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SEISMICITY AND LATE CENOZOIC FAULTING IN THE BROWNLEE DAM AREA-- OREGON - IDAHO: A PRELIMINARY REPORT

by

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ABSTRACT

At least 17 separate small to strong earthquakes have occurred in a relatively small area in or adjacent to part of the Snake River Canyon south of Hells Canyon, on the Idaho - Oregon border. Earthquake locations prior to 1964 are based on intensity data and the accuracy is poor. At least one shock in 1983 is well located and intensity patterns strongly reinforce the confidence in the accuracy of the instrumented location. The epicenter is in Idaho about 10 km south-southeast of Brownlee Dam. A cluster of well located microseismic earthquakes recorded in 1984 fall within 5 km of Brownlee Dam. In addition, large normal faults with a significant amount of Late Cenozoic displacement are located in this area. Moderate levels of seismicity and regional structural relations suggest that the area west, southwest, and around the dam site is an active pull-apart basin. Instrumented epicenter locations and intensity data from the 1981 - 1984 earthquake sequence suggest that two or more areas of subsurface faulting are active. This earthquake-prone area remains unmonitored by a local seismic net and further studies are necessary to evaluate the potential hazards of earthquakes and faulting to Brownlee Dam.

INTRODUCTION

The Cuddy Mountains of west-central Idaho are unique because four Paleozoic - Mesozoic tectonostratigraphic terranes (Brooks, 1979a; Dickinson, 1979; Silberling and others, 1984; Vallier, in press) of the accreted Blue Mountains Island Arc (BMIA) are juxtaposed in fault-bounded belts in a narrow eight-km-wide area. The terranes strike northeast along the west side of the Cuddy Mountains and are subsequently covered by the middle Miocene Columbia River Basalt Group. These terranes are highly metamorphosed and tectonized where they reappear at the West Idaho Suture Zone (WISZ) which forms the border between the Idaho Batholith and the accreted BMIA. The narrow and well-exposed belts along the western side of the Cuddy Mountains contain the most easterly exposures of rocks that are characteristic of the terranes. These exposures hold keys to unlocking many of the complex stratigraphic and structural problems encountered in the WISZ and other parts of the BMIA.

Original work in 1986 and 1987 focused on Mesozoic structural, stratigraphic, and petrologic investigations. During the 1987 field season it was recognized that a major terrane-bounding thrust fault, named the Cuddy Mountain fault in this report, had as much as 700 m (2000+ ft) of retrograde dip-slip offset of the Miocene basalt section, and could be a regionally extensive structure. Earthquake epicenter data for the area was obtained in early 1988 in an attempt to correlate the Cuddy Mountain fault reactivation with recent local seismicity. A suprising number of moderate to strong shocks have occurred in and adjacent to the Pine Valley - Cuddy Mountain area, which is surrounded by otherwise regional seismic quiescence. Furthermore, recent seismological investigations of Zollweg and Jacobson (1986), describe a discrete seismic zone in the area. Gaps in geologic mapping, coupled with deficiencies in regional and local Cenozoic structural relationships, rendered any specific correlations from seismicity to faulting inconclusive in their study. The focus of this research, therefore, was shifted to provide geologic data on the Late Cenozoic deformation which could help narrow structural correlations to patterns of seismicity. The Brownlee Dam and Cuddy Mountain 7.5 minute quadrangles (plates 1 and 2) were mapped in some detail, with attention focused on Late Cenozoic structure. Structural and seismicity data have been compiled for a broad area in an effort to isolate possible tectonic influences on the complexly deformed local area.

REGIONAL GEOLOGY

The location of this investigation is within the Blue Mountains physiographic subprovince of northeastern Oregon and west-central Idaho near the mutual boundary of the Basin-and-Range Province, the Snake River Plain, The Idaho Batholith, the Columbia Plateau, and the Blue Mountains Uplift (fig. 1). This area is also part of the Columbia Embayment, a postulated paleo-cratonic bight into which the allochthonous Blue Mountains Island Arc terranes were accreted.

The Paleozoic and Mesozoic rocks within the Blue Mountains province are divided into five tectonostratigraphic terranes (Brooks, 1979a; Dickinson, 1979; Silberling and others, 1984), four of which

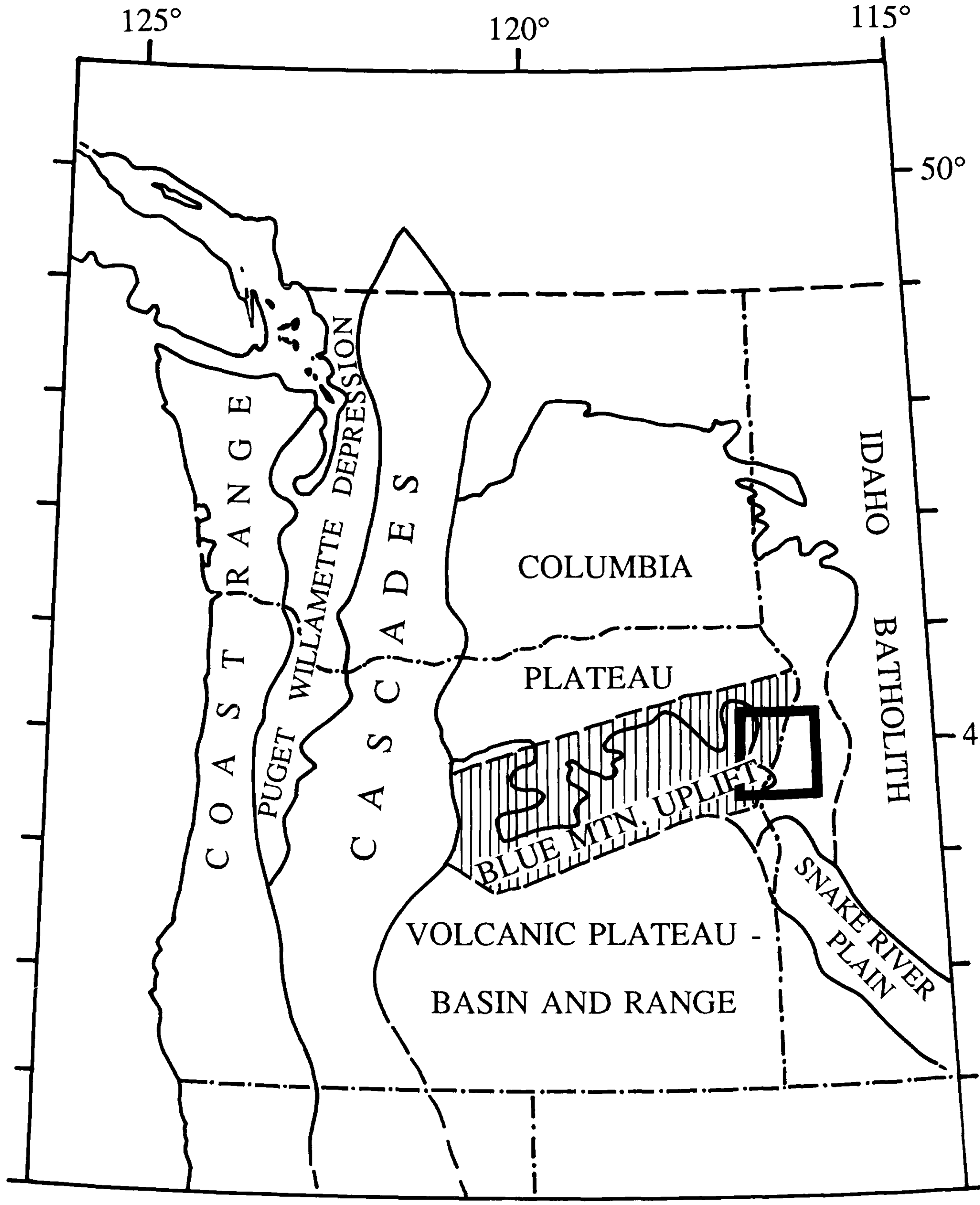

Figure 1 : Generalized phsiographic boundary map of the Pacific Northwest. Boxed area encloses region of this investigation. (modified from Hill, 1972)

are exposed in the Brownlee area. Rocks in these terranes evolved in an oceanic island arc and were subsequently accreted to North America in the Early Cretaceous. Rocks of the terranes crop out in broad east trending zones in eastern Oregon, and gradually thin and bend more northerly as they approach the WISZ (fig. 2). The terrane thinning phenomenon has been variously attributed to: (1) strike-slip slivering and northwest tectonic escape along an oblique convergence (Wernicke and Klepacki, 1988); (2) progressive eastward overthrusting during accretion approaching the region of greatest compression (Vallier, pers. comm., 1987), and (3) original arc morphology, possibly coupled with some combination of the above. Perhaps all of these factors played roles in the evolution of the present terrane morphology.

The Cuddy Mountains and Sturgill Peak-Hitt Mountain area of western Idaho are geographically and geologically independent from the Idaho Batholith and mountainous region to the east (fig. 3). The Long Valley fault system marks the transition zone between the BMIA accreted terranes and the Idaho Batholith, east of the Cuddy Mountain-Hitt Mountain area (plate 4). The ERTS satellite image (fig. 3) shows the location of the Cuddy Mountains, Pine Valley, Brownlee Reservoir, and other principal geographic entities referred to in this text.

The Tertiary Columbia River Basalt Group is extensive throughout the area, and is locally as much as 700 m (2000+ ft) thick. The older Mesozoic rocks are largely exposed by Late Cenozoic normal faulting and regional uplift, resulting in unroofing of the basalt cover, and subsequent exhumation of the underlying older strata. These basalts belong to the Weiser embayment (Fitzgerald, 1983), and are part of the southeasternmost extension of the Columbia Plateau (fig. 1).

The pattern of Late Cenozoic faulting in northeastern Oregon and extreme west-central Idaho is dominated by northwest-trending faults (plate 4). Lawrence (1976) proposed that strike-slip faulting on the Brothers Fault Zone and the Vale Zone terminated the Basin-and-Range Province (fig. 4, plate 4). These approximately N 50°W trending fault zones are typically manifested by discontinuous swarms of subparallel ***en echelon*** normal faults. A broad and somewhat segmented northwest-trending zone of fault-bounded pull-apart basins, ***en echelon*** normal faults, and intervening areas of uplift were found to strike into the study area (plate 4). This structural zone is referred to in this report as the Grande Ronde-Pine Valley zone (GPZ) (fig. 4, plate 4). The GPZ parallels other major northwest-trending fault zones in southeastern Oregon, western Nevada, and northeastern California that have been described as right-lateral strike-slip fault zones (Donath, 1962; Wise, 1963; Stewart and others, 1968; Pease, 1969; Lawrence, 1976; Walker and Nolf, 1981; Anderson and Hawkins, 1984). Similar tectonic mechanisms probably are involved in the Late Cenozoic deformation and recent seismicity in the Pine Valley-Cuddy Mountain area.

REGIONAL STRATIGRAPHY

The most inboard or northwesterly terrane is the Wallowa Terrane (plates 1, 2 and 3). It is an arc-axis igneous massif with a thick sedimentary and volcanic veneer. The igneous massif is largely composed of Permian to Triassic age plutons similar to those in the Sparta Complex (Vallier, in press). In the study area, diorite and quartz-diorite outcrops (PTrqd) occur in a small inlier in the Snake River Canyon below Brownlee Dam. These diorites are intruded by numerous fine-grained diabase dikes. In Wildhorse River Canyon, very similar plutonic rocks are in tectonic contact with Wallowa(?) Terrane carbonates and volcaniclastic breccias to the north, and Baker Terrane metasediments to the south (plate 1).

Locally, the Wallowa(?) Terrane sediments are comprised of a basal volcanic tuff with minor limestone lenses and conglomerate (Trvs) gradationally overlain by an approximately 350 m (1000 ft) thick limestone and carbonate flysch unit (Trl). These carbonate rocks are, in turn, depositionally overlain by a thick sequence of volcaniclastic breccia and conglomerate (Trvb) primarily composed of extrusive andesite and basalt clasts. Fossils (*Halobia sp.*) collected from the limestones indicate a Triassic (Norian) age (Silbering, written comm. to Vallier, 1986). This Triassic sedimentary sequence is informally referred to in this report as the Wildhorse Formation. These same three units comprising the Wildhorse Fm. are present as chaotic olistostrome in the inlier below Brownlee Dam, but are well exposed and relatively undeformed in Wildhorse River Canyon.

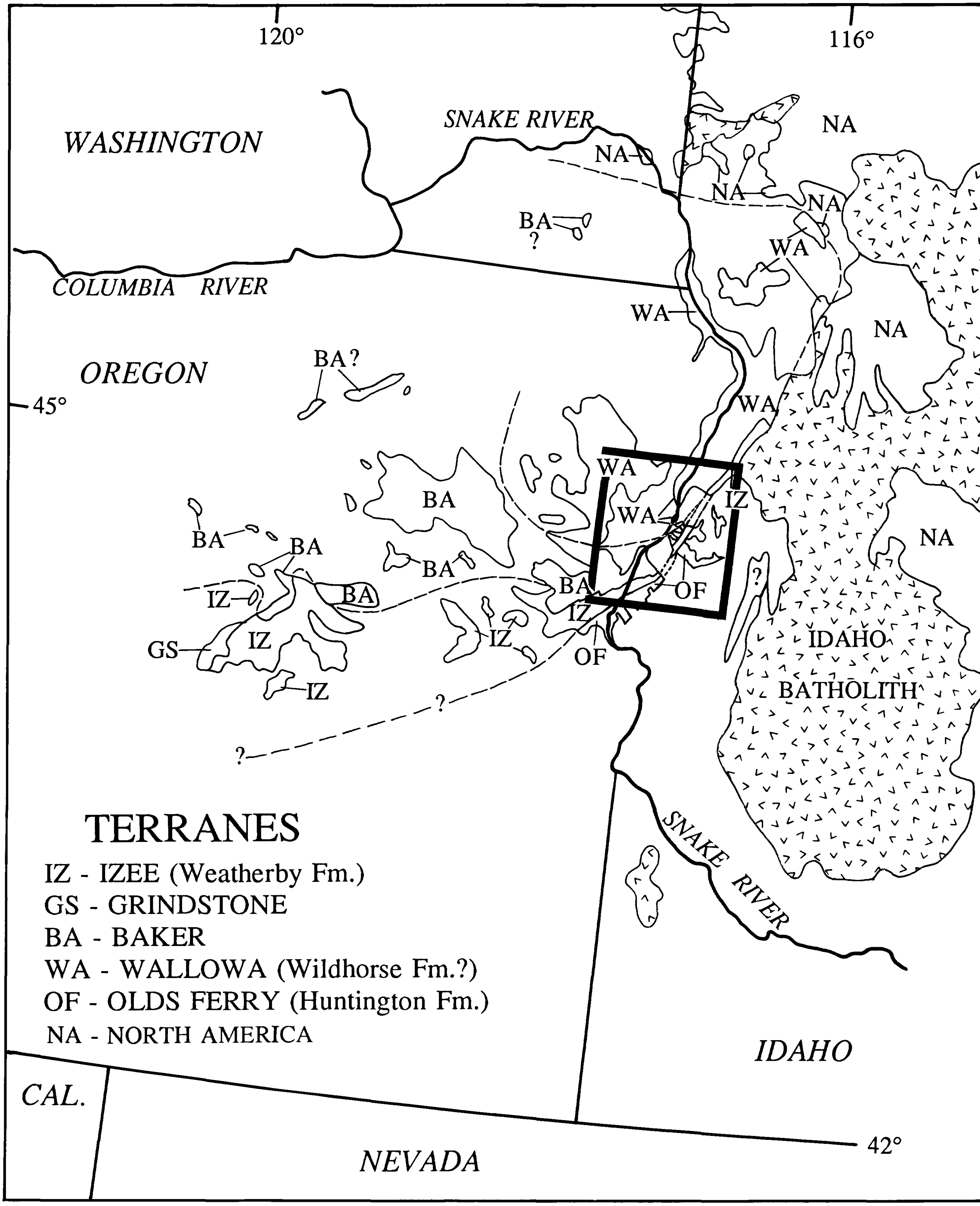

Figure 2 : Tectonostratigraphic terranes of the Blue Mountains Island Arc and adjacent North America cratonic rocks. Boxed area encloses region of this investigation. (from Vallier, in press)

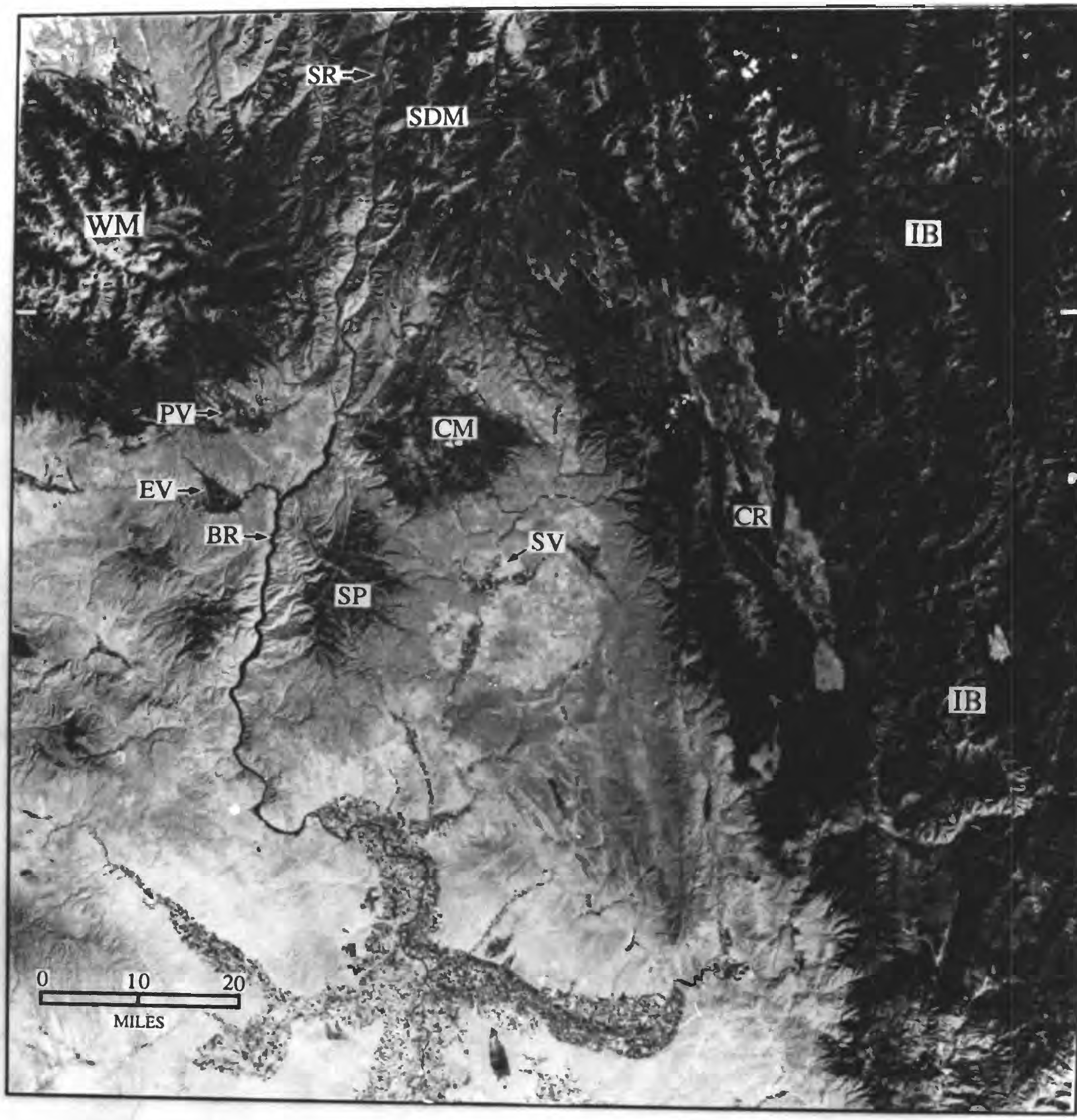

Figure 3 : ERTS satellite image recorded on September 11, 1973, of extreme east-central Oregon and west-central Idaho, showing principal geographic features of the region. BR = Brownlee Reservoir; CR = Cascade Reservoir; CM = Cuddy Mountains; EV = Eagle Valley; IB = Idaho Batholith Province; PV = Pine Valley; SDM = Seven Devils Mountains; SP = Sturgill Peak - Hitt Mountain area; SR = Snake River - Hells Canyon area; SV = Salubria Valley; and WM = Wallowa Mountains. Developed area at lower-center is part of Weiser, Idaho, and Ontario, Oregon.

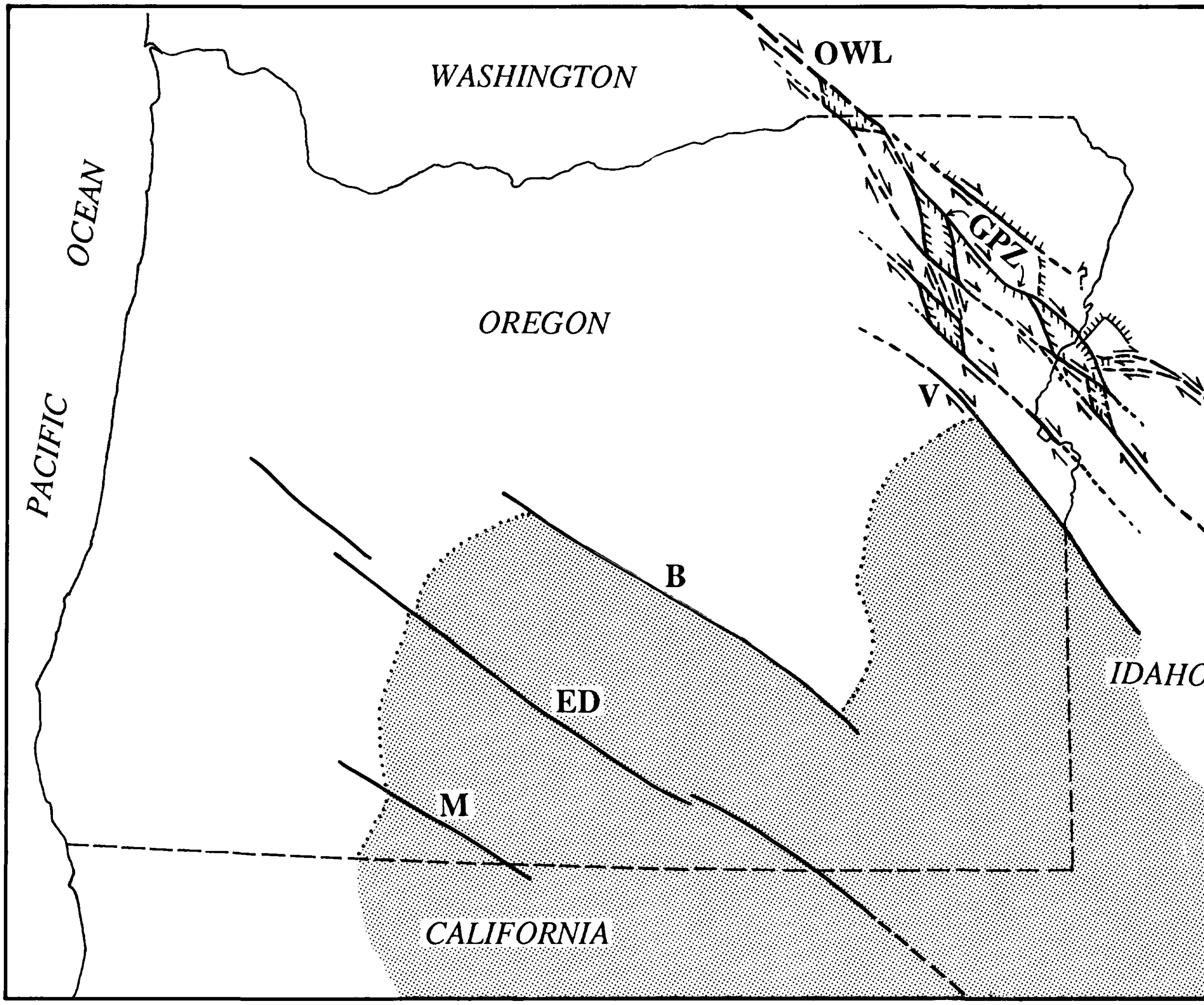

Figure 4 : Northwest limits of Basin and Range province (shaded) with major northwest trending right-lateral strike-slip fault zones. Tectonic basins (hachered interiors) and uplifts (hachered exteriors) are greatly generalized. M = McLoughlin zone, ED = Eugene - Denio zone, B = Brothers fault zone, V = Vale zone, and GPZ = Grande Ronde - Pine Valley zone (proposed), OWL = Olympic - Wallowa Lineament of Raisz (1945). (modified from Lawrence, 1976)

The rocks of the Baker Terrane (PzMzu) (plates 1 and 2) are among the most deformed and metamorphosed of the four terranes. This terrane has a very complicated history; some parts may be related to accretionary wedge and forearc sedimentary-basin deposits. Based on fossils collected in similar rocks in eastern Oregon, its age ranges from Devonian to Late Triassic (Mullen and Wardlow, 1986). In eastern Oregon, there are several areas of serpentinite-matrix melange, tectonized greenstones, and other deformed strata within the Baker Terrane. However, in the Cuddy Mountain area the internal structure is less chaotic, with deformation mainly restricted to low-angle bedding-plane faults, small-scale isoclinal folding, or slump folds. Some individual chert beds can be followed for about a mile.

The Baker Terrane rocks in the Cuddy Mountains are dominantly composed of massive chert, ribbon chert, cherty phyllite, phyllitic argillite, and cherty argillite. Recrystallized limestone and cherty limestone are common, although they tend to be laterally discontinous or lensoid in outcrop. Metagabbro has been identified in some locations. Because of the highly siliceous composition of the metasedimentary rocks the effects of metamorphism are not always pervasive; however metamorphic grade is typically of the greenschist facies in the less silicic units.

The Baker Terrane is in tectonic contact with the Weatherby Fm. (Izee Terrane of eastern Oregon) along a high-angle Mesozoic thrust fault (Cuddy Mountain fault). The Cuddy Mountain fault has experienced significant Late Cenozoic retrograde dip-slip displacement. A Jurassic age for the Weatherby Fm. (Js), is based on regional and local fossil identifications, especially ammonite fauna (Brooks, 1979b). This formation is composed primarily of sandstone and shale, some with flute casts and grooves suggesting deposition by turbidity currents. Sandstones frequently contain a large percentage of chert detritus suggesting erosion from the Baker Terrane. These rocks are postulated to be remnants of an intra-arc basin between the Baker (and Wallowa?) and Olds Ferry Terranes (Vallier, in press).

Several conglomerate units are present, particularly in the lower part of the Weatherby Fm. One conglomerate unit exhibits stretched clasts and much shearing in the matrix, which may be the result of low-angle faulting along its contact. This same horizon has been mapped approximately ten miles to the south in the Sturgill Peak area in a similar stratigraphic position (Hendricksen, 1975).

A recrystallized light-green andesitic tuff unit (Jv) was mapped within the Weatherby Fm. (plate 2). Some contacts are gradational and the lack of thermal alteration in bounding sediments suggests subaqeous deposition of an ash-fall tuff. Near the base of the Weatherby Fm. a rhyodacite crystal tuff unit is regionally extensive. It has been isotopically dated at 200 m.y. (C. Field, Oregon State Univ., written comm., 1988). Parts are both welded (ash-flow), and non-welded (air-fall).

A distinctive red and green conglomerate lies unconformably below the rhyodacite crystal tuff (Fankhauser, 1969; Bruce, 1971). Clasts are mainly of extrusive basaltic and andesitic composition, and were eroded from the underlying Huntington Fm. The tuffaceous matrix is in places highly sheared and talc-rich as a result of low-angle faulting and shear metamorphism along bedding planes. This unit is most likely correlative with the Jett Creek Member described near Lime, Oregon (Brooks, 1979c), although in Oregon according to Brooks (1979c) the Jett Creek Member is overlain by sedimentary Weatherby Fm. rocks. The rhyodacite crystal tuff and the basal red and green conglomerate are included in the Weatherby Fm. in this study.

The Weatherby Fm. overlies the Triassic Huntington Fm. in the southeast corner of the region (Brooks, 1979b), and is well exposed in the southeast corner of the Cuddy Mountain quadrangle (plate 2). Mapping by Slater (1969) and Bruce (1971) identified repeated sections along contacts of these two formations in the upper No Business Creek area (Cuddy Mtn. quad). They attributed these repetitions to low-angle reverse faulting adjacent to and along the unconformity between the two formations. This contact was described as a thrust fault during early regional reconnaissance (Livingston, 1932).

The Huntington Fm. as identified by Brooks (1979b) in eastern Oregon is essentially identical to the rocks in the same structural-stratigraphic position in the Cuddy Mountains and Sturgill Peak area. The Huntington Fm. is composed mainly of metasedimentary and metavolcanic rocks. Volcanic conglomerate and sandstone are most abundant in the Cuddy Mountain area; less common are limestone, chert, and bedded barite. The metavolcanic rocks consist of tuff, flow breccia, spilite, andesite, keratopyhere, and dacite porphyry (Fankhauser, 1969; Slater, 1969; Bruce, 1971; King, 1971; Skurla, 1974; Hendricksen, 1975; C. Field, written comm., 1988). These rocks are in places intruded by quartz diorites, porphyritic granodiorite, and other more mafic bodies. They were altered to the greenschist facies by hydrothermal, contact, and incipient regional metamorphism (C. Field, written comm., 1988).

All of the BMIA rocks are unconformably overlain by basalt flows of the Columbia River Basalt Group. The angular unconformity is a product of original moderate to rugged Miocene relief, with subsequent flow onlap against and over the irregular topography. The thickest sequence of flows encountered is about 700 m (2000+ ft) in the vicinity of Black Canyon on the Oregon side of the Snake River (plate 1). The maximum Miocene relief, before normal faulting and monoclinal flexure was about 1000 m (3000 ft) in the Cuddy Mountains.

Two distinct Miocene basalt formations are present in the region. The lower is the Imnaha Fm. (originally referred to as Picture Gorge Fm.). Flows in the Imnaha Fm. are typically porphyritic and easily weathered; however some of the lower canyon-filling flows are more aphyric and resistant. The upper Grande Ronde Fm. (originally referred to as Yakima Fm.) flows are more dense, aphyric, and are commonly cliff formers. These formations were mapped separately by some workers (King, 1971), however the distinction is not made in maps accompanying this report. The basalt flows in the western Cuddy Mountains are dominantly of the Imnaha Fm.

Tuff and flow breccia are abundant in both basalt formations. Channel breccia and tuffaceous sediments are also common, and diatomite beds occur locally. Numerous dikes of the Columbia River Basalt Group are present throughout the area, and may have been feeders for local fissure eruptions. Most of these intrusions are oriented nearly north-south, and in a few cases follow small vertical normal faults.

STRUCTURAL GEOLOGY

Mesozoic Structure

The tectonic juxtaposition of accreted terranes and formations in the Cuddy Mountains was mainly achieved by thrust faulting, although strike-slip faulting may have been responsible for various amounts of translation along some structures. Wernicke and Klepacki (1988) proposed both right-lateral and left-lateral tectonic escape of BMIA fragments away from the Columbia Embayment. Ave Lallemant and others (1985, in press) present evidence for left-lateral motion in the Triassic period on major shear zones within the Wallowa Terrane based on detailed field measurements, isotopic age determinations, and oriented petrofabric investigations.

The major Mesozoic shear zones and structural grains trend approximately N 40°E, and bend more northerly as the terranes approach the West Idaho Suture Zone (fig. 2). A regional strike of bedding with the same general trend occurs in rocks of the Weatherby Fm., and is also well expressed in the Baker Terrane rocks. Huntington Fm. and Wildhorse Fm. rocks were deformed by folding prior to thrust faulting, and exhibit less uniform bedding attitudes. Because the unconformably overlying Weatherby Fm. is relatively undeformed at Cuddy Mountain compared to the complexly deformed Huntington Fm., a considerable depositional hiatus is inferred. The surface of the unconformable contact was later subjected to thrust faulting. The rupture surface is only partially coincident with the surface of the unconformity, and in places passes into the bounding formations (plates 2 and 3). From geologic mapping in the Huntington quadrangle in eastern Oregon, Brooks (1976) suggested that the total amount of movement is undefinable. However, in the overlying Weatherby Fm. there is evidence of imbricate thrusting to the southeast, and the total displacement is likely very large (Brooks, 1976). On the basis of evidence such as vertical separation of beds and repetition of a rhyolite tuff unit, Hendricksen (1975) suggests that movement of 1000 feet or more was possible (along a single structure). A similar range in Mesozoic displacement is probable at Cuddy Mountain, but may be as much as 5000 feet in the vicinity of No Business Canyon, with many other small low-angle thrusts likely throughout the Weatherby Fm.

The Weatherby Fm.-Baker Terrane tectonic contact in the Dukes Creek and No Business Canyon areas (plates 2 and 3) is a large Mesozoic thrust fault. This structure is referred to as the Connor

Creek Fault in eastern Oregon (Brooks, 1979b), the Cuddy Mountain fault (Mann, 1988) in this report, and as the Lick Creek Fault (Fitzgerald, 1983). Unlike the presumed limited displacements along the Huntington Fm.-Weatherby Fm. contact, the Paleozoic-Triassic Baker Terrane was thrust over the Jurassic Weatherby Fm. for possibly tens of miles. A zone (approximately 50-500 feet wide) of serpentinite, serpentinized fault gouge, and remobilized Baker Terrane and Weatherby Fm. rocks is present in the fault zone (MzCzum) (plates 1 and 2). The Cuddy Mountain fault was reactivated in the Late Cenozoic as a normal fault.

Another major Mesozoic structural discontinuity exists between the Baker Terrane and rocks of the Wildhorse Fm. (Wallowa Terrane?) in the vicinity of Wildhorse River Canyon (plates 2 and 3). Unlike the Cuddy Mountain - Connor Creek faults, this fault system is composed of several anastomosing and subparallel shear zones which juxtapose a wide variety of rock types. This zone, referred to here as the Wildhorse shear zone, includes a belt of dismembered igneous rocks composed mostly of cataclastic quartz diorite (PTrqd), and mylonite (PzMzm) (plates 1 and 2). It probably is the sole (duplex?) of a major thrust that brought (Wallowa Terrane?) igneous basement into tectonic contact between two very different terranes.

Rock units of limited areal extent such as a silicic tuff (PTrv), and a chert conglomerate (PTrc), are believed to be tectonic fragments related to the Baker Terrane (plate 2). On the north side of the sole a tuffaceous unit with thin limestone and ribbon chert beds (PTrvs) is of unknown terrane affinity (plate 2). A fusulinid of unknown genus was identified in thin section and indicates a Paleozoic age for part of the unit. Ribbon cherts suggest a relationship to the Baker Terrane, although its tectonic juxtaposition on the north side of the inferred sole places it in tectonic contact with the Wildhorse Fm. Just east of Brownlee Dam in lower Dukes Creek, an isolated outcrop of garnet mylonite (PzMzm) is in tectonic contact with rocks of the Baker Terrane (plate 1). Silicic tuff (PTrv?) may occur between the Baker Terrane rocks and the mylonites at this location. The mylonite was originally a clinopyroxene gabbro. The euhedral garnet crystal habit interspersed in the mylonitic fabric indicates post-mylonite metasomatic alteration. This mylonite body is on strike with the mylonite-bearing dismembered igneous sole in the Wildhorse shear zone, and is probably related to it. This structural trend and related lithologies are projected along strike under Miocene basalt cover to the line of section for the accompanying cross-section (plate 3).

Cenozoic Structure

Lawrence (1976) interpreted the pattern of faulting in southeastern Oregon as consisting of four major zones of right-lateral strike-slip faults that separate blocks broken by basin-and-range normal faulting (fig. 4). He proposed that the sum total of east-west extension decreased in the block north of each strike-slip fault zone, and that these dextral-slip fault zones formed to compensate for the decrease in extension across them. The two southernmost zones, the Eugene-Denio Zone and the McLoughlin Zone, offset the Pliocene-Holocene trend of the High Cascades some 10 to 20 km.

Lawrence (1976) also correlated clustered seismicity in eastern Oregon to locations along these northwest-trending zones (fig. 5) including the seismic zone described in this report. Strongly clustered seismicity in the Milton-Freewater area and the Brownlee Dam area, with several events scattered in between, indicates that this is an active northwest-trending seismic belt (fig. 5). This seismic belt is separated by a vast nearly aseismic area to the south from the Basin-and-Range Province of southeastern Oregon, and appears to be related to a different stress regime. Algermissen and others (1982) referred to this regional northwest-trending zone (zone P005) of seismicity, and noted that it appears to be only part of a more regional belt of moderate strain release that extends to the southeast into the western Snake River Plain.

Possibly the best studied and most conspicious of these fault zones is the Brothers Fault zone (fig. 4). This zone is mainly a series of discontinuous *en echelon* fractures trending about N 40°W. These fractures are mostly normal faults, about 10-20 km long, with intervening minor horsts and grabens (Lawrence, 1976). A second subordinate trend striking about N 30°E is composed of less abundant faults about 5 km long. This pattern is well demonstrated along strike-slip shear zones on many scales (Tchalenko, 1970). The geometric relationship of the faults suggests that the *en echelon* fractures are

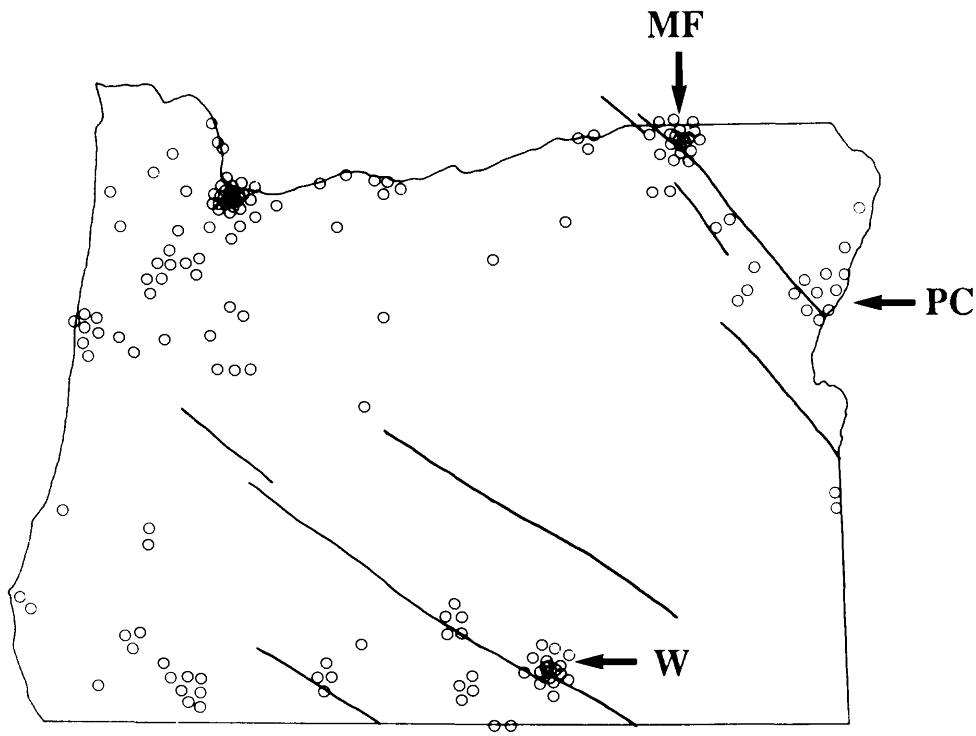

Figure 5 : Map of Oregon showing locations of major northwest-trending right-lateral strike-slip fault zones (fig. 4), with locations of earthquake epicenters between 1841 and 1970. W = Warner Valley sequence, MF = Milton-Freewater group, and PC = Pine Valley-Cuddy Mountain group (Oregon events only-approximately located). (slightly modified from Lawrence, 1976)

Riedel shears and the shorter fractures are conjugate Riedel shears (Tchalenko, 1970). The acute angle between the overall shear zone (N 60° W) and the *en echelon* fractures is consistent with right-lateral motion.

Walker and Nolf (1981) noted that, at the surface, the Brothers Fault zone is dominated by closely spaced *en echelon* normal faults of moderate to small displacement (mostly 10-100 meters) that localized many basaltic and rhyolitic vents of Late Miocene through Pleistocene age. The abundance and age distribution of these vents indicate that recurrent crustal breaking has taken place along the zone for a considerable span of time (Walker and Nolf, 1981). Walker and Nolf (1981) suggested that these normal faults and volcanic vents along the Brothers Fault zone represent only the surface manifestation of deformation on a large deeply buried structure with some lateral sense of displacement, most likely right-lateral. Walker (personal comm., 1989) believes that the pattern of surface faulting in the Brothers Fault zone (and similar zones) may be largely the result of brittle unloaded fracture response in the Late Cenozoic volcanic carapace which mantles the older, more ductile metasedimentary basement.

Many faults with a northwest trend similar to the trends in southeast Oregon are located in this study area. Figure 6 is a compilation of known and suspected Late Cenozoic faults in this area overlain on topography. Figure 7 is also a compilation of Late Cenozoic faults at a different scale, utilizing the same goegraphic window as the seismicity compilation shown later. The most prominent feature is a regional complex graben trending approximately N 55° W. This extensive graben has two distinct components; the Pine Valley pull-apart basin to the northwest (in Oregon), and the Brownlee-Salubria graben to the southeast (in Idaho) (fig. 7), which together constitute the Pine Valley graben (plate 4). Two areas of extensive uplift are also recognized (Fitzgerald, 1983) and referred to as the Cuddy Mountain uplift, and the Sturgill Peak uplift. The intervening complex Pine Valley graben may be coincident with a wide zone of deep-seated right-lateral strike-slip shear similar to others in eastern Oregon.

Most Late Cenozoic faults on the Oregon side were mapped by Brooks and others (1976), Walker (1977), and Brooks (1979a, 1979c). Much of the Late Cenozoic structure compiled in this report for the Pine Valley-Richland area relied on air-photo interpretation and limited reconnaissance (Walker, personal comm., 1989). Late Cenozoic structure in the Sturgill Peak-Hitt Mountain area and the southeast and northeast Cuddy Mountains area are compiled from several Oregon State University theses completed under the direction of Cyrus Field, and the mapping is in good detail, with important identifications of large Late Cenozoic faults (King, 1971; Skurla, 1974; and Hendricksen, 1975). Detailed fault mapping on the southeastern side of the Cuddy Mountains and Salubria Valley is provided by McIntyre (1976). Additional regional fault mapping in Idaho is compiled from Mitchell and Bennett (1979), and Fitzgerald (1983). Other faults along the west and northwest Cuddy Mountains of Idaho, and extreme eastern Pine Valley, Oregon are compiled from Mann (1988), and this report.

There is some confusion among workers regarding which side of the Wallowa Mountains the Olympic-Wallowa Lineament (OWL) is located. Zollweg and Jacobson (1986) stated that their 1981-1984 earthquake locations are coincident with the OWL, and refer to Lawrence (1976, fig. 3). Lawrence, however, does not refer to this group of northwest-trending faults as the OWL. The northern segments in the vicinity of the Milton-Freewater area are coincident with the OWL as originally proposed by Raisz (1945). He interpreted the lineament to pass into the Grande Ronde Valley and then bend or step northeast to form the northern rim of the Wallowa Mountains, obliquely chopping off the north-south trending ridges and valleys of the Wallowa Mountains. This segment of the OWL is coincident with the extensive Wallowa Fault system on the north side of the Wallowa Mountains. However, no significant northwest trending fault system can be found projected southeast along strike in the Snake River Canyon just north of the Oxbow Dam area. In this report, the lineament along the south side of the Wallowa Mountains noted (but unnamed) by Lawrence (1976) is referred to as the Grande Ronde-Pine Valley zone (GPZ), named after the pull-apart basins found along trend. The GPZ may then be related to the southeastern extension of the OWL.

The structural style of the GPZ is not identical to the other northwesterly trending zones of Lawrence (1976). The southern end of this zone is dominated by the extensive Pine Valley graben, described in this report to be of pull-apart basin origin. At its northern end it is dominated by the

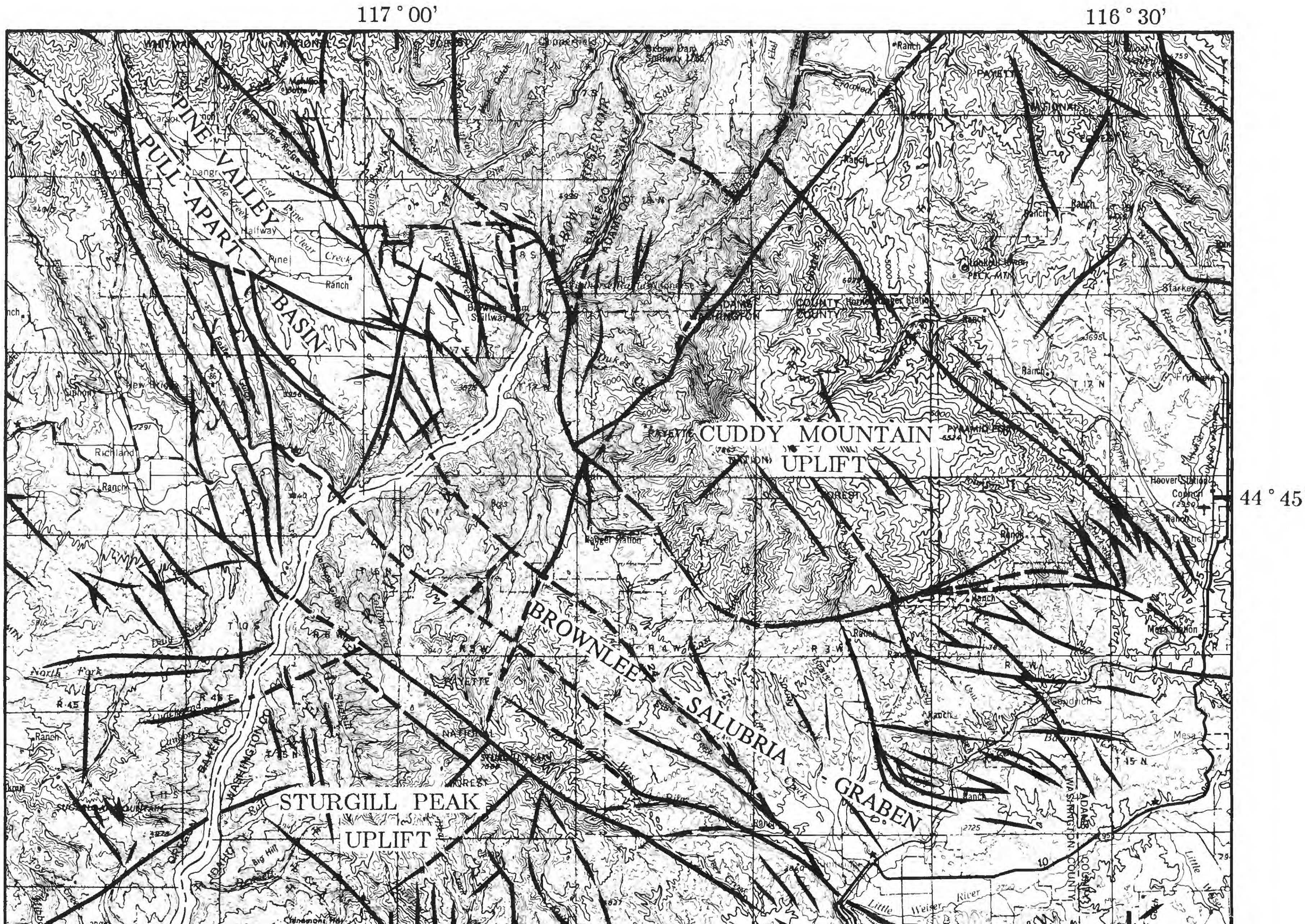

Figure 6 : Generalized Cenozoic fault map with topography. Structural and geomorphic names are informal. Compiled from Slater (1969), Bruce (1971), King (1971), Skurla (1974), Hendricksen (1975), Brooks and others (1976), McIntyre (1976), Walker (1977), Brooks (1979), Mitchell and Bennett (1979), Fitzgerald (1983), and Mann (1988). See figure 7 for fault names and towns.

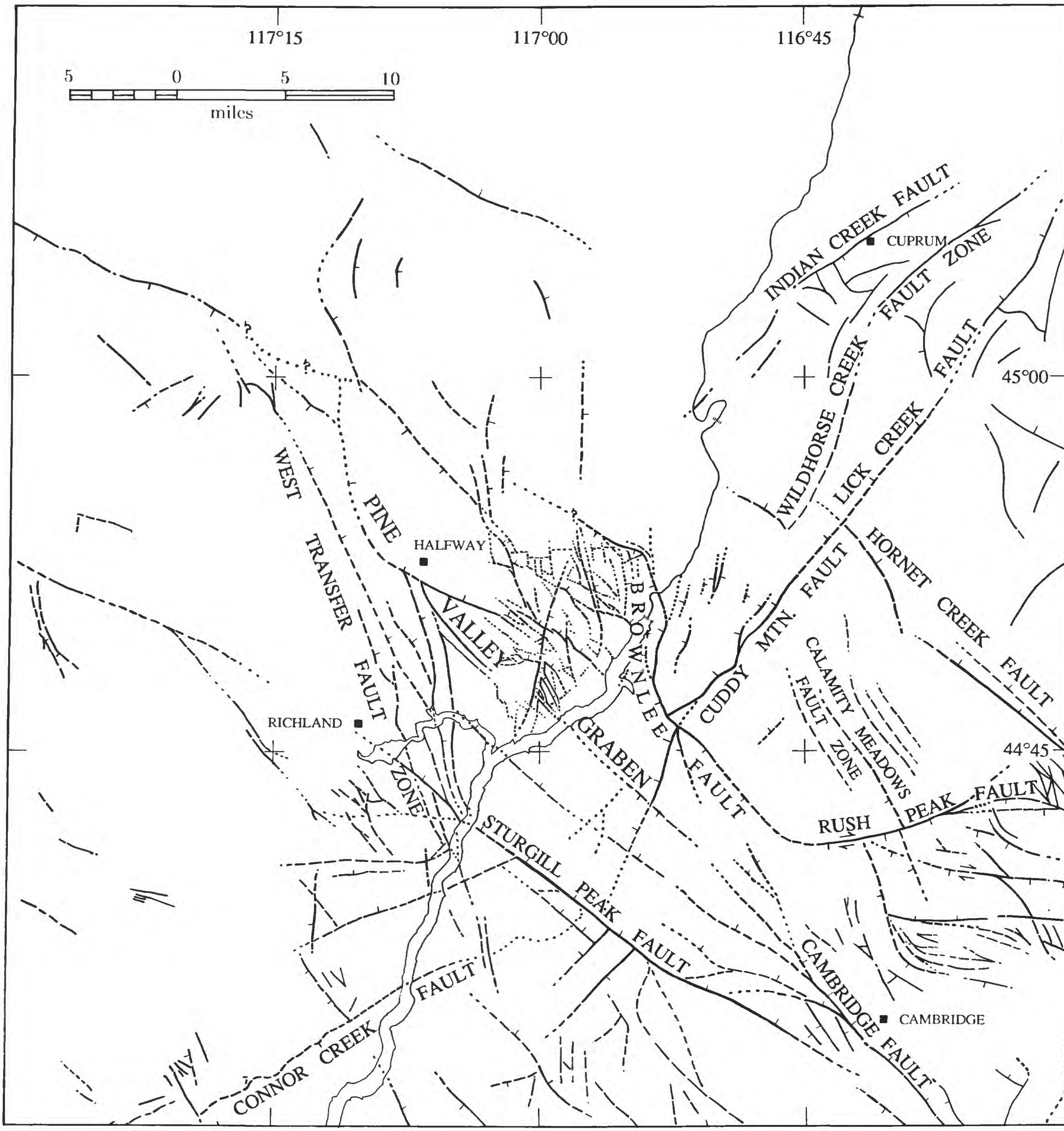

Figure 7 : Cenozoic fault map (hatchures on downthrown side). Compiled from Slater (1969), Bruce (1971), King (1971), Skurla (1974), Hendricksen (1975), Brooks and others (1976), McIntyre (1976), Walker (1977), Brooks (1979), Mitchell and Bennett (1979), Fitzgerald (1983), and Mann (1988).

large Grande Ronde graben, demonstrated as a complex pull-apart basin by Gehrels and others (1980). Gehrels and others (1980) state that the Grande Ronde graben owes its origin to limited northwest-southeast extension along bounding fault systems with right-lateral strike-slip displacements. Exposed fault surfaces are typically vertical and show a strong predominance of horizontal slickenside striae locally overprinted by vertical striae (Gehrel and others, 1980).

Gehrels and others (1980) further state that local Quaternary faulting is indicated by faulted colluvium west of La Grande and by the 1969 North Powder earthquake (M=3.6), with an epicenter 30 km southeast of the Grande Ronde graben and along structural trend from it. This earthquake is located in the north end of Baker Valley, consistent with intensity patterns (Couch and Whitsell, 1969). Mark Ferns (personal comm., Oregon Dept. Geol. & Min. Ind, 1989) has noted that a minimum of 200 feet of normal displacement of Quaternary colluvium has been reported in a mine along a segment of fault bounding the Elkhorn Mountains and Baker Valley near the town of Baker, Oregon. Barrash and others (1979) state that a weak northwest-trending photo-linear extending from the trace of the Foothill Ronde fault south of La Grande into the Grande Ronde Valley fill coincident with a topographic break in slope may be an indication of recent fault movement. The U. S. Army Corps of Engineers (Design Memorandum No. 26, 1983, p. 25) state that at least one fault in the valley cuts Holocene colluvium and displaces it. However, the U. S. Army Corps of Engineers show recent faulting located on the east side of the valley (Design Memorandum No. 26, 1983, plate 19), not on the west side along the trace of the Foothill Ronde fault.

The discovery that the Grande Ronde graben is a pull-apart basin on structural trend with the Pine Valley graben suggests that this feature and associated seismicity may be related to similar tectonic mechanisms. An examination of the geometry and style of faulting in the Pine Valley graben lends support to this possibility. The axes of Pine Valley and Brownlee-Salubria graben are offset several km (fig. 6 and 7). If this composite Pine Valley graben structure is controlled by dextral-slip faulting in the basement, then the axial discordance may be related to a right-stepping bend or offset in that basement structure. A right step along a right-lateral strike-slip fault system constitutes a releasing bend (Crowell, 1974). Pull-apart basins are known to nucleate at releasing bend fault segments along active strike-slip fault zones (Mann and others, 1983). Mann and others (1983) further state that "in rigid intracontinental settings, larger active pull-aparts do not occur singly but rather as part of a series on a segment of the principle displacement fault zone..." Thus, the overall Late Cenozoic fault geometry in the Brownlee area may be related to pull-apart development at a releasing bend.

The trend of the graben structure is approximately N 55°W; however no through-going master fault segment with this trend has been conclusively identified in this region. Discontinuous fault segments ranging in length from about 30 km to less than 1 km are present along the length of the GPZ. There may be several reasons why a through-going master fault segment is not developed. In some cases, the amount of strike-slip faulting at depth may be sufficiently small or deep that a through-going fracture fails to develop at the surface (Naylor and others, 1986). Instead only a long zone of *en echelon* normal faults or Riedel shears form (Erdlac and Anderson, 1982; Sylvester, 1988). Discontinuities are characteristic of dip-slip as well as strike-slip faults and are independeant of length, total fault slip, rock properties, and tectonic settings (Segall and Pollard, 1980). It is possible that the dextral-slip fault zones in northeastern Oregon are in an incipient or immature stage of development, and through-going surface ruptures may evolve during more advanced stages of deformation. It is also possible that the brittle unloaded fracture response of the overlying basalt veneer suggested by Walker (personal comm., 1989) may play a large role in masking through-going basement faulting.

The geometry of Late Cenozoic faulting in the Brownlee Dam area is complex, and structural relationships between subparallel groups of faults with a wide assortment of trends are not immediately straightforward. Figure 8 is an air photo of the northeast corner of the Pine Valley pull-apart basin, immediately west of Brownlee Dam. Figure 9 is a companion lineament map compiled from features observed on the air photo. Two structural trends are apparent, the regional N 50°-60°W trend, and a group with an approximate northerly trend. Some of these northwest-trending faults are exposed on the west walls of the Snake River Canyon, where they have apparent normal displacements of small to modest throw, generally between 10-60 m. However some segments, such as the system bounding the southwest side of Pine Valley have larger throws, roughly corresponding to the scarp relief. Those

Figure 8 : Air photo taken on October 8, 1951, of the east end of Pine Valley, Oregon. The Snake River before inundation by Brownlee Reservoir is at lower right. Note profusion of strong lineaments throughout the area, which are faults in most cases. This area is entirely covered by thick flows of the Miocene Columbia River Basalt Group. See figure 9 for interpretations of lineaments.

Figure 9 : Lineament map compiled from features observed on air photo in figure 8 Two dominant trends are evident; an approximately N 50° W trending set, intersected by a more northerly trending group. While many of the smaller northwesterly trending features may be only coincident with drainage, most of the larger escarpments are structural. The northerly trending set has more coherent expression than the segmented and anastomosing northwesterly group.

faults that bound the valley to the west and southwest have some expression in free-air gravity anomalies (Berg, 1967). No detailed studies of these faults has been attempted; therefore there is no documented strike-slip or oblique-slip component. The N 50°-60° W trend dies out rapidly to the east in this area (fig. 8), and the Brownlee fault becomes the dominant structure on the Idaho side of the reservoir. This area is therefore a highly deformed zone of structural transition.

The Tertiary basalt flows dip progressively more to the southwest moving north along the west shore of Brownlee Reservoir, in response to increased proximity to the Brownlee fault (plate 1). The Brownlee fault crosses the Snake River below Brownlee Dam near the mouth of Wildhorse River. The fault, according to air photo examinations and limited reconnaissance, crosses the mouth of Black Canyon on the Oregon side, progresses up the west side of Black Canyon and then abruptly turns west and crosses the basalt flats along the northeast corner of Pine Valley (fig. 6 and 7). In section 24 (plate 1) the throw of the Brownlee fault increases rapidly, and steep footwall slopes in the Mesozoic rocks are predominantly composed of cherty phyllites.

The Brownlee fault is a hinge fault resulting from the apparent breach of a monoclinal flexure. This flexure is actually extensional in nature; the area west of the fault being downdropped into the Pine Valley pull-apart basin as a result of crustal thinning below the basin. The Miocene basalt flows on the west side of the Brownlee fault dip at about 10°-20° to the west and southwest, and are essentially flat or dip gently to the west on the footwall. The map view of the fault trace bends slightly to the west (fig. 6, plate 1) down stream channels, therefore the fault plane is interpreted to dip at about 75° to the west along its north-trending reservoir-parallel segment. The fault possibly has a more shallow dip along its northwest-trending segments. In section 6 at the extreme southwest corner of the Cuddy Mountain quadrangle (plate 2), a splay of the Brownlee fault offers an excellent exposure of the fault plane. Figures 10 and 11 show slickensides and mullion structures on a plane locally striking N 11° W, and dipping 55° WSW. The orientation of these fault-plane features implies some component of oblique motion along this segment; however the sinous trend of the fault overall would seem to preclude a through-going oblique component along its entire length.

The Brownlee fault forms the northeast boundary of the Pine Valley pull-apart basin. Figure 12 is a composite of two overlapping photos, with a view looking northwest at the Brownlee fault, showing the relationship of the west dipping hanging wall and nearly horizontal footwall. The Oregon extension, interpreted from air photos and limited reconnaissance, appears to have a decreasing dip-slip throw, and may pass into a monoclinal hinge to the northwest. The surface trend of the Brownlee fault is the most sinuous of any in the region. To the southeast it appears to merge with the Rush Peak fault, however it may pass into a monoclinal hinge before rounding the bend in the Cuddy Mountain uplift (fig. 6 and 7). To the south in section 25 (plate 1), footwall slope gradients in some places exceed 60 percent in soft sheared phyllitic rocks with little chert. Footwall scarp retreat rates in these soft metasedimentary rocks *may* be evidence for vigorous Late Quaternary displacement.

The Brownlee fault's helicoidal surface trend is analagous to a tension gash or Riedel shear. Dewey (1978) suggested that pull-apart basins nucleate on rotated large-scale tension gashes or Riedel shears, which formed during the initial stages of fault offset. However, Mann and others (1983) note that such well-shaped structures are rare in nature. The overall shape of the Pine Valley graben is somewhat intermediate between a spindle shaped "lazy S" basin shape of Mann and others (1983, fig. 11c) and a more mature rhombic shape (Mann and others, 1983, fig. 11d).

If the Brownlee fault and subparallel smaller faults at the northeast corner of the Pine Valley pull-apart basin (fig. 6, 7, 8, 9) are part of a zone of structural transition, or transfer of stress at the east side of a right-stepping offset, then a similar relationship is required at the west end of the step. An approximately N 30° W trending group of subparallel normal faults between the towns of Richland and Halfway, referred to informally in this report as the west transfer fault zone (fig. 7), may represent such a zone. Crowell (1974) depicted pull-apart basins as bounded on their sides by two, subparallel, overlapping strike-slip faults, and at their ends by perpendicular or ***diagonal*** dip-slip faults, which he termed "transfer faults". These dip-slip faults link the ends of the strike-slip faults and allow slip to be transferred through the discontinuity (Segall and Pollard, 1980). The locations of the groups of faults that are analogous to these transfer zones, and the approximate locations of the hypothesized right-lateral strike-slip fault zones are shown in figure 13.

Figure 10 : Slickensides on a fault plane oriented N 11°W and dipping 55°WSW. Location of exposure is at extreme southwest corner of the Cuddy Mountain quadrangle, section 6, on a splay of the Brownlee Fault.

Figure 11 : Mullion structures on a fault plane oriented N 11°W and dipping 55°WSW. Location of exposure is at extreme southwest corner of the Cuddy Mountain quadrangle, section 6, on a splay of the Brownlee Fault.

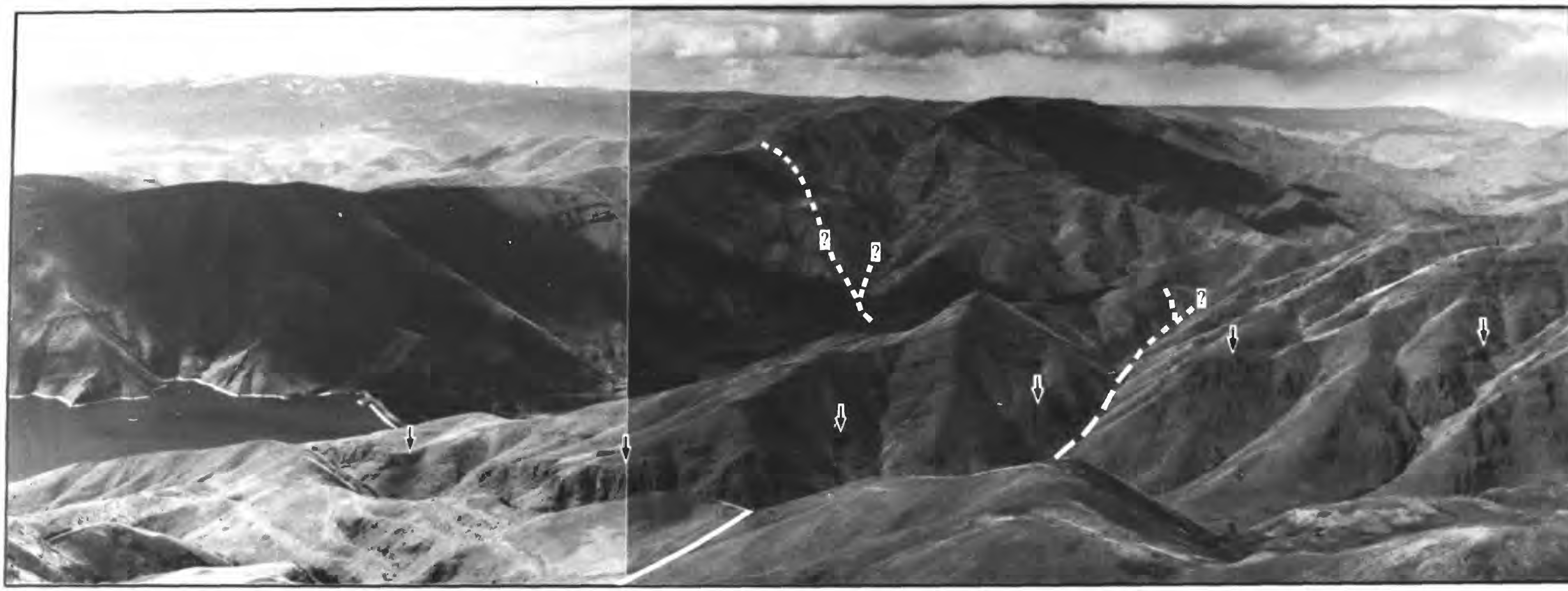

Figure 12 : Photomosaic taken from section 18 looking northwest across lower Dukes Creek into Oregon and the northeast corner of Pine Valley. The Brownlee Fault (shown in white at center-right) separates westward-tilted basalt flows on left from uplifted nearly horizontal flows on the right. Fault passes as indicated (dashed line) through small tributary on north side of Dukes Creek. Fault then crosses over hill and bends northwest (**plate** 1) towards mouth of Wildhorse River, crossing the Snake River below Brownlee Dam, and reemerging in the background (center) on the west side of Black Canyon, Oregon. Note similar relationship of westward tilted basalt section on canyon walls on the Oregon side. Arrows point to approximate location of top of resistant basalt canyon fill flow sequence.

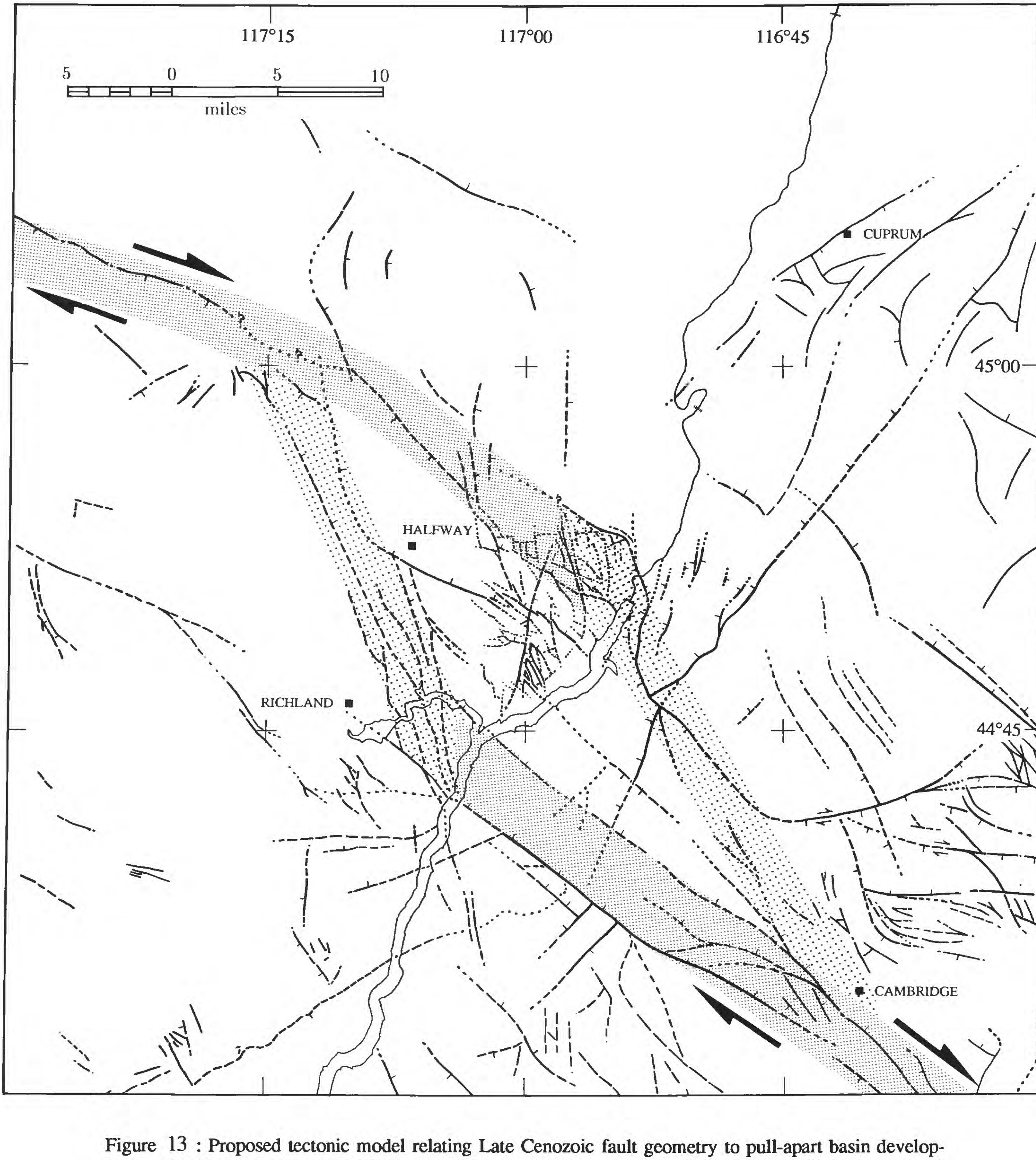

Figure 13 : Proposed tectonic model relating Late Cenozoic fault geometry to pull-apart basin development. Overlapping segments of basement right-stepping dextral strike-slip fault zones (light dot pattern) are connected by normal dip-slip (and oblique slip?) transfer fault zones (coarse dot pattern). Faults compiled from Slater (1969), Bruce (1971), King (1971), Skurla (1974), Hendricksen (1975), Brooks and others (1976), McIntyre (1976), Walker (1977), Brooks (1979), Mitchell and Bennett (1979), Fitzgerald (1983), and Mann (1988).

The pull-apart model is perhaps overly simplistic and fails to properly explain such features as the Cuddy Mountain uplift (fig. 6). The GPZ may be a more concentrated zone of dextral shear within a broad regional zone. The magnitude and rate of shear strain could be expected to vary locally within a broad regional zone; the entire domain of strike-slip may be composed of many subzones, each undergoing simple shear at a particular rate (Aydin and Page, 1984; Sylvester, 1988). Figure 14 is a hypothetical diagram showing the variety of extensional and compressional features that can be generated in a broad strike-slip environment (Aydin and Nur, 1982). Areas of compressional uplift or "push-ups" are typically of a rhombic shape, and form at restraining bends--or left-stepping offsets on a dextral-slip fault. This model helps explain the unusual fault-bounded rhombic shape of the Cuddy Mountain uplift, and may then imply a bordering strike-slip zone to the southeast interacting with the GPZ.

A broad zone of strike-slip faulting with more concentrated zones or lanes of deformation interrupted by dilational and antidilational jogs may help explain the subparallel scatter of northwest oriented major pull-aparts, suspected pull-aparts, and faulted uplifts in northeastern Oregon such as Pine Valley, Grande Ronde Valley, Baker Valley, the Wallowa Mountains, and the Cuddy Mountains. Figure 15 is a proposed tectonic model superimposed on the regional Late Cenozoic fault compilation (plate 4) that attempts to correlate the pattern of faulting in this area to modern theories and observations of broad strike-slip fault belts.

Evidence for dextral-slip faulting southeast of, and bordering the Cuddy Mountain uplift is presented by McIntyre (1976) and King (1971). McIntyre has mapped an apparent dextral offset of 0.5 km along a northwest-trending fault by an east-west trending fault approximately 12 km (8 miles) northeast of Cambridge, Idaho (fig.6 and 7). This apparent dextral strike-slip fault is also mapped as a normal fault downdropped on the south side, and the strike-slip motion arrows are queried. The apparent offset of a northwest trending fault by the Rush Peak fault (fig. 7) prompted King to speculate that this fault may have a minor right-lateral strike-slip component of motion. It is notable that these approximately east-west-trending faults project eastward into a disrupted zone in the Long Valley fault system (plate 4). This disrupted zone composed mostly of short northwest-trending faults may be the result of penetration of the Idaho Batholith by a deep seated basement fault zone. Faults projected along trend to the northwest, such as the Hornet Creek fault (fig. 7) apparently do not rupture Late Cenozoic basalt cover along strike in the Oxbow area of the Snake River Canyon. An antidilational jog, west of the disrupted zone around the Cuddy Mountains through the fault system mapped by King (1971) and McIntyre (1976) may be the preferred course of a deep basement, dextral-slip shear zone. Such an antidilational jog, if it exists, could control the uplift of the Cuddy Mountains. It is also interesting to note that this disrupted zone and its projection to the southeast is on trend with the Central Idaho Seismic Belt, a west-northwest trending appendage of the Intermountain Seismic Belt.

The Crane Creek fault zone (plate 4) is composed mainly of short normal faults 3 to 15 km in length, averaging about 6 km (Fitzgerald, 1983). Fitzgerald (1983) notes that although fault density is great in this zone, displacements are least compared to other northwest trending fault zones in west-central Idaho where fault density is lower. Fitzgerald (1983) interpreted this zone to widen to about 50 km to the northwest, encompassing the Dead Indian Ridge fault zone to the west, north to the Pine Valley graben, together comprising the Paddock Valley fault system. An average length of 8 - 10 km occurs in this wider belt of lower fault density and larger offsets. Fitzgerald (1983) attributed these differences to both magnitude of deformation and response of the different near-surface basalt units to stresses in the two areas (brittle unloaded fracture response). In the southeast (Crane Creek fault zone) the near-surface units are the less competent Weiser Basalt flows and interbedded and abutting sediments, whereas in the northwest the near-surface units are the more competent flow-on-flow Grande Rhonde and Imnaha flows (Fitzgerald, 1983). The Crane Creek fault zone appears to die out to the northwest along the southern side of Shoe Peg Valley (plate 4), which may be the result of a dilational jog in a controlling basement strike-slip fault zone to the north, perhaps stepping to the Cambridge fault- Sturgill Peak fault system (plate 4, fig. 7 and 15).

The Dead Indian Ridge fault zone is on trend with faults in Oregon that appear to be associated with a 3 km dextral offset of the Connor Creek fault (plate 4). However, a gap in the Connor Creek

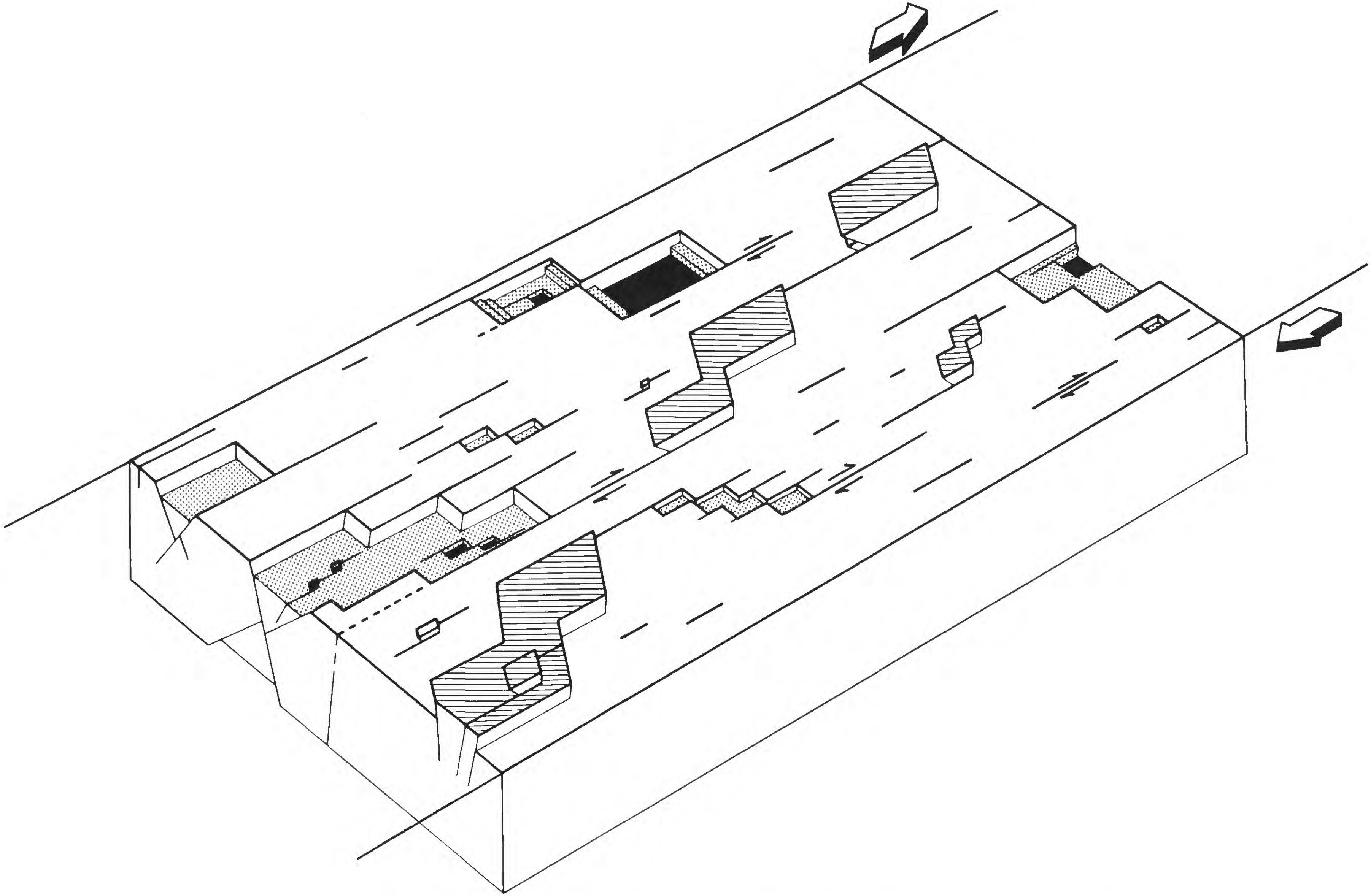

Figure 14 : Hypothetical diagram illustrating a broad right-lateral strike-slip environment. Right-stepping and left-stepping offsets on discontinuous subparallel strike-slip faults interact to form pull-apart basins and rhombic horsts of various sizes. (modified from Aydin and Nur, 1982)

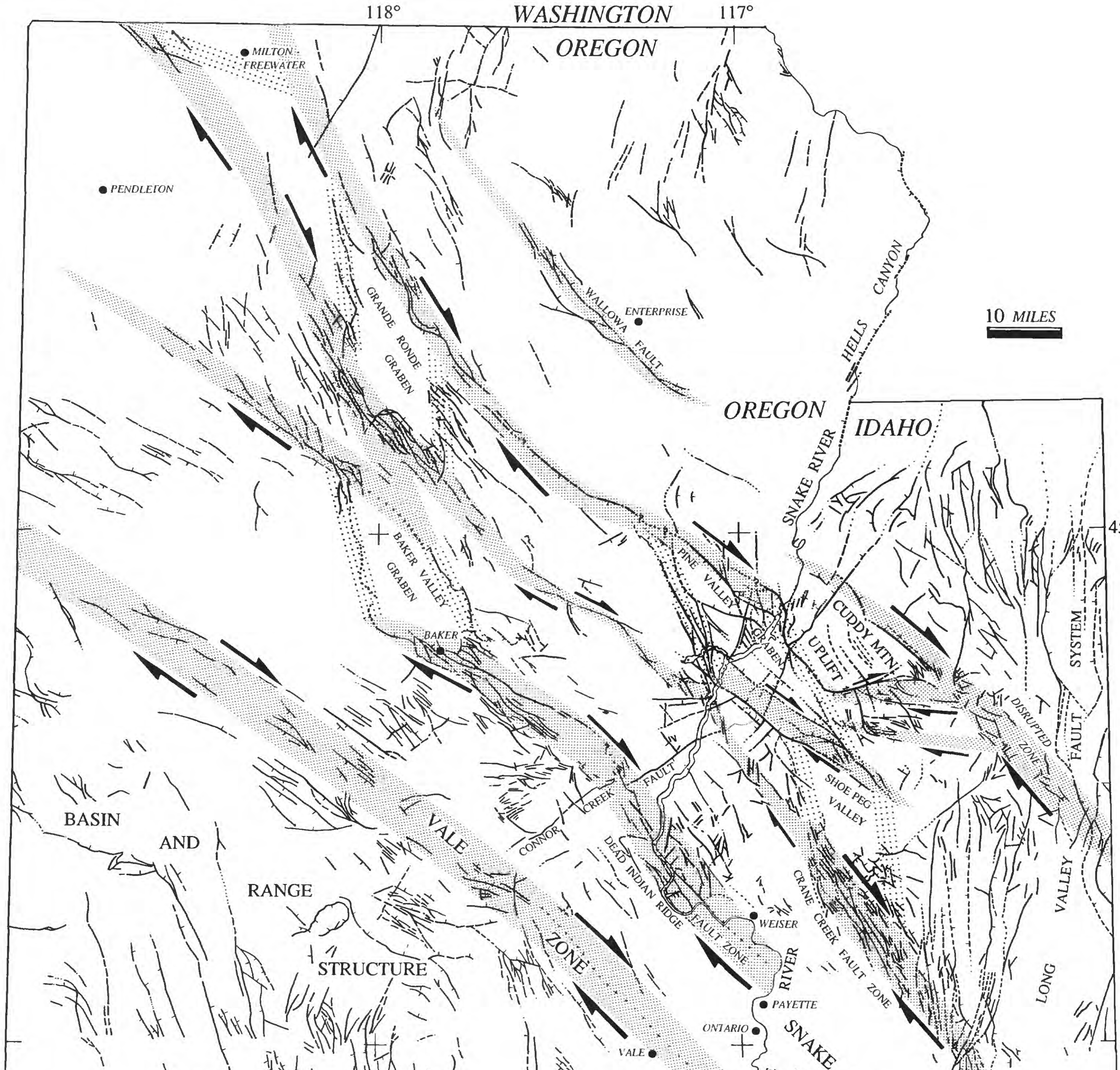

Figure 15 : Regional Cenozoic fault map with proposed tectonic model showing relationship of discontinuous and anastomosing dextral strike-slip fault zones (light dot pattern) to pull-apart basin (graben) and uplift development. Dip-slip (and oblique-slip ?) transfer fault zones bounding pull-apart basins shown in coarse dot pattern. Faults compiled from Slater (1969), Bruce (1971), King (1971), Skurla (1974), Hendricksen (1975), Brooks and others (1976), McIntyre (1976), Walker (1977), Brooks (1979), Mitchell and Bennett (1979), Fitzgerald (1983), and Mann (1988).

fault, astride the apparent offset, is intruded by an Upper Jurassic-Lower Cretaceous quartz diorite pluton (Brooks and others, 1976), and a through-going Late Cenozoic surface rupture is not presently identified. The apparent offset may be the result of irregular Mesozoic thrust fault geometry; Mesozoic right-lateral offset; or post-Mesozoic deformation. The Oregon extension of the Dead Indian Ridge fault zone is approximately on strike with an extensive northwest-trending fault system extending to the south side of Baker Valley (plate 4). A through-going basement fault system, if it exists, could link the two fault systems (fig. 15).

Rodgers (1980) questioned the relevance of theoretical models to actual pull-aparts, noting that fault patterns predicted by models might only apply to the initiation of faulting, and not to subsequent faulting which may reactivate older structures rather than form new ones. The Cuddy Mountain fault is a reactivated Mesozoic high-angle reverse fault (fig. 7, plate 2). The Paleozoic - Triassic Baker Terrane was thrust over the Jurassic Weatherby Fm. probably during Early Cretaceous accretionary processes. Sometime after extrusion of the last flows of the Columbia River Basalt Group (in the Late Miocene?), this structure was reactivated as a normal fault with as much as 700 m (2000+ ft) of retrograde dip-slip displacement. The Cuddy Mountain fault borders the northwest side of the Cuddy Mountain uplift, and is part of a group of normal faults which bound most (if not all) of the Cuddy Mountains (figs. 6 and 7). Unlike the other faults that border the uplift, the Cuddy Mountain fault appears to be a regionally extensive structure. To the southwest in Oregon, this same Mesozoic terrane boundary fault is referred to as the Connor Creek fault, where it displaces Late Cenozoic tuffaceous sedimentary rocks (Brooks and others, 1976).

The largest measured Cenozoic retrograde normal displacement on the Cuddy Mountain fault is in the vicinity of No Business Canyon on the northwest side of the Cuddy Mountains (plate 2). Dip-slip offset measured at the basal flow of the Columbia River Basalt Group is approximately 700 m (2000+ ft) (fig. 16). A plug of serpentinite (MzCzum) which intrudes the fault zone along the banks of No Business Creek is surrounded by a mineralized zone of tectonized Baker Terrane and Weatherby Fm. sedimentary rocks (MzCzum) (plate 2). A splay on the northwest side of the main trace may mark the actual terrane boundary. If that is the case, then the Baker Terrane rocks are thinned to a belt only about 2 km wide. The Baker Terrane is projected to tectonically pinch-out only a few more miles along strike (to the northeast). Structural relationships are obscured by a massive landslide covering nearly one square mile of the valley floor in lower No Business Canyon, and thick Miocene basalt cover to the northeast. No known Baker Terrane rocks crop-out along the WISZ to the northeast (T. Vallier, oral comm., 1988).

Late Cenozoic throw along the Cuddy Mountain fault decreases gradually to the southwest. At the center of the Cuddy Mountain quadrangle (plate 2), at a small pass at the 6000 ft (2000 m) level known locally as Blue Bunch Saddle, the Late Cenozoic offset diminishes to approximately 365 m (1100 ft). Excellent exposures of the fault plane in a ravine to the northwest of this saddle yield a dip of 65° NW. Geomorphic features suggest Late Quaternary rupture along this segment of the Cuddy Mountain fault. A prominent scarp composed of Jurassic Weatherby Fm. sedimentary rocks protrudes as much as 10 m (30 ft) above a linear depression marking the main fault trace (fig. 17). Stream erosion does not appear to have contributed significantly to modify or otherwise effect a fault-line scarp origin for this feature, although lower in the Dukes Creek drainage the fault zone is modified by erosion. An unusual flat area approximately 5 m wide and about 30 m long truncates the top of this scarp and may be the remnants of an uplifted erosional surface (fig. 17). Shallow linear depressions in the fault zone gouge and soils can be found in this vicinity. These features may be remnants of surface rupture traces that have been smoothed by weathering, or may be the result of differential settling of loose material above sheared zones. This fault scarp (fig. 17) diminishes in height rapidly to the northeast in the extreme upper Dukes Creek drainage, and is essentially absent at Blue Bunch Saddle, then reappears as a more subdued scarp in the upper Starveout Creek drainage. Because the scarp is discontinuous, evidence for Holocene rupture is inconclusive, however fresh surface rupture traces may be discontinuous, especially for smaller scarps. The preservation of nearly vertical to steep slopes on some of the fault scarp faces, composed mostly of fissile Jurassic shales (Jw) and soft serpentinized sediments, is strongly suggestive of Late Quaternary displacement.

The difficulty in identifying Holocene (<10,000 year) displacement on faults in the area is greatly

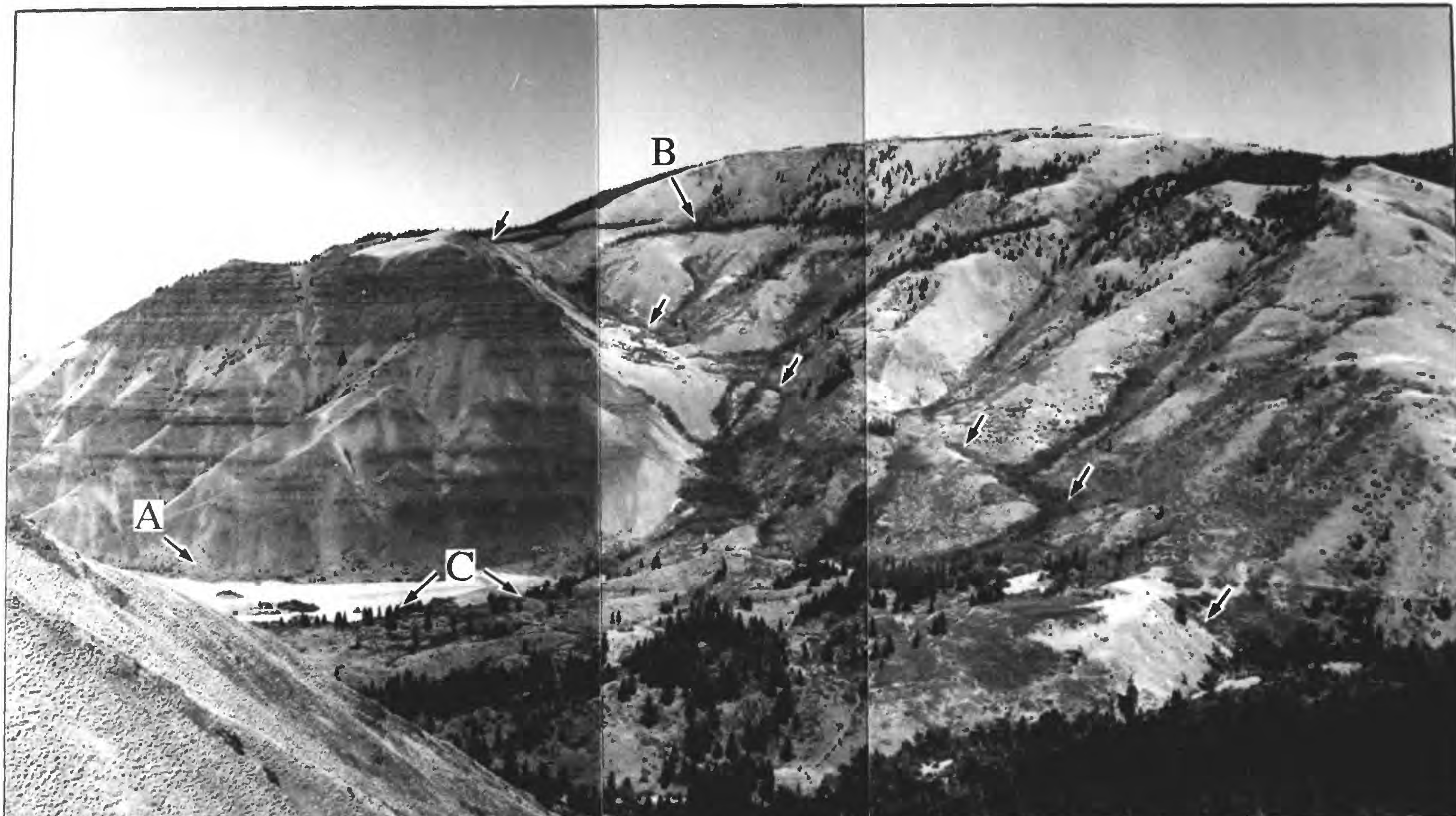

Figure 16 : Photomosaic taken from section 10 (Cuddy Mtn. quad.) looking east across lower No Business Canyon into sections 1, 2, and 3. Lowest Miocene basalt flow (A) and overlying 2000 foot section are displaced by 2000 feet of dip-slip offset to location (B) by Cuddy Mtn. Fault. Fault trace, shown by arrows, is marked by a series of low scarps. Massive basalt boulder landslide originated at terrace near (B) in Lost Basin, and traveled over two miles, turning northwesterly down No Business Canyon, and depositing wide zone of hummocky boulder debris (C).

Figure 17 : View looking south in southeast corner of section 16, upper Dukes Creek. Approximately 30 foot tall late Quaternary fault scarp composed of altered and partially serpentinized Weatherby Fm. shales and sandstones. Jurassic Weatherby Fm. sedimentary rocks are juxtaposed against Miocene basalts in this vicinity. Note flat area on top of scarp, which could be an uplifted erosional surface.

Figure 18 : View looking west across lower Grade Creek (section 31-Cuddy Mtn. quad.) at trace of Brownlee Fault. 600 foot thick basalt section on left is tilted southwesterly about 20°, and is abruptly truncated against Weatherby Fm. sedimentary rocks. Dip-slip displacement is estimated to be nearly 1500 feet. Trace of fault shown in white.

hindered by the lack of recognizable Holocene deposits; and perhaps also associated with Holocene magnitudes of seismic events. "Surface faulting associated with earthquakes in the magnitude 5 to 6 range is difficult to identify immediately after the event, and evidence of a prehistoric faulting event that may have occurred hundreds or thousands of years ago is even more difficult to identify... Besides being of limited extent, the evidence is apt to be subtle. Field evidence of the smaller displacements, some of only a few millimeters, will be very obscure, particularly for strike-slip faulting... Such conditions may explain why so few active faults have been recognized in some regions..." (Bonilla, 1988). Earthquakes in this range are frequently not accompanied by coseismic surface rupture. For these reasons minor Holocene rejuvenation of this scarp is possible, and cannot be ruled out without more detailed examination (i.e., trenching).

Farther southwest at the drainage divide between the north fork of Grade Creek and Dukes Creek, the Cuddy Mountain fault has approximately 200 m (600 ft) of throw. The Brownlee fault has its maximum throw of about 500 m (1500 ft) near this intersection, whereas the Cuddy Mountain fault has its Late Cenozoic throw diminished to less than half this amount (fig. 18). These two faults may have a syntectonic Late Cenozoic origin. Because of the nearly orthogonal angle of intersection, the geometry of their fault planes at depth would appear to be cross-cutting. One fault may offset the other at depth, or the plane of the Brownlee fault may sole into the plane of the Cuddy Mountain fault. The Cuddy Mountain fault is projected to maintain its surface dip of approximately 65° NW at depth, or perhaps it may become more listric because of its original thrust fault origin. This complex fault plane geometry may result in the concentration of stress near this intersection.

A splay of the Cuddy Mountain fault may pass into the Weatherby Fm. shales (Jw) and extend southwest down Grade Creek (plates 1 and 2), crossing highway 71 and extending to a fault present in the middle fork of Brownlee Creek (fig. 6). Weatherby Fm. shales are faulted against Late Cenozoic basalt in the middle fork of Brownlee Creek, and the Late Cenozoic displacement may be truncated by a northwest-trending fault approximately 4 km south of Highway 71 (fig. 6). The middle fork of Brownlee Creek maintains its linear north-northeast trend south of this intersection to within 1 km of the Sturgill Peak fault (fig. 6). There is no Mesozoic outcrop south of the aforementioned intersection, with the exception of an isolated outcrop of Baker Terrane metasediments and juxtaposed Weatherby Fm. shales at Lime Spring, approximately 5 km south of the last documentable Late Cenozoic offset. This Mesozoic juxtaposition marks the same tectonostratigraphic terrane boundary as the Cuddy Mountain fault, however, Late Cenozoic reactivation could not be documented during limited reconnaisance along this segment of fault.

In the middle fork of Brownlee Creek, the projected extension south of the Mesozoic tectonostratigraphic terrane boundary exposed at Lime Spring, does not align with the boundary as mapped by Skurla (1974) on the south side of the Sturgill Peak fault. Figure 7 shows the Idaho extension of the Connor Creek fault dotted to an apparent truncation along the Sturgill Peak fault, because Late Cenozoic offset is not demonstrated along this part of the structure. The apparent 2 km right-lateral offset of this Mesozoic terrane boundary may be the result of Mesozoic age fault offset; connection of this segment to the Lime Spring outcrop under Cenozoic basalt cover; or some component of right-lateral strike-slip displacement along the Sturgill Peak fault.

The Sturgill Peak fault was largely mapped by Skurla (1974), and is described as a high-angle normal fault with over 700 m (2000+ ft) of displacement near Sturgill Peak. Other smaller northeast-trending normal faults have throws between 400 and 600 ft and intersect, but do not displace the Sturgill Peak fault (Skurla, 1974). They appear to be controlled by the northeast-trending Mesozoic structural fabric. Several faults branch from the Cambridge fault and may be linked to the Sturgill Peak fault (fig. 6 and 7). A prominent topographic scarp that bounds the south side of Salubria Valley south of Cambridge, Idaho (fig. 6) marks the main trace of the Cambridge fault. This scarp stands an impressive 200-400 m above the west side of Salubria Valley, and the Pleistocene(?) Weiser River drainage is antecedent through an approximately 350 m tall section of the scarp. Numerous shear zones are present in roadcuts along Highway 95 near the base of the scarp.

Bounding the southeast corner of the Cuddy Mountain uplift is the Rush Peak fault (fig. 7). This large structure is described as a high angle normal fault with at least 700 m (2000+ ft) of dip-slip throw along much of its length (King, 1971), and closer to 1000 m below Rush Peak. The fault may have followed pre-existing structures because it parallels a rhyolite intrusive body (King, 1971). The Rush Peak fault splinters into several smaller normal faults to the east near Council, Idaho. The Hornet Creek fault bounds the northeast side of the Cuddy Mountain uplift, and appears in a structural-lithologic compilation by Hendrickson (1975). The southeastern part of this fault appears on a published map by Mitchell and Bennet (1979), and along the length of the northeastern part of the Cuddy Mountain uplift by Fitzgerald (1983). The Rush Peak fault and the Hornet Creek fault together form the wedge-shaped eastern part of the Cuddy Mountain uplift which points southeastward into the disrupted zone in the Long Valley fault system (plate 4).

A small normal fault is located in Dukes Creek between the Cuddy Mountain fault and the Brownlee fault (plate 1, fig. 19). This fault offsets the Miocene basalts about 65 m, is downdropped on the west, and has a small companion antithetic fault to the west, producing a keystone graben. A large backrotated basalt landslide block has calved on its east side from the fault zone (fig. 19). This fault can only be followed for about 1 km, and may be a local structure related to tensile stresses associated with the diverging trends of the two much larger normal faults to the east and west.

The entire region is laced with a great number of small normal faults, most with a northerly trend. One set of faults with a nearly perfect north-south orientation may be related to the multitude of Columbia River Basalt Group dikes that intrude the area in this orientation. They may be related to a period of regional crustal extension during and after the outpouring of these basalts.

According to Vallier (oral comm., 1989) in the Seven Devils-Hells Canyon region to the north, faults with north-south trends are as long as tens of kilometers. Other large faults with a northerly trend lie along the deepest part of Hells Canyon, controlling the drainage pattern of straight segments of the Snake River (plate 4). Cook (1954) has reported that Columbia River basalt is present at 8,150 ft elevation in the Seven Devil Mountains on the Idaho side of the Snake River. These remnants are mostly Imnaha flows, suggesting that Grande Ronde units may have been eroded so that more structural than topographic relief occurs (Fitzgerald, 1983). Comparable exposures of Imnaha basalts on the Oregon side of the Snake River are present at about 4,000 ft elevation.

Other significant fault systems include the Indian Creek Fault and the Wildhorse Creek fault zone, north of the Cuddy Mountain uplift (fig. 7). These northeast-trending structures are probably controlled by Mesozoic BMIA structure, similar to the Cuddy Mountain-Lick Creek faults. The Indian Creek fault is on trend with the Oxbow-Cuprum shear zone, a major mylonized Mesozoic shear zone in the Wallowa Terrane (Vallier, pers. comm., 1988). The Wildhorse Creek fault zone is on trend with the Wildhorse shear zone (plate 3), which is interpreted to be a tectonostratigraphic terrane boundary between the Wallowa Terrane and the Baker Terrane. Late Cenozoic offset in western Wildhorse Canyon was not observed in this shear zone, although minor offsets may exist.

Large active landslides have also been identified in this study along the shores of Brownlee Reservoir, and are shown on the accompanying quadrangle maps (plates 1 and 2). One of these slides is a reactivation of an ancient failure and was first noticed in 1984. This landslide is located about 2 km south of Brownlee Dam on the Idaho side in section 14 (plate 1). It is composed of approximately 25 million yds^3 of basalt sliding on westward inclined cherty Baker Terrane rocks (aquitard) along the Tertiary-pre-Tertiary unconformity. There is some speculation that this slide was reactivated by precipitation levels of about 170 per cent above normal during the very wet 1982-1983 El Nino conditions (S. Wood, oral comm., 1988). Because its toe is submerged in the reservoir, this additional saturation may contribute to future instability. Some minor damage to Highway 71 can be attributed to this landslide.

SEISMICITY

Historical Seismicity

The Pine Valley-Cuddy Mountain-Oxbow area has experienced intermittent seismicity throughout this century. Table 1 is a compilation of all known earthquakes occurring between 1913-1988, reported in the region bounded by 117°30'-116°30'W and 44°30'-45°30'N. With the exception of the 1981 and 1983 events, and the microseismic events recorded in 1984 by Zollweg and Jacobson (1986), the magnitudes, intensities, and locations of these earthquakes are not well constrained. The Blue Mountains

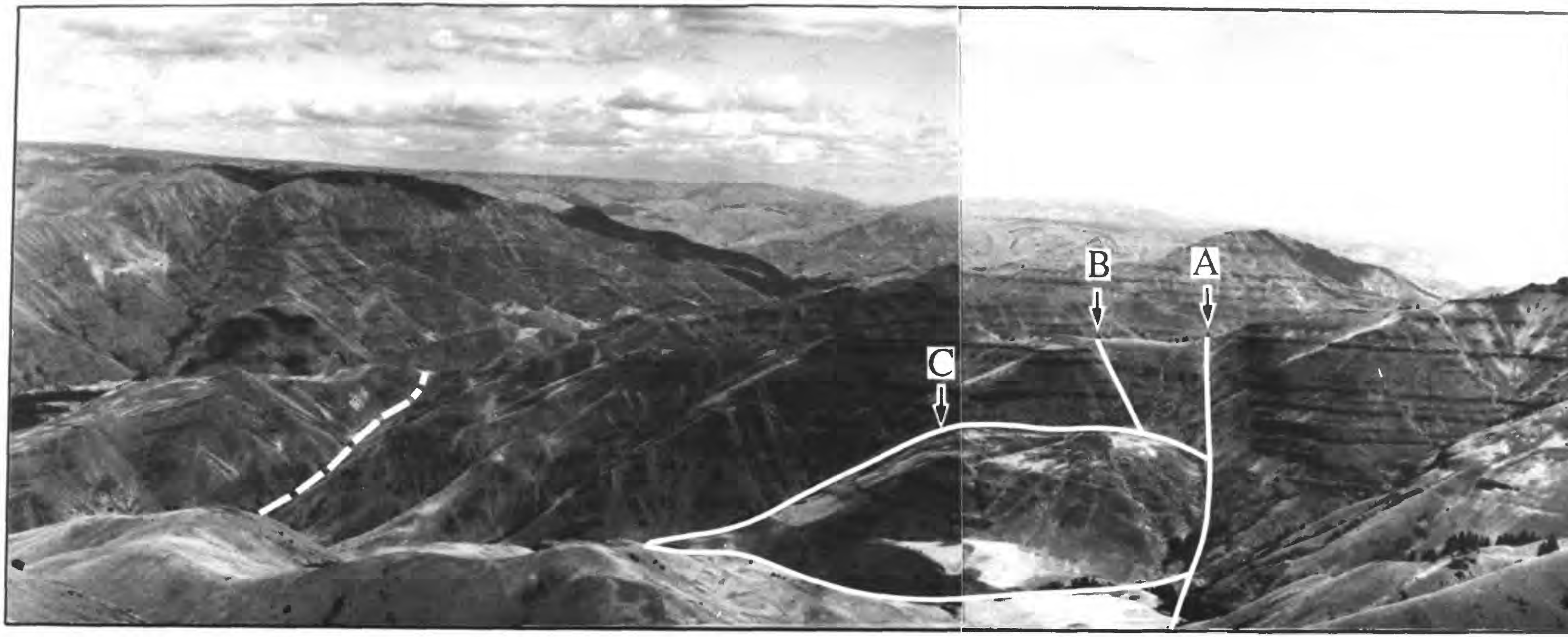

Figure 19 : Photomosaic taken from section 18 looking north across Dukes Creek. A steeply westward dipping normal fault (A) displaces the Miocene basalt section approximately 175 feet (right side up). A small antithetic fault (B) down-drops section between faults about 25 feet producing a small graben. Large back-rotated coherent basalt landslide block (C) has calved from the hanging wall of the larger fault. Basal rupture plane of landslide block is coincident with cherty Baker Terrane sequence. Brownlee Fault is visible on the left. Faults and landslide outline shown in white.

Table 1 : Historical seismicity in the region bounded by: 116°30 - 117°30W, 44°30 - 45°15N.

Year	Date (m/d)	Origin Time (UTC)	Latitude (North)	Longitude (West)	Magnitude Intensity
1913	10/13	2300	45.10	116.70	VI
1927	4/9	0500	44.88	117.20	V
1927	4/9	0700	44.83	117.32	IV
1927	4/9	0930	44.82	117.08	IV
1927	4/9	1400	44.75	117.23	IV
1941	12/23	1748	44.75	117.00	IV
1942	6/12	0930	44.92	117.00	V
1942	6/12	0935	44.92	116.87	III
1942	6/14	0600	44.83	116.92	III
1949	3/15	2053	44.80	116.60	III
1955	5/31	2335	44.70	116.70	IV
1963	9/6	1924	44.80	117.10	IV
1965	11/7	1642	44.97	116.93	4.3
1966	12/30	0352	44.87	116.94	4.2
1981	9/29	0540	44.693	116.993	3.6
1983	3/29	0137	44.790	116.881	3.4
1984	8/10	0727	44.981	116.970	3.8
1984	9/19	0132	45.079	116.766	3.8

Seismological Observatory (BMO) operated northeast of Baker between 1961 and 1975 and their instrumental determinations of epicenters probably contributed to more accurate locations for the 1963, 1965, and 1966 events. Because of the inadequate seismic instrumentation coverage, all of the other epicenter locations relied heavily on intensity patterns in the sparsely populated area.

Population is unevenly distributed in the Blue Mountains region, and is mostly concentrated in small, widely spaced communities and scattered ranches. Large mountainous areas and semiarid tracts are unpopulated, while river valleys have attracted most of the settlement. These factors strongly influence intensity patterns, especially for earthquakes having magnitudes less than 4 (Zollweg and Jacobson, 1986). Zollweg and Jacobson (1986) stated that these intensity-assigned epicenters are accurate to no better than 25-50 km. It is also possible that some intensity designations are underestimated because the intensities are based on descriptions of felt reports in the small towns, and might therefore be a minimum if the earthquake actually occurred many km away in an uninhabited area, as opposed to within the town limits. Caution should therefore be exercised in assuming a maximum *expected* intensity or magnitude for the area based on this short seismic history.

Figure 20 shows the locations of the earthquakes compiled in Table 1, together with microseismic aftershock events recorded in 1984 (Zollweg and Jacobson, 1986). A cluster of earthquakes is evident, roughly centered in the vicinity of the microseismic events. Figure 21 shows the epicenter locations composited on the Cenozoic fault map compilation used in figure 7. It is clear from this compilation that the majority of earthquakes in the region corresponds well with the areas of most intense deformation associated with pull-apart basin development.

The earliest recorded earthquake in this area occurred on October 14, 1913, at about 3:00 PM (PST). With a Modified Mercalli intensity of VI this is the strongest shock known to have occured in this region. Based on newspaper accounts obtained during the preparation of this report, the epicenter has been revised to a location at 45.1 ° N, 116.7 ° W, from the previous location at 44.8 ° N, 117 ° W (C. Stover, USGS, written comm., 1989). The normal method of assigning locations based on newspaper accounts is to pick coordinates in increments of 0.1 ° near the place of highest intensity, while accommodating patterns of lower intensity distribution from surrounding areas.

Segments of the newspaper article from the October 14, 1913, Baker (Oregon) Democratic Herald read as follows: *"It is reported that at Homestead chairs were rocked and dishes rattled, while at Landore the shock was more severe... The shock was over a minute in duration. At Landore windows were broken, stoves rocked and dishes shaken from their shelves. At Cuprum, Idaho, the inhabitants of the village were badly scared, running from their houses... Four years ago a similar shock was felt in this locality but was not of such long duration nor as severe. Telephone communications with Landore cannot be established... Ballards Landing plainly felt the shock."*

An important observation regarding this earthquake is its long duration, reported to be about one minute. Increased duration is known to greatly augment the destructive capacity for even moderately strong earthquakes (Page and others, 1975). No record of the 1909 quake has been found other than the mention in the newspaper article. Based on the accounts of the damage in the town of Landore, the epicenter may have been close by, although it would be more helpful to obtain accounts to the north and east (C. Stover, written comm., 1989). It is therefore possible that this quake may have reached an intensity of VII (or greater) towards the north in the Hells Canyon area. There are apparently no felt reports from the remote, and at the time essentially uninhabited area in the canyon.

An intensity VII earthquake that struck southwest Idaho on May 13, 1916, was previously located at 44 ° 45'N, and 117 ° 00'W on National Geophysical Data Center records. A search of local newspaper accounts yielded good descriptions of the event in the (Cambridge, Idaho) News. Segments of the article are as follows: *"Last Friday evening at about 7:30 the most distinct earthquake shock ever felt in Cambridge was experienced by our people. According to descriptions given the News, it came with a rumbling sound not unlike a sudden strong wind, and lasted from two to three seconds. Most people agree that the vibrations were from southwest to northeast. Doors rattled and lamps swung. No damage of any kind was done, so far as can be learned. Boise people, at least a few of them think that the vast amount of water accumulated at Arrow Rock dam is responsible for the shocks. Boise has always claimed everything in sight, so it is not surprising that they should try to supercede God Almighty in the matter of earthquakes. Then, again, Saturday evening another light shock was felt by those who were*

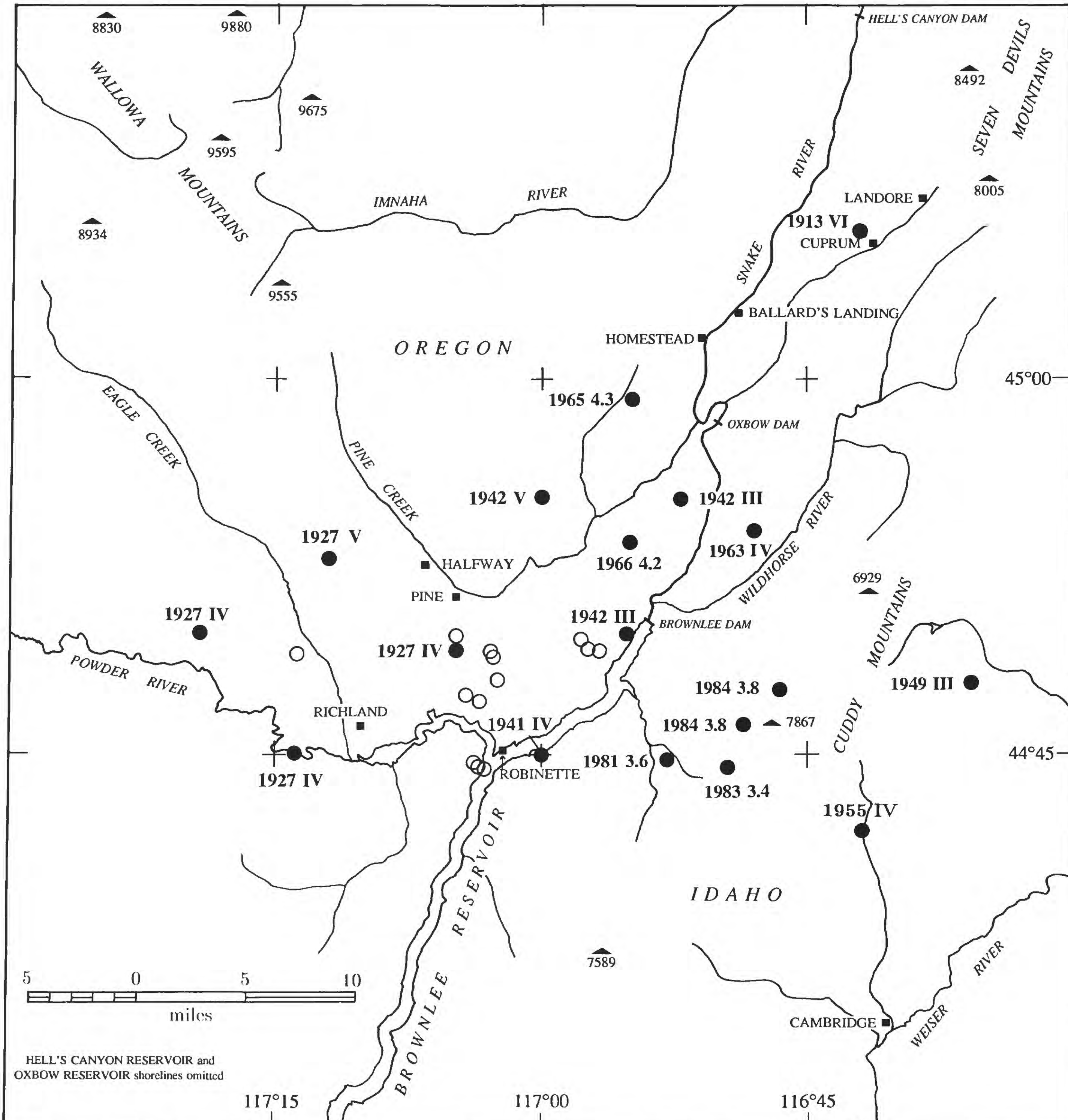

Figure 20 : Location of all earthquake epicenters (1913 - 1984) in the region of this investigation. Open circles = microseismic epicenters recorded in 1984 (Zollweg and Jacobson, 1986); solid circles = all other earthquakes. Richter magnitude or MM (modified Mercalli) intensity follows year.

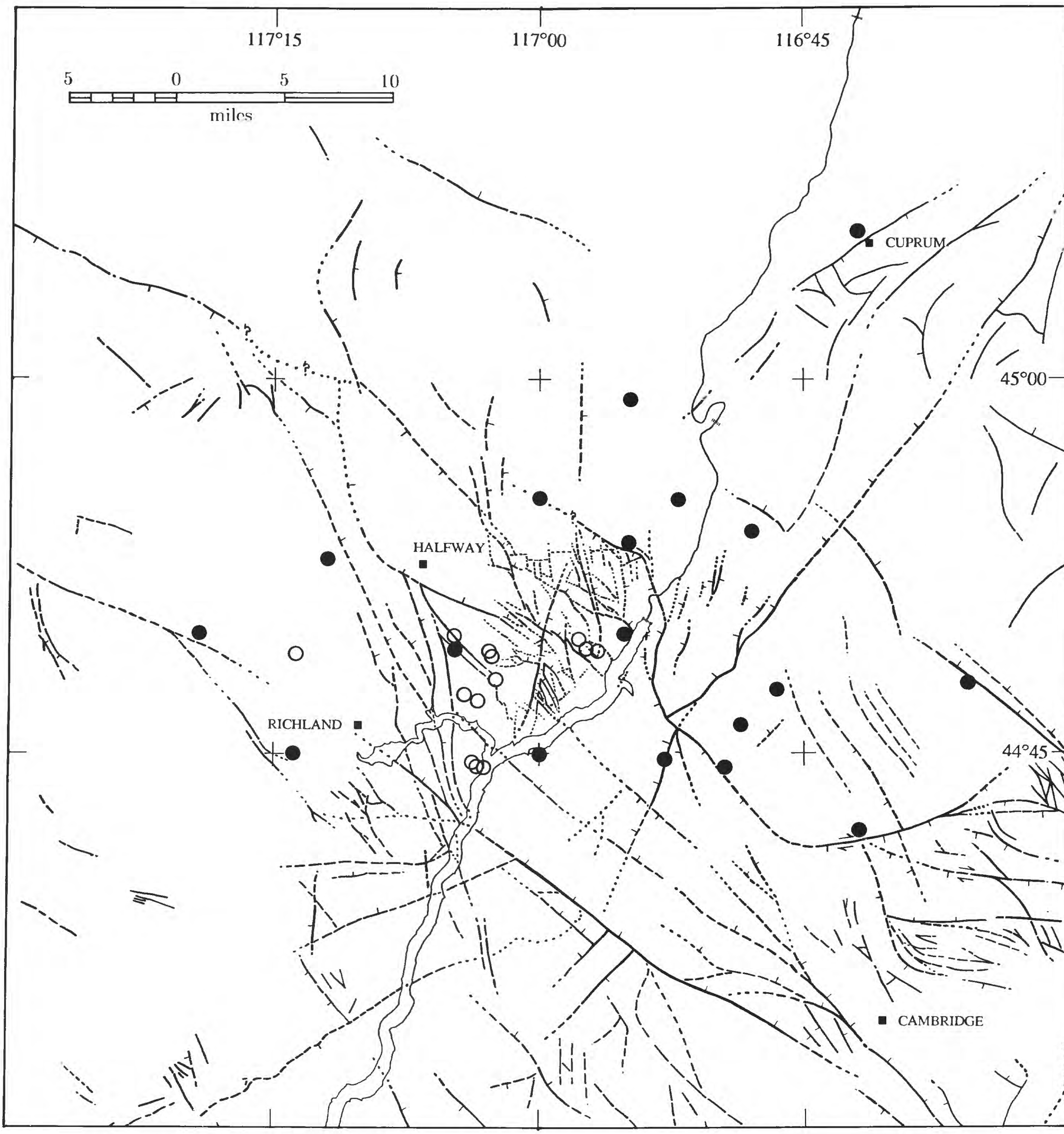

Figure 21 : Cenozoic fault map with all known historic and recent earthquake epicenters (1913 - 1984). Open circles = microseismic epicenters recorded in 1984 (Zollweg and Jacobson, 1986); solid circles = all other earthquakes.

awake at 11:30. This one was not so distinct and was only noticed by a few."

The News also relates the following account from Boise: *"At 7:26 Friday evening Boise experienced the most violent earthquake shock in the history of the city. The tremor lasted about three seconds and was more in the nature of an upheaval than a wave. In the downtown district people rushed from the buildings to the streets. No damage has been reported. The quake was the second in a fortnight, the last one having been recorded on April 30. At Weiser, 80 miles west, the quake was felt with exceptional violence. A new gas well, in which a flow was struck ten days ago, showed remarkable increases in pressure immediately after the shakeup. It caught fire and hundreds of people were out watching the shooting flames. In Boise several brick chimneys were partly wrecked. The tremor was not felt to the east of Boise, either at Twin Falls or at Hailey. Twenty-five miles north, at Emmet, the quake was violent and alarmed the inhabitants. Nampa, to the south, also felt the quake, as did Idaho City, 36 miles north. Windows rattled at Payette. Friday nights violent earthquake here was followed Saturday evening by a slight shock at 4 minutes after 9 o'clock. It was noticed only in the cities higher buildings and not at all at street level. No damage was done."*

The Baker (Oregon) Democratic Herald reports in the May 14, 1916 issue that in the vicinity of Arrow Rock Dam, Idaho, irrigation canals were damaged. The speculation from Boise residents (Cambridge, Idaho, News, May 13, 1916) that the accumulation of large amounts of water at Arrow Rock Dam coupled with the moderate damage reported to irrigation works near the dam indicates the possibility that this shock may have been a reservoir-induced earthquake. The date of closure of Arrow Rock Reservoir was October 22, 1914, and a high-water mark (or filling) was recorded on June 12, 1915 (J. Hanson, Bureau of Reclamation, oral comm., 1989). Because the main shock occurred over 11 months after filling of the reservoir, the temporal correlation is not very conclusive. This quake was felt at Baker, Oregon; Spokane, Washington; and Anaconda, Montana. Seismographs at Pasadena, California recorded this earthquake at a magnitude of 5.3 (C. Stover, USGS, pers. comm., 1989). The short duration of this earthquake probably helped keep damage to a minimum. Based on these newly aquired newspaper accounts this earthquake's previous location has been revised to a location at 44.2° N, 116.5° W (C. Stover, USGS, pers. comm., 1989). This location is roughly coincident with the intersection of the Crane Creek fault zone and the Long Valley fault system (plate 4).

On April 8, 1927 at approximately 9:00 PM (PST) an intensity V earthquake struck the Pine Valley - Richland area (fig. 20). This main shock was followed by three aftershocks at about midnight that night and at 1:30 AM and 6:00 AM (PST), April 9, all were reported at about intensity IV. The main shock was reported as felt widely over a sparsely settled area. At Richland and Halfway entertainments were interrupted (Earthquake History of the United States, 1982). Older local inhabitants recall that the earthquake struck during a school play at the Eagle Valley High School, resulting in the quick evacuation of the audience and participants. No other readily available accounts of these earthquakes have been located. Because the main shock was felt strongly in both Halfway and Richland, it is possible that the epicenter was located between those two towns, possibly along the N 30° W trending group of faults that comprise the proposed west transfer fault zone (fig. 7 and 13).

An intensity IV earthquake was reported felt on December 23, 1941, at 9:48 AM (PST) in the small town of Robinette, Oregon (fig. 20), at the confluence of the Powder and Snake Rivers (United States Earthquakes, 1941). Robinette was razed prior to inundation by Brownlee Reservoir in 1958. This shock may have been precursory to an intensity V event on June 12, 1942, at 2:30 AM (PDST) felt strongly at Halfway where dishes were broken. A table was reported upset at Eagle Valley (Richland area), and the shock was also felt at Pine, Richland, and as distant as Baker, Oregon. Two intensity III aftershocks were reported felt in Pine on June 12 at 2:35 AM (PDST), and on June 13, shortly after 11:00 PM (PDST). These earthquakes were felt in the same general vicinity as the earthquakes in 1927. Again, it is posible that the N 30° W trending west transfer fault zone could have harbored the causative fault. Because slight damage was reported, and the main shock was felt in Baker, Oregon, about 35 miles southwest of Halfway, Oregon, this main shock may have been stronger than the 1927 main shock.

One intensity III earthquake in 1949, and one intensity IV earthquake in 1955 are located approximately along the northeast and southeast flanks of the Cuddy Mountain uplift respectively (fig. 20). No pertinent information regarding these quakes appears in catalogues, however, a few local inhabitants in the town of Cambridge, Idaho, remember the May 31, 1955 event. The editor of the local newspaper remembers it as a strong shaking. This epicenter was located (probably by coincidence) in the immediate vicinity of the Rush Peak fault (fig. 7). There are many faults with large Late Cenozoic displacements in the Cambridge area. These earthquakes may be related to extensive normal (and strike-slip ?) faulting along the northeast and southeast sides of the Cuddy Mountain uplift.

The first earthquake to strike the region in 21 years occurred on September 7, 1963, at 12:24 PM (PDST) and was assigned an intensity IV. It was reported felt at Brownlee Dam and Oxbow Dam, and is located between these facilities (fig. 20). There is some confusion about assignment of a Richter magnitude of 4.0 to this event. In the September 7, 1963 edition of the Baker, (Oregon) Democratic Herald the manager of the Blue Mountains Seismological Observatory (BMO), near Sparta, Oregon, is quoted as saying that the shock registered an intensity of four on the Richter scale. Richter readings are recorded as magnitudes, whereas intensities are rated according to felt reports on the Modified Mercalli (MM) scale. It is therefore uncertain if the event had an instrumented Richter magnitude assignment. The article also states inhabitants in the community of Oxbow, Oregon felt the quake, and that it rattled dishes there.

On November 7, 1965 at 16:41 (GCT) a magnitude 4.3 quake was recorded in the Oxbow-Brownlee area. On December 31, 1966 at 03:51 (GCT) a magnitude 4.2 quake was recorded in the same general area. There are apparently no local newspaper accounts of these quakes, and there is some doubt concerning the magnitudes. The Blue Mountains Seismological Observatory, as well as most others at this time, used M_b or body wave determinations from short-period, P-arrival amplitudes. In practice, M_b for small events prior to about 1975 was often erroneously determined from P^* or P_g amplitudes at 1 or 2 stations, with P_n or P commonly being too weak or noisy to be observed in the appropriate distance range (Zollweg and Jacobson, 1986). Valid P or P_n magnitudes may have been averaged together with the erroneous values. These types of determinations commonly resulted in inflated magnitudes. These earthquakes may have been in the same magnitude range as the 1981-1984 events (i.e. 3.4 - 3.8), and probably did not exceed a magnitude of 4.0 (Zollweg, USGS, oral comm., 1988).

Records of the 1965 and 1966 events have been tentatively re-examined and epicenters have been relocated. Analysis of records from high-gain stations at Bozeman, Dugway, Eureka, BMO, and others suggest that the epicenter for the 1965 event is slighty northeast of the old location at 44.9°N, 117°W, and the 1966 event was slightly southeast of the previous location at 44.7°N, 117°W (C. Stover, USGS, written comm., 1989). The new relocated epicenter for the 1965 event is 44.97°N, 116.93°W; and for 1966 the relocation is to 44.87°N, 116.94°W (Table 1, fig. 20). These relocated epicenters are accurate to about 15 kilometers (C. Stover, USGS, personal comm., 1989).

Recent Seismicity -- The 1981-1984 Earthquake Sequence

The first locatable earthquake to occur in the area in almost 15 years was recorded at 0539 (UTC) on September 29, 1981. It had a magnitude of 3.6 M_L and was felt at Cambridge, Idaho (Stover and others, 1982). Another earthquake occurred in this area at 0136 (UTC) on March 29, 1983, and registered a magnitude of 3.4 M_L. This earthquake was widely felt throughout the Brownlee - Oxbow area. Seismograms of these earthquakes are somewhat similar (Zollweg and Jacobson, 1986), and may be related to the same source area. For a more detailed description of the geophysical aspects of these events, the reader is referred to Zollweg and Jacobson (1986).

Figure 22 is an intensity map for the 1983 event, constructed from interviews of local inhabitants conducted in 1988 (Baumhoff and Smith, Boise State Univ., written comm. from S. Wood, 1989). Although some people had difficulty remembering the earthquake, most people close to the instrumental epicenter remembered specific details related to intensity. In all cases people who felt the March 29 event also felt the October 28, 1983 Borah Peak M_s 7.3 quake, and could temporally distinguish between the two. The desciptions in and immediately adjacent to the intensity IV zone are very similar, with most reporting several sharp up-down jolts; in the Oxbow area the quake is reported as "one sharp jerk", and "a little tremble" (Baumhoff and Smith, Boise State Univ., written comm. from S. Wood, 1989). Zollweg and Jacobson (1986) reported that the quake was felt in the Halfway area,

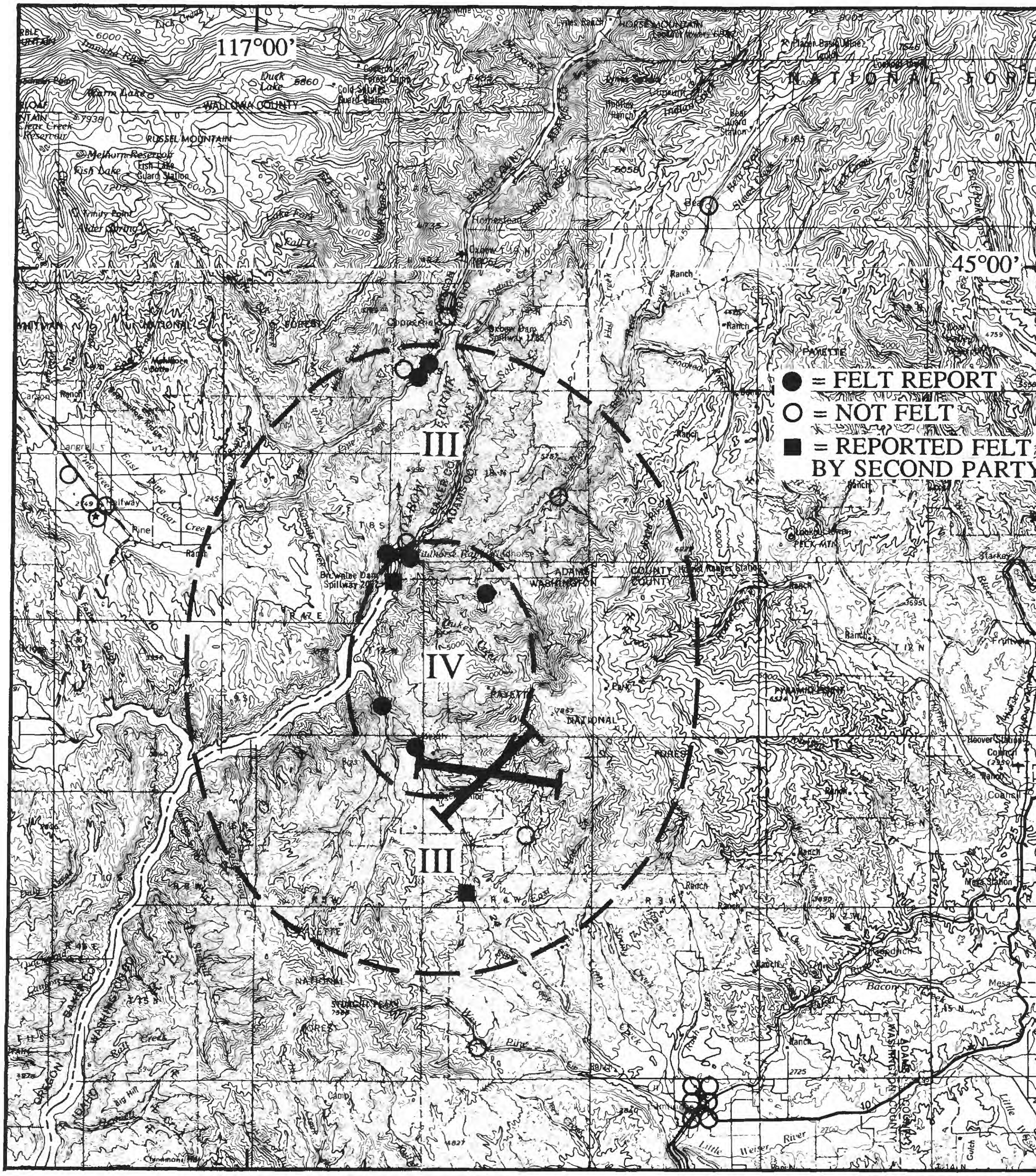

Figure 22: Intensity map for March 29, 1983 Richter = 3.4 earthquake. The X corresponds to the replotted epicenter and error axes for the event. The "intensity center" suggests that this quake occurred near the intersection of the Cuddy Mountain and Brownlee faults. Intensity III - slight, resembling vibrations caused by heavy traffic, felt by some; Intensity IV - moderate, rocking of free-standing objects, felt by most. (Interviews conducted by A. Baumhoff and R. Smith, Boise State Univ. 1988. Intensity contours compiled and felt reports reviewed by T. Toppozada, Cal. Div. Mines and Geology, 1989).

however several people interviewed in 1988 in this area had no recollection of that quake. The "intensity center" suggests that the epicenter may have been about two miles northwest of Zollweg and Jacobson's replotted epicenter, although the deviation is not too significant (T. Toppozada, Cal. Div. Mines & Geol, written comm., 1989). This intensity center is very near the intersection of the Cuddy Mountain and Brownlee faults. The good instrumental epicenter location coupled with the intensity patterns clearly shows that this earthquake occurred on the Idaho side of the reservoir, and not in the Pine Valley-Richland area. Figure 23 shows the original NEIS locations and the redetermined epicenters for the 1981-1984 main shocks, and the 95 percent error bounds for each event.

On August 10, 1984, at 0726 (UTC), a magnitude 3.8 M_L earthquake occurred, preceeded by a magnitude 2.7 event at 0750 (UTC) on August 9 and by magnitude 2.8 events at 0337 (UTC) and 0614 (UTC) on August 10 (Zollweg and Jacobson, 1986). The foreshocks were not locatable, and no aftershocks were recorded. Only the main shock was reported felt, with most reports coming from the Richland-Halfway area. On September 19, 1984, at 0132 (UTC), a second magnitude 3.8 M_L quake occurred in the same general vicinity as the August 10 main shock. Felt reports again are from the Halfway-Richland area (Zollweg and Jacobson, 1986). It is uncertain how these reports were collected or if residents in the Brownlee area were interviewed. These earthquakes were felt west of Richland, Oregon by crews monitoring the Hole-In-The-Wall Gulch landslide (Jacobson and others, 1985).

Zollweg and Jacobson (1986) stated that the seismograms of the 1984 earthquakes are so similar as to suggest sources within a few kilometers of each other. The 1984 recordings share only very general characteristics with the 1981 and 1983 quakes. They also concluded that all of the events recorded regionally came from a relatively small source area. The 1983 event is the best-located of this group, with the major horizontal axis of the 95 percent confidence ellipsoid having a half length of about 5 km, roughly corresponding to the location accuracy (Zollweg and Jacobson, 1986). The 1981 location is more intermediate in accuracy, being accurate to about 20 - 25 km; the 1984 events are poorly located and are accurate to about 20 - 50 km (Zollweg, oral comm., 1988). Focal depths could not be determined with any reliability for these earthquakes.

Recording with up to five portable seismographs was maintained from October 3-7, 1984, during Zollweg and Jacobson's microseismic aftershock survey. A station was installed at Sparta, Oregon, on October 9, and was run an additional 3 weeks. Thirty four tectonic earthquakes were recorded. Phases of 15 events were recorded on three or more stations, with magnitudes ranging from M_L -0.8 - 1.8. Focal depths tended to be between 5 and 11 km below the datum of +1 km above mean sea level, with ninety-five percent confidence limits on these depths between 1 - 2 km (Zollweg and Jacobson, 1986).

The better located events clustered in three areas (fig. 24) that are too far apart too be due to hypocentral uncertainties (Zollweg and Jacobson, 1986). The most active area was group A, accounting for about half of the located epicenters. Other events with S-P signatures comparable to group A were noted throughout their survey, but were unlocatable. Group B had the largest event (M_L) 1.8, and represents a sequence of earthquakes lasting only 14 minutes. Group C represents another earthquake sequence lasting 37 minutes.

FAULT PLANE GEOMETRY AND PATTERNS OF SEISMICITY

The main shocks of 1981-1984 were not sufficiently well recorded at regional stations to justify independent or composite fault-plane solutions. Also, the small number of stations in the portable network prevented single-event, fault-plane solutions for any of the microseismic events (Zollweg and Jacobson, 1986). However, a composite fault-plane solution can be determined if it is assumed that all of the events were generated by the same applied tectonic stress field. Figure 25 shows such a composite fault-plane solution using the portable network data and regional data from three of the larger events. This solution appears to be fairly well controlled, indicating nearly pure normal faulting on a plane oriented about N 30°W (Zollweg and Jacobson, 1986).

Because of the independent clustering of the microseismic events on the Oregon side, coupled with good evidence that at least the 1983 event was centered on the Idaho side, it appears unlikely that there is only one causative fault. The microseismic event groups A and B are located very near the

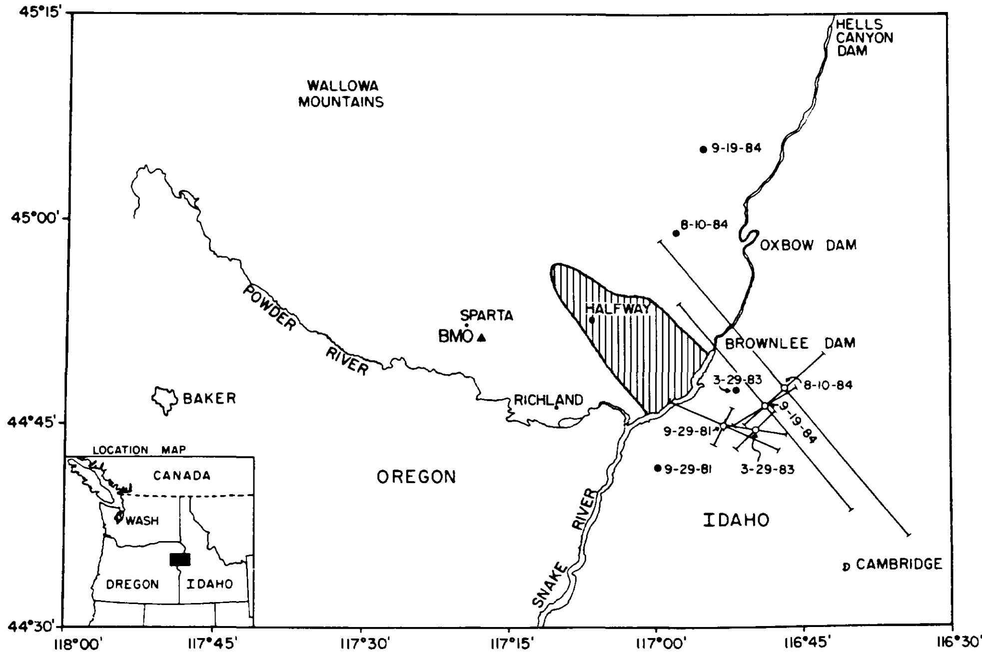

Figure 23 : Regional map showing preliminary epicenter locations for the 1981-1984 sequence (solid dots), and redetermined epicenters (open dots) with 95 per cent error bounds. Stripped area is Pine Valley. (adapted from Zollweg and Jacobson, 1986)

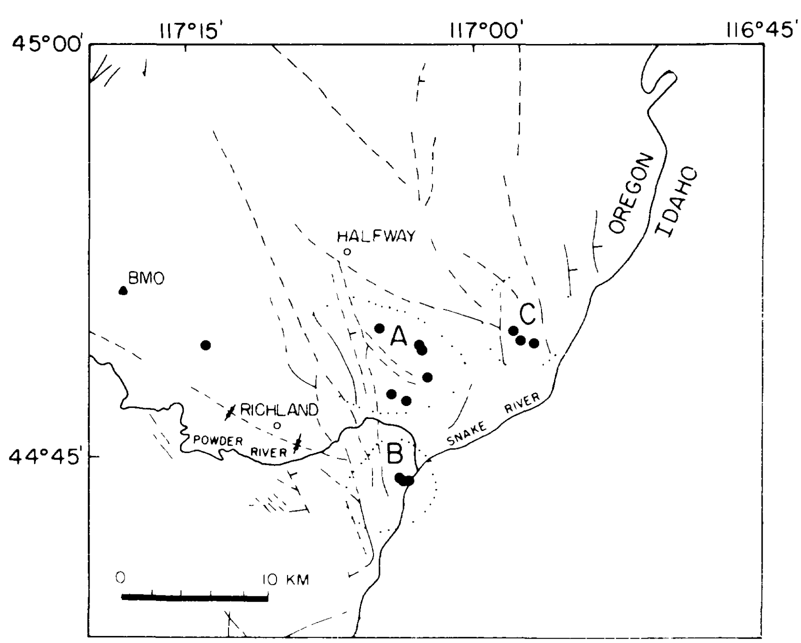

Figure 24 : Location of microearthquakes determined in 1984 portable instrument survey. Cenozoic fault traces shown in Oregon are generalized. These earthquakes are clustered within three discrete groups, labeled A, B, and C. Dotted boundaries serve to differentiate the locations of the earthquake clusters. (adapted from Zollweg and Jacobson, 1986)

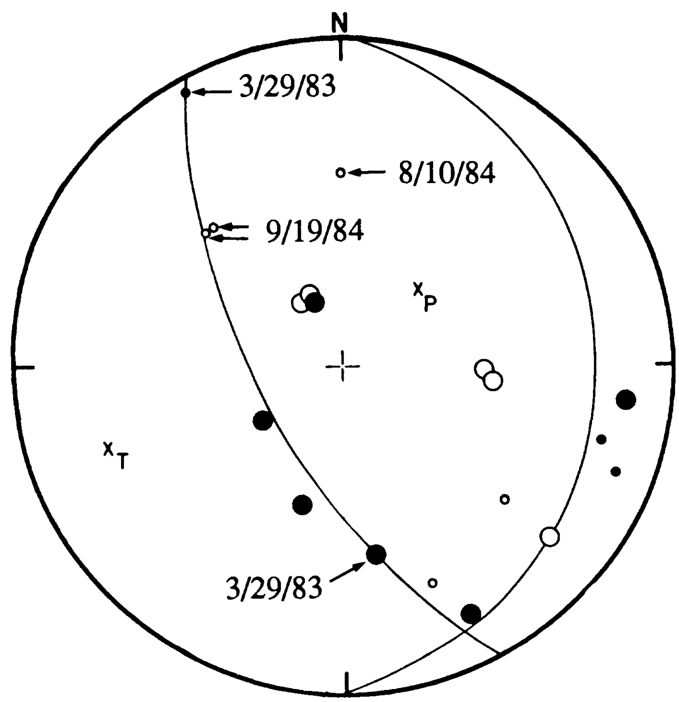

Figure 25 : Composite lower hemisphere fault plane solution. Regional data from larger earthquakes in 1983-1984 contributed as indicated, the remainder of the solution incorporates portable instrument data from microearthquake survey of aftershocks in October, 1984. Solid circles represent compressional first motions, open circles represent dilitations, and larger circles represent better-quality polarity determinations. The P axis is located at N45°E, plunging 65°, and the T axis is at W20°S, plunging 20°. Normal faulting is indicated along a plane orientated N29°W dipping at 70°. (adapted from Zollweg and Jacobson, 1986)

northwest-trending group of faults between Richland and Halfway proposed to be pull-apart basin bounding transfer faults in this report (fig. 6 and 13). Zollweg and Jacobson (1986) stated that these could only be the causative faults if their dips were high at the surface and much lower to the east (if the faults planes were listric at depth). It is assumed that Zollweg and Jacobson were referring to the epicenter locations for the larger events, as well as the microseismic events. However, all of the events in groups A and B (fig. 24) are 2 km or less from the most easterly location of the proposed west transfer fault zone (fig. 6 and 13). These transfer zone faults could therefore be the causative faults for the microseismic aftershocks in groups A and B even if they had moderately steep easterly dips (about 70°-80°).

It is known that the Cuddy Mountain fault dips at about 65° NW at the surface to the northeast of these faults (plates 2 and 3). It is unlikely that the west transfer zone faults would have listric east dipping planes at depth because the very heterogenous northwest dipping Mesozoic basement rocks would present strong structural and lithologic barriers to low angle faulting cross-cutting at that orientation (plate 3). A fault plane geometry requiring both the east and west transfer zone faults to be listric or shallow normal faults may be contradictory to deep-seated lithospheric extension expected below pull-apart basins. It is therefore more likely that the west transfer zone faults, as well as the east transfer zone fault(s) (i.e., Brownlee fault) are high to moderate-angle dipping (70°-80°) normal faults.

The development of the Pine Valley pull-apart basin at its particular location may be related to controlling lithologic and structural elements in the Paleozoic-Mesozoic basement fabric. The greatly heterogeneous lithologic belts in the northeast trending basement fabric are nearly perpendicular to the proposed northwest trending basement strike-slip fault zones related to the Pine Valley pull-apart basin (figs. 2, 13). This strong structural-lithologic trend may be related to the right-step in the right-lateral strike-slip fault zone (GPZ)-- perhaps directly contributing to mechanical barriers and asperities which have resulted in a refraction or stepping of the basement dextral strike-slip fault zone. Indeed, clustered zones of seismicity have been directly related to boundaries between "suspect" or tectonostratigraphic terranes in the southwestern United States (Wheeler and Bollinger, 1984), and the possible reactivation of Mesozoic basement structures. It is notable that the strike-slip--transfer zone junctions (fig. 13) are located approximately along the tectonic contact between the rocks of the Wildhorse Fm. (Wallowa Terrane?) and the hard and brittle cherty rocks of the Baker Terrane (fig. 2). Microseismic earthquake locations are also approximately coincident with these junctions (figs. 21 and 24). It is also interesting to note that this same tectonostratigraphic terrane boundary swings west and then north (fig. 2) under Baker Valley and Grande Ronde Valley.

Microseismic earthquake group C (fig. 24) is several kilometers east of the group A events and does not appear to be related to the same group of faults. Group C is located in the extremely deformed area shown in figure 8 that is proposed to be an area of structural transition between the terminus of a basement right-lateral strike-slip fault and right-stepping dip-slip transfer zone faults. The ends of strike-slip faults probably play an important role in the concentration of stresses and in the consequent formation of secondary structures (Aydin and Page, 1984). It is therefore possible that groups C and A correspond to such areas of concentrated stress. It is also possible that this group could be related to faulting at seismogenic depths on the Brownlee fault, whose surface exposure is about 5 km east. Again, for this fault to be the causative structure, the dip of the Brownlee fault would have to become increasingly low-angle at depth, about 45° for a hypocenter at 10 km depth. While this is possible, it is not the most simple explanation.

Based on intensity patterns and instrumental epicenter locations for the 1981 and 1983 earthquakes it appears that these events were located on fault(s) on the Idaho side of the reservoir. Instrumental location error boundaries and uncertainties in intensity patterns for the two main 1984 shocks raise questions about which side of the reservoir these events actually were located. Felt reports come mainly from the Richland-Halfway area, whereas the instrumentally determined epicenters are well to the east in Idaho. Microseismic earthquake groups A and B are very proximal to the west transfer fault zone between Halfway and Richland. It is possible that these groups are aftershocks of the 1984 main shocks that may have been located on these same faults (or in Idaho), while the 1981 and 1983 events may have been located on a northwest trending fault(s) on the Idaho side (i.e., Brownlee fault?).

This would explain why the fault-plane solution appears well controlled for the N 30°W trend, whereas intensity data and instrumental epicenter locations strongly suggest at least two source areas on opposite sides of the reservoir.

CONCLUSIONS

Recurrent and recent moderate levels of seismicity have occurred in the vicinity of Brownlee Dam. Large Late Cenozoic normal faults have been mapped and described on the Idaho side of the reservoir. Other significant Late Cenozoic structures have been identified by several workers in the region and are compiled on maps in this report. A tectonic model is proposed that explains the regional fault geometry within a plausible framework, and raises important questions regarding possible correlations to patterns of seismicity.

A limited search of readily available historical accounts of some of the earlier earthquakes was attempted. Intensity data from accounts of the 1913 and 1916 events were used to relocate their approximate epicenters. These relocations should be recognized as still poor in relative accuracy, and these events could actually have been larger events in less populated areas. No new information was found regarding the earthquake sequences of 1927 and 1941-1942 in the Pine Valley - Richland area. These earthquakes may have been located on a northwest-trending group of faults located between the towns of Halfway and Richland.

Local earthquakes assigned Richter magnitudes in 1965 and 1966 are probably overestimates. Body-wave (M_B) formulas used to compute magnitudes *may* have resulted in inflated magnitudes. Based on tentative reexamination of some regional records for these two earthquakes, the epicenters have been slightly relocated.

General observations regarding the 1981-1984 earthquake sequence have been summarized and some new data has been presented. Intensity data acquired for the 1983 M_L 3.4 earthquake clearly show that this shock was on the Idaho side of the reservoir. This event has the best instrumented location of any of the larger events recorded regionally. Together, the intensity data and the instrumented epicenter offer compelling evidence for active faulting on the west or southwest side of the Cuddy Mountains. The two available compressional first motions for this quake are compatible with normal faulting on the northeast trending Cuddy Mountain fault, as well as the northwest trending Brownlee fault (T. Toppozada, written comm., 1989), (when considered independently of the rest of the composite solution). The choice of fault planes is therefore not constrained. It should be emphasized that these two faults are large and proximal structures.

The 1913 earthquake is too remote to be associated with seismicity in the Brownlee Dam area. This was a strong earthquake of apparent long duration. This earthquake is poorly located and could have occurred farther north in the Hells Canyon area or more easterly in the Seven Devils Mountains. It is also possible that the source may have been more local, perhaps associated with the Indian Creek fault. There should be some concern should such an earthquake strike in that vicinity in the future. Also, earthquakes in 1949 and 1955 (and others to the east) suggest seismicity on the east side of the Cuddy Mountains, in the vicinity of the towns of Council and Cambridge, Idaho. Large Cenozoic faults with appreciable displacements have been identified in these areas. If these faults were capable of producing strong shocks, then there would be some geohazard implications for a populated area.

It has been noted that earthquakes in this area occurring between 1927 and 1984 may have a temporal pattern of recurrence. Zollweg and Jacobson (1986) described a historical record suggesting periodicity with intervals of 11 to 21 years between cycles of events consisting of several main shocks within a 3 year span. While it may be tempting to speculate that this pattern could hold in the future, the historical record is too short to begin to define a periodicity with reliability. However, based on the short historical record, the earthquake recurrence interval in this area is one of the highest in the region.

Some workers have suggested that pull-apart basins tend to have recurrent or frequent moderate levels of seismicity, typically without strong or large earthquakes ($M_L > 5.0$) (Koide and Bhattacharji, 1977; Mann, 1983). However, the Cuddy Mountain uplift is a horst block not directly related to pull-apart development, but probably to the same broad zone of right-lateral strike-slip faulting. The

Brownlee fault borders the east side of the Pine Valley pull-apart basin, and bounds the Cuddy Mountain uplift east of the Cuddy Mountain fault. This dual-character makes fault classification difficult, and renders any attempt at suggesting possible magnitudes that could be generated a dubious proposition. Also, possible syntectonic relationships with secondary structures, such as the Cuddy Mountain fault, further complicate the interactions. Although historically intensities have apparently not exceeded V in the Pine Valley - Richland area, there is evidence that activity may now have shifted to faults on the Idaho side of the reservoir. It is therefore unclear if historical patterns of seismicity will persist, or if large faults in Idaho have passed into an active phase.

Only a limited field search of some segments of a few faults has been conducted for evidence of Holocene surface rupture. Conclusive evidence of Holocene rupture along the Brownlee fault and Cuddy Mountain fault, if it exists, probably can not be established without more detailed examination, such as trenching. Steep to nearly vertical gradients on local fault scarps and footwall scarps in soft Mesozoic sedimentary and metasedimentary rocks may be evidence of vigorous Late Quaternary activity along both of these structures. Large Late Quaternary displacements in analogous regional structural settings such as Baker Valley and Grande Ronde graben (Holocene) have been reported.

The lack of identification of Holocene rupture may be largely the result of the near absence of recognizable Holocene deposits in the area, and may also be related to the magnitude of Holocene seismic events. Bonilla (1988) concluded that faults that only produce infrequent earthquakes having magnitudes less than 6 may be very hard to identify using near-surface geologic methods. Shaking from earthquakes having magnitudes between 5 and 6 can damage structures, particularly if the structures are near the seismogenic fault (Bonilla, 1988), and shaking from earthquakes in this range can cause landslides (Keefer, 1984). Shaking from earthquakes in the lower end of this range can cause high horizontal accelerations, especially if near the source and the focus is shallow. Peak horizontal accelerations for an earthquake of magnitude m_b 5.0 and M_s 5.4 have been recorded at 0.72 g and 56 cm/sec respectively (Shakal and others, 1986).

Based on evidence from the 1981-1984 earthquake sequence, it is shown that seismic events of moderate strength can occur proximally to Brownlee Dam. If the Pine Valley basin is indeed an active pull-apart basin, then all faults related to the pull-apart are potentially active.

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