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SURFACE FAULTING ACCOMPANYING THE BORAH PEAK EARTHQUAKE AND SEGMENTATION OF THE LOST RIVER FAULT, CENTRAL IDAHO

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ABSTRACT

On the morning of 28 October 1983, the M_S 7.3 Borah Peak earthquake struck central Idaho and formed a Y-shaped zone of surface faults that is divided into a southern, a western, and a northern section. The total length of the surface faults is 36.4 ± 3.1 km, and the maximum net throw is 2.5 to 2.7 m. The near-surface net slip direction, determined from the rakes of striations in colluvium, averaged 0.17 m of sinistral slip for 1.00 m of dip slip.

The 20.8-km-long southern section is the main zone of surface faulting and coincides with the Thousand Springs segment of the Lost River fault. It has the largest amount of net throw, most complex rupture patterns, and best evidence of sinistral slip. The surface faults include zones of ground breakage as much as 140 m wide, *en echelon* scarps with synthetic and antithetic displacements, and individual scarps that are nearly 5 m high.

The 14.2-km-long western section diverges away from the Lost River fault near the northern end of the southern section. The net throw on this section is generally less than 0.5 m but locally is as much as 1.6 m. The new ruptures are poorly developed across the crest and north flank of the Willow Creek hills; they are mostly downhill-facing, arcuate scars, perhaps incipient landslides, that may overlie a deeper zone of tectonic movement.

The northern section, at least 7.9 km long, is on the Warm Spring segment of the Lost River fault and has a maximum net throw of about 1 m. The pattern of surface faulting on this section is simple compared to the other sections. A 4.7-km-long gap in 1983 surface faults separates the northern and southern sections but contains an older scarp of late Pleistocene age.

Geologic, seismologic, and geodetic data from the earthquake suggest that barriers confined the primary coseismic rupture to the Thousand Springs segment of the fault. The rupture propagated unilaterally to the northwest from a hypocenter near the southeastern end of the segment. The southeastern boundary of the segment is marked by an abrupt bend in the range front, a 4-km-long gap in late Quaternary scarps, and transverse faults of Eocene age that intersect the Lost River fault.

The northwestern boundary of the Thousand Springs segment is at the junction of the Willow Creek hills and the Lost River fault. Here, the southern and western sections of surface faults diverge and there is a gap in the 1983 scarps. During the first few weeks after the main shock, the large-magnitude and large stress-drop aftershocks clustered near this barrier. Later, aftershocks were mainly northwest of the barrier on the Warm Spring and Challis segments, and showed that strain adjustments eventually affected the entire northern part of the Lost River fault. Fault-scarp morphology and the bedrock geology suggest that the boundary between the Thousand Springs and Warm Spring segments has probably ruptured less frequently and had less net slip during much of the late Cenozoic than the interior of the adjacent segments. The 1983 faulting shows that although segment boundaries can stop or deflect primary ruptures, secondary surface faulting can occur on adjacent segments of the main fault. A late

Pleistocene scarp in the 1983 gap suggests that infrequent earthquakes, perhaps larger than the 1983 event, might break through a segment boundary and thus release strain on two adjacent segments.

INTRODUCTION

At 8:06 a.m. (MDT) on 28 October 1983, an M_S 7.3 earthquake struck east-central Idaho, caused two deaths in the town of Challis, and produced about $12.5 million of damage in the sparsely populated surrounding area (Stover, 1985). A northwest-trending zone of surface faults 36.4 ± 3.1 km long formed during the earthquake. Most of the 1983 fault scarps are along the surface trace of the Lost River fault, the southwestern boundary of the Lost River Range, but a prominent west-trending splay diverges from the range-front fault at the western base of Dickey Peak (Figure 1). This splay extends across the Willow Creek hills that separate Thousand Springs Valley on the south from Warm Spring Valley on the north (Figure 2).

The Borah Peak earthquake was the first earthquake to produce surface faulting in the intermountain west since the 1959 earthquake at Hebgen Lake, Montana. It

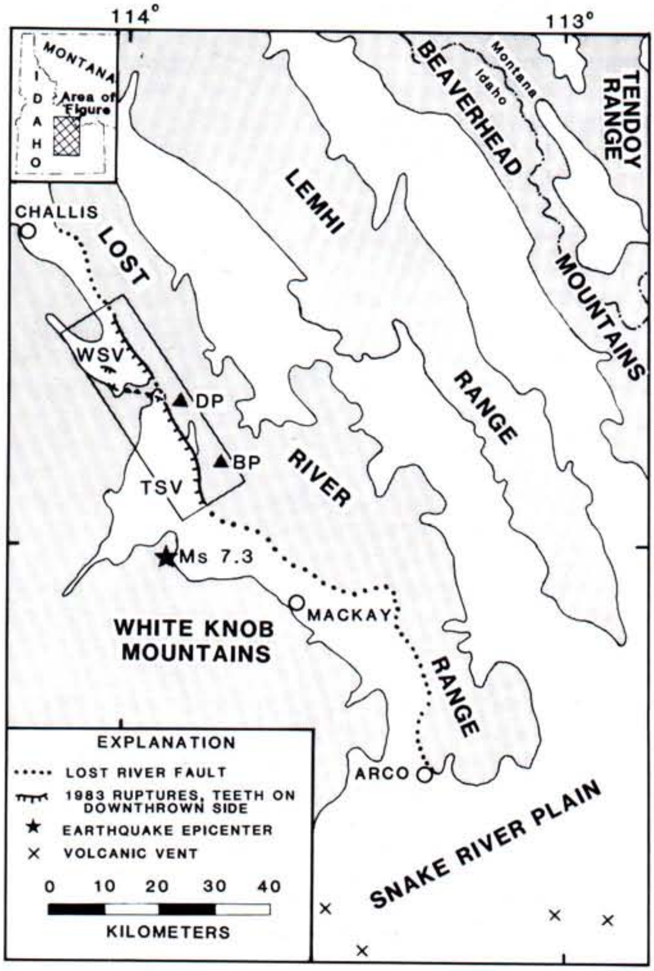


FIG. 1. Location of Borah Peak earthquake. Mountainous terrain is shaded. Solid triangles are Borah Peak (BP) and Dickey Peak (DP); Warm Spring Valley is WSV and Thousand Springs Valley is TSV. Rectangle shows area of Figure 2. Earthquake epicenter is location of main shock from Dewey (1985).

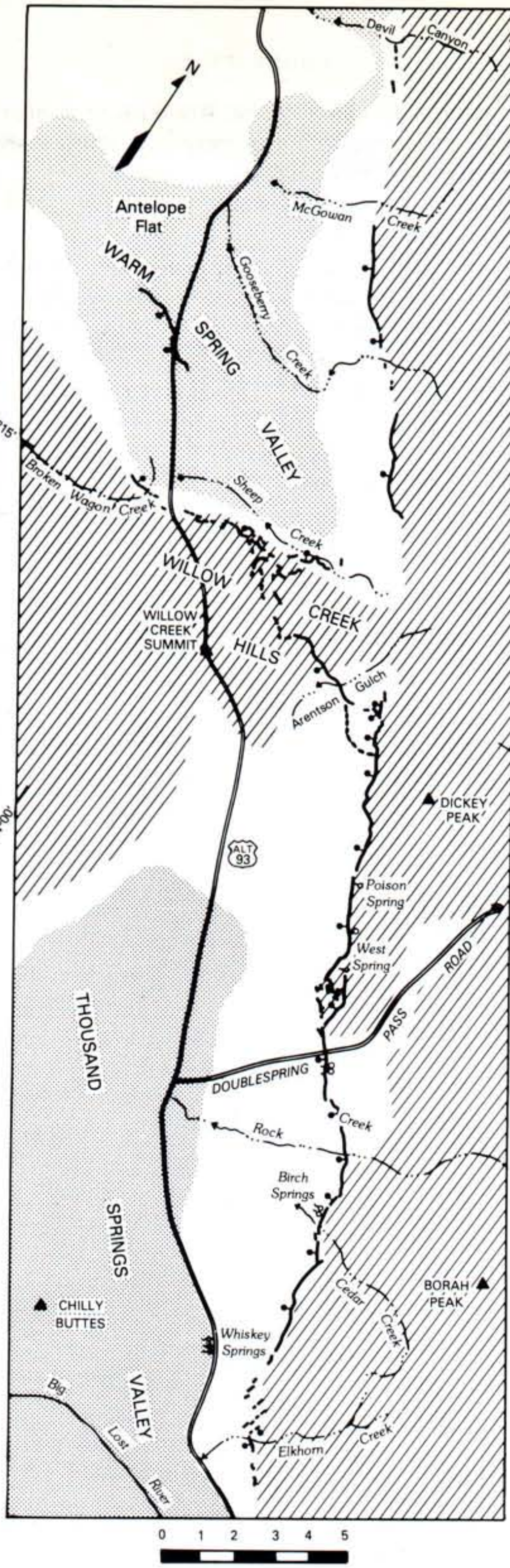


FIG. 2. Generalized map of fault scarps and ground ruptures associated with Borah Peak earthquake. eavy lines are prominent scarps, bar and ball on downthrown side of fault; dashed lines are poorly fined scarps or cracks. Stippled areas are valley bottoms; hatchured areas are mountainous parts of st River Range and Willow Creek hills. Plates 1 to 3 are detailed maps of scarps.

provides a rare opportunity to evaluate the behavior of a Basin and Range normal fault during a major earthquake using current geologic, seismologic, and geodetic methods. The surface faulting, aftershocks, and ground deformation associated with the earthquake show how individual segments (as expressed at the surface) of a major range-front fault responded to the propagation of a large coseismic rupture.

The Borah Peak earthquake was a seismologically unexpected but geologically predictable event. It occurred in a region of relatively low seismicity between the most active part of the Idaho seismic zone to the west and the Yellowstone-Hebgen region of the Intermountain seismic belt to the east (Smith and Sbar, 1974). Relocation of instrumentally recorded earthquakes in central Idaho for the 20 yr preceding the 1983 earthquake shows that the area within 25 km of the 1983 epicenter had no earthquakes of magnitude 3.5 or larger (Dewey, 1985). In addition, no foreshocks greater than magnitude M_L 2.0 were recorded within 50 km of the surface faults in the 2 months before the main shock (Richins *et al.*, 1985). Thus, the seismological record did not indicate the seismic potential of the Lost River fault. In contrast, prominent Holocene and upper Pleistocene scarps along the Lost River and adjacent range-front faults (Nakata *et al.*, 1982; Scott *et al.*, 1985) attest to large prehistoric earthquakes. Furthermore, a trench excavated in 1976 across the scarp on the Lost River fault just north of Doublespring Pass road documented a surface-faulting event of Holocene age (Hait and Scott, 1978). A section of the backfilled trench was exposed in the 1983 fault scarp. Thus, before the earthquake, the geological evidence indicated a potential for movement on the fault that was not apparent from the few decades of seismological data.

Geologic Setting

The Lost River Range and nearby ranges north of the Snake River Plain and east of the Idaho Batholith have a structural style and topographic expression (Figure 1) similar to those elsewhere in the Basin and Range province (Reynolds, 1979). In east-central Idaho, the ranges are composed of allocthonous Precambrian and Paleozoic sedimentary rocks that were complexly folded and faulted as they were transported east to northeastward on gently dipping thrust faults. Movement on the thrust faults began perhaps in the Early Cretaceous and continued into the early Eocene (Ross, 1947; Skipp and Hait, 1977; Ruppel, 1978). Later, tectonic events included the formation of northeast-striking Eocene normal faults (Baldwin, 1951), Eocene and Oligocene volcanism, and late Cenozoic regional uplift associated with the formation of the Snake River Plain. Finally, late Cenozoic extensional block faulting defined the modern basins and ranges (Reynolds, 1979). Movement on the normal faults bounding one or both flanks of the ranges (Baldwin, 1951; Skipp and Hait, 1977; Ruppel, 1982) probably started in middle Miocene time (Reynolds, 1979; Scott *et al.*, 1985); the movement postdates the Challis Volcanics which are mostly Eocene (McIntyre *et al.*, 1982). However, much of the present topography probably results from late Pliocene and Pleistocene displacements on the normal faults (Baldwin, 1951; Scott *et al.*, 1985).

The amount of throw on many of these range-bounding normal faults is uncertain; along the southwest side of the Lost River Range, it may be as much as 6.1 km (Skipp and Hait, 1977). Near Thousand Springs Valley, where the 1983 faulting occurred, the net Cenozoic throw on the Lost River fault is at least 2.5 km, based on the present 1.9 km of topographic relief between Borah Peak and the floor of the valley, plus an estimated 0.6 to 0.9 km of fill in the valley (Crosthwaite *et al.*, 1970). The bedrock geology here suggests more than 5 km of structural throw, of which perhaps 3 km may be Neogene (Skipp and Harding, 1985).

Characteristics of 1983 Surface Faults

The surface faults and ground ruptures (Figure 2; also see Plates 1 to 3 at the back of the issue) associated with the earthquake form a Y-shaped pattern about 36 km long that can be divided into three parts: (1) a 20.8-km-long southern section that coincides with the surface trace of the Lost River fault (Baldwin, 1951) from near Elkhorn Creek to east of Arentson Gulch; (2) a 14.2-km-long western section that extends from east of Arentson Gulch discontinuously west-northwestward across the Willow Creek hills, which separate Thousand Springs and Warm Spring Valleys, and into Antelope Flat; and (3) a 7.9-km-long northern section that extends from 2 km north of Sheep Creek to just south of McGowan Creek, which also coincides with the surface trace of the Lost River fault. North of McGowan Creek, cracks with little or no displacement are present locally for another 5 km, to just south of Devil Canyon. The northern and southern sections are separated by a 4.7-km-long gap where no new scarps formed along the Lost River fault adjacent to the Willow Creek hills (Figure 2); subdued fault scarps of probable late Pleistocene age show that ground breakage has occurred during past earthquakes in the 1983 gap. The western section of surface faults diverges from the range-front fault at the south end of the gap. These relationships (discussed in the section "Segmentation of the Lost River Fault") suggest that the gap marks the boundary between two segments of the Lost River fault.

The observed surface faulting characterizes the slip at depth only in a general manner. At places, the apparent dip and strike of the fault determined by the relationship of the scarps to the topography are dramatically different than the dip and strike measured in natural and man-made exposures. These discrepancies can be caused by irregularities in the fault plane, multiple fault strands, refraction of the fault in surficial materials, and slumping. For example, the strike and dip of faulting, and amount and sense of slip deduced from surface evidence differ from measurements derived from seismologic and geodetic data. Along the southern section of the fault, the throw averages 1.1 m and exceeds 1 m along 58 per cent of its length. The average coseismic displacement at depth was estimated as 1.4 m from the seismic moment (Doser and Smith, 1985) and 2.1 m from modeling geodetic data (Stein and Barrientos, 1985). From these data, it appears that some of the slip at depth failed to reach the surface, as noted by Barrientos *et al.* (1985) and Stein and Barrientos (1985).

Most of the ground breakage we describe is believed to be tectonic, although some small cracks with minor displacement may have formed by differential settling, by slumping, or by oscillatory motion associated with the passage of seismic waves.

Determining the total length of the 1983 surface faults is complicated by the difficulty of identifying tectonic versus nontectonic ruptures at the ends of the ground breakage. Tectonic ruptures near the northern end of the surface faults certainly extend to about 0.6 km southeast of McGowan Creek (S, Plate 3) because ruptures there have both sinistral-slip (left-lateral slip) and dip-slip components. However, discontinuous fractures extend 5 km farther north to just south of Devil Canyon (T, Plate 3). These fractures have as much as 4 cm of throw and may be tectonic or merely the result of shaking on moderately steep slopes. They coincide with an inflection in the slope of the range front that marks the bedrock-alluvium contact.

At the southern end of surface faults near Elkhorn Creek, the extent of the tectonic ruptures is uncertain because they merge with a feature that may be an incipient landslide near point B (Plate 1). Additional ground breakage located about 1 km farther to the southeast (A, Plate 1) introduces more uncertainty because here a nearly continuous, 0.2-km-long zone of fractures crosses a broad spur and nearly reaches the adjacent drainage to the southeast. Another fracture about 30 m long crosses a small spur farther southeast, making the whole zone 0.3 km long. The open fractures mainly indicate extension but, at places, the southwest (valley) side is downdropped as much as 5 cm, and minor sinistral or dextral (right-lateral) displacements are present locally. We do not know if these fractures are tectonic or if they are from minor downslope movement of surficial materials caused by the shaking.

The minimum span of 1983 surface faults along the Lost River fault is 33.3 km, the distance from 0.6 km southeast of McGowan Creek to 0.5 km southeast of Elkhorn Creek. The maximum length is 39.5 km if the northernmost cracks near Devil Canyon and the open fractures 1.6 km southeast of Elkhorn Creek are included. The average of the minimum and maximum lengths is 36.4 km. For comparison, the length of surface faults associated with the three major historic earthquakes on normal faults in the Basin and Range are: 48 km for the M_S 7.2 Fairview Peak, Nevada, earthquake; 26 km for the M_S 7.5 Hebgen Lake, Montana, earthquake (Bonilla *et al.*, 1984; Doser, 1985); and 59 km for the M_S 7.6 Pleasant Valley, Nevada, earthquake (Wallace, 1984a; Bonilla *et al.*, 1984).

Most of the fault scarps produced by the Borah Peak earthquake are at or near the alluvium-bedrock contact and displace unconsolidated alluvium and colluvium; this relation holds true for most fault scarps associated with major historic earthquakes in the Basin and Range (Slemmons, 1957; Witkind, 1964; Wallace, 1984a). We found bedrock exposed in the scarps at four locations: (1) directly downslope from Birch Springs (D, Plate 1); (2) on the hillslope north of Doublespring Pass road (G, Plate 1); (3) on the west flank of Dickey Peak near the junction of the southern and western sections (L, Plate 2); and (4) in the modern channel of Arentson Gulch (near M, Plate 2). At the Dickey Peak site, striations on the bedrock record the slip directions of two pre-1983 displacements (discussed in the section "Recurrent Surface Faulting"). In a few locations such as near West Spring (Plate 1) and near Elkhorn Creek (Plate 1), cracks in outcrops of shattered bedrock were reopened, and rock fragments fell from the outcrops. We believe that these features are related to seismic shaking and not to surface displacement on deep-seated faults.

The strike of the surface faults and their southwesterly dips are consistent with the seismologic data. Displaced man-made features and striations in alluvium and colluvium show that movement during the earthquake was primarily normal slip with subordinate amounts of sinistral slip. Four analyses of teleseismic data for the main shock yielded focal mechanisms with preferred nodal planes that strike N14°W, N20°W, N22°W, and N42°W, and that dip 45° to 60°SW (respectively, Barrientos *et al.*, 1985; Dewey, 1985; National Earthquake Information Service, 1983; Dosier and Smith, 1985). The strikes of these nodal planes average N25°W and compare favorably to the N23°W strike and southeast dip of the new scarps in the Thousand Springs Valley. The tension axes from the focal mechanisms are nearly horizontal and strike northeast-southwest, consistent with the trend of Neogene basins and ranges in east-central Idaho (Figure 1).

The focal mechanisms and modeling of the geodetic data (Barrientos *et al.*, 1985) both indicate a sinistral-slip component that generally agrees with the field measurements of slip. However, the seismologic data generally show a greater proportion of lateral to dip slip than the field data. These relationships suggest that, as the rupture extended upward from the hypocenter, the proportion of the lateral slip to dip slip diminished.

At most places, we found it difficult to measure the amount of lateral slip across the fault although reliable measurements of net throw could be made in many locations. Our measurements of net throw are the tectonically induced vertical displacement of the ground across the entire zone of obvious surface deformation associated with the 1983 earthquake (Plates 1 to 3). Our most accurate measurements were made on previously unfaulted Holocene deposits such as the colluvium on hillslopes and the alluvium of modern streams. Our values compensate as much as possible for antithetic faults, local warping and backrotation, and complex ground breakage adjacent to the fault, and are the same as the net cumulative vertical tectonic displacement of Swan *et al.* (1980). Reliable measurements of the 1983 slip were difficult to obtain where the new breakage was superposed on preexisting scarps. On Plates 1 to 3, we also include measurements of scarp height (Bucknam and Anderson, 1979), sinistral and dextral slip, and other pertinent characteristics of the 1983 surface faulting.

Vertical accelerations in the vicinity of the 1983 fault scarps apparently did not exceed 1 *g*. We found no evidence of stones or other objects on nearly flat surfaces being thrown from depressions that marked their original location; such features would be good indicators of very high vertical accelerations.

Southern section. The southern section is the main section of surface faults and coincides with the Thousand Springs segment of the Lost River fault as defined by Scott *et al.* (1985). Here, the surface faults have the largest throw, the best evidence of sinistral slip, and the most complex rupture patterns; they form the highest and most continuous of the 1983 scarps.

On the southernmost 3 km of this section, throw is generally less than 0.5 m, but it increases rapidly northward to more than 1.0 m (Figures 3 and 4). From this point to just south of West Spring (Figure 4), a distance of about 10 km, the throw is commonly more than 1 m and locally more than 2 m. Just south of West Spring, the surface faults enter a 1.5-km-wide block of shattered limestone, here called the West Spring block, where widely spaced, east- and west-facing scarps form a complex pattern. This block has been downdropped to the west at least 200 m relative to the Lost River Range, but it is not clear if it is a tectonic sliver within a range-front fault zone or a slide block that rests on valley fill (Skipp and Harding, 1985). Northward from the West Spring block, the throw is generally about 1.0 m to the gap in 1983 surface faulting (0, Plate 2).

The maximum net throw we measured on the 1983 fault scarps is 2.55 to 2.70 m just south of Rock Creek (F, Plate 1) at the western base of Borah Peak, the highest point in Idaho. The spatial relation between the maximum net throw and the high peak may not be fortuitous in as much as the large topographic and structural relief along this part of the Lost River fault may be indicative of high late Neogene slip rates (Scott *et al.*, 1985).

Our most reliable measurements of net slip directions are from large scarps on the southern section of the fault. Here, clasts dragged along the fault plane formed grooves and corrugations on slightly smoothed gouge surfaces near the base of the free face of some large scarps in colluvium (Figure 5). Ten measurements of the rake of these grooves indicates a sinistral-slip component that averages about 17 per cent of the dip-slip component (i.e., 0.17 m of sinistral slip per 1.0 m of dip slip)

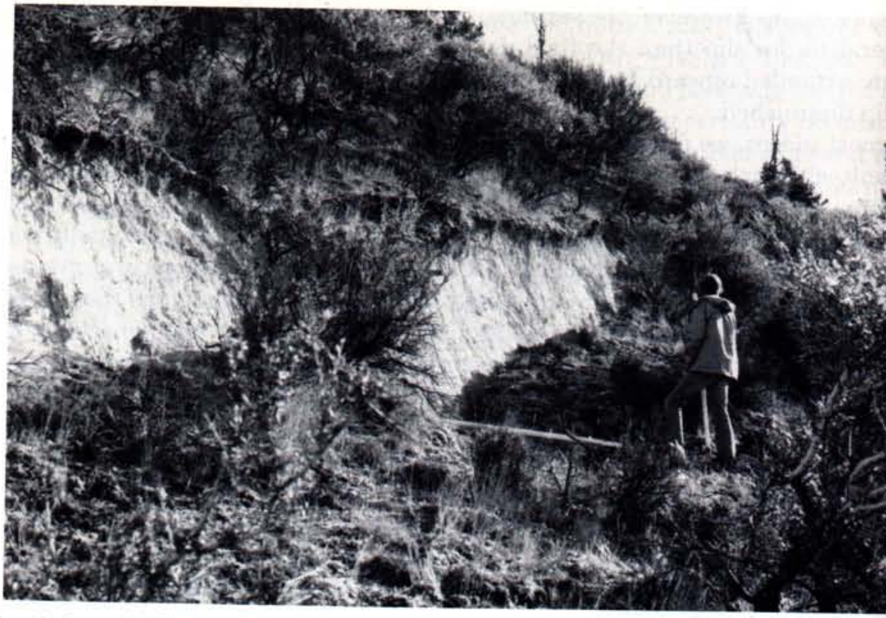


FIG. 3. Large fault scarp about 5 km north of Elkhorn Creek (see Plate 1). The net throw is 1.20 m, and the scarp height is 1.73 m. Overhanging upper part of the free face is anchored by plant roots. Details of corrugations and grooves on smoothed surface near base of scarp are shown in Figure 5. Photograph by A. J. Crone, 3 November 1983.

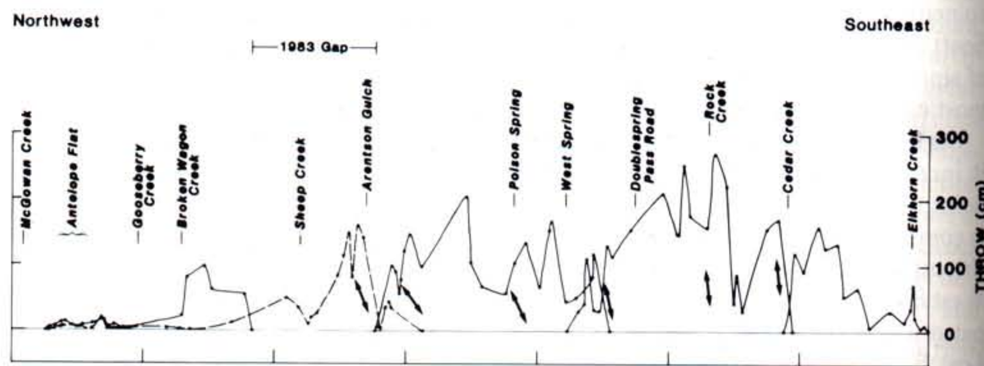


FIG. 4. Amount of net throw on surface faults associated with Borah Peak earthquake. Solid line is throw along Lost River fault; dashed line is throw along western section (measurements shown by solid dots). Throw measurements for western section are located by their distance westward from junction of the southern and western sections (point L; Plate 2). Gap in 1983 surface faults on Lost River fault is shown. Arrows show net-slip direction from rake of slickenlines or displaced cultural features. Vertical exaggeration is 2000×.

along fault planes having near-surface dips of 60° to 90°. The upper part of the free face of large scarps is commonly vertical to overhanging, owing to increased coherency from soil development and plant roots (Figure 3; Wallace, 1977).

Many man-made features that cross the fault also show a sinistral-slip component. The most impressive of these features was at Rock Creek where a concrete irrigation ditch (now removed) with a strike 37° oblique to the main scarp had a throw of 2.90 m and a geometrically corrected left-lateral displacement of 0.43 m (Figure 6). Sinistral slip is visible along three fence lines that cross the fault: (1) about 30 m

FIG. 5. Corrugations and grooves near the base of the large scarp shown in Figure 3. Grooves were formed by clasts dragged along the fault plane. Bars on scale show inches (above) and centimeters (below). Photograph by A. J. Crone, 3 November 1983.

north of Rock Creek (Plate 1); (2) at the east edge of West Spring block (H, Plate 1); and (3) 0.3 km northwest of Poison Spring (K, Plate 1). An estimated 0.5 to 1.0 m of sinistral slip has misaligned the old Doublespring Pass road. On the west flank of Dickey Peak near the northwestern end of the southern section, a jeep trail was displaced 0.90 m vertically and 0.65 m sinistrally (Figure 7).

The complexity in the patterns and amount of displacement on individual scarps along this section varies greatly. At many places, *en echelon* scarps show both synthetic and antithetic displacement. The most extensive breakage is along the 8 km between West Spring and Cedar Creek (Figure 2; Plate 1). The violent shaking, combined with differential movement on individual faults, shattered the ground surface into randomly tilted blocks several meters in width (Figure 8). The zone of ground breakage is up to 140 m wide and commonly has 4 to 8 *en echelon* scarps, some of which can be as much as 1 to 2 m high (Figure 9). Along strike, these scarps typically diminish in height until the ground surface is warped but not broken; farther along strike, the warping dies out completely. As the displacement on one scarp decreases, it usually increases on an adjacent *en echelon* scarp. In plan view,

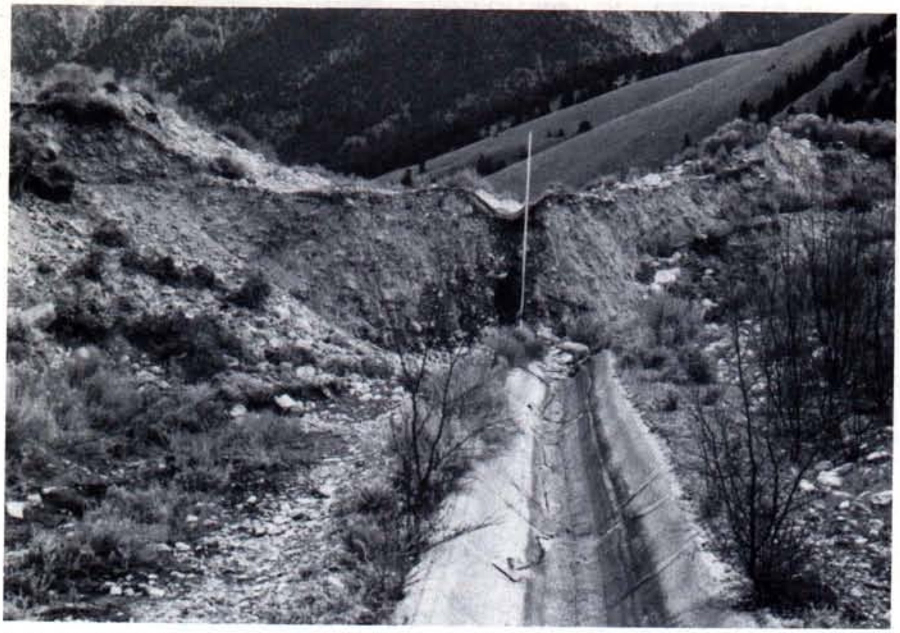


FIG. 6. Faulted concrete irrigation ditch at Rock Creek showing component of sinistral slip (see Plate 1). Vertical displacement of the ditch is 2.90 m, and sinistral slip is 0.43 m. Stadia rod is 4 m long. Photograph by A. J. Crone, 30 October 1983.

FIG. 7. Faulted jeep trail near western base of Dickey Peak at northwest end of southern section (see Plate 2). Net throw is 0.90 m, and sinistral slip is 0.65 m. Stadia rod shows separation of two previously connected points. Rod is 1.5 m long. Photograph by W. E. Scott, 2 November 1983.

the *en echelon* scarps and cracks commonly form a right-stepping pattern, which is characteristic of sinistral movement (Figure 9). The highest scarps commonly bound the upthrown (footwall) block and are best developed where adjacent antithetic faults form grabens. The highest individual scarp, 4.7 m high, is 0.3 km northwest of the Doublespring Pass road (G, Plate 1).

FIG. 8. Ground breakage caused by 1983 faults at Doublespring Pass Road. Zone of faults is 20 to 30 m wide with ground surface broken into numerous blocks. View is to the northeast. Photograph by M. N. Machette (U.S. Geological Survey Archive No. 40), 9 November 1983.

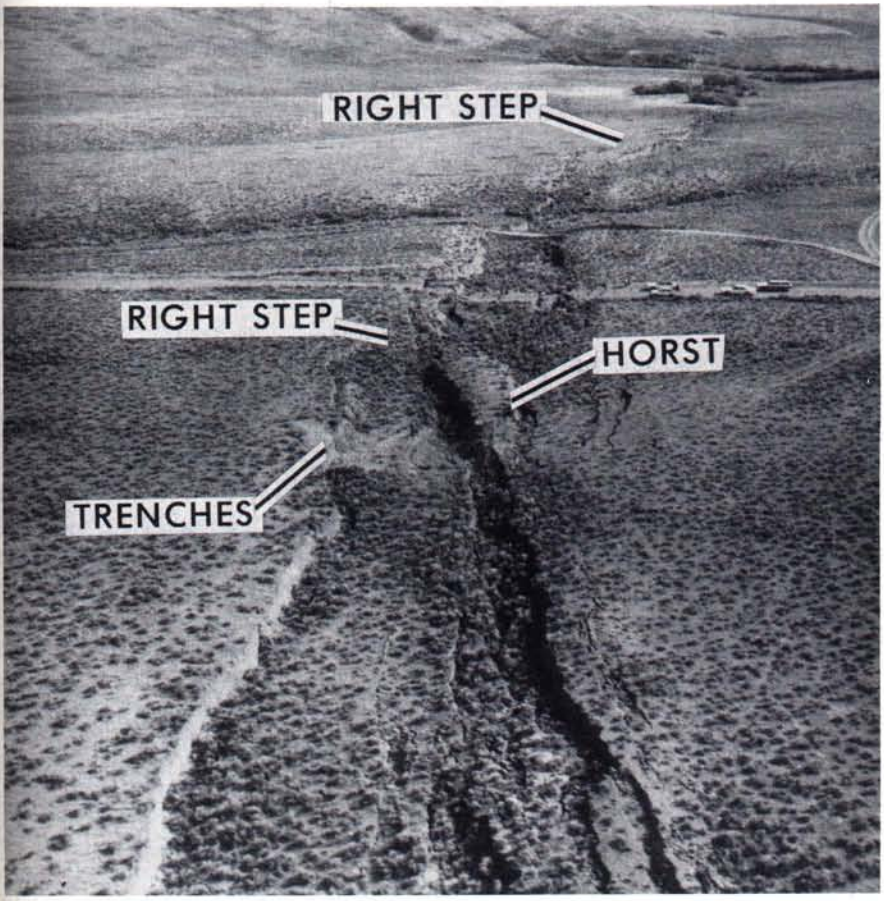


FIG. 9. Oblique aerial photograph of 25- to 50-m-wide zone of complex surface ruptures associated with Borah Peak earthquake near Doublespring Pass Road (see Plate 1). Exploratory trenches (Hait and Scott, 1978; Schwartz and Crone, 1985) and horst are discussed in text. Right-stepping pattern of extensional scarps is evidence of sinistral slip. Figure 10A shows detailed map of scarps in this area. View is to the southeast. Doublespring Pass road is in middle distance, and Rock Creek Canyon is in far distance. Photograph by R. E. Wallace, 3 November 1983.

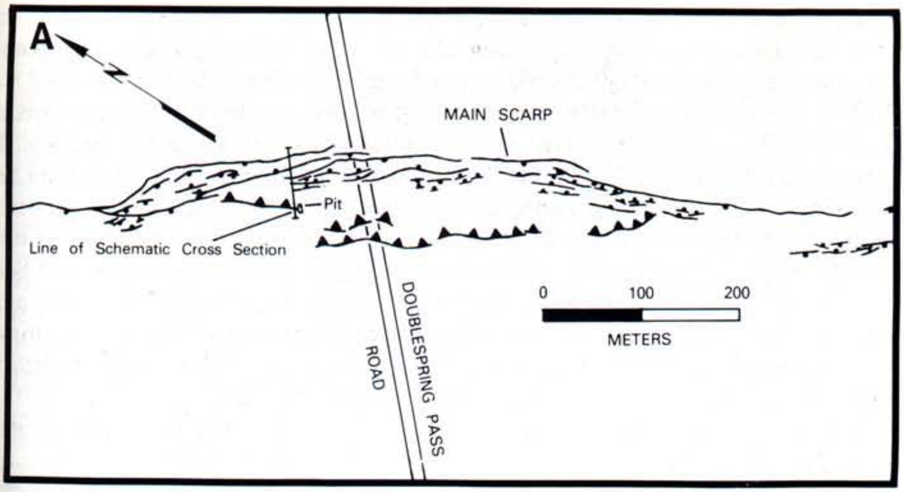
Between Cedar Creek and the West Spring block, most of the 1983 scarps coincide with preexisting scarps, not only along the main fault, but along major graben-bounding faults as well. At a major unnamed drainage about 0.7 km south of Rock Creek (E, Plate 1), a broad graben formed by pre-1983 surface faulting has diverted the westward-flowing, intermittent stream northwestward along the graben. The 1983 surface faulting reactivated the main fault and antithetic faults in this area, accentuating the topographic depression of the graben in which the stream is confined. Another example of new ruptures following preexisting scarps is just north of the Doublespring Pass road at the site of two trenches (Figure 9; Hait and Scott, 1978; Schwartz and Crone, 1985). Hait and Scott's trench and pre-1983 aerial photographs show a well-defined horst block within the graben adjacent to and west of the main fault scarp. During the Borah Peak earthquake, the faults bounding the horst were reactivated, enhancing the topographic expression of the block.

A 0.4-km-long zone of small thrust faults near Doublespring Pass road (Plate 1) shows that minor compressional features can form locally in a tectonic setting that is dominated by extension. The thrust faults are marked by scarps as much as 15 cm high, 30 to 90 m downslope from the main fault scarp. These faults formed in the hanging-wall block adjacent to a complex graben (Figure 10A). A 1-m-deep pit across one of the thrust-fault scarps shows that the near-surface dip of the fault is 28° to the northeast, toward the graben. If this dip does not change appreciably with depth, it suggests that the thrust faults are shallow features which could extend only 20 to 40 m deep. Minor reverse faults also formed during the 1954 Dixie Valley-Fairview Peak, Nevada, and 1959 Hebgen Lake, Montana, earthquakes (Slemmons, 1957; Myers and Hamilton, 1964).

The origin of the thrust faults is uncertain but they seem to be secondary features that do not represent the sense of movement on the main fault zone at depth. Several observations provide some insight into the origin of these thrust faults. Adjacent to the thrust faults, the main scarp has a broad arcuate form in plan view (Figure 10A), similar to a scarp at the head of a landslide. Furthermore, the thrust faults are confined to the central part of the glacial-outwash fan which occupies the valley of Willow Creek, one of the few broad alluvial valleys that cross the fault. Where the 1983 fault scarps cross the fan, the faults probably cut a locally thick section of unconsolidated alluvium. Lastly, seismic-refraction data (Pelton *et al.*, 1985) suggest that the top of the water table may be 27 to 35 m deep near the main scarp and approximately coincident with the projection of the thrust faults into the main fault (Figure 10B). Thus, geologic factors at this site may have contributed to the formation of shallow thrust faults, which are not characteristic of an extensional tectonic setting.

Other small local thrusts on the south side of the West Spring block (G, Plate 1), near West Spring (I, Plate 1) and near the junction of the southern and western sections (near L, Plate 2), are believed to be the result of local near-surface geologic conditions and are not direct indicators of the primary tectonic movement.

The net throw along the southern section is consistently down-to-the-valley except for two areas of uphill-facing scarps near the south end; one area is located on the hillslopes directly north of Elkhorn Creek (Figure 11) and the other is about 1-km farther north. These scarps are visible on pre-1983 aerial photographs. Individual uphill-facing scarps are commonly a few tens to slightly more than 100 m long and are distributed over distances of up to several hundred meters along the slopes. They have a maximum throw of 0.71 m and show no evidence of lateral slip. These scarps have formed small benches as much as a few tens of meters wide on



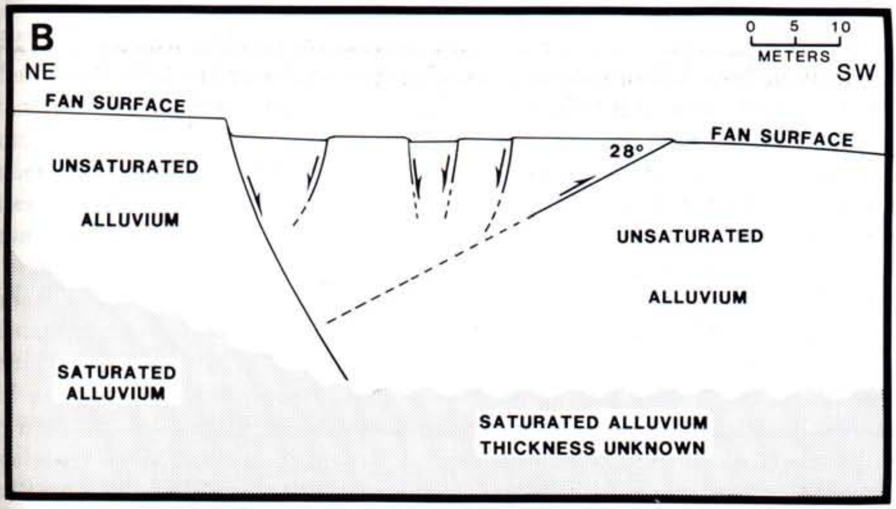


FIG. 10. Detailed map of fault scarps (A) and schematic diagram (B) of thrust faults near Doublespring Pass Road (see Plate 1). (A) Normal faults have ball on downthrown side; thrust faults have barbs on upper plate. Thrust fault exposed in pit dips 28° to the northeast. Note arcuate form of main scarp. (B) Schematic diagram showing geologic relations associated with the thrust faults. Maximum height of thrust faults is about 15 cm. Depth to top of saturated alluvium from Pelton *et al.* (1985).

the steep hillslopes. Intensely jointed and fractured Kinnikinic Quartzite (Ross, 1947) forms small, jagged spires on the hillslopes immediately below the scarps. At places, fresh cracks in the slope colluvium extend into and through the spires, indicating that some of the fracturing is caused by late Quaternary displacements.

Along the trace of the Lost River fault near point C (Plate 1), we saw coarse angular talus concentrated at benches on mountain-slope and alluvial-fan surfaces. The valleyward (downhill) slope of these surfaces projects 1 to 2 m higher than their mountainward (uphill) slope, thus forming catchments for the coarse slope debris and fan alluvium. We found no 1983 ruptures on these benches, but they show that uphill scarps (now largely buried) formed along the Lost River fault prior to 1983.

FIG. 11. Uphill-facing scarp on hillslope directly north of Elkhorn Creek (see Plate 1). Maximum scarp height is 0.71 m. Scarps may be related to downslope lateral spreading of the hillslope. View is to the south-southwest. See Plate 1 for exact location. Photograph by A. J. Crone, 6 November 1983.

The origin of the uphill-facing scarps is unclear; they may be true tectonic features related to fault dislocations at depth or they may be gravity-induced failures not directly related to fault movements at depth. The benches on the hillslopes and the anomalous spires of shattered rock immediately below the scarps both favor a mechanism of gravity-induced slope failure, but this interpretation cannot be proved. If the scarps are gravity-induced, they may be the product of deep-seated lateral spreading and downslope creep of the entire hillslope, perhaps initiated by violent seismic shaking (Radbruch-Hall *et al.*, 1976; Bovis, 1982). Wallace (1984a) proposed such an origin for the Stillwater scarp that formed during the 1915 earthquake at Pleasant Valley, Nevada. Similarly, seismic shaking has been proposed as the triggering mechanism for uphill-facing scarps and ridge trenches in the Carpathian Mountains of eastern Czechoslovakia and southern Poland (Jahn, 1964) and in the Alpine Mountains of New Zealand (Beck, 1968). Alternately, if the scarps are tectonic features, their restricted distribution and sense of displacement (opposite to the displacement of virtually all of the other scarps associated with the earthquake) imply an anomalous and complex local stress field.

Western section. The scarps of the western section are generally smaller, less continuous, and have less evidence of lateral slip than those of the southern section. Westward from the junction with the southern section, displacement on the scarps diminishes rapidly. Near the junction (L, Plate 2), scarps (with as much as 0.85 m of displacement) and cracks are concentrated along two separate strands that merge 0.4 km southeast of Arentson Gulch (M, Plate 2). These strands locally show evidence of dextral and reverse movement. East of Arentson Gulch, the new ruptures locally coincide with older scarps in the alluvium. Immediately east of the Arentson Gulch road, a small but prominent graben formed adjacent to the main scarp. The new scarps across the road have 1.4 m of throw and 0.6 m of sinistral slip (Figure 12). At the road, two hunters saw the main scarp form 20 m in front of them during

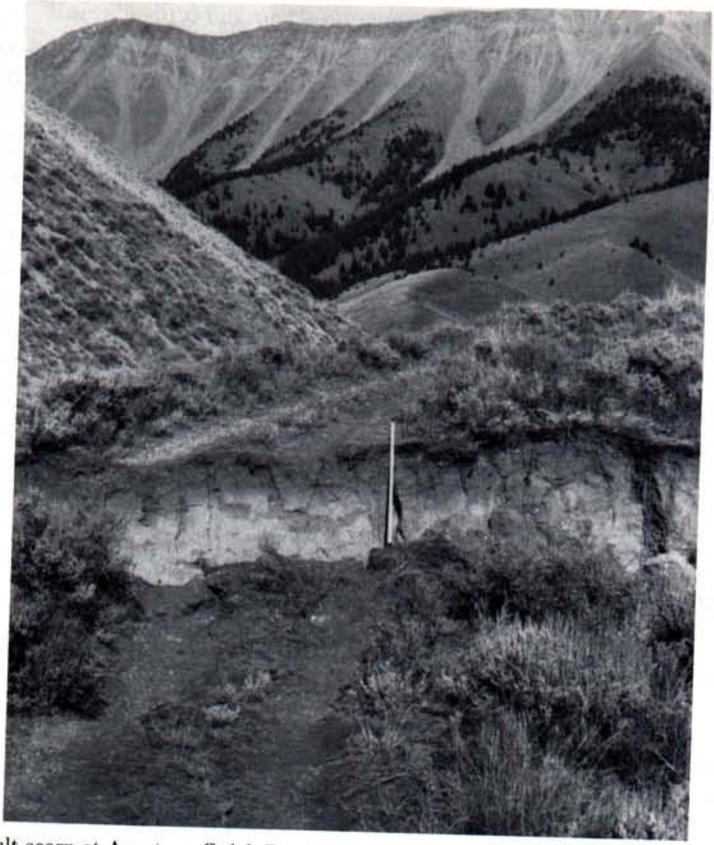


FIG. 12. Fault scarp at Arentson Gulch Road (see Plate 2). Net throw is 1.4 m; 0.6 m of sinistral slip can be seen by misalignment of the road. This is approximately the view that two hunters had as they watched the scarp form during the earthquake (Pelton *et al.*, 1984). Stadia rod is 1.5 m long. Photograph by W. E. Scott, 3 November 1983.

the earthquake (Pelton *et al.*, 1984) and nearby, another hunter saw scarps form on the hillslopes to the west (Wallace, 1984b). Where the main scarp crosses the modern channel of Arentson Gulch, brecciated limestone is exposed in the footwall of the fault.

The maximum net throw on the western section is near Scratching Post Spring west of Arentson Gulch where we measured 1.6 m on a single scarp. Westward, this scarp is generally continuous for about 1.8 km toward a ridge crest in the Willow Creek hills (N, Plate 2). Near the ridge crest, severe local shaking either uprooted healthy trees or snapped them off several meters above the ground; this is the only such site we found along the fault. Local site conditions apparently focused the seismic energy and amplified the ground motion.

Northwestward from the ridge, the scarps are discontinuous, and the net throw is commonly less than 0.5 m. On valley slopes, most new ruptures form a series of aligned, northwesterly trending, downhill-facing, arcuate scars that may represent incipient landslides. Because these scars are generally aligned with the northwesterly projection of the scarps at Arentson Gulch, we suspect that they may be nonsystematic surface breakage above a deeper zone of tectonic movement. The displacement on this deeper zone was insufficient to produce distinct, continuous fault scarps.

Along the north flank of the Willow Creek hills (Plate 2) and in Antelope Flat (Plate 3), the 1983 ground breakage is expressed as discontinuous cracks and small scarps less than 20 cm high. The deformation is limited to small displacement, down-to-the-north cracks on the north slope of the hills and just west of Broken Wagon Creek. Much of the vertical displacement on these cracks may be the result of differential compaction in the alluvium and thus may not reflect tectonic slip. However, the cracks near Broken Wagon Creek do show minor evidence of sinistral slip which implies a tectonic origin. An isolated series of discontinuous 5- to 20-cm-high, down-to-the-south scarps and small cracks near the highway south of Antelope Flat (Plate 3) define the western end of this section. In places, these cracks show evidence of sinistral slip and on a small scale are aligned in a right-stepping *en echelon* pattern. Their tectonic significance is not known, but they lie just west of several isolated outcrops of Mississippian Scott Peak Formation that are surrounded by alluvium (Hays *et al.*, 1978).

Northern section. The northern section of surface faults coincides with the Warm Spring segment of the Lost River fault as named by Scott *et al.* (1985). The scarps on this section are generally superposed on older scarps, and overall, the faults on this section have the smallest, least continuous scarps of those formed in 1983. The pattern of ruptures is simple, and there is evidence of lateral slip in only a few places. The most prominent new scarps on this section extend 3 km northwestward from the north end of the gap (P, Plate 2) and show a maximum throw of about 1 m on the faults. The rupture pattern is simple, and the displacement is commonly concentrated on a single scarp (Figure 13); rarely is the zone of ground breakage

FIG. 13. Largest scarp on the northern section of surface faults (see Plate 2). Net throw is 1.0 m, and scarp height is 1.2 m. Pattern of ground breakage on this section is generally simple compared to the patterns on the other sections. Photograph by W. E. Scott, 29 October 1983.

more than 10 m wide. Farther to the northwest, the 1983 displacement decreases rapidly and the scarp ends about 1 km south of Gooseberry Creek (Q, Plate 2).

Near Gooseberry Creek, pre-1983 aerial photographs show a complex graben about 0.5 km wide that contains as many as four antithetic and two synthetic scarps. The graben extends about 2.2 km beyond the northern end of the prominent new scarps (Q, Plate 2). The largest 1983 displacements we found on old scarps in the graben were 5 cm on an antithetic scarp and 3 cm on a synthetic scarp. In this area, the 1983 fault displacements at depth were probably distributed among several fault strands in the graben system.

Northwest of the graben, 1983 throw along the range-front fault is a maximum of 17 cm. A sinistral-slip component can be recognized locally in this area; our only accurate measurement shows a lateral slip of 40 per cent of the throw (R, Plate 2). Here, there is 5 cm of 1983 breakage superposed on an older 4.5-m-high scarp. The most recent prior surface faulting here is thought to be Holocene because the scarp has steep slope-angles (25° to 28°) (Bucknam and Anderson, 1979) relative to its height. The minor 1983 breakage suggests that this large scarp resulted from earthquakes on the Warm Spring segment, rather than earthquakes similar to the Borah Peak event on the Thousand Springs segment.

Northwest of the unnamed canyon (R, Plate 2), the throw gradually decreases until no new breakage was recognized in a talus slope about 0.6 km southeast of McGowan Creek (S, Plate 3). At the mouth of McGowan Creek Canyon, a 3.4-m-high scarp, also thought to be Holocene in age, has no new ruptures. A careful search of the old scarps further to the northwest revealed no new ruptures, except for small discontinuous cracks in the vicinity of Devil Canyon (T, Plate 3). These cracks generally have no measurable lateral slip, but locally have up to 4 cm of throw. The cracks near Devil Canyon lie along the bedrock-alluvium contact where subdued scarps are preserved. On 22 August 1984, an M_L 5.8 aftershock occurred about 4 km west of Devil Canyon (Zollweg and Richins, 1985; Plate 3), near the northern end of the Warm Spring segment. Subsequent field checks and selected remeasurements of the 1983 scarps on this part of the northern section showed no new breakage or additional slip caused by the aftershock.

Recurrent Surface Faulting

The imposing mountain front of the Lost River Range is distinct evidence of recurrent Quaternary movement on the Lost River fault. Trenching studies, compound scarps, and bedrock slickenlines have provided valuable information about late Quaternary, recurrent surface faulting on the Thousand Springs segment.

In 1984, an exploratory trench was excavated across the complex zone of scarps near Doublespring Pass road at the site of Hait and Scott's (1978) earlier trench (Schwartz and Crone, 1985). The trenches contained evidence of one pre-1983 surface-faulting event that is thought to be early Holocene or latest Pleistocene in age (Scott *et al.*, 1985). The pattern of surface faults in 1983 was remarkably similar to that from the earlier event; small faults with only a few tens of centimeters of displacement were reactivated with similar amounts of slip (Schwartz and Crone, 1985). Similarities in the amount and distribution of slip for the 1983 event and the prior earthquake suggest that the near-surface rupture dynamics at this site were similar during both earthquakes. These similarities support the characteristic earthquake model of Schwartz and Coppersmith (1984).

Most of the 1983 surface faults are superposed on older scarps, thereby forming composite scarps (Mayer, 1984; Nash, 1984). This is especially apparent on the southern section where the scarps now are composed of a steep free face and a wedge of fresh colluvial debris from the 1983 displacements, plus part of the degraded older scarps preserved as vegetated bevels adjacent to the new breakage (Figure 14). In most places, the new breakage makes it very difficult to confidently reconstruct the morphology of the older scarps. Therefore, comparing the 1983 displacements with those from earlier movements based on reconstructions is speculative.

About 2 km west of the summit of Dickey Peak, at least two sets of superposed striae are preserved on shale in the footwall of the new scarps (L, Plate 2). Rakes of the striae record the net-slip directions during two pre-1983 displacements on this part of the Thousand Springs segment (Figure 15). Oxidation and minor weathering of the striated surface show that neither set formed during the Borah Peak earthquake. The relative age of the sets was determined by cross-cutting relationships and their degree of preservation. The younger striae are very fine, shallow, and cover most of the exposed surface. They are best developed on a thin film of secondary calcium carbonate which was deposited on much of the exposed fault plane. The older striae are preserved as a few coarse, deep grooves which, in places, have been partially filled by the secondary calcium carbonate. The older set of striae have a rake of 40°SW on the fault plane that strikes N12°W and dips 53°W; this relation indicates 1.2 times as much sinistral slip as dip slip. The rake 82°NW for the younger set indicates 0.14 times as much dextral slip as dip slip. In 1983, the sinistral slip was 0.17 times as much as the dip slip. Thus, the lateral component of slip during three geologically recent movements on the fault included: (1) a large amount of sinistral slip; (2) a minor amount of dextral slip; and (3) a minor amount of sinistral slip in 1983.

FIG. 14. Composite scarp near Doublespring Pass road showing vegetated bevel (old eroded scarp) above the free face of the new scarp (see Plate 1). Extensive tilting and complex ground breakage in 1983 make it difficult to accurately reconstruct the configuration of the old scarp. Stadia rod is 3 m long. Photograph by A. J. Crone, 8 November 1983.

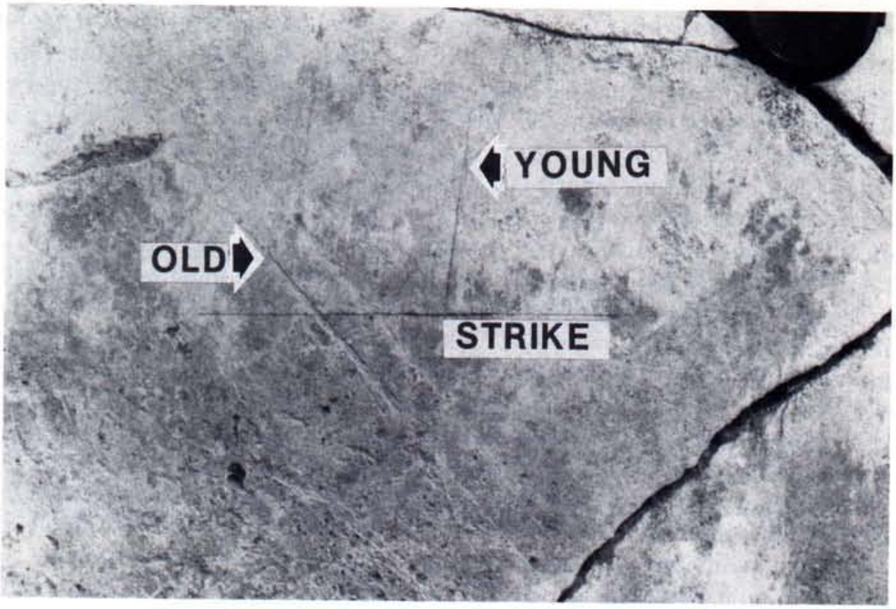


FIG. 15. Shale bedrock exposed in footwall of the 1983 scarp on northwest flank of Dickey Peak showing two sets of pre-1983 striae (see L on Plate 2). Fault plane strikes N12°W and dips 53°W. Older set of striae shows more sinistral slip than dip slip; younger set shows mostly dip slip with minor dextral slip. Camera lens cover in upper right is 5.5 cm in diameter. Photograph by A. J. Crone, 27 August 1984.

These striae probably record only late Quaternary movements on the fault that presumably occurred within the current stress regime. Small-scale features such as the striae probably would not be preserved very long (10,000 to 100,000? yr) in the easily weathered shale. This is especially true for the younger striae in the secondary calcium carbonate. Variations in net-slip directions recorded by striae at a point on a fault may not necessarily indicate major changes in the regional stress field. The changes in slip directions we measured at this site may simply reflect local stress variations on the fault near a segment boundary.

MORPHOLOGY OF THE 1983 SCARPS

Scarp-morphology data, which are usually plotted as scarp height versus maximum scarp-slope angle, have been applied frequently to estimate the relative age of scarps associated with normal faults (Bucknam and Anderson, 1979; Mayer, 1984; Nash, 1984; Machette, 1986). An important assumption in this technique is that the new scarp is formed by a simple rupture that produces an abrupt step in the topography. The new scarp has a morphology similar to that described by Wallace (1977). Morphologic dating of composite scarps (recurrent movement on a fault strand) and multiple scarps (composed of two or more smaller scarps) may yield erroneous age estimates (Machette and McGimsey, 1982; Mayer, 1984; Nash, 1984). Absolute age estimates have been made by modeling the time-dependent processes which dominate erosion of the scarp after it has reached the angle of repose (Nash, 1980; Colman and Watson, 1983; Hanks *et al.*, 1984; Pierce and Colman, 1986).

Local variations in the morphology of the 1983 scarps show how the distribution of fault strands, sorting of sediment, and content of interstitial water influence the morphology of the new scarps; these variations have important implications in using morphologic data as a means of estimating the age of prehistoric fault scarps. We found unusual morphologies along about 5 to 10 per cent of the 1983 scarps.

These variations emphasize the need to select sites that are typical of simple fault scarps; poor site selection can substantially reduce the quality of the data and influence the reliability of age estimates based on morphometric data.

About 300 m north of the Doublespring Pass road, a cumulative scarp height of more than 3 m is distributed on several synthetic faults to form a stair-step group of scarps (Figure 16). Directly to the north, the group merges into a single scarp having a similar height. When the stair-step scarps erode, they will coalesce into a broad, integrated scarp that has a more gentle slope than the adjacent simple scarp. Thus, in the future, the integrated scarp will seem anomalously degraded and, therefore, appear older than the adjacent simple scarp (Nash, 1984).

The grain-size distribution of the faulted materials can have a major effect on the morphology of the newly formed scarps (Wallace, 1980). This effect is apparent near Rock Creek where scarps on colluvium are nearly vertical to overhanging and have prominent free faces. A few tens of meters away, the same scarps on stream alluvium had lost most of their free faces and had degraded to slopes with angles of 30° to 37° within 1 day after the earthquake (Figures 17 and 18). Particle-size analyses of the colluvium and stream alluvium indicate that the silt and clay in the matrix of the colluvium may play a role in maintaining a vertical free face. Both the colluvium and alluvium are calcareous and have about 85 per cent gravel; the remaining matrix (15 per cent) consists of sand, silt, and clay. However, the matrix in the colluvium contains about 10 per cent more silt plus clay than the alluvium (47 versus 37 per cent) which apparently provides enough cohesion to retard degradation of the scarps (Figure 18). In addition, the angularity and subhorizontal fabric of clasts in the colluvium may contribute to its greater cohesion.

Another variation in scarp morphology is found near springs where the newly formed scarps cut fine-grained, wet sediments. At two locations, an unnamed spring 0.9 km northwest of West Spring (J, Plate 1), and the unnamed springs about 0.8 km southeast of Doublespring Pass road (Plate 1), the net throw across the fault zone exceeds 1 m. The scarps on poorly sorted, gravelly colluvium adjacent to the springs have sharp, well-defined free faces, but where the scarps extend into the fine-grained, wet sediment, the ground surface is warped into a broad monocline having a comparatively low slope angle and virtually no free face. Near the crest of

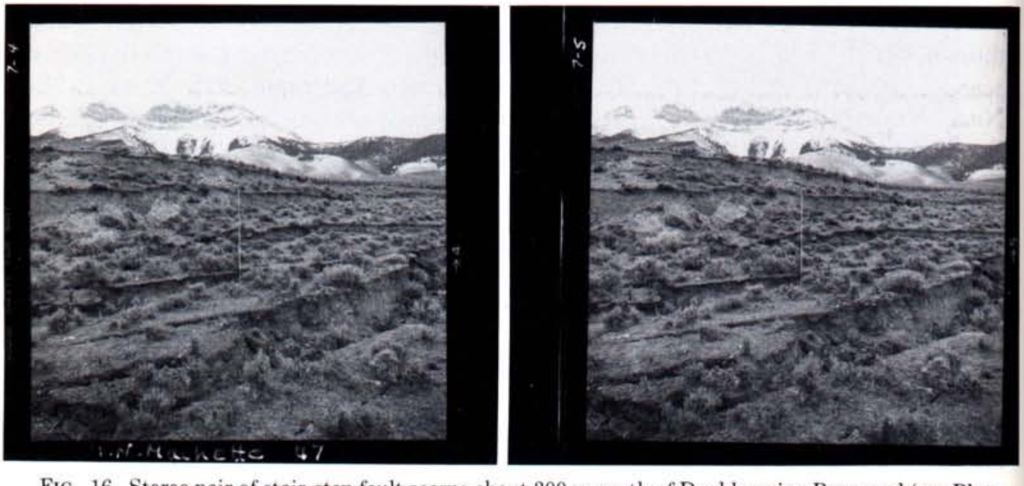
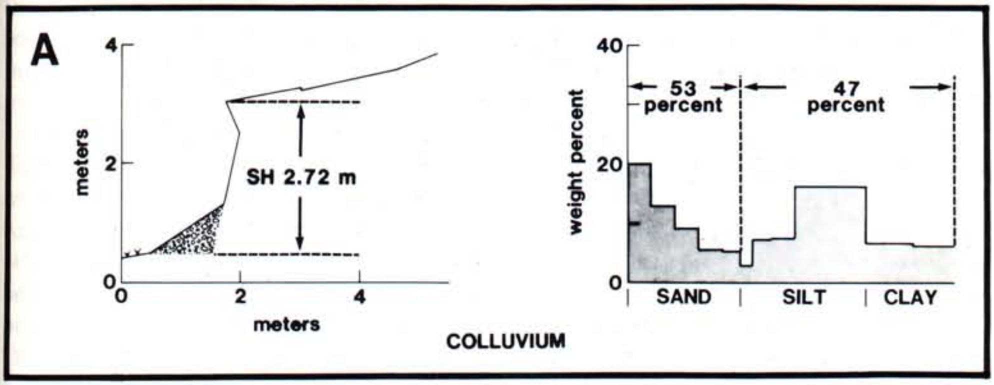


FIG. 16. Stereo pair of stair-step fault scarps about 300 m north of Doublespring Pass road (see Plate 1). View is to the southeast. Stadia rod is 2.5 m long. Photographs by M. N. Machette (U.S. Geological Survey Archive No. 47), 8 November 1983.

FIG. 17. Scarp formed on slope colluvium and stream alluvium near Rock Creek (see Plate 1). Scarp on colluvium in the background is near vertical, with large free face whereas scarp in stream alluvium in foreground had degraded to angle of repose within 1 day of the earthquake. Figure 18 shows profiles of these scarps. Photograph by A. J. Crone, 17 August 1984.



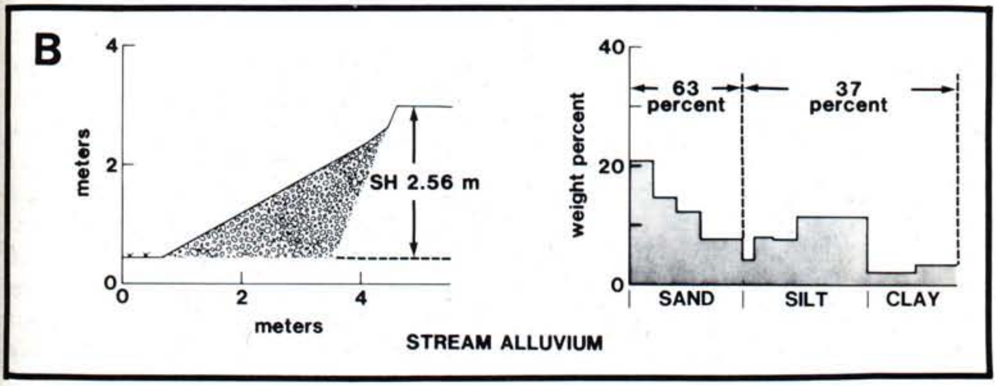


FIG. 18. Slope profile and particle-size distribution of matrix (<2 mm) from scarp formed on colluvium (A) and stream alluvium (B) near Rock Creek (Figure 17). Colluvium and stream alluvium both contain about 85 per cent gravel and 15 per cent matrix, but matrix of the colluvium contains 10 per cent more silt and clay. SH is scarp height in meters.

the monocline, large tension cracks with little or no net vertical displacement formed in the area of maximum flexure (Figure 19). The wet sediment deformed plastically, whereas adjacent unsaturated gravelly colluvium consistently ruptured in a more brittle fashion. In the future, the monocline will appear to be significantly more degraded and therefore apparently older than the scarps on colluvium.

Segmentation of the Lost River Fault

Geologic studies and historic surface faulting in the Basin and Range province show that only a fraction of major normal faults rupture during a single large earthquake. Geologic data have been used to subdivide some normal faults in the Western United States into segments (Swan *et al.*, 1980; Schwartz and Coppersmith, 1984; Scott *et al.*, 1985; Machette, 1986). Confirming the existence of segments, and understanding their coseismic behavior and interaction are important aspects of earthquake-hazard studies in the province. It is important to determine if segments tend to behave independently during large earthquakes and if coseismic ruptures are generally confined to individual segments. Equally as important is determining

FIG. 19. New scarp formed on wet sediment at unnamed spring 0.9 km northwest of West Spring (J, Plate 1). Wet sediment warped into a broad monocline with virtually no free face whereas nearby dry slope colluvium (in background) was more brittle and formed scarps with prominent free faces. View is to the north; stadia rod is 3.2 m long. Photograph by M. N. Machette, 1 November 1983.

what features create or control segment boundaries and if these features are geologically persistent and repeatedly inhibit or halt propagating ruptures.

The Borah Peak earthquake provides an excellent opportunity to evaluate the coseismic behavior of a segmented normal fault in the Basin and Range province and provides information on how a specific segment of a fault responded geologically and seismologically to a major earthquake. Studies of the earthquake suggest that the primary coseismic rupture in 1983 was mainly confined to the Thousand Springs segment (Scott *et al.*, 1985). Limited geologic data and structural relationships both suggest that, repeatedly, the fault at the segment boundaries has responded differently to earthquake ruptures than it has along the interior of the Thousand Springs segment.

Scott *et al.* (1985) divided the Lost River fault into six segments on the basis of geomorphic expression and structural relief of the range, and the age of last movement on the fault. From southeast to northwest, the segments are the Arco, Pass Creek, Mackay, Thousand Springs, Warm Spring, and northernmost segments. Unlike the other segments, the name "northernmost" does not provide an adequate geographic reference for this part of the fault; therefore, we propose that the "northernmost" of Scott *et al.* (1985) be renamed Challis (Figure 20).

During the Borah Peak earthquake, the main zone of surface faulting (the southern section) was on the Thousand Springs segment, although minor faulting (northern section) occurred on the Warm Spring segment and on the western section of scarps. No 1983 faulting occurred on the Mackay segment despite the high rate of late Quaternary slip inferred from prominent scarps and other geomorphic evidence (Crone *et al.*, 1985; Scott *et al.*, 1985).

The Thousand Springs-Mackay segment boundary. The boundary between the Thousand Springs and Mackay segments is marked by a gap in late Quaternary scarps between the southern end of the 1983 surface faults (near Elkhorn Creek) and the prominent scarps about 4 km to the southeast, near Lone Cedar Creek (Crone *et al.*, 1985). Southeast of Lone Cedar Creek, fault scarps show clear evidence of recurrent late Quaternary movement along the entire 24 km length of the Mackay segment. The gap in the scarps also coincides with an abrupt 55° change in the trend of the range front. This change probably reflects a similar change in the strike of the subjacent Lost River fault.

Bends in faults can be important in the initiation and termination of coseismic ruptures (Aki, 1979; King and Yielding, 1984; King and Nábělek, 1985). The epicenter of the main shock was on the down-dip projection of the Lost River fault near the southern end of the 1983 surface faults (Dewey, 1985). The epicentral location suggests that the rupture nucleated near this major bend in the fault and propagated upward and unilaterally to the northwest, a model that is consistent with the teleseismic body-wave data (Doser and Smith, 1985), the northwesterly elongation of the isoseismal contours (Stover, 1985), and the distribution of the surface faulting. Because the rupture extended only to the northwest, the part of the fault southeast of the hypocenter must have been more resistant to breaking, and displacements directed to the southeast were inhibited or arrested. The greater resistance of the fault to breaking in this area creates the segment boundary. If the fault were not more resistant to rupturing near this boundary, the 1983 rupture should have extended at least a short distance to the southeast, along the Mackay segment. Considering the length of time since the last surface faulting on the Mackay segment (about 4300 to 6800 yr ago) and the high late Quaternary slip rate inferred from geologic data (Scott *et al.*, 1985), some strain has probably accumulated

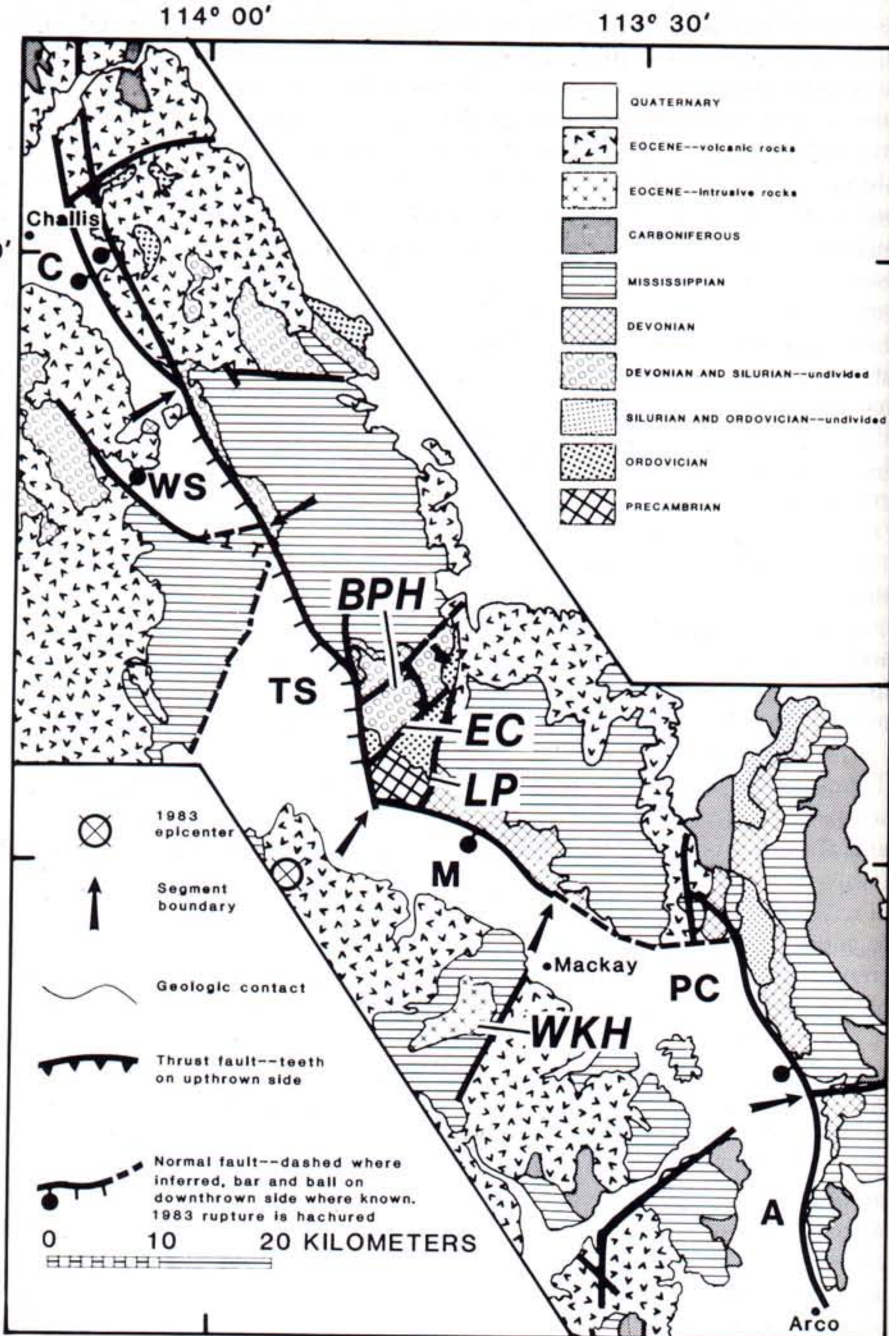


FIG. 20. Generalized geologic map of the Lost River Range, Idaho (modified from Bond, 1978). Arrows show segment boundaries of Lost River fault. Bold letters are segment names (modified from Scott *et al.*, 1985): A = Arco; PC = Pass Creek; M = Mackay; TS = Thousand Springs; WS = Warm Spring; C = Challis. Hatchured part of the fault ruptured in 1985. EC and LP are Elkhorn Creek and Leatherman Pass faults, respectively; BPH and WKH are Borah Peak and White Knob horsts, respectively.

on the Mackay segment but was not released in 1983. This implies that a barrier decoupled the Mackay and Thousand Springs segments and allowed them to respond independently to the 1983 rupture.

The coincidence of a major bend in the fault with this boundary suggests that the segment boundary may be controlled by a geometric barrier as defined by Aki (1979), although he applied the term to earthquakes on interplate strike-slip faults rather than intraplate normal faults. Aki (1979), and later King and Nábělek (1985), noted that small bends (5° in the 1966 Parkfield, California, earthquake) can play important roles in the earthquake rupture process. Because of different fault mechanics, small bends in normal faults may not affect propagating ruptures the same way they do in strike-slip faults. However, the response of the Thousand Springs-Mackay segment boundary during the Borah Peak earthquake suggests that prominent bends (55° in this case) in normal faults can affect propagating coseismic ruptures in extensional environments.

Three large northeasterly and north-northeasterly striking faults of Eocene age intersect the Lost River fault near the end of the Thousand Springs segment (Figure 20; Ross, 1947; Baldwin, 1951; Skipp and Harding, 1985). The Eocene faults bound the Borah Peak horst (Skipp and Harding, 1985) and have been cut by the Lost River fault. Two of these faults, the Elkhorn Creek and Leatherman Pass faults (labeled EC and LP, Figure 20) have as much as 3 km of throw and are near the Thousand Springs-Mackay segment boundary. From regional gravity data, Skipp and Harding (1985) infer that a northeast-trending ridge of shallow bedrock may lie beneath the valley fill and may be the southwesterly extension of the Borah Peak horst. They also speculate that the core of the horst may be an Eocene pluton, similar to the White Knob horst about 25 km to the south (Figure 20). If correct, this analogy implies that the boundary faults of the Borah Peak horst formed during emplacement of the pluton and, therefore, they penetrate the Mesozoic thrust sheets and may extend to hypocentral depths. Major, deep penetrating, transverse faults, such as the Leatherman Pass and perhaps the Elkhorn Creek faults, could create irregularities or obstructions on the Lost River fault plane that might inhibit propagating ruptures and form a boundary between the two segments.

The temporal pattern of aftershocks near the hypocenter supports the presence of a structural barrier at the boundary and suggests that a substantial difference in strain now may exist across the boundary. The few aftershocks near the hypocenter all occurred within several days of the main shock. Later, the aftershocks moved to the northwest as the Lost River fault adjacent to the epicenter became seismically quiet (Richins *et al.*, 1985; Zollweg and Richins, 1985). This pattern implies nearly complete strain release near the hypocenter with virtually no residual stress concentrations (Boatwright, 1985). Of the 421 aftershocks recorded in the 3 weeks following the main shock, only 23 events (mostly $M_L \leq 2$; Richins *et al.*, 1985). were located along the Mackay segment. As we have noted, an unknown but perhaps significant amount of strain may have accumulated on the Mackay segment before the earthquake. Thus, much of the strain of the Lost River fault northwest of the boundary was relaxed in 1983, while southeast of the boundary virtually no strain was released. If this strain contrast exists, it is further evidence of the resistance of the fault to rupturing at the boundary and demonstrates the independent behavior of adjacent segments.

The Thousand Springs-Warm Spring segment boundary. The boundary between the Thousand Springs and Warm Spring segments is at the junction of the Willow Creek hills and the Lost River fault (Figure 2). There are three significant changes in the 1983 ruptures at this boundary: (1) the main (southern) section of surface faults on the Lost River fault ends; (2) the western section diverges from the range front; and (3) a 4.7-km-long gap in surface faults on the Lost River fault separates the southern and the northern sections. The gap in the new surface faults and the formation of three separate sections of scarps that converge near the bedrock hills (Plate 2) both suggest a complex interaction of faults that ruptured in 1983.

The nature of the barrier associated with the Warm Spring-Thousand Springs segment boundary is uncertain. At this boundary, the Lost River fault changes strike by about 9°, but it is unknown if such a small change in strike is sufficient to create an effective geometric barrier on a normal fault. Alternatively, this segment boundary may be controlled by an inhomogeneous barrier (Aki, 1979) in which there is no obvious geometric obstruction to stop an earthquake rupture. Inhomogeneous barriers are related to variations in the strength of the fault, that is, parts of the fault are more resistant to breaking and therefore, can inhibit or arrest propagating ruptures.

The Willow Creek hills separate the Thousand Springs and Warm Spring Valleys. At the hills, normal faults that bound the western sides of the two valleys converge with, and may intersect, the Lost River fault (Figure 20; Baldwin, 1951; Skipp and Harding, 1985). These intersecting faults could create irregularities in the Lost River fault that would locally increase the Lost River fault's resistance to breaking; this locally strong part of the fault could create an inhomogeneous barrier that would inhibit propagating ruptures and thus form a segment boundary. It is worth noting that the intersecting faults that might be responsible for the barrier at the Thousand Springs-Warm Spring segment boundary are expressed only in the hanging wall of the Lost River fault; by contrast, the transverse faults at the Thousand Springs-Mackay segment boundary are exposed in the footwall and may extend into the hanging-wall block.

Focal mechanisms and the pattern of aftershocks near the Willow Creek hills show that the fault movements were more complex near this boundary than they were along the central part of the Thousand Springs segment. In cross-section, aftershocks on the Thousand Springs segment define a southwest-dipping, planar fault consistent with the location of the surface faults and the focal mechanism (Boatwright, 1985; Richins *et al.*, 1985). In contrast, the cross-sectional pattern of aftershocks near the Willow Creek hills and the area immediately to the north is diffuse but suggests a plane dipping steeply to the northeast. However, the nodal planes of a composite focal mechanism for these events are incompatible with the cross-sectional distribution of hypocenters (Boatwright, 1985). Focal mechanisms for individual aftershocks here indicate strike-slip and normal-slip movements on north-south to east-west-striking nodal planes (Richins *et al.*, 1985). Together, these data suggest that slip during the aftershocks occurred on a complex network of faults in the vicinity of the Willow Creek hills.

The temporal pattern of aftershocks suggests that the large strain release associated with the main shock on the Thousand Springs segment caused strain adjustments on the entire northern part of the Lost River fault. Intense aftershock activity near the Willow Creek hills started about 12 hr after the main shock and indicates that initially strain was concentrated at a barrier here. Most large-magnitude aftershocks were near the barrier, as were virtually all of those events with large stress drops (Boatwright, 1985). These large-stress-drop earthquakes suggest that strong barriers or asperities were present locally on the faults. Starting in late November 1983, aftershock activity increased dramatically on the Warm

Spring segment northwest of the Willow Creek hills (Zollweg and Richins, 1985), suggesting that the strain, which initially was concentrated at the Thousand Springs-Warm Spring segment boundary, shifted to the northwest. The strain adjustment affected the entire Warm Spring segment because, almost 10 months after the main shock, one of the largest aftershocks in the entire Borah Peak sequence, an M_L 5.8 event on 22 August 1984, occurred close to the proposed boundary between the Warm Spring and Challis segments (Figure 20; Plate 3). The temporal pattern of aftershocks and slip on a complex network of faults strongly support the presence of a fault-bounded salient near the Willow Creek hills that forms a structural barrier. At first, the barrier retarded strain adjustments on the northern part of the Lost River fault, but eventually the adjustments extended beyond the barrier and produced aftershocks along the Warm Spring and Challis segments.

The morphology of the pre-1983 fault scarps near the segment boundary shows how the boundary has influenced surface faulting during the late Pleistocene. On the Thousand Springs segment, the morphology of the pre-1983 fault scarps indicates a latest Pleistocene or early Holocene age for the surface faulting (Scott *et al.*, 1985). At two sites on this segment near Doublespring Pass road, the old scarp was 3.7 and 4.1 m high with maximum slope angles of 18.5° and 17.5°, respectively (Figure 21). On the Warm Spring segment and along the northern 1.9 km of the 1983 gap, scarps in alluvium have morphologies that suggest an age as young as or

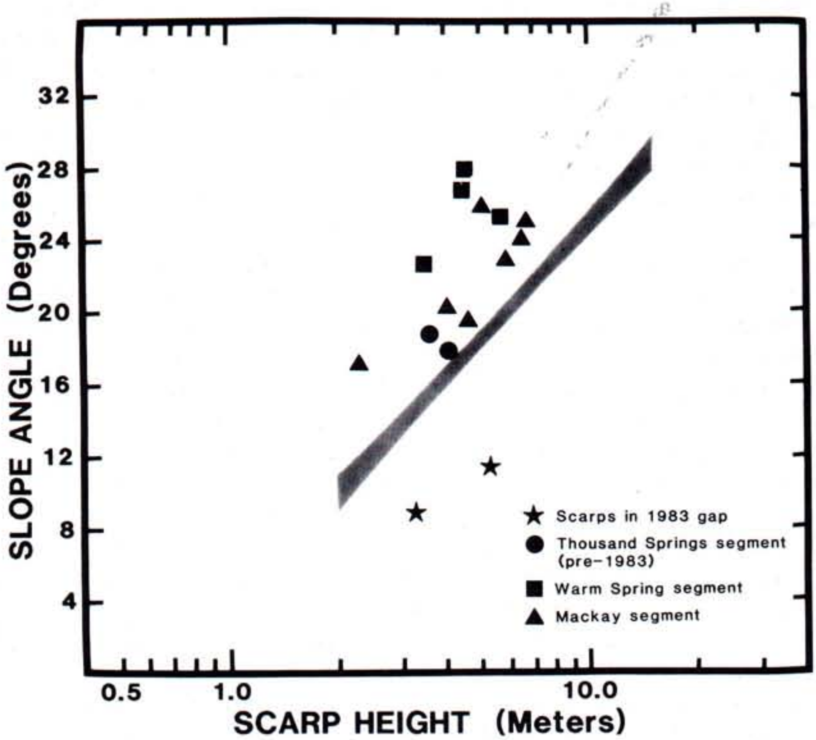


FIG. 21. Semi-log plot of scarp-height versus maximum slope-angle for selected fault scarps on the Mackay, Thousand Springs, and Warm Spring segments of the Lost River fault. Location of segments is in Figure 20. Some of the higher scarps may be composite scarps. Young scarps plot toward the upper left of the diagram and older scarps toward the lower right. Scarps in 1983 gap are those on the Lost River fault adjacent to the Willow Creek hills. Shaded area shows 95 per cent confidence interval around regression line of shoreline scarps of Lake Bonneville from Bucknam (1980). Shoreline scarps are about 15,000 yr old.

younger than the pre-1983 scarps on the Thousand Springs segment based on the relations described by Bucknam and Anderson (1979). At McGowan Creek and a nearby unnamed canyon to the southeast, scarps about 3.4- and 4.5-m-high have maximum slopes of 22.5° and 25° to 28°, respectively (near S, Plate 3; R, Plate 2). In the northern part of the 1983 gap near Sheep Creek, a 5.7-m-high multiple-event scarp also has a maximum slope of 25° (near P, Plate 2). In contrast, at the southern end of the 1983 gap, a subdued scarp on the alluvium of Arentson Gulch is 3 to 5 m high with a maximum slope of only 9° to 11° (0, Plate 2). The comparatively low slope angles indicate that this scarp is significantly older than the nearby scarps on the Warm Spring and Thousand Springs segments. Assuming that it is a degraded simple scarp, the old scarp is probably greater than 15 ka because it has lower slope angles and thus is apparently older than equally high, wave-cut scarps of Lake Bonneville (Bucknam and Anderson, 1979). Thus, the fault-scarp morphology suggests that the gap in surface faulting on the Lost River fault in 1983 may be typical of the behavior of this segment boundary during the latest Quaternary.

Although poorly constrained, the available geologic data indicates that the cumulative Cenozoic throw on the Lost River fault is probably less at this segment boundary than it is on the adjacent segments. The bedrock geology shows at least 1.7 km of cumulative Cenozoic throw on the Warm Spring segment near Gooseberry Creek (Mapel *et al.*, 1965). As noted, the structural relief on the Thousand Springs segment near Borah Peak may be more than 2.5 km, although this cannot be confirmed by the available mapping and subsurface data. By comparison, the throw on the fault is only about 1.2 km at the Willow Creek hills (Skipp and Harding, 1985). Thus, the Lost River fault at this boundary either has ruptured less frequently (as suggested by the 1983 faulting) or as often but with less throw each time compared to the fault along the adjacent segments. These relations suggest that the boundary is a long-lived geologic feature that probably has affected displacements on the Lost River fault throughout much of its history.

The Pleistocene scarp in the 1983 gap raises some intriguing questions about how this boundary in particular and perhaps segment boundaries in general respond to propagating ruptures. If the gap marks a locally strong part of the fault, do earthquakes larger than the 1983 event sometimes break through the barrier, possibly releasing the strain that has accumulated on two adjacent segments? This scenario implies that earthquakes larger than a characteristic earthquake of Schwartz and Coppersmith (1984) might occur infrequently on Basin and Range normal faults. Alternatively, does surface faulting occur in the gap only when a rupture propagates southeastward from a hypocenter on the Warm Spring segment? If so, this implies that barriers might have different apparent strengths, depending on the direction a rupture enters the barrier. Fault-scarp morphology data indicate that the last surface faulting on the Warm Spring segment postdates the last surface faulting in the gap. Finally, does surface faulting at this segment boundary occur only after several events have ruptured the fault on both of the adjacent segments? Perhaps several cycles of surface faulting are needed on the Warm Spring and Thousand Springs segments before the boundary is sufficiently strained to fail during an earthquake. The relationship of the Pleistocene scarp in the gap to surface faulting on the adjacent segments is still unclear, but the scarp does identify a part of the fault that behaves unusually.

Discussion

Segmentation and the 1983 earthquake. We have combined geologic evidence with seismologic and geodetic data to show that the past and present behavior of the

Lost River fault is different at boundaries of the Thousand Springs segment than it is along the interior of the segment. The asymmetry of the 1983 scarps with respect to the epicenter and aftershock data suggest that a major barrier at the Thousand Springs-Mackay segment boundary inhibited or arrested the southeastward propagation of the rupture. Similarly, the morphology of fault scarps, the bedrock geology, and the pattern of aftershocks near the Willow Creek hills all indicate an important change in the behavior of the fault at the Thousand Springs-Warm Spring segment boundary. We speculate that transverse faults in both the hanging wall and footwall at these boundaries may increase the resistance of the Lost River fault to rupturing. Many details of how geologic structures affect the rupture process at segment boundaries remain unclear, but the Borah Peak earthquake has provided us with the opportunity to better characterize the coseismic behavior of segments of a major range-front fault.

Even without the detailed seismologic and geologic data from the 1983 earthquake, the boundaries of the Thousand Springs segment are marked by distinctive geologic characteristics. At the southern end of the segment, the abrupt change in the strike of the Lost River fault and the absence of prominent scarps southeast of Elkhorn Creek are conspicuous signs of anomalous behavior. At the northern end of the segment, the reduced throw as suggested by the bedrock geology near the Willow Creek hills requires that this part of the fault has had a lower late Cenozoic slip rate. The late Pleistocene scarp in the gap in 1983 surface faulting compared to the more youthful scarps on the nearby segments is another important sign of unusual late Quaternary behavior at the northern segment boundary. Our experience at Borah Peak supports the use of late Quaternary fault scarps as guides to identifying segments of normal faults in the Basin and Range and thus delimiting the probable extent of surface faulting during future earthquakes on those faults.

The range-front geometry, bedrock geology, and fault-scarp morphology are useful criteria that help identify the segments of the Lost River fault. These criteria may be equally as useful in identifying segments of other major range-front faults in the Basin and Range, thereby contributing to a better understanding of possible fault behavior during future earthquakes.

Interpretation of events in 1983. The following scenario summarizes our interpretation of the sequence of events associated with the Borah Peak earthquake. On the morning of 28 October 1983, the main shock nucleated on the Lost River fault near the Thousand Springs-Mackay segment boundary. At the major bend in the fault southeast of Elkhorn Creek, a geometric barrier detached the Thousand Springs and Mackay segments, and confined the rupture in this area to the Thousand Springs segment. On the Thousand Springs side of the barrier, virtually all of the strain on the fault was released, whereas on the Mackay side, no strain was released. From the hypocenter, the rupture propagated upward and northwestward along the fault (Smith *et al.*, 1984; Boatwright, 1985), reached the surface near Elkhorn Creek, and continued unimpeded on the Thousand Springs segment to the junction with the Willow Creek hills. At the junction, a barrier on the fault plane either halted the primary rupture and immediately triggered a secondary rupture on the western section, or diverted the primary rupture away from the Lost River fault and onto other faults along the western section (Arentson Gulch fault of Wallace, 1984b). The height of the scarps and the continuity of the surface ruptures on the western section diminish rapidly to the west, suggesting that the displacement at depth also decreases. It is uncertain if the surface ruptures along the entire western section truly represent a continuous, through-going subsurface structure. The termination of the rupture in the Willow Creek hills may have resulted from motion on the Lost River fault being distributed among an array of subsidiary faults on the western section in a manner similar to that described by King and Yielding (1984) and King and Nábělek (1985).The scarps along the Lost River fault north of the gap (northern section) are thought to be only indirectly related to the primary rupture that originated at the hypocenter. As noted by Boatwright (1985, p. 1271), the mechanics of aftershocks near and north of the Willow Creek hills are inconsistent with the surface faulting on the northern section. This suggests that the scarps may be surficial features, not directly related to significant movement on deep faults. If our proposed scenario is correct, then it appears that segment boundaries can stop or deflect propagating ruptures during large earthquakes in the Intermountain Seismic Belt. However, as demonstrated by the scarps along the northern section, coseismic energy release and shaking can trigger secondary surface faulting on segments adjacent to the main rupture.

Following the main shock, strain adjustments occurred on most of the northern part of the Lost River fault. Within a few hours of the main shock, numerous aftershocks at the Thousand Springs-Warm Spring segment boundary show that this area was the locus of most of the strain adjustments. During the following months, further strain adjustments affected the Warm Spring and Challis segments. An M_L 5.8 shock on 22 August 1984 associated with these adjustments shows that potentially damaging aftershocks can occur tens of kilometers away from and many months after the main shock.

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References

Aki, K. (1979). Characterization of barriers on an earthquake fault, *J. Geophys. Res.* **84,** 6140–6148.

Baldwin, E. W. (1951). Faulting in the Lost River Range area of Idaho, *Am. J. Sci.* **249,** 884–902.

Barrientos, S. E., S. N. Ward, J. R. Gonzalez-Ruiz, and R. S. Stein (1985). Inversion for moment as a function of depth from geodetic observations and long period body waves of the 1983 Borah Peak, Idaho earthquake, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 485–518.

Beck, A. C. (1968). Gravity faulting as a mechanism of topographic adjustment, *N. Z. J. Geol. Geophys.* **11,** 191–199.

Boatwright, J. (1985). Characteristics of the aftershock sequence of the Borah Peak, Idaho, earthquake determined from digital recordings of the events, *Bull. Seism. Soc. Am.* **75,** 1265–1284.

Bond, J. G. (1978). Geologic map of Idaho, *Idaho Geol. Surv.*, 1:500,000.

Bonilla, M. G., R. K. Mark, and J. J. Lienkaemper (1984). Statistical relations among earthquake magnitude, surface rupture length, and surface fault displacement, *Bull. Seism. Soc. Am.* **74,** 2379–2411.

Bovis, M. J. (1982). Uphill-facing (antislope) scarps in the Coast Mountains, southwest British Columbia, *Geol. Soc. Am. Bull.* **93,** 804–812.

Bucknam, R. C. (1980). Characteristics of active faults in the Great Basin, in *Summaries of Technical Reports*, Volume IX, M. L. Turner, Compiler, *U.S. Geol. Surv., Open-File Rept. 80-6*, 94–95.

Bucknam, R. C. and R. E. Anderson (1979). Estimation of fault-scarp ages from a scarp-height-slope-angle relationship, *Geology* **7,** 11–14.

Colman, S. M. and K. Watson (1983). Ages estimated from a diffusion equation model for scarp degradation, *Science* **221,** 263–265.

Crone, A. J., K. M. Haller, R. C. Bucknam, S. Gou, Y. Liao, Y. Li, and K. R. Vincent (1985). Late Quaternary faulting along the Mackay segment of the Lost River fault, central Idaho (abstract), *EOS, Trans. Am. Geophys. Union* **66,** 1067.

Crosthwaite, E. G., C. A. Thomas, and K. L. Dyer (1970). Water resources in the Big Lost River Basin, south-central Idaho, *U.S. Geol. Surv., Open-File Rept.*, 109 pp.

Dewey, J. W. (1985). Instrumental seismicity of central Idaho, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 264–284.

Doser, D. I. (1985). Source parameters and faulting processes of the 1959 Hebgen Lake, Montana, earthquake sequence, *J. Geophys. Res.* **90,** 4537–4555.

Doser, D. I. and R. B. Smith (1985). Source parameters of the 28 October 1983 Borah Peak, Idaho, earthquake from body wave analysis, *Bull. Seism. Soc. Am.* **75,** 1041–1051.

Hait, M. H., Jr. and W. E. Scott (1978). Holocene faulting, Lost River Range, Idaho, *Geol. Soc. Am Abstracts with Prog.* **10,** 217.

Hanks, T. C., R. C. Bucknam, K. R. Lajoie, and R. E. Wallace (1984). Modification of wave-cut and fault-controlled landforms, *J. Geophys. Res.* **89,** 5771–5790.

Hays, W. H., D. H. McIntyre, and S. W. Hobbs (1978). Geologic map of the Lone Pine Peak quadrangle, Custer County, Idaho, *U.S. Geol. Surv., Open-File Rept. 78-1060*, 1:62,500.

Jahn, A. (1964). Slopes morphological features resulting from gravitation, *Zeitschrift für Geomorph.* (Supplement Band 5), 59–72.

King, G. and G. Yielding (1984). The evolution of a thrust system—Processes of rupture initiation, propagation and termination in the 1980 El Asnam (Algeria) earthquake, *Geophys. J. R. Astr. Soc.* **77,** 915–933.

King, G. and J. Nábělek (1985). Role of fault bends in the initiation and termination of earthquake rupture, *Science* **228,** 984–987.

Machette, M. N. (1986). History of Quaternary offset and paleoseismicity along the La Jencia fault, central Rio Grande rift, New Mexico, *Bull Seism. Soc. Am.* **76,** 259–272.

Machette, M. N. and R. G. McGimsey (1982). Quaternary and Pliocene faults in the Socorro and western part of the Fort Sumner 1° × 2° quadrangles, New Mexico, *U.S. Geol. Surv. Misc. Field Invest. Map, MF-1465-A*, 1:250,000.

Mapel, W. J., W. H. Read, and R. K. Smith (1965). Geologic map and sections of the Doublespring quadrangle, Custer and Lemhi Counties, Idaho, *U.S. Geol. Surv. Geol. Quad. Map, GQ-464*, 1:62,500.

Mayer, L. (1984). Dating Quaternary fault scarps formed in alluvium using morphologic parameters, *Quat. Res.* **22,** 300–313.

McIntyre, D. H., E. B. Ekren, and R. F. Hardyman (1982). Stratigraphic and structural framework of the Challis Volcanics in the eastern half of the Challis 1° × 2° quadrangle, Idaho, in *Cenozoic Geology of Idaho*, B. Bonnichsen and R. M. Breckenridge, Editors, Idaho Bureau Mines and Geology, Bulletin **26,** 3–22.

Myers, W. B. and W. Hamilton (1964). Deformation accompanying the Hebgen Lake earthquake of August 17, 1959, *U.S. Geol. Surv. Profess. Paper 435-I*, 55–98.

Nakata, J. K., C. M. Wentworth, and M. N. Machette (1982). Quaternary fault map of the Basin and Range and Rio Grande rift provinces, Western United States, *U.S. Geol. Surv., Open-File Rept. 82-579*, 1:2,500,000.

Nash, D. B. (1980). Morphologic dating of degraded normal fault scarps, *J. Geol.* **88,** 353–360.

Nash, D. B. (1984). Morphologic dating of fluvial terrace scarps and fault scarps near West Yellowstone, Montana, *Geol. Soc. Am. Bull.* **95,** 1413–1424.

National Earthquake Information Service (1983). Preliminary determination of epicenters, monthly listing, *U.S. Geol. Surv.* (October 1985), 20 pp.

Pelton, J. R., C. W. Meissner, and K. D. Smith (1984). Eyewitness account of normal surface faulting, *Bull. Seism. Soc. Am.* **74,** 1083–1089.

Pelton, J. R., C. W. Meissner, C. J. Waag, and S. H. Wood (1985). Shallow seismic refraction studies across the Willow Creek fault rupture zone and the Chilly Buttes sand boils, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 161–181.

Pierce, K. L. and S. M. Colman (1986). Effect of height and orientation (microclimate) on geomorphic degradation rates and processes, late-glacial terrace scraps in central Idaho, *Geol. Soc. Am. Bull.* **97,** 869–885.

Radbruch-Hall, D. H., D. J. Varnes, and W. Z. Savage (1976). Gravitational spreading of steep-sided ridges ("Sackung") in western United States, *Internatl. Assoc. Eng. Geol. Bull.* **14,** 23–35.

Reynolds, M. W. (1979). Character and extent of Basin-Range faulting, western Montana and east-central Idaho, Rocky Mountain Association of Geologists and Utah Geological Association, 1979 Basin and Range Symposium, 185–193.

Richins, W. D., R. B. Smith, C. J. Langer, J. E. Zollweg, J. J. King, and J. C. Pechmann (1985). The

1983 Borah Peak, Idaho earthquake—Relationship of aftershocks to the main shock, surface faulting, and regional tectonics, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 285–310.

Ross, C. P. (1947). Geology of the Borah Peak quadrangle, Idaho, *Geol. Soc. Am. Bull.* **58,** 1085–1160.

Ruppel, E. T. (1978). Medicine Lodge thrust system, east-central Idaho and southwest Montana, *U.S. Geol. Surv. Profess. Paper 1031*, 23 pp.

Ruppel, E. T. (1982). Cenozoic block uplifts in east-central Idaho and southwest Montana, *U.S. Geol. Surv. Profess. Paper 1224*, 24 pp.

Schwartz, D. P. and K. J. Coppersmith (1984). Fault behavior and characteristic earthquakes—Examples from the Wasatch and San Andreas fault zones, *J. Geophys. Res.* **89,** 5681–5698.

Schwartz, D. P. and A. J. Crone (1985). The 1983 Borah Peak earthquake—A calibration event for quantifying earthquake recurrence and fault behavior on Great Basin normal faults, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 153–160.

Scott, W. E., K. L. Pierce, and M. H. Hait, Jr. (1985). Quaternary tectonic setting of the 1983 Borah Peak earthquake, central Idaho, *Bull Seism. Soc. Am.* **75,** 1053–1066.

Skipp, B. and M. H. Hait, Jr. (1977). Allochthons along the northeast margin of the Snake River Plain, Idaho, *Wyoming Geological Association 29th Annual Field Conference Guidebook*, 499–515.

Skipp, B. A. and S. T. Harding (1985). Preliminary report of geology of the Borah Peak area, Idaho, including interpretation of seismic and gravity data, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 657–671.

Slemmons, D. B. (1957). Geological effects of the Dixie Valley-Fairview Peak, Nevada, earthquakes of December 16, 1954, *Bull. Seism. Soc. Am.* **47,** 353–375.

Smith, R. B. and M. L. Sbar (1974). Contemporary tectonics and seismicity of the western United States with emphasis on the Intermountain Seismic Belt, *Geol. Soc. Am. Bull.* **85,** 1205–1218.

Smith, R. B., W. D. Richins, D. I. Doser, J. C. Pechmann, and C. Langer (1984). The 1983 M_s 7.3 Borah Peak, Idaho, earthquake—A model for active crustal extension, *Geol. Soc. Am. Abstracts with Prog.* **16,** 661.

Stein, R. S. and S. E. Barrientos (1985). The 1983 Borah Peak, Idaho, earthquake—Geodetic evidence for deep rupture on a planar fault, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 459–484.

Stover, C. W. (1985). Isoseismal map and intensity distribution for the Borah Peak, Idaho, earthquake of October 28, 1983, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 401–408.

Swan, F. H., D. P. Schwartz, and L. S. Cluff (1980). Recurrence of moderate to large magnitude earthquakes produced by surface faulting on the Wasatch fault zone, Utah, *Bull. Seism. Soc. Am.* **70,** 1431–1462.

Wallace, R. E. (1977). Profiles and ages of young fault scarps, north-central Nevada, *Geol. Soc. Am. Bull.* **88,** 1267–1281.

Wallace, R. E. (1980). Degradation of the Hebgen Lake fault scarps of 1959, *Geology* **8,** 225–229.

Wallace, R. E. (1984a). Faulting related to the 1915 earthquakes in Pleasant Valley, Nevada, *U.S. Geol. Surv. Profess. Paper 1274-A*, 33 pp.

Wallace, R. E. (1984b). Eyewitness account of surface faulting during the earthquake of 28 October 1983, Borah Peak, Idaho, *Bull. Seism. Soc. Am.* **74,** 1091–1094.

Witkind, I. J. (1964). Reactivated faults north of Hebgen Lake, *U.S. Geol. Surv. Profess. Paper 435-G*, 37–50.

Zollweg, J. E. and W. D. Richins (1985). Later aftershocks of the 1983 Borah Peak, Idaho, earthquake and related activity in central Idaho, in *Proceedings of Workshop XXVIII on the Borah Peak, Idaho, Earthquake*, R. S. Stein and R. C. Bucknam, Editors, *U.S. Geol. Surv., Open-File Rept. 85-290*, 345–367.

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