, and Coast Ranges, are rock units older than Cenozioc not in evidence.

Bedrock in the Columbia Plateau consists of a thick sequence of Miocene basalt flows with minor amounts of interflow sediments. These rocks are locally mantled by younger sediments of Pliocene to Holocene age (Figure 2.5-2). The basalts, particularly in the western parts of the Columbia Plateau, have been folded into a series of east-west anticlines.

The Pacific Northwest gravity map (Figures 2.5-9, 2.5-9a, and 2.5-9b) includes Oregon, Washington, and parts of

British Columbia and Idaho. The aeromagnetic map for the central part of the area is presented in Figures 2.5-8, 2.5-8a, and 2.5-8b.

Within the Pacific northwest area, most gravity anomalies appear to correlate well with mapped rock masses and geologic structures. Regional trends present in the Bouguer gravity fields are different west of the axis of the northern and central Cascade Mountains. A prominent north-south gravity gradient coincides with the Puget-Willamette Trough, whereas the Columbia Plateau is characterized by a broad northeast trending 40 mgal gravity high centered in the area of the WNP 1-2-4 site.

2.5.1.1.1 Geologic History

The tectonics and geologic history of the Pacific Northwest are the consequences of complex and poorly understood interactions among the Pacific, North American, and Juan de Fuca plates. The Juan de Fuca, 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 Explorer, Juan de Fuca, and Gorda Ridges and the transform faults (fracture zones) which bound them. The Juan de Fuca-North American plate boundary is one of past convergence and one along which subduction probably still continues at a rate of several centimeters per year (Riddihough and Hyndman, 1976).

In the Mesozoic, new oceanic crust that was generated along the East Pacific Rise, spread westward as part of the Pacific plate and eastward as part of the Farallon plate. The Farallon plate was subducted at a trench along the western margin of the North American plate. About 29 million years ago, the relative motion of the Farallon and North American plates resulted in the oblique impingement of the East Pacific Rise along the marginal trench (Atwater, 1970). Consequently, the Farallon and Pacific plates and the North American and Pacific plates were in direct contact along a right lateral transform fault system. Triple junctions of the Farallon, Pacific, and American plates then migrated along the zone of transform shear both northward and southward. The San Andreas Fault grew in length as more of the East Pacific Rise and bordering fracture zones reached the trench along the North American plate boundary. Continued spreading resulted in subduction of most of the Farallon plate to the extent that only a few relict pieces remain. The remaining segments of the Farallon plate began to interact with the North American plate as small, independent plates. The Cocos, Rivera, and Juan de Fuca

plates are small, relict pieces of the Farallon plate that have behaved as independent plates with directions and rates of spreading different from those of the parent Farallon plate.

In the Pacific Northwest during the early Tertiary, the coastline shifted from the western side of the present Cascade Mountains to the western side of the Coast Ranges with the accretion of the Siletz/Crescent volcanic assemblage. Early Tertiary deposits in the Cascade Range include thousands of feet of strata of the brackish Puget Group and continental Swauk, Ohanapecosh, Chumstick, and Roslyn Formations. In Eocene-Oligocene time, volcanism and continental sedimentation were active in much of the Cascade Volcanic Belt, the Intermontane Belt, the Omineca Crystalline Belt, and the Blue Mountains. During this early Tertiary episode of widespread volcanism and sedimentation, erosion was unroofing the Mesozoic batholiths. Granitic intrusions were emplaced in north-central Washington and northeastern Oregon during this period.

During Miocene-Pliocene time, large parts of southeastern Washington and northeastern Oregon were covered by the flood basalts of the Columbia River Basalt Group. On the Columbia Plateau the flows generally advanced from east to west. Deformation of the Columbia River Basalts commenced in the Miocene and continued into Pleistocene.

In the Cascade Volcanic Belt and Blue Mountains, minor uplifts caused the flanking and intermontane basins to receive early and middle Pliocene sediments of the Dalles and Ellensburg Formations. Throughout the Pliocene, the Cascade Volcanic Belt was marked by local volcanism. The most recent uplift of the Cascade Range and Blue Mountains took place predominantly during the Plio-Pleistocene.

Large stratovolcanoes were formed during the early Pleistocene in the Cascade Mountains (Figure 2.5-22). Major active stratovolcanoes included Mt. Baker, Glacier Peak, Mt. Rainer, Mt. St. Helens, Mt. Adams, and Mt. Hood. Some of these vents have been active into the Holocene, notably Glacier Peak and Mt. St. Helens and have erupted large volumes of ash. The ash falls from these eruptions were deposited over a large part of the region and provide time stratigraphic horizons for dating Quaternary geologic events. The latest significant eruption occurred in 1980 from Mt. St. Helens.

Vast sheets of continental ice covered most of the northern part of the region during the Pleistocene. The glaciers

advanced and receded several times, sculpturing the landscape and depositing till and glaciofluvial sediments. In the Columbia Plateau, the Channeled Scablands resulted from the great outpourings of floodwater after breaching of ice-dammed lakes in Montana (Bretz, 1923). Sub-basins of the Columbia Plateau, like the Pasco Basin, received glaciofluvial sediments. In eastern portions of the Columbia Plateau windblown, silty sediments (Palouse soils) were deposited over large areas. Erosion and deposition by rivers and wind continued to be significant in the Columbia Plateau in Holocene time.

The evolution of the region in light of current plate tectonics theory is discussed by Dr. G. A. Davis in Appendix 2.5-N. This discussion of late Cenozoic tectonics of the Pacific Northwest is primarily concerned with Miocene and younger events in the Columbia Plateau and adjacent geologic provinces. An earlier review (Davis in Washington Public Power Supply System, 1977a) treated evolution of the entire Pacific Northwest from Precambrian time to the present. The analysis by Davis, presented in Appendix 2.5-N, relies heavily on data and concepts developed by earth scientists since 1977, including the structural model for the Columbia Plateau developed between 1977 and 1980 by Dr. Hans P. Laubscher, University of Basel, Switzerland (Appendix 2.5-O).

Specific topics treated by Davis are: (1) the nature of the present stress state and strain patterns in the Pacific Northwest; (2) present North American-Pacific-Juan de Fuca plate relationships; (3) Quaternary and late Tertiary geometric and kinematic interrelations between the Basin-and-Range province and the Pacific Northwest; (4) the geologic significance of the Olympic-Wallowa lineament; and (5) the implications of Quaternary deformation in the Columbia Plateau.

2.5.1.1.2 Provinces

The Pacific Northwest physiographic provinces are shown on Figure 2.5-1. The source of the province boundaries and descriptions are taken from McKee (1972), Washington Public Power Supply System (1977b), and Rockwell (1979). No distinction is made between geologic and physiographic provinces due to their coincidence throughout the Pacific Northwest.

From the Pacific Ocean eastward, the first major physiographic feature is the Coast Range of Washington and Oregon. These mountains extend northward from the Klamath Mountains of southern Oregon to the Strait of Juan de Fuca.

East of these mountains lies the Puget-Willamette Trough, a series of topographic lowlands that extend parallel to the Coast Range from the Willamette River valley on the south to the Strait of Georgia on the north. East of the the Puget-Willamette Trough are the Cascade Mountains. The Cascade Mountains extend from northern California to southern British Columbia where they merge with the Coast Mountains. East of the Cascades, the north-south grain of the regional physiography gives way to an east-west grain. From north to south, the principal elements are the Okanogan Highlands, the Columbia Plateau, the Blue Mountains, and the High Lava Plains. To the east and north in Idaho, western Montana, and British Columbia, the north to northwest regional grain returns in the form of the Northern Rocky Mountains.

The WNP 1-2-4 site (Figure 2.5-1) lies in southeastern central Washington within the Columbia Plateau province. The site is situated near a north-south stretch of the Columbia River, the major watercourse in the region. The Pasco Basin contains the site and comprises approximately 1600 square miles of undulating semiarid plain with low-lying hills, dunes, and intermittent streams. The northern and southern boundaries of the Pasco Basin (Figures 2.5-4, 2.5-6a, 2.5-6b, and 2.5-6c) are defined by the Saddle Mountains and Rattlesnake Mountain, respectively. The easterly ends of Umtanum and Yakima Ridges mark the western boundary of the basin. To the east the basin merges into a vast expanse of dunes, dissected flatlands, and coulees northwest of the Snake River. A detailed discussion of the Columbia Plateau province is contained in Section 2.5.1.2. A detailed discussion of the Pasco Basin is contained in Section 2.5.1.2.4.

2.5.1.1.2.1 Columbia Plateau

The Columbia Plateau province is bounded by the Blue Mountains and High Lava Plains on the south, the northern Rocky Mountains-Idaho batholith on the east, the Okanogan Highlands on the north, and the Cascade Mountains Province on the west.

The Columbia Plateau is drained by the Columbia River which flows westward toward the Pacific Ocean. The Snake River joins the Columbia River after draining the eastern Columbia Plateau and parts of the adjoining provinces to the east and south. Most of the Plateau (see Figure 2.5-7) has gentle topographic relief. Exceptions to this gentle relief are the deep gorge of the Columbia River, the many steep-walled

coulees north and east of the Columbia River, and the series of linear, generally west to northwest-trending, anticlinal ridges in the vicinity of Yakima.

The Channeled Scabland of Washington covers the Columbia Plateau from Spokane on the northeast to the Snake River on the south and to the Columbia River on the west. The scabland topography was formed in Pleistocene time by the action of glacial meltwaters and catastrophic floods due to breaching of ice-dammed lakes in western Montana.

The Columbia Plateau formed between 16.5 and 6 m.y.b.p. (Watkins and Baski, 1974; McKee and others, 1977) when large volumes of basalts were erupted from north-northwest trending linear vent systems in northeastern Oregon and southeastern Washington (see Figure 2.5N-2)(Waters, 1961; Taubeneck, 1970; Swanson and others, 1975; Fruchter and Baldwin, 1975; Price, 1977; Swanson and others, 1977).

The lavas of the Columbia Plateau cover an area of approximately 78,000 square miles and have an estimated volume of 41,000 cubic miles (Figures 2.5-3 and 2.5N-2) (Swanson and Wright, 1978). The Columbia Plateau is surrounded by topographically higher areas. The character of the pre-Tertiary rocks covered by basalt is visible only within highlands surrounding the plateau.

Individual basalt flows are voluminous, generally 2 to 6 cubic miles, with a maximum known volume of 145 cubic miles. Flows range in thickness from a few inches to more than 300 feet, with an average thickness of 90 to 120 feet (Swanson and Others, 1979a). The thickest flows are interpreted as showing ponding in pre-basalt valleys, in structurally controlled basins that developed during volcanism, or in narrow canyons previously eroded into older flows (termed intracanyon flows) (Swanson and Wright, 1978).

Flows of the Columbia River Basalt Group are interbedded with and overlapped by Miocene-Pliocene epiclastic and volcaniclastic sediments, especially along the margin of the province. Invasive flows, formed when lava "burrowed" into surficial deposits of unconsolidated sediments, are also common (Byerly and Swanson, 1978). The youngest suprabasalt sedimentary units on the plateau are fluvial, lacustrine, glaciofluvial, and eolian deposits of Pliocene to Holocene age. Localized accumulations of Pliocene to Pleistocene lavas are also present within the western and southern portions of the province.

2.5.1.1.2.2 Coast Range of Oregon and Washington

The Coast Range of Washington and Oregon lies west of the Puget-Willamette Trough. It consists of a 50 to 80 mile wide belt of mountains which extends more than 600 miles from the Klamath Mountains in southwestern Oregon to the Strait of Juan de Fuca in northwestern Washington. The Olympic Mountains are part of the Coast Range and form the highest and most rugged part of the province. Among the oldest rocks in the region are a 15,000 foot sequence of basaltic volcanics (Siletz/Crescent volcanics) that probably represent an oceanic seamount chain that was accreted to the continental margin in early Tertiary time. South of the Straits of Juan de Fuca, the principal geologic units are Eocene through Miocene marine volcanic and sedimentary rocks. Faults in the Olympic Mountains are predominantly early and middle Tertiary in age. South of the Olympic Mountains, the structure of the Coast Range can be generalized as a system of folds of Miocene and Pliocene age.

2.5.1.1.2.3 Puget-Willamette Trough

The Puget-Willamette province is an elongate structural and topographic low that lies between the Cascade Mountains on the east and the Coast Range on the west. It extends from about Eugene, Oregon on the south to Vancouver, British Columbia, on the north. From south to north its major features are the Willamette Valley, the Puget Lowlands, and the Georgia depression. These troughs are the surface expressions of a large downwarp. This province is a depositional site containing thick fills of Tertiary and Quaternary sediments. From the Puget lowlands northward, Pleistocene glacial sediments are the dominant geologic unit present. South of the Puget lowlands, the sediments are predominantly alluvial. Low hills occur in Washington between the Columbia River and the Puget Lowlands. On the north, the Puget Sound area was extensively modified during Pleistocene time by the Puget Lobe of the continental glacier.

2.5.1.1.2.4. Blue Mountains

The Blue Mountains province lies immediately south of the Columbia Plateau in northeastern Oregon. This province includes the Ochoco Mountains on the west and Wallowa Mountains on the east. The overall structure can be characterized as a 200-mile long, northeast-trending series of anticlines with moderately steep northern flanks and gentle southern flanks. Superimposed upon these asymmetric structures is a north to northwest trend of folds and high

angle faults of late Cenozoic age. Initial development of the Blue Mountains anticlinal system predates eruption of the Columbia Plateau basalts. The Blue Mountains are capped by essentially the same Cenozoic volcanic rock sequence which fills the Columbia Plateau to the north. Uplift and resulting erosional stripping have exposed older rock strata which do not crop out in the Columbia Basin. These pre-Tertiary rocks consist of Paleozoic and lower Mesozoic ophiolitic rocks, and associated sedimentary and volcanic rocks that have been intruded by Mesozoic granitic plutons (c.f. Davis in Washington Public Power Supply System, 1977a).

2.5.1.1.2.5 High Lava Plains Province

The High Lava Plains province, which is characterized by the presence of youthful volcanic rocks, is located south of the Blue Mountains and east of the Cascade Mountains provinces. It is a transitional province between the Blue Mountains and the Basin and Range province farther south. Most rocks exposed within the province are high-alumina basalt flows of Miocene to Pliocene age or younger. Younger basaltic and rhyolitic ash flows (as young as Holocene) were erupted from cinder cones and volcanic vents found within the topographically subdued province. The most prominent structural feature within the province is the Brothers fault zone, a diffuse northwest-trending zone of deformation first described in detail by Lawrence (1976) and discussed by Davis in Appendix 2.5-N.

2.5.1.1.2.6 Northern Rocky Mountains and Idaho Batholith

The Northern Rocky Mountains are located to the east of the Okanogan Highlands province and to the northeast of the Columbia Plateau province. Major drainages flow toward the Columbia Plateau including the Spokane, Clearwater, and Salmon Rivers. Rocks of the Northern Rocky Mountains province are predominantly slates, argillites, and quartzites of the Precambrian Belt Supergroup.

Northwest-striking vertical faults and parallel folds of Tertiary age exist north of a west-northwest trending fracture belt known as the Lewis and Clark Line. The Rocky Mountain Trench is a north-south valley at least 800 miles long which is formed along lines controlled by faulting. Eastward is the northwest-trending Rocky Mountain thrust belt.

The Idaho Batholith lies east of the Columbia Plateau and Blue Mountains and south of the northern Rocky Mountains. It is a major batholithic intrusion exposed over an area of

approximately 16,000 square miles. The rocks are predominantly granodiorite and quartz monzonite of Mesozoic age. Eocene plutons and minor amounts of other granitic rocks are also present. Much of the batholith appears to have been intruded as a unit or as a number of large, related, and possibly gradational units. Within the border zone of the batholith, the sedimentary rocks have been metamorphosed to gneiss and schist, within which are bodies of igneous rocks. These marginal rocks are deformed as well as metamorphosed. The dominant types of deformation along the margins of the province are folding and thrust faulting. Internally the batholith is essentially undeformed.

2.5.1.1.2.7 Okanogan Highlands

The Okanogan Highlands province lies east of the Cascade Range and north of the Columbia Plateau. The province consists of broad, rounded hills. The topography of the Okanogan Highlands has been largely molded by glacial and glaciofluvial processes active during late Pleistocene time. The major drainage in the Okanogan Highlands province is the Columbia River, which flows north-south through the province and east-west along its southern boundary. The river divides the highlands into eastern and western regions (Yates and others, 1966).

The eastern region is underlain by Precambrian Beltian rocks and by a lower Paleozoic miogeosynclinal sequence of quartzites, phyllites, and carbonates. The Kootenay Arc, a fold belt trending northward in a convex eastward arc from the Columbia Plateau to the Columbia Mountains, is the dominant structural feature of the eastern region. The folds of this belt trend approximately northeast and are cut by pre-Tertiary faults. The western region is composed of upper Paleozoic and Mesozoic rocks. These rocks constitute a eugeosynclinal assemblage of graywackes, greenstones, shales, cherts, conglomerates, and limestone. Late Mesozoic granodiorites and quartz monzonites intrude these sedimentary sequences, and lower Tertiary volcanic rocks unconformably overlie all of the older rocks. In the western region, north-trending, high-angle normal faults of Cretaceous and early Tertiary age are the most conspicuous structures.

Within the Okanogan Highlands the major stuctural element is the Republic graben, a northeast-trending structure about 4 to 10 miles wide, that extends approximately 50 miles north to the Canadian Border. The graben is bounded on the west by the Bacon Creek and the Scatter Creek fault zones and to

the east by the Stevens fault and several smaller faults. Apparent displacements along these bounding faults range from several hundred feet or less to over 17,000 feet (Yates and others, 1966). Geologic observations at Grand Coulee Dam and reconnaissance observations by J. A. Blume and Associates (1972) for Washington Public Power Supply System in the region between the dam and the southern most mapped exposures of the eastern border faults showed that no discernible evidence of faulting exists in the Columbia Plateau along the southern projection of the graben. Blume's reconnaissance study made this conclusion based on the following observations: no fault offset of the glacial till resting on the granitic rocks of the Colville batholith or of the granitic rocks themselves; continuity, both vertically and horizontally, of stream patterns across the projected traces of the Republic graben faults; no faulting or disturbance of the Columbia River basalts south of the Columbia River except by the Coulee and Barker Canyon monoclines; and concordance of Columbia River terraces across postulated traces of the bounding faults. These terrace surfaces and the underlying terrace deposits of the Nespelem Formation can be traced across the fault projections without interruption. The most prominent terrace surface, the Nespelem Terrace, is at least 12,000 years old and the base of the Nespelem Formation is in excess of 27,000 years old (Blume, 1972). These last two observations have been verified by mapping and extensive exploratory drilling along the east bank of the Columbia River downstream of Grand Coulee Dam, and by geologic mapping of the Columbia River basalts in the vicinity of Banks Lake (BUREC, 1974).

The age of the Republic graben is not precisely known although Staatz (1964) felt the last movement on the bounding faults was Miocene. Muessig (1967) in his mapping of the northern Republic graben states:

> "The evidence for later and probably periodic movements along the marginal faults...indicates that the major movement on the Bacon Creek fault probably had terminated before the deposition of the basalt member of the Klondike Mountain Formation [late Oligocene - early Miocene]. There is no limiting upper date for movement along the Stevens fault but it seems logical to suppose that the history of movement was not very much different from that of the other graben faults."

The lack of offset of glacial deposits (kame terraces and lake sediments related to Pleistocene glaciation) indicates a lack of recent fault activity (BUREC, 1974).

2.5.1.1.2.8 Cascade Mountains

The Northern Cascades are separated from the Coast Mountains of British Columbia by the Fraser River Valley. Snoqualmie Pass, in west-central Washington, forms the approximate boundary between the Northern Cascades and the rest of the range.

The topographic character of the province changes progressively to the north, reflecting the increasing erosional effects of alpine glaciation at higher altitudes, and the overriding of much of the Northern Cascades by Cordilleran ice sheets.

The Northern Cascades are somewhat higher overall than comparable areas to the south of Snoqualmie Pass. Similarly, the Coast Mountains of British Columbia are somewhat higher overall than the Northern Cascades. Towering several thousand feet above the average crestline of the Cascade Mountains are the principal young stratovolcanoes: Mt. Garibaldi, Mt. Baker, Glacier Peak, Mt. Rainier, Mt. St. Helens, Mt. Adams, and Mt. Hood (Figure 2.5-22).

The Northern Cascades consist of Paleozoic and Mesozoic marine strata that have been severely deformed, regionally metamorphosed, and invaded by numerous igneous intrusives of Cretaceous and Tertiary age. The core of the Northern Cascades is composed of metamorphosed sedimentary and volcanic rocks and granitic intrusives. These are flanked by younger sedimentary and volcanic rocks and are tightly and intricately folded along north-northwesterly trending axes. Paleozoic and Mesozoic rocks are also tightly folded in places, but they are less intensely deformed than the metamorphic rocks. In general, the Tertiary rocks are not tightly folded, although locally they are faulted and deformed. The Central Cascades consist almost entirely of nonmarine volcanic and sedimentary rocks of Cenozoic age, and are free of large-scale granitic intrusions. Pre-Tertiary rocks in the Central Cascades are exposed in only a few locations, notably Tieton Reservoir to the west of Yakima. A thick sequence of Paleocene to Eocene volcanic and sedimentary strata comprises the core of the central Washington Cascades. These rocks are folded and faulted along a northwesterly trend. They are unconformably overlain by a series of Oligocene to Miocene volcanic rocks and associated sediments. These strata are faulted and folded, although not as strongly, along the same northwest trend as the underlying strata.

North of Mount Hood, the present Cascades were formed by late Cenozoic upwarping along a north-south axis, producing over 5,000 feet of structural relief. Late Cenozoic volcanic deposits unconformably overlie the middle Tertiary strata. Erosional remnants of Miocene to Pliocene Columbia River Basalt are found near the margins of the province, but the bulk of the late Cenozoic deposits consist of the extensive Plio-Pleistocene and Quaternary outpourings of the active stratovolcanoes Mt. St. Helens, Mt. Rainer, Mt. Hood, and Mt. Adams. These deposits were extruded from numerous vents and fissures and include pyroclastic and mudflow deposits and lavas of a variety of compositions.

These stratovolcanoes account for a large portion of the late Cenozoic volcanic rocks in the province. Quaternary glacial deposits are less abundant in the Cascade Mountain province south of Snoqualmie Pass than in that part of the province to the north. Holocene alluvial deposits mantle the older rocks in major stream drainages.

The structural trends within the Central Cascade Mountains of Washington are primarily northerly and northwesterly. They contrast with the more westerly trends to the east in the Columbia Plateau.

Three major structural elements are present within the northern Cascades. They are the Straight Creek-Fraser River fault system, the Methow Graben, and the Chiwaukum Graben. Each of these prominent elements is discussed separately below.

Straight Creek and Fraser River Fault Zones

The Straight Creek and Fraser River fault (Figure 2.5-3) zones are major north-northwest trending zones of faulting in northwestern Washington and southwestern British Columbia. As presently mapped, the Fraser River fault zone extends from Lillooet, British Columbia to the latitude of Chilliwack, British Columbia. The Straight Creek fault zone extends from just north of the Skagit River south to the vicinity of Snoqualmie Pass.

In British Columbia the zone is named the Fraser River fault zone and is dominated at its southern end by two major subparallel structures, the Hope and Yale faults. These two faults and their associated structures form a continuous zone approximately 3 miles wide that extends at least 160 miles north from the Canadian border. This zone is well-defined along most of its length by prominent

alignments of hillside notches, linear talus deposits, and faceted spurs along both walls of the deep, linear canyon of the Fraser River (Woodward-Clyde Consultants, 1978).

Although not as conspicuously defined as the Fraser River fault zone, the Straight Creek fault zone can be traced more or less continuously through pre-Oligocene rocks of the North Cascades. Along most of its length the Straight Creek fault zone separates a complex terrane of unmetamorphosed to slightly metamorphosed, pre-Tertiary volcanic and sedimentary rocks west of the fault zone from a pre-Mesozoic terrane of medium grade schist, orthogneiss, and migmatite intruded by Mesozoic granitic plutons east of the zone. Within the fault zone itself, narrow strips of downfaulted Eocene continental sediments, highly deformed blocks of the lithologies observed both east and west of the zone, and small serpentinized bodies are common. The dip of the fault zone ranges from approximately 65° E to vertical. These steep dips are reflected in the linear trace of the zone (Figure 2.5-7).

The Straight Creek fault zone represents a profound structural discontinuity. West of the fault zone, rocks of low temperature, high pressure, metamorphic conditions (the Shuksan Metamorphic Suite) are involved in a series of imbricate thrust faults. East of the zone, large thrust faults are absent and the rocks are typical of high temperature and low pressure conditions. Misch (1966) points out that only dip-slip displacements on the Straight Creek fault cannot easily explain this close juxtaposition of contrasting metamorphic facies and terranes, and that considerable strike-slip displacement is likely. He has suggested that the geology exposed west of the Fraser River fault zone near Harrison Lake in British Columbia (Lowes, 1971) is very similar to the geology in the Kachess Lake area. About 120 miles of right-slip displacement on the Straight Creek and Fraser River fault zones would be suggested by these relations (Misch, 1977b).

Although the age of latest movement of the Straight Creek and Fraser River fault zones is not precisely known, several geologic relationships put contraints on the age. The Straight Creek fault zone cuts the post-Early Cretaceous Shuksan thrust, showing that some of the displacement on the fault zone was younger than Early Cretaceous. The Kachess Lake fault, a possible southern continuation of the Straight Creek-Fraser River fault system, includes tectonic slivers of the Late Cretaceous Mount Stuart batholith, indicating that displacement occurred on it after the Late Cretaceous (Shannon and Wilson, 1977a). Several mid-Tertiary

batholiths intrude the fault zone and show little or no displacement across its general trend (Misch, 1966; Washington Public Power Supply System, 1977c; Tabor and Frizzell, 1979; and Shannon and Wilson, 1977a). These intrusions are the Chilliwack and Mount Barr batholithic phases of the Chilliwack composite batholith, the Monte Cristo pluton and the Snoqualmie batholith. K-Ar age determinations of these batholiths are, respectively, 29-16 m.y., 25 m.y., and 18 m.y.b.p., indicating that extensive strike-slip displacement has not occurred along the fault zone since Oligocene time.

Scattered exposures of Eocene continental sandstone, shale, and conglomerate of the Chuckanut and equivalent Swauk formations are found along the fault zone. Many of these exposures are faulted, demonstrating that displacement on the Straight Creek fault zone continued at least into Eocene time. In the Kachess Lake area, the Naches Formation of Eocene age is also involved in faulting. Also, Late Oligocene ash flow tuffs equivalent to the Stevens Ridge Formation overlie strongly deformed Naches Formation along an angular unconformity, indicating that major movement ended on the Kachess Lake fault within Oligocene time (in Washington Public Power Supply System, 1977c). These same ash flow tuff units also lie across the projection of the Straight Creek fault in the Monte Cristo area, suggesting a similar age relationship (in Washington Public Power Supply System, 1977c).

South of the Kachess Lake the Goat Peak segment of the Straight Creek fault becomes the northwest-striking Taneum Lake fault. 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 (Washington Public Power Supply System, 1977d), its age relationship to Miocene plateau basalts is unclear. However, Tabor and Frizzell (1979) conclude that Miocene or younger movement along either the southern Straight Creek fault "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 fault."

The Methow Graben

The Methow graben is a large northwest-trending structural low, lying between the Mesozoic crystaline plutonic belts of the North Cascade crest region to the west and the Okanogan Highlands to the east. It is occupied largely by a

stratigraphic section of marine and continental sedimentary and volcanic rocks of Jurassic-Cretaceous age (Barksdale, 1975). The Pasayten fault, the northeastern boundary of the graben, is a single surface that strikes about N30°W. The western border fault consists of a number of short segments interrupted by late Mesozoic and Tertiary granitic intrusions. The segments are not colinear and have an average trend of about northwest. Both border faults extend into the British Columbia. The Mesozoic sedimentary sequence is absent at the southern end of the graben, where the border faults converge, and the nature of their continuation through the crystalline rocks toward the Columbia Plateau has not been established.

The graben sedimentary sequence displays several episodes of folding and internal faulting which may in part have been synchronous with movement on the border faults. The ages of plutons which cut the border faults indicate that significant movement was largely completed by early Tertiary time.

The Pasayten fault, also referred to as the Chewack-Pasayten fault (Barksdale, 1975) and the Eightmile Creek fault (Lawrence, 1968; Staatz and others, 1971) is well-exposed north of Eightmile Creek where it dips steeply west. The southern mapped segment of the fault is very straight, but becomes sinuous toward and north of the British Columbia border.

The Pasayten fault separates crystalline plutonic rocks of the Okanogan terrane on the east from the thick supra-crustal sedimentary succession of the graben to the west. Lawrence (1968) postulates major stike-slip movement on the fault on the basis of the straightness of the fault trace. Interpretation of structural data from the rocks adjacent to the fault led him to postulate right-slip displacement. Barksdale (1975) interprets movement on the fault as largely dip-slip. Movement on the Pasayten fault predates the Island Mountain Volcanics just south of the Canadian border which lie across the fault with angular unconformity. The Island Mountain Volcanics are Tertiary, but, in the absence of radiometric dating, their precise age is not known. Attempts to trace the Pasayten fault south through the crystalline rocks toward the Columbia Plateau (Menzer and Swanberg, 1969; Washington Public Power Supply System, 1977e) have not proven successful. One point is clear: the fault does not continue directly on strike as a well-developed linear shear or mylonite zone through the crystalline rocks.

The western boundary fault system, is comprised of several irregular, non-colinear segments which collectively separate the Skagit terrane on the west from the Methow graben succession (or, locally, the Hozameen Formation on the east). The two northern segments mapped by Misch (1966) are referred to by him as the Ross Lake fault zone. The two southern segments studied by Barksdale (1975) have been called the Twisp River and Foggy Dew faults. The overall trend of the fault zone is approximately N45°W. Only the Twisp River segment shows strong physiographic expression. The fault zone continues north of the British Columbia border with a more northerly trend, as the Yale fault of the Frazer fault system. The southern segment, the Foggy Dew fault, is truncated by the Cooper Mountain batholith. Reconnaissance mapping by Shannon and Wilson (in Washington Public Power Supply System, 1977e) south of the Cooper Mountain batholith between Pateros and Chelan failed to identify the southern continuation of the fault in the crystalline rocks south of the graben. The Ross Lake segments of the fault zone are interrupted and separated from the Twisp River segment by several late Cretaceous and Tertiary plutons.

The Chiwaukum Graben

The Chiwaukum graben is a major structural and geographic feature in the Cascade Mountains (Willis, 1950b). It extends as a topograhically low region for about 62 miles NNW from Wenatchee, on the Columbia River, to near the crest of the Cascade Range. The graben is filled primarily by the Chumstick Formation, a thick sequence of folded and faulted continental sedimentary rocks of middle to upper Eocene age. Two major border fault systems, the Leavenworth on the west and the Entiat on the east, juxtapose the Chumstick against pre-Tertiary metamorphic and plutonic igneous rocks. Middle Miocene flows and intercalated sediments of Yakima Basalt lie unconformably across the graben at its southern end and post-date major faulting. Faults at the northern end of the graben are truncated by intrusive granitic rocks of the early Miocene Cloudy Pass batholith.

The Leavenworth fault system, at the western border of the graben, has been mapped as a zone of parallel faults at the southern end (Tabor and others, 1977) and as a single fault farther north. It trends generally north-northwest, but is irregular with several north-south segments, the longest of which is along its contact with the Mount Stuart batholith. The Entiat fault has been mapped as a single fault trace (Laravie, 1976). It is nearly straight and trends about N35°W.

An uplifted block of pre-Tertiary gneiss within the graben is bounded on the east and locally on the west by faults of the Eagle Creek system (Whetten and Laravie, 1976). These are all high-angle faults. Studies of sedimentary facies variations in the Chumstick Formation, which occupies the graben (Cashman, 1974; Laravie, 1976; and Whetten, 1977), demonstrate the coarse fanglomerates in the Chumstick were derived from local uplift of the pre-Tertiary crystalline rocks adjacent to and, in part, within the graben. These fanglomerates abut against and thin away from the contacts with the border faults, indicating that the faults were active during upper Eocene sedimentation. This early faulting was essentially tensional and probably represents a continuation of the extensional tectonic regime which led to formation of the Teanaway dike swarm in the area just to the south.

Movement on the Leavenworth and Entiat faults at the northern end of the graben was completed prior to emplacement of the Cloudy Pass batholith in early Miocene time. The latest fault movement at the southernmost end of the Entiat fault predates the early Oligocene Wenatchee Formation, which was deposited across the fault. Movement of the southern end of the Leavenworth fault ended before deposition of the lower Yakima Basalt in middle Miocene time.

Post-lower Yakima Basalt faulting has been recognized on two structures which parallel the master boundary faults of the Chiwaukum Graben. In the area between the northern part of Table Mountain and Mission Ridge, just outside the southwestern border of the Chiwaukum graben, Rosenmeier (1968) mapped a 16 km long, northwest-trending high-angle fault between Yakima Basalt on the west and Swauk Formation on the east. This fault has not been recognized within the Yakima Basalt southeast of the area of upper Nameum Creek. Shannon and Wilson (Washington Public Power Supply System, 1977c) suggest that the structure mapped by Rosenmeier as a fault also could result from a sharp monoclinal flexure.

Laravie (1976) recognized a major 30 km long fault just east of the Entiat fault along the western front of the Entiat Mountains. He initially postulated this fault, the Chumstick fault, largely on the basis of physiograhic evidence.Movement on the Chumstick fault may have been related to, and synchronous with, folding of the Yakima Basalt on the Columbia Plateau to the southeast. Slip on the Chumstick fault probably represents a structural adjustment in the basement rocks which accompanied folding of the basalts in Pliocene or earliest Pleistocene time. The Chumstick fault dies out or merges with the Entiat fault

both to the north and south where the bench and adjacent fault scarp disappear. The presence of a small erosional remnant of Wenatchee Formation sediments at an altitude of 3,800 feet on Burch Mountain southeast of the Chumstick fault was noted above. This indicates that post-Wenatchee dip-slip movement, synchronous with movement on the Chumstick fault, has occurred on this segment of the Entiat fault. This young movement on the Entiat fault, however, dies out before the fault reaches the Columbia River where the Wenatchee Formation lies at the same elevation on both sides of the fault (Washington Public Power Supply System, 1977f, Appendix 2RD).

2.5.1.2 Columbia Plateau Province and Site Geology

The WNP 1-2-4 site is located within the Pasco Basin of the Columbia Plateau province. A summary discussion of the Columbia Plateau province has been previously presented in section 2.5.1.1.2.1. The discussion that follows will present an expanded version of the geologic history of the Columbia Plateau with emphasis on the major structures surrounding the Pasco Basin. For additional detailed discussions of the geology, stratigraphy, lithology, and major folds of the Columbia Plateau refer to Washington Public Power Supply System (1977f) and Rockwell (1979).

2.5.1.2.1 Geologic History of Columbia Plateau

The Columbia Plateau is underlain by a thick sequence of plateau basalt flows of Tertiary age which, in turn, are overlain locally by Pliocene fluvial and lacustrine sediments, by Pleistocene glaciofluvial and eolion sediments, and by Holocene alluvium and eolian deposits. East of the Pasco Basin, the basalt flows are essentially horizontal with only slight regional dips. To the west and south, the lava flows have been folded into a series of prominent, easterly-trending, linear, anticlinal ridges (Figure 2.5-4). The geologic history of the Columbia Plateau has been synthesized by Davis (Washington Public Power Supply System, 1977a; and Appendix 2.5-N). The following discussion is taken primarily from Davis' analysis.

Pre-Cenozoic

The pre-Cenozoic rocks of the Columbia Plateau are buried by the Columbia River basalts. The northern Columbia Plateau boundary area is underlain by a pre-Tertiary basement complex of plutonic and metamorphic rocks. These crystalline rocks form the mountainous highlands to the north and west of the Plateau, and are exposed below the

Tertiary strata of the Plateau itself in the Columbia River canyon. These rocks include the Swakane Biotite Gneiss, Chelan Batholithic Complex, Methow Gneiss, and the Okanogan Batholithic Complex.

The absence of exposure of pre-Tertiary rocks in the plateau area has contributed to the concept of the Columbia Embayment, a convex-eastward embayment centered on the Columbia River between Oregon and Washington within which no "continental" or pre-Tertiary basement rocks should be found. The concept stems originally from deep resistivity studies in the area by Cantwell and Orange (1965), although the resistivity boundaries they drew between inferred basaltic oceanic and granitic continental crust do not coincide closely with the postulated Columbia Embayment.

The northwest-trending northern edge of the embayment was accepted by Skehan (1965) as a boundary between oceanic crust to the south and continental crust to the north, because it coincided approximately with a topographic lineament noted by Raisz (1945) and called by him the Olympic-Wallowa lineament (OWL). The two inferred features--the surficial topographic lineament and a deeper crustal boundary--have been linked together in the literature. This is so despite the fact that regional gravity surveys across the trace of the proposed lineament do not support its existence (Danes, 1969; Washington Public Power Supply System, 1977); that it cannot be confirmed by regional aeromagnetic studies (Zietz and others, 1971; Washington Public Power Supply System, 1977g); and that no offsets or disruptions of any geologic units along its hypothesized trace in the bedrock areas of northeastern Oregon have been described (Shannon and Wilson, 1980; Kendall and others, 1981; Davis in Appendix 2.5-N).

Previous analyses of the Columbia Embayment and the nature of the basement beneath it and the basaltic Columbia Plateau relied almost exclusively on the geologic map distribution of pre-Tertiary units. Much can be learned, however, about the distribution of such units in Washington state when the disruptive effects of late Cretaceous strike-slip faulting in the northern part of the state are considered. The Straight Creek fault is particularly important in this regard. As previously discussed, Misch (1977a) proposed that a distinctive suite of Cascade metamorphic rocks exposed east of the fault in the Stevens Pass area has been offset 120 miles to the Harrison Lake area of southern British Columbia. Pre-latest Cretaceous rocks extend 140 miles south of Harrison Lake on the west side of the fault. If Jurassic ophiolitic rocks exposed southeast of

Mt. Rainier at Rimrock Lake (Tieton Reservoir) lie west of the fault as well, as seems likely, then pre-latest Cretaceous rocks extend 200 miles to the south of Harrison Lake.

When right-slip along the Straight Creek fault is reversed by 120 miles to bring the Stevens Pass and Harrision Lake areas into juxtaposition, it becomes apparent that pre-latest Cretaceous rocks on the east side of the fault must extend as far south of Stevens Pass as they do on the other side south of Harrison Lake, i.e., 140 to 200 miles. In other words, rocks of this age must underlie the western Columbia Plateau at least as far south as the present Columbia River, and apparently well into north-central Oregon (Washington Public Power Supply System, 1977a).

Thus the portion of the Columbia Embayment east of the Cascade Range is apparently underlain by basement oceanic crust and associated sedimentary rocks. Additional evidence for this conclusion is the widespread distribution in western and central Washington of such rocks. Pre-Swauk Mesozoic units southeast of Stevens Pass (east side of Straight Creek fault) consist of the east-striking Ingalls "ophiolite" containing serpentinite, late Jurassic gabbro, and structurally associated chert, slate, and mafic volcanic rocks (Miller, 1977a, oral communication, and Miller, 1977b). The chert yields upper Jurassic radiolaria. This oceanic assemblage is believed to have been thrust northward (obducted) across the crystalline core of the Cascades. Structural emplacement occurred prior to intrusion of the Mt. Stuart batholith, 88 million years ago (Miller, 1977b). Similar late Jurassic oceanic rocks occur farther south on Manastash Ridge, near Cle Elum (also probably east of the Straight Creek fault) (Hopson and Mattinson, 1973; Southwick, 1974).

On the west side of the Straight Creek fault in Snohomish, Skagit, and King Counties, assemblages of marine clastic rocks, radiolarian ribbon chert, volcanic rocks, and minor lenses of limestone yield Tithonian and lowermost Cretaceous (Newcomian) fossils (Danner, 1966). These rocks originally lay southwest of Hanford but have been offset to their present location by movement along the Straight Creek fault. Their offset counterparts east of the south-projected Straight Creek fault should now lie just north of the Columbia River in the general vicinity of Goldendale Washington. The Rimrock Lake Jurassic ophiolitic assemblage has been described by Vance and others (1980). In north-central Oregon, Jurassic argillites were encountered in Chevron's Kirkpatrick well just north of

Condon. The argillites underlie a Tertiary Section (Columbia River Basalts, John Day Formation, Clarno Formation) that is approximately 2200 feet thick (S. Reber, Chevron Oil Corporation, personal communication).

Although strike-slip displacement along the Straight Creek fault appears to clarify the nature of basement beneath the Columbia Plateau, the extent south of Snoqualmie Pass is unclear. In extending major right-lateral strike-slip fault zones from northern Washington into the Columbia Plateau, the northeastern trend of pre-Tertiary inliers and structural elements appears to be unbroken across the southern edge of the embayment. Regional gravity anomalies in central and northeastern Oregon parallel these geologic trends and also appear to be unbroken (Thiruvathukal, 1970). The Columbia Embayment probably did not exist at the time of Straight Creek, Ross Lake, and related strike-slip faulting. Prior to latest Cretaceous time, the continental margin in Oregon is thought to have had a more southerly trend than its present NE-SW position. Clockwise rotation of this margin is judged to have occurred after regional strike-slip faulting to the north.

Cenozoic

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; e.g. Compton and others, 1977). Some of these complexes are now dated as Mesozoic in age, e.g. the Whipple Mountains of southeastern California (Washington Public Power Supply System, 1977a). Loring (1976) has documented widespread examples of Oligocene normal faulting of variable trend (north-south to east-west) within Utah and Nevada. Topography suggests that tectonic activity was 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 during that time (Christiansen and McKee, 1979). Sheet-like late Oligocene - early Miocene (25 to 20 m.y.b.p.) ignimbrites mantled a topography of low relief over extensive areas.

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 Blue Mountain province was deformed at some time after Eocene arc volcanism (Clarno Fm.), but prior to Columbia River basaltic volcanism. Before deposition of overlying Picture Gorge basalts, Clarno strata were locally folded as steeply as 60° on the northern limb of the east-west trending Aldrich Mountain anticline, the structural axis of the western Blue Mountains (Thayer & Brown, 1966). Robyn (1977) suggests that this deformation occurred between 25 and 20 m.y.b.p., but his older age limit is not well constrained. If folding was earlier, deposition of the Oligocene (36 m.y.b.p.) 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 folding. 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) is inferred by the occurrence of the Blue Mountains anticline along the northern edge of the John Day basin. This major fold constituted a topographic barrier in middle Miocene time and separated flows of the lower Yakima and Picture Gorge Basalts (Nathan and Fruchter, 1974).

Plateau folds and parallel faults, usually with thrust or reverse fault geometries began to develop during the extrusion of 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 (Camp and Hooper, 1980), is the 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 Naneum Ridge anticline (Rockwell, 1979, p. IV-16).

Evidence for deformation during Wanapum time, approximately 14.5 to 13.6 m.y.b.p., 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 began to form in late-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 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.b.p.), enabling Reidel and others (1980) to conclude that the rate of uplift of this structure between 14.5 and 10.5 m.y.b.p. was approximately 39 meters/million years.

Some of the clearest evidence for late Miocene deformation on the Columbia Plateau comes from the Yakima and Umtanum anticlinal structures. Bentley (1977, p. 339) describes "locally substantial deformation...14 to 12 m.y.b.p." 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 (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 by Rockwell, (1979, p. III-166) from the Filey Road area of eastern Umtanum Ridge. Here, 2 miles 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 and 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 1200 ft. of structural relief has developed at the eastern end of Umtanum 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.b.p. of the widespread tuff member of the Rattlesnake Volcanics (Thayer and Brown, 1966; Robyn and others, 1977; Robyn, 1977). Robyn (1977) assigns an age of 10 to 7 m.y.b.p. for this compressional event. Renewed folding and rupturing of the north flank of the Aldrich Mountain anticline generated the east-west trending John Day 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.b.p. (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.b.p. age of the member. Rockwell (1979, p. IV-17, IV-20, 21) conclude that most deformation in the Pasco Basin area occurred between 10.5 and approximately 5. m.y.b.p., although the younger age limit is not well-controlled.

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; accordingly, there is a problem in establishing an upper limit for most plateau folding and associated faulting because of the incomplete Pliocene and Quaternary stratigraphic record of the plateau area.

Pliocene stratigraphic units (ca. 5 to 1.8 m.y.b.p. 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). Geologists working on the plateau have differed on the duration 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 Ringold Formation (5.1 - 3.3 m.y.b.p., Rockwell, 1979) are very gentle (Figures 2.5-36 and 2.5-37). This relation may indicate waning deformation (Rockwell, 1979, IV-21) since lower Ringold strata on the flanks of Saddle Mountain to the north dip as steeply as 40° (Washington Public Power Supply System, 1977h, 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 lower and middle Ringold sediments that lie across the buried southeastern end of the Gable Mountain anticline 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 into 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, 1977h) 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". Elsewhere Bentley (1977, p. 352) states 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 track and K-Ar ages of 3.7 to 4.8 m.y.b.p. (Waitt, 1979). From such relations, Bentley (1977, p. 339) derives the conclusion that "the majority of the 'ridges' rose in Pliocene-Pleistocene times" (6 to 1.5 m.y.b.p.).

Thoms and others (in Washington Public Power Supply System, 1977h, p. 2RH 8-3), however, although favoring local deposition of upper Ellensburg deposits "during or after" ridge uplift, take a different 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 the west end of Smyrna Bench (Saddle Mountain) and in the Kittitas Valley do small dip slip faults cut these deposits".

Elsewhere, (Washington Public Power Supply System, 1977h, p. 2RH 8-12) Thoms and others conclude 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, questioned 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 to 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 deformation 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.b.p. (Shannon and Wilson, 1973d). Shannon and Wilson (1973c) 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).

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 $\pm$ 0.1 m.y.b.p.) lies in an

erosional reentrant across the truncated northern flank of the anticline (Rockwell, 1979, p. II-77). Bentley (1977a, p. 354) concluded that most of the deformation along nearby Manastash Ridge occurred prior to the development of the one million year old Thrall pediment surface on its north flank.

Fault displacements of Quaternary age, unknown in 1977 at the time of submittal of Amendment 23 to the WNP 1/4 PSAR, have since been recognized in three areas of the Plateau: 1) Toppenish Ridge, approximately 53 miles west of the WNP 1-2-4 site (Campbell and Bentley, 1980); 2) about 28 miles southeast of the site, from the vicinity of Wallula Gap on the Columbia River southeastward to the Walla Walla/Milton-Freewater area (Shannon and Wilson, 1980); and 3) approximately 11 miles north of the site on the eastern end of Gable Mountain.

2.5.1.2.2 Stratigraphy and Lithology

The stratigraphy of the site region is shown in Figure 2.5-10. The nomenclature is taken from Rockwell (1979). The rocks in the site region (Figure 2.5-4) consist predominantly of basalt flows but include some interflow sediments, and a sequence of fluvial, lacustrine, and eolian sediments overlying the basalts. The following is a brief lithologic description of the principal stratigraphic units. For more detailed discussion, refer to Rockwell (1979).

Volcanic Rocks

The Yakima Basalt subgroup (Miocene) is the principal exposed rock unit in the Columbia Plateau (Figure 2.5-2). It consists of individual flows ranging in thickness from 50 to 150 feet. The older Picture Gorge Basalt unconformably underlies the Yakima basalts in northeastern Oregon. Some Yakima basalt flows are extensive, one covering an estimated 20,000 square miles (Bingham and Grolier, 1966). The maximum thickness of the Columbia River Group is unknown. Only the upper 2,000 feet of Yakima Basalts have been studied in detail (Bingham and Grolier, 1966). An exploratory well near Odessa, 100 miles north of Richland, penetrated 4,500 feet of basalt. Raymond and Tillson (1968) reported 10,000 feet of volcanics in the Rattlesnake Hills well on Rattlesnake Mountain, most of which may not be Columbia River basalts.

The general character of the Yakima Basalt is dark gray to black, very dense, fine-grained, and commonly vesicular or scoreaceous at the bottom and top of individual flows. Some

flows have physically distinctive features, but generally these features do not extend laterally for great distances. The homogeneous appearance of the individual flows and the limited areal extent of some have made correlations difficult and sometimes tenuous. However, correlation based on geophysical logging and chemical composition appears to have successfully delineated some upper flow members in the Pasco Basin.

Columbia River Basalt Group

This section contains a brief description of the five formations generally recognized within the Columbia River Basalt Group (Swanson and others, 1979b).

Imnaha Basalt and Picture Gorge Basalt

The two basal formations of the Columbia River Basalt Group, as defined by Swanson and others (1979b), are the Imnaha Basalt and Picture Gorge Basalt (Figure 2.5-10). Flows of Picture Gorge Basalt are confined to the vicinity of the John Day Basin in northcentral Oregon and were extruded from feeder dikes comprising the Monument swarm (Waters, 1961; Wilcox and Fisher, 1966; Fruchter and Baldwin, 1975). Outcrops of Imnaha Basalt are confined to southeastern Washington, northeastern Oregon, and adjacent Idaho, where known source dikes are exposed (Taubeneck, 1970; Kleck, 1976).

Flows of Imnaha Basalt are generally coarse grained and phyric. In the vicinity of the type locality (Taubeneck, 1970; Hooper, 1974; Kleck, 1976; Swanson and others, 1979b), Imnaha Basalt has a thickness of about 1,500 feet and is composed predominantly of flows of normal polarity.

Grande Ronde Basalt

The Grande Ronde Basalt (Figure 2.5-10), the oldest formation of the Yakima Basalt Subgroup, was extruded between 14.5 and 15.4 million years age. Its type locality is a prominent west-trending spur ridge in the lower part of the Grande Ronde River valley, Asotin County, extreme southeastern Washington (Camp and others, 1978; Swanson and others, 1979b). The character of Grande Ronde Basalts at the type locality was discussed in detail by Camp (1976), Price (1977), Reidel (1978b), and Ross (1978).

The Grande Ronde Basalt is the most extensive unit of the Columbia River Basalt Group and underlies most of the Columbia Plateau. It is also the most voluminous unit,

comprising 85 percent of the basalt group (Swanson and Wright, 1978). Known feeder dikes for these basalts (Taubeneck, 1970; Price, 1977; Swanson and others, 1977) are exposed in the southeastern part of the plateau. Along the plateau margin, the formation laps onto an irregular erosional surface.

The known thickness of the Grande Ronde ranges from tens of feet along the margin of the plateau to more than 3,000 feet in the Rattlesnake Hills well (Raymond and Tillson, 1968; ARHCO, 1976). Within the plateau, the thickest exposures of Grande Ronde average 2,500 feet. Sections generally comprise 30 to 40 flows (Swanson and others, 1979b).

Most Grande Ronde flows are fine grained, aphyric, black, and dense. Rarely, plagioclase and clinopyroxene phenocrysts are visible in hand specimen, and microphenocrysts of those minerals are apparent in thin section.

To date, the only consistent "regional" subdivisions of the Grande Ronde are four magnetostratigraphic units defined on the basis of dominant magnetic polarity (Swanson and Wright, 1976). These units, in ascending order are: reversed-one (R_1), normal-one (N_1), reversed-two (R_2), and normal-two (N_2). This magnetic stratigraphy, first determined from detailed and reconnaissance mapping in southeastern Washington (Swanson and others, 1977), has been found to hold throughout the area.

The contact of the Grande Ronde with the underlying Imnaha Basalt is well exposed in the Tri-state area. The contact is conformable, and there is no evidence of a major time break (Bond, 1963; Holden, 1974; Camp, 1976; Price, 1977; Reidel, 1978b; Swanson and others, 1979b).

The contact between the Grande Ronde Basalt and the older Picture Gorge Basalt is exposed only in central Oregon. The relationship between these formations has been studied in detail by Nathan and Fruchter (1974). They have shown that upper flows of the Picture Gorge Basalts interfinger with lower Grande Ronde flows of the N_1 magnetic unit. Interfingering of Grande Ronde Basalt and Picture Gorge Basalts indicates that the formations are, in part, coeval. Consistent with this interpretation, K-Ar dates indicate an age of about 16 million years for each unit (Watkins and Baski, 1974).

The contact between the Grande Ronde and the younger Wanapum Basalt commonly is marked by a zone of weathering, or by a

thin sedimentary interbed (Mackin, 1961; ARHCO, 1976; Swanson and others, 1977).

Wanapum Basalt

The Wanapum Basalt (Figure 2.5-10) is the middle formation of the Yakima Basalt Subgroup (Swanson and others, 1979b). Its designated type locality is the nearly continuous exposure along the east side of the Columbia River near Wanapum Dam. The type Wanapum has been described by Mackin (1961) and in more detail by Myers (1973).

The Wanapum Basalt is the second most voluminous of the basalt formations, comprising about 15 percent of the Columbia River Basalt Group. Although its volume is considerably less than that of the Grande Ronde, the Wanapum is the most extensively exposed of the five formations and covers 80-90 percent of the central part of the plateau (Swanson and Wright, 1978).

The Wanapum Basalt consists mainly of medium grained, olivine-bearing, and slightly to moderately plagioclase-phyric flows. Known Wanapum feeder dikes are exposed in the southeastern part of the plateau (Price, 1977; Swanson and others, 1977).

The Wanapum currently is described in terms of four members: Eckler Mountain; Frenchman Springs; Roza; and Priest Rapids (listed oldest to youngest) (Swanson and others, 1979b). All but the Eckler Mountain Member cover large parts of the region.

Saddle Mountains Basalt

Saddle Mountains is the youngest formation in the Columbia River Basalt Group and comprises 10 recognized members (Figure 2.5-10). Saddle Mountains flows in the vicinity of the type locality, an anticlinal ridge located in south-central Washington, have been described by Mackin (1961), Bingham and Grolier (1966), and Schmincke (1967b).

The Saddle Mountains Basalts were extruded between 13.5 and 6 m.y. ago, predominantly from fissures in the eastern and east-central part of the plateau (Swanson and others, 1979b). Despite the 7 m.y. accumulation period of this formation, it comprises less than one percent of the volume of the Columbia River Basalt Group. Saddle Mountains time was marked not only by waning volcanism, but by the development of thick, local, sedimentary deposits between flows, and by folding and canyon cutting (Swanson and

Wright, 1978). Consequently, the contact between Saddle Mountains Basalt flows and older basalt units is generally a regional disconformity with local angular and erosional unconformities in areas of earlier ridge or basin development.

The Saddle Mountains have been subdivided into 10 members, listed from oldest to youngest as: Umatilla, Wilbur Creek, Asotin, Weissenfels Ridge, Esquatzel, Pomona, Elephant Mountain, Ward Gap, Buford, Ice Harbor, and Lower Monumental (Swanson and others, 1979b). Members of Saddle Mountains Basalt display the greatest petrographic, chemical, and paleomagnetic variability of any formation of the Columbia River Basalt Group.

Ellensberg Formation

The thickness and areal extent of basalt interbeds within the Columbia River Group are variable. Interbeds, which are more common in the upper portion of the group, range in thickness from 5 to 130 feet and are generally composed of tuffaceous sand, clay, or silt. Some interbeds probably represent lacustrine environments during interflow periods. Other interbeds are characterized by highly altered and weathered basalts displaying poorly developed soil profiles which reflect periods of weathering and landscape stability between succeeding flows. The Ellensburg Formation underlies and intertongues with the Yakima Basalts (Figure 2.5-10) (Bentley in Washington Public Power Supply System, 1977d).

The Ellensburg contains material derived from three distinct sources: volcaniclastic material from the Cascade volcanic range; plutonic, metamorphic, and older sedimentary material derived from the northern Cascade Range and adjacent areas north of the plateau; and basaltic material derived from within the plateau. Volcaniclastic materials are dominant in the western lithofacies and the middle and upper part of the section. Plutonic and metamorphic arkosic materials are most common in the northern and central lithofacies, and dominate the lithology of the lower part of the section. Basaltic materials dominate the lithology of the uppermost Ellensburg, and occur in middle Ellensburg units adjacent to basalt ridges. Heavy mineral suites (Coombs, 1941) and current directions indicate a complex south and southeast drainage system involving four major system changes through time. In early Ellensburg time, marginal streams moved materials southward along the western edge of the plateau. In middle Ellensburg time, the ancestral Yakima and Columbia Rivers appear to have

developed on the plateau and moved materials southward and southeastward across the plateau forming clastic wedges along their courses. In late Ellensburg time, the rising Cascades Range and Yakima Ridges trapped large volumes of locally derived materials in subsiding basins. The two later changes probably are related to continuing uplift.

Ellensburg lithologies include conglomeratic sandstone, siltstone, claystone, laharic breccia, airfall tuff, loess, and colluvial breccia. There is no evidence of a single large fan of Ellensburg materials extending over the western margin of the Columbia Plateau.

The lowest Ellensburg probably is about 17 m.y. old based on K-Ar dating of lavas. The middle Ellensburg ranges in age from about 13 to 6 m.y.b.p. The age of the youngest Ellensburg is Pliocene and possibly early Pleistocene. Fission-track dating methods yield ages of 4 to 2 m.y.b.p. (Waitt, 1979a). The younger Ellensburg is correlative with Ringold.

The Lower Ellensburg

The lower Ellensburg consists of all sedimentary material intertongued with the Wanapum Basalt (middle Yakima) and Grande Ronde Basalt (lower Yakima) and conformably underlying the Grande Ronde Basalt. Three members are well known from the work of Mackin (1961): Quincy Member, a diatomite and claystone; Squaw Creek Member, a diatomite; and Vantage Member, a coarse lithic and tuffaceous - quartz mica sandstone with interbeds of tuffaceous claystone. These members are intertongued or underlie the Wanapum Basalt. Other similar interbeds occur at several levels within the Grande Ronde in the margin area of the Grande Ronde. The dominant lithologies in the lower Ellensburg are volcanic and tuffaceous siltstones. On the plateau margin, hyaloclastites and peperite breccias are common.

The Middle Ellensburg

The middle Ellensburg is a sedimentary sequence intertongued with the members of Saddle Mountains Basalt (Upper Yakima). It includes the Rattlesnake Ridge, Selah, Beverly, and Mabton members (Schmincke, 1967a).

The middle Ellensberg consists of five lithofacies: volcaniclastic facies in the western basins; Columbia River facies near the ancient Columbia River; Yakima River facies

near the ancient Yakima River; local sidestream and slope facies near rising ridges; and interstream or flood plain facies away from major stream channels.

The Upper Ellensburg

The upper Ellensburg consists of a sedimentary sequence deposited in local basins across the western plateau. It is characterized by basalt detritus derived locally from within the plateau as the Yakima Ridges reached a stage of maximum development. Each basin in the western Columbia Plateau has a different local stratigraphy with different lithofacies distributions which depend on stream size (mainstreams vs. sidestreams), local slope conditions, relative tectonic activity, and underlying rock types. Intrabasin stratigraphy has been partially established in the Kittitas, Ahtanum, and Pasco Basins.

Airfall tuffs (tephras) derived from the Cascade Range are interlayered with all facies in all basins and probably offer the best method of intra and interbasin correlation. In Kittitas Valley, the Thorp Gravel contains tuffs dated by the fission-track and K-Ar method at 3.7 to 4.8 m.y.b.p. (Porter, 1976; Waitt, 1979).

Ringold Formation

In much of the Pasco Basin the Columbia River Basalt is overlain by Pliocene sediments (Figure 2.5-10) of the Ringold Formation. These deposits, found mainly in the Pasco Basin, range in thickness from a few feet to more than 700 feet, based on surface outcrops along the White Bluffs east of the WNP 1-2-4 site.

Basal and Lower Members

In the Pasco Basin a basal sequence of sandy gravels lies below "blue clays" and above the basalts. The gravels reach a thickness of 125 feet in some places, thickening near the west-central part of the basin and thinning to the margins (Rockwell, 1979). Brown and Brown (1961) suggest that these gravels are not equivalent to the Ringold but may be stratigraphically related to the lower-to-middle Pliocene Ellensburg (Figure 2.5-10) found in the Yakima Basin to the west.

The lower member of the Ringold is a "blue-gray" to greenish, sandy and clayey silt. Similar blue clays reach a

thickness of 400 feet in the Pasco Basin subsurface. The silt is typical of deposition in a lake environment (Newcomb, 1958).

Middle Member

Overlying the "blue clays" are beds of sandy gravel and coarse sand. This member is approximately 340 to 400 feet thick and probably represents old gravel-train or channel-fill of the ancestral Columbia River (Newcomb, 1958). It is exposed along the lower portion of the White Bluffs. Similar gravels overlie the "blue clays" in the Pasco Basin subsurface.

Upper Member

Above the middle member are light-colored silts and fine-grained sands that form the crest of the White Bluffs. Approximately 500 feet of the upper member have been mapped in the White Bluffs (Newcomb, 1958). These beds probably were deposited in a lacustrine environment.

In the subsurface of the Pasco Basin, sand and silt similar to that at White Bluffs has been encountered. It appears only as remnants up to 145 feet thick in the buried topographic-low areas along Cold Creek Valley. Because of its local occurrence, correlation with the upper member of the White Bluffs has been difficult (Brown and Brown, 1961).

Recent paleontologic evidence (Beck, 1956; and Gustafson, 1973, 1976) suggests a mid to late Pliocene age for the upper Ringold. Interpretation of paleomagnetic data by Packer and others (1979) suggests an age of between 3.3 and 5.1 m.y.b.p.

Simcoe Lavas

The Simcoe Lavas or Simcoe Mountains Volcanics cover approximately 300 square miles, mostly to the northwest of Goldendale, Washington. These lavas unconformably overlie the Yakima Basalt and Ellensburg Formation and predate the Pleistocene and Holocene lavas from Mount Adams and its subsidiary volcanoes. K-Ar dates of Simcoe Lavas indicate extrusion between 4.5 and 0.9 m.y. ago (Shannon and Wilson, 1973d).

Simcoe Lavas are predominantly olivine basalt, with minor amounts of pyroxene-olivine basalt, rhyolite, and dacite. Most olivine basalt occurs as thin flows erupted from northwest trending fissures or from the flanks of numerous

small, shield volcanoes and spatter cones in the area. Rhyolite, dacite, and pyroxene-olivine basalts occur in a few small flows and in cinder and spatter cones (Shannon and Wilson, 1973d).

Tieton Andesite

The Tieton Andesite consists of several flows of hypersthene-augite andesite which were erupted from vents in the Cascade Range (Swanson, 1967). Only one flow extends into the region. This is an intracanyon flow restricted to the valleys of Cowiche Creek and the Tieton and Naches Rivers (Rigby and Othberg, 1979; Swanson and others, 1979a). A K-Ar date obtained for the Tieton Andesite by Dymond (in Shannon and Wilson, 1973d) indicates that this lava was extruded about one million years ago.

Older Eolian Deposits (including Palouse Formation)

The older eolian deposits can be divided into three units. The two oldest loess units have been correlated with pre-Bull Lake glaciation (pre-Illinoian, older than 125,000 years) (Pierce and others, 1976; Richmond and others, 1965). These units are dark reddish-orange and are best exposed in deep "cuts" in the Palouse Hills area. The next higher loess has been defined as Bull Lake (Illinoian) in age (Richmond and others, 1965). It comprises most of the Palouse Hills and has been designated as the Palouse Formation. The eolian origin of the Palouse is demonstrated by the predominant northeast alignment of the Palouse Hills, parallel to the direction of the prevailing winds, the "fining" of component particles northeastward, and an east-northeast progressive overlap of loess sub-units (Rigby and Othberg, 1979). The Palouse is tan to brown, very fine grained (silt to clay size), and up to 250 feet thick.

Glaciofluvial Sediments

In the Pasco Basin, glaciofluvial sediments of late Pleistocene age, including glacial outwash, catastrophic flood gravels, and slackwater deposits (Figure 2.5-10) were deposited on topography developed on the basalt and the Ringold Formation.

Glacial till, drift, and outwash deposits are present in the northern and westernmost portions of the Columbia Plateau, and occur most extensively in the northwestern part which was overridden by the Okanogan Lobe of the Cordilleran Ice Sheet. The terminal position of this late Wisconsinan (15,000 to 20,000 y.b.p.) ice sheet is marked by the Withrow

moraine (Easterbrook, 1976a). The area is characterized by eskers, ground moraine deposits, and large basalt erratics ("haystack" rocks). Glacial drift, a product of Wisconsinan alpine glaciation, occurs in the Kittitas, Wenatchee, and Chelan Valleys (Rigby and Othberg, 1979).

Catastrophic flood deposits are widespread in the Columbia Plateau. These formed when ice dams in western Montana and northern Idaho were breached and massive volumes of water spilled abruptly across eastern and central Washington. The timing and number of the floods remain unknown (Bretz and others, 1956; Bretz, 1959; Patton and Baker, 1978; Waitt, 1979a). Most of the flood deposits are late Pleistocene in age, and the last major flood sequence is dated at about 12,500 y.b.p. (Mullineaux and others, 1977). Older flood gravels in the Pasco Basin were dated at approximately 200,000 y.b.p. using the thorium/uranium method (Tallman and others, 1978). Within the region, catastrophic flood deposits are divisible into two main facies -- flood gravel facies (including the Pasco gravels) and slackwater facies (Touchet Beds).

Flood Gravel Facies (Pasco Gravels)

This facies consists of poorly sorted, sub-rounded to angular clasts and has foreset bedding locally. Distribution is limited to: (a) scabland channels, (b) depositional basins into which these channels emptied (primarily the Quincy and Pasco Basins), (c) Columbia River Gorge below Wallula Gap, and (d) Snake River Canyon (as far up as Lewiston, Idaho). Within the scabland channels and along the Columbia River, this facies occurs principally as flood bar deposits (Baker, 1973; Rigby and Othberg, 1979). The grain size of the gravels ranges from boulders to medium- to fine-grained sands. Clasts of basalt, granite, diorite, and porphyritic volcanic rock are dominant; clasts of gneiss, schist, metasedimentary rock, and chert-pebble conglomerate are subordinate.

Deposits of sand and gravel derived from the Pleistocene-age Lake Bonneville floods also are included in this unit. These were transported to the plateau via the Snake River drainage, and presently are thought to be confined to the Snake River Canyon (Rigby and Othberg, 1979).

Slackwater Facies (Touchet Beds)

This facies comprises the rhythmically bedded and graded silt, sand, and minor gravel deposits originally referred to by Flint (1938a) as the Touchet Beds (Figure 2.5-10). In

general, the slackwater facies is limited to peripheral areas of the Pasco and Quincy basins, and to tributary valleys such as the Touchet and Walla Walla (Rigby and Othberg, 1979). Slackwater conditions occurred at those localities during impoundment of floodwaters due to postulated hydraulic damming at Wallula Gap. Locally, slackwater deposits contain ice-rafted erratics, ash beds, clastic dikes, and minor gravel-filled channels. At several localities, slackwater deposits approach a thickness of 300 feet (Rigby and Othberg, 1979). At Burlingame ditch in the Walla Walla Valley, a thick sequence of 40 separate rhythmite cycles is exposed. Waitt (1979b) proposed that each cycle records a catastrophic flood event. Also included within this unit are minor deposits in the Snake River Canyon thought to be associated with Lake Bonneville catastrophic flood(s) (Rigby and Othberg, 1979).

Clastic Dikes

Clastic dikes are common in the Touchet Beds but less abundant in the Pasco Gravel, the Ringold Formation, and the uppermost basalts. They do not appear in materials younger than the Pleistocene glaciofluvial deposits. The dikes range in width from a fraction of an inch to about four feet and generally consist of vertically-laminated, fine-grained sand and silt. In some places, the dikes describe polygonal networks (in plain view) with cell diameters of 50 to 400 feet (Jones and Deacon, 1966; Newcomb, 1962; Black, 1979).

Several mechanisms have been suggested to explain emplacement of the clastic dikes in the Pasco Basin. Lupher (1944) believes that the dikes formed when pre-existing fractures were filled by glaciolacustrine sediment and that "the clastic dikes were formed only under conditions prevailing during glacial stages and are restricted to proglacial deposits." Newcomb (1962) postulates that the dikes were formed by upward injection of groundwater during sudden lowerings of the impounded lake (Lake Lewis) and concommitant releases of hydrostatic confining pressure.

It appears from these suggested mechanisms that formation of the dikes was limited to conditions extant only in Pleistocene times. The principal condition was repeated inundation of the Pasco Basin by Lake Lewis with a depth of 1,000 feet or more. The lowering of this lake caused groundwater responses which, in certain conditions, could have formed the clastic dikes and their particular features.

Ash Deposits

The most widespread ash in the Columbia Basin is a thin, but widespread and locally continuous unit in late-glacial lacustrine sediments. It occurs in both the Touchet beds and Pasco gravels throughout the Pasco Basin. Commonly this ash occurs as two subparallel units separated by as much as 2 inches of lacustrine sediment. Although, to date, neither the source nor the absolute age of these ash layers has been definitely established, studies by the U.S. Geological Survey (R. E. Wilcox, personal communication, 1978) indicate that the ash is petrographically distinct from the Glacier Peak ash, and that it more closely resembles the cummingtonite-bearing ash of tephra set S from Mount St. Helens. The most voluminous layers of tephra set S are about 13,000 years old (Mullineaux and others, 1975).

Another ash occurs interbedded with Holocene eolian sediments along the margin of the Pasco Basin. This ash layer locally has been correlated with the catastrophic eruption of Mount Mazama in Oregon about 6,600 years ago (Fryxell, 1965).

Ash from the Glacier Peak volcano has not been positively identified in the Pasco Basin, but it has been identified in other parts of the Columbia Basin to the north and east of the site. (Powers and Wilcox, 1964; Fryxell, 1965). Radiocarbon dates indicate that the Glacier Peak ash was erupted about 12,500 years ago (Fryxell, 1965).

Alluvium and Terrace Deposits

Holocene alluvium deposited by the Columbia and Yakima Rivers occurs along the present channel courses, but it is often difficult to distinguish from the glaciofluvial deposits. Fan deposits exist along the rim of the Pasco Basin and may be as much as 50 feet thick.

Fanglomerates

This unit occurs as small isolated outcrops of cemented basalt gravel on the toe of the Horse Heaven Hills near Yellepit (Washington Public Power Supply System, 1977). The fanglomerate, up to 25 feet thick, consists mainly of rounded basalt clasts having rinds between 1/4 and 3/4 inch thick. Gravels of the Kennewick fanglomerate are compositionally distinct as compared to gravel of glaciofluvial deposits in the Pasco Basin. The latter contain up to 60 percent of non-basaltic clasts, and commonly have thinner weathering rinds or none at all.

Caliche rinds on Kennewick fanglomerate cobbles have been dated at approximately 55,000 years (Woodward-Clyde Consultants, 1978). However, this Uranium-thorium technique is being reevaluated.

Talus, Slopewash & Landslike Deposits

Talus and slopewash occur along the bases of slopes around and within the Pasco Basin. These are deposits of sand, gravel, and debris developed through processes of creep, rockfall, and slopewash.

Landslides as large as two square miles in area exist in the Ringold Formation along the White Bluffs, north and east of the project site (Bingham and Grolier, 1966). The major sliding has occurred along interlayered pyroclastic or sedimentary materials of low shear strengths, caused by saturation by subsurface water and oversteepening of slopes. Surficial failures of slope debris occur along the steep flanks of the anticlinal ridges in the region.

Eolian Deposits

Loess (wind-deposited fine-grained sand and silt) mantles much of the Columbia Plateau. Loess deposits of at least four different ages have been described (Richmond and others, 1965). For discussion of three older eolian deposits, see discussion above.

The youngest recognized loess is a light-tan unit variously interpreted as Pinedale (late Wisconsinan, older than 10,000 years) (Pierce and others, 1976) and as recent in age (Richmond and others, 1965). In general, this loess is capped by a weakly developed petrocalcic layer and is separated from the Palouse by an unconformity. There is evidence to suggest that this loess is essentially a post-catastrophic-flood loess (younger than 13,000 years) (Rigby and Othberg, 1979).

2.5.1.2.3 Columbia Plateau Structural Geology

2.5.1.2.3.1 Badger Mountain, Beezley Hills, Moses Stool Folds

The Badger Mountain-Beezley Hills is a broad, asymmetrical, anticlinal structure in the northwest Columbia Plateau east of Wenatchee. This structure plunges southeastward from the Columbia River at about 100 feet per mile (Figure 2.5-3).

The northern limb dips steeply into a broad downwarp known as the Waterville syncline, whereas the southern limb typically dips less than about five degrees.

The main development of the Badger Mountain structure was clearly a post-Grande Ronde event, as indicated by deformation of the Vantage Member (Washington Public Power Supply System, 1977e). Although a minimum age for the structure has not been determined with certainty, no evidence was found to indicate faulting of late Cenozoic sediments which mantle much of the anticline (Washington Public Power Supply System, 1977e; Rigby and Othberg, 1979).

The Moses Stool anticline is en echelon with Badger Mountain and merges with the Beezley Hills east of Moses Coulee. West of Moses Coulee, the Moses Stool anticline merges with Badger Mountain (Swanson and others, 1979a). At the approximate location of the juncture of the Beezley Hills and Moses Stool anticlines, the Beezley Hills curves north and becomes the Coulee monocline, which extends a distance of approximately 30 miles to the northeast. The terminus of the monoclinal axis of Beezley Hills coincides with a thrust fault developed in Wanapum Basalt. Southward, a second thrust fault justaposes Grande Ronde Basalt above Wanapum Basalt (Swanson and others, 1979a).

2.5.1.2.3.2 Kittitas Valley

The Kittitas Valley (Figures 2.5-11, 2.5-3, and 2.5-12) is a topographic and structural low along the northwestern margin of the Columbia Plateau. It is bounded on the west by the Cascade Mountains, on the north by the Wenatchee Mountains uplift, on the east by the Naneum-Hog Ranch anticline, and on the south by the Manastash-Thrall structure. The geology of the area has been the subject of investigations by Bentley (1977, and in Washington Public Power Supply System, 1977d), Tabor and others (1977), Martin (in Washington Public Power Supply System,1977j), Waitt (1979b), and Hanson (in Rigby and Othberg, 1979).

Grande Ronde flows and Ellensberg strata, having dips of 10° to 50°, crop out on the flanks of Kittitas Valley. These same units have dips of about 5° near the middle of the valley (Waitt, 1979a). Rocks exposed within the valley also include pre-Tertiary gneiss, early Tertiary volcanics, Thorp Gravel, and glacial drift (Rigby and Othberg, 1979).

Most faults within the Kittitas Valley appear older than the late Cenozoic sediments of the area. However, Martin (in Washington Public Power Supply System, 1977j) and Waitt

(1979b) described three east-striking faults in the valley which they interpreted as younger than Thorp Gravel (less than 3.7 m.y.b.p.) and older than Kittitas Drift (120,000 y.b.p.). These faults were later examined by Rigby (Rigby and Othberg, 1979), who concluded that evidence for two of the faults is tenuous.

2.5.1.2.3.3 Frenchman Hills

This east-trending anticlinal fold extends from the eastern edge of Kittitas Valley (Figure 2R G-15 of Washington Public Power Supply System, 1977o) on the west to the Potholes Reservoir on the east. East of the Columbia River the Frenchman Hills anticline is subdued and symmetrical.

2.5.1.2.3.4 Manastash Ridge - Hanson Creek

The Manastash Ridge-Hanson Creek anticlines are structures extending from east of the Columbia River near Priest Rapids Dam to the Ellensberg area (Figures, 2.5-12 and 2.5-13). Manastash ridge separates Kittitas Valley to the north from the Squaw Creek syncline to the south.

In the vicinity of Yakima River, (Figure 2.5-12 and 2.5-13), the Manastash structure is a broad, low, asymmetrical anticline having steeper dips on the north limb. Two second-order folds, the Thrall syncline and the Thrall anticline, are superimposed on the north flank of the structure. Axes of the Manastash anticline and the Thrall anticline are parallel.

Most faults associated with the Manastash structure are near vertical and located along the steep north limb. The most continuous of these may be part of a longer fault system, termed the Manastash-Hanson Creek fault system (Bentley, 1977). One north-south cross fault displaces the structure at Badger Gap (Shannon and Wilson, 1978; Swanson and others, 1979a). On the basis of remote sensing imagery, this feature cannot be traced westward into Kittitas Valley or eastward across Manastash Ridge (Rockwell, 1979). Bentley (1977a) concluded that most deformation on Manastash Ridge occurred after sediments interpreted as Thorp Gravel were deposited, and before development of the Thrall pediment surface.

2.5.1.2.3.5 Yakima Ridge

Yakima Ridge extends from west of the city of Yakima (Figures 2.5-15 and 2.5-16) to the western border of the

Hanford Reservation. Yakima Ridge is divided into western, middle, and eastern segments on the basis of topography (Washington Public Power Supply System, 1977k).

The western segment of Yakima Ridge (Figure 2.5-15) extends from Sedge Ridge eastward along Cowiche Mountain to Selah Gap. The ridge in this area is a box-fold with sharp hinge lines along both flanks. The minimum age of deformation on Yakima Ridge is established by the Tieton andesite which onlaps the eroded north flank of the anticline along Cowiche Creek. K-Ar determinations by J. Dymand (in Washington Public Power Supply System, 1977k) indicate an age of 1.0 and ± 0.1 m.y.b.p. for the andesite.

The middle segment of Yakima Ridge (Figure 2.5-16) extends eastward from Selah Gap to Cairn Hope Peak. This segment is a box-fold anticline with sharp hinge lines (Waters, 1955; Bentley, 1977). Faults are present locally in the hinge area and core of the anticline.

The eastern segment of the Yakima Ridge structure (Figure 2.5-16) has been studied by Rockwell (1979), Bond and others (1978), and Kienle (in Washington Public Power Supply System, 1977k). The dominant fold within the eastern segment of the Yakima Ridge is the N70-75°W-trending, southeastward plunging Cairn Hope Peak anticline. The shorter and steeper north limb of this anticline dips 30° to 40° north, and its wider southern limb dips 10° to 15° south. The southern limb of the anticline contains two monoclines. The northernmost monocline, the Cairn Hope Peak monocline, trends N60°W and is interpreted (Bond and others, 1978) as merging with the Silver Dollar Fault of Goff and Myers (1978). The southern monocline (unnamed) trends N70°E.

The Silver Dollar fault (Figure 2.5-16) reportedly offsets the Frenchman Springs against Umatilla and Pomona across a fault breccia zone 165 to 230 feet wide (Goff and Myers, 1978). Field reconnaissance by Shannon and Wilson (1980) confirmed this fault west of the Yakima-Benton county line. Based on stratigraphic displacements, Shannon and Wilson (1980) interpreted the fault as a scissor fault having a hinge to the east and increasing displacement westward.

The Yakima Ridge structure is separated from Umtanum Ridge by the Cold Creek syncline which plunges 3° to 5° east-southeastward into the central Pasco Basin.

Based on his geologic mapping of Yakima Ridge, Bentley suggested at least two periods of deformation (Bentley,

1977b). The first period involved uplift along a north-trending axis during post-Priest Rapids and pre-Pomona time (12 to 14 m.y.b.p.). The second period involved uplift along N50°-80°W-trending axes in post-Elephant Mountain time (about 10.5 m.y.b.p.). Uplift as indicated by undisturbed Tieton Andesite along Cowiche Creek on the west end of Yakima Ridge ceased by 1 m.y.b.p. (Washington Public Power Supply System, 1977k).

On the basis of remote sensing data, the slopes of Yakima Ridge appear more subdued and symmetrical than the slopes of the anticlines farther south. Landslides appear to cover the flanks.

2.5.1.2.3.6 Ahtanum Ridge

Ahtanum Ridge (Figure 2.5-16), an east-trending anticline, is essentially the western extension of the Rattlesnake Hills. In general, the structure is asymmetrical, having steeply dipping and overturned basalt flows and thrust faults exposed on both flanks (Bentley, in Swanson and others, 1979a).

Late Cenozoic sediments on Ahtanum Ridge were mapped by Campbell (in Rigby and Othberg, 1979). He found no evidence of faulting in recent loess and stream alluvium overlying the surface of the ridge. Faulting of Ellensburg sediments and cemented basalt gravels was observed at two points on the lower slopes of Ahtanum's north flank. The age of deformation of these older sediments post-dates deposition of the Ellensburg strata and pre-dates deposition of the loess and alluvium. Farther west, the most recent age of deformation appears to be limited by undeformed Tieton Andesite (1 m.y.b.p.) (Shannon & Wilson, 1978).

2.5.1.2.3.7 Toppenish Ridge

Toppenish Ridge is an east-trending anticline. It has been traced on aerial photos from State Highway 22 to the Klickitat River. This anticline has been traced on aerial photos under Quaternary volcanics on the west that exhibit no evidence of displacement (Washington Public Power Supply System, 1977o). Toppenish Ridge has recently been mapped by Campbell (in Rigby and Othberg, 1979) and by Bentley (in Swanson and Others, 1979a).

Because of restricted access to this area, Washington Public Power Supply System has been unable to conduct field studies

and has been limited to imagery analyses. Therefore, the work of Campbell and Bentley, who did have ground access, is presented here.

Columbia River basalt flows exposed along Toppenish Ridge range from R_2 Grande Ronde through Elephant Mountain. The east end of the ridge consists of an anticline in Wanapum and Saddle Mountains Basalts. A small syncline near the crest of the ridge produces a double fold in the basic structure. Two thrust faults also are present on this part of the ridge. Westward, a high-angle reverse fault is present on the south side of the ridge. Additionally, an anticline-syncline fold system is developed on the north side of the ridge in Saddle Mountains Basalt. Farther westward, N_2 and R_2 Grande Ronde Basalt flows underlie much of Toppenish Ridge and are disrupted by a low-angle thrust fault. The lower anticline-syncline dies out in this area.

During geologic reconnaissance mapping by Campbell and Bentley, it was discovered that sediments mantling Toppenish Ridge had been disrupted in the vicinity of the Tecumseh-Pumphouse Road Junction (T9N, R19E) (Campbell, in Rigby and Othberg, 1979; Bentley, in Swanson and Others, 1979a). Late Cenozoic sediments consisting of upper Ellensburg units (defined as post-Elephant Mountain in age), Touchet-like beds (dated elsewhere as 13,000 y.b.p.), landslides of several ages, loess, and alluvial fans of several ages are also deformed. Surface mapping of these features suggests that they may be fault scarps. Other theories of origin, such as mass landsliding or down-slope creep are difficult to justify on the basis of current field data. The scarps appear to be Holocene in age (younger than Touchet-like sediments) and appear to have developed independently of the older thrust faulting along the ridge. Volcanic ash, organic material, and caliche samples from Toppenish Ridge have been submitted for dating.

Toppenish Ridge is structurally and topographically similar to several other nearby ridges--Horse Heaven, Ahtanum, Rattlesnake, and Manastash. After discovery of the scarps on Toppenish Ridge, other ridges were examined for similar scarps that might have been overlooked earlier. Although high-angle faults were observed to be associated with the auxiliary folding on the northern flanks of these ridges, scarps similar to those on Toppenish Ridge were not found (Campbell, in Rigby and Othberg, 1979).

Although surface rupture on and adjacent to Toppenish Ridge may represent Quaternary tectonic activity, as suggested by

Campbell and Bentley, an alternative explanation by Dr. G.A.Davis has been proposed. Davis (Appendix 2.5-N) states that:

> "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 2.5N-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. 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."

2.5.1.2.3.8 Horse Heaven Hills

The Horse Heaven Hills is essentially an anticlinal structure extending from east of Wallula Gap to and beyond the Quaternary basalt terrane north of The Dalles to the White Salmon River. The structure is asymmetrical with gentle dips on the south limb and a steep northern escarpment. The overall structure can be divided into western, middle, and eastern segments.

The western segment of the ridge contains a northeast-trending fault downthrown to the north (Swanson and others, 1979a) and appears to be displaced along several northwest-trending cross-faults.

The middle segment is a broad anticlinal arch with associated thrust faults (Anderson in Swanson and others, 1979a). Between Bickleton and Chandler, the middle segment of the structure is dominated by a series of northeast-trending anticlines. Near the crest of the Horse Heaven Hills, dip slopes and gentle folds underlie the intervening flat terrain (Gardner, in Swanson, 1978; Gardner in Swanson and others, 1979a). Thrust faults and high-angle faults associated with these folds cut the youngest (Elephant Mountain) basalt unit present.

Along the middle segment, between Prosser and Bickleton, Campbell (in Rigby and Othberg, 1979) observed a scarp developed in late Cenozoic material. On the southwest, the scarp terminates at a landslide, and it dies out northeastward in alluvium. Campbell suggested that this scarp represents a dip slope on a resistant unit within ancestral Columbia River gravels.

The eastern segment of the Horse Heaven Hills is a broad northwest-trending anticlinal ridge (Figure 18). Several faults, including the Zintel Canyon fault, are associated with this segment of the Horse Heaven Hills. The Zintel Canyon fault is interpreted as a high-angle, southward-dipping reverse fault (Rockwell, 1979). The fault appears to merge with the anticlinal fold to the east and to decrease in displacement westward.

2.5.1.2.3.9 Columbia Hills

The Columbia Hills anticline (Figure 2.5-3) extends from The Dalles, Oregon, to the Horse Heaven Hills near Wallula Gap. The structure comprises a series of asymmetrical, doubly-plunging anticlines. The amplitude of the structure decreases progressively eastward. The western half of this anticline is intersected by a series of northwest-trending thrust faults, and high-angle and low-angle faults, especially along the steep south limb of the structure (Swanson and others, 1979a).

Right lateral displacement along northwest-trending faults is suggested by the en echelon arrangement of small anticlines southeast of Camas Prairie (Swanson and others,

1979a). Vertical displacements for some northwest-trending structures are estimated to be on the order of tens of feet (Shannon and Wilson, 1973d).

Shannon and Wilson (1973d) examined the age of the northwest-trending fault system, and on the basis of K-Ar dating of Simcoe Lavas, concluded that the age of last movement on the structures was between 3.5 and 4.5 m.y.b.p. However, they also concluded that alignment of volcanic cones (some less than 100,000 years old) along the trend of these structures suggests that the fault system has continued to be a zone of crustal weakness providing conduits for late Cenozoic magmas.

2.5.1.2.4 Pasco Basin Structural Geology

The WNP 1-2-4 site lies within the Pasco Basin (Figure 2.5-4), one of several physiographic basins within the western Columbia Plateau. The Pasco Basin is partly surrounded by large anticlinal ridges, the Yakima Folds, developed during Miocene and Pliocene time. The Saddle Mountains form the northern boundary of the Pasco Basin; Umtanum Ridge and Yakima Ridge form the western boundary; Rattlesnake Hills and Horse Heaven Hills form the southern boundary; and a broad zone of gradually increasing westward dip (Figure 2.5-20) forms the eastern boundary. Umtanum Ridge and Yakima Ridge plunge eastward, decrease in relief, and die out within the basin interior.

Folds within the Pasco Basin trend eastward in the northwestern part of the basin and northwestward in the southeastern part. The anticlines generally correspond to narrow linear ridges, and the synclines to broad alluvium-filled valleys. The folds typically are asymmetrical, with the anticlines steeper to overturned on the north. Faulting is common along the ridges, particularly in areas where folding is most pronounced. Second-order folds are known to occur on both flanks of the principal structures.

2.5.1.2.4.1 Saddle Mountains

The Saddle Mountains structure (Figures 2.5-13 and 2.5-14) is an asymmetric anticlinal ridge having a faulted steep northern flank that extends from southeast of Ellensburg to east of Othello. Maximum structural relief is about 2,000 feet on the west and diminishes to less than 500 feet in the eastern basin.

The hinge area of the Saddle Mountains structure was mapped by Grolier and Bingham (1971), Washington Public Power Supply System, (1974), Taylor (1976), Washington Public Power Supply System, (1977h), and Reidel (1978a). The hinge area is subdivided on the basis of fold geometry into western, central, and eastern segments (Grolier and Bingham, 1978; Reidel, 1978b). Two major structural trends (west and northwest) and one minor trend (northeast) are present.

On the west side of Sentinel Gap (Figure 2.5-13), the Saddle Mountains fault is a complex zone containing several low-angle thrust faults. Neither fault displacements nor attitudes have been measured, but slickensides suggest low-angle, reverse movement. On the east end of the Saddle Mountains (Figure 2.5-14), faulting is suggested by a change in trend of the Saddle Mountains anticline. A change in attitude of the beds was interpreted by Glass and Slemmons (in Washington Public Power Supply System, 1977l) as a fault, but no indication of faulting was found during field examination.

In the western segment of the Saddle Mountains structure near Sentinel Gap (Figure 2.5-13), the anticline bifurcates into northern and southern axial segments.

West of the Columbia River (Figures 2.5-13 and 2.5-11) a brecciated zone along the flank of the Saddle Mountains can be traced for several miles. Immediately southeast of Doris, overturned basalts are brecciated and in contact with the Ellensburg Formation along a high-angle reverse fault (Washington Public Power Supply System, 1977d). At Foster Springs, where the Saddle Mountains become the Boylston Mountains, a 45° south-dipping fault places brecciated basalt over near-horizontal undisturbed columnar basalts. Farther west, the fault plane decreases in dip to become a near-horizontal thrust through the Boylston Mountains. The fault appears to retain this character into Kittitas Valley where it disappears beneath alluvium at Badger Pocket (Washington Public Power Supply System, 1977d).

Between Sentinel Gap and the western edge of Smyrna Bench (Figure 2.5-13), the Saddle Mountains anticline is asymmetrical. The north limb is near-vertical where exposed and the south limb dips less steeply (Grolier and Bingham, 1978).

Changes in trend of the axial trace and fold geometry on the side of Smyrna Bench (Figure 2.5-14) mark the boundary between the central and eastern segments of the Saddle Mountains structure. Style of folding changes from a tight

box-fold on the west to a more open eastward-plunging fold on the east. On the east side of Smyrna Bench, the fold becomes more open and subdued.

The Smyrna monocline is approximately parallel to the trend of the Saddle Mountains anticline south of Smyrna Bench and forms a northern hinge zone in the central portion of the anticlinal fold. A monocline similar to the Smyrna monocline extends along the north side of the Saddle Mountains from Corfu east to about where Washington State Highway 240 crosses the Saddle Mountains (Figure 2.5-14). Near Washington State Highway 240, the monocline merges with the Saddle Mountains anticline. This has been inferred to be the same monocline on both sides of the landslide block south of Corfu (Rockwell, 1979).

The most recent deformation of the Saddle Mountains apparently occurred after Ringold time but prior to development of the pediments on the flanks of the Saddle Mountains.

2.5.1.2.4.2 Umtanum Ridge

The Umtanum Ridge anticline (Figure 2.5-16) is a nearly continuous feature from the vicinity of Vernita Bridge northwestward to the margin of the Columbia Plateau. Umtanum Ridge is separated from the Hansen Creek anticline to the north by the Wahluke syncline and separated from the Yakima Ridge anticline to the south by the Cold Creek syncline. The distance from the WNP 1-2-4 site to the nearest surface expression of Umtanum anticline is about 22 miles.

The western segment of the Umtanum Ridge anticline near the Yakima River trends east-southeast and is faulted on the northeast side. A cross fault near the Yakima River shifts the asymmetry to the south. From the Yakima River to the east, the middle segment of the Umtanum Ridge anticline trends east-southeast to Hog Ranch Buttes. This segment appears to be bounded on the south by a reverse fault (Bentley, 1977). East of Hog Ranch Buttes the anticline again changes symmetry and the steep limb is on the north. Between Hog Ranch Buttes and Priest Rapids Dam, the area of maximum structural relief is reached. An imbricated thrust fault zone is exposed along this segment (Bentley, 1977). From near Priest Rapids Dam to Juniper Springs the anticline trends east-southeastward. East of Juniper Springs the anticline trends eastward and plunges beneath the surficial sediments (Goff and Myers, 1978) (Figure 2.5-16).

Near Priest Rapids Dam, where the hinge area can be delineated, its width is relatively narrow in relation to the straight southern limb of the fold. In this area, the fold has been described as "kink-like" (Rockwell, 1979). Dips on the planar northern limb range from about 35°N to overturned and 35°S.

Mapping near Priest Rapids Dam indicates a thrust fault at the base of the northern flank of the fold (Rockwell, 1979). The thrust is indicated by a discordance in the flow layering and stratigraphic sequence on the north side of the structure relative to a concordance in the flow layering and stratigraphic sequence on the south side of the structure. The age of deformation of the Umtanum Ridge structure appears to be post-Pomona and pre-upper Ellensburg fanglomerate.

A structural analysis of Umtanum Ridge in the vicinity of Priest Rapids Dam is being conducted by E. H. Price, of Rockwell Hanford. The results of his preliminary analysis are contained in Rockwell (1979, p. III-150 to III-166). Field studies are now being conducted by Golder Associates for Northwest Energy Service Company (NESCO) in this area.

2.5.1.2.4.3 Gable Mountain

Gable Mountain and Gable Butte are topographically isolated, anticlinal ridges of basalt and interbedded sediments that are the only bedrock outcrops in the central Pasco Basin (Figure 2.5-17). Portions of Gable Mountain and Gable Butte have been mapped by Newcomb and others (1972), Bingham and others (1970), Washington Public Power Supply System (1974), Brooks (1974), Fecht (1978), and NESCO (1981, unpublished). Gable Mountain and Gable Butte are composed of westerly trending, doubly plunging, en echelon anticlines and synclines interpreted as folds within the closure of larger asymmetric folds (Fecht, 1978). Gable Mountain and Gable Butte folds are flanked by the Wahluke syncline to the north and Cold Creek syncline to the south.

Three faults have been mapped on Gable Mountain by Northwest Energy Services Corporation: one fault near the west end, a second on the south flank, and a third across the central section of the mountain. Previous workers had originally mapped the fault on the south flank and the fault in the central section as a single structure with a sinuous, i.e., S-shaped trend (Bingham and others, 1970; Newcomb and others, 1972; and Fecht, 1978).

Trenching has revealed that the fault at the west end is a reverse fault that strikes N30°E and dips northwestward 55° to 60°. Displacement on this fault dies out within 200 feet of its exposure near the crest of the anticline. It appears to be a tear fault restricted to an area close to the anticlinal axis at the western end of Gable Mountain.

The fault on the south flank of the mountain is a reverse fault that strikes east-west and dips southward 30° to 45°. Trenches across this fault showed slickensides and striations on some clastic dikes present along the fault plane. Striations on polished surfaces within the dike material plunge parallel to the dip of the fault. This dike material is believed to be derived from overlying glaciofluvial sediments. Earlier investigation of the fault indicated that the overlying glaciofluvial sediments were undisturbed. These sediments were interpreted to be 40,000 years old (Bingham and others, 1970) on the basis of radiocarbon dating of wood fragments taken from the sediments. However, comparison of these sediments with other glaciofluvial sediments in the Pasco Basin suggests that they are correlative with Missoula flood deposits and are, therefore, probably 12,000 to 18,000 years in age. These sediments were found to be undisturbed where they overlie the fault approximately 300 feet east of where the slickensided and striated clastic dikes were observed.

The fault in the central section of Gable Mountain strikes N45°-55°E and dips southeastward 30° to 50°. It appears to die out to the northeast into the anticlinal axis on the north side of the mountain. Trenches excavated by NESCO in 1980 revealed that the central fault is overlain by glaciofluvial deposits that showed small displacements. These sediments are believed to be about 13,000 years old based on the presence of probable St. Helen's S ash near the top of the glaciofluvial sequence. Near the most northeastern end of the fault, slickensides and striations occur in clastic dikes (believed to be derived from the glaciofluvial sediments) within a thin (6- to 8-inch) zone of bedrock faulting. Trenches excavated at three points over a distance of approximately 1100 feet to the southwest along strike exposed minor displacements (1 to 4 inches) in the lowermost glaciofluvial deposits. Trenching several hundred feet farther southwestward along strike exposed undisplaced flood deposits. The distance over which displacement of the glaciofluvial sediments may be inferred is 1100 feet.

In addition to the mapped faults, corehole DB-10 (Rockwell, 1979) penetrated a reverse fault on the south limb of the

Gable Mountain structure. Studies by Northwest Energy Services Company indicate that this fault is relatively minor and probably oblique to the axis of folding in the area.

2.5.1.2.4.4 Rattlesnake-Wallula Alignment

The Rattlesnake-Wallula alignment is part of a topographic and structural alignment that trends northwest from near Milton-Freewater, Oregon, to the northwest end of the Rattlesnake Mountain near Horse Thief Point (Figure 2.5-18). This feature is expressed by an alignment of topographically defined anticlines between Wallula Gap and Horse Thief Point (Rattlesnake-Wallula alignment) and by the alignment of parallel and subparallel linear structural features from Wallula Gap south-eastward. The axis of this alignment approaches to within 12.5 miles southwest of the WNP 1-2-4 site.

This alignment, as defined by remote sensing, begins at Wallula Gap (Figure 2R K-17 of Washington Public Power Supply System, 1977p) and extends in a northwest direction to the north edge of Rattlesnake Ridge. The alignment cannot be traced on Landsat imagery across Dry Creek to Yakima Ridge. The lineament is very gently curved and appears as a northeast-facing break in slope. On remote sensing imagery (Washington Public Power Supply System, 1977p), the Rattlesnake-Wallula alignment appears to have a dextral strike-slip component of displacement.

On the basis of remote sensing analysis, recent faulting does not appear to have occurred along the Rattlesnake segment of the alignment as indicated by a lack of scarps across the abandoned Yakima River channel at Vista. The Rattlesnake anticline cannot be traced as a single continuous structure beyond Snively Ranch.

The Rattlesnake-Wallula alignment between the Yakima River and Wallula Gap is defined by a series of ten elongate (northwest-southeast) hills. These hills appear as distinct, subparallel, doubly-plunging anticlines generally aligned along a N50° W trend. The folds are commonly overlain by loess above elevations of 1000 feet, and by undifferentiated loess and Touchet beds below this elevation. Basalt bedrock crops out generally near the crest of the hills, while the flanks are mantled by colluvium and silt.

The portion of the Rattlesnake-Wallula alignment mapped from Badger Coulee northwest to the Yakima River (Figure 2.5-18)

contains four discontinuous, sub-parallel, doubly-plunging anticlines generally aligned along a N50°W trend. From southeast to northwest the structures include an unnamed mountain southeast of Badger Mountain, Badger Mountain, an unnamed mountain northwest of Badger Mountain, and Red Mountain.

These anticlines are structurally similar in that the northeast limbs generally dip slightly more steeply than the topographic slope and somewhat less steeply than the southwest limbs. The axes of the individual folds generally parallel the overall Rattlesnake-Wallula alignment. The Umatilla Basalt is locally exposed in the core of the folds, while the Pomona, Elephant Mountain-Ward Gap, and Ice Harbor flows are widespread.

The structure from the Yakima River gap northwestward to the north end of Rattlesnake Mountain (Figure 2.5-18), where it adjoins the east trending section of Rattlesnake Hills, is a box-shaped anticline with an over-steepened north limb along much of its length. A thrust fault or high angle, reverse fault has been interpreted to exist along the steep, northern limb (Rockwell, 1979).

The following sections include discussions of the Rattlesnake-Wallula alignment, by geographic or structural section, commencing with the Rattlesnake Mountain area at the northwestern extent of the alignment.

2.5.1.2.4.4.1 Rattlesnake Hills and Rattlesnake Mountain Anticlines

In the area of Rattlesnake Mountain, the east-west Rattlesnake Hills anticline changes direction and merges with the northwest trending section of the Rattlesnake-Wallula alignment (Figure 2.5-18). The Rattlesnake Hill Anticline area (Figure 2.5-18) extends from the north end of Rattlesnake Mountain westward to Horse Thief Point (T11N, R23E). The Rattlesnake Hills alignment in this area is defined by the sinous, east-west trending anticline. The anticline is asymmetrical with dips of 2 to 10 degrees on the south flank and dips of 10 to 25 degrees on the north flank.

Known faulting in this area is confined to the western portion of the Rattlesnake Hills anticline. The 3926 fault is a 2.5 mile long, north-south trending, vertical scissor fault located in Sec. 8, T11N, R24E, (Rockwell, 1979). The western side of the fault is up and juxtaposes Pomona against Priest Rapids flows with an estimated 300 feet of

vertical displacement. The Maiden Spring fault occurs about one mile south of the axis of the Rattlesnake Hills anticline, is parallel to it for about 1.5 miles in Sec. 18, T11N, R24E, and appears to connect with the southern end of the 3926 fault (Figure 2.5-18). It is a high angle, north dipping, reverse fault that juxtaposes Elephant Mountain and Umatilla flows; having up to 200 feet of displacement. The Rattlesnake Hills anticline at the west end (Horse Thief Point), is a structurally complex area that includes several folds and an unnamed high angle fault that juxtaposes Elephant Mountain and Umatilla flows. Much of the structure of the eastern portion in the Horse Thief Point area is hidden by landslide debris of the Snively Basin and by extensive loess cover.

The structure of this segment from the Yakima River to the northwest end of Rattlesnake Mountain is described primarily as a box-shaped anticline with an oversteepened north limb along most of its length (Rockwell, 1979). The central part of the anticline is breached and is estimated to have 3000 feet of structural relief. About one mile northwest of the Yakima River, the anticline plunges gently to the southeast and dies out on the flank of a northeast trending, southeast dipping monocline.

Faulting along this segment occurs along the northeastern limb of the anticline (Figure 2.5-18). This steeply dipping fault is interpreted as a reverse fault (Washington Public Power Supply System, 1977k; Rockwell, 1979). Displacement is estimated to be 1,300 feet (Rockwell, 1979). The fault is inferred to be about 4.5 miles in length. One mile to the northwest and on the projection of the Rattlesnake Mountain fault, a south dipping thrust fault has been inferred by Rockwell (1979). Neither the Rattlesnake Mountain fault nor the thrust fault appear to displace mapped Quaternary units.

2.5.1.2.4.4.2 Red Mountain

Red Mountain is a doubly plunging anticline about 3.5 miles in length located southeast of the Rattlesnake Mountain anticline. Pomona basalt is exposed in its core. No faults have been mapped on this structure. Remote sensing analysis indicates that Red Mountain is on-line with the major Rattlesnake trend. Several warped, abandoned stream channels and terraces suggest that to the north of Red Mountain the only Quaternary deformation has been a gentle warping.

2.5.1.2.4.4.3 Badger Mountain

In the area between the Yakima River and Badger Coulee, there are four structurally similar anticlines. Their southwest limbs generally dip slightly steeper than the topographic slope and somewhat less steep than the northeast limbs. The axes of the individual folds are subparallel to the overall Rattlesnake-Wallula alignment. The Umatilla Basalt is locally exposed in the cores of the folds, while the Pomona, Elephant Mountain-Ward Gap, and Ice Harbor flows are widespread.

Badger Mountain is defined as the three elongate hills (Figures 2.5-18 and 2.5-21) in T9N, R28E trending northwesterly across the area between Badger Coulee and the Richland-Prosser Highway. All three hills are underlain by the Ice Harbor, Elephant Mountain, Pomona and Umatilla Members of the Saddle Mountains Basalt (Washington Public Power Supply System, 1974; Rockwell, 1979; Bond and others, 1978). They are commonly overlain by loess above an elevation of approximately 1000 feet, and by Touchet beds and loess below this elevation. Structurally, the hills consist of three elongate, doubly-plunging, slightly en echelon, sinuous anticlines.

The Badger Mountain fault is a prominent structure associated with anticlines to the southeast. It extends about 4.5 miles from O-Hill to near the unnamed hill northwest of Badger Mountain (Figures 2.5-18 and 2.5-21). It appears to be a high angle reverse fault with the northeastern block downthrown. The stratigraphic displacement appears to be greatest at the southeastern end (approximately 200 feet) and decreases to the northwest where it appears to die out near the unnamed hill (Bond and others, 1978; Rockwell, 1979). The existence and extent of the fault is based on inferred stratigraphic offset, tight fold geometries, steep reverse dips, and topographic escarpments. Reconnaissance mapping (Shannon and Wilson, 1980) could not confirm the continuation of the fault between O-Hill and Badger Mountain nor the proposed northwestern projection beyond Badger Mountain. Thus, these sections of the fault are queried on Figures 2.5-18 and 2.5-21. Where the existence and location of the fault are constrained, no evidence of Quaternary displacement has been observed.

Mapping of the southeast hill (Shannon and Wilson, 1980) shows an overturned fold with a possible fault along the north flank. The fault interpretation was based on the tight geometry of the fold and steep to reverse dips along

the north flank. Mapping of the middle and northwest hills, which are comparatively broader folds, showed no evidence of faulting. (Shannon and Wilson, 1980).

2.5.1.2.4.4.4 O-Hill

The mapping of O-Hill showed an overturned fold with a possible rupture near the axial plane. The ruptured nature of the fold is inferred from the tight geometry and the overturned dip of the north limb of the fold. Ice Harbor and Elephant Mountain Basalts are steeply dipping 70° south into the Pomona Basalt along the north flank (Figure 2.5-21).

2.5.1.2.4.4.5 N-Hill/M-Hill

N-Hill and M-Hill are tightly folded, doubly plunging anticlines. On N-Hill, a few outcrops of breccia within the Umatilla basalt occur near the crest of the anticline. This breccia is overlain by the Pomona basalt with no indication of displacement. It appears that the breccia in this hill was developed by the tight folding of the Umatilla basalt. Ringold (?) sediments are steeply folded on the north flank of this anticline (Figure 2.5-21). On M-Hill no faults have been detected.

2.5.1.2.4.4.6 L-Hill

L-Hill is the smallest hill of the Rattlesnake-Wallula alignment. Its topographic relief appears to be due to a steep, north-dipping monoclinal fold. Several 2-to 5-inch wide north-dipping normal faults in the Umatilla basalt are exposed in quarries. These minor faults are cross-cut by clastic dikes that are not offset.

2.5.1.2.4.4.7 K-Hill

K-Hill, also known as Game Farm Hill, is a doubly-plunging anticline. Jones and Landon (1978) mapped a 1/2-mile long fault (Game Farm Hill fault) along the south flank of the anticline. The fault is prominently exposed in a quarry at the southeastern end of the hill (Table 1, Appendix 2.5-N). Shear zones in the quarry are cross-cut by clastic dikes that show no fault displacements.

2.5.1.2.4.4.8 The Butte

The Butte is a tightly folded, east-plunging anticline. The Finley Quarry fault is exposed at the west end of the Butte

(Figure 2.5-21). It consists of a 35-foot wide zone and includes fragmented and pulverized slivers of basalt and interbeds. The fault zone is bounded by two fault planes.

The south bounding fault strikes N65°W and dips vertically. This fault is overlain by a thin veneer of undisturbed loess. The north bounding fault strikes approximately east-west and dips 55° to the south. It juxtaposes an older colluvial unit against the fault zone. A younger undisturbed colluvial unit overlies the north fault and is in turn overlain by up to 10 feet of loess. The age of this loess is estimated to be at least 6,600 y.b.p. based on the presence of what is probably Mazama ash within similar loess on slopes in nearby areas.

Mapping by Rockwell (1979) has extended the Finley Quarry fault approximately two miles southeast of Finley Quarry (Figure 2.5-21). This is based on inferred stratigraphic offset (Jones and Landon, 1978; Rockwell, 1979). The Finley Quarry fault lies along the projected trend of the Wallula fault zone (Shannon and Wilson, 1979; 1980).

2.5.1.2.5.4.9 Molly Hill

Molly Hill is a northwest-trending, tightly folded, southeast- plunging anticline. Both the north and south limbs are faulted (Figure 2.5-21). The two faults define a small horst that parallels the trend of the hill. The fault on the north limb is marked by a linear breccia zone that trends about N60°W. Offset is interpreted to be less than 100 feet (Jones and Landon, 1978). This fault may be related to the Finley Quarry fault (Jones and Landon, 1978), and it is the northwest extension of the Wallula fault zone (Shannon and Wilson, 1979a; Washington Public Power Supply System, 1977n) described below. The southern fault merges with the dip slope on its northwest and southeast ends.

2.5.1.2.4.4.10 Kennewick, Horn Rapids, Cold Creek Lineaments

The Kennewick, Horn Rapids, and Cold Creek lineaments are subparallel to the Rattlesnake-Wallula lineament from the vicinity of Wallula Gap to near the east end of Yakima Ridge (Figures 2.5-18 and 2.5-21). It can be seen on topographic maps, aerial photographs, LANDSAT imagery, and on various geologic maps.

The Kennewick and Cold Creek lineaments were recognized and described by Glass (Washington Public Power Supply System, 1977p) as follows:

"The Kennewick lineament parallels the Rattlesnake structure from Wallula Gap to approximately Kennewick. The lineament is characterized by an abrupt vegetation contrast and an east-facing break in slope (Figure 2R K-15 of Washington Public Power Supply System, 1977p). A number of factors have led me to interpret this feature as a terrace. At its northern end the Kennewick lineament turns to the northeast and joins several similar features originating to the west and northwest. At its southern end the lineament gradually (sic) decreases in height and cannot be followed south of Finley. Several older terraces appear upslope to the west of the Kennewick lineament and trend roughly parallel to it. The Cold Creek lineament extends from Columbia Camp to beyond Benson Ranch (Figure 2R K-24, Sheet 2 of Washington Public Power Supply System, 1977p). The lineament is eminently detectable on LANDSAT imagery and parallels the Rattlesnake structure."

Glass in Washington Public Power Supply System (1977p) describes the lineament as paralleling the Rattlesnake trend, but about 1.5 miles east of it, and as being formed by a 6 to 9 foot-high topographic break combined with a striking contrast in vegetation. He also noted several ponds resembling sag ponds along the lineament's trend and suggested the lineament may extend onto the Hanford Reservation in the vicinity of Cold Creek.

Although the lineament may appear to be a single, continuous feature on high-altitude imagery or on larger-scale topographic maps, detailed analysis shows that it actually consists of three distinct, nonaligned segments, each having its own peculiar geomorphic, geologic, and geographic limits.

The Cold Creek lineament consists of two features. The first is the Cold Creek valley, a topographically low linear area parallel to Rattlesnake Mountain. The second feature is the low scarp, seen along the west side of the valley near the confluence with the Yakima River. This feature was probably produced by the lateral planation of Cold Creek. Both of these features are probably post-Wisconsin in age.

The Horn Rapids lineament consists of a series of six low hills trending about N35°W along the southwest side of the Yakima River to just north of Horn Rapids (Jones and Deacon, 1966; and Washington Public Power Supply System, 1974). The hills consist of shallow dipping flows of the Elephant Mountain and Ice Harbor members of the Saddle Mountains Basalt mantled by glaciofluvial deposits (Figure 2.5-18). This group of low, basaltic hills is interpreted to be a

series of low amplitude folds subparallel to the Rattlesnake-Wallula alignment. These hills have been modified by the plucking action of Spokane flood waters and subsequently nearly buried under glaciofluvial deposits. No evidence for faulting was observed (Washington Public Power Supply System, 1974).

The Kennewick Lineament is a distinct feature on topographic maps and can be plainly seen as a slope break and vegetative change on aerial photographs (Figures 2R K-12 and 2R K-15 of Washington Public Power Supply System, 1977p) as well as on LANDSAT photos of the area. It is a portion of a broader pattern of terraces (Washington Public Power Supply System, 1977h).

The Kennewick lineament chiefly consists of the break in slope between terrace levels T_3 and T_2, and was probably produced as an erosional feature during one of the later episodes of Spokane flooding (Wisconsinan age) (Shannon and Wilson, 1980). The linear nature of the feature, although long and distinct, is not unique among erosional or depositional features produced by Spokane flooding. For example, between Umatilla and Boardman Junction in Oregon, a 400 foot-terrace can be traced as a near-straightline feature for about 11 miles, while in the Portland, Oregon area a 200 foot-terrace can be traced for about 6 miles. Thus, it appears that the Kennewick lineament may represent a streamlined erosional feature produced by flood waters rushing toward Wallula Gap. The dramatic, vegetational change along the lineament coincides with the proximity of groundwater to the surface. In fact, the ponding of spring water issuing from the gravel near the base of the break in slope between terrace T_2 and T_3 was noted by Glass in Washington Public Power Supply System (1977p), who originally thought they might be sag ponds. Ponding appears to be caused in part by the pervious lining of the irrigation ditch, which parallels the break in slope. The possibility of right-lateral motion along this trend was hypothesized and then discarded by Glass in Washington Public Power Supply System (1977p), Figures 2R K-10 and 2R K-11.

According to Glass' interpretation given above, the Kennewick lineament would have been produced in late Wisconsin time. Alluvial fans produced since that time, by sidestreams debouching onto terrace T_2, show no features that can be considered part of the alignment. No surficial evidence could be found for either horizontal or vertical movement along the Kennewick-Cold Creek lineament.

2.5.1.2.4.5 Wallula Fault Zone

The Wallula fault zone (Figure 2.5-21) extends southeastward from slightly west of Wallula Gap to the vicinity of Milton-Freewater, Oregon (Shannon and Wilson, 1979a). This zone has topographic expression from Wallula Gap southeastward to a point about two miles east of Warm Springs Canyon. Beyond this point the fault zone is covered by the Touchet Beds of glacial age (approximately 13,000 y.b.p.) and is not clearly defined (Shannon and Wilson, 1979a).

Near Wallula Gap, the Wallula fault zone apparently consists of two major strands. The northern inferred strand may continue to the doubly-plunging anticlines and domes of the Rattlesnake-Wallula alignment. Bingham and others (1970) first proposed this link between the Wallula fault zone and the doubly-plunging folds. This link has not been verified because of a lack of bedrock exposures between Molly Hill and the inferred junction of the two strands. Additional discussion of the connection between the Wallula fault zone and the Rattlesnake-Wallula alignment is contained in Appendix 2.5-N.

The southern strand is the Wallula Gap fault (Jones and Deacon, 1966). It extends approximately 5.5 miles west-northwest from the inferred junction of the two strands. West of Yellepit the terminus of the fault is in dispute (Farooqui in Washington Public Power Supply System, 1977i; Shannon and Wilson, 1979a; Foundation Sciences, 1980). A trench excavated across the Wallula Gap fault near Yellepit (see Washington Public Power Supply System, 1977i) revealed that the fault does not displace the Kennewick fanglomerate. Caliche in the fanglomerate was dated for Woodward-Clyde Consultants (1978) and found to be approximately 55,000 y.b.p. The methodology used in this dating technique is presently being re-evaluated. In the Yellepit trench both vertical and horizontal striations were observed; the former were prominent. Remote sensing analysis of the area between Vansycle Canyon and Wallula Gap indicates that the Wallula Gap fault could extend as far west as Jumpoff Joe.

East of the gap, the Wallula Gap fault is not exposed but its presence is inferred from narrow shear zones and a series of faceted spurs. A shear zone along the highway on the east side of Wallula Gap shows prominent near-horizontal striae. Approximately 2.5 miles east of Wallula Gap, the Wallula Gap fault may merge with the inferred northern strand and continue southeastward across Vansycle Canyon to

the vicinity of Warm Springs Canyon. Scarp-like features along the fault appear at the mouth of Vansycle Canyon but appear not to displace recent alluvium. Aerial photo analysis of the Wallula Gap fault indicates that it does not appear to be active Glass in Washington Public Power Supply System (1977p).

Bingham and others (1970), while investigating this segment of the Wallula fault zone, observed a youthful, curved linear feature they interpreted to be the result of recent faulting. Trenching across this linear feature in 1981 by Woodward-Clyde Consultants does not confirm the conclusions of Bingham and others (1970). Although minor diplacements associated with clastic dikes in the Touchet deposits were exposed in these trenches, continuous strata (including the St. Helen's S ash) across the lineament zone indicate the absence of post-Touchet faulting having enough vertical displacement to account for the topographic relief (locally as much as about 3 feet) associated with the lineament.

Faulted colluvial deposits of probable Quaternary age are exposed approximately one mile east of Warm Springs Canyon. A discussion of this exposure can be found in Shannon and Wilson (1979a). Additional geologic studies including trenching are in progress at this location.

Approximately two miles east of Warm Springs canyon, the Wallula fault zone is concealed by Touchet deposits and cannot be clearly defined to the southeast (Shannon and Wilson, 1979a). Although the topographically-defined trace of the fault zone becomes indistinct, the continued presence of northwest-striking normal (?) faults in the hills to the south as far east as Milton-Freewater implies an extension of the Wallula fault zone into that region. Because of extensive surficial cover, neither these en echelon fault planes nor their intersections with the Wallula fault are exposed. The geometry of the escarpments produced by these en echelon faults show southeastward decreasing displacement.

In the area of the Walla Walla Basin the location of the Wallula fault zone is uncertain but several occurences of faulting have been identified. South of Umapine, Oregon, north-dipping (30° to 60°) faults of tectonic or slump origin displace Touchet beds and cross-cutting clastic dikes having maximum offset of 1.5 feet (Shannon and Wilson, 1979b). Faulting 6.2 miles southeast of the Umapine locality is suspected by Shannon and Wilson (1979b) but not demonstrated.

Two other Quaternary fault localities are the Buroker thrust fault east of Walla Walla and the Little Dry Creek fault south of Milton-Freewater. The base of the Pleistocene Palouse Formation is offset approximately three feet along the former fault, a west-dipping (26°) reverse fault that strikes north-south. Loess deposits (Holocene?) overlying the Palouse appear to be unfaulted (Shannon and Wilson, 1980). Near Little Dry Creek, basalt and Palouse soil are downdropped about 1.5 feet along a steep (75°) northwest-striking, northeast dipping fault. This fault lies south of the eastward projection of the Wallula fault zone and is not in alignment with it.

The southern extent of the Wallula fault zone is inferred to extend along the South Fork of the Walla Walla River. Raisz (1945) believed that the South Fork of the Walla Walla River formed part of the Olympic-Wallowa lineament which continues southeasterly across the Blue Mountains to the Wallowa Mountains. However, mapping by D. Swanson (U.S.G.S., unpublished), by Shannon and Wilson (1979b), and by R. Dale and J. Kendall (Kendall and others, 1981) has demonstrated the continuity of major vertical faults (including the Hite fault)and the west-dipping dikes of Frenchman Springs basalt across the South Fork of the Walla Walla River. Thus, the extent of this inferred segment of the Wallula fault zone can continue no farther southeast than Milton-Freewater, Oregon.

2.5.1.2.5 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. Davis (in Washington Public Power Supply System, 1977a) proposed 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 1980, however, Hill concluded that the crust beneath the Pasco Basin is indeed thin (ca. 15 miles 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 (Appendix 2.5-O) has proposed that the basement is genuinely continental, but was thinned during early Tertiary regional doming -- expression of the doming and resultant crustal thinning is represented by 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. Davis (Appendix 2.5-N) still believes that exposures of ophiolitc rocks and associated marine sediments north (Ingalls' ophiolite of Washington Cascades) and west (Rimrock Lake) of the plateau imply an accreted, oceanic basement beneath central areas of the plateau.

Recent geophysical studies support both Davis' (Washington Public Power Supply System, 1977a) and Laubscher's (Appendix 2.5-O) 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) (Figure 2.5-9) that a change in basement rocks or crustal character occurs along the lineament. Furthermore, a pronounced gravity gradient separates the Yakima and Pasco Basins and trends north-south across OWL. 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 N45°W and S45°W that cross the gradient would have horizontal displacements less than two-three km. This conclusion supports Laubscher's contention (1977 and Appendix 2.5-O) that any strike-slip displacement along Raisz's lineament (CLEW) must be less than 2 km. The north-south gravity gradient referred to above is interpreted by Laubscher to

delineate the western edge of a master north-south trending graben of Eocene age developed longitudinally along the crest of the regional dome he has postulated.

A recent study (Systems, Science, and Software, 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 revealed no changes in crustal or mantle structure at depths greater than 6 miles 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.

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 prominent horseshoe-shaped outcrop pattern of the Eocene Crescent Volcanics around the northern, eastern, and southern margins of the Olympic Mountains defines this antiformal structure (Tabor and Cady, 1978b). A recent U.S.G.S. report on the geology of the Olympic Peninsula (Tabor and Cady, 1978b) 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 the displacement history of the faults is difficult to evaluate, they appear to be unlikely representatives of a hypothesized 390 mile-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. However, mapping by D. Swanson, U.S.G.S. (unpublished), by Shannon and Wilson (1979b), and by R. Dale

and J. Kendall (Kendall and others, 1981) has demonstrated the continuity of vertical major faults and west-dipping dikes across the South Fork of the Walla Walla River. It is, therefore, clear that this segment of the topographically defined Olympic-Wallowa lineament cannot be related to faulting in rocks of Miocene age.

Davis (in Washington Public Power Supply System, 1977a, p. 2RC-34) states that the Olympic-Wallowa lineament is "as originally defined a fictional structural element of the Pacific Northwest." In 1981 Davis (Appendix 2.5-N) is still inclined to this 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. This is the CLEW zone of Laubscher (Appendix 2.5-O).

2.5.1.2.6 Hazards

Actual or potential hazards or problems related to natural geologic features or to man's activities in the region and site vicinity have been assessed. It was determined that there are no potential hazards or problems to the site due to any of these phenomena.

2.5.1.2.6.1 Volcanic Hazards

There are several major volcanoes in the Cascade Range west of the WNP1-2-4 site. The closest is Mt. Adams about 102 miles distant; the most active is Mt. St. Helens approximately 137 miles west-southwest of the site. Because most volcanic activity is confined to the immediate area of the volcano, mud flows, avalanches, pyroclastic rock flows, lava flows, and shock waves that may be associated with such activity do not pose a hazard to the site. The only potential hazard to the site is ash fall resulting from a major eruption of one of these volcanoes.

Several historical ash deposits have been detected in the Columbia Basin but only two are known to be present in the vicinity of the site. Although the ash falls may be widespread in extent, commonly they are not found everywhere owing to the nature of their depositional environment and to subsequent erosion. The oldest ash layer identified in the site vicinity occurs in the upper Touchet Beds. In the past, investigators have considered it to be a Glacier Peak ash with an age of approximately 12,000 y.b.p. (Brown, 1970). However, recent studies by the U.S.G.S. indicate that it may be from a Mt. St. Helens eruption that occured