Planar High-Angle Faulting in the Basin and Range:
Geodetic Analysis of the 1983 Borah Peak, Idaho, Earthquake

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Geodetic elevation changes record the broad-scale deformation associated with the M = 7.0 October 28, 1983, Borah Peak, Idaho, earthquake on the Lost River fault. The crest of the Lost River Range rose 0.2 m, and adjacent Thousand Springs Valley subsided 1.0 m, in relation to reference points 45 km from the main shock epicenter. The deformation was modeled by dislocations in an elastic half-space. A planar fault with uniform dip slip of 2.05 ± 0.10 m, dipping 47° ± 2°SW and extending to a vertical depth of 13.3 ± 1.2 km, fits the geodetic data best and is also consistent with the main shock hypocenter and fault plane solution. The geodetic moment is 2.6 ± 0.5 × 10¹⁹ N m (2.6 ± 0.5 × 10²⁶ dyn cm), and the estimated static stress drop is 2.9 ± 0.4 MPa (29 ± 4 bars). Tests for coseismic slip on listric faults (which flatten with depth and on detachments (horizontal faults or shear zones) showed fits to the geodetic data that are inferior to those for planar high-angle faults. No detectable coseismic slip occurred on the Mesozoic White Knob thrust fault, although the low-angle thrust sheet intersects the south end of Lost River fault near the 1983 mainshock epicenter. If the high-angle Lost River fault abuts a flat-lying detachment fault or shear zone, such a structure must lie at depths of > 12 km, near the brittle-ductile transition, where stick-slip behavior gives way to creep. The depth and geometry of faulting at Borah Peak is similar to that inferred from seismic and geodetic evidence for the 1954 M = 7.2 Fairview Peak, Nevada, and the 1959 M = 7.3 Hebgen Lake, Montana, events, suggesting that if detachments are active at these localities, they are deep and most likely slip by creep.

INTRODUCTION

Today, more than a century after LeConte [1878] and Gilbert [1884] first studied the earthquake potential of the Basin and Range province following the M = 7.8 1872 Owens Valley, California, earthquake, we still depend on large earthquakes and the permanent deformation they produce at the surface to investigate the mechanics of faulting. Six large historical earthquakes (M > 7) have struck the Great Basin (Table 1); of these, only the 1954 M = 7.2 Fairview Peak, Nevada, the 1959 M = 7.3 Hebgen Lake, Montana, and the 1983 M = 7.0 Borah Peak, Idaho, shocks have left a seismic and geodetic record that enables study of the subsurface fault slip (moment magnitudes cited; see Hanks and Kanamori [1979]. From this limited sample and from geologic observations, we must assess seismic hazards in the seven western states shown in Figure 1. The damage that can be inflicted by large earthquakes in this region depends, first, on the proximity of the active faults to population centers and, second, on the subsurface geometry of these faults.

Whether active faults in the Basin and Range province commonly become listric (decrease in dip to become flat lying) or slip steeply at depth is not known. Resolution of this question is crucial not only to evaluate earthquake hazards but also to understand the mechanism and amount of extension across the Great Basin. Rupture of listric and detachment (flat-lying) faults suggests a shallower seismic source (3–10 km) that can locate farther from the fault trace than does rupture at the base of the seismogenic crust (15 km deep) on planar high-angle faults.

Estimates of total extension across the 750-km-wide Great Basin range from 10% [Stewart, 1971] to 100% [Hamilton, 1985]. The range-bounding normal faults within the Great Basin dip 45°–70° at the surface, but planar faults with these dips cannot accommodate more than about 25% extension unless the faults are rooted to horizontal slip surfaces or ductile shear zones [Davis, 1980]. Mesozoic thrust sheets may be underly many of the high-angle faults, however, supplying the requisite horizontal slip surfaces. Reactivation of the midcrustal thrust sheets as sites of normal slippage requires greatly elevated pore pressures, which have yet to be measured in situ. Great extension may also be accommodated by ductile shear. Ductile shear zones should lie below the depth of peak crustal strength, which occurs at approximately 10–15 km [Sibson, 1982]. Recently, geological mapping and deep seismic reflection profiling in the Great Basin [e.g., Wernicke, 1981; Allmendinger et al., 1983; Crone and Harding, 1984; Smith and Bruhn, 1984] and teleseismic waveform modeling in other regions of continental extension [Eyidoğan and Jackson, 1985] have been offered as new support for listric faults that merge into detachments and shear zones in the upper crust (3–10 km), a geometry first proposed by Longwell [1945].

The 1983 Borah Peak earthquake affords a rare opportunity to delineate the subsurface fault geometry of an active normal fault, particularly since none of the listric or horizontal reflectors identified as faults have been observed to slip seismically. We use the geodetic record of the Borah Peak earthquake to distinguish between planar and listric fault slip and to identify the maximum depth of seismic slip. We find that high-angle faulting extended to the base of the seismogenic crust and was accompanied by no detectable detachment slip. We also compare the coseismic model with the earthquake source parameters measured from seismic radiation at much shorter periods than the geodetic observations and find these measurements to be compatible. Finally, we compare the
coseismic deformation with the cumulative deformation preserved in the geologic structure.

**GEODETIC DATA**

The precision and spatial coverage of the geodetic data equal or surpass all records of vertical elevation changes associated with large normal faulting earthquakes. The 70-km-long first- and second-order leveling route that crosses the Lost River fault permits estimation of the fault geometry and slip independent of the seismic and geological data.

The coseismic elevation changes were obtained by subtracting the relative heights of bench marks (BM's) surveyed after the earthquake in July 1984 from the heights of the same BM's surveyed before the earthquake in 1933 and 1948 (Figure 2a). The top panel of Figure 3 shows the profile of coseismic elevation change, projected onto an azimuth normal to the Lost River fault. Note that what we refer to as the "coseismic period" includes the 50 years preceding the earthquake and the 8 months after it, under the assumption that no artificial subsidence and no preearthquake deformation took place during this period. Tests of this assumption are discussed in the next section. All leveling was conducted by the National Geodetic Survey (NGS); the 1984 survey was carried out at the request of the U.S. Geological Survey (USGS).

In spirit leveling, height differences are measured between adjacent BM's by sighting a horizontal telescope on graduated

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**TABLE 1. Large Historical Basin and Range Earthquakes**

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>Moment Magnitude M</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan. 9, 1872</td>
<td>Owens Valley, California</td>
<td>7.8</td>
<td>Hanks and Kanamori [1979]</td>
</tr>
<tr>
<td>Oct. 2, 1915</td>
<td>Pleasant Valley, Nevada</td>
<td>7.2</td>
<td>Wallace [1984a]</td>
</tr>
<tr>
<td>Dec. 10, 1932</td>
<td>Cedar Mountain, Nevada</td>
<td>7.0</td>
<td>Doser [1985a]</td>
</tr>
<tr>
<td>Dec. 16, 1954</td>
<td>Fairview Peak, Nevada</td>
<td>7.2</td>
<td>Doser [1985a]</td>
</tr>
<tr>
<td>Aug. 19, 1959</td>
<td>Hebgen Lake, Montana</td>
<td>7.3</td>
<td>Doser [1985a]</td>
</tr>
<tr>
<td>Oct. 28, 1983</td>
<td>Borah Peak, Idaho</td>
<td>7.0</td>
<td>Doser [1985a]; this paper</td>
</tr>
</tbody>
</table>

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Fig. 1. Sketch of Basin and Range province of the western United States, after Nakata et al. [1982], Wallace [1984a, b], and Scott et al. [1985], showing locations of earthquakes discussed in text.
Fig. 2. (a) Schematic map of Borah Peak earthquake site, showing locations of epicenter from Richins et al. [1985], surface rupture from Crone and Machette [1984], and BM's (crosses); contours of elevation change predicted by the coseismic model (dotted rectangle with rounded corners representing the base of the fault); and the line of projection for the profiles in Figures 3, 4, and 6. (b) Map showing surface rupture and locations of the October 28, 1983, main shock and aftershocks from October 29 to November 19, 1983 [Richins et al., 1985], coseismic fault model. The solid rectangle is the surface projection of the fault plane; the dashed rectangle is the largest fault plane compatible with the observations, representing a 10% increase in model residuals over the best fit.
Fig. 3. (Top) Profile of coseismic elevation change projected normal to the fault. (Bottom) Leveling route topography.

Survey Errors

Potential sources of noise in the data include random and systematic surveying errors and nontectonic subsidence due to groundwater withdrawal. Estimated cumulative random errors for the observed elevation changes nowhere exceeds 20 mm. Briefly, BM's are spaced 1–2 km apart (a pair of adjacent BM's is called a section). The elevation difference over each section can be obtained once or twice during a survey, by surveying forward and backward. If double run, the section is rejected and rerun if the two leveling do not agree within the prescribed section tolerance. The mean differences for all (forward-backward) runnings must not differ significantly from zero, and the standard deviation of the difference, \( \beta \), gives a good indication of the random error per section. The random error \( \sigma \) should propagate with the square root of the distance leveled: \( \sigma = \alpha L^{1/2} \) where \( \sigma \) is in millimeters, \( \alpha \) is a function of \( \beta \), and \( L \) is the distance in kilometers [Bomford, 1971]. In principle, if \( \beta \) is Gaussian distributed and the survey is double run with BM's spaced 1 km apart, \( \alpha \) should be about one third of \( \beta \). Because some of the 1984 survey was single run using the procedure of Whalen and Balazs [1977], we set \( \alpha = \beta \) (Table 2 lists the values for \( \alpha \) and \( \beta \)). Thus, the cumulative random error for the 70-km-long postearthquake survey is 1.38 mm \( (L^{1/2}) = 12 \text{ mm. For the 1933 survey, } \beta = 1.83 \text{ mm, slightly larger than } \beta \text{ for 1983 or 1984. Because the 1948 leveling segment was single run, } \beta \text{ cannot be accurately assessed. We assume here that a value of } \alpha = 2.0 \text{ mm provides a fair estimate for the combined 1933 and 1948 surveys; the cumulative error for the preearthquake leveling is then 17 mm. The random error for elevation change differenced from the preearthquake and postearthquake surveys becomes } (\alpha_1^2 + \alpha_2^2)^{1/2}L^{1/2}, \text{ where the subscript 1 denotes 1984 and the subscript 2 denotes}

### TABLE 2. Leveling Specifications

<table>
<thead>
<tr>
<th>NGS Line</th>
<th>Survey Period</th>
<th>Segment, Start BM–End BM</th>
<th>Double or Single Run</th>
<th>Order of Leveling</th>
<th>Reurns, %</th>
<th>Rejection Tolerance, ( \beta ), mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>L24832</td>
<td>June 19–July 12, 1984</td>
<td>L51-6345; USGS-E15–E15</td>
<td>single</td>
<td>first</td>
<td>4.7</td>
<td>0.4, 1.38</td>
</tr>
<tr>
<td></td>
<td></td>
<td>L51–B15</td>
<td>double</td>
<td>first</td>
<td>6.8</td>
<td>0.4</td>
</tr>
<tr>
<td>L24812</td>
<td>Nov. 14–23, 1983</td>
<td>L1–E15</td>
<td>double</td>
<td>first</td>
<td>1.2</td>
<td>0.4, 1.09</td>
</tr>
<tr>
<td>L2785</td>
<td>Aug. 18–23, 1948</td>
<td>L51–B15</td>
<td>single</td>
<td>second</td>
<td>4.0</td>
<td>0.8</td>
</tr>
<tr>
<td>L951</td>
<td>Oct. 4–Nov. 24, 1933</td>
<td>L51–USGS-B15</td>
<td>double</td>
<td>first</td>
<td>4.8</td>
<td>0.4, 1.83</td>
</tr>
</tbody>
</table>

Indices are explained in the text and by Federal Geodetic Control Committee [1984].

All data were corrected for level collimation (horizontal plumb of the level telescope), rod excess (calibration of rod graduations), rod thermal expansion (rod length change as a function of temperature), and atmospheric refraction (bending of the line of sight through vertically stratified air). The observed rod temperatures were used for all surveys to compute rod expansion. For the 1933 and 1948 surveys the thermal coefficient of expansion was assumed; for the 1984 survey it was measured at the National Bureau of Standards. The 1933 and 1948 surveys were corrected for refraction by using the REDUC4 program of Holdahl [1981], in which temperature gradients are modeled as a function of time and location, whereas the 1984 observations were corrected following Kukkamaki [1988], using observed temperature differences made at every instrument setup.
We examine all surveys for elevation-dependent error, using the method of Stein [1981], and found none. Inaccurate rod calibration corrections and, under some circumstances, unmodeled refraction error cause a dependence of elevation change on the elevation. In these cases, elevation change and elevation profiles tend to mimic or mirror each other and thus disclose a systematic error unaccounted for by random error estimates. If tilt and slope were correlated, this error could be identified and removed. Regression coefficients were calculated, but the fact that none of these coefficients departs significantly from zero indicates no systematic errors larger than the limits of detection, 20 ppm times the maximum topographic height difference (600 m), or 12 mm. Stein and Barrientos [1985] plot tilt as a function of slope for this data and give the detailed statistics. Although tilt and slope can be related tectonically, McTigue and Stein [1984] showed that topographic amplification of stress produces elevation-dependent correlations of only a few parts per million. Because the end points of the coseismic survey lie at nearly the same elevation, these BM's are essentially unaffected by slope-dependent errors. Two isolated BM's that showed elevation changes about 400 mm larger than adjacent BM's were eliminated from the model fitting as unstable.

**Subsidence Errors**

Hydrologic data demonstrate near stasis of the water table during the coseismic period (1933–1984), making subsidence due to fluid withdrawal unlikely. A reconnaissance water resource survey of the valley [Stearns et al., 1938, Plate 31] indicates that the water stood 3–6 m below the ground surface in 1930–1937 in the vicinity of BM 6345 USGS (BM 22, Figure 2a). USGS observation well 9N22E-06CCA1, which was drilled about 150 m west of BM 22 (6345 USGS) in 1967, was observed in 1967–1968, 1978, and 1983–1984 (see Figure 2a). The water table stayed 5.4 ± 0.9 m below the surface during 1967–1984, indistinguishable from the 1930–1937 level obtained by reconnaissance. USGS observation well 9N21E-144BBC1, located 4 km west of P15 (BM 23, Figure 2a), provides a continuous hydrograph during 1966–1984. This well also shows a constant hydraulic head with only 1-m seasonal fluctuations (see the record of Whitehead et al. [1984]). The water table remained static because the alluvial aquifer beneath the leveling route has been recharged by canal diversions from the Big Lost River since 1931 [Crosthwaite et al., 1970].

The 1933 leveling (BM's 1–23) and the 1948 leveling (BM's 22–38) overlap one section at BM's 22–23 (6345 USGS–P15). We used the junction section to test for tilt during 1933–1948. The 1933 and 1948 measurements of height difference over the junction section differ by only 1.56 mm, less than the expected random error for either survey. The absence of measurable tilt during this 15-year period in Thousand Springs Valley supplies a weak test of our assumption that deformation or subsidence preceding the earthquake was negligible.

To summarize, the cumulative uncertainty of BM displacements at each end of the leveling route is 20 mm; the uncertainty of points of greatest relief (km 30–km 40, Figure 3) is nearly the same, owing to smaller random error and a larger potential for slope-dependent error. The uncertainty in elevation change of adjacent BM's is less than 2 mm.
Fig. 4. (Top) Profile of observed coseismic elevation change for BM's in Figure 2a (dots), together with predicted changes (line) for the coseismic model. "Kinks" in the model profile result from changes in azimuth of the leveling route shown in Figure 2a. (Middle) Coseismic model, with main shock and aftershocks beneath the leveling line from Richins et al. [1985]. (Bottom) Schematic geologic cross section from Bond [1978]. Thickness of quaternary alluvium (Qa) is based on gravity modeling by Mabey [1985]. Also shown are Tertiary volcanic deposits (Tev), Mississippian (M) and Devonian (D) carbonates, and the inferred position of the White Knob thrust sheet, based on surface exposure at the northern fault end.

From the displacement spectra of teleseismic P waves, Bourtwick and Choy [1985] estimate a dynamic stress drop of 2.8 MPa for the initial rupture event and a static stress drop of 1.7 MPa (17 bars) for the entire fault. Following Kanamori and Anderson [1975] for dip-slip faults, the static stress drop for the coseismic model is 2.9 ± 0.4 MPa (29 ± 4 bars). The geodetic moment $M_0$ is $(2.4-3.5) \times 10^{19}$ N m (minimum at 2.6 \times 10^{19} N m), under the assumption that the shear modulus is...
Planar Versus Listric Normal Faulting

The leveling data can be simply explained by a planar fault rupture extending from the surface to a depth of 13.3 ± 1.2 km (Figure 6). Diminished slip at a slightly greater depth is possible, because the model prescribes uniform slip everywhere on the fault. The model is consistent with analysis of the main shock epicenter. Slip must also greatly diminish about 7-8 km northwest of the main shock, close to the south end of the model fault (Figure 5). This result suggests that for the Borah Peak earthquake the hypocenter is not a site of significant fault slip but rather the site where rupture initiated, with slip increasing both updip and toward the north.

Interpretation

Planar Versus Listric Normal Faulting

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We examined a suite of listric faults that pass through the aftershocks and the main shock and dip 45° (listric 1) to 80° (listric 4) at the surface (Figure 7, bottom), consistent with field observations of 60°–80° dips on the 1983 Borah Peak scarp face (Crone and Machette, 1984). The curved fault shape was approximated by three planar segments. The fault slip and the absolute elevation change were estimated simultaneously by least squares minimization of the residuals. The model residual for these faults is shown in Figure 7 (middle). Listric 3 (short dash), the best of the faults, shows residuals 2.0 times larger than the best fit planar fault model. Next, slip was varied in each of the three segments on the faults to determine the best fit slip distribution with depth in each case. Coseismic slip confined to the deepest portions of the model faults represents detachment faulting. Under these circumstances, listric 2a provided superior (shown in Figure 6), with uniform slip to a depth of 14.3 km and essentially no slip below this depth. Nevertheless, listric 2a has a residual 1.6 times that for the simpler planar fault. We conclude that for uniform slip, a planar or nearly planar fault fits the data best. Even a negligible amount of concave upward fault shape (listric 2a) degrades the fit to the geodetic observations. Independent support for a planar fault geometry is supplied by a study of the 43 largest aftershocks by Smith et al. [1985], which showed no significant variation in the dip of the nodal planes with depth.

**TABLE 3. Parameters of Basin and Range Earthquakes Deduced From Geodetic Measurements**

<table>
<thead>
<tr>
<th>Event</th>
<th>Dip Slip, m</th>
<th>Dip, deg</th>
<th>Length, km</th>
<th>Width, km</th>
<th>Depth, km</th>
<th>Width of Subsidence, km</th>
<th>Geodetic Moment $M_0$*, $10^{19}$ N m</th>
</tr>
</thead>
<tbody>
<tr>
<td>1983 M = 7.0 Borah Peak†</td>
<td>2.05 ± 0.10</td>
<td>47 ± 2</td>
<td>23.5</td>
<td>18 ± 1.7</td>
<td>13.3 ± 1.2</td>
<td>15-20</td>
<td>2.6 ± 0.3</td>
</tr>
<tr>
<td>1954 M = 7.2 Fairview Peak§</td>
<td>2.8</td>
<td>38</td>
<td>50</td>
<td>8.5</td>
<td>5</td>
<td>5</td>
<td>5</td>
</tr>
</tbody>
</table>

*The shear modulus assumed to be $3 \times 10^9$ MPa (40 kbars). Barrientos et al. [1985] find $M_0 = 1.85 ± 0.17 \times 10^{19}$ N m by modeling the P and SH waves from synthetic seismograms, Boatwright and Choy [1985] found $M_0 = 1.7 ± 0.2 \times 10^{19}$ N m, and Doser and Smith [1985] obtained $M_0 = 2.1 \times 10^{19}$ N m. The calculated geodetic and seismic moments are thus in substantial agreement.

†The end points of the trace of the model fault are $44^\circ 17' 15" N$, $113^\circ 56' 04" W$ on the north end and $44^\circ 04' 08" N$, $113^\circ 48' 12" W$ on the south end. §From Savage and Hastie [1966] and Doser [1985b].

Variation of Slip Along Strike

A 16-km-long segment of the leveling route that is aligned parallel to the Lost River fault (BM's 21-27, Figure 2a) affords a comparison of surface slip and buried fault slip. The fault throw measured by Crone and Macbette [1984] along this segment (from Poison Spring southward to Elkhorn Creek) is shown in Figure 5 (top) together with the subsidence on the downthrown block measured from releveling (middle). The string of BM's lie 4–5 km basinward from the fault trace; the locations of the BM's and surface break are shown on the accompanying map (Figure 5, bottom). Both the subsidence and the displacement on the surface rupture decrease toward the main shock epicenter. Slip must also greatly diminish about 7–8 km northwest of the main shock, close to the south limit of surface rupture (the end of the model fault is situated at km 17 in Figure 5). This result suggests that for the Borah Peak earthquake the hypocenter is not a site of significant fault slip but rather the site where rupture initiated, with slip increasing both updip and toward the north.
The key result of this analysis is that the Lost River fault must be nearly planar to a depth of at least 12 km. The deformation data yield no information as to the fault geometry below this depth, but they preclude any coseismic slip on low-angle or listric faults above this depth. Mesozoic thrust sheets have been proposed as the likely candidates for low-angle slip surfaces in the seismogenic crust, with the high-angle faults cut by or merging into the flat-lying faults. The White Knob thrust sheet (Figure 4) is in close proximity to the Lost River fault but did not slip and produced no aftershocks during the earthquake sequence. A number of investigators have proposed that low-angle faults are active at depth of 2–10 km [Wernicke, 1981; Smith and Bruhn, 1984; Hamilton, 1985]. The Borah Peak earthquake provides no support for this view, although low-angle faults may creep between earthquakes or may be active elsewhere.

The character of the deformation produced by slip on the planar high-angle fault has been previously attributed to listric fault geometries. Coseismic surface deformation steepens near the fault on both the upthrown and downthrown blocks (Figure 4); this is a dominant feature of the profiles of the data and both the planar and listric models. The pattern of increasing dip of strata toward the fault in the post-Tertiary structure (Figure 4, bottom) has been interpreted as being caused by block rotation or void accommodation resulting from a listric fault geometry [Longwell, 1945; Anderson et al., 1983]. Increasing stratal dips toward the fault may, instead, be the product of elastic deformation associated with repeated earthquakes and thus be unrelated to block rotation.

**Absolute Elevation Changes**

Leveling across the Lost River Range and Thousand Springs Valley permits tentative assignment of absolute elevation changes. The end points (BM's 1 and 39, Figure 4) give the same elevation change within the measurement uncertainty (20 mm). Because the end points of the survey extend outside the region of observed and modeled coseismic deformation, they can be assumed to be fixed. Very broad wavelength tilts with magnitudes of less than 1 ppm would go undetected, however, as would uniform vertical displacement over distances greater than 50 km. Absolute uplift associated with the earthquake is 0.3 m (at BM 17); the crest of the Lost River Range in the vicinity of Borah Peak rose about 0.2 m (Figure 4). Absolute uplift of the ranges bounding a normal faulting earthquake has never before been demonstrated, because leveling was ambiguous or absent on the upthrown fault block. The maximum coseismic subsidence measured at Borah Peak was 1.2 m at the east margin of Thousand Springs Valley.

From the topographic relief, the geomorphic expression of the fault scarp, and the inferred thickness of the basin deposits, Scott et al. [1985] argue that 3 km of dip slip has accumulated over the past 4–7 m.y. on the Lost River fault, at a rate of 0.2–0.4 mm/yr. Repeat of events equal in size to the 1983...
Fig. 7. (Top) Comparison of observed elevation changes and listric fault model 2. Note systematic misfit on both sides of the fault. (Middle) Residuals for four listric fault models. (Bottom) Depth cross section of the four listric fault models with seismicity.
We interpret these fundamental similarities to indicate that the Borah Peak earthquake typifies the events which built the profile across Thousand Springs Valley [Smith et al., 1985].

Comparison of the Borah Peak, Fairview Peak, and Hebgen Lake Events

Although the 1954 Fairview Peak, Nevada, $M = 7.2$ event [Savage and Hastie, 1966; Snay et al., 1985] and the 1959 Hebgen Lake, Montana, $M = 7.3$ event [Meyers and Hamilton, 1964; Savage and Hastie, 1966] may have occurred on listric faults, the seismic and available geodetic evidence favors a similar planar geometry for all three faults. The fault dips estimated from geodetic elevation changes for the events and for the Idaho shock average about $45^\circ$. Doser [1985] found the $M = 7.3$ Hebgen Lake main shock to lie at a depth of $15 \pm 3$ km, with the principal nodal plane dipping $45^\circ$–$60^\circ$.

Table 3 lists the geodetically determined fault parameters for these three events. Neither the Nevada nor the Montana co-seismic model was fitted to geodetic observations from the upthrown fault block, a circumstance that greatly hampers resolution of the fault geometry. The geodetic record for the Montana earthquake also lacks observations in the region 20–30 km from the fault on the downthrown block. The Nevada data are subject to probable systematic leveling surveying errors that have yet to be removed and that limit their utility [see Savage and Church, 1974, Figure 5].

The few well-studied shocks of $M < 7$ also conform to a geometry of planar faulting. The 1954 $M = 6.8$ earthquake on the Dixie Valley fault (Figure 1) also nucleated at a depth of 15 km on a fault dipping about $50^\circ$–$62^\circ$ [Okaya and Thompson, 1985; Snay et al., 1985]. Arabasz et al. [1981] found a 13- to 14-km maximum depth of aftershocks associated with the 1975 Pocatello Valley, Idaho, $M = 6.0$ earthquake (Figure 1).

Implications for Earthquake Prediction in the Great Basin

The base of a seismic fault is a plausible site for slippage before an earthquake. Detection of preseismic fault slip at a depth of 12–15 km is a much more challenging task than if premonitory slip were to take place on a detachment surface located at much shallower depths. To monitor slip at a depth of 12–15 km on a fault dipping 45$^\circ$, geodetic surveys must extend from the fault trace for about 30 km on the downthrown block, and they must have sufficient precision to detect broad-wavelength deformation. A 4- by 4-km dislocation patch situated at the base of the fault would cause the ground to subside less than 20 mm per meter of fault slip. If the zone of preseismic slippage extended 35 km along strike (for example, a 4- by 35-km strip beneath the entire Borah Peak rupture surface), then as much as 90 mm of subsidence per meter of buried slip would take place (Figure 8). For this hypothetical preseismic slip the greatest subsidence would occur within the basin, making discrimination of tectonic deformation from groundwater-related subsidence very difficult. In any case, leveling restricted to within 10 km of the fault trace will be insufficient to detect preseismic slip at the base of the fault.

CONCLUSIONS

One of the best records of coseismic elevation change ever measured for a large normal faulting event is most compatible with a planar high-angle fault rupture. The earthquake parameters deduced from geodetic observations are consistent with those gaged from seismic measurements. The 0.3 mm/yr slip rate and 5000- to 10,000-year earthquake repeat time estimated from the geodetic data and the recent geology is commensurate with geological estimates averaged over the entire 4–7 m.y. life of the Lost River fault. Absolute uplift of ranges accompanying an earthquake in a region of continental extension has been recorded.

Similarity of the fault dip ($40^\circ$–$50^\circ$) and the maximum depth...
of fault slip (12–15 km) for the Borah Peak, Heben Lake, Fairview Peak, and Dixie Valley events suggests that steeply dipping planar faults pose the dominant seismic hazard in the Great Basin. The maximum depth of faulting for these events lends evidence to the hypothesis of Sibson [1982] and Smith and Brunn [1984] that a depth of 12–15 km corresponds to the brittle-ductile transition in the crust at the center of the ranges, and many characteristics with the Lost River fault: Both faults existence of low-angle Mesozoic thrust sheets among the high-contemporaneous slip or Quaternary geologic displacement. The pattern of deformation associated with the Borah Peak earthquake, with increasing stratal dips toward the fault, has previously been interpreted as evidence for listric faulting. The existence of low-angle Mesozoic thrust sheets among the high-angle faults has given rise to the hypothesis that the listric thrust faults now undergo normal slip. Acceptance of listric faults and detachments in the upper crust as the principal sites accommodating Great Basin extension must await evidence of contemporaneous slip or Quaternary geologic displacement.

The heavily populated Wasatch fault zone in Utah shares many characteristics with the Lost River fault: Both faults show greatest cumulative surface displacement and shortest measurable repeat times near the center of the ranges, and both fault zones appear to be typified by 2-m slip events [Schwartz and Coppersmith, 1984; Scott et al., 1985]. Whether the Wasatch fault zone also shares the subsurface geometry of the Hebgen Lake and Lost River faults is uncertain. Seismic reflection profiles across the Wasatch zone should be reexamined for deep planar bounding fault structures, as Okaya and Thompson [1985] have done in Dixie Valley, Nevada. Irrespective of its subsurface geometry, efforts to detect premonitory deformation surveys that span the fault and the downthrown block for at least 30 km, or twice the probable depth of the fault.

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