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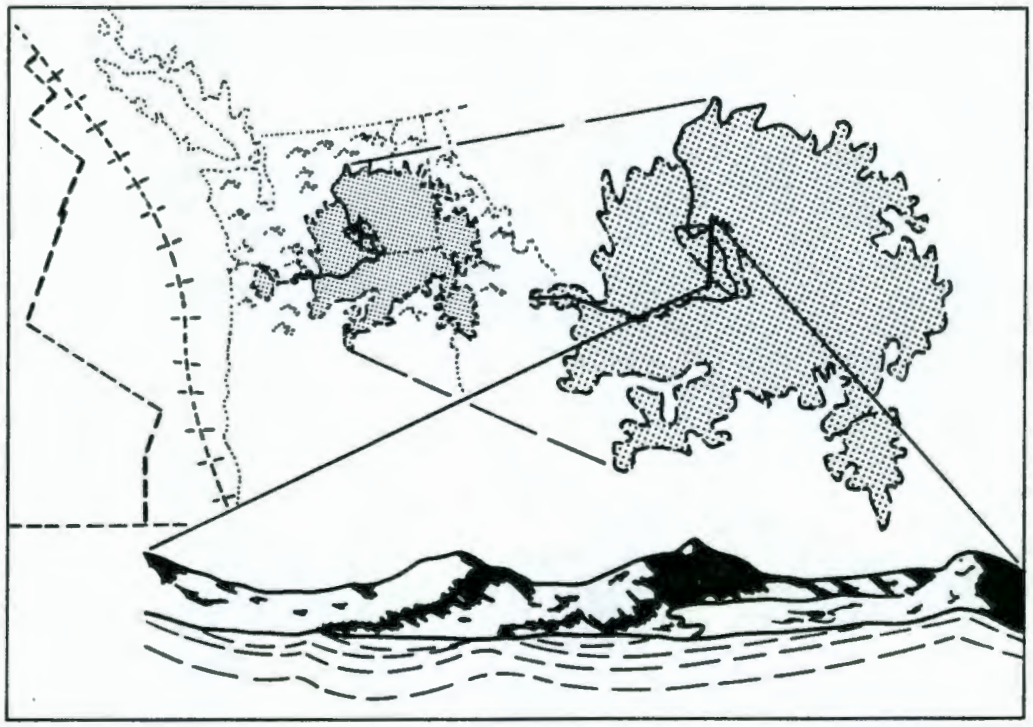

March 1983

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Preliminary Interpretation of the Tectonic Stability of the Reference Repository Location, Cold Creek Syncline, Hanford Site

March 1983

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EXECUTIVE SUMMARY

Staff

The Columbia Plateau is a tholeiitic flood basalt province that occurs west of the Rocky Mountains near the western margin of the North American plate. Columbia River basalt, which underlies the plateau, was erupted between at least 16.5 and 6 million years ago from nearly north-south linear vent systems mostly in eastern Washington, northeastern Oregon, and west-central Idaho. Basaltic lavas flowed generally westward and consolidated in an intermontane basin between the Rocky Mountains and the ancestral Cascade Range. Columbia River basalt has been divided into five formations based on stratigraphic relationships, chemistry, and paleomagnetic polarity. Three of these formations, the Grande Ronde, Wanapum, and Saddle Mountains Basalts, occur in the Pasco Basin where the maximum known thickness of Columbia River basalt exceeds 3 km.

Rocks ranging in age from Precambrian to Quaternary are exposed along the margins of the plateau. Many of these rocks along the margin have been structurally deformed along northwest, northeast, and north-south trends that extend into the plateau. The age, rock types, and structures that occur beneath Columbia River basalt in the central Columbia Plateau can only be inferred from terrains and structural trends observed along the plateau margin.

Deformation of varying styles has enabled subdivision of the plateau into three structural subprovinces: the Yakima Fold Belt, the Blue Mountains, and the Palouse subprovinces. The Pasco Basin is a Tertiary basin that occurs in the central Columbia Plateau near the junction of the Yakima Fold Belt and Palouse subprovinces. Asymmetric anticlines characteristic of the Yakima Fold Belt border the Pasco Basin on the north, west, and southwest. These folds have generally steeper north limbs in which thrust or reverse faults subparallel to the folds axes are common. Strata on the south limb of these anticlines dip gently into broad synclines containing Neogene and Quaternary sediments.

Geologic investigations and interpretations of available data have been used to select a candidate site for a nuclear waste repository in the Cold Creek syncline of the Pasco Basin. The reference repository location is a 47-km^2 area located in the western part of the syncline in the broad axial trough where the basalt strata are nearly flat lying.

A site for a nuclear waste repository should be in a tectonically stable area. The Basalt Waste Isolation Project defines a tectonically stable area as one where the late Cenozoic tectonic processes have been identified and their rates of operation determined so that, when these rates are projected into the future, they do not adversely affect repository operation or waste isolation.

A preliminary assessment of the tectonic evolution of the reference repository location has established the following pattern of deformation:

- The stress regime suggested by nearly east-west folds and subparallel thrust and reverse faults and steeply dipping northwest and northeast-trending faults in the Yakima Fold Belt is one of nearly north-south, nearly horizontal compression, with a nearly vertical axis of least compression.

- Focal mechanism solutions of earthquakes in the central Columbia Plateau indicate that this stress regime exists today and suggests that this regime has been relatively unchanged for >14 million years. No change of this stress regime is anticipated over the next 100,000 yr.

- Deformation was in progress in the late Grande Ronde time (~14.5 million years before present) and continued at an average low rate of uplift (vertical strain) from 14.5 to 10.5 million years before present as determined from the aerial and thickness distribution of basalt flows.

- Projection of average low uplift rates in the Saddle Mountains and Rattlesnake Hills, which were deforming in the late Miocene over a long term, can account for the present elevation of basalt flows in these anticlinal structures. Similarly, subsidence at these long-term, average low rates can account for the development of the Cold Creek syncline relative to the stable Palouse paleoslope.

- Deformation, once initiated, appears to have continued on developing first-order structures. Strain appears to be concentrated in steeply dipping strata and on major structures. New first-order structures do not appear to have developed in the Quaternary nor are they anticipated to develop in the next 10,000 to 100,000 yr.

- Seismicity in the central Columbia Plateau is confined to a thin 28-km crust and is characterized by temporally and spatially limited swarms of low magnitude (magnitude ≤3.5) shallow (≤6 km) earthquakes that may be characteristic of brittle deformation in basalt. Deep earthquakes (>6 km) respond to the same stress regime, occur as single events or mainshock-aftershock sequences, appear more diffuse in spatial distribution, and do not generally appear to be related to shallow events. Differences in recurrence (b value) and pattern of activity suggest that basalt may behave in a more brittle manner than rocks beneath basalt in which some aseismic deformation may be occurring.

- The largest earthquake recorded during 12 yr of instrumental monitoring in the central Columbia Plateau was a magnitude 4.4 event on Royal Slope, the south flank of the Frenchman Hills, a Yakima fold several kilometers north of the Pasco Basin.

- Earthquakes in the central Columbia Plateau are not presently associated with mapped geologic faults, nor do hypocenters align in a manner that suggests the presence of unmapped faults. Swarms have occurred on the flanks of the Saddle Mountains, a first-order structure which is faulted, but the events do not correspond with mapped faults. However, swarms have also occurred elsewhere where there are no mapped geologic structures. Some small alignments are indicated by the migration of swarm events in the Saddle Mountains.

- Focal mechanism solutions and the diffuse spatial pattern of hypocenters for swarms suggest that slip occurs on several fault planes and not one principal fault.

- The geometric pattern of essentially east-west or northwest fold axes and subparallel thrust or reverse faults and nearly straight, lengthy, steeply dipping faults of northwest and northeast-trend suggest deformation in a wrench tectonic regime that is anticipated to continue.

- The northwest-trending Cle Elum-Wallula zone of deformation appears to be a zone of dextral shortening that may relate to deformation in the basement beneath the basalt. Gravity and magnetotelluric data suggest that rocks of contrasting density or resistivity are not juxtaposed across the Cle Elum-Wallula zone of deformation along the Rattlesnake Hills on the margin of the Pasco Basin. Seismicity is not concentrated along this zone as might be expected if it were a major zone of weakness in the crust.

- Magnetotelluric data suggest the presence of three thick laterally continuous superposed units in the Pasco Basin: (1) an upper resistive unit assumed to be Columbia River basalt which overlies (2) a conductor (interpreted to be sediments) that is underlain by (3) a resistor (interpreted to be crystalline basement). A preliminary interpretation suggests high relief at the top of the conductor and the lower resistor. At Umtanum Ridge, an interpreted basement high corresponds with an anticlinal structure in the basalt; elsewhere a correlation between structure in basement and in basalt is more ambiguous with available data. Detachment surfaces could be present within the conductive or resistive units or at the base of folds; however, a major regional décollement at the base of the basalt is not supported by the high relief at the contact of the basalt and conductor or the conductor and lower resistor.

- Several tectonic models may explain the kinematic development of Yakima folds. Differences in the models relate to the mechanical growth of folds and the sequential development of folds and faults. Several models are based on properties or structures hypothesized to occur within, at the base, or below the basalt, which are inferred from the gross spatial disposition and geometry of folds and faults.

This preliminary interpretation of available geologic, seismologic, geophysical, and geodetic data indicates that the reference repository location in the Cold Creek syncline of the Pasco Basin meets the U.S. Nuclear Regulatory Commission and Basalt Waste Isolation Project proposed criteria for tectonic stability. Uncertainties regarding the timing, sequence, and mechanics of fold and fault development will be addressed during the program of detailed site characterization. This preliminary assessment of tectonic stability does not appear to preclude development of a repository in the reference repository location.

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CHAPTER 1 - RATIONALE AND APPROACH FOR A TECTONIC STABILITY ASSESSMENT OF THE REFERENCE REPOSITORY LOCATION

J. A. Caggiano

INTRODUCTION

In February 1976, the U.S. Energy Research and Development Administration, currently the U.S. Department of Energy (DOE), expanded its commercial radioactive waste management programs and established the National Waste Terminal Storage (NWTS) Program. The NWTS Program mission was defined to provide multiple repository facilities in various deep geologic formations within the United States for the storage of nuclear waste. The scope of the program includes development of technology necessary for designing, licensing, constructing, operating, and decommissioning a repository, and identification of a number of geologic sites suitable for location of radioactive waste repositories. The Columbia River basalts beneath the Hanford Site were among those media selected for initial study and characterization.

In July 1978, the NWTS Program was restructured and expanded. The DOE contracted with Battelle Memorial Institute to develop the technology for mined geologic disposal of nuclear wastes. The Office of Nuclear Waste Isolation was estabished with the Project Management Division of Battelle Memorial Institute to adapt and augment existing NWTS programs. As part of the NWTS Program, the Basalt Waste Isolation Project (BWIP) was established with an objective to assess the feasibility of disposal of commercial radioactive spent fuel/high-level waste in deep geologic formations in basalt beneath the Hanford Site. The BWIP mission is to identify potential geologic repository sites in basalt beneath the Hanford Site and to develop the associated facilities and technology required for the permanent isolation of radioactive wastes in basalt formations. The BWIP is one element of the NWTS Program and is presently managed by the DOE-Richland Operations Office. Rockwell Hanford Operations (Rockwell) is the prime contractor responsible for this work.

HISTORY OF GEOLOGIC AND TECTONIC STUDIES AT THE HANFORD SITE

Geologic and tectonic studies have been conducted at the Hanford Site for several decades to characterize and monitor the flow of groundwater and to support design and construction of facilities. More than 1,500 boreholes have been drilled since the 1940s to characterize the sediments overlying Columbia River basalt within the Hanford Site and to define the rate and direction of groundwater flow in the unconfined aquifer. Initially, little work was done to map basalt flows beneath the sediments, owing to a lack of diagnostic criteria by which flows could be identified.

Concern for the structural stability of facilities in the 1960s led to investigations of major geologic structures and their earthquake potential. Interpretation of aerial photographs and reconnaissance fieldwork were used to identify faults, evaluate their age of activity, and their potential for damaging earthquakes (Jones and Deacon, 1966; Jahns, 1967; Waldron and Bonilla, 1968). This included geologic mapping and trenching at Gable Mountain and at the Smyrna Bench area of the Saddle Mountains, and reconnaissance geologic mapping along Rattlesnake and Horse Heaven Hills (Bingham and others, 1970) (Fig. 1-1). Most of the faults were shown to be the result of folding associated with development of the anticlinal ridges of the central and western Columbia Plateau (Brown, 1968).

In 1969, a network of seismometers was installed around the Hanford Site by the U.S. Geological Survey for the Atomic Energy Commission. Hundreds of microearthquakes were detected with this network (Pitt, 1971, 1972); however, these events were not clearly related to mapped geologic structures. Focal mechanism solutions suggested that rupture was occurring as a result of nearly north-south compression. This network of seismometers was expanded to the eastern Washington region in 1975 and is currently operated for the DOE by the University of Washington (Chapter 6).

In 1972, a trilateration array was established and measured for the first of six times with a precision geodetic survey. Nineteen monuments were established and 29 lines have been surveyed since 1972 by the U.S. Geological Survey (Savage and others, 1981; Chapter 6).

In 1968, a study was initiated to determine the feasibility of permanently storing radioactive defense waste in caverns mined in the Columbia River basalt underlying the Hanford Site. To accomplish this study, it was first necessary to distinguish individual basalt flows. Boreholes were drilled and cored, and the basalt in cores was characterized using state-of-the-art techniques. The resulting stratigraphy was then used to evaluate the confined groundwater flow system and the structure of the Columbia River basalts (LaSala and Doty, 1971; Newcomb and others, 1972). Basic data were collected on the geochemistry, stratigraphy, structure, and hydrology of the basalts, but interpretation of these data was suspended in 1972. In 1976, the Office of Waste Isolation requested that these data on basalt be documented in support of the feasibility study for terminal storage of commercial high-level nuclear waste (ARHCO, 1976).

Since 1968, geologic and tectonic studies have been performed in support of construction of the Fast Flux Test Facility and for nuclear powerplants for the Washington Public Power Supply System and Puget Sound Power and Light Company, all located on the Hanford Site. These investigations focused on developing seismic design parameters and site stratigraphy and foundation conditions for these nuclear facilities (Blume and Associates, 1970, 1971; WPPSS, 1974, 1977, 1981; PSPL, 1982).

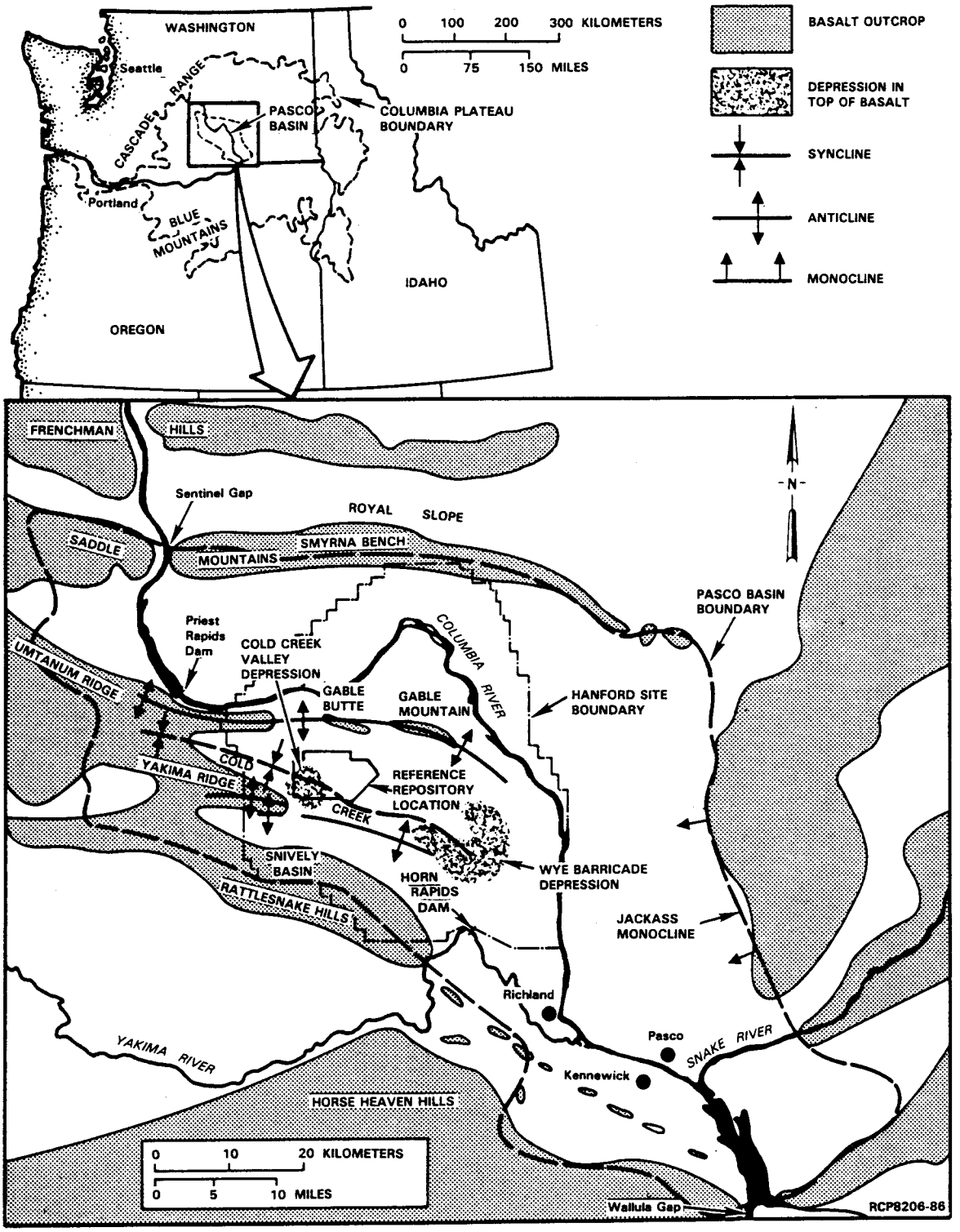

FIGURE 1-1. Extent of the Columbia River Basalt Group, Pasco Basin, and Reference Repository Location.

Since 1977, the BWIP investigations have focused on assessing the suitability of Columbia River basalt beneath the Hanford Site in the Pasco Basin as a site for a nuclear waste repository. These investigations have included reconnaissance level investigations throughout the Columbia Plateau (defined for this report as the outcrop extent of the Columbia River Basalt Group (see Fig. 1-1; Myers/Price and others, 1979). More detailed geologic mapping of exposed basalt and late Cenozoic sediments within the Pasco Basin has been supplemented by borehole, geodetic, and geophysical studies. These investigations were among those used to identify several candidate sites for a nuclear waste repository in the Cold Creek syncline of the Pasco Basin. A 47-km^2 area in the western part of the Cold Creek syncline was chosen as the preferred site (WCC, 1981b) and has been designated the reference repository location (see Fig. 1-1). The geology of the Cold Creek syncline with emphasis on the reference repository location was described in Myers and Price (1981); plans to further characterize the reference repository location and a summary of geologic and geophysical studies leading to its selection are contained in the Site Characterization Report for the BWIP (Rockwell, 1982).

Geologic and tectonic investigations completed to date have not revealed any conditions that would preclude development of a repository in Columbia River basalt beneath the Hanford Site. Uncertainties regarding the details of the geologic and tectonic evolution of the reference repository location and vicinity will be addressed during the program of detailed site characterization as outlined in the Site Characterization Report for the BWIP (Rockwell, 1982). The data and investigations used to reach this preliminary assessment of the tectonic stability of the reference repository location are reviewed in this report.

RATIONALE FOR TECTONIC STUDIES

The tectonic features (i.e., folds, faults, shear zones) and tectonic processes (i.e., folding, faulting, fracturing) at the Hanford Site are being evaluated in order to determine whether a repository can be designed and operated to meet performance criteria (NRC, 1981). Tectonic processes affecting surface and subsurface facilities are being identified so that their possible effects on repository design can be evaluated. Since faulting in or very close to a repository could lead to failure of rock in tunnel or shaft walls, the potential for faulting is being assessed. The rate and magnitude of faulting and folding and their potential effects on nearly flat-lying basalt strata in the reference repository location are being evaluated. The location of faults and probable vibratory ground motion produced by nearby earthquakes are also being evaluated for repository design.

After closure of a repository located in basalt, thermal stresses induced by the waste along with tectonic stresses in the host rock could potentially lead to the modification of existing fractures or the development of new fractures that could accelerate the transport of radionuclides in groundwater to the accessible environment. The nature and probability

of potential effects produced by tectonic stresses after repository closure are being ascertained for use in computer models to assess repository performance during the period of operation and over the suggested 10,000-yr period (NRC, 1981) following repository closure.

Scenarios of disruptive tectonic events have been identified that could create pathways by which groundwater flow paths and transport time to the accessible environment are reduced. These scenarios will be analyzed to assess the possible consequences of tectonically produced effects on waste isolation (Arnett and others, 1980; Caggiano, 1982).

Predictions of future tectonic events (for a 10,000-yr period) that will be used in the assessment of repository design and performance can be obtained by developing a conceptual tectonic model, which is the goal of tectonic studies of the BWIP. A tectonic model is a theory or hypothesis that satisfactorily explains the evolution of geologic formations and their structural development during the recent geologic past. Tectonic processes and their estimated rates of operation during the geologic past can then be extrapolated to the future to predict both potential short-term effects needed for design and potential long-term effects required for performance assessment.

TECTONIC PROCESSES AND SITING CRITERIA

Because tectonic processes could affect the design and operation of a repository, requirements have been established in various siting criteria (NRC, 1981; NAS, 1978; NWTS, 1981; IAEA, 1977). Areas that have been tectonically stable for hundreds of thousands of years (i.e., undergoing slow rates of change) are preferred, especially sites where stability can be adequately demonstrated from the geologic record. The criteria for tectonic stability generally specify characteristics or features to be avoided, including:

- Structurally unstable tectonic blocks and tectonic boundaries (NAS, 1978)
- Faults that have been active in Quaternary time or that have the potential to rupture during repository operation (NAS, 1978; NRC, 1981)
- Faults or other geologic structures that could compromise repository design, operations, or containment (NWTS, 1981)
- Historic earthquakes of a size that could, if repeated, affect the operation of the repository (NRC, 1981)
- Seismic zones with a higher rate of occurrence of earthquakes compared to surrounding areas or with the potential for more frequent, large events in the future (NRC, 1981)

- Rapid uplift or subsidence rates that could compromise repository integrity or lead to increased erosion that could jeopardize isolation (NWTS, 1981)

- Quaternary volcanism, the potential for renewed volcanism, or high geothermal gradient (NRC, 1981; NAS, 1978; NWTS, 1981).

Using this list of potentially adverse conditions, the BWIP defines a tectonically stable area as one where the late Cenozoic (especially Quaternary) tectonic processes have been identified and their rates of operation determined so that, when these rates are projected into the future, they do not adversely affect repository operation or waste isolation. This definition agrees with and amplifies the definition of stability given in 10 CFR 60 (NRC, 1981). Since there are no areas on earth that are static and unchanging, stable areas are those where tectonic processes operate at slow rates compared to active, orogenic areas.

BASALT WASTE ISOLATION PROJECT STRATEGY FOR TECTONIC STUDIES

The BWIP is focusing on assessing the tectonic stability of the reference repository location in the Cold Creek syncline in the Pasco Basin (see Fig. 1-1). This assessment entails integrating geologic, geophysical, geodetic, and seismologic data obtained by the BWIP with data and interpretations reported by others. Geologic data provide information on the chronological development of geologic structures and the stresses responsible for their development. Geophysical surveys provide data on gross crustal properties and subsurface anomalies that may reflect structures not expressed at the ground surface. Seismologic and geodetic data provide information on contemporary processes. The objective of this effort is the development of a conceptual tectonic model that can (1) satisfactorily explain the tectonic evolution of the region in the recent geologic past, and (2) be used to predict potential effects of tectonic processes continuing to operate over the period of containment of the nuclear waste.

The tectonic stability of the reference repository location is also being assessed by considering its regional setting, because the evolution of the reference repository location must be compatible with the known or hypothesized tectonic evolution of the region. In evaluating the tectonic stability of the reference repository location, emphasis is placed on the development of the Pasco Basin and Columbia Plateau since the reference repository location is not a tectonically distinct unit. Adoption of this approach approximately conforms to the suggested radius of geologic investigation for a candidate area of ~100 km (NRC, 1980; NWTS, 1981).

Confidence in tectonic models that will be used to predict possible stress and strain for 10,000 or more years after closure of a repository is a function of how well the model explains the evolution of the crust over the recent geologic past. The major emphasis on assessing tectonic stability for the Pasco Basin is on the last 15 million years (m.y.)

because a stratigraphic record of that interval is present in outcrops and cores from boreholes, and because it is assumed that processes that have operated over the past 15 m.y. will continue to operate at similar rates over the next 10,000 yr. The greater the knowledge of these rates, the greater the confidence in predictive models used to assess isolation.

SCOPE AND ORGANIZATION OF REPORT

This is a progress report summarizing interpretations and data that were available prior to May 1982. Geologic, geophysical, seismologic, and geodetic studies are continuing as part of an ongoing assessment of the tectonic stability of the area to support design and performance assessment.

This preliminary assessment of the tectonic stability of the reference repository location examines the regional geologic and tectonic setting of the Pasco Basin and reference repository location (Chapters 2 and 3), reviews geologic and geophysical data on possible rock types and structures below basalt (Chapter 4), and discusses geologic evidence for slow, continuous deformation during the last 15 m.y. (including contemporary deformation, Chapters 5 and 6). Following this discussion, tectonic models are reviewed and evaluated (Chapter 7) as to how well they explain the geologic and geophysical data (Chapters 2 through 6). Chapter 8 then summarizes this assessment of tectonic stability for a nuclear waste repository in Columbia River basalt in the reference repository location at the Hanford Site.

CHAPTER 2 - GEOLOGIC SUMMARY OF THE COLUMBIA PLATEAU

J. A. Caggiano

INTRODUCTION

The Columbia Plateau, a flood basalt province, is underlain by Miocene tholeiitic basalts that were erupted from north-northwest trending linear vent systems exposed mostly in southeastern Washington, northeastern Oregon, and west-central Idaho (Waters, 1961; Taubeneck, 1970; Swanson and others, 1977). These highly fluid lavas spread great distances from their source vents and flowed generally westward along a gentle slope before consolidating in a basin between the ancestral Cascade and Rocky Mountain Ranges in the interval 16.5 to 6 m.y. ago. Some voluminous flows extended down an ancestral Columbia River gorge and are now found along the Oregon coast (Beeson and Moran, 1979). The total areal extent of the Columbia River basalt is ~200,000 km^2. The basalt reaches a maximum known thickness of >3 km in the Pasco Basin (Reidel and others, 1982).

The Columbia Plateau occurs on the western margin of the North American plate (Fig. 2-1) in the Cordilleran orogen (the area west of the Rocky Mountain Front). The western margin of the plateau is >250 km east of the consumptive plate boundary between the North American and Juan de Fuca plates. The Juan de Fuca plate is a relict of the larger Farallon plate, which has been subducting beneath the North American plate during the Mesozoic and Cenozoic Eras (Atwater, 1970). The plateau occurs east of the north-south-trending Cascade Range, which is a calc-alkalic volcanic arc arising from partial anatexis of oceanic crust of the Juan de Fuca plate where it descends beneath continental crust of the North American plate. Lithospheric plate interactions are reviewed in Atwater (1970) (Fig. 2-2) and Coney (1978). Effects of plate and microplate interactions with emphasis on the Cenozoic Era (particularly the last 15 m.y.) are tabulated and discussed briefly in Chapter 7.

Basalt of the Columbia River Basalt Group has been divided into five formations, based on field mapping, chemical composition, and paleomagnetic polarity (Fig. 2-3; ARHCO, 1976; Swanson and others, 1979b). The Imnaha and Picture Gorge Basalts are the oldest units and have been mapped in relatively limited areas in northeastern and north-central Oregon, respectively. Imnaha Basalt has also been mapped in southeastern Washington and west-central Idaho. The Grande Ronde Basalt was erupted in the interval 16 to 14.5 m.y. ago and represents >85% of the total volume of Columbia River Basalt Group. The Wanapum Basalt, which forms the surface of much of the plateau, was erupted between 14.5 and 13.6 m.y. ago and consists of four members. The overlying Saddle Mountains Basalt was erupted intermittently between 13.5 and 6 m.y. ago. The Saddle Mountains Basalt is volumetrically small and its flows, for the most part, are areally limited to the central Columbia Plateau. Flows of both the Wanapum and Saddle Mountains Basalts in the Pasco Basin are in places separated by tuffaceous clastic sediments.

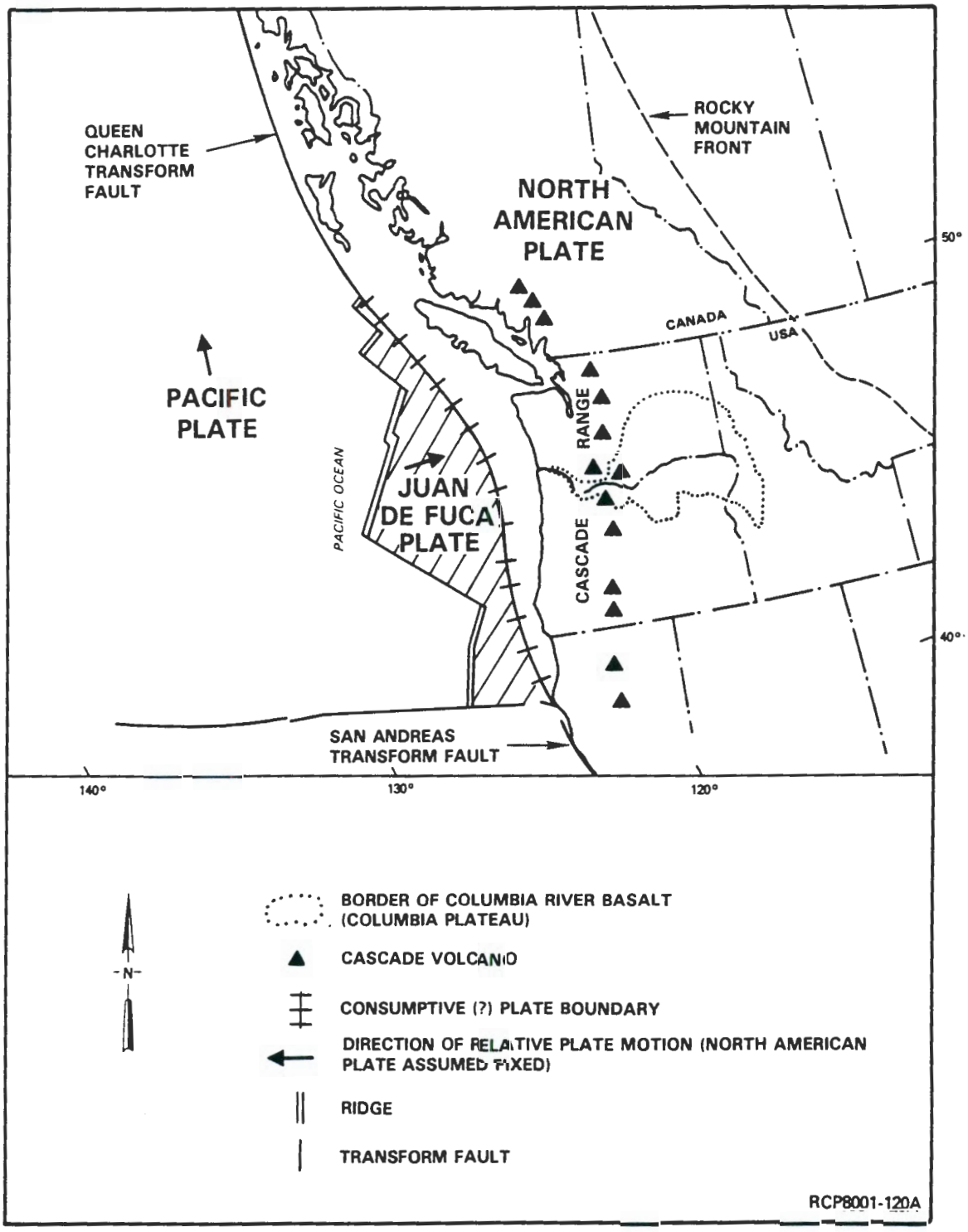

FIGURE 2-1. Tectonic Setting of the Columbia Plateau (after Riddihough, 1978).

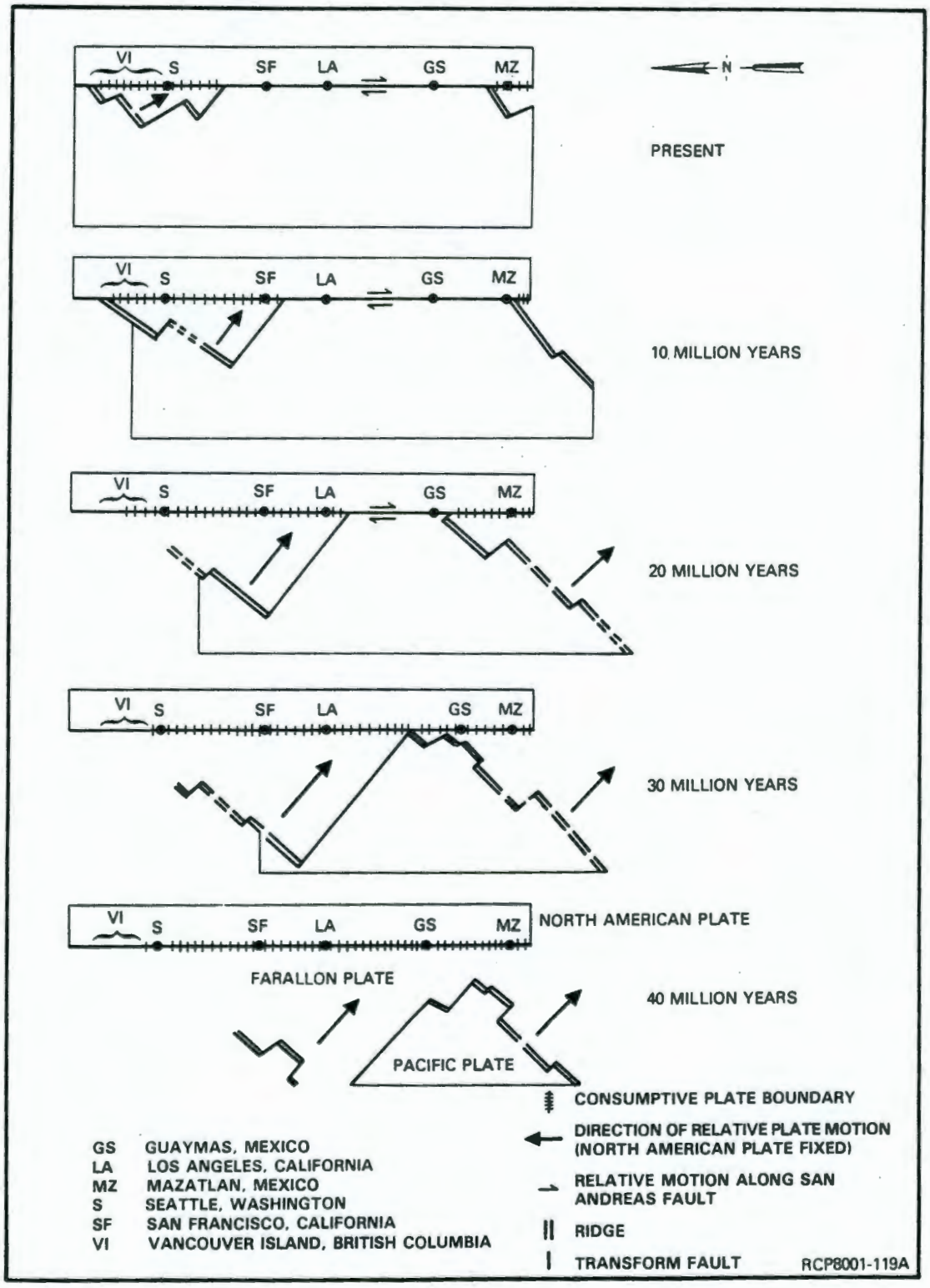

FIGURE 2-2. Schematic Illustration of Lithospheric Plate Interactions in Western North America During Part of the Cenozoic Era (after Atwater, 1970).

SERIES		GROUP	SUB-GROUP	FORMATION	MEMBER	K-Ar AGE (10^6 yr)	MAGNETIC POLARITY
MIOCENE	Upper Miocene	COLUMBIA RIVER BASALT GROUP	YAKIMA BASALT SUBGROUP	Saddle Mountains Basalt	LOWER MONUMENTAL MEMBER	6[a]	N
					EROSIONAL UNCONFORMITY		
					ICE HARBOR MEMBER		
					BASALT OF GOOSE ISLAND	8.5[a]	N
					BASALT OF MARTINDALE	8.5[a]	R
					BASALT OF BASIN CITY	8.5[a]	N
					EROSIONAL UNCONFORMITY		
					BUFORD MEMBER		R
					ELEPHANT MOUNTAIN MEMBER	10.5[a]	N,T
					EROSIONAL UNCONFORMITY		
	Middle Miocene				POMONA MEMBER	12[a]	R
					EROSIONAL UNCONFORMITY		
					ESQUATZEL MEMBER		N
					EROSIONAL UNCONFORMITY		
					WEISSENFELS RIDGE MEMBER		
					BASALT OF SLIPPERY CREEK		N
					BASALT OF LEWISTON ORCHARDS		N
					ASOTIN MEMBER		N
					LOCAL EROSIONAL UNCONFORMITY		
					WILBUR CREEK MEMBER		N
					UMATILLA MEMBER		N
					LOCAL EROSIONAL UNCONFORMITY		
				Wanapum Basalt	PRIEST RAPIDS MEMBER		R_3
					ROZA MEMBER		T,R_3
					FRENCHMAN SPRINGS MEMBER		N_2
					ECKLER MOUNTAIN MEMBER		
					BASALT OF SHUMAKER CREEK		N_2
					BASALT OF DODGE		N_2
					BASALT OF ROBINETTE MOUNTAIN		N_2
	Lower Miocene			Grande Ronde Basalt		14-16.5[b]	N_2
							R_2
				Picture Gorge Basalt[d]		(14.6-15.8)[b,c]	N_1
				?			R_1
				Imnaha Basalt[d]			R_1
							T
							N_0
							R_0?

[a]DATA FROM McKEE AND OTHERS (1977)
[b]DATA MOSTLY FROM WATKINS AND BAKSI (1974)
[c]INFORMATION IN PARENTHESES REFERS TO PICTURE GORGE BASALT
[d]THE IMNAHA AND PICTURE GORGE BASALTS ARE NOWHERE KNOWN TO BE IN CONTACT. INTERPRETATION OF PRELIMINARY MAGNETOSTRATIGRAPHIC DATA SUGGESTS THAT THE IMNAHA IS OLDER.

RCP8202-49A

FIGURE 2-3. Regional Columbia River Basalt Group Stratigraphic Nomenclature (after Swanson and others, 1979b).

PASCO BASIN AND REFERENCE REPOSITORY LOCATION GEOLOGIC SUMMARY

The Pasco Basin, in which the reference repository location has been sited, is one of several Tertiary synclinal basins in the central Columbia Plateau that contain Neogene and Quaternary sediments. These basins are bordered mostly by anticlinal ridges. Grande Ronde Basalt is the oldest formation of the Columbia River Basalt Group known in the Pasco Basin and is overlain by Wanapum and Saddle Mountains Basalts. Intercalated with the basalts are fluvial, tuffaceous sediments of the Ellensburg Formation, which are common in the Wanapum and Saddle Mountains Basalts. These basalts are in turn overlain by mostly fluvial sediments of the late Miocene and Pliocene Ringold and Quaternary Hanford formations. Narrow, asymmetrical anticlines of the Yakima Fold Belt border the basin to the north (Saddle Mountains), west (Umtanum and Yakima Ridges), and the southwest (Rattlesnake Hills) (see Fig. 1-1). Structural relief on these folds decreases eastward from the western margin of the basin where the axes of Umtanum and Yakima Ridges plunge beneath the basin. Basalt crops out in the basin on Gable Butte and Gable Mountain, eastward continuations of Umtanum Ridge, and at Horn Rapids Dam on the Yakima River. Yakima Ridge extends eastward where its extension is recognized in the subsurface from drilling and geophysical data. The eastern margin of the Pasco Basin is essentially the north-trending Jackass monocline, east of where the basalt thins and rises gently eastward onto the Palouse slope. Thrust and reverse faults occur on steeply dipping limbs of the anticlinal folds; minor folds and faults oblique to the axes of first- and second-order folds occur on gentle dip slopes of these structures. Details of the geology of the Pasco Basin along with geologic maps can be found in Myers/Price and others (1979).

The reference repository location is in the axial region of the western part of the southeast-trending Cold Creek syncline, one of several synclines in the Pasco Basin. The Cold Creek syncline is asymmetrical with strata rising more gently to the north on the south limb of the Umtanum Ridge-Gable Mountain structure and more steeply onto the buried extension of Yakima Ridge to the south. The axial region is a broad, open fold where strata are nearly flat lying (Fig. 2-4). Two depressions in the top of basalt occur at the western and eastern ends of the Cold Creek syncline (Myers, 1981). Second- and third-order folds are present on the large first-order folds, which bound the syncline. Borehole data and geophysical anomaly patterns suggest the possibility of northwest- and, possibly, northeast-trending structures oblique to the axis of the Cold Creek syncline as well as small monoclinal structures along the syncline subparallel to its axes. These possible cross structures have been used to separate the "inner" part of the syncline into large areas in which basalt bedrock is apparently free of major structures. The reference repository location is in one of these masses of intact rock. Within the reference repository location, thick Grande Ronde Basalt flows with dense flow interiors such as the Umtanum, McCoy Canyon, and the Middle Sentinel Bluffs flows are considered prime candidate horizons in which a repository could be constructed. Details of the geology of the Cold Creek syncline can be found in Myers and Price (1981). Characteristics of the Grande Ronde Basalt are described by Long and Landon (1981) and Long and Davidson (1981).

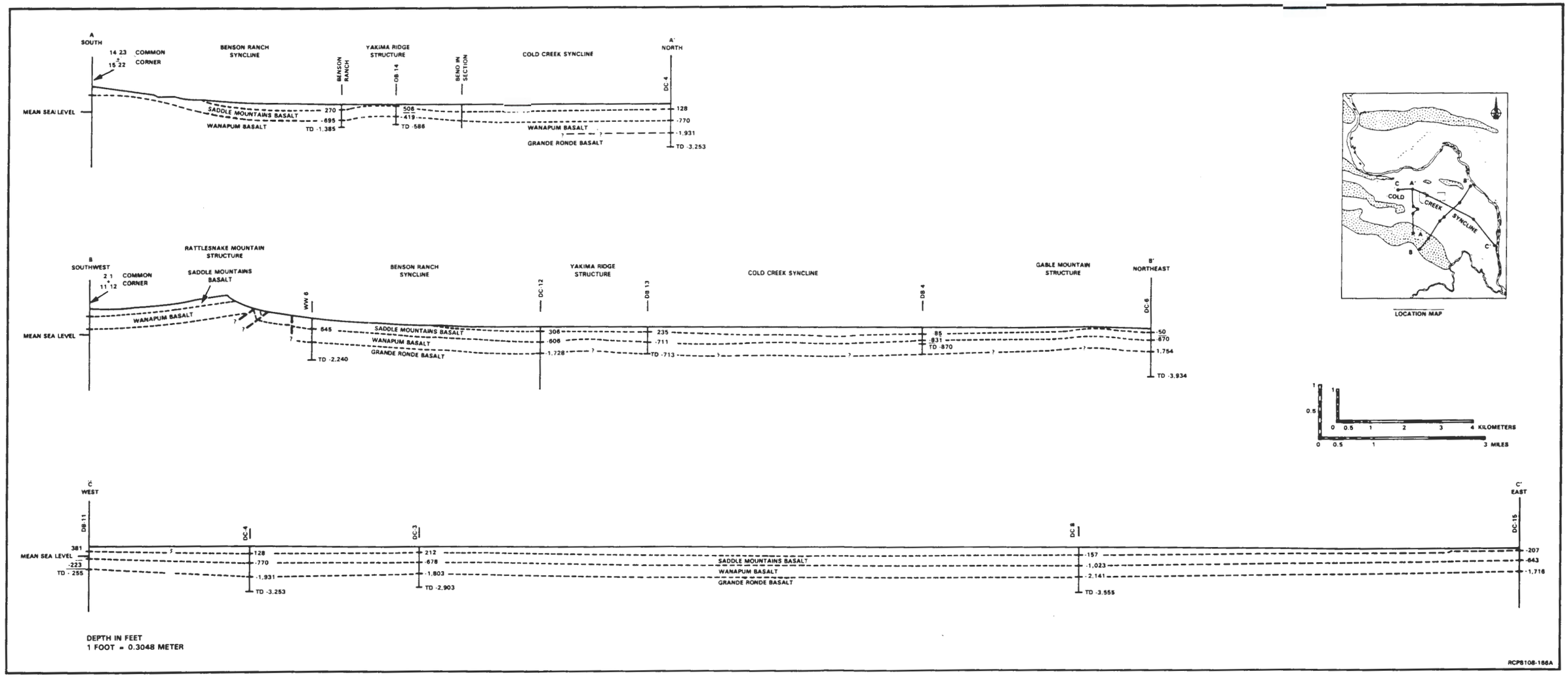

FIGURE 2-4. Structure Cross Sections of the Cold Creek Syncline.

STRUCTURE OF THE CRUST IN THE COLUMBIA PLATEAU

A long, north-south refraction line from British Columbia to Oregon through the eastern Pasco Basin indicates that the crust of the continent in eastern Washington thins southward from ~35 km near the granitic Okanogan Highlands to ~25 to 28 km beneath the Pasco Basin before thickening to the south in the Blue Mountains (Hill, 1972, 1978). Velocity models derived from compressional wave traveltimes suggest the same trends (UWGP, 1977; Eaton in Hill, 1978) as do magnetotelluric data (Myers/Price and others, 1979) and inversion of gravity and traveltime residuals (Rodi and others, 1980). Recent magnetotelluric data support a thin crust beneath the central Columbia Plateau, but with substantial relief of the geoelectric basement (Chapter 4). North-south-trending, regional gravity gradients west of the Pasco Basin at about 120° W. and 121° W. longitude may reflect changes in thickness of Columbia River basalt (Koníček, 1974; Robbins and others, 1975). The thin crust of the plateau contrasts with a substantially thicker crust of ~40 km beneath Puget Sound west of the Cascade Range (Crosson, 1972, 1976).

Although gravity anomalies are negative over all of the Columbia Plateau, there is a relative gravity high beneath the Pasco Basin (Weston, 1981) (Fig. 2-5) supporting the thin crust interpretation of refraction, seismic, and magnetotelluric data. Beneath the Columbia River basalt, the crust may be continental (Laubscher, 1981) or oceanic (Davis, 1977). However, the basalt has yet to be penetrated by boreholes except near the margins where basalt is thinner than the 3.2-km thickness in the Rattlesnake Hills well in the Pasco Basin (Reidel and others, 1982). Exploratory drilling by Shell Oil Company at Roza Dam north of Yakima and on Whiskey Dick Ridge between Ellensburg and Vantage has presumably penetrated the basalt in a search for hydrocarbons in unknown, but probably sedimentary, rock types beneath basalt.

The thin crust of the Columbia Plateau is divisible seismically and electrically into strata of different compressional wave velocity and conductivity. Velocities indicate the existence of three crustal units (layers) north of the Frenchman Hills (see Fig. 1-1), while to the south the central Columbia Plateau is divisible into five units (UWGP, 1977) overlying the mantle. Resistivity data from magnetotelluric surveys indicate that there are five layers (including the mantle) in the Pasco Basin area of the central Columbia Plateau (Myers/Price and others, 1979; Chapter 4). These crustal units are summarized in Figure 2-6.

Unlike the Great Basin in the Basin and Range province where a thin crust is accompanied in places by high heat flow (Eaton and others, 1978), the thin crust beneath the central Columbia Plateau is an area of relatively low heat flow (Blackwell, 1978) (Fig. 2-7). Heat flow in the southern Cascade Range west and south of the plateau is high (Blackwell, 1978). Direct measurements of heat flow in the Columbia Plateau are sparse, but geothermal gradients have been logged or calculated from spot measurements of fluid temperature in several deep wells on the Hanford Site. The sparse data suggest a varying geothermal gradient ranging from ~0.8° to ~1.5°C/30 m (29° to 50°C/km). The wide range in reported gradients is interpreted as a reflection of measurement techniques as well as groundwater circulation.

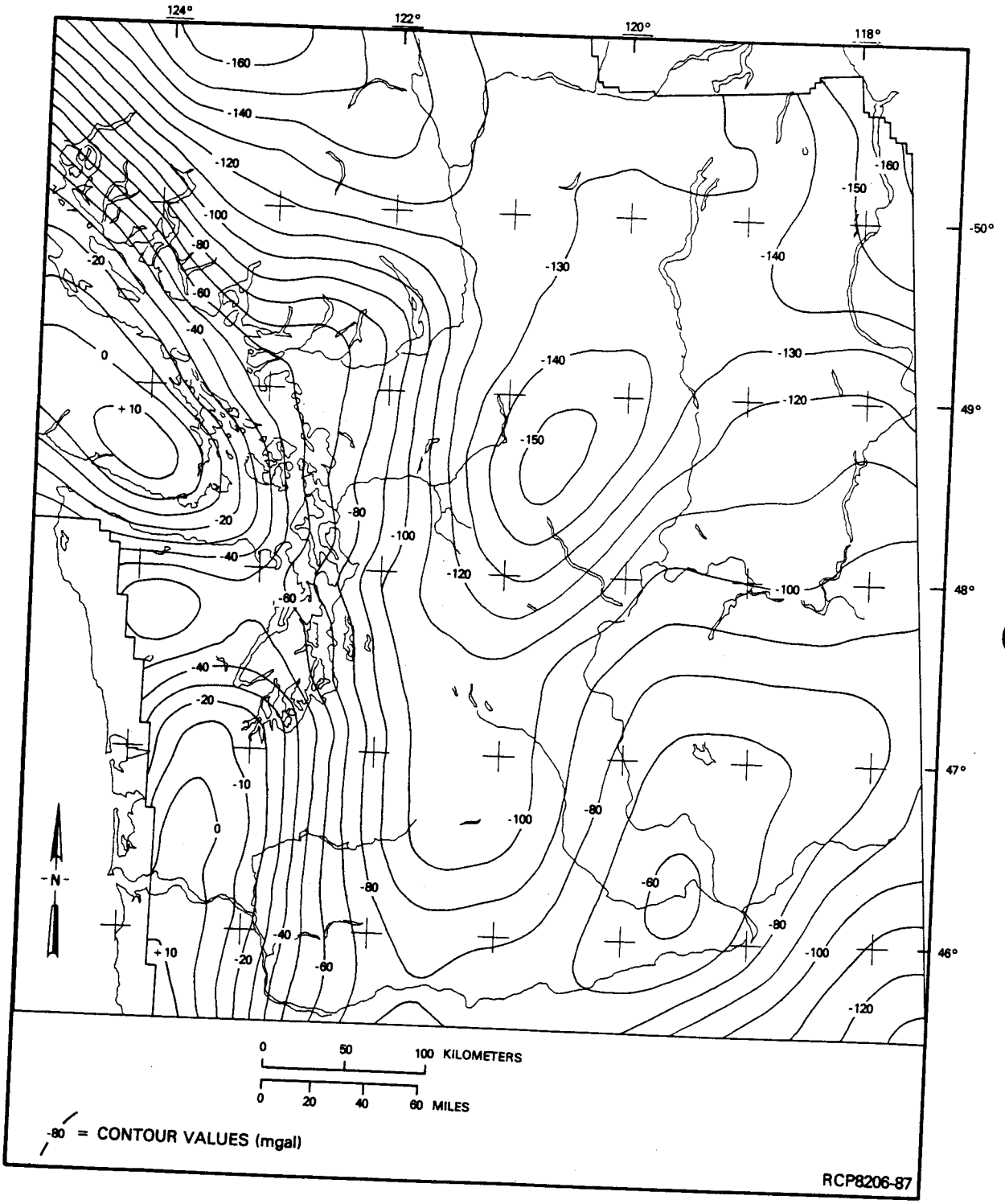

FIGURE 2-5. Regional Gravity Map of the Western United States and Canada (after Weston, 1981; WPPSS, 1981).

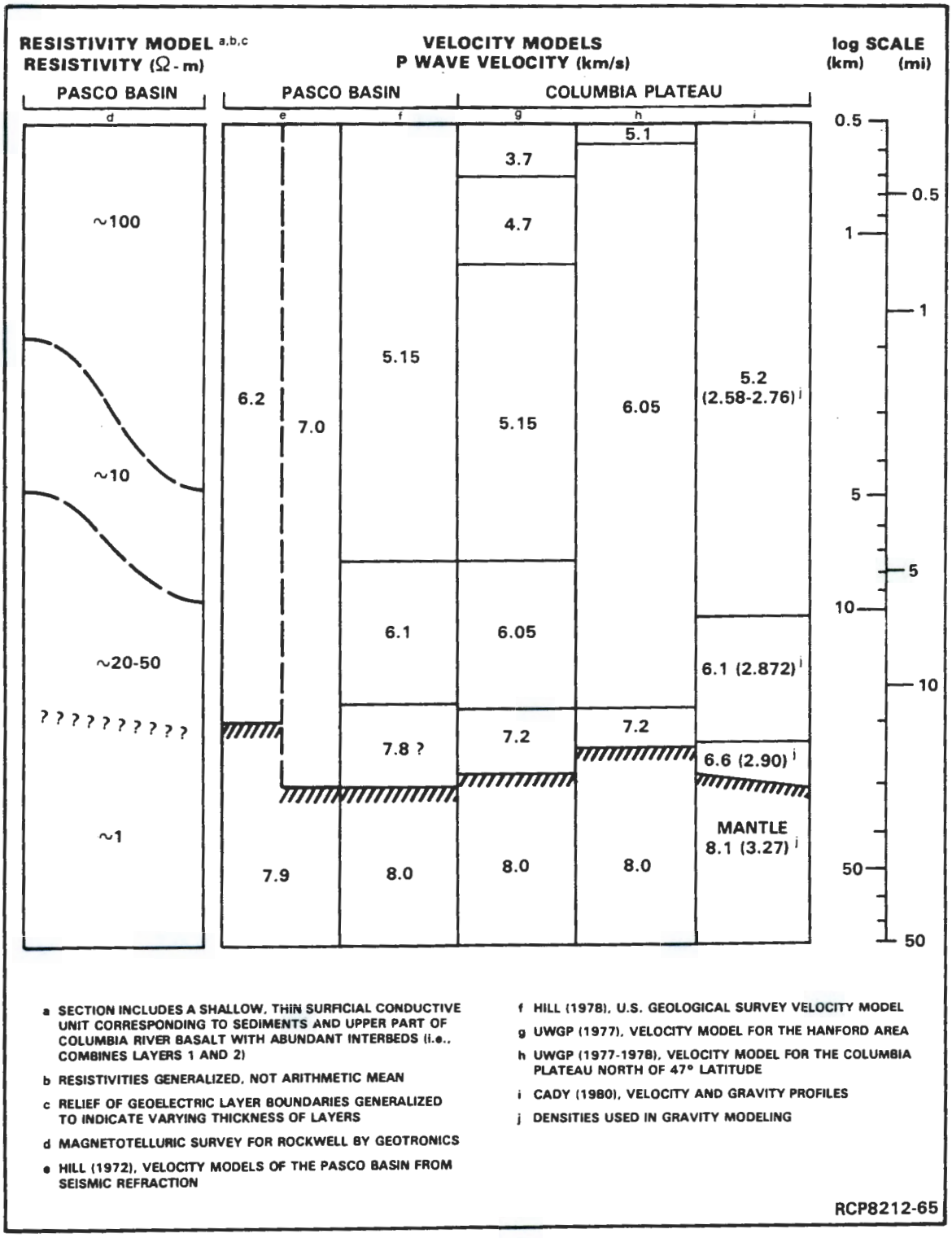

FIGURE 2-6. Crustal Models for the Pasco Basin and Columbia Plateau.

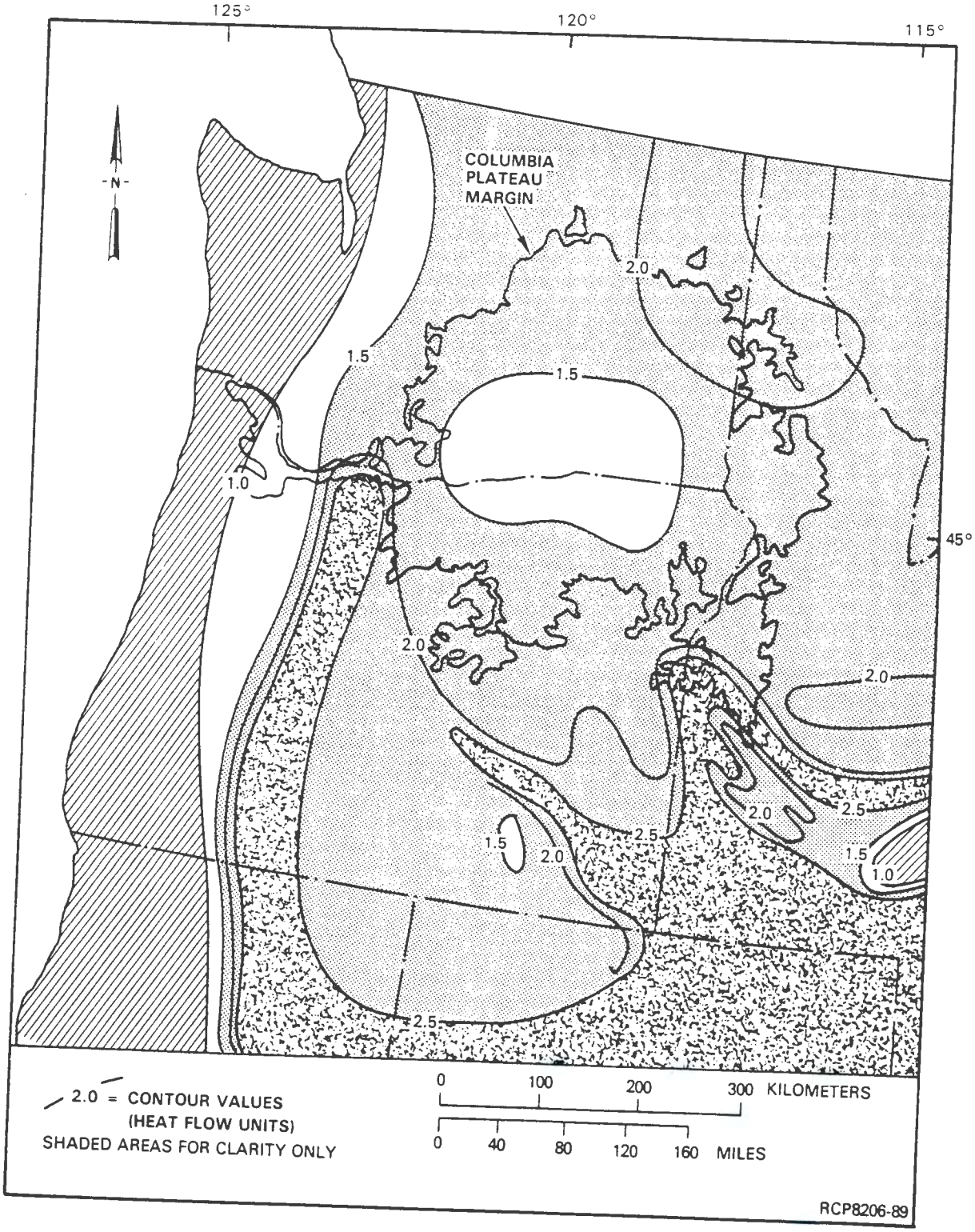

FIGURE 2-7. Heat Flow in the Northwestern United States (after Blackwell, 1978).

The Columbia Plateau is an area of low seismicity where the relatively few, generally small, crustal earthquakes have not been related to mapped faults (Fig. 2-8 and 2-9). Swarms of generally shallow events have been recorded on the flanks of the Saddle Mountains, but they have not been associated with known geologic faults. By contrast, earthquakes in the Puget Sound area range up to magnitude 7.0, are generally larger and more frequent, and occur to depths of ~70 km (Crosson, 1972, 1976). Historically, there have been a few moderate earthquakes in the plateau in the 140 yr of earthquake records (Rasmussen, 1967). Events of modified Mercalli intensity VII have been reported in Umatilla, Oregon in 1893 and in the Milton-Freewater, Oregon area in 1936 (Coffman and von Hake, 1973; Chapter 6). The nearest moderate-sized historic event to the reference repository location was a November 1, 1918 event near Corfu, Washington, which is currently listed as a modified Mercalli intensity V to VI (Coffman and von Hake, 1973). Instrumentally monitored microseismicity in the plateau is discussed in Chapter 6.

STRUCTURAL DEFORMATION IN THE COLUMBIA PLATEAU

Columbia River basalt has been folded and faulted in Miocene and younger time (Reidel and others, 1980; Reidel and Fecht, 1981, 1982; Chapter 5; Plates 2-1 and 2-2). Differences in the style of deformation have led to a subdivision of the Columbia Plateau into informal subprovinces whose boundaries are not clearly defined (Myers/Price and others, 1979; Figure 2-10). Most of the known deformation has occurred in the Yakima Fold Belt (Bentley and others, 1980; Bentley, 1977; Grolier and Bingham, 1978) and the Blue Mountain subprovinces (Swanson and others, 1977; Ross, 1978; Ross and Reidel, 1982; Camp, 1981; Camp and Hooper, 1981; Hooper and Camp, 1981). Only minor deformation has been mapped on the Palouse subprovince (Griggs, 1973; Swanson and others, 1977).

Two tectonic maps (Plates 2-1 and 2-2) have been prepared to illustrate deformation in the Columbia Plateau. Geologic structures are from geologic maps of the Columbia Plateau prepared by Rockwell (Myers/Price and others, 1979) for the Pasco Basin and mapping and compilation by the U.S. Geological Survey (Swanson and others, 1979a, 1980, 1981) for the BWIP. For the entire Columbia Plateau, a 1:500,000 scale base was chosen (Plate 2-1); for the Pasco Basin area, structures are shown at 1:250,000 scale. Summaries of work on structures in the Yakima Fold Belt are in Myers/Price and others (1979) and Myers and Price (1981). General characteristics of structures in each of the three structural subprovinces follows.

Yakima Fold Belt Subprovince

Asymmetric anticlines with wavelengths of a few kilometers and amplitudes of <1 km are characteristic of the Yakima Fold Belt in the western Columbia Plateau (see Fig. 2-10). Synclines are commonly broad, sediment-filled valleys or basins containing several synclines such as the Pasco Basin. Anticlinal axes trend generally east-west, but range from west-southwest to northwest. The folds extend westward to the eastern flank

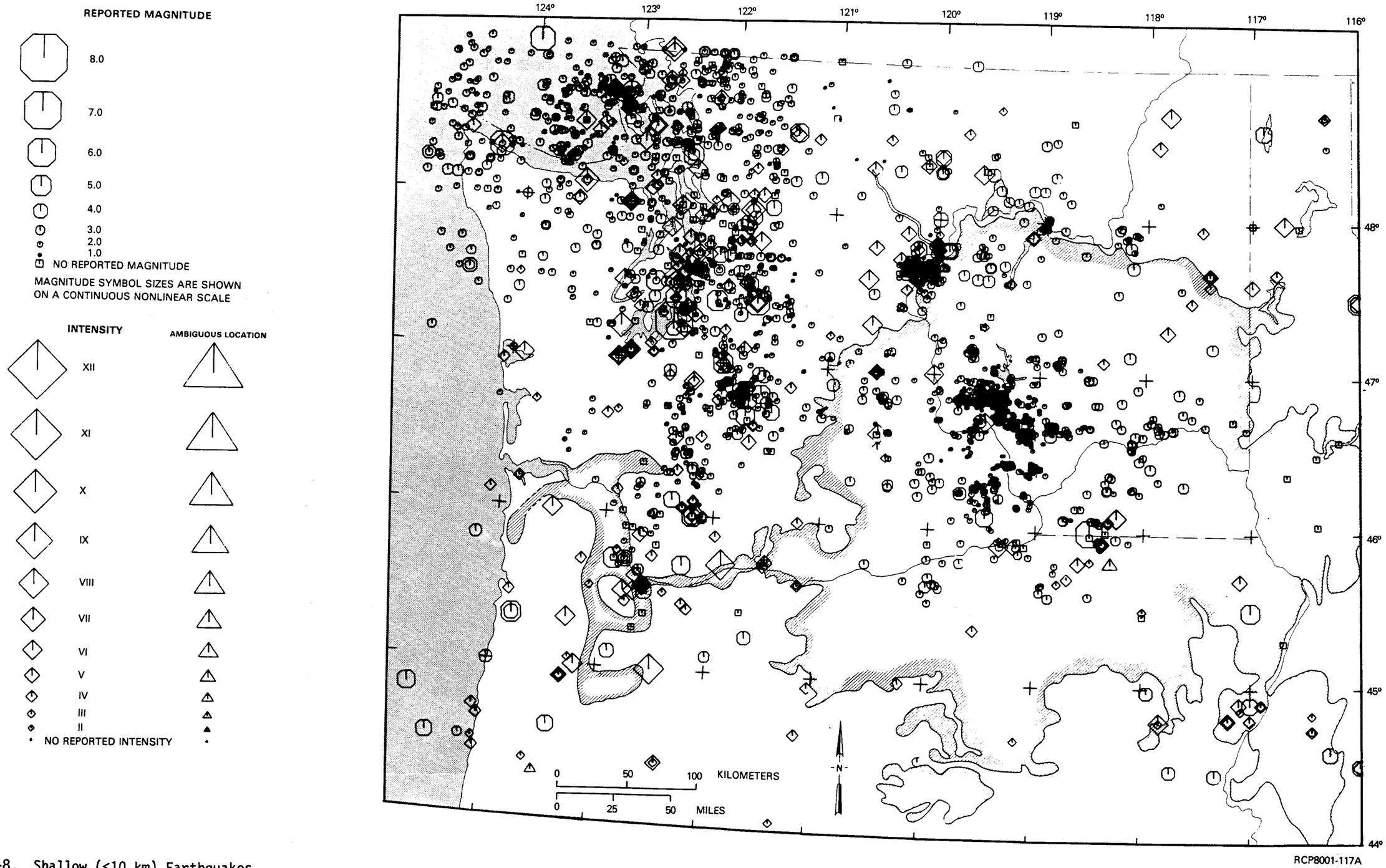

FIGURE 2-8. Shallow (<10 km) Earthquakes in the Pacific Northwest, 1965-1979.

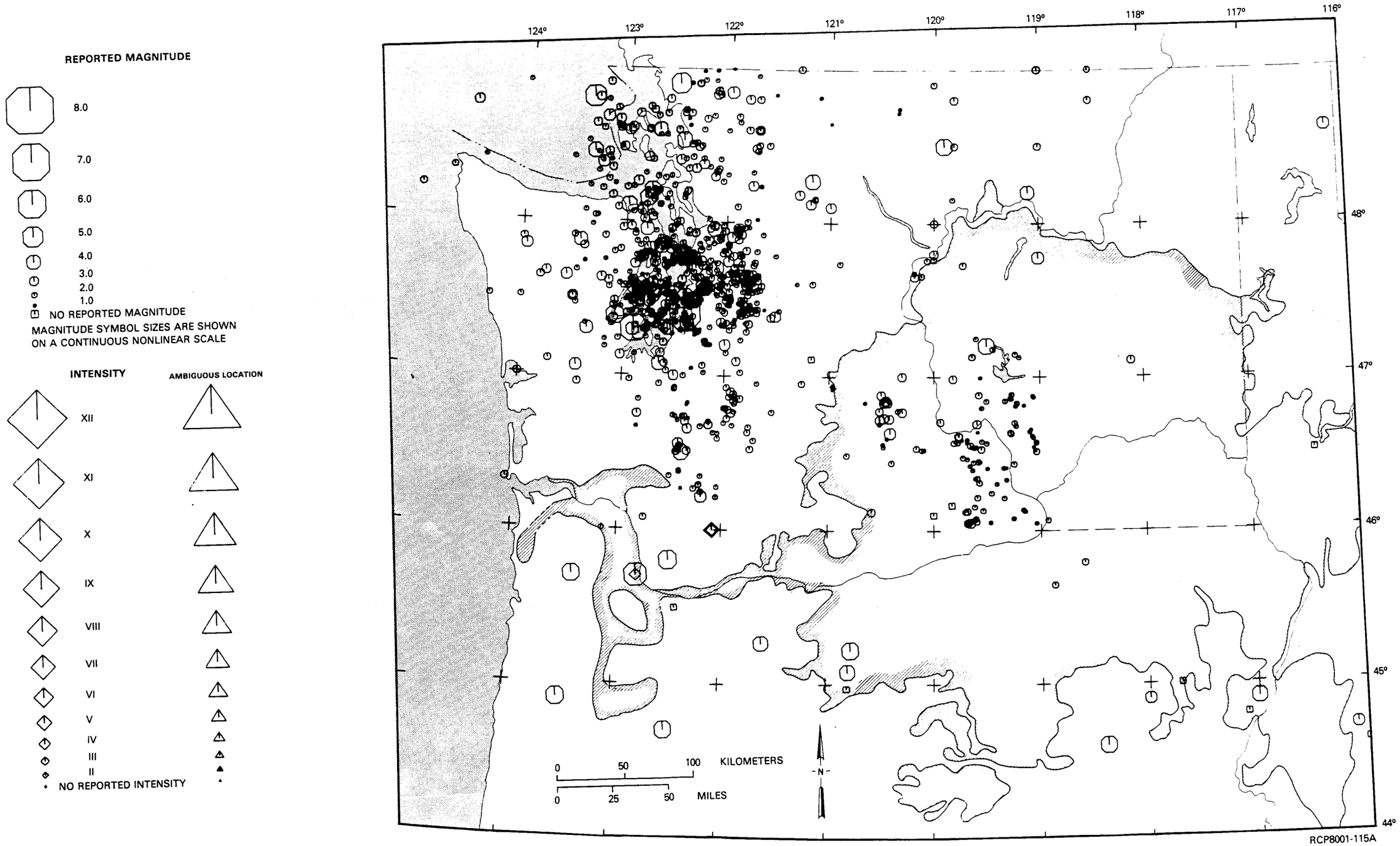

FIGURE 2-9. Deep (>10 km) Earthquakes in the Pacific Northwest, 1965-1979.

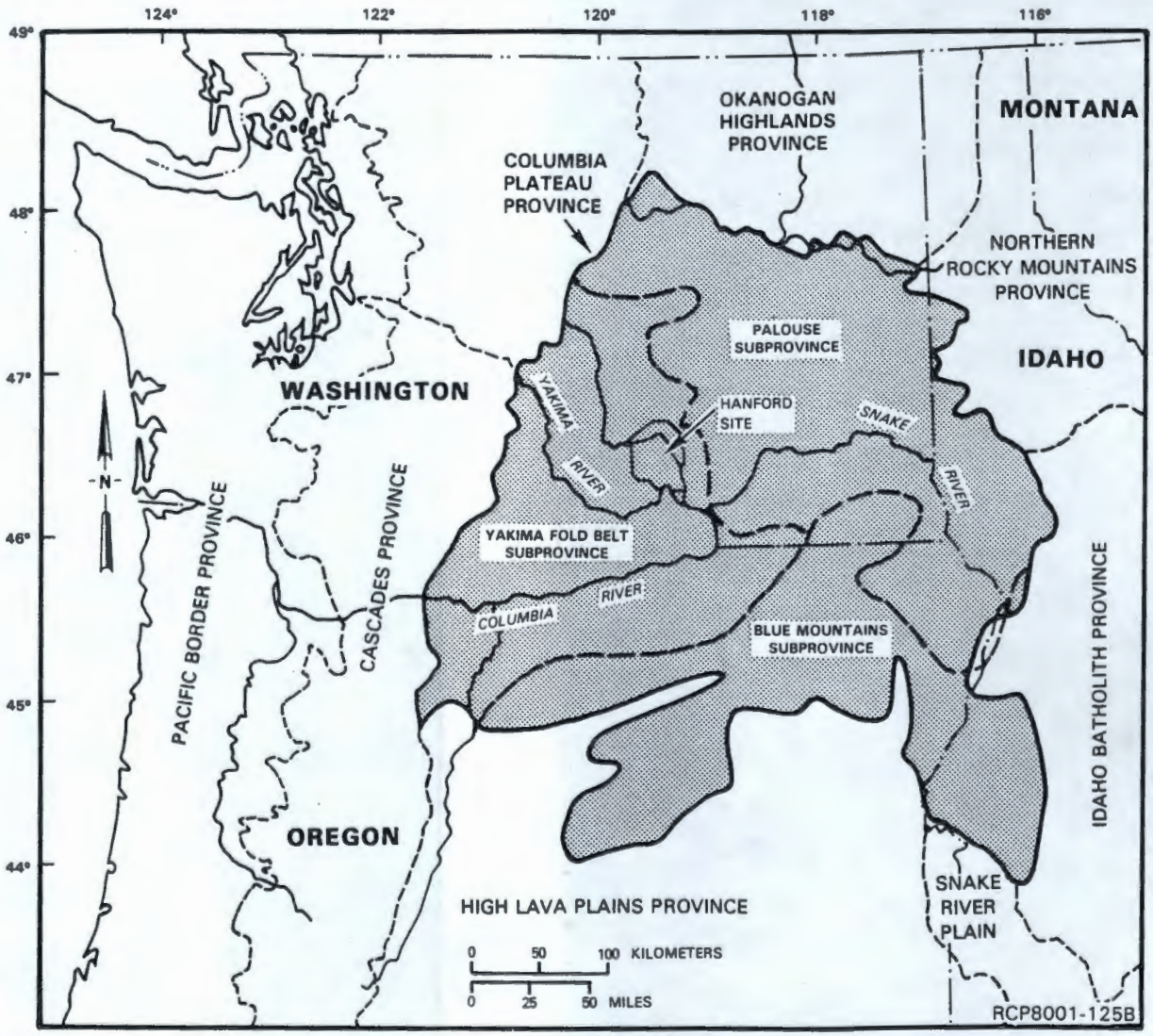

FIGURE 2-10. Tectonic Provinces and Subprovinces of the Pacific Northwest.

of the Cascade Range where, in some cases, the fold axes change strike. Eastward, the folds decrease in amplitude and are unrecognizable on the Palouse slope. Axes of some folds of eastward-diminishing amplitude plunge beneath Pliocene and younger sediments in late Tertiary basins of the central Columbia Plateau.

In cross section, asymmetrical anticlines of the Yakima Fold Belt commonly have a steeper north limb and a long, gentle, dip slope on the south limb. However, in some anticlines, the south limb is steeper, while in others, the asymmetry changes along strike from a steep north flank to a steep south flank, often where north-south or northwest-trending folds or steeply dipping faults separate segments of the anticlines with different cross-sectional geometry (e.g., in the Saddle Mountains (Reidel, 1978), Gable Mountain (Fecht, 1978), and along Toppenish Ridge (Bentley and others, 1980)). Thrust and reverse faults with fault planes striking subparallel to fold axes are common on the steeper limbs and are interpreted to reflect folding strain (Price, 1982). Estimates of displacement on these dip-slip faults range from a few meters to tens of meters to, in places, a few hundred meters (Gardner, 1977; Reidel, 1978; Gardner and others, 1981; PSPL, 1982). Northwest- and northeast-trending, steeply dipping faults have been mapped with several meters to several tens of meters of dip slip displacement (Bentley and others, 1980; Gardner and others, 1981; Swanson and others, 1979a). Strike-slip displacement of varying amounts has been postulated for these northeast- and northwest-trending faults along the Columbia gorge near The Dalles, Oregon (Plate 2-1). The length, straightness, seeming reversal of dip slip along strike, and the spatial relationship to the axes of folds and other faults suggest that they are wrench faults (Bentley, 1980; Anderson, in press). Without steeply dipping strata or marker horizons, the amount of strike-slip on these faults, if any, has not been firmly established. Smaller, second- and third-order folds with axes subparallel or oblique to first-order fold axes may be present. Low amplitude, third-order folds of northwest trend have been mapped on the long backslopes of folds (Myers/Price and others, 1979). Possible Holocene normal faults subparallel to ridge crests have been interpreted on Toppenish Ridge >50 km west of the Pasco Basin (Campbell and Bentley, 1981; WCC, 1981c).

The long narrow anticlines of the Yakima Fold Belt are separated by broad synclines with limited bedrock exposure because of overlying Neogene and Quaternary sediments. Consequently, information on synclines in the central plateau is based primarily on interpretation of borehole and geophysical data. Cross sections of synclines in the Pasco Basin, where there is subsurface data, indicate that strata have low dip and are minimally deformed (Myers/Price and others, 1979; Myers and Price, 1981).

A northwest-trending topographic lineament >400 km long is well expressed in the Yakima Fold Belt and may express structural development. Originally noted and described on small-scale physiographic and topographic maps by Raisz (1945), who named it the Olympic-Wallowa lineament, this feature is displayed as a zone of comparatively greater deformation in the Yakima Fold Belt from Cle Elum to Wallula Gap. This segment of the Olympic-Wallowa lineament, called the Cle Elum-Wallula zone of deformation (Kienle and others, 1978), is a narrow zone in which folds change trend

from west-northwest or east-west to northwest and in which faults, especially of northwest trend, may be more common. In the Pasco Basin, this trend is topographically well displayed by the steep northeast face of Rattlesnake Mountain* and a northwest-trending series of brachyanticlines between Rattlesnake Mountain and Wallula Gap. East of Wallula Gap, the west-northwest-trending fault(s) of the Wallula fault system, which are approximately parallel to those of the Brothers fault zone (Chapter 3), continue along this trend. Regional gravity anomalies do not support the existence of a boundary between oceanic and continental crust along the Olympic-Wallowa lineament as suggested by Wise (1963) and Skehan (1965). Swanson (in Zietz and others, 1971) interpreted magnetic anomalies from a high-altitude aeromagnetic survey as suggestive of a continental/oceanic crustal boundary along the Olympic-Wallowa lineament (following Skehan, 1965), but magnetotelluric data do not indicate the juxtaposition of rocks of highly contrasting resistivity along Rattlesnake Hills. Structural interpretations of the Cle Elum-Wallula zone of deformation are discussed further in Chapter 7.

Blue Mountains Subprovince

The Blue Mountains are a broad anticlinal uplift that extends east-northeast from the foothills of the Cascades to west of Pendleton, Oregon where the range turns to assume a more northeast trend. The Blue Mountains are the major geologic structure in the Blue Mountains subprovince (see Fig. 2-10). The Blue Mountains anticlinorium (Newcomb, 1970) is very broad compared with anticlines in the Yakima Fold Belt and is generally asymmetrical with flows dipping generally up to 8° on the north flank and up to 3° south on the south flank (Kendall, 1981). Strata in the broad axial region dip very gently and appear almost undeformed. The geometry of the Blue Mountains anticlinorium suggests that it has resulted from uplift of a structural block of Mesozoic and older rocks in the basement beneath the capping basalt. The Blue Mountains apparently began uplifting in late Grande Ronde time (Ross, 1978) as they are a source for Neogene sediments in basins to the north (Farooqui and others, 1981). They also strongly influenced the distribution of Columbia River basalt (Ross, 1978; Camp and Hooper, 1981). Thus, deformation of Columbia River basalt and uplift of the Blue Mountains were under way by middle Miocene time with coincident folding and faulting (Pigg, 1961; Ross, 1978; Shubat, 1979).

The broad east-west-trending Blue Mountains uplift has been affected by northwest-trending folds and faults with dip slip displacement that are especially prominent near La Grande, Oregon where a graben occurs (Walker, 1977; Kienle and others, 1979; Barrash and others, 1980). Northeast-trending folds and faults are also present. Gehrels and others (1980) interpret the La Grande Basin as a rhomb graben formed by dextral slip on a N. 30° W.-trending fault system.

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*In this report, the segment of the Rattlesnake Hills between the change in azimuth of the Rattlesnake Hills at Snively Basin to the Yakima River will be termed Rattlesnake Mountain to distinguish it from other parts of the Rattlesnake Hills.

The thick section of Grande Ronde Basalt in the Stites Basin along the margin of the Idaho batholith attests to its early development during Grande Ronde time, perhaps by movement on the north-south trending Stites fault (Camp and Hooper, 1981). Large north-south faults in pre-basalt rocks south of the Stites Basin (Plate 2-1) that continue beneath the basalt (Hamilton, 1962) suggest reactivation as a possible cause for development of the Stites Basin. Other basins such as the Troy and Lewiston Basins (Plate 2-1) are bounded by the northeast-trending Limekiln fault or its monoclinal extension to the northeast. Distribution of Grande Ronde flows suggest development of the Limekiln fault (Plate 2-1) and the Lewiston (Reidel, 1978) and Troy Basins (Ross, 1978) by late Grande Ronde time, with continued development of these basins during Wanapum, Saddle Mountains, and later time (Price, 1977; Ross, 1978; Reidel, 1978; Camp and Hooper, 1981).

North-northeast trending, steeply dipping faults along the northwest edge of the Blue Mountains between Prineville, Oregon and the Snake River in Washington belong to the Hite fault system (Newcomb, 1970; Kienle and others, 1979; Kienle, 1980). Other subparallel faults (Plate 2-1) that trend essentially north-south have a similar geologic history and are included as part of the Hite fault system (Plate 2-1; Kienle, 1980). Two major episodes of movement have occurred on the Hite fault: (1) primarily down-to-the-west dip-slip motion in pre-Frenchman Springs time, and (2) primarily strike-slip motion in post-Frenchman Springs time (Kienle and others, 1979; Kendall, 1981). The Hite fault is expressed as a gouge zone of up to 150 m wide or as a stratigraphic discontinuity of up to 120 m of pre-Frenchman Springs dip slip (Kendall, 1981). Kienle and others (1979) feel that dextral movement has occurred on the Hite fault in post-Frenchman Springs time, but Kendall (1981) considers motion on the fault in post-Frenchman Springs time to be sinistral and the Hite fault to be one member of a conjugate fault system (the La Grande fault system being the dextral member).

Different thicknesses of Grande Ronde Basalt on opposite sides of the Hite fault suggest uplift had begun in pre-Wanapum time (Kendall, 1981). Kienle and others (1979) consider the Hite fault to be older than the west-northwest-trending Wallula fault, which they suggested was capable by U.S. Nuclear Regulatory Commission criteria (NRC, 1980b) because of offset of late Pleistocene units. Trenching at Warm Springs at a locality known as "Bingham's linear" (Bingham and others, 1970) along the Wallula fault suggests that the most recent movement pre-dates the deposition of late Wisconsinan rhythmites (known informally as the Touchet member of the Hanford formation) and may be Sangamonian or pre-Wisconsinan (i.e., ~70,000 yr before present; WCC, 1981d). The Hite fault may be the terminus of the topographic feature described by Raisz (1945) as the Olympic-Wallowa lineament. Davis (1981) suggested that this topographic feature and the Wallula fault system may be terminated at the Hite fault.

The Palouse Subprovince

The major feature of the Palouse subprovince (see Fig. 2-10) is a gentle westward slope that developed during the Miocene (Chapter 5). Superimposed on this area of very gently dipping (<5°) strata are long wavelength, low amplitude folds that trend north-northwest to northwest (Plate 2-1; Swanson and Wright, 1978).

Northwest-trending fractures spaced ~10 km apart and extending for ~40 km or more are found in the northeastern Columbia Plateau near Cheney, Washington (Griggs, 1973). Named (informally) the Cheney fracture zone, these fractures are broad, linear, topographic depressions believed to result from differential erosion along zones of closely spaced joints. These linear depressions continue on the trend of folds and faults that lie beyond the edge of the basalt (45 km or more away) and that are presumed to continue beneath the basalt (see Chapter 3 for a discussion of structural trends that may be present below the basalt). Minor adjustments of the thin basalt overlying basement structures presumably have resulted in development of zones of closely spaced fractures that have been eroded by catastrophic floods. Kiver and Stradling in Rigby and Othberg (1979) found no evidence of any late Quaternary movement along any of these fractures.

Linear Vent Systems

Grande Ronde, Wanapum, and Saddle Mountains Basalts were extruded from fissures in the eastern portion of the Blue Mountains and the Palouse subprovinces (Waters, 1961; Taubeneck, 1970; Price, 1977). Recently, a Grande Ronde dike has been found near Heppner, Oregon west of previously mapped source vents (Swanson and others, 1981). These linear vent systems are more prevalent in a zone extending from the eastern margin of the Pasco Basin to the eastern edge of the plateau. This zone is up to a few hundred kilometers long (Swanson and others, 1975; Taubeneck, 1970). Individual dikes range from a few centimeters to a few meters wide and trend north-northwest.

The consistent north-northwest trend of the dikes indicates that they are an important tectonic element that need to be considered in assessing structural mechanics and the development of tectonic models. Eruptions from north-northwest-trending fissures were occurring penecontemporaneously with development of folds in the Yakima Fold Belt and the Blue Mountains (Reidel and others, 1980; Ross, 1978). These folds suggest compression in a stress regime characterized by a nearly horizontal principal axis trending slightly west of north (Ross, 1978; Hooper and Camp, 1981), a trend which agrees with Nakamura's (1977) observation that dikes intrude along a direction parallel with the axis of maximum compression (i.e., σ_1). Contemporary deformation (Chapter 6) also seems to be occurring under a similar stress regime suggesting that nearly horizontal, nearly north-south compression has prevailed in the Columbia Plateau for at least the past 16.5 m.y.

DETERMINATION OF A CHRONOLOGY OF DEFORMATION

Using the thickness and distribution of lava flows, the development of structurally controlled topography can be demonstrated by at least middle Miocene time in the Columbia Plateau (Reidel and others, 1980; Camp and Hooper, 1981; Camp, 1981). Highly fluid Columbia River basalt lavas spread great distances from their linear vents along paleoslopes filling topographic lows and thinning over topographic highs. Each lava flow thus constitutes a "cast" of topography at an "instant" of Miocene time. Differences in thickness of noneroded basalt flows and their distribution provide a quantitative estimate of relief at the time of solidification. Lavas from eruptions of low volume commonly remained fairly close to sources and filled local basins and valleys; many late Miocene flows were channeled along the ancestral Snake River and ponded in more distant low areas such as the Pasco Basin. Flows of great volume spilled out of canyons and ponded in major basins so that major Miocene topographic highs are presently recognized by the relative thicknesses of flows.

Structural development in Pliocene and younger time has been assessed from the thickness and distribution of the terrestrial sediments overlying the basalt. Thick accumulations of sediments are present only in basins such as the Pasco, Umatilla, and Quincy basins, which are bordered by anticlinal folds. Differences in provenance and resulting lithic dissimilarity have made correlation of units between basins difficult, but some progress has been made (Myers/Price and others, 1979; Farooqui and others, 1981). The age of the sediments along with the attitude of bedding as the units thin onto anticlinal ridges suggest the location and timing of deformation. Lithic composition of gravels suggest changes in drainage patterns that probably reflect growth of structurally controlled topography. Since exposures are often poor along the flanks of anticlines, the relationship of sedimentary units to anticlinal ridges must be deduced from borehole data that are more abundant in the Pasco Basin.

Initiation of Deformation

Structurally controlled topography was apparently developing throughout the Columbia Plateau by late Grande Ronde to early Wanapum time and may have begun earlier. The restriction of certain basal pillow complexes of youngest Grande Ronde flows to synclinal troughs in the Cascades of northern Oregon with reciprocal thinning of these flows over anticlines suggests that deformation, at least in the western Columbia Plateau, had begun by ~16 m.y. ago (Beeson and Moran, 1979). Long and others (1981) and Landon and others (1982) have suggested that farther east in the Pasco Basin tectonic deformation may have been under way in late Grande Ronde time as evidenced by the westward thinning of flows onto the developing Hog Ranch axis (see also Price, 1982; Reidel and Fecht, 1981). Reanalysis of chip samples from the RSH-1 exploratory hole drilled in 1957-58 on the back slope of Rattlesnake Mountain indicates that the entire 3,250 m of the well is in Columbia River basalt and there appears to be no repeated stratigraphic section (Reidel and others, 1982). Therefore, some crustal extension or downwarping is suggested to accommodate this extremely thick stratigraphic section of Grande Ronde Basalt (Reidel and others, 1982).

In the eastern Columbia Plateau, Camp and Hooper (1981) have demonstrated that the Stites Basin began to develop in Grande Ronde time during a period of paleomagnetic reversal designated R_2 (see Fig. 2-3), possibly by movement on the bordering north-south Stites fault. The Lewiston Basin was developing in R_2 Grande Ronde time (Reidel, 1978) by subsidence along the northeast-trending Limekiln fault and its monoclinal extension to the northeast. Thinning of the Vantage member of the Ellensburg formation on to Naneum Ridge of the northwest-trending Wenatchee Mountains suggests that this structurally controlled topographic feature had been initiated in late Grande Ronde time (Tabor and others, 1982). Thus, deformation of Columbia River Basalt in the western, central, eastern, and northern Columbia Plateau had been initiated by at least late Grande Ronde time. If earlier deformation had taken place, it is masked by the relatively rapid rate of eruption of flows.

The timing and mechanics of deformation in each ridge of the Yakima Fold Belt and in the entire Columbia Plateau are still being refined. The initiation of deformation of Columbia River basalt at widely separated localities in the Columbia Plateau in late Grande Ronde time suggests that the process was regional and penecontemporaneous, but it has not been established that all ridges began to develop at this time. However, the indication that deformation across the entire plateau was penecontemporaneous in Grande Ronde time does not support the sequential south to north development of Yakima folds proposed by Laubscher (1981; Chapter 7). The style and relative sequence of fold and fault development in each of the Yakima folds are a function of the kinematics of deformation and are discussed in Chapter 7 where tectonic models are reviewed.

Duration of Deformation

In places, the stratigraphic sequence suggests that deformation on local structures within fold ridges developed during a brief period of time and had ceased during Miocene or Pliocene time (Bentley, 1977; Shannon and Wilson, 1973). Elsewhere, the evidence suggests that deformation continued through the Pliocene and into the Quaternary and is currently ongoing (Chapters 5, 6; Caggiano and others, 1980).

The westward offlap of progressively younger flows and the continuity of flows from sources in the eastern Columbia Plateau to the central and western Columbia Plateau and beyond attests to continuing uplift and maintenance of a paleoslope and continuing subsidence of the Pasco Basin in the Miocene (Swanson and Wright, 1978; Reidel and others, 1982; Chapter 5). Whether this continuing uplift is due to isostatic disequilibrium and the rise of the Idaho batholith, as suggested by Hooper and Camp (1981), remains to be confirmed.

The initiation of plateau-wide deformation can be inferred from Miocene topographic development that probably reflects folding. However, whether faulting developed penecontemporaneously with folding is a function of the interpretation of the mechanics of fold growth. If the folds develop where faults ramp upward from a buried detachment (décollement),

then fault slip slightly precedes fold development (Laubscher, 1981). If faults develop late in the folding process when folds begin to "lock up," then fault development is much later than the initiation of folding (Davis, 1981; Price, 1982). The mechanics and style of deformation are discussed in Chapter 7.

Folding in the Saddle Mountains east of Sentinel Gap can be demonstrated in Grande Ronde time; folding in Rattlesnake Mountain south of Snively Basin cannot be demonstrated until early Wanapum time, but may have been occurring earlier. Both ridges continued to grow through Wanapum and Saddle Mountains time at a long-term average low rate of uplift (vertical strain) (Reidel and others, 1980; Chapter 5). If this average low rate of strain is projected from 14.5 m.y. ago to the present, it accounts for the present structural relief of flows on Rattlesnake Mountain and the Saddle Mountains suggesting that deformation has been occurring at a slow, relatively continuous rate over the past 14.5 m.y. in the Pasco Basin (Reidel and Fecht, 1981).

Folding in other ridges of the Yakima Fold Belt was probably developing penecontemporaneously with the Saddle Mountains. However, the aerial distribution and thicknesses of stratigraphic units known from the present degree of detailed study do not allow this hypothesis to be demonstrated. Folding on eastern Umtanum Ridge began in pre-Umatilla time and continued at least through and beyond Pomona time (Goff, 1981). In the Filey Road area of Umtanum Ridge, steeply dipping flows of Grande Ronde and Wanapum Basalts are overlain with angular unconformity by essentially undeformed gravels that have been correlated with the Selah interbed (pre-Pomona) by Bentley (1977). Similar relationships are found on Yakima Ridge east of the Yakima River (Bentley, 1977). On Gable Mountain, the eastern continuation of Umtanum Ridge, the earliest suggestion of deformation is in post-Umatilla time as evidenced by thinning of the Selah interbed (Fecht, 1978). However, in the absence of stratigraphic control from boreholes, this does not preclude deformation at an earlier time.

In the Blue Mountains of central Oregon, folding and faulting along the northern flank of the Aldrich Mountain anticline occurred between solidification of the ~16-m.y.-old Picture Gorge Basalt and the essentially undeformed overlying 6.6-m.y.-old Rattlesnake volcanics (Robyn, 1977). Under north-south compression, the east-west striking John Day reverse fault and northwest- and northeast-striking strike slip faults developed.

Terrestrial sediments and volcanics in Neogene synclinal basins along the northern flank of the Blue Mountains in Oregon suggest uplift of this source area, which presumably resulted from deformation. The sediments in the Dalles Basin (Chenoweth Formation of Farooqui and others, 1981), the Tygh Basin (Tygh Valley Formation of Farooqui and others, 1981), and the Madras Basin (Deschutes Formation of Farooqui and others, 1981) rest on the Priest Rapids and Frenchman Springs, the Frenchman Springs and Grande Ronde, and the Grande Ronde and older rocks, respectively. Each of these

synclinal basins is bordered to the north and south by anticlines developed in Columbia River basalt, suggesting that deformation had been initiated by Priest Rapids and possibly earlier time. Epiclastic and volcaniclastic rocks in these basins were derived from the Cascade Range indicating that it was topographically higher, probably due to tectonic uplift, volcanism, or both.

The Alkali Canyon and McKay Formations (Farooqui and others, 1981) occur in the Arlington and Agency basins, respectively, which are in the Dalles-Umatilla and Agency synclines (Plate 2-1). The Alkali Canyon Formation rests conformably and disconformably on flows ranging from Frenchman Springs to Elephant Mountain ages. The McKay Formation rests conformably atop the Frenchman Springs member and the Grande Ronde Basalt, but an unconformity is inferred where undeformed sediments rest on deformed strata on the limbs of anticlines. Sediments in these basins are derived from the Blue Mountains suggesting that the Blue Mountains anticline was a relative topographic (structural) high by Wanapum time as suggested by Ross (1978). Trenches through Neogene sediments at the former site of the proposed Pebble Springs nuclear powerplant near Arlington, Oregon revealed no evidence of tectonic deformation in strata as old as the Pomona (Shannon and Wilson, 1975a). The Pebble Springs site is located near the axis of the northeast-trending, northeast-plunging Dalles-Umatilla syncline (Shannon and Wilson, 1975b).

Since tectonic deformation of the youngest sediments in the synclines is minimal or absent, it has been assumed (in the absence of subsurface data) that the oldest sediment and the youngest basalt in the synclines are relatively undeformed. In the Pasco Basin where subsurface data are available, some folds of smaller wavelength and amplitude are interpreted from subsurface data indicating deformation of the youngest basalt and perhaps of the overlying oldest sediments. Some deformation of old Ringold sediments along the flanks of the Southeast anticline (the eastward subsurface continuation of the Gable Mountain structure) has been reported (PSPL, 1982). Smaller second- or third-order folds have been interpreted in the Cold Creek syncline in the Pasco Basin from subsurface data (Myers, 1981). Geophysical anomalies or gradients with trends oblique to the inferred axes of synclines suggest the presence of cross structures, but structural interpretations have not been confirmed. Nonstructural interpretations of some of these anomalies (e.g., flow edges, pinch-outs) are possible. The attitudes of the oldest sediments appear conformable with those of the youngest basalt as interpreted from subsurface data in the Cold Creek syncline. Basalt strata are essentially undeformed, but dips of both basalt and sediments increase slightly onto the flank of the subsurface extension of Yakima Ridge. Thus, in the Pasco Basin, it appears that the low rate of deformation that affected the youngest basalt in the synclines continued into post-basalt time and may have resulted in some deformation of late Miocene or Pliocene sediments directly overlying the basalt.

A stratigraphy for Neogene and Quaternary sediments is being developed (see Tallman and others, 1981; Myers/Price and others, 1979), and this has greatly assisted in establishing a timetable of deformation.

For example, the Ringold Formation was previously assumed to be middle to upper Pleistocene (Newcomb and others, 1972) thus dating deformation of Columbia River basalt as geologically very recent and leading to much higher rates of deformation. Analysis of faunal and paleomagnetic data has established that Ringold sedimentation began in the late Miocene and extended to middle to late Pliocene (i.e., ~8.5 to ~3.5 m.y.; Tallman and others, 1981). Thus, new dates on the Ringold Formation have led to a revision in the chronology and rates of deformation.

Analysis of core samples and geophysical logs of the Ringold Formation of the southeast anticline (the eastern subsurface continuation of the Gable Mountain structure on the Hanford Site) enabled identification of correlative units interpreted to be sedimentary cycles (Webster and Crosby, 1981). These units thin over the anticline and thicken toward the Cold Creek syncline. Older strata on the flanks of the anticline dip more steeply than younger strata suggesting that continuous slow strain resulted in older strata experiencing greater deformation. This supports the interpretation that deformation in the Pasco Basin area has been slow and continuous at an average low rate since ~14 m.y. ago to the present (Caggiano and others, 1980).

Trenches through the central fault on Gable Mountain on the Hanford Site revealed minor offset of glaciofluvial sediments by a reverse fault that is continuous with one in the underlying basalt (PSPL, 1982). Maximum displacement was ~6 cm along a 334-m stretch of a fault that is ~3 to 4 km long. The central fault is interpreted to be a tear fault between the east and west second-order anticlines on Gable Mountain and this limits its length to 3 to 4 km. Drill holes through the hanging wall and into the footwall revealed progressively decreasing displacement from 64 m at the base of the Selah interbed to 55 m at the base of the Rattlesnake Ridge interbed to 0.06 m in the ~13,000-yr-old sediments. These displacements suggest a long-term average slip rate of ~0.005 mm/yr or ~5 mm/1,000 yr (5 m/m.y.). This rate is in order of magnitude agreement with vertical uplift rates between 12 and 10.5 m.y. ago (PSPL, 1982).

Deformation along structures of the Cle Elum-Wallula zone of deformation may have ceased or is on the wane (Davis, 1981; Chapter 7). Some deformation along the Wallula fault system (Farooqui, 1979; WCC, 1981c) may have occurred in pre-Wisconsinan time at Finley Quarry and at Warm Springs (WCC, 1981d), but major displacement has not been demonstrated. The Cle Elum-Wallula zone of deformation is relatively aseismic, suggesting that deformation in this zone is infrequent, has ceased, or is at least partially aseismic creep. If a major fault in this zone were locked (and thus be aseismic) geomorphic or stratigraphic evidence of major displacement might be expected similar to other locked faults (e.g., parts of the San Andreas fault, the Median Tectonic Line in Japan, the Anatolian fault prior to 1939; Allen, 1975). Such geomorphic evidence has not been observed along the Cle Elum-Wallula zone of deformation.

Deformation in Manastash Ridge apparently occurred between deposition of presumably Thorp gravel, which is folded, and the Thrall pediment, which truncates folded strata (Bentley, 1977). Thorp gravel in the Kittitas Valley has been dated at 3.7 m.y. by the fission track method (Waitt, 1979). Bentley (1977) assigns the Thrall pediment a date of ~1 m.y. old. Therefore, deformation in Manastash Ridge is apparently limited to <2.7 m.y. of Pliocene and early Quaternary age. However, this chronology is based on the assumptions that the folded gravel is correlative with the Thorp and that the Thrall pediment is ~1 m.y. old.

Scarps on Toppenish Ridge that offset topographic features and alluvial fans have been interpreted as tectonic (Campbell and Bentley, 1981; WCC, 1981c). Toppenish Ridge is a Yakima fold with some east-west trending reverse faults and some northwest-trending fracture zones interpreted as strike-slip faults (Bentley and others, 1980). Without data from the subsurface, it is not possible to unequivocally dismiss these features as evidence of Holocene tectonic faulting. The occurrence of steeply dipping tensional structures near and along the crest of the ridge with "horst and graben" along the slope and a low angle "thrust" at the base is suggestive of a gravitational influence in the development of these features.

In the western Columbia Plateau near Goldendale, Washington, the Simcoe volcanics provide a datable time stratigraphic horizon to which deformation can be related. The Simcoe volcanics overlie a number of northwest-trending fracture zones observed on imagery that have been interpreted as dextral wrench faults in The Dalles area (Plate 2-1; Bentley and others, 1980; Campbell and Bentley, 1981). One northwest-trending fault cuts a 4.5-m.y.-old flow but is overlain by an undeformed 3.5-m.y.-old flow (Shannon and Wilson, 1973). Other northwest-trending faults cut or appear to affect Simcoe volcanics that have not been radiometrically dated (Bentley and others, 1980), but others of these fractures are overlain by Simcoe lavas that are not cut or affected.

SUMMARY

The Columbia Plateau is a flood basalt province located in an intermontane basin on the western margin of the North American plate. It is an area of thin crust, relatively low heat flow, and low seismicity in contrast with the Cascade Range and Puget Sound to the west. The youngest three Columbia River basalt formations occur in the Pasco Basin of the central Columbia Plateau and are overlain by late Miocene, Pliocene, and Quaternary sediments. The reference repository location has been sited in the western part of the Cold Creek syncline, one of several synclines in the Pasco Basin that is bordered by asymmetrical anticlines to the north, west, and south.

The style of deformation of Miocene Columbia River basalt allows an informal division into three structural subprovinces. Relatively narrow, asymmetric anticlines of the Yakima Fold Belt trend generally east-west, but range from west-southwest to northwest. Reverse or thrust faults that strike subparallel to the axes of these folds are common on the steeply dipping limbs; northwest and northeast-trending, steeply dipping faults and some north-south faults are present on which dip-slip and some strike-slip movement has been reported. Synclines between anticlines are relatively broad depressions containing sediments whose geometry and structure have been interpreted using limited exposure and subsurface data. Strata in asymmetrical anticlines dip steeply ($\leq 90^{o}$) on short, steeply dipping limbs and more gently ($\leq 30^{o}$) on dip slopes. Strata in broad, open synclines dip much less steeply ($\leq 10^{o}$) and near the troughs may appear nearly flat lying. Strain appears to be concentrated in anticlinal folds, especially on steeper limbs; however, some strain can be anticipated in any dipping strata including those of synclines.

The Blue Mountains subprovince is characterized by a major northeast-trending, asymmetrical anticlinorium (the Blue Mountains) that is cut by northwest- and northeast-trending faults. Strata on the north side of the Blue Mountains dip $\leq 8^{o}$; strata on the south side dip $\leq 3^{o}$. Near the broad axial region of the Blue Mountains, strata are nearly flat lying. Northwest-trending faults are especially common near La Grande, Oregon where both dip- and strike-slip movement has occurred along border faults of the La Grande graben. Northwest-trending folds are also present in the Blue Mountains. West-northwest-trending faults of the Wallula fault system are parallel to those of the Brothers fault zone (Chapter 3) in the younger High Lava Plains to the south of the Columbia Plateau. The Wallula fault has been interpreted as a right lateral fault system with northwest-trending extensional dip-slip faults to the south. Activity on faults of this system is as young as late Quaternary. The north-northeast to northeast-trending Hite fault system has experienced dip as well as strike-slip motion and is interpreted as older than the Wallula fault system.

The Palouse subprovince is characterized by a gentle slope that developed in the Miocene. Superimposed on this slope are low amplitude, long wavelength folds of north-northwest to northwest trend. The northwest-trending Cheney fracture zone in the northern Columbia Plateau is expressed as a series of linear depressions that reflect joint-controlled erosion.

Structural deformation of Columbia River basalt began in late Grande Ronde time and has continued to the present. Strain, at least in the Pasco Basin area, appears to have developed under a regime of nearly north-south compression at an average low rate such that deformation has continued on first- and lower-order structures formed in the Miocene or Pliocene. Thrust or reverse faults developed with trends subparallel to fold axes. These east-west structures as well as northwest- and northeast-trending faults with steep dip may be related to deformation in a wrench tectonic regime. New structures do not appear to have developed in the last few million years. By siting the reference repository location in a syncline and by avoiding known structures, the potential effects of strain on a repository should be minimized over the next few thousand years.

CHAPTER 3 - REGIONAL GEOLOGIC SETTING OF THE COLUMBIA PLATEAU

J. A. Caggiano and K. R. Fecht

INTRODUCTION

Columbia River basalt overlies a geologically diverse terrain as judged from the age, lithology, and structure of rocks exposed along the plateau margin. In assessing the tectonic stability of a site in the central Columbia Plateau, it is important to consider the ages, rock types, and structures that occur beneath the basalt to assess the tectonic development of the area and to assess the potential for reactivation of older structures. If structures formed under an ancient stress regime are present and they are oriented favorably with respect to the present stress regime, the potential for reactivation is suggested. The potential for displacement on such possible structures beneath the basalt needs to be assessed.

Because there are no exposures of rocks beneath the thick basalts in the central Columbia Plateau and because there are no available reports of older rocks having been penetrated by drilling in that area, the age, rock types, and structures that might be present beneath the central Columbia Plateau must be inferred from extrapolation of terrains along the margins of the plateau. However, the diversity of lithology and structure along the plateau margins does not permit a simple extrapolation. The geologic history of the areas surrounding the Columbia Plateau is reviewed briefly as background for the interpretion of geophysical data on sub-basalt rock types and structures (Chapter 4), which are fundamental to evaluating hypotheses of the tectonic evolution of the central Columbia Plateau (Chapter 7). A generalized geologic map (Plate 3-1) is used to illustrate and locate features discussed in this chapter.

THE PRECAMBRIAN CRATON

The Precambrian margin of the North American plate lies east of the Columbia Plateau. Phanerozoic sedimentary and volcanic rocks west of this margin have been accreted to the continent over the past 600 m.y. Strontium isotopic ratios ($^{87}Sr/^{86}Sr$) in igneous rocks of Mesozoic and Cenozoic age suggest that the buried westernmost edge of the Precambrian basement of the contiguous North American plate (Armstrong and others, 1977) trends north-south just east of the Idaho border. This boundary between older and younger rocks turns abruptly westward in southeastern Washington and becomes buried beneath the Miocene basalts of the Columbia Plateau. The Precambrian margin of North America is assumed to emerge north of the plateau along the western edge of the Kootenay arc* (Davis, 1977), but

*The western boundary of the Kootenay arc is approximated by the course of the south-flowing Columbia River north of the Columbia Plateau (see Plate 3-1).

Precambrian rocks with high $^{87}Sr/^{86}Sr$ ratios typical of cratonic rocks are exposed west of the longitude of the Kootenay arc in the Okanogan Highlands and North Cascades (Armstrong and others, 1977). Davis (1977) suggests that the Pacific Northwest west of the Precambrian craton is a mosaic of allochthonous crustal blocks. These blocks have become juxtaposed by Mesozoic and early Cenozoic tectonic processes along the convergent margin of the North American plate. Thus, Precambrian, Paleozoic, and Mesozoic terrains west of the craton may be exotic as a result of tectonic transport. Hypotheses to explain the tectonic evolution of terrains that seem to have resulted from different geologic processes should consider this possibility and not assume such rocks are at or close to the place where they were formed.

PRECAMBRIAN AND PALEOZOIC

Late Precambrian and early Paleozoic clastic and carbonate rocks are found in northern and eastern Washington and adjacent Idaho and British Columbia (Stewart and Suczek, 1977; Griggs, 1973; Hooper and Webster, 1982). Strata of the Precambrian Belt Supergroup (United States terminology) and Purcell Group (Canadian terminology) rest unconformably on older Precambrian metasediments. These are overlain, in places unconformably, by clastics, carbonates, volcanics and diamictites of the late Precambrian Windermere Group which, in places, grades conformably upward to Lower Cambrian fossiliferous, terrigenous, detrital sediments. Overlying Middle and Upper Cambrian carbonates were succeeded by deposition of carbonates and interbedded fine terrigenous sediments which continued at least into middle Paleozoic time (Greenman and others, 1977). Similar lithologies with some volcanics are reported in the Chilliwack Group in the North Cascades (Misch, 1966). This sedimentary environment persisted from late Precambrian time through most of the Paleozoic throughout the North American Cordillera. West of the north-trending Purcell trench (Plate 3-1), Precambrian Belt Supergroup strata have been highly metamorphosed; east of the trench (a possible fault zone), the metamorphic grade is much lower.

Paleozoic and possibly older rocks are found to the north in the North Cascades and the Okanogan Highlands and to the south in the Blue Mountains. In the North Cascades, these rocks are highly metamorphosed, structurally complex, and can be dated only as pre-Devonian (Misch, 1966). Some basement rocks may have been displaced during faulting of late Paleozoic or Mesozoic age. Sedimentary and volcanic rocks of probable Paleozoic age have been metamorphosed to schists and gneisses. Suggested correlations of rocks in the North Cascades with rocks exposed tens to hundreds of kilometers away in the Blue Mountains imply that these rocks have been tectonically transported to their present location (Danner, 1977).

Limited radiometric dates and exposures restrict detailed interpretation of Paleozoic rocks in the Blue Mountains. Terrigenous sediments of middle to late Paleozoic age were deposited on a shallow shelf and these are presumably transitional to an oceanic assemblage. Oceanic and shallow shelf rocks are presently in structural contact (Vallier and others, 1977).

A Permo-Triassic volcanic arc in the Blue Mountains of eastern Oregon and western Idaho (Brooks and Vallier, 1978) may be the southern extent of Wrangellia--a formerly continuous terrain of Paleozoic and younger rocks now occurring as fragmentary, exotic terrains displaced long distances by major strike-slip motion (Jones and others, 1977). An Early Permian arc may be represented by volcanic and related rocks of the Chilliwack Group on the western flank of the North Cascades (Monger and others, 1972), but relationships to Blue Mountain volcanics are enigmatic.

MESOZOIC

Mesozoic rocks of different metamorphic grade and structural complexity occur to the northwest, north, east, and south of the Columbia Plateau. Deformed eugeoclinal graywackes, mudstones, breccias and volcanics of Late Triassic to Early Cretaceous age occur in the western foothills of the North Cascades. The relationship of these strata in the western foothills to a thick and locally unique sequence of late Mesozoic marine and non-marine strata in the Methow graben is not clear. In the Methow graben, marine, fossiliferous, cherty, and arkosic shale, sandstone, and conglomerate are overlain by terrestrial clastic and volcanic rocks (Barksdale, 1975). Deposition was on submarine detrital fans derived initially from an eastern and then a western source (Tennyson and Cole, 1978). The section is >15 km thick and is a paleogeographic enigma requiring a narrow arm of the sea, erosion of accompanying Cretaceous rocks that do not now occur in the vicinity, or major post-depositional translation by faulting to explain the presence of these strata. Cretaceous plutonic rocks that were exposed to erosion by Tertiary streams and thick saprolites indicate that a lengthy subaerial exposure occurred in Late Cretaceous-early Tertiary time (Vallier and others, 1977). The gently folded marine sedimentary rocks in the Methow graben are bordered by major northwest-trending faults.

A conformable sequence of Carboniferous to Triassic eugeoclinal clastics and some cherty limestone occurs in the western Okanogan Highlands (Yates and others, 1966). These rocks were subsequently folded, faulted, and intruded by plutonic rocks of Triassic to early Tertiary age. Sedimentation may have occurred in island arcs or back arc basins that were subsequently overridden by the North American plate (Fox and others, 1977).

Volcanic and volcaniclastic rocks of Permian to Upper Triassic age occur along the western flank of the Idaho batholith and the eastern end of the Blue Mountains of Oregon. These rocks originated in a volcanic arc terrain, the Seven Devils volcanic arc, that has since been metamorphosed and deformed (Brooks and Vallier, 1978; Brooks, 1979). Carbonates and clastics overlying Seven Devils volcanics are succeeded by volcaniclastic flysch of Late Triassic to Late Jurassic age. Many of these rocks, especially those near the Idaho batholith, have been metamorphosed and complexly deformed.

Synorogenic plutonic rocks that were intruded during various stages of the late Mesozoic to early Cenozoic Laramide orogeny occur in the region to the northwest, north, and east of the plateau (Hamilton, 1978). Most of these are granite, granodiorite, or quartz diorite of Jurassic to early Tertiary age. Batholiths of this age and composition comprise much of the North Cascades and also occur in the Okanogan Highlands, the Blue Mountains (Wallowa and Bald Mountain batholiths), and in the northern Rocky Mountains (e.g., the Idaho and Kaniksu batholiths).

CENOZOIC

Terrigenous molasse and volcanics of Cenozoic age are extensive in the Cascades to the west and the High Lava Plains to the south (Armstrong, 1978). Terrestrial sediments and volcanics occur to the north and south of the Columbia Plateau in restricted basins (Dickinson, 1979; Cole and Armentrout, 1979; Nilsen and McKee, 1979). These sediments and volcanics reflect widespread orogenesis in the region in late Mesozoic to early Cenozoic time. Paleogene (Paleocene, Eocene, Oligocene) sediments and volcanics form the major part of the South Cascades west of the plateau. These rocks reflect development of a volcanic arc produced by subduction of oceanic crust of the Farallon plate beneath continental crust of the North American plate. This arc was a source for lower to middle Tertiary sediments deposited as eastward and westward thickening wedges along its margin (Armstrong, 1978; Dickinson, 1979). Alluvial fans and plains extended eastward (Swauk, Chumstick, Roslyn, and Wenatchee Formations) and westward (Puget Group) where they interfingered with marine sediments on a marginal coastal plain. Volcanics are common in both eastern and western clastic wedges (Snavely and Wagner, 1963; Gresens and others, 1977). The ancestral Cascades of Eocene/Oligocene age were uplifted in Pliocene time to form the present Cascade Range. The high stratovolcanoes that cap the present range were developed and glaciated in the Quaternary.

In the North Cascades, Paleogene sediments and volcanics are restricted to structural basins. Some Paleogene sediments cap a thick section of Cretaceous rocks in the Methow graben, but an extensive, thick sequence of fluvial arkosic sandstones and conglomerates is found in the northwest-trending Chiwaukum graben (McKee, 1972). Coal-bearing sediments of the Roslyn Formation overlie the Teanaway basalt, which rests upon a thick sequence of fluvial sandstone and conglomerate--the Swauk Formation. The Swauk Formation is >7 km thick near Wenatchee, Washington, indicating that the Chiwaukum graben was subsiding relative to source areas during sedimentation and volcanism. Eocene dacite to quartz latite and interbedded detrital sediments fill the northeast-trending Republic graben north of the plateau and are but part of an extensive section of Paleogene volcanics in the Okanogan Highlands. Isolated remnants elsewhere in northern Washington indicate that a former, more extensive Paleogene volcanic section has been largely removed by erosion.

To the south, Eocene and Oligocene bimodal volcanics and volcaniclastic sediments of the Clarno and John Day Formations indicate volcanic activity in the Blue Mountains and the Cascades (Armstrong, 1978; Nilsen and McKee, 1979). Several hundred meters of fossiliferous sandstones, breccias, tuffs and volcanics of the Clarno Formation are unconformably overlain by several hundred meters of brightly colored ash, basaltic flows (near the base), and tuffaceous sandstone of the John Day Formation, which contains a rich flora and vertebrate fauna. These volcanics and sediments derived from them are part of the Challis volcanic arc (Armstrong, 1978).

Neogene (Miocene and Pliocene) sediments and volcanics occur principally in the Columbia Plateau and to the south and west. The Columbia Plateau is an areally extensive accumulation of Miocene tholeiitic basalt (the Columbia River Basalt Group) with some interbedded tuffaceous sediments (the Ellensburg Formation, the Latah Formation, Payette Formation, etc.). Miocene and possibly pre-Miocene basalt and rhyolite occurs south of the plateau in the High Lava Plains of Oregon (Walker, 1977). Calc-alkalic volcanism continued in the Cascades to the west while basaltic volcanism occurred in the plateau. Sediments resulting from erosion of the Cascade chain spread eastward and westward. The intermontane Columbia basin received sediments and volcanics from surrounding highlands and continued slow subsidence. Many of these sediments interfinger with or overlie Columbia River basalt. Bimodal basalt-rhyolite volcanism to the south in the High Lava Plains both preceded and succeeded Columbia Plateau basaltic fissure eruptions. Basaltic volcanic activity of mostly Quaternary age occurred along the Snake River Plain to the south. Youthful volcanism of the High Lava Plains is evident from the landscape of cinder cones, some of which are aligned along fractures. Aligned rhyolitic domes generally decrease with age to the northwest culminating in the Quaternary Newberry Crater (Armstrong, 1978).

STRUCTURAL RELATIONSHIPS

Structures and the rocks in which they occur trend into or under the Columbia Plateau along its margin. In places, structures continue beneath the basalt and are expressed in the plateau (e.g., the Wenatchee Mountains extend beneath the plateau and continue as the Hog Ranch axis). The similarity of structural trends in plateau and older rocks suggests a genetic relationship of plateau structures to structures in areas marginal to the plateau (Tolan, 1982) (Plate 3-1). The similarity of trends and style of folding and faulting between rocks along the margins of the plateau and rocks beyond the plateau margin suggests that deformation of the plateau has been similar to that in the region. Such regional deformation probably affected rocks in the central Columbia Plateau also. Such regional deformation may reflect the movement of lithospheric plates or crustal blocks. An exception is the east-west trend of folds and faults in the Yakima Fold Belt, a trend not seen in major structures surrounding the plateau.

Faults and fold axes have variable orientations in terrains marginal to the plateau. Sub-parallel, northwest- to north-northwest-trending broad open folds in Precambrian rocks north of Spokane, Washington are parallel to younger normal faults, but all deformation is probably Laramide or older (Yates and others, 1966). East of Spokane, Belt Supergroup rocks are cut by a series of northwest- to west-northwest-trending faults of the Lewis and Clark alignment of which the Hope and Osburn faults are examples. Upwards of a few tens of kilometers of strike slip as well as some dip slip has occurred on these faults. Northwest- to north-northwest-trending listric thrust faults along the Purcell anticlinorium displace Belt strata eastwards, a change from earlier interpretations in which these strata were considered autochthonous or parautochthonous (Harrison and others, 1980).

The arcuate, nearly north-south zone separating slightly metamorphosed Precambrian Belt strata to the east and highly metamorphosed strata to the west corresponds to the Purcell trench. The narrow curvilinear trend and contrast in metamorphic grade along with a steep gravity gradient suggests that this zone may be a fault (Cady, 1980), but the contact may be an unconformity between substantially older Precambrian strata and younger Precambrian Belt rocks (Harrison and others, 1980). A north-south alignment of metamorphic core complexes extending from southern Canada to northern Mexico (Crittenden and others, 1978) passes to the west of the Purcell trench. High-grade metamorphics and gravity data suggest a buried gneiss dome of this alignment near the Canada-United States border (Cady, 1980).

Rocks east of the Columbia River and west of the Purcell trench are deformed along northeast trends. West of the Columbia River, the trend of folds and faults is north to northwest but northeast-trending faults are present (Yates and others, 1966).

Remnant Eocene volcanic rocks in northern Washington are preserved in the northeast-trending Republic graben (Plate 3-1) (Staatz, 1964). Several hundred meters of ash, tuff, breccia, rhyodacite, and quartz latite in the Republic graben are in abrupt contact with older metamorphic rocks along the northeast-trending Scatter Creek and Sherman fault zones that border the Republic graben (Vance, 1977). Some of these post-Eocene faults terminate in Mesozoic gneisses; others appear to continue to the margin of Columbia River basalt. Grand Coulee continues along the trend of the Republic graben in the northern Columbia Plateau (Plate 3-1), suggesting possible control of Pleistocene catastrophic erosion by fractures developed where the basalt overlies this older structure. A similar relationship exists with the west-northwest-trending Cheney fracture zone (Plate 3-1) (Griggs, 1973) in the northern Columbia Plateau. This zone continues along the projected strike of faults in older rocks beyond the plateau margin.

Northwest and northeast structural trends are common in and along the margins of the northeast-trending Blue Mountains (Walker, 1977). While capped by Grande Ronde Basalt, the Blue Mountains have been uplifted and dissected to expose pre-Miocene rocks. The broad Blue Mountains anticlinal arch is flanked on the north by northeast-trending faults of the Hite fault

system (Plate 2-1) (Newcomb, 1970) where dip slip has been mapped. An episode of strike slip after dip-slip offset has been recently proposed; however, Kienle (1980) suggests that this slip is dextral, while Kendall (1981) suggests that it is sinistral. Northwest-trending faults, which cut Mesozoic gneisses and Eocene volcanics, are parallel to border faults of the La Grande graben (Walker, 1977).

North-south, steeply dipping faults with dip-slip displacement are common along the western margin of the Idaho batholith in Mesozoic rocks. Older rocks are exposed closer to the batholith suggesting that the attitude of the fault plane changes with depth and that these blocks have been rotated during displacement (Hamilton, 1963).

A north-south structural trend is well represented in the Cascade Range by the alignment of the range, marginal thrust faults, and the Straight Creek fault (Vance, 1977) (Plate 3-1). Pliocene uplift of an early Tertiary volcanic arc in the Cascades south of Snoqualmie Pass and an older terrain of Paleozoic schists and gneisses intruded by granitic plutonic rocks to the north of Snoqualmie Pass have produced the present Cascade Range. The High Cascade volcanoes of Quaternary age and the Late Cretaceous and early Tertiary plutons generally align along a north-south trend as well. North-south-trending thrust faults on the margins of the North Cascades reflect compression during the Laramide orogeny. The Shuksan thrust on the west and the Jack Mountain thrust (Plate 3-1) on the east dip toward the axis of the range and displace rocks from the core toward its lesser deformed margins.

One of the major faults in the Pacific Northwest, the Straight Creek fault, trends generally north-south through the Cascade Range of northern Washington, but turns slightly west of north in British Columbia where it joins the Fraser fault zone. Up to 200 km of right lateral offset of Cretaceous rocks has been suggested for this fault (Misch, 1966, 1977). Movement on the Straight Creek fault has apparently been episodic (Vance, 1977), but plutons dated radiometrically at 16 and 35 m.y. cut the fault and are not offset indicating cessation of activity by the middle Tertiary (Tabor and Frizzell, 1979).

Southeast of Kachess Lake and the Yakima River, structures of the Straight Creek fault zone turn southeastward (Tabor and Frizzell, 1979). The Eocene, north-south-trending Goat Peak segment of the Straight Creek fault becomes the northwest-trending Taneum Lake segment ~15 km south of the Yakima River. The Taneum Lake fault is approximately coincident with a major northwest-trending lineament, the Olympic-Wallowa lineament (Raisz, 1945; Chapter 2), which extends from the Olympic Peninsula of Washington to the Wallowa Mountains of Oregon. The Taneum Lake reverse fault offsets only pre-Miocene rocks (Bentley, 1977); plutonic rocks of the 17 to 18-m.y.-old Snoqualmie batholith are not offset at the south end of the Straight Creek fault. Thus, activity on the Straight Creek and related faults is pre-Miocene (Tabor and Frizzell, 1979).

Folds and faults of northwest trend are common in both the North and South Cascades. Miocene and older rocks north and south of Mount Rainier are deformed in linear and curvilinear folds of generally northwest trend, the axes of which plunge northwest and southeast (Fiske and others, 1963). Major northwest-trending fault zones of late Mesozoic age border the Methow and Chiwaukum grabens. The Chewack-Pasayten fault and a series of subparallel faults of the Skagit fault system (Shannon and Wilson, 1977) border the Methow graben (Barksdale, 1975), which extends from near the border of the Columbia Plateau into southern British Columbia. North of Chelan, Washington, the border faults merge and the Mesozoic strata within the graben pinch out west of the edge of the plateau. Thus, the graben may not extend beneath the plateau. In addition to dip slip, a substantial amount of strike slip motion is suspected on the Skagit and Chewack-Pasayten faults (Vance, 1977), especially since they appear to join the Straight Creek fault near the crest of the North Cascades. Activity on segments of these fault zones had essentially ceased by mid-Cenozoic time as indicated by the lack of offset of early Tertiary plutonic rocks where they cross these faults.

The northwest-trending Leavenworth and Entiat faults border the Chiwaukum graben on the west and east, respectively (Gresens and others, 1977). These faults were active in Eocene and Oligocene time. Some post-Miocene deformation on these faults is suggested by gently dipping Grande Ronde Basalt that overlies folded rocks within the graben, but aerial and ground reconnaissance failed to reveal any evidence of Quaternary displacement (Vance, 1977). The Entiat and Leavenworth faults appear to extend to and beneath Columbia River basalt along the plateau margin suggesting that the graben continues beneath the plateau. Gresens and Stewart (1981) suggest the Chiwaukum may be a rhomb graben that continues southeastward beneath the plateau.

The High Lava Plains to the south of the plateau may be characterized by major northwest- (N. 60° W.) trending, right lateral, transcurrent faults that segment the area into blocks and form the northern boundary of the Basin and Range Province to the south (Lawrence, 1976). Within each of these blocks, N. 40° W.- and N. 30° E.-trending fractures interpreted from satellite imagery as normal faults (the northwest trend predominating) form a set of Riedel and conjugate Riedel shears (Tchalenko, 1970) indicative of right lateral displacement. East-west extension characteristic of the Great Basin to the south apparently diminishes progressively northward across these fault zones and terminates north of the Brothers (Plate 3-1) and Vale fault zones. Davis (1981), from field study, alternatively interprets some N. 50° W.-trending, en echelon features between Route 20 and Prineville, Oregon as faulted anticlinal ridges, features that Lawrence interpreted as normal faults. Compressional (rather than tensional) features of this orientation suggest to Davis (1981) that they resulted from sinistral slip along a system of nearly east-west faults. A series of rhyolitic domes trending northwest with decreasing age westward suggests growth of a fracture system westward. These faults offset Pliocene and younger rocks (Christiansen and McKee, 1978).

STRUCTURAL AND TECTONIC DEVELOPMENT OF THE PACIFIC NORTHWEST

The earliest tectonic development of the Pacific Northwest is interpreted from the similar stratigraphic succession of late Precambrian rocks that occurs throughout much of the Cordillera. Miogeoclinal strata of the Precambrian Belt Supergroup were deposited along the western margin of the continent. Belt rocks were succeeded by diamictites and volcanics of the late Precambrian Windermere Group (Stewart and Suczek, 1977), which grade upward into Lower Cambrian clastics and carbonates. Regional rifting and the development of an ocean basin are suggested, but there is disagreement as to whether rifting began prior to deposition of Belt strata (Monger and others, 1972; Burke and Dewey, 1973) or after (Stewart, 1976). Stewart and Suczek (1977) proposed that diamictites and volcanics of the Windermere Group accumulated in rift valleys prior to the opening of a marginal ocean basin. Cambrian and younger shallow marine sediments of the North American Cordillera accumulated on a shelf in this developing marginal ocean basin.

Evidence of orogenesis in the Pacific Northwest during the Paleozoic Era is limited. Late Silurian and Middle Devonian conglomerates in northeastern Washington suggest nearby uplifted sources, but little other evidence of major deformation remains (Greenman and others, 1977). Thrusting of oceanic assemblages eastward over the continent in the Cordillera of the southern United States during the Devonian-Mississippian Antler and Permo-Triassic Sonoman orogenies may indicate obduction during closure of Paleozoic ocean basins (see Dickinson, 1977, for discussion). Effects of the Antler orogeny are seen in Idaho (Poole and Sandberg, 1977) along trends which can be projected into southeastern Washington. West-northwest-trending faults of the Lewis and Clark alignment were active at least in Paleozoic and younger times.

Subduction along the Mesozoic western edge of the North American continent and subsequent right lateral displacement has produced a "collage" of terrains, many of which may be allochthonous (Davis and others, 1978; Hamilton, 1978; Monger and others, 1972). Juxtaposed fragments of oceanic and volcanic arc terrains that were accreted to North America mostly during the Mesozoic have been variously interpreted. Igneous activity in the Pacific Northwest has been dated as Permian, Late Triassic, Jurassic, and Cretaceous, but it is not clear whether the dated rocks arose from arc volcanism or plutonism and orogenesis along plate margins. Plutonic rocks of Cretaceous age are found in the North Cascades, the Blue Mountains, and the Idaho batholiths as evidence of late Mesozoic orogenesis. Right lateral motion along the northwest-trending Ross Lake fault and the north-south Straight Creek fault in the Cascades during the Cretaceous and early Tertiary time juxtaposed rocks of varying types.

Major geologic structures in the northwest such as the Straight Creek, Ross Lake, and Entiat faults are interpreted as a result of plate interaction during the Late Jurassic to early Tertiary Laramide orogeny. Northwest, west-northwest, and north-south structures in pre-Tertiary rocks north, east, and south of the plateau generally date from this

period during which diverse terrains were accreted to the continental margin. Much metamorphism, especially along the margins of plutons, presumably is also Laramide.

Deformation in the Cenozoic continued along previously established trends. Northwest-trending structures common in pre-Tertiary rocks north, east, and south of the plateau also occur in early Tertiary rocks in the Cascades. Northwest-trending folds and the faults that border the Chiwaukum graben developed in early Tertiary time and deformation continued along similar trends in older rocks. The northeast-trending Sherman and Scatter Creek fault zones that border the Republic graben along the northern margin of the plateau are penecontemporaneous with Eocene volcanics in the graben.

SUMMARY

Columbia River basalt overlies a diverse assemblage of rock types and structures along its margin that can not be easily extrapolated into the central Columbia Plateau. Paleozoic and older metasedimentary and metaplutonic rocks are present in the North Cascades and Okanogan Highlands to the northwest and north of the plateau and in the Blue Mountains in the southern Columbia Plateau. In the North Cascades, the rocks include eugeoclinal, highly metamorphosed graywackes, shales, sandstones, carbonates, and volcanics and metamorphosed calc-alkalic and acidic plutonic rocks. Slightly metamorphosed miogeoclinal clastics and carbonates as well as more strongly metamorphosed eugeoclinal clastics and volcanics occur in the eastern and western Okanogan Highlands, respectively. Both have been subsequently intruded. Paleozoic rocks in the Blue Mountains are metamorphosed and structurally complex. Eugeoclinal clastics and spilites rest on older high-grade schists and gneisses thought to be an older Paleozoic eugeoclinal clastic and volcanic assemblage.

Mesozoic rocks are found to the east in the Idaho batholith and its western flank and also in the Blue Mountains to the south and the Okanogan Highlands and Cascade Range to the north. The Upper Triassic portion of the Permo-Triassic Seven Devils Group in the Blue Mountains of western Idaho and eastern Oregon consists of volcanics and clastics that unconformably overlie Permian eugeoclinal strata. These units represent old island arc systems accreted to the continent at a convergent plate boundary. The western part of the Okanogan Highlands is also underlain by a eugeoclinal clastic and volcanic sequence consisting of sandstones, slates, and greenstones of Carboniferous to mid-Mesozoic age. Jurassic to early Tertiary monzonite to granodiorite of the synorogenic Idaho batholith occurs east of the plateau; similar plutonic rocks also occur in the Okanogan Highlands, Blue Mountains, and Cascade Range to the north, south, and west of the plateau. A thick sequence of Cretaceous marine clastics are preserved in the Methow graben in the North Cascades.

Tertiary sedimentary and volcanic rocks comprise the South Cascades and the High Lava Plains to the west and south; Tertiary sediments and volcanics also occur in the North Cascades, Okanogan Highlands, and Blue Mountains. Paleogene rocks in the Cascades consist mostly of volcanic and volcaniclastic rocks and terrestrial clastic sediments derived therefrom. Calc-alkalic volcanic rocks of the Cascade volcanic arc were eroded by streams draining eastward into the intermontane Columbia Basin and westward to the sea. These streams deposited clastic sediments with some interbedded volcanics that thicken away from the ancestral Cascade crest. Volcanics and volcaniclastics of the Challis and ancestral Cascade volcanic arcs occur to the north and south of the plateau in the Okanogan Highlands and Blue Mountain and High Lava Plain provinces. Calc-alkalic and bimodal rhyolitic and basaltic volcanics of Neogene age are evidence of continuing volcanism west of the plateau. South of the plateau, bimodal volcanism in the High Lava Plains continued into the Neogene with tuffs and basalts (Walker, 1977). Thick Miocene tholeiitic basalts comprise the Columbia Plateau, with interbedded clastic sediments derived from surrounding highlands, especially during late Miocene time. Pliocene and Quaternary terrestrial sediments accumulated in intermontane basins as streams drained from surrounding highlands where relief increased in the late Neogene and Quaternary. Bimodal volcanism continued during the Quaternary in the High Lava and Snake River Plains to the south and to the west where the large stratovolcanoes of the High Cascades are evidence of ongoing volcanism.

The diversity of rock types surrounding the plateau suggests candidates for rocks beneath the basalt in the Pasco Basin area that are interpreted from geophysical data (Chapter 4). Rocks of low interstitial porosity and permeability that appear to be present beneath the plateau could be metasediments or metavolcanics that occur to the north in the Okanogan Highlands and Northern Rocky Mountains, to the south in the Blue Mountains, to the east along the margin of the Idaho batholith, and to the northwest in the North Cascades. Cretaceous and Tertiary sedimentary rocks of relatively high porosity and permeability that surround the plateau to the northwest and south may also extend beneath the plateau.

West-northwest-, northwest-, north-south-, and northeast-trending folds and faults are found in rocks beyond the margins of the Columbia Plateau. Northwest-trending structures occur in the Cascade Range, in the Blue Mountains, and between the plateau and the Idaho Batholith. West-northwest-trending faults are common in the Lewis and Clark alignment northeast of the plateau. Dip and strike slip displacement has occurred on faults of this trend. North-south-trending dip slip faults are common along the western margin of the Idaho batholith. The Straight Creek fault is a major strike slip fault in the North Cascades that extends well into British Columbia as part of the Fraser fault zone. Northeast-trending folds and faults are present in the Okanogan Highlands where northeast normal faults border the Republic graben. Some northeast trends also occur in the Cascades as well as along the margin of the Idaho batholith.

Structures that may extend beneath the plateau are the north-south-trending Wenatchee Mountains, the northeast-trending Republic graben, and the northwest-trending La Grande, Methow, and Chiwaukum grabens. The Hog Ranch axis continues along the trend of the Wenatchee Mountains and influences the thickness and distribution of flows of Columbia River basalt. Grand Coulee continues along the strike of the Republic graben and, along with the parallelism of major northeast-trending coulees, suggests the possibility of structural control. North-south- and northwest-trending major faults such as the Straight Creek, Ross Lake, and Entiat faults along which major right-lateral displacement has been postulated in late Mesozoic and early Tertiary time suggest that terrains displaced by this dextral motion may be present beneath the plateau. Thus, Cretaceous and Paleogene rocks in the Methow and Chiwaukum grabens may continue on strike or with a dextral en echelon offset beneath the plateau. The interpretation of the La Grande as a rhomb graben developed along northwest-trending structures similarly suggests continuation of this trend beneath the plateau. Whether any of these structures extend beneath the Pasco Basin can only be hypothesized at present since sub-basalt rock types and structures can only be interpreted from limited geophysical data (Chapter 4).

CHAPTER 4 - PRE-COLUMBIA RIVER BASALT GROUP STRATIGRAPHY AND STRUCTURE IN THE CENTRAL PASCO BASIN

T. H. Mitchell and K. A. Bergstrom

INTRODUCTION

The pre-Columbia River Basalt Group stratigraphy and structure in the central Pasco Basin is difficult to study due to the thick volcanic cover that has buried all rocks older than Columbia River basalt. Any interpretation of stratigraphy and structure of the pre-basalt rocks is limited by the absence of boreholes or eroded "windows" in the basalt that might expose these older rocks within the Pasco Basin. Exploratory wells near Union Gap and Odessa, Washington and Condon, Oregon, outside of the Pasco Basin, have penetrated through sections of Columbia River Basalt Group (Myers/Price and others, 1979), but are not directly correlatable to the Pasco Basin studies, due to their distance from the basin.

A genetic relationship between structures beneath Columbia River basalt and the development of Yakima folds has been hypothesized in several conceptual tectonic models for the Columbia Plateau (Chapter 7). These hypotheses infer basement structures from the geometry and spatial distribution of first-order structures observed in basalt. To test these hypotheses, it is necessary to have information on the rock types, thickness, and elevation of the top of units beneath the basalt. From this information it may be possible to interpret deep structures and evaluate their relationship to the location and geometry of Yakima folds.

If Yakima folds are surficial expressions of deep structures, then the stability of the reference repository location can be assessed by studying structures in the basalt. If Yakima folds are not related to deep structures, then a stability assessment must consider whether these interpreted deep structures present any potential hazard to a repository under the present stress regime or a potential hazard if the stress regime were to change and cause possible reactivation.

APPROACH

Six steps are currently being used to assess the presence of structures beneath the basalt and their possible effects on the stability of the reference repository location. They are as follows:

1. Compile geologic and geophysical data pertinent to the tectonic evolution of the Columbia Plateau.

2. Integrate and interpret the data.

3. Collect additional data where the preliminary evaluation shows a need.

4. Evaluate and/or develop conceptual tectonic models based on steps 1 through 3.

5. Quantitatively model and test the above interpretations in order to generate constraints, and limit the number of possible tectonic models.

6. Apply the constraints developed in the previous five steps to the tectonic models identified and assess features that could have an effect on the stability of the reference repository location.

The BWIP's progress to date is proceeding into step 4.

This chapter is primarily centered around a qualitative analysis and integration of three geophysical investigations. Figure 4-1, the study-area location map, delineates the general boundary of the study area and the geographic areas discussed. Quantitative analysis and thorough integration of the data with geology in and around the Pasco Basin is being conducted to adequately test the results and assess the tectonic stability of the reference repository location. Some preliminary geologic interpretations of the structure and stratigraphy of the pre-Columbia River Basalt Group and their possible relationships to the Yakima folds are discussed. The geophysical investigations include the following surveys:

- Magnetotelluric (44 stations, Plate 4-1)
- Gravity (regional Bouguer, Plate 4-2)
- Aeromagnetic (4,000 ft (1,220 m) constant barometric altitude, Fig. 4-2)

The pre-Columbia River Basalt Group stratigraphy and deep structures discussed in this chapter are interpreted primarily from magnetotelluric data collected by Geotronics Corporation, Austin, Texas for Rockwell in 1981 and 1982. The capabilities of the magnetotelluric method make it an effective geophysical tool for providing geologic information beneath the Columbia River basalt. See Appendix A for a summary of the magnetotelluric methodology.

The qualitative integration of magnetotelluric, aeromagnetic, and gravity surveys is represented on five maps depicting the interpreted deep configuration within the central Pasco Basin. These maps include: (1) isopach of the volcanic sequence, Figure 4-3; (2) isopach of the conductor (interpreted sedimentary sequence below the volcanics), Figure 4-4; (3) contour map of the top of the conductor (interpreted sedimentary sequence below the volcanics), Figure 4-5; (4) contour map of the interpreted basement below the sedimentary sequence, Figure 4-6; and (5) a geophysical anomaly map, Figure 4-7. A description of the geophysical anomalies shown in Figure 4-7 is given in Table 4-1.

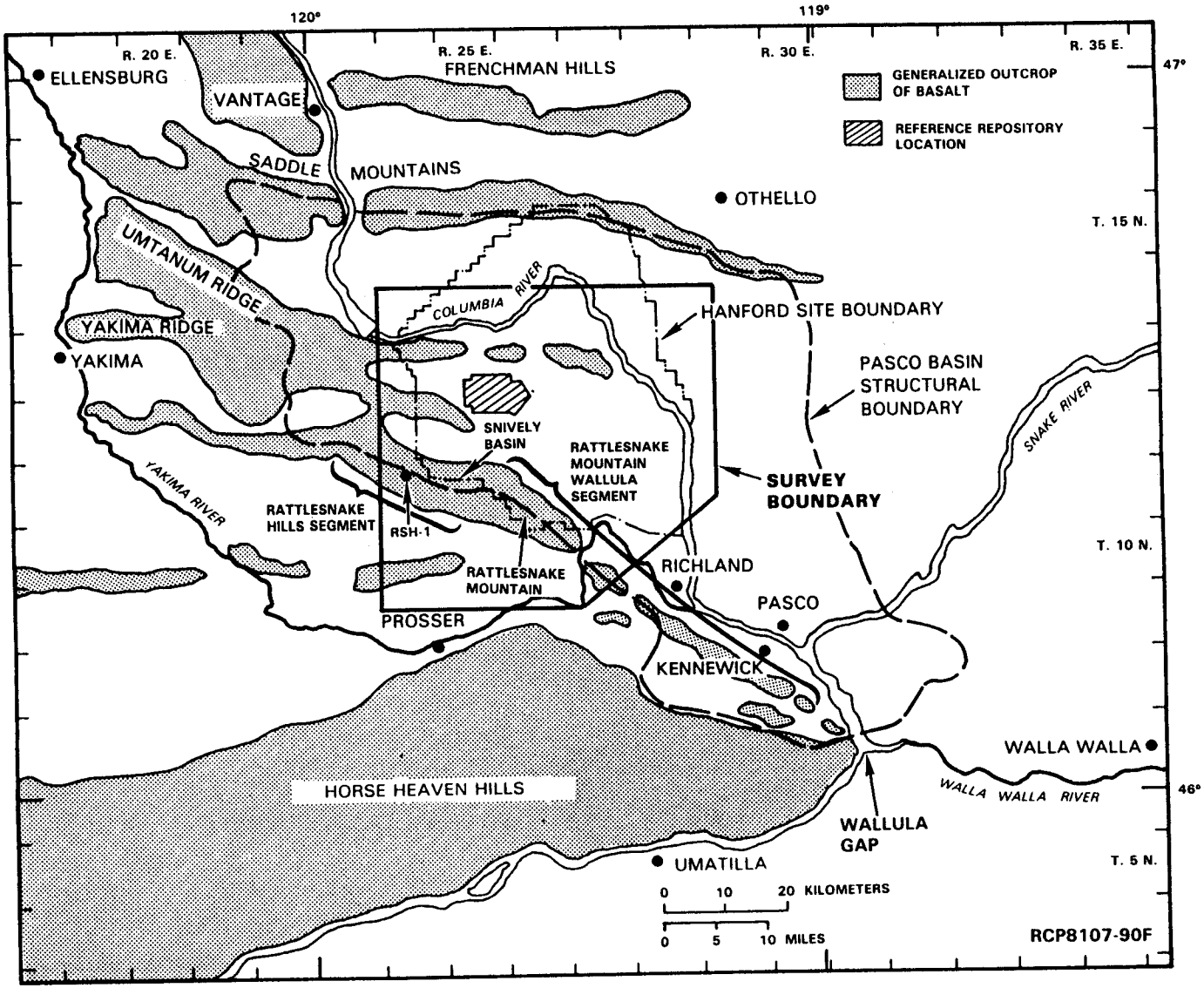

FIGURE 4-1. Study-Area Location Map.

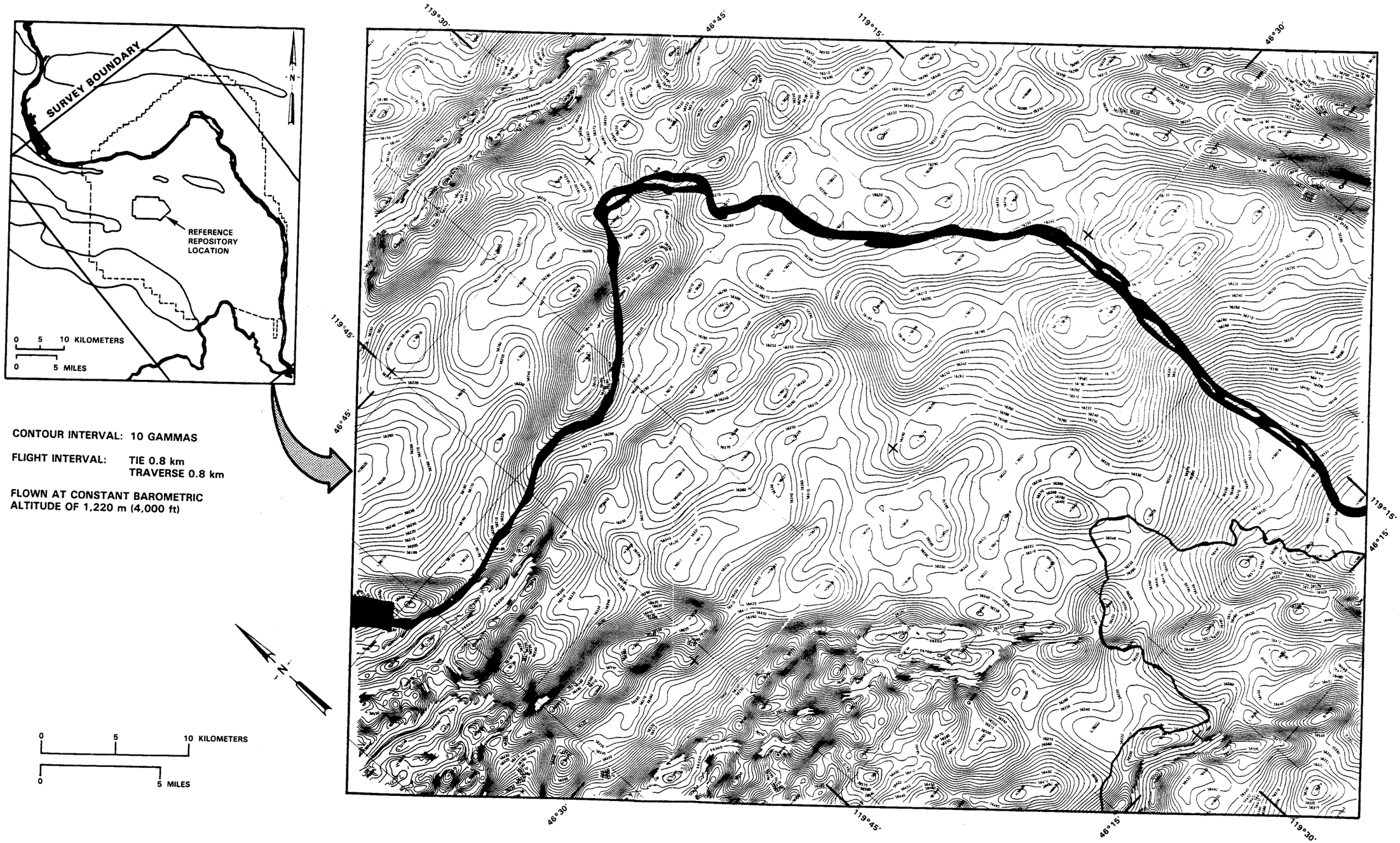

FIGURE 4-2. Aeromagnetic Total-Field Countour Map (1,220 m (4,000 ft)).

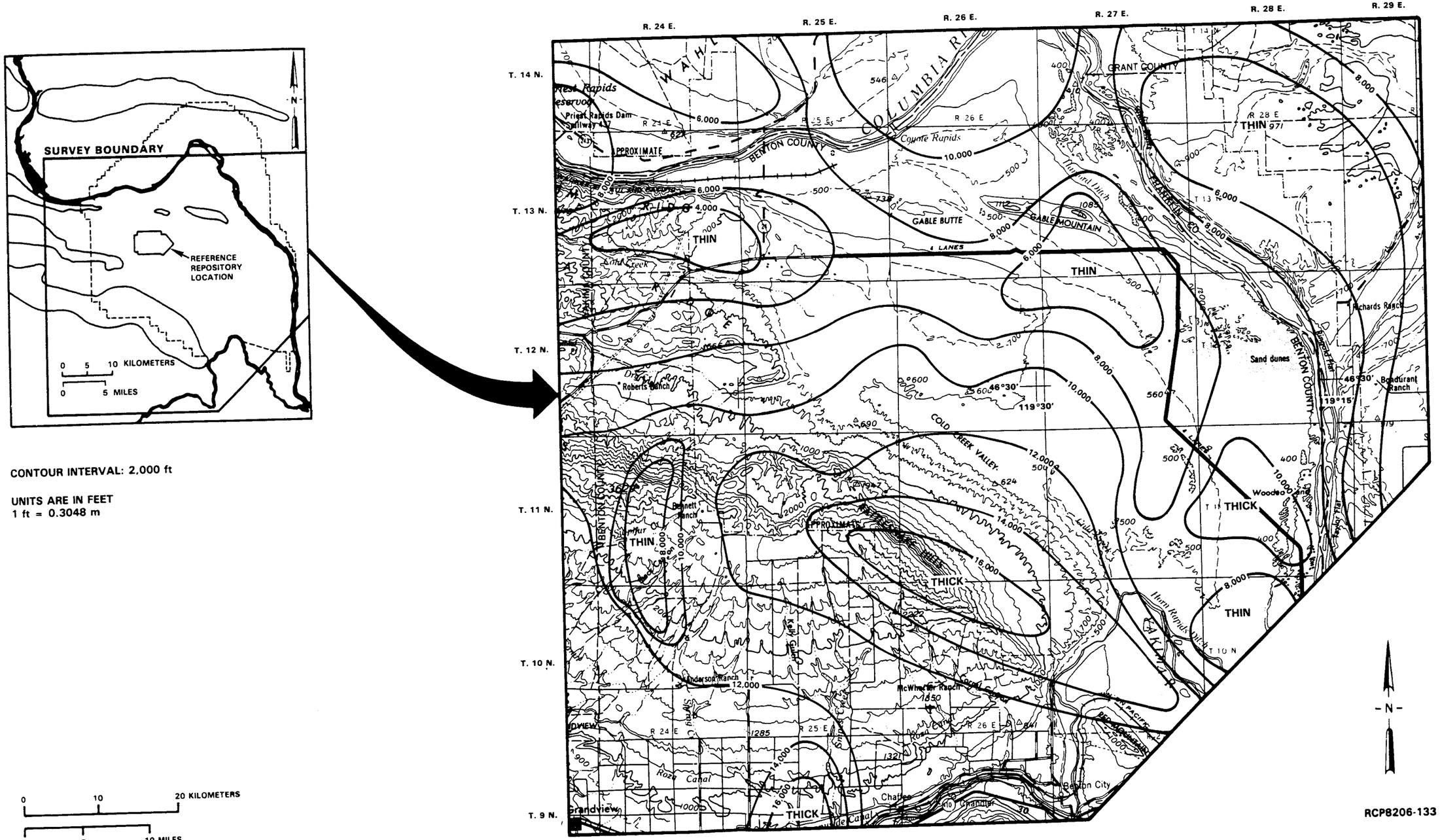

FIGURE 4-3. Isopach of the Volcanics, Geoelectric Layer 2.

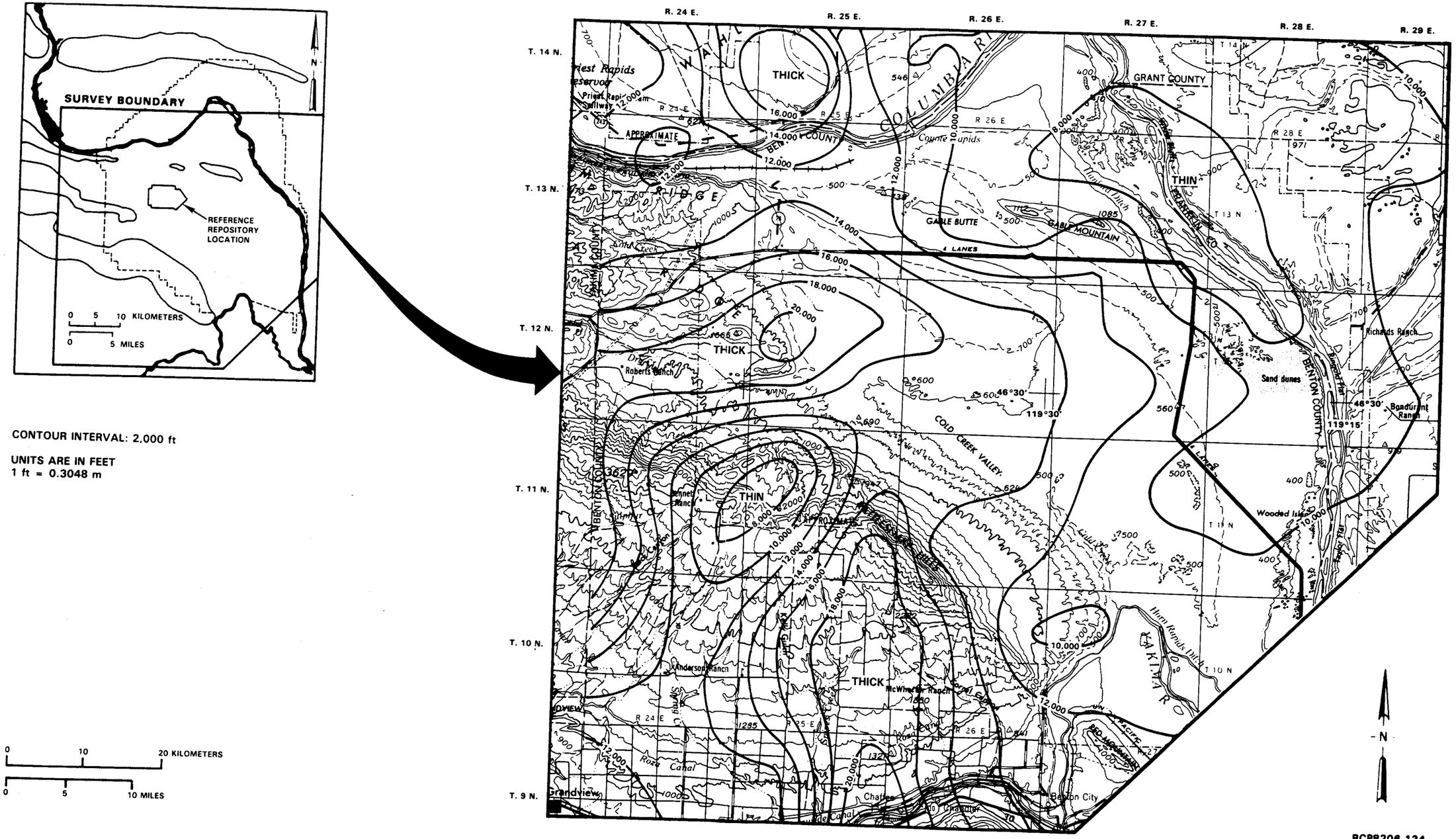

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FIGURE 4-4. Isopach of the Conductor, Geoelectric Layer 3.

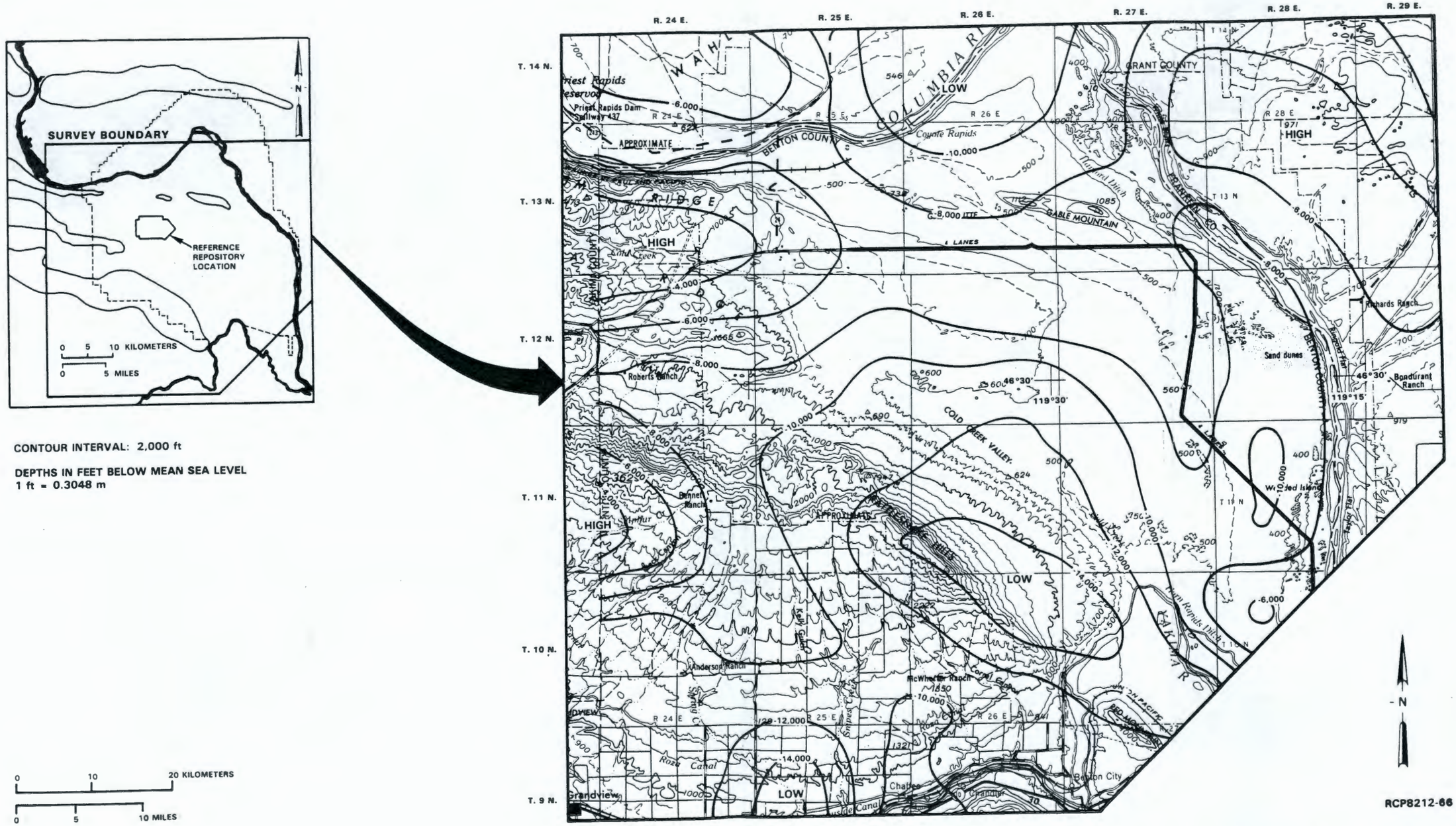

FIGURE 4-5. Contour Map of the Top of the Conductor, Geoelectric Layer 3.

FIGURE 4-6. Contour Map of The Top of the Interpreted Basement, Geoelectric Layer 4.

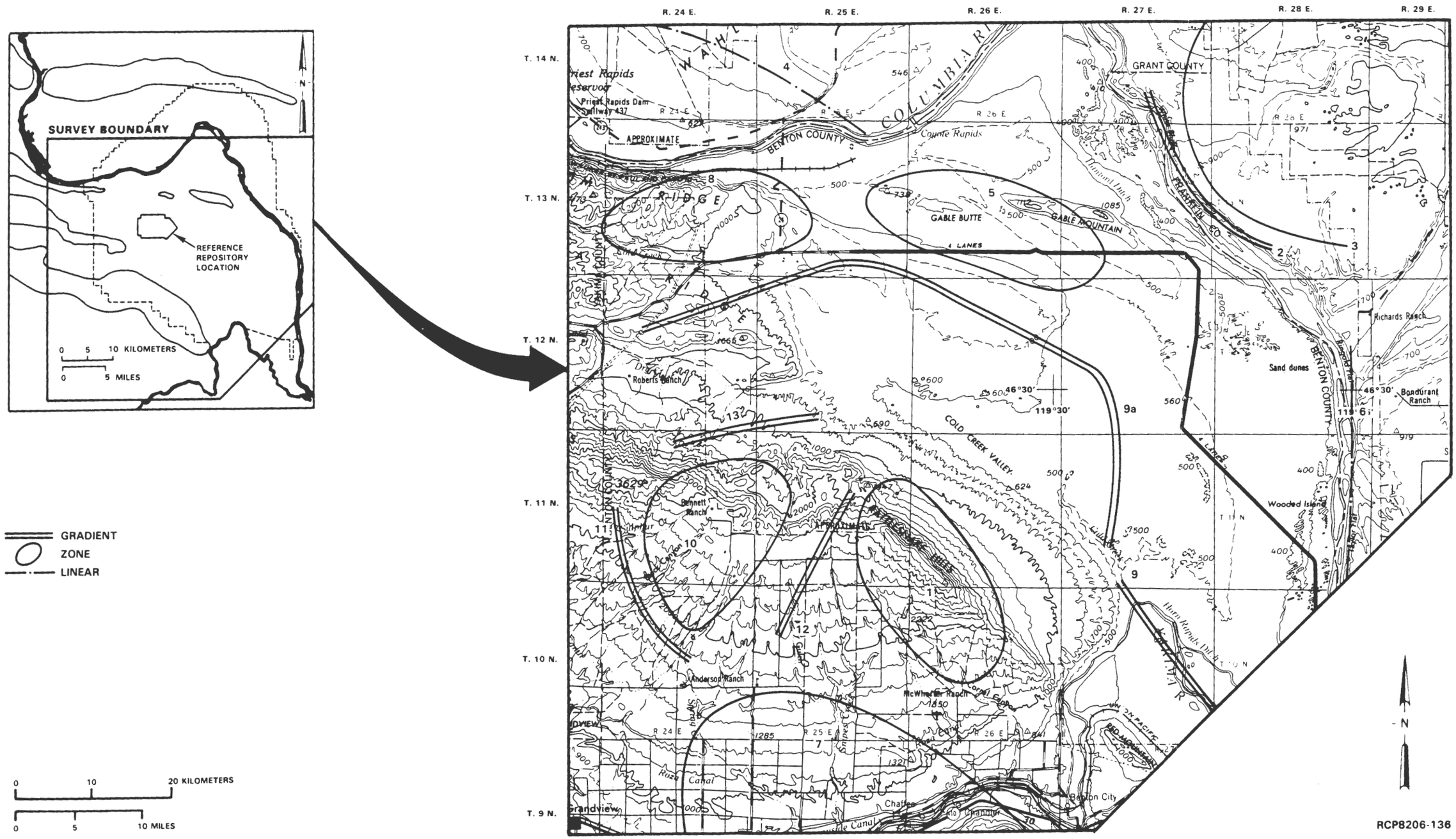

FIGURE 4-7. Geophysical Anomalies. A compliation of anomalies from Plate 4-2 and Figures 4-2 through 4-6. See Table 4-1 for the descriptions of anomalies.

TABLE 4-1. Selected Geophysical Anomalies.

Anomaly (see Fig. 4-7)	Location	Type*	Anomaly description
1	T. 10, 11 N., R. 25, 26 E.	II	Zone - 8-mgal NW-trending gravity low, gravity and magnetics are both affected by topography and are in part representing terrain effects; thick volcanics (12,000 to 16,000 ft); low-resistive basement (-32,000 ft); high magnetic gradients. Corresponds with Rattlesnake Mountain structure.
2	T. 13, 14 N., R. 27 E.	II or III	Gradient - NW-trending gravity linear caused by 6 to 8 mgal gravity gradient interpreted by Weston (1981) to be terrain effected due to the White Bluffs; volcanics thin SW to NE from 8,000 to 6,000 ft; 50 gamma corresponding magnetic gradient.
3	T. 13, 14 N., R. 28 E.	III	Zone - Gravity low; thin volcanics (6,000 ft); basement high (-16,000 ft); corresponds with magnetic high. Corresponds with edge of Palouse Slope.
4	T. 14 N., R. 24, 25 E.	III	Linear - NW-SE-trending gravity low; volcanics thinning to the SE.
5	T. 13 N., R. 25, 26 E.	I or II	Zone - W-NW-trending gravity high; volcanics thin to the E and W of the anomaly; moderate electrical anisotropy in SE portion; anomaly terminates to the E at abrupt N-S magnetic linear caused by steep magnetic gradient. Corresponds with Umtanum Ridge-Gable Mountain structure.
6	T.11, 12 N., R. 28 E.	III	Linear - N-S-trending gravity linear caused by 8 to 10 mgal gravity gradient; volcanics thin from 8,000 to 6,000 ft from west to east; corresponding N-S magnetic linear.
7	T. 8, 9, 10 N., R. 24, 25, 26 E.	III	Zone - Highest gravity values within the survey area; interpreted by Koníček (1974) to be due to thickness of volcanics; no magnetotelluric data for this area.
8	T. 12, 13 N., R. 24 E.	I	Zone - Volcanics thin to 4,000 ft along a E-W trend; basement is high (-16,000 ft); electrical anistropy corresponding with the western edge of the anomaly. Corresponds with Umtanum Ridge structure.
9	T. 10 N., R. 27, 28 E.	I or II	Gradient - Volcanics thin 6,000 ft from SW to NE, corresponding to 4,000 ft basement gradient; paralleling magnetic linear to the SW of the anomaly. Parallels Olympic-Wallowa lineament.
9a	T. 11-13, N., R. 24-27 N.	III	Gradient - 4,000- to 8,000-ft basement gradient deepening from N to S in western half of anomaly and E to W in the eastern half; the anomaly turns abruptly to the S at a magnetic low.
10	T. 11 N., R. 24 E.	II	Zone - Volcanics thin to 8,000 ft; basement high (-20,000 ft), magnetotelluric data indicates a geologically complex area; sparse gravity data. Corresponds with Snively Basin area.
11	T. 10, 11 N. R. 24 E.	III	Gradient - Volcanics thin by 4,000 ft from SW to NE into the anomaly; steep basement gradient (8,000 ft) with basement deepening to the W-SW.
12	T. 10, 11 N., R. 25 E.	II	Gradient - An 8,000- to 12,000-ft basement gradient connecting anomaly 10 (basement high) and anomaly 1 (basement low).
13	T. 12 N., R. 25 E.	III	Gradient - An 8,000-ft basement gradient deepening to the N of anomaly 10; corresponding moderate electrical boundary; corresponding E-W magnetic linear caused by N-S magnetic gradient.

To convert feet to meters, multiply by 0.3048.

*This refers to the type of relationship between deep and surficial structures as discussed in the text.

Techniques employed in the map construction include:

- All the inversions of the magnetotelluric data (see Appendix A) from which depths to geoelectric boundaries were selected, were honored (analogous to borehole data points). In the case of poor data quality, the depth information was used with discretion.

- The widely spaced magnetotelluric data coverage allowed freedom in the contouring. Bias in the spacing and orientation of the contour lines was applied based on significant trends in gravity, aeromagnetic, magnetotelluric electrical strike, and resistivity information with no intended bias towards the observed surficial geology.

- The maps were contoured in a conservative manner based on suggestions of Low (1977). This method avoids a highly interpretive contour scheme displaying faults, major fold systems, or other interpreted structures.

CENTRAL PASCO BASIN STRATIGRAPHY

The stratigraphy within the Pasco Basin is divisible into five geoelectrical layers based on magnetotelluric results (Myers/Price and others, 1979). Four of the five geoelectrical layers are interpreted to be pervasive throughout the central part of the Pasco Basin. From the surface, the five layers include: (1) a thin surficial electrically conductive layer (the nonpervasive unit), (2) an electrically resistive volcanic sequence corresponding to the Columbia River Basalt Group, (3) an electrically conductive (interpreted sedimentary) sequence, (4) an electrically resistive basement complex, and (5) a deep, electrically conductive zone (interpreted to be lower crust or upper mantle).

Layer 1 consists primarily of electrically conductive surficial sediments that overlie the more resistive volcanics. This unit, consisting primarily of the Hanford and Ringold Formations, is discussed in Tallman and others (1981) and Myers/Price and others (1979). This layer includes some upper Columbia River Basalt Group flows and interbedded Ellensburg Formation.

Geoelectric layer 2, the volcanics, is an electrically resistive unit composed primarily of the Columbia River Basalt Group and the Ellensburg Formation (see Chapter 2). The thickness of layer 2 in the Pasco Basin is quite variable and is interpreted to range from 1,220 to over 4,875 m (~4,000 to over 16,000 ft) (see Fig. 4-3). This thickness as well as all of the layer thicknesses discussed later is based on one-dimensional inversions of the magnetotelluric data, which, despite the limitations discussed in Appendix A, is a widely accepted first approximation performed on most magnetotelluric data; however, it should be emphasized that absolute depths and relief may be in error at this stage of interpretation. The volcanics generally thicken from north to south within the boundaries of the study area (see Fig. 4-3). The thickest section,

near Rattlesnake Mountain, trends northwest-southeast and is interpreted in places to be in excess of 4,875 m (16,000 ft) thick. Borehole RSH-1 reached a depth in excess of 3,050 m (10,000 ft) without penetrating the base of the basalt (Reidel and others, 1982). Koníček (1974), using gravity data, suggests a thick accumulation of basalts in the vicinity of Prosser, Washington (Fig. 4-1 and Plate 4-2) south of RSH-1. The basalt is interpreted to be thinnest (1,200 m (~4,000 ft)), in an area between Yakima Ridge and Umtanum Ridge in the western portion of the Pasco Basin (see Fig. 4-3).

The conductor, geoelectric layer 3 (see Fig. 4-4) is interpreted to be a sedimentary sequence. This layer is pervasive throughout the Pasco Basin with an estimated thickness ranging from 1,830 to 5,485 m (6,000 ft to >18,000 ft). This unit generally thickens from east to west. It is interpreted to be thinnest in an area to the northeast of Gable Mountain. The thickest sections are apparently on the southern side of Rattlesnake Hills and to the north and south of Umtanum Ridge. A relatively thin section (3,660 m (12,000 ft)) is located beneath the vicinity of Snively Basin in the Rattlesnake Hills area.

There are 2,440 m (8,000 ft) of interpreted relief on the top of layer 3 (see Fig. 4-5). This is more relief than interpreted on the top of basalt reported in Myers and Price (1981). The lowest area is 4,270 m (~14,000 ft) below sea level and the highest area is 1,830 m (6,000 ft) below sea level.

Layer 3 has been postulated to consist of mostly Mesozoic and early Tertiary sedimentary rocks (Myers/Price and others, 1979). A reconnaissance examination of lithologic units peripheral to the Columbia Plateau reveals a variety of geologic terrains that might be represented in the Pasco Basin in layer 3 (see Chapter 3 for more detail). To the west of the Columbia Plateau, in the area of the present Cascades, a volcanic arc was evolving in the lower Tertiary accompanied by thick accumulations of volcanic sediments and related extrusives. Concurrently, to the northwest, terrestrial sandstones with interbedded shales and conglomerates were being deposited (Frizzel, 1979). To the north, the Okanogan Highlands may have been a sediment source. The Idaho Batholith and Belt Series Supergroup, east and northeast of the Columbia Plateau, are also potential sources for the sediments within the Pasco Basin, due to an east to west paleoslope that was present at the onset of the eruption of westward tilting Columbia River basalts (Reidel and Fecht, 1981, 1982). South of the Columbia Plateau, 160 km (<100 mi) from the Pasco Basin, the early Tertiary was dominated by volcanism. These volcanic units were the source of sedimentary rocks in Oregon basins and may have contributed to the Pasco Basin. These are the major geologic terrains that are being considered to be directly or indirectly contributing to layer 3.

Layer 4, the resistive basement, is the least understood of all the layers. Its interpreted surface configuration is depicted on a contour map shown in Figure 4-6. The resistive basement is shown to have relief on the order of 4,875 m (16,000 ft) across the central Pasco Basin, but this amount of relief is speculative and possibly in error, due to limitations in the present level of magnetotelluric evaluation. Its minimum

depth, 4,875 m (16,000 ft) below sea level, is in the northeast and northwest corners of the central Pasco Basin and its maximum depth, 9,750 m (32,000 ft) below sea level, is located in the southcentral portion of the Pasco Basin. Two highs in the western portion of Figure 4-6 are paralleled by a north-south trending series of lows in the central portion of Figure 4-6. An east-west trending high appears to cross the north-south trends in the northern half of Figure 4-6.

Lithologic interpretation of the resistive layer 4 is tenuous. Some potential lithologies include: (1) Precambrian Belt Supergroup metasediments present in northeastern Washington and western Idaho to the north and northeast of the Columbia Plateau, (2) Paleozoic to Mesozoic metasediments and volcanics of the North Cascades and the Blue Mountains to the northwest and southeast of the central Columbia Plateau, and (3) plutonic and related metamorphic rocks of the northern Cascades and southern British Columbia, Omineca Province to the north of the Columbia Plateau.

Geoelectric layer 5, the conductive basement, is identified below 21 km (70,000 ft). This layer is interpreted to be lower crust or upper mantle (Myers/Price and others, 1979).

DISCUSSION

Surface topographic highs and lows do not consistently correspond to highs and lows on the top of layer 3. However, the lowest elevation in the top of layer 3, located in the vicinity of Rattlesnake Mountain (see Fig. 4-5), coincides with the thickest accumulation of basalts (see Fig. 4-3). This suggests that there was subsidence in the central Pasco Basin contemporaneous with the deposition of the Grande Ronde Basalt causing pooling of the earlier Grande Ronde Basalt, as suggested by Landon and others (1982) and Reidel and others (1982). A pre-Grande Ronde Basalt topographic low may also have been present beneath borehole RSH-1, but the magnitude of relief suggests that the entire thickness of Grande Ronde Basalt is not due to just paleotopographic relief but also to downwarping in Miocene time, perhaps relating to crustal extension.

The relief on the top of layer 4 is much greater than the relief observed on layer 3 (see Fig. 4-5 and 4-6). Many of the highs and lows expressed in layer 4 spatially correspond with the highs and lows in layer 3. These corresponding highs and lows do not, however, clearly correlate with surface topography and structural trends observed in the volcanics, layer 2. There does not, as presently contoured, appear to be any dominant structural trends in the basement similar to the predominant northwest-southeast and east-west structural trends mapped at the surface. This suggests a change in character of the regional deformation between the development of some of the pre-Columbia River Basalt Group structures and the development of the Yakima folds.

An extensional environment has been suggested as a past stress regime prior to Yakima fold development (Waters and others, 1981; Davis, 1977, 1981; Laubscher, 1981; Reidel and others, 1982). Transcurrent stresses have also been suggested as a past stress regime. The high relief of the pre-volcanic rocks (see Fig. 4-5 and 4-6), if correct, may indicate vertical tectonics or a block faulted basement. The lack of consistency in trends and in the relief of layer 4 suggests that some structures may have been altered by later tectonic processes.

Based on integrated geophysical data, 13 anomalies that may have deep structural significance have been identified. They are summarized on Table 4-1 and shown in Figure 4-7. The observed anomalies have been grouped into three categories: (1) linears, (2) gradients, and (3) zones. These are broad categories with overlap. Linears represent changes in the character of contour lines or a series of highs or lows that form an alignment. The gradients represent areas of high rates of change of the contour lines relative to the surrounding area. The absolute amplitude of several of these gradients may be questionable due to limitations in the magnetotelluric depth inversion process as discussed in Appendix A. Zones are highs and lows in the gravity and magnetic fields (see Fig. 4-2 and Plate 4-2), anomalously thin or thick sections in the isopachs (see Fig. 4-3 and 4-4), or highs or lows in the top of the basement contour map (see Fig. 4-6). These depicted anomalies could be the result of a wide variety of geologic structural configurations such as folds or faults. They may also reflect erosional features, thickening and thinning of geoelectric units, or lithologic variations within geoelectric layers.

Identification of the geologic cause of the geophysical anomalies in Figure 4-7 may not be possible, because it is not practical to adequately drill these anomalies. A discussion of the interpreted geologic significance of these anomalies follows.

Three fundamental relationships between deep structures and structures observed at the surface are recognized. One relationship, designated here as Type I, is characterized by anomalies interpreted to reflect deep structures that occur at about the same locality as surface structures and that are interpreted to be vertically continuous with surface structures. These are considered to be a single structure. A second relationship, Type II, includes interpreted deep structures that spatially correspond with surface deformation but are not vertically continuous structures. However, these deep structures may have been influential in the development of the surface structure, although the manner of interaction is not clearly understood. For example, these deep structures may have acted as buttresses around which the basalt and the Yakima folds have been translated as suggested by Price (1982). These deep structures could also represent boundaries for microplates or tectonic blocks analogous to those proposed by Bates and others (1981), Magill and Cox (1981), and Magill and others (1981, 1982).

The third relationship, Type III, is where there is no apparent correlation or direct relationship between interpreted deep structures or anomalies and structures observed at the surface. In the Pasco Basin, all three types of relationships are observed. None of the three appear to be predominant. Some examples of these types are discussed below.

From qualitative analysis of the geophysical data presented in this chapter, it appears that there is a Type I relationship at Umtanum Ridge where the interpreted deep structure spatially correlates with the southern flank of the Umtanum Ridge anticline. The deep structure is best depicted on the isopach of the volcanics (see Fig. 4-3 and 4-7, anomaly 8) where the volcanics are interpreted to be anomalously thin and on the contour map of geoelectric layer 3 (see Fig. 4-5), which shows a high along the southern flank of Umtanum Ridge.

Reidel and Fecht (1981) have shown from observations of the thickness variations in the basalt flows, that Umtanum Ridge was an actively developing topographic feature by Wanapum time. They were able to calculate the rate of uplift since Wanapum time, but were unable to draw conclusions on its earlier development because of the lack of data.

Some conclusions may be drawn from the geophysical observations at Umtanum Ridge. One possibility is that the thinning of the volcanics was due to the existence of Umtanum Ridge as an asymmetric topographic high with a south plunging fold axis earlier in Grande Ronde time. If Umtanum Ridge began developing in early Grande Ronde time and the interpreted relief at the base of layer 2 is reasonable, then higher average rates of uplift in Grande Ronde time, than those reported for Wanapum and Saddle Mountains time, are essential to accommodate the relief observed at the base of layer 2 at Umtanum Ridge. Higher rates of differential uplift and subsidence have been suggested in the Pasco Basin for Grande Ronde time relative to Wanapum and Saddle Mountains time (see Fig. 2-3, Chapter 2) (Landon and others, 1982). Higher rates of uplift or subsidence in Grande Ronde time would explain the greater amount of relief observed at the base of the basalt than at the top. Another interpretation for the relief observed at Umtanum Ridge might be that the ridge was a major topographic high at the onset of the volcanic deposition followed by continued deformation at a constant rate. In either case, with the present data, Umtanum Ridge appears to be an example of a Type I relationship.

In the Snively Basin area, a Type II relationship is illustrated by an assemblage of anomalies (anomalies 10, 11, 12 and 13, Table 4-1; Fig. 4-7) that are spatially related to complex structural deformation mapped at the surface. Snively Basin is located where the Rattlesnake Hills structure changes trend from a northwest-southeastern trend (Rattlesnake Mountain-Wallula segment) to an east-west trend (Rattlesnake Hills segment, Fig. 4-1 and Plate 2-2). A corresponding change in geologic character between the segments has been reported (Myers/Price and others, 1979). The Wanapum and Saddle Mountains Basalts have been shown to thin over Rattlesnake Mountain in the Rattlesnake-Wallula segment, but do not significantly thin over Snively Basin (Reidel and Fecht, 1981). Correspondingly, layer 2 is interpreted to be much thinner beneath Snively Basin than beneath Rattlesnake Mountain (see Fig. 4-3). Beneath Snively Basin, considerable relief is also interpreted in layer 4 (see Fig. 4-6).

Based on the above observations, some preliminary interpretations may be made for Snively Basin. The data indicate a gradient that could represent some type of structural discontinuity between Snively Basin and the Rattlesnake Mountain-Wallula segment (see the anomaly 12 description in

Table 4-1). The gradient apparently existed prior to the deposition of the earliest basalts and appears to be related to a deep structure beneath Snively Basin that is observed in layer 4. This interpreted deep structure may have influenced the development of the bend of the Rattlesnake Hills structure.

The Snively Basin area lies within a diffuse zone known as the Cle Elum-Wallula zone of deformation. The Cle Elum-Wallula zone of deformation is an area of relatively intense deformation lying within the eastern portion of the Yakima Fold Belt subprovince. Various interpretations have been proposed for the origin of the Cle Elum-Wallula zone of deformation, which is generally considered to be a fundamental element in several tectonic models for the Columbia Plateau (see Chapter 7). Several models suggest involvement of pre-Columbia River basalt rock in the deformation of basalt along the Cle Elum-Wallula zone of deformation, but due to limited data, the amount of basement influence on structural deformation is not well understood. Data for the Snively Basin area suggest that some of the structural complexities (i.e., abrupt changes in structural trends) along the Cle Elum-Wallula zone of deformation may best be explained by sub-basalt Type II relationships.

The Yakima folds are characteristically long narrow anticlines as exemplified by the Umtanum Ridge structure. If the Umtanum Ridge structure is typical of the Yakima folds, then it is possible that Type I may be the predominant relationship between the pre-Columbia River basalt structures and the anticlinal ridges. To the east of the Pasco Basin, in the Palouse subprovince, there is little surficial structural deformation (Plate 3-1). This suggests that there may be minor relief in the pre-Columbia River basalt strata to the east of the Pasco Basin; however, there is limited data to substantiate this hypothesis.

Examples of Type III relationships are listed in Table 4-1.

SUMMARY

The BWIP investigations to date include a qualitative integration of magnetotelluric, gravity, and aeromagnetic data with a preliminary geologic assessment of the integrated interpretation. Magnetotelluric data have proven to be the most useful of the data sets, supplying information on the electrical stratigraphy and approximated depths to electrical horizons. Gravity, electrical strikes, and aeromagnetic data are used to interpret between magnetotelluric stations. Five electrical layers are observed in the magnetotelluric data. Layer 1 is interpreted as a conductive layer composed primarily of sediments overlying the Columbia River basalt. Layer 2 is a resistive unit interpreted to be primarily the Columbia River Basalt Group and appears to vary significantly in thickness with the maximum thickness in the vicinity of Rattlesnake Hills and minimum thickness on the south side of Umtanum Ridge. Layer 3, a conductive layer, is interpreted to consist primarily of sedimentary rock. Layer 4 is considered to be the resistive basement complex, and layer 5 is a conductive zone, which is interpreted to define the lower crust/upper mantle boundary.

Layers 2, 3, and 4 are detected at all of the magnetotelluric stations. Layer 5, the conductive basement, is observed at several magnetotelluric stations.

Interpreted relief on the top of the geoelectric layers generally decreases to the east within the survey area. Layer 4 shows the greatest amount of relief, with layers 3 and 2 showing progressively less. There are no unequivocal structural trends interpreted in the basement similar to the east-west and northwest-southeast trends observed at the surface.

Several geophysical anomalies were identified that may have deep structural sources. These anomalies were examined qualitatively and preliminary interpretations were made in order to assess possible relationships between surface structures and basement structures. Umtanum Ridge is an area of Type I relationship, where surface and basement structures appear directly related and are interpreted to be one continuous structure. Snively Basin is interpreted to be an area of Type II relationship where basement structures may have significantly influenced the development of surface structures, but are not the same vertically continuous structure. Type III relationships are where interpreted basement structures have no apparent influence on surface structure. No predominant type of relationship is apparent in the central Pasco Basin.

The data presented in this chapter suggesting potential structures underlying the basalt sequence in the Pasco Basin do not yet permit a choice among the tectonic models discussed in Chapter 7. Interpretation of possible basement structures is preliminary and warrants further refinement due to (1) relatively broad magnetotelluric station spacing, requiring much inference between some data points, (2) the location and spacing of stations relative to several interpreted deep structures, permitting only equivocal interpretation, (3) the restricted resolution inherent in the geophysical methods used (see Appendix A), (4) the present level of qualitative integration without quantitative modeling support, and (5) the difficulty in "geophysically recognizing" strike-slip faulting where there is a juxtaposition of rocks having little contrast in rock properties. Strike-slip faulting is of particular interest because it is an integral element in many of the most recently proposed tectonic models (Davis, 1981; Laubscher, 1981; Price, 1982; Bentley, 1982).

Several observations from this qualitative interpretation suggest some limitations on tectonic models for the Columbia Plateau:

- A throughgoing northwest-southeast feature has not been detected at the base of or below the volcanics along the trend of the Cle Elum-Wallula zone of deformation.

- The possible difference in the type of deep structure beneath Snively Basin and Umtanum Ridge, both of which occur within the Cle Elum-Wallula zone of deformation (i.e., different interpreted-type relationships), suggests that either a diversity of mechanisms was influential in the development of the Cle Elum-Wallula zone of deformation or Cle Elum-Wallula zone of deformation in the Pasco Basin area is not a function of major continuous basement structures.

- The relief of the pre-volcanic strata decreases across the survey area from west to east and this generally corresponds with the decreasing structural relief observed from west to east in the surface structures. This changing relief suggests that the pre-Columbia River Basalt Group structure and stratigraphy may have in some way influenced the development of folds observed at the surface.

- The relief on the top of layer 3 is incompatible with any large regional décollement at the base of layer 2, although zones of limited detachment cannot be precluded.

CHAPTER 5 - CONSTRAINTS ON TECTONIC MODELS AS PROVIDED FROM STRAIN RATES

S. P. Reidel, R. W. Cross, K. R. Fecht

INTRODUCTION

The Pasco Basin (Fig. 5-1) is one of several structural basins in the central and western Columbia Plateau and occurs at the boundary between the Yakima Fold Belt subprovince and the Palouse subprovince (see Fig. 2-10). These subprovinces are defined by specific structural trends within each region. The Yakima Fold Belt subprovince is dominated by east-west-trending anticlines and synclines, whereas a westward-tilting regional paleoslope characterizes the Palouse subprovince.

Tectonic models for the Columbia Plateau can be constrained by the geologic history of areas in the Pasco Basin that have undergone deformation and the rates at which deformation occurred. Emphasis in this paper is not on structural style and the effect of fold geometry, rather it is on uplift rates of the anticlines, subsidence rates of the synclines, and crustal shortening. Data used to assess uplift and subsidence are from studies of the lateral extent and thickness variations of basalt flows and sedimentary interbeds (Reidel and others, 1980; Reidel and Fecht, 1981, 1982) and from the suprabasalt sediments (Tallman and others, 1981). These studies will only be summarized here and the reader is referred to the original work for a more detailed discussion of the data and interpretations.

THICKNESS AND LATERAL EXTENT OF BASALT FLOW AND SEDIMENTS

The lateral extent and thickness variations among flows in the Pasco Basin are the product of a collective interplay of four factors (Reidel and others, 1980; Reidel and Fecht, 1981): (1) the volume of each flow, (2) the location of flow margins (constructional topography), (3) the effect of uplift and subsidence, and (4) the influence of a regional paleoslope. The flow eruption and emplacement rate and the rate of uplift and subsidence are so closely related that the evolution of the basin through the Miocene can be reconstructed (Reidel and others, 1980; Reidel and Fecht, 1981, 1982; Long and Landon, 1981; Landon and others, 1982; Fecht and others, 1982). A summary of the results of studies on the Columbia River Basalt Group and intercalated sedimentary units in the Pasco Basin follows.

The Columbia River Basalt Group consists of five formations: Imnaha, Grande Ronde, Picture Gorge, Wanapum, and Saddle Mountains Basalts. Imnaha Basalt is the oldest known Columbia River basalt; the Picture Gorge Basalt physically resembles it but is equivalent in age to the older Grande Ronde Basalt. Picture Gorge and Imnaha Basalts comprise ~11 vol% of Columbia River basalt and thus far have not been observed in boreholes or outcrop in the Pasco Basin.

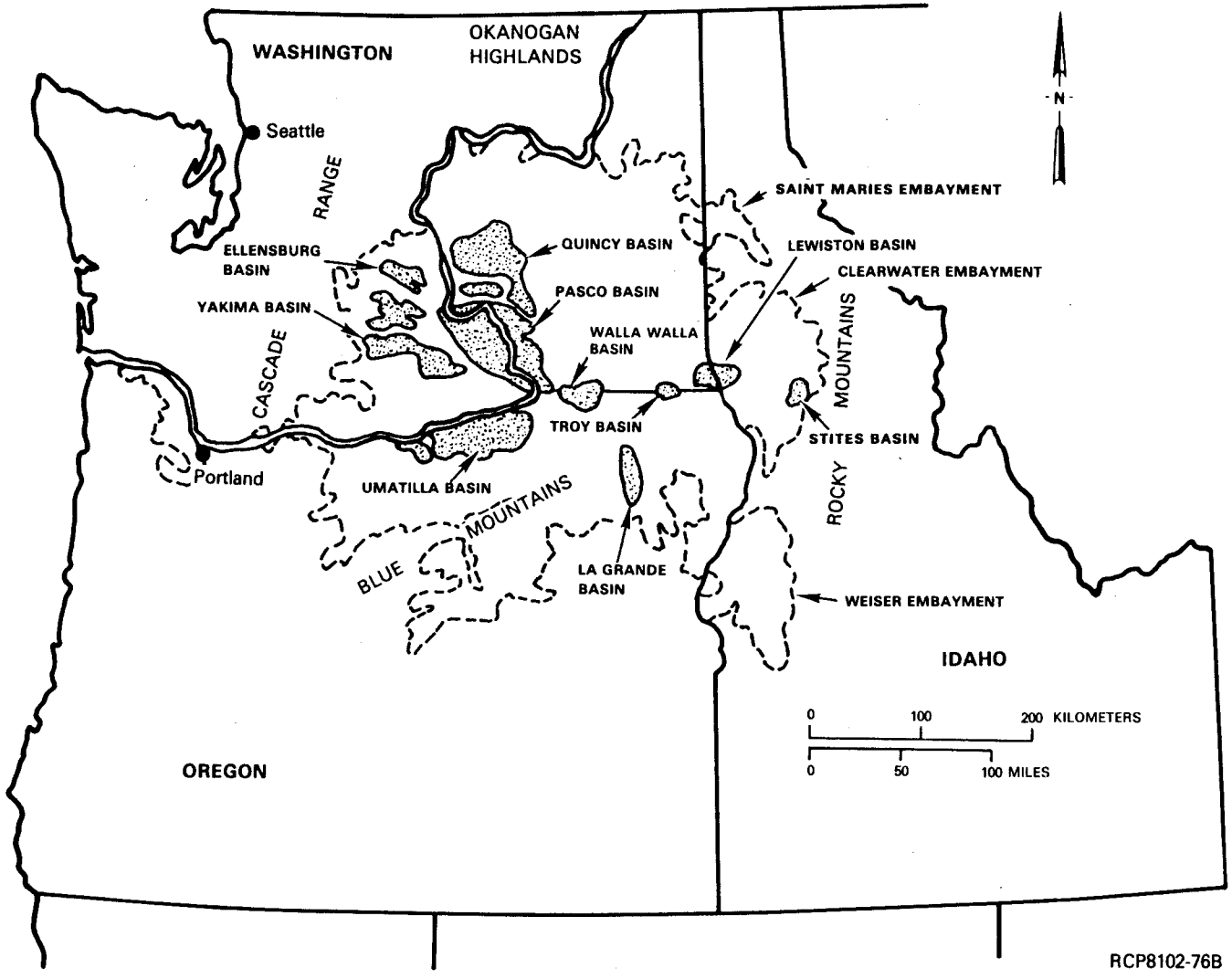

FIGURE 5-1. The Columbia Plateau Showing Locations of Major Structural Basins.

The Grande Ronde Basalt is 85 vol% (325,000 km^3, Reidel and others, 1982) of the Columbia River Basalt Group and was erupted over about a 2-m.y. period (16.5 to 14.5 million years before present (m.y.B.P.); Watkins and Baksi, 1974). The greatest known thickness of basalt occurs in the Pasco Basin with >3 km of Grande Ronde Basalt in borehole RSH-1. Flow distribution and individual isopach maps for the late Grande Ronde Basalt indicate that its lateral extent and thickness are controlled principally by flow volume and topography at the time of emplacement. According to Long and Landon (1981) and Landon and others (1982), topography at the time of flow emplacement was principally constructional with some structural control. Flow volume varies; some flows cover the entire basin and others have only a limited extent.

The Wanapum Basalt consists of ~3 vol% of the Columbia River Basalt Group and was erupted over about a 1-m.y. period (14.5 to 13.5 m.y.B.P.). Consequently, there was a greater length of time between the eruption of flows, and thus, the lateral extent and thickness variation of flows were more strongly influenced by uplift and subsidence and the regional paleoslope (Reidel and Fecht, 1981).

The Saddle Mountains Basalt consist of <1 vol% of the Columbia River Basalt Group, yet was erupted over a 7-m.y. period (~13.5 to 6 m.y.B.P.). Consequently, more time elapsed between eruptions allowing the accumulation of thick sedimentary deposits and erosion by rivers and streams. The lateral extent and thickness of flows in the Pasco Basin are controlled primarily by uplift and subsidence, and secondarily, by flow volume, constructional topography, and the regional paleoslope. The thickness and lateral extent of interbedded sedimentary units of the Ellensburg Formation are similarly controlled (Reidel and Fecht, 1981).

Reidel and others (1980) and Reidel and Fecht (1981, 1982), using thickness variations and the lateral extent of basalt flows and intercalated sedimentary units, demonstrated that the Pasco Basin was structurally evolving during the Wanapum and Saddle Mountains times. The great volume and rapid eruption rate of the Grande Ronde Basalt, combined with limited surface exposures, masked any small developing tectonic feature; however, several broad tectonic features, such as the Hog Ranch anticline (Reidel and Fecht, 1981; Landon and others, 1982) and structural subsidence in the Pasco Basin (Reidel and others, 1982; Landon and others, 1982), were evident.

The growth of several Yakima folds during Grande Ronde time can be demonstrated, however (Price, 1982; Reidel and others, 1982). The Saddle Mountains were growing in late Grande Ronde time as evidenced by the thinning and absence of flows of the high-Mg chemical type (Fig. 5-2) near Sentinel Gap. Price (1982) documented evidence for thinning of the Umtanum flow near Priest Rapids Dam (see Fig. 1-1), but was uncertain as to whether thinning is related to growth of the Umtanum Ridge or the Hog Ranch anticline to the west. The absence of the Vantage interbed (Fig. 5-2) along portions of the Saddle Mountains, Umtanum Ridge, and other Yakima folds suggests that most Yakima folds were growing in Grande Ronde time.

PERIOD	EPOCH	GROUP	SUBGROUP	FORMATION	K-Ar AGE YEARS x 10^6	MEMBER OR SEQUENCE	GEOLOGIC MAPPING SYMBOL	SEDIMENT STRATIGRAPHY OR BASALT FLOWS	
QUATERNARY	Pleistocene/ Holocene					SURFICIAL UNITS	Ql	LOESS	
							Qd	SAND DUNES	
							Qa, Qaf	ALLUVIUM AND ALLUVIAL FANS	
							Qld	LANDSLIDES	
							Qt	TALUS	
							Qco	COLLUVIUM	
	Pleisto-cene			Hanford		TOUCHET BEDS/ PASCO GRAVELS	Qht/Qhp		
TERTIARY	Pliocene			Ringold			Trs	UPPER RINGOLD	FANGLOMERATE
							Trc	MIDDLE RINGOLD	
							Trls	LOWER RINGOLD	
							Trg	BASAL RINGOLD	
	Miocene	Columbia River Basalt Group	Yakima Basalt Subgroup	Saddle Mountains Basalt	8.5	ICE HARBOR MEMBER	Ti: Tig	GOOSE ISLAND FLOW	ELLENSBURG FORMATION Tel
							Tim	MARTINDALE FLOW	
							Tib	BASIN CITY FLOW	
								LEVEY INTERBED	
					10.5	ELEPHANT MOUNTAIN MEMBER	Tem: Tem2	UPPER ELEPHANT MOUNTAIN FLOW	
							Tem1	LOWER ELEPHANT MOUNTAIN FLOW	
								RATTLESNAKE RIDGE INTERBED	
					12.0	POMONA MEMBER	Tp: Tp2	UPPER POMONA FLOW	
							Tp1	LOWER POMONA FLOW	
								SELAH INTERBED	
						ESQUATZEL MEMBER	Te: Te2	UPPER GABLE MOUNTAIN FLOW	
								GABLE MOUNTAIN INTERBED	
							Te1	LOWER GABLE MOUNTAIN FLOW	
								COLD CREEK INTERBED	
						ASOTIN MEMBER	Ta	HUNTZINGER FLOW	
						WILBUR CREEK MEMBER	Tw	WAHLUKE FLOW	
						UMATILLA MEMBER	Tu: Tus	SILLUSI FLOW	
							Tuu	UMATILLA FLOW	
				Wanapum Basalt	13.6			MABTON INTERBED	
						PRIEST RAPIDS MEMBER	Tpr: Tpri	LOLO FLOW	
							Tprr	ROSALIA FLOWS	
								QUINCY INTERBED	
						ROZA MEMBER	Tr: Tr2	UPPER ROZA FLOW	
							Tr1	LOWER ROZA FLOW	
								SQUAW CREEK INTERBED	
						FRENCHMAN SPRINGS MEMBER	Tf: Tfa	APHYRIC FLOWS	
							Tfp	PHYRIC FLOWS	
				Grande Ronde Basalt	14.5			VANTAGE INTERBED	
						SENTINEL BLUFFS SEQUENCE	Tsb	UPPER FLOWS	
								MIDDLE SENTINEL BLUFFS FLOW	
								LOWER FLOWS	
								McCOY CANYON FLOW	
						SCHWANA SEQUENCE	Ts	INTERMEDIATE-Mg FLOW	
								LOW-Mg FLOW ABOVE UMTANUM	
								UMTANUM FLOW	
								HIGH-Mg FLOW BELOW UMTANUM	
								VERY HIGH-Mg FLOWS	
					16.5			AT LEAST 30 LOW-Mg FLOWS	

RCP8108-1D

FIGURE 5-2. Stratigraphy of the Pasco Basin.

Thickness variations and the lateral extent of all Wanapum and Saddle Mountains Basalt flows and some Grande Ronde Basalt flows indicate that the present Yakima folds were actively growing through much of Miocene time (Reidel and others, 1980; Reidel and Fecht, 1981, 1982). Deformation was concentrated along present first-order structural trends, as shown by areas of maximum thinning consistently corresponding to anticlinal axes and areas of maximum thickening corresponding to synclinal troughs.

REGIONAL WESTWARD TILTING PALEOSLOPE

The westward tilting paleoslope (Fig. 5-3) represents a major regional feature that has dominated the eastern Columbia Plateau since at least the early Miocene. Swanson and Wright (1976) have convincingly argued that the paleoslope has had a nearly constant slope since the late Miocene (12 m.y.B.P.). We calculate this gradient to be ~1.5 to 2.0 m/km. The gradient of the paleoslope was not constant with respect to one datum prior to 12 m.y.B.P., however (Swanson and Wright, 1976). From thinning of flows onto the paleoslope, a gradient of 2 to 5 m/km can be calculated for most flows in the late Grande Ronde (Long and others, 1980; Reidel, in press; and Landon and others, 1982), and a gradient of ~2 m/km can be calculated for Wanapum flows. The change from subsidence in the west, producing an increasing gradient with time for any particular horizon in Grande Ronde, Wanapum, and early Saddle Mountains times, to a constant gradient from the middle Saddle Mountains time to present suggests a change from overall subsidence to the west and uplift to the east to either reduced rates of differential subsidence and uplift or more restricted areas of uplift and subsidence with broad unaffected areas.

In any event, the westward tilting paleoslope represents a major regional feature from the early Miocene to present. The stability of the gradient of this paleoslope since at least 12 m.y.B.P. permits its use as a reference plane in order to estimate areas of relative uplift and subsidence. The relative uniformity of the surface gradient at the time of eruption of basalt flows prior to 12 m.y.B.P. also suggests that this paleoslope can be used to define areas of uplift and subsidence during the early and middle Miocene.

RATES OF UPLIFT AND SUBSIDENCE

Previous estimates of the rate of uplift have been based on topographic relief that developed after the basalts were emplaced (Brown, 1970; Kienle and others, 1978; Glass, 1977; Shannon and Wilson, 1978). Brown (1970) estimated a slow rate of basining that may still be occurring today, but argued that anticlinal uplift began at the end of the basalt eruptions. He estimated a growth rate of 1 m/10,000 yr (0.1 mm/yr) for the anticlinal ridges.

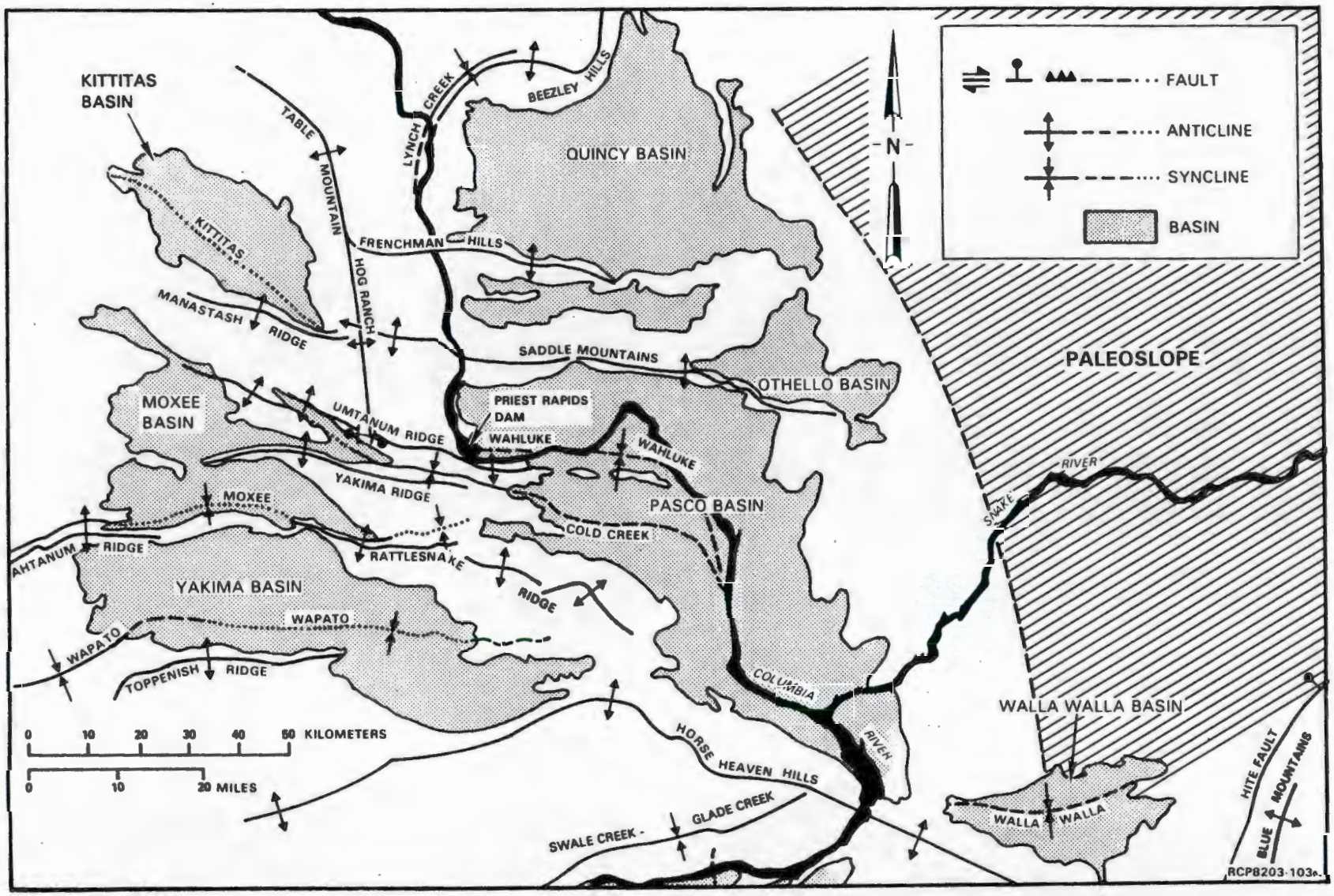

FIGURE 5-3. Major Structural Elements of the Pasco Basin and Surrounding Area.

The difference in thickness of non-eroded flows from the synclinal valleys to anticlinal ridges can be used in conjunction with the age of flows to estimate the amount of paleorelief on structurally controlled topography during the Miocene (Reidel and others, 1980). Uplift and subsidence are assumed to be principally caused by folding with some undetermined amount of faulting. Because the flows thin over ridges and less often pinch out on the flank, they reflect the relief at the time they were emplaced. By determining the relief for each time interval, as recorded by the flows, the rate of relief development can be calculated. Taking the maximum variation observed in the Pasco Basin, which occurs between Rattlesnake Mountain and the Cold Creek syncline (Fig. 5-4), the combined rate of uplift and subsidence was found not to exceed 150 m/m.y. (0.15 mm/yr) over the time interval of 14.5 to 10.5 m.y.B.P. (Fig. 5-5) and decreases from Wanapum to Saddle Mountains time.

In order to differentiate rates of uplift and rates of subsidence, it is necessary to have a regional frame of reference. Because the regional paleoslope is the dominant feature east of the Pasco Basin and has been relatively stable since 12 m.y.B.P., it was chosen as a frame of reference in order to differentiate uplift and subsidence.

Extrapolation of the paleoslope into the Pasco Basin, using the 2 m/km gradient that has prevailed since at least 12 m.y.B.P., is shown in Figure 5-6. The extrapolated paleoslope approximately intersects the basalt-sediment contact observed today suggesting that this extrapolation is a good approximation of areas that have not undergone any net uplift or subsidence with reference to the regional slope.

The rate at which respective areas were subsiding and undergoing uplift during the Miocene relative to the regional paleoslope can be calculated two ways. The first is to take measured thicknesses of basalt flows that occur near this extrapolated paleoslope and compare them to the measured thicknesses of flows from the crest of anticlines and troughs of synclines. The second method is to first calculate the combined uplift and subsidence rate for the Miocene, then take the present observed structural relief and use the extrapolated paleoslope to determine the amount of relief that has occurred by uplift and by subsidence. Using the proportions that have been uplifted and subsided, the combined uplift and subsidence rate is then divided using these same proportions.

In order to use these methods, several assumptions are necessary. First, it is assumed that the measured thicknesses at the intersection with the paleoslope represent the true thicknesses that would have occurred in the area, had uplift or subsidence not occurred. Second, it is assumed that the structural relief resulting from both uplift and subsidence was occurring at the same time and the relative proportions remained constant. These assumptions are difficult to assess, especially because very little data can be collected for flow thicknesses near the extrapolated intersection of the paleoslope.

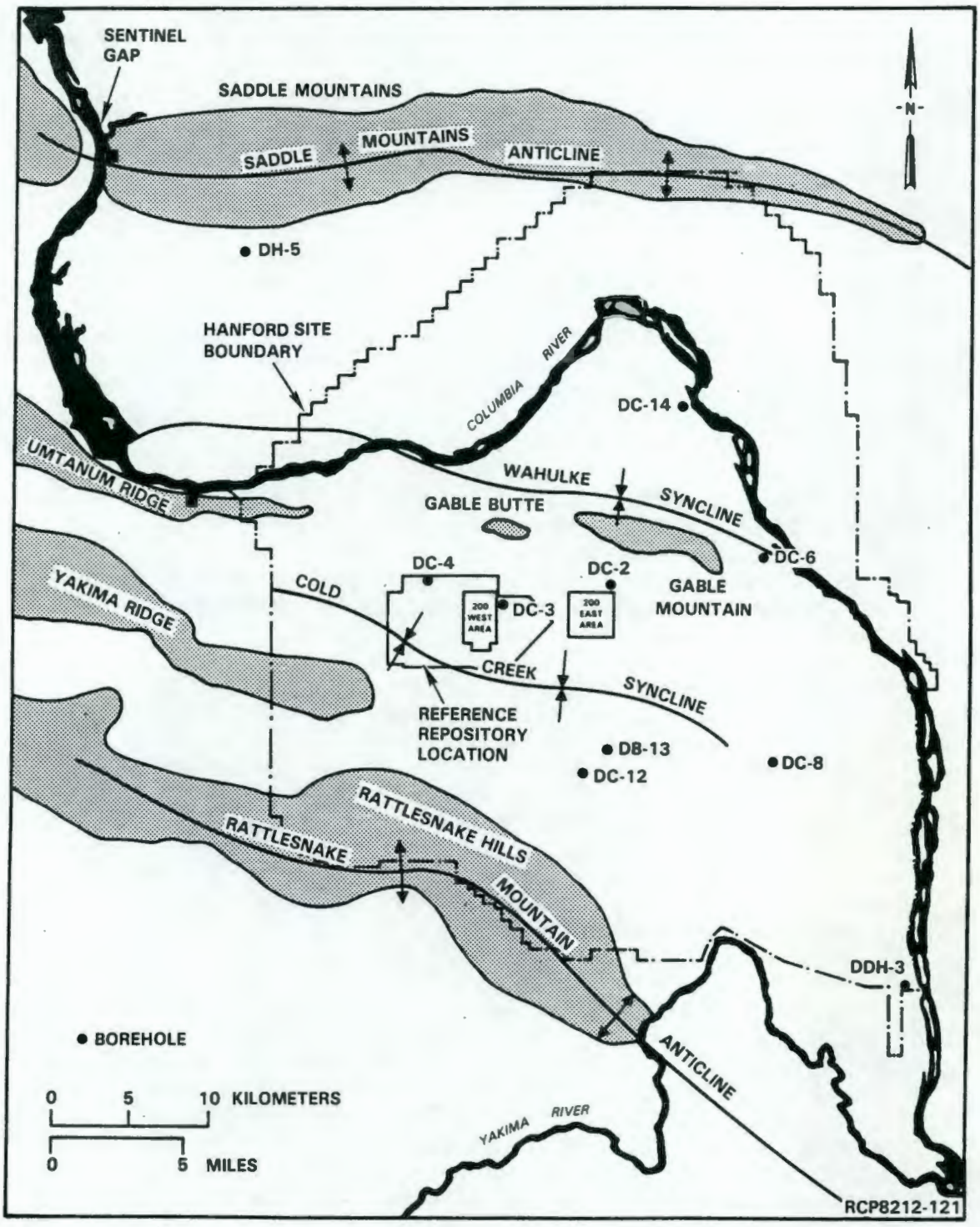

FIGURE 5-4. Borehole Location Map.

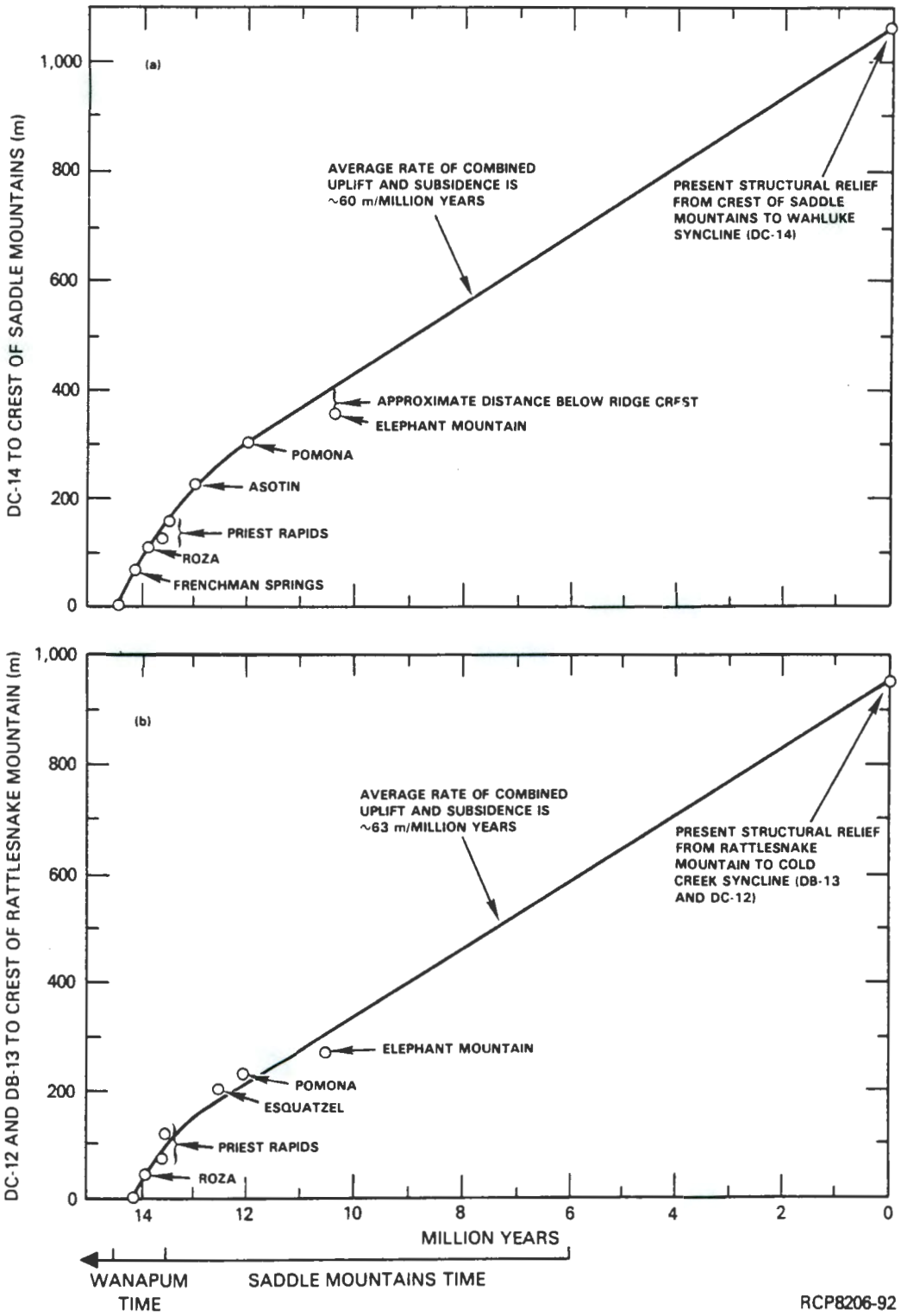

FIGURE 5-5. Rate of Combined Uplift and Subsidence for (a) the Saddle Mountains and (b) Rattlesnake Mountain.

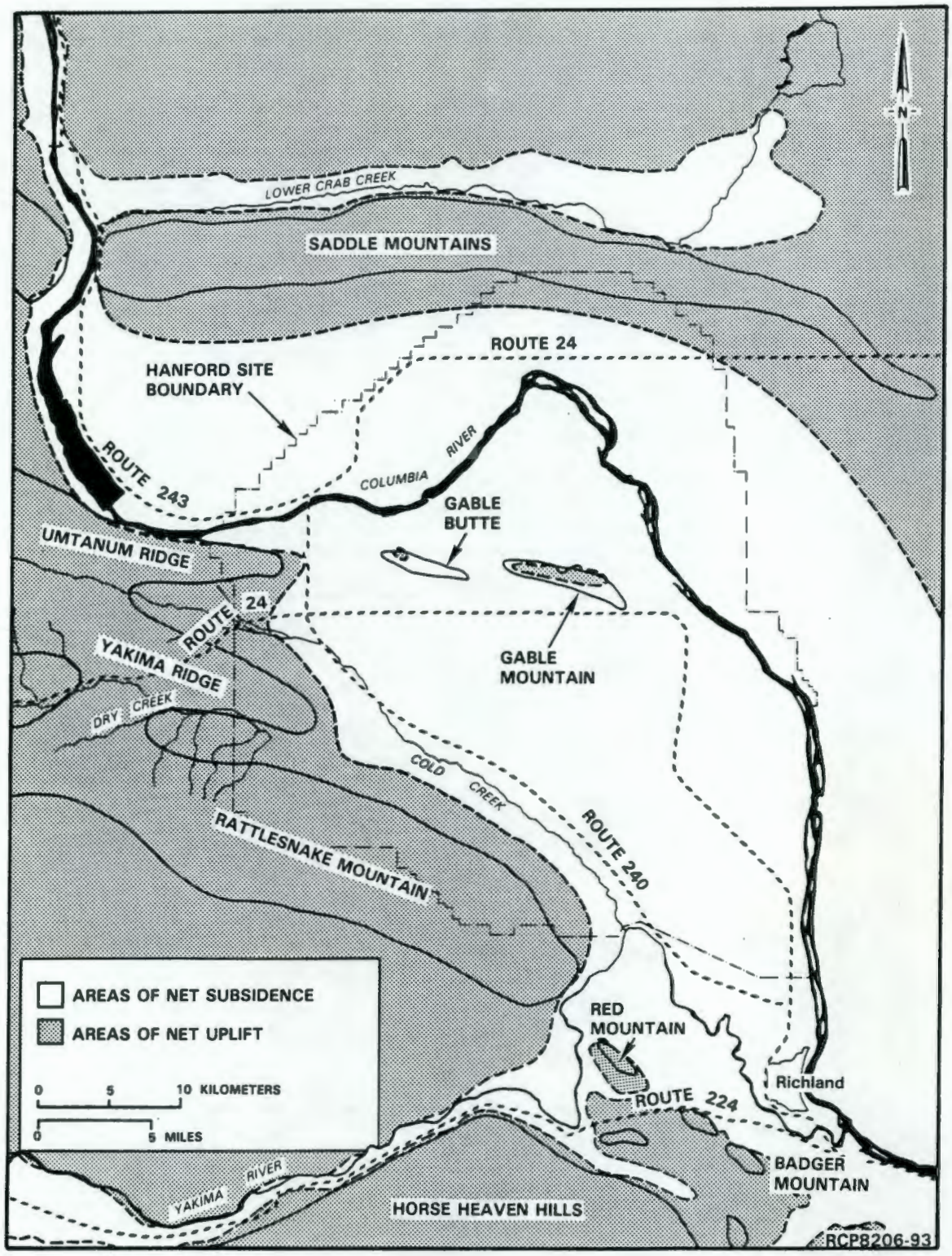

FIGURE 5-6. Areas of Net Uplift and Net Subsidence Derived from Extrapolation of the Regional Westward Tilting Paleoslope into the Pasco Basin.

One such area where thicknesses are known for flows near the extrapolated paleoslope is in the Saddle Mountains near Sentinel Gap. Borehole DH-5 (Fig. 5-4) is near the extrapolated paleoslope. Rates of uplift and subsidence for the Saddle Mountains are shown in Figure 5-7. These are calculated for flow thicknesses along the crest of the Saddle Mountains relative to DH-5 for uplift; flow thicknesses in borehole DC-14, near the trough line of the Wahluke syncline are used to calculate subsidence. Over the period of time from 14.5 to 10.5 m.y.B.P., the rates decrease to ~40 m/m.y. In both cases, it is necessary to have <40 m/m.y. of uplift and 40 m/m.y. of subsidence since the Miocene to account for the present structural relief.

In other areas where data have been collected, uplift and subsidence rates yield similar results. The area that shows the greatest uplift rate is at Rattlesnake Mountain, the structurally highest point in the Pasco Basin area. Growth rates calculated for this ridge indicate that the uplift rate decreases to ~34 m/m.y. from 13 m.y.B.P. to present based on the second method of calculation. These calculations also indicate that subsidence was occurring at <29 m/m.y. for this same time period in the Cold Creek syncline.

Little information is currently available, however, to estimate the rate of uplift or subsidence during the Grande Ronde (16.5 to 14.5 m.y.B.P.). In the eastern Columbia Plateau where the Grande Ronde Basalt is well exposed, Reidel (in press) estimates that the rate of subsidence in the Lewiston graben, based on rate of movement of the Limekiln fault, was <40 m/m.y. In the Pasco Basin, however, where information comes primarily from boreholes, the base of the sequence has not been penetrated so the only information available on the thickness is the indirect geophysical data presented in Chapter 4. The basalt layer is primarily Grande Ronde Basalt (Reidel and others, 1982) and the isopach (see Fig. 4-3) of this geophysical unit indicates considerable relief at base (see Fig. 4-5), at least an order of magnitude greater than surface relief. Although it is difficult to estimate how much relief is pre-basalt, much less the rate of growth, some estimates can be made for uplift if thinning across ridges is assumed to be due to structural growth. For example, the volcanic pile roughly thins 600 to 1,200 m across Umtanum Ridge (see Fig. 4-3). If growth occurred at a regular rate over the 2-m.y. eruptive period, then the rate of growth on Umtanum Ridge during Grande Ronde time is between 300 and 600 m/m.y. This growth rate is consistent with growth rates in Wanapum and late Grande Ronde times as suggested by the rate of growth curves (see Fig. 5-5 and 5-7) indicating a more rapid rate of growth in Grande Ronde time than in Wanapum and Saddle Mountains time. The uncertainty in the method, the nature of the base of the basalt, and anomalous areas such as Rattlesnake Mountain where the basalt thickens rather than thins, suggest that rates of uplift and subsidence for the Grande Ronde time period are more complex than indicated by this simple treatment.

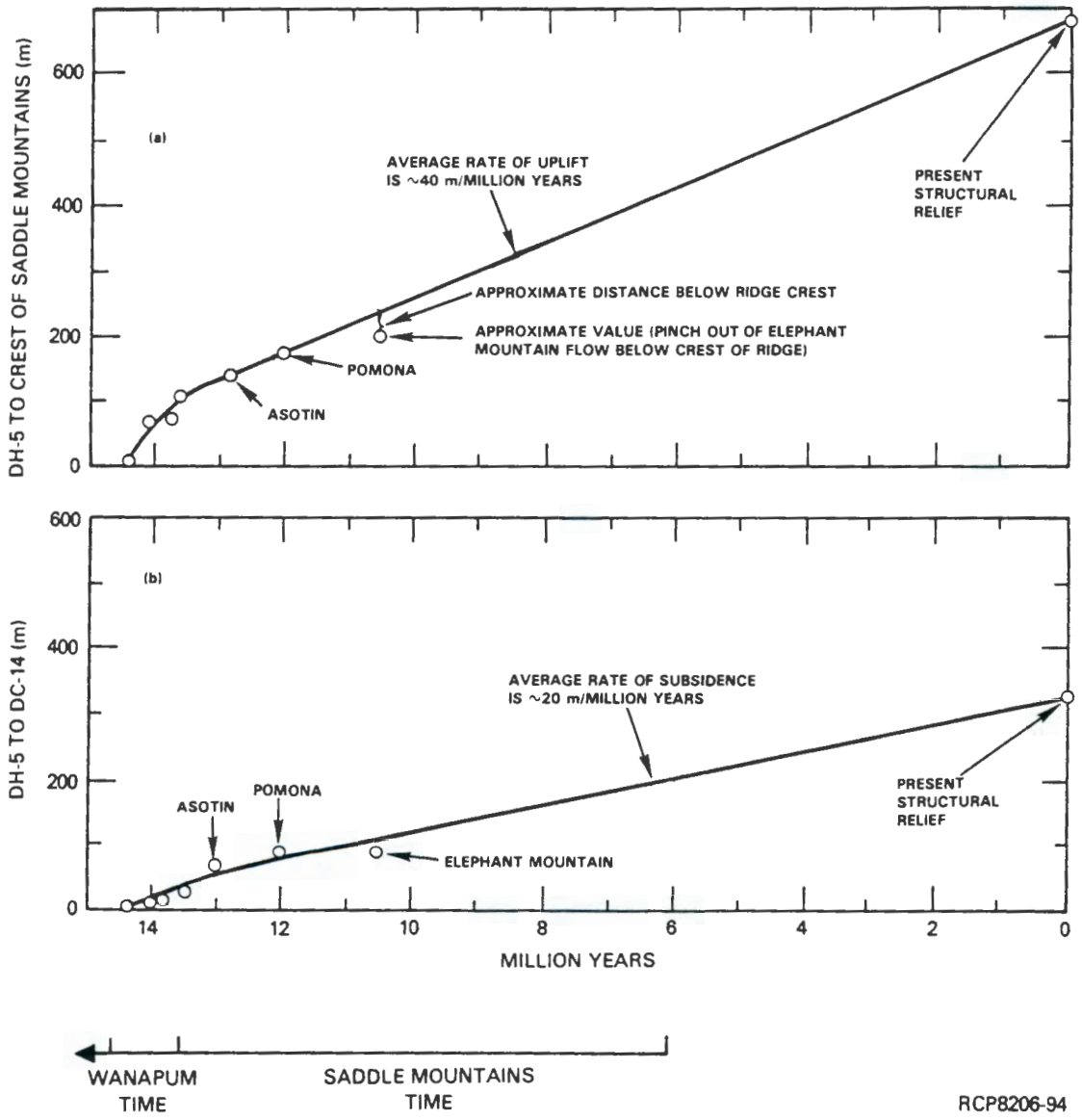

FIGURE 5-7. Rate of (a) Uplift and (b) Subsidence for the Saddle Mountains Anticline and Wahluke Syncline.

CRUSTAL SHORTENING AND CONSTRAINTS ON THE RATE OF SHORTENING

Crustal shortening is more difficult to calculate for the Yakima folds because, for one reason, faults along which some of the shortening occur are generally poorly exposed. Laubscher (1981) estimated that shortening is generally <500 m for an individual fold while Bentley (1980) estimated that shortening ranged from 0.5 to 2.5 km.

In the Pasco Basin, the best exposed structure is Umtanum Ridge, which was studied in detail by Price (1982). He found that shortening was accomplished by: (1) reverse displacement along the Umtanum thrust, (2) reverse movement along the Buck thrust, and (3) folding of the basalt flows. He calculated a total of 965 m of shortening over the entire structure or 25% horizontal strata shortening.

The rate of shortening for structures such as Umtanum Ridge is even more difficult to estimate due to the lack of marker beds, which record shortening through time. However, because shortening is in part due to folding of the basalt flows such as at Umtanum Ridge, some constraints can be placed on the rate. For example, at Umtanum Ridge, the youngest unit present, which is also deformed in the fault zone (Price, 1982), is the Priest Rapids flow (~13.5 m.y.B.P.). Yet farther east along the same structure at Gable Mountain, gravels probably <13,000 yr old are faulted (PSPL, 1982). If it is assumed that the 965 m of shortening took place over the length of time from Priest Rapids time to present (~13.6 m.y.) at an approximately constant rate, then the rate of shortening would be ~72 m/m.y. The lack of data with which to bracket the movement beyond these limits, however, makes this rate speculative. This figure also represents a local strain rate. Data from other structures are necessary before this can be argued to be representative of all folds.

EXTRAPOLATION OF UPLIFT AND SUBSIDENCE INTO THE PLIOCENE AND PLEISTOCENE

Continued subsidence and uplift from the eruption of basalts until the present is based on four lines of evidence: (1) thinning onto ridges and continued deposition and accumulation of sediments in those areas that are structurally below the extrapolated position of the regional paleoslope, (2) cumulative deformation in the sediments, (3) the fact that rates of uplift and subsidence extrapolated from the Miocene to the present can account for the present structural relief, and (4) the results of trilateration measurements (Savage and others, 1981; Chapter 6). After extrusion and emplacement of the basalts, the basin continued to aggrade thick sequences of Neogene and Quaternary sediments (Tallman and others, 1981). As the sediment aggraded, deformation also continued. The older sediments (lower Ringold) exhibit greater dip on the flanks of anticlines than the younger sediments, a fact that we interpret to mean that deformation was continuous throughout the deposition of the Ringold sediments and thus older sediments were deformed more. The rate of subsidence in lower Ringold time is approximately the same rate as in late Saddle Mountains time based on thickness variations of these units between core hole DDH-3

in the Pasco Basin and the north flank of the Rattlesnake Hills along the south basin margin. The trilateration studies (Savage and others, 1981) support a slow continued deformation at a rate <0.8 mm/yr (Chapter 6); although this is an order of magnitude greater and a different type of measurement, it is essentially in agreement with the concept that a low rate of uplift and subsidence has continued since the Miocene.

POTENTIAL VALIDITY OF EXTRAPOLATION OF UPLIFT AND SUBSIDENCE RATES

One test of the validity of extrapolating the rates of uplift and subsidence to the present, is the ability of the rates, as calculated for specific points, to account for the present top of basalt. That is, if a theoretical, present day, top-of-basalt surface is calculated from rates in the Miocene and matches the present observed top of basalt (Fig. 5-8), then the present structures can be accounted for by continued uplift and subsidence from the Miocene at these rates. However, if it does not, then some other explanation must be sought to explain any anomalies.

A preliminary attempt to calculate a theoretical top of basalt from uplift and subsidence rates extrapolated to the present is shown in Figure 5-9. The method used to accomplish this is given in Appendix B and summarized below. The regional paleoslope was projected into the Pasco Basin trigonometrically with the aid of a computer projection. The intersections of this paleoslope with boreholes in the Pasco Basin were derived, and using rates of deformation previously calculated per each borehole, values were obtained for the amount of uplift or subsidence for each. Applying these individual rates over a time interval of 14.5 m.y., numbers were produced for a theoretical top-of-basalt elevation for each borehole. Contouring these values produced the theoretical top-of-basalt map (Fig. 5-9). Since the boreholes used in this process were not located on ridges, the theoretical map does not produce an accurate representation of the structures. A comparison to the observed top of basalt is shown as a plot of residual differences in Figure 5-10.

Several interpretations can be made from the plot of residual differences between the theoretical top-of-basalt map based on rates of deformation and the observed top-of-basalt map based on borehole data and geologic investigations. Obviously apparent are anomalies that appear on the residual plot because of their presence on either the theoretical top of basalt or the observed top of basalt, but not on both. These anomalies then appear on the residual plot as either a prominent negatively or positively contoured feature. Gable Butte, Gable Mountain, and Yakima Ridge are among several prominent postively contoured features (Fig. 5-10) that are present because the theoretical top-of-basalt map, which was constructed, lacked rate of uplift data in these areas and thus a representation of them was not produced. Due to better data density (borehole locations) in synclinal areas, the residual map is most valuable for interpretation of features within the Cold Creek syncline.

THE LIGHT NUMBERS SHOWN ON THE TOP AND LEFT-HAND SIDE OF THE FIGURE ARE LABELS FOR EVERY 5,000 m OF THE UNIVERSAL TRANSVERSE MERCATOR PROJECTION GRID IN THIS MAP AREA.

CONTOURS IN FEET FROM MEAN SEA LEVEL
1 ft = 0.3048 m

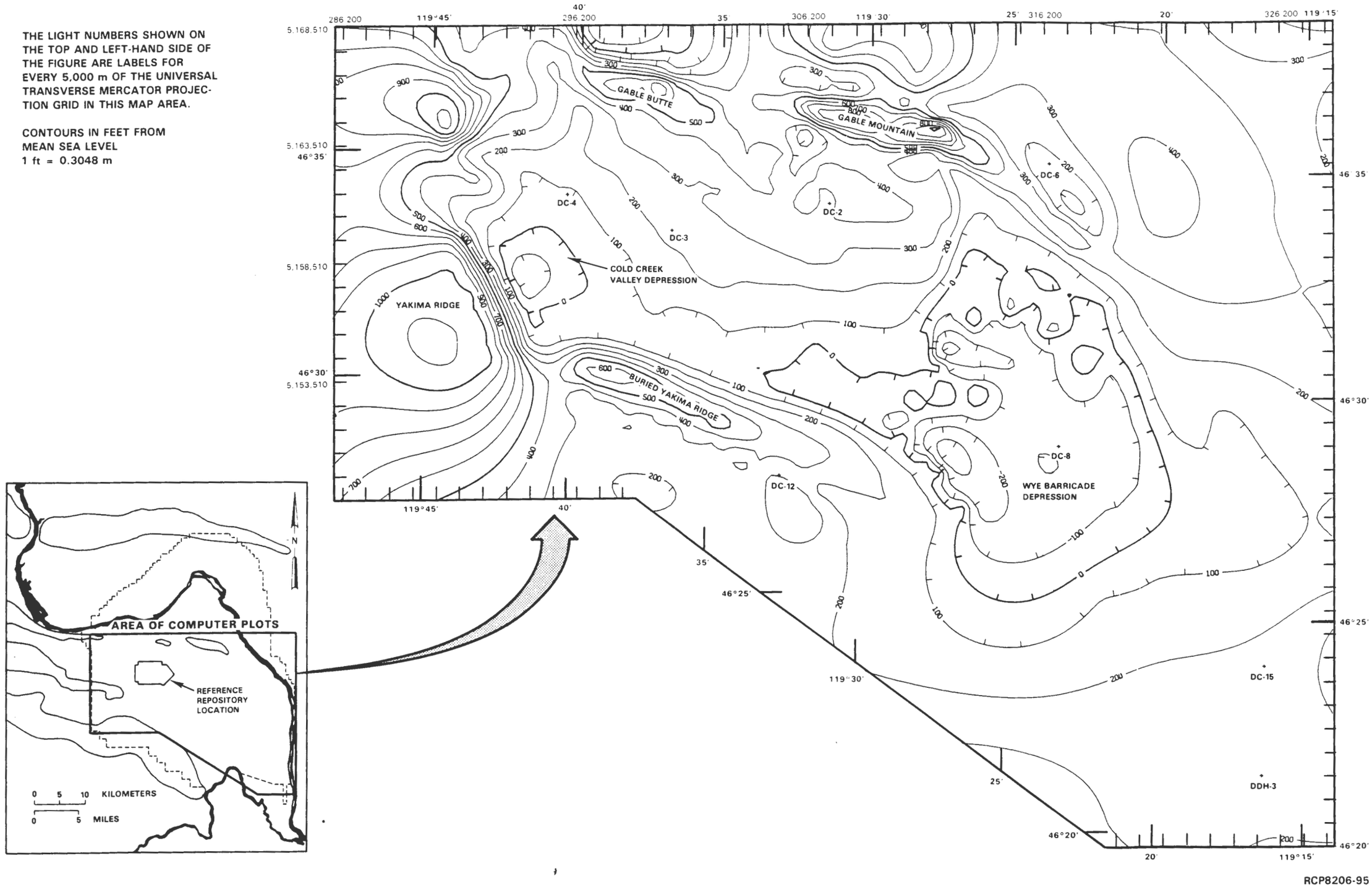

FIGURE 5-8. Observed Top-of-Basalt Map for the Cold Creek Syncline (after Myers, 1981).

THE LIGHT NUMBERS SHOWN ON THE TOP AND LEFT-HAND SIDE OF THE FIGURE ARE LABELS FOR EVERY 5,000 m OF THE UNIVERSAL TRANSVERSE MERCATOR PROJECTION GRID IN THIS MAP AREA.

CONTOURS IN FEET FROM MEAN SEA LEVEL
1 ft = 0.3048 m

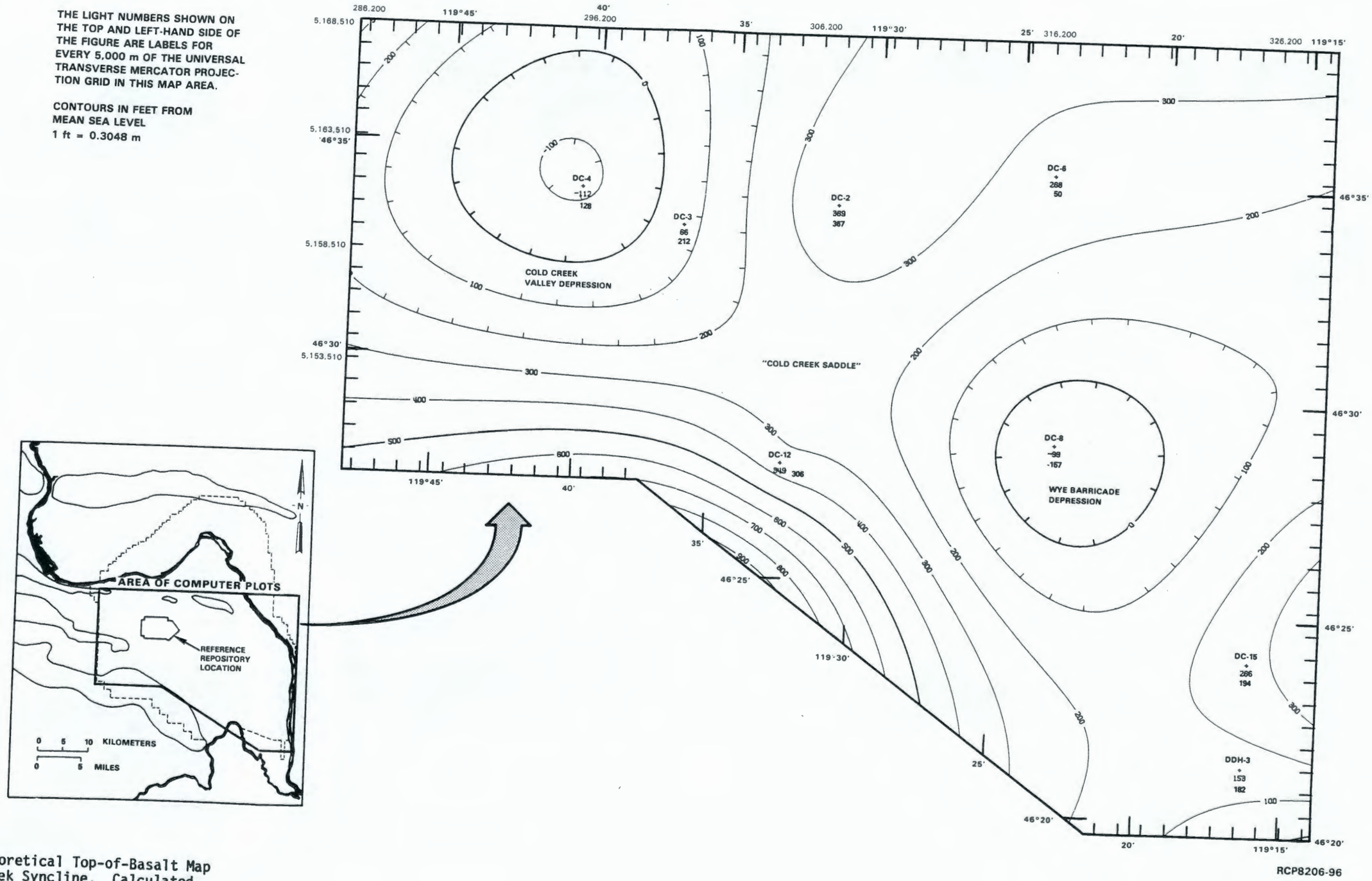

FIGURE 5-9. Theoretical Top-of-Basalt Map for the Cold Creek Syncline. Calculated from rates of uplift and subsidence.

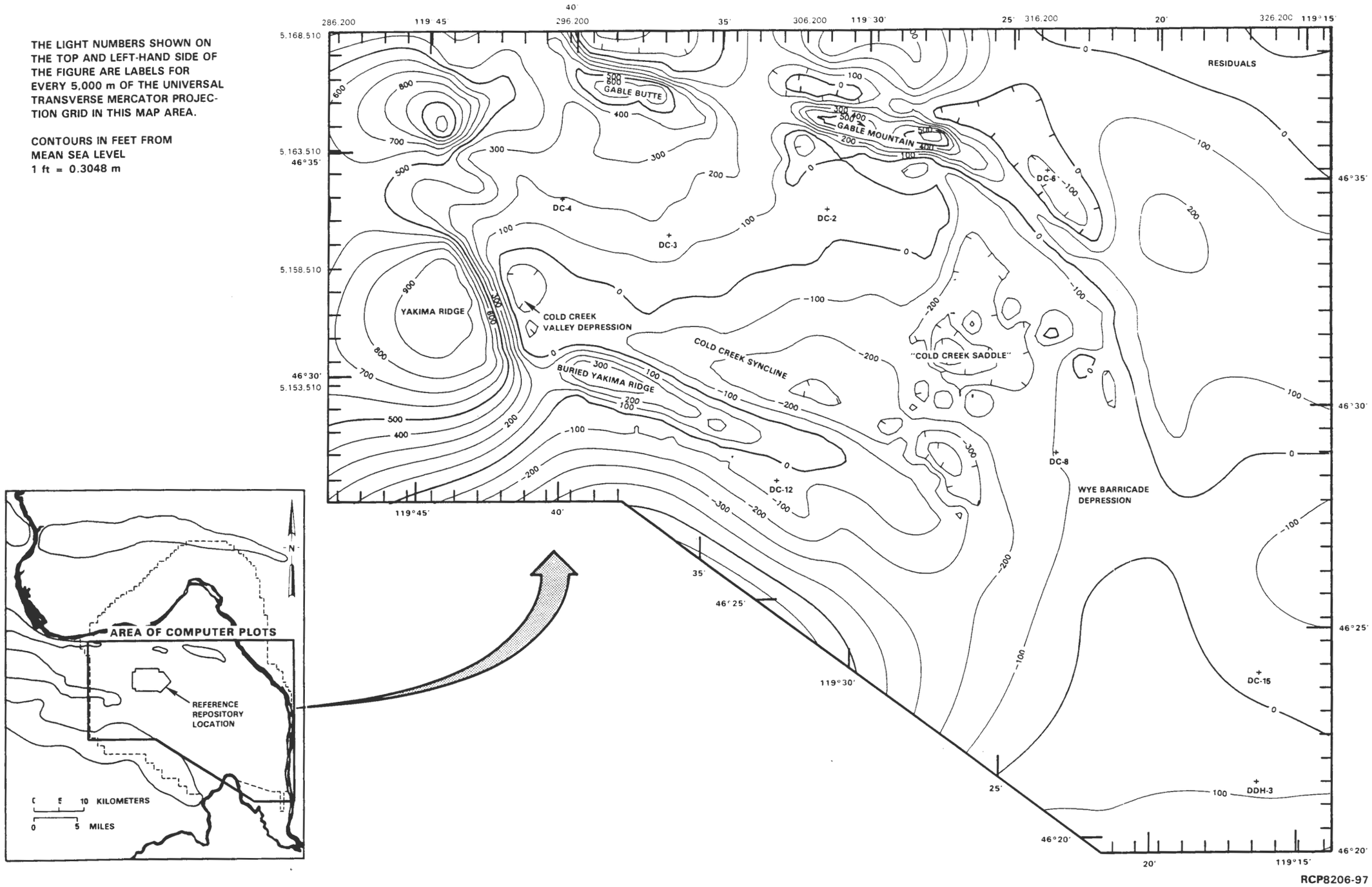

FIGURE 5-10. Residual Top-of-Basalt Map for the Cold Creek Syncline. Calculated by subtracting the theoretical top of basalt (see Fig. 5-9) from the observed top of basalt (see Fig. 5-8).

The zero contour on the residual map indicates areas of general agreement between the theoretical and observed top-of-basalt maps. In general, the major structural elements of the Cold Creek syncline can be explained by extrapolating the uplift and subsidence rates as calculated from nearby boreholes and measured sections to the present time. Of particular interest is the topographic and possible structural high between the Wye barricade depression and the Cold Creek Valley depression, which also appears on the theoretical top of basalt. This model suggests that this anomaly, in the observed top of basalt, originated from deformation that has been continuous since the Miocene and is not a recent superimposed structure resulting after emplacement of the basalt.

Several minor anomalies between the observed and theoretical top-of-basalt maps occur in the Cold Creek syncline. One area extending from Gable Butte south to borehole DC-3 and from boreholes DC-4 to -2 (see Fig. 5-4) on the theoretical top-of-basalt map (see Fig. 5-9) is structurally too low based solely on the subsidence rates that were used. A second area enclosed by a polygon encompassing boreholes DC-2, -6, -8, and -12 as corners (see Fig. 5-4 and 5-9) is also of interest. This anomaly suggests that subsidence, as implied from the observed top-of basalt map, was greater in this area than the rates suggest. In both examples, however, the magnitude of the anomaly falls well within the margin of error of the technique (see Appendix B).

SUMMARY

Studies of the lateral extent and thickness variations of basalt flows in the Pasco Basin provide information as to the age and rate of uplift and subsidence of the folds and basin. These data can then be used to model the growth of the Pasco Basin and provide constraints on tectonic models proposed to explain the central Columbia Plateau. The primary constraints include:

- The central Columbia Plateau surrounding and including the Pasco Basin consists of three structural elements: (1) east-west- and (2) northwest-trending anticlinal ridges and associated faults and synclinal valleys, and (3) a relatively undeformed westward-tilting regional paleoslope. The paleoslope has been a relatively stable feature since 12 m.y.B.P. and can be used as a regional reference plane to differentiate uplift and subsidence.

- The Pasco Basin, an area within the east-west-trending folds and bordered on the east by the paleoslope, was subsiding in Grande Ronde time. By late Grande Ronde time, subsidence declined and continued to decline to a rate of <40 m/m.y. by the late Miocene.

- The growth of Yakima folds was also under way in Grande Ronde time. Uplift rate declined from early Miocene into the middle and late Miocene and was <40 m/m.y. by the end of Columbia River basalt volcanism.

- Growth was concentrated on first-order structures that probably formed in the early to middle Miocene and these structures continued to grow as deformation continued.

CHAPTER 6 - CONTEMPORARY DEFORMATION IN THE PASCO BASIN AREA OF THE CENTRAL COLUMBIA PLATEAU

A. C. Rohay and J. D. Davis

INTRODUCTION

The low average rates of deformation during the period of time 14.5 to 10.5 m.y.B.P. when projected to the present account for the present structural relief of the Saddle Mountains and Rattlesnake Mountain anticlines (see Chapter 5). This extrapolation suggests that deformation is presently occurring. Contemporary deformation is indicated by the pattern of seismicity and the change in length of precisely measured geodetic survey lines. Changes in the crust resulting from ongoing tectonic processes as indicated by seismology and geodesy are reviewed and compared in this chapter.

The capability to instrumentally record and locate earthquakes since 1969 has permitted an assessment of Columbia Plateau seismicity that is more quantitative than that provided by the previous record of felt earthquakes. The largest historical and instrumentally located earthquakes of the Columbia Plateau are noted in Table 6-1 and Figure 6-1. The U.S. Geological Survey seismic network was originally concentrated in the Hanford Site area from 1969 to 1975 (Pitt, 1971, 1972). The network was expanded to cover the entire eastern Washington region in 1975 when network operation was assumed by the University of Washington, Seattle, Washington (Malone, 1976; UWGP, 1977, 1978, 1979, 1980, 1981). The network has had a station spacing of ~25 km in the Hanford area since its initial installation. This spacing has provided the capability to locate all earthquakes larger than magnitude 1.5 to 1.8 in the Hanford area. The coverage has gradually increased and currently permits all earthquakes of magnitude 2.0 or larger to be located throughout eastern Washington. Earthquake epicentral location accuracy is estimated to be 1 to 2 km, but determination of hypocentral depths may be in error by 2 to 4 km. The development of the network and its location and detection thresholds have previously been described (WCC, 1981a; Myers/Price and others, 1979). Magnitudes are determined using a coda-length-magnitude relationship developed for this region (UWGP, 1979) that has been approximately calibrated with Wood-Anderson magnitudes. Prior to 1972, Richter magnitudes were determined from trace amplitudes (Pitt, 1971).

Earthquakes above magnitude 3.0 located since 1969 in eastern Washington are shown in Figure 6-2. The seismic activity is concentrated around the northern and western Columbia Plateau and the southern borders of the State of Washington. Most of these earthquakes have occurred where significant historical earthquakes occurred prior to the installation of the instrumental network. A concentration of seismic activity above magnitude 3.0 has occurred in the central Columbia Plateau region (associated with earthquake swarms), but activity above magnitude 3.5 is concentrated around the northern and western Columbia Plateau and the southern border of the State of Washington.

TABLE 6-1. Felt[a] and Recorded[b] Earthquakes Within the Columbia Plateau and Surrounding Area Through 1980. (Sheet 1 of 3)

Date	Year	Time	Intensity/ magnitude	Coordinates		Location/ remarks
05 Mar.	1892	l.t.	VI	46.6° N.	120.5° W.	North Yakima, Washington
05 Mar.	1893	l.t.	VI	45.9° N.	119.3° W.	Umatilla, Oregon
15 Dec.	1897	l.t.	V	47.8° N.	120.0° W.	Lakeside, Washington
18 Oct.	1905	23:l.t.	V	47.8° N.	120.0° W.	Chelan, Washington
18 Feb.	1907	12:20l.t.	V	47.84° N.	120.02° W.	Chelan, Washington
12 Jun.	1908	Unknown	V	45.0° N.	117.25° W.	Cornucopia, Oregon
24 May	1909	22:l.t.	V	47.73° N.	120.36° W.	Chelan-Leavenworth, Washington
05 Jul.	1911	08:00	V	47.00° N.	120.54° W.	Ellensburg, Washington
14 Oct.	1913	23:00	V	45.7° N.	117.1° W.	Seven Devils, Idaho
28 Feb.	1918	23:15	V	46.5° N.	120.5° W.	Yakima, Washington
12 Mar.	1918	03:26	V	47.7° N.	117.0° W.	Spokane, Washington
01 Nov.	1918	17:20	VI	46.8° N.	119.5° W.	Corfu, Washington
07 Oct.	1920	02:l.t.	V	47.6° N.	120.1° W.	Waterville, Washington
28 Nov.	1920	Unknown	IV-V	45.7° N.	121.5° W.	Hood River, Oregon
14 Sep.	1921	11:00	VI	46.1° N.	118.25° W.	Dixie-Walla Walla, Washington
06 Jan.	1924	23:10	V	45.8° N.	118.3° W.	Milton and Weston, Oregon
17 Oct.	1926	02:45	V	45.73° N.	121.48° W.	White Salmon, Washington
30 Dec.	1926	17:57	VI	47.7° N.	120.2° W.	Chelan, Washington
03 Jan.	1927	04:58	VI	47.59° N.	120.66° W.	Leavenworth, Washington
09 Apr.	1927	05:00	V	44.8° N.	117.2° W.	Richland, Oregon
03 Sep.	1930	13:00	V	47.3° N.	117.8° W.	Lamont, Washington
18 Sep.	1934	24:l.t.	V	47.0° N.	120.54° W.	Ellensburg, Washington
26 Sep.	1934	16:15l.t.	V	47.0° N.	120.54° W.	Ellensburg, Washington
26 Sep.	1934	16:45	V	47.0° N.	120.54° W.	Ellensburg, Washington

TABLE 6-1. Felt[a] and Recorded[b] Earthquakes Within the Columbia Plateau and Surrounding Area Through 1980. (Sheet 2 of 3)

Date	Year	Time	Intensity/ magnitude	Coordinates		Location/ remarks
26 Sep.	1934	21:15	V	47.0° N.	120.54° W.	Ellensburg, Washington
19 Oct.	1934	23:31l.t.	V	47.0° N.	120.54° W.	Ellensburg, Washington
01 Nov.	1934	07:28	V	47.0° N.	120.54° W.	Ellensburg, Washington
02 Nov.	1934	15:17l.t.	V	47.0° N.	120.54° W.	Ellensburg, Washington
09 Jul.	1935	22:45	V	47.7° N.	120.0° W.	Chelan Falls, Washington
12 Oct.	1935	01:03	V	47.66° N.	120.22° W.	Entiat, Washington
16 Jul.	1936	07:07:49:0	VII $5.75M_s$	46.21° N.	118.23° W.	Milton-Freewater, Oregon (WCC Relocated)
18 Jul.	1936	16:30	V	45.93° N.	118.38° W.	Milton-Freewater, Oregon
04 Aug.	1936	09:19	V	44.8° N.	118.6° W.	Helix, Oregon
28 Aug.	1936	04:39	V	45.93° N.	118.38° W.	Milton-Freewater, Oregon
29 Nov.	1939	09:15	V	47.7° N.	120.0° W.	Chelan Falls, Washington
07 Apr.	1941	09:25	VI $4.5M_L$	48.3° N.	119.6° W.	Mazama, Washington (Okanogan)
23 Feb.	1942	14:03	V	47.6° N.	120.2° W.	Wenatchee-Chelan Falls, Washington
12 Jun.	1942	09:30	V	44.9° N.	117.1° W.	Halfway and Pine, Oregon
24 Apr.	1943	00:10:46.0	VI	47.3° N.	120.6° W.	Leavenworth, Washington
31 Oct.	1944	11.34:28.7	V	47.8° N.	120.6° W.	Entiat, Washington
04 Jan.	1945	02:34:48.7	V	47.7° N.	120.2° W.	Entiat, Washingston
13 Jan.	1948	06:55:00.0	V	47.9° N.	120.3° W.	Lucerne, Waterville, Washington
28 Aug.	1948	22:25	VI-V	47.8° N.	117.25° W.	Deer Park, Denison, Washington
15 Mar.	1949	20:53:11.0	$4.8M_L$	45.5° N.	117.0° W.	Joseph, Oregon
04 Jan.	1951	13:45.00.0	V	47.7° N.	120.0° W.	Chelan-Waterville, Washington
07 Jan.	1951	22:45:00.0	V	45.9° N.	119.2° W.	McNary, Oregon
04 Mar.	1952	19:42:00.0	V	47.67° N.	117.40° W.	Spokane, Washington
23 May	1954	13:41:42.0	V	48.34° N.	120.14° W.	Twisp, Winthrop, Washington

TABLE 6-1. Felt[a] and Recorded[b] Earthquakes Within the Columbia Plateau and Surrounding Area Through 1980. (Sheet 3 of 3)

Date	Year	Time	Intensity/ magnitude	Coordinates		Location/ remarks
24 Feb.	1956	22:00:00.0	V	47.9° N.	119.1° W.	Electric City, Washington
01 Nov.	1957	10:12:02.0	$4.2M_L$	46.7° N.	121.5° W.	Mount Rainier, Washington
12 Apr.	1958	22:37:11.0	VI $4.1M_L$	48.0° N.	120.0° W.	Chelan, Washington
20 Jan.	1959		V	46.2° N.	118.2° W.	Milton-Freewater, Oregon
06 Aug.	1959	03:44:32:0	VI $4.4M_L$	47.8° N.	119.9° N.	Chelan, Washington
15 Jan.	1962	05:29.00.0	$4.3M_L$	47.8° N.	120.2° W.	Chelan, Washington
22 Dec.	1963	02:54:	V $4.4M_B$	48.0° N.	119.3° W.	Discrepancy in location
07 Nov.	1965	16:41:47.4	$4.3M_B$	44.9° N.	117.0° W.	5-km depth
23 Jul.	1966	01:47:08.8	$4.3M_B$	47.2° N.	119.5° W.	Ephrata, Washington
30 Dec.	1966	03:51:40.3	$4.2M_B$	44.9° N.	117.0° W.	10-km depth
20 Dec.	1973	01:08:28.2	V $4.4M_C$	46.87° N.	119.35° W.	2.4-km depth Corfu, Washington (UW)
13 Apr.	1976	00:47:17.1	VI $4.8M_L$ $4.5M_B$	45.22° N.	120.77° W.	15-km depth constrained (NOAA)
17 Apr.	1976	02:11:44.4	$4.2M_L$	45.08° N.	120.80° W.	15-km depth constrained (NOAA)
19 Jan.	1979	14:55:15.4	V $3.6M_C$	47.92° N.	119.68° W.	Chief Joseph, Washington (UW)
08 Apr.	1979	07:29:37.8	$4.15M_C$	45.99° N.	118.45° W.	Walla Walla, Washington (UW)
18 Feb.	1980	06:00	$4.1M_C$	47.25° N.	120.22° W.	Cle Elum, Washington (UW)

l.t. = Local time.

M_B = body-wave magnitude.

M_C = coda-length magnitude.

M_L = local magnitude.

M_S = surface wave magnitude.

NOAA = National Oceanic and Atmospheric Administration.

UW = University of Washington.

WCC = Woodward-Clyde Consultants.

[a]Greater than IV intensity.

[b]Greater than 4.0 magnitude.

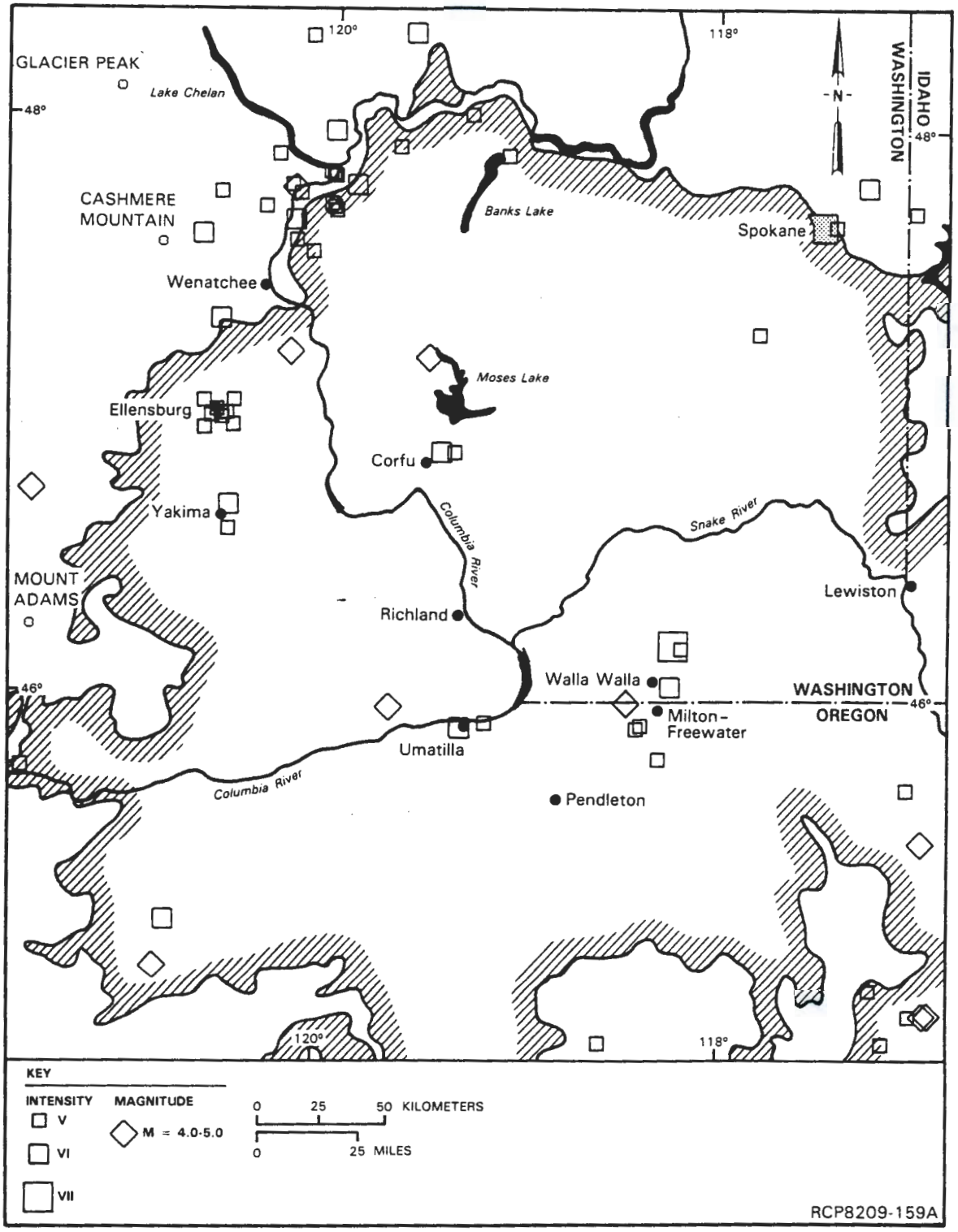

FIGURE 6-1. Historical Seismicity of the Columbia Plateau Tectonic Province. All earthquakes with modified Mercalli intensity greater than IV are shown (see Table 6-1). Instrumental magnitudes are shown from some events above magnitude 4.0.

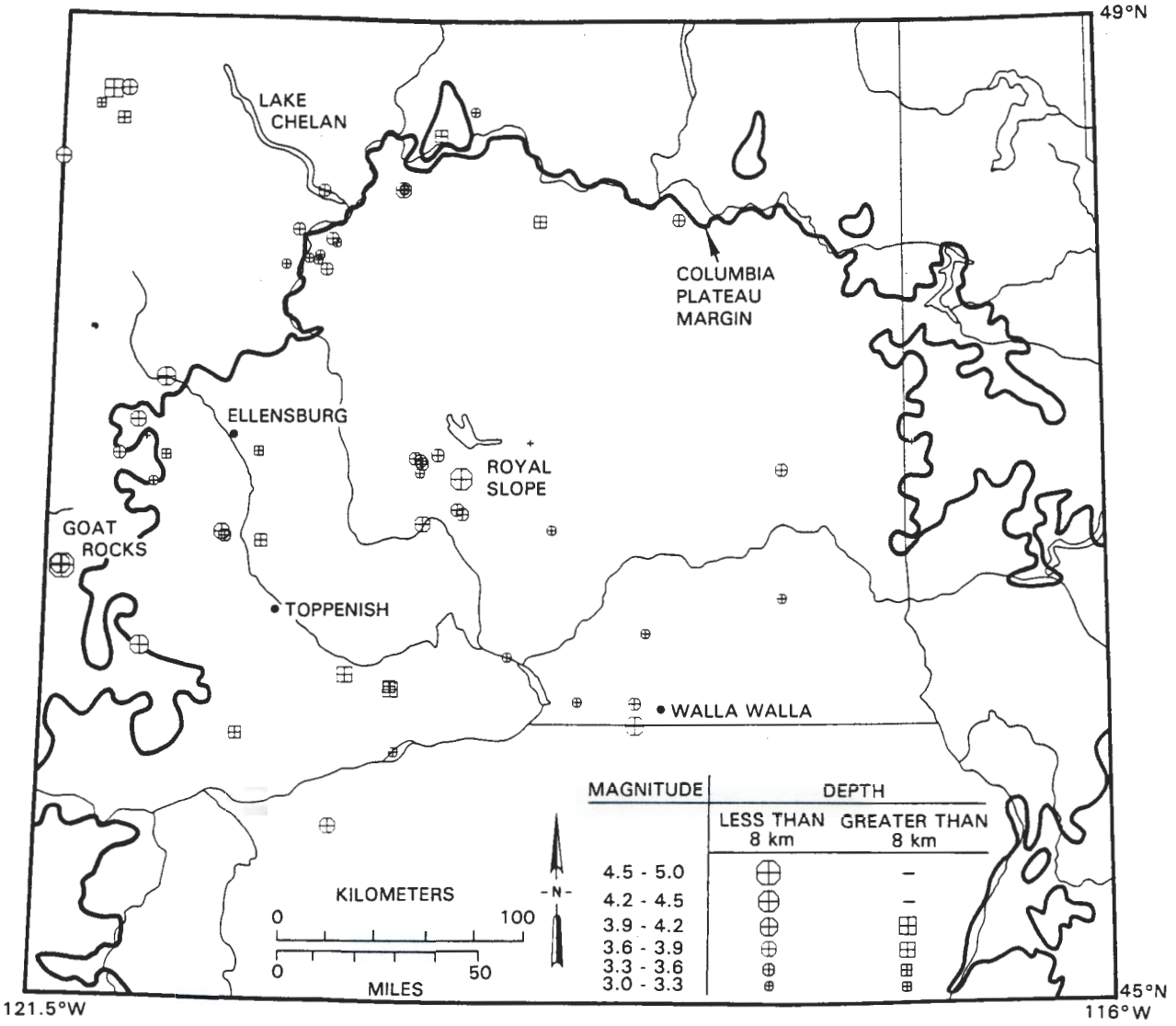

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FIGURE 6-2. Instrumental Seismicity (post 1969) above Magnitude 3.0, Columbia Plateau Tectonic Province. Activity not interpreted to be affected by the changing location threshold due to gradual network development.

EARTHQUAKE SWARMS AND SHALLOW SEISMIC ACTIVITY

Earthquake swarms are the predominant characteristic of the Columbia Plateau seismicity. These earthquake sequences typically last a few days to several months and occur in a volume of rock with typical dimensions of ~5 km. Earthquake swarms (as detected and located by the regional seismic network) may contain several to ~100 locatable earthquakes of magnitude 1.0 to 3.5, but most are smaller than magnitude 2.0. During a swarm, there is no distinctively large event that is followed by a generally decreasing level of seismicity, as is typical of a mainshock-aftershock sequence. Swarm earthquakes tend to gradually increase and decay in frequency, but not in magnitude.

Shallow earthquakes are occurring at a significantly greater rate than deep earthquakes. Approximately 75% of the earthquakes are located shallower than 4 km deep, and 80% of these shallow events occur during earthquake swarms (WCC, 1981a). A distinction between shallow and deep seismicity is also reflected in the recurrence relations or size distribution of the earthquakes. A causal relationship between the rheological behavior of the basaltic layers and the characteristic of earthquake swarm activity is also suggested.

Earthquake swarm activity is concentrated in the central portion of the Columbia Plateau, principally north and east of the Hanford Site between the Saddle Mountains and Frenchman Hills structures (Fig. 6-3). It is within this region that most events (in swarms and as individual events) larger than magnitude 3.0 occur, including the largest magnitude swarm earthquake instrumentally recorded, the December 20, 1973, magnitude 4.4 Royal Slope earthquake. There is an apparent alignment of smaller earthquakes along the Saddle Mountains structure (Fig. 6-3). However, the larger events, and consequently, the greatest implied deformations, are not aligned with mapped faults of the Saddle Mountains or Frenchman Hills anticlines.

The swarm earthquakes are concentrated north of the Saddle Mountains structure (Fig. 6-3) between Saddle Gap and the Columbia River at Sentinel Gap. This swarm activity gradually shifts to the south of the Saddle Mountains structure towards its eastern end. This shift in the location of seismic activity coincides with the approximate location of a slight bend in the axial trace of the anticline. Swarms south of the Saddle Mountains are separated by aseismic areas, so that no preferred alignment of these swarm areas is apparent.

Other notable earthquake swarms have occurred at Wooded Island and Coyote Rapids (Fig. 6-3). The swarm area near Coyote Rapids is relatively inactive compared to other swarm areas, but was the site of the October 25, 1971, magnitude 3.8 Coyote Rapids earthquake, the second largest instrumentally recorded earthquake in the central Columbia Plateau. The swarms near Wooded Island have been some of the most intense in terms of number of events, but all of these events have been less than magnitude 3.0 (Pitt, 1972; Rothe, 1978).

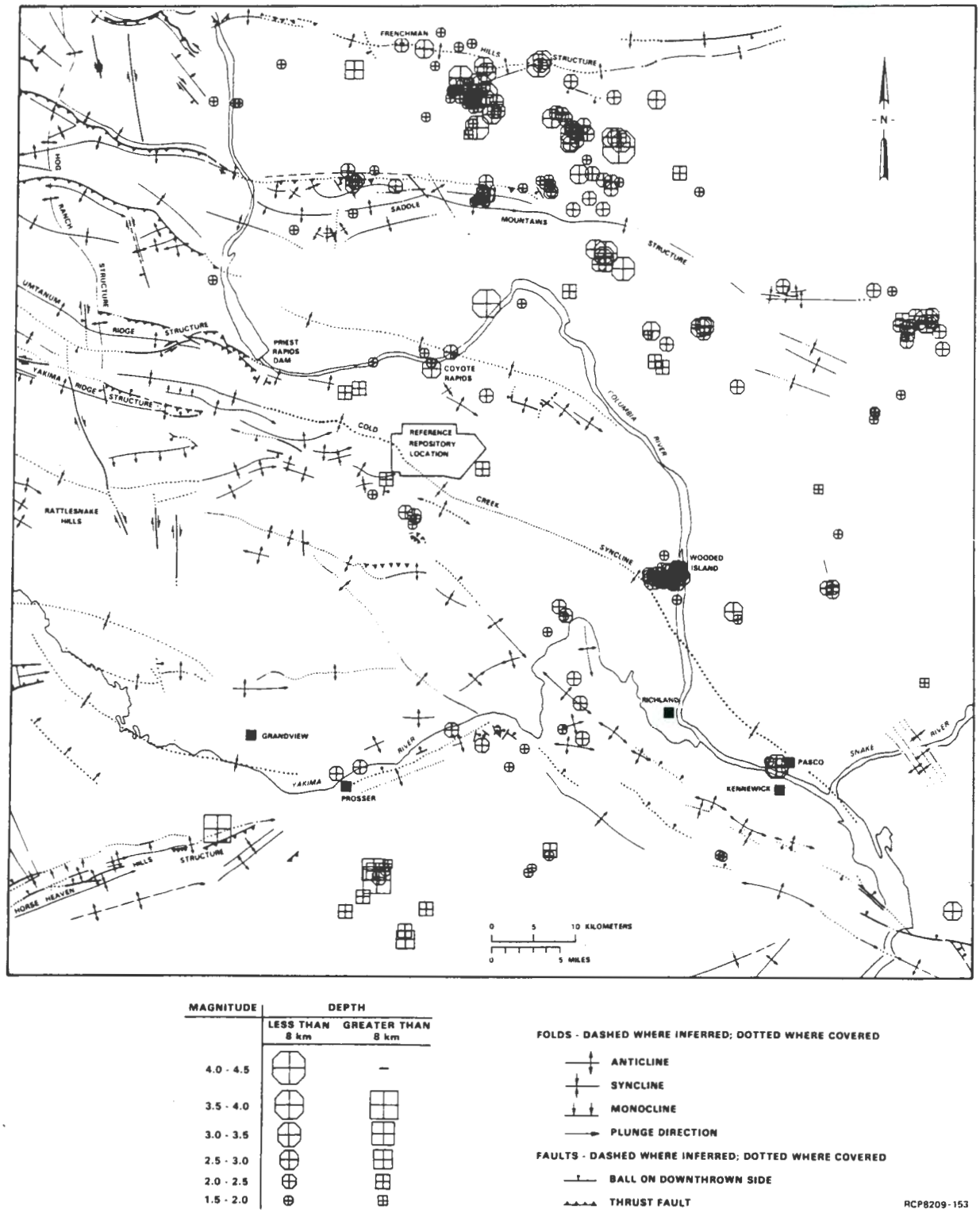

FIGURE 6-3. Instrumental Seismicity of the Central Columbia Plateau. The largest events are concentrated near the Saddle Mountains and Frenchman Hills.

Linear trends of epicenters are sometimes apparent within individual swarms. These trends are ~5 km long in several swarm areas and are oriented east-west, as in the Wooded Island swarm, or northwest-southeast as in the swarms on Royal Slope north of the Saddle Mountains (see Fig. 6-3). Earthquake swarms have often occurred more than once within the same small areas. In a few cases, successive swarms have been centered in about the same location, but the general tendency is for migration of the activity in a particular direction, producing some of the alignments mentioned above.

Several earthquake swarms have been studied using a dense, temporarily deployed networks of portable seismic-monitoring stations, which provided epicenter location accuracies of ~0.2 to 0.5 km. Based on these more accurate locations, shorter alignments can be detected. East-west trends of epicenters, ~2 km long, were found in the 1969-1970 Wooded Island swarm (Pitt, 1971) and the 1975 Wooded Island swarm (Rothe, 1978). The spatial distribution of epicenters determined by Malone and others (1975) for the Royal Slope sequence in 1973 also showed an apparent 2-km east-west alignment near the location of the December 20, 1973, magnitude 4.4 event. Malone and others (1975) found that three different alignments of hypocenters within the 1973 Eltopia swarm could be interpreted. (These correspond to three different focal mechanism classes as well; see Table 6-2.) The overall patterns of these accurately located events indicates that swarm earthquakes occur in a relatively diffuse pattern, without significant concentration of events on a single plane. This observation indicates that the events are not occurring exclusively on unmapped or buried fault segments (Malone and others, 1975). The inferred 2- to 3-km-long fault segments are not connected into throughgoing faults.

Aftershocks are often associated with eastern Washington earthquakes of a magnitude 3.5 or larger. Two shallow earthquakes in the central Columbia Plateau have been larger than magnitude 3.5: the magnitude 4.4 Royal Slope event and the magnitude 3.8 event on Wahluke Slope northeast of Coyote Rapids. The Royal Slope event was 2.4 km deep and was immediately followed by numerous shallow earthquakes that decreased in frequency, but not in magnitude, with time. The earthquake sequence lasted much longer than expected for aftershocks of an event this size. The Royal Slope sequence thus exhibited characteristics of both swarms and aftershocks. The October 25, 1971, magnitude 3.8 earthquake (<1 km deep) occurred without any significant increase in smaller swarm-type events.

DEEP EARTHQUAKES

Earthquakes occur to a depth of ~28 km in the Columbia Plateau region, but at a lower frequency than the shallow (≤6 km) swarm earthquakes (UWGP, 1979). This depth is the approximate thickness of the crust as determined from seismic refraction studies (see Chapter 2).

Two deep earthquakes have occurred in the central Columbia Plateau with magnitudes greater than 3.5. Both occurred south of the Hanford Site near the Horse Heaven Hills structure (see Fig. 6-3). A series of foreshocks and aftershocks accompanied the June 28, 1975, magnitude 3.8 event.

TABLE 6-2. Eastern Washington Focal Mechanism Solutions. (Sheet 1 of 2)

Location	Compression		Tension		Remarks	Source
	Azim.	Dip	Azim.	Dip		
Central Columbia Plateau Deep Events						
Prosser/Horse Heaven Hills	180°	01°	082°	86°	Composite	WCC (1980a)
Rattlesnake Mountain	174	24	295	51	Composite	WCC (1980a)
Yakima/Umtanum/Gable Ridges	160	16	292	66	Composite	WCC (1980a)
Eltopia north	293	43	032	09	Composite, poorly constrained	WCC (1980a)
Corfu/Royal/Othello	172	05	058	78	Composite	WCC (1980a)
NW. Hanford/Midway	218	04	097	74	Two event composite	Weston (1977)
NW. Hanford/Midway	032	04	292	54	Alternative	Weston (1977)
Horse Heaven Hills (06/28/75)	326	00	235	74	M=3.8, z=10.2 km	Weston (1977)
Regional Events						
Walla Walla (04/08/79)	269	11	166	48	M=4.1	WCC (1980a)
Milton-Freewater (1936)	056	22	161	32	M=6.1, historical	WCC (1980a)
Cle Elum mainshock (06/26/78)	334	23	223	52	M=3.7	UWGP (1978)
Cle Elum aftershocks (6 months)	344	19	119	64	Composite	WCC (1978)
Cle Elum (02/18/81)	025	34	151	42	M=4.2	UWGP (1981)
Goat Rocks mainshock (05/28/81)	178	03	270	32	M=5.0	UWGP (1981)
Goat Rocks foreshock (05/28/81)	180	00	270	00	M=4.0	UWGP (1981)
Toppenish (02/02/81)	182	24	088	09	M=4.0	UWGP (1981)
Goldendale (06/14/81)	179	16	059	60	M=3.3	UWGP (1981)
Lake Chelan	321	03	058	58	Composite, 0/9 km	UWGP (1978)
Chief Joseph Dam (01/19/79)	129	03	024	68	M=3.9, z=7 km	UWGP (1979)
Toppenish (02/02/81)	196	04	100	60	M=3.9, z=5 km	UWGP (1981)

TABLE 6-2. Eastern Washington Focal Mechanism Solutions. (Sheet 2 of 2)

Location	Compression		Tension		Remarks	Source
	Azim.	Dip	Azim.	Dip		
Central Columbia Plateau Shallow Events						
Eltopia swarm (1)	346	17	244	34	Composite	Malone et al. (1975)
Eltopia swarm (2)	333	30	189	54	Composite	Malone et al. (1975)
Eltopia swarm (3)	358	05	136	81	Composite	Malone et al. (1975)
Royal mainshock (12/20/73)	201	09	310	61	M=4.4	WPPSS (1981)
Royal sequence	001	04	198	85	Week after mainshock, composite	Malone et al. (1975)
Royal sequence	172	34	326	63	Week after mainshock, composite	Malone et al. (1975)
Royal (09/04/74)	184	23	000	67	M=2.8, swarm event	WPPSS (1981)
Smyrna (09/22/72)	072	05	334	62	M=2.5, swarm event	WPPSS (1981)
Scooteney Reservoir (02/15/73)	100	10	320	77	M=1.4, swarm event	WPPSS (1981)
Wahluke Slope (10/18/72)	182	05	340	85	M=1.7, swarm event	WPPSS (1981)
Wahluke Slope (10/19/72)	286	13	106	77	M=1.4, swarm event	WPPSS (1981)
Wahluke Slope (11/30/72)	343	04	069	57	M=2.0, swarm event	WPPSS (1981)
Wahluke Slope (12/03/72)	356	05	135	84	M=1.4, swarm event	WPPSS (1981)
Wahluke Slope (12/15/72)	278	04	037	80	M=1.0, swarm event	WPPSS (1981)
Coyote Rapids (10/25/71)	340	03	200	84	M=3.8, swarm event	Rasmussen and Rohay (1982)

M = magnitude.

z = depth.

All of these events were 8 to 11 km deep. The February 17, 1979, magnitude 3.6 event was not accompanied by either foreshocks or aftershocks. However, as stated previously, earthquakes in eastern Washington less than magnitude 3.5 are not accompanied by foreshocks or aftershocks.

Except for these two large (for the central Columbia Plateau), deep events, in the area of the Horse Heaven Hills structure, no correlation is observed between presently mapped geologic structures and deep seismicity (see Fig. 6-3). Notably, there is no concentration of activity along the Rattlesnake Hills structure that has been proposed as the surface expression of a deep structure (Chapter 7).

Generally, there is no apparent concentration of deep seismicity in the areas of intense shallow swarm activity. One exception is the several relatively low-level swarms near the eastern end of the Saddle Mountains (see Fig. 6-3) during which earthquakes as deep as 12 km occurred (WCC, 1981a). Otherwise, deep seismicity occurs in a seemingly random pattern, associated neither with known geologic structures or areas of shallow seismicity.

FOCAL MECHANISMS

Focal mechanisms of earthquakes in eastern Washington (see Table 6-2 and Fig. 6-4) indicate a response to a nearly horizontal principal compression oriented north-south. The minimum compression (or extension) is generally vertical, so that thrust or reverse faulting on east-west striking planes is indicated. These principal stress directions are in good agreement with mapped east-west orientations of fold axes and associated thrust or reverse faults (see Chapters 2 and 5). Recent in situ stress measurements using the hydrofracturing method also indicate that the maximum horizontal stress at a 1-km depth is oriented north-northeast (Kim and Haimson, 1982). Such north-south compression is found for focal mechanisms of earthquakes throughout the western United States (Crosson, 1972; Smith and Lindh, 1978).

Earthquake focal mechanisms for eastern Washington are summarized in Table 6-2 and Figure 6-4. The small magnitude of most earthquakes often precludes the determination of a focal mechanism for individual events; but, composite focal mechanism solutions have been performed on several groups of earthquakes within defined regions or within particular earthquake swarms. There appears to be a scatter in the maximum compression axes (see Fig. 6-4) of $\pm 20^{\circ}$ and a scatter in the minimum compression (or tension) axes of $\pm 30^{\circ}$. The smaller, shallow events show the greatest scatter. The larger events, which are generally better determined show much more consistency.

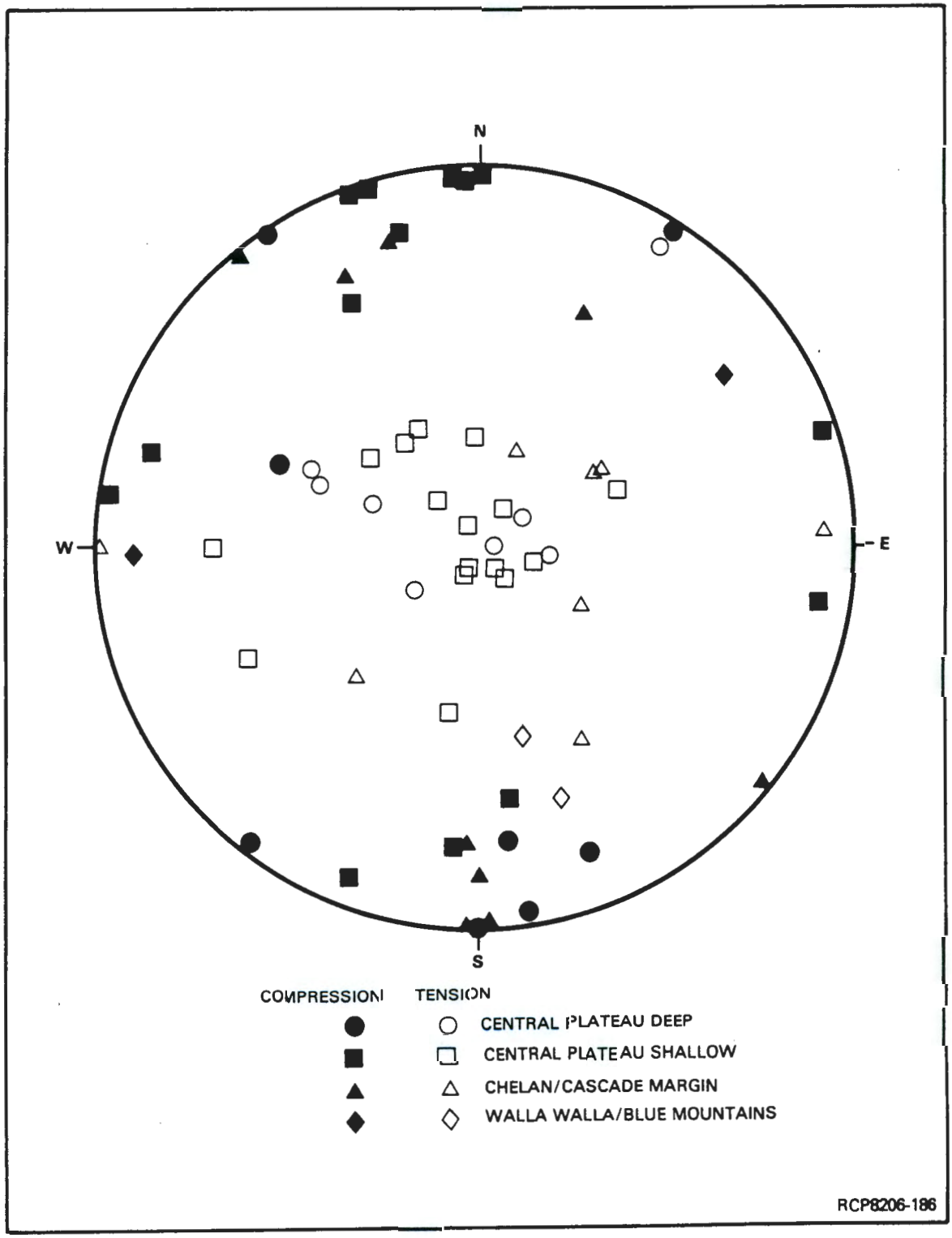

FIGURE 6-4. Focal Mechanisms of Eastern Washington. Lower hemisphere, equal area projection of compression and tension axes derived from focal mechanism solutions.

This state of stress is found for both shallow events in swarms and for deep events. The orientation of stresses does not change with depth. The vertically oriented, least principal stress is expected due to the proximity to the surface of the shallow earthquakes. The deep events, where up to 25 km of overburden pressure acts, still consistently exhibit vertically oriented tension.

Focal mechanism solutions for the earthquakes on the northern, western, and southern margins of the Columbia Plateau also indicate principally north-south compression, except for events in the Walla Walla and Milton-Freewater area (WCC, 1980b). The April 1979 and July 1936 events have maximum compression axes oriented nearly east-west and non-vertical tension axes, indicating mixed thrusting and strike slip faulting possibly on north-south oriented planes.

Earthquakes in the southeastern Cascades of Washington and western border of the Columbia Plateau have a tendency to have east-west rather than vertical tension, inferring strike-slip faulting (UWGP, 1981). This stress orientation is particularly evident in the two large Goat Rocks events (see Fig. 6-2) north of Mt. Adams and the Toppenish event, all of which occurred in 1981. In general, the tension axes of other regional earthquakes is not as concentrated in the vertical direction as is observed for the central Columbia Plateau events.

The composite focal mechanism solutions (see Table 6-2) represent only those for which consistent data were obtained. Detailed studies of several earthquake swarms have shown that no single focal mechanism will fit all events (Malone and others, 1975; Rothe, 1978). There is sufficient variation in the focal mechanisms of small swarm earthquakes to indicate that they occur on a variety of fault orientations, which suggests some inhomogeneity in the stress field. This qualitatively supports the previous observation that the epicenters of swarm earthquakes do not exclusively align along planar structures, but occur in diffuse patterns. Several individual swarms exhibit an east-west elongated pattern but focal mechanism solutions suggest rupture on several, rather than one plane. The predominance of north-south compression and vertical tension supports a conclusion that most of the seismic deformation in swarms occur as thrust and reverse faulting on a number of very small east-west oriented planes, rather than a single, larger fault plane.

The difference in orientation of the axis of least compression from vertical in the central Columbia Plateau to east-west on the margin of the southern Washington Cascades suggests that deformation in the southern Cascades is predominantly from strike-slip movement on northwest-southeast oriented fault planes (UWGP, 1981). These recently determined focal mechanisms may assist in refining the tectonic models of the Pacific Northwest (Chapter 7). In some of these tectonic models, right lateral strike-slip faulting has been presumed to occur on the Rattlesnake alignment in the central Columbia Plateau, but earthquake hypocenters and focal mechanisms do not support this assumption. The composite focal mechanisms for deep events in proximity to this structure indicate roughly equal parts of strike slip and thrust or reverse movement.

The east-west compression noted above for events in the Walla Walla and Milton-Freewater areas is anomalous compared to central Columbia Plateau earthquakes. These earthquakes may be associated with the Hite fault system (Plate 2-1), which is oriented roughly northeast-southwest, parallel to one of the inferred fault planes of the focal mechanism solutions (WCC, 1980b). Aftershocks of the 1936 event appeared to migrate in a north-south direction (WCC, 1980b). The magnitude 4.1 event in 1979 was not associated with any aftershocks (which is unusual for an event of this size), so that selecting the Hite fault orientation for the actual fault plane is tentative. Detailed, temporary monitoring of this area did not detect any small earthquakes in this region (WCC, 1980b) nor is there any significant concentration of events located using the regional network since 1969 that could assist in defining the fault plane for these events.

RECURRENCE RELATIONSHIPS

A plot of magnitude versus total number of events above that magnitude of a large sample of earthquakes reveals their size distribution. The slope of the recurrence curve (Fig. 6-5), called the b value, is normally found to be roughly equal to 1.0. This indicates that ten times more earthquakes of a given magnitude occur than earthquakes one magnitude unit larger. Earthquakes below magnitude 1.5 are below the location threshold, causing the curve to taper off at lower magnitudes. The b values <1.0 indicate a greater proportion of larger earthquakes compared to smaller earthquakes; b values >1.0, indicate a larger proportion of smaller earthquakes.

In the central Columbia Plateau, several distinctions between shallow (≤6 km) and deep events (>6 km) have been noted. The largest number of earthquakes and the largest magnitude earthquakes have been recorded in the shallow zone, with fewer, generally smaller events in the deep zone. Earthquakes above and below a 6-km depth are shown in Figure 6-5.

The b value for deep events is 0.8, compared to 1.15 for the shallow events (UWGP, 1979). Earthquakes prior to 1972 had magnitudes computed using a different technique, and the relationship between the two magnitude scales may not be 1:1. The shallow seismicity is characterized by proportionally more small earthquakes, compared to the deep seismicity.

In a more recent study (WCC, 1981a), events shallower than 3 km were found to have a b value of 1.0, while events deeper than 3 km were found to have a b value of 0.72. Earthquake swarms (which sometimes contain events deeper than 3 km) have a b value of 1.15, in agreement with the b value found for events <6 km (UWGP, 1979).

The difference in b values between the shallow events and deep events suggests that the seismogenic processes for the two groups of earthquakes are different. Focal mechanism solutions indicate that the rocks respond to the same stress regime, but the mechanics of deformation may be different. Lower b values (≤0.9) are characteristic of tectonic earthquakes in the North American Cordillera, while higher b values are characteristic

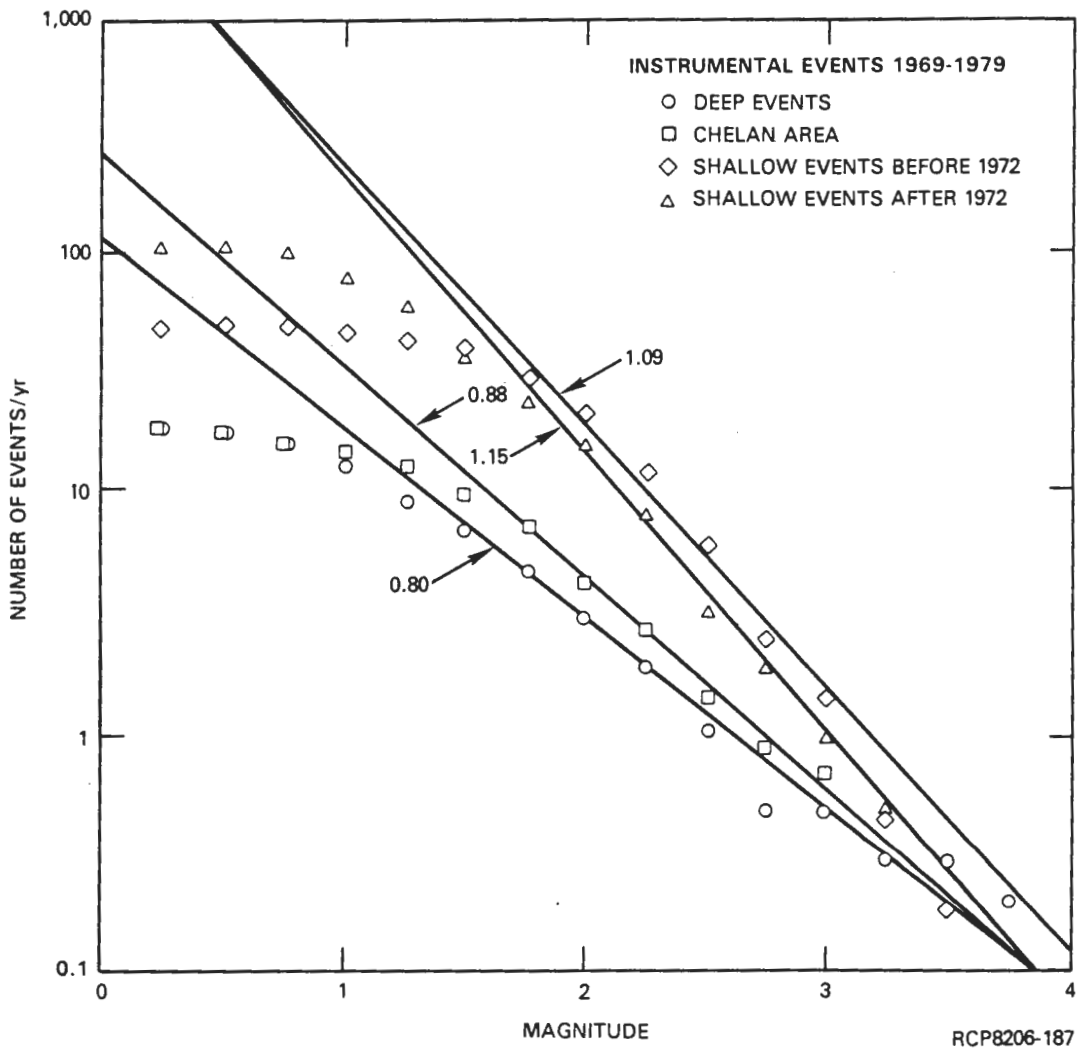

FIGURE 6-5. Recurrence Relationships (after UWGP, 1979). The b values, or magnitude distribution of shallow seismicity, are different from deep or regional seismicity.

of regions that exhibit swarm activity (UWGP, 1979). The b values may reflect the nature of the stress, strain rate, mechanical properties of the rock, and the data sample, for example, but differences in b values between areas are not clearly understood.

This division of events into shallow and deep earthquakes was based on the seismic velocity structure of the central Columbia Plateau, which shows a 6-km thickness of material with velocity of 5.15 km/s. Magnetotelluric studies (Chapter 4) suggest that the upper crustal layer is actually composed of two units. The upper resistive layer averages 3 km in thickness and is considered to be the basalt. A more conductive zone beneath the basalt, which also averages 3 km in thickness, is considered to be a sedimentary sequence. This zone could have a seismic velocity similar to the basalts, or the velocity could be lower in the sediments. A lower velocity layer beneath basalt is difficult to detect with either refraction data or earthquake data. (An undetected low velocity zone beneath the basalts could introduce larger errors in the interpreted depth distribution of the earthquakes based on the present location model.)

The layered nature of the basalts has been suggested by Rothe (1978) to account for the higher b values characteristic of the shallow swarm activity. According to Rothe (1978), swarm earthquakes occur within, and are bounded by individual basalt flows. He further suggested that the limited thickness of the flows controls b values by increasing the number of events with a small characteristic magnitude (~1.5). If events were limited to one thick flow, this would tend to increase the number of small events compared to larger events, thereby increasing the b value. In support of this hypothesis, he cited the concentration of the events at Wooded Island, in a depth range correlated with a thick basalt flow interpreted from geophysical logs. Rothe also hypothesized that the dense interiors of basalt flows could be expected to be more competent and could support higher stresses before fracturing during a seismic event. The weathered zones, interflows, and breccias would not support such stresses, and deformation could occur in an aseismic mode.

In contrast, WCC (1981a) suggests that the high b values are characteristic of low stress and strain rates based on the dependence of b values on those parameters in laboratory experiments. However, they accepted Rothe's conclusion that differences in thickness and resulting strength of flows may also control the size distribution of earthquakes.

ESTIMATION OF SEISMIC DEFORMATION RATE

Knowledge of two relationships is needed to translate the observed frequency and size distribution of earthquakes into deformation rates. The seismic moment of earthquakes (M_0) is related to the area of the rupturing fault (A) and the average displacement of the fault (D) by the relationship:

$$M_0 = \mu DA \qquad (1)$$

where

μ = the rigidity of the rock (taken to be 3×10^{11} dynes/cm^2) (Lee and Stewart, 1981).

The empirical relationship between magnitude and seismic moment (Lee and Stewart, 1981) is:

$$\log M_0 = 16 + 1.5M_L \qquad (2)$$

where

M_L = local (Richter) magnitude.

Thus, earthquake size and frequency can be used to determine an average rate of seismic slip on particular structures, or as deformation distributed throughout a region (Aki and Richards, 1980).

Such estimates are based on several assumptions, the most critical of which are accuracy of magnitude measurement and stability of the earthquake frequency with time. The deformation rates estimated are most dependent on the largest earthquakes in the data sample; therefore, relatively long time periods provide better estimates. However, this requires the use of historical data for which earthquake magnitudes are less reliable. Thus, historical data may not improve the reliability of the estimates significantly.

Two volumes of rock are considered in estimating deformation rates: the first volume includes the basalts and the second, a larger volume, includes rock below basalts. The shallow seismicity in the basalts is considered from 46 to 47° N., 118.75 to 120° W., and at depths <3 km. In this volume, the two main events are the magnitude 4.4 Royal Slope event and the magnitude 3.8 earthquake near Coyote Rapids. Using the moment-magnitude relationship above for all events in this volume between 1969 and 1979, a seismic moment release of 5×10^{21} dyne-cm/yr is calculated. Approximately 80% of this quantity is contributed by the magnitude 4.4 event, 10% is from the magnitude 3.8 event, and the remaining 10% is from events in the magnitude range 3.0 to 3.35.

All of the recorded shallow seismic activity greater than magnitude 3.0 occurs between the Frenchman Hills and Umtanum Ridge-Gable Mountain anticlines. For the purpose of making a seismologic estimate, which can be compared to the geologic estimates of deformation (Chapter 5), this activity is assumed to correlate with the Saddle Mountains structure.

The Saddle Mountains structure and the earthquake epicenters both are found in the 100-km-wide region considered for this analysis. The greatest geologic deformation and seismic activity are concentrated from Saddle Gap westward to Sentinel Gap (see Fig. 1-1). The anticline and the earthquake epicenters both appear to be shifted to the south at a position

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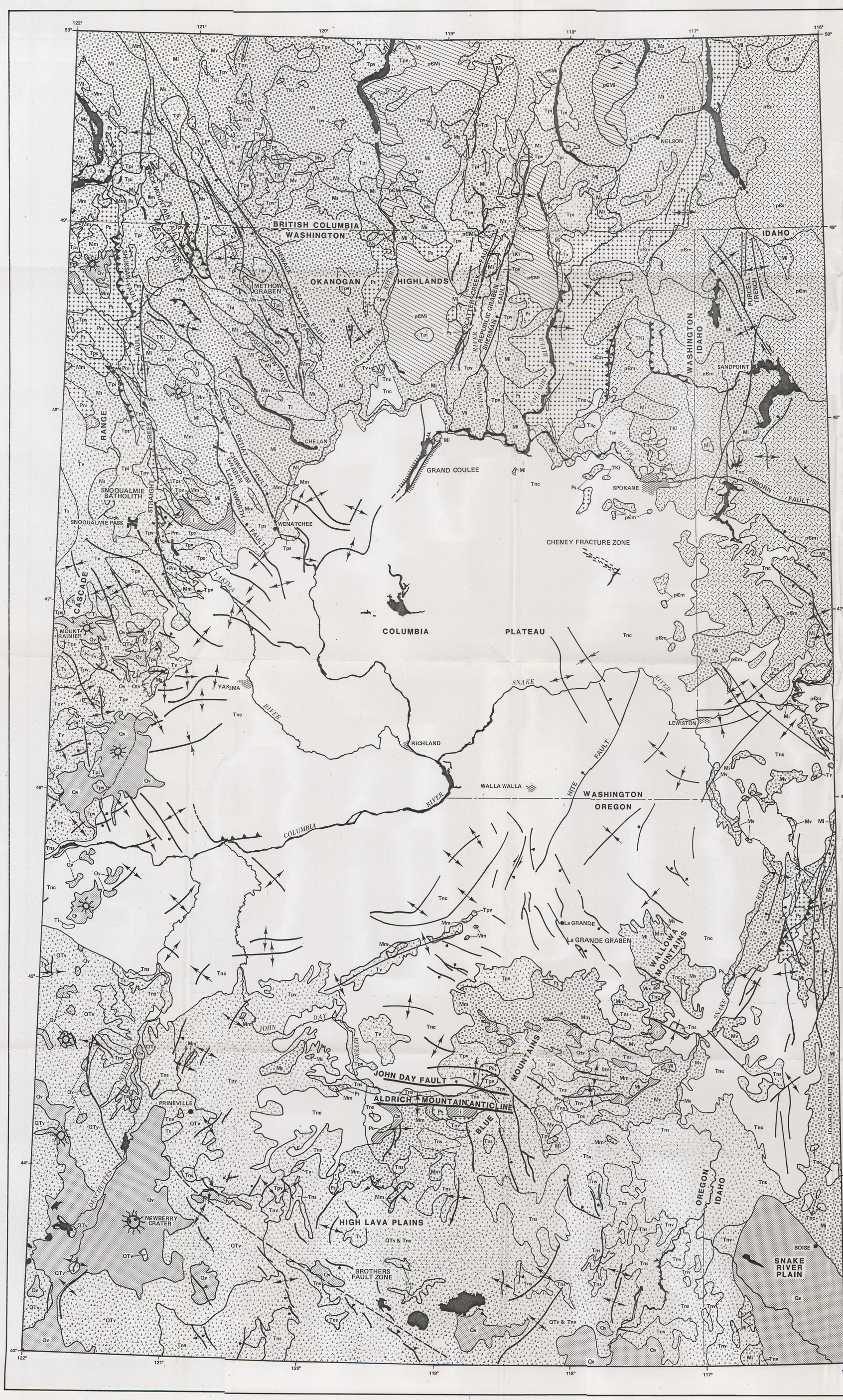
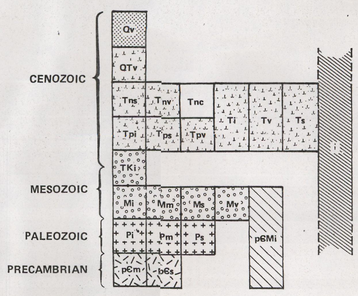
PLATE 2-1
Columbia Plateau Tectonic Map



AGE CORRELATIONS



DESCRIPTION OF MAP UNITS

Qv – QUATERNARY VOLCANIC ROCKS – ANDESITIC AND BASALTIC FLOWS AND PYROCLASTIC DEPOSITS.

QTv – PLIOCENE AND PLEISTOCENE VOLCANIC ROCKS – PREDOMINANTLY ANDESITIC AND BASALTIC FLOWS AND FLOW BRECCIA AND MINOR ASSOCIATED MUD FLOWS.

Ti, Tv, Ts – TERTIARY INTRUSIVE IGNEOUS, VOLCANIC AND SEDIMENTARY ROCKS OF PALEOGENE AND/OR NEOGENE AGE UNDIVIDED.

Tnc – BASALTIC AND ANDESITIC FLOWS OF THE COLUMBIA RIVER BASALT GROUP; NEOGENE SEDIMENTARY ROCKS.

Tnv – NEOGENE VOLCANIC ROCKS – BASALTIC FLOWS AND FLOW BRECCIAS WITH ASSOCIATED PILLOW PALAGONITE; SOME TUFFACEOUS SEDIMENTARY INTERBEDS.

Tns – NEOGENE SEDIMENTARY ROCKS–SANDSTONE, SHALE, CONGLOMERATE AND TUFF, INCLUDES SOME VOLCANIC ROCKS. ALSO INCLUDES SOME QUATERNARY SEDIMENTS.

Tpv – PALEOGENE VOLCANIC ROCKS– ANDESITIC AND BASALTIC FLOWS AND FLOW BRECCIA AND ASSOCIATED PYROCLASTIC DEPOSITS, MUD FLOW AND SEDIMENTARY ROCKS.

Tps – PALEOGENE SEDIMENTARY ROCKS– SANDSTONE, SHALE, CONGLOMERATES, AND TUFF, MINOR COAL. INCLUDES SOME VOLCANIC ROCKS INCLUDES MARINE AND NONMARINE SEDIMENTARY ROCKS.

Tpi – PALEOGENE GRANITIC ROCKS– INTRUSIVE IGNEOUS ROCKS. PRIMARILY GRANITE, BUT RANGE IN COMPOSITION FROM GRANITE TO GABBRO, INCLUDES SOME ASSOCIATED PERIDOTITE ROCKS.

TKi – EARLY TERTIARY TO LATE CRETACEOUS GRANITIC ROCKS– INTRUSIVE IGNEOUS ROCKS. PRIMARILY GRANITE, BUT RANGE IN COMPOSITION FROM GRANITE TO GABBRO.

Mv – MESOZOIC VOLCANIC ROCKS –ANDESITIC AND BASALTIC FLOWS, TUFF, ASSOCIATED PYROCLASTIC DEPOSITS AND INTERBEDDED SEDIMENTARY AND METAVOLCANIC ROCKS. INCLUDES LATE ~~MESOZOIC~~ PALEOZOIC VOLCANIC AND VOLCANICLASTIC ROCKS.

Ms – MESOZOIC SEDIMENTARY ROCKS – SHALE, SANDSTONE, LIMESTONE, CONGLOMERATE. INCLUDES MARINE AND NONMARINE SEDIMENTARY ROCKS.

Mm – MESOZOIC METAMORPHIC ROCKS – PHYLLITE, SCHIST, MIGMATITE, GNEISS, AND SOME METASEDIMENTARY AND METAVOLCANIC ROCKS.

Mi – MESOZOIC GRANITIC ROCKS– IGNEOUS INTRUSIVE ROCKS. PRIMARILY GRANITIC, BUT RANGE IN COMPOSITION FROM GRANITE TO GABBRO.

Ps – PALEOZOIC ROCKS UNDIVIDED – SANDSTONE, SHALE, GREENSTONE, LIMESTONE, AND PHYLLITE, PRIMARILY MARINE SEDIMENTARY AND LOW–GRADE METAMORPHIC ROCKS.

Pm – PALEOZOIC METAMORPHIC ROCKS – SCHIST, QUARTZITE, MARBLE, AND GNEISS. PRIMARILY MEDIUM TO HIGH–GRADE METAMORPHIC ROCKS.

Pi – PALEOZOIC GRANITIC ROCKS– PRIMARILY GRANITE, INTRUSIVE ROCKS. INCLUDES SOME METAMORPHIC ROCKS OF SIMILIAR COMPOSITION.

p€s – PRECAMBRIAN ROCKS UNDIVIDED– SANDSTONE, SHALE, CHERT, DOLOMITE, GREENSTONE, SCHIST, AND QUARTZITE. PRIMARILY SEDIMENTARY AND LOW TO INTERMEDIATE GRADE METAMORPHIC ROCKS. INCLUDES SOME CAMBRIAN SEDIMENTARY ROCKS.

p€m – PRECAMBRIAN METAMORPHIC ROCKS – SCHIST, GNEISS, AND MIGMATITE, PRIMARILY HIGH–GRADE METAMORPHIC ROCKS.

p€Mi – PRECAMBRIAN TO MESOZOIC GNEISS DOME COMPLEXES – PARAGNEISS, ORTHOGNEISS, AND ASSOCIATED GRANITIC ROCKS.

i – PALEOZOIC TO TERTIARY MAFIC INTRUSIVE ROCKS – INTRUSIVE IGNEOUS ROCKS, OPHIOLITE COMPLEXES AND ALPINE PERIDOTITES. PRIMARILY MAFIC OR ULTRAMAFIC GABBRO, DUNITE, AND SERPENTINITE.

GEOLOGIC MAP SYMBOLS

CONTACT

FOLDS

APPROXIMATE OR INFERRED LOCATION OF AXIAL TRACE. ARROW INDICATES DIRECTION OF PLUNGE. DASHED WHERE INFERRED, DOTTED WHERE CONCEALED.

ANTICLINE

SYNCLINE

MONOCLINE

FAULTS

DASHED WHERE INFERRED, DOTTED WHERE CONCEALED.

NORMAL FAULT

BALL AND BAR ON DOWN THROWN SIDE

STRIKE SLIP

ARROWS INDICATE RELATIVE HORIZONTAL MOVEMENT

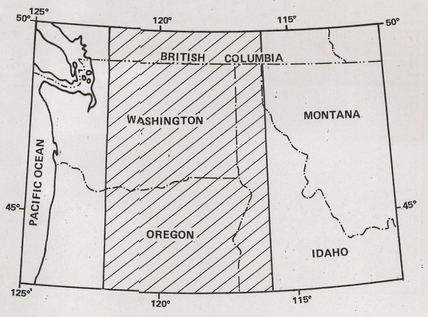
THRUST FAULT

SAWTEETH ON UPPER PLATE

MAJOR QUATERNARY ERUPTIVE VENT

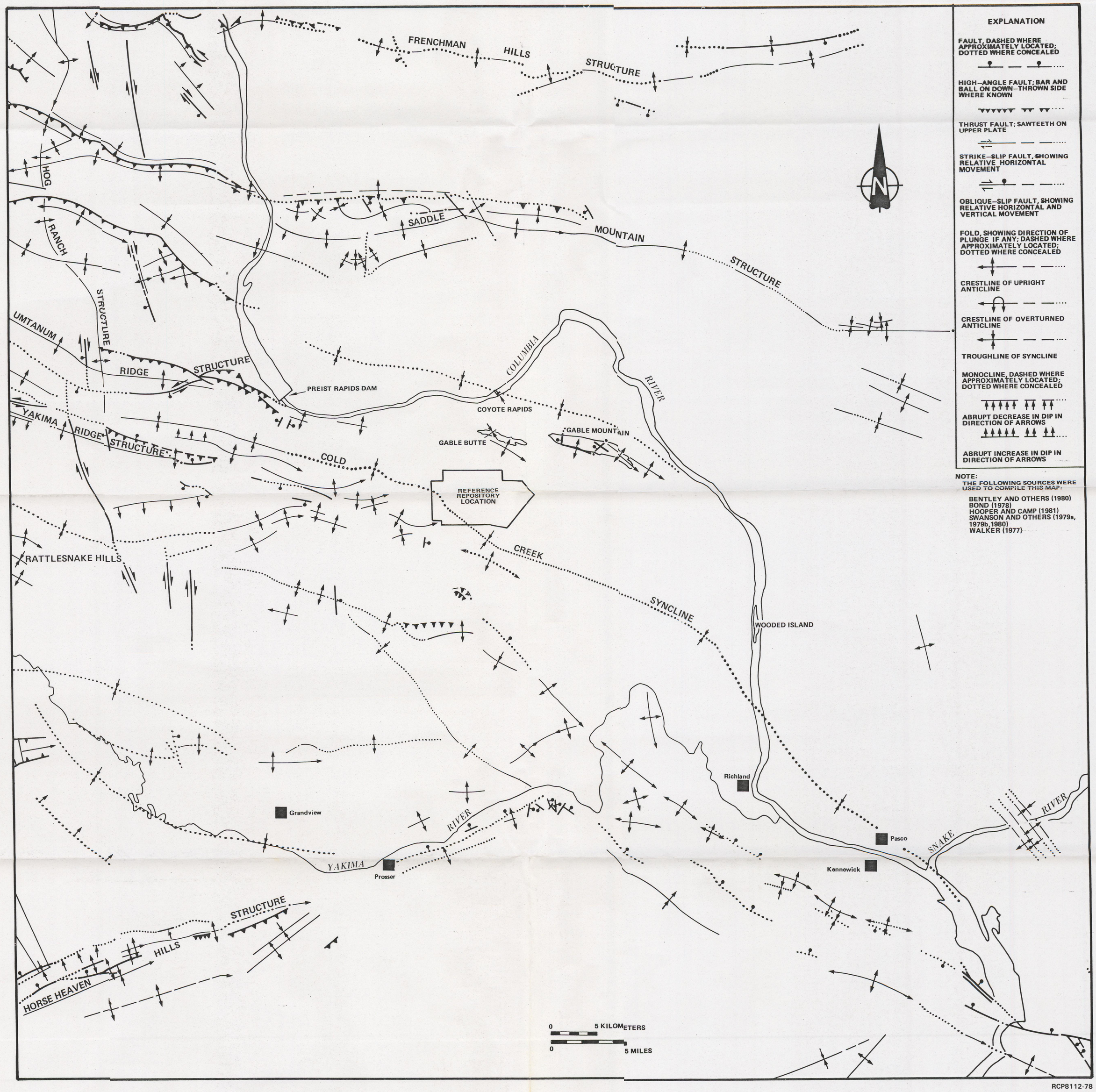
FOR STRUCTURES IN THE CENTRAL COLUMBIA PLATEAU SEE PLATES 2-1 AND 2-2

INDEX MAP



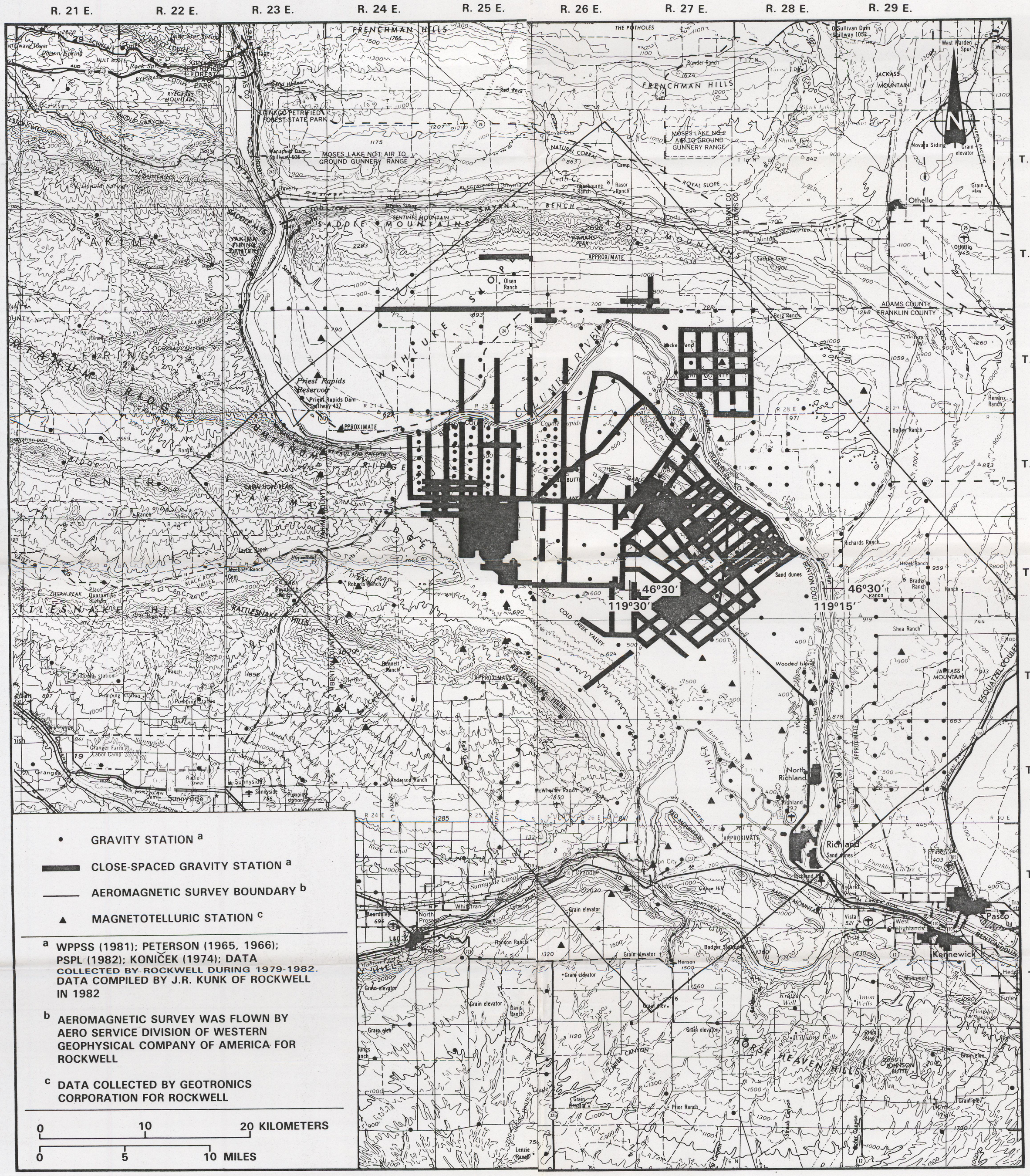
NOTE:
THE FOLLOWING SOURCES WERE USED TO COMPILE THIS MAP: BOND (1976), WALKER (1977), AND WPPSS (1981).

RHO-BW-ST-19 P
PLATE 3-1
Generalized Geologic Map of the Region



RCP8112-78

RHO-BW-ST-19 P
PLATE 2-2
Pasco Basin Tectonic Map



• GRAVITY STATION [a]

▬ CLOSE-SPACED GRAVITY STATION [a]

—— AEROMAGNETIC SURVEY BOUNDARY [b]

▲ MAGNETOTELLURIC STATION [c]

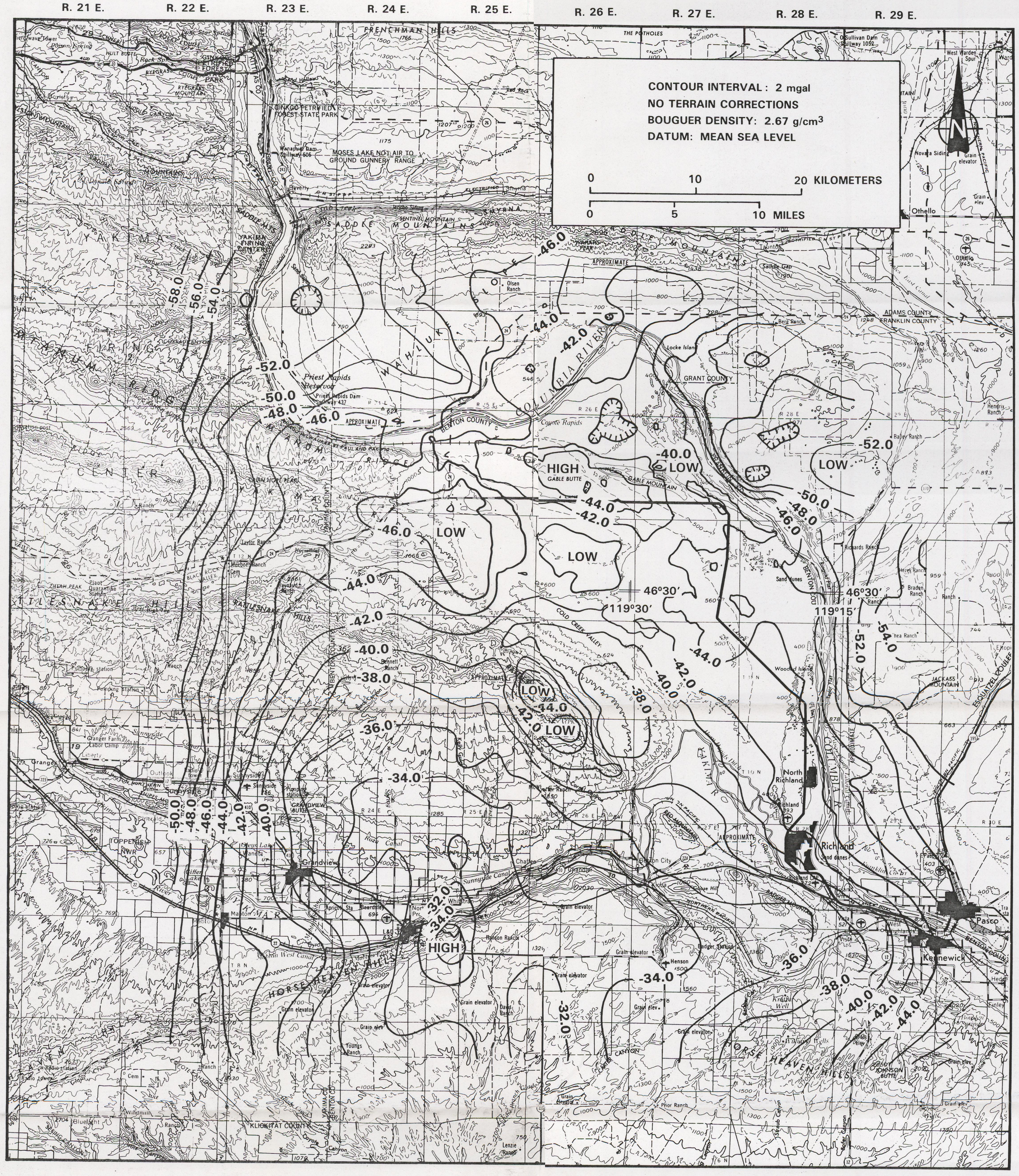
[a] WPPSS (1981); PETERSON (1965, 1966); PSPL (1982); KONIČEK (1974); DATA COLLECTED BY ROCKWELL DURING 1979-1982. DATA COMPILED BY J.R. KUNK OF ROCKWELL IN 1982

[b] AEROMAGNETIC SURVEY WAS FLOWN BY AERO SERVICE DIVISION OF WESTERN GEOPHYSICAL COMPANY OF AMERICA FOR ROCKWELL

[c] DATA COLLECTED BY GEOTRONICS CORPORATION FOR ROCKWELL

RCP8206-90

RHO-BW-ST-19 P
PLATE 4-1
Geophysical Survey Location Map



RCP8206-91

RHO-BW-ST-19 P
PLATE 4-2
Simple Bouguer Gravity

coincident with the change in the intensity of geologic deformation and seismic activity. Earthquake focal mechanisms are consistent with the observed style of deformation. However, there is only minor spatial association of the epicenters with the actual crest of the anticline or associated faults.

In order to make an estimate of the geologic deformation implied by the seismic activity, a simple geologic model must be assumed to estimate the total fault area. However, there is no difference between this estimate and the result of alternative methods that distribute the deformation throughout an area or volume of rock. For the purpose of making an estimate of deformation only, the displacement due to earthquakes is assumed to occur on an east-west-striking reverse fault that extends to a depth of 3 km, the estimated thickness of the basalts. The fault is assumed to dip 45°, so that the fault plane has a 5-km-long (down dip) surface. The length of the assumed fault is taken to be 50 to 100 km, since the seismic activity has an approximate east-west trend for 100 km, but most of the large events occur in the western 50 km of this zone of seismicity.

An estimate of seismic deformation can now be made using the estimated cumulative seismic moment (per year) and a fault area of 250 to 500 km^2. Using Equation 1, an average displacement rate of 0.03 to 0.06 mm/yr is calculated. This amount of displacement on a 45° reverse fault represents equal parts of north-south contraction and vertical uplift, each equivalent to 0.02 to 0.04 mm/yr. It should be emphasized that while a fault model has been assumed, this estimate could equally well represent the total deformation associated with a wider zone north and south of the crest of the Saddle Mountains structure.

A comparative deformation rate estimate can be made for the region of 45.5 to 47.5^{o} N., 118 to 121° W., and at depths down to the base of the crust, 28 km. In this volume, the cumulative annual seismic moment is 11×10^{21} dyne-cm/yr. Half of this quantity is represented by the two large, shallow earthquakes in the basalt. The remainder of the events in this crustal volume contribute 6×10^{21} dyne-cm/yr. However, the volume of sub-basalt rock is over ten times larger than that of the overlying basalt, indicating that the amount of seismic deformation per unit volume is higher in the basalts than in the rocks below the basalts. This analysis suggests that deformation of the basalts may be more seismogenic than the deformation in the deeper zones.

This analysis suggests that the basalts either behave in a more brittle fashion than deeper crustal layers, or the basalts are mechanically decoupled from the underlying layers. Because of the increased pressure and temperature in the deeper crust, it seems more likely that a greater proportion of the total deformation in the deeper zones may occur in an aseismic mode.

These estimates are based on the assumption that the coda-length magnitudes calculated by the University of Washington are equivalent to local (Richter) magnitudes. The coda-length magnitudes are probably only accurate to ± 0.25 units. These magnitudes were compared to Wood-Anderson

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These estimates are based on the assumption that the coda-length magnitudes calculated by the University of Washington are equivalent to local (Richter) magnitudes. The coda-length magnitudes are probably only accurate to ± 0.25 units. These magnitudes were compared to Wood-Anderson

instruments to determine their relationship to the classical local magnitude scale, M_L (UWGP, 1979). This comparison showed ± 0.5 magnitude units of scatter. If the coda-length magnitudes were in error by this amount, the estimates of seismic moments of the few, large events could be in error and would be approximately accurate to a factor of 5. Because of the other assumptions and empirical relationships used to estimate the seismic deformation in the Saddle Mountains area, the two estimates are not significantly different.

Another limitation of these estimates is that they have been based only on the last 10 yr of seismic activity. The historical record of seismic activity should also be considered in order to evaluate the stability of such estimates over a longer time period. In the central Columbia Plateau, the only other event of significance to the estimate is the 1918 Corfu earthquake (see Table 6-1), which had a location and magnitude similar to the 1973 Royal Slope event (WCC, 1981a). This Corfu earthquake occurred over 50 yr ago suggesting that a magnitude 4.4 event occurs on the Saddle Mountains structure only once every 30 to 60 yr, instead of once every 10 years as suggesting by using only the post-1969 instrumental period. However, there is an even greater uncertainty in the magnitude of the 1918 event, so that estimates using instrumental data since 1969 are probably more reliable.

It should be noted that if the 1936 Milton-Freewater event is considered ($M_L = 6$) in an analysis for the central Columbia Plateau, then the average rate of seismic deformation in the larger region is ~10 times higher than previously estimated from the post-1969 instrumental data. This is a further illustration of how larger events influence such an analysis. The events in the Milton-Freewater (Blue Mountains) region have focal mechanisms that distinguish them from the focal mechanisms of central Columbia Plateau seismicity and that are inconsistent with the style of deformation in the Columbia River basalts in the Pasco Basin area.

Over the last 10 yr, the observed seismic activity in the vicinity of the Saddle Mountains has contributed an estimated 0.02 to 0.04 mm/yr of both uplift and north-south contraction. Estimates of the uplift and north-south shortening on this structure based on geologic evidence (see Chapter 5) suggest a comparable rate of 40 m/m.y. (0.04 mm/yr). Although these estimates are not significantly different, there is a suggestion that some proportion of the observed geologic deformation may be occurring in an aseismic mode, if these two estimates are correct. A significantly smaller rate of seismic deformation is indicated for deeper deformation. This lower rate of seismicity in the sub-basalt layers indicates that a greater proportion of the total deformation is likely occurring as aseismic (or ductile) deformation.

Rothe (1978) suggested that the heterogeneous physical properties of the Columbia Plateau basalts could dominate the observed characteristics of the shallow seismicity. The often observed slow migration of activity indicates that interaction of discrete events comprising the swarms occurs principally in an aseismic mode. Clearly, a causal interaction is required

to explain the broad spatial and temporal grouping of such small earthquakes. Rothe's hypothesis suggests that this observed migration is due to slow, aseismic deformation in weaker parts of the flows or interbeds than that which occurs in the stronger and more brittle interiors of the flows.

GEODETIC MEASUREMENT OF DEFORMATION

Introduction

Contemporary deformation of the Earth's surface can be measured using repeated measurements of relative distances or elevations between points. In the United States, repeated measurements of elevation have been made by the U.S. Coast and Geodetic Survey (Whitten, 1966). The U.S. Geological Survey has developed an extensive trilateration program for measuring changes of horizontal positions in selected locations of the western United States (Savage and Prescott, 1973; Prescott and others, 1979; Savage and others, 1981).

This section first reviews an earlier analysis of elevation changes using U.S. Coast and Geodetic Survey leveling data (Tillson, 1970). These data are reported to show long-term subsidence of the central portion of the Columbia Plateau, relative to surrounding regions on the order of 1 mm/yr. The results of U.S. Geological Survey trilateration studies at the Hanford Site since 1972 are then described. These studies suggest minimal horizontal strain accumulation that barely exceeds the limitation of the technique. Supplementary geodetic networks have recently been installed at selected areas near the Hanford Site by the BWIP and are described later.

Leveling Surveys

A reanalysis of geodetic data in the Columbia Plateau was reported by Tillson (1970) who analyzed repeated leveling surveys and triangulation surveys. U.S. Coast and Geodetic Survey leveling data for four lines were utilized: Seattle, Washington to Pasco, Washington; Pasco, Washington to Spokane, Washington; Pasco, Washington to Ontario, Oregon; and Pasco, Washington to Portland, Oregon. Results of Tillson's (1970) analysis of leveling data suggest that the Pasco Basin is undergoing a gradual, contemporary basining, which is proceeding at an average rate on the order of 1 mm/yr. Only first-order leveling surveys were used, which are defined as surveys with leveling loop closure of at maximum 4 mm times the square root of the length of the level line. The most consistent and accurate data came from the Portland to Pasco line, which showed 33 mm of subsidence in 22 yr or 1.5 mm/yr.

The Seattle to Pasco and Pasco to Spokane level lines also showed subsidence of between 1 and 2 mm/yr. The leveling surveys did not indicate any significant movements where they crossed mapped structural discontinuities. In particular, the leveling data do not indicate significant vertical displacement rates at Wallula Gap, where the Olympic-Wallowa

lineament (see Chapter 2) crossed the Pasco to Ontario level line. This line showed <1 mm/yr of subsidence as far south as Pendleton, Oregon. The 1-mm/yr rate of basinal subsidence observed is believed by Tillson (1970) to be comparable with other measured areas where crustal movement is minor or negligible.

Tillson's (1970) analysis of triangulation data for the Columbia Plateau provides no evidence for horizontal crustal displacement. All position changes were less than the magnitude of the errors associated with the measurements or were attributable to the instability of particular surveying monuments.

TRILATERATION SURVEYS

A trilateration network to measure strain accumulation at the Hanford Site was emplaced in 1972. This network consists of 19 named monuments shown in Figure 6-6. Distances between these monuments have been measured by the U.S. Geological Survey along 29 lines (Fig. 6-6) with an average line length of 19 km. Nearly all of these lines have been surveyed seven times: twice in 1972 and once in 1973, 1975, 1978, 1979, and 1981.

During 1981, additional trilateration survey monuments were emplaced by the BWIP (Fig. 6-6), which the U.S. Geological Survey has surveyed once. The strain calculations for the 9-yr measurement period are described below. No determination of strain can yet be made for the recently installed networks.

The units used in geodetic surveys are termed strains or microstrains (10^{-6} strains). These units are dimensionless measures of line length changes ΔL, over different total line lengths, L. It is assumed that the deformation within a geodetic network is uniform, so that longer line lengths change proportionally the same as shorter line lengths in a given time. In this sense:

$$1 \text{ strain} = \Delta L/L \qquad (3)$$

where

L and ΔL = the same units (e.g., meters).

The unit "1 microstrain" thus corresponds to 1 mm/1 km; i.e., line lengths change 1 mm for each kilometer of their length.

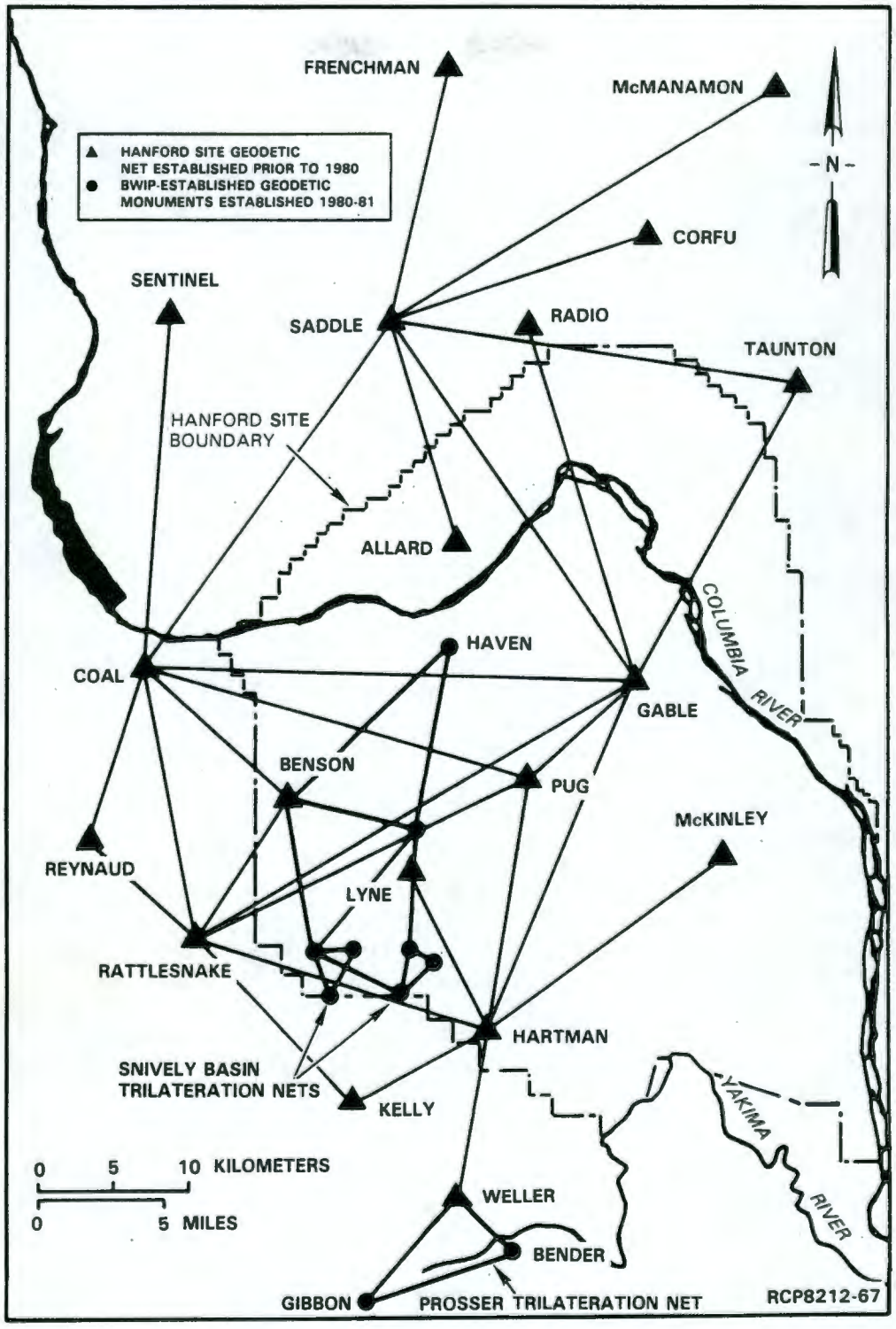


FIGURE 6-6. Hanford Trilateration Network, 1972 to Present. Geodetic monuments and survey lines are shown.

All measurements from 1972 to 1981 have been made using U.S. Geological Survey calibrated surveying equipment. The distance measuring instrument utilized was a Spectra Physics Geodolite 3G. Air density was determined by measurement of temperature and humidity from aircraft flown along the lines and from end-point pressure observations. Current estimates of random and systematic errors of distance measurements have not changed since an earlier error analysis of the Hanford data (Savage and Prescott, 1973). The standard deviation of a distance measurement in millimeters is given by:

$$\sqrt{a^2 + b^2L^2} \tag{4}$$

where

$a = 3$ mm

$b = 2 \times 10^{-7}$ (a unitless constant)

L = the line length in millimeters.

In the above formula, the $a = 3$ mm basic accuracy limitation is attributed to the stability of station monuments, error in centering the instrument, and the limited resolution of the instrument. The distance dependent term (b) is due to the limited accuracy of the atmospheric corrections previously mentioned. In terms of strain, lines longer than 20 km have nearly optimal accuracy of 2×10^{-7} or 0.2 microstrains. The fixed error of 3 mm is proportionately larger for strain measurements at shorter distances.

Determination of the strain within a network of distance observations is more accurate than that of the component lines making up the network. The random errors associated with individual lines tend to average out in a highly redundant network during each survey. However, systematic errors that may be caused by the limited accuracy of instrument calibration, including meteorological instruments, may bias all line lengths. Because different instruments are used during different surveys, each survey may be contaminated by such systematic errors, but over a long time period with repeated surveys, these should also tend to average out.

For the Hanford Site network, which has an average line length of 19 km, the average absolute error is 5 mm, and the error in terms of strain is 0.25 microstrains. These random errors average out and produce an error for the network as a whole of 0.01 to 0.02 microstrains. Systematic errors are estimated to produce the equivalent of 0.01 to 0.02 microstrains/yr.

The analysis of the Hanford Site geodetic data for the seven surveys through 1981 resulted in the principal strain rates (in units of microstrains per year):

$$E_1 = -0.016 \pm 0.018, \text{ N. } 5^{o} \text{ E.} \quad (5)$$

$$E_2 = -0.041 \pm 0.018, \text{ N. } 85^{o} \text{ E.} \quad (6)$$

Negative strains denote a contraction or shortening of lines. Thus, the significant strain rate for Hanford is an east-west contraction. For the average line length of 19 km, the average absolute change in line length is -0.78 mm $\pm$0.34 mm. While the magnitude of the principal strain rates is unchanged from the 1972 to 1979 surveys reported by Savage and others (1981), the greatest contraction determined in that study (from 1972 to 1979) was oriented northeast-southwest. However, the magnitude of the principal strain rates are only marginally greater than the standard errors, and the change in orientation of the axes is not significant.

The network strain components are plotted as a function of time in Figure 6-7. In this figure, E_{11} and E_{22} measure linear strain on east-west- and north-south-oriented lines. Shortening of east-west lines is observed. The difference, $E_{11} - E_{22}$, is given by γ_1, which measures shear on vertical planes oriented at 45° to E_{11} and E_{22}. Left-lateral shear on northwest-southeast-oriented vertical planes or right-lateral shear on northeast-southwest-oriented planes is indicated by the negative trends.

Also shown is γ_2, which is twice the tensor shear on lines parallel to E_{11} and E_{22}. Left-lateral shear on planes oriented north-south or right-lateral shear on planes oriented east-west is positive. The recent change in γ_2 from a negative to a positive trend observed as a result of the 1981 survey is responsible for the change in the orientation of the principal strain axes between 1979 and 1981. Also shown is the total areal dilation ($E_{11} + E_{22}$), which indicates that the total area is decreasing. It could be hypothesized that this implies that the region as a whole is being uplifted, although the areal dilation is particularly sensitive to systematic errors.

In view of the magnitude of the random and systematic errors inherent in the method and the low strain rates indicated by the observations, the absolute values of strain rates determined from geodetic data at Hanford are only marginally significant. Significant is the low rate of strain accumulation compared to tectonically active areas and the order of magnitude agreement with geologically determined rates of strain. For example, principal strain rates measured on similar networks along the San Andreas fault are in the range of 0.3 to 0.5 microstrains/yr. Conversely, principal strain rates in Nevada are comparable to those found at Hanford, and yet the State of Nevada is known for significant seismic and tectonic activity (Prescott and others, 1979). It was noted by Savage and others (1981) that annual strain accumulation at Hanford is no greater than the strain induced by the semidiurnal earth tides, so the lack of correlation

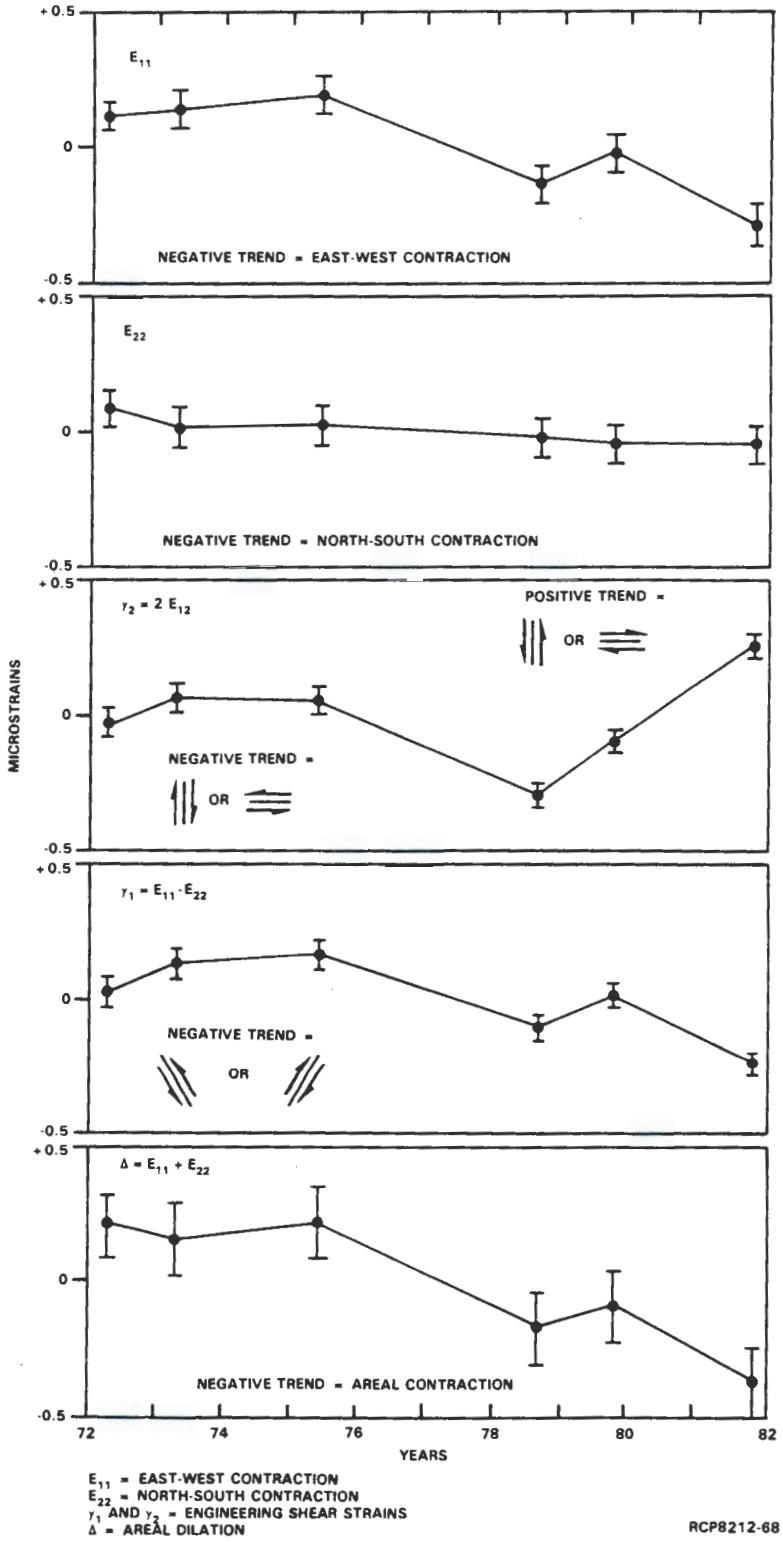


FIGURE 6-7. Hanford Network Strain Components, 1972 to Present.

of marginally significant strain rates with seismicity is not surprising. However, both seismicity and geodetic surveying suggest continuing deformation at low rates of strain similar to those suggested by geologic analysis over a longer period of time.

The geodetic strain measurements imply a maximum compression in the east-west direction, instead of north-south as expected from the east-west-trending folds and from earthquake focal mechanisms. It was noted by Savage and others (1981) that the east-west compression determined from geodetic measurements agrees with the focal mechanisms of earthquakes in the Milton-Freewater area. The July 15, 1936 Milton-Freewater earthquake is the largest historical earthquake in the Columbia Plateau region (surface wave magnitude of 5.75), but its focal mechanism is clearly different from most focal mechanism solutions in eastern Washington (WCC, 1980b). Although focal mechanisms farther east (near Milton-Freewater) indicate east-west compression in agreement with the geodetic data, the Hanford Site geodetic data may not yet have sufficient accuracy to detect the low strain rates in the Hanford region.

The BWIP has installed additional geodetic networks in order to measure the rate of strain accumulation in the Milton-Freewater area, along the Wallula fault system and along the Rattlesnake Hills anticline. The Milton-Freewater earthquake has been assumed for purposes of licensing nuclear powerplants at the Hanford Site to occur on the Rattlesnake-Wallula topographic alignment. This structure was chosen for additional geodetic observation. Geomorphic features interpreted to be surface fault-line ruptures have locally been noted in association with the inferred location of the Wallula fault system and the Rattlesnake Hills anticline (Bingham and others, 1970; Bond and others, 1978; Farooqui, 1979; Kienle and others, 1979), but trenching has not confirmed tectonic displacement.

Three triangular networks independent of one another were established along the inferred trace of the Wallula fault system (Fig. 6-8). Two existing monuments were added to the existing Hanford Site network near Prosser, and seven additional monuments were installed in the Snively Basin western Cold Creek syncline area (see Fig. 6-6). These supplementary networks have been established to assess whether deformation is localized along particular structures. If deformation is localized, long-term average rates determined for the entire basin by averaging local effects over a larger volume would be an unrealistic model of deformation.

During 1981, lengths of the newly established lines were measured for the first time under the direction of the U.S. Geological Survey. The geodetic trilateration network (see Fig. 6-6 and 6-8) now consists of 53 lines. However, for the independently measured networks (shown in Fig. 6-8), average line lengths are 3-7 km, so that the accuracy of measurements in these networks in terms of average strain rate is significantly less than for the longer lines of the previously existing network. This is because the fixed error represents a larger proportion of measured line length. Significant displacement on the targeted structure would tend to be averaged over a larger network, however.

FIGURE 6-8. Geodetic Trilateration Survey Lines Along the Wallula Fault System. These independent networks were installed in 1981 to detect episodic strain on the Wallula fault system.

In addition to distance measurements, elevations were determined to first-order precision with a leveling survey across the Cold Creek syncline network (Fig. 6-9). Subsequent surveys of this line will permit detection of differential vertical strains in addition to the horizontal strains.

SUMMARY

Historical and instrumental data indicate that the Hanford Site and the reference repository location are areas of relatively low seismicity compared to eastern Washington in general. The observed seismicity in the Pasco Basin is tentatively associated with the Saddle Mountains structure, but not with mapped faults along the feature. No concentration of activity is found on the Rattlesnake-Wallula alignment, which has been inferred by others (Chapter 8) to be a seismogenic structure. The regional seismicity also appears to be unrelated to mapped faults.

Focal mechanisms indicate that the basalt and rocks below are responding to a nearly north-south, nearly horizontal axis of compression. This orientation is consistent with the east-west oriented anticlines that began developing in Miocene time (see Chapter 5). The axis of least compression is generally near vertical, indicating thrust or reverse faulting due to horizontal stresses greater than vertical stresses. The rate of activity, size distribution (recurrence relationship), and the tendency to occur as swarm sequences suggest that the layered basalts deform in a manner differing from the sub-basalt rocks.

Most of the seismic activity within the Columbia River basalts in the central portion of the Columbia Plateau occurs in the vicinity of the Saddle Mountains structure. This seismic activity indicates a deformation rate of 0.02 to 0.04 mm/yr of both north-south contraction and vertical uplift. This estimate is in order-of-magnitude agreement with the larger estimates of 0.04 to 0.08 mm/yr determined from geologic data (see Chapter 5). However, no seismic deformation of comparable magnitude is observed for any structure other than the Saddle Mountains.

The geodetic data indicate a shortening of ~0.8 mm/yr, in an east-west direction, and a shortening of ~0.4 mm/yr in a north-south direction for the average line length of the survey (20 km). This is an order of magnitude greater than estimated from geological or seismological data over comparable distances. The east-west shortening of distances is inconsistent with the geologic and seismologic data in the vicinity of the Pasco Basin.

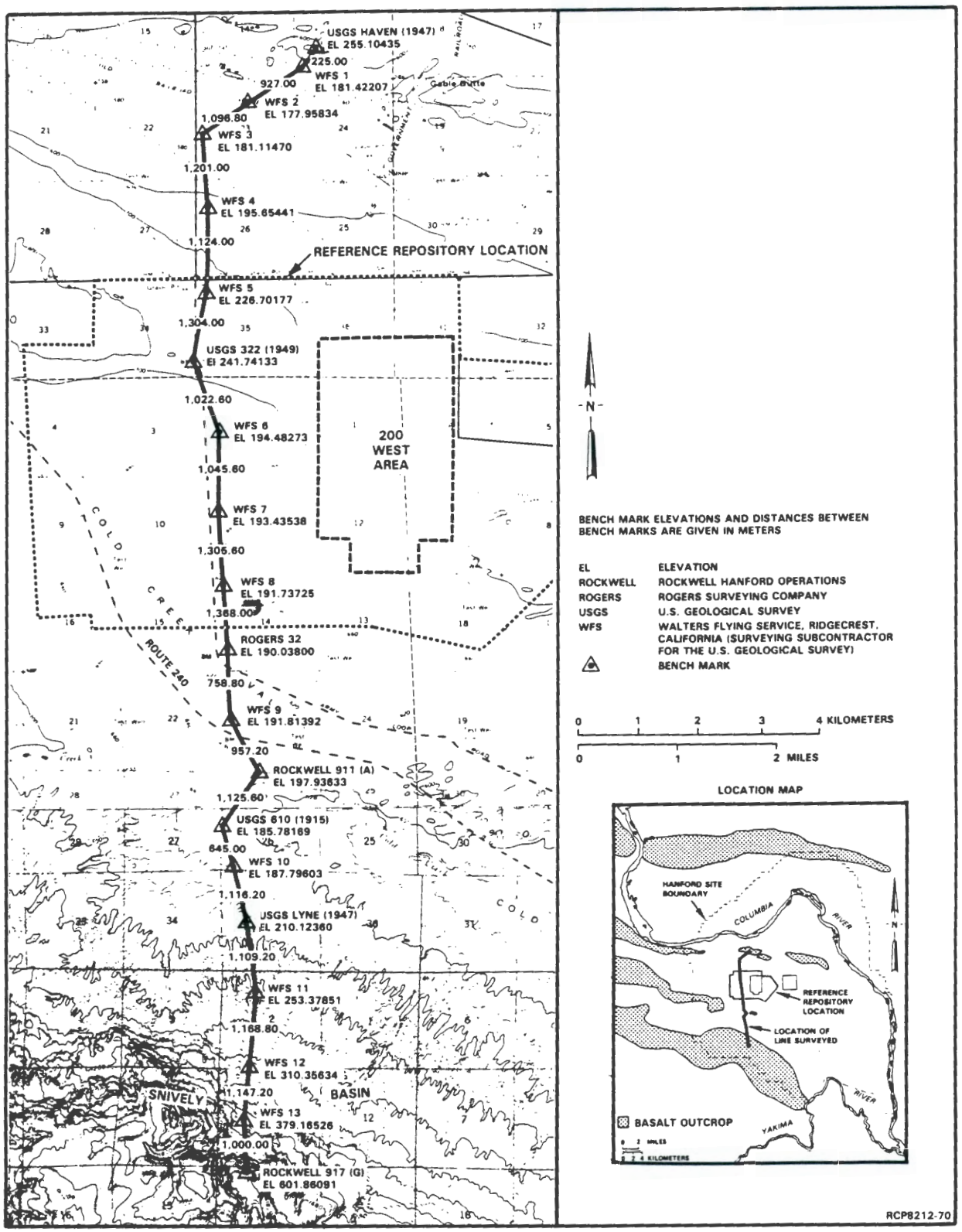


FIGURE 6-9. Location of Monuments and Lines; First-Order Leveling Survey of the Cold Creek Syncline.

CHAPTER 7 - A REVIEW OF TECTONIC MODELS OF THE COLUMBIA PLATEAU AND PASCO BASIN

D. W. Duncan

INTRODUCTION

The purpose of developing a conceptual tectonic model for the geologic past is to enable evaluation of the tectonic regime in the future. Such a model allows one to predict the location and rate of strain in the future, and thus, determine areas of probable relative stability.

The approach in this chapter is to compare available tectonic models and evaluate them on the basis of deformation style and rate data reported in preceding chapters for the reference repository location. Regional models, which include major parts of the Pacific Northwest, are examined first. The objective is to identify major tectonic movements and features, such as plate or microplate boundaries, distributed strain zones, or rigid blocks, and establish their impact on the tectonic setting of the Columbia Plateau. Local models, which deal primarily with the Columbia Plateau, are addressed next. The objective is to establish the details of Columbia Plateau tectonism with respect to details of deformation style and rate in the Pasco Basin area. Both regional and local models are limited primarily to the Cenozoic with emphasis on the Neogene and Quaternary.

REGIONAL TECTONIC MODELS

Thirty-three tectonic models of the Pacific Northwest are summarized in Table 7-1. Many models encompass the entire Cordilleran region and are included owing to their references to Pacific Northwest tectonics. Table 7-1 is designed to summarize the nature of geological observations or tectonic relationships used as a basis for the model, the interpreted rate of deformation in some part of the region, the tectonic pattern or driving force, and prediction for tectonic style in the future. Whereas some of the deformation rates are representative of the margins of the Pacific Northwest, they establish the tectonic style in other areas with which the Columbia Plateau tectonics can be compared. Figure 7-1 shows many of the locations of geologic and geographic features mentioned in Table 7-1.

The models of Table 7-1 express the range and diversity of explanations for the tectonic evolution of the Pacific Northwest. However, seven general classes of models can be differentiated, and the table is organized according to the classification. Inclusion of a given model in a class does not necessarily mean that the author disagrees with other models, nor that the classes of models are necessarily mutually exclusive. However, the classes provide a convenient framework for a discussion of current thinking about Pacific Northwest tectonics.

TABLE 7-1. Tectonic Models of the Cordillera with Emphasis on the Pacific Northwest. (Sheet 1 of 5)

Source	Area modeled; time	Geologic observations or tectonic relationships	Rate of deformation	Tectonic pattern or driving force	Prediction for future
			Active Subduction		
Ando and Balazs (1979)	Western Washington; present	Regional eastward tilt of western Washington	Uplift of +3 mm/yr at western Olympic Peninsula and subsidence of -3 mm/yr at western base of Cascades define regional eastward tilt	Juan de Fuca plate is continuously aseismically underthrusting the North American plate	Not given
Carson and others (1974)	Washington offshore; Pleistocene to present	Anticline formation and landward thrusting of marine sediments at base of continental slope	Not given	Active subduction off Washington coast	Not given
Hyndman (1976)	Southwestern British Columbia; recent	A zone of low heat flow from the coast to ~200 km inland is paralleled further inland by a zone of high heat flow	Not given	Subduction of the Juan de Fuca plate beneath the North American plate is active at present (or recently north of 51° N. latitude)	Not given
Keen and Hyndman (1979)	Southwestern Canada; 10 m.y. ago to present	Seismic data indicate that motion on Nootka transform fault (perpendicular to coast offshore from Vancouver Island) is sinistral	Perpendicular convergence between North American and Juan de Fuca plates is ~40 mm/yr	Active subduction of Juan de Fuca plate	Not given
Riddihough (1979)	British Columbia and Washington; present	Gravity high-low pair in arc-trench gap	Not given	Active subduction along margin of British Columbia and Washington	Not given
			Subducted Spreading Center		
McKee (1971)	Great Basin; Tertiary	Widespread Basin and Range faulting accompanied initiation of extensive basaltic magmatism (including Columbia River basalts) at ~16 m.y. ago	Not given	The Farallon-Pacific spreading center was subducted beneath the North American plate and caused crustal expansion and basaltic eruptions	The subducted spreading center has been active from 16 m.y. ago to the present
			Back Arc Regime		
Davis (1977)	Pacific Northwest and Columbia Plateau; late Cenozoic	Alternation of regional maximum stress from north-south to east-west, and back to north-south allowed north-south diking and east-west folding to develop since ~20 m.y. ago. Columbia Plateau tectonic block is separated from Basin and Range terrane by a major fault zone	Subduction of Juan de Fuca plate proceeds at 40 mm/yr	Subduction of Juan de Fuca plate creates back arc regime. Juan de Fuca and North American plates are uncoupled, accounting for lack of east-west compressional features in Columbia Plateau. Right-lateral transform occurs at plate boundary	Not given
Dickinson (1976)	Western North America; middle Miocene to present	Incipient crustal separation in back-arc (Columbia Plateau) possibly related to ductility induced by thermal effects of Oligocene arc magmatism	Not given	Continuing subduction of Farallon plate west of Cascades creates back arc terrane	Not given
Eaton (1979)	Western United States; late Cenozoic	Backarc extension in Columbia Plateau began at 16 to 17 m.y. ago and was oriented northeast-southwest. In the Basin and Range between 11 and 7 m.y. ago, spreading turned to eastwest, and after 7 m.y. ago changed to west-northwest-east-southeast. Spreading apparently ceased in the Columbia Plateau after 11 m.y. ago.	Not given	Hydrodynamic convection in asthenosphere drags the lithosphere apart, causing tension and eventual failure.	Not given

TABLE 7-1. Tectonic Models of the Cordillera with Emphasis on the Pacific Northwest. (Sheet 2 of 5)

Source	Area modeled; time	Geologic observations or tectonic relationships	Rate of deformation	Tectonic pattern or driving force	Prediction for future
Eaton and others (1978)	Western Cordillera; Cenozoic	Early Cenozoic northeast-southwest spreading in southern Washington to southern Nevada. Late Cenozoic east-west spreading, initiated between 9 and 7 m.y. ago, overprinted earlier spreading regime and accelerated in rate	Not given	Rise and divergent flow of asthenosphere accompanying mantle upwelling caused spreading	Continuing today
Stewart (1978)	Basin and Range and Columbia Plateau; middle and late Cenozoic	Basin and Range province characterized by low crustal seismic velocities, thin crust, high heat flow, uplift and extension, previous deformational history, and inland position relative to transform plate boundary	Not given	Back arc spreading in Basin and Range and probably Columbia Plateau	Not given
Mantle Plume, Diapir, or Hot Spot					
Armstrong (1978)	Northwestern United States; Cenozoic	On the Columbia Plateau rapid basalt extrusion from 16 to 13 m.y. ago (Columbia episode) was followed by waning magmatism. Dissection, subsidence, and deformation of the plateau followed	Not given	Subduction on continental margin and crustal disruption from the Yellowstone hot spot account for tectonism since the Columbia episode. Origin of Columbia episode is unexplained	Not given
Smith and Sbar (1974)	Western United States; present	Fault plane solutions from the Intermountain Seismic Belt suggest the Northern Rocky Mountain subplate (north of the Snake River Plain) is moving northward away from the Great Basin subplate and westward away from the remainder of the North American plate	Juan de Fuca plate is subducting to the east-northeast at 28 mm/yr	Lithospheric rifting along the Snake River Plain is caused by westward migration of North American plate over a mantle plume (Yellowstone). Extension occurs to west of Intermountain Seismic Belt, owing to radial stress away from Yellowstone plume	Relative east-northeastward migration of Yellowstone plume can be expected to continue
Suppe and others (1975)	Western United States; late Cenozoic (10 to present m.y. ago)	Fanning pattern of Yakima fold and thrusts indicates that the Western United States subplate (to the south) has been moving north and west against northeastern Oregon and Washington. The Columbia Plateau clockwise sense of rotation is related to distributed extension to the south	1 to 10 mm/yr extension between Western United States and North American plates	Western United States subplate is rotating counterclockwise about a center in southern California. Rising asthenosphere at plate boundary or heating of lithosphere from below or some combination of processes may be the cause of regional uplift and extension	Continuing today
Thompson (1977)	Western United States (Columbia-Snake River-Yellowstone Province); Cenozoic	The volume and rate of magmatism in the Columbia Plateau and western Snake River Plain greatly exceed the eastern Snake River Plain. This is evidence against the Yellowstone deep mantle plume producing province-wide magmatism	Rate of magmatism: Columbia River Plateau, 100,000 km^3/m.y.; western Snake River Plain, 31,000 km^3/m.y; eastern Snake River Plain, 3.3 km^3/m.y.	Tectonic and magmatic features produced by shallow diapiric mantle upwelling in both the Basin and Range and the Columbia Plateau. Constiction of Columbia Plateau diapir by shallowly subducting Farrallon plate led to "run-away" mantle fusion and massive basaltic magmatism	Mantle below Columbia Plateau has been semistagnant since 13 m.y. ago

TABLE 7-1. Tectonic Models of the Cordillera with Emphasis on the Pacific Northwest. (Sheet 3 of 5)

Source	Area modeled; time	Geologic observations or tectonic relationships	Rate of deformation	Tectonic pattern or driving force	Prediction for future
			Shear Between Plates		
Atwater (1970)	Western North America; late Cenozoic	Pacific Northwest undergoing slight east-west compression and regional dextral shear on northwestern-trending zones	Oblique motion of 60 mm/yr	Subduction of Juan de Fuca plate and transform of Pacific plate relative to North American plate	Not given
Beck (1976, 1980)	Western edge of North America; Cenozoic	Northwestward translation and clockwise rotation of crustal blocks on western margin of North America. Oregon-Washington Coast Range Province rotated as independent microplate. Columbia Plateau basalts unrotated	15 mm/yr of relative northward movement in early and middle Tertiary for western Washington and Oregon	Transform faulting, oblique subduction, or oblique rifting between Pacific or Farallon plates and North American plate	Not given
Christiansen and McKee (1978)	Great Basin and Columbia Intermontane regions; late Cenozoic	Extension of a few kilometers north of High Lava Plains accompanied by basaltic magmatism, increased heat flow, lowered lithospheric rigidity, and regional uplift	Extension inactive since 14 m.y. ago	Partial coupling of transform between Pacific and North American plates caused east-west extension	Not given
Coney (1978)	North American Cordillera; early Miocene to present	East-west extension and cessation of subduction related to wide transform zone in thermally weakened Basin and Range Province	Not given	Subduction of Farallon plate virtually ceases; transform of Pacific plate predominates	Continued minimal subduction and predominant transform
Crosson (1972)	Puget Sound region; present	North-south compression determined for contemporary earthquakes	Not given	Subduction recently ceased and transitional tectonics are active	Not given
Ewing (1980)	Pacific Northwest; Paleogene	Eocene right-lateral motion on major strike-slip faults, by grabens and reset terranes (e.g., metamorphic core complexes) suggest that a "large, concealed extensional area" exists beneath the Columbia Plateau	Not given	Transcurrent fault system resulting from oblique subduction or an on-land transform margin. The second model is preferred	Not given
Livaccari (1979)	Western United States; late Cenozoic	Northwest-trending dextral faulting in Oregon, California, and Nevada; no similar faulting in British Columbia. Dextral faulting minimal in Washington	Not given	San Andreas transform is not buttressed to north, so dextral slip occurs. Queen Charlotte transform is buttressed, so no slip occurs. No transform exists to west of Washington, so driving force for dextral shear is absent in Washington	Not given
Magill and others (1982)	Southern Washington, 12 m.y. ago to present	Pomona Member basalts rotated clockwise ~16° to the west of Cascades, but are essentially undisturbed on the Columbia Plateau	Not given	Southwest Washington rotated as a rigid block, while Columbia Plateau remained unrotated but possibly sheared on northwest-trending faults. A distinct zone of unspecified nature (the Cascades) separates the tectonic domains. Cause of rotation is possibly shear on North American plate imposed by the Pacific plate, or unbuttressed sliding off of the Great Basin asthenospheric high (to the north this would cause north-south compression of Columbia Plateau)	Not given

TABLE 7-1. Tectonic Models of the Cordillera with Emphasis on the Pacific Northwest. (Sheet 4 of 5)

Source	Area modeled; time	Geologic observations or tectonic relationships	Rate of deformation	Tectonic pattern or driving force	Prediction for future
Sbar (1982)	Western North America; present	Six seismotectonic domains delineated on basis of crustal stress and regional seismicity. Pacific Northwest domain has stress field similar to San Andreas, east-central California, and Mendocino Triple Junction domains	Not given	Four domains listed are generated by shear between Pacific and North American plates. Transform connection between San Andreas and Queen Charlotte Island fault systems may be initiating	Not given
Wise (1963)	North American Cordillera; Paleozoic to present	East-west extension and north-south compression for Columbia Plateau. Regional northwest-trending dextral shear	Not given	Not given	Not given
		Mantle Diapir and Interplate Shear			
Smith (1977, 1978)	Western North America; Cenozoic	Western United States is composed of subplates, each with its own motion in overall regime of northwest-southeast extension. General northwest movement of all subplates is accomplished by right-lateral oblique shear and extensional faulting. North-south compression in Columbia Plateau is possibly due to buttressing against the rigid North American plate to the north	Not given	Obliquely convergent plate interaction, extension over a mantle diapir, and stress relaxation following end of subduction account for features observed	Not given
Barrash and Venkatakrishnan (1982)	Western North America; late Cenozoic	Tectonic and magmatic events began, changed intensity, or ceased at ~16, 10, and 5 m.y. ago. Magmatism at 16 m.y. ago, shift of maximum compressive stress 25° clockwise to north-south at 10 m.y. ago, and cessation of folding at 5 m.y. ago affected the Columbia Plateau	Not given	16-m.y. events were related to mantle diapir; 10-m.y. events were related to rotation of Pacific plate spreading direction; 5-m.y. events were related to oceanic plate reorganizations. Farallon plate was uncoupled since 16 m.y. ago, so continental tectonic forces were derived from Pacific plate	Increasing volcanism in Cascades and Yellowstone may be precursors of transition in Pacific Northwest tectonic regime
		Microplate Rotations			
Bates and others (1981)	Western Washington; Oligocene to present	Paleomagnetic data for Washington Coast Range and southern Cascades show similar clockwise rotations of ~34°	1.5°/m.y. from early Eocene to early Miocene	Oregon and Washington Coast Ranges rotated together since late Eocene, owing to no specific driving mechanism	Not given
Hammond (1979)	Pacific Northwest; Cenozoic	Coast Range and Cascade arc blocks rotated clockwise from Olympic-Wallowa lineament about Olympic Peninsula center. Blue-Ochoco Mountains dragged behind along northwest-trending dextral faults. Mesozoic rocks below Columbia Plateau are thinned tectonically. Yakima folds superimposed on northern Cascades are from 12 to 5 m.y. (This is essentially the second model of Simpson and Cox, 1977)	Rotation apparently ceased by 20 m.y. ago	Rotation of subduction zone (Juan de Fuca) and related arc and back arc extention remain unexplained	Not given

TABLE 7-1. Tectonic Models of the Cordillera with Emphasis on the Pacific Northwest. (Sheet 5 of 5)

Reference	Area modeled; time	Geologic observations or tectonic relationships	Rate of deformation	Tectonic pattern or driving force	Prediction for future
Heptonstall (1977)	Western Cordillera; middle Tertiary (30 m.y. ago) to present	Three oroclines between Vancouver, British Columbia, and northern California, including one located near the Oregon-Washington-Idaho border disrupt the linearity of the Cordilleran orogenic belt	14° rotation from 29 to 4 m.y. ago; inactive after 4 m.y. ago	Cordillera north of Vancouver, British Columbia remained fixed as Cordillera south of the Klamath Mountains moved northwest with the Pacific plate. Two segments linked by Oregon-Washington-Idaho orocline acted as fragmented segments linking the major plates	Not given
Magill and others (1981)	Oregon Coast Range; 44 m.y. ago to present	Paleomagnetic data show clockwise rotations of ~46° of the entire Oregon Coast Range block. The Washington coast is characterized by smaller discrete blocks of variable clockwise rotation	Two phases: Eocene, 40° to 50°/5 to 6 m.y. (6°/m.y.) and Miocene to present, 30°/20 m.y. (1 1/2°/m.y.)	Fragmentation of Farallon plate results in Eocene rotation and accretion of Coast Ranges to North America. Eocene phase involves southern pivot linked to northern Klamath Mountains. Initiation of Basin-Range extension caused Miocene phase rotation about northern pivot. Northwest-trending strike-slip faulting and microplate adjustments cause north-south compressive strain in relatively stable Columbia Plateau of Washington	Not given
Simpson and Cox (1977)	Pacific Northwest; Eocene to present	Rotation of Oregon Coast Range clockwise into alignment with present coastline with either southern or northern pivot. Northern pivot requires rifting and extension in back arc regime east of Coast Ranges, along Olympic-Wallowa lineament, with extension increasing southward	50° to 75° of rotation since Eocene; 28° since Oligocene	Rotating oceanic plates during subduction or back arc extension causes rotation of Coast Range block	Rotation may continue today in response to differential extention in back arc
Watkins and Baksi (1974)	Oregon, Washington, Idaho; 16 to 13 m.y. ago	Paleomagnetic data show an orocline extending across southeastern Washington, western Idaho, and eastern Oregon. Magnetic poles for contemporaneous lava sections indicate a clockwise rotation of at least 15° of the southern sections relative to the northern sections. Compression in central Washington and Oregon is required	Average rotation rate of 1°/m.y.	Not given	Not given

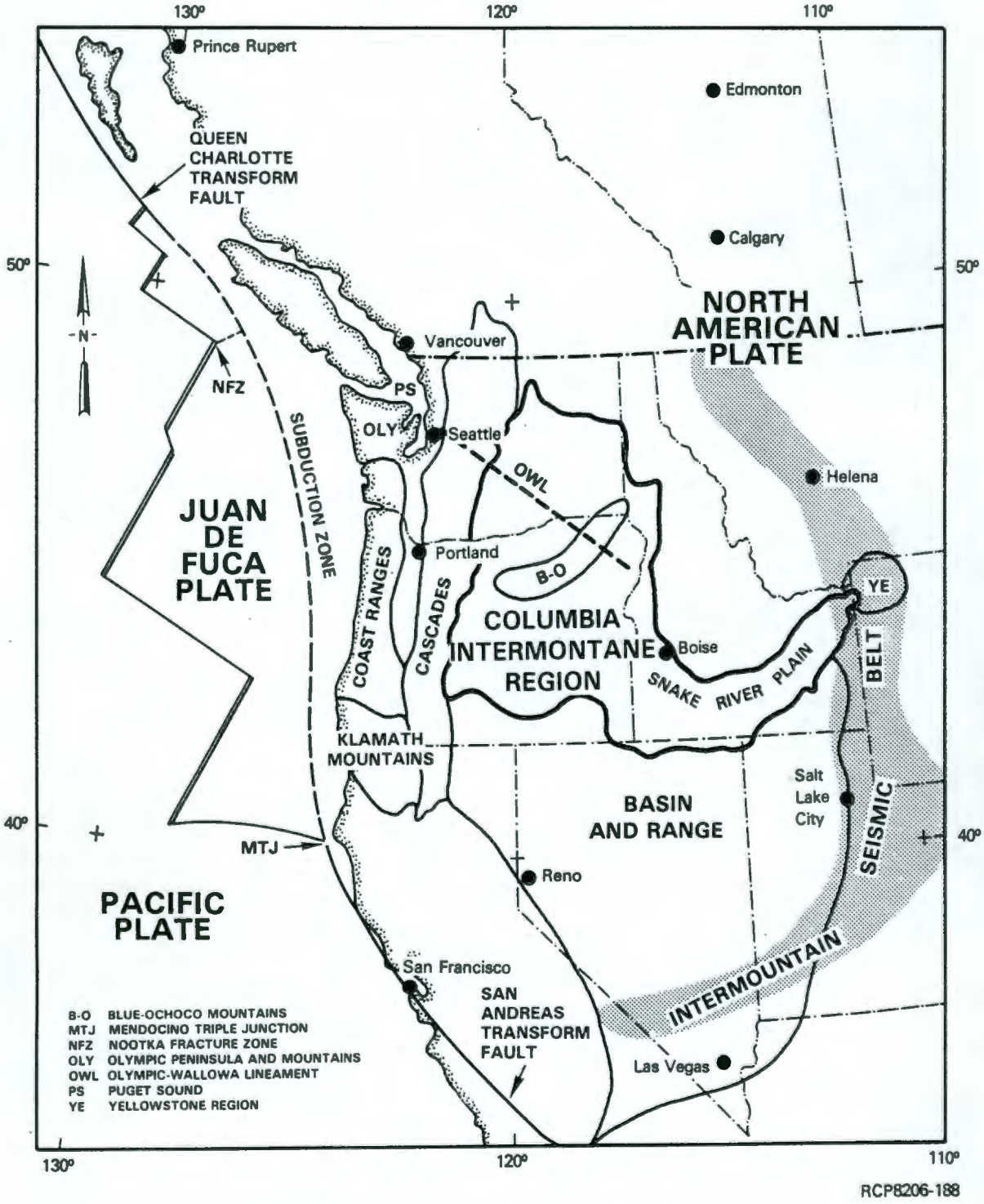


FIGURE 7-1. Location Map for Geologic and Geographic Features Referred to in Table 7-1.

The first class of models is in agreement that subduction is presently active off the Washington coast. A diverse set of observations is cited as evidence that subduction of the Juan de Fuca plate beneath the North America plate presently continues, although largely aseismically. The contemporary seismicity data of Crosson (1979) neither support nor reject ongoing subduction, but indicate that slip of a downgoing slab, if it occurs, is limited to a seismically quiet zone. Crosson interprets an east-northeast-dipping zone of seismicity below the quiet zone as stress release within a subducted slab. Some of the models in the remaining six classes address subduction in the context of other regional tectonic features and are not included in this class.

The second class of models explains Basin and Range faulting and Columbia River basaltic magmatism as a result of the subducted Farallon-Pacific spreading center beneath North American continental lithosphere. Most other authors reviewed here accept the theory that spreading centers are destroyed rather than subducted intact at consumptive plate boundaries.

The third class of models holds that crustal extension and Columbia Plateau magmatism is best explained by the development of a continental, back arc environment. Asthenospheric convection behind the Cascade volcanic arc is related to subduction (at ~40 mm/yr) of the Farallon plate and its remnant, the Juan de Fuca plate.

The fourth class of models calls for a mantle plume, diapir, or hot spot under the Columbia Plateau, the Great Basin, or Yellowstone. These types of thermal disruption caused rifting, crustal thinning, basaltic magmatism, and rotations of crustal blocks within plates. Subduction of the Juan de Fuca plate is permitted, but not related to the mantle disturbances.

The fifth class of models postulates a major shear between the Pacific plate (and other oceanic plates) and the North American plate. This shear is a result of oblique subduction, transform motion, oblique rifting, partial plate coupling, or transitional tectonics between subduction and transform. Whatever the mechanism, the effect is dextral shear on northwest-trending zones, northwestward translation, clockwise rotation of crustal blocks, east-west extension, north-south compression, and basaltic magmatism.

The sixth class of models is essentially a combination of the mantle diapir and Pacific-North American plate shear models. An obliquely convergent plate margin coupled with a mantle diapir produce the basaltic magmatism and the north-south compression observed on the Columbia Plateau.

The seventh group of models shares the concept of microplate rotations within the Pacific Northwest. Extension between the Cascades and the Rocky Mountains is cited as either the cause or an effect of the rotating blocks to the west. Most evidence suggests clockwise rotations of the Washington and Oregon Coast Ranges, but one model calls for a counterclockwise rotation of much of Washington. Several models call

for rotation of the Blue Mountains and parts of the southeastern Columbia Plateau. The clockwise rotation models generally reflect rotations of the coastal blocks of ~1.5°/m.y. since the Miocene.

In summary, the tectonic models of the North American Cordillera during the Cenozoic portray the Pacific Northwest as a region disturbed by subduction, volcanic arc construction, extension in the back arc area, production of voluminous flood basalts, rotation of the Coast Ranges and central Oregon blocks, northwest-trending dextral faulting, and north-south shortening in central Washington. The effects of most of these major processes on the Columbia Plateau since cessation of basaltic magmatism (~6.5 m.y. ago) have been minor. Specifically, subduction is either inactive or proceeding aseismically with few effects east of the Cascades, and Cascade volcanism has minor geological impact on the Columbia Plateau. Extension of the back arc region apparently has not affected the Columbia Plateau in post-basalt times as compressional structures are dominant in the basalt strata. Rotation of the Coast Ranges may be continuing, but the Columbia Plateau has undergone little or no rotation (Magill and others, 1982; Beck, 1980). The only processes of regional tectonic significance that may affect the stability of the central Columbia Plateau are northwest-trending dextral faulting and north-south shortening. These processes and related features are discussed in detail in the section on Columbia Plateau tectonics, but the general picture for the western United States is one of waning subduction and increasing dextral transform motion. The change from subduction to transform is not expected to be noticeable in a 100,000-yr time period, but it does provide the basis for evaluating tectonic and structural models of the Columbia Plateau.

LOCAL TECTONIC MODELS

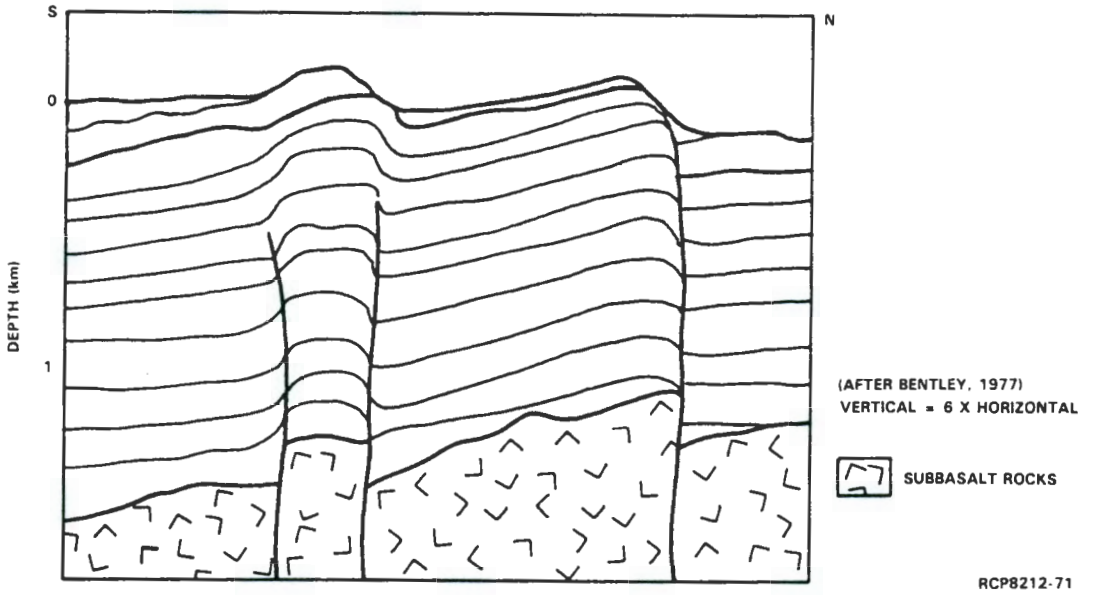
Explanations of the tectonics of the Columbia Plateau must consider the Yakima folds and wrench faulting of the western Columbia Plateau, the Blue Mountain structures of the southeastern Columbia Plateau, and the relatively minimal deformation of the Palouse Slope in the northeastern Columbia Plateau (see Chapters 2 and 5). Because the Hanford Site lies within the Yakima Fold Belt subprovince, the following discussion focuses on the Yakima folds in the context of the surrounding tectonic subprovinces. Within the Yakima fold subprovince, models must address two main structural features: generally east-west-trending folds and thrusts and northwest-trending strike-slip faults. Both features are consistent with the North American Cordilleran tectonics discussed in the previous section, but an adequate model must successfully explain the interrelationship between the folding and faulting. Further, the probable location and rate of present and future deformations, the extent of the basement (pre-basalt rock) involvement in deformation, and the timing of folding and faulting must be identified. Adequate models of the structural style and kinematics of plateau deformation, integrated with the data of previous chapters, will allow prediction of the stability of the reference repository location for the next 100,000 yr.

Bentley (1977) reported that the Yakima folds are faulted drape folds formed over rotated basement blocks (Fig. 7-2). Anticlinal deformation is concentrated above basement weakness zones during both horizontal compression and vertical movement, and the adjacent broad, synclinal basins subsided with the downward-rotating portions of the basement blocks. He conceived the Olympic-Wallowa lineament as a basement zone of weakness along which vertical movements have concentrated surface structures. "No evidence of strike-slip movement was found in any faults cutting the Miocene and Pliocene rocks" (Bentley, 1977, p. 343). The basic model of Bentley, then, is that basalt deformation is confined to narrow zones controlled by basement structures.

Ross (1978) documented the formation of east-west-trending folds and northeast-trending sinistral faults in the Blue Mountains region of northeast Oregon and southeast Washington. Folding was active since Grande Ronde time, but accelerated after Wanapum time, and was superimposed on regional westward tilting. In Ross' model, north-south compression formed the folds and faults while Columbia River basalt fissures formed along N. 5° W. to N. 20° W. zones of basement weakness (possibly related to Basin and Range structures). Dip-slip reactivation on the faults is explained as the result of relaxation of compression.

Hooper and Camp (1981) and Camp and Hooper (1981) expanded on Ross' work and proposed a model for the formation of structures in the southeast Columbia Plateau. It includes (1) a regional westward tilting owing to isostatic adjustments of the Idaho batholith, (2) continuous north-northwest-south-southeast horizontal compression and horizontal tension at right angles during basalt eruption, and (3) an older structural grain of northwest-southeast, north-south, and northeast-southwest faults. The Olympic-Wallowa lineament is explained as the border between different crustal rocks: older, thicker, continental rocks to the northeast and thinner, younger crust to the southwest. The difference in structural style recognized between the northeast and southwest portion of the plateau was attributed to this basic crustal contrast.

There is substantial evidence of northwest-trending dextral wrench faults of up to 30 km in length in the southwestern Columbia Plateau. Newcomb (1970) was one of the first to recognize the pattern of strike-slip faulting in this area. Lawrence (1976) interpreted four major northwest-trending dextral-slip zones from Landsat imagery. The zones separate the Columbia Plateau from the Basin and Range, but Lawrence considered the plateau to be largely unfaulted. Compilation maps by Swanson and others (1977, 1980, 1981) document the distribution of northwest-trending faults and aerial photograph lineaments throughout the western and southern Columbia Plateau. Mapping by Bentley and others (1980) on the Ahtanum, Toppenish, and Horse Heaven Hills structures of the western Columbia Plateau shows many northwest-trending fracture zones interpreted as strike-slip structures which have aggregate right-lateral displacement. Mapping by J. L. Anderson (Swanson and others, 1981) has revealed numerous parallel strike-slip faults in north-central Oregon and south-central Washington. The more numerous dextral faults strike N. 20° to 60° W. and the fewer sinistral faults strike N. 60° to 80° E. The sinistral faults generally parallel the axes of Yakima folds in the area.



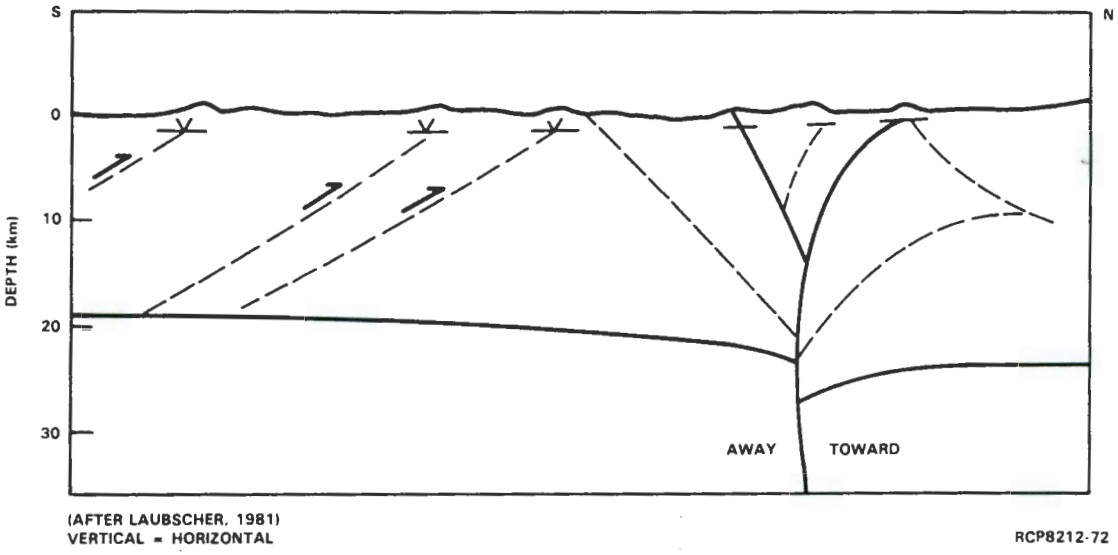


FIGURE 7-2. Interpretive Cross Sections of Columbia Plateau Tectonic Models: Bentley (1977) and Laubscher (1981).

Anderson considers all to be a part of a wrench fault system based on geometric similarity with known wrench fault zones. Lawrence (1979) reported consistent patterns of conjugate joint sets, east-west-trending folds, and northwest-trending faults based on imagery of the north-central Oregon region. The structural pattern is interpreted to be the result of north-south compression during the late Pliocene. The known extent of northwest-trending strike-slip faulting on the plateau is shown in the tectonic map (Plate 2-1). Recognition of the dextral faulting role has led to several new models of Columbia Plateau deformation.

Bentley (1980) proposed a wrench fault system composed of hundreds of parallel strike-slip faults distributing shear southwest of the Olympic-Wallowa lineament. With an average spacing of several kilometers and aggregate offset of 100 to 150 km, faults of the type Bentley proposed can be expected to be found anywhere on the plateau southwest of the Olympic-Wallowa lineament. The Olympic-Wallowa lineament is conceived as a zone of dextral motion abutting more rigid rocks to the northeast. This model is based on at least one reconnaissance traverse across part of the plateau and the data have not been published.

The structural analysis of Laubscher (1981), based primarily on a review of the literature, concluded that deformation of the Yakima Fold Belt subprovince is related to the motions of the "Yakima block" of the "Idaho-Olympic block belt." The Yakima block includes the Yakima folds and the Blue Mountains; the Blue Mountains are conceived as the stable hinterland of the Yakima folds to the north. The Idaho-Olympic block belt is a northwest-trending mosaic of crustal blocks bounding the Basin and Range Province. The Idaho-Olympic block mosaic and the Basin and Range are "the product of dextral-convergent interaction between the Pacific and North American plates" (Laubscher, 1981, p. 2.5 0-55). The Idaho-Olympic mosaic displays the effects of distributed dextral-convergent shear, north-south compression, and east-west extension as a result of the plate interaction.

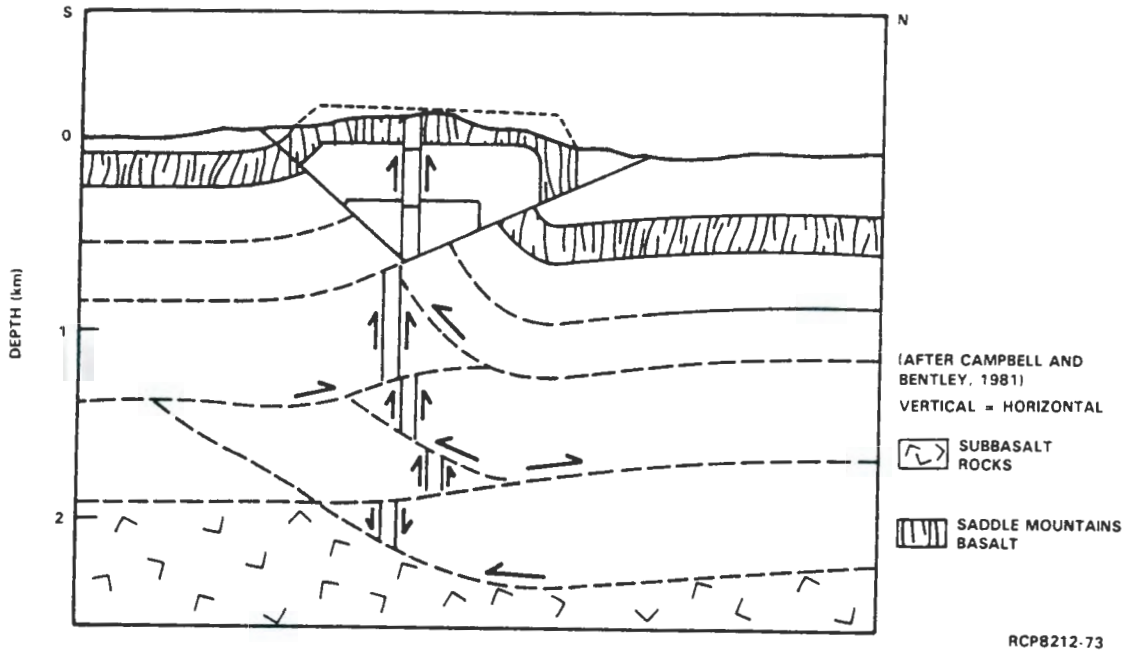
According to the model of Laubscher (1981) (see Fig. 7-2), the deformation of the Yakima block began along the Cle Elum-Wallula zone, which defines the northeastern edge of the block. The motion represents a major, Miocene, lithospheric break with dextral-convergent motion. At this time, gentle, periodic, en echelon brachyanticlines formed in the partially decoupled, superficial basalts along the zone (these are the Yakima anticlines within the Cle Elum-Wallula zone). After <2 km of movement, the Yakima block stopped and internal compression built up. Then, a décollement developed at the base of the Columbia Plateau crust (~20 km deep). From this base, further ramps developed, propagating northward, which extended to the base of the basalts. These upper ramps served as loci for shallow (1 to 3 km) detachments and Yakima anticline formation to the southwest of the Cle Elum-Wallula zone. The Yakima folds to the southwest formed with the same periodicity as the earlier en echelon Yakima anticlines of the Cle Elum-Wallula zone and may have propagated from them. Finally, the décollement cut across the Cle Elum-Wallula boundary and formed the eastern extensions of the Yakima folds (Gable Mountain, Gable Butte, Saddle Mountains, and Frenchman Hills).

Laubscher's model requires only 2 km of dextral motion on the Cle Elum-Wallula zone and 7 km across the entire Idaho-Olympic block mosaic. The idea of distributed shear within the Yakima block is rejected in favor of dextral motion along the block boundaries. Further, the model of Hammond (1979) (see Table 7-1), in which the Olympic-Wallowa lineament represents a major rift and crustal discontinuity, is rejected.

Bentley (1982) proposed that the Columbia Plateau was undergoing thin-skinned deformation under north-south compression while the dextral wrench system primarily affected rocks to the west of the plateau. In this model, based primarily on megascopic surface features, the Yakima folds reflect drag on ramps from sub-basalt and interbasalt décollements. The décollement surfaces can have variable slip directions; beneath Toppenish Ridge the basalt is depicted as faulting southward over the pre-basalt, while intrabasalt décollements have northward and southward slip directions (Campbell and Bentley, 1981; Fig. 7-3). The Cle Elum-Wallula segment of the Olympic-Wallowa lineament is conceived as a zone of "piling on" of the surface detachments.

Bruhn (1981) studied the mechanics of folding in Umtanum Ridge and Gable Mountain. He agrees with Price's (1982) conclusion that Umtanum Ridge represents a kink-type fold; however, the lack of bedding plane slip in steeply dipping beds on Gable Mountain supports dissimilar styles of folding on the two structures. Bruhn (1981) explains the long, gentle limbs on some of the Yakima folds by a fault ramp-flexure model with a décollement or group of localized detachments at 3 to 5 km deep, near the base of the basalts (Fig. 7-3).

Price (1982) proposed a model in which compressional strain is concentrated in the anticlinal ridges and is limited in the synclinal valleys. His mechanical model suggests that primary folding was followed by late reverse faulting required by the concentric geometry of the folds (Fig. 7-4). "The rocks below the fold must accommodate the surface shortening by...folding, faulting, or...ductile deformation" (Price, 1982, p. 115). Price argued that major thrusts cannot underlie the Yakima anticlinal folds owing to the lack of expected strain features related to folding over a ramp from a décollement. He concluded that the Yakima folds formed under the same north-south regional stress responsible for deformation elsewhere in the plateau; differences in fold orientations are explained by the clockwise rotation of basalt around a rigid basement buttress under the Palouse Slope (paleoslope of Chapter 5). Local dextral shear along the northwest-southeast-trending portions of the Rattlesnake and Horse Heaven Hills structures is required to accommodate the folding strain. Price's buttress model rules out a décollement within the rigid buttress, but permits a décollement at the base of the basalts under the Yakima folds. In summary, Price proposes that there is a basic rheological difference between sub-basalt rocks on either side of the Cle Elum-Wallula lineament. This contrast results in clockwise rotation and shortening of the basalt sequence under north-south compression. High-angle basement faults or ramps from a décollement are probably not connected to the fold-related thrust faults, but local detachments under the folds are permitted.



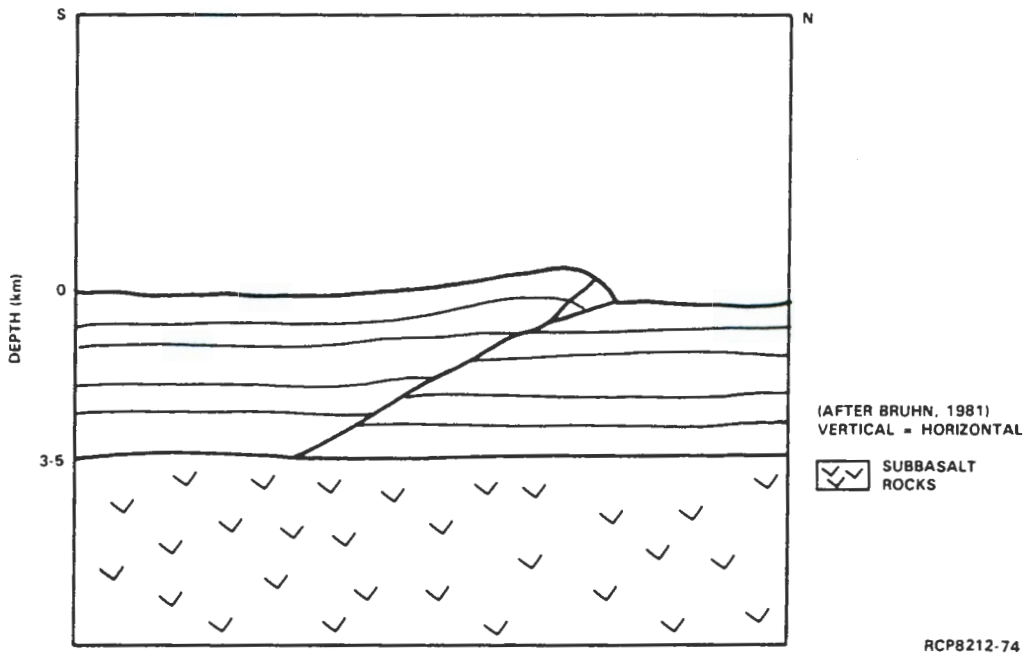
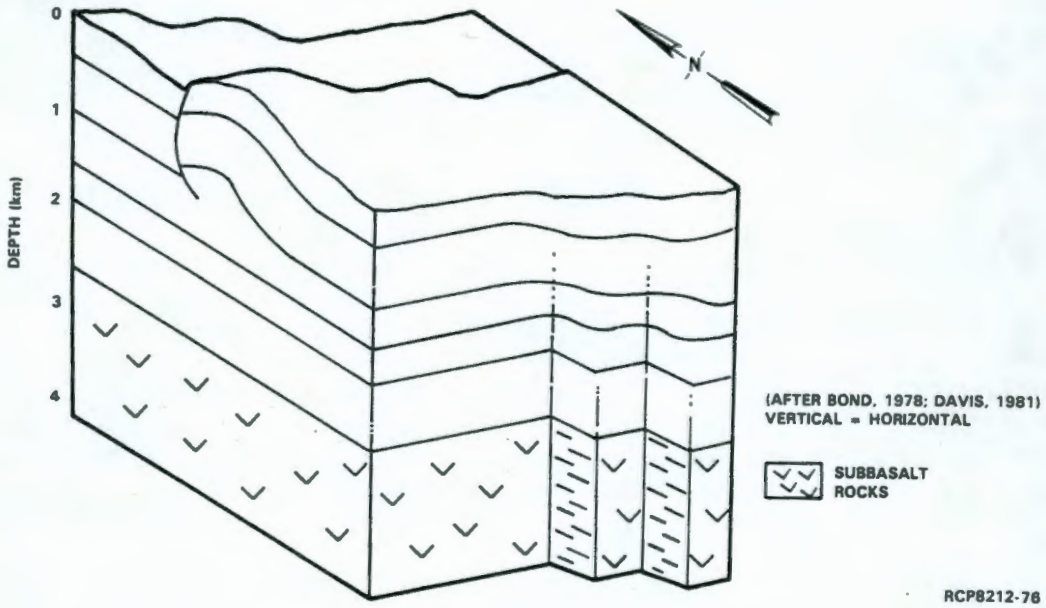


FIGURE 7-3. Interpretive Cross Sections of Columbia Plateau Tectonic Models: Campbell and Bentley (1981) and Bruhn (1981).



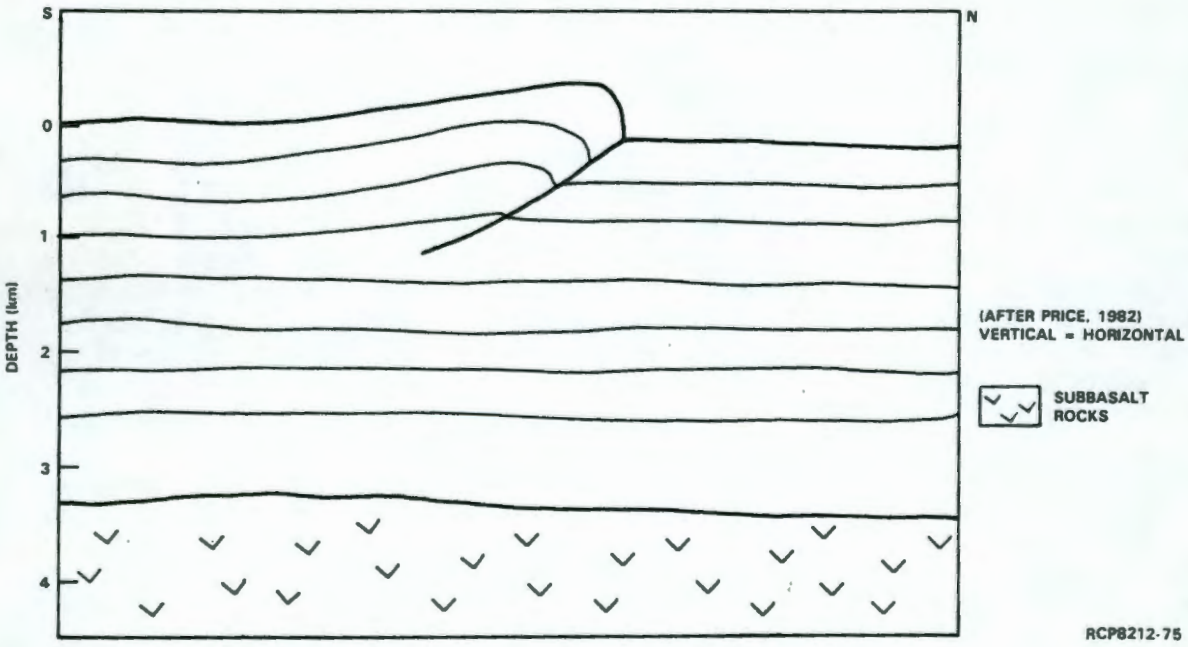


FIGURE 7-4. Interpretive Cross Sections of Columbia Plateau Tectonic Models: Bond (1978), Davis (1981), and Price (1982).

A summary of Columbia Plateau tectonics has been presented by Davis (1981). Davis' report includes discussions of the timing of plateau folding and the origin of plateau structures including the Cle Elum-Wallula lineament. Davis interpreted from a review of the literature that deformation has been active since late Grande Ronde time with most occurring after Pomona time. In the Blue Mountains subprovince, major deformation is interpreted to have ceased by 7 m.y. ago; Yakima fold deformation is reported to be continuing.

Davis' (1977) notion of the nature of the Olympic-Wallula lineament as a "fictional structural element" was amended by his feeling that the Cle Elum-Wallula segment is definitely a "disturbed structural zone" (Davis, 1981, p. 2.5 N-24). The Cle Elum-Wallula zone is conceived as a region of dextral movement along a deep fault or zone of anisotropy. North-south shortening produced the Yakima fold and fault structures by buckling above local detachments, while the Cle Elum-Wallula zone folds were the result of a complex interaction of a deep-seated fault and the shallow buckling and thrusting (see Fig. 7-4). Davis accepted Laubscher's (1981) concept that the underlying wrench fault is a Miocene structure rather than a reactivated pre-Miocene structure.

On the nature of basement involvement in the Yakima folds, Davis (1981) rejected Laubscher's (1981) concept of a regional, throughgoing detachment connecting each fold-associated thrust ramp. Instead, like Price (1982), he accepted the concept of local, shallow detachments required by the concentric nature of the folding. Davis (1981, p. 2.5 N-28) cited the small amount of shortening compared to the distance between structures and the alternating north-to-south vergence of some structures as evidence against folding over regional thrusts.

In summary, Davis (1981) accepted Laubscher's (1981) basic idea of the formation of Columbia Plateau structures within a broad belt of regional dextral shear, of which the Cle Elum-Wallula zone is a part. Most of the surface structures were interpreted to be the result of minor shortening on originally very flat and horizontal plateau strata. Deep-seated thrusts are not required below the Yakima folds, and basement involvement (probably without reactivation of basement structures) is postulated beneath the Cle Elum-Wallula structurally disturbed zone. All Yakima fold structures were reported to have developed synchronously.

Bond and others (1978), based on mapping in the Hanford area, postulated that the Yakima folds formed in two sets that responded differently, but contemporaneously, to a regional N. 20° W. compression. One set, including the folds from the Blue Mountains to Cowiche Mountain, trends about N. 70° E. The other set, extending from the Rattlesnake Hills northwest to the western Saddle Mountains, trends about N. 70° W. and lies within the Cle Elum-Wallula zone. The first set is considered to have formed perpendicular to the maximum regional stress and the second set to have formed in semirigid basalts above a basement dextral transcurrent zone corresponding to the Olympic-Wallowa lineament. The Olympic-Wallowa lineament is speculated to be a reactivated plate suture zone between oceanic and continental crust. The hypotheses of omnidirected compression from basin subsidence, gravity sliding from the Blue Mountains, drape folding over basement faults, and intersecting compressional fields are rejected.

Cowan (1981) reviewed the tectonic models of Price (1982), Laubscher (1981), and Davis (1981) with reference to the Umtanum Ridge-Gable Mountain structure. Cowan concluded from his review that Yakima fold deformation is best explained by buckling without underlying imbricate thrusts from a décollement. Further, he questioned the continuation of the Cle Elum-Wallula zone of structural disturbance beyond the inflection in Rattlesnake Hills. To the southeast of the inflection, structures can be explained by buckling without dextral slip along a fault or a sub-basalt discontinuity. Shortening within the basement is postulated to be accommodated by diffuse rather than discrete structures. Based on his review, Cowan generally accepts Price's (1982) model, but Laubscher's (1981) is rejected.

Barrash and others (in press) proposed a three-stage deformation of the Columbia Plateau based on a literature review. Stage one, between 17 and 10 m.y. ago, was characterized by the north-northwest-striking basalt dikes, mild warping across the plateau, formation of open folds in the Blue Mountains, and north-northwest-trending maximum stress. Stage two, between 10 and 4 m.y. ago was characterized by formation of most of the Columbia Plateau structural relief in the Yakima folds and Blue Mountains and a change in the maximum stress direction to north-south. Stage three, from 4 m.y. ago to the present, was characterized by a possible deformational hiatus followed by minor faulting, practically no folding, and a north-south maximum stress. An ancestral continental-oceanic crustal boundary is assumed to lie below the basalts, possibly beneath the Olympic-Wallowa lineament. Basement control of the orientation of the Yakima fold structures near the Olympic-Wallowa lineament and of the Chief Joseph dike swarm is also postulated. The three-stage development of plateau structures, with deformation culminating in stage two, is related to major plate tectonic events at the western edge of the North American plate (Barrash and Venkatakrishnan, 1982; see Table 7-1).

SYNTHESIS

The tectonic models summarized above are similar in that the major structural features of the Columbia Plateau were formed as a result of north-south compression. Most workers also agree that the majority of the strain in the basalts is represented by the anticlines and associated faults, but comparisons with basalt exposures within synclines have so far been limited. The Cle Elum-Wallula zone of deformation, a part of the Olympic-Wallowa lineament, is recognized as a structurally disturbed zone, which possibly reflects some degree of dextral movement. Workers in the southwestern part of the plateau generally accept the presence of the parallel, northwest-trending, dextral, wrench faults. Workers in the eastern and southeastern plateau support basement involvement and control of basalt structural features and a regional westward tilting of the structurally stable Palouse Slope block.

Areas of disagreement among models are more numerous and range from the minor to the critical. The areas of disagreement that have direct bearing on the stability of the reference repository location are discussed below. These areas are summarized as follows: (1) degree of sub-basalt structural involvement, (2) location of strain, and (3) rate of deformation. Other topics with secondary impact on repository stability, such as the degree of horizontal rotation of plateau rocks during deformation, are not discussed.

One major disagreement concerns the degree of sub-basalt involvement in basalt structures. In thin-skinned tectonic models, thrust faults, which branch upward and cause secondary folding, are connected to regional detachments (décollements); rocks below the décollement are undeformed or deform independently. Thin-skinned models generally preclude involvement of rocks below the décollement, but structures or buttresses in the sub-décollement rocks may affect the location of the décollement. In thick-skinned models, rocks are deformed without the formation of a regional detachment, but the style of deformation may vary with depth because of varying rheological properties. In thick-skinned deformation, either folding or faulting can be the primary strain element. Thus, the distinction between the two types of deformation can be vague.

Because of the difficulties in strictly defining thin- and thick-skinned deformation, the terms are not very useful in categorizing Columbia Plateau tectonic models. For example, Laubscher's (1981) thin-skinned model postulated a décollement 20 km deep connecting major thrusts below the Yakima anticlines; 20 km is well below the top of geoelectric layer 4 (see Chapter 4). In contrast, Bentley (1982) called for décollement within or at the base of geoelectric layer 2, the basalt sequence. Both models are thin-skinned, but one requires extensive sub-basalt rock involvement and the other requires none. To aid clarity, the remainder of this discussion of sub-basalt involvement will avoid the thin- and thick-skinned classification and focus on two questions:

- Does surface folding and faulting extend below the basalt?
- What is the relationship between folding and faulting?

Different models have different explanations of the subsurface extent of intrabasalt faults. The models of Laubscher (1981), Bentley (1977, 1982), Davis (1981), and Price (1982) support sub-basalt involvement in basalt deformation in the Cle Elum-Wallula zone. For the Yakima folds exclusive of the Cle Elum-Wallula zone, the models of Bentley (1982), Davis (1981), Bruhn (1981), and Price (1982) do not require sub-basalt involvement for the development of surface structures. Laubscher (1981) and Bentley (1977) extend sub-basalt involvement to Yakima folds in general. In contrast, Cowan's (1981) model does not support sub-basalt involvement in any of the Yakima structures. Hooper and Camp (1981) and Ross (1978) supported sub-basalt control of surface structures in southeastern Washington.

Owing to the thinness of the basalt observed on the plateau margins, most workers agree that structures in pre-basalt rocks influenced the deformation of the overlying basalt. The irregular pattern of plateau margin folding and the continuation of nonplateau structural trends onto the plateau (see Chapter 3) also support the basement control hypothesis. There is good evidence for major sub-basalt topography under the Pasco Basin (see Chapter 4), which is evidence against a major décollement at the base of the basalt sequence (unless block faulting later offset the décollement surface). The possible correlation of structures in the basalt and sub-basalt of Umtanum Ridge interpreted from geophysical studies (see Chapter 4) also argues for sub-basalt structural involvement. The pattern of folds in the central Columbia Plateau can be argued to be the result of folding over similarly spaced basement structures, and the deflection of fold trends in the Cle Elum-Wallula zone of deformation can be most easily explained by sub-basalt involvement. Seismic data show similar focal mechanism solutions for both shallow and deep events (see Chapter 6), indicating that contemporary deformation is similar in basalt and sub-basalt rocks without decoupling between them. Also, the large relief on geoelectric layers 3 and 4 is permissive of vertical block faults such as hypothesized by Bentley (1977). However, with current magnetotelluric station distribution and resolution, correlation of relief in layers 3 and 4 with surface folds is poor. Finally, the observation that the basalt sequence thins to the east onto a paleoslope and structural relief of surface structures and interpreted basement decreases eastward (see Chapter 4) may also be evidence for pre-basalt influence on basalt structures.

Independent deformation of basalt and sub-basalt is supported by the regularly spaced folds in much of the central Columbia Plateau. It may be possible to explain the regularity as the expected fold wavelength for a horizontally isotropic multilayer undergoing horizontal compression, although mechanical modeling has yet to be used to demonstrate this. Also, the lack of consistent correlation between surface structures and interpreted features in the subsurface geoelectric layers argues against a simple model of basalts deforming over similar sub-basalt structures. The Cle Elum-Wallula zone of deformation is most commonly thought to reflect a degree of sub-basalt structural control, but no corresponding feature was detected in the sub-basalt geoelectric layers (see Chapter 4). In summary, sub-basalt influence or involvement in basalt structures is problematic. The example of Snively Basin may be typical of the central Columbia Plateau. At Snively Basin, structural complexity in the basalts seems to coincide with geophysically interpreted complexity in the sub-basalt rocks without evidence for direct involvement between the two layers (see Chapter 4). Thus, some degree of sub-basalt involvement seems possible, but the nature of the interaction is not clear.

There is also disagreement about the second question: What is the relationship between folding and faulting? The models of Laubscher (1981), Bentley (1982), and Bruhn (1981) support the concept of anticlinal development over major thrust faults. In contrast, Price (1982), Cowan (1981), and Davis (1981) reported that major reverse faults developed after folding proceeded to a stage where no more shortening could be accommodated by folding. The evidence for both points of view is linked

to the interpretations of the tectonic models and so far remains inconclusive. However, if basement faults can be shown to control surface structures, then primary faulting with secondary folding of the surficial basalts would be strongly supported.

Determination of the extent of faulting and folding in the sub-basalt rocks under Yakima folds is needed to predict the style of future deformation. In particular, the predicted areal extent of the faults underlying the Yakima folds varies considerably depending on which model is chosen. At present, there is not enough information to choose one preferred model. Performance of a mechanical analysis of the basalt and sub-basalt rocks under regional, horizontal compression may be able to determine the theoretical behavior of the basalts. It may be that the basalt sequence is too weak to buckle at the surface independently of sub-basalt warping, or it is possible that the sub-basalt and deeper basalt may deform passively while faulting and flexural folding deforms the surface basalt. Also, the expected fold wavelength could be calculated and compared to the observed wavelength; a close agreement would argue for no sub-basalt involvement. Further structural analyses coupled with geophysical modeling may help resolve the disagreement.

The second major area of disagreement is the location of strain. Most workers agree that the Yakima folds have concentrated the majority of strain in anticlinal buckles and reverse faults. Price (1982), for example, argues that the synclines are practically undeformed. However, Bentley (1980) has postulated pervasive distribution of strike-slip faulting south of the Olympic-Wallowa lineament. Owing to the lack of documentation for dextral faults every kilometer or so across the Pasco Basin, it is difficult to evaluate Bentley's postulated faults as a problem for the stability of the reference repository location. However, focal mechanism solutions suggest that contemporary deformation in the central Columbia Plateau is by thrusting on east-west-striking planes, with little evidence for strike-slip movement on northwest-southeast-striking planes (see Chapter 6).

The problem of predicting future strain involving the east-west thrusts and reverse faults is more tractable because the issue of basement involvement in surface folding does not affect predictions for future locations of strain. If master basement structures such as fault ramps, décollements, or warps are responsible for Yakima folding and faulting, then those structures or zones of weakness will remain the most probable sites for localizing future strain. On the other hand, if the basalt deformed independently of the rocks beneath the basalt, then that style of deformation can be expected to continue. In the second case, it is well known that folds in initially horizontally layered strata, once formed, will localize future strain (Johnson, 1977). The reasoning is that the horizontal stress necessary to deform multilayered strata is much less for slightly dipping layers than for perfectly horizontal layers. In the case of the Columbia Plateau, the expected response to continuing north-south, horizontal compression is continuing deformation of the more steeply dipping anticlinal strata and relatively little deformation of the nearly horizontal synclinal strata. This does not preclude the possibility that

the first-order (and lower-order) anticlinal structures could lock up and that stresses could be localized elsewhere resulting in the development of a new first-order anticline in an area between existing anticlines. However, in the central Columbia Plateau, stratigraphic data suggest synchronous growth of several first-order folds during Wanapum and Saddle Mountains (and possibly Grande Ronde) time. New first-order folds do not appear to have developed in a "second round" of folding as suggested by the distribution of basalt flows and the continuity and attitude of Ringold strata in the northern part of the Cold Creek syncline (see Chapter 5). Therefore, while the possibility of development of a new first-order fold exists, the likelihood of growth of a new first-order anticlinal structure in the next 10,000 to 100,000 yr is small, based on the geologic record of fold growth in the last 15 m.y. in the central Columbia Plateau.

A final area of disagreement among models is the rate of deformation. The predictions for future rates of structural growth from work performed are presented in Chapters 5 and 6 and are summarized in Chapter 8. These rates are in disagreement with those reported by Barrash and others (in press). Barrash and others (in press), as discussed earlier in this chapter, interpret a major increase in deformation rate at 10 $\pm$ 2 m.y. ago, which continued until 4 m.y. ago. Before 10 $\pm$ 2 m.y. ago and after 4 m.y. ago, deformation rates are reportedly much lower. In contrast, the data in Chapter 5 show high deformation rates beginning in late Grande Ronde time (before 14.5 m.y. ago) and declining rates in Wanapum time. Deformation rates were lowest from Saddle Mountains time (~13.6 m.y. ago) to the present (see Chapter 6). The data of Chapter 6 show that the contemporary rate of deformation appears comparable to that estimated for the period after Wanapum time, and the hypothesis of Barrash and others (in press) of high strain rates in the Pasco Basin from 10 to 4 m.y. ago is not supported.

SUMMARY

The tectonic development of the Pacific Northwest since at least the early Miocene, if not earlier, is characterized by volcanic arc construction and extension in the back arc region, clockwise rotation of the Coast Ranges and central Oregon, voluminous basaltic magmatism, and north-south shortening and possibly rotation in central Washington. Within the Yakima Fold Belt subprovince of the Columbia Plateau, the dominant structural features are the Yakima folds and faults and northwest-trending dextral faults. The Yakima folds and related faults have absorbed regional north-south shortening and possibly rotation of the basalts, but the extent of basement involvement in the near-surface structures is not yet resolved. The Cle Elum-Wallula zone seems to reflect some dextral shearing and compression with involvement of sub-basalt rocks; the extent of strike-slip motion is not thought to be large. The northwest-trending faults common in the southwestern part of the plateau are recognized as a wrench fault system, but the hypothesis of many parallel wrench faults distributing strain within the Pasco Basin is presently unsupported. The Palouse

Slope appears to have been relatively undeformed except for regional westward tilting and may have acted as a buttress against which Yakima Fold Belt subprovince structures formed. Finally, the models suggest that strain within the Yakima Fold Belt subprovince is concentrated near the anticlines whereas the synclines are relatively undeformed. Tectonic conditions prevailing over the past 15 m.y. are expected to continue over at least the next 1 m.y.

CHAPTER 8 - A PRELIMINARY ASSESSMENT OF THE TECTONIC STABILITY OF THE REFERENCE REPOSITORY LOCATION

J. A. Caggiano

INTRODUCTION

Both folding and faulting have occurred in the Pasco Basin area over the last 15 m.y. The rate and mechanics of deformation are a function of tectonic processes and the rheological properties of the rocks. Preliminary data and interpretations suggest a general chronology and a long-term average rate of deformation.

Deformation in the central Columbia Plateau may be episodic, continuous, or some combination of the two. Folding apparently results from continuous, long-term compressive strain, perhaps with some slip along layer boundaries. Depending on the mechanism of faulting, displacement along faults may occur as continuous aseismic fault creep, as episodes of discrete seismic slip, or some combination of the two.

Predictions of potential future vibratory ground motion and displacement during the period of operation and isolation of a repository are functions of whether folding or faulting predominates, the distance of geologic structures from a repository, and the rate at which each process occurs. If faulting predominates, predictions of possible future vibratory ground are contingent on determining whether displacement along faults is by continuous fault creep or by discrete episodes of seismic slip (stick-slip) separated by intervals when the juxtaposed rocks are locked and strain energy is accumulating. If faulting in the Pasco Basin area occurs by the stick-slip mechanism, the amount of displacement per episode of slip, its return period, and the size of any accompanying earthquakes should be assessed. If folding predominates over faulting, deformation is likely to be mostly aseismic with minimal displacement. The detailed mechanics of the deformation process remain to be determined during detailed site characterization.

THE PATTERN AND RATE OF TECTONIC DEFORMATION

Available structural and tectonic data permit a preliminary interpretation of the pattern and chronology of deformation in the Pasco Basin and reference repository location. Deformation of basalt was in progress during late Grande Ronde time (~14.5 m.y.B.P.) as evidenced by the thickness and distribution of flows (see Chapter 5). Uplift on Miocene topography that is presumably of structural origin proceeded at low average rates of between 40 to 80 m/m.y. (0.04 and 0.08 mm/yr) during the period 14.5 to 10.5 m.y. ago and this can be interpreted as the vertical component of strain. If these low average rates of uplift are projected to the present, they account for the presently observed structural relief and the elevation of flows on Rattlesnake Hills and the Saddle Mountains.

Subsidence in the Cold Creek syncline in the central Pasco Basin relative to the projected Palouse slope of southeastern Washington proceeded at similar rates (see Chapter 5). The decreasing dip of progressively younger strata of the Miocene-Pliocene Ringold Formation on the flanks of anticlines similarly suggests that deformation at relatively low rates of vertical strain was continuous during Neogene and Quaternary time and resulted in greater deformation of older units. Contemporary deformation appears to be continuing at these long-term, low average rates as evidenced by the level and distribution of earthquakes (see Chapter 6) and the measured shortening of lines of a trilateration survey over the period of 1972 to 1981 (see Chapter 6). Horizontal strain indicated by six geodetic surveys is of the order of 0.02 to 0.04 mm/km/yr compression.

Geologic (see Chapters 3 and 5) and seismologic (see Chapter 6) data indicate that the axes of crustal shortening are oriented generally north-south and are nearly horizontal. The nearly east-west strike of anticlinal and synclinal axes and accompanying subparallel thrust and reverse faults suggest nearly north-south, nearly horizontal compression as do the steeply dipping northwest (and some northeast) trending faults. Focal mechanism solutions (for individual events and composites of several presumably related events) suggest that the axis of maximum compression is nearly horizontal and oriented slightly west of north. The axis of minimum compression from fault plane solutions is nearly vertical. Focal planes trend generally east-west, dip steeply, and result from essentially reverse faulting, with little, if any, strike slip component (see Chapter 6). This pattern holds for shallow as well as deeper events (>6 km). Two notable exceptions are focal mechanisms for the July 15, 1936 and April 17, 1979 earthquakes along the margin of the Blue Mountains >50 km east of the Pasco Basin. For these events, the axis of maximum compression is nearly west-east and slip was probably on northeast-trending planes of rupture. There were no foreshocks or aftershocks for the 1979 magnitude 4.1 event (Malone, 1979), but aftershocks for the 1936 magnitude 5.8 event align in a generally north-northeast to northeast direction approximately parallel to faults in the Hite fault system. Measurement of in situ stress by hydraulic fracturing near the reference repository location in the Umtanum flow resulted in a mean (of six measurements) principal compression of N. 23° E. and a near vertical axis of least compression (Kim and Haimson, 1982). Geodetic data also suggest nearly horizontal compression, but directed from the west (see Chapter 6).

Deformation (i.e., vertical strain) in the central Columbia Plateau appears to have established zones of weakness (first-order structures) relatively early. Subsequent deformation appears to have occurred along these established first-order structures with subsequent development of lower-order structures in relation to first-order structures, but with no obvious development of very geologically youthful structures. Deformation appears to have been in progress at least 14.5 m.y. ago and appears to have continued, at least in some structures, to the late Pleistocene or Holocene.

RELATIONSHIP OF STRUCTURES IN BASALT TO SUB-BASALT STRUCTURE

The assumed structural relief of the top of geoelectric layers beneath the basalt in the Pasco Basin, as presently interpreted (see Chapter 4), only suggests several possible relationships of basalt and sub-basalt structures. A basement high beneath Umtanum Ridge suggests continuity of that structure with depth. Basement complexity beneath the structurally complex Snively Basin suggests a possible relationship of basement and surface structure. However, other fold ridges exhibit no obvious relationship to basement structure as it is presently interpreted.

Northwest-, north-south-, and northeast-trending structures in pre-Miocene rocks beyond the margins of the Columbia Plateau (see Chapter 3) would be expected to relieve stress under north-south compression. These zones of weakness should be areas where stress is relieved as earthquakes under a stress regime of north-south compression. Structures of similar trend may be present beneath the basalt in the central Columbia Plateau, but they are not expressed as zones where stress is preferentially relieved as earthquakes--including the Cle Elum-Wallula zone of deformation, which has been hypothesized as playing a key role in deformation of the Yakima folds. (See discussion and maps of earthquakes in Chapter 6.)

Shell Oil Company has drilled two exploration holes along folded ridges in the basalt in the hope of discovering commercial quantities of hydrocarbons in closed structures in presumably sedimentary rocks below the basalt. No success has been reported for either the Roza Dam (Yakima Ridge) or Whiskey Dick Mountain exploratory holes.

Focal mechanism solutions for the central Columbia Plateau (see Chapter 6) suggest that both the basalt and rocks below the basalt are responding to the same nearly north-south, nearly horizontal compression. The absence of any detectable relationship between earthquakes and mapped geologic faults and the absence of any clear alignment of hypocenters in rocks below basalt suggest that stress is not being relieved along major structural zones of weakness below the basalt. Instrumental earthquake data do not suggest the presence of active structures beneath the basalt.

Differences in the pattern of seismicity, recurrence, and cumulative strain between basalt and rocks below the basalt suggest that basalt is deforming in a brittle manner, while the response of rocks below basalt to the same stress field is more ductile and possibly aseismic (see Chapter 6). The contrast between the rheological properties and environment of deformation of the shallow, anisotropic layered basalt and rocks beneath basalt may account for this observed behavior.

In summary, rock type and structure that exist below basalt can only be hypothesized at present based on current interpretations of geophysical surveys and the gross geometrical relationships observed in structures of the Yakima Fold Belt. Major structures beneath the basalt that would relieve stress under north-south compression do not appear to be present

as judged by the pattern of distribution of instrumentally recorded earthquakes that have occurred below basalt. The structural relief of geoelectric layers below basalt interpreted from geophysical data suggests that major detachment surfaces (décollements) are not present at the contact of units of major resistivity contrast; however, detachments within such units or local detachments beneath anticlinal folds can not be eliminated as possible structures at this time.

FUTURE DEFORMATION IN THE CENTRAL COLUMBIA PLATEAU

Columbia River basalt in the central Columbia Plateau has apparently been deforming over the past 14.5 m.y. under nearly horizontal, nearly north-south compression at a long-term average low rate and should continue to do so over the next 10 to 100 millenia. There are no known geologic reasons to suggest that this pattern of deformation will change during the period of repository operation and waste isolation. The location and amount of deformation predicted for the Pasco Basin area are a function of the mechanics of deformation, which have been explained somewhat differently in several proposed tectonic models (see Chapter 7). Regardless of the detailed mechanics of deformation that remain to be established, the long-term rate of deformation is not anticipated to change as suggested by the geologic data from the past 14.5 m.y.

The north-south, nearly horizontal compression is expected to continue and result in deformation on structures with trends compatible with this stress field. Folding in the central Columbia Plateau is anticipated to continue on existing and principally east-west trends. Reverse slip on thrust and reverse faults that are subparallel to the axes of these folds should continue and is suggested by focal mechanism solutions of earthquakes (Chapter 6). Dextral slip on northwest-trending and sinistral slip on northeast-trending, steeply dipping faults is compatible with north-south compression in the hypothesized wrench tectonic regime and may occur, although focal mechanism solutions indicate a minimal strike-slip component during earthquakes in the Pasco Basin area. Dextral slip on north-south to northwest-trending faults in the western Columbia Plateau and in the Cascades is indicated by recent focal mechanism solutions. Strike-slip faulting on northwest or northeast-trending focal planes is also suggested for the July 1936 and April 1979 events along the front of the Blue Mountains some 50 or more kilometers east of the Pasco Basin.

Whether deformation by folding or faulting will predominate is currently being investigated. Price (1982) indicates that folding strain should be concentrated in areas of steeply dipping strata on the limbs of anticlinal folds, with some slip between layers. In this model, faults develop after the folds begin to "lock up." Deformation in the synclines should be minimal according to Price (1982), but some strain can be anticipated in any dipping strata.

Deformation, by whatever mechanical processes, is anticipated to continue at a long-term average rate of <1 mm/yr and be concentrated on existing first- or second-order structures throughout the basin.

Available data can not preclude development of new first-order structures, but the probability of their development is considered very low in view of the known record of deformation in the last 15 m.y. Slip rates on faults in the central Columbia Plateau are mostly unknown at present, but are anticipted to be low in view of the pattern and level of seismicity and the basin-wide low rate of strain suggested by geologic and geodetic data. The low rate of slip determined for the central fault on Gable Mountain (0.005 mm/yr from decreasing displacement of the basalt contact of three geologic units from 12.5 m.y. to ~13,000 yr ago) supports this assessment.

The size and depth of instrumentally recorded earthquakes in the Pasco Basin have been used to estimate seismic moments of small events, which in turn suggest minimal displacements on small faults (Caggiano, 1982). Analysis of local digitally recorded seismic data is expected to provide a local moment-magnitude relationship that will allow further evaluation of the rupture area of faults and displacement per event. The size of commonly recorded, local earthquakes suggests that strain does not build to high thresholds before rupture occurs on short (≤5 km), probably shallow (≤3 km) faults producing small earthquakes (generally less than magnitude 3.0). Determination of local moment-magnitude and recurrence relationships from a locally deployed network of seismometers will contribute to an understanding of strain and failure processes occurring in Columbia River basalt in the Pasco Basin and reference repository location.

While it is anticipated that future strain will be concentrated on existing first- and second-order structures that developed millions of years ago, it is not possible to predict on which structures any deformation will be localized. The displacement per event and its return period for faults also remain to be identified. Ongoing deformation is indicated by small earthquakes, but these are not associated with specific mapped faults nor do they align in a manner suggesting the presence of buried, seismogenic faults.

PRELIMINARY ASSESSMENT OF THE TECTONIC STABILITY OF THE REFERENCE REPOSITORY LOCATION

Based on the trends of folds and faults and their chronological development, the seismicity, and the rate of deformation summarized above, the reference repository location appears to comply with the proposed draft criteria for tectonic stability (NRC, 1981). The reference repository location is in a volume of rock that presently appears free of major structures and occurs in the Pasco Basin of the central Columbia Plateau where a low, average rate of strain seems to have been operative for at least the past 14.5 m.y. Several uncertainties in this preliminary assessment are currenly being studied, including:

- Detailed mechanics and timing of deformation in Yakima folds
- Continuity and role of the Cle Elum-Wallula zone of deformation in development of Yakima folds

- Continuity of surface structures with depth and their relationship, if any, to structures below basalt

- Location and characteristics of earthquake sources, their relationship to geologic structure, and ground motion from maximum probable events.

Although these uncertainties remain to be resolved, data and completed studies allow preliminary intepretations that do not appear to preclude development of a nuclear waste repository in the reference repository location at the Hanford Site. The reference repository location in the Cold Creek syncline contains relatively undeformed strata in an environment where bounding anticlines apparently have been deforming slowly at long-term average rates of strain. This pattern and rate of deformation has apparently been in continuous operation for at least 14 m.y. and can be assumed to continue over the next few thousand to tens of thousand years. Major first- and second-order structures have been identified, and it is assumed that future strain will be concentrated on these structures; however, the detailed mechanics of the deformation process remain to be determined.

While deformation may be more episodic than indicated in this preliminary assessment, the long-term, low average strain rate that has been operative in the central Columbia Plateau since at least late Grande Ronde time is at least an order of magnitude less than that occurring in active orogenic areas. Because no areas on earth are static and unchanging, those areas where diverse evidence suggests deformation at low rates of strain for a long period are "tectonically stable" and appear to meet the proposed criteria for tectonic stability.

CHAPTER 9 - REFERENCES

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APPENDIX A - GEOPHYSICAL METHODOLOGY AND OBSERVATIONS

METHODOLOGY

Understanding the principles of aeromagnetic, gravity, and magnetotelluric methods is essential to properly comprehend the geologic implications of their interpretations. Geophysical techniques detect discontinuities in rock properties such as magnetic, density, or electrical conductivity. These discontinuities in rock properties are not always easily or directly correlatable to geologic structure and stratigraphy. Several assumptions and approximations used in geophysical methods place additional limitations on the data and interpretations presented. But, by combining the results of several geophysical methods with the known geology of the study area, anomalies may generally be assessed for their impact on subsurface configuration. The magnetotelluric method is discussed here in more detail than gravity and aeromagnetics because it generally is the least understood of these three methods.

The magnetotelluric method determines the electrical conductivity of the subsurface based on the measurement of natural, varying electromagnetic fields within the Earth (Vozoff, 1972). The magnetotelluric method depends on the deep penetration of electromagnetic energy into the Earth where the depth of penetration is inversely proportional to the frequency and directly proportional to the resistivity of the underlying strata.

Data are obtained by simultaneously measuring the orthogonal magnetic and electrical fields at the Earth's surface over a wide frequency range (e.g., 0.001 to 200 Hz). From these data, the magnetotelluric surface impedence as a complex function of frequency is determined. These frequency-domain-impedance functions are uniquely derived from field measurements at each site and are the basis for an electrical interpretation. The electrical strike, which is the direction of transitional invariance or strike along which the electrical conductivity is essentially constant, is also derived from these functions (Gamble and others, 1982). In practice though, some uncertainties in the geoelectric configuration do exist due to data accuracy, which is influenced by recording noise, near-surface geologic distortions, and finite sampling.

Depths to the geoelectric boundaries, observed in the magnetotelluric data, were picked by the Bostick (1977) one-dimensional inversion process, which is subject to interpretation and adjustment based on two-dimensional model studies. The one-dimensional inversion of the parallel electric component of the magnetotelluric tensor impedance function is normally used for an estimate of the resistivity-depth profile at each site. For any geometry that is not one-dimensional, this method represents an approximate solution, and the accuracy of the approximation is dependent on the dimensional complexity of the structure. For two-dimensional geometry, it is well established that this method provides a solution that approximates a lateral average of the conductivity below the site. The same is normally

true for large-scale three-dimensional features that represent a grossly two-dimensional appearance to the magnetotelluric response. Low values for the tensor skew and ellipticity parameters offer reasonable (but not absolute) assurance of the validity of the approximation.

Significant error can be produced by some three-dimensional geometry, especially in the case of local small-scale three-dimensional features caused by current channeling. This error is usually manifested as amplitude scale factors on the impedance functions. If three-dimensional errors occur, they are normally suggested by several parameters, including elevated skew and ellipticity, high parallel electric/perpendicular electric impedance mode separation, and laterally abrupt changes in low-frequency impedance. First-order corrections for this error effect are made in the interpretation. The corrections for this survey were based on the assumption that layer 3 has lateral continuity with relatively constant conductivity throughout the survey area. This is a rather speculative assumption, but it is useful for a first approximation due to the absence of borehole and electric log data beneath the basalt. It is possible though that erroneous shallowing of the deep conductive layer still occurs at some of the magnetotelluric sites.

The subsequent geological interpretation based on the magnetotelluric electrical interpretation is subjective. A range of subsurface geological configurations could produce any given subsurface geoelectrical structure. Therefore, an integration of all geophysical and geological information available in an area is essential for determining the most credible geologic interpretation. In general, the porosity of the rock, which can vary with lithology, and the interstitual fluid chemistry, in particular salinity, are the primary governing parameters affecting the bulk resistivity. Porosity also varies with compaction, which is generally a function of age (Dobrin, 1976; Telford and others, 1976). The conductivity of mineral grains is generally of secondary importance. In general, sedimentary rocks are electrically more conduc- tive than igneous or metamorphic rocks.

GEOPHYSICAL OBSERVATIONS IN THE CENTRAL PASCO BASIN

The five-layer geoelectric stratigraphy in the central Pasco Basin was first reported in Myers/Price and others (1979) based on a magnetotelluric survey conducted in the Pasco Basin during 1979. The five-layer geoelectric stratigraphy is considered to be a realistic representation of the Pasco Basin, although the absolute depths and overall configuration reported in Myers/Price and others (1979) should be reevaluated. The magnetotelluric method has benefited in the past few years from significant development and advancement in the data acquisition and data quality phases as well as in the interpretation techniques. Therefore, more credence and reliability is placed on the recently acquired magnetotelluric data because of the overall higher data quality.

The five geoelectric layers were defined based on magnetotelluric apparent resistivity curves. The apparent resistivities range from 1 to greater than 500 Ω-m with values on the order of 100 Ω-m for the volcanics (layer 2), 20 to 50 Ω-m for the basement complex (layer 4), and around 5 to 10 Ω-m for the conductive section (layer 3) between these relatively more resistive sections. The values represent averages for thick stratigraphic sections and do not preclude the existence of thin anomalous layers within a generally electrically uniform sequence. An example is the electrically more conductive, thin sedimentary interbeds (Ellensburg Formation) within the overall thick resistive Columbia River basalt sequence, jointly referred to as volcanics. The magnetotelluric method cannot typically resolve these thin layers, but instead maps the entire sequence with a bulk apparent resistivity that is slightly lower than it would be if the entire sequence were basalt. Electrical logs of boreholes that penetrate layer 3 are necessary to establish good depth and resistivity control.

The volcanics (layer 2) have a high magnetic intensity because basalts, known to be the predominant volcanic rock on the plateau, are highly magnetic. The magnetic character of the basalts is composed of two components, susceptibility and natural remanent magnetization. Both properties can have variable intensities within basalt flow units and often very high variability between different flow units. Susceptibility, natural remanent magnetization, and the geometric configuration of the individual units are parameters typically used in magnetic modeling programs.

Numerous polarity reversals occur within the basalts on the Columbia Plateau. The magnetostratigraphic units and the general thicknesses of the units within the Saddle Mountains and Wanapum Basalts are reasonably well understood. The Grande Ronde Basalts are less well understood.

Layer 3 is not expected to exhibit significant magnetic intensity because sediments, in general, have a magnetic intensity that is considerably lower than extrusive volcanics. This unit of low magnetic intensity may provide a useful magnetic contrast in the interpretation and modeling of the full section, though. Magnetic parameters for layer 4 are unknown. It is assumed, based on interpreted lithology, that this layer may have a higher magnetic intensity than the overlying layer 3.

Investigations of deep structures using gravity data make assumptions on the density values of the geoelectric layers. In the Pasco Basin, density values measured directly or from reasonable assumptions based on lithologic observations are available for layers 1 and 2. For layer 2, the volcanic sequence (Weston, 1981) reports densities ranging from 2.6 to 2.8 gm/cm^3. Cady (1980) used densities of 2.58 and 2.76 gm/cm^3 for the volcanics in his crustal models. These appear to be reasonable estimations.

Data on the density of layer 3 are presently not available. Densities must be inferred from conjectured basement lithologies. Therefore, geologic interpretations of the basement relying on gravity must be used cautiously. Weston (1981) has done two-layer gravity models across the Pasco Basin. They have used a density contrast of 0.187 gm/cm^3 between the volcanics and their interpreted underlying basement complex. Their models do not include the resistive basement, layer 4, depicted by magnetotelluric data.

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APPENDIX B - METHODOLOGY FOR COMPUTER ASSISTED DETERMINATION OF RATES OF DEFORMATION

METHODOLOGY

A paleoslope was calculated using data from geologic mapping (Swanson and others, 1981) east of the Pasco Basin. The Roza-Frenchman Springs contact and Pomona intracanyon flow were used as the reference horizons for calculating that paleoslope. The Roza-Frenchman Springs contact at Almota and Devil's Canyon sections (Swanson and others, 1980) yielded elevations of 518 m and 366 m, respectively. The vertical distance between the two contacts is 152 m and the map distance between these two points is 79.8 km. Thus, the contact is inferred to have a slope of 1.9 m/km.

In comparison, the gradient of the present Snake River between Little Goose and Lower Monumental dams is 1.3 m/km. The gradient of the ancestral Snake River based on the Pomona intracanyon low flow yields 1.5 to 2.0 m/km. This gives a dip value, but no direction. A dip direction was visually selected based on drainage patterns on the Palouse slope due east of the Pasco Basin. The dip direction selected was 245° (with a range of 240° to 255°).

To project the paleoslope west into the Pasco Basin, the 1.9 m/km dip and 245° declination was divided into two components, a north-south component and an east-west component. Trigonometric calculation yields a north to south component of 0.80 m/km and an east to west component of 1.72 m/km.

Using these components of dip, three points were selected within the Pasco Basin and the elevation of the paleoslope plane was determined for each of those points. These three points were used to effectively define the plane of the paleoslope within the Pasco Basin using computer generated graphics. The geographic location and elevation of the paleoslope for each of these three points were input into a computer graphic surface gridding routine (marketed by Dynamic Graphics) and Trend Surface Analysis, which produced a planar surface with a dip of 1.9 m/km to the west-southwest. These gridded plane data were then subtracted from a file of gridded topographic information for the same geographic area (Pasco Basin). The resultant residual data were plotted as a contour map of the area with contours depicting topography above and below the paleoslope intersection with the existing topography. The zero contour line repre sents the line of intersection of the paleoslope with topography, and identifies areas of no net deformation. It is also possible to identify areas of net subsidence (negative isopleths) and net uplift (positive isopleths).

The average thicknesses of basalt members from the areas of no relative deformation (the zero isopleth) were used as a standard to compare with the measured thickness of members from sections and boreholes. The resulting differences in member thicknesses for a given section or borehole were graphed against time stratigraphic intervals to determine uplift or subsidence rates.

From the gridded paleoslope plane data, it was possible to determine the elevation of the hypothetical paleoslope at each borehole and measured section. Starting with this elevation and applying the rates determined from the graphs, the elevation of a theoretical top of basalt for each borehole and measured section used was produced. Using the theoretical top-of-basalt values at each borehole and measured section, it was possible to construct a theoretical top-of-basalt contour map using computer graphics. This map was based solely on rates of deformation over the last 14.5 million years starting from a base elevation of the plane of the projected paleoslope.

Finally, the top-of-basalt map based on rates of deformation was compared with the observed top-of-basalt map (Myers/Price and others, 1979, Plate III-4a; and Myers, 1981). A computer file of the theoretical top-of-basalt map was subtracted from a file of the observed top-of-basalt map to produce a map of residuals. This resultant map graphically portrays deformation anomalies.

Potential errors in the method are of several types. First, the lack of borehole control limits the accuracy of the detail of the isopleths. Second, the rates of uplift and subsidence are dependent on the number and location of measured field sections and boreholes. Unless located at a point of maximum subsidence or uplift, the rates are apparent values. Potentially the greatest source of error is the lack of data from 10.6 million years before present to the present.

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