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Pocatello Valley Area, Idaho-Utah

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QUATERNARY GEOLOGY AND TECTONIC GEOMORPHOLOGY
OF THE POCATELLO VALLEY AREA, IDAHO-UTAH

by

John D. Garr

A thesis submitted in partial fulfillment
of the requirements for the degree
of
MASTER OF SCIENCE
in
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1988
ABSTRACT

Quaternary Geology and Tectonic Geomorphology of the Pocatello Valley Area, Idaho-Utah.

by

John D. Garr, Master of Science
Utah State University, 1988

Major Professor: Dr. James P. McCalpin
Department: Geology

Pocatello Valley in southeastern Idaho and northern Utah is a structural and topographic basin bounded on all sides by mountains composed of Paleozoic platform carbonates and clastics. In the late Pleistocene it contained pluvial Lake Utaho, which, prior to 1981, was considered to have been an arm of Lake Bonneville. This study corroborates the finding of Currey (1981) that the two lakes were separate.

The Quaternary deposits examined in this study are divided into two broad groups: those that were deposited prior to the last pluvial lake cycle, and those that were deposited during and after the pluvial lake maximum, (since approximately 16 ka) when the area was occupied by
Lake Utaho and Lake Bonneville. Pediment gravels, alluvial fans, piedmont colluvium, and talus comprise the older group; the younger deposits include stream channel deposits, lacustrine sediments, and loess.

Quantitative geomorphological techniques (mountain front sinuosity ratios and valley floor width-valley height ratios) indicate that the bounding ranges on the east and west margins of Pocatello Valley are slightly to moderately active tectonically. Precise surveying of the Lake Utaho highstand shoreline revealed significant deviations from the smooth, isostatically rebounded shoreline elevation curve of Crittenden (1963). The greatest deflections occur where the sinuous shoreline crosses the more linear inferred range front faults at the base of Samaria Mountain, on the east margin of the valley. The deflections (as much as 6.4 m over a horizontal distance of 900 m) suggest that movement has occurred along the range front faults since the shorelines were created approximately 16 ka, but no fault scarps were formed. A buried colluvium estimated to be 95 ka ± 15 ka that was exposed in a trench at the range front is monoclinally draped over the inferred fault, and dips as much as 49° toward the valley. There are no fractures in the colluvium, which suggests that, although relative movement between the mountain and valley blocks has occurred, the displacement has only warped the colluvium. This further
suggests that any earthquakes accompanying the movements must have been below the magnitude threshold (M_L 6.2-6.3) necessary for surface rupture.
INTRODUCTION

General Statement

This report summarizes an investigation of the surficial deposits and Quaternary tectonics of the Pocatello Valley area, Idaho-Utah. The area was chosen for study because: 1) its surficial deposits had not previously been mapped in detail; 2) Pocatello Valley is a closed depression; 3) it contained one of the northernmost Pleistocene pluvial lakes associated with Lake Bonneville; and 4) it was the site of a damaging magnitude 6.0 earthquake in March, 1975.

Purpose of Investigation

This study had three primary objectives: 1) to investigate and map the surficial deposits of Pocatello Valley and determine the general Quaternary geologic history of the area; 2) to locate and map Quaternary-age faults, determine their geometries and rates of activity, and evaluate the effect of Quaternary tectonism on the geomorphology of the valley and its surrounding hills and mountains; and 3) to compare the Quaternary and historic records of seismicity in the Pocatello Valley area in an attempt to test the validity of earthquake recurrence
rates derived through the extrapolation of historic seismicity data alone.

Evidence of individual, large magnitude earthquake faulting events could not be found in the subsurface deposits examined. Moreover, material suitable for absolute-age dating by radiocarbon or amino acid techniques was either absent or present in only minuscule amounts. For these reasons, the third objective was not accomplished.

Previous Investigations

Many early explorers visited the Bonneville Basin and recognized the shoreline features that suggest the area was once occupied by a large ancient lake (Fremont, 1845; Gunnison, 1856; Simpson, 1859). The most comprehensive study of the Lake Bonneville region was conducted by G. K. Gilbert between 1873 and 1883, and resulted in the classic first monograph of the United States Geological Survey (Gilbert, 1890).

Although Gilbert studied Lake Bonneville shoreline features in the Curlew and Malad valleys (west and east, respectively, of Pocatello Valley), no mention is made in his field notes of his visiting Pocatello Valley (Hunt, 1982). One of his assistants, Gilbert Thompson, traversed the length of Pocatello Valley in the fall of 1880, however, and mapped a shoreline there that he believed to
have been produced by a bay of Lake Bonneville (Gilbert, 1890).

Early in the shoreline surveying phase of the present study, it was determined that the Pleistocene lake which occupied Pocatello Valley was not a bay of Lake Bonneville, as thought by Gilbert and subsequent investigators, but was a lake unto itself, separated from the main body of Lake Bonneville by a loess-mantled ridge at the south end of what is locally known as "Little Pocatello Valley." Several weeks after our "discovery" it was learned that Currey (1981) had already determined the two lakes were separate, and had named the "new" former body of water "Lake Utaho."

The bedrock geology of the central Blue Springs Hills (in the northern portion of the present study area) and the Paleozoic stratigraphy of Samaria Mountain (along the east margin of Pocatello Valley) have been studied by Beus (1963; 1968). Low-angle faults and the Paleozoic stratigraphy of Pocatello Valley area have been investigated by Allmendinger and Platt (1983), who have also produced geologic maps of the area (Platt, 1977; and Allmendinger, 1983).

Interest in Pocatello Valley was renewed following the March 28, 1975 (UTC) magnitude 6.0 Utah-Idaho Border earthquake. The earthquake had its epicenter in the west-central part of Pocatello Valley, and was the strongest
earthquake in the United States since the 1971 San Fernando Valley, California, earthquake. Immediately following the Utah-Idaho Border event, the area was visited by personnel from the Utah Geological and Mineral Survey, the University of Utah Seismograph Stations, and the United States Geological Survey. Reports on the effects of the earthquake (snowcracks, damage to structures, and other phenomena) were soon published (Platt, 1975, 1976; Rogers and others, 1975; Kaliser, 1975, 1976; Cook and Nye, 1975, 1979). The seismology of the earthquake was studied by Arabasz and others (1975, 1979). Stover and others (1976) used seismic data from the event in a study of earthquake intensities and attenuation. USGS benchmarks first surveyed in the early 1960s were re-leveled after the earthquake by Bucknam (1976). A gravity survey of the area was conducted by Harr and Mabey (1976).
PHYSIOGRAPHIC AND GEOLOGIC SETTING

Location and Accessibility

Pocatello Valley lies astride the Utah-Idaho border, approximately midway between the towns of Malad, Idaho, and Snowville, Utah. The study area, which includes all of the Pocatello Valley floor and parts of the surrounding mountains and hills, is located between latitudes 41° 57' 18" and 42° 08' 16" north and longitudes 112° 23' 00" and 112° 34' 00" west, and covers approximately 310 square kilometers (120 square miles) (Figure 1). Elevations range from 2308 m (7695 ft) at the top of Samaria Mountain to 1490 m (4968 ft) at the lowest point on the valley floor.

The area is easily reached by driving west from Tremonton, Utah, on Interstate Highway 84 for 31 km (19 miles), then turning north at Exit 24. From there, a paved county road leads 16 km (10 miles) north into the southeast corner of Pocatello Valley.

The entire valley floor is served by a grid of unimproved dirt and gravel roads built along section lines. Access to the foothills and canyons above the valley floor is limited to a few rough dirt roads, most of which are not suited to passenger car travel. Because of the high
Figure 1. Index map showing the location of the study area.
clay content of the soils in the valley, all but the gravel-surfaced roads become impassable during rain or snow storms.

Climate, Vegetation, and Soils

The Pocatello Valley area has a semi-arid, continental climate. Summers are hot and dry, with rather cool nights. Daytime summer temperatures rarely exceed 38°C (100°F). Winter temperatures seldom drop below -18°C (0°F). Much of the area's precipitation falls as winter snow and spring rain. The average annual precipitation ranges from 36 cm (14 in.) on the valley floor to more than 51 cm (20 in.) on the highest mountains (Chadwick and others, 1975).

On the few non-cultivated areas on the valley floor, native vegetation is comprised of bluebunch wheatgrass, Sandberg wheatgrass, big sagebrush, cheatgrass, yellowbrush, phlox, balsamroot, and annual grasses. The foothills and mountains support the native plants of the valley floor, as well as slender wheatgrass, yarrow, Great Basin wildrye, serviceberry, and snowberry (Chadwick and others, 1975). Sparse stands of Utah juniper occupy on some of the south-facing slopes of Samaria Mountain.

The soils on the central part of the valley floor are Aridisols formed on the most recent lake bottom sediments (Chugg and others, 1968). All the remaining soils in the
Geologic Setting

The Pocatello Valley area lies near the eastern margin of the Basin and Range Province, the earth's largest intracontinental rift system (Allmendinger, 1983). The regional topography and structure is dominated by north-south trending block uplifts of late Cenozoic age, separated by structural basins, either graben or "half graben" (Stewart, 1971; Hintze, 1973; Eaton, 1982; Allmendinger and Platt, 1983). Pocatello Valley is a structural and topographic depression surrounded by uplifts: the North Hansel Mountains to the west, Samaria Mountain to the east, the Blue Springs Hills to the north, and an unnamed, isolated mountain to the south which is bounded by low-lying ridges on both the east and west. These mountain blocks are dominantly composed of limestone, sandy limestone and sandstone of the Permian-Pennsylvanian Oquirrh Formation, and small exposures of conglomerate, tuffaceous sandstone, and diamictite of the Tertiary Salt Lake Formation.
STUDY METHODS

Field Methods

Mapping

Standard 1:28,000-scale vertical black-and-white aerial photographs were examined with a mirror stereoscope to produce a preliminary geologic map. Geologic units and geomorphic features noted on the aerial photographs were later checked in the field.

Shoreline Surveying

The elevations of Pleistocene lake shorelines were determined at 46 locations around the valley margin with a Leitz TM-10E theodolite and a Leitz RED-2 electronic distance meter (EDM) obtained from the Civil and Environmental Engineering Department at Utah State University. The surveying procedure required two people: one to operate the EDM and theodolite, the other to carry the reflecting prism and place it on the shoreline scarp. Because many of the surveying shots were greater than 1km distant, two-way radios were used for communication.

The EDM and theodolite were set up over or near a benchmark or spot elevation on the valley floor. In most instances several shoreline elevation points could be shot
from a single instrument station. After the assistant had placed the tripod-mounted reflecting prism at the base of a shoreline scarp, the instrument operator measured the horizontal and vertical angles and distance to the prism relative to the instrument station. While this was being done, the assistant measured and sketched a profile of the slope perpendicular to the shoreline scarp being measured, using a 4.5m telescoping fiberglass rod and an Abney level, a method similar to that used by Bucknam and Anderson (1979). In most instances the profile measurement was made from 30m above to 30m 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°) (Figure 2).

Because the Pleistocene lake shorelines were developed in unconsolidated deposits in all but a few locations around the valley margin, the shoreline angle has been covered 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 (Figure 2). The intersection of the two angles is assumed to be a more accurate approximation of the Pleistocene shoreline angle. Twenty-one of the 46
Figure 2. Diagram showing method used in estimating the true shoreline angle from the surveyed shoreline scarp elevation.
surveyed shorelines were profiled and corrected in this way (it was not until midway through the shoreline surveying phase that it was realized the profiling method would yield more accurate shoreline elevation data). Two valley margin trenches excavated after the shoreline surveying phase of the study provided an opportunity to check 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 highstand was horizontal, shorelines formed by the lake may have been above, at, or below the mean water plane (Currey, 1982). All of the surveyed shoreline scarps along the base of Samaria Mountain were developed on similar drift-aligned gravel beaches. Data from Rose (1981) show that the shoreline angle of such beaches is usually within 0.5 m of the mean water surface. Lake Utaho had a diameter of only 15 km at its maximum; large waves needed to deposit gravel far above the mean water plane would have been difficult to generate in such a small lake.

The precision of the theodolite used in this study is stated by the manufacturer to be ±5" of arc, or ±2.4 cm of elevation per kilometer of horizontal distance. The precision of the EDM used is stated to be ±1 mm per kilometer. The longest survey shot made in this study was
1.5 km long; most of the shots were less than 1 km long. Elevation error due to imperfect leveling of the theodolite was checked by doubling six survey shots (making a "normal" shot, then rotating the theodolite 180°, rotating the telescope 180°, making a "reversed" shot, and averaging the two values). The maximum error due to imprecise leveling was found to be +10.7 cm of elevation per kilometer of horizontal distance. The combined uncertainties of operator and instrument error (0.13 m), projected shoreline angle (0.3 m), and relation to the mean water plane (0.5 m) yield an uncertainty of +0.93 m (3.05 ft) for each surveyed shoreline point.

**Trenching**

Three trenches were excavated with a large track-mounted backhoe to look for direct or indirect evidence of Quaternary faulting at selected sites within Pocatello Valley. Trench 1 was cut across a prominent scarp at the mouth of Elevator Hollow, on the east side of the valley in the NW1/4NW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). The trench had a bearing of N75°E, was 23 m long, 0.9 m wide, and reached a maximum depth of 4.5 m. Because Trench 1 yielded no tectonic information, Trench 2 was cut perpendicular to the base of a faceted spur approximately 400 m south of Trench 1, in the NE1/4SW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). The trench was 35 m long, 0.9 m wide, 4.5 m deep at its maximum, and had a bearing of
In order to examine Lake Utaho lake bottom sediments for evidence of earthquake-induced soft sediment deformation features, Trench 3 was excavated near the center of the valley in the center of the NE1/4NE1/4 of Sec. 9, T16S R34E (Plate 1). It was 20 m long, 0.9 m wide, 4.25 m deep, and had a bearing of N90°E.

To prevent caving, ladder-type shoring was installed in each trench. Roughsawn 5 cm by 15 cm (2 in by 6 in) fir planks were laid vertically against the trench walls on 2 m centers, and were cross-braced with 10 cm by 10 cm (4 in by 4 in) fir posts every meter or so. The cross-braces also served as scaffold supports during logging of the trench walls (Figure 3).

The north and south walls of trench 1 and trench 2 were stratigraphically identical, so the straightest wall of each was selected for study. After cleaning the trench walls with a variety of scraping tools and brooms, a 1 m by 1 m horizontal and vertical grid of nylon line was laid out on each wall to facilitate mapping. The trench walls were then mapped at 1:20 scale, and the units were described and sampled.

Because of the high clay content of the lacustrine sediments exposed in Trench 3, the walls had been smeared by the backhoe shovel during excavation. To expose fresh material, the wall had to be scraped and cleaned by hand,
Figure 3. Photograph of Trench 3 showing the method used to shore trench walls.
a task both difficult and time consuming. A total of 18 hours were spent preparing a vertical column 0.5m wide and 4.5m long. Spot checks along the length of the trench showed little lateral variation in the stratigraphic sequence of the lacustrine sediments. It was decided that a single column was representative. The column was then photographed, described, and sampled.

**Laboratory Methods**

**Grain Size Analysis and Determination of Water Content**

Most of the samples taken from the trenches had a high clay mineral content and could not be oven-dried prior to grain size analysis. Oven drying would effectively bake the clays into large clumps, which would thereby distort the size analyses. Separate split samples were used for grain size analysis and for determining the water content of the samples.

Moist 50gm samples were placed in a drying oven for 24 hours at 100°C, and were then removed, cooled, and weighed. The samples were then returned to the oven for 8 hours, removed, cooled, and re-weighed to ensure that all the water had been driven off. Results are shown in Appendix A.

Fifty-gram portions of the remaining splits were wet-sieved through a 40 stainless steel screen to separate the silt and clay size fractions from the coarser grains. The
portions coarser than 40 were then dry-sieved through -4, -2, 0, and +20 screens. These grains were later examined with a binocular microscope to determine their mineralogy and organic matter content. The portions finer than 40 were dispersed in a solution of sodium hexametaphosphate and analyzed according to the pipette method described by Folk (1968). Results are shown in Appendix B.

**Acid-Insoluble Residues**

The samples previously used in the determination of water content were each completely digested with 10% HCl solution to remove carbonate minerals. The residues were then washed repeatedly, oven dried, and weighed. Results are shown in Appendix C.

**Mineralogy**

The silt and clay size portions of each sample used in the pipette size analyses were completely digested with 10% HCl solution to remove carbonate minerals. Slurries were made of the residues, which were then spread on glass plates and allowed to air-dry. Each plate was then placed in a Siemens Crystalloflex x-ray diffractometer and scanned from 2 to 45° 2θ. Peak intensities were compared with published values (Chen, 1975) to determine the mineralogy of the residues. Results are shown in Appendix D.
PRE-QUATERNARY GEOLOGY

Stratigraphy

All of the mountains within the study area are underlain by limestones, sandy limestones, and sandstones of the Permian-Pennsylvanian Oquirrh Formation deposited in the Cordilleran miogeosyncline (Beus, 1968). Platt (1977) subdivided the Oquirrh Formation of Samaria Mountain (along the east side of the present study area) into 3 units: an upper unit of calcareous sandstone with subordinate ledges of Wolfcampian fusilinid-bearing limestone, a sandy middle limestone unit also containing Wolfcampian fusilinids, and a lower limestone unit bearing Atokan and Morrowan fusilinids.

In the North Hansel Mountains (approximately 10km west of Samaria Mountain) a more complete section of the Oquirrh Formation has been differentiated into 6 units by Allmendinger and Platt (1983). This section contains fusilinids ranging in age from Middle Pennsylvanian (Atokan) to Early Permian (Leonardian).

Fanglomerates, tuffaceous conglomerates, and sandstones of the Tertiary Salt Lake Formation have been mapped in a few rather small, isolated portions of the southern part of the study area (Doelling, 1980; Platt,
1977). On top of a hill mapped as Salt Lake Formation (Doelling, 1980) in the NW1/4 Sec. 12, T14N R6W (Plate 1) early settlers had dug a cellar 4 m deep and 5 m in diameter. The excavation is lined with cobbles of Oquirrh Formation sandstone and sandy limestone, the common rock types found at the surface throughout the southern portion of the study area. The cobbles have thick (2-4 mm) carbonate coatings and are pre-dominantly round to subround. The cellar was excavated in unconsolidated colluvium, and no outcrops of conglomerate, fanglomerate, or tuffaceous sandstone could be found nearby, or anywhere in the south-central portion of the study area. In this study, the area is mapped as undifferentiated debris (map unit Qd, Plate 1).

Exposures of Salt Lake Formation calcareous to tuffaceous conglomerate, siltstone, sandstone, and diamictite are present just north of the Utah-Idaho border at the south end of Samaria Mountain in the SE1/4 of Sec. 29, T16S R34E (Platt, 1977).

**Structure**

The Pennsylvanian section of the Oquirrh Fm. is 500 m thick at Samaria Mountain, but 2200 m thick in the North Hansel Mountains, a westward thickening of more than 4 times through a distance of 10-15 km. This thickening is ascribed to west-to-east low angle faulting during the
Mesozoic, which has resulted in a "telescoping" of the Oquirrh Formation (Allmendinger and Platt, 1983). The allochthon is exposed in the North Hansel Mountains, the "unnamed mountain", and on the west face of Samaria Mountain. Throughout the study area, outcrops of Oquirrh Formation are intensely folded, fractured, and jointed.

The mountains and valleys of the study area are typical of those found in the Basin and Range Province, and are thought to be the result of extensional block faulting initiated in the Miocene (Allmendinger and Platt, 1983). Structural and geophysical studies conducted within the Pocatello Valley area suggest that the North Hansel Mountains and Samaria Mountain are horsts; that "Little Pocatello Valley" is underlain by a small graben; and that the "unnamed mountain" may be the upper corner of an eastward tilted block which plunges northeastward beneath valley fill (Allmendinger and Platt, 1983; Harr and Mabey, 1976).

Gravity data show Pocatello Valley to be a closed structural basin, with much thinner valley fill than the adjacent Curlew and Malad valleys (Harr and Mabey, 1976). The North Hansel Mountains, Samaria Mountain, and the intervening Pocatello Valley are inferred to have been parts of the same uplifted block, which subsequent crustal extension has broken into several smaller blocks. Recent earthquake activity and apparent valley floor subsidence
Figure 4) suggest the process is continuing, and lend support to the conclusion of Harr and Mabey (1976, p.8) that "Pocatello Valley is a relatively young basin developing in what has been a major mountain mass".

**Gravity Survey Data**

Figures 5 and 6 show four transects through the study area that combine topography from USGS 7-1/2' quadrangles with inferred valley fill thicknesses derived from a gravity survey by Harr and Mabey (1976).

The results of the gravity survey show Pocatello Valley to be an enclosed structural basin partially filled with Cenozoic sediments derived from the surrounding highlands. In their study, gravity values were measured at 76 locations within the valley. A density contrast of 0.55 g per cm$^3$ was assumed between the Paleozoic bedrock (2.67 g/cm$^3$) and the valley fill sediments (2.12 g/cm$^3$). Figure 7 shows the Bouguer anomaly map of the present study area. The inferred valley fill thicknesses are shown on Figure 8. Closed depocenters exist in the southeast and northwest portions of the valley, on either side of a relative gravity high presumed to represent the northeast-plunging "unnamed mountain" half-graben. The deepest depocenter within the valley is adjacent to the southern portion of the Samaria Mountain front. The thickness of valley fill there is estimated to be 570 m.
Figure 4. Map showing apparent subsidence (in centimeters) of benchmarks that occurred between initial leveling in the early 1960's and re-leveling after the March, 1975, earthquakes (After Bucknam, 1976).
Figure 5. Index map of surface/subsurface transects through Pocatello Valley. The transects were constructed by combining topographic data from USGS 1:24,000 topographic quadrangles with inferred valley fill thicknesses from Harr and Mabey, (1976). (See also Figures 7 and 8).
Figure 6. Surface/subsurface transects through Pocatello Valley. No vertical exaggeration.
Figure 6. (continued).
Figure 7. Bouquer gravity anomaly map of the study area. After Harr and Mabey (1976).

Figure 8. Inferred valley fill thickness map of the study area. After Harr and Mabey (1976).
(Harr and Mabey, 1976). The gravity gradient from the mountain to the deepest part of the depocenter is approximately -3 mgals per km, almost double the gradient around most of the valley margin.
QUaternary geology

Pre-Lake Utaho/Lake Bonneville Deposits

Alluvium

Pediment gravels—The oldest Quaternary deposit mapped in this study is an alluvial gravel that mantles a remnant of a pediment surface on the east flank of the North Hansel Mountains in the west-central part of the study area (map unit Qag, Plate 1). The surface of the alluvial gravel is incised by stream channels and has undergone normal fault uplift relative to the adjacent valley floor. The pediment remnant forms the upper surface of a NNW-trending, 86-m high escarpment. The pediment surface is buried beneath an unknown thickness of valley fill on the downthrown side of the fault. The gravel-covered pediment surface is generally planar, and slopes approximately 3.5° toward the valley. In plan view, the deposit appears as isolated irregular patches on Oquirrh Formation bedrock, which strikes approximately N30°W and dips 30-45°E (Allmendinger, 1983). A test pit in the SE1/4NW1/4NW1/4 of Sec. 1, T16S R33E showed the gravel unit to be 50 cm thick and composed of 60% clasts of Oquirrh Fm. limestone ranging from 2 mm to 20 cm in diameter, and 40% clay, silt
and sand matrix. The limestone clasts are angular to subangular, and are predominantly irregular and equant. The amount of uplift and the degree of stream incision of the gravel suggest the age of the pediment surface is greater than the alluvial fan deposits (map unit Qaf, Plate 1).

Alluvial fan deposits- Alluvial fans (map unit Qaf, Plate 1) occur around the entire margin of Pocatello Valley, but are most numerous and well-developed along the east side of the valley at the base of Samaria Mountain. There, alluvial fans are present at almost every canyon mouth. Alluvial fans on the north, south, and west margins are not strongly expressed; most are clearly visible only on aerial photographs.

In 1983 and 1984, the northern Utah region received precipitation that exceeded 150% of normal (Arnow, 1984). Debris flows, mudflows, and other fan-building processes were active on several of the alluvial fans along the Wasatch Front, but no noticeable change occurred in the erosion-deposition regime of the Pocatello Valley area. Under present-day climatic conditions, the porous alluvial deposits at the canyon mouths around Pocatello Valley seem competent to accommodate even the anomalously high precipitation amounts of 1983 and 1984 through groundwater underflow alone.

Trench 1 exposed alluvial fan deposits to a depth of
Table 1. Descriptions of units exposed in Trench 1.

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<td>GRAD</td>
<td>CLRWVY</td>
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<sup>1</sup> Color designation from Munsell Soil Color Chart.
<sup>2</sup> CLYSLT = clayey silt; SLTCLY = silty clay.
<sup>3</sup> LS = gray Oquirrh Fm. limestone.
<sup>4</sup> Diameter of lithic fragments. Longest dimension of elongated clasts.
<sup>5</sup> MOD = moderately well sorted; W = well sorted; P = poorly sorted.
<sup>6</sup> IRR = irregular; PRO = prolate; EQ = equant; TAB = tabular.
<sup>7</sup> SA = subangular; R = round; SR = subround.
<sup>8</sup> The degree of carbonate cementation. MOD = moderate; VWEAK = very weak; ST = strong.
<sup>9</sup> CLR = clear; CLRWVY = clear wavy; GRAD = gradational.
4.5m, at the mouth of Elevator Hollow in the NW1/4NW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). A total of 13 stratigraphic units were differentiated (Table 1). All the units contain clasts of Oquirrh Formation limestone and sandy limestone. All but units C, H, and J are poorly sorted and matrix-supported, and appear to be of debris flow origin. Units C and H are moderately to well sorted and clast supported, and probably were deposited by clear-water streams. Unit J is a Lake Utaho highstand beach gravel lens (Figure 9).

In general, alluvial fans on the east side of the valley (adjacent to steep, high Samaria Mountain) have a more clearly conical shape, and are more areally restricted than fans on the north, south, and west margins. A total of 17 fans are present along the 8 km length of the west flank of Samaria Mountain. They range in area from less than 0.02 km$^2$ to greater than 1.5 km$^2$, but 14 of the 17 are less than 0.1 km$^2$ in area. The mean slope of fans on the east side of the valley is 6.7°, the mean length (here defined as the distance from the mountain front to the toe of the fan) is 552 m. In contrast, only 4 fans occur on the west side of the valley at the base of the North Hansel Mountains. Two of these fans have coalesced to form a bajada 3 km broad at the toe, and now cover an area of approximately 4.5 km$^2$. The other 2 fans are each approximately 0.75 km$^2$ in area. The mean slope of the
Figure 9. Log of Trench 1, located at the mouth of Elevator Hollow in the NW1/4NW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). Descriptions of units in Table 1.
alluvial fans on the west side of the study area is 1.9°, the mean length is 1330 m.

The bulk of alluvial deposition is assumed to have occurred prior to the time the valley was occupied by Lake Utaho. Fans with heads above the Utaho highstand shoreline bear both Utaho and Pocatello shoreline scarps, and fans with heads below the Utaho highstand but above the Pocatello shoreline bear only Pocatello shoreline scarps (Figure 10).

The alluvial fans and stream channel deposits in Pocatello Valley seem to be relics from the late Pleistocene. Depositional and erosional processes have been relatively inactive since the time Lake Utaho left the valley. If surface runoff has occurred in the stream channels during the Holocene, erosion and deposition have been so minor that the evidence has been obliterated by farming.

Pierce and Scott (1982) pointed out that alluvial processes in southeastern Idaho were much more active during the late Pleistocene than at present. They suggested that the increased alluviation was due not to increased precipitation, but to peak seasonal discharges up to 10 times present-day levels, a result of mean annual temperatures 10-15°C cooler than present. The cooler temperatures presumably caused decreases in evaporation, transpiration, and sublimation, and allowed more of the autumn and spring
Figure 10. Aerial photograph (view to west) of alluvial fans on the west margin of Pocatello Valley. Note stream channels graded to the Pocatello shoreline.
precipitation to occur as snow, and thus resulted in an increased cold season snowpack. Cooler temperatures were inferred to have delayed the snowmelt to a part of the season when the sun is more nearly vertical and the days are longer, which caused snowmelt to occur over a shorter time interval, and resulted in greater peak discharges. According to Pierce and Scott (1982), little of the significantly increased seasonal stream discharge could be accommodated through groundwater underflow, so most of the increase would occur as surface runoff.

Mass Movement Deposits

Colluvium- Bedrock outcrops are uncommon in the Pocatello Valley area. Almost all of the upland areas are covered with at least a thin veneer of colluvium composed of clasts and other weathering products of Oquirrh Formation limestone, sandy limestone, and sandstone. These three lithologies are found in varying relative proportions throughout the study area. The colluvium (map unit Qd, Plate 1; lithologic unit I, Figure 11 and Table 2) is generally composed of 50% coarse sand and finer groundmass, and 50% clasts of Oquirrh Fm. rocks ranging in size from granules to boulders 1 m in diameter. The Oquirrh limestone clasts are typically subangular, the sandy limestone clasts are subround, and the sandstone clasts are subround to round.

Loessial colluvium deposits (units II and IV, Figure 11


Figure 11. Log of Trench 2, located in the NE1/4SW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). Descriptions of units in Table 2. Lighter lines indicate soil contacts, heavier lines show lithologic contacts.
Table 2. Descriptions of units exposed in Trench 2.

<table>
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<th>Utah ho</th>
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<tr>
<td>LOWER CONTACT</td>
<td>CLR</td>
<td>CLRWVY</td>
<td>CLR</td>
</tr>
</tbody>
</table>

1 Color designation from Munsell Soil Color Chart.
2 SNDSLT = sandy silt; CLYSLT = clayey silt; SLTCLY = silty clay.
3 LS = gray Oquirrh Fm. limestone.
4 Diameter of lithic fragments. Longest dimension of elongated clasts.
5 MOD = moderately well sorted; W = well sorted; P = poorly sorted.
6 IRR = irregular; PRO = prolate; EQ = equant; TAB = tabular.
7 SA = subangular; R = round; A = angular; SR = subround.
8 The degree of carbonate cementation. MOD = moderate; W = well; P = poor; VWEAK = very weak.
9 CLR = clear; CLRWVY = clear wavy; GRAD = gradational.
and Table 2) were exposed in Trench 2, which was cut at the base of a faceted spur in the NE1/4SW1/4NE1/4 of Sec. 12, T16S R34E (map unit Qd, Plate 1). Although unit IV is of post-Utaho age, it is included here for the sake of simplicity. Both loess units are discussed further in the section on loess exposed in Trench 2. Figure 11 shows the log of Trench 2, and the units are described in Table 2.

**Talus slopes and "scree patches"**—Talus slopes occur on some of the west-facing slopes of faceted spurs near the base of Samaria Mountain (Figure 12). The largest of these is in the El/2NE1/4 of Sec 13, T16S R34E (Plate 1), and is composed of equant, angular clasts of Oquirrh Formation limestone ranging in size from 10 to 25 cm in diameter. Ledges of the limestone bedrock lie at the top of the talus body, which is approximately 100 m wide and extends 75 m downslope to the valley floor. The remaining talus slopes are small and discontinuous. All talus is included in map unit Qd (Plate 1), which also contains minor slopewash and colluvial deposits.

Other talus bodies exist on the west slope of Samaria Mountain as small, isolated "scree patches". These patches are here defined as talus that has the same slope as adjacent colluvium, has no convex-up or lobate form, and appears to have undergone little or no downslope transport (Figure 13). The scree patches are typically less than 10 m in diameter, and are composed of equant,
Figure 12. Aerial photograph (view to east) showing faceted spurs, talus slopes, and anomalously small alluvial fans at the base of the southern part of Samaria Mountain.
Figure 13. Photograph of a "scree patch" in the SW1/4SW1/4NE1/4 of Sec. 1, T16S R34E. See text for description.
angular clasts of Oquirrh Formation limestone ranging from 5 to 20 cm in diameter. There are no bedrock outcrops near the tops of the scree patches, although small outcrops usually present nearby suggest that bedrock is near the surface. The patches may be relict periglacial features produced by congelification (Bryan, 1946).

Soil slumps- Step-like soil slumps are present on the north-facing slopes of a few of the canyons on the west side of the valley in the North Hansel Mountains. Because of their small size, they are included in map unit Qd (Plate 1). Slopes that were actively slumping in November, 1983, were smaller than 100 m² in area. Soil slabs up to 0.5 m thick were cracked into steps 3-4 m long and 20-50 cm wide, with "risers" up to 0.5 m high. The roots of sagebrush growing on the slumping masses trailed uphill and the plants were rotated downslope. Adjacent portions of the slopes showed a similar staircase form, but cracked soil had been filled and plant growth had returned to normal. Apparently, the slopes had been inactive for several years.

Deposits of Lake Utah/Lake Bonneville Age

Alluvium

Stream channels are incised in most of the fans around the valley. All are graded to either the Lake
Utaho highstand or the Pocatello shoreline. No recent or modern surface drainage reaches the valley floor. According to a local farmer (Lee Furhiman, oral communication, 1983), for the first time in over 20 years a lake persisted year-round during 1983, in portions of Secs. 2, 3, 10, and 11, T16S R34E (Plate 1). The lake has remained since, apparently fed exclusively by groundwater underflow. In October, 1985, the lake covered an area of approximately 5 km$^2$ (Figure 14).

Poorly sorted stream channel deposits of sand, gravel and boulders up to approximately 0.5 m in diameter lie on the floors of canyons around the margin of the valley. The internal sedimentary character of the stream channel deposits is unobservable due to the lack of good exposures, but the deposits are comprised of erosion products of upstream Oquirrh Formation limestone, sandstone, and sandy limestone bedrock. The channels (map units Qau and Qab, Plate 1) are abruptly truncated at the Lake Utaho and Lake Bonneville highstand shorelines or at a lower, subsequent shoreline (map unit Qap, Plate 1). All of the stream channels in the study area are graded to Pleistocene lake shorelines, which suggests that little or no stream sedimentation or erosion has occurred since the lakes receded.

**Lacustrine Deposits**

**Shoreline deposits**— Lacustrine deposits of sand and
Figure 14. Aerial photograph (view to east) of the lake that occupied the valley in April, 1986, inundating several km$^2$ of grain acreage.
gravel (near shore) and silt and clay (lake bottom) that lie between the Lake Utaho highstand shoreline scarp and the Pocatello shoreline are shown as map unit Qlu; those lacustrine deposits below the Pocatello shoreline are shown as map unit Qlp (Plate 1). In the southern portion of the study area, Lake Bonneville deposits are shown as map unit Qlb (Plate 1).

Shoreline features typical of those found in the Lake Bonneville basin (bars, spits, and beach terraces) are not strongly developed in the Pocatello Valley area. The wave energy of Lake Utaho was limited by a short (10-15km) fetch, and the mountains surrounding the valley would have provided some protection from strong, persistent winds. In addition, the only sources of the detritus needed for the development of large constructional features were local canyon streams that drained rather small basins. Shoreline morphologic features are not differentiated from lacustrine units Qlu, Qlp, and Qlb (Plate 1).

The shoreline scarp of the Lake Utaho highstand, while subtle in most places, can be traced around the entire valley margin. It lies at 1573 m (5160 ft) elevation in the southern portion of the study area, on a bar at the south end of "Little Pocatello Valley" which separated Lake Utaho from the main body of Lake Bonneville, and at 1564 m (5130 ft) in the northern part of the study area (Plate 1). At its maximum, Lake Utaho covered an area of
approximately 114 km$^2$ to a maximum depth of about 61 m.

A second, even less prominent shoreline scarp lies 40 m below the Utaho highstand scarp, and was created when the lake was approximately 30% smaller in area than it was at the highstand. At this post-highstand level the lake covered approximately 80 km$^2$ to a depth about 21m. The shoreline scarp (here called the Pocatello shoreline) lies on land that is less steep and less rocky than that of the Lake Utaho highstand scarp, and in most instances it has been tilled repeatedly by grain farmers. This has obscured what features may have been visible prior to agriculture in the valley. Although it is visible on aerial photographs, the only places the Pocatello shoreline scarp can be clearly seen on the ground are at a promontory located in the NE1/4 of Sec. 35, T15S R34E, and on the north flank of the unnamed mountain in Secs. 20 and 21, T16S R34E (Plate 1). Figure 15 shows a photograph of both the Utaho highstand and Pocatello shoreline scarps in profile.

Six lower, post-Pocatello shorelines are evident in the southwest part of the study area, in "Little Pocatello Valley". They are seen only as subtle tonal variations on aerial photographs, at vertical intervals of 2 to 5 m below the Pocatello shoreline, and are included in map unit Q1p (Plate 1).

One of the best-defined Lake Utaho beach terraces in
Figure 15. Photograph (view to south) of a promontory located in the NW1/4 of Sec. 35, T15S R34E, showing the Lake Utaho highstand and Pocatello shorelines in profile.
Pocatello Valley is on the north flank of the unnamed mountain in the south-central part of the study area, in the SE1/4NW1/4NE1/4 of Sec. 21, T16S R34E (Plate 1). Here, beach deposits have been exposed in a gravel pit operated by Oneida County, Idaho. The pit was the only place other than the trenches dug for this study where Utaho-age shoreline deposits could be examined in section. The numerous units of well sorted sand and gravel units exposed showed successive landward onlapping, typical of transgressive shoreline deposits found in the main body of Lake Bonneville (J.P. McCalpin, oral communication, 1983). The uppermost portion of the excavation is more than 5 m below the Utaho highstand elevation, so a correlative highstand beach gravel could not be found for comparison with beach gravel lenses found in Trenches 1 and 2. The gravel pit was reclaimed in early 1986.

A lens of poorly sorted Utaho highstand beach gravel was exposed in Trench 1 in the NW1/4NW1/4NE1/4 of Sec. 12, T16S R34E (Plate 1). The gravel (unit J, Figure 9) is composed of prolate, tabular, and minor equant round to subround clasts of Oquirrh Formation limestone. The clasts range from 3 mm to 10 cm in diameter and are very loosely packed. Cementation is minimal, and occurs primarily as carbonate coatings on the undersides of the clasts. The lens body is crescentic, concave-up, approximately 2.5 m long, and 50 cm thick at its center.
The top of the lens is at 1570 m (5151 ft) elevation.

In Trench 2, at the base of a faceted spur approximately 400 m south of Trench 1, only the upslope portion of the Utaho highstand beach gravel was exposed (Figure 11). Of the 3.5 m length that was exposed, the lens has a maximum thickness of nearly 1 m. The gravel has the same characteristics as that found in Trench 1, except the size range is somewhat larger (from 1 to 16 cm) and the carbonate coatings on the undersides of the clasts are slightly thinner. The elevation of the top of the beach gravel lens in Trench 2 is 1569 m (5148 ft).

Lake bottom sediments exposed in Trench 3- Lake bottom sediments were exposed in Trench 3, in the center of the NE1/4NE1/4 of Sec. 9, T16S R34E (map unit Q1p, Plate 1). The trench was sited so as to intersect snowcracks mapped by Kaliser (1976) following the Pocatello Valley (Utah-Idaho Border) earthquake of March 28, 1975. The infilled cracks extend approximately 0.5 m below the ground surface, and are probably produced by the shrinking and swelling of highly plastic clays, not groundshaking. Similar cracks were observed several times during the present field investigation, and usually appeared 2 or 3 days after heavy rain showers. A vertical column 0.5 m wide and 4.5 m deep was subdivided into 11 units on the basis of subtle textural, color, and bedding charac-
teristics (Table 3). Because no clear unconformities were observed, age differentiation of the lake bottom beds was not attempted. Additionally, no soft sediment deformation was observed, which suggests that earthquakes of magnitudes sufficient to deform the lake bottom beds did not occur in the area during periods it was occupied by pluvial lakes.

All lake bottom sediments below the Pocatello shoreline are mapped as Qlp (Plate 1). All the samples from Trench 3 are comprised of particles smaller than 3\(\phi\), with 50-75\% of the particles smaller than 8\(\phi\) (Appendix B). The weight percents of acid-insoluble residues of Trench 3 samples are listed in Appendix C. All the samples have similar mineralogy, except chlorite is present in the 26-122 cm portion (Appendix D). Microscopic examination of the 8-3\(\phi\) fractions (silt and very fine sand) revealed the presence of shell fragments in each of the 11 sampled units, but no whole specimens were found in any of the samples and identification was not attempted.

Lagoonal deposits- Surficial deposits of silt and clay were found behind a Bonneville-age bar in the N1/2NW1/4 of Sec. 12, T14N R6W (map unit Q11, Plate 1). A moderately well defined Lake Bonneville shoreline scarp is approximately 4 m above the top of the bar, and can be traced around the margin of the lagoon. During the time
Table 3. Descriptions of units exposed in Trench 3.

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1 Color designation from Munsell Soil Color Chart.
2 CLY = clay; S LTCLY = silty clay.
3 BLKY = blocky; PRISM = prismatic; POLYG = polygonally faceted particles; GRAN = granular.
4 WEAK = only faintly visible; CRDWYY = clear, wavy. Thickness of individual beds shown in mm.
5 GRAD = gradational; CLR = clear.
of the Bonneville highstand, the lagoon covered an area of 0.3 km$^2$ to a depth of approximately 8 m.

Evidence of a larger Bonneville-age lagoon can be found in portions of Secs. 2, 3, 10, and 11, T14N R6W. Here, the top of the bar which separated the lagoon from Lake Bonneville is 6 m below the highstand shoreline scarp. Within the lagoon, a knoll of Pennsylvanian-Permian Oquirrh Fm. limestone is located in the SW1/4SE1/4 of Sec. 3. A faint shoreline scarp encircles the knoll, the top of which protruded about 4 m above the water surface during the Bonneville highstand. The lagoon covered an area of approximately 5 km$^2$. On the northeast margin of the lagoon there is a loess-mantled ridge (Figure 16), the lowest elevation of which is 1582 m (5196 ft), 10 m above both the Lake Utaho highstand shoreline scarp (on the north) and the Lake Bonneville-age lagoon shoreline scarp (on the south).

**Eolian deposits**

Surficial loess—The only surficial loess found in the Pocatello Valley area mantles a low lying, west-trending ridge at the south end of "Little Pocatello Valley" (Figure 17) in Sec. 3, T14N R6W (map unit Qel, Plate 1). On the north slope, a Lake Utaho highstand shoreline scarp lies 10 m below the lowest point on the ridge. As mentioned, the shoreline of a Bonneville-age lagoon occurs at the same elevation (1573 m; 5160 ft) on the south slope.
Figure 16. Aerial photograph (view to southeast) of the barrier ridge shown in Figure 17.
Figure 17. Transect through the barrier ridge at the south end of "Little Pocatello Valley" (Plate 1) that separated Lake Utaho from the main body of Lake Bonneville.
of the ridge (Figure 16). Both shorelines can be clearly seen on aerial photographs. A test pit in the SW1/4SW1/4NE1/4 of Sec. 3, T14N R6W (Plate 1) exposed yellow-buff structureless silt to a depth of at least 1 m. The loess contained approximately 20% randomly oriented angular clasts of Oquirrh Formation limestone up to 2 cm in diameter. The presence of the loess (although the lithic fragments suggest it may be re-worked, and not primary loess) further supports the idea that Lake Utaho and Lake Bonneville were not directly connected. It is unlikely that unconsolidated silt could remain on a topographically high structure such as the ridge after being submerged for even a short time.

Loess exposed in Trench 2- Two colluvial loess units (units II and IV, Figure 12) were exposed in Trench 2 (Plate 1). The lower unit (II) has an exposed thickness of 1.6 m, and overlies buried soil unit Blb (although unit II is of pre-Utaho age, it is discussed here for simplicity). The upper loess unit (IV) has a maximum exposed thickness of 1.5 m, and underlies modern soil unit Bca. Both loess units are truncated by unit Bca, and probably thicken downslope. A lens of Utaho highstand beach gravel is situated between the two loess units, indicating that loess deposition occurred both prior to and following the Lake Utaho highstand. Both units are
apparently unstratified, but a few scattered pebbles are present in the lower unit, which suggests it may have been reworked to a greater degree than the upper unit. The loess units are described in Table 2.

Loess deposits between 50 and 100 cm thick cover much of southeastern Idaho (Lewis and Fosberg, 1982). Most of this loess is presumed to have been derived through the deflation of alluvial sediments from the Snake River Plain during the late Pleistocene (Pierce and others, 1982; Lewis and Fosberg, 1982; Pierce and Scott, 1982). Deflation of pluvial lake beds during periods of desiccation has also been suggested as a possible source of loess found in areas adjacent to Pleistocene lakes Lahonton and Bonneville (Morrison, 1965). Loess deposits on the west flank of the North Hansel Mountains (approximately 3 km west of the northwest corner of the present study area) are described by Allmendinger (1983) as probably consisting of Bonneville lake bottom sediments reworked by wind.

A loess section described by Lewis and Fosberg (1982) near Malad, Idaho, (approximately 15 km northeast of the present study area) is the southernmost sample site of three taken along a 73 km-long traverse originating on the Snake River Plain just west of Pocatello, Idaho. The 1.8 m thickness of the loess at this site showed a decrease in the very fine sand and coarse silt size fractions and an
increase in the medium and fine silt fractions that was considered by the researchers to be consistent with the trend of decreasing loess thickness and decreasing loess particle size that they expected and observed with increasing distance from the Snake River Plain. Because the loess body appears to be continuous from the Plain, they concluded that little if any of the loess at Malad was derived through the deflation of Lake Bonneville sediments, although the Bonneville shoreline is just a few kilometers south of the sample site.

Approximately 1000 km$^2$ of Lake Bonneville lake bottom sediments lie to the west and north of the Pocatello Valley (Currey, 1982, Sheet 1). During periods of desiccation, it is likely that winds from the north and west (the directions assumed by Lewis and Fosberg, 1982) would have carried Lake Bonneville lake bottom sediments into Pocatello Valley, which may have combined with airborne silt from the Snake River Plain and deposited as post-Utah loess in the present study area.
Range Front Morphology

Samaria Mountain

Gravity survey data (Harr and Mabey, 1976) show the deepest Cenozoic depocenter in Pocatello Valley to be adjacent to the southern part of Samaria Mountain. The steep gravity gradient that occurs between the Paleozoic bedrock of the mountain and the Cenozoic valley fill of the depocenter is perpendicular to a portion of the range front in which geomorphic evidence of Quaternary tectonism can be seen in the form of a steep mountain front, triangular faceted spurs, and anomalously small alluvial fans at canyon mouths (Figure 13). Platt (1976) inferred that subsidence of the valley floor near the base of Samaria Mountain has resulted in alluvial fans of much smaller volume than would be expected considering the size of the canyons from which the alluvium originated.

Samaria Mountain is much steeper and more elevated above the valley floor than the North Hansel Mountains across the valley to the west. Samaria Mountain has been uplifted at least 1420 m, the difference between the elevation at the top of the mountain and the base of
valley fill inferred by gravity survey (Harr and Mabey, 1976). Active stream channel downcutting has produced steep, high canyon walls, and canyon embayment and pedimentation of the range front are minimal, which suggests a young, active mountain front.

To obtain a general quantitative indication of the tectonic activity of the western front of Samaria Mountain, mountain front sinuosity and valley width-valley height ratios were determined using methods outlined by Bull and McFadden (1977). The mountain front sinuosity ratio (S) is defined by the equation:

\[
S = \frac{L_{mf}}{L_{s}}
\]

where \( L_{mf} \) equals the length of the mountain-piedmont junction (this term is admittedly somewhat subjective, and is defined by Bull and McFadden (1977) as the junction between the erosional and depositional subsystems), and \( L_{s} \) equals the overall length of the mountain front. Tectonically inactive mountain fronts tend to have high (e.g., 3.0 to 7.0) \( S \) values, due to extensive pedimentation and embayment of canyon mouths. In contrast, tectonically active mountain fronts have low (e.g., 1.2 to 1.6) values of \( S \), because the fall of base level has been great enough that streams maintain active channel downcutting, which allows for little pedimentation and embayment. The western front of Samaria Mountain has an \( S \) value of 1.7 (Figure 18), which is indicative of
Figure 18. Map of the south-central front of Samaria Mountain. Heavy line is the range front, lighter line is the mountain/piedmont junction used in the calculation of mountain-front sinuosity. Ball-and-bars indicate where $E_{ld}$, $E_{rd}$, and $E_{sc}$ were measured for valley floor width/valley height ratios.
slightly active to active tectonism (Bull and McFadden, 1977, p. 115).

The valley floor width-valley height ratio \( V_f \) is another indicator of the rate of base level fall due to tectonism, and is defined by the equation:

\[
V_f = \frac{V_f}{(Eld-Esc) + (Erd-Esc)}
\]

where \( V_f \) equals the width of the valley (or canyon) floor, Eld and Erd are the elevations of the left and right divides, respectively, and Esc is the elevation of the stream channel. In this study, all of these factors were measured at a point 0.5 km upstream from the mountain front. (The canyons studied by Bull and McFadden are much longer, so they measured the factors at a point 1 km upstream). The ratio \( V_f \) gives a general indication of whether the stream is engaged in channel downcutting (higher rate of base level fall, lower value of \( V_f \)) or lateral erosion (lower rate of base level fall, higher value of \( V_f \)). 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 base of Samaria Mountain have stream channels long enough for use in valley floor width-valley height calculations, and these yielded values of 0.15 and 0.27 (Figure 18), characteristic values for deep, strongly v-shaped canyons.
The mountain front sinuosity ratio and the valley floor width-valley height ratios calculated for the western front of Samaria Mountain quantify the geomorphic evidence from field observation, topographic maps, and aerial photographs: the mountain front is steep, the canyons are deeply incised and strongly v-shaped, and pedimentation and canyon embayment are minimal, all of which suggest a young, tectonically active mountain front.

North Hansel Mountains

On the west side of the study area, the North Hansel Mountains slope gradually up from the valley floor. The eastern flank of the North Hansel Mountains is not a "mountain front" in the sense that the western front of Samaria Mountain is, rather, it is the eroded top of an eastward-tilted block.

Over 30 normal faults have been mapped in the portion of the North Hansel Mountains which lies within the present study area (Allmendinger, 1983). Many of these faults are only a few hundred meters long, have small displacements, and strike predominantly east and northwest. The longest and apparently most recently active fault lies in the west-central part of the study area (Plate 1). It is 8 km long and strikes N20°W. The southern 5 km of the fault is most distinct, and displaces an older Quaternary pediment surface (map unit Qag, Plate 1) by as much as 86 m. The northern 3 km of the fault can
be traced into Oquirrh Fm. bedrock, and is much less distinct than the southern section. The northern portion of the fault does not define a range front, so it could not be used in the determination of mountain front sinuosity (S). The S value of the southern portion is 2.33 (Figure 19), indicative of moderate to slightly active tectonism (Bull and McFadden, 1977, p.115). The valley floor width-valley height ($V_f$) values of two gullies along the range are 1.0 and 1.25. These values are 5 to 10 times those calculated for the Samaria Mountain range front, and suggest that the uplift rate of the western margin fault is less than that of the eastern margin fault. Most of the displacement along the western margin fault has probably occurred during the Quaternary, following a period of tectonic quiescence that allowed the formation of the pediment gravel surface (map unit Qag, Plate 1). Uplift of Samaria Mountain may have begun as early as the Miocene, when extensional block faulting in the Basin and Range was initiated (Allmendinger and Platt, 1983).

Shoreline Deformation

Because no fault scarps displacing Quaternary deposits could be found in the study area, it was reasoned that indirect evidence of tectonic deformation might be found in anomalous deflections of the originally horizontal Lake
Figure 19. Map of the western margin fault zone on the east flank of the North Hansel Mountains. (See caption of Figure 18 for explanation).
Utaho highstand shoreline. Figure 20 shows the locations of three shoreline elevation profiles through Pocatello Valley that were constructed using shoreline elevations that had been corrected for post-Lake Utaho highstand colluviation. The shoreline elevations used in the profiles are listed in Table 4.

Profile A-A' (Figure 21) shows shoreline elevations that deviate as much as 6.4 m from each other and up to 4.5 m from elevations predicted by Crittenden (1963, Figure 3) based on isostatic rebound of the Bonneville Basin. It is interesting to note how closely the elevation of the north shoreline of Lake Utaho (1581 m; 5138 ft) coincides with the elevation interpolated from Crittenden, (approximately 5140 ft) although his nearest surveyed shoreline elevation is over 30 km to the southwest. The most significant deviations from a smooth profile between the north and south shorelines of Lake Utaho occur where the shoreline crosses faults at the base of Samaria Mountain. Shoreline point 17, (Figure 21) at the southern end of the profile, is cut onto a faceted spur on the upthrown side of the inferred range front fault, whereas shoreline points 4, 5 and 6 are clearly on the downthrown side of two parallel range front faults (Figure 20). Point 17 lies very near the elevation predicted by Crittenden (1963, Figure 3), whereas points 4, 5 and 6 plot 4.2 to 4.5 m too low in relation to their...
Figure 20. Index map showing the locations of Lake Utaho highstand shoreline elevation profiles. Numbered shoreline elevation points are keyed to points listed in Table 4.
Table 4. Surveyed shoreline data for the Lake Utah highstand shoreline in Pocatello Valley. Twenty-five additional shoreline elevation points are not included here because their slopes were not profiled.

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<th>Colluvial Wedge Correction (ft.)</th>
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1 Reference points of known elevation where the surveying base station was located: a, SW corner, Sec. 16, T15S R34E; b, BM O' FNL, Sec. 26, T15S R34E; c, SW corner, Sec. 1, T16S R34E; d, SW corner, Sec. 13, T16S R34E; e, SW corner, Sec. 30, T16S R34E; f, BM SE corner, Sec. 20, T15S R34E; g, NW corner, Sec. 17, T16S R34E.

2 Shoreline points shown on Figures 13 and 14.

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

4 Calculated by adding column 3 to column 1.

5 The elevation difference between: a) the projected shoreline angle (see Figure 2), and b) the point where the shoreline elevation was measured; precision = 1.0 ft.

6 Calculated by adding column 5 to column 4. This elevation should be within 1.6 ft. of the mean elevation of the formative water plane (Rose, 1981, Table 5.6).
Figure 21. Lake Utah highstand shoreline elevation profiles.
predicted post-rebound elevations.

The highly variable shoreline elevations between points 16 and 7 probably result from the shoreline crossing repeatedly from the upthrown to the downthrown side of the more linear buried fault or fold hingeline. For example, point 9 is a surveyed shoreline cut into bedrock at the base of a faceted spur, while points 8 and 10 are shoreline scarps cut into unconsolidated colluvium farther valleyward, on the downthrown block. The position of a linear fault at the range front would cross somewhere between point 9 and points 8 and 10. To the north, a parallel branch of the north-trending range front fault intersects the north-northwest-trending Lake Utaho shoreline somewhere between points 6 and 9. Within this area, the shoreline drops 6.4 m in elevation in a horizontal distance of only 900 m. All of the shoreline profiles are very similar, which suggests they were formed with a constant relation to the mean water plane. It is important to note that there are no fault scarps between points 6 and 9.

East-west shoreline profiles suggest the Lake Utaho shoreline has been tilted eastward from 1.8 m (Profile B-B', Figure 21) to as much as 7.4 m (Profile C-C', Figure 21) in the most active-appearing portion of the range front. According to Crittenden (1963, Figure 3), approximately 1 m of west-to-east tilt of shorelines
across Pocatello Valley is expected, due to differential isostatic rebound. If this tilt is taken into account, the small residual displacement on profile B-B' (1.8 m - 1.0 m = 0.8 m) is within the range of surveying error and does not suggest that significant tectonic eastward rotation has occurred north of the mapped eastern margin fault. The 6.4 m residual down-to-the-east displacement along profile C-C' strongly suggests that post-Utaho slip has occurred along the eastern margin range front fault. The overall implication is that late Quaternary displacements along the range front fault have resulted in very little absolute elevation change of the upthrown block, because points on the upthrown block closely match the smooth profile shoreline elevations predicted by Crittenden. However, points on the downthrown block have been lowered 4.6 - 6.4 m, presumably the result of subsidence of the Pocatello Valley block. Re-leveling of benchmarks by Bucknam (1976) following the 1975 M_L 6.0 earthquake showed no absolute uplift of surrounding mountain blocks, (Figure 4, this paper) but a maximum of 13 cm of valley floor subsidence was measured near the epicenter in western Pocatello Valley. The closed topography of Pocatello Valley also suggests that valley block subsidence has been the dominant tectonic trend during the Quaternary, as previously suggested for the late Cenozoic by Harr and Mabey (1976).
Fault Zone Trenching

Because Trench 1 yielded no tectonic information, Trench 2 was cut at the base of a faceted spur approximately 400 m south of Trench 1 (Plate 1). Excavation was begun at the base of the spur directly on an outcrop of Permian-Pennsylvanian Oquirrh Formation limestone, which was encountered at depths ranging from 0 to 1.5 m in the upslope 15 m of the trench (Figure 11). The bedrock contact beneath the colluvium then dropped 3 m at an angle of $63^\circ$ into the bottom of the trench, 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 deformed colluvial deposit in Trench 2 (unit IV) contains a paleosol with well-developed argillic horizons. An attempt was made to date the unit IV paleosol by its degree of clay formation and its relation to late Pleistocene lacustrine and eolian features (McCalpin, Robison, and Garr, in press). Comparison of characteristics of the unit IV paleosol with a dated buried soil from Jordan Valley, Utah, (Scott and others, 1982, p.42) suggests that the unit IV soil represents approximately 66+ 15 ka of weathering (Figure 22). Overlying loessial colluvia (units II and IV) respectively
Figure 22. Comparison of paleosols in Trench 2 with buried soils in Jordan Valley, Utah. The amount of pedogenic clay of the buried soil in Trench 2 is 70% of that in the Jordan Valley soil, which has been estimated as having undergone 94±22 ka of development, based on calcium carbonate accumulation (Scott and others, 1982, Table 5 and p. 42). Assuming that the loessial parent materials of both soils contained 25% clay (as suggested for late Pleistocene loess by Shroba, 1987), that the bulk densities of the two soils are similar and that clay formation rates are constant through time, the Trench 2 soil represents roughly 66±15 ka of soil formation. This age, when added to the age of overlying loess unit C2ca (approximately 32 ka, if deposited during the early Bonneville transgression) yields an age of 98±15 ka, an age that is generally compatible with the age inferred if tectonic deflection has occurred at a constant 0.25°/ka rate over the past 95 ka.
under- and overlie a lens of Lake Utaho highstand beach gravel (unit III) which is assumed to be contemporaneous with the Lake Bonneville highstand shoreline (15.5-17.0 ka; Scott and others, 1983). The modern colluvium (unit V) overlies an erosion surface that truncates all underlying units.

The presence of older, steeper colluvial wedges overlain by successively younger, more gently dipping layers could be expected, but the very steep tilt of the unit I paleosols (up to $49^\circ$) is anomalous. Colluvial wedges below fault scarp free faces typically have initial slopes of $35^\circ$ (Wallace, 1977). This slope is maintained only while deposition is rapid. By the time the slope is declining slowly enough to allow soil development on the colluvial wedge, the slope is lower ($8-25^\circ$; Wallace, 1977, Figure 3, stage E). Buried soil horizons within unit I must have required slope stability to form, but they now dip $35^\circ$ to $49^\circ$ west downslope of the tilt axis shown in Figure 11. The unit II/unit IV contact is bent slightly along this axis, whereas younger units are not deflected. The amount of angular change on the contact between unit I and unit II ($8^\circ$; age 32 ka if deposited early during the Utaho transgression) versus the angular change on the unconformity between units II and IV ($4^\circ$; age 15-17 ka) suggests progressive monoclineal flexure at a mean rate of $0.25^\circ$/ka. If this flexure rate is applied to the even
more steeply dipping buried soil horizons B3b and B4b, an age of approximately 88 ka is derived. This age is roughly compatible with the age of 98 ± 15 ka for unit II estimated from the degree of soil development (Figure 22). There are no discrete faults or shear zones in any of the colluvial units exposed in Trench 2.

Monoclinal flexure of surficial materials in narrow zones 2 to 50 m wide have been described by Clark and others (1972, p. 118) and Bonilla (1982, p. 18). Although the evidence found in Trench 2 is not conclusive, it does suggest that progressive monoclinal flexure of the colluvium has occurred during the past 95 ka within a zone approximately 15 m downslope of the inferred range front fault in Paleozoic bedrock, and that the displacement has occurred without surface rupture. It is not known whether the flexure is the result of coseismic deformation or aseismic creep, but the absence of fractures or shears within the colluvium supports the latter mechanism.
Depositional evidence of surficial geologic processes that were active prior to the time Pocatello Valley was occupied by Lake Utah is limited. There is an uplifted, dissected pediment surface remnant (map unit Qag, Plate 1) on part of the east flank of the North Hansel Mountains, and a gently sloping, colluvium-mantled piedmont (map unit Qd, Plate 1) that lies between the Paleozoic bedrock highlands and the late Pleistocene lacustrine deposits on the valley floor. The presence of the elevated pediment remnants on the upthrown side of the western margin fault suggests a long period of tectonic quiescence passed prior to movement along the fault. No evidence as to the precise timing of the uplift was found, but the mountain front sinuosity ratio calculated for the southern portion of the fault indicates moderate to slightly active tectonism. Because the Utaho highstand shoreline is so poorly developed along much of the west side of the valley, shoreline deflections due to tectonism since the time Lake Utaho occupied the valley were not detected.

The bulk of alluvial fan deposition occurred prior to the last pluvial lake maximum approximately 15 ka (Currey and Oviatt, 1985). Stream channel deposits graded to the
Lake Utaho and Lake Bonneville highstand shorelines (map units Oau and Qab, respectively, Plate 1) were deposited while the lakes occupied their highest positions. The fact that present-day stream channels are still graded to the Utaho highstand shoreline suggest that alluvial processes have been relatively inactive since the lake regressed.

It is not known whether a hydraulic connection existed between Lake Utaho and Lake Bonneville. The highstand shorelines of both bodies occur on either side of a barrier ridge at the south end of "Little Pocatello Valley", at 5160 ft, (1587 m) elevation, 36 ft (11 m) below the lowest point on the ridge. A surface connection between the two lakes is unlikely, because the ridge is mantled with loess (map unit Qel, Plate 1) which would have been quickly eroded had the ridge been submerged for even a short time. It seems more probable that the lakes communicated through the barrier ridge, perhaps through unconsolidated older Quaternary materials. If such a connection existed, it would explain what would otherwise be an extraordinary coincidence: that climate-controlled Lake Utaho maintained the same highstand elevation as the exterior-threshold-controlled Lake Bonneville. Also, if a subsurface hydraulic connection existed, the Pocatello shoreline at 80 ft (25 m) below the Utaho highstand may represent the level to which Lake Utaho dropped in
response to the catastrophic Bonneville Flood, as the main lake level passed below the threshold of the hydraulic connection.

The lake may have remained at the level of the Pocatello shoreline until approximately 14 ka, when climatic change caused the pluvial lakes throughout the region to recede (Currey and Oviatt, 1985).

Only 7 of the 21 stream channels around Pocatello Valley regraded to the Pocatello shoreline (map unit Qap, Plate 1). There are no stream channels graded to the valley floor, which suggests that seasonal snowmelt has been through groundwater underflow since Lake Utaho desiccated. The primary cause of the regression from the Pocatello shoreline was probably a shift in the precipitation-evaporation balance due to climatic change.

Although there are no fault scarps in late Quaternary deposits on the east margin of Pocatello Valley, evidence of small but measurable late Quaternary displacement is seen in the rapid elevation changes (up to 6.4 m) of the 15 ka Utaho highstand shoreline, and in the warped but unfaulted range front colluvium in Trench 2. The vertical shoreline deflections are greatest along a 7 km-long segment of the central part of the eastern margin fault zone. Most of the displacement during the late Quaternary here seems to be due to subsidence or eastward rotation of the valley block, as evidenced by the anomalously small
alluvial fans along the eastern margin of the valley, and the observation that most of the shoreline elevations measured at points on the downthrown side of the faults fall below the shoreline elevations predicted by Crittenden (1963), while points measured on the upthrown block lie very near the predicted rebounded shoreline elevations.

The surface rupture threshold for shallow focus earthquakes on normal faults is estimated to be $M_L 6.2-6.3$ (Bonilla, 1982; Slemmons, 1982). Apparently, no earthquakes of this magnitude or greater have occurred anywhere along the eastern margin of Pocatello Valley in the last 15 ka. Colluvium estimated to be 98+15 ka (McCalpin, Robison, and Garr, in press) exposed in Trench 2 shows no disruption within 20 m of the inferred range front fault, which suggests that no events of $M_L 6.2-6.3$ or greater have occurred in that time.
SUMMARY AND CONCLUSIONS

The Pocatello Valley area provides a unique opportunity to examine a well-preserved late Pleistocene geologic record. Because it is a closed basin, the valley has undergone a minimum of outside influence. Aside from wind-borne loess, little if any material has been introduced or removed from the valley during the Quaternary.

The surficial deposits and their associated landforms seem to be relics of the late Pleistocene, when mean annual temperatures, evaporation, transpiration and sublimation were decreased, and peak seasonal stream discharge was increased relative to present-day levels. Under modern climatic conditions, erosional and depositional processes are nearly inactive. The porous materials on the piedmont surrounding the valley apparently are competent to conduct seasonal precipitation and snowmelt through groundwater underflow. There is no evidence of significant surface drainage to the valley floor since Lake Utaho regressed from the Pocatello shoreline.

The youthful appearance of the western front of Samaria Mountain, anomalously small alluvial fans, and warped pluvial-lake shorelines suggest that the Pocatello Valley area has been tectonically active during the
Quaternary. The lack of fault scarps in Quaternary deposits suggests that relative movement of the mountain and valley blocks may be achieved through earthquakes below the threshold necessary for surface rupture. Buried well-developed argillic soil horizons examined in Trench 2 are monoclinaly draped and dip as much as 49° over an inferred range front fault, and are presumed to have been deflected by either coseismic flexure or aseismic creep over the buried fault trace. The earthquake sequence of March-April, 1975, and continuing low-magnitude seismicity during the past 13 years suggest the Pocatello Valley area is still tectonically active.

Further study, perhaps including boreholes, geophysical techniques (shallow, high-resolution seismic reflection, for example) and additional trenches could bring forth more information on this unique and interesting area.
REFERENCES CITED


Fremont, John C., 1845, Report of the exploring expedition to the Rocky Mountains in the year 1842 and to Oregon and northern California in the years 1843-44: Washington, D.C., Gales and Scaton, 693p.


Appendix A. Water Content of Samples From Trenches 2 and 3 (T2 and T3).
## Trench 2 Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Moist Wt. (gms)</th>
<th>Dry Wt. (gms)</th>
<th>Wt. % Water</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>50.0</td>
<td>47.24</td>
<td>5.52</td>
</tr>
<tr>
<td>Bca</td>
<td>50.0</td>
<td>46.19</td>
<td>7.62</td>
</tr>
<tr>
<td>C1ca</td>
<td>50.0</td>
<td>42.64</td>
<td>14.72</td>
</tr>
<tr>
<td>C2ca</td>
<td>50.0</td>
<td>40.91</td>
<td>19.62</td>
</tr>
<tr>
<td>Blb</td>
<td>50.0</td>
<td>45.13</td>
<td>9.74</td>
</tr>
<tr>
<td>B2b</td>
<td>50.0</td>
<td>39.92</td>
<td>20.16</td>
</tr>
<tr>
<td>B3b</td>
<td>50.0</td>
<td>39.94</td>
<td>20.12</td>
</tr>
<tr>
<td>B4b</td>
<td>50.0</td>
<td>41.70</td>
<td>16.60</td>
</tr>
<tr>
<td>Cox</td>
<td>50.0</td>
<td>42.37</td>
<td>15.26</td>
</tr>
<tr>
<td>Altered PPo</td>
<td>50.0</td>
<td>43.97</td>
<td>12.06</td>
</tr>
</tbody>
</table>
### Trench 3 Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Moist Wt. (gms)</th>
<th>Dry Wt. (gms)</th>
<th>Wt. % Water</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-26 cm</td>
<td>50.0</td>
<td>40.81</td>
<td>18.38</td>
</tr>
<tr>
<td>26-46 cm</td>
<td>50.0</td>
<td>39.51</td>
<td>20.98</td>
</tr>
<tr>
<td>46-73 cm</td>
<td>50.0</td>
<td>41.56</td>
<td>16.88</td>
</tr>
<tr>
<td>73-112 cm</td>
<td>50.0</td>
<td>42.13</td>
<td>15.74</td>
</tr>
<tr>
<td>112-160 cm</td>
<td>50.0</td>
<td>42.83</td>
<td>14.34</td>
</tr>
<tr>
<td>160-192 cm</td>
<td>50.0</td>
<td>41.17</td>
<td>17.66</td>
</tr>
<tr>
<td>192-240 cm</td>
<td>50.0</td>
<td>41.70</td>
<td>16.60</td>
</tr>
<tr>
<td>240-272 cm</td>
<td>50.0</td>
<td>40.23</td>
<td>19.54</td>
</tr>
<tr>
<td>272-339 cm</td>
<td>50.0</td>
<td>41.10</td>
<td>17.80</td>
</tr>
<tr>
<td>339-380 cm</td>
<td>50.0</td>
<td>36.92</td>
<td>26.16</td>
</tr>
<tr>
<td>380-424 cm</td>
<td>50.0</td>
<td>37.35</td>
<td>25.30</td>
</tr>
</tbody>
</table>
Appendix B. Grain Size Analysis of Samples From Trenches 2 and 3 (T2 and T3).
Trench 2, Unit A

GRAIN SIZE (PHI UNITS)

CUMULATIVE WT %

GRAIN SIZE (PHI UNITS)
Trench 2, Unit Bca

Grain Size (Phi Units)

Cumulative wt %
Trench 2, Unit Clca

GRAIN SIZE (PHI UNITS)

Weight %

Cumulative wt %

GRAIN SIZE (PHI UNITS)
Trench 2, Unit C2ca

GRAIN SIZE (PHI UNITS)

GRAIN SIZE (PHI UNITS)
Trench 2, Unit Blb

GRAIN SIZE (PHI UNITS)
Trench 2, Unit B2b

GRAIN SIZE (PHI UNITS)

CUMULATIVE WT %

GRAIN SIZE (PHI UNITS)
Trench 2, Unit B3b

GRAIN SIZE (PHI UNITS)

Cumulative wt.

GRAIN SIZE (PHI UNITS)
Trench 2, Unit B4b

GRAIN SIZE (PHI UNITS)

GRAIN SIZE (PHI UNITS)
Trench 2, Unit Cox

![Graph 1: Weight % vs. Grain Size (Phi Units)]

![Graph 2: Cumulative WT % vs. Grain Size (Phi Units)]
Trench 2, Altered PPo

GRAIN SIZE (PHI UNITS)

CUMULATIVE WT %

GRAIN SIZE (PHI UNITS)
Trench 3, 0-26 cm
Trench 3, 26-46 cm

GRAIN SIZE (PHI UNITS)

GRAIN SIZE (PHI UNITS)
Trench 3, 46-73 cm

GRAIN SIZE (PHI UNITS)

CUMULATIVE Wt. %

GRAIN SIZE (PHI UNITS)
Trench 3, 73-112 cm

GRAIN SIZE (PHI UNITS)
Trench 3, 112-160 cm

GRAIN SIZE (PHI UNITS)
Trench 3, 160-192 cm

GRAIN SIZE (PHI UNITS)

GRAIN SIZE (PHI UNITS)
Trench 3, 192-240 cm

GRAIN SIZE (PHI UNITS)
Trench 3, 240-272 cm

GRAIN SIZE (PHI UNITS)

Cumulative wt %
Trench 3, 272-339 cm

GRAIN SIZE (PHI UNITS)

GRAIN SIZE (PHI UNITS)
Trench 3, 339-380 cm

GRAIN SIZE (PHI UNITS)
Trench 3, 380-424 cm

GRAIN SIZE (PHI UNITS)

Cumulative weight %

GRAIN SIZE (PHI UNITS)
Appendix C. Weight % of Acid-Insoluble Residues in Samples From Trenches 2 and 3.
### Trench 2 Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Initial Wt. Weight (gms)</th>
<th>Digested Wt. Weight (gms)</th>
<th>Weight % Insolubles</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bca</td>
<td>106.4</td>
<td>91.8</td>
<td>86.3</td>
</tr>
<tr>
<td>B2b</td>
<td>85.0</td>
<td>81.7</td>
<td>96.1</td>
</tr>
<tr>
<td>Cox</td>
<td>104.3</td>
<td>32.6</td>
<td>31.2</td>
</tr>
<tr>
<td>Altered PpO</td>
<td>106.7</td>
<td>46.9</td>
<td>44.0</td>
</tr>
</tbody>
</table>

### Trench 3 Samples

<table>
<thead>
<tr>
<th>Depth Range</th>
<th>Initial Wt.</th>
<th>Digested Wt.</th>
<th>Weight % Insolubles</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-26 cm</td>
<td>63.8</td>
<td>52.8</td>
<td>82.8</td>
</tr>
<tr>
<td>26-46 cm</td>
<td>110.5</td>
<td>88.0</td>
<td>79.6</td>
</tr>
<tr>
<td>46-73 cm</td>
<td>121.7</td>
<td>90.7</td>
<td>74.5</td>
</tr>
<tr>
<td>73-112 cm</td>
<td>129.8</td>
<td>97.3</td>
<td>75.0</td>
</tr>
<tr>
<td>112-160 cm</td>
<td>120.2</td>
<td>91.4</td>
<td>76.0</td>
</tr>
<tr>
<td>160-192 cm</td>
<td>81.7</td>
<td>69.3</td>
<td>84.8</td>
</tr>
<tr>
<td>192-240 cm</td>
<td>84.4</td>
<td>69.4</td>
<td>82.2</td>
</tr>
<tr>
<td>240-272 cm</td>
<td>82.5</td>
<td>66.6</td>
<td>80.7</td>
</tr>
<tr>
<td>272-339 cm</td>
<td>97.5</td>
<td>82.8</td>
<td>85.0</td>
</tr>
<tr>
<td>339-380 cm</td>
<td>73.3</td>
<td>60.3</td>
<td>82.3</td>
</tr>
<tr>
<td>380-424 cm</td>
<td>57.3</td>
<td>50.5</td>
<td>88.1</td>
</tr>
</tbody>
</table>
Appendix D. Mineralogy of Acid-Insoluble Residues of Selected Samples From Trenches 2 and 3.
**Trench 2 Samples**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Quartz</th>
<th>Plagioclase</th>
<th>Illite &amp; Micas</th>
<th>Kaolinite</th>
<th>Smectite</th>
<th>Chlorite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bca</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>0</td>
</tr>
<tr>
<td>Blb</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>0</td>
</tr>
<tr>
<td>Cox</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>0</td>
</tr>
<tr>
<td>Altered PPo</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>0</td>
</tr>
</tbody>
</table>

X  Mineral is present.

?  Mineral may be present in small amount.

0  Mineral is not present.
Trench 3 Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Quartz</th>
<th>Plagioclase</th>
<th>Illite &amp; Micas</th>
<th>Kaolinite</th>
<th>Smectite</th>
<th>Chlorite</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-26 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>26-46 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>46-73 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>73-112 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>?</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>112-160 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>160-192 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>192-240 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>240-272 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>272-339 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>339-380 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
<tr>
<td>380-424 cm</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>O</td>
</tr>
</tbody>
</table>

X  Mineral is present.

?  Mineral may be present in small amount.

O  Mineral is not present.