NEOTECTONICS OF THE HANSEL VALLEY-POCATELLO VALLEY CORRIDOR, NORTHERN UTAH AND SOUTHERN IDAHO

By JAMES MCCALPIN, ROBERT M. ROBISON,[1] and JOHN D. GARR

ABSTRACT

The Hansel Valley-Pocatello Valley corridor on the Utah-Idaho border has experienced intense historical earthquake activity, including damaging earthquakes in 1934 (M_L 6.6) and in 1975 (M_L 6.0). Range fronts flanking both valleys display youthful morphology, yet valleys contain few well-preserved fault scarps in Quaternary deposits that might allow reconstruction of fault histories. Range-front trenching, pluvial lake shoreline surveying, and arroyo wall mapping were used to reconstruct the neotectonic history of each valley. In Pocatello Valley, 15-ka shorelines are locally warped vertically as much as 6.4 m where they cross en echelon splays of the eastern range front fault, but no scarps in unconsolidated deposits are found. A trench at the range front exposes unfaulted colluvial wedges up to 95 ka old that may have been rotated down toward the valley center and locally warped. The absence of either fault scarps or liquefaction features in late Quaternary deposits implies that displacement on the range-front fault in eastern Pocatello Valley occurs by creep or by small displacements that do not propagate as discrete fractures to the surface.

Two short fault scarps on the eastern range-front fault of Hansel Valley indicate that four to five surface-faulting events have occurred in the last 65 to 140 ka. A splay fault in an Interstate 84 roadcut shows only one recent 2.6-m displacement in the last 100 ka, not compatible with the scarp evidence. Along the southwestern valley margin, one of several scarps created by the 1934 earthquake merges with an 8-km-long prehistoric scarp up to 9 m high that displaces Lake Bonneville shoreline gravels. Scarp profiles and geomorphic relations with shorelines indicate that the scarp was formed by one or two subaqueous displacements totaling 1.3 to 2.5 m between 26 and 12 ka. Supporting evidence of recurrent faulting based on gully exposures and five thermoluminescence dates show that deposits have been multiply faulted between 138 and 72 ka; no events occurred from 72 to 58 ka, one event occurred between 58 and 26 ka (nearer the latter), one event occurred around 15 to 21 ka, and a possible event occurred at 13 ka. Displacements of up to 2.5 m seem to be clustered in time when large pluvial lakes existed in the basin (oxygen isotope stages 6 and 2) rather than when either no lake or only shallow lakes existed (oxygen isotope stages 5, 4, and 3). We suggest that the 1934 M_L 6.6 earthquake may be a typical interpluvial maximum event (long recurrence time, small displacement) in comparison with the larger, more frequent surface-faulting events presumably triggered by pluvial lake water loading.

INTRODUCTION

The Hansel Valley-Pocatello Valley corridor on the Utah-Idaho border has been the site of intense historical seismicity, including the 1934 M_L 6.6 and the 1975 M_L 6.0 earthquakes (fig. 1). The 1934 event, the largest in Utah's history, is one of only three historical earthquakes in the Intermountain Seismic Belt to rupture the surface and produce a fault scarp. The high rate of historical seismicity in the corridor contrasts with a scarcity of prehistoric fault scarps offsetting late Quaternary deposits, in spite of the youthful range-front morphology of bounding mountain ranges. The primary objective of this research was to determine why such a seismically active area lacked abundant geologic evidence of late Quaternary surface faulting. A secondary objective was to compare earthquake frequency-magnitude relations deduced from geologic evidence with those from historical time to help assess the current potential of this region to generate large, damaging earthquakes. Finally, if damaging earthquakes are found to have occurred without leaving significant geologic evidence, do the methods currently used on the Wasatch fault need to be modified for estimating earthquake potential?

The study area is located within the Hansel Mountains-West Hills part (Stokes, 1977) of the Great Basin section of the Basin and Range physiographic province (Stokes, 1977) and is composed of north- to northeast-trending valleys separated by linear mountain ranges (fig. 2). The Hansel, North Hansel, and North Promontory Mountains are primarily composed of the Pennsylvanian-Permian Oquirrh Formation (Adams, 1962; Allmendinger, 1983; Jordan, 1985), while Samaria Mountain

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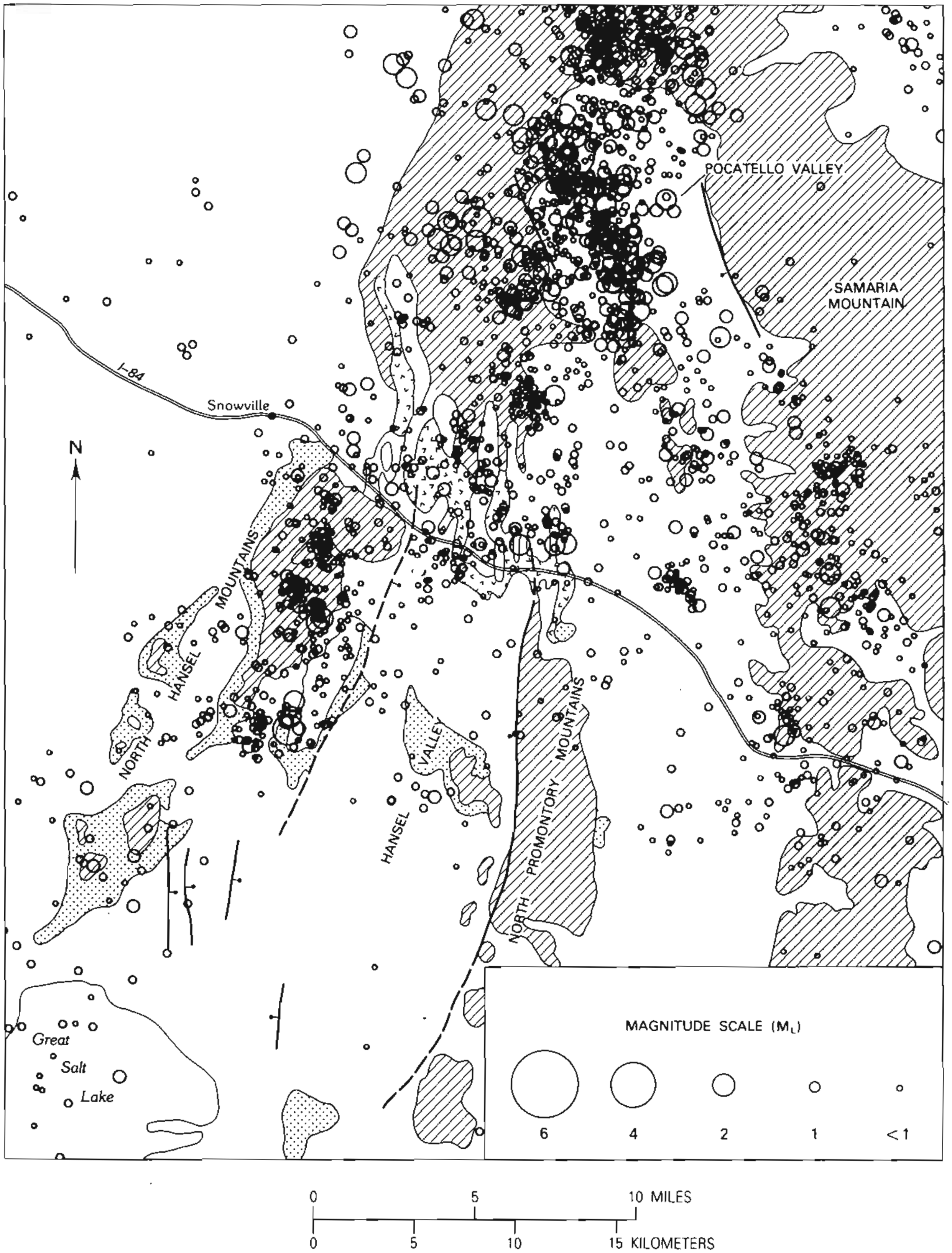
[1]Now at Sergent, Hauskins, and Beckwith Engineers, 4030 South 500 West, Salt Lake City, UT 84123.


POCATELLO VALLEY
SAMARIA MOUNTAIN
I-84
Snowville
N
NORTH HANSEL MOUNTAINS
HANSEL VALLEY
NORTH PROMONTORY MOUNTAINS
Great Salt Lake
MAGNITUDE SCALE (M_L)
6
4
2
1
<1
0 5 10 MILES
0 5 10 15 KILOMETERS

◀ FIGURE 1.—Earthquake epicenters of the Hansel Valley-Pocatello Valley corridor for the period March 28, 1975, to June 30, 1985. Data provided by the University of Utah Seismograph Stations. Diagonal patterns represent mountain blocks; stippling represents Lake Bonneville shoreline gravels. Normal faults displaying Quaternary displacement are shown by heavy solid lines (bar and ball on downthrown side). Dashed lines show inferred fault locations at the base of steep range fronts.

contains carbonates and clastic rocks of Ordovician through Pennsylvanian age (Beus, 1968; Platt, 1977). In the valleys, Lake Bonneville sediments occur below about 1,585 m; higher valley margins are covered by alluvial fans. The valley fill beneath Lake Bonneville sediments is as much as 570 m thick in Pocatello Valley (estimated from gravity data by Harr and Mabey (1976, p. 6–7)), presumably composed mainly of pre-Bonneville Quaternary deposits and of the Tertiary Salt Lake Formation, which crops out on valley margins. Fill in Hansel Valley is considerably thinner, as shown by smaller gravity anomalies (fig. 3) (from Peterson, 1974) and by outcrops of Paleozoic rock within the valley (Hood, 1971, p. 5).

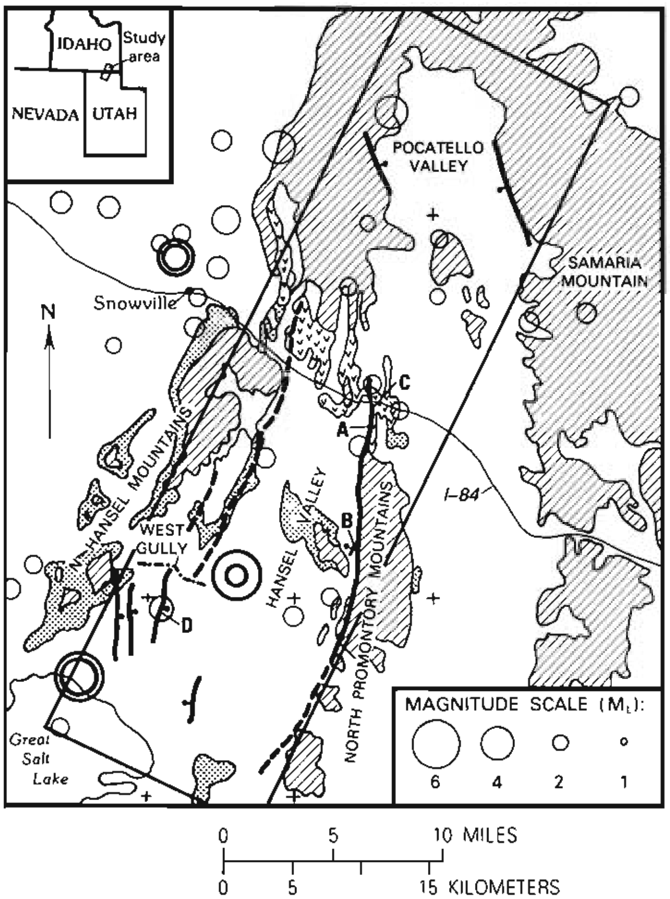


FIGURE 2.—Location and general geology of the Hansel Valley-Pocatello Valley corridor (rectangle). Marginal normal faults discussed in text are shown by heavy lines (dashed where inferred, bar and ball on downthrown side). Earthquake epicenters from 1850 through June 1962 are shown by heavy circles; epicenters from June 1962 through March 27, 1975, are shown by lighter circles. Paleozoic rocks, diagonal pattern; Tertiary volcanic rocks, small v's; Quaternary shoreline gravels, stippling; Quaternary lake-bottom deposits and alluvium, no pattern. A through D locate faults discussed in text.

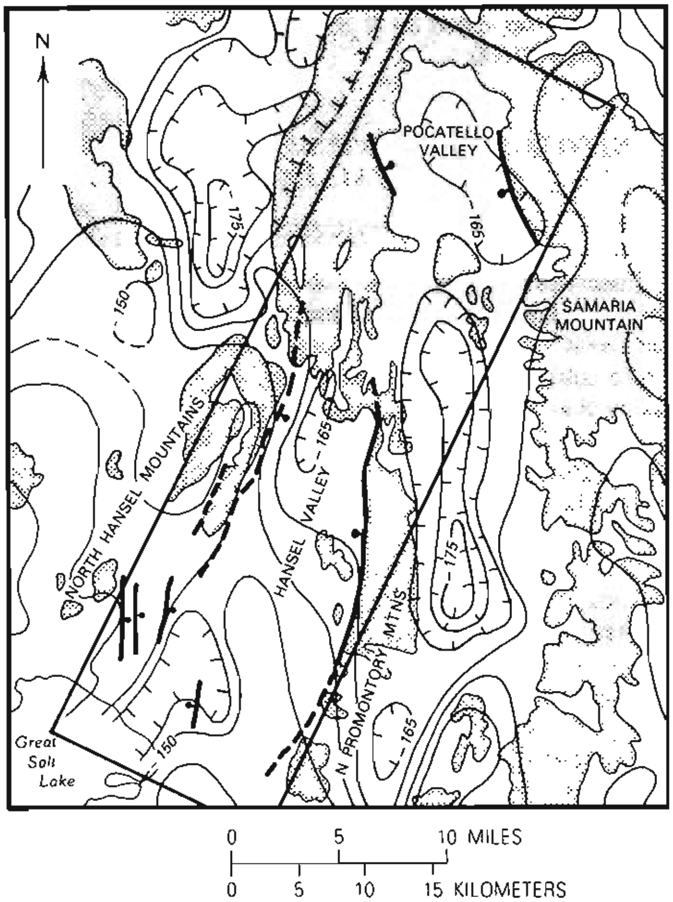


FIGURE 3.—Bouguer gravity anomalies of the Hansel Valley-Pocatello Valley corridor (rectangle) and adjacent areas (from Peterson, 1974). Gravity values are in milligals; contour interval 5 mGal. Stippling represents mountain blocks; heavy lines represent Quaternary faults (dashed where inferred, ball and bar on downthrown side). The gravity low of Pocatello Valley seems to continue southward to Blue Creek Valley rather than southwestward to Hansel Valley.

ACKNOWLEDGMENTS

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NEOTECTONIC FEATURES IN POCATELLO VALLEY

TECTONIC SETTING

Pocatello Valley is a topographic and structural depression bounded on the east by Samaria Mountain, on the west by the North Hansel Mountains, and on the north and south by low hills (fig. 4). Cenozoic normal faults bound the eastern side of the valley (Beus, 1968; Platt, 1977) and part of the western side (Allmendinger, 1983). Platt (1975, 1976) inferred late Quaternary activity on the eastern margin fault from indirect geomorphic evidence such as steepness of faceted spurs. Both gravity data and tilt of strata suggest that this fault forms the boundary between two east-tilted blocks. Another inferred north-trending fault may bound the western side of a gravity saddle in the center of the valley (fig. 3). Cenozoic displacement on the eastern margin fault may be up to 1,420 m, as shown by the elevation difference between the top of Samaria Mountain (2,440 m) and the bottom of inferred valley fill (1,020 m) (Harr and Mabey, 1976). A shorter (8 km long) normal fault of lesser displacement bounds the central western margin of Pocatello Valley. Allmendinger (1983, section *C–C'*) showed a down-to-the-east displacement of about 122 m where it offsets Paleozoic rocks on its northern end. The smooth gravity gradients and the lack of large range-front spurs suggest that the western valley margin is dominantly a dip slope of an east-tilted block (Allmendinger and Platt, 1983, fig. 3) marked by a few short, low-displacement normal faults. In late Quaternary time, this topographic depression was occupied by pluvial Lake Utaho (Currey, 1981), a small body of water contemporaneous with but physically separate from Lake Bonneville.

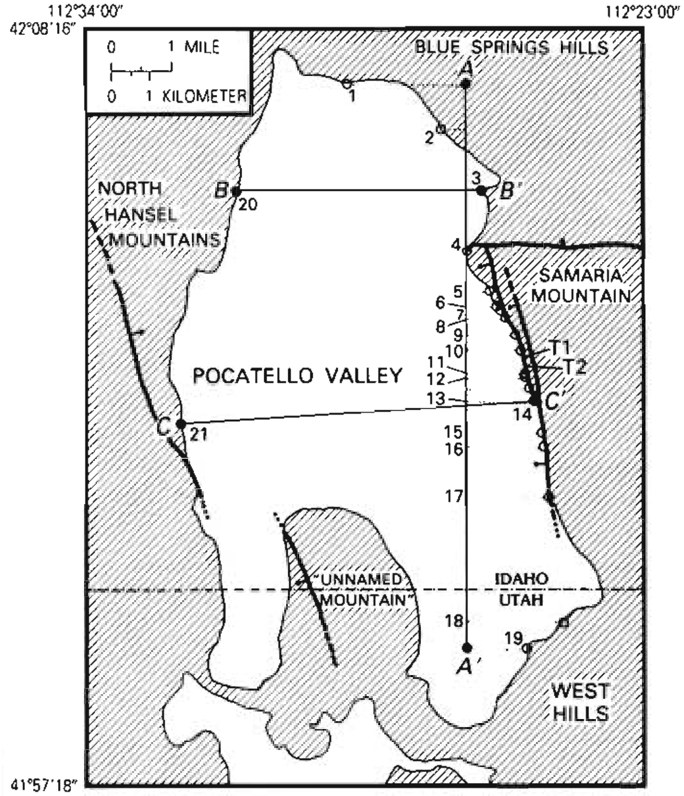

FIGURE 4.—Pocatello Valley area, Idaho-Utah. Diagonal pattern indicates highlands above the Lake Utaho and Lake Bonneville shorelines. Open circles are shoreline elevation data points; solid circles are data points and (or) endpoints of profiles shown in figure 9. Dotted lines indicate the position of shoreline elevation data points used in transect *A–A'* (fig. 9). Normal faults are shown by heavy lines (dashed where inferred, bar and ball on downthrown side). T1 and T2 indicate the locations of trench 1 and trench 2, respectively.

EASTERN MARGIN FAULT

The youthful range-front morphology cited by Platt (1977) as evidence of late Quaternary activity was quantified by using mountain-front sinuosity ratios (S) and ratios of valley floor width to valley height (V_f) (after Bull and McFadden, 1977). S is the ratio of the length of the mountain-piedmont junction to the overall length of the mountain front. Tectonically inactive fronts have high (3.0–7.0) S values caused by extensive embayment and pedimentation, whereas tectonically active fronts have lower values (1.2–1.6). The eastern margin range front has a sinuosity ratio of 1.7, suggestive of slightly active to active tectonism. The ratio of valley floor width to valley height (V_f) gives a general indication of whether streams draining the mountain block are engaged in channel downcutting (high rate of tectonic base-level fall, low value of V_f) or lateral erosion (low rate of base-level fall, high value of V_f). The width of a canyon floor is compared with the mean height of the canyon at a given distance (0.5 km in this study) upstream from the mountain front. Canyons and valleys in several ranges in the northern Mojave desert have V_f values ranging from 0.055 to 47.0 (Bull and McFadden, 1977). Only two canyons along the eastern margin of Pocatello Valley have stream channels long enough to use in calculating ratios of valley floor width to valley height; they yielded low values of 0.22 and 0.24 that indicated active tectonism. The mountain-front sinuosity ratio and the ratios of valley floor width to valley height quantify what can be seen in the field: the mountain front is steep, the canyons are deeply incised and strongly V shaped, and embayment and pedimentation are minimal, all suggestive of a young, active mountain front. In addition, Platt (1976) inferred that subsidence of the valley floor near the base

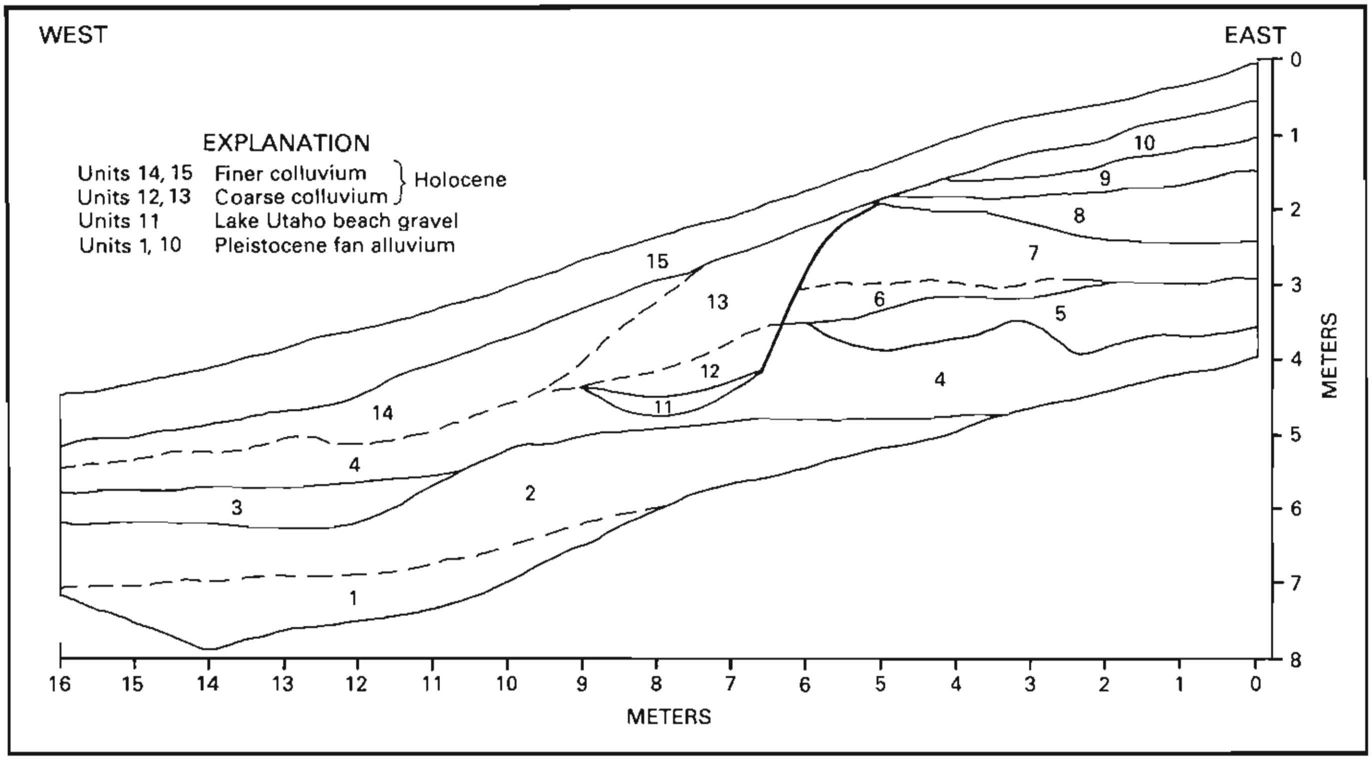

FIGURE 5.—Log of trench 1, excavated across the Lake Utaho shoreline on the eastern valley margin (T1, fig. 4). Lenticular late Pleistocene alluvial-fan gravels (units 1–10) are truncated by an erosional scarp, at the base of which lies Lake Utaho highstand beach gravel (unit 11). Bouldery scarp-derived colluvium (unit 12) is overlain by a wedge of gravelly colluvium (unit 13), the 35° W.-dipping fabric of which suggests a growing colluvial wedge. Finer colluvial unit 14 appears to be mixed with loess (early Holocene?), and unit 15 contains a weak A/Cca soil profile. Alluvial-fan strata beneath the buried scarp face (units 1–4) are not displaced, proof that no surface faulting has occurred at this location in the last 15 ka.

of Samaria Mountain has resulted in alluvial fans of much smaller volume than one would expect, given the size of the canyons from which the alluvium originated. According to gravity data (Harr and Mabey, 1976), the deepest depocenter within the valley (570 m of fill) is adjacent to the southern portion of the Samaria Mountain front. The gravity gradient from the mountain to the depocenter is approximately 3 mGal/km, almost double the gradient around most of the valley margin. Despite this indirect evidence for Quaternary fault activity, no fault scarps displacing Quaternary deposits could be located at the range front, either by earlier workers (Rogers and others, 1975) or in this study.

STRATIGRAPHY AT THE RANGE FRONT

Two trenches were excavated by a large track-mounted backhoe to look for evidence of Quaternary faulting at the range front. Trench 1 was cut across the most prominent scarp in Quaternary deposits along the range front (T1, fig. 4). This scarp was located at the same elevation as the Lake Utaho shoreline and could have been (1) a fault scarp, (2) an abnormally well developed pluvial lake shoreline scarp, or (3) a fault scarp later occupied and reworked by a shoreline. Trench logging revealed a thin lens of beach gravels at the base of the buried scarp free face, overlain by a fining- upward colluvial wedge (fig. 5). Moderately well stratified alluvial fan gravels beneath the beach gravel were continuous across the entire length of the trench, indicating no tectonic offset beneath the scarp and thus disproving a tectonic origin.

Because trench 1 yielded no tectonic information, trench 2 (T2, fig. 4) was cut through a colluvial apron at the base of a faceted spur approximately 400 m south of trench 1. Excavation was begun at the base of the spur directly on an outcrop of Pennsylvanian-Permian Oquirrh Formation limestone, which was encountered at depths ranging from 0 to 1.5 m in the upslope 15 m of the trench (fig. 6). The bedrock contact beneath colluvium then abruptly dropped 3 m into the trench floor at an angle of 63°, beyond the limit of extension of the backhoe. The limestone in and around the inferred range-front fault zone is intensely fractured and locally hydrothermally altered. The oldest colluvial deposit (unit I) carries a paleosol displaying well-developed argillic horizons. Comparison of this soil with a dated buried soil from

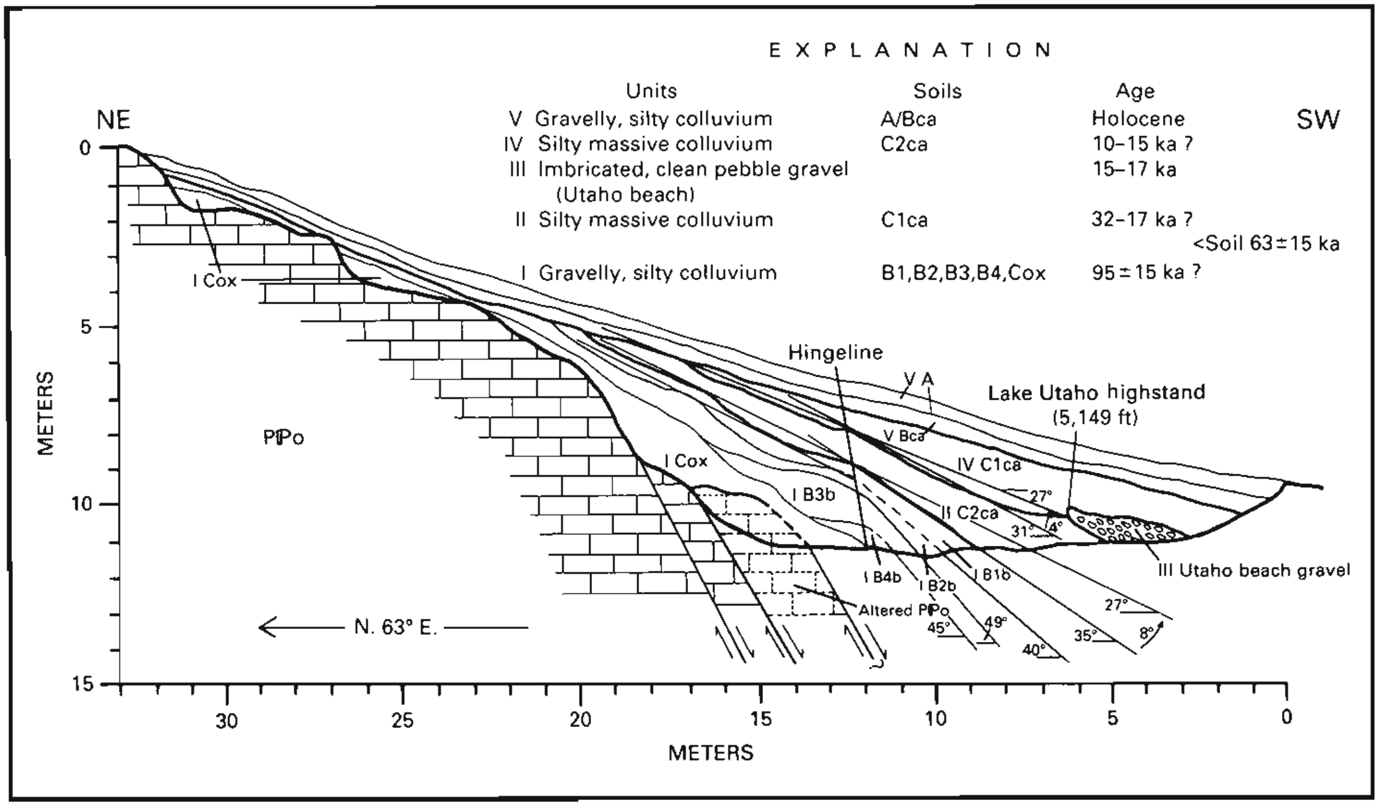

FIGURE 6.—Log of trench 2, located at the base of a faceted spur on the eastern valley margin (T2, fig. 4). Heavy lines separate lithologic units I through V; light lines separate soil horizons. Dips of units II through IV increase downslope of the hinge line by 4° to 8°; dips in unit I soil horizons increase up to 49°. Ages of units are estimated by correlating the Lake Utaho shoreline gravels (unit III) with the Lake Bonneville shoreline and by assessing the degree of soil development (fig. 7). The block pattern represents limestone bedrock (dashed where bedrock has been fractured and hydrothermally(?) altered to a variegated clay).

Jordan Valley, Utah (Scott and others, 1982, p. 42), suggests that this soil represents about 63±15 ka of weathering (fig. 7). Overlying stone-free (loessial?) colluviums (units II and IV) both underlie (unit II) and overlie (unit IV) Lake Utaho highstand beach gravel (unit III), which is assumed to be contemporaneous with the Bonneville shoreline (15.5–17 ka) (Scott and others, 1983). The modern colluvium (unit V) truncates all underlying units.

Although the overall geometry of older, steeper wedges overlain by successively younger, gentler layers is expectable, the very steep tilt of unit I paleosols (up to 49°) is anomalous. Colluvial wedges below fault scarp free faces typically have initial slopes of 35° (Wallace, 1977); this slope is maintained only while deposition is rapid. By the time the slope is declining slowly enough to allow soil formation on the colluvial wedge, the wedge slope is considerably lower (8°–25°) (Wallace, 1977, fig. 3, stage E). Buried soil horizons within unit I must have required slope stability to form, but now dip 35° to 49° west downslope of the hinge line shown in figure 6. The unit II-unit IV contact is bent about 4° along this hinge line, whereas younger units are unbent. The amount of angular change on the unit I-unit II contact (8°, age ~32 ka if deposition occurred early during the Utaho transgression) versus that on the unconformity between units II and IV (4°, age 15–17 ka) suggests progressive monoclinal flexure with time (0.25°/ka). If this rate is applied to the even more steeply dipping buried soil horizons B3b and B4b, an age of approximately 88 ka is obtained. This date is roughly compatible with an age of 95±15 ka for unit I estimated by adding the time of soil formation (63±15 ka) to the inferred age of the unit I-unit II unconformity (fig. 7). There are no discrete faults or shears within the colluvial wedge.

Monoclinal flexure of surficial materials in narrow (2–5 m wide) zones overlying fault traces has been previously described by Clark and others (1972, p. 118) and by Bonilla (1982, p. 18). Although the evidence in trench 2 is not conclusive, it suggests that some kind of progressive flexure has occurred over the last 95 ka within 20 m downslope of the inferred range-front fault in bedrock, unaccompanied by any surface faulting. It is unknown whether the warping resulted from coseismic deformation or aseismic creep, but the absence of fractures supports the latter explanation.

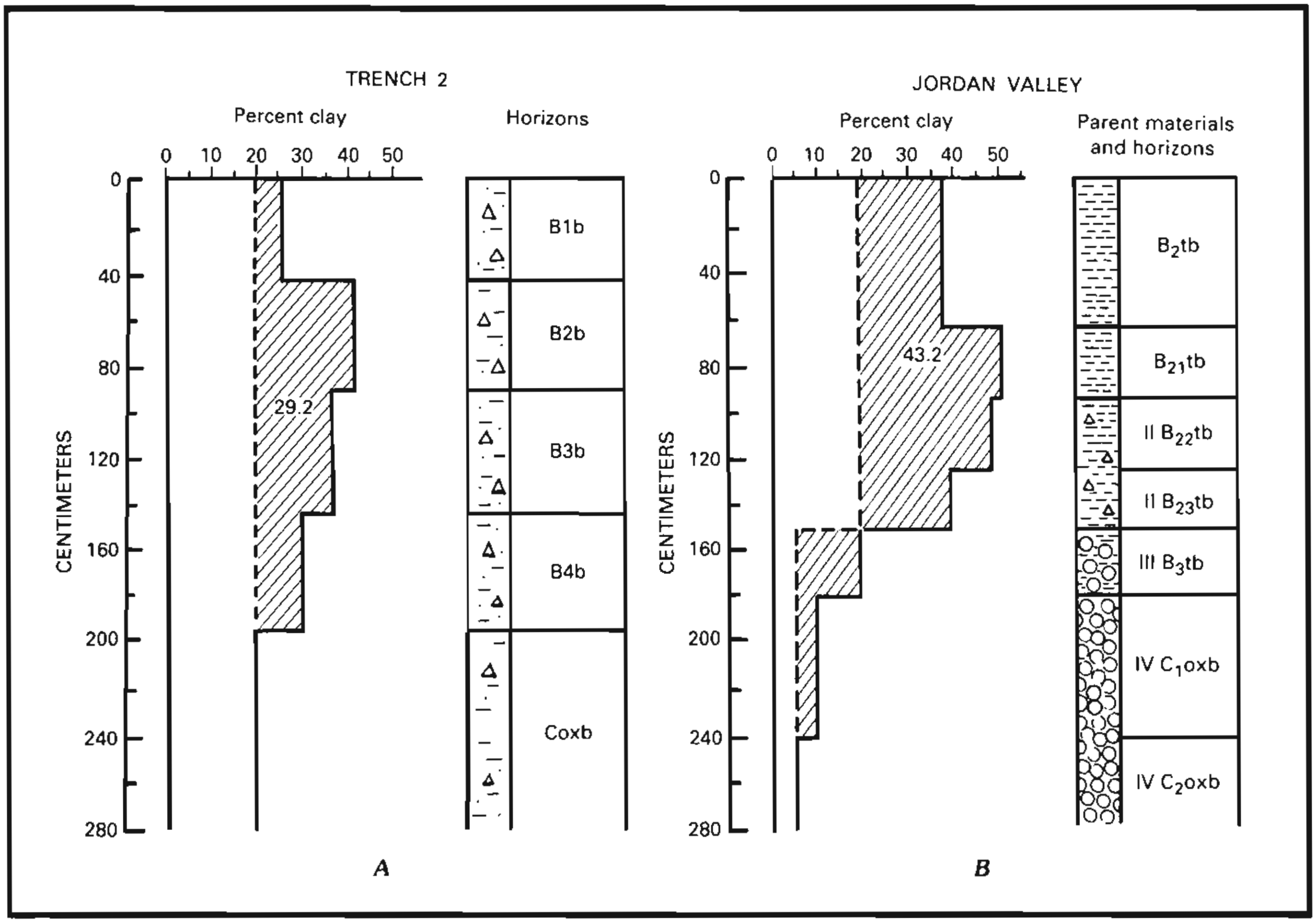

FIGURE 7.—Comparison of soil profile properties between (*A*) the trench 2 paleosol developed on unit I and (*B*) a buried soil developed on pre-Bonneville lacustrine gravel in Jordan Valley, Utah (Scott and others, 1982, table 5, p. 42). Total pedogenic clay in each horizon is shown by the diagonal pattern, computed by multiplying the percentage of pedogenic clay by weight in each soil horizon (heavy lines) by the horizon thickness. Two assumptions are made in the comparison: (1) the original clay content of both parent materials (dashed lines) was about 20 percent for loess and colluvium and 5 percent for lacustrine gravels, and (2) bulk densities of comparable horizons in both soils are similar. Numbers in the patterned areas show dimensionless values for comparing the amount of clay in each soil. The soil shown in *A* contains roughly 67 percent as much pedogenic clay as the soil shown in *B*. If clay formation rates are similar for the soil in *A* (Pocatello Valley) and the soil shown in *B* (Jordan Valley), the soil in *A* should represent roughly 67 percent of the time of soil formation represented by the soil in *B*. Scott and others (1982) calculated the formation time of the calcareous facies of the soil in *B* as 94±22 ka. Comparison of pedogenic clay amounts suggests that the soil in *A* in trench 2 developed over a period equivalent to 67 percent of 94±22 ka or roughly 63±15 ka. Horizontal dashes, silt; dots, sand; triangles, angular pebbles and small cobbles; open circles, rounded pebbles and cobbles. Horizon abbreviations follow U.S. Department of Agriculture conventions (b indicates buried horizons).

WARPED SHORELINES

The elevation of the Lake Utaho shoreline was determined at 46 points around the margin of Pocatello Valley to detect indirect evidence of tectonic deformation. Survey points were concentrated along suspected fault zones at the base of steep mountain fronts to look for anomalous warps or tilts in the originally horizontal shorelines. Surveying was performed with a Leitz TM10E theodolite and Leitz RED–2 electronic distance meter. The meter and the theodolite were set up over a benchmark or spot elevation on the valley floor, and a tripod-mounted reflecting prism was placed at the base of the shoreline scarp. While measurements were taken at the instrument station, a profile of the shoreline scarp was made by using a 4.5-m rod and Abney level (after Bucknam and Anderson, 1979, p. 12). In most instances, the profile measurement was made from 30 m above to 30 m below the shoreline scarp. Typically, three distinct slope components were evident: a beach platform (3°–4°), a colluvial wedge (8°–12°), and a wave-cut scarp (18°–28°) (fig. 8).

Because the Pleistocene lake shorelines were developed in unconsolidated deposits in all but a few locations

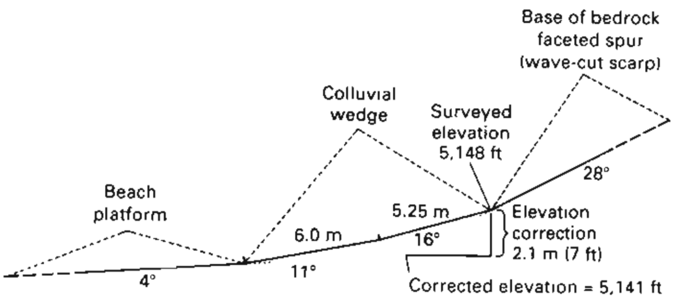

FIGURE 8.—Method used in estimating actual elevations of pluvial lake shorelines covered by Holocene colluvium. The beach platform and wave-cut scarp components are projected beneath the colluvial wedge. The resulting intersection is presumed to be a more accurate representation of the actual shoreline angle.

around the valley margin, the shoreline angle has been covered subsequently by Holocene colluvium. The amount of colluvial cover varies depending on slope, aspect, deposit grain size, and local drainage patterns. To eliminate the effect of colluvial deposition from the surveyed shoreline scarp elevations, the beach platform and wave-cut scarp angles were projected under the colluvial wedge component on the profiles (fig. 8). The intersection of the two angles is assumed to be a more accurate approximation of the shoreline angle. Because trenches 1 and 2 were excavated after the shoreline surveying phase of the study, they provided a means of checking the accuracy of the method; in both instances, the projected beach platform and wave-cut scarp angles intersected within 30 cm of the tops of Lake Utaho beach gravel lenses.

Although the water surface at the Lake Utaho high-stand was horizontal, shorelines could have been super-elevated, coincident, or subelevated with respect to the formative water plane (Currey, 1982, p. 21–22). All survey points along the eastern range front were in similar drift-aligned gravel beaches; data from Rose (1981, table 5.6) showed that the shoreline angle of such beaches is usually within 0.5 m of the mean water surface. Large waves that might deposit gravel far above the mean water surface would be hard to generate in Lake Utaho, which was only 10 km in diameter at its maximum extent.

The precision of the instrumentation used in this study was calculated to be ±1.31 cm elevation per kilometer of horizontal distance. The longest survey shot made was 1.78 km, and most shots were less than 0.48 km long. The combined uncertainties of surveyed elevation (13 cm), projected shoreline angle (1.0 ft), and relation to water surface (50 cm) yield a ±93-cm uncertainty for each measured shoreline point (table 1).

Figure 9 (transect *A–A′*) shows shoreline elevations that deviate as much as 6.4 m from one another and up to 4.5 m from elevations predicted by Crittenden (1963) on the basis of basin-scale isostatic rebound. The most significant deviations from a smooth profile occur where the north-northwest-trending shoreline crosses the more northerly trending faults at the base of Samaria Mountain. The shoreline at the southern end of the transect (point 17) is cut onto a faceted spur on the upthrown side of the fault, whereas, at points 4, 5, and 6, the shoreline is clearly on the downthrown side of the two parallel range-front faults (fig. 4). Point 17 falls at almost the exact elevation predicted for the Bonneville shoreline by Crittenden (1963, fig. 3), whereas points 4, 5, and 6 on the downthrown block plot 4.2 to 4.5 m, too low in relation to their expected post-rebound elevations. Highly variable shoreline elevations between points 16 and 7 probably result from the sinuous shoreline's repeatedly crossing from the upthrown to the down-thrown side of a more linear but buried fault or fold hinge line. For example, point 9 is a shoreline cut onto bedrock at the base of a spur, whereas the shoreline at points 8 and 10 is cut into colluvium farther valleyward. The position of a linear fault paralleling the range front would cross between point 9 and points 8 and 10. Farther north, a parallel branch of the north-trending range-front fault intersects the north-northwest-trending Utaho shoreline somewhere between points 6 and 9 (fig. 9). Within this area, the shoreline loses 6.4 m of elevation in a horizontal distance of 900 m. The great similarity of all shoreline profiles suggests that they were formed with a constant relation to the mean water plane. Importantly, there are no fault scarps visible in this area.

East-west profiles suggest that the Utaho shoreline has been tilted eastward from 1.8 m (fig. 9, transect *B–B′*) to as much as 7.4 m (fig. 9, transect *C–C′*) in the most active-appearing range-front area. Crittenden (1963, fig. 3), showed an expected west-to-east shoreline elevation drop of approximately 1 m across Pocatello Valley resulting from differential isostatic rebound. The small residual displacement on transect *B–B′* (1.8 m − 1.0 m = 0.8 m) is within the range of surveying error and suggests that no significant tectonic eastward rotation has occurred north of the mapped eastern margin fault. The 6.4-m residual down-to-the-east displacment along transect *C–C′* strongly argues for post-Utaho slip along the central range front.

The overall implication is that late Quaternary slip on the range-front fault has resulted in very little absolute elevation change of the upthrown block, but points on the downthrown block have been lowered 4.5 to 6.4 m, presumably by subsidence of the Pocatello Valley block. Geodetic data collected by Bucknam (1976) after the 1975 M_L 6.0 earthquake showed no absolute uplift of surrounding mountain blocks but a maximum of 13 cm of valley floor subsidence near the epicenter in western

TABLE 1.—*Surveyed elevation data for the Lake Utaho shoreline, Pocatello Valley, Idaho*

Reference elevation[1] (ft a.s.l.)	Shoreline elevation point[2]	Height of shoreline above reference elevation[3] (ft)	Calculated shoreline elevation[4] (ft)	Colluvial wedge correction[5] (ft)	Corrected shoreline elevation[6] (ft a.s.l.)
5,032 (a)	1	110	5,142	−3	5,139
5.022 (b)	2	111	5,133	+3	5,136
	3	105	5,127	+2	5,129
	4	113	5,135	−2	5,133
4,977 (c)	5	157	5,134	0	5,134
	6	161	5,138	−3	5,135
	7	168	5,145	−3	5,142
	8	166	5,143	−2	5,141
	9	182	5,159	−3	5,156
	10	164	5,141	0	5,141
	11	174	5,151	0	5,151
	12	176	5,153	−3	5,150
	13	172	5,149	0	5,149
4,994 (d)	14	148.5	5,142.5	−1.5	5,141
	15	149	5,143	−1.5	5,141.5
	16	146.5	5,140.5	−4	5,136.5
5,063 (e)	17	88	5,151	−2	5,149
	18	98.5	5,159.5	−13	5,148.5
	19	96.5	5,159.5	−10	5,149.5
5,006 (f)	20	127	5,133	0	5,133
4,997 (g)	21	160	5,157	+9.5	5,166.5

[1]Reference points of known elevation where the surveying base station was located: a, SW1/4 sec. 16, T. 15 S., R. 34 E.; b, benchmark O'FNL, sec. 26, T. 15 S., R. 34 E.; c, SW1/4 sec. 1, T. 16 S., R. 34 E.; d, SW1/4 sec. 13, T. 16 S., R. 34 E.; e, SW1/4 sec. 30, T. 16 S., R. 34 E.; f, benchmark SE1/4 sec. 20, T. 15 S., R. 34 E.; g, NW1/4 sec. 17, T. 16 S., R. 34 E.

[2]Shoreline points located on figures 4 and 9.

[3]Calculated by trigonometric leveling from reference elevation; precision 0.5 ft.

[4]Calculated by adding the third column to the first column.

[5]The elevation difference between the projected shoreline angle (see fig. 8) and the point where the shoreline elevation was measured; precision 1.0 ft.

[6]Calculated by adding or subtracting the fifth column from the fourth column. This elevation should be within 1.6 ft of the mean elevation of the formative water plane (Rose, 1981, table 5.6).

Pocatello Valley. The closed topography of the valley suggests that subsidence has been the dominant long-term tectonic trend during the Quaternary, as Harr and Mabey (1976) previously suggested for the late Cenozoic.

WESTERN MARGIN FAULT

Over 30 normal faults have been mapped by Allmendinger (1983) in the portion of the North Hansel Mountains in our study area. The longest fault strikes N. 20° W., is 8 km long (fig. 4), and displaces older Quaternary pediment gravels by at least 55 m down to the east. The S value for the southern portion of the fault is 2.33, indicative of moderate to slightly active tectonism (Bull and McFadden, 1977). The V_f ratios of two gullies along the western margin fault are 5.0 and 4.0, values which are approximately 20 times those calculated for the eastern margin fault. The difference in V_f values suggests that the western margin fault is less active than the eastern margin fault.

NEOTECTONIC FEATURES IN HANSEL VALLEY

TECTONIC SETTING

Hansel Valley, Utah, lies southeast of Pocatello Valley and is bounded on the east by the North Promontory Mountains and on the west by the Hansel (or Summer Ranch) Mountains (fig. 2). The fault at the western base of the North Promontory Mountains has created an extremely linear north-trending range front, complete with faceted spurs, about 25 km long. The eastward tilt of Oquirrh Formation strata on either side of the fault near Bull's Pass suggests eastward rotation of fault blocks (Jordan, 1985). The inferred Cenozoic throw on this fault near Bull's Pass (425 m) (Jordan, 1985, section *A–A'*) is considerably less than the 1,420 m inferred for the eastern fault of Pocatello Valley. The linear range front at the eastern base of the Hansel Mountains is inferred to be underlain by a normal fault (Adams, 1962, pl. 1; Doelling, 1980, pls. 1, 3, section *G–G'*). However, the fault does not offset Quaternary deposits and has not

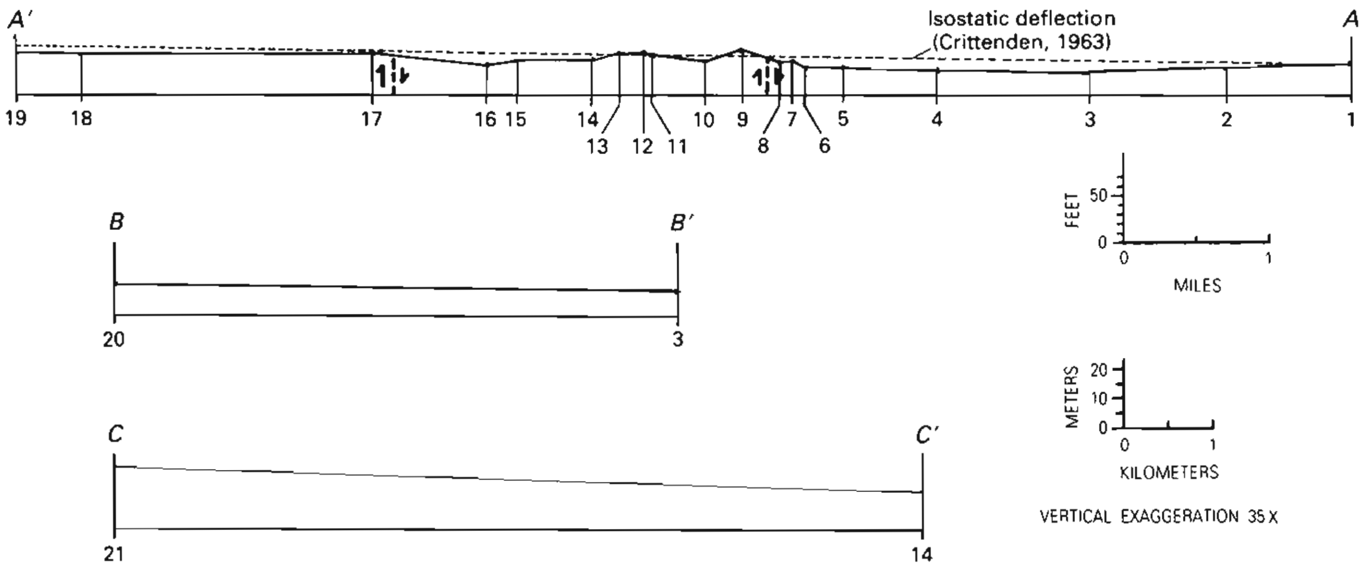

FIGURE 9.—Lake Utaho shoreline elevations. Locations of profiles are shown in figure 4. The amount of error owing to instrument inaccuracies and projecting of the shoreline angle is approximately represented by the width of the profile line (1.4 ft). On profile *A–A'*, significant deviations from the expected elevation occur where the shoreline crosses inferred branches of the range-front fault (points 17-16, points 9–6). Profiles *B–B'* and *C–C'* show slight tilting of shorelines toward the eastern range-bounding fault.

created the sharp faceted spurs that exist on the eastern valley margin. Outcrops of Oquirrh Formation on the valley floor suggest that the throw on this fault is no greater than that on the eastern margin fault; it may be a minor fault on a hinged valley side, similar to the western margin fault in Pocatello Valley. The southwestern margin of Hansel Valley is a subdued and irregular range front veneered by Lake Bonneville deposits. The east-facing 1934 fault scarp and a prehistoric scarp up to 9 m high trend northerly across this zone. Very low gravity gradients in this region (fig. 3) and the lack of appreciable topographic relief suggest that this fault has a low Cenozoic slip rate. Tertiary and early Pleistocene sediments exposed in arroyos on the valley floor are extensively faulted. These faults usually have displacements of less than 5 m and do not displace overlying Lake Bonneville bottom sediments. The segmented linear nature of arroyos in northern Hansel Valley may be controlled by these north- to northeast-trending faults in underlying Tertiary sandstones (Salt Lake Formation). Normal faults of similar strike offset Tertiary basalts immediately north of Interstate 84 and tilt them into a series of east-dipping blocks (Adams, 1962).

EASTERN MARGIN FAULT

Fault scarps occur at only two locations along the eastern margin fault, where the trace is not covered by Holocene talus. At the northern location (A, fig. 2), a 13-m-high scarp offsets a delta graded to the Bonneville shoreline. Farther south, a branch fault (B, fig. 2) diverges southwesterly from the range front and creates a scarp up to 12.9 m high across pre-Bonneville alluvial fans. Both scarps are probably the result of multiple events; because each is limited to a single geomorphic surface and displays no multiple crests, reliable evidence of recurrent movement is lacking. Two profiles across the northern scarp and three across the southern scarp yield scarp surface offsets and maximum scarp slope angles that can be compared with those of dated single-event scarps elsewhere in Utah (fig. 10). The sparse data suggest that the scarps are roughly comparable in age to the Bonneville shoreline (15.5–17 ka) (Scott and others, 1983).

A graben bounded by two high-angle faults in Quaternary deposits is exposed on the southern roadcut of Interstate 84 across Rattlesnake Pass (C, fig. 2). The faults offset seven distinct units of locally derived alluvium and colluvium deposited in a broad swale cut into Tertiary basalts (fig. 11). Of the four paleosols within the sequence, the lowest one is better developed than 140-ka-old noncalcareous soils developed elsewhere in the Bonneville Basin (Scott and other, 1982, p. 42), whereas soils on units II, IV, and VI plus VIII each resemble calcareous soils developed elsewhere on 17-ka deposits (Scott and others, 1982, p. 36). Units I through V are clearly faulted; unit VI may be faulted, whereas soil horizon K1 is clearly superposed across the fault, an indication that faulting occurred between the deposition of unit V and the formation of soil K1 (latest Pleistocene–early Holocene?).

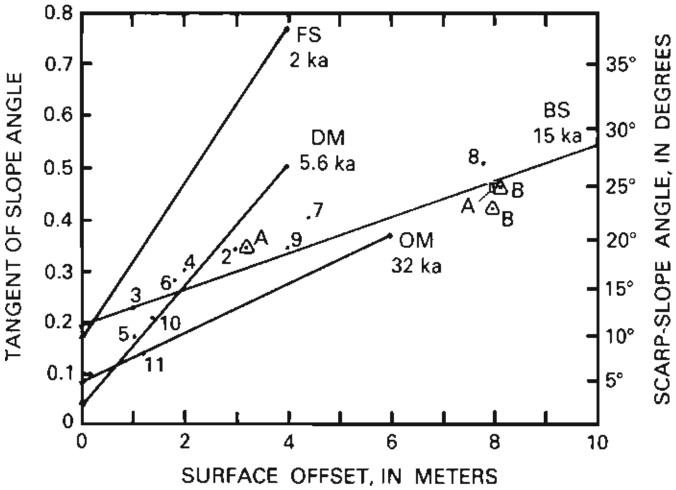

FIGURE 10.—Scarp surface offset plotted against the tangent of the maximum scarp slope angle for Hansel Valley fault scarps and other dated scarps within Utah (modified from Hanks and others, 1984). Solid lines show the relationship between offset and slope angle for scarps at Fish Springs (FS), the Drum Mountains (DM), the Bonneville shoreline (BS, many geographic locations), and the Oquirrh Mountains (OM). Hansel Valley data are shown by open triangles marked B (from scarp area B in fig. 2), an open square marked A (from scarp area A in fig. 2), and numbered solid circles (from the southwestern margin scarp, marked D on fig. 2). Numbers refer to profile locations shown in figure 12. Comparison of Hansel Valley scarp data with other Utah data suggests that (1) the scarp at location A is roughly contemporaneous with the Bonneville shoreline (because this scarp offsets a delta graded to the Bonneville shoreline, faulting must have occurred immediately after construction of the delta); (2) scarps at location B are older than the Bonneville shoreline, which is consistent with their displacing pre-Bonneville alluvial fans above the Bonneville shoreline; and (3) the scarp at location D is slightly younger than the Bonneville shoreline (this age is consistent with age estimates based on geomorphic data (fig. 14) that faulting occurred during the regression from the Provo shoreline before about 12 ka).

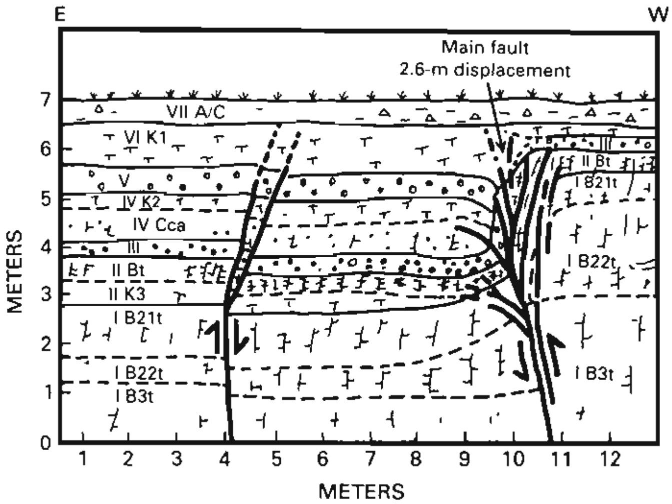

FIGURE 11.—Fault exposed on the southern side of the Rattlesnake Pass (Interstate 84) roadcut (C, fig. 2). Four paleosols are exposed in the eastern two-thirds of the cut, developed on parent materials of unit I (horizons B21t, B22t, and B3t), unit II (horizons Bt and K3), unit IV (horizons K2 and Cca), and unit VII (horizon A/C) plus unit VI (horizon K1). The main fault displaces units I through III by 2.6 m. Units IV and V have evidently been eroded from the upthrown block west of the main fault, and unit VI is considerably thinner on the upthrown block, probably the result of scarp relief. Unit VIII (Holocene? colluvium) is not faulted. Pattern intensity indicates the degree of secondary carbonate or clay enrichment in soil horizons.

NORTHWESTERN MARGIN FAULT

The northwestern margin fault flanks the eastern margin on the Hansel Mountains (fig. 2). The range front is linear and exhibits subdued talus-covered faceted spurs. Bonneville-cycle shoreline deposits abut the range front, but no evidence of surface faulting exists in these sediments. Any fault scarps that may have been formed before the transgression of Lake Bonneville were presumably completely eroded. A shorter fault, about 4 km in length, parallels the northwestern margin fault along the east side of a small valley at the southern end of the Hansel Mountains (fig. 2) (Adams, 1962; Doelling, 1980). This fault forms a horst at the southern end of the mountain and decreases in displacement toward the north. This smaller fault is aligned with the trace of the 1934 scarp and may be an extension of that fault zone.

SOUTHWESTERN MARGIN FAULT

The fault on the southwestern margin of Hansel Valley is markedly different from the valley-bounding faults previously described. Here the mountain block is located 3 to 4 km west of an 8-km-long east-facing fault scarp offsetting Bonneville-cycle shoreline gravels. Eleven profiles across the scarp (fig. 12) reveal an increase in scarp height from 1.7 m to 9.0 m along the central part of the scarp, where it ascends a 60-m-high bedrock-cored escarpment. The increased scarp heights probably result from the addition of a fault scarp plus a landslide scarp produced by syntectonic sliding (fig. 13), as the arcuate plan shape of the scarp suggests. Because the material involved in the sliding (beach gravel) is not particularly susceptible to landsliding when it is unsaturated, we infer that pore pressures were very high at the time that faulting and sliding occurred, either from ground-water saturation or from submersion under Lake Bonneville.

Geomorphic relations (described below) between the fault scarp and the shorelines indicate that the scarp must have formed after the Bonneville transgression rose above 1,402 m (about 26 ka) (Scott and others, 1983, fig. 5) but before the post-Provo regression fell below 1,326 m (about 12 ka). Our reasoning is as follows. During a pluvial lake cycle, surface-faulting events can

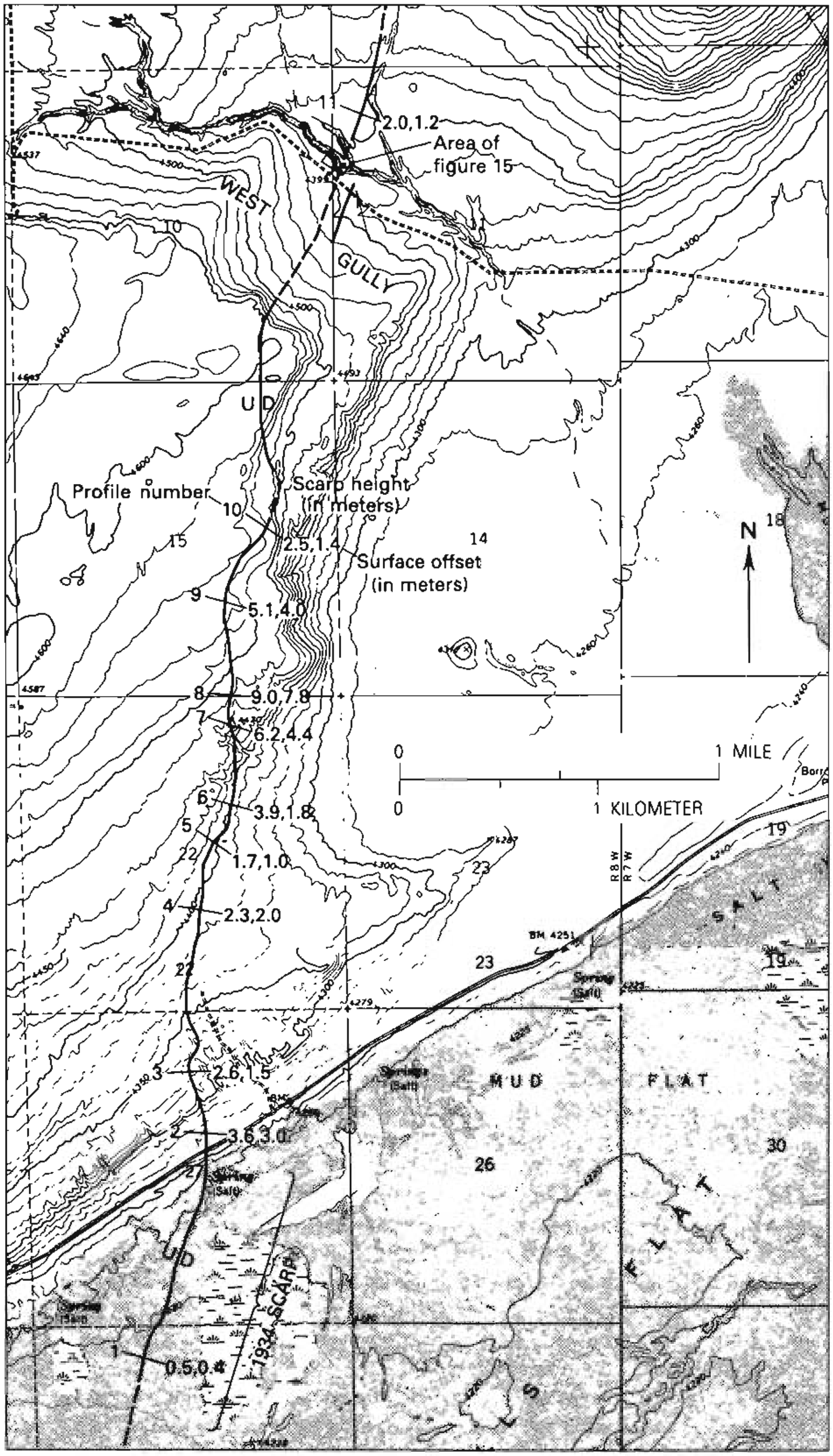

FIGURE 12.—Profile locations, scarp heights, and surface offsets on the Hansel Valley southwestern margin fault scarp (scarp D, fig. 2). The 1934 scarp appears on the southern end of the prehistoric scarp. Scarp heights increase as the scarp ascends a larger escarpment; the arcuate scarp plan shape strongly indicates syntectonic slumping (fig. 13). If slump effects are disregarded, most surface offsets are in the 1.0- to 2.0-m range.

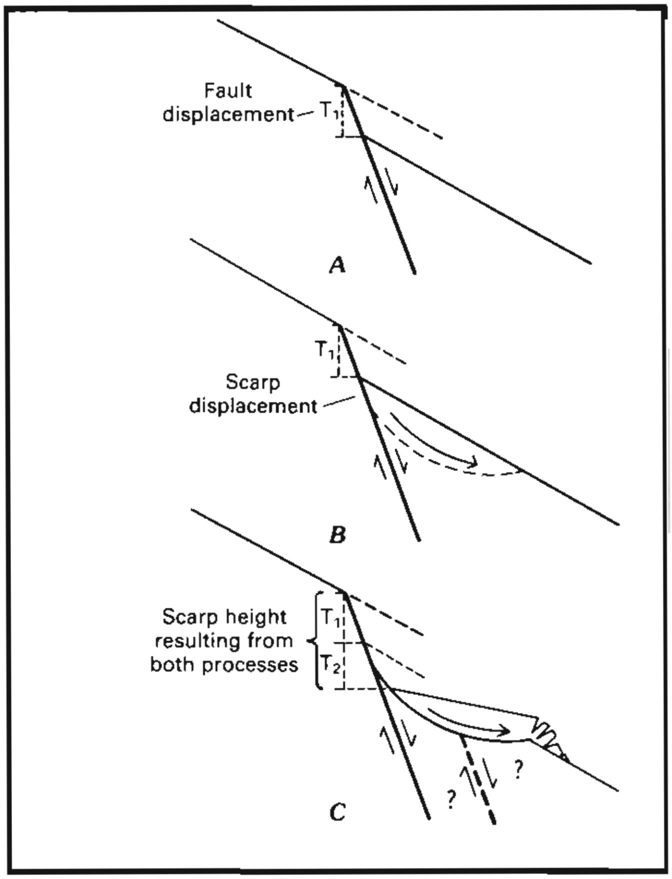

FIGURE 13.—Conceptual model showing how fault scarps on the southwestern margin might be heightened by a coseismic rotational slump. *A*, Initial fault displacement creates a scarp of height T_1. *B*, During seismic shaking, a rotational slump begins to form on the downthrown block. Slumping would be likely if surface materials were saturated (see text). *C*, After slumping, the total scarp height includes components from fault displacement (T_1) and from the slump headscarp (T_2). The dashed and queried fault in *C* indicates that the causative fault may actually underlie the slump, in which case the scarp preserved today is entirely of slump origin. Because slumps were large and slump toes indistinct, profiles used to calculate surface offsets in figure 12 were not extended downslope to include the entire slump deposit.

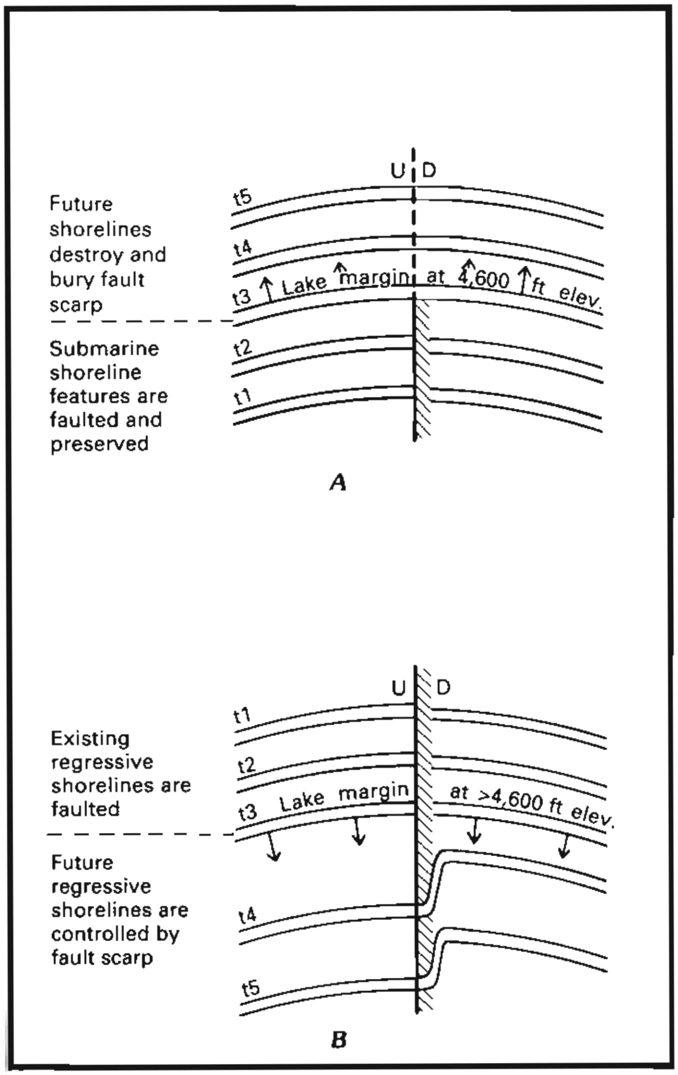

FIGURE 14.—Geomorphic setting of a fault scarp formed during a lacustrine transgression (*A*) and regression (*B*). Time of shoreline occupation is shown by t and progresses from t1 to t5. In both instances, hypothetical faulting occurs when the shoreline is at t3.

occur (1) before the lake rose, (2) during the transgression, (3) during the lake highstand, (4) during the regression, or (5) after the lake had completely receded. Small fault scarps (1–2 m high) that formed before a major transgression would be reworked and destroyed, judging from the thickness (>5 m) and coarseness of Bonneville transgressive bars. Surface faulting during a transgression would displace submerged shorelines already formed; the scarp might be preserved if it were below the effective wave base. That part of the scarp above the shoreline would be destroyed during the continued transgression. A scarp formed during a transgression therefore should displace all shorelines below the one occupied at the time of faulting and terminate at that shoreline, being obscured at higher elevations by wave reworking and longshore deposition (fig. 14*A*). Any fault scarp formed during the lake highstand would displace all transgressive shorelines that it intersected. If faulting occurred during a regression, small regressive shorelines already formed as the lake receded should be displaced. The position of later (lower elevation) regressive shorelines would be controlled by the position of the fault scarp, because only minor deposition (less than 1 m) accompanies regression (except for the Provo shoreline) in the Bonneville Basin. In this case, a scarp would displace regressive shorelines above a certain elevation; below that elevation, shorelines would wrap around or abut a modified scarp of similar height (fig. 14*B*). The

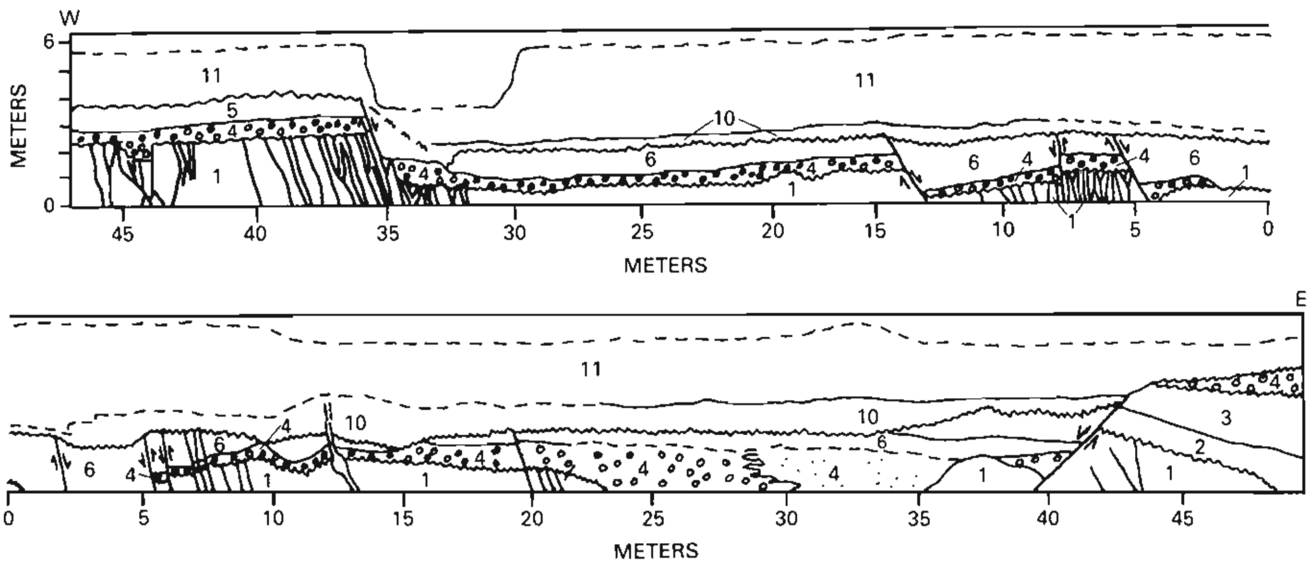

FIGURE 15.—Wall log of the West Gully showing fault patterns. Numbered units 1 through 6 refer to the stratigraphic section in figure 16; units 10 and 11 (not shown in fig. 16) are Holocene alluvium. The zero point on the horizontal scale is the section 10-section 11 boundary, which crosses the arroyo (fig. 12). The chronology of faulting deduced from this log is discussed in the text.

elevation of this transition in morphology marks the position of the shoreline at the time of surface faulting. Surface faulting after the lake had completely receded would result in displacement of all transgressive and regressive shorelines.

Most fault scarps in the Bonneville Basin trend parallel to shorelines near range fronts and therefore are not favorably oriented for intersecting multiple shorelines. The Hansel Valley scarp, however, displaces numerous shorelines as it ascends from the valley floor (1,286 m) to 1,402 m (fig. 12). The displacement of all transgressive shorelines up to 1,402 m indicates that surface faulting occurred after the Bonneville transgression had risen to this elevation (about 26 ka) (Scott and others, 1983, fig. 5). Small regressive bars at 1,326 m terminate against the scarp, an indication that the scarp was already formed by the time the post-Provo regression reached that elevation (about 12 ka). Therefore, the scarp must have formed after 26 ka but before 12 ka. Scarp surface offset versus tangent of slope angle data suggest that the latest surface-rupture event occurred between roughly 15 and 5.6 ka (closer to the older date) (fig. 10), although scarps formed underwater are not strictly comparable to subaerial scarps. The overlap between these age estimates (12–15 ka) probably closely dates the latest faulting event. If these dates are correct, then the entire southwestern margin scarp was produced by subaqueous displacement of the lake floor.

Near its northern end, the prehistoric scarp is intersected by a 10- to 15-m-deep arroyo (informally termed the West Gully) (figs. 2, 12). Two deformation zones are well exposed on arroyo walls: (1) the main fault zone, consisting of 11 normal faults defining a 90-m-wide graben (fig. 15), and (2) a subsidiary fault zone approximately 100 m to the east, composed of three major normal faults. Detailed logging and sampling of 200 m of arroyo walls reveal that the faults offset a stratigraphic section 19 m thick composed of nine lacustrine units (fig. 16) from three lake cycles. Physical stratigraphy, sedimentology, ostracode assemblages, and thermoluminescence (TL) dating (described by McCalpin, 1986) indicate that deposits of two pre- Bonneville ages are present. The older deposit (unit 1, fig. 16) appears discontinuously in arroyo bottoms, consists of very compact, laminated silt and clay beds, and is disrupted by numerous high-angle fractures and faults of small displacement. Ostracode fauna and a TL date of 138 ka suggest that this lower unit is correlative with the Little Valley deposits described by Scott and others (1983). Unconformably overlying this unit are shoreline gravels and lake-marginal marsh sediments (units 2 and 3), which are also faulted (40–45 m E., fig. 15). TL dates of 76 and 82 ka indicate that these units were deposited during a previously undated lake cycle (informally termed the Hansel Valley cycle) in late oxygen isotope stage 5 or early stage 4 (Morley and Hays, 1981). A lake cycle of similar inferred age has been postulated recently by Oviatt and others (1987). Deposits of the Bonneville cycle (units 4–9) include transgressive gravel and lake-bottom silts and clays. Lake-bottom beds (unit 6) approximately 1.5 m above the transgressive gravel are dated at 14 ka by amino acids and at 15.5 ka by ^{14}C on gastropod shells. Strata 4 m below the top of the stratigraphic section (upper unit 8) are dated at 13 ka by TL. These dates are roughly correlative with the Bonneville-cycle lake elevation curve of Scott and others (1983, fig. 5).

Although they are not shown on figure 15, faults extend through unit 6 but cannot be traced through units 7 or 8. Both units 6 and 8 are internally deformed, the former by diapirs, roll structures, and convolutions and the latter by rotational faulting of intact blocks along listric faults (lateral spreading?). Although such deformation could be caused by nontectonic mechanisms, the coincidence of discrete deformed beds with a known multiple-event Quaternary fault suggests earthquake-induced liquefaction. The difference in deformation styles may be owing to great difference in water depths (and pore pressures) between the time when unit 6 was deformed (at roughly 245 m) and the time when unit 8 was deformed (at less than 60 m).

Post-Bonneville alluvium (units 10 and 11) truncates all faults exposed in the West Gully, demonstrating that no Holocene events (not even the 1934 earthquake) have induced surface rupture at this location. Although the prehistoric scarp cannot be traced to the edge of the West Gully (because of agricultural disturbances), the net displacement (1.3 m, down to the east) of the transgressive gravel (unit 4) across the graben is similar to scarp heights north and south of the arroyo.

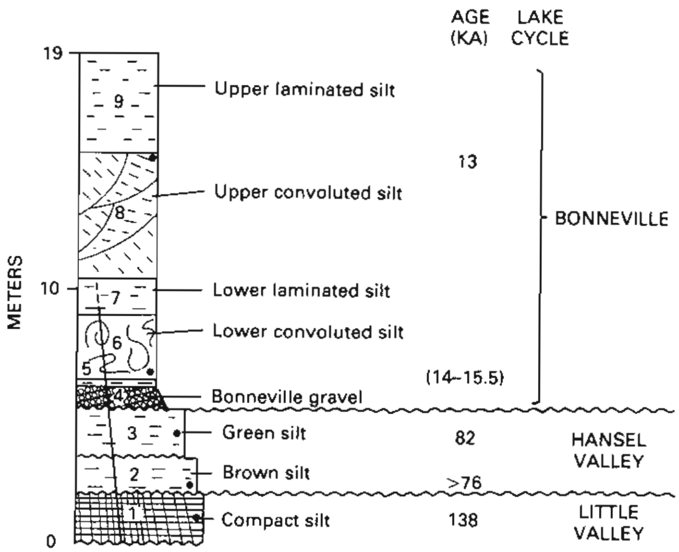

FIGURE 16.—Quaternary deposits exposed in the West Gully. Numbers correspond to those in figure 15, except that post-Bonneville alluvium (units 10 and 11) is not shown. Ages are from thermoluminescence dates on bulk sediment (McCalpin, 1986) or from ^{14}C dates on gastropod shells (in parentheses). Sample locations are shown by solid circles.

LATE QUATERNARY TECTONIC HISTORY

POCATELLO VALLEY

Three lines of indirect evidence suggest a small but measurable amount of late Quaternary displacement on the eastern margin of Pocatello Valley: (1) lack of fault scarps in 15-ka deposits, (2) rapid elevation changes (up to 6.4 m) of a 15-ka shoreline, and (3) tilted but unfaulted Quaternary colluvium at the range front. Shoreline elevation data further suggest that deformation has been dominated by valley subsidence along the 7-km-long central segment of the eastern margin fault. Compilations by Bonilla (1982) and Slemmons (1982) showed that the threshold of surface rupture for shallow-focus earthquakes on normal faults is roughly M_L 6.2 to 6.3. Therefore, it appears that no earthquakes of this magnitude or larger have occurred anywhere along the range-front fault in the last 15 ka. Within 20 m of the inferred range-front fault, the absence of tectonic displacements disrupting 95-ka colluvium suggests that no earthquakes of $M_L>6.2$ to 6.3 have occurred since that time. If this 7-km-long fault segment is capable of generating surface-faulting earthquakes, then recurrence times between them must be at least 15 ka and possibly as much as 95 ka.

The cause of the rapid 4.5- to 6.4-m vertical changes in shoreline elevation is not fully understood. The magnitude of change appears to be significantly greater than that of the combined surveying errors, whereas the locations of greatest change coincide with shoreline-fault intersections. However, the calculated post-Utaho subsidence rate (4.5 m in 15 ka, or 0.3 mm/yr) seems excessive in light of the overall structural relief and age of the valley. At that rate, the 1,420-m inferred displacement across the fault could have been created in only 475 ka. The presence of Salt Lake Formation (Miocene-Pliocene) within the downthrown block shows that inception of block faulting must be at least Miocene, so the inferred late Quaternary rate must be much higher than the post-Miocene average. This inference is compatible with the observed high rates of historical seismicity.

The 1975 M_L 6.0 earthquake we tentatively assign to the western margin fault, on the basis of epicentral location and focal mechanisms derived by Arabasz and McKee (1979). Like the eastern margin fault, this fault does not displace any Lake Utaho deposits, an indication that no $M_L>6.2$ to 6.3 events have occurred in the last 15 ka. An early(?) Quaternary pediment (500 ka?) is displaced at least 55 m down to the east, yielding a slip rate of 0.11 mm/yr. This rate is roughly one-third of the slip rate calculated for the eastern margin fault, compatible with the differences in their tectonic geomorphology values quoted earlier.

HANSEL VALLEY

Paleoseismic recurrence and magnitude data from the eastern margin fault are scanty and somewhat contradictory. The 8-m surface offset of the scarp across a 15-ka delta implies multiple ruptures (three to four if displacements per event were 2.0–2.5 m) in the last 15 ka. The resulting recurrence intervals (3.7–5.0 ka) seem too short in view of the total range-front relief and tectonic geomorphology in comparison with those of nearby normal faults such as the Wasatch fault (Machette and others, this volume). Evidence of only a single recent 2.6-m displacement on the Rattlesnake Pass splay fault within the last 100 ka is not compatible with three to four events on the main trace in 15 ka, unless the main fault has ruptured many times without involving this particular splay. However, multiple post-Bonneville displacements are seen nowhere else along the length of the eastern margin, although the inferred fault trace emerges from under its talus cover for several long stretches. Only a single branch fault farther south offsets pre-Bonneville (oxygen isotope stage 4 or 6) alluvial fans 9.5 m. Depending on the age assigned to these fans (roughly 65 or 140 ka), the recurrence intervals for 2.0- to 2.5-m events are between 16 and 35 ka, although not every main trace rupture may have involved this branch. Given the poor geomorphic evidence, all that can be stated with certainty is that this fault has sustained surface rupture at least once since Bonneville time and several times since either oxygen isotope stage 4 or 6.

Although temporal control is better for the well-exposed faulted strata in the West Gully, the inability to trace any subsurface faults to the surface (and thus connect them to the surface scarp) poses a large obstacle to fault-history reconstruction. The older history of faulting is ambiguous; multiple events of unknown displacement occurred along many traces, some of which were truncated by the oxygen isotope stage 4 transgression about 72 ka. On blocks rotated by post-26-ka faulting, the lack of angular unconformity between oxygen isotope stage 6 and 4 deposits indicates that no tectonic rotation occurred there from 72 to 58 ka (35 m E., fig. 15). The 26-ka Bonneville transgression truncates many small-displacement fractures of unknown age in unit 1 but also fills in an open tectonic fissure (42 m W., fig. 15) that could not have remained open for too long after its creation; this fact implies an event not far in advance of 26 ka. Although units 4, 5, and 6 are displaced along major faults, these faults cannot be traced through overlying units owing to poor exposures. The only later evidence of tectonism is the rotated blocks of unit 8, which could be a result of either tectonic or nontectonic lateral spreading. Taken together, the evidence argues for multiple events between roughly 140 and 72 ka, no events from 72 to 58 ka, at least one event between 58 and 26 ka (nearer the latter), an event around 14 to 15 ka (the age of unit 6) and possibly another at 13 ka (unit 8), and a M_L 6.6 event in 1934.

This history implies widely varying recurrence intervals (RI's), from one event from 72 to 26 ka (RI=46 ka) to two events at 14 to 15 ka and 13 ka (RI=1–2 ka). Although the field evidence is ambiguous, it appears that recurrence intervals have been shorter when large lakes existed in the Bonneville Basin (oxygen isotope stages 6 and 2) than they have been during times of small or no lakes (oxygen isotope stages 5, 4, 3). An exception is the 1934 M_L 6.6 earthquake, which occurred in the middle of an interpluvial episode. However, its maximum displacement (0.5 m) is small in comparison with the displacements seen in stage 2 time (2.2–2.6 m, in either one or two events) and may indicate that interpluvial earthquakes are smaller than pluvial earthquakes. Crustal stresses imposed by the rapid filling and draining of pluvial lakes (especially the catastrophic drop from the Bonneville shoreline to the Provo shoreline) may be the main triggering mechanism for the Quaternary surface-faulting events experienced by the southwestern margin fault, as others (Swan and others, 1983; Machette and others, this volume) have suggested for the nearby Wasatch fault.

CONCLUSIONS

Quaternary geologic mapping, shoreline surveying, arroyo wall logging, and trenching indicate that late Quaternary displacements have occurred on both margin faults in Pocatello and Hansel Valleys. The lack of fault scarps in pluvial lake deposits in Pocatello Valley suggests that no surface faulting has occurred in the last 15 ka, yet shorelines and colluvium at the range front appear warped. The 6.4-m maximum vertical shoreline warping may be caused by monoclinal flexure over a buried fault trace; examples of similar warping of surficial deposits near faults have been given by Bonilla (1982, p. 18), although none are for normal faults. Colluvium in a range-front trench appeared warped but unfractured, suggestive of creep flexure rather than flexure around an upward-propagating rupture. Recent trenching on the West Valley fault zone in Salt Lake County, Utah, by Keaton and others (1986) has exposed monoclinal warps along Quaternary normal faults. Given the limited data on hand, it appears that surface flexure could arise from creep or from earthquakes occurring slightly below the threshold magnitude necessary to create a fault scarp. Geomorphic evidence from Pocatello Valley suggests that significant vertical displacements may occur without the creation of fault scarps, over horizontal distances of hundreds of meters, and that these displacements are not detectable in the field without careful surveying. Such a conclusion has implications for displacement studies on the Wasatch fault, where it has been assumed that most (if not all) cumulative net vertical tectonic displacement can be measured solely by fault-scarp profiling.

Marginal faults bounding Hansel Valley have generated multiple surface-faulting earthquakes in the late Quaternary, despite the seemingly small structural relief of this valley in comparison with that of the Pocatello Valley. Although a few short scarps occur on the steep eastern margin, the best record of activity is on the southwestern margin, where an 8-km-long scarp is intersected by a 15-m-deep arroyo. The multiple events deduced from logging a 90-m-wide complex graben that displaces lacustrine sediments 138 to 13 ka old show that even faults exhibiting minimal topographic and structural expression can have complex recurrence histories. TL dating of silty lacustrine deposits in the graben yielded ages compatible with stratigraphic and paleoecologic evidence and showed that this technique holds promise for dating noncarbonaceous, nonfossiliferous lake beds. Finally, the concentration of deformation in the oldest lacustrine deposit (oxygen isotope stage 6) and in the deposits of the Bonneville cycle (oxygen isotope stage 2) suggests that surface faulting was enhanced when deep pluvial lakes occupied Hansel Valley, as opposed to when no lakes (oxygen isotope stages 3 and 5) or shallow lakes (oxygen isotope stage 4) existed. This temporal clustering has been described by others (Swan and others, 1983) and may result from increased pore pressures and water loading of the crust.

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