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STRUCTURAL EVOLUTION OF THE COLUMBIA PLATEAU IN WASHINGTON AND OREGON

WARREN BARRASH,* JOHN BOND,** and
RAMESH VENKATAKRISHNAN***

DEDICATION

We dedicate this paper to Jim Fitzgerald who was completing a study of the geology of the Weiser Basaltic Embayment of the Columbia Plateau when he was killed in the eruption of Mt. St. Helens on May 18, 1980.

ABSTRACT. Columbia River Basalt flows were extruded episodically between 17 and 6 m.y.b.p.; they partially filled a pre-existing topographic basin in eastern Washington, western Idaho, and northern Oregon. At least 99 percent of the basalt volume was erupted by 13.5 m.y.b.p. Geophysical properties define a unique Columbia Plateau crustal unit north of the Columbia Arc. Deformation styles differed on either side of the Columbia Arc during plateau evolution. In addition, within the Columbia Plateau crustal unit: (1) a subsurface crustal discontinuity appears to separate a northern continental terrane from a southern oceanic(?) terrane, and (2) the plateau surface is transected by the Olympic-Wallowa (topographic) Lineament which is defined by the structural anomalies of the Cle Elum-Wallula (deformational) Lineament.

Deformation of the Columbia Plateau produced a wide range of structural styles and orientations during three tectonic stages. Stage one (17 to 10 ± 2 m.y.b.p.) was characterized by: (1) north-northwest-trending dike swarms, (2) mild warping along several orientations in numerous locales across the plateau, and (3) formation of open folds of the Blue Mountains anticlinorium. Orientation of $\sigma_{H \max}$ was dominantly north-northwest. Stage two (10 ± 2 to ~ 4 m.y.b.p.) reflects a change of $\sigma_{H \max}$ orientation to north-south; it was characterized by: (1) formation of asymmetrical anticlinal ridges and associated structures north of the Columbia Arc, (2) uplift of the Blue Mountains with strike slip at the northwestern margin and extension in the eastern Blue Mountains region, and (3) thrusting, block faulting, and strike slip in the John Day transition zone. Stage-two deformation produced most of the structural and topographic relief evident in the Columbia Plateau today. Stage three (~ 4 m.y.b.p. to the present) has been characterized by relatively weak deformation of somewhat different pattern than stage two. Cur-

* Geoscience Research Consultants and Department of Geology, University of Idaho, Moscow, Idaho 83843. Present address: Conservation and Survey Division, IANR, Panhandle Station, Scottsbluff, Nebraska 69361

** Geoscience Research Consultants, 641 N. Hayes, Moscow, Idaho 83843

*** Department of Geophysical Sciences, Old Dominion University, Norfolk, Virginia 23508

rent seismicity and Quaternary structures indicate that most stage-three strain mechanisms are consistent with north-south-oriented

$\sigma_{II \max}$.

The three stages of deformation in the Columbia Plateau may be correlated with changing tectonic patterns at the western margin of the North American plate. Stage one in the Columbia Plateau area began at the time that a widespread thermal disturbance caused uplift, east-northeast-west-southwest extensional tectonism, and basaltic and bimodal volcanism in the Great Basin region. Stage two began when the Pacific plate spreading direction rotated approx 25° in a clockwise direction and the Pacific plate significantly increased its rate of motion relative to the North American plate. Stage three began as part of a culmination of inter- and intraplate tectonic readjustments in the Pacific Northwest region.

INTRODUCTION

In this paper the term Columbia Plateau refers to the area east of the Cascade Mountains that is underlain by Columbia River Basalt. The plateau is largely an intermontane topographic basin which covers portions of Washington, Oregon, and Idaho (fig. 1). In the last 15 yrs detailed stratigraphic data have been accumulating from all portions of the plateau as individual flows, and stratigraphic units have become recognized and mapped by a combination of field characteristics, paleomagnetic properties, and geochemistry. Generalized stratigraphic information on the Columbia River Basalt and on some associated units of the Columbia Plateau is summarized in figure 2.

Structural evolution of the Columbia Plateau has long been a subject of debate due to the variety of structural features, styles, and orientations developed in Columbia River Basalt and associated rocks. However, spatial and temporal patterns of events are emerging from the steadily growing data base. From these data, we interpret the structural evolution of the portion of Columbia Plateau that lies in Washington and Oregon to be a three-stage process influenced by pre-existing topography, basement character, regional stress orientations, and local stress configurations. Although structural patterns that developed in the plateau differ from adjacent regions, the plateau cannot be fully understood until it is placed into the larger picture of late-Cenozoic tectonism of the western United States. The nature and timing of deformation in the Columbia Plateau appear to be linked to variable interactions of the Pacific, North American, and remnant Farallon plates.

DISTRIBUTION

Columbia River Basalt volcanism occurred episodically between 17 and 6 m.y.b.p. with more than 99 percent of the total volume of Columbia River Basalt erupted by 13.5 m.y.b.p. (McKee, Swanson, and Wright, 1977; McKee, Hooper, and Kleck, 1981). Hence, the Columbia Plateau was essentially constructed by 13.5 m.y.b.p. The regional distribution of Columbia River Basalt units was largely controlled by a pre-existing, westward-draining topographic basin in eastern Washington, northern

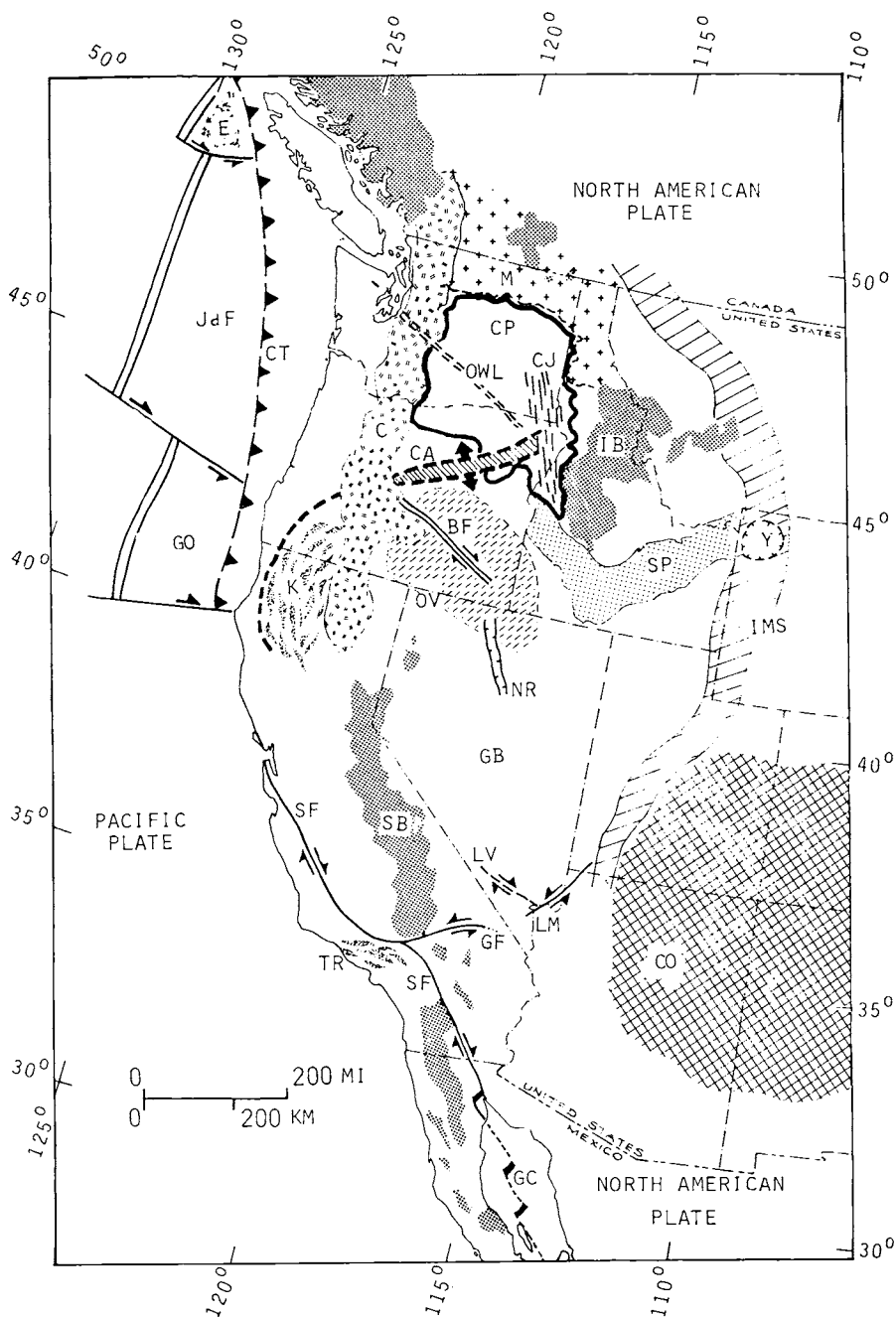


Fig. 1. Tectonic elements of the western United States. E, Explorer plate; JdF, Juan de Fuca plate; GO, Gorda plate; CT, Cascade trench; M, metamorphic and plutonic terrane; CP, Columbia Plateau; CJ, Chief Joseph dike swarm; OWL, Olympic Wallowa Lineament; IB, Idaho batholith; C, Columbia Arc; CA, Columbia Arc; BF, Brothers fault zone; SP, Snake River Plain; Y, Yellowstone; K, Klamath Mountains; OV, South-eastern Oregon volcanic plateau; NR, Northern Nevada rift; IMS, Intermountain Seismic Belt; GB, Great Basin; SF, San Andreas fault zone; SB, Sierra batholith; LV, Las Vegas shear zone; LM, Lake Mead fault zone; GF, Garlock fault; CO, Colorado Plateau; TR, Transverse Ranges; GC, Gulf of California.

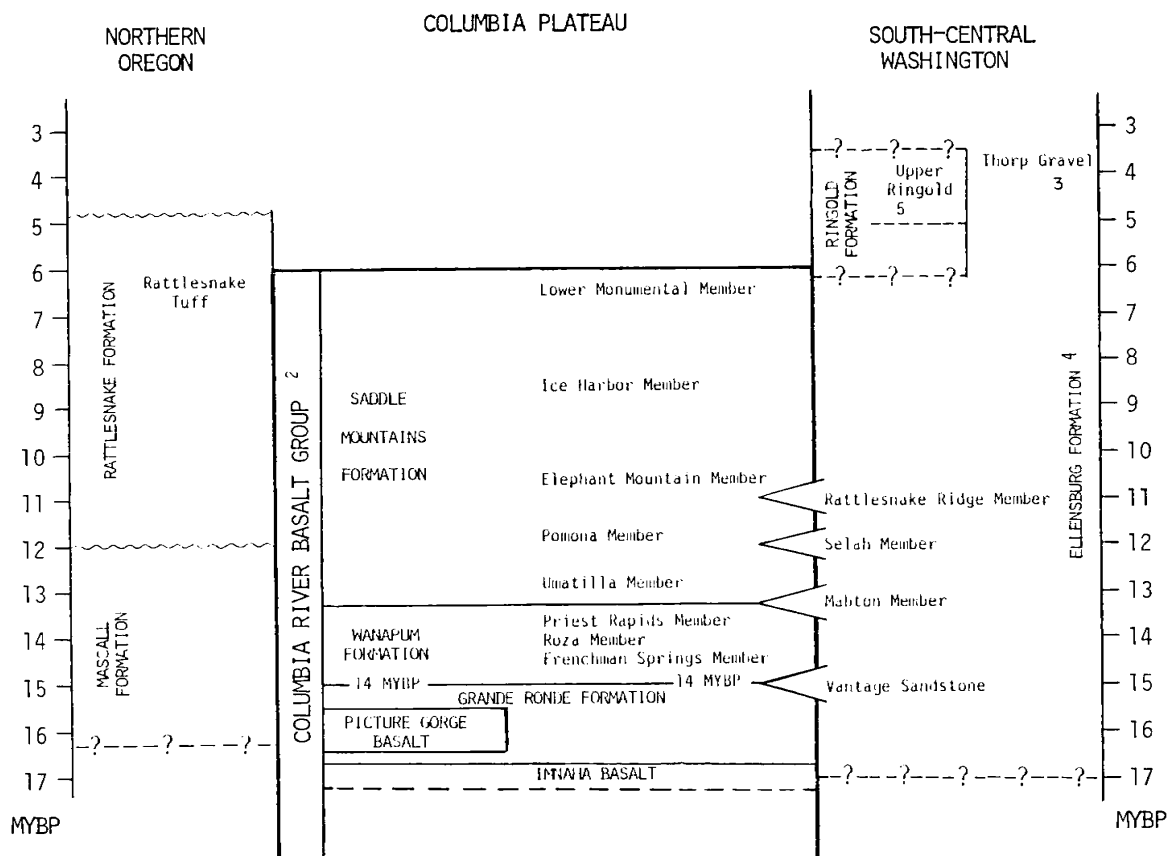


Fig. 2. Generalized stratigraphy of the Columbia River Basalt Group and some associated units of the Columbia Plateau. (1) after Beaulieu, 1972, and Enlows, 1973; (2) after Swanson, Wright, and others, 1979, and Hooper, 1982; (3) and (4) after Shannon & Wilson, 1977; (5) after Gustafson, 1978, and Packer and Johnston, 1979.

Oregon, and western Idaho. The boundary of this pre-existing basin and the Columbia Plateau is well defined on the east, north, and west where Columbia River Basalt flows terminate against ancestral high-relief mountains. The southern boundary of the Columbia Plateau was not and is not well defined at the surface by a mountain front. Rather it is a transition zone that has been tectonically active numerous times in the past (Bud-denhagen, 1967; Fisher, 1967; Vallier, Brooks, and Thayer, 1977). Columbia River Basalt flows lapped onto and over a series of east- to northeast-trending topographic and structural highs that occupied the area of the Columbia Arc (Taubeneck, 1966) or present-day Ochoco-Blue Mountains uplift (Rogers, 1966; Nathan and Fruchter, 1974).

Distribution of Columbia River Basalt within the Columbia Plateau has been influenced by centripetal subsidence during and following the extrusion period. Off-lap contact relationships occur within (Grolier and Bingham, 1978) and at the margins (Mackin, 1961; Bond, 1963) of the plateau. Less voluminous members of the youngest basalt formation (Saddle Mountains Basalt) are restricted to the central portion of the plateau, even though most of the flows originated near the southeastern margin (Price, ms; Swanson, Wright, and others, 1979). The greatest reported thickness of Columbia River Basalt occurs at the edge of the Pasco Basin near the center of the plateau where the Rattlesnake 1 well penetrates more than 3050 m of volcanic rocks, with thin coal layers in portions of the section (Raymond and Tillson, 1968). Reidel and others (1982) reported that Grande Ronde Basalt occurs to the base of the well; they cited magnetotelluric data which suggest the base of the volcanic sequence (perhaps entirely Columbia River Basalt) may be 4500 m deep. Subsidence, centered in the region near Pasco, was likely due to isostatic readjustment to crustal loading (Bond, 1969; Swanson, Wright, and Helz, 1975) and may be continuing at present (Waters, 1955; Brown and McConiga, 1960).

CRUSTAL PROPERTIES

Regional geophysics indicates that the Columbia Plateau north of the Columbia Arc is a distinct crustal province in addition to being a distinct lithologic and deformational province. The areal extent of Columbia River Basalt north of the Columbia Arc is coextensive with high Bouguer gravity values (> -110 milligal) which center in the Pasco Basin (Bonini, Hughes, and Danes, 1974; Eaton and others, 1978). Passing through the Columbia Arc is a zone of transition to the regional gravity low (< -110 milligal) which is characteristic of the Great Basin (Eaton and others, 1978). The Columbia Arc divides the Columbia Plateau into regions of lower elevation (< 1 km) to the north and higher elevation (> 1 km) to the south and east (Blackwell, 1978; Eaton and others, 1978). Smith (1978, p. 117) presented a map of crustal thickness which shows thin crust (closed 25-km contour) within the Columbia Plateau and crustal thickening (to 30 km) coincident with the Columbia Arc and with the crystalline border regions to the east, north, and northwest of the plateau. Smith (1978, p. 118) also showed that higher P_n velocity (8.1 km/sec) oc-

curs beneath Columbia Plateau crust north of the Columbia Arc than occurs beneath the Cascade Mountains (7.8-8.0 km/sec) and regions south of the Columbia Arc (7.8-8.0 km/sec). Heat flow through the Columbia Plateau is high relative to average continental crust, but the central portion of the plateau is a region of low heat flow (< 1.5 heat-flow units) relative to surrounding regions and to the Cordilleran thermal anomaly zone as a whole (Blackwell, 1969, 1978).

Despite the above-mentioned unifying properties of the crust beneath the Columbia Plateau north of the Columbia Arc, several lines of evidence suggest that a continental(sialic)-oceanic(simatic) crustal transition zone is present beneath the plateau north of the Oregon-Washington border. Also, the plateau surface is transected by the northwest-trending Olympic-Wallowa Lineament (OWL) which was originally defined as a regional topographic anomaly (Raisz, 1945). Some workers have suggested that the OWL is the surface manifestation of basement structure and/or a continental-oceanic crustal transition. Some geophysical data are compatible with interpretations of basement structural control of the OWL and with a crustal transition beneath the plateau. However, the crustal transition may not be beneath or parallel to the OWL.

Cantwell and Orange (1965) and Cantwell and others (1965) found that deep resistivity measurements suggested the presence of: (1) highly resistive granitic basement north of the OWL in north- and east-central Washington, (2) more conductive, perhaps oceanic, crust south of the OWL in northern Oregon, and (3) a zone of transition in resistive properties in the region of the OWL. Rocks hosting the variable resistive properties were interpreted to occur below a relatively uniform geophysical unit of 10 to 15 km thickness (Cantwell and others, 1965). Hill (1972) interpreted long-range seismic refraction data across central Washington to indicate strong lateral variations in the crust and upper mantle beneath the Pasco Basin, north of the Oregon-Washington border. Rodi and others (ms) interpreted the joint inversion of gravity and earthquake travel-time data to indicate that a transition from continental (north) to oceanic (south) crust and upper mantle occurs across an east-trending zone at about lat 47 (north of the OWL).

Skehan (1965, 1966), following Cantwell and others, hypothesized that the OWL reflects a deep-seated continental-oceanic boundary which may be a major fault beneath the overlying units. Swanson (*in* Zeitz and others, 1971) interpreted aeromagnetic data to be compatible with either Skehan's (1965, 1966) hypothesis or with strike-slip movement along the OWL. Swanson noted that a pronounced northwest trend was present in broad negative anomalies north of the OWL and in broad positive anomalies south of the OWL. This northwest trend was attributed to a pre-Miocene structural grain in basement rocks beneath the Columbia Plateau. Hammond (1979) interpreted the OWL to be a former continental margin that is a zone between laterally immobile continental crust (north of the OWL) and rifted, stretched, and rotated Columbia Arc crust (south of the OWL).

Although there is a common theme to most interpretations of the tectonic significance of the transition zone of crustal geophysical properties and the OWL, large uncertainties about their fundamental natures and relationships remain.

BASEMENT INFLUENCES ON COLUMBIA PLATEAU DEFORMATION

Basement features have, in part, controlled deformation patterns within the Columbia Plateau. As noted above, the Columbia Arc (1) coincides with the transition from the Columbia Plateau crustal unit to Basin and Range-type crust and (2) has been tectonically active in the geologic past. Early in Columbia Plateau history the Columbia Arc was reactivated as the locus of broad folding and minor uplift. Later, after 10 ± 2 m.y.b.p., more significant uplift was localized along this axis, which also was the boundary between compression (to the north) and extension (to the south).

Within the plateau, the topographic OWL results from aligned structural features which are anomalous because of orientation, degree, and type of deformation. "Basement influence" appears to be responsible for OWL structural anomalies, as will be discussed later.

Structural fabric within the Wallowa batholith may have influenced the orientations of dikes and faults. Structural ramps and grabens have been produced at the margins of the uplifted Wallowa batholith.

A large fraction of Columbia River Basalt was erupted from the Chief Joseph dike swarm which may have been localized by the presence of a Paleozoic-Mesozoic arc-trench suture zone (Davis, 1977, 1980).

The northeastern portion of the plateau is relatively undeformed (Griggs, 1976; Swanson and Wright, 1978). This region is underlain (butressed, pinned) at relatively shallow depths by granitic and metamorphic rocks with considerable sub-basalt topographic relief.

STRUCTURAL STAGE ONE: NORTH-NORTHWEST-ORIENTED $\sigma_{H \max}$ DURING THE PERIOD 17 M.Y.B.P. TO 10 ± 2 M.Y.B.P.

During stage one (fig. 3) Columbia River Basalt erupted from predominantly north-northwest-trending dikes and constructed the Columbia Plateau. Tectonic joints of similar orientation to the dikes were imparted to the plateau in southern Washington and northern Oregon. Mild warping in the western and east-central portions of the plateau followed northwest and east trends. Relatively mild, episodic anticlinal folding was localized in the Yakima Ridges area. Broad open folding of the Blue Mountains anticlinorium followed northeast to east trends along the Columbia Arc; strike-slip faulting and incipient uplift accompanied the folding in northeastern Oregon.

Dikes and joints.—Starting about 17 m.y.b.p., Columbia River Basalt erupted from the Monument and Chief Joseph dike swarms. Fisher (1967) noted that the Monument syncline was forming concurrently with eruption from the Monument dike swarm, and he attributed the deformation to north-northwest compression which had been active prior to Columbia River Basalt extrusion. The predominant orientation of Columbia River

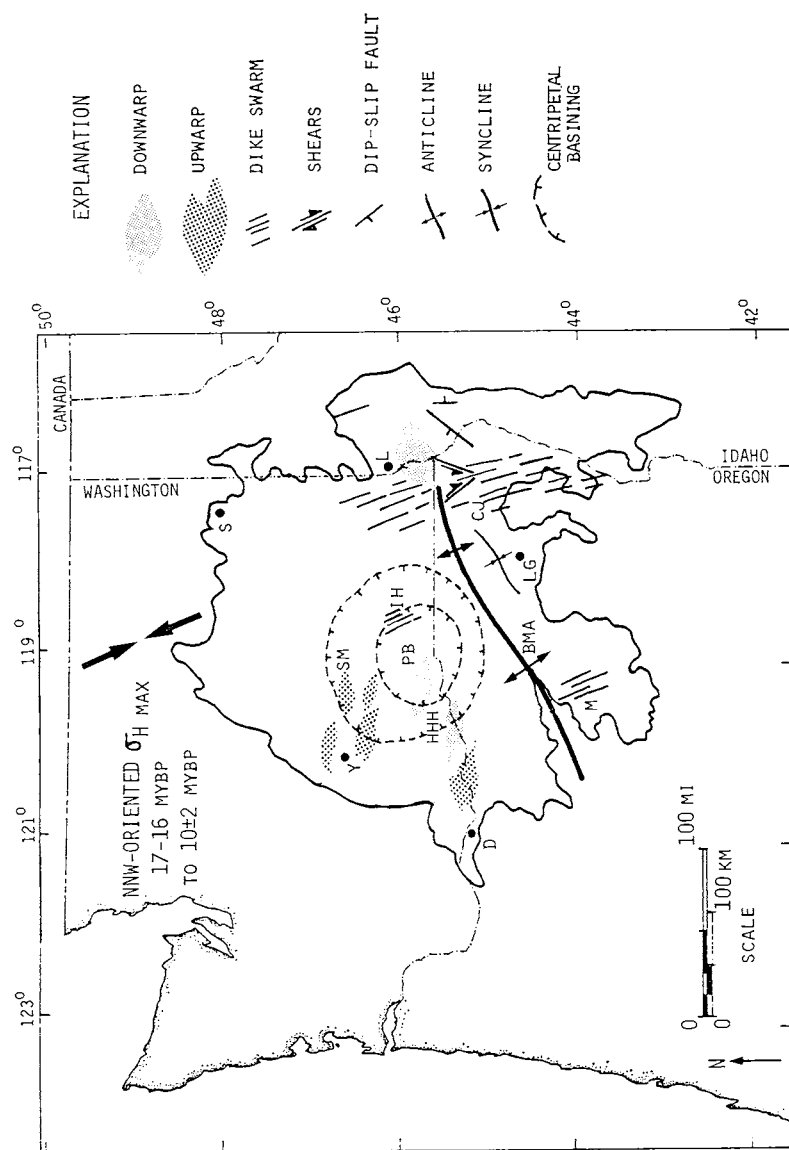


Fig. 3. Generalized map of stage-one structural deformation in the Columbia Plateau, S, Spokane; Y, Yakima; L, Lewiston; D, The Dalles; LG, LaGrande; SM, Saddle Mountains area; PB, Pasco Basin area; IH, Ice Harbor dike swarm; M, Monument dike swarm; BMA, Blue Mountains anticline; HHHH, Horse Heaven Hills area.

Basalt dikes are N 30° W in the Monument swarm (Brown and Thayer, 1966) and N 10° to 30° W in the Chief Joseph swarm (Gibson, 1969; Taubeneck, 1970; Hooper, 1982). Smedes (ms) suggested that the N 10° W orientation of dikes in the Wallowa Mountains was due to the influence of pre-existing joints in the plutonic rocks there. Near the center of the Pasco Basin, Ice Harbor Basalt dikes trend N 30° W (Swanson, Wright, and Helz, 1975). The Ice Harbor flows (fig. 2) lie near the top of the Columbia River Basalt stratigraphic sequence and have been dated at 8.5 m.y.b.p. (McKee, Swanson, and Wright, 1977).

If we consider basalt dikes to be extension fractures oriented approximately parallel to the regional maximum compressive stress (Nakamura, 1977; Nakamura, Jacob, and Davies, 1978), we can conclude that $\sigma_{H \max}$ was oriented approx N 20° to 30° W from at least 17 to 8.5 m.y.b.p. However, indications of a change to relatively strong north-south compression in the Columbia Plateau at about 10 m.y.b.p. (Shannon & Wilson, 1977; Robyn, 1979) suggest reactivation of north-northwest-trending fractures established in stage one, or a slightly later change to north-south compression in the Pasco Basin (see stage two discussion) to explain the orientation and timing of the Ice Harbor dike swarm.

Vertical fractures with a predominant N 30° W trend have been recognized in the Columbia Plateau of northern Oregon. Lawrence (1979) interpreted northwest-striking tectonic joints in Columbia River Basalt in north-central Oregon to be the earliest set of tectonic joints in Columbia River units. Master joints of similar trends are exposed in the western portion of The Dalles-Umatilla syncline (Shannon & Wilson, 1973). In northeastern Oregon, some N 30° W fractures were reactivated later (Venkatakrishnan and others, in preparation) during uplift of the Blue Mountains (stage two) and are delineated as dip-slip faults on geologic maps of this area (Walker, 1977; Swanson and others, 1981).

Mild warping and anticlinal folding.—Mild warping in the west-central portion of the Columbia Plateau occurred during stage one. Laval (ms) recognized downwarping and mild local warping during the accumulation of Saddle Mountains Basalt (fig. 2) in the southwestern Pasco Basin. Thinning of the Saddle Mountains Basalt occurred in the Rattlesnake Hills area while thickening of Saddle Mountains units occurred in the Horse Heaven Hills area (Bond and others, 1978). In north-central Oregon and south-central Washington, minor pre-Roza warping followed northwest trends, and minor pre-Selah warping followed an eastern trend (Farooqui and Kienle, 1976; Farooqui and Kent, 1978). A similar pattern of northwest warping followed by east-trending warping was noted by Reidel (1978) in the Saddle Mountains area.

Controls on the localization and orientations of these early structures are not easily explained with north-northwest-oriented $\sigma_{H \max}$. The warps and folds may reflect local responses to rapid crustal loading over pre-existing northwest-trending structural grain. Alternatively, or in addition, north-northwest compression may have induced limited dextral shear on pre-existing, northwest-trending basement structure(s) beneath the OWL

and thereby influenced the orientation of surface folds as in Riedel-type deformation (Laubscher, ms and 1981; Davis, 1981; Koide and Bhattacharji, 1977, fig. 3F).

Despite the structural adjustments in the Columbia Plateau during stage one, topographic relief was minor. In the western portion of the plateau minor folding was in progress in Wanapum and early Saddle Mountains time, but trunk rivers shifted widely across an alluvial plain that was interrupted only locally with basalt ridges (Mackin, 1961; Bentley, Anderson, and others, 1980). Similarly, relief and elevations were low in the Rattlesnake Hills and Horse Heaven Hills areas during stage one. Coarse ancestral Columbia River gravels of dominantly non-basaltic origin overlie Elephant Mountain Basalt on the Rattlesnake Hills and Horse Heaven Hills anticlines; hence, growth of these structures and major diversion of the Columbia River did not occur until after 10.5 m.y.b.p. (Bond and others, 1978; Bentley, Anderson, and others, 1980).

In the Yakima Ridges area, Waters (1955) interpreted depositional patterns and unconformities within Ellensburg sediment tongues intercalated with Columbia River Basalt flows to indicate that anticlinal deformation was concurrent with deposition of the Ellensburg Formation. Also in the Yakima Ridges area, Saddle Mountains valley-fill basalt flows occur in a broad synclinal valley formed in earlier, widespread units (Bentley, 1977; Kauffman, Bond, and Barrash, ms). However, these same valley-fill flows form flood sheets of relatively even thickness in adjacent areas and the central portion of the Columbia Plateau (Schmincke, 1967; Swanson, Wright, and others, 1979; Bentley, Anderson, and others, 1980). Thus, stage-one folding north of the Columbia Arc was mostly localized in the Yakima Ridges area; only a few folds had enough topographic relief to stand above the Elephant Mountain Basalt sheet which was extruded at 10.5 m.y.b.p. From analysis of stratigraphic data Reidel and others (1980) determined the maximum rate of uplift during stage one (between 14.5 and 10.5 m.y.b.p.) to be 0.07 mm/yr in the eastern Yakima Ridges area and 0.04 mm/yr in the Saddle Mountains.

Bentley, Powell, and others (1980) reported that extensive thrust faulting occurred in the Columbia Hills area during Grande Ronde time (15 m.y.b.p.) and also occurred in parts of the Yakima Ridges area between 16 and 9 m.y.b.p. (Bentley, Powell, and others, 1980). This reported thrusting contrasts in style (extent of shortening and mode of strain) with the mild up- and down-warpage typical of stage one over most of the western Columbia Plateau.

Blue Mountains anticlinorium.—Broad open folds of the Blue Mountains anticlinorium were formed during stage one. These folds trend northeast to east and are coextensive with the Columbia Arc. Hogenson (1964) and Newcomb (1967) suggested that the Blue Mountains anticline formed prior to a more intense stage of deformation which created the Horse Heaven Hills structure (stage two). Folding began in the Grande Ronde Canyon area in late-Grande Ronde time; spreading of post-Grande Ronde flows was increasingly confined to structural downwarps until the growth of an east-trending segment of the Blue Mountains anticline

blocked the northward spread of flows erupted 10.5 m.y.b.p. and later (Walker, 1973; Ross, ms).

Conjugate strike-slip faults developed prior to 10.5 m.y.b.p. in the Grande Ronde Canyon area (Ross, ms). These shears were oriented about N 40° W and N 20° E and formed north of the Wallowa batholith where the Blue Mountains anticlinorium has an approximately eastern trend. Apparently, $\sigma_{H \max}$ was oriented about N 10° W in the Grande Ronde Canyon area during stage one. It is important to note that where stage-one strike slip did occur, the strike-slip fault orientations were approx 30° from the orientations of dikes (that is, extension fractures) in the same area.

Geomorphological evidence along the course of the Grande Ronde River is compatible with fracturing and open folding (stage one) followed by post-10.5 m.y.b.p. major uplift (stage two). The Grande Ronde River follows a northeast course east of the Blue Mountains anticline (Walker, 1977). West of the LaGrande graben, some linear segments of the upper reaches of the Grande Ronde River and most tributaries to the river follow N 30° W linear trends; these linear segments developed along N 30° W-trending extension fractures. Below its upper reaches, the Grande Ronde River follows the northeast, arcuate trend of the Grande Ronde synclinal hinge before spilling into the LaGrande graben. This river course was consequent within the openly folded Grande Ronde syncline which was formed prior to uplift (stage two) as indicated by the presence of meanders which were subsequently incised in the gorge above LaGrande (Barrash and others, 1980). Similarly, during stage one in the Grande Ronde Canyon near the Oregon-Washington border the Grande Ronde River meandered across a relatively low-relief surface within a broadly warped region that later (post-10.5 m.y.b.p. or stage two) was uplifted rapidly enough to entrench the meanders (Walker, 1973; Ross, 1980). Such broad stage-one warping within the basalt pile can be seen in structures as far east as the Lewiston Basin in Idaho (Bond, 1963; Camp, 1976; Camp and Hooper, 1981).

It should be noted, however, that some uplift in the Blue Mountains area occurred during at least one pulse within stage one. The north-northeast-trending Hite fault system bounds the Blue Mountains east of the Walla Walla area (Newcomb, 1965, 1970); displacement across the fault system of > 100 m occurred prior to Frenchman Springs time, and oblique slip occurred after Frenchman Springs and Umatilla times (stage two?) (Shannon & Wilson, ms, 1979a).

STRUCTURAL STAGE TWO: NORTH-SOUTH-ORIENTED $\sigma_{H \max}$ DURING THE PERIOD 10 ± 2 M.Y.B.P. TO ~ 4 M.Y.B.P.

Aside from the folds in the Yakima Ridges and Blue Mountains regions, at the close of stage one (10 ± 2 m.y.b.p.) the Columbia Plateau was a relatively featureless surface with minor undulations distributed over much of the plateau. The second stage in the structural evolution of the plateau resulted in a varied suite of higher amplitude structures (fig. 4) which, as in stage one, locally suggest basement influence. Maximum hori-

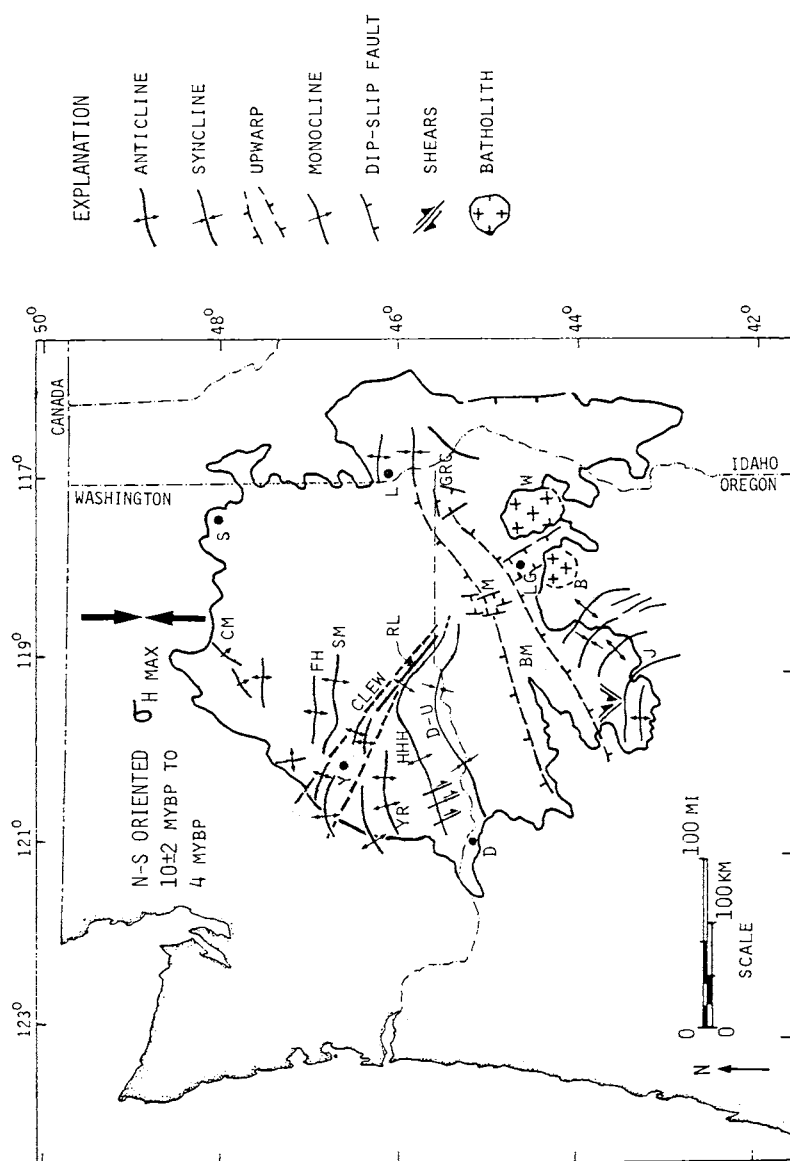


Fig. 4. Generalized map of stage-two structural deformation in the Columbia Plateau. Centripetal subsidence continued during stage two. S, Spokane; L, Lewiston; D, The Dalles; LG, LaGrande; CM, Coulee monocline; FH, Frenchman Hills anticline; SM, Saddle Mountains anticline; CLEW, Cle Elum-Wallula Lineament; RL, Rattlesnake Lineament; YR, Yakima Ridges area; HHH, Horse Heaven Hills structure; D-U, Dalles-Umatilla syncline; BM, Blue Mountains uplift; M, Meacham Creek trough; GRC, Grande Ronde Canyon area; W, Wallowa batholith; B, Bald Mountain batholith; J, John Day fault area.

zontal compressive stress was oriented approximately north-south for this stage of deformation. The plateau in central Washington and north-central Oregon deformed in response to compression. In the Blue Mountains region uplift was accompanied by extensional faulting.

Stage-two deformation may have begun as early as 12 m.y.b.p. (Swanson and Wright, 1978) or at different times down to 8 m.y.b.p. at different localities. Stage two lasted at least until Ringold time (Grolier and Bingham, 1978) and, in places, upper Ringold time (Brown and McConiga, 1960), or about 5 to 3 m.y.b.p. (Gustafson, 1978; Packer and Johnston, 1979).

North Columbia Plateau structures.—Structures in the northern Columbia Plateau are dominated by east-trending folds and thrust faults with northward vergence. The deformation that produced these structures post-dated Wanapum and Elephant Mountain Basalts and, in places, thrusts post-date Ringold sediments (Grolier and Bingham, 1971, 1978). Magnitude of deformation decreases northward across the plateau, and deformation intensity along individual structures generally decreases eastward.

The Frenchman Hills anticline in the northern Columbia Plateau trends east and is asymmetrical with steeper north limb; the west end of the anticline is thrust northward over Ringold sediments (Grolier and Bingham, 1978). North of the Frenchman Hills anticline are the east-trending synclinal Quincy Basin and several low-amplitude anticlines and synclines including the Lind Coulee anticline, the Pinto Ridge anticline, and the Soap Lake anticline (Newcomb, 1970; Grolier and Bingham, 1971, 1978). North of the Lind Coulee anticline is the Lind Coulee fault, a north-directed thrust which, in one locality, places Wanapum Basalt over Ringold sediments (Grolier and Bingham, 1971, 1978).

Near the northern margin of the Columbia Plateau, south and west of the Grand Coulee Dam, two monoclines have dominantly northeast trends (Grolier and Bingham, 1971, 1978; Staatz and Morris, 1976). Structural relief of the Coulee monocline exceeds 300 m (Swanson and Wright, 1978). The orientation of the two monoclines contrasts with the eastern and northwestern trends which predominate in the western half of the Columbia Plateau. These structures likely formed during stage-two deformation, but their orientations suggest local basement influence.

Saddle Mountains.—The Saddle Mountains trend N 80° to 90° W from Sentinel Gap on the west to the center of the plateau on the east. Structurally, the Saddle Mountains are an asymmetrical anticline with a steeper and locally overturned (Laval, ms; Grolier and Bingham, 1978) northern limb. North of and parallel to the anticlinal trend is the Saddle Mountains fault which has been variously interpreted to be a thrust fault (Laval, ms; Reidel, 1978) and a high-angle reverse fault (Grolier and Bingham, 1978). Structural relief on the anticline decreases eastward.

Average rates for development of stage-two structural relief in the Saddle Mountains may be estimated from structure-contour data presented by Reidel (1978). The data indicate that up to 640 m of post-

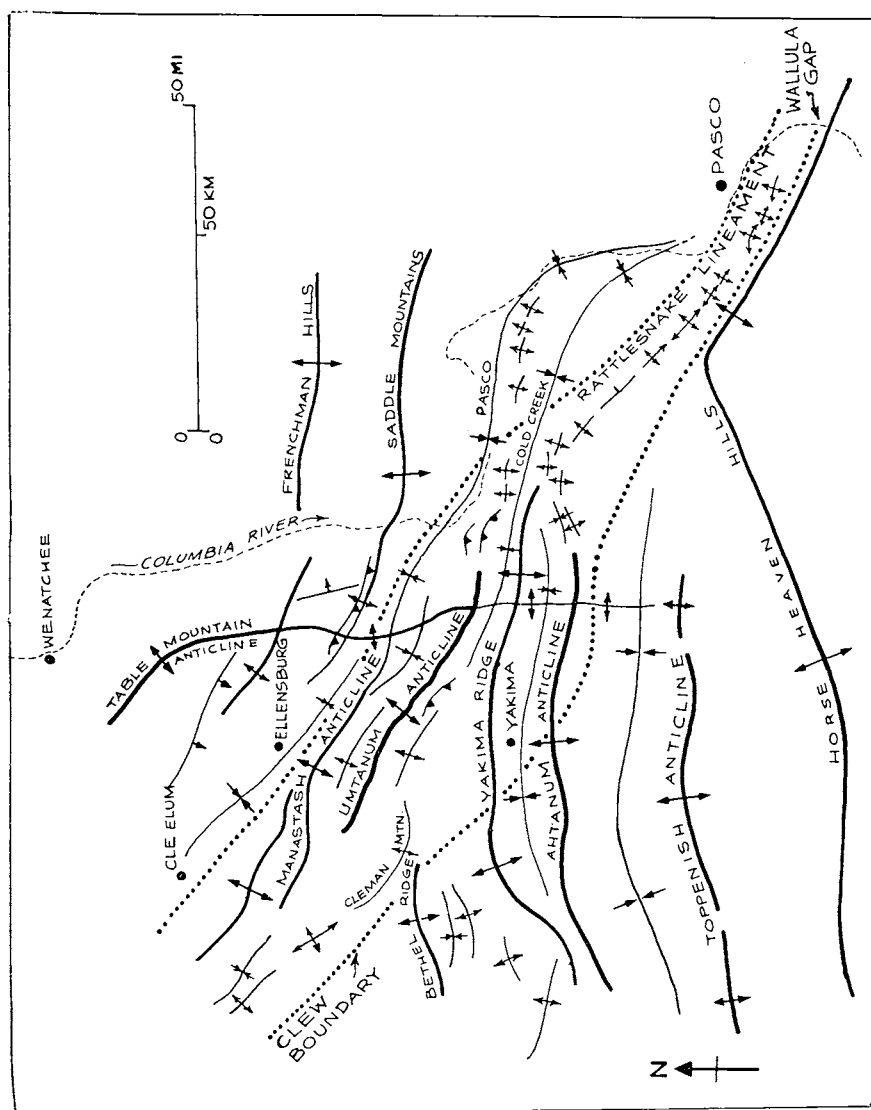


Fig. 5. Generalized map of stage-two structures in the Yakima Ridges area (after Kienle, Bentley, and Anderson, 1977; Bond and others, 1978; Myers and Price, 1979, pl. II-20). Frenchman Hills, Saddle Mountains, and Horse Heaven Hills structures are shown for perspective, but are discussed separately in the text.

Elephant Mountain (post-10.5 m.y.b.p.) uplift occurred south of Smyrna Bench. The stage-two average uplift rate likely fell within the range of 0.1 mm/yr (assuming deformation occurred between 10.5 and 4 m.y.b.p.) and 0.14 mm/yr (for the period of 8.5-4 m.y.b.p.). These rates are 2.5 to 3.5 times greater than the 0.04 mm/yr rate estimated for stage-one uplift in the same area (Reidel and others, 1980).

Yakima Ridges.—The Yakima Ridges occur in the west-central portion of the Columbia Plateau (fig. 5). Waters (1955) and Mackin (1961) recognized that deformation occurred sporadically in the Yakima Ridges area during Columbia River Basalt eruption (stage one). However, relief on stage-one deformation was generally minor as indicated by the extensive, sheet-like distribution of Saddle Mountains flows in the Yakima Ridges area (Schmincke, 1967; Swanson, Wright, and others, 1979). The present structural and topographic relief in the region is dominantly due to post-basalt, stage-two deformation.

The Yakima Ridges are long, linear, asymmetrical anticlines which dominantly trend east and/or northwest. South of Yakima, some individual structures which follow the dominant trend(s) commonly trend northeast for 5 to 15 km at their western terminations near the Cascade Mountains. Broad synclinal valleys separate the narrow anticlinal ridges. The synclines generally have simple, open geometry, but the anticlines generally have thrust and/or reverse faults parallel to fold axes and cross folds and/or faults where there are changes in parent-fold tightness, symmetry, or orientation (Kienle, Bentley, and Anderson, 1977; Shannon & Wilson, 1977). Progressively stronger northward vergence is observed from the center to the western margin of the plateau (Bentley, 1981; Price, 1982).

Within the Yakima Ridges structural province is a 30 to 40-km-wide, N 45° to 50° W-trending zone dominated by N 60° W-trending, en echelon folds and faults (fig. 5), many of which enter or leave this zone with an east trend and with less-pronounced structural relief and complexity (Kienle, Bentley, and Anderson, 1977; Shannon & Wilson, 1977). Laubscher (ms) introduced the term Cle Elum-Wallula Lineament (CLEW) to include both the above-mentioned northwest-trending zone and the Rattlesnake Lineament (discussed below) which follows the N 45° to 50° W trend from the Yakima Ridges to Wallula Gap along the southern margin of the Pasco Basin. The CLEW is along the central 200 km of the Olympic-Wallowa (topographic) Lineament (OWL) of Raisz (1945). The CLEW may overlie and be influenced by basement terrane with pre-existing northwest-trending structural grain (Swanson, *in* Zietz and others, 1971). The increased intensity of structural deformation within the CLEW reflects greater relative mobility of the crust there. Despite the differences in structural orientation and deformation intensity within and outside the CLEW, it appears that the Yakima Ridges structural province, as a whole, deformed contemporaneously within a north-south-oriented compressive tectonic system (Kienle, Bentley, and Anderson, 1977; Kienle and others, 1978; Bentley, Anderson, and Farooqui, 1980; Bentley, Powell, and others, 1980; Price, 1982).

Laubscher (ms and 1981) and Davis (1981) presented models of Riedel-type, en echelon fold development above a northwest-trending basement discontinuity which is favorably oriented for right-lateral strike-slip movement. Davis (1981) cited gravity data which may constrain the magnitude of lateral slip across CLEW to 2 to 3 km. However, Davis (1981) states this limit is compatible with (1) Laubscher's model if depth to the underlying fault increases with increasing width of the affected surface zone, or (2) Davis' model if the width of the underlying fault zone increases with increasing width of the affected surface zone.

Price (1982) presented a model to explain the radiating orientations of folds in the western Columbia Plateau (including the Yakima Ridges and the Horse Heaven Hills, Saddle Mountains, and other folds north of the Yakima Ridges). In this model, the eastern portion of the plateau is underlain by a relatively immobile crustal wedge; basalt in the western plateau is transported northward but is continuous with basalt pinned in the east. This kinematic relationship (fig. 6) results in slight clockwise rotations of fold trends with progressively greater northward translation west of the pinned crust. This model is attractive but may require basement influence, perhaps Reidel-type deformation as suggested by Laubscher (ms and 1981) and Davis (1981), to account for increased deformation intensity in the CLEW and changes of fold trend on entrance to and exit from the CLEW.

Accurate dating of the onset of stage two in the Yakima Ridges region is difficult due to the lack of precise post-basalt time markers and because the uppermost Saddle Mountains flows were not voluminous

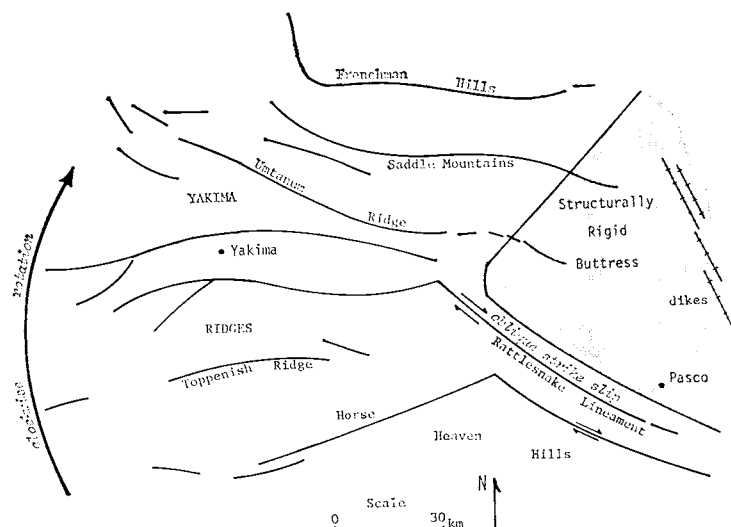


Fig. 6. Price's (1982) kinematic model for stage-two deformation in the west-central part of the Columbia Plateau. Basalt is pinned over the rigid buttress in the northeastern part of the plateau. To the west, folding increases in amplitude and amount of northward translation, thereby resulting in slight clockwise rotation. Southern limit of the buttress is parallel to the Rattlesnake Lineament of the CLEW.

enough to reach this portion of the plateau. However, stage-two deformation of the Yakima Ridges did not begin before 10.5 m.y.b.p. in those areas covered by the Elephant Mountain Basalt (Bentley, 1977; Kauffman, Bond, and Barrash, ms). Farther southeast along the trend of the CLEW, significant stage-two deformation did not begin until post-Ice Harbor time (8.5 m.y.b.p.) (Bond and others, 1978).

In the Yakima Ridges, Kienle and others (1978) estimated fold-amplitude growth rates of 0.75 to 1.5 mm/yr for the period between 8 or 6 m.y.b.p. to 4 m.y.b.p. These estimated ranges suggest that the stage-two average uplift rates were between 6.5 and 21 times as great as the 0.07 mm/yr average maximum stage-one uplift rate in the same area of the CLEW (Reidel and others, 1980). We estimate the average uplift rate for stage-two folding of the Rattlesnake Hills anticline to be 0.14 to 0.20 mm/yr. This estimate is based on the 885 m elevation difference between the Elephant Mountain-Pomona contact at the crest of the Rattlesnake Hills anticline near the Yakima-Benton county line and at the Yakima River to the south (Bond and others, 1978) and on time periods of 10.5 to 4 m.y.b.p. (Elephant Mountain to pre-Thorp Gravel) and 8.5 to 4 m.y.b.p. (Ice Harbor to pre-Thorp Gravel).

Rattlesnake Lineament.—From the eastern end of the Rattlesnake Hills anticline to the Wallula Gap area a series of structures follows a linear N 45° to 50° W trend (Newcomb, 1970; Bond and others, 1978). This line of structures is the Rattlesnake Lineament (fig. 5). From northwest to southeast, these structures include a steeply dipping reverse fault that changes laterally to a plunging anticline, and doubly plunging anticlines (elongated domes), some of which are associated with steeply dipping reverse faults (Shannon & Wilson, 1977; Bond and others, 1978). Davis (1981, p. 32) cited unpublished aeromagnetic data which are consistent with a steep, north-side-down fault beneath and between some domes. A lower limit for the onset of the deformation that created these structures is set by the inclusion of previously undeformed Ice Harbor Basalt (8.5 m.y.b.p.) in the Rattlesnake Lineament structures (Bond and others, 1978).

The difference in basement influence between the wide zone of asymmetric anticlines of the Yakima Ridges and the linearly focused, reverse faults and elongated domes along the Rattlesnake Lineament may be due to (1) a wider sub-basalt zone of discontinuity under the Yakima Ridges and (2) dip-slip movement beneath the Rattlesnake Lineament but strike-slip movement, with or without dip-slip, beneath the Yakima Ridges. Also, a wedge of Cascade volcanic and volcanoclastic rocks under the Columbia River Basalt in the Yakima Ridges region may have affected the surface manifestation of a mobile zone at depth. Regardless, the deformation that produced the present configuration (degree of deformation, orientation, and location) of the CLEW structures occurred concurrently with the formation of the east-trending folds and thrust faults in the northern portion of the plateau and with the Horse Heaven Hills and Dalles-Umatilla structures to the south (see below). A north-south-oriented

$\sigma_{H \max}$ is required to create the structures north and south of the structures along the OWL; conversely, a mechanism involving several different regional stress orientations to deform the structures along the OWL simultaneously with the structures north and south of the OWL is difficult to imagine.

Horse Heaven Hills structure.—Basaltic river gravels atop the Horse Heaven Hills (HHH) confirm the low elevation of that area in stage one, or early Saddle Mountains time (Bond and others, 1978; Bentley, Anderson, and others, 1980). Swanson and Wright (1978) noted that the water gap (Wallula Gap) through the HHH began to form 12 to 8 m.y.b.p.; inclusion of Ice Harbor Basalt in the eastern segment of the HHH structure indicates deformation began after 8.5 m.y.b.p. there (Bond and others, 1978). The HHH structure consists of three anticlinal segments with different orientations (fig. 7). On the east is a northwest-trending anticline, and in the center is a northeast-trending anticline; both of these anticlines dip more steeply on their north than south limbs (Bond and others, 1978). The western segment of the HHH structure trends east-west between Bickleton and the Cascade Mountains (Newcomb, 1967, 1970; Swanson, Anderson, and others, 1979, 1981).

The north-pointing salient of the HHH structure east of Bickleton exhibits monoclinic symmetry about a vertical, north-trending symmetry plane which bisects the intersection of the northeast- and northwest-trending anticlinal segments (fig. 7). The east-west components of length of the northwest- and northeast-trending anticlinal segments are com-

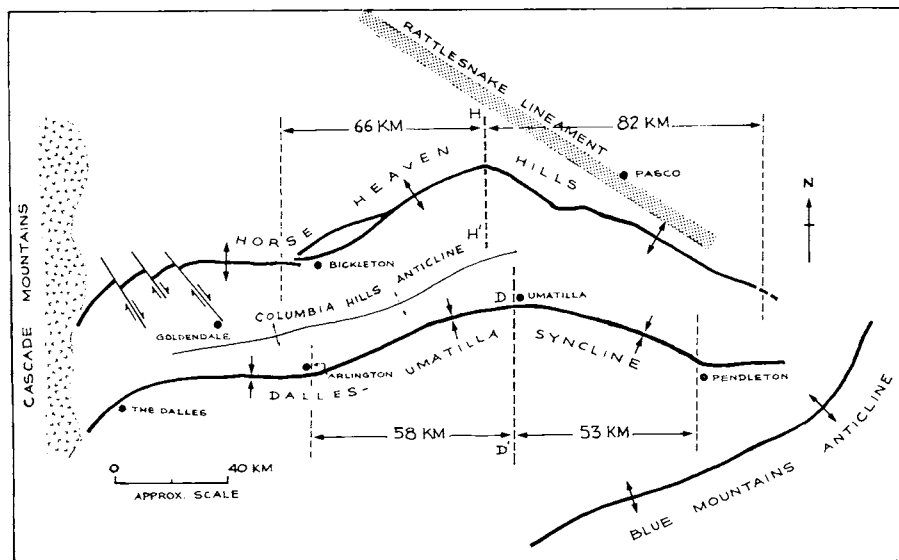


Fig. 7. Generalized map of structural symmetry and geometrical relationships of the Horse Heaven Hills structure and The Dalles-Umatilla syncline. H-H' and D-D' are vertical, north-south-oriented symmetry planes of the monoclinic salients of the Horse Heaven Hills anticline and The Dalles-Umatilla syncline, respectively. Map after Newcomb, 1967.

rable in magnitude. This style of structurally symmetrical deformation is compatible with a north-south-oriented $\sigma_{H \max}$ (Billings, 1946). The arcuate shape and the steeper north limbs indicate relative northward translation for the structure as a whole.

A possible explanation for the arcuate shape of the HHH structure is derived from the proximity of the northwest-trending eastern segment of the structure to the Rattlesnake Lineament portion of the OWL of approximately the same orientation (Newcomb, 1970; Myers and Price, 1979, pl. II-20). We suggest that a buttressing effect at depth across steeply dipping structures along the Rattlesnake Lineament (Bond and others, 1978; Davis, 1981) influenced the initial deformation pattern at the surface, and that the lateral continuity of the basalt flows allowed HHH area to respond as a continuous unit with similar volumes of rock displaced east and west of the symmetry plane.

Price (1982) included the HHH as one of the folds generated by the buttressing effect of a northwest-trending, material and mechanical discontinuity near to, but northeast of the OWL (fig. 6). However, Price's model predicts dextral shear along the northwest-trending eastern segment of the HHH and within the Rattlesnake Lineament. There is little evidence to support lateral slip along either of these zones (Shannon & Wilson, 1977 and ms 1979b; Bond and others, 1978), although lateral slip has occurred along trend (from the Wallula Gap area to the southeast) where the structural style also changes from folds to en echelon normal faults (Shannon & Wilson, ms 1979a and ms 1979b; Farooqui, 1980; Davis, 1981) at the southeastern termination of the OWL (Kendall, Dale, and Davis, 1981).

The western, east-trending (Bickleton to Cascades), anticlinal segment of the HHH structure is offset by north-northwest-trending, right-lateral strike-slip faults which were active during HHH folding. These faults also occur southward to the Columbia River and offset other fold structures including the Columbia Hills anticline. The age range of these faults falls within the period of about 10 to 4 or 5 m.y.b.p. based on the ages of faulted units and on unfaulted Simcoe volcanics (3.5 or 4.5 m.y.b.p.) which overlie some of these faults (Shannon & Wilson, 1977; Bentley and Anderson, 1979; Hammond, 1979; Bentley, Anderson, and others, 1980; Bentley, Powell, and others, 1980). The east trend of the HHH structure to the west of Bickleton (beyond the influence of the monoclinic salient) and the associated north-northwest-trending faults are in accordance with a north-south-oriented $\sigma_{H \max}$.

The Dalles-Umatilla syncline and Columbia Hills anticline.—The Dalles-Umatilla syncline follows a pattern approximately parallel to the HHH structure; the Columbia Hills anticline closely parallels the western two-thirds of The Dalles-Umatilla syncline (fig. 7). The hinge trace of The Dalles-Umatilla syncline trends approximately east between The Dalles on the west and Arlington on the east; this segment closely parallels the east-trending anticlinal segment of the HHH structure west of Bickleton. East of Arlington The Dalles-Umatilla syncline follows a

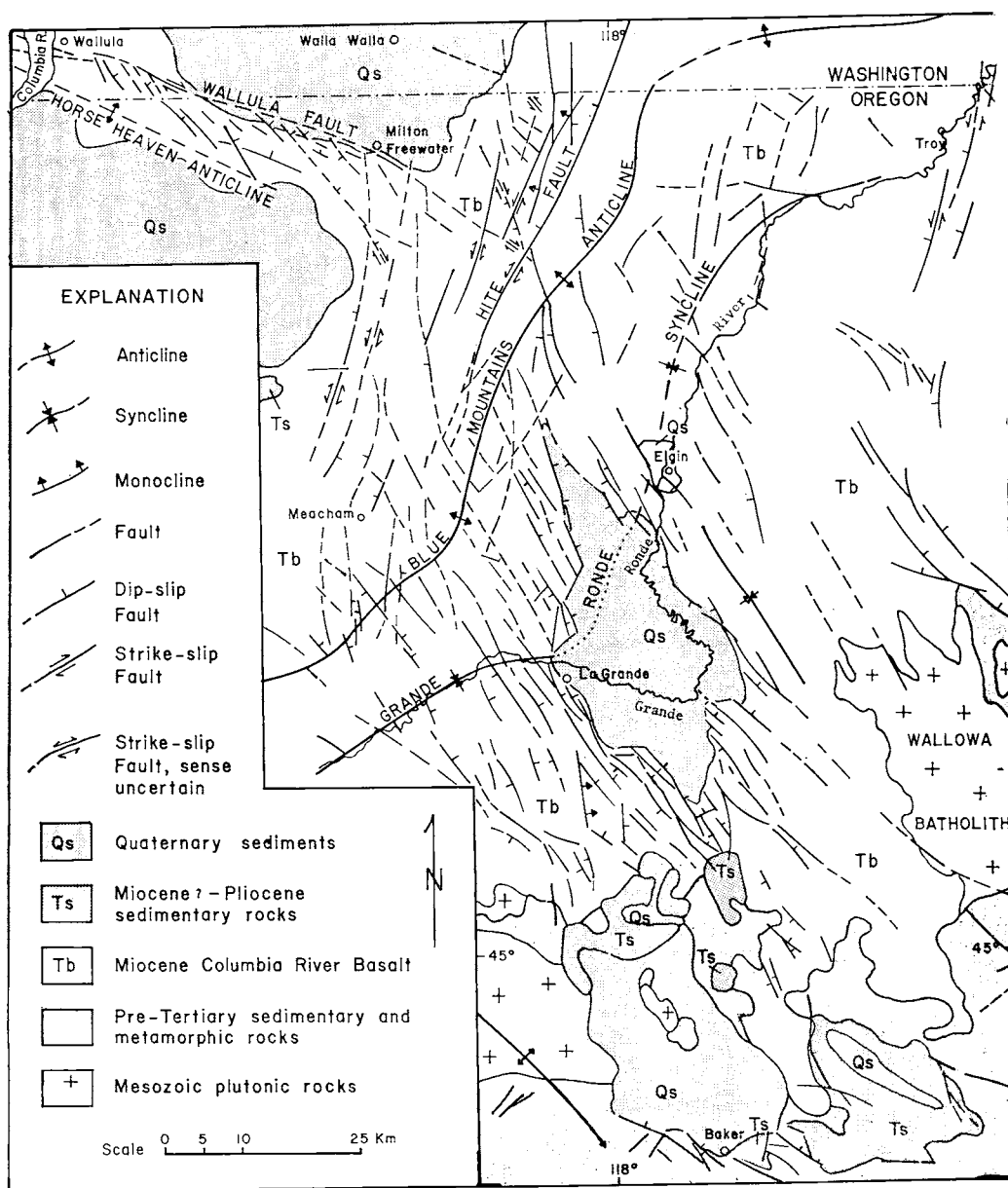


Fig. 8. Geologic map of the eastern Blue Mountains area (after Walker, 1977, 1979; Ross, ms; Shannor & Wilson, ms 1979a; Barrash and others, 1980; Swanson and others, 1981).

northward-convex arc with the point of maximum curvature slightly east of the north-south symmetry plane that bisects the HHH salient. As a whole, the HHH structure and The Dalles-Umatilla syncline form an anticlinal uplift and synclinal downwarp couple created concurrently in response to north-south compression.

Hogenson (1964) and Newcomb (1967) stated that the main warping of The Dalles-Umatilla syncline post-dated the initial (stage-one) warping of the Blue Mountains anticline. They noted that deformation of The Dalles-Umatilla syncline and HHH structure was concurrent with uplift of the Blue Mountains (stage two), and that the folding lasted until mid-late Pleistocene time based on fossils from Ringold sediments in the Pasco Basin. However, recent paleontological (Gustafson, 1978) and paleomagnetic (Packer and Johnston, 1979) work places upper Ringold time within the 5 to 3 m.y.b.p. period in the Pasco Basin.

Blue Mountains region.—Stage-two deformation in the Blue Mountains region resulted primarily in uplift of the Blue Mountains anticlinorium and in extensional faulting within and east of the uplift (fig. 8). Down-faulting of the LaGrande graben occurred concurrently with (1) major uplift of the Blue Mountains (Barrash and others, 1980), (2) uplift of the Wallowa Mountains to the east (Livingston, 1928; Robyn and Hoover, 1982), and (3) possibly with the uplift of the Elkhorn Mountains to the southwest (Gilluly, 1937). Shear zones along the northwestern margin of the Blue Mountains may have been active during stage two. Post-stage-one deformation along trend of the Blue Mountains is clearly demonstrated as far east as the Washington-Idaho border by pronounced east-west-trending structures within the Lewiston Basin (Bond, 1963; Camp, ms; Camp and Hooper, 1981).

West of LaGrande the Grande Ronde River closely follows the hinge of the Grande Ronde syncline and flows in incised meanders (fig. 8). As mentioned previously, the river found its consequent meandering course in the broad Grande Ronde syncline during stage one. In stage two the meandering course was incised during uplift of the Blue Mountains and lowering of local base level by depression of the LaGrande graben. In the Grande Ronde Canyon area near the Washington-Oregon border, Ross (1980) noted that the Grande Ronde River was able to maintain or re-establish meandering during early stages of folding and following the eruption of post-10.5 m.y.b.p. flows in that area (stage one). With later rapid uplift of the Blue Mountains (stage two), the meandering Grande Ronde River entrenched its course (Ross, 1980).

Extensional faulting within and east of the Blue Mountains uplift (Hampton and Brown, 1964; Hogenson, 1964) commonly occurred on re-activated north-northwest-trending extension fractures (Venkatakrishnan and others, in preparation) parallel to dikes that were established during stage one (Shannon & Wilson, ms 1979a, fig. 2; Swanson and others, 1981). The LaGrande graben is largely bounded by high-angle north-northwest-trending faults with vertical displacement occurring after the emplacement of Columbia River Basalt and, in places, after emplacement

of 7.3 to 6.5 m.y.b.p. intermediate volcanic units (Shannon & Wilson, ms 1979a). West and north of the LaGrande graben, structural blocks that contain numerous step-faulted slices form components of the larger Blue Mountains uplift. These structural blocks are bounded by nearly vertical north-northwest-trending dip-slip faults (Hampton and Brown, 1964; Swanson and others, 1981) and by north-trending monoclinical and normal fault structures that overprint the northwest trend (Shannon & Wilson, ms 1979a; Barrash and others, 1980).

Vertical offset also occurred during stage two in the Grande Ronde Canyon area north of the LaGrande graben. Ross (ms) noted that strike slip occurred on a shear-couple fault system in stage one (prior to 10.5 m.y.b.p.); he also reported post-10.5 m.y.b.p. dip-slip movement on some of the shears and normal movement on new faults in this area (fig. 9).

In his study of the Umatilla River basin, Hogenson (1964) noted the presence of a broad trough, the Meacham Creek trough, that cross cuts the anticlinal axis of the Blue Mountains uplift. The trough is an extensional feature that post-dates basalt eruption in this area; faults of the trough trend north-northwest and north (fig. 8).

Uplift of the Wallowa Mountains followed basalt extrusion and occurred across northwest-trending faults on the northern, western, and southern sides of the uplift; the western side of the Wallowa Mountains uplift is the eastern boundary of the LaGrande graben. Furthermore, northwest-trending faults disrupted pre-existing drainages at the southern margin of the Wallowa Mountains uplift and the LaGrande graben (Livingston, 1928).

Although the LaGrande graben has a north-northwest-trending grain imparted by boundary faults, it is one of four such grabens that together form a north-trending zone of post-Columbia River Basalt intermontane basins (fig. 9; Walker, 1977). In particular, Baker Valley is similar to the LaGrande graben in dimensions and orientation and perhaps in time and mode of formation. Gilluly (1937) attributed the formation of Baker Valley to vertical crustal adjustments which produced a combination of folding and faulting, and he suggested that the valley is probably bounded almost entirely by dip-slip faults. The string of basins along a north trend suggests that this area may be a northern extremity of Basin and Range-type deformation (Thayer and Wagner, 1969; Lawrence, 1976). This north-trending string of basins may be explained with σ_1 vertical and σ_2 (or $\sigma_{H \max}$) oriented north-south in the plane of Earth's surface.

Swanson and Wright (1981) observed that the greatest diversity in composition of volcanic products in the Columbia Plateau occurs in northeastern Oregon. They speculated that this variety was related to extensional tectonism (north-south and northwest-southeast normal faulting). This concept is also consistent with the localization of middle to late Miocene platy andesite source areas at the margins of the LaGrande graben.

Strike-slip displacement in the eastern Blue Mountains and the east-central portion of the Columbia Plateau.—Approximately horizontal slickensides have been recognized on some north-northwest trending verti-

cal faults and dikes north and south of the Snake and Columbia rivers (Taubeneck, 1978), in the west-central plateau (Bentley, Anderson, and Farooqui, 1980), and on north-northeast-, west-northwest-, and north-northwest-trending structures in the eastern Blue Mountains region (Shannon & Wilson, ms 1979a; Gehrels, White, and Davis, 1980). The slickensides have been interpreted as evidence for strike-slip movement. Shannon & Wilson (ms 1979a) and Gehrels, White, and Davis (1980) proposed that the LaGrande graben formed as a pull-apart basin within a dextral shear zone. This model is based on the N 30° W orientation and en echelon configuration of the majority of the basin-bounding faults and on the presence of dominantly sub-horizontal slickensides which are overprinted by vertical slickensides in some of the faults. However, we believe the LaGrande graben formed by extension between the rapidly uplifting Blue Mountains and Wallowa Mountains because several lines of evidence (see above) suggest: (1) the stage-two north-south compressive stress field dominantly affected the eastern Blue Mountains region with a verti-

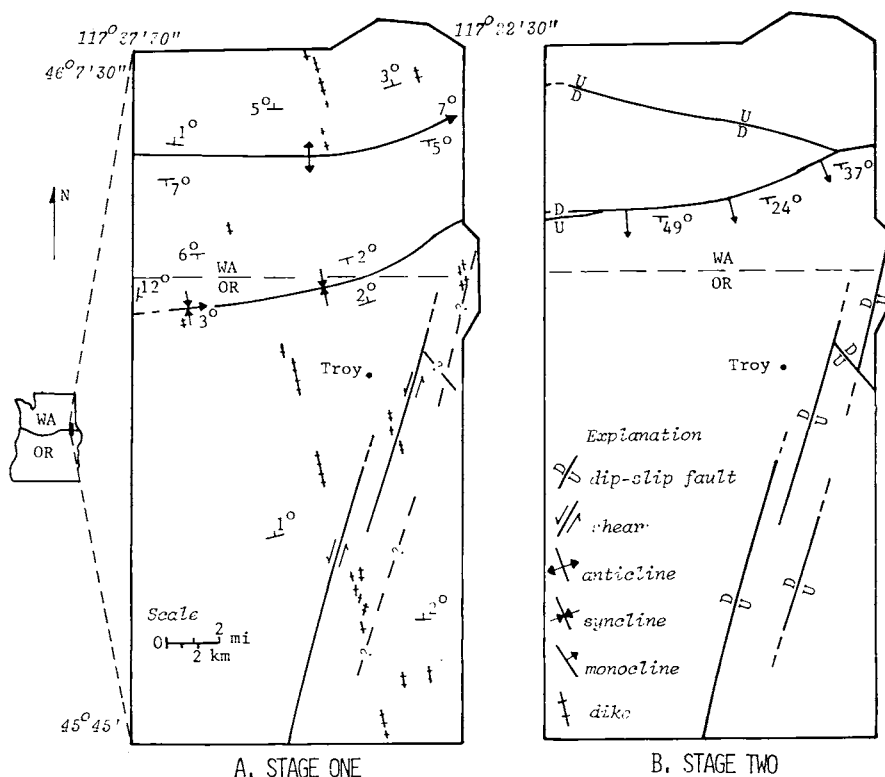


Fig. 9. Deformation in the Grande Ronde Canyon area changed style from stage one to stage two. (A) Stage one was characterized by open folding, dike emplacement, and shear development conjugate about dike orientation. (B) Stage two (post-10.5 m.y.b.p.) was characterized by vertical adjustments such as regional uplift, dip-slip movement on stage-one shears, and formation of new (steep) monoclinial and dip-slip fault structures. Maps after Ross (ms).

cal maximum compressive stress, (2) pre-existing (stage-one) structures accommodated stage-two dip-slip offset, and (3) new structures produced during stage two were predominantly tensional in origin.

The LaGrande area shows evidence that significant dip-slip movement occurred across north-northwest-trending, predominantly vertical structures. Because vertical displacement predominated in this area (Hampton and Brown, 1964; Shannon & Wilson, ms 1979a, p. 41-49; Barrash and others, 1980; Swanson and others, 1981), and because dip-slip predominantly occurred across vertical faults (only a few new north-trending 60°-70°-dipping normal faults were created) (Shannon & Wilson, ms 1979a; Barrash and others, 1980), the horizontal slickensides in the LaGrande area may indicate that limited right-lateral strike-slip movement accommodated extension that could not be accommodated across the vertical dip-slip faults. This kinematic relationship may be expected in a north-south-oriented stress regime where $\sigma_{H \max}$ is σ_2 (not σ_1), and where pre-existing north-northwest-trending vertical fractures accommodate the lateral strain (Bott, 1959; Price, 1966, p. 137 and 141). In our model, to the north, east, and west of the eastern Blue Mountains region, stage-two extension perpendicular to $\sigma_{H \max}$ occurred across pre-existing north-northwest-trending extension joints and dikes which were favorably oriented for limited, distributive right-lateral shear (Bott, 1959; Price, 1966, p. 137 and 141-142) when the orientation of $\sigma_{H \max}$ changed from north-northwest-south-southeast (stage one) to north-south (stage two).

It should be observed that strike-slip displacement has occurred across some north-northeast- and northwest-trending faults in or near the Blue Mountains region since stage one (fig. 8). However, the magnitude, sense of motion, timing, and effects of strike slip are disputed. Shannon & Wilson (ms 1979a and b) and Farooqui (1980) presented evidence for right-lateral oblique slip on the Wallula fault system northwest of the Blue Mountains. Kienle and Hamill (1980) reported 1.7 to 3.5 km of *right-lateral* displacement on the north-northeast-trending Hite fault system. This offset was largely prior to more limited right-lateral strike slip on some faults of the north-northwest- and west-northwest-trending LaGrande and Wallula fault systems which, in turn, have not laterally offset the Hite system noticeably (Shannon & Wilson, ms 1979a). Kendall, Dale, and Davis (1981) concluded that motion across the Hite system was *left-lateral* and that the Hite and LaGrande systems were conjugate shears with limited simultaneous movement. However, other recent mapping between the LaGrande graben and the Hite fault system does not support the presence of a strike-slip fault zone (LaGrande system) there (Swanson and others, 1981).

We suggest that the orientation of maximum compressive stress was transitional across the Blue Mountains uplift during stage two. In this model, maximum compressive stress was horizontal and north-south directed to the north and west of the Blue Mountains uplift, but maximum compressive stress was vertical to the east and south of and within much of the Blue Mountain uplift. This model is compatible with the conjugate shear interpretation for the Hite and LaGrande fault systems at the

northwestern margin of the Blue Mountains uplift (Kendall, Dale, and Davis, 1981) and with right-lateral oblique slip on the Wallula fault system (Shannon & Wilson, ms 1979a and b; Farooqui, 1980). However, more precise dating of geologic events and documentation of sequences and *magnitudes* of different types of offset will be needed to clarify relationships in this portion of the plateau.

John Day region.—Along the southern margin of the Columbia Plateau a transition occurs between different structural provinces. The plateau north of the Columbia Arc is dominated by compressional fold structures, and southern Oregon is dominated by Basin and Range faults (Fuller and Waters, 1929; Thayer, 1957). North and northeast of the John Day fault, folds change trend from east to northwest as they enter the structural transition zone (Thayer, 1957; Brown and Thayer, 1966; Walker, 1977). Farther to the southeast along fold trends, the folds are broken into north-northwest-trending block-faulted segments; Robyn, Hoover, and Thayer (1977) interpreted the north-northwest-trending normal faults to be related to north-south compressional deformation. Thayer (1957) interpreted the overall progressive change of structural orientation and style to the east and southeast to be analogous to the more abrupt eastward change in the John Day fault from east-trending, north-directed reverse faults to the north-northwest-trending strike- and dip-slip faults (fig. 4).

Thayer (1957) stated that folds in the Columbia Plateau and faults in the John Day area were contemporaneous and formed in response to north-south compression. A period of strong north-south compression in the John Day area began at 10 m.y.b.p. (Robyn, 1979) as indicated by formation of northwest- and northeast-trending conjugate shears and the east-trending John Day (reverse) fault (Robyn and Hoover, 1982). In the Mitchell, Oregon area to the west, stage-one folds were uplifted across east-trending step faults following a post-folding erosion period (Oles and Enlows, 1971) but prior to Rattlesnake Tuff time, or 6.4 m.y.b.p. (Robyn, Hoover, and Thayer, 1977). The John Day fault was active both before and after emplacement of the Rattlesnake Tuff (Thayer, 1957; Thayer and Brown, 1966). Thus stage-two deformation along the southern margin of the Columbia Plateau began about 10 m.y.b.p. and continued until about 6 m.y.b.p. in the John Day area (Robyn and Hoover, 1982), although uplift in the Mitchell area had been accomplished by 6.4 m.y.b.p.

STRUCTURAL STAGE THREE: NORTH-SOUTH COMPRESSION DURING THE PERIOD ~ 4 M.Y.B.P. TO PRESENT

Stage three began about 4 m.y.b.p. with the cessation or marked waning of folding and associated deformation in the western half of the Columbia Plateau. The lack of significant post-4 m.y.b.p. folding is demonstrated by the unfolded nature of Thorp and post-Ringold Pliocene, Pleistocene, and Holocene sediments (Shannon & Wilson, 1977) and the general absence of seismicity along major structures (Myers and Price, 1979). Although stage three $\sigma_{II \max}$ orientation is the same as that of stage two, the deformation pattern is noticeably different in the western half

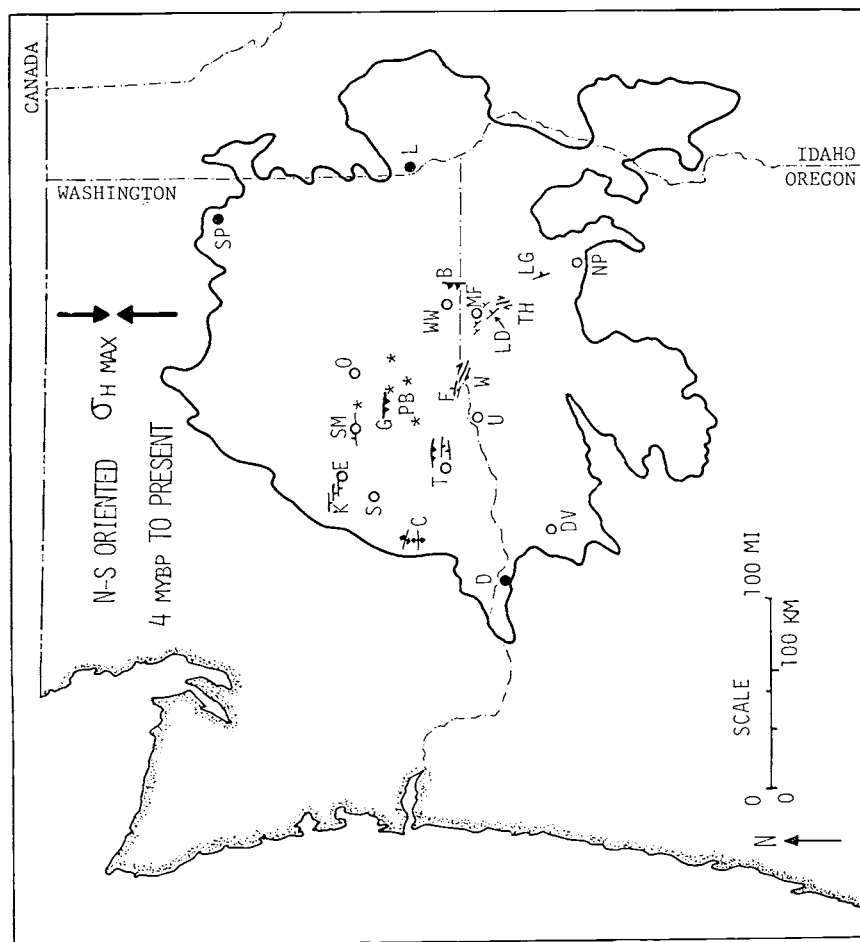


Fig. 10. Generalized map of stage-three structures and seismicity in the Columbia Plateau. Centripetal subsidence continued during stage three. Descriptions of structures and seismic events are given in tables 1 and 2, respectively. L, Lewiston; NP, North Powder; LG, LaGrande; MF, Milton-Freewater; WW, Walla Walla; B, Buroker; D, The Dalles; DV, Deschutes Valley; U, Umatilla; TH, Thorn Hollow; W, Wallula Gap; F, Finley Quarry; SM, Saddle Mountains; O, Othello; S, Selah; SP, Spokane; T, Toppenish Ridge; PB, Pasco Basin; G, Gable Mountain; C, Cowboy Parking Lot and Sedge Ridge anticlines, LD, Little Dry Creek; K, Kittitas Valley; E, Ellensburg.

of the plateau and less clearly so in the eastern Blue Mountains region (fig. 10).

The U.S. Department of Energy and the Washington Public Power Supply System have sponsored geologic studies to determine the nature and timing of regional deformation and the characteristics of individual structures, and to interpret and monitor seismic activity in the Columbia Plateau. Recent reviews by Myers and Price (1979) and Davis (1981), and work cited therein, discuss evidence for and against the presence of stage-three structural features. Table 1 is an attempt to list and briefly describe these features. Seismic activity is similarly listed in table 2. Both surface deformation and seismic activity are located schematically on figure 10.

Regardless of the debate over the existence or exact nature of some listed structures or seismic events, collectively these features suggest to us a deformation pattern consistent with a regional north-south-oriented $\sigma_{H \max}$ stress regime (see also Couch and Lowell, 1971; Smith and Lindh, 1978), although not uniquely so (Davis, 1981). Also, overall strain across stage-three structures indicates that post-4 m.y.b.p. deformation has been markedly less intense than during stage two. Folding since 4 m.y.b.p. is minor or questionable. Earthquake swarms within, north, and east of the Pasco Basin are not primarily associated with major stage-two structures in that area.

At present there is no evidence for or against a deformational hiatus or change in strain rate at about 4 m.y.b.p. in the eastern Blue Mountains region. Quaternary faulting and recent seismicity associated with the Walhalla fault system appear to be well documented. Also, minor Quaternary movement in the Hite fault system and at the margin of the LaGrande graben seems consistent with the style of stage-two deformation in this region.

Neither widespread nor major surface deformation appears to have occurred in the Columbia Plateau during stage three. Much of the stage-three tectonic activity is Recent but would not have been recognized without modern instrumentation, historical records, or detailed field work. It is not clear whether the above-mentioned surface deformation and seismic activity represent elements of a continuum of less intensity than that of stage two or a new (< 2 m.y.b.p.) weak(?) tectonic cycle following a lull in activity at about 4 m.y.b.p.

REGIONAL TECTONIC PATTERNS

The timing of Columbia Plateau deformation stages may be related to major late Cenozoic tectonic events at the western margin of the North American plate (Barrash and Venkatakrishnan, 1982; Robyn and Hoover, 1982). Stage one (fig. 11A) began synchronously and on trend with the widespread thermal disturbance at 16 ± 1 m.y.b.p. which resulted in uplift, east-northeast-west-southwest-directed extensional deformation, and basaltic and bimodal volcanism in the Great Basin (McKee, 1971; Noble, 1972; Zoback and Thompson, 1978) and southern Idaho and Oregon (Armstrong, Leeman, and Malde, 1975; MacLeod, Walker, and McKee, 1976). Calc-alkaline volcanic activity unrelated to subduction began at

TABLE 1
Stage three structures

Location	Description	Age estimate	Reference
Kittitas Valley	3 reverse faults with east trends, dextral en echelon pattern	within 3.7 to 0.13 m.y.b.p.	2, 6, 11, 12, 13, 17
Toppenish Ridge	nearly 100 surface ruptures; crestal normal faults, flank reverse and thrust faults; faults parallel ridge trend of N 90° E $\pm$ 30°	13000-500 y.b.p.	3, 5
Cowboy Parking Lot and Sedge Ridge anticlines	possible post-Simcoe volcanics minor uplift and tilting	late Pliocene to Pleistocene?	3
Saddle Mountains	east-trending dip-slip fault	Pleistocene	4, 6, 10, 11
Gable Mountain	2 south-dipping thrust faults cut Missoula-age flood gravels	late Pleistocene	4, 6, 7, 11
Wallula Gap	northwest-trending dextral shears offset Palouse and Touchet Formations and post-Touchet loess	late Pleistocene, 13000-7000 y.b.p.	4, 6, 14
Finley Quarry	2 faults; north fault attitude N 90° E 55° S, south fault attitude N 65° W 90° E; dip-slip offset	late Pleistocene, pre-7000 y.b.p.	6, 16
Milton-Freewater	4 northwest-trending normal (down to north) faults offset Touchet beds	late Pleistocene	6, 15
Buroker	north-trending, 26° W-dipping reverse fault offsets Palouse Formation less than 1 meter	Pleistocene	6, 11, 12, 16
Little Dry Creek	N 20° to 35° W 75° NE dip-slip fault drops basalt and Palouse Formation about 0.5 meter	Pleistocene	6, 15
Thorn Hollow	N 15° to 5° W, 85° W fault, near horizontal slickenside strike, possible Palouse Formation offset	late Pleistocene	15
LaGrande	fault contact between alluvium and colluvium on trend with north-northwest-trending fault	late Pleistocene Holocene?	1, 8, 9

(1) Barrash and others, 1980; (2) Bentley, 1977; (3) Bentley, Anderson, and others, 1980; (4) Bingham and others, 1970; (5) Campbell and Bentley, 1981; (6) Davis, 1981; (7) Fecht, 1978; (8) Gehrels and others, 1980; (9) Hampton and Brown, 1964; (10) Howard and others, 1978; (11) Myers and Price, 1979; (12) Rigby and Othberg, 1979; (13) Shannon & Wilson, 1977; (14) Shannon & Wilson, ms 1979a; (15) Shannon & Wilson, ms 1979b; (16) Shannon & Wilson, 1980; (17) Waitt, 1979.

TABLE 2
Historical seismicity in the Columbia Plateau; major earthquakes
and recorded earthquakes since 1969 with magnitude >3 (Richter scale)

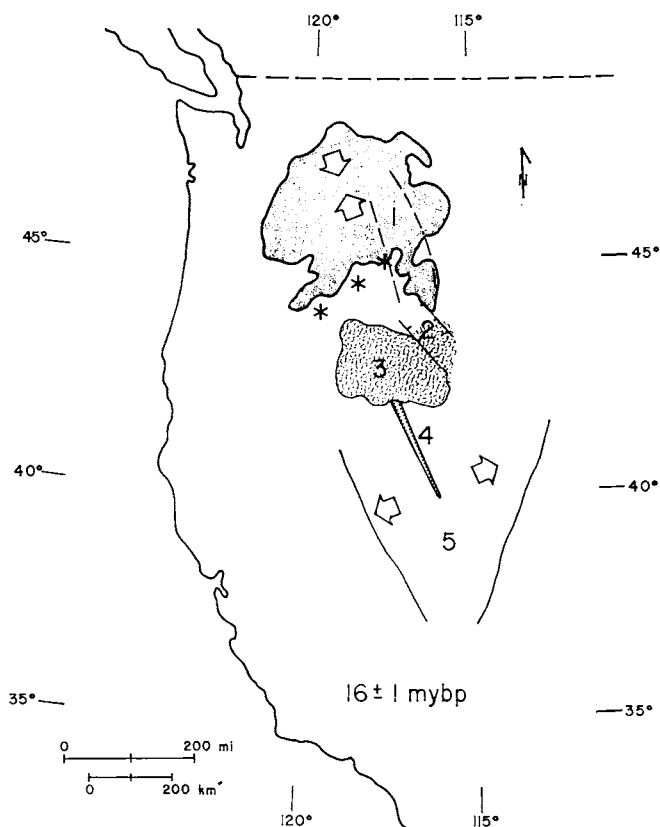
Location	Magnitude*	Year	Description	Reference
Milton-Freewater	R=5.75-6.1	1936	principal compression axis N 56° E 22° related to Wallula fault system related to Blue Mountain uplift-Hite fault system other data	10, 20 13 19 2, 4, 15, 16
Deschutes Valley	R=4.8	1976	reverse or thrust faulting across fault system with attitude N 72° ± 4° W 32° or 58° S, depth about 15 km	7, 9
Othello	R=4.1	1973	depth about 1 km	6
Pasco Basin area	R=0-4 esp. R<2		earthquake swarms, depth generally <3 km, reverse faulting across variable- to steep-dipping east-trending structures	2, 14, 15, 17
North Powder	R=3.9 R=3.6	1971 1969	depth about 33 km first motion consistent with right-lateral strike slip on northwest-trending fault, depth 32 ± 3 km	5 8
Toppenish Ridge	R=3.8	1981	either right-lateral offset on N 47° W 90° NE fault or left-lateral offset on N 43° E 70° NW fault, depth about 0.1 km	3
Walla Walla	R=3.2-4.1	1979	depth about 5 km	18
Selah	R=3.1-3.7 R=3.2-3.4	1979 1979	depth about 0 km depth about 5 km	18 18
Ellensburg	M=V	1934		15, 19
Umatilla	M=VI-VII	1893		1, 4, 15, 19
Saddle Mountains	M=IV-VI	1918		2, 4, 11, 12, 13, 15

Note: WPPSS (1974) presents a list of felt earthquakes through 1973.

* R is Richter scale, M is Modified Mercalli scale.

(1) Berg and Baker, 1963; (2) Bingham and others, 1970; (3) Campbell and Bentley, 1981; (4) Coffman and von Hake, 1973a; (5) Coffman and von Hake, 1973b; (6) Coffman and von Hake, 1975; (7) Coffman and Stover, 1978; (8) Couch and Whittset, 1969; (9) Couch and others, 1976; (10) Davis, 1981; (11) Fifer, 1966; (12) Jahns, 1967; (13) Jones and Deacon, 1966; (14) Malone and others, 1975; (15) Myers and Price, 1979; (16) Rasmussen, 1967; (17) Rothe, 1978; (18) Stover and von Hake, 1981; (19) WPPSS, 1974; (20) Woodward-Clyde Consultants, 1980.

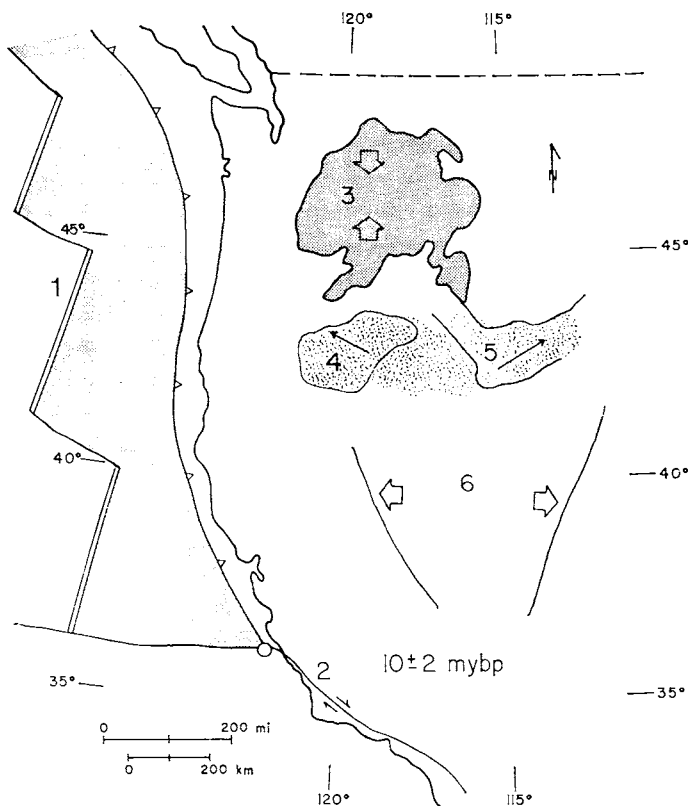
Fig. 11. Regional tectonic patterns.



A. Starting at 16 ± 1 m.y.b.p.: (1) eruption of Columbia River Basalt, (2) development of western Snake River Plain structural depression, (3) bimodal volcanism in southeastern Oregon and southwestern Idaho, (4) development of Northern Nevada rift, (5) uplift, tensional deformation and basaltic and bimodal volcanism in Great Basin. Asterisks locate areas of calc-alkaline volcanic activity which became active at 16 ± 1 m.y.b.p.

this time east of the Columbia Arc in eastern Oregon (Robyn, 1979; Robyn and Hoover, 1982).

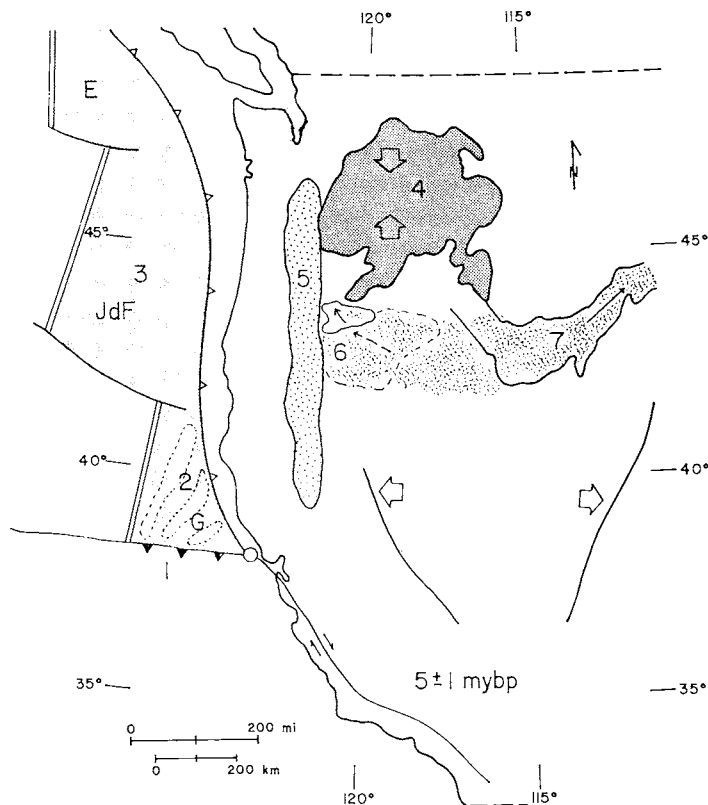
Stage two (fig. 11B) began when (1) the Pacific plate spreading direction rotated about 25° in a clockwise direction from east-west to southeast-northwest (Vine, 1966; Pitman and Hayes, 1968) and (2) relative motion between the North American and Pacific plates increased significantly, as measured by the slip rate across the San Andreas fault (Atwater and Molnar, 1973; Hein, 1973). At this time (10 ± 2 m.y.b.p.) the stress orientation in the Columbia Plateau and elsewhere in the western North American plate also rotated clockwise 20° to 45° from north-northwest-south-southeast- to north-south- and north-northeast-south-southwest-oriented regimes (Suppe, 1970; Smith, Mabey, and Eaton, 1976; Zoback and Thompson, 1978; Eaton, 1979). Also a relatively strong response to compression



B. Starting at 10 ± 2 m.y.b.p.: (1) change Pacific plate spreading direction, (2) increase relative motion across San Andreas fault, (3) change stress orientation and style of deformation in Columbia Plateau, (4) start west-northwestward migration of southern Oregon volcanism, (5) northeastward migration of Snake River Plain volcanism and increase in volcanic production, (6) change stress orientation in Great Basin.

started at this time in the western half and at the southern margin of the Columbia Plateau, and uplift occurred in the Blue Mountains (Columbia Arc). At 10 ± 2 m.y.b.p. silicic volcanic activity began migrating to the west-northwest in southern Oregon, and northeastward-migrating bimodal volcanic activity in the Snake River Plain increased production. Calc-alkaline volcanic activity in eastern Oregon ceased at this time (Robyn, 1979; Robyn and Hoover, 1982).

Stage three (fig. 11C) began about 4 m.y.b.p. in the Columbia Plateau. At this time (5 ± 1 m.y.b.p.) the regional stress orientation did not change, but deformation style and intensity and related volcanic activity were altered in tectonic subdomains at the western edge of the North American plate. In particular, in the Pacific Northwest region at 5 m.y.b.p. the Pacific plate began overthrusting the Gorda plate across the Mendocino fracture zone, and the Gorda plate began bending internally (Silver, 1971). In turn, between 5 and 4 m.y.b.p. the Juan de Fuca and Explorer plates slowed their convergence rates toward North America (Riddihough, 1977).



C. Starting at 5 ± 1 m.y.b.p.: (1) Pacific plate overthrusts Gorda Plate (G), (2) internal bending of Gorda plate, (3) slow convergence of Gorda, Juan de Fuca (JdF) and Explorer (E) plates toward North American plate, (4) change deformation style and intensity in Columbia Plateau, (5) change nature of volcanism in Cascade Arc, (6) change direction and rate of southern Oregon volcanism, (7) increase volcanic production in Snake River Plain.

At about 5 m.y.b.p. Cascade volcanism changed from dominantly intermediate calc-alkaline activity to dominantly high-alumina olivine basalt activity (McBirney, 1978; Hammond, 1979). Also, an increase in rate of volcanic production in the Cascades occurred at 5 ± 1 m.y.b.p. (McBirney and others, 1974; Kennett, McBirney, and Thunell, 1977). South of the Columbia Plateau, westward migrating silicic volcanism in central Oregon changed to a more northerly course and slowed its rate of migration at about 5 m.y.b.p. (Walker, 1974; MacLeod, Walker, and McKee, 1976).

We do not propose a mechanism to account for the change of structural style and intensity in the Columbia Plateau at about 4 m.y.b.p. However, in recognition of the approximately synchronous inter- and intraplate rearrangements in adjacent regions, we believe that the change in plateau behavior was part of the regional culmination in tectonic and volcanic activity at 5 ± 1 m.y.b.p.

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