IN THE FOOTSTEPS OF G.K. GILBERT--

LAKE BONNEVILLE AND NEOTECTONICS OF THE EASTERN BASIN AND RANGE PROVINCE

GUIDEBOOK FOR FIELD TRIP TWELVE

THE GEOLOGICAL SOCIETY OF AMERICA
100th Annual Meeting, Denver, Colorado
October 31-November 3, 1988

UTAH GEOLOGICAL AND MINERAL SURVEY
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Editor's Note: The papers in the Guidebook were solicited by the organizers of this field trip for the Geological Society of America's Annual Meeting and have been edited and given a common format. However, their style and content have not been formally reviewed by the Utah Geological and Mineral Survey.

Cover illustration: Gilbert, 1890, plate XLIII
Frontispiece: U.S.G.S. Photographic Library Portrait 129, G.K. Gilbert

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GUIDEBOOK FOR FIELD TRIP TWELVE

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CONTENTS

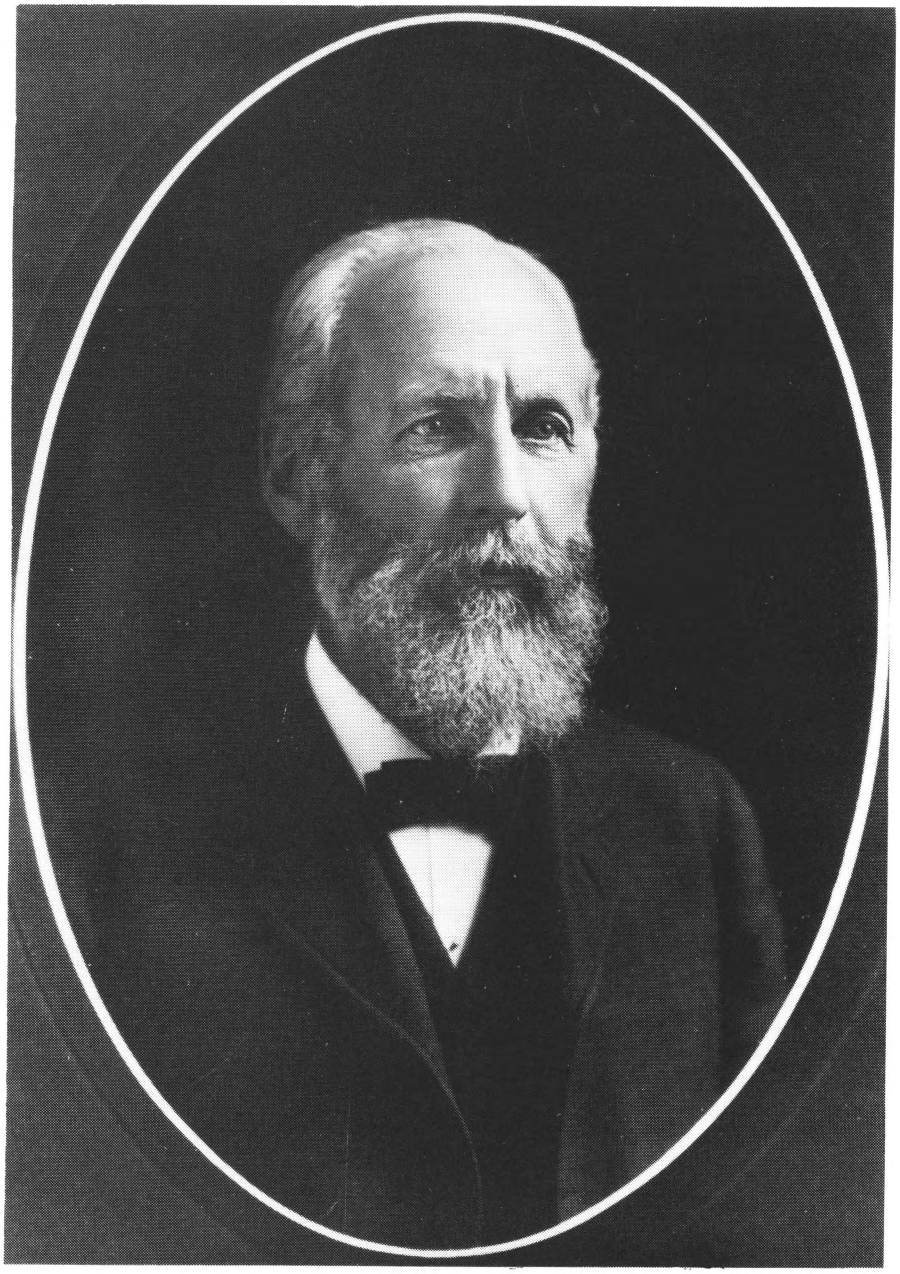

Figure 1. *G.K. Gilbert. U.S. Geological Survey Photographic Library Portrait 205 (no date).*

PROLOGUE: GROVE KARL GILBERT, EDUCATOR BY EXAMPLE

by

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INTRODUCTION

G.K. Gilbert (figure 1) is widely regarded as the greatest American geomorphologist. Gilbert lived from 1843 to 1918, a period of tremendous advancement in the earth sciences. A significant part of the progress made during this period is directly attributable to Gilbert's contributions. The four scientific papers most often cited as Gilbert's best are **Report on the Geology of the Henry Mountains** (1877), **Lake Bonneville** (1890), **Hydraulic-Mining Debris in the Sierra Nevada** (1917), and the posthumously published **Studies of Basin-Range Structure** (1928). Three of these papers concern Utah field areas. The reports on Lake Bonneville and basin-range structure deal specifically with the principal topics of interest in this guidebook--lacustrine and neotectonic features in the eastern Basin and Range physiographic province, western Utah (figure 2).

GILBERT'S BACKGROUND

Three earth scientists, Mendenhall (1920), Davis (1927), and Pyne (1980), have written the most comprehensive biographies of Gilbert. According to these authors Gilbert was physically frail during his youth, which he spent near the shores of Lake Ontario in Rochester, New York. He graduated from the University of Rochester in 1862 after completing a curriculum that emphasized Greek, Latin, mathematics, and English; he apparently took only one course in geology. Gilbert later confided to a colleague that while in college he had been much more interested in mathematics and engineering than geology (Mendenhall, 1920, p. 29).

Following graduation Gilbert accepted a public school teaching job in Michigan, but resigned before the end of the academic year. After five years of museum work in the Ward Natural Science Establishment (Cosmos Hall) at the University of Rochester, Gilbert secured his first geology position in 1869 as a volunteer for the Ohio State Survey. In 1871 the director of the Ohio Survey, J.S. Newberry, recommended Gilbert for the geological assistant position on Wheeler's U.S. Geographical Surveys West of the One-Hundredth Meridian. Three years later, Gilbert joined the Powell Survey. Upon formation of the U.S. Geological Survey in 1879, Gilbert became head of the Great Basin Division. From 1889 to 1892 he served the Survey as Chief Geologist, and may have declined the directorship during this period (Mendenhall, 1920, p. 38). Besides his professional geology positions, Gilbert was a founder and the only person to be twice elected president of the Geological Society of America.

Although for almost 50 years Gilbert was employed as a geologist, he may also be considered a geographer and even an engineer (Pyne, 1979, p. 226) because he worked at the interface of these three disciplines. Andrews (1920, p. 60 and 68), Mendenhall (1920, p. 35), Davis (1927, p. 37), and Hunt (1980, p. 46) called Gilbert a geographer or physiographer. The title geographer is additionally merited because Gilbert helped establish the Association of American Geographers in 1904, and served as its president in 1908. In his presidential address to that organization, Gilbert (1909a, p. 121) referred to himself as a geologist and geographer. Gilbert also served on the National Geographic Society board of managers from 1891 to 1900, excluding 1896 and 1897 when he was its vice president. He was elected to membership in the Royal Geographical Society of London, the Royal Scottish Geographical Society

Figure 2. Vertical aerial photograph of the southwestern portion of the House Range and adjacent piedmont, Millard County, Utah (north is to the left). A piedmont fault scarp (f) and Lake Bonneville shorelines (s) are visible in geomorphically inactive sections of the alluvial piedmont. Gilbert (1928) studied the House Range as an example of basin-range structure. Photo by Eros Data Center.

of Edinburgh, and the Geographical Societies of Berlin, Leipzig, and Geneva (Davis, 1927, p. 294). In 1893 he taught physical geography at Columbia University (Davis, 1927, p. 195). The only textbook that Gilbert wrote was a high school physical geography text, which he coauthored with A.P. Brigham (Gilbert and Brigham, 1902).

GILBERT'S INFLUENCE IN AMERICAN GEOMORPHOLOGY

Gilbert's contemporaries, both in North America and abroad, held him in very high esteem (Chamberlin, 1918, p. 376; Andrews, 1920, p. 60-61; Mendenhall, 1920, p. 42; Davis, 1927, p. 38 and 292; Gilluly, 1963, p. 218). Modern geomorphologists respect him to at least the same degree. This is evidenced, for example, in edited volumes written in tribute of Gilbert (Yochelson, 1980; this volume) and in the recent establishment of the G.K. Gilbert award for excellence in geomorphic research by the Geomorphology Specialty Group of the Association of American Geographers. However, from approximately early to mid-20th century, American geomorphologists generally neglected the contributions of Gilbert and concentrated on those of W.M. Davis. Eventually, many geomorphologists became dissatisfied with the constraints imposed by working within the Davisian framework (e.g., Strahler, 1952; Hack, 1960; Chorley, 1962). For an alternative approach they looked back to the masterful work of Gilbert. His style of process-oriented geomorphology is now considered to be the dominant model for American geomorphic research.

In searching for reasons why the work of this genius of American geomorphology was eclipsed for a time by the Davisian Geographical Cycle, modern researchers have focused on the differences between these two men. Keeping in mind that Gilbert and Davis were not adversaries, modern authors have tended to dichotomously contrast (1) Davis the academician with Gilbert the nonacademic investigator (Chorley and Beckinsale, 1980, p. 131; Pyne, 1980, p. 165), and (2) Davis's evolutionary concept of landscape development with Gilbert's use of physics to analyze landforms as the manifestation of balance of forces (Hack, 1960, p. 81; Pyne, 1975, p. 291; Chorley and Beckinsale, 1980, p. 130). As a way of expressing the second comparison, Chorley (1962, p. B2-B4) further categorized the Davisian and Gilbertian geomorphic approaches as closed- and open-systems thinking, respectively. Organizing the roles and contributions of Gilbert and Davis into antithetical categories helps us understand their differences. However, such generalizations are valuable only as long as we keep in mind that they greatly simplify reality.

By projecting generalizations concerning Gilbert and Davis onto the backdrop of prominent 20th century intellectual trends, modern researchers have attempted to explain the varying popularity of the geomorphic approaches of Gilbert and Davis (e.g., Chorley, 1965; Pyne, 1975; Chorley and Beckinsale, 1980). In the late 19th and early 20th centuries, the biological theory of evolution was applied by analogy to many disciplines, including soil science, ecology, cultural geography, and social science. Davis (e.g., 1884, 1889, 1899) adapted the fashionable life-cycle analogy to geomorphology. Besides the general popularity of evolutionary theories and historical approaches, in being nonquantitative the Geographical Cycle appealed to a wide audience, and therefore had a certain advantage over Gilbert's technique of analyzing forces. In addition, as a professor at Harvard University, Davis occupied a position from which he could champion his geomorphic approach to his peers and influence numerous students. Through his considerable pedagogic skill, Davis developed a cadre of ardent disciples. Even Gilbert (1905, p. 29), a firm believer in the value of analogies (1886, 1896a), indicated advantages to be gained from characterizing topographic stages with the terms youthful, mature, and senile. He further suggested that the appropriateness of these terms facilitated the widespread acceptance of Davis's geomorphic approach (1905, p. 29):

> *The aptness and familiarity make the terms permanently mnemonic, so that the use of any one of them brings to mind not only the sequence, but relative position within the sequence. Davis's generalization had such merit that it would probably have found eventual appreciation, whatever its mode of expression, but I think that the promptness and universality of its acceptance and assimilation were in large measure due to the felicity of the associated terminology.*

Gilbert (1905, p. 29) also admitted that people could violate the narrow limits of the analogy:

> *... humanistic analogy ... has sometimes been carried too far.... The stream valley resembles the human being in that from an early stage it evolves normally through a definite sequence of stages; and in most other respects the two differ.*

Whereas Davis's Geographical Cycle employed a favored intellectual trend of the early 20th century, Gilbert's tendency to apply physics principles to the surface of the earth was somewhat out of vogue (Pyne, 1975, p. 295; Baker and Pyne, 1978, p. 97; Pyne, 1979, p. 226). Moreover, in contrast with Davis, Gilbert did not expressly advocate his mechanistic approach. This was probably more a reflection of his modest personality (Andrews, 1920, p. 60; Pyne, 1975, p. 284) than the lack of an academic forum. Had Gilbert purposefully tried to convert people to his geomorphic approach, he could probably have done so in the nonacademic setting, employing to that end his influence at the U.S. Geological Survey, prominence in scientific societies, and published papers. In addition, had he wanted an academic forum he likely would have accepted the offers of a permanent faculty position made to him by Brown, Cornell, and perhaps other universities (Davis, 1927, p. 195).

GILBERT AS EDUCATOR

Although Gilbert never held a permanent university teaching position, he was nevertheless a teacher as well as an investigator. Gilbert apparently neither shunned teaching nor considered investigators and educators to represent mutually exclusive categories. For example, Gilbert (1886, p. 288) asserted that:

> *The investigator becomes an educator when in giving his work to the world he describes the route by which his end was reached. It is not denied that the publication of sound conclusions is in itself educational, but it is maintained that the publication of the concrete illustration of a good method is educational in a higher sense.*

Gilbert taught courses at Columbia and Johns Hopkins and lectured at numerous other universities. In describing the young Gilbert's first teaching experience in Michigan, Mendenhall (1920, p. 29) said that Gilbert "was neither happy nor successful as a teacher, not being equipped temperamentally to deal with those unruly pupils...." This does not necessarily translate to Pyne's (1980, p. 12) more abbreviated conclusion that the young Gilbert was "temperamentally unsuited to teach." In any case, people mature, and writing later of his Columbia University teaching experience the middle-aged Gilbert declared that he had enjoyed lecturing (Davis, 1927, p. 195). Gilbert also taught those who accompanied him in the field, worked with him in the office, listened to his presentations at scientific meetings, or read his reports. Gilbert's contemporaries described him as a teacher (Gregory, 1918, p. 129; Davis, 1927, p. 292; Andrews, 1920, p. 67) and as having disciples (Andrews, 1920, p. 67).

Gilbert continues to educate earth scientists in at least three general ways. We can learn (1) about the earth from his scientific papers, (2) from his views concerning research and education, and (3) by using him as a role model.

Gilbert greatly improved our understanding of a phenomenal range of geomorphic topics. For example, he contributed to our knowledge of fluvial geomorphology (e.g., 1877, 1914, 1917), coastal geomorphology (e.g., 1885, 1890), glacial geomorphology (e.g., 1903, 1906a, 1906b), hillslopes (e.g., 1877, 1909b), structure (e.g., 1876, 1928), faulting and earthquakes (e.g., 1907, 1909a, 1928), intrusive bodies (e.g., 1877), and isostasy (e.g., 1890, 1895). Gilbert also investigated problems in Quaternary geology, most notably the chronology of Pleistocene Lake Bonneville (e.g., 1882, 1890). Although his Lake Bonneville work stressed the geomorphic evidence, few people who have studied Monograph 1 (Gilbert, 1890)—or, for example, his Arkansas Valley groundwater reports (Gilbert, 1896b, 1897)—would agree with Pyne's (1979, p. 229) statement that "Gilbert...never practiced stratigraphy." Contrasting Gilbert's tendency to analyze forces with the historical-evolutionary approach that was popular in the early 20th century does not require us to overlook Gilbert's use of stratigraphy or to downplay his significant contributions to the historical geology of the Quaternary Period (figure 3).

Figure 3. The Gilbert Shoreline (G). Eardley and others (1957) named the youngest Pleistocene shoreline in the Bonneville basin for G.K. Gilbert in recognition of his outstanding contributions to the study of Lake Bonneville. The Provo Shoreline (P) and Stansbury Shoreline (S) are prominently displayed in the background on the Silver Island Range, Tooele County, Utah. Photograph by D.R. Currey, University of Utah.

Whereas primarily earth scientists would study the content of Gilbert's reports, modern students from most scientific disciplines can benefit from reading his principal statements concerning research and education (Gilbert, 1886, 1896a). The following brief summary of Gilbert's comments regarding this topic is drawn from his presidential address to the American Society of Naturalists (Gilbert, 1886).

Gilbert (1886, p. 284-290) described the teacher's role as consisting of two parts, storing students' minds with factual knowledge and training students to be investigators so that they can store their own minds. The educational process is most efficient if teachers store pupils' minds in a way that also trains them. In order to teach students how to investigate, educators should present examples of investigations, showing how they proceeded; investigators become teachers by example when they present their methods and reasoning along with their conclusions. In both cases it is important to emphasize the role of multiple working hypotheses and to illustrate how an investigation's "successes are achieved through series of failures" (Gilbert, 1886, p. 288). In order to explain the investigative procedure, Gilbert analyzed his own thinking (Gilluly, 1963, p. 222). As a stepping stone to understanding relationships between observed phenomena, scientists first group together similar phenomena. Researchers work to discover how the phenomena are interrelated by testing various hypotheses as to the nature of the relationships. It is most efficient and effective to test several hypotheses at once. The successful investigator, then, "must be fertile in the invention of hypotheses and ingenious in the application of tests" (Gilbert, 1886, p. 286). Is the ability to devise hypotheses and tests an innate creative talent or something that people can learn? In Gilbert's view, hypotheses are generated by analogy from similar relationships. Therefore, we can improve upon our given amount of inherited ability by storing our minds with knowledge and through practice. With this notion, Gilbert reassures not only young investigators who wonder if they have the talent to be successful, but also established scholars who worry if they can maintain their success. Gilbert did not originate all of these ideas (Gilbert, 1886, p. 284; Gilluly, 1963, p. 220-222), and the role of analogy in generating hypotheses has been questioned (e.g., Kitts, 1980). Nevertheless, he very skillfully assembled his views on education and investigation into a clear statement that has practical appeal to a wide audience.

The third way in which we can continue to learn from Gilbert is to take his advice regarding education by example, but choose him for our example of the consummate scientific investigator. Gilbert (e.g., 1886, 1890, 1896a, 1904) provided some very effective examples of the method of multiple working hypotheses. After carefully describing various hypotheses in an impartial way, he subjected them to exhaustive tests. If a single hypothesis remained, he tended to accept it cautiously, as a hypothesis that has not yet been disproven; if all the hypotheses are rejected, it indicates the need for more observation, hypothesizing, and testing. In addition to his methods, praise abounds for Gilbert's consistently clear and straightforward writing style (e.g., Chamberlin, 1918, p. 375-376; Mendenhall, 1920, p. 42; Davis, 1927, p. 198; Chorley and Beckinsale, 1980, p. 133; Pyne, 1980, p. 102; White, 1980, p. 15), and Hunt (1959) has urged field scientists to emulate Gilbert's report-writing techniques. In research as well as in written communication Gilbert tended to be comprehensive and thorough. His work reveals integrity and an effort to minimize egotism (e.g., Gilbert, 1886, 1896a, 1904). He was generous in giving credit to others (Davis, 1927, p. 63) and avoided controversy (Merriam, 1919, p. 392). Perhaps by following Gilbert's excellent example, as preserved in his written works, we can approach Gilbert's skill in some of these areas.

CONCLUSIONS

Gilbert was clearly well respected in his own time, but his technique of solving geomorphic problems by analyzing physical forces is more popular today than in the early 20th century when Davisian geomorphology overshadowed his contributions. Besides the popularity of historical-evolutionary analogies in that era, the mechanics approach to geomorphology may have been eclipsed because Gilbert did not champion it, he merely used it. Gilbert was not on the bandwagon of that era. The staying power of his contributions illustrates the importance of pursuing individual interests and using intellectually satisfying techniques, regardless of their popularity. The continued significance of Gilbert's work also illustrates the value of working at the disciplinary interface between geology, geography, and engineering.

According to Gilbert (1886, p. 288), investigators teach by communicating to others the conclusions of their work, but they educate in a higher sense when they state their methods and reasoning along with their conclusions. Presenting the steps that led to the conclusion not only shows what alternative hypotheses have been tested, it also educates by example. Gilbert both communicated results and detailed his methods and reasoning. He also analyzed his methods, formulated his views on the origin of hypotheses, and communicated them to others. Modern earth scientists can learn from his scientific work, from his views concerning research and education, and by following his example.

Several writers have reviewed the technical and methodological aspects of Gilbert's work and have noted the value of his excellent example (e.g., Hunt, 1959; Hack, 1960; Gilluly, 1963; Baker and Pyne, 1978; Pyne, 1978; Yochelson, 1980). Biographies, memoirs, and other secondary sources often provide useful interpretations and fresh insights concerning Gilbert. However, to some extent, secondary sources suffer as well as benefit from generalization. Especially in Gilbert's case, one should not underestimate the value of reading the original.

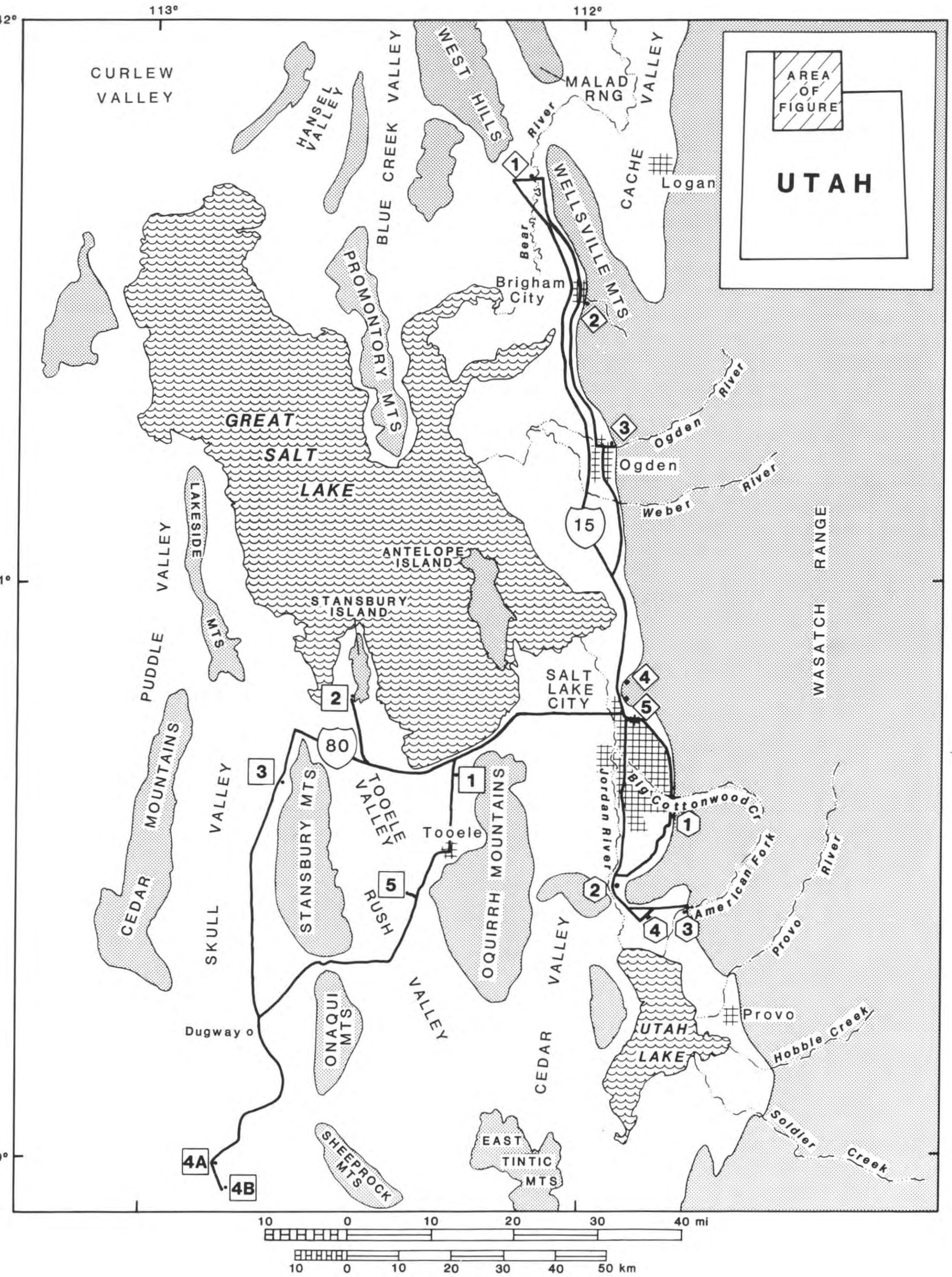

Figure 1. *Index map for field trip. The route and stops are shown by the following symbols: day 1, diamonds; day 2, squares; day 3, hexagons.*

FIELD TRIP INTRODUCTION: A Brief Review of Research on Lake Cycles and Neotectonics of the Eastern Basin and Range Province

by

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The Geological Society of America's 100th Annual Meeting, being held in Denver, Colorado this year, has a Centennial theme that is carried forth in symposia and field trips. The trips emphasize and revisit areas and topics covered by the great western surveys of the late 19th century, including the Powell, Wheeler, and Hayden Surveys, and the early pioneer geologists who assembled many of the basic geologic frameworks we now take for granted. We seize this opportunity to focus on the work of G.K. Gilbert, whose basic field research in the eastern Basin and Range Province has provided a framework for the past 100 years of research on Lake Bonneville and the neotectonics of the eastern Basin and Range Province.

Our field trip consists of three one-day excursions from Salt Lake City--the first day north, the second day west, and the third day south (figure 1)--to revisit some of G.K. Gilbert's classic stratigraphic localities, to discuss data from several new sites that have modified his original work, and to examine his work in light of modern stratigraphic concepts and a regional tectonic framework (see table 1). The trip emphasizes three topics that Gilbert studied a century ago: the geomorphology of Lake Bonneville deposits (shorelines, bars, spits, and deltas), the stratigraphy of the lacustrine cycles of Lake Bonneville, and neotectonics of the eastern Basin and Range Province, especially the Wasatch fault zone.

Although this field trip log and the papers in the guidebook address some of the many aspects of research conducted on these subjects in the past 35 years, our main focus will be on G.K. Gilbert, his pioneering studies of a century ago, and his legacy to contemporary earth scientists.

The discussions of stops on this field trip are organized by day and stop number, and are collected at the end of the road log for each day. For example, the discussion of the Cutler Dam Alloformation by Oviatt and McCoy is referenced as "day 1-stop 1," and follows the road log for day 1. In addition to the field trip stops, papers at the end of the guidebook discuss various aspects of Gilbert's work that we cannot witness on the trip. These papers are referenced by authors, for example Van Horn and Varnes, this volume.

Table 1. SUMMARY OF FIELD TRIP STOPS

(Abbreviations: WFZ, Wasatch fault zone; Bnv., Bonneville; l.c., lake cycle; fms, formations)

Field trip locality	Lake Bonneville stratigraphy	Lake Bonneville geomorphology	Neotectonic features
DAY 1 (to the north):			
Stop 1	Cutler Dam lake cycle		
Stop 2		Box Elder delta	WFZ at Brigham City trenching site
Stop 3			WFZ at East Ogden trenching site
Stop 4A	Transgressive Bnv. lake cycle		
Stop 4B	Alpine, Bonneville, and Draper fms.		
Stop 5 (optional)			Holocene faulting at Becks Hot Springs
DAY 2 (to the west):			
Stop 1			Faulting along Oquirrh Range
Stop 2	Early transgressive Bnv. lake cycle	Stansbury shoreline	
Stop 3			Faulting along Stansbury Range
Stop 4A	White Marl Bluff at Old River Bed		
Stop 4B	The Shutoff at Old River Bed	Old river channel and gravel bars	
Stop 5	Late transgressive Bnv. lake cycle	Stockton Bar, bay-head barrier	
DAY 3 (to the south):			
Stop 1A	Glacial/lacustrine history	Glacial moraines, terraces, deltas	
Stop 1B			Fault history at Little Cottonwood and Bells Canyons
Stop 2	Bnv. l.c., Little Valley l.c.	Point of the Mtn. spit	
Stop 3	Keg Mountain oscillation	Bonneville fan-delta complex	Faulting at American Fork Canyon
Stop 4	Complete? Bnv. lake cycle	Provo fan-delta complex	Paleoliquefaction of Bnv. l.c. sediment

A SUMMARY OF RESEARCH ON LAKE CYCLES IN THE BONNEVILLE BASIN

Beginning with G.K. Gilbert, investigators have sought evidence for lake cycles that preceded the last (that is, the Bonneville) lake cycle, whose traces are prominently displayed on the landscape of the northeastern Great Basin. They used a variety of techniques to identify and correlate deposits of older lake cycles and were influenced by prevailing concepts of changes in climate during the Quaternary. These approaches have led to quite different interpretations of lake history and chronology. In this part of the Introduction, we trace the development of ideas about older lake cycles and end with a discussion of contemporary methods for identifying and correlating deposits of different lake cycles. These methods are well illustrated at two field-trip stops: at day 1-stop 1 (Bear River) where deposits of an intermediate-level lake cycle are differentiated from the Bonneville and Little Valley lake cycles and at day 3-stop 2 (Point of the Mountain) in southern Salt Lake Valley where deposits of the last two deep-lake cycles and an interlacustrine episode are well exposed in an active gravel quarry.

DEVELOPMENT OF CONCEPTS

Research aimed at unraveling the Quaternary lacustrine history of the Bonneville basin may be divided into four eras that encompass three generations of workers and span almost 115 years of scientific research.

1. The reconnaissance studies of the late 1800s, which include mainly the work of G.K. Gilbert, one of the few investigators to study the entire basin. Little detailed work was conducted in the following half century.
2. The studies between 1945 and 1960, which consisted of detailed (1:62,500 scale) geologic mapping in several parts of the Bonneville basin by C.B. Hunt and associates from the U.S. Geological Survey, and by scientists from several Utah universities.
3. The detailed stratigraphic work of R.B. Morrison and others during the 1960s, which was greatly influenced by new methods of stratigraphic subdivision, radiometric dating, qualitative pedology, and correlation with Quaternary climatic records derived from other geologic studies.

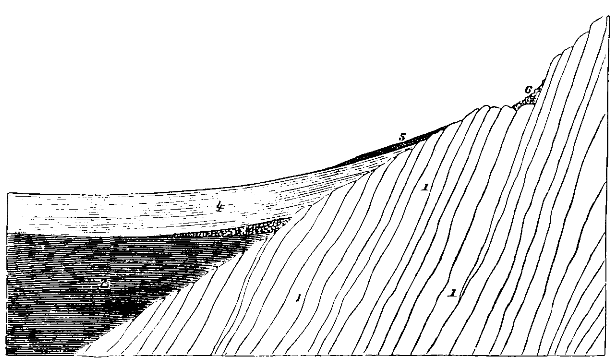


Figure 2. Relations near Leamington showing two sequences of lake deposits separated by a wedge of alluvial gravel (from Gilbert, 1890, fig. 28).

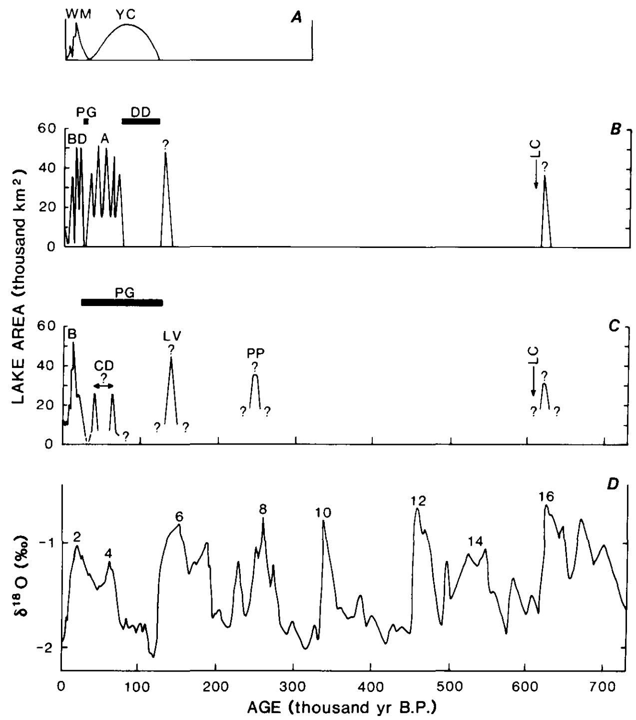


Figure 3. Selected reconstructions of the lake cycles in the Bonneville basin: A, Gilbert (1890; fig. 34); B, Morrison (1965c); C, combined from Scott and others (1983), Currey and Oviatt (1985), McCoy (1987), and Oviatt and others (1987). Black bars represent times of formation of the following geosols; PG, Promontory; DD, Dimple Dell. Lake cycles: WM, White Marl; YC, Yellow Clay; BD, Bonneville-Draper; A, Alpine; B, Bonneville; CD, Cutler Dam; LV, Little Valley; PP, Pokes Point. LC, Lava Creek ash bed. For comparison, we have included the marine oxygen-isotope record (D) from core V28-238 of Shackleton and Opdyke (1973) as inverted (vertically) from Imbrie and Imbrie (1980; fig. 10B). Numbered peaks are inferred glacial periods.

4. After a hiatus of about 10 years, another generation of scientists initiated research that relied on new dating techniques and correlation methods, as well as the application of Quaternary geologic and climatic concepts to Lake Bonneville's history.

GILBERT ERA

Recognition of separate lake cycles in the Bonneville basin began in 1879 with Gilbert's discovery of a wedge of alluvial gravel that he thought separated the Yellow Clay from the White Marl in the Old River Bed area (Hunt, 1982a, p. 128; day 1-stops 4A and 4B). In 1890, Gilbert published the following conclusions on the basis of stratigraphic studies at the Old River Bed and near Leamington (figure 2), and from the study of shoreline deposits.

1. The epoch of Lake Bonneville followed a long period of low water during which large alluvial fans were built (sic, throughout the basin; figure 3A).

   Gilbert left open the possibility that deposits of older lake cycles were buried beneath the fans, but concluded that they were very old compared with the length of Lake Bonneville time (Gilbert, 1890, p. 220-221).
2. The first cycle of Lake Bonneville is represented by the Yellow Clay and by some of the intermediate shorelines (day 1-stop 4A), which lie at elevations between the Provo and Bonneville shorelines.

The lake surface rose to a level about 90 feet (27 m) below the later Bonneville shoreline as determined from the altitude of shoreline deposits in Preuss (Wah Wah) Valley that Gilbert interpreted as being of the age of the Yellow Clay. Figures 4A and 4B show the relations between deposits of these shorelines. "Having learned from the sediments ... that the water rose at least twice from the lower to higher parts of the basin, besides undergoing many minor oscillations, it was not difficult to see that certain of the shore embankments were referable to an earlier flood than certain others. The most important locality ... shows a series of curved bars (bbbb), overlapped by a series of spit-like embankments massed together into a few sloping terraces (tttt). The source of the shore drift was at the north, and the beaches which conveyed it to the curved bars are hidden by the later embankments. It would be impossible for the bars to originate under the lee of spits. Moreover, the spits everywhere exhibit their gravelly constitution, but the curved bars are half buried by lake deposits" (Gilbert, 1890, p. 151-152). Gilbert concluded that the bars which we now call barriers were formed during the Yellow Clay lake cycle and the spits during the White Marl lake cycle.

Reasons for the marked difference in the shoreline features of the two cycles, small bars during the earlier and large embankments during the later, are not discussed. We think a more plausible explanation is that, at a given lake level, a bar would form along the shoreline and, for reasons of coastal shape and wave, and current conditions, a spit would begin to form to the north of the places marked by "t." The spit would prograde to the southwest and ultimately block the shore drift to the bar in a manner analogous to a growing baymouth bar cutting off the sediment supply to a bay-head bar. The example from Preuss Valley is complicated because the lake continued to rise and some of the bars shown in figure 4A were overlapped by younger spits that relate to a higher lake level; however, all the bars need not predate all the spits in the manner envisioned by Gilbert. Whether deposits of two separate lake cycles are present in this sequence cannot be resolved strictly from their geomorphic relations.

3. A period of low water following deposition of the Yellow Clay is represented by an erosional unconformity, locally marked by alluvial gravel (Gilbert, 1890, p. 192, p. 194-196; see day 2-stops 4A and 4B).
4. During the younger lake cycle, many of the intermediate shorelines were formed and the White Marl was deposited as the lake rose to the Bonneville shoreline (Gilbert, 1890, p. 170 and 198).
5. Climate changes caused the lake to fluctuate and glaciers should have advanced and retreated as lake levels rose and fell (Gilbert, 1890, p. 316-318).

To help place Gilbert's contributions in perspective, one should note that his field work and writing of reports on Lake Bonneville (1871-1890) coincided with development of ideas about multiple glaciation in the midwestern United States and the tracing of two drift units over large areas of the country by T.C. Chamberlin and his colleagues (Flint, 1965, 1971; Frye and others, 1965). Identification of multiple glaciations in the Midwest relied on such evidence as nonglacial sediments or fossil-bearing beds lying between two till deposits and differences in the degree of weathering of adjacent drift units. The recognition of multiple oscillations of closed-basin lakes lagged behind glacial studies; the only evidence came from inferences about tufa formation in the Lahontan basin (King, 1878) and from Russell's (1885) discovery of two sequences of Lake Lahontan beds separated by alluvial deposits. Although the latter discovery occurred after Gilbert completed his field

A

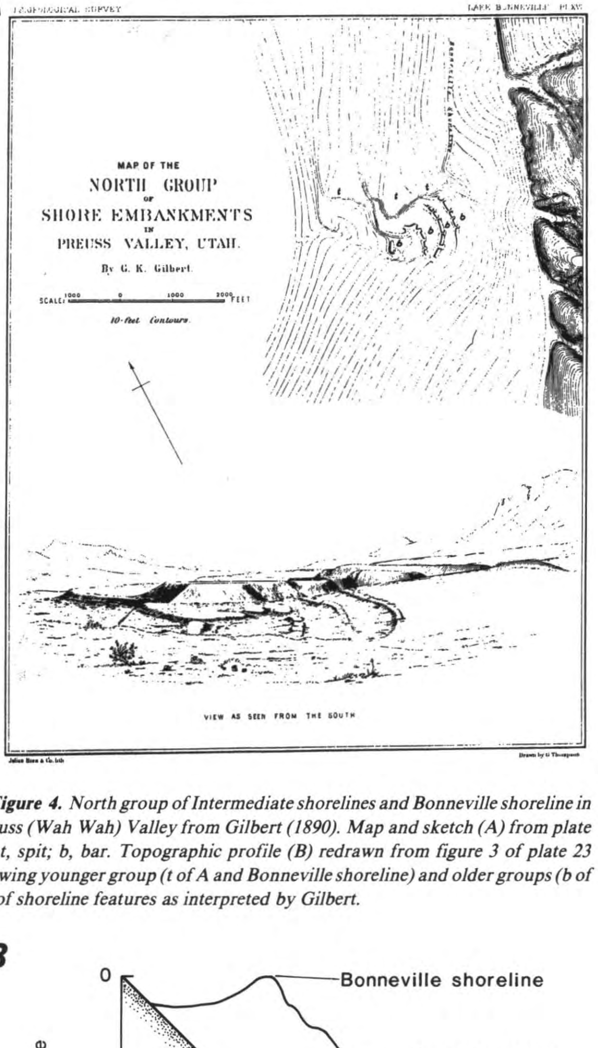
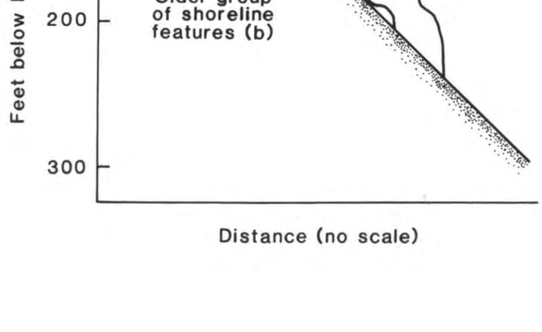


Figure 4. *North group of Intermediate shorelines and Bonneville shoreline in Preuss (Wah Wah) Valley from Gilbert (1890). Map and sketch (A) from plate 16; t, spit; b, bar. Topographic profile (B) redrawn from figure 3 of plate 23 showing younger group (t of A and Bonneville shoreline) and older groups (b of A) of shoreline features as interpreted by Gilbert.*

B



investigations, Russell had assisted Gilbert in the Lake Bonneville area and was instrumental in obtaining some of the evidence of older lake cycles (see day 3-stop 3). Thus, Gilbert was certainly aware of environmental changes during the Quaternary and thought he had evidence for separate lake cycles at the Old River Bed and Leamington. But, this evidence was less compelling than the weathering and fossil evidence used at that time for differentiating drifts and interpreting multiple glaciations in the Midwest.

Little detailed study of Lake Bonneville deposits occurred during the first half of the 20th century. Atwood (1909) mapped glacial deposits of two ages in the Wasatch Range, and believed that the younger deposits were contemporaneous with the younger (White Marl) cycle of Lake Bonneville. In contrast, Blackwelder (1931) concluded that the younger moraines at the mouths of Little Cottonwood and Bells Canyons (day 3-stop 1) were deposits of an older (Tahoe) glaciation that preceded the White Marl lake cycle. Atwood's view was subsequently resurrected in the mid-1970s (McCoy, 1977; Madsen and Currey, 1979; Scott and others, 1983).

HUNT ERA

In 1946, detailed mapping of Lake Bonneville deposits and related sediments was initiated by C.B. Hunt in northern Utah Valley (Hunt and others, 1953). This work was accompanied over the next decade by the mapping and stratigraphic studies of many other geologists (Bissell, 1963; Eardley and others, 1957; Feth and others, 1966; Jones and Marsell, 1955; Varnes and Van Horn, 1961, 1984; Williams, 1962). Hunt and others (1953) followed the lake history developed by Gilbert and mapped deposits of two lake cycles; the older represented by the Alpine Formation and the younger by the Bonneville and Provo Formations. The Alpine Formation was considered equivalent in age to the Yellow Clay and the intermediate shorelines of Gilbert (1890).

Influenced by Gilbert's interpretation of the great duration of the older lake cycle, Hunt and others (1953) included a thick sequence of mainly fine-grained sediment in the Alpine Formation that is widely exposed at the surface between the Provo and Bonneville shorelines and in a few deep exposures where it underlies the Provo Formation (figure 5). However, by their usage the Bonneville Formation was restricted mostly to deposits of gravel and sand at the highest of the Bonneville shorelines. This interpretation differed from Gilbert's concept of the White Marl (Bonneville) lake cycle in which deposits from the transgressive phase (which includes sediment of many of the intermediate shorelines), as well as from the high stand (Bonneville shoreline), were identified through most of the altitudinal range of the lake.

Hunt and others (1953) did not identify a clear unconformity between the Bonneville and Alpine Formations. Bissell (1963) sought evidence for an unconformity between the formations in the southern Utah Valley, but found only local disconformities between gravel of the Bonneville Formation and silt and clay of the Alpine Formation. He described sparse evidence of a submature soil and thin subaerial deposits between them. Other workers at the time (Eardley and others, 1957; Feth and others, 1966; Jones and Marsell, 1955) also found little evidence of a break between Bonneville and Alpine Formations, and several of them mapped deposits between the Bonneville and Provo shorelines as undifferentiated Bonneville and Alpine formation. Williams (1962) and Bissell (1963), who were mapping in Cache Valley and southern Utah Valley (respectively) at the same time as Hunt and associates, also followed this strategy. Of William's interpretation, Hunt (*in* Eardley and others, 1957, p. 1166) says "He feels that the Bonneville formation represents the culmination of the Alpine, that the unconformity between the Bonneville and Alpine is merely a minor diastem due to overlap of coarse clastic onto fine grained sediments."

Thus at the conclusion of the Hunt era, deposits of two major lake cycles were widely recognized and mapped by some geologists, but others were unsure of the accuracy of this interpretation. Firm, widely accepted evidence of a break of interlacustral significance had not yet been found.

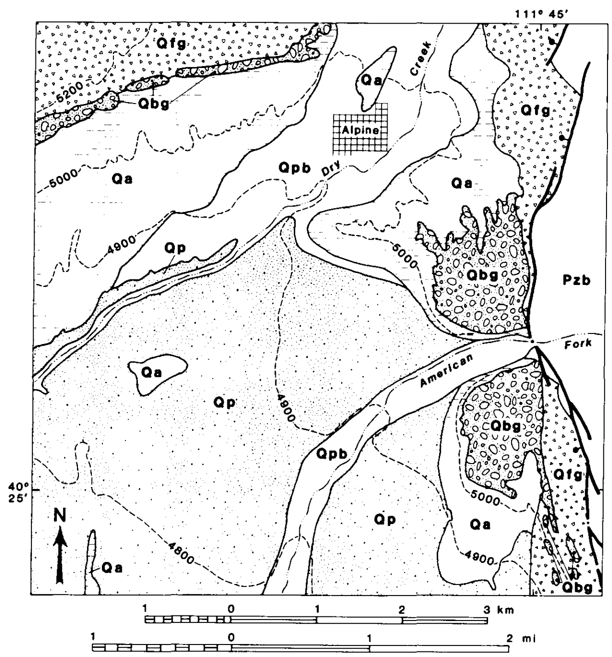


Figure 5. Map of the Alpine area in northern Utah Valley (simplified from Hunt and others, 1953). Qpb, post-Lake Bonneville alluvial and colluvial deposits; Qp, Provo Formation; Qb, Bonneville Formation; Qa, Alpine Formation; Qfg, pre-Lake Bonneville alluvial-fan deposits; Pzb, Paleozoic bedrock. Dashed lines are elevation contours, in feet (5100 ft contour and eastern part of 5200 ft contour are omitted for clarity).

MORRISON ERA

Research in the Bonneville basin during the late 1950s and 1960s relied heavily on radiocarbon dating (Broecker and Kaufman, 1965; Morrison, 1965a), on detailed stratigraphic and soil studies (Richmond, 1964; Morrison, 1965a, b), and on coring of the basin floor (Eardley and others, 1973). These techniques led to the development of dated stratigraphic sequences, and papers from this era stress correlations at regional to global scales.

Morrison's (1965a, b) studies in the eastern Jordan Valley and at Little Valley near Promontory Point added much detail to the record of lake history (figure 3B). His work and interpretations of sediment cores by Eardley and others

(1973) showed that numerous lake fluctuations had occurred through much of Quaternary time, and not just the two major cycles proposed by Gilbert and recognized by some during the Hunt era. In the large excavations at Little Valley, Morrison found marked unconformities between deposits of separate lake cycles; unconformities characterized by substantial disconformity, subaerial deposits, and well-developed buried soils.

Except for the evidence at Little Valley, mapping in areas along the Wasatch Front by Morrison (1965b) and Van Horn (1972) did not rely on marked unconformities between deposits mapped as Alpine and Bonneville. Bright (1963) searched for marked unconformities as a a basis for identifying Alpine deposits in northern Cache Valley, but found little evidence for them. After several decades of study by many geologists, we feel that the only undisputed older lake-cycle deposits exposed at the surface are those in Little Valley. They were found in an extensively quarried area that is characterized by a modest sedimentation rate. There is little wonder that investigators working along the Wasatch Front, where poor natural exposures and high sedimentation rates are typical, had trouble finding convincing evidence of older lake cycles.

PRESENT ERA

We perceive the present era of Lake Bonneville studies as beginning in the mid-1970s with the development of several new dating techniques, including amino-acid geochronology and quantitative soil studies, and with renewed efforts to better understand lake history through stratigraphic and geomorphic studies. Detailed geomorphic studies of the shorelines had largely been ignored since Gilbert's time. This era coincided with a period of rapid population growth along the Wasatch Front. Many quarries were opened, exposing ever-enlarging sections that could be followed over a period of years, and the large exposures at Little Valley were still intact. Thus, a degree of exposure was available that was unknown to previous workers, especially Gilbert, who appears to have had few extensive exposures to study other than those at the Old River Bed and along the Sevier River near Leamington (Van Horn and Varnes, this volume).

Two other major efforts in Quaternary studies greatly influenced work of the present era. First, the construction of a marine oxygen-isotope record (Shackleton and Opdyke, 1973) provided a general framework of climate change (figure 3D) that helped establish a perspective for the probable timing of expansions and contractions of Lake Bonneville. Second, concepts about the timing of Rocky Mountain glacial events were changing significantly (Pierce and others, 1976), which initiated a rethinking of the chronology of Lake Bonneville (Morrison, 1975).

During the past decade, concepts have changed about the timing of the last two deep-lake cycles and the character and duration of the intervening period of no lakes or low lake levels (Scott and others, 1982, 1983; Currey and Oviatt, 1985; McCalpin and others, 1987; several papers in this volume). These changes are summarized below and in figure 3C

1. Both the Yellow Clay and White Marl of Gilbert were deposited during the Bonneville lake cycle. Radiocarbon dating, aminostratigraphy, and recently developed facies models were needed to prove this hypothesis. Gilbert probably never saw deposits of an older lake cycle in section. Some intermediate shorelines are deposits of an older lake cycle, but he never saw the stratigraphic relations necessary to support such an interpretation.
2. As mapped by most workers along the Wasatch Front, much of the Alpine Formation is not separated from the overlying Bonneville Formation by a marked unconformity; both were deposited during the Bonneville lake cycle. The presesnt landscape provides a model for characterizing the unconformity one would expect to find between deposits of two lake cycles. Since the end of the Bonneville lake cycle, streams cut valleys into the previously smooth piedmont surfaces; fluvial, eolian, and colluvial sediment has been deposited over large areas of the piedmont; and new soils have formed across essentially the entire landscape. A rising lake would modify some of theses geologic features, but most should be preserved beneath a cover of new lacustrine sediment.

   The presence of a disconformity alone, even if accompanied by subaerial deposits, does not necessarily signify a major interlacustral break. Because Lake Bonneville was a closed-basin lake for all but a brief instant of its history, numerous second-order fluctuations would have occurred in all the lake cycles (Gilbert, 1890). These brief second-order fluctuations formed numerous disconformities, such as the one between the Yellow Clay and White Marl (see day 2-stops 4A and 4B). But where well exposed, few of these disconformities are traceable over altitude ranges of more than a few tens of meters (30-60 feet).

   If a buried soil is preserved (completely) between two lake deposits, the degree of soil development provides an estimate of the time interval separating the deposits, but it says nothing about the absolute age of deposits. Other techniques are needed to date or at least to determine time equivalence of deposits--techniques such as radiocarbon dating, aminostratigraphy, and thermoluminescence dating.
3. The Little Valley lake cycle, which replaces the term Alpine as the penultimate deep-lake cycle, probably correlates with oxygen-isotope stage 6 of the marine record, which culminated about 140,000 years ago. This correlation is made on the basis of estimates of the time needed to form the Promontory Geosol of the last interlacustral episode (see day 3-stop 2). The Bonneville lake cycle occurred between about 32,000 and 10,000 yr B.P., during oxygen-isotope stage 2. The Cutler Dam lake cycle is represented by an expansion of the lake to a relatively modest level (maximum altitude of 4400 ft, 1340 m) that occurred between the Bonneville and Little Valley lake cycles. By several lines of reasoning, the Cutler Dam lake cycle occurred during oxygen-isotope stage 4 (see day 1-stop 1).

A SUMMARY OF RESEARCH ON NEOTECTONICS AND THE WASATCH FAULT ZONE

After working along the Wasatch fault zone for a number of years, we have concluded that G.K. Gilbert's studies of faults, scarps, and earthquakes qualifies him as the *de facto* Father of

Paleoseismology. We believe that the following list of but a few of Gilbert's observations and conclusions support this title.

1. The topographic and structural relief of mountains is produced incrementally along range-bounding faults, which are evidenced by "piedmont scarps" (Wallace, 1980, p. 30).
2. Piedmont scarps, in turn, are evidence of former earthquakes. The morphological nature of the scarps (degree of rounding, height versus steepness, etc.) are indicators of the age of scarps (Wallace, 1980, p. 40) and, thus, the timing of the associated paleoearthquake.
3. He recognized that lack of scarps along an otherwise active fault system could suggest a large earthquake might eventually occur there (Wallace, 1980, p. 38). Within this observation lies the heart of modern research on seismic gaps and fault segmentation.
4. In 1883, in what might be considered an earthquake warning, Gilbert published a popular article describing the processes of mountain building, the constant accumulation of strain, and the likelihood of major earthquakes in the Salt Lake City area (Wallace, 1980, p. 38).
5. In 1884, he suggested that fault displacements caused earthquakes, which was a radical departure from the commonly accepted idea the faulting accompanied but didn't cause earthquakes. It was almost 25 years before Gilbert's concepts on earthquakes found a theoretical basis in Reid's application of elastic-rebound theory (Wallace, 1980, p. 38).
6. In his 1890 monograph on Lake Bonneville, he presented compelling evidence that the strength of the earth's crust could not support the load of water as the basin filled; it subsided, but recovered when the load was removed (Mabey, 1980, p. 67). In this observation, Gilbert realized one of the first real opportunities to study the dynamic behavior of the crust.
7. His observations of the position and characteristics of surface ruptures of the 1906 earthquake on the San Andreas fault remain among the best records of such features ever recorded (Wallace, 1980, p. 41). In 1908, Gilbert presented a simple and clear analysis of strike-slip faulting that included a recognition of the importance of uplift and tilting.
8. In 1909, he presented a balanced approach to the reduction of earthquake hazards by assessing earthquake risk and exposure. In addition, he evaluated the value of earthquake insurance and decried the policy of concealing of earthquake hazards from the public (Wallace, 1980, p. 38).

Gilbert's interest in young geologic structures and the processes of faulting (later to be known as neotectonics) was stimulated by his exploration of the West with the 1871 Wheeler Survey. The exceptional degree of exposure in the fault-bounded mountain ranges of Utah and Nevada obviously made quite an impression on Gilbert, having previously studied the folded terrain of the heavily vegetated Appalachians.

> *Indeed, I entered the field with the expectation of finding in the ridges of Nevada a like structure, and it was only with the accumulation of difficulties that I reluctantly abandoned the idea.*
> (Gilbert, 1875, p. 41).

This statement demonstrates one of Gilbert's fundamental strengths: his emphasis on facts and observations and the discovery of their relations (Wallace, 1980, p. 43).

Gilbert's investigations of structure in the Basin-Range System (later to become known as the Basin and Range Province) started in 1871 and continued until 1881, when he was transferred to Washington, D.C. for an administrative assignment. In 1900, he became embroiled in a heated debate with another member of the U.S. Geological Survey --Josiah Spurr--over the basic geologic structure of the basin ranges (Basin and Range Province). Spurr (1901) proposed that the basin ranges, like the Appalachians, consisted of folded mountains and were formed by compression. The fire rekindled, Gilbert returned to the field in the summer of 1901, this time to the House and Fish Springs Ranges of western Utah to prove his hypotheses. However, the maps and topographic records from this field session were destroyed in an unfortunate accident (see Pyne, 1980; Burstyn, 1984) and, sickened by the loss, Gilbert turned his attention to other scientific problems for the next decade.

In 1914 on his way to San Francisco, Gilbert stopped in Ogden, Utah for two weeks to study again the Wasatch fault zone (see day 1-stop 3). With his studies of hydraulic mining and sediment transport nearly complete, he turned his attention again to basin-range structure. Although incomplete, Gilbert's major and final paper on the Basin-Range System was published posthumously in 1928. With its publication, his debate with Spurr over the tectonic origin of the Basin-Range System was largely settled.

THE NEXT HALF CENTURY

The next half century, from 1917 when Gilbert last visited the Wasatch fault zone until about 1970, was marked by a number of significant studies of faulting in the eastern Basin and Range Province. However, none of these studies had the regional perspective or the far-ranging effects of Gilbert's 1928 Professional Paper.

In this period, focused studies of neotectonics in the eastern Basin and Range Province ranged widely in scope and nature, from general reconnaissance studies of the Wasatch fault zone (see Marsell, 1964) to studies of faults in natural and artificial exposures. Improved understanding of the location and timing of movement along the Wasatch fault zone and other faults resulted from the stratigraphic studies and geologic mapping of Hunt and others (1953), Bissell (1963), Williams (1963), and Morrison (1965b). In addition, detailed mapping, such as in the Salt Lake City area by Marsell and Threet (1964), identified strands of the Wasatch fault zone or faults that are entirely within the adjacent basins. But still, without the aid of dating tools (such as radiocarbon analysis) workers were frustrated in attempts to decipher the timing of individual movements on faults.

MODERN RESEARCH--THE ERA OF PALEOSEISMOLOGY

Although the pioneering studies of Gilbert (1884, 1890, 1928) provided the basic framework for study of the Wasatch fault zone and neotectonics of the region, the first modern studies of the fault zone were not completed until the 1970s (Cluff and others, 1970, 1973, 1974). These studies used low-sun-angle photography to map the surface trace of the Wasatch fault zone from Malad City, Idaho, to Gunnison, Utah. Their work was the first systematic investigation of the fault zone, but was not field checked and did not include stratigraphic data (such as mapping of geologic units).

As an extension of the reconnaissance, geologists at Woodward-Clyde Consultants conducted a series of detailed investigations of four trench sites along the Wasatch fault zone from 1978 to 1982. Their investigations resulted in many detailed reports and culminated in two major synthesis reports that marked a new era of research on ancient earthquakes and faulting. The study of these subjects has become known collectively as paleoseismology (see Crone and Omdahl, 1987). In the first report, Swan and others (1980) speculated on the number of possible segments that comprise the Wasatch fault zone; fault segments being the fundamental structural elements of long fault zones. They suggested at least 6 segments on the basis of historic microseismicity to as many as 10 segments on the basis of geometric variations along the fault zone and on a common rupture length of 18-25 miles (30-40 km) for normal faults worldwide. In the second report, Schwartz and Coppersmith (1984) proposed that the Wasatch fault zone has six major segments on the basis of a combination of geomorphic, geophysical, paleoseismic, and geodetic data. In addition, they presented strong evidence that large-magnitude earthquakes along the Wasatch fault zone are associated with a characteristic amount of displacement at any one site.

In 1984, the U.S. Geological Survey chose Utah's Wasatch Front Urban Corridor (Brigham City to Nephi) as the focus of research for the National Earthquake Hazards Reduction Program (NEHRP). During the past five years, the combined efforts of geologists, geophysicists, and seismologists has led to a better understanding of the processes, timing, and hazards associated with major earthquakes in the region (see Gori and Hays, 1987).

A major effort of the U.S.G.S.'s program was to map the surficial geology in detail along the Wasatch fault zone and to establish a firm, well-dated stratigraphic framework for assessing times and rates of faulting. For the most part, the mapping of the urban area has been completed by Personius (1988), Nelson and Personius (unpubl. mapping), Scott and Shroba (1985), and Machette (unpubl. mapping). These maps provide a uniform overview of the Quaternary geology along the Wasatch fault zone. The mapping has led Machette and others (1987) to suggest that the Wasatch fault zone is comprised of 10-12 fault segments; the increased number of segments comes from subdividing some of the long segments defined by Schwartz and Coppersmith (1984).

A NEW EFFORT TO EXPLORE THE HISTORY OF THE WASATCH FAULT ZONE

The Wasatch fault zone has been the focus of extensive study since the initial trenching investigations in the late 1970s (see Schwartz and Coppersmith, 1984). However, beginning in 1985, the U.S. Geological Survey and the Utah Geological and Mineral Survey joined forces to explore the Holocene history of the Wasatch fault zone at a number of critical sites. The cooperative team of principle investigators includes David Schwartz (U.S.G.S.-Menlo Park), William Lund (U.G.M.S.-Salt Lake City), Steve Personius, Alan Nelson, and Michael Machette (all U.S.G.S.-Denver). Including the studies completed by 1987, 25 new trenches and 3 natural exposures have been logged and described from 14 sites on 8 segments of the Wasatch fault zone (see Machette and others, 1987, table 1). Most of the trenches have provided some constraints on the time of most recent faulting and set limits on recurrence intervals and slip rates for faulting events. These constraints and limits are supported by about 60 radiocarbon dates (both conventional and accelerator methods on charcoal and soil organic matter) and 12 experimental thermoluminescence age estimates that have been obtained in the past two years.

SUMMARY OF TIMING AND RECURRENCE OF HOLOCENE MOVEMENT ON SEGMENTS OF THE WASATCH FAULT ZONE

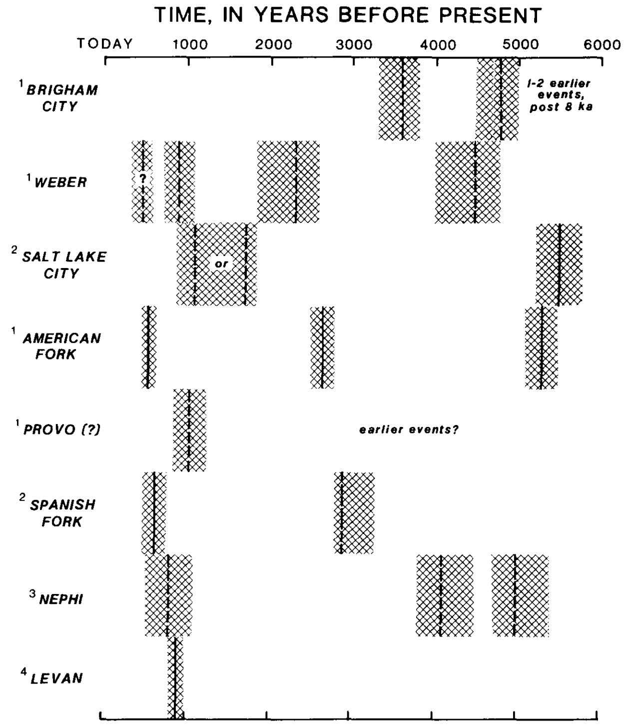
As a result of the new trenching investigations and geologic mapping, the Wasatch fault zone is probably now the most intensively studied Quaternary normal-slip fault in the world. Obviously a review of the new data is beyond the scope of this introduction, but a brief synthesis of the timing and recurrence of Holocene movement on the fault zone will help us appreciate Gilbert's hypotheses and conclusions about normal faulting.

The timing of major Holocene faulting events on the Wasatch fault zone is shown diagrammatically in figure 6. Several interesting patterns have evolved from these chronology. One gets a sense that the timing of faulting events was random prior to the flurry of events in the past 1200 years. Within the 1200-6000 year record, there seems to be a spacial difference in recurrence intervals; the southern segments (Salt Lake City to Nephi) have fairly long intervals, typically 2000-5000 years, whereas the Weber and Brigham City segments have intervals of less than 2000 years.

Secondly, there seem to be two periods of temporal clustering in the past 6000 years, the strongest cluster being in the past 1200 years. If one assumes that the most recent event on the Salt Lake City segments was about 1100 years ago (rather than 1800 years ago), then faulting has occurred in the past 1100 years on seven of the eight segments of the Wasatch that have Holocene movement. Although these relations point strongly to a process of temporal clustering on the Wasatch fault zone, the process seems to be active intermittently (that is, no clustering from 1200-4000 yr B.P.). In addition, a case can be made for an earlier (4000-5500 yr B.P.) episode of clustering, although it is not well defined nor are the timing of events well constrained).

CONCLUSIONS

G.K. Gilbert was a consummate observer of geomorphic, stratigraphic, and structural relations in the field. During the period from 1871 (when he was first introduced to the region while a member of the Wheeler Survey) until 1917 (when he made his last trip to study faulting), he traversed most of the Bonneville basin and the Wasatch fault zone. His 1890 U.S. Geological Survey Monograph and 1928 Professional Paper are widely regarded as his benchmark papers for this region, but he also published widely in scientific journals on topics as



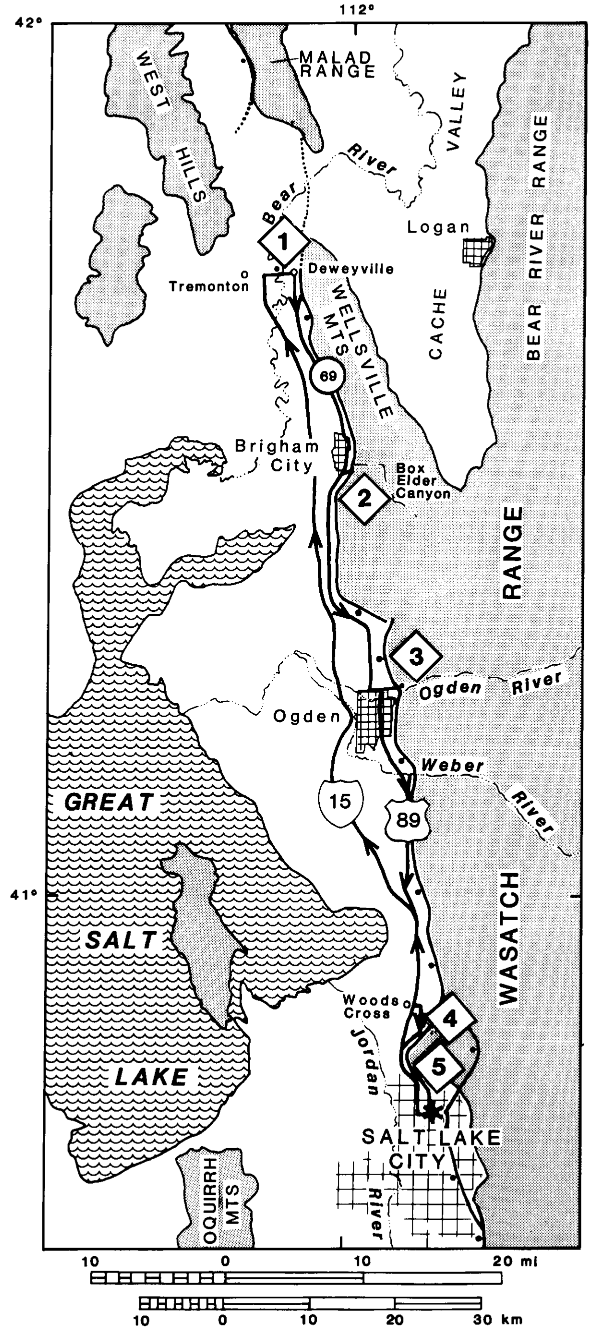
Sources of information:
[1] Machette and others (1987)
[2] Schwartz and others (1988); Schwartz (oral commun., 1988)
[3] Schwartz and Coppersmith (1984)
[4] M.E. Jackson (written commun., 1988)

Figure 6. *Estimated timing of major surface ruptures on segments of the Wasatch fault zone during the past 6,000 years (as known by investigators in June, 1988). Vertical lines indicate our estimated times of faulting; solid line where confident of timing or dashed (or queried) where timing is less certain. Cross-hachure pattern indicates permissible limits for faulting events as determined from calendar-calibrated ^{14}C ages and TL age estimates.*

varied and contemporary as "Plans for deep scientific drilling" (cited *in* Benson and Stehli, 1988) to "A theory of the earthquakes of the Great Basin, with a practical application" (Gilbert, 1884).

Gilbert was endowed with a clear and factual writing style, largely a result of his admirable practice of writing his notes fully in the field (Hunt, 1982a, p. vii), and he faithfully employed and tested multiple-working hypotheses to solve stratigraphic and structural problems. We agree completely with Dorothy Sack's proposal (this volume) that G.K. Gilbert was and shall continue to be an "Educator by example" and with C.B. Hunt's (1982a, p. vii) assertion that G.K. Gilbert "should be rated the greatest geologist this country ever produced."

Although few workers found fault with Gilbert's ideas and observations about Lake Bonneville and neotectonics in the eastern Basin and Range Province, many felt compelled to modify or embellish his framework over the past 100 years. The general consensus among contemporary researchers is that Gilbert had it "mostly right" and that if he had our analytical tools he would have had it "completely right."


42°
112°
41°
MALAD RANGE
WEST HILLS
CACHE VALLEY
BEAR RIVER RANGE
Bear River
Logan
Deweyville
Tremonton
WELLSVILLE MTS
69
Brigham City
Box Elder Canyon
RANGE
Ogden
Ogden River
Weber River
15
89
GREAT SALT LAKE
WASATCH
Woods Cross
Jordan River
SALT LAKE CITY
OQUIRRH MTS
1
2
3
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5
10 0 10 20 mi
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DAY 1

Road Log From Salt Lake City to the Bear River Near Deweyville, Utah, and Return

by

Michael N.Machette and Donald R. Currey

Field-trip leaders: Charles G. Oviatt, Stephen F. Personius, Alan R. Nelson, William E. Scott, and Richard Van Horn

The first day of our field trip takes us 65 miles (100 km) north of Salt Lake City to examine lacustrine deposits of the Cutler Dam lake cycle along the Bear River (stop 1). The road log is brief for this long run north and only points out some of the more obvious geomorphic and geographic features along the route. From stop 1, we turn south along the Wasatch Front, stopping east of Brigham City to see faulted sediments of the Bonneville lake cycle, which form a prominent delta at the mouth of Box Elder Canyon (stop 2). This stop will include an overview and discussion of lacustrine and tectonic features along the length of the Brigham City segment of the Wasatch fault zone.Lunch will be at stop 3, just north of the Ogden River at the mountain front on the east edge of the city of Ogden. Here, we will see evidence for several faulting events in the Holocene and discuss Gilbert's perceptions of faulting along the Wasatch fault zone. After lunch, the trip continues south along the Wasatch Front to a large active gravel pit (stop 4) that exposes dated transgressive deposits of the Bonneville lake cycle. If time permits, we will also stop near Becks Hot Spring to review Gilbert's observations about the times and amounts of movement on the Warm Springs fault, just north of Salt Lake City.

Mileage	Description of features along route
0.0	Intersection of 4th South and 7th East (Residence Inn). Proceed west on 4th South.
1.5	Turn left (south) on 2nd West.
1.6	Turn right (west) on 5th South (onramp to I-15).
2.3	Merge onto northbound I-15.
4.0	6th North exit. From here, there is a nice view to the east of the State Capitol, and to the north of the Becks Hot Spring (near refineries) along the Warm Springs fault, which bounds the west edge of the Salt Lake salient (see day 1-optional stop 5).
6.0	Here the highway is founded on the bed of Hot Spring Lake, which was drained sometime after 1934.
7.8	Tertiary conglomerates dip gently north in gravel pits along west edge of Salt Lake salient.
9.0	To the east, the Warm Springs fault breaks into several splays and merges with another major fault along the north side of the Salt Lake salient.
10.0	Woods Cross overpass.
12.0	To the east, the Wasatch Range is composed of Precambrian Farmington Canyon Complex.
17.8	Lagoon and Farmington exits. This area is tectonically backtilted, which causes Great Salt Lake almost to impinge on the Wasatch fault zone.
19.0	Junction with U.S. Hwy 89N; stay on I-15 northbound.
20.0	The bumpy, irregular terrain of the golf course on right (east) is part of massive lateral spreads that postdate the Gilbert shoreline (10.3 ka). These were first recognized by Van Horn (1975), who stated "To Utah belongs the dubious distinction of having what are probably the United States' largest landslides of the type known as failure by lateral spreading....Two landslides of this type, the Farmington Siding landslides, occur in Davis County, Utah between Farmington and Great Salt Lake. The younger covers about 9 km^2 and is probably less than 2000 years old. The older covers at least 8 km^2, but an unknown amount is hidden under the younger landslide. The older landslide is between 2000 and 5000 years old."
28.5	Here we skirt the western (distal) edge of the Weber delta. Oviatt (1984) notes that this feature is not a true "Gilbert delta" (gravel-cored), but rather is an underflow fan composed of fine sand and silt.
32.8	Near Roy, we start to see a number of regressive subdeltas graded to progressively restricted phases of Lake Bonneville (below the Provo level). The irregular (northerly) meadering course of the Weber is controlled by these subdeltas and possibly by eastward tilting of the basin.
37.2	Light brown, fine-grained sand and silt of the Weber delta (underflow fan) are exposed between 31st and 24th Streets near the old Pillsbury Mill.
38.5	21st Street exit. Proceeding down onto the Holocene floodplain of the Weber River.
40.0	Overpass. From here you can see the Pleasant View salient (spur) with shorelines of Lake Bonneville (to north) and fault scarps along northern part of the Weber segment of the Wasatch fault zone (to east).
45.0	Approaching the Pleasant View salient. Gilbert recognized this outlier as a bedrock spur and considered it to be a significant structural element of the Wasatch fault zone. More recent studies show that the Pleasant View salient is a structural barrier of the fault zone, and as such represents the boundary between the Weber (to the south) and the Brigham City (to the north) segments of the Wasatch fault zone (see day 1-stop 2).
48.5	Approaching Willard Reservoir (to northwest of I-15), a diked portion of Great Salt Lake that holds irrigation water diverted from the Weber River.
51.0	To the right (east) are spectacular north-dipping beds of Paleozoic Tintic Quartzite, which is unconformable on the Precambrian Farmington Canyon Complex. This part of the range is disturbed by complex Mesozoic thrust faulting.
67.5	Pass exit 375 to Honeyville.
69.1	Cross Holocene floodplain of the Bear River for next mile. Note intermediate-level deltas (terraces here) which are graded to the Gilbert level of Lake Bonneville.
71.6	Take exit 379 northbound on Utah Hwy 39 (to Tremonton and Pocatello).
74.1	Turn right at Crossroads, the small town at the junction with U.S. Hwy 30S. Proceed east to the Bear River.
75.0	West edge of the Bear River valley. The road log was constructed for a major stop on the bluff just east of the Bear River and on the north side of U.S. Hwy 30S. However, permission to visit this site had not been obtained in May 1988, so Oviatt and McCoy (day 1-stop 1) have included a number of alternate stops north along the Bear River. We will visit site 2.

DAY 1-STOP 1 discussion by Oviatt and McCoy.

Mileage	Description of features along route
0.0	Bluff on west bank of Bear River, U.S. Hwy 30S. Proceed east to Deweyville (trip will detour north to Oviatt and McCoy's site 2 along the Bear River).
0.5	Channel of the Bear River.
1.8	Turn right (south) on Utah Hwy 69, and proceed south through Deweyville.
5.0	Entering city limits of Honeyville (main part of town 2 miles south). To the left is the Madsen spur with landslides on the north end.
5.7	Crystal Hot Springs; fault scarp on hillslope to left (east) of highway.
6.3	View to east of alluvial-fan sequence and shorelines at Jim May Canyon. This area is the boundary between two segments of the Wasatch fault zone: the Collinston segment to the north and the Brigham City segment to the south.
7.0	Entering residental part of Honeyville. Note headscarp of large landslide east of town (see day 1-stop 2, figure 1).
9.2	Call's Fort gravel pit to the east.
14.9	Fault scarps in the alluvial fans at the mouth of Hanson Canyon (near Wheatland Seed, Inc. building). Continue south into Brigham City.
15.9	After junction with Utah Hwy 13 on north side of Brigham City, turn left (east) onto 900 North. Note scarps on alluvial fan of Waterfall Canyon (slightly south of due east).
16.5	Turn right (south) on Highland Drive (Boulevard) which parallels the Wasatch fault zone for the next 1 mile.
16.9	Cross Sunset Drive. The Bowden Canyon trench site (see day 1-stop 2) is located just to the southeast of Sunset Drive.
17.3	Intersection with Bott Avenue. Continue south on Highland Boulevard. Road climbs up alluvial fans of Bott and Flat Bottom Canyons. Note fault scarps on these alluvial fans.
17.6	Turn right (west) onto Beecher Avenue (150 North). The high surface to the south is the north end of the Box Elder delta, which is graded to the Provo level of Lake Bonneville. From here we will return to Main Street and proceed south through Brigham City.
18.2	T-intersection with 6th East. Turn left (south).
18.3	Turn right (west) on 1st North. Cross channelized floodplain of Box Elder Creek.
18.8	Turn left (south) on Main Street (Utah Hwy 69). Continue south through town.
20.6	Intersection with U.S. Hwys 89-91; turn left (east) toward Logan.
21.8	As you near the crest of the delta (and deep road cuts), take first right on dirt road.
21.9	Take left fork of dirt road, proceed up steep but short hill. Bear right at wooden sign marked "Brigham Wildlife Management Area..."
22.2	Climb up escarpment on the delta, road turns to right, then climbs a large fault scarp.
22.5	Park on upthrown block of Wasatch fault zone for an overview and discussion of stop 2.

DAY 1-STOP 2 discussion by Personius.

0.0	Retrace route to U.S. Hwy 89-91.
0.6	Turn left (west) across median of highway: be cautious about oncoming traffic.
1.8	Turn left (south) onto U.S. Hwy 89 (to Ogden).
2.0	As the highway curves to the right (southwest), you can see a large scarp that diagonally crosses the Provo and Bonneville shorelines. This scarp ends above the Bonneville shoreline in bedrock near Evans and Mathias Canyons (about 1 mile south of highway intersection at mileage 1.8).
3.0	Note Bonneville delta complex of Perry Canyon high on the slope to the left (southeast). Dissection on the north side of the delta obscures evidence of faulting.
5.3	Huge bedrock-cored landslide between Perry Canyon and Willard is pre-Bonneville lake cycle in age.
6.1	Gravel pit on north side of Willard Canyon is in a Bonneville-level delta. Faulting of the delta is not obvious from here. The larger, lower delta is graded to the Provo level.
7.7	Large fault scarps are visible south of Willard Canyon.
8.7	Holmes Canyon to left (east); note broad alluvial fan and distinct Holocene debris-flow deposits.
9.3	Wasatch fault zone trends southeast away from the highway and crosses the upper end of the Pole Patch (surface of the Pleasant View salient). Look back to northeast to see fault scarps in Holocene to middle-Pleistocene alluvial deposits.
11.3	Exit right onto old U.S. Hwy 89 (marked to Ogden).
12.0	Pass under I-15.
12.5	Utah Hot Springs. Southeast-trending fault on hillslope (bedrock on upthrown, south side) extends across highway to the springs.
15.3	Road to left (2550 North, Utah Hwy 235) provides a good point to view the Pleasant View salient from a distance.
18.0	Intersection with Utah Hwy 204 (Wall Avenue). Stay on U.S. Hwy 89.
18.9	Intersection with Washington Boulevard. Proceed into Ogden on Washington Boulevard.
20.0	Turn left (east) on 12th Avenue. (Canyon Road, Utah Hwy 39).
21.2	Intersection with Utah Hwy 203 (Harrison Boulevard). Proceed east toward mouth of the Ogden River canyon.
21.7	Turn left (north) on 1600 East and climb hill (¼ mile before Utah Power and Light substation).
21.8	Turn right (east) on 1350 South.
22.0	Turn left (north) on Maxfield Drive and climb hill.
22.2	Turn right (east) on Hislop Drive. Pavement ends in about 150 feet. Continue on dirt road about 0.2 miles, bearing to the left to an open area where we can park our vehicles for stop 3 (mileage 22.4). This is also our lunch stop.

DAY 1-STOP 3 discussion by Nelson.

Mileage	Description of features along route
0.0	Retrace route to Utah Hwy 39 (Canyon Road). As you descend into the Ogden River valley, note the large landslide complex and fault scarps on the south bank of the river to the southeast. Gilbert's photograph of this view was published in his 1928 Professional Paper as Plate 8B.
0.7	Turn right (west) onto Utah Hwy 39.
1.1	Turn left (south) onto Harrison Boulevard.
1.4	Cross Ogden River and climb bluff to crest of Provo-level delta formed by the Ogden River.
3.7	Intersection with 32nd Ave; proceed south.
8.4	Entrance to Weber State College on left (east). The main campus is constructed on fine-grained sand and silt of the Weber delta (underflow fan of Oviatt, 1984).
9.7	Intersection with 56th Street; proceed south. Road traverses the upper edge of the Provo delta, which here is cut into older lacustrine sediment of the Bonneville-level delta.
11.2	Intersection with Washington Boulevard. Proceed to left (southeast) on U.S. Hwy. 89 (southeast) and descend into the valley of the Weber River. Exposures in bluff created by extensive landsliding reveal a thick, monotonous section of fine-grained sand that comprises the delta. Many landslides are present on both the north and south bluffs of the valley.
13.1	Pass under I-84, cross Weber River, and continue south on U.S. Hwy 89. At the mouth of the Weber Canyon (to the east), late Holocene fault scarps extend northward almost to the river.
15.2	As we climb up onto the Bonneville-level delta, the Wasatch fault zone forms large scarps and grabens at and east of the highway. The choicest home sites along this part of the fault are on the crest of the main scarp. Extensive debris flows from the canyons have extended well west of the highway many times during the late Holocene.
19.9	Mountain Road (to east) leads to the Kaysville (Fruit Heights) trench site. Westminster Presbyterian Church on the left (east). From here we get a good view of the Kaysville site. A large (8-9 m) scarp with prominent graben is formed on alluvial fans graded to the Provo level.
22.0	Intersection with Utah Hwy 273.
24.1	Merge with I-15 and continue south toward North Salt Lake (stops 4A and 4B). Watch for large lateral spread (mile 20.0 of the first segment of day 1) on west side of highway.
32.7	Take Woods Cross exit. Turn left and proceed east on 2600 South.
33.1	Turn right (south) onto US. Hwy 89.
35.1	As you crest the hill just beyond Orchard Drive, get into inside (left) traffic lane, and turn left (east) into CPC North Salt Lake gravel pit.
36.2	Turn left inside the entrance gate and traverse the north and then the east side of pit. Park vehicle in uppermost pit near the end of conveyor belt for stops 4A and 4B.

DAY 1-STOP 4 discussion by Scott (4A) and Van Horn (4B).

Mileage	Description of features along route.
0.0	Retrace route to entrance of CPC gravel pit.
1.1	Turn left onto U.S. Hwy 89 (southbound), which crosses over I-15 and parallels it on the west. On left (east) side of highway is northeast-dipping Tertiary conglomerate (Van Horn, written commun., 1988) with a thin veneer of Bonneville lake cycle sediment.
2.2	U.S. Hwy 89 and I-15 diverge, then converge as Hwy 89 passes under I-15 (mileage 2.5). For the next 2 miles this highway is coincident with Beck Street. Stay on Hwy 89 into Salt Lake City. Cambrian bedrock to the east dips southeast.
3.5	Large gravel pit of Monroc Rock Products on left (site of optional stop--see day 1-stop 5). Mining of the gravel in this pit has revealed spectacular exposures of the Warm Springs fault (fault plane), with southeast-dipping Paleozoic rocks in the footwall.
4.0	Prominent tufa-cemented beach gravels of the Stansbury shoreline are to the left above the highway. In the basin one mile to the west, the top of the bedrock surface is 3500 feet (1670 m) below the land surface (Van Horn, written commun., 1988).
4.4	Take right fork in highway; Victory Road is the left fork. Proceed on U.S. Hwy 89.
4.9	U.S. Hwy 89 merges with 3rd West and turns due south.
6.7	Turn left (east) on 4th South.
7.5	The large Salt Lake City and County buildings to the south, between 3rd and 4th East, are founded in a lateral spread landslide).
8.3	Intersection with 7th East. End of log for day 1.

THE CUTLER DAM ALLOFORMATION: DEPOSITS OF A PROBABLE EARLY WISCONSIN LAKE IN THE BONNEVILLE BASIN

by

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INTRODUCTION

At this stop we will examine stratigraphic evidence (the Cutler Dam Alloformation and the Fielding Geosol) for a relatively shallow lake of probable early Wisconsin age in the Bonneville basin. A brief review of previous work on lakes of this age in the Bonneville basin, beginning with G. K. Gilbert's interpretations, is presented below to help place this stop in proper context.

Gilbert (1890, p. 259-260) summarized his interpretations of the history of lake fluctuations in the Bonneville basin as follows:

> *The relation of the alluvial cones to the shore-lines, and the condition of the low passes on the rim of the basin, show that before the Bonneville flooding the water level was low. This we may call the pre-Bonneville low-water epoch. It was of great duration compared with those enumerated below.*
>
> *The first Bonneville epoch of high water is stratigraphically represented by the Yellow Clay. Peculiarities of the shore-lines, and the phenomena at Red Rock and other passes, show that the water did not rise to the rim of the basin and was not discharged [elsewhere (p. 200) Gilbert suggests that the "Yellow Clay" lake rose to within 90 ft. of the level of the Bonneville shoreline].*
>
> *After the deposition of the Yellow Clay the water subsided, and the basin was nearly or perhaps completely dessicated. The stratigraphic evidence of this subsidence is found in the unconformity between the Yellow Clay and the White Marl and in the alluvial deposits occurring at that horizon. The possibility of complete dessication is suggested by the difference in character between the antecedent and subsequent deposits, which difference may have been occasioned by a change in the conditions of sedimentary precipitation. This may be called the inter-Bonneville epoch of low water.*
>
> *The second Bonneville epoch of high water is represented stratigraphically by the White Marl. Before the close of the epoch the water overflowed at Red Rock Pass ...*

Gilbert's interpretation was accepted for many years by subsequent workers on Lake Bonneville history. In 1953, Hunt and others (1953, p. 17) defined the Alpine Formation as lake deposits laid down during the time of development of Gilbert's (1890, p. 135-153) "Intermediate Shore-Lines." The Alpine Formation was apparently intended to be equivalent to Gilbert's (1890, p. 190) yellow clay at the Old River Bed, although Hunt did not explicitly call for this correlation. Later, Varnes and Van Horn (1961) correlated the Alpine Formation with the yellow clay (see day 2-stops 4A and 4B).

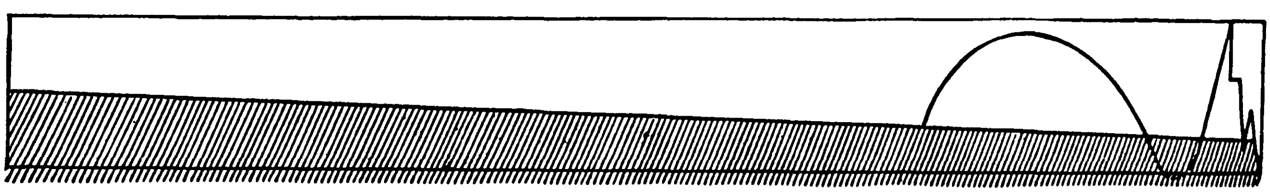


Figure 1. *Gilbert's (1890) fig. 34—rise and fall of water in Lake Bonneville. This figure depicts his interpretation of lake fluctuations in the Bonneville basin "... where the upper and lower horizontal lines represent the horizons of the Bonneville shore and the surface of Great Salt Lake. Horizontal distances represent time, counted from left to right. The curve represents the height of the oscillating water surface, and the shaded area indicates ignorance" (Gilbert, 1890, p. 261-262). See day 2, stop 4A for further discussions of Gilbert's interpretations of lake history.*

Further work by Eardley and others (1957), Williams (1962), Bissell (1963), Morrison (1965a, b, 1966), and others, modified the interpretations of the Alpine Formation. From the middle 1960s through at least the late 1970s, the Alpine Formation was generally interpreted as a stratigraphic unit that represented deposition in a deep lake (thought by some to be the deepest late Pleistocene lake in the basin), which expanded and contracted in a number of subcycles between about 75,000 and 30,000 yr B.P. (Morrison, 1966). The Alpine Formation, therefore, was interpreted as the sedimentary record of an early Wisconsin pluvial lake, or series of lakes, in the Bonneville basin (Morrison and Frye, 1965). Importantly, this interpretation of the formation was a direct descendent of Gilbert's interpretation of the yellow clay. However, new data and interpretations at the Old River Bed (Oviatt and McCoy, day 2-stops 4A and 4B) suggest that the yellow clay was not deposited during a separate, much earlier, major lake cycle as Gilbert thought, but that it was deposited early in the transgressive phase of the last lake cycle. In addition, the unconformity between the yellow clay and the white marl is found within only a narrow altitudinal range near the Stansbury shoreline, and the lake did not desiccate completely during the "inter-Bonneville epoch of low water."

Research by a new generation of workers during the last decade has resulted in many new interpretations of the lacustrine stratigraphy of the Bonneville basin. New radiocarbon dates, amino-acid analyses, and descriptions of paleosols (Scott and others, 1983; McCoy, 1987) strongly suggest that sediments previously mapped as the Alpine Formation in many areas are either much younger (part of the Bonneville Alloformation), or much older (part of the Little Valley Alloformation), than the presumed age of the Alpine. In addition, Scott and others (1983) showed that the early Wisconsin lake(s) in the Bonneville basin had to be confined to relatively low levels (less than 4530 feet (1380 m)). Exposures along the Bear River, which we will examine in this stop, have yielded evidence that the probable early Wisconsin lake reached a maximum altitude of 4400 feet (1340 m) (Oviatt and others, 1987).

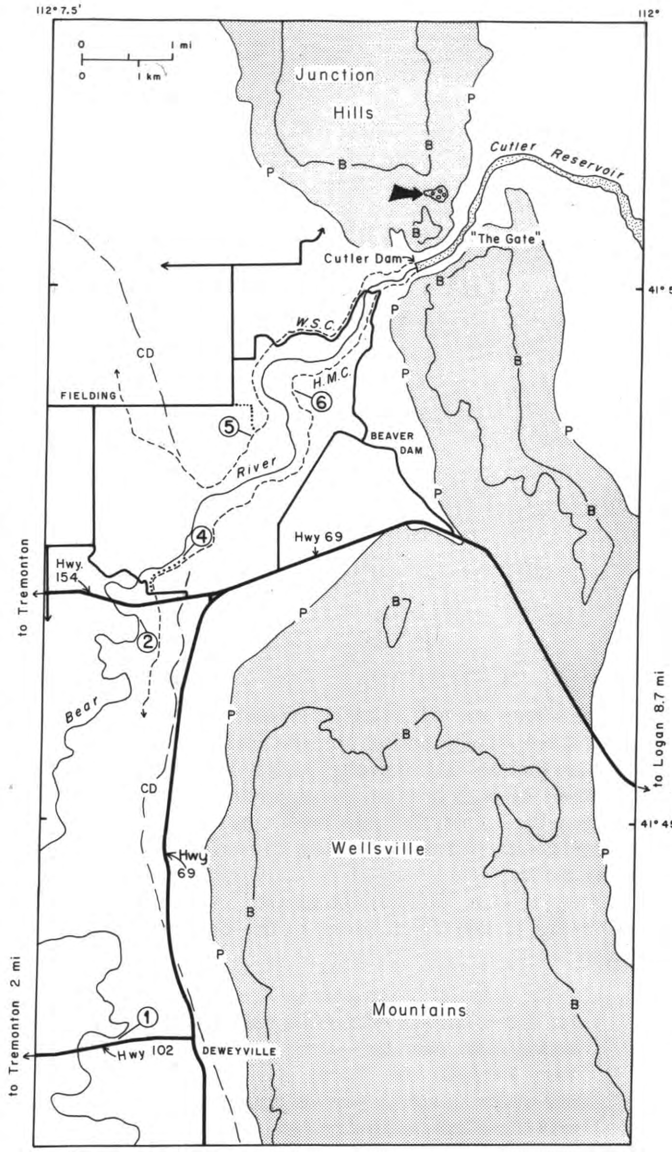


Figure 2. Location map showing highways (solid lines), dirt roads (dotted lines), and numbered stratigraphic sections (shown in figures 5-9). The approximate upper limit of the Cutler Dam lake (CD) and the Bonneville (B) and Provo (P) shorelines are shown. The large arrow north of "The Gate" of the Bear River represents the direction of strong currents that formed giant ripple marks in this gap during the Bonneville Flood. W.S.C., West Side Canal; H.M.C., Hammond Main Canal; area above Provo shoreline is shaded.

Table 1.
Amino Acid Analyses of Mollusks from Measured Sections in the Cutler Dam Area

Allostratigraphic unit [1]	Measured section	Lab number	Genus (n)[2]	alle/Ile[3] ratio
Bonneville	6	AGL-298	*Amnicola* (3)	0.105±0.005
Bonneville	6	AGL-297	*Sphaerium* (1)	0.11
Cutler Dam (l)	6	AGL-295	*Sphaerium* (3)	0.15 ± 0.01
Cutler Dam (ml)	5	AGL-269	*Valvata* (3)	0.14 ± 0.01
Cutler Dam (ml)	5	AGL-268	*Helisoma* (3)	0.11 ± 0.01
Cutler Dam (ml)	5	AGL-270	*Lymnaea* (2)	0.12 ± 0.01
Cutler Dam (ml)	5	AGL-274	*Lymnaea* ? (3)	0.14 ± 0.01

[1] (l), lacustrine facies; (ml), marginal lacustrine facies

[2] n, number of independent preparations and analyses of subsamples from a single collection

[3] alloisoleusine/isoleucine in the total acid hydrolysate

Gilbert did not visit the localities along the Bear River described in the figures below because many of the exposures did not exist in the late 1800s. Additionally, his field notes do not mention passing along the Bear River Valley in this area. However, he did travel through the area and was impressed by the course of the Bear River from Cache Valley into the Bear River Valley through a narrow, precipitous gorge, which he referred to as the "Gate of the Bear River" (figures 2 and 3; Gilbert, 1890, p. 178-180; Hunt, 1982a, b). During the last lake cycle, the main body of Lake Bonneville was connected to Cache bay, and thus to the overflow threshold (Red Rock Pass), through the Bear River "gate" and through two smaller passes, one north of the "gate," and one south of the "gate." The pass north of the "gate" has large gravel bars (giant ripples) on its eastern side, which were probably deposited as water discharged at a high velocity from the main body into Cache bay during the catastrophic Bonneville Flood (Oviatt, 1986a).

STOP 1, DAY 1

Refer to figures 2 through 9 for information at stop 1, day 1. Depending on the time available and the accessibility to private property, we will stop at one or several of the sections described in these figures. Further information concerning the Cutler Dam Alloformation and the Fielding Geosol can be found in Oviatt (1986a, 1986b) and Oviatt and others (1987).

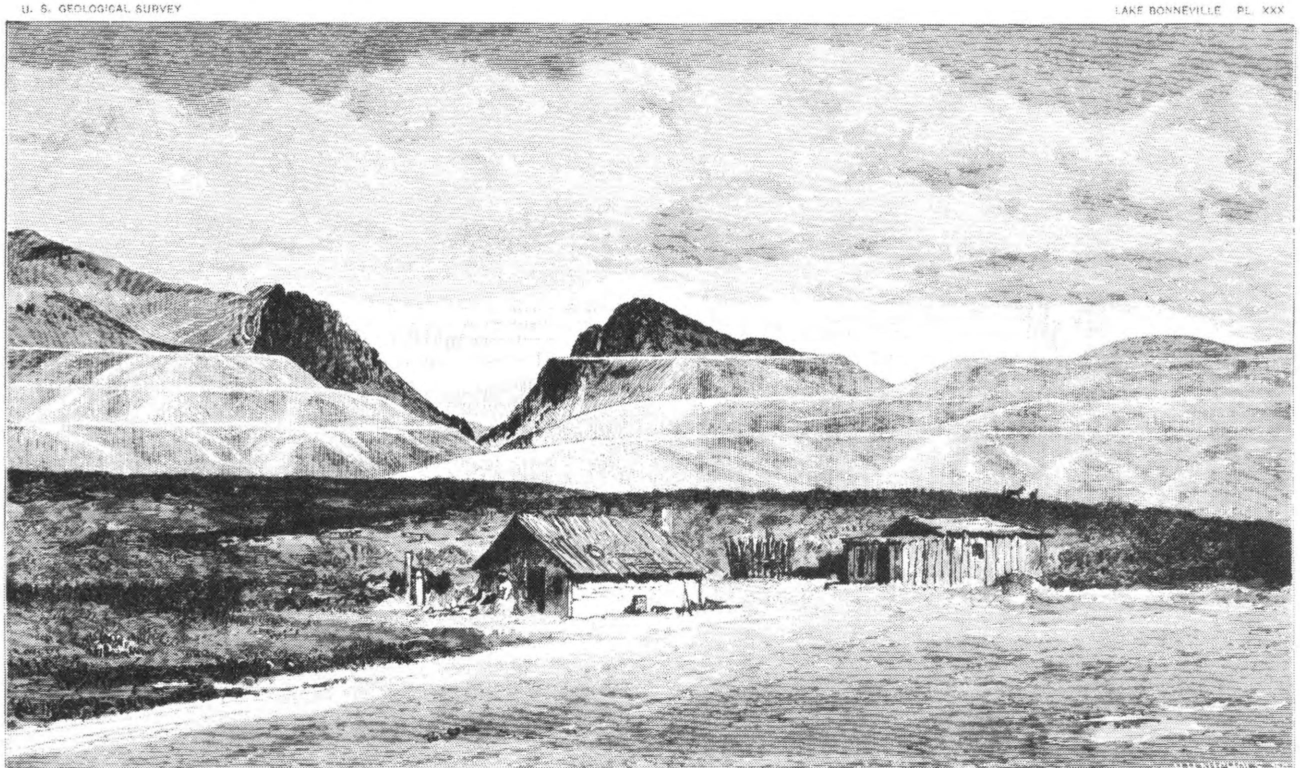
Two and possibly three, lacustrine stratigraphic units are exposed along the Bear River (figure 4). The Bonneville Alloformation is exposed at the top of each section and typically overlies pre-Bonneville lacustrine deposits, which we refer to as the Cutler Dam Alloformation. We refer to the buried soil that is developed in the Cutler Dam deposits as the Fielding Geosol.

The major topics of interest at the stops along the Bear River are the age of the Cutler Dam Alloformation, and the maximum altitude of the lake in which it was deposited. The age of the Cutler Dam Alloformation is constrained by radiocarbon dates and amino-acid ratios. A radiocarbon date greater than 36,000 yr B.P. (Beta-9845) on wood collected near the base of the Cutler Dam Alloformation provides a minimum limit on the age of the deposits. Amino-acid analyses of molluscs from the Cutler Dam Alloformation place a maximum limit on the age of the deposits (table 1). The alloisoleucine/isoleucine (aIle/Ile) ratios of *Lymnaea* shells are significantly lower than those from the Little Valley Alloformation collected elsewhere in the basin (table 2)[1]. The average aIle/Ile ratios of *Lymnaea*

[1]*Keep in mind that rates of isoleucine epimerization are taxonomically and temperature dependent, and that only aIle/Ile ratios of like genera and like environmental settings should be compared.*

Figure 3. *"The Gate" of the Bear River, from the east (Gilbert, 1890, plate XXX). The highest shoreline in the drawing is the Bonneville. During the Bonneville Flood, water discharged to the east through both "The Gate" and through the smaller pass to the right (north) of the peak in the center of the drawing. Large gravel bars are present in this smaller pass (figure 2; Oviatt, 1986a).*

U. S. GEOLOGICAL SURVEY LAKE BONNEVILLE PL. XXX



THE GATE OF BEAR RIVER, FROM THE EAST.

and *Sphaerium* shells from the Cutler Dam Alloformation are significantly greater than those from the Bonneville Alloformation but suggest that the Cutler Dam may be closer in age to the Bonneville than to the Little Valley Alloformation. Based on these data, and on the assumption that the Cutler Dam lake was approximately synchronous with a period of glaciation, we suggest that the Cutler Dam Alloformation is early Wisconsin in age, or broadly correlative with marine oxygen isotope stage 4.

From the available data (see figures 5-9), it appears that the Cutler Dam lake rose no higher than 4400 feet (1340 m) in altitude. In contrast, Lake Bonneville reached an altitude of 5092 feet (1552 m) at the margin of the basin, and the Little Valley lake probably rose no higher than about 4900 feet (1490 m; Scott and others, 1983). The differences in lake depth and surface area are partly due to differences in climate, which is the main factor, but are probably also due to the diversion of the upper part of the Bear River drainage into the Bonneville basin in late Pleistocene time (Bright, 1963; McCoy, 1987). The diversion probably occurred during the low-water interval between the Cutler Dam cycle and the Bonneville cycle, and may thus account for the much greater size of Lake Bonneville and the fact that it overflowed, whereas the other lakes did not.

ACKNOWLEDGMENTS

We are grateful to David McConnell and Michael Machette for constructive comments on the manuscript.

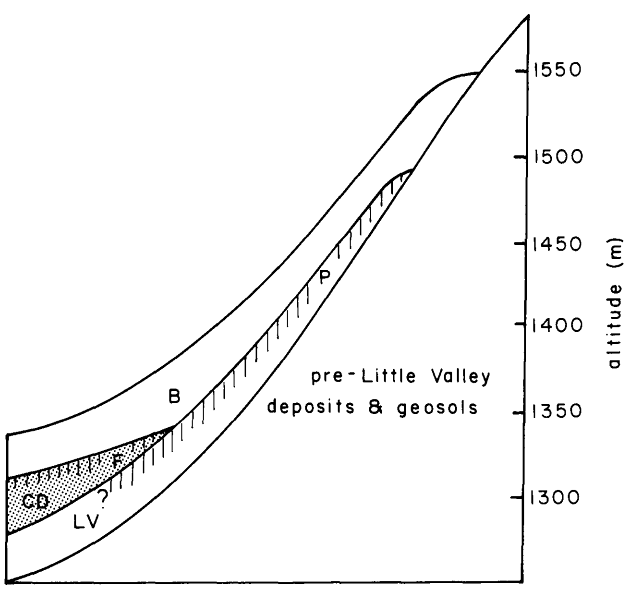


Figure 4. Schematic cross section through onlapping lacustrine deposits of the Bonneville basin. B, Bonneville Alloformation; C.D., Cutler Dam Alloformation; LV, Little Valley Alloformation; F, Fielding Geosol; P, Promontory (Dimple Dell) Geosol. Modified from Oviatt and others (1987, fig. 2). Query indicates that the lower altitudinal limit of the Promontory Geosol is unknown.

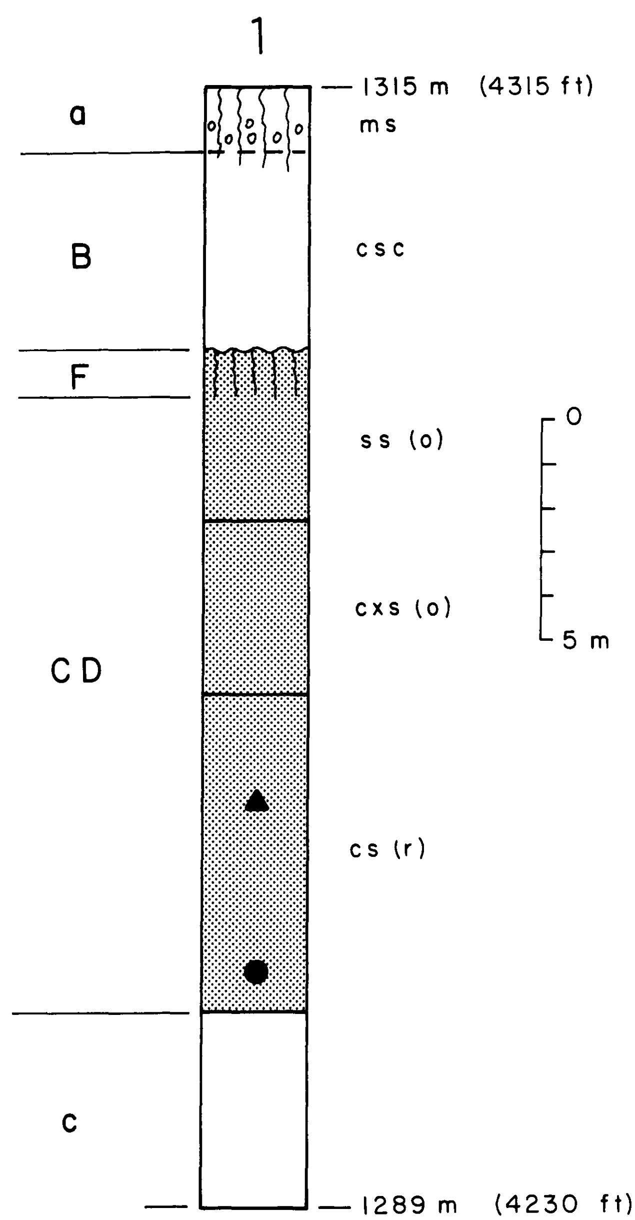


Figure 5. Measured stratigraphic section at location 1 (figure 2), a slump scarp overlooking the Bear River. Symbols: a, post-Bonneville alluvium; B, Bonneville Alloformation; F, Fielding Geosol; CD, Cutler Dam Alloformation; c, covered; ms, modern soil; csc, calcareous silt and clay; ss (o), oxidized sand and silt; cxs (o), oxidized cross-bedded sand interbedded with clay; cs (r), reduced clay and fine sand. The solid triangle marks the location of a sample of organic-rich clay that yielded a radiocarbon date of 28,180 ± 1120 yr B.P. (Beta-9483), which we interpret as a minimum age because of the likelihood of contamination with young carbon. The solid circle marks the location of a sample of clay containing the ostracode Limnocythere staplini *(R.M. Forester, pers. commun., 1984),* Picea *(spruce) needles (B.J. Albee, pers. commun., 1984) and* Salmo *sp. (trout) bones (G.R. Smith, pers. commun., 1984). A detailed description of the Fielding Geosol at this section is published in Oviatt and others (1987).*

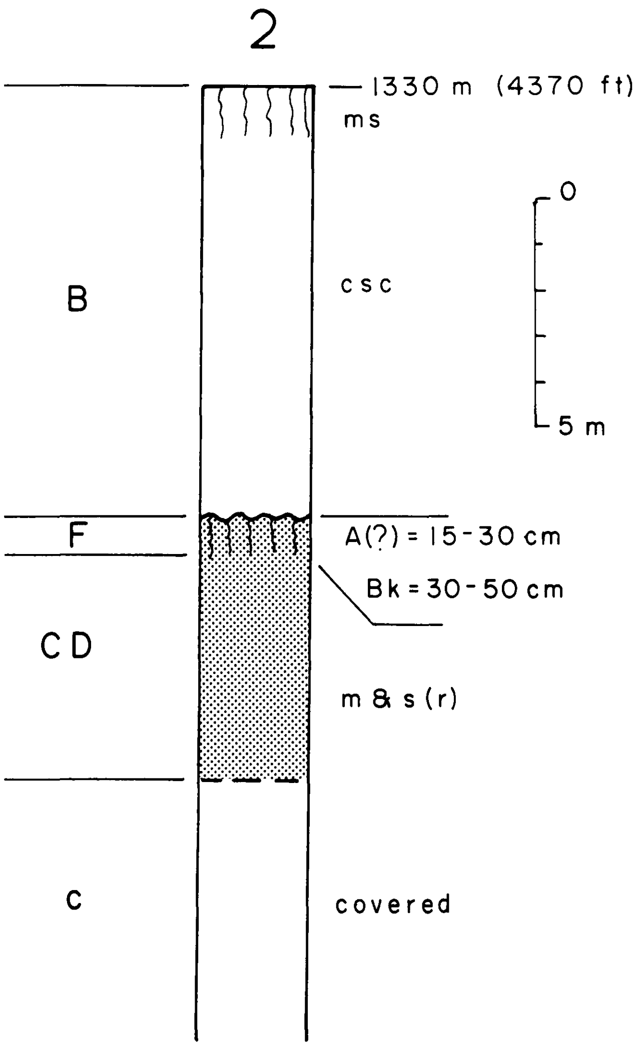


Figure 6. Measured stratigraphic section at locality 2 (figure 2), exposed in a slump scarp overlooking the Bear River. Symbols are the same as in figure 5, except m & s (r) is interbedded reduced mud and sand. In addition, the thicknesses of A and Bk horizons in the Fielding Geosol are given.

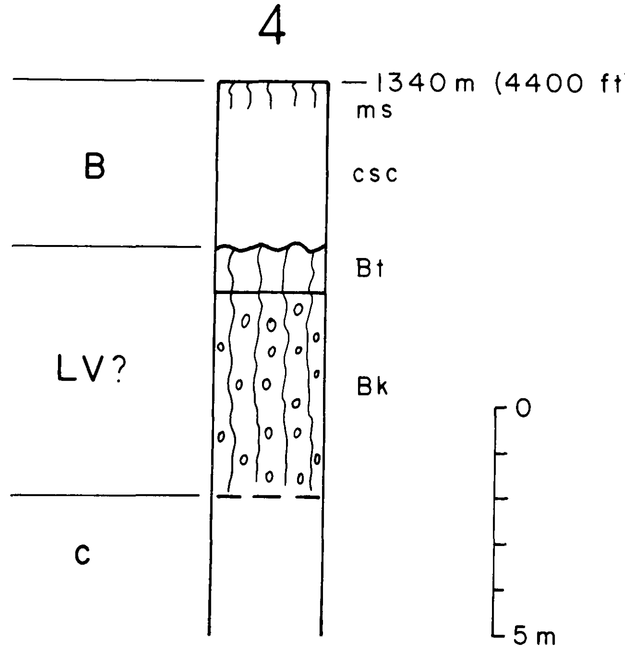


Figure 7. Measured stratigraphic section at locality 4 exposed in a slump scarp overlooking the Bear River directly west of the Hammond Main Canal road (dotted line in figure 2). Symbols are the same as in figure 5, except LV? is Little Valley Alloformation?; Bt and Bk are argillic and calcic horizons of buried soil (Promontory Geosol?). A detailed description of this buried soil profile is published in Oviatt and others (1987).

Table 2
Summary Table of Average alle/Ile Ratios in the Total Hydrolysate of Shells from Deposits in the Bonneville Basin

Allostratigraphic unit[1]	*Llymnaea*	*Sphaerium*	*Amnicola*
Bonneville (average)[2]	0.11 ± 0.03	0.12 ± 0.01	0.15 ± 0.03
Bonneville (CD area)	0.06	0.11	0.10 ± 0.005
Cutler Dam (l)	—	0.15 ± 0.01	—
Cutler Dam (ml)	0.13 ± 0.01	—	—
Little Valley (average)[2]	0.27 ± 0.03	—	0.32 ± 0.03

[1](l), lacustrine facies; (ml), marginal lacustrine facies

[2]from McCoy (1987) Note: The preparation method of samples for amino-acid analysis was changed after the research reported by McCoy (1987). The revised preparation method yields generally lower alle/Ile ratios than those reported earlier. Therefore, the ratios listed above as being from McCoy (1987) should be reduced by about 10-20% in order to be directly comparable to the other data in the table.

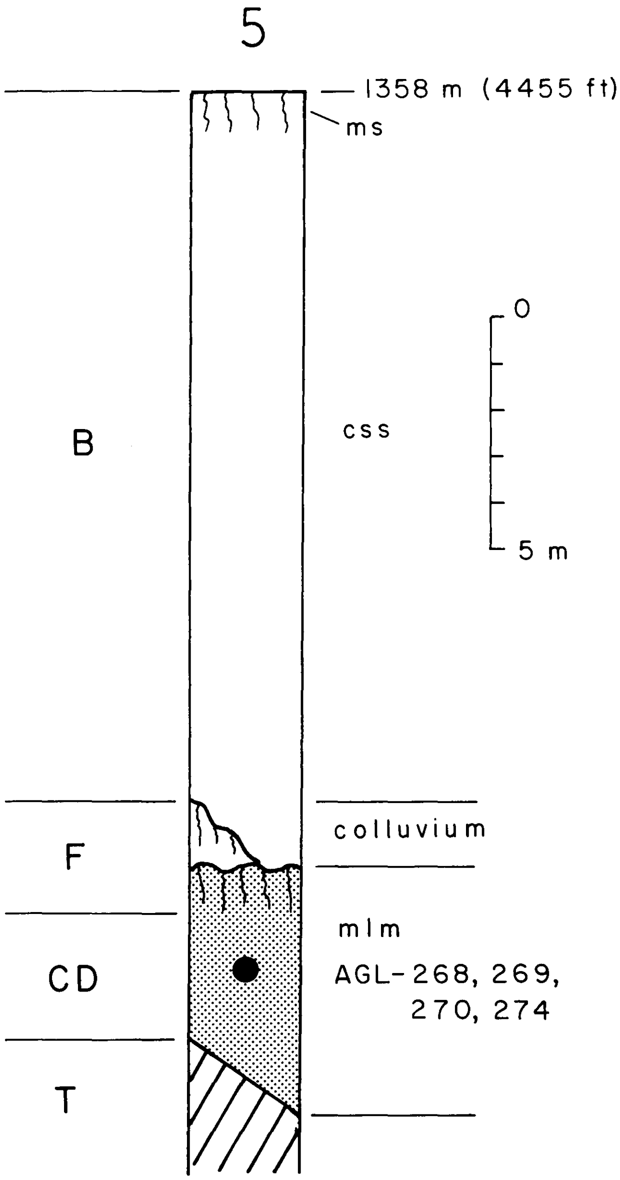


Figure 8. Measured stratigraphic section at locality 5 exposed in the bank of the West Side Canal (figure 2). Symbols are the same as in figure 5, except T is Tertiary lacustrine deposits; css is calcareous silt and sand; mlm is marginal lacustrine or fluvial/marsh muds. The solid circle marks the location of collections of shells for amino-acid analyses (table 1), and trumpeter swan bones (Feduccia and Oviatt, 1986). The marginal lacustrine deposits (mlm) at this, and nearby localities, fill shallow paleovalleys that were tributary to the Cutler Dam lake, and suggest that this is near the upper altitudinal limit of the Cutler Dam Alloformation, about 4400 feet (1340 m).

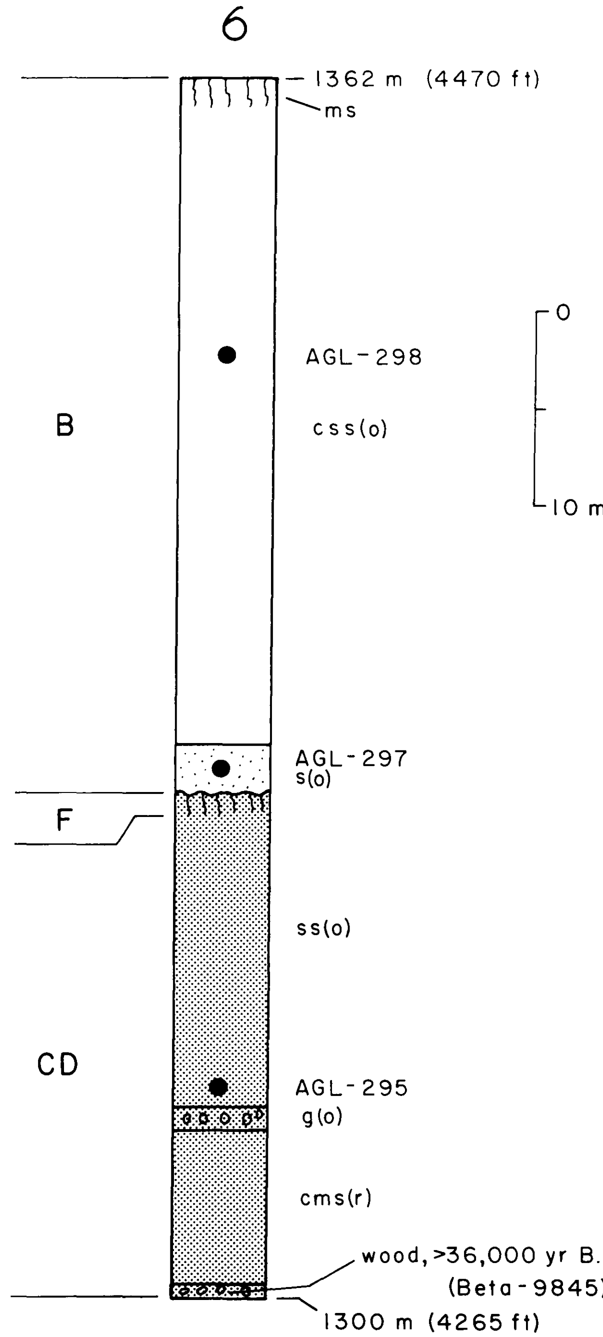


Figure 9. Measured stratigraphic section at locality 6, exposed in an excavation for a siphon along the Hammond Main Canal in 1984. The exposure is no longer open, but the section is important because wood for a radiocarbon date and shells for amino-acid analyses (table 1) were collected here. Symbols are the same as in figure 5, except css (o) is oxidized calcareous silt and sand, including volcanic ash reworked from Tertiary deposits; s(o) is oxidized gravel cms(r) is reduced calcareous mud and sand. The Fielding Geosol at this locality consists of a thin discontinuous A horizon over a weak BK, with maximum total thickness of 1 m.

A BRIEF SUMMARY OF THE SURFICIAL GEOLOGY ALONG THE BRIGHAM CITY SEGMENT OF THE WASATCH FAULT ZONE, UTAH

by

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INTRODUCTION

Recent studies of Quaternary geology along the Wasatch fault zone (WFZ) have concluded that the fault zone is composed of 6-12 seismically independent pieces or segments (Schwartz and Coppersmith, 1984; Machette and others, 1987). This paper is a summary of the surficial geology along the Brigham City segment, the northernmost segment of the WFZ that exhibits recurrent Holocene surface faulting (Personius, 1988). The discussion begins with a description of evidence of late Quaternary faulting, continues with descriptions of the segment boundaries, and ends with a description of Bonneville-lake-cycle deltas found along the segment.

Where appropriate, the discussion includes references to the pioneering work of G.K. Gilbert and his landmark studies of Lake Bonneville (Gilbert, 1890) and Basin-Range faulting (Gilbert, 1928). Although Gilbert spent very little time in the Brigham City area, his studies here and in other parts of the Bonneville basin have laid the foundation upon which recent studies of Quaternary geology and paleoseismology of this region are based.

LATE QUATERNARY FAULTING ALONG THE BRIGHAM CITY SEGMENT

Honeyville to Brigham City

The Brigham City segment has experienced repeated Holocene surface faulting along most of its length (Personius, 1986; 1988), but fault scarps are less well developed on the northern part of the segment, from near Honeyville (figure 1) to Brigham City. However, the apparent decrease in size and degree of preservation of fault scarps on the northern part of the segment may be more related to the geomorphology of the area than to a lack of late Quaternary tectonic activity. The west flank of the Wellsville Mountains is extremely steep, with only a thin veneer of Quaternary deposits preserved at the base of the range; these Quaternary sediments are especially thin on the piedmont interfluves between alluvial fans. No deltas or other lacustrine deposits of significant thickness are present along the range front between Honeyville and Brigham City. In addition, the fault zone commonly is coincident with the Bonneville and(or) Provo shorelines, making recognition of fault scarps very difficult. However, my detailed mapping of this section of the Brigham City segment has shown that fault scarps younger than the Provo shoreline (less than 13.5 ka) are present on uppermost Pleistocene and lower to middle Holocene surficial deposits (Personius, 1988). Topographic profiles across some of these scarps yield surface offsets of 26-30 feet (8-9 m), an amount that strongly suggests they were formed by recurrent surface faulting. In addition, an exposure in a gravel pit near Calls Fort 2.2 miles (3.5 km) southeast of Honeyville showed evidence of three post-Provo-shoreline surface-faulting events (Personius, 1988). These relations indicate that even though fault scarps are poorly developed along the west flank of the Wellsville Mountains, recurrent Holocene faulting has occurred on the northern part of the Brigham City segment of the WFZ.

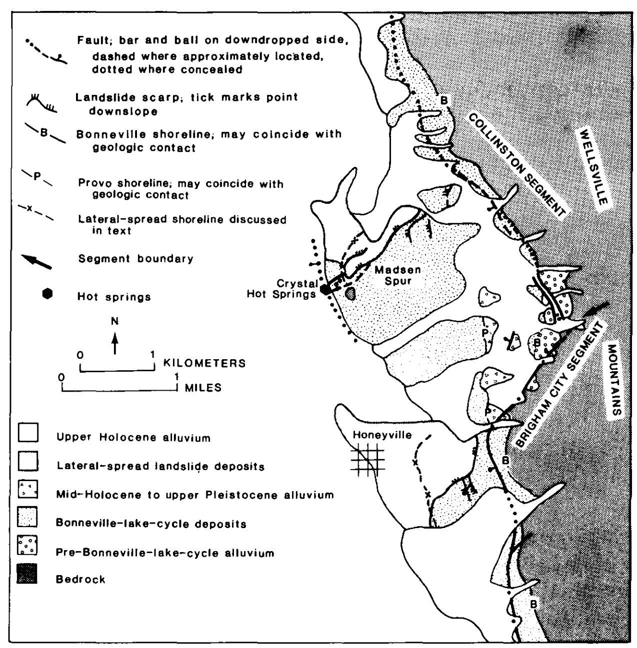

Figure 1. Surficial geologic map of the Honeyville area, simplified from Personius (1988). The arrow marks the boundary between the Brigham City and Collinston segments of the Wasatch fault zone.

Brigham City

Large, easily recognized fault scarps are present along the northern and eastern limits of Brigham City. The fault zone is clearly marked by large (16-50 ft, 5-15 m high) scarps and a series of springs along the eastern edge of the city, and by a series of scarps on Provo-level delta remnants on both sides of Box Elder Creek (figure 2). Fault scarps are so well developed in the Brigham City area that comments by Gilbert (1928, p. 39) about the lack of scarps must reflect the short amount of time he spent in the area:

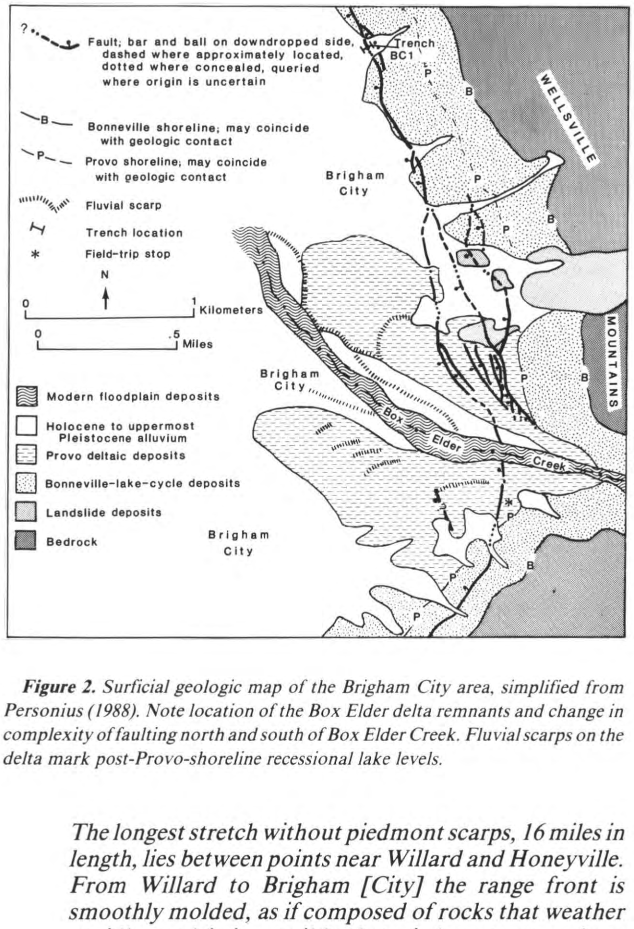

Figure 2. Surficial geologic map of the Brigham City area, simplified from Personius (1988). Note location of the Box Elder delta remnants and change in complexity of faulting north and south of Box Elder Creek. Fluvial scarps on the delta mark post-Provo-shoreline recessional lake levels.

> *The longest stretch without piedmont scarps, 16 miles in length, lies between points near Willard and Honeyville. From Willard to Brigham [City] the range front is smoothly molded, as if composed of rocks that weather rapidly, and it is possible that piedmont scarps have been less durable here than elsewhere; but this suggestion does not apply to the front between Brigham and Honeyville, which is bold and rugged. It is not to be assumed that the orogenic forces are there dormant, but rather that strains are slowly growing, to be relieved at some time in the future by a renewal of faulting.*

This statement is a bit confusing because in his earlier report, Gilbert (1890, plate XLV, facing p. 352) shows "line of recent faulting" extending continuously along the base of the range front from Willard northward beyond Brigham City. Although Gilbert apparently did not see the scarps along this section of the Brigham City segment, his statement clearly shows that he recognized that the lack of scarps did not preclude the presence of active faulting. Gilbert's statement also suggests that he recognized the presence of seismic gaps and the importance of strain accumulation between surface-faulting events. These are topics that have only recently gained attention in studies of active faults.

Several small (16-26 ft, 5-8 m high) scarps are developed on lower and middle Holocene alluvial-fan deposits east and northeast of Brigham City (figure 2; Personius, 1988). A trench (BC1) excavated across one of these scarps in 1986 revealed evidence for three or four surface-faulting events, the last two of which occurred about 3.6 and 4.8 ka (Personius and Gill, 1987; S.F. Personius, unpub. data, 1988). Evidence for one or two undated but earlier events was exposed in the lower part of the trench. Nearby fault scarps on Provo-level deposits are approximately twice as high as those in the Holocene fan deposits, suggesting that as many as six to ten surface-faulting events may have occurred on the central part of the Brigham City segment in the past 13,000-14,000 years (Personius, 1988).

Scarps on the Provo-level delta remnants at the mouth of Box Elder canyon are especially large and complex (figure 2). This may reflect some slumping associated with faulting or seismic shaking, but two lines of evidence point to a fault origin for most scarps on the Box Elder delta. On the north remnant of the delta, the fault zone is a complex of fault strands and grabens, whereas on the south remnant, the fault zone is marked by a single large scarp, and possibly by a second scarp of questionable origin (figure 2). Despite these differences in the pattern of surface faulting, topographic profiles measured across both remnants indicate that offset is roughly the same on both sections of the fault zone. In addition, the scarps on both remnants of the Box Elder delta extend onto thinner non-deltaic deposits, also suggesting that these scarps were formed by surface faulting.

Brigham City to Pleasant View

A complex belt of fault scarps marks the Brigham City segment from Brigham City to the southern segment boundary at the Pleasant View salient (Personius, 1988). Just as Gilbert (1928, p. 39, see quote above) had speculated, fault scarps are present on deeply dissected sandy Bonneville-lake-cycle deposits 3 miles (5 km) south of Brigham City near Perry. South of Perry, however, the fault zone is commonly marked by a very large (50-65 ft, 15-20 m high) scarp or scarps on thinner, less dissected lacustrine deposits. These scarps are preserved almost continuously from Perry south to the southern boundary of the segment.

BEDROCK SALIENTS AND SEGMENT BOUNDARIES

Gilbert (1928) discussed in great detail the bedrock spurs that protrude into the hanging wall at several locations along the Wasatch fault zone. We now know that in most cases these spurs or salients are fault-bound bedrock blocks that mark segment boundaries (Schwartz and Coppersmith, 1984; Machette and others, 1987; Wheeler and Krystinik, 1987). However, this conclusion has been derived from modern studies of Quaternary faulting that used numerical-dating techniques unavailable to Gilbert. Although the structural complexities within bedrock salients are still poorly understood, Gilbert's simple hypothesis about the formation of these structures (figure 3) closely resembles modern models of salient formation.

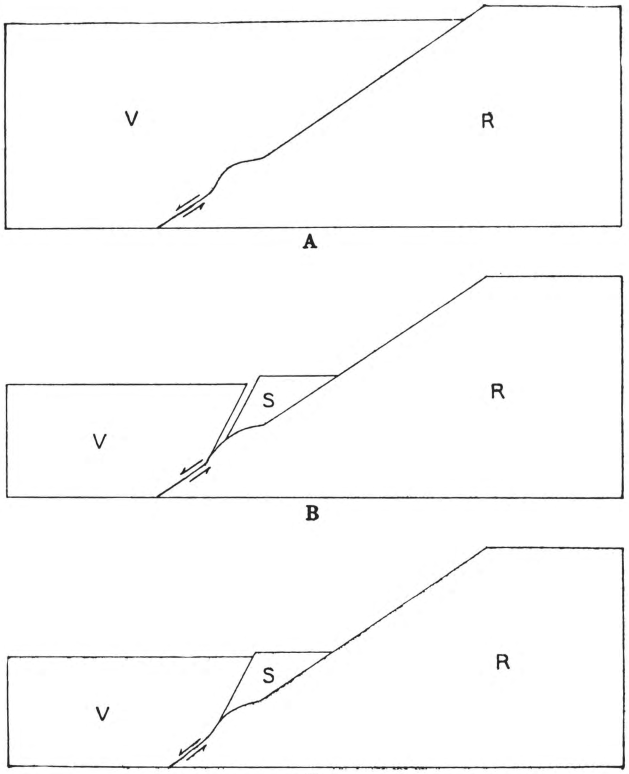

Figure 3. *Idealized, sequential cross sections showing formation of a spur (salient) block of the Wasatch fault zone (from Gilbert, 1928, fig. 27, p. 32).*

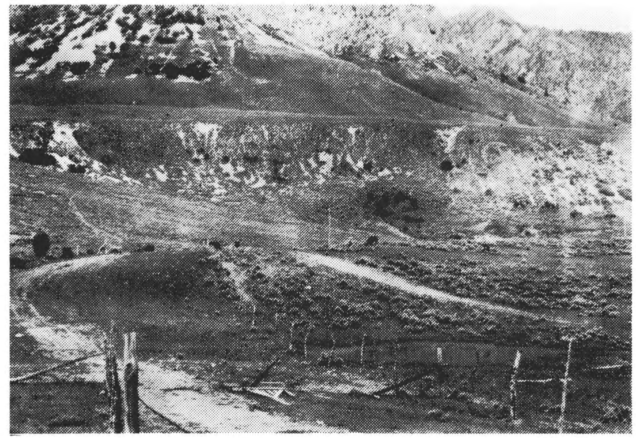

Figure 4. *Photograph of the headwall of the Honeyville lateral-spread landslide, view looking southeast (from Gilbert, 1928, plate 11B; USGS Photographic Library Gilbert Archive no. 3487). Note undrained depression and hummocky topography in the middle ground; prominent bench in the background is the Provo shoreline.*

Madsen Spur

The Honeyville area marks the boundary between the northern end of the Brigham City segment and the southern end of the Collinston segment of the Wasatch fault zone (figure 1; Personius, 1988) and is also the location of the two northernmost bedrock spurs discussed by Gilbert (1928). Gilbert's Madsen and Honeyville spurs are north and east of Honeyville, respectively. As Gilbert eventually concluded, the Honeyville spur (figure 4) is a lateral-spread landslide that underlies part of the town of Honeyville (Gilbert, 1928, p. 30). Although there is still some disagreement among workers about the origin of the Madsen spur, Gilbert's classification of the spur as a fault-bound bedrock block agrees with my interpretation of this area (Personius, 1988). Oviatt (1986b) attributed the bedrock exposures on the Madsen spur to a huge landslide that originated in the Wellsville Mountains, but my mapping suggests that landslide deposits are restricted to bedrock erratics located close to the mountain front. Regardless of the origin of the Madsen spur, evidence from distribution and amount of offset of fault scarps in several ages of deposits clearly suggests that a segment boundary exists in this area (Personius, 1988).

Two large lateral-spread landslides, the previously mentioned one underlying the town of Honeyville and a second northeast of Crystal Hot Springs (figure 1), may provide clues to the timing of at least one major earthquake in the Honeyville area. Gilbert recognized both these features, although he first classified the Honeyville landslide (figure 4) as a bedrock spur. With the help of airphotos, Oviatt (1986b) and Personius (1988) determined that these landslides were simultaneously deposited into the receding waters of Lake Bonneville as the lake was retreating from the Provo shoreline. This relation is evident because a shoreline at an altitude of 4420-4440 feet (1347-1353 m) divides both these deposits into an upper, undisturbed part and a lower part that has been slightly reworked by wave action (figure 1). This shoreline marks the level of Lake Bonneville at the time of lateral-spread emplacement, and its altitude is bracketed between the Provo shoreline (4820-4840 ft, 1470-1474 m in this area) and the altitude of modern Great Salt Lake (4210 ft, 1283 m). Lake Bonneville is known to have been at the Provo shoreline about 13.5 ka, and had retreated to the level of modern Great Salt Lake by about 11 ka (Scott and others, 1983; Currey and Oviatt, 1985), so the lateral-spread landslides in the Honeyville area, and several others to the north (Oviatt, 1986a, b; Personius, 1988) apparently were deposited into Lake Bonneville simultaneously about 12 ka.

The apparent simultaneous deposition of several lateral-spread landslides near Honeyville strongly suggests that they formed as a result of seismic shaking from a large-magnitude earthquake. Even though most of these landslides are located adjacent to the Collinston segment of the WFZ, this earthquake probably occurred somewhere on the Brigham City segment because fault scarps that post-date the Provo shoreline are abundant along the Brigham City segment, but are present only at the southern end of the Collinston segment (figure 1; Personius, 1988). These relations suggest that the distribution of the Honeyville lateral-spread landslides may be more closely related to the greater thickness of Bonneville-lake-cycle deposits in this area than to proximity of surface faulting. Because historic lateral-spread landslides are associated with earthquakes of magnitude 5.0 (M_L) or greater (Keefer, 1984, p. 409), the landslides near Honeyville probably formed during a large-magnitude (M 5-7.5) earthquake on the Brigham City segment of the WFZ about 12 ka (Oviatt, 1986a, b; Personius, 1988).

The shoreline on the lateral-spread landslide northeast of Crystal Hot Springs is offset across two short faults near the western edge of the Madsen spur (figure 1). These relations indicate that at least two surface-faulting events (the 12 ka event discussed above, and a later one that formed these faults) have occurred at the northern end of the Brigham City segment since abandonment of the Provo shoreline about 13.5 ka. Crystal Hot Springs appears to be localized at the intersection of these two faults and an older buried strand of the Wasatch fault zone (Davis, 1985) that forms the west flank of the Madsen spur (figure 1). Gilbert (1928, p. 30) presented evidence for this now-inactive fault strand in a discussion of hot springs and bedrock salients found along the WFZ.

Pleasant View Salient

The Pleasant View salient marks the boundary between the southern end of the Brigham City segment and the northern end of the Weber segment (figure 5). Like the Madsen spur, this bedrock salient is suspended at an intermediate structural level between the main trace of the Wasatch fault zone and an older buried strand (Davis, 1985) of the fault zone. Gilbert (1928, p. 24-27, 31-32) recognized several important features on the salient (figure 6):(1) the presence of pre-Bonneville lake cycle alluvial-fan deposits and scarps of probable fault origin on them, (2) the presence of fault scarps along the main trace of the Wasatch fault zone at the mountain front, (3) the existence of a fault (Gilbert's "Spur fault") along the southwestern flank of the salient that controls the location of Utah Hot Springs, and (4) the gap in surface faulting between the Brigham City and Weber segments (figure 6, C-D). These features characterize what is now considered an outstanding example of a normal fault segment boundary.

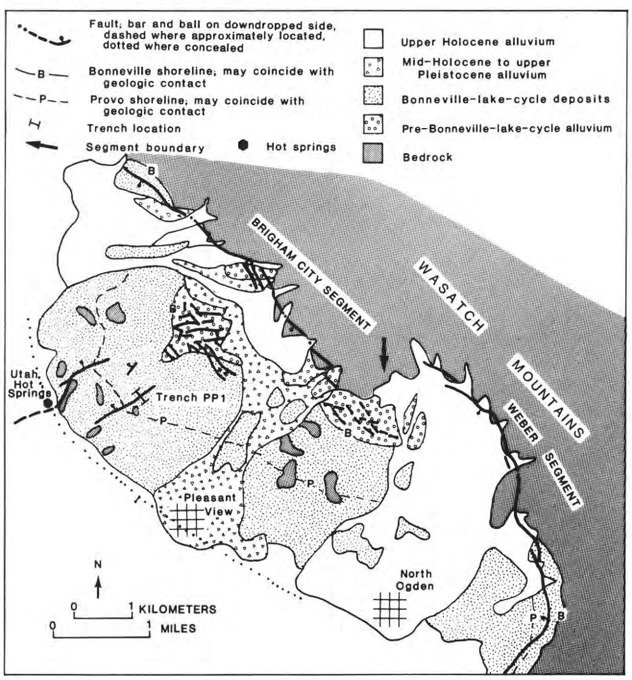

Figure 5. Surficial geologic map of the Pleasant View salient, simplified from Personius (1988) and unpublished mapping by A.R. Nelson. The arrow marks the boundary between the Brigham City and Weber segments of the Wasatch fault zone. Compare with Gilbert's map of the same area shown in figure 6.

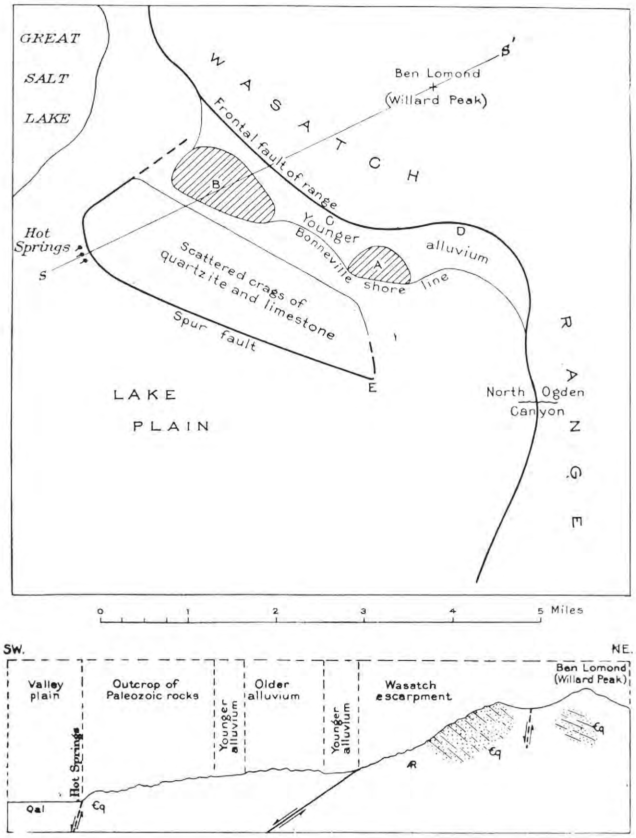

Figure 6. Map and cross section of the Pleasant View salient (from Gilbert, 1928, figs. 18 and 19, p. 25). On the map, A and B are tracts of pre-Bonneville lake-cycle alluvium; C and D mark the ends of a gap in Quaternary fault scarps; E presumably marks the eastern extent of bedrock on the salient; line S-S' marks the line of the cross section shown below. Compare with figure 5.

Although numerous short normal fault scarps are present on Quaternary deposits on the Pleasant View salient, only five of these scarps are present on Bonneville lake-cycle deposits (figure 5; Personius, 1988). A short trench (PP1) excavated across one of these scarps in 1985 revealed evidence of three surface-faulting events and about 16 feet (5 m) of offset since abandonment of the Bonneville shoreline about 15 ka. Radiocarbon analysis on organic-rich sediment deposited in a tectonic crack exposed in the trench indicates that the last event occurred 4.5-5.0 ka (S.F. Personius, unpub. data, 1988). This event may be the same as the penultimate faulting event recorded in the Brigham City trench (discussed above). Similar-aged deposits along the main trace of the Brigham City segment nearby are offset 2-3 times as much as those exposed in trench PP1, suggesting that the fault scarps on the Pleasant View salient may be reactivated during some, but not all, of the surface-faulting events that occur on the main trace of the Brigham City segment.

The longest of the five post-Bonneville lake-cycle faults on the Pleasant View salient appears to cut Holocene fluvial and lacustrine deposits near Utah Hot Springs (figure 5; Personius, 1988). The springs may be localized at the intersection of this fault and the buried fault strand that forms the southwestern flank of the salient. Although it may be a coincidence, the only hot springs on the Brigham City segment (Crystal and Utah Hot Springs) appear to be localized by fault intersections at the segment boundaries.

DELTAS OF THE BONNEVILLE LAKE CYCLE

Although Gilbert only briefly mentioned the Box Elder delta in his published work (Gilbert, 1890, p. 163), this delta (figure 2) exhibits most of the characteristics of other classic Provo-level deltas that he described in more detail. The internal stratigraphy of the Box Elder delta is typical of other Provo-level deltas: the bulk of the delta consists of moderate to steeply dipping foreset beds of sand and sandy gravel, which are capped by less-well-sorted, gently dipping topset alluvium. The geomorphic characteristics of the Brigham City delta include the slightly inclined, fan-like form of the delta top, the steep face of the outer delta margin, the central channel that bifurcates the delta, and inset terraces and regressional deltas that represent erosion and deposition at lower topographic positions as the lake retreated from the Provo shoreline (figure 2). Although not elaborated on in his published reports, Gilbert's field notes of August 25, 1876 on the Box Elder delta (Hunt, 1982a, p. 72) indicate that he recognized most of the important aspects of the geomorphology and history of the delta and Box Elder Canyon:

> *The history of the [Box Elder] canon during Bonneville times is clear. It preexisted with substantially its present form. Its bottom had a low grade and when the water rose it set back several miles at least making a firth [a narrow arm of the sea] less than a mile wide. ... During BB [Bonneville beach] time this bay was not filled up. If it had been the formation of a delta in the lake would have begun, and the well preserved beaches show that such was not the case. But during PB [Provo beach] time the filling of the cañon to the then water level was completed and a gravel delta was heaped into the lake. Its form was not that of an alluvial cone but was rather tabular. The gravel was carried as far as the water was stirred by waves and currents and then dropped. It does not appear to have been swerved by a general current but was thrown straight out into the valley [this statement was amended the next day by the following (Hunt, 1982a, p. 73): "I can see too this morning that the bars [deltas] of both creeks [Box Elder and Threemile] were deflected northward by a gentle current from the south"]. ... As the water subsided the waves excavated beaches in the gravel heap, and ... the creek began the cutting of its channel to successively lower levels, building new deltas seaward from the material of the old. A dozen of them could be counted. At last the lake retreated beyond reach of the gravel, and an alluvial cone was formed by the stream. Into this the stream has cut a sluice 20 or 30 feet deep and leaving it as a plain of planation is building a new one to the north. The town stands on the two alluvial cones.*

Gilbert (1890, p. 154) clearly understood that the restricted preservation of large deltas at the Provo shoreline reflected the pre-Provo-shoreline depositional history of the Bonneville lake cycle:

> *In the case of Lake Bonneville, the number of streams competent to project deltas from the shores of the open lake or of the larger bays, was small; ... With very few exceptions, they enter the lake basin through mountain gorges so deeply eroded before the lake epoch that the rising water set back into them, forming narrow estuaries. Knowing as we do ... that the water rose slowly as it approached the highest level [the Bonneville shoreline], we can not doubt that the stream drift was contemporaneously accumulated into a series of deltas within the mountain gorges. Afterward, when the water fell rapidly to the Provo level and there rested, the streams attacked the deltas in the defiles and carried their substance farther lakeward to form new structures. ... The material furnished by the older deltas in the defiles was close at hand, and in a condition peculiarly favorable for removal ... and we need not be surprised that the traces of its original forms are nearly obliterated.*

Gilbert (1890, p. 166-167) also speculated on climatic changes as reflected in the restricted distribution of Bonneville-lake-cycle deltas:

> *A certain significance attaches likewise to the absence of deltas from the greater portion of the coast of the old lake. ... In the western portion of the basin, there are catchment districts of considerable extent which furnish little or no waters to the lowlands by reason of the scantiness of rainfall. If the rainfall in Bonneville times was very great, as compared to the modern, these catchment districts should have furnished tributary streams; and such streams, flowing over tracts of alluvium, the accumulation of ages, should have transported large quantities of it to the margin of the lake and constructed deltas of it. We seem thus to have an intimation that the climatic change, whatever its nature, did not affect the rainfall in a degree commensurate with the difference in area of lake surface.*

Although considerable disagreement still exists today about climatic changes during the late Quaternary, many workers now agree that very cold, relatively dry conditions in late Pleistocene time accompanied glaciation and the growth of "pluvial" lakes in the western United States (McCoy, 1981; Galloway, 1983; Porter and others, 1983).

Deltas formed during the transgressive (Bonneville) phase of the Bonneville lake cycle are rare along the Wasatch Front (Gilbert, 1890, p. 154) but remnants of one delta of the transgressive phase are still preserved above a larger Provo-level delta at the mouth of Willard Canyon, 6 miles (10 km) south of Brigham City (Personius, 1988). This is the only delta of the transgressive phase still preserved along the Brigham City segment. The top of the transgressive delta is at an altitude of 4940 feet (1506 m), suggesting that most of this delta was deposited at an intermediate lake level, before the lake reached the Bonneville shoreline (5200 ft; 1585 m in this area). Although Gilbert apparently never made note of this delta, he did describe a similar delta complex at the mouth of American Fork Canyon, 25 miles (40 km) south of Salt Lake City (Gilbert, 1890, p. 155-159; see day 3-stop 3), and determined that the intermediate delta pre-dated both the Bonneville and Provo deltas.

CONCLUSIONS

The Brigham City segment of the Wasatch fault zone is marked by recurrent Holocene faulting along most of its length, but fault scarps are higher and better preserved from Brigham City south to the southern segment boundary. The segment boundaries near Honeyville and Pleasant View are marked by bedrock salients or spur blocks that indicate bifurcation of the main trace of the fault zone. These boundaries were recognized by G.K. Gilbert as features that controlled the geometry of the Wasatch fault zone, long before the concept of normal-fault segmentation had been formulated. Some of the most conspicuous geomorphic features left by Lake Bonneville are deltas deposited at the mouths of major canyons; most of these deltas were deposited at the Provo level, although a single transgressive delta is preserved on the southern part of the Brigham City segment.

The classic studies of Lake Bonneville and Basin-Range faulting by G.K. Gilbert have been an inspiration to our modern studies of paleoseismicity of the Wasatch fault zone. In many cases, recent advances in knowledge have come through the use of techniques not available to Gilbert, such as airphoto interpretation and use of numerical-dating methods. His remarkable insights into the geology of this region have truly stood the test of time.

ACKNOWLEDGMENTS

This research was conducted under the auspices of the Earthquake Hazards Reduction Program of the U.S. Geological Survey; the exchange of ideas with colleagues Michael N. Machette, Alan R. Nelson, and David P. Schwartz has added substantially to the author's understanding of Quaternary faulting. The trenching operation in the Brigham City area was made possible through the support of William R. Lund and his staff at the Utah Geological and Mineral Survey. The manuscript benefited from constructive comments by Theodore P. Barnhard and Alan R. Nelson.

THE NORTHERN PART OF THE WEBER SEGMENT OF THE WASATCH FAULT ZONE NEAR OGDEN, UTAH

by

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GILBERT'S STUDIES OF THE WASATCH FAULT NEAR OGDEN

G.K. Gilbert spent several weeks studying the Wasatch fault zone near Ogden during trips in the late 1870s, in 1901, and in 1914. Many of the arguments he used to support his views on the origin of fault scarps in unconsolidated deposits ("piedmont scarps" of Gilbert, 1928, p. 12, 34) and of the Basin and Range structure were based in part on evidence he found near Ogden. For example, he used the orientation and distribution of shear zones and fault gouge in bedrock to support his notion of incremental, normal displacement on range-bounding faults. He also recognized the importance of pre-existing structures in controlling the trace of the fault (at both large and small scales) and he anticipated the recent development of concepts of fault segmentation by correctly interpreting the origin of the large bedrock salients that bound the Weber segment on its north and south ends (see index map, 1).

Gilbert's most pertinent contributions to present studies along the Weber segment were his ideas about piedmont scarps, which he found almost continuously along the base of the Wasatch Range near Ogden (figure 1.). Gilbert found grabens along many of the scarps he studied and suggested several possible styles of graben formation (Gilbert, 1890, p. 355). He also knew that *en echelon* scarps represent lateral transfer of slip along the fault (Gilbert, 1928, p. 34). More importantly, he recognized that the relative age of scarps can be determined by comparing the height of scarps in alluvial deposits of different ages, by comparing scarp mid-slope angles, by comparing the degree of dissection and gullying on scarps, and by comparing the degree of rounding of the scarp crest (figure 2; Wallace, 1980). Considering his observations about differing scarp heights, one wonders whether Gilbert had come to a conclusion about the size of the characteristic earthquake (Schwartz and Coppersmith, 1984) for the Wasatch fault. Finally, Gilbert (1928, p. 34) noted that "In several places it is evident that two or more movements have occurred on the same line." Thus, Gilbert was familiar with almost all of the basic concepts that we use today to map and interpret fault (piedmont) scarps.

Figure 1. *Photograph of the East Ogden trench site (figure 3) looking east, taken by Gilbert in 1901 (USGS Photographic Library Gilbert Archive no. 3470). Fully-developed residential areas now extend almost as far east as the lowest prominent scarp in the photo. All of the scarps in the southern part of the site (right part of the photo) have now been destroyed by gravel mining operations.*

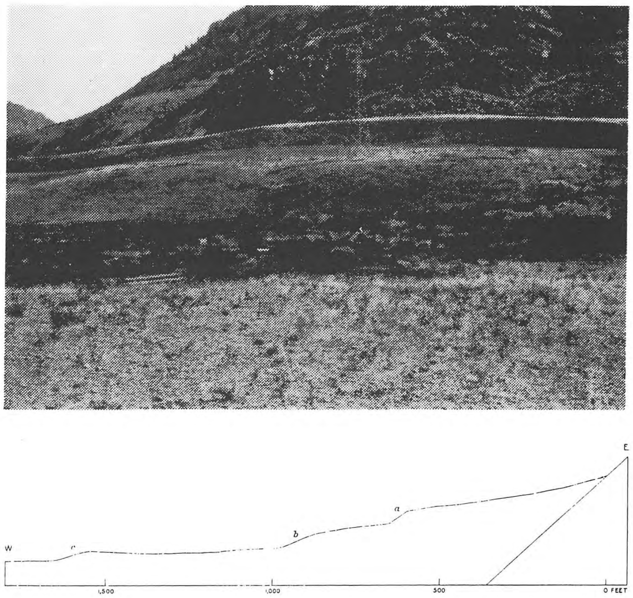


Figure 2. *(A) Photograph taken by Gilbert (1928, plate 17B; USGS Photographic Library Gilbert Archive no. 3419) in 1901 of the present site of a reservoir for the City of Ogden at the east end of 23rd St. in Ogden. (B) Gilbert's profile of the piedmont scarps (scarps a, b and c) made at the site in this photograph (A). View of scarps a and b is to the east from the crest of a third, westernmost scarp (scarp c in B). Gilbert interpreted the sharp crest of the easternmost scarp in the photo as evidence that it was younger than the lower scarp with the rounded crest. He also pointed out that both scarps were younger than the Bonneville shoreline, which can be seen in the upper left corner of the photograph.*

FAULT VERSUS LANDSLIDE ORIGIN FOR PIEDMONT SCARPS

The problem of distinguishing landslide scarps from fault scarps was not a problem that worried Gilbert in his earlier studies--the uniform morphology and the continuity of the piedmont scarps he observed parallel to the base of many ranges made it obvious to him that these scarps were produced by faulting (Gilbert, 1928, p. 34). Later criticism of his lack of discussion of possible alternative hypotheses of scarp origin (such as landsliding) made Gilbert more cautious, as illustrated in the following statement:

> *The preceding statement of facts and considerations bearing on the origin of the piedmont scarps has a regrettable prolixity because it is a record of an endeavor to free my own mind from an initial bias in favor of one hypothesis by a thorough examination of its rival. When the piedmont scarp was first observed in the Great Basin, in 1876, it was at once ascribed to movement on a frontal fault, the existence of which had already been inferred from independent physiographic evidence. During the succeeding decade the same association of characters was found in many places, and the piedmont scarp was thought, especially by Mr. Russell and myself, to afford strong corroborative evidence of the fault-block theory of Basin Range structure. When, therefore, one of my colleagues in discussing the structure treated the evidence of the scarp as "no evidence," and others suggested that the scarp might be due to slumping of the piedmont deposits, I was led to suspect that I had been overconfident of the credibility of the witness whose testimony had so strengthened my preconceptions and to review the subject with special care in later field studies.*
> Gilbert, 1928, p. 36-37).

Gilbert (1928, p. 36-37) recognized that landsliding was a normal consequence of earthquake ground motions, especially in unconsolidated sediments with a shallow water table. He also reasoned that surface ruptures would produce scarps that were very susceptible to slumping and that, therefore, in some areas landslide and fault scarps might coincide. Despite the fact that we have extensive, detailed air photographic coverage that Gilbert didn't have, we have experienced most of the same problems he had with distinguishing landslide scarps from fault scarps (Nelson and Personius, unpub. mapping, 1987).

One aspect of the problem that Gilbert didn't fully understand was the origin and extent of the large lateral-spread deposits found in some of the thick sequences of lacustrine sediments along the Weber segment (mileage 20.0 on way to day 1-stop 1). Gilbert identified large areas of arcuate landsliding, such as on the bluffs along the Ogden and Weber Rivers. He also recognized that some large subparallel ridges, such as those on and just north of the Provo-age Weber River delta south of Ogden (near mile 10.8 after this stop; Sec. 15, T 5 N, R 1 W), were the

> *products of dislocation, and the only plausible explanation I have been able to suggest is that they have resulted from the quaking of the delta at a time when movement on the near-by fault was creating the piedmont scarps.*
> (Gilbert, 1928, p. 36).

He deduced that the ridges represented "the breaking of the terrane into blocks and the separation of the blocks," but was confused because all the ridges he observed near Ogden were not systematically tilted into a slope like the much smaller earthquake-induced spread deposits he had observed in California after the 1906 San Francisco earthquake. I think the ridges that confused Gilbert are probably parts of lateral-spread headscarps developed in silty lake sediments that predate the Bonneville shoreline. Ridges in the northern part of this area may be the tops of blocks from a large, pre-Bonneville landslide that slid off of the mountain front (Pashley and Wiggins, 1972). The origin of all these ridges is hard to interpret because they have been extensively eroded by waves during the rise, and perhaps the fall, of Lake Bonneville.

Gilbert (1928, p. 36) ventured in the wrong direction in his discussion of landslides and earthquake-induced features along the Wasatch Front when he repeatedly referred to very small (less than 1 m high) ridges produced by the liquefaction of tidal-flat muds in Tomales Bay, California, during the 1906 earthquake. This was, perhaps, part of his effort (described in the above long quote) to appear more objective to his critics in considering scarp origins. Gilbert was intrigued by the smooth morphology and irregular orientation and spacing of the ridges perpendicular to the fault just north of Jump Off Canyon (Sec. 4, T6N, R1W), about 2.8 miles (4.5 km) north of the Ogden River. He knew these ridges were the result of a large-scale ground disturbance (which he inferred was caused by an earthquake), but it is not clear from his discussion whether he thought the ridges had formed primarily by liquefaction (topographic undulations preserved from the waves of liquefied sediment produced by earthquake ground motions) or whether liquefaction had only helped small blocks of sediment to slump and slide (lateral spreading). Gilbert didn't recognize that some of these ridges are parts of large lateral-spread deposits and that others are old landslide deposits from the mountainside above the fault, which have been later smoothed by the waves of Lake Bonneville (my interpretation). However, even modern observers would have a very difficult time determining the correct origin of these features from horseback.

MIDDLE AND LATE HOLOCENE HISTORY OF THE WASATCH FAULT AT THE EAST OGDEN SITE

From the number of photographs that Gilbert took of the scarps at the area that is now at the east end of 9th and 12th streets in Ogden, it is clear that he was just as intrigued with the record of the history of faulting here as I am (figures 3 and 4). He recognized that several events of different ages had occurred and that the graben along the lower, 26-foot (8 m) scarp had developed through antithetic faulting. He also used the fact that the main scarp curves to the west at the north end of the site and wraps around the spur of dark gneiss as an example of how pre-existing structural features control the trace of the fault. However, he left some questions unanswered:

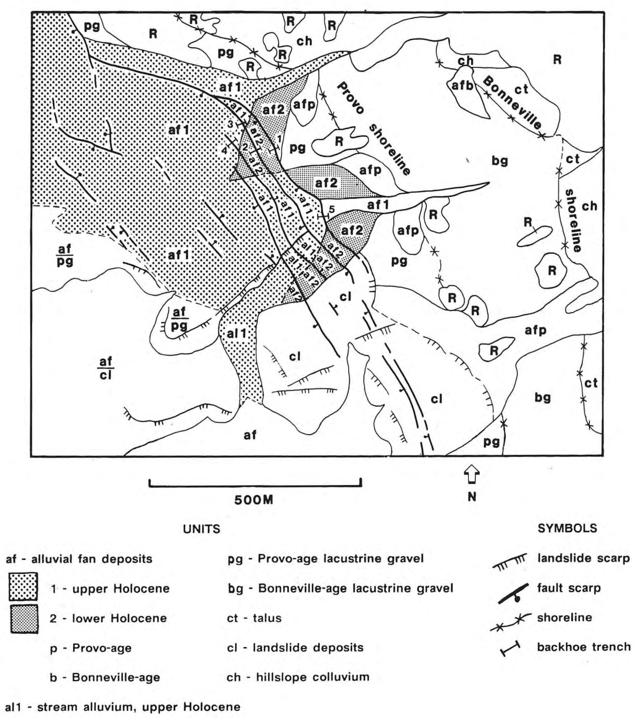
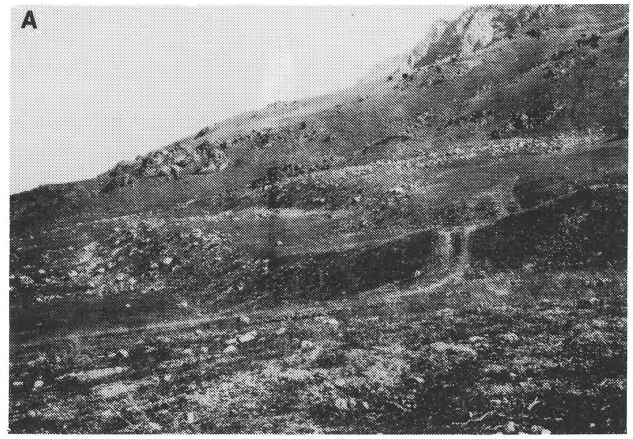


Figure 3. Surficial geologic map of the East Ogden trench site showing the prominent fault scarps crossing the site (figures 1 and 4) and the location of the five trenches (numbered 1 through 5, 1 = EO-1, 2 = EO-2, etc.) dug across three of these scarps. Trenches EO-1 and EO-5 are on the 5-m scarp and trenches EO-2 and EO-3 are on the 8-m scarp. Trench EO-4 is on a small antithetic scarp. Unit af/pg designates thin alluvial fan deposits overlying Provo-age gravel.

I did not satisfy myself whether the buttress [spur] should be classed as an independent crust block or as a knob projecting from the Wasatch block, a knob so strong and hard that it has plowed its way through the hanging wall as the valley block descended.
(Gilbert, 1928, p. 31)



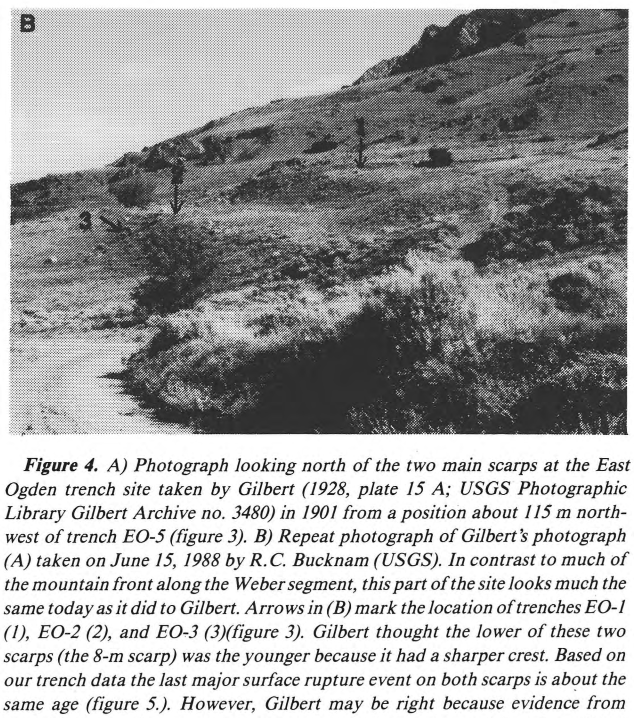


Figure 4. *A) Photograph looking north of the two main scarps at the East Ogden trench site taken by Gilbert (1928, plate 15 A; USGS Photographic Library Gilbert Archive no. 3480) in 1901 from a position about 115 m northwest of trench EO-5 (figure 3). B) Repeat photograph of Gilbert's photograph (A) taken on June 15, 1988 by R.C. Bucknam (USGS). In contrast to much of the mountain front along the Weber segment, this part of the site looks much the same today as it did to Gilbert. Arrows in (B) mark the location of trenches EO-1 (1), EO-2 (2), and EO-3 (3)(figure 3). Gilbert thought the lower of these two scarps (the 8-m scarp) was the younger because it had a sharper crest. Based on our trench data the last major surface rupture event on both scarps is about the same age (figure 5.). However, Gilbert may be right because evidence from trench EO-3 on the same scarp suggests a small event (event d, less than 60 cm) may have ruptured this scarp less than 500 years ago.*

Gilbert (1928, p. 38-39) tried hard to explain why piedmont scarps, like those at this site (figure 4), were parallel to but at various distances from the bedrock escarpment of the fault. Along the eastern edge of the city of Ogden the scarp closely follows the steep north face of the spur of gneiss north of the East Ogden site, but on the south side the escarpment slopes more gently and the scarp is farther away from it. But at sites with multiple scarps of different ages, Gilbert observed that at some sites the youngest scarp was the closest scarp to the escarpment and at other sites the youngest scarp was the farthest away. He suggested that the geometry of the bedrock-alluvial contact, the seasonally variable elevation of the water table, and (perhaps) the intensity of the earthquake ground motions combined to determine the point at which the fault ruptured the alluvial surface from one event to the next. Thus, Gilbert may have inferred that the steeper (younger), 26-foot (8 m) scarp at East Ogden was produced by a late summer or fall earthquake (lower water table).

Through the expenditure of several orders of magnitude more time and money than Gilbert spent we have developed a somewhat more detailed history of faulting for this site (figure 5; Nelson and others, 1987). On middle Holocene fan deposits (^{14}C dated at about 5.5 ka) two main scarps record about 16 and 26 feet (5, 8 m) of total displacement; on upper Holocene fan deposits, the displacements on these scarps are only 4 and 6 feet (1.2, 1.8 m), respectively. As Gilbert recognized, this shows that there has been recurrent displacement on both scarps.

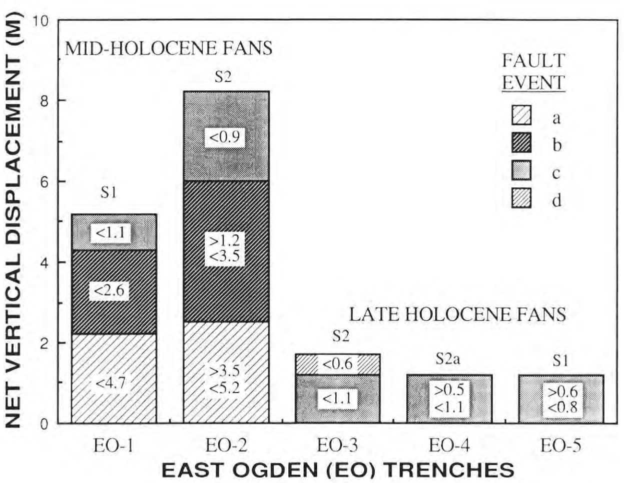


Figure 5. *Displacements and age control for surface faulting events recorded in the five trenches at the East Ogden site. Numbers within boxes are minimum or maximum age estimates in thousands of years based on radiocarbon and thermoluminescence analyses. Event c is recorded in all trenches on both the 8-m scarp (S2) and the 5-m scarp (S1). Scarp S2a is antithetic to scarp S2, so most of the displacement for event c in trench EO-4 should be subtracted from the displacement for this event on the adjacent main scarp (trench EO-2) to obtain the net displacement for this event near the 8-m scarp. Trench EO-3 contains evidence of a small, latest event, event d, for which no evidence was found in trench EO-2 on the same scarp. Our limited age control cannot confirm that events a and b are the same events in trenches EO-1 and EO-2, but TL analyses suggest the correlation shown.*

Trenches across the scarps of 16 and 26 feet (5, 8 m) exposed thick (greater than 10 feet; 3 m) sequences of bouldery stream- and debris-flow deposits overlying deltaic sands and gravels that were probably deposited near the Provo level of the Bonneville lake cycle about 14 ka. The thickness and geometry of multiple colluvial wedges adjacent to the faults exposed in the trenches allowed us to estimate the relative size of surface faulting events. Radiocarbon analyses of organic concentrates from soil A-horizons (apparent mean residence time or AMRT ages) developed on the wedges provided age estimates for faulting events, but uncertainties in the AMRT age estimates are at least ± 400 years. The colluvial stratigraphy and age estimates from the 16 foot (5 m) scarp suggested that two events each of about 7 feet (2.2 m) of displacement during the middle Holocene were followed by an event of 3 feet (0.9 m) in the late Holocene (figure 5). Stratigraphy and age estimates in trench EO-2 on the 26 foot (8 m) scarp revealed two middle Holocene events of 8 and 11 feet (2.5, 3.5 m) displacement, respectively, followed by a 4 foot (1.2 m) event during the late Holocene.

Mixing of near-surface stratigraphic units by burrowing and the interpretive problems with the AMRT age estimates make it difficult to determine if the 16 foot (8 m) scarp at the site has also been displaced by a small event within the past 500-600 years. Trench EO-3 showed stratigraphic evidence of a latest displacement event of less than 0.6 m (event d on figure 5), but burrowing has obscured some unit contacts and AMRT ages on burrowed units are hard to interpret. Trench EO-2 on the same 16 foot (8 m) scarp did not show any evidence of a small, latest event, but evidence of this event may have been destroyed by burrowing. The conflicting evidence from the trenches raises questions about the minimum size of displacement events that can be recognized in trenches, about whether large later events destroy evidence of small earlier events, and about the limits of AMRT ages in resolving fault events that may be spaced only 500-1000 years apart.

The results of radiocarbon analyses of three charcoal samples and 16 A-horizon concentrate samples from the five trenches at the site highlight many of the problems with using A-horizon AMRT ages on slopes to estimate the ages of normal faulting events. Of the 10 samples collected within 3 feet (1 m) of the present surface, two pairs of sample ages were inverted (lower sample younger than a stratigraphically higher sample) and four samples yielded greater than 112 percent modern carbon. These results are probably primarily due to incorporation of modern "bomb" carbon into the A-horizon samples by unrecognized mixing of surface sediment into the lower parts of the A horizon by burrowing. Several other inconsistent ages may also reflect reworking of old A-horizon sediment exposed in fault scarps into new horizons developing on colluvial-debris wedges adjacent to the scarps. Thus, faulting events recorded by near-surface units (less than 3 feet or 1 m depth) are difficult to accurately date using AMRT ages, at least at sites like East Ogden, which are favored by rodents.

Thermoluminescence (TL) analyses of fine-grained distal colluvium in trenches EO-1 and EO-2 (by S.L. Forman and J.P. McCalpin) also helped to constrain the age of the faulting events, particularly the earlier events that are poorly dated by radiocarbon analyses. A TL age estimate of 2500 ± 300 years is probably a maximum age for the second event in trench EO-1. Three age estimates from samples above the 2500-year sample suggest a non-uniform sedimentation rate for the distal scarp colluvium near the foot of the 16 foot (5 m) scarp, a result we would expect for colluvium deposited following multiple fault events. TL ages from trench EO-2 also constrain the maximum age of the second event on the 16 foot (8 m) scarp and agree with AMRT ages from stratigraphically equivalent samples. Thus, the TL age estimates suggest that the second rupture event in both trenches is the same event -- a conclusion we could not draw solely from our radiocarbon dates.

Trench stratigraphy and dating show that there have been at least three, and possibly four, surface faulting events across the two main scarps since the 5.5 ka debris-flow units were deposited. This conclusion compares favorably with Gilbert's (1928, p. 39) estimate of "one to at least four" post-Bonneville fault movements at different sites along the central segments of the Wasatch fault. Both of the main faults at East Ogden probably ruptured the ground surface during the three largest events (a, b, and c on figure 5). Stratigraphic displacements on the scarps range from less than 2 to 11 feet (0.6-3.5 m) displacement during event c seems to have been only about half that of events a and b. The recurrence of events in the last 6000 years has ranged from 400 to 2200 years, averaging about 1400 years. The overall slip rate for the fault zone at the site is about 6.5 feet/1000 years (2m)(post 5.5 ka).

CONCLUSIONS

Gilbert correctly interpreted most of the fault-related landforms in the Ogden area and must have worked out a relative history of faulting at selected sites. Most of the assumptions and methods that we used in our more detailed mapping of the area (Nelson and Personius, unpublished mapping, 1987) do not differ significantly from his, with the exception of our use of air photographs. In determining a detailed history of faulting events for the East Ogden site, including estimating the numerical age and relative size of events, and correlating the events at different sites along the fault, we have relied on radiocarbon analysis, thermoluminescence analysis, and trenching--techniques that have been developed only very recently in the century-long history of paleoseismic investigations of the Wasatch fault.

ACKNOWLEDGMENTS

Work at the East Ogden site would not have been possible without extensive logistical and financial support from the Utah Geological and Mineral Survey. Bob Klauk (UGMS), Mike Lowe (Weber County Planning Commission), and John Garr (Earthfax) helped log and interpret the trenches. Bill Lund (UGMS) and his staff provided advice and logistical support. We especially thank the Fife Equipment and Investment Company, Brigham City, for granting us permission to excavate on the site. Comments by Stephen Personius and Tim Sullivan substantially improved this paper.

TRANSGRESSIVE AND HIGH-SHORE DEPOSITS OF THE BONNEVILLE LAKE CYCLE NEAR NORTH SALT LAKE, UTAH

by

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INTRODUCTION

Extensive gravel-quarrying operations in two areas on the north side of the Salt Lake salient have intermittently exposed Lake Bonneville deposits from an altitude of about 165 feet (50 m) above the level of Great Salt Lake to the Bonneville shoreline, a vertical distance of about 820 feet (250 m) (figure 1). G.K. Gilbert (1890, 1928) did not make detailed observations in these specific areas, although he did spend much time studying other parts of the Salt Lake salient. He camped at the base of the west end of the salient for a short time and no doubt observed the piedmont on numerous train rides between Salt Lake City and Ogden. My studies of these exposures between 1979 and 1983 while quarrying was underway yielded stratigraphic information and numerous radiocarbon ages that aid in reconstructing aspects of the transgressive phase and high stand of the last, or Bonneville, lake cycle. The interpretations discussed here reinforce several of Gilbert's (1890) conclusions about lake history: (1) the intermediate shorelines (those lying between the Bonneville and Provo shorelines) pre-date the Bonneville shoreline, and many were deposited during the transgressive phase of the Bonneville lake cycle; (2) the lake level oscillated during the rise to and the stand at the Bonneville shoreline; and (3) the lake fell so rapidly from the Bonneville to the Provo shoreline that it left no record of its recession. In addition, new data from one of the localities discussed here are difficult to explain in terms of Currey and Oviatt's (1985) proposed chronology for the Bonneville lake cycle.

The following discussion of the key relations exposed in the quarries proceeds from the lower to the upper pits of the Concrete Products Corporation (CPC) and to cuts in Bonneville shoreline deposits in the MacNeish pit (figure 1). Bedrock in both localities consists of sediments of Tertiary age; alluvial-fan and other subaerial deposits of Quaternary age are also present.

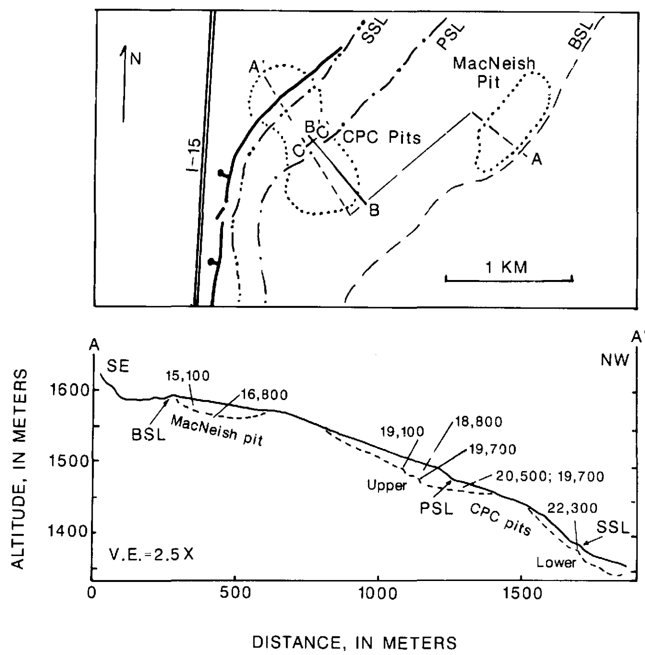


Figure 1. *Simplified map and schematic topographic profile of the piedmont on the northwest margin of the Salt Lake salient near North Salt Lake showing the Concrete Products Corporation (CPC) and MacNeish gravel pits in relation to the Stansbury (SSL), Provo (PSL), and Bonneville (BSL) shorelines. Post-Bonneville scarp of Warm Springs fault is heavy line with ball and bar on downthrown side. The MacNeish pit lies east of the CPC pit and is projected southwest onto the profile. Approximate locations of collection sites of radiocarbon-dated wood from deposits of the Bonneville Alloformation are also shown. Figure 2 shows cross sections B-B' and C-C' in upper CPC pit.*

CPC PITS

Bonneville Alloformation

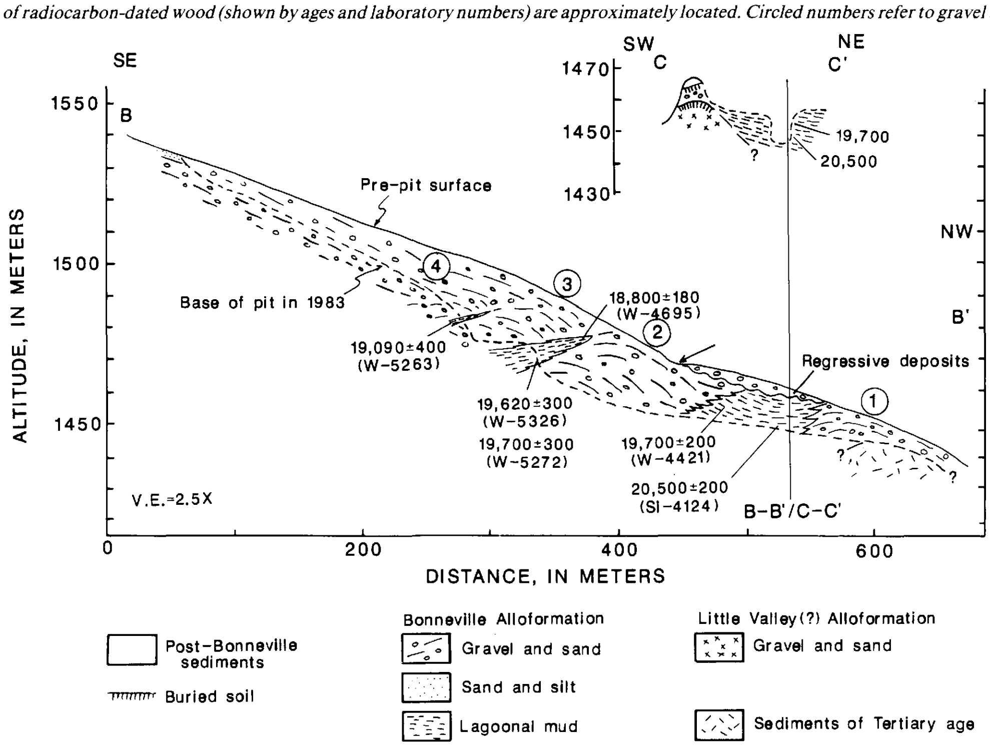
The Bonneville Alloformation (Pleistocene) exposed in the CPC pits consists mostly of gravel and sand deposited in bars and spits and of fine-grained sediments deposited in lagoons formed by the bars and spits. Minor sand and silt deposited in littoral and deeper-water zones are also exposed locally. Stratigraphic and age information indicate that most of the Bonneville Alloformation in the pits was deposited during the transgressive phase of the Bonneville lake cycle; very little sediment can be demonstrably related to the short-lived regressive phase.

Deposits of the Stansbury shoreline were exposed at the heads of small landslides that were active in the lower CPC pit during the wet spring of 1983. The sediments consist of a greater than 11.5 feet-thick (3.5 m) sequence of well-bedded and laminated brown to dark-gray mud that contains abundant wood including logs up to 15 inches (40 cm) in diameter. The mud also contains thin interbeds of rippled sand and pebble gravel and is overlain by beach gravel and sand, most of which has been quarried. The base of the mud is not exposed. These deposits lie at about the altitude of the Stansbury shoreline, which is defined by a discontinuous tufa-encrusted abrasion platform at 4495-4527 feet (1370-1380 m) on the west end of the Salt Lake salient. Evidently, the mud was deposited in a lagoon that formed behind a gravel bar at the Stansbury level and was subsequently buried by gravel of higher shorelines as lake levels rose. Although Gilbert (1890) interpreted the Stansbury as a recessional shoreline of the last lake cycle, stratigraphic evidence on Stansbury Island (Currey and others, 1983; see day 2-stop 2) and at the Old River Bed (Oviatt, 1987; see day 2-stop 4) shows clearly that the major erosional and depositional features of the Stansbury shoreline were formed during the trangressive phase of the Bonneville lake cycle. The age of two small branches collected from the mud, 22,300±400 yr B.P. (W-5269), is consistent with ages from other Stansbury shoreline deposits and suggests that the lagoon was probably related to the first of two recognized Stansbury oscillations (see day 2-stop 2).

Deep exposures immediately below and above the altitude of the Provo shoreline in the upper CPC pit reveal deposits of a complex of gravel bars and lagoons that date from the Bonneville lake cycle, beach deposits of an older lake cycle, and soils and subaerial deposits that both predate and postdate the Bonneville lake cycle. Figure 2 shows my interpretation of the stratigraphy in this area based on several visits between 1979 and 1983 as the pit was greatly enlarged.

The Provo shoreline formed when the level of Lake Bonneville stabilized after its rapid drop during the Bonneville Flood (Gilbert, 1890) and is clearly marked both north and south of the pit as a cliff and beach without significant bar or spit deposits. Although pre-quarry photographs show the shoreline crossing the pit site, Provo shoreline deposits are not clearly recognizable in the pit. I suspect the Provo deposits here are thin and were largely removed during early quarrying. Coarse-grained beach gravels exposed in the southwest part of the pit are not overlain by other lake sediments and may well be deposits of the Provo shoreline (Van Horn, 1982).

Figure 2. *Sketches of stratigraphic relations in the upper CPC pit. Line of cross sections located on Figure 1. Section B-B' shows pre-pit surface measured from topographic map and approximate base of pit in 1983 (dashed line). Most of the units and stratigraphic relations shown were viewed as excavations proceeded from 1979-1983; the regressive deposits at the Provo shoreline (PSL) are inferred. Section C-C' is normal to B-B; and shows only 1983 surface. Orientation of lines in stippled units and orientation of dashes in mud units indicate bedding attitudes. Jagged contact indicate intertonguing of units; wavy contacts, disconformities. Collection sites of radiocarbon-dated wood (shown by ages and laboratory numbers) are approximately located. Circled numbers refer to gravel bars discussed in text.*



On the basis of stratigraphic relations and radiocarbon ages, most of the deposits in the upper CPC pit must relate to the transgressive phase of the Bonneville lake cycle. In contrast to the thin deposits of the Provo shoreline, the transgressive deposits are thick and consist of a sequence of at least four gravel bars and interbedded lagoonal muds exposed over an altitude range of 165-197 feet (50-60 m). Each bar consists of both lakeward- and shoreward-dipping beds of mostly well-rounded and sorted gravel and sand that interfinger with lagoonal muds. Interbeds of rippled sand and minor gravel are also present in the muds. The lagoonal deposits occupy a small valley cut into older lake and subaerial deposits; these relations suggest that the rising lake flooded a dissected piedmont similar to today's.

Stratigraphic relations and six radiocarbon ages of well-preserved wood in the lagoonal muds indicate that the four gravel bars were deposited from lowest to highest (bars 1 to 4) between about 20,500 and 19,000 yr B.P. (figure 2). In two places, muds on both sides of a bar (bars 2 and 3) contain wood of similar age; therefore multiple lagoons existed at these times. The tops of the adjacent bars were at slightly different altitudes, and, in the process envisioned by Gilbert (1890, p. 141-142), they alternated as main sites of transport and deposition as the lake level fluctuated in response to minor climatic changes. In this process, a lagoon would form behind a bar fed by sediment carried in the shore drift. If the lake level were to rise rapidly enough, deposition on the bar would not keep pace and waves would cross the bar and begin to form another bar at the rear of the lagoon. The former lagoon would then be offshore and perhaps occasionally receive some coarser grained sediment from the shore drift. Ultimately the lagoon might be completely buried by the new bar. A subsequent fall to near the initial level would then activate deposition on the original bar. Consequently, the lagoon formed behind the second bar would be stranded and, depending on local hydrologic conditions, might exist as a pond lying slightly above lake level. Repetition of this process as lake level oscillated could have formed the complex sequence of overlapping and interbedded bar and lagoonal deposits that we see in the upper CPC pit.

If the shoreline deposits above the Provo shoreline in this pit had topographic expression, Gilbert would have called them intermediate shorelines. Clearly the stratigraphic and dating evidence here, and that collected in other areas in the Bonneville basin, support an interpretation that most of the intermediate shoreline features are deposits of the transgressive phase of the Bonneville lake cycle. This view differs from that of Hunt and others (1953) and many subsequent workers who interpreted the intermediate shorelines as deposits of a separate, older lake cycle (see Introduction to this volume). Gilbert didn't clearly state any interpretation other than that the intermediate shorelines pre-date the Bonneville shoreline. But did he think they were deposits of the transgressive phase of the White Marl (Bonneville) lake cycle, or deposits of an older lake cycle? In his discussions, he typically presented hypotheses, tested them, and came to some conclusion. The lack of this approach regarding the age of the intermediate shorelines is puzzling; perhaps he hadn't made up his mind. In most cases he doesn't mention a major drop in lake level between deposition of the units. For instance, Gilbert notes (1890, p. 170),

> *The order of sequence of the shores to which names have been given is: first, Intermediate; second, Bonneville... During the period of the formation of the Intermediate embankments, there were no persistent water stages; but the water surface oscillated up and down. The last additions to the embankments were made during a general advance of the water. The oscillation of the water surface continued through the Bonneville epoch, the Bonneville shore representing the combined results of wave action at a series of water levels having a vertical range of 20 feet.*

Similarly, in his discussion of the units that form the delta of American Fork, Gilbert (1890, p. 157; see day 3-stop 3) concludes,

> *In the order of time the Intermediate comes first and the Provo last. The Intermediate was built; the Bonneville was spread over its back, but failed to cover it completely; the lake fell, and the two were eroded by the creek, the Provo being formed at the same time.*

From these and other similar statements, I think Gilbert believed that many intermediate shorelines formed during the transgressive phase of the lake cycle that reached the Bonneville shoreline. The Introduction to this volume discusses the one case in which Gilbert identified intermediate shorelines that he thought dated from an earlier lake cycle.

Shore gravels with generally lakeward dips extend above bar 4 and are overlain by a sequence of fine sand and silt that was deposited in deeper water as the lake level stood near and at the Bonneville shoreline. Lack of evidence that a beach facies was deposited over these deeper water deposits supports Gilbert's and later workers' views that the fall from the Bonneville shoreline to the Provo was very rapid.

Older Lake Bonneville deposits

Deposits of an older lake cycle are exposed widely in the south-central part of the upper pit; however, no shells were found for amino-acid analysis, so their correlation is not known with certainty. They are probably deposits of the penultimate deep-lake cycle, the Little Valley, in as much as numerous other localities at altitudes close to the Provo shoreline contain deposits of this age (Scott and others, 1983; McCoy, 1987; see day 3-stop 2). The Bonneville and older deposits are separated by subaerial deposits, a disconformity with more than 32 feet (10 m) of relief, and a well-developed buried soil (Van Horn, 1982). This unconformity defines a landscape similar to that formed in post-Bonneville time, except that the buried soil is more strongly developed than the soil at the present land surface. An unconformity with these characteristics is seldom seen in a piedmont locality along the Wasatch front, unless exposure is deep, because of the thick cover of Bonneville Alloformation and post-Bonneville sediments.

Some previous workers along the Wasatch front mapped the Alpine Formation, the name given by Hunt and others (1953) to the deposits of the penultimate deep-lake cycle, over large areas at the surface. In the few areas in which the contact mapped between Bonneville and Alpine Formations is exposed, the units are essentially conformable. Considering the degree of dissection and modification of deposits of the Bonneville lake cycle and the nature of the unconformity between Bonneville and older lake deposits as seen here, the earlier mapping and interpretations appear to be erroneous (Scott and others, 1983; see Introduction to this volume; see day 3-stop 3).

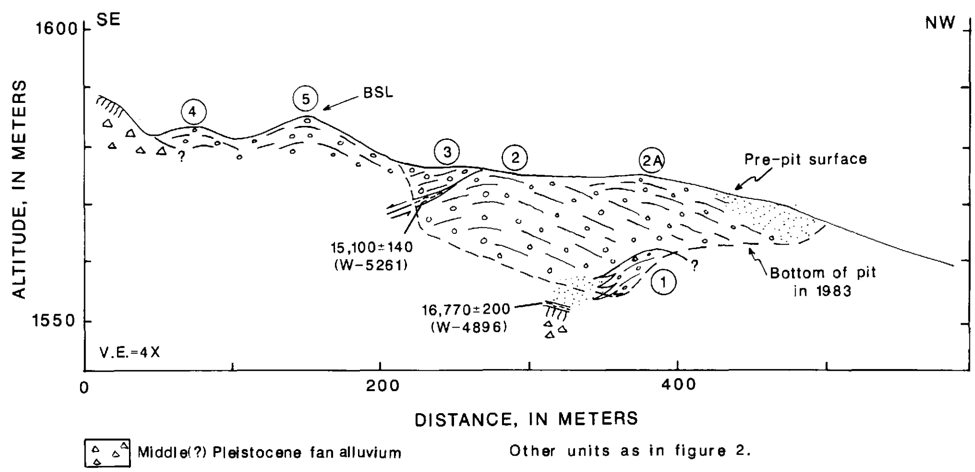
MACNEISH PIT

Deposits at and as much as 130 feet (40 m) below the Bonneville shoreline are well exposed in the MacNeish pit, one mile (1.5 km) east of the CPC pit (figure 1). Figure 3, an idealized cross section perpendicular to the shoreline through the central part of the pit, shows deposits of several overlapping gravel bars. Sections drawn parallel to the shoreline would show that the Bonneville sediments bury an incised and terraced topography formed in pre-Bonneville alluvial-fan deposits and sedimentary rocks of Tertiary age.

Stratigraphic relations and two radiocarbon ages place some constraints on the timing of lake stands at the Bonneville shoreline. Five major bars and spits are identifiable. The oldest bar (bar 1; figure 3), which was poorly exposed, dams a lagoon whose organic-rich mud overlies a strongly developed soil. Wood collected from the mud in the spoil of an exploratory trench that extended into the buried soil has an age of 16,770±200 yr B.P. (W-4896). Bar 2 and a bar on its lakeward flank (2a) overlie bar 1 and its lagoonal deposit. Bar 2 is overlain on its southeast side by a thin lagoonal mud containing wood that yielded an age of 15,100±140 yr B.P. (W-5261). Bar 3 overlies the mud. Sheared gravel lenses in the mud indicate that bar 3 slid downward and southeastward some unknown amount after its deposition. If substantial, this displacement would require that a depression (probably a lagoon) existed southeast of bar 3 into which the bar could slide. Bar 3 is apparently overlain by the highest depositional feature, bar 5, although the contact relations were not well exposed. Bar 4 is largely buried by bar 5 and is only locally distinct, while the relation between bar 4 and bars 1-3 is not known. The simplest interpretation of this bar sequence is that following the deposition of bar 1 and its related lagoon about 16,800 yr B.P., the lake level rose and a large compound bar (2 and 2a) prograded across the lagoon and bar 1. Bar 3 was deposited into the lagoon behind bar 2 sometime after 15,100 yr B.P. and may represent the continued enlargement of bar 2. The lake level continued to rise, and bars 4 and 5 formed. Bars 1-5 all lie above the altitude of the point of initial overflow of Lake Bonneville, the Zenda threshold in southeastern Idaho (Currey and others, 1983). Therefore, the rise of the lake level recorded in this bar sequence may partly reflect continued isostatic subsidence of the basin in response to earlier loading even after outflow had begun through the Zenda threshold (Currey, 1980; Scott and others, 1983).

The lake-level history revealed in the MacNeish pit does not fit well into a recently proposed model for the culminating events of the Bonneville lake cycle, summarized by Currey and Oviatt (1985), if the radiocarbon ages are interpreted literally (figure 4). In their model, the lake level reached the Bonneville shoreline at the Zenda threshold about 16,400 yr B.P., remained there until 15,900 B.P., then fell about 150 feet (45 m) by 15,300 B.P., and returned briefly to the Bonneville shoreline about 15,000 yr B.P., completing the Keg Mountain oscillation. One result of this oscillation in areas with great water loads was that some isostatic uplift occurred during the lake-level drop that was not completely reversed during the loading associated with the brief second maximum. Therefore, deposits of the second maximum are lower in altitude than those of the first. The second maximum ended shortly after 15,000 yr B.P., when the outlet was downcut catastrophically more than 330 feet (100 m) to the level of the Provo shoreline. Using this model, the age of 15,100 years suggests that bars 1 and 2 represent the first maximum and bars 3 and 5 the second.

Figure 3. *Sketch of stratigraphic relations in a section through the MacNeish pit. During 1981-1983, the pit consisted of numerous deep excavations separated by unquarried remnants that extended to or near the pre-pit surface. The sketch is composed of exposures located within 100 m of the line of section. The relations above the buried soil were reconstructed from a hole augered next to a backfilled exploratory trench and from interviews with the person who dug the trench. Circled numbers refer to gravel bars discussed in text.*



Two problems emerge from this model. First, the very brief second maximum is represented by a sizable volume of shore deposits, much more than in other areas (for instance the Great Bar at Stockton (Currey and others, 1983); see day 2-stop 5). Second, the deposits of the second maximum are higher than deposits of the first, inconsistent with the isostatic rebound argument mentioned above. How can the differences between the model and observations at the MacNeish pit be reconciled?

Obviously an age of 15,100 years for bar 3 would fit the model well in terms of timing, height, and volume, if it were the only deposit of the second maximum. Although the stratigraphic relations between bars 3 and 5 are not well exposed, geometric relations suggest that bar 5 overlies bar 3. The downward displacement of bar 3 to the southeast, if substantial, requires a depression to exist in the vicinity of bar 5 after 15,100 yr B.P. Therefore, this solution appears inappropriate.

One interpretation of the evidence in the pit is consistent with two lake-level maxima, the second of which is represented by deposits of smaller volume and lower altitude than the first. In this interpretation, bar 5 and some of the lower bars date from the first maximum and an erosional surface on bars 2 and 3 dates from the second. This surface truncates the bedding of bars 2 and 3 and displays a discontinuous lag of cobbles; the surface and lag were formed by wave erosion while the lake stood about 26-29 feet (8-9 m) below the top of bar 5. An argument based on differential isostatic deformation suggests that this 26-29 feet (8-9 m) of difference in altitude between the suggested maxima at the MacNeish pit is consistent with similar evidence from other areas in the Bonneville basin (Currey and others, 1983; Oviatt, 1987). For instance, the 1.5 ratio of the altitude difference between the two maxima at Stockton Bar (42.5 feet; 13 m) and at the MacNeish pit (26-29 feet; 8-9 m) is about the same as the 1.3 ratio of the difference between local and Zenda-threshold altitudes of the Bonneville shoreline at Stockton (144 feet; 44 m) and MacNeish (115 feet; 35 m). The similarity of these ratios suggests that differential isostatic deformation between the MacNeish pit and Stockton could account for the difference in the altitudes of the maxima at the two localities. If so, the sets of shorelines could be correlative.

If valid, the above reasoning suggests that estimates of the timing of the culmination of the Bonneville lake cycle need to be revised. Admittedly, the reasoning relies heavily on a single radiocarbon age and on stratigraphic and isostatic relations that rely on interpretation; however, this is the only radiocarbon age on wood from the Bonneville shoreline, and its importance should be evaluated. The 15,100-year age of the wood should probably be regarded as a maximum limiting age for the lagoonal mud, because it grew for some time before becoming incorporated in the deposit and the chances of contamination with young organic matter seem remote. The wood was found in an excellent state of preservation, contained no modern roots, and was collected from a fresh excavation. However, it was not oven-dried until 48 hours after being collected, so a chance existed for some bacterial fixing of atmospheric carbon while it remained sealed in a plastic bag.

The data from the MacNeish pit suggest that the Keg Mountain oscillation began after 15,100 yr B.P., perhaps significantly after in order to allow for deposition of bar 5, rather than at 15,900 years as estimated by Currey and Oviatt (1985). Their 15,300-year age of the low point of the Keg Mountain oscillation is based on six radiocarbon ages of carbonate shells (figure 4; Currey and Oviatt, 1985; Oviatt, 1987). The ages average 15,200 years and most have large errors. The carbonate ages have other potential sources of error, including contamination with younger carbon and low initial ^{14}C contents that would lead to apparent ages that are too old. Broecker and Kaufman (1965) originally subtracted 500 years from the ages to account for initially low ^{14}C contents. Considering such uncertainties, the wood could be older than the shells. The age of the second maximum at Stockton is limited by an age on tufa of 14,730±100 yr B.P. (SI-4227C), which is regarded as a minimum age because of the possibility of contamination with younger carbon (Currey and others, 1983; Currey and Oviatt, 1985; see day 2-stop 5). However, if the Keg Mountain oscillation is indeed younger than 15,100 years, the 14,700-year age may be closer to the true age than thought previously (see Currey and Burr, this volume). The timing proposed here restricts the duration of the 147-foot (45 m) oscillation greatly, perhaps to less than several centuries. Considering the full range of uncertainties in radiocarbon dating, including possible variations in radiocarbon content of the atmosphere during this time interval and stratigraphic interpretations, we may never be able to decipher the timing of these events accurately. Perhaps future excavations at the MacNeish pit can help to better define the stratigraphic position of the 15,100-year wood and provide additional radiocarbon samples.

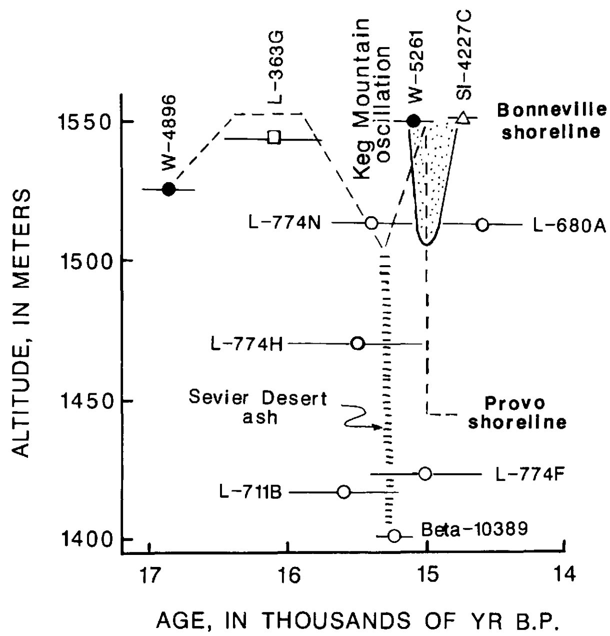


Figure 4. Reconstruction of lake-level fluctuations (dashed line) at culmination of Bonneville lake cycle (Currey and Oviatt; 1985). Possible position of Keg Mountain oscillation, as indicated by data in MacNeish pit, shown by stippled area. Key radiocarbon ages (with 1-sigma error ranges) that relate to these events are identified by laboratory number and the following symbols: solid circles, wood from MacNeish pit; square, tufa from near Bonneville shoreline; open circles, carbonate shells from Sevier Desert volcanic ash or stratigraphic units of Keg Mountain oscillation; triangle, tufa from deposits of latest occupation of Bonneville shoreline at Stockton Bar. Sample ages are plotted at altitudes that have been corrected for estimated isostatic effects. Shell ages based on 95 percent NBS modern standard; L-dates of shells were reported originally with 500 yr subtracted from them to account for a postulated low initial $^{14}C/^{12}C$ content (Broecker and Kaufman, 1965).

SUMMARY

Regardless of the refinements of lake history on which current workers are concentrating, stratigraphic evidence from the extensive exposures in the North Salt Lake area supports several of Gilbert's 100-year-old interpretations.

1) The intermediate shorelines pre-date the Bonneville and Provo shorelines and, although Gilbert's views are ambiguous, some, if not most, of the intermediate shorelines were formed during the transgressive phase of the Bonneville (White Marl) lake cycle.

2) The lake fell quickly from the Bonneville to the Provo shoreline during the Bonneville Flood and left essentially no record of its recession in that altitudinal range.

3) The lake level oscillated during times of no outflow, due primarily to climatic causes, and the prominence of the Bonneville and Provo shorelines reflects their formation during times that lake level was stabilized by outflow.

AN ALTERNATE INTERPRETATION OF DEPOSITS AT THE NORTH SALT LAKE CITY GRAVEL PIT, UTAH

by

Richard Van Horn
U.S. Geological Survey, Retired

Description and Interpretation of Measured Section S-379 (Modified from Van Horn, 1982)

A former outcrop near this stop (locality S-379, NW¼ SW¼ Sec. 12, T 1 N, R 1 W) displayed evidence of a possible significant fall and rise of Lake Bonneville after the last rise to the Bonneville shoreline (see measured section and figs. 1 and 2 in Van Horn, 1982). My interpretation of sediments exposed in this older part of the gravel pit conforms to generally accepted older theories on the geologic history of Lake Bonneville. They differ, however, from the newly proposed theories of Scott (1980; day 1-stop 4A) which, in part, depend on interpretations of sediments in this active gravel pit, which is about 330 feet (100 m) north to 1315 feet (400 m) northeast of my locality S-379.

According to my interpretation of locality S-379, the Alpine Formation was deposited in a lake that was higher than 4770 feet (1453 m) above sea level (units 15-23). It then fell below 4750 feet (1448 m) altitude and a strong soil formed (unit 14). The lake then rose above 4785 feet (1458 m), (presumably to 5220 feet (1590 m)) after which the younger and older members of the Bonneville Formation were deposited (units 11, 9, and 8) as the lake again receded. The lake fell to some unknown level below 4785 feet (1458 m) and another soil formed (unit 6). The lake then rose above this soil and the Draper Formation (unit 4) was deposited above 4790 feet (1460 m). After this the lake again receded. This interpretation follows the conventionally used (older) history of Lake Bonneville.

The eroded soil (unit 14, the Promontory Soil) passes under a lagoonal deposit about 355 feet (200 m) north of the locality S-379. A log collected in the lagoonal deposit by W.E. Scott of the U.S. Geological Survey yielded a ^{14}C date of 19,700 ± 200 yr B.P. (W-4421; Meyer Rubin, U.S. Geological Survey, written commun. to W.E. Scott, Dec. 17, 1979). I was not able to trace the lagoonal deposit into the exposure at S-379, but Scott believes the gravel overlying the lagoonal deposit can be traced around the head of the new gravel pit (this field trip stop) and into S-379 where it is represented by what I have called the younger and older members of the Bonneville Formation (units 11, 9, and 8) and what he would call the transgressive stage of the last lake cycle (see day 1-stop 4A). Scott does not recognize any deposits related to the Provo shoreline in these exposures. In addition, Scott does not believe that unit 4 is a lake deposit; if his interpretation were correct, unit 4 would not be the Draper Formation.

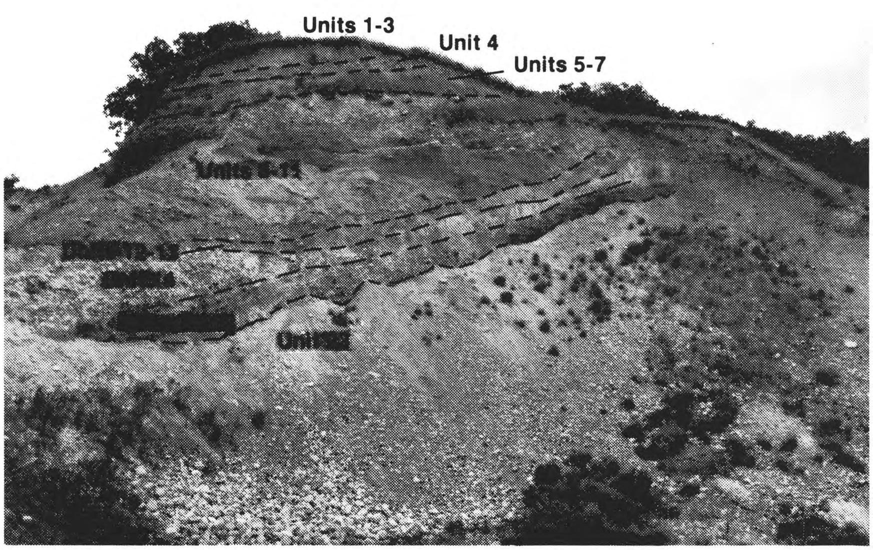


Figure 1. *Measured section S-379 showing units mentioned in the text. The outcrop is 55 feet (17 m) high. Units 15-22 are lake deposits of the Alpine Formation indicating a relatively high-level lake at this time. Units 12-14 are colluvium indicating that the Alpine lake had dropped in altitude and that a strong soil had developed on the terrestrial deposits (Promontory Soil, unit 14) while the lake was at the low level. Units 8-11 are mostly lacustrine deposits of the older and younger members of the Bonneville Formation (although there are some terrestrial deposits near the middle); these indicate a generally rising lake. Units 5-7 are terrestrial showing the lake had again receded. Unit 4 is a lacustrine deposit of the Draper Formation indicating the lake had risen again. Units 1-3 are terrestrial deposits that were deposited as the lake receded for the last time in this area.*

G.K. GILBERT'S OBSERVATIONS OF POST-BONNEVILLE MOVEMENT ALONG THE WARM SPRINGS FAULT, SALT LAKE COUNTY, UTAH

by

William E. Scott
U.S. Geological Survey, Vancouver, Washington

INTRODUCTION

Extensive gravel quarrying at the west end of the Salt Lake salient (City Creek spur of Gilbert, 1890, 1928) has locally revealed the spectacularly striated footwall of the north-trending Warm Springs fault. The exposed bedrock consists of a southeasterly dipping sequence of Mississippian, Devonian, and Cambrian limestones and fault breccias derived from them (Van Horn, 1981). The striations record two slip directions, a southwest set superposed on an older northwest set. These slip directions, combined with fault-plane solutions of nearby earthquakes, suggest that the fault is favorably oriented within the present stress field for generating future surface faulting (Pavlis and Smith, 1979).

The extensive industrial development in this area and the strike of the fault south toward the center of Salt Lake City (figure 1) emphasize the need for detailed information about the rupture characteristics and recent slip history of the Warm Springs fault. Unfortunately, the quarrying that has exposed the fault plane has also removed most of the Quaternary deposits and fault scarps in unconsolidated materials that could be used to provide such information. Fortunately, perhaps due to the Powell Survey's low *per diem* rates, a desire to elude the bright lights and fast pace of big-city living, or an appreciation of a hot-spring bath at the end of a hard day, Gilbert chose to camp near the Warm Springs fault, between Becks and Wasatch thermal springs. During a few days in early July, 1877 he made some key observations that help to unravel the fault's recent behavior.

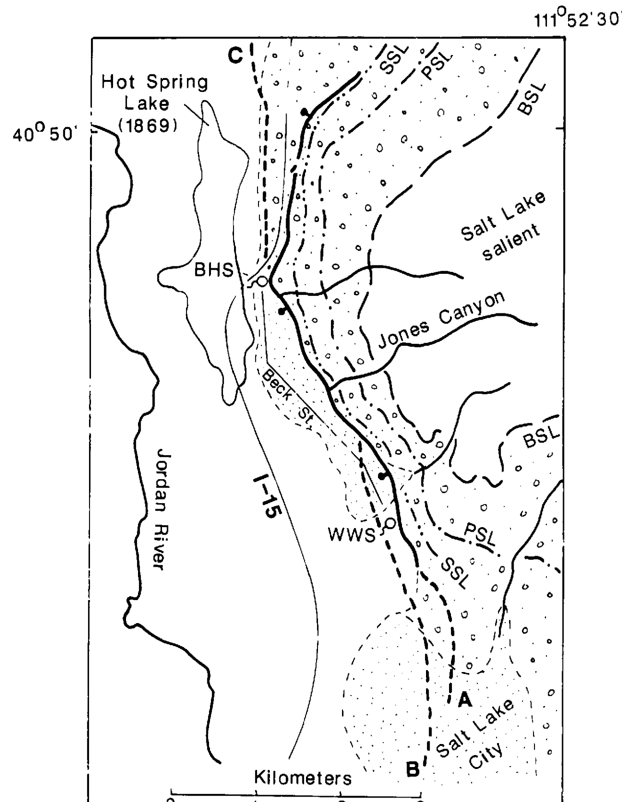


Figure 1. *Map of the Warm Springs fault showing extent of known post-Bonneville surface faulting (heavy solid line; ball on downthrown side) and inferred southern extent (heavy dashed line) as mapped by (A) Kaliser (1976) and (B) Marsell and Threet (1964). Scarp (heavy dashed line C) trending north from Becks Hot Spring (BHS) is interpreted by Van Horn (1982) as extension of Warm Springs fault, but is probably an erosional scarp at Gilbert shoreline. Quaternary deposits generalized from Van Horn (1982) and Scott and Shroba (1985) Hot Spring Lake is now drained. WWS, Wasatch Warm Springs; BSL, Bonneville shoreline; PSL, Provo shoreline; SSL, Stansbury shoreline.*

GILBERT'S OBSERVATIONS

The following observations by Gilbert (1890, 1928; Hunt, 1982a), which were made in a very short time period and were supplemented by several views while passing on a train, provide an understanding of the post-Bonneville history of surface offsets along the Warm Springs fault.

He judged that the steep bedrock scarps at the base of the salient, from which he calculated a throw of 46 feet (14 m), and the gentler slopes above indicated a recent interval of more rapid displacement that followed a time of relatively little slip during which the slope was worn back. Gilbert's photograph of the scarp (figure 2) shows a steep exposed face of limestone breccia, which could be mistaken for the free face of a fault scarp formed in unconsolidated colluvium. Gilbert envisioned that the total throw on the west end of the salient (1475 feet, 450 m minimum) was accomplished in alternating episodes of rapid and slow uplift.

He attributed the 0.6-mile-wide (1 km), shallow Hot Spring Lake and marsh (now drained; figure 1) between the Jordan River and the Warm Springs fault to recent subsidence and noted that the river would tend to quickly fill such a depression with sediment (Gilbert, 1928, p. 23; Hunt, 1982a, p. 27). It's not clear if or how he thought the subsidence here might be related to recent fault activity. He frequently described the backtilting of small blocks adjacent to piedmont fault scarps; however, his only mention of backtilting of a block several kilometers long toward the Wasatch fault that I've found is in his notes (Hunt, 1982a, p. 71), not in any published work.

Figure 2. Gilbert's photograph (1928, plate 5A) of a fault scarp formed in limestone breccia at the west end of the Salt Lake salient. Mule for scale; calculated throw on fault is about 46 feet (14 m).

His key observation along the Warm Springs fault was made at the fan of Jones Canyon, just south of his camp, where he could trace the bedrock scarp into piedmont scarps formed in fan alluvium of post-Lake Bonneville age (figure 3; Gilbert, 1890, p. 348-349; field notes *in* Hunt, 1982a, p. 29). Gilbert (1890) wrote (p. 349):

> *The portion of the alluvial cone that lies above the fault scarp is channeled by the stream, and a study of the system of terraces bordering this channel shows that the total displacement of 30 feet was produced by at least three independent movements, the measures of the parts being 15 feet, 5 feet, and 10 feet.*

Gilbert's work probably represents the inaugural application of a now classic technique that uses tectonically produced stream terraces to deduce recurrent fault activity. However, he did not account for, and may not have recognized, the likelihood of partial burial of a scarp in this geomorphic position occurring between faulting events. Therefore the total 30-foot (9-m) offset is a minimum value for post-Bonneville slip, which is probably similar to the 40 to 46 feet (12-14 m) that he measured along the bedrock scarp. In addition, the larger individual offsets that he estimated may have been the products of more than one event.

EXTENT OF POST-BONNEVILLE SURFACE FAULTING

Gilbert's work clearly shows that the Warm Springs fault has generated at least three surface-faulting events since the end of the Bonneville lake cycle; however, the extent of surface faulting north of Becks Hot Spring and south of Wasatch Warm Springs is less clear. Gilbert (1890, p. 348; Hunt, 1982a, p. 20) implied that the fault scarp of post-Bonneville age begins near Wasatch Warm Springs and continues north with a throw as great as 46 feet (14 m) to Becks Hot Spring at the western point of the salient. He notes the scarp then turns northeast and "fades and tapers", but it's not clear how far north Gilbert traced the scarp. Aerial photographs taken in 1952, before the gravel pits north of Becks Hot Spring consumed most of the piedmont, and mapping by A.R Nelson and S.F. Personius (written communication, 1988) show that the fault scarp does turn northeast at the hot springs for a short distance, but then continues north and northeast for another 1.8 miles (3 km). A gap containing a few short scarps extends for several kilometers north of the end of the Warm Springs fault scarps. This gap marks the boundary between the Salt Lake and Weber (Ogden) segments of the Wasatch fault zone (Schwartz and Coppersmith, 1984; Ogden segment renamed the Weber segment by Machette and others (1987)).

Van Horn (1982) mapped the Warm Springs fault for more than 4.4 miles (7 km) north of Becks Hot Spring along a scarp that lies west of Interstate Highway 15 and follows a pronounced gravity gradient that was interpreted as a fault by Cook and Berg (1961). This scarp does not join the bedrock fault scarp at Becks Hot Spring, which clearly turns northeast at the spring. If the scarp mapped by Van Horn is a fault scarp, it would represent a left-stepping trace of the Warm Springs fault. However, the base of this scarp is horizontal except where younger alluvium has been deposited against it and it lies at an altitude of about 4250 feet (1295 m); both observations suggest that it is a wave-cut scarp of the 10,000-year-old Gilbert shoreline.

The extent of the Warm Springs fault south of Wasatch Warm Springs is not as well known as its northern extent. Maps by Marsell and Threet (1964) and Kaliser (1976) show inferred faults extending into the center of Salt Lake City. However, from Gilbert's descriptions and the rapidly decreasing scarp height south of Wasatch Warm Springs, it's unlikely that significant post-Bonneville surface faulting extends much south of 6th North Street (figure 1; Van Horn, 1982; Scott and Shroba, 1985).

IMPLICATIONS FOR EARTHQUAKE HAZARDS

The Warm Springs fault represents a significant hazard in the Salt Lake City area. The total length of demonstrable post-Bonneville surface rupture along the fault is about 4 to 5 miles (7-8 km). From Gilbert's observations the total post-Bonneville slip occurred in at least three events and is as much as 40 to 46 feet (12-14 m), which is similar to that on other parts of the Salt Lake City segment of the Wasatch fault zone (Scott and Shroba, 1985; Schwartz and Lund, this volume), and occurred in at least three events.

An intriguing problem is whether surface-faulting events on the Warm Springs fault occur independently of events on the East Bench and main trace of the Wasatch faults to the south, or whether all three form a pattern of *en echelon* breaks during a single event on the Salt Lake City segment. Movement of the Warm Springs fault might also be triggered by surface-faulting events on the Weber (Ogden) segment to the north. Gilbert's (1890) interpretations at Jones Canyon suggest that individual slip events on the Warm Springs fault have measured 5 feet (1.5 m) or more. Although comparative data are sparse, this amount of displacement is greater than that expected for a surface fault of less than 6 miles (10 km), based on statistical relations between rupture length and displacement for historical events (Bonilla and others, 1984). Displacements of 5 feet (1.5 m) or more are typically related to ruptures tens of kilometers long. Although interpretation of the rupture length-displacement relations involve large uncertainties, including conditions that control the surface expression of displacements at depth, surface-faulting events on the Warm Springs fault may occur as part of a longer break that involves adjacent faults.

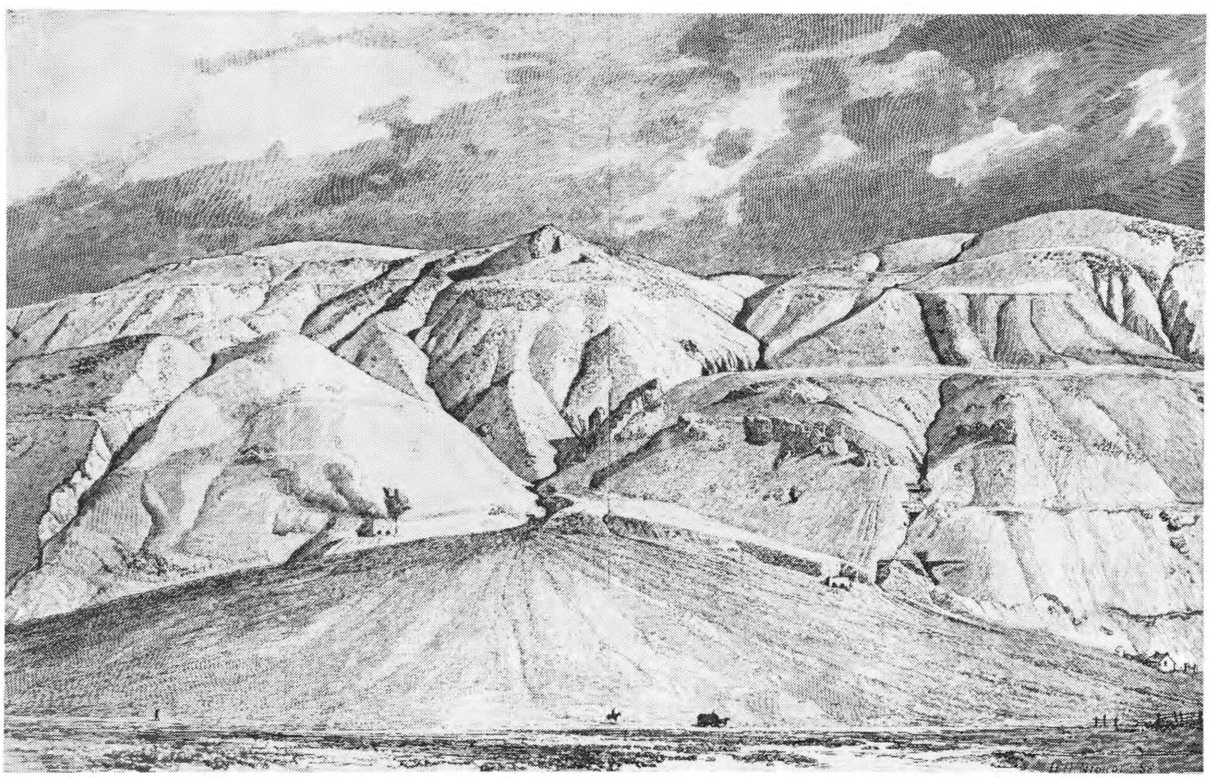
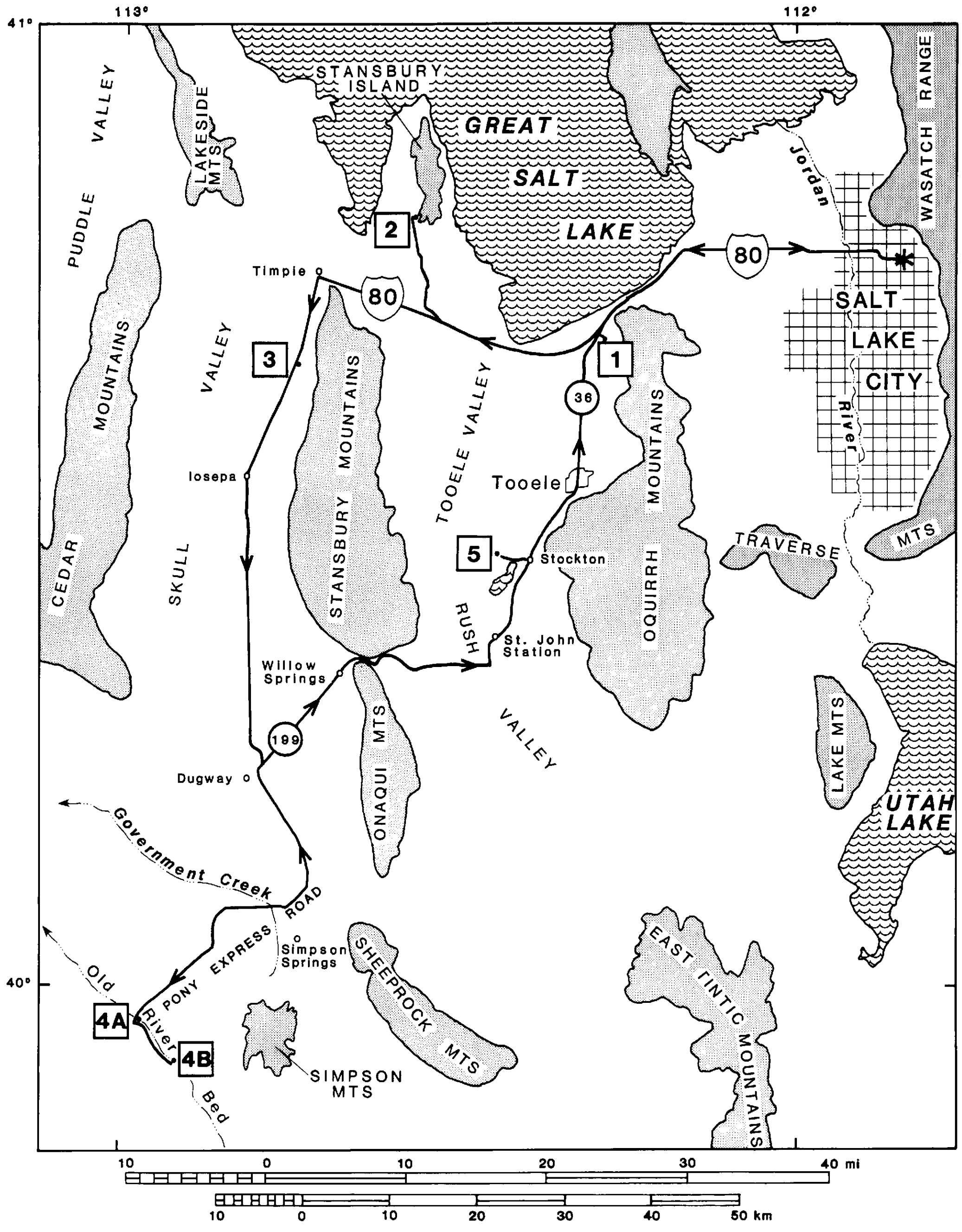


Figure 3. *Sketch by W.H. Holmes of the faulted alluvial fan of Jones Canyon (Gilbert, 1890, plate 44). Note lime kiln on scarp to left of canyon and scarp above house to right of canyon. From bottom to top on right side of sketch are the Stansbury, Provo, Intermediate, and Bonneville shorelines.*



41°
113°
112°
40°
STANSBURY ISLAND
GREAT SALT LAKE
PUDDLE VALLEY
LAKESIDE MTS
WASATCH RANGE
Jordan River
SALT LAKE CITY
Timpie
80
2
3
1
36
CEDAR MOUNTAINS
SKULL VALLEY
STANSBURY MOUNTAINS
TOOELE VALLEY
Tooele
OQUIRRH MOUNTAINS
TRAVERSE MTS
Iosepa
5
Stockton
St. John Station
RUSH VALLEY
Willow Springs
199
ONAQUI MTS
LAKE MTS
UTAH LAKE
Dugway
Government Creek
ROAD
PONY EXPRESS
Simpson Springs
Old River Bed
4A
4B
SHEEPROCK MTS
SIMPSON MTS
EAST TINTIC MOUNTAINS
10 0 10 20 30 40 mi
10 0 10 20 30 40 50 km

DAY 2
Road Log from Salt Lake City to the Old River Bed and Return

by

Michael N. Machette and Donald R. Currey

Field trip leaders: Theodore P. Barnhard, Donald R. Currey, Ted N. Burr, Susan A. Green, Charles G. Oviatt, and William D. McCoy

The second day of our field trip takes us west of Salt Lake City (figure 1) to examine neotectonic features and lacustrine deposits of the Bonneville lake cycle. Our first stop (1) is at the northwest end of the Oquirrh Mountains where, in the early morning sun, we will see fault scarps on lake sediments. From there, we continue west on I-80 across the south end of Great Salt Lake. We cross a causeway to the south end of Stansbury Island (stop 2) to examine early transgressive deposits related to the Stansbury shoreline of the Bonneville lake cycle. Next is a long trek to the south, through the Skull Valley (stop 3), past Dugway Proving Grounds, and to Gilbert's classic sections in the Old River Bed (stop 4A). After lunch, we will examine the white marl and yellow clay near The Shutoff, our turn-around point for this day and stop 4B.

Returning eastward, we will cross thick post-Bonneville eolian sands that are piled up against the west side of the Stansbury Range. After crossing the Stansbury Range at Johnson Pass, we will descend to Rush Valley and visit one of the world's largest non-marine barrier-spit complexes—the Great Bar at Stockton, Utah (stop 5).

Mileage Description of features along route

0.0 Intersection of 4th South and 7th East (Residence Inn). Proceed west on 4th south.

1.5 Turn left (south) on 2nd West.

1.6 Turn right (west) on 5th South (onramp to I-15).

2.6 Take feeder road north to I-80 west (parallels I-15).

3.4 Merge with I-80 westbound.

8.4 International Center (business park) on right (north). This complex is built on the Holocene floodplain and delta of the Jordan River. Lake levels during the Holocene have been as high as 4221 feet (Currey and others, 1988), whereas the historic levels have ranged from highs of 4211.5 feet in 1873 and 4211.8 feet in 1847 to a low of 4191.4 feet in 1963. In August of 1988 the lake level was about 4208 feet.

12.8 Morton Thiokol salt processing plant on left (south). A short distance to the north is the drill site of the Saltair core (Eardley and Gvosdetsky, 1960).

18.0 Overpass at the new Saltair Resort (on right). Ahead you can see the Magna Smelter at the north end of the Oquirrh Mountains (on left) and Great Salt Lake (on right).

22.9 Take exit 99 (Utah Hwy 36 to Stansbury and Tooele) and proceed south to Lake Point.

24.2 Turn left (east) into Oquirrh Motor Inn and proceed to rear of their parking lot for stop 1.

DAY 2-STOP 1 discussion by Barnhard.

Mileage Description of features along route.

0.0 Oquirrh Motor Inn. Retrace route to I-80 westbound.

0.7 Enter I-80 westbound. For the next several miles we will traverse the south edge of Great Salt Lake, which is retained by a large dike built on the north side of the Western Pacific Railroad alignment.

11.4 Burmester Siding north of the overpass. This is the drill site of the Burmester core, the longest published record of Bonneville basin paleolake sediments (Eardley and others, 1973).

15.8 Take exit 84 (marked Utah Hwy 138 to Grantsville) and turn right (north). Turn left on frontage road, cross railroad tracks that pass under I-15. Salt evaporation ponds in this area are operated by the American Salt Company.

16.6 Turn right (north) on causeway road and pass over main alignment of Western Pacific Railroad. In 1877, Gilbert noted that "the bar to Stansbury Island was first covered by water 2 years ago (sic, 1875). It is now belly deep to a horse" (Hunt, 1982a, p. 45). However, according to Currey and others' (1984) compilation of historic levels of Great Salt Lake, the lake had peaked at an altitude of about 4212 feet in 1873 and by 1877, when Gilbert travelled through this region, it had receded to 4210 feet.

20.2 Fork in causeway road, take left fork, which leads around west edge of Stansbury Island. Near this site Gilbert wrote "I have noticed for two days in approaching Stansbury Island that the most conspicuous beach on it is the Provo. The BB (Bonneville beach) can be detected at a few points only" (Hunt, 1982a, p. 45). To the south, Gilbert measured a long profile across the shorelines on the north end of the Stansbury Mountains and refers to the Oquirrh beach (now known as the Stansbury beach). Although he never visited Stansbury Island, he states that the Oquirrh (Stansbury) beach is better developed than the Bonneville. This impression may have influenced Gilbert's cautious belief that the Oquirrh (Stansbury) was younger than the Bonneville and Provo beaches.

22.1 Four-way intersection with dirt road marked by white sign. Turn right (east) and drive through large gravel pit that provided material for rebuilding the causeway and dikes during the recent high lake stand.

22.9 Park above the east end of gravel pit and walk up to diatomite quarry at the mouth of Stansbury Gulch (new formal geographic name) for stop 2.

DAY2-STOP 2 discussion by Green and Currey.

Mileage	Description of features along route.
0.0	Retrace route to causeway road.
0.8	Turn left (south) on causeway and retrace route to I-80.
6.8	Enter I-80 westbound toward Wendover, Nevada. Proceed around the north end of the Stansbury Mountains (on left).
14.3	Timpie. Take exit 77 marked Rowley and Dugway. Proceed south toward Dugway. This unnumbered highway skirts the eastern edge of the Skull Valley and western edge of the Stansbury Mountains. There are four major springs that issue from the base of alluvial fans just west of the highway. Burnt Springs, the northernmost of the springs, is aligned with the fault that we will see at stop 3.
20.3	Turn left (east) onto dirt road. Park just off highway for stop 3.

DAY 2-STOP3 discussion by Barnhard.

Mileage	Description of features along route.
0.0	Return to highway, turn left and proceed south.
9.4	Iosepa. This community was established as a leper colony for Polynesian Mormon converts. Iosepa is pidgin English for Joseph.
17.7	Entering Skull Valley Indian Reservation.
30.8	Junction with Utah Hwy 199, which leads west to the main entrance to Dugway Proving Grounds. Continue south on gravel road which parallels the Proving Grounds perimeter fence.
32.5	BLM road marker "9 miles to Pony Express Road." Road turns southeast and departs from Proving Grounds perimeter fence.
38.2	Fork in road. Take right (south) fork to Simpson Springs (22 miles ahead).
40.5	Crest gravel bar graded to highest shoreline of Lake Bonneville. This feature is known as the Davis Mountain tombolo.
41.2	Intersection with Pony Express Road. Turn right (southwest).
42.9	Crossing Government Creek, an ephemeral stream that drains a large part of the Sheeprock Mountains to the southeast and Simpson Mountains to the south. We will cross at least three more transgressive bayhead barriers in the next several miles.
46.7	To the north (right), the large barrier at the Provo level has been breached on its northern end by Government Creek.
47.2	Provo shoreline wraps around a bedrock knob north of road. To the southwest the Provo shoreline is cut on bedrock and alluvial-fan deposits.
47.7	Simpson Buttes to the west. This inselberg is partially buried by underflow-fan deposits of the ancient Sevier River.
50.7	Simpson Springs, site of a restored Pony Express Station.
58.1	Road bends to left (south) and descends to Old River Bed channel. The sediments exposed along this edge of the valley, which has been formally named White Marl Bluff, are typical of those studied by Gilbert in November of 1879 (see day 2-stop 4A).
58.6	Cross channel of Old River Bed valley. Turn south on dirt road (note sheet metal sign marked Old River Bed road.)
61.1	Crest of hill is known as The Shutoff. To the right (west), Slow Elk Wash has built an allluvial fan across the Old River Bed, thereby segmenting the channel.
61.9	Follow minor dirt road to the left (east).
62.2	Small stock reservoir. Park here for lunch and a short discussion of the Old River Bed. Stop 4B will be on exposures to the north and east.

DAY 2-STOPS 4A and 4B discussion by Oviatt and McCoy.

Mileage	Description of features along route.
0.0	Retrace route north to Pony Express Road.
3.6	Turn right (east) and proceed to Simpson Springs.
11.5	Simpson Springs--Pony Express Station. Water and toilet facilities are available here.
21.0	Turn left (north) on road to Dugway (same route as before).
31.0	Junction with Utah Hwy 199. Turn right (east) and proceed to Willow Springs.
39.0	As we approach the base of the Stansbury Mountains, you can see fault scarps north of the road (day 2-stop 3)
39.7	Road cuts along the highway expose a thick section of eolian sands that postdate the Bonneville lake cycle. These sands form a thick ramp at the mountain front, not unlike a a small-scale version of the Great Sand Dunes of the San Luis Valley (Colorado).
40.5	Willow Springs. Paleozoic rocks are exposed just east of here. Proceed east on Utah Hwy 199 (toward Clover and Rush Valley).
43.0	Johnson Pass. From here you get a nice view of the Rush Valley.
47.8	Emerge from Clover Creek Canyon onto the piedmont slope east of the Stansbury Mountains. Rush Valley is a large closed basin and Rush Lake, at the north end, is the remnant of a formerly large, late Pleistocene lake.
53.0	Junction with Utah Hwy 36. Turn left (north).
55.2	Cross railroad tracks at Saint John Station.
56.8	Junction with Utah Hwy 73, proceed north on Utah Hwy 36. When Lake Bonneville dropped from its highest level about 14.5 ka, the Stockton Bar (see day 2-stop 5) at the northern end of Rush Valley became a natural dam to the remaining lake in the valley. This stranded remnant of Lake Bonneville is termed Lake Shambip by Currey after usage by Mormon pioneers.
62.0	Entering Stockton, Utah. Turn left (west) on Silver Avenue, (at the gas station and Stockton Cafe). About 1.5 miles to the north, deep cuts in the Stockton Bar for the railroad provide excellent exposures of transgressive sands and gravels of the Bonneville lake cycle.
62.2	Proceed west and cross railroad tracks. The road curves to the north around Rush Lake, and the prominent concave-to-the-south shorelines were formed by Lake Shambip.
62.7	At the northern extent of the Lake Shambip shorelines, turn right (north) on the dirt track.
62.9	Fork in road; take left (northwest) fork. Note the two cross-valley spits which curve to the north. These spits were formed during transgression of Lake Bonneville into the Rush Valley.
63.3	Climb onto the Stockton Bar. Continue northwest to gate in fence (mileage 23.0). From here we will walk to the top of the small hill which affords an excellent view of the Stockton Bar and the Tooele and Rush Valleys.

DAY 2-STOP 5 discussion by Burr and Currey.

Mileage	Description of features along route.
0.0	Retrace route to Utah Hwy 36 at Stockton.
1.3	Stockton, Utah. Turn left (north) onto Utah Hwy 36 and proceed to Tooele.
2.8	Crest of Stockton Bar.
3.7	South end of steep high bluff in alluvial fans that predate the Bonneville lake cycle. This bluff has been interpreted both as an erosional shoreline and as a fault scarp; it is the former (see day 2-stop 5, fig. 6).
5.4	North end of bluff.
6.5	Abrasion platform of Lake Bonneville cut in quartzite bedrock.
8.0	Downtown Tooele. Proceed north on Utah Hwy 36 towards I-80.
9.6	We cross the Provo shoreline on the north side of Tooele. The railroad tracks cross the highway about 0.5 mile north of the Provo shoreline.
16.1	To left (west) at Stansbury Park you will see a gulch that heads down to an old restored mill, which is built on the Gilbert shoreline.
19.7	Onramp to I-80 eastbound to Salt Lake City. The northern end of the Oquirrh Mountains has a well-preserved sequence of tufa-cemented beach gravels at the Stansbury and Provo shorelines.
40.6	Take offramp to I-15 southbound, but get in the exit lane for 6th South. Proceed east on 6th South.
43.3	Turn left (north) on 7th East.
43.6	Intersection of 4th South and 7th East. End of road log for day 2.

FAULT-SCARP STUDIES OF THE OQUIRRH MOUNTAINS, UTAH

by

Theodore P. Barnhard
U.S. Geological Survey, Denver, Colorado

INTRODUCTION

The field of paleoseismology has become an important player in the roll of assessing earthquake hazards. Much of paleoseismology is a new and developing science and many current paleoseismology concepts were only developed in the last several decades. Recent technological advances in dating materials, remote sensing, and seismic reflection profiling have afforded geologists many tools to decipher the histories of past geologic events.

Geologists also rely on previous studies by other geologists to advance the state of knowledge of the science. Basin and Range studies by G. K. Gilbert in the late 1800s and early 1900s have stood the test of time with many of his concepts and conclusions still valid today. Part of Gilbert's studies involved examining fault scarps, determining their origin, and the general age of faulting that produced the scarps. Gilbert used crosscutting relations of fault scarps with geomorphic features of Lake Bonneville as well as the morphology of the scarp itself to infer the age of faulting that produced the scarp.

G. K. Gilbert's pioneering work in the Basin and Range led him to use fault scarp morphology as an indicator of age in studying paleoseismicity as indicated by the following statement (Gilbert, 1928, p. 39).

> *The relative ages of piedmont scarps are shown in part by stages of weathering and in part by relation to other features. The most recent have scarcely or perhaps not wholly attained the slope of rest (sic, angle of repose). Their upper edges are sharply defined, and their faces are not fully occupied by vegetation. In those of earlier origin the edges are rounded, the faces have gentler slopes, and benches above them have been trenched by the drainage, so that faces are divided into facets. On the aprons of the ranges are seen scarps in all stages of progress toward obliteration, but illustrations of the more advanced stages are not readily to be found along the Wasatch front because the greater part of the piedmont tract has been remodeled by the shore work of the waves of Lake Bonneville. Below the highest shore line all pre-Bonneville scarps have been either obliterated or so nearly effaced that their identification is problematic, and the only unquestioned records are those of post-Bonneville time. Fault-block ranges without piedmont scarps and also parts of such ranges that lack the scarps have presumably felt no paroxysm of growth for tens of thousands of years.*

Gilbert (1928) also drew some of the first profiles of fault scarps and used them to determine the amount of throw on the fault (figure 1). He realized that the amount of throw (a-b on figure 1) on the fault is often less than the scarp height. Gilbert's scarp profiles were the forerunners of modern scarp morphology techniques.

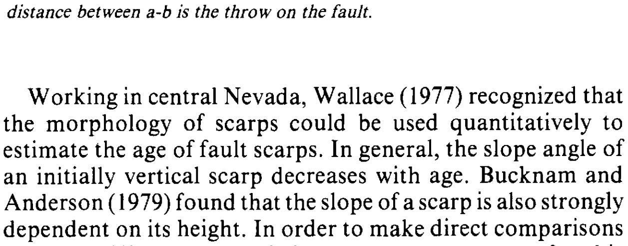


Figure 1. *Fault scarp profile from Gilbert (1928, figure 28, p. 34). The vertical distance between a-b is the throw on the fault.*

Working in central Nevada, Wallace (1977) recognized that the morphology of scarps could be used quantitatively to estimate the age of fault scarps. In general, the slope angle of an initially vertical scarp decreases with age. Bucknam and Anderson (1979) found that the slope of a scarp is also strongly dependent on its height. In order to make direct comparisons between different scarps it is necessary to account for this dependence on scarp height by normalizing data from various scarps to the same height. If profiles for each scarp or group of scarps are measured over a broad range of scarp heights, then an empirical morphometric relation of maximum scarp-slope angle (θ) against scarp height (H) can be defined. The most commonly used relation is between θ and log of H. The resulting regression equation can then be used to compare scarps of different ages or to help clarify scattered data.

The regression equation derived from the highest Bonneville shoreline profiles (R.C. Bucknam, written communication, 1982) is used in this study as a benchmark against which regression equations from profiled fault scarps can be compared. The Bonneville shoreline is about 15 ka (Scott and others, 1983; Currey and Oviatt, 1985), so if a regression equation plots above the Bonneville shoreline regression line it is considered to be younger than 15 ka and if it plots below the Bonneville line it is considered to be older than 15 ka. The profiles of the Drum Mountains fault scarps are also used to compare regression lines. The Drum Mountains scarps are considered to be about 9 ka by Pierce and Colman (1986) (figure 2).

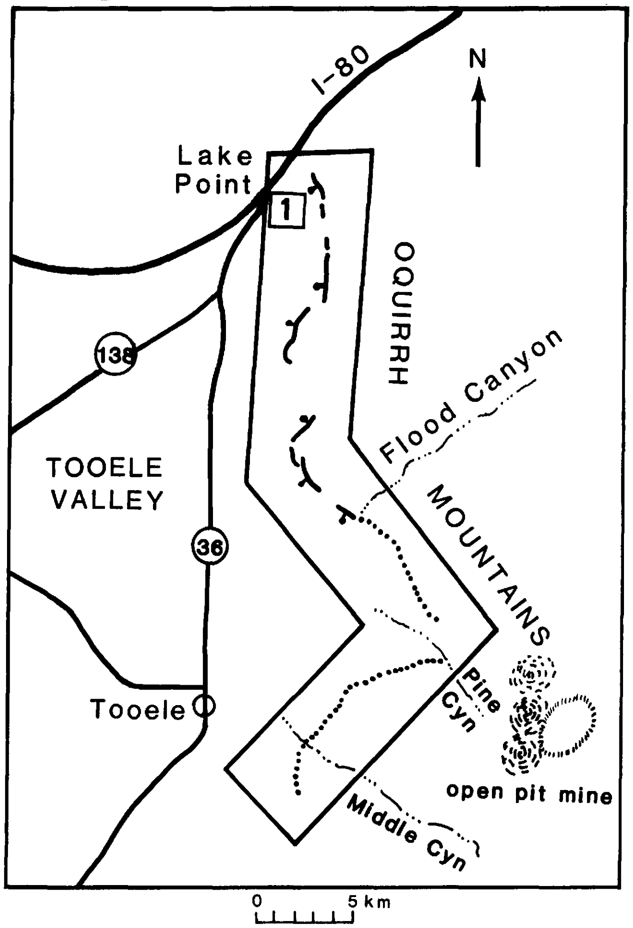


Figure 2. General location map for scarps of the Oquirrh fault zone. This stop is shown by the square and #1. Solid lines in boxed area are fault scarps in alluvium with bar and ball on down thrown side. Dotted line is fault trace at bedrock-alluvium contact.

SCARPS OF THE OQUIRRH FAULT ZONE

Gilbert's Studies

Gilbert visited the Tooele area several times, first in July of 1880 (Hunt, 1982a) and later in July and September of 1901 (Gilbert's unpublished field notebooks, numbers 90 and 94). In 1880, Gilbert approached the town of Tooele from the north and noted (Hunt, 1982a, p. 170):

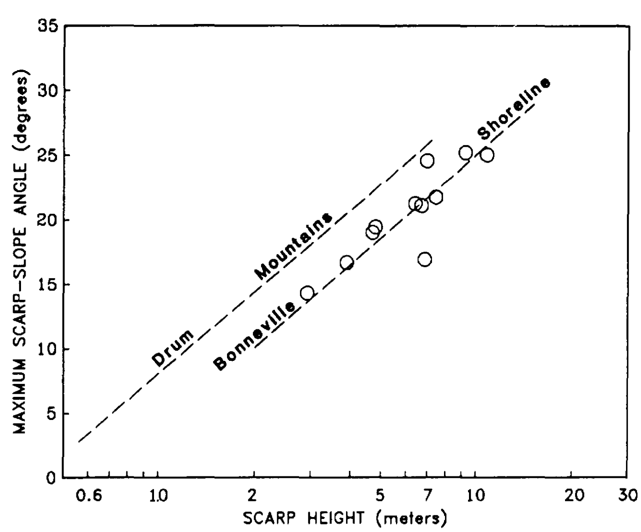
Approaching Tooele there is a fine view of the fault scarp in the embayment of the mts (mountains). It is high up on the flank of the all. (alluvial) cones running steeply down to B.B. (Bonneville beach). It is distinctly marked to West cañon (Canyon) and a mile beyond and then it disappears by running behind foothills of rock. In two miles it reappears at the south for short distance at least. I am not sure there have been any post-Bonneville movement but think there is.

Recent Studies

I have mapped a series of prominent down-to-the-west faults that offset Quaternary alluvial deposits on the west side of the northern end of the Oquirrh Mountains (figure 2 in Barnhard and Dodge, 1988). A zone of prominent scarps 9.5-35 feet (2.9-10.8 m) high on alluviium extends discontinuously for 10.5 miles (17 km) from north of Lake Point to just north of Flood Canyon. The northern part of the fault zone lies close to the range front and is buried in many places by active alluvial fans issuing from the Oquirrh Mountains. In the middle part of the fault zone, scarps trend away from the range front and extend northwest into Tooele Valley, where they are less modified by alluvial fans that postdate the surface faulting. At the southern end of the zone, the fault scarp trends back to the range front and forms the bedrock-alluvium contact near Flood Canyon.

Fault scarps along most of the Oquirrh fault zone are compound scarps, i.e., they are formed by several discrete displacement events. However, at three locations near the northern end of the fault zone, the trace of the most recent surface faulting diverges from an older scarp; here four scarp profiles were measured on the most recent scarp (figure 3). The compound scarps are as much as twice as high (15.7 versus 35.4 feet (4.8 vs 10.8 m)) as the most recent single-event scarp and are clear evidence of recurrent faulting.

Figure 3. Scarp-morphology data from the Oquirrh fault zone. Circles represent individual measurements of scarp-height and maximum-slope-angle. Line of best fit from the regression-equations for 9-ka Drum Mountains scarps (Pierce and Colman, 1986) and for 15-ka Lake Bonneville highstand shoreline scarps (R.C. Bucknam, written communication, 1982) are shown as dashed lines.



The fault scarp generally lies topographically below the prominent Provo shoreline of Lake Bonneville (4850 feet; 1478 m in this area) in the northern and middle parts of the scarp zone. However, at three locations, the fault scarp crosses the abrasion platform of the Provo shoreline, indicating that the latest surface-faulting event was post-Provo in age (13.5 ka from Scott and others, 1983; Currey and Oviatt, 1985).

Data for the four profiles of the most recent surface faulting, which is defined by single-event scarps with scarp heights less than 16 feet (5 m), plot parallel to and above the line for the 15-ka Lake Bonneville shoreline data, but below the line for the approximately 9-ka-old Drum Mountains, Utah fault scarps (figure 3). This relation suggests that the age of the most recent surface faulting for the Oquirrh fault zone is between 9 and 15 ka. Although the data plot closer to the 15-ka regression, the scarps are clearly less than 13.5 ka because the most recent faulting offsets the Provo level abrasion platform. Therefore, I favor an age between 9 ka and 13.5 ka for the latest faulting event on the Oquirrh fault zone.

Gilbert's assessment of this fault zone was that there was post-Bonneville movement. Assuming that Gilbert meant post-high stand for "post-Bonneville" his evaluation of the youthfulness of faulting on this zone was correct.

THE STANSBURY SHORELINE AND OTHER TRANSGRESSIVE DEPOSITS OF THE BONNEVILLE LAKE CYCLE

by

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INTRODUCTION

The Stansbury shoreline, with the Bonneville and Provo shorelines, is one of the three most conspicuous former lake levels in the Bonneville basin. G.K. Gilbert observed the Stansbury shoreline at several localities, including the western salient of the Wasatch Mountains immediately north of Salt Lake City, the north ends of the Oquirrh and Aqui (now Stansbury) Mountains, and the south end of Stansbury Island. Gilbert (1890, plate III) viewed from a distance but never visited the latter locality. Nevertheless, it was the basis for the name "Stansbury shore-line" that he (Gilbert, 1890, p. 134) eventually selected for this lacustrine landmark. The illustrious I.C. Russell, while employed as Gilbert's principal assistant, did visit the south end of Stansbury Island in July, 1880.

At each of the above localities the Stansbury shoreline clearly displays the geomorphic features that distinguish it at many steeply sloping sites, and especially in re-entrants at steep sites: a bouldery or rocky shore platform bearing an apron of tufa-cemented gravel (tufaglomerate). The higher part of the apron is a nearly flat tufaglomerate caprock that paves the platform surface to its landward limit, about the level of the formative water plane. The lower part of the apron is a tufaglomerate drapery that formed as a lakeward extension of the caprock, within photic depths on the steeply shelving near-shore bottom. At some localities features such as barriers and lagoons occur in close spatial and temporal association with the Stansbury tufaglomerate; at other localities beaches and the tufaglomerate caprock are subordinate, and fluviodeltaic features are dominant. The complete set of Stansbury-related littoral features can be regarded as the Stansbury shoreline complex (SSC); the spatiotemporal range of the SSC has been referred to as the Stansbury oscillation or oscillations (e.g., Currey and Oviatt, 1985). For purposes of regional mapping and neotectonic analysis, the tufaglomerate caprock serves well as the definitive Stansbury paleolake datum. For field trip purposes, the tufaglomerate caprock and adjacent deeper-water lithofacies will serve to introduce fundamental stratigraphic relations at what is, by Gilbert's choice of nomenclature, the Stansbury shoreline type area. At this stop, in Stansbury Gulch on the south end of Stansbury Island, the field trip will focus on evidence that SSC deposits, and the topographically expressed Stansbury tufaglomerate caprock in particular, were laid down in a relatively early phase of the Bonneville transgression, and not during the regressive phase of the Bonneville lake cycle.

Basin-wide delineation of the SSC, partly in detail (Currey, 1980, plate 1) and partly in reconnaissance, indicates that at the Stansbury stage the lake had an area of about 9200 square miles (24,000 km^2) and a maximum depth of about 425 feet (130 m). Maximum differential hydro-isostatic rebound among points on comparable morphostratigraphic facies of the SSC seems to be no more than about 33 feet (10 m), which is probably only a minimum estimate of Stansbury-stage hydro-isostatic depression; as much as 85 feet (26 m) or more of later depositional loading (e.g., Eardley, 1962, figure 1) has probably prevented the central basin from fully rebounding to a pre-Stansbury configuration.

From the time of Gilbert, workers have implicitly or explicitly tended to presume that Stansbury shore features formed during the regressive phase of the Bonneville lake cycle. Partial exceptions to that presumption were registered by Antevs (1948, p. 173), who attributed the Stansbury shoreline in large part to a "Lake Stansbury" episode postdating Lake Bonneville and predating "Lake Provo," and Eardley and others (1957, p. 1186), who held open the possibility that earlier "hesitations" at the Stansbury level may have preceded a dominant stand in post-Provo time.

STRATIGRAPHY

Stratigraphic relations in Stansbury Gulch are represented in measured sections 1 through 6 of figure 1. In section 1 two thin sand layers that occur 4 inches (10 cm) apart near midsection are important stratigraphic markers. The sand layers contain shells of the gastropod *Amnicola* and are conformably underlain and overlain by laminated micritic carbonate that varies from very sandy to slightly sandy. The underlying micritic sequence is aragonitic, contains ostracodes, and has laminae with abundant filaments of charophyte chalk. The lowest package of charophyte-rich laminae has yielded a $^{13}C/^{12}C$-adjusted ^{14}C age of 24,870 $\pm$ 410 yr B.P. (Beta-8343). Below the dated horizon the sequence is extremely sandy, but excavations so far have not been deep enough to expose the beach gravels that almost certainly mark the base of the Bonneville Alloformation. The micritic sequence that overlies the pair of sand layers is calcitic, contains ostracodes and diatoms, and becomes increasingly sandy near the top of the section, where the Bonneville Alloformation is overlain by post-Bonneville colluvium.

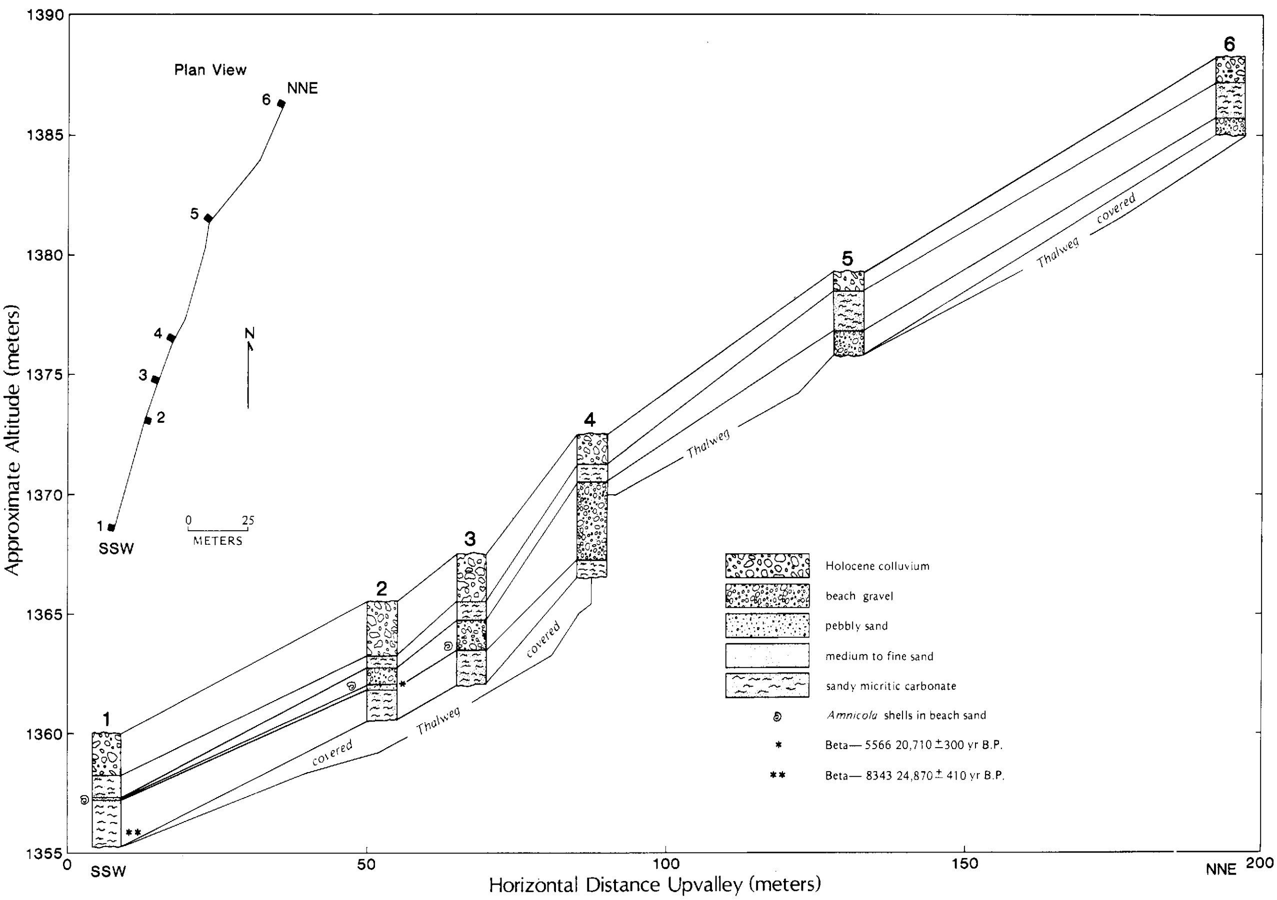

Figure 1. Measured stratigraphic sections at the type Stansbury shoreline in Stansbury Gulch, near the south end of Stansbury Island. Survey by Glenn B. Plyler (Currey and others, 1983, fig. 3).

Up-gulch, from section 1 to section 2, the two thin sand layers thicken into a 4-inch (10-cm) bed of pebbly sand at the base of a 20-inch (50-cm) bed of locally derived limestone and quartzite cobbles and small boulders. The sand at the base of the coarse gravel contains abundant *Amnicola* shells, which have yielded a $^{13}C/^{12}C$-adjusted ^{14}C age of 20,710 ± 310 yr B.P. (Beta-5566); because the dated material was 100 percent aragonite, with no calcite detected by x-ray diffraction, there is little reason not to interpret the age literally. The sand and gravel in the middle of section 2 is underlain and overlain by laminated micritic carbonate sequences that are similar to those at section 1.

Farther up-gulch, the gravel bed thickens dramatically, from about 3 feet (1 m) at section 3 (figure 2) to over 10 feet (3 m) at section 4. This shoreward lateral change is accompanied by a substantial increase in tufa cementation, which at section 4 culminates in a resistant caprock of tufaglomerate at the top of the gravel. Sessile algae in the shore zone were probably important as food for gastropods and as bio-inducers of tufa deposition. Later, diagenetic carbonate probably contributed to the induration of the caprock. The gravel at section 3 (figure 2) and section 4 is underlain and overlain by essentially the same sequences of laminated micritic carbonate that occur in the lower and upper portions of sections 1 and 2.

It should be noted in the field that the tufaglomerate caprock in section 4 is the topographic platform that is perceived as the Stansbury shoreline when Stansbury Island, as well as many other islands and headlands, is observed from a distance. More importantly, stratigraphic relations in Stansbury Gulch clearly indicate that Lake Bonneville: (1) transgressed to above the level of section 4, perhaps about 25,000 yr B.P., prior to deposition of the tufa-cemented gravel; (2) regressed from above the level of section 4 to one or more stages near that level, but probably not much lower, perhaps about 20,700 yr B.P.; (3) transgressed to levels well above section 4 after about 20,700 yr B.P.; and (4) eventually regressed to levels below sections 1 through 6 without pausing long enough to deposit or erode any feature that can be construed as a shoreline, let alone a shoreline of basin-wide prominence.

ACKNOWLEDGMENTS

The research summarized here was supported in part by U.S. Geological Survey Contracts 14-08-0001-18289 and 14-08-0001-G1536 and by the Research Committee of the University of Utah.

Figure 2. *The west wall of Stansbury Gulch, near measured section 3 in figure 1. The 3-foot (1-m) thick nearshore gravel in the middle of the section grades shoreward (upslope) into a 10-foot (3-m) thick tufa-cemented boulder bench which is paved with the tufaglomerate caprock that is so typical of the Stansbury shoreline as a regional landform. Underlying the nearshore gravel are alternating layers of ripple-laminated micritic carbonate and cross-laminated sandy micritic carbonate that contain variably abundant aragonite, ostracodes, and charophyte debris; convolute lamination occurs in the layer 8 to 12 inches (20-30 cm) below the gravel. Overlying the nearshore gravel is laminated micritic carbonate with diatoms and ostracodes, including an ostracode coquina 27 inches (70 cm) above the gravel. Above the coquina the section becomes increasingly sandy and then is overlain by post-Bonneville colluvium. Neither at this locality nor at any other Stansbury shoreline locality do deposits of a distinctive regressive shoreline occur at the top of the section that was laid down during the Bonneville lake cycle. Photography by Susan A. Green, 1987.*

FAULT-SCARP STUDIES OF THE STANSBURY MOUNTAINS, UTAH

by

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U.S. Geological Survey, Denver, Colorado

GILBERT'S STUDIES

Gilbert first observed fault scarps on the west flank of the Stansbury Mountains, in Skull Valley, on July 21, 1877 (Hunt, 1982a, p. 46):

> *The fault bench is unmistakable along here but it is the first I have felt sure of (it) in this range. There is no special evidence of recency.*

On a subsequent visit to Skull Valley in July of 1901 (Gilbert's unpublished field notebook number 90, p. 13) he wrote:

> *5 miles S. (south) I find a fault scarp, a double scarp. It first appears at the base of a small low rock spur and runs obliquely SE to the main mt base, skirting and crossing an all. (alluvial) cone. I trace the scarp, with some doubt, for about 2 miles, to a grove, where low ls. (limestone) spurs run far into the valley. (p. 15, 16) Later I am satisfied of a fault scarp for several miles beyond Matthews (now Deseret) ranch.*

RECENT STUDIES

Fault scarps along the north-trending Stansbury fault zone are formed on Quaternary basin-fill deposits along the west flank of the Stansbury Mountains. The zone of scarps is nearly continuous for 19 miles (30 km) from Deadman Canyon northward to beyond Broons Canyon (figure 1). The main down-to-the-west faulting is accompanied by antithetic faulting, thereby forming a narrow (65 ft/20 m wide) graben along much of the southern part of the zone. It is this feature which Gilbert described as a double scarp.

At Chokecherry Canyon, the main and antithetic faults bifurcate; the antithetic fault scarp curves westward and the main fault scarp curves eastward, creating a structural graben as wide as 1.5 miles (2.5 km). The throw on the antithetic fault diminishes and dies out at Box Canyon, about 3 miles (5 km) north of Chokecherry Canyon. The main fault continues northward to just north of Broons Canyon, where it intersects

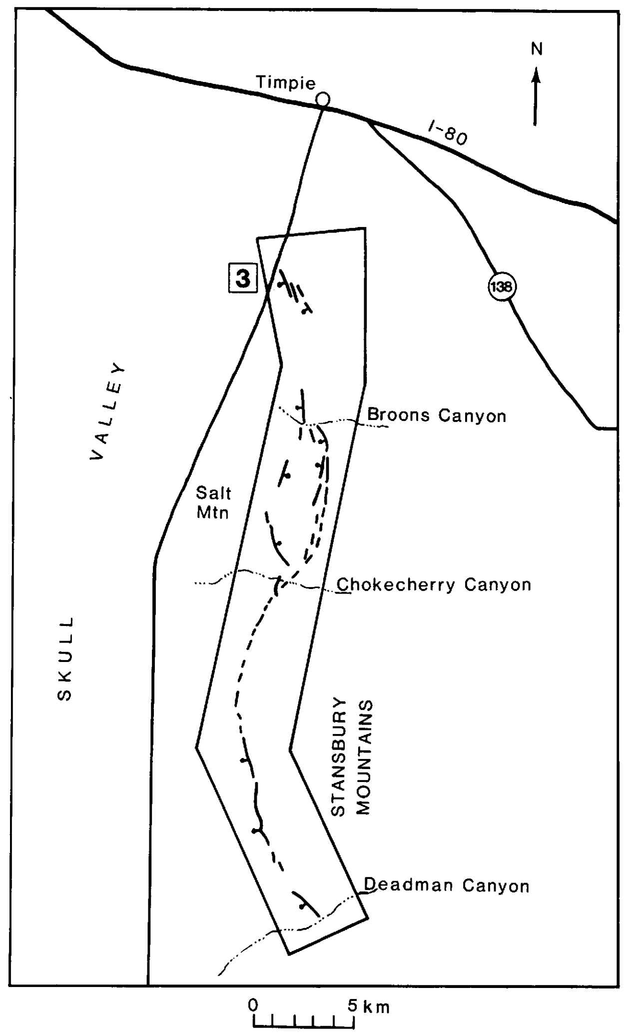

Figure 1. *General location map for scarps of the Stansbury fault zone. This stop is shown by the square and #3. Solid lines in the boxed area are fault scarps on alluvium with bar and ball on the downthrown side.*

the highest shoreline of Lake Bonneville. The shoreline-fault intersection is at a bedrock outcrop, so no interpretations of relative-age relations can be established. At the northernmost end of the zone, a series of en echelon fault scarps trend north-northwest and are the only scarps in the zone topographically below the highstand shoreline of Lake Bonneville.

Scarp heights are larger in older alluvial deposits than in younger deposits along the length of the fault zone, indicating recurrent movement on the Stansbury fault zone. Fault scarp data from nine profiles along the Stansbury fault zone plot subparallel to and below the Bonneville regression line (figure 2; also see day 2-stop 1), suggesting that the most recent surface-faulting event probably occurred before occupation of the Bonneville highstand shoreline (i.e., before 15 ka).

Gilbert's assessment of the surface faulting along the Stansbury Mountains was that it showed "no special evidence of recency." If recency meant that the faulting was pre-Bonneville highstand in age, he again was correct. Gilbert's ability to correctly infer the age of faulting along the Stansbury Mountains is remarkable since most of the trace of the scarp lies above shoreline features of Lake Bonneville. Therefore, he must have used the morphology of the fault scarp to infer the relative age of surface faulting.

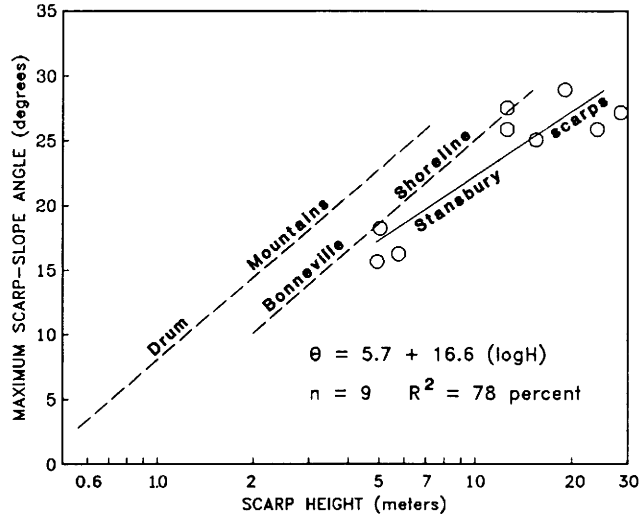

Figure 2. Scarp-morphology data from the Stansbury fault zone. Circles represent individual measurements of scarp-height and maximum-slope-angle. Solid line is the line of best fit from the regression equation derived from scarp profiles from the Stansbury fault zone. Line of best fit from the regression-equations for 9-ka Drum Mountains scarps (Pierce and Colman, 1986) and regression equation for 15 ka Lake Bonneville highstand shoreline scarps (R. C. Bucknam, written commun., 1982) are shown as dashed lines.

THE OLD RIVER BED

by

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INTRODUCTION

Gilbert (1890, p. 181-182) described the Old River Bed in the following passage:

> *The overland stage road which, before the day of Pacific railroads, carried the mail across the Great Basin, skirted the southern margin of the Great Salt Lake Desert. From Salt Lake City to Canyon Station, at the eastern base of the Deep Creek Mountains, its route lay almost entirely upon the bed of Lake Bonneville. Midway it crossed a broad channel, which every one recognized as an ancient river bed. Here a stage station was established and a change of horses was kept. The horses were not watered by the river, nor even by a diminutive modern representative of it, but by means of a well sunk to a depth of 100 feet. Now that the road has fallen into disuse and earth has clogged the neglected well, the chance traveller finds nothing to quench his thirst from Simpson spring to Fish Spring, a distance of 40 miles. One who stands here in the midst of a desert, where the only vegetation is a scattering of low bushes, and looks on an ancient river course 2,000 feet broad and more than 100 feet deep, can not fail to be deeply impressed.*

Gilbert recognized that the Old River Bed was carved by a river that flowed northward from a shallow overflowing lake in the Sevier Desert, late in the history of Lake Bonneville. More recent work shows that the Sevier Desert lake (Lake Gunnison) overflowed at the Old River Bed threshold (figure 1) from about 12,000 to 10,000 yr B.P. (Currey, 1982; Oviatt, 1988).

The Old River Bed is significant in Lake Bonneville stratigraphic studies because Gilbert spent approximately seven days studying the exposures of Lake Bonneville deposits along the channel margins (Hunt, 1982a, p. 125-139, 182). He spent at least another seven days in the Old River Bed area exploring, mapping, and leveling shorelines. There is no question that Gilbert spent more time studying Lake Bonneville stratigraphy at the Old River Bed than at any other place in the basin. As a result, he proposed the Old River Bed as the type section of the Bonneville beds (Gilbert, 1890, p. 189-190) and based many of his interpretations of the lake's history on his work in this area. At this field trip stop we will visit both Gilbert's Lower River Bed ("typical") section, and his Upper River Bed section.

STOP 4A — LOWER RIVER BED SECTION (type section)

Gilbert (1890, p. 189-190) described the Lower River Bed Section as follows:

> *The deepest section of the lake beds, or more strictly the section representing the largest fraction of the Bonneville Period, is exposed in the walls of the Old River Bed near the point where it is crossed by the Overland Stage-road. It has some title to be regarded as the typical section, and exhibits the following members:*

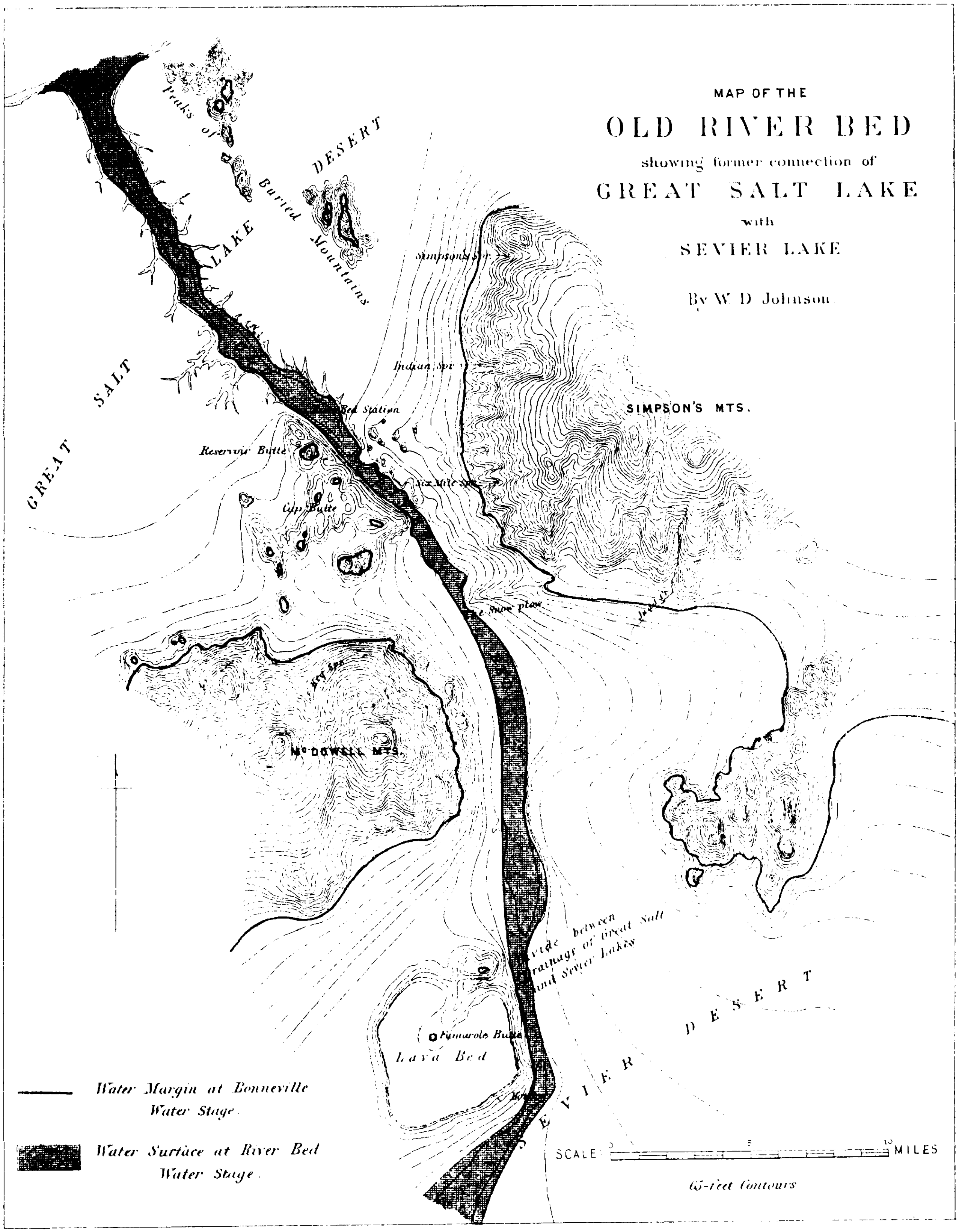


Figure 1. Map of the Old River Bed (from Gilbert, 1890, p. 182, plate XXXI). Gilbert's Lower River Bed section (stop 4A) is near the River Bed Station; the Upper River Bed section (stop 4B) is at the constriction in the channel about 2.5 miles (4 km) southeast of the River Bed Station; the Old river Bed threshold, as determined from modern topographic maps, is about 5 miles (8 km) north of where Gilbert placed the divide.

1. (At base.) The Yellow Clay, a fine argillaceous deposit, laminated throughout, olive gray on its fresh exposure, but weathering to a pale yellow. In this are occasional passages of sand, but these are local and discontinuous. Nodules of selenite, consisting of grouped arrow-head crystals, are abundant; and jointage cracks sometimes contain rosettes of recrystallized gypsum. Bivalve shells of several species are included. The base is not seen; a thickness of 90 feet is exposed.

2. The White Marl, a fine calcareous clay or argillaceous marl, light gray or cream-colored on fresh exposure, nearly white on weathered surface. Contains some gypsum, but less than No. 1. Overlies No. 1 with unconformity by erosion, and is at its base crowded with shells representing nearly the same fauna. Thickness, 10 feet.

3. The marl passes upward into a fine sand, the transition being gradual and the continuity perfect. The sand contains also the same species of shells. Thickness, about 10 feet, the upper limit being obscured by a recent eolian deposit of similar texture.

It is unclear exactly where Gilbert measured this section because the measurements do not match any of the sections recorded in his field notes (Hunt, 1982a, p. 125-139, 182). Probably any well-exposed section near the Overland Stage Road is an acceptable choice as the Lower River Bed section because the stratigraphic sequence is similar for long distances in this area. However, Gilbert's interpretation of the pebbly sand channel-fill deposits, which are exposed along the east side of the Old River Bed, must be considered. Gilbert (1890, p. 183) interpreted the pebbly sand as post-Bonneville alluvium deposited on top of the white marl at low altitudes, and within the partially entrenched Old River Bed channel at higher altitudes (Hunt, 1982a, p. 125-126, figure 11.3 and 11.4). Subsequent work (Varnes and Van Horn, 1961) shows that the pebbly sand at the stage-road crossing is stratigraphically within the yellow clay (figure 2), and that the pebbly sand Gilbert observed above the white marl at low altitudes is a younger stratigraphic unit (Oviatt, 1984, 1987).

The white marl and lower yellow clay at the Lower River Bed section (figure 2) can be physically traced to the Upper River Bed section where they also appear, but their lithologic character changes gradually in this distance. The lithologic changes were caused by differences in the local geomorphology and, thus, in the depositional settings between the two measured sections.

STOP 4B — UPPER RIVER BED SECTION

Figures 3 and 4 show Gilbert's descriptions of the stratigraphic sequence at the Upper River Bed section. These figures will form the basis for our discussions at Stop 4B.

Gilbert referred to the Upper River Bed section as "the wedge locality" in his field notes (Hunt, 1982a, p. 128-131, 135-136, 137-138, 182). The "wedge" refers to a truncated gravel unit (FG, First Gravel) underlain by the yellow clay and overlain by the white marl (figure 3). In his field notes Gilbert repeatedly wrote of the First Gravel as a lacustrine bar, and there is little doubt that in the field he interpreted the gravel in that way (Hunt, 1982a). However, in Monograph 1 (Gilbert, 1890, p. 195) he noted that "... it is possible that an interlacustrine river was the agent of transportation." He reasoned that regardless of the origin of the First Gravel, it represented an episode of relatively low lake levels between periods of deeper water. We interpret the First Gravel as part of a sequence of gravel beaches and spits that were deposited when Lake Bonneville first reached altitudes between about 4540 and 4600 feet (1385-1400 m).

INTERPRETATIONS OF LAKE HISTORY

Gilbert's (1890) interpretations, and the interpretations of lake history of subsequent authors, are summarized in figure 5. In each case the interpretations are based on observations at a number of localities in the Bonneville basin. However, the Old River Bed exposures have been significant in formulating all three summaries.

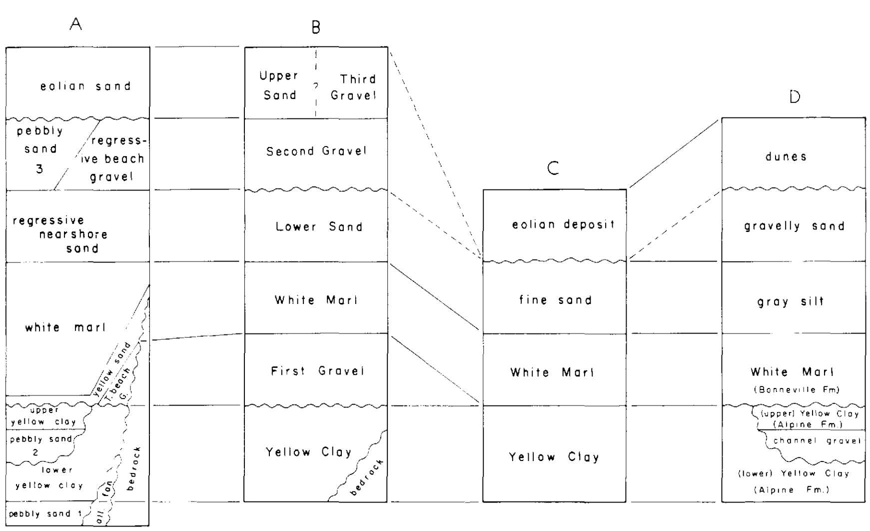


__Figure 2.__ Diagram showing different stratigraphic interpretations of the Bonneville beds at the Old River Bed. A, composite stratigraphic column for the Old River Bed area after Oviatt (1984, 1987, this guidebook); B, Gilbert's (1890) interpretation of the Upper River Bed section; C, Gilbert's (1890) interpretation of the Lower River Bed section; D, Varnes and Van Horn's (1961) interpretation of the Lower River Bed section. Wavy lines indicate unconformities.

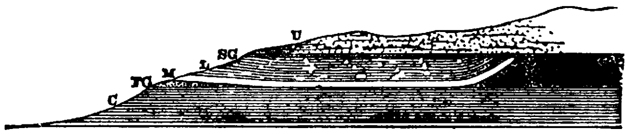


Figure 3. Cross section through the Bonneville beds at the Upper River Bed section (Gilbert, 1890, figure 29). this place is named The Shutoff on the U.S.G.S. Coyote Springs 7.5 minute quadrangle. We interpret this section as follows (Oviatt, 1984, 1987): U, eolian sand; SG, regressive-phase beach gravel (RBG); L, regressive-phase near-shore sand (RNS); M, white marl; FG (dark wedge on right), transgressive-phase beach gravel (TBG) ; C, lower yellow clay (LYC). Samples for radiocarbon dates L-774Q and L-672J (table 1) were collected from L, and from the base of M, respectively.

The significance of the yellow clay has been a matter of some controversy in recent years. Gilbert (1890), Ives (1951), and Varnes and Van Horn (1961) interpreted the yellow clay as representing a long period of deposition in a deep lake (up to about 90 feet (27.4 m) below the Bonneville shoreline) during a major lake cycle(s) much earlier than the Bonneville ("white marl") lake cycle. Stratigraphic, geomorphic, and geochronometric data, however, support the alternative interpretation that the yellow clay was deposited in shallow water during a relatively short period early in the transgressive phase of Lake Bonneville. Substantiating data are included in reports by McCoy (1981, 1987) and Oviatt (1984, 1987). The following observations support our interpretations. 1. The lithology of the yellow clay consists of silt, silty clay, and sand, whereas the white marl is fine-grained calcium carbonate. 2. The lower contact of the yelllow clay is gradational with locally derived alluvium, and beds within the yellow clay abut the underlying steep slopes of bedrock or colluvium (in contrast, the white marl is draped over underlying topographic irregularities). 3. The yellow clay and the white marl are conformable and gradational at low altitudes. 4. The fossil record in the yellow clay includes mollusks typical of shallow-water environments, ostracodes typical of marsh-pond and marginal lacustrine environments (R.M. Forester, 1983, personal communication), and impressions of rooted aquatic plants (B.J. Albee, 1983, personal communication). 5. The radiocarbon ages and the extent of amino-acid racemization in mollusks from the yellow clay (table 1). The amino-acid ratios are significant in that they show that the yellow clay is not much older than the overlying lacustrine deposits (i.e., the white marl). Thus, the yellow clay was deposited during the last lake cycle, and not during the penultimate deep-lake cycle, which is represented by lacustrine deposits that would underlie the Promontory Soil in Little Valley (in the Promontory Range of northern Utah).

Collectively these data suggest that the yellow clay was deposited early in the transgressive phase of Lake Bonneville (figure 5A) when the Sevier Desert basin was flooded and overflowing. The resulting river transported a large volume of fine-grained sediment northward to the southern margin of the Great Salt Lake Desert, and deposited it at the shore of the lake as a fine-grained delta or, more appropriately, an underflow fan (Oviatt, 1984, 1987).

Overlying the yellow clay is a variety of near-shore deposits, which range from beach or spit gravel (such as the "First Gravel" (TBG, figure 5C) at the Upper River Bed section), to well-sorted beach sand. In many places moderately sorted yellow sand (YS) containing gastropods overlies the yellow clay, and it forms the basal Bonneville facies at altitudes above the upper altitudinal limit of the yellow clay (about 4540 feet, 1385 m). The white marl is draped over all these near-shore facies.

We interpret the white marl as the deep-water facies of Lake Bonneville, and as such it represents about 8000 years of deposition in the Old River Bed area (figure 5C). Calcium carbonate (white marl) was deposited at a rate of approximately 10-12 cm per 1000 years. The Pavant Butte basaltic ash (16,000-15,300 yr B.P.; Oviatt and Nash, 1988) is exposed in the white marl at the Old River Bed and at many localities in the Sevier Desert to the south (Oviatt, 1984; Oviatt and Nash, 1988).

Table 1.
Selected Geochronometric Data on Samples From the Old River Bed Sections

Lab No.	Method[1]	Results[2]	Material Analyzed	Stratigraphic Unit[3]	Reference[4]
L-774Q	C-14	11,900 ± 300	ostracodes	**RLS**	**1**
L-672J	C-14	19,800 ± 400	gastropod shells	**YS**	**1**
Beta-5038	C-14	23,190 ± 1360	*Sphaerium* shells	**LYC**	**2,3,4**
AAL1803	AA	0.14 ± 0.01	*Amnicola* shells	**RLS**	**5,6**
AAL1809	AA	0.10 ± 0.005	*Amnicola* shells	**RLS**	**5,6**
AAL1448	AA	0.12 ± 0.01	*Sphaerium* shells	**PS2**	**5,6**
AGL189	AA	0.076 ± 0.005	*Sphaerium* shells	**LYC**	**3,4,7**
AGL516	AA	0.129	*Pisidium* shells	**LYC**	**7**
AGL517	AA	0.12	*Valvata* shells	**LYC**	**7**
AGL518	AA	0.11	*Anodonta* shells	**LYC**	**7**
—	T	16,000-15,300	Pavant Butte - basaltic ash	**WM**	**3,8**

[1]C-14, radiocarbon dating; AA, amino acid epimerization; T, tephrochronology.

[2]Radiocarbon dates in yr B.P.; amino-acid ratios (alloisoleucine/isoleucine) in the total acid hydrolysate; age of Pavant Butte ash in radiocarbon yr B.P.

[3]After Oviatt (1984, 1987), and figure 2. RLS, regressive-phase littoral sand; YS, transgressive-phase littoral sand; LYC, lower yellow clay; PS2, pebbly sand between lower and upper yellow clay; WM, white marl.

[4]References: 1, Broecker and Kaufman, 1965; 2, Currey and others, 1983; 3, Oviatt, 1984; 4, Oviatt, 1987; 5, McCoy, 1981; 6, McCoy, 1987; 7, McCoy, unpublished data, 1987; 8, Oviatt and Nash, 1988.

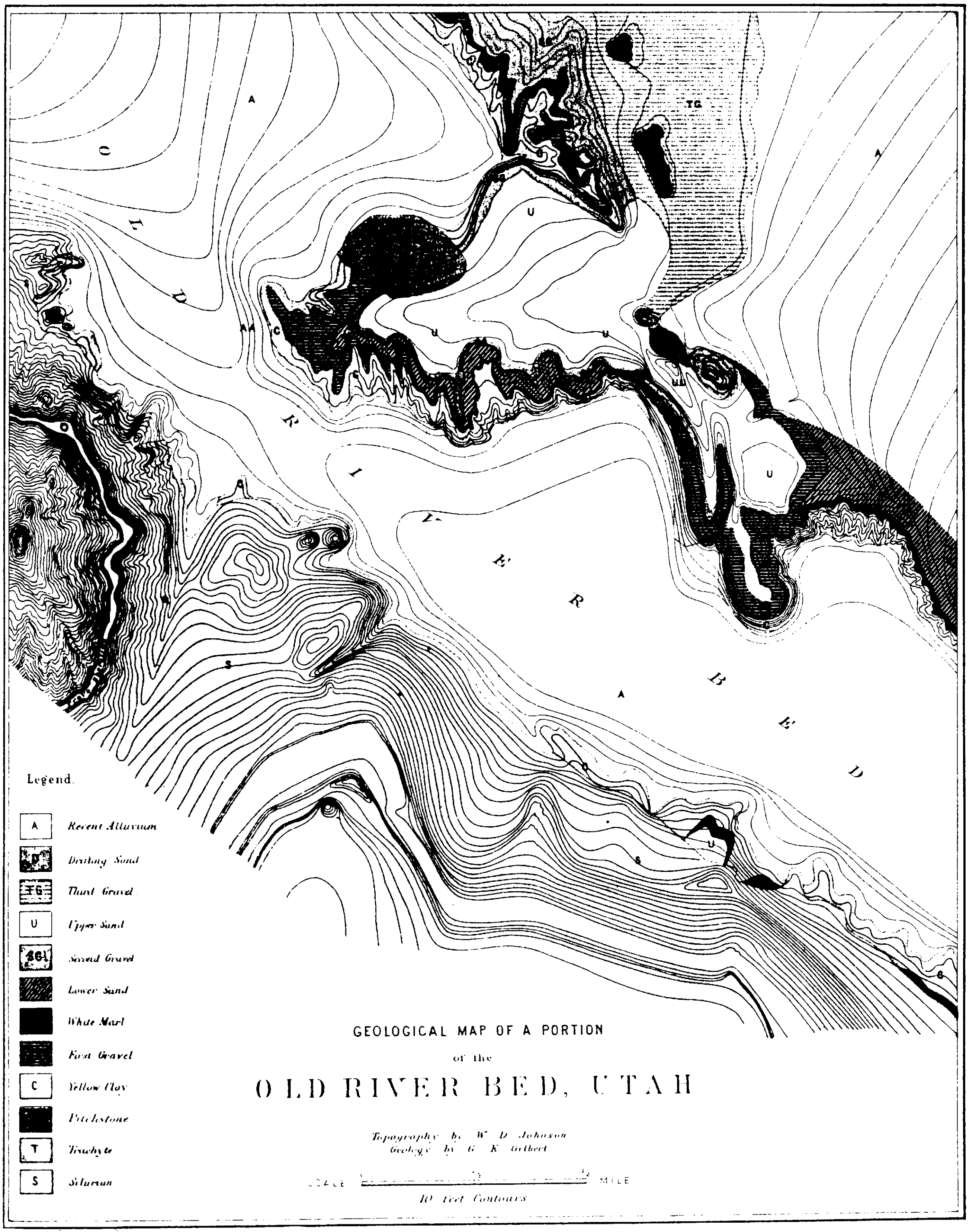


Figure 4. *Geologic map of the Upper River Bed section (Gilbert, 1890, p. 194, plate XXXII). See figures 2, 3, and 5 for further interpretation of map symbols. A sample of* Sphaerium *shells from the lower yellow clay, which has yielded a radiocarbon date of 23,190 ± 1360 yr B.P., and an alloisoleucine/isoleucine ratio of 0.076 ± 0.005 (AGL189; table 1), was collected on the west side of the Old River Bed southwest of the "L" in "Old".*

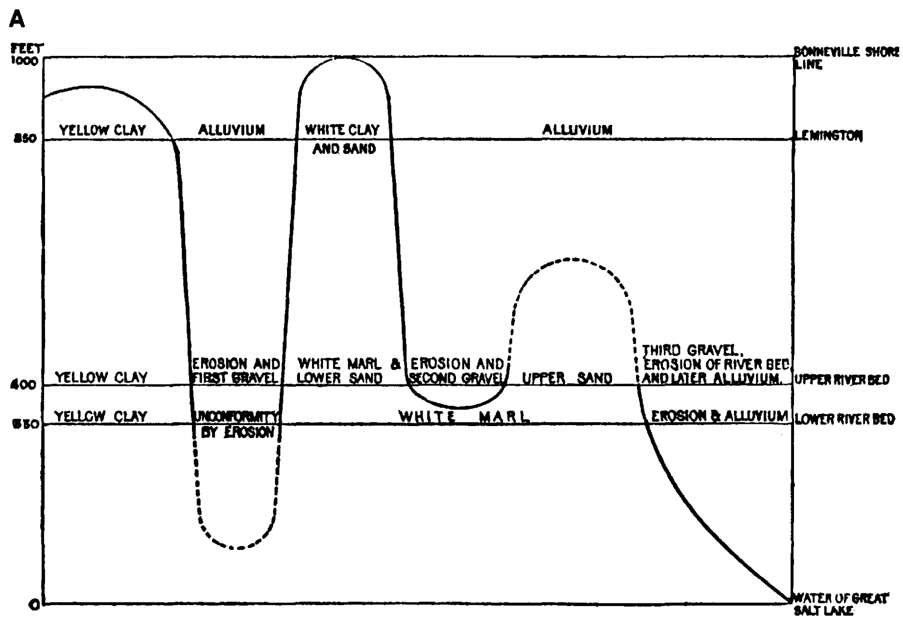
We interpret Gilbert's "Lower Sand" and "Second Gravel" as regressive-phase facies deposited as the lake dropped from the Provo shoreline (RNS and RBG, respectively). Waves again became effective in the Old River Bed area at lake levels below the Old River Bed threshold (4600 feet, 1400 m). The Old River Bed channel was carved as fluvial discharge continued northward from the Sevier Desert between about 12,000 and 10,000 yr B.P. During part of this overflow period the Gilbert shoreline formed in the Great Salt Lake basin (figure 5C). With the exception of the Gilbert shoreline at 4285 feet (1306 m), at the mouth of the Old River Bed, no lacustrine deposition took place in the area after 11,000 yr B.P. Later deposition of fans of Slow Elk Wash filled the channel of the Old River Bed at The Shutoff.

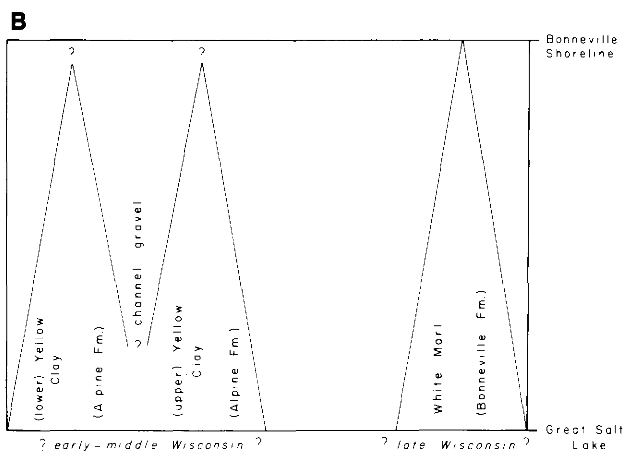
CONCLUSION

The deposits at the Old River Bed are well exposed over a larger area and through a greater altitudinal interval than at most other places in the Bonneville basin. Therefore, the area has great potential for future descriptive and interpretive work on Lake Bonneville stratigraphy and geomorphology. G.K. Gilbert was the first to work at the Old River Bed and he posed the major hypotheses about Lake Bonneville stratigraphy, geomorphology, and history that are still under discussion today. Thus, Gilbert's work is at the core of any research on Lake Bonneville and of many other aspects of Quaternary studies in the Bonneville basin. This, however, is not simply because he was the first. Through clear thinking Gilbert was able to identify the major questions in the Lake Bonneville scientific problem, and in uncomplicated terms he proposed the most reasonable answers to those questions in light of his available evidence. A student could find no better way to learn the methods and content of earth science than by studying Gilbert's work.

ACKNOWLEDGMENTS

We are grateful to Julie Brigham-Grette and Michael Machette for constructive comments on the manuscript.





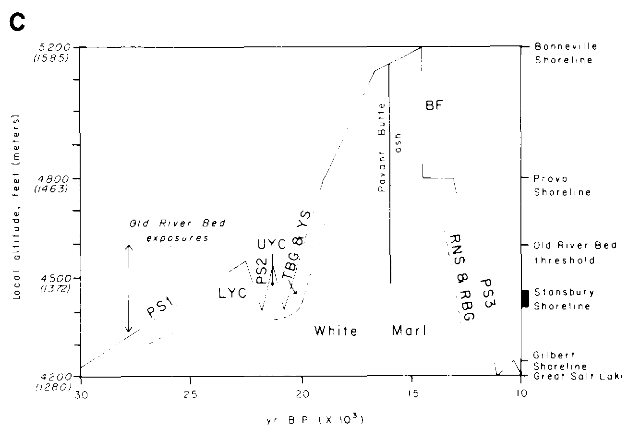


Figure 5. *Alternative interpretations of Lake Bonneville history based partly on observations at the Old River Bed by A, Gilbert (1890, figure 30); B, Varnes and Van Horn (1961); and C, Oviatt (1984, 1987, this guidebook). In B, ages and altitudinal limits of lake fluctuations are inferred from Varnes and Van Horn (1961) and Hunt and others (1953). In C the following symbols are used: PS1, lower pebbly sand (fluvial); LYC, lower yellow clay; PS2, middle pebbly sand (fluvial); UYC, upper yellow; TBG, transgressive-phase beach gravel; YS, nearshore sand; RBG, regressive-phase beach gravel; PS3, upper pebbly sand (fluvial). Altitudes shown are those for the Old River Bed area; no corrections have been made for isostatic rebound.*

THE STOCKTON BAR

by

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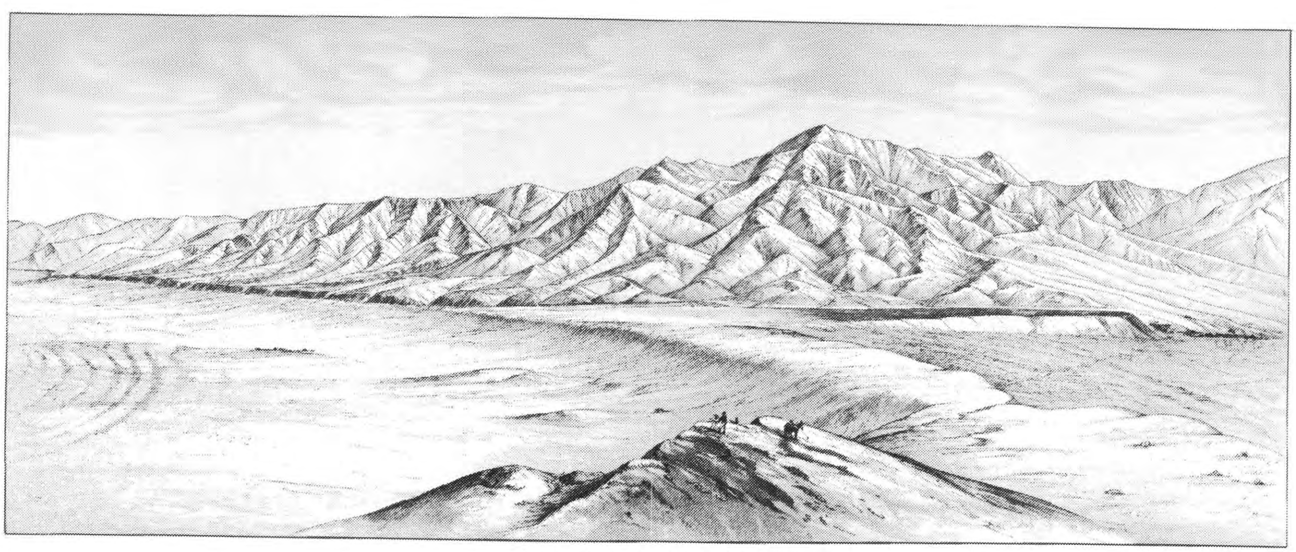
INTRODUCTION

Shore features that formed during the last deep-lake cycle are well developed and well preserved at many localities in the Bonneville basin. However, the pass between Rush and Tooele Valleys--formerly a Lake Bonneville strait--near Stockton, Utah, is exceptional, if not world class, in the details of paleo-lake history that are recorded in its suite of littoral deposits. G.K. Gilbert first crossed the pass on July 14, 1877, on which occasion he made a preliminary sketch of the "Bonneville Bar" (Hunt, 1982a, p. 38). Gilbert returned to the area three years later, at which time he made a detailed drawing (figure 1) of "The Great Bar at Stockton, Utah" (retouched version in Gilbert, 1890, plate IX; original field version in Hunt, 1982a, p. 170). Gilbert's topographers also prepared a detailed map of the area (figure 2; Gilbert, 1890, plate XX). Gilbert's field notes (Hunt, 1982a, p. 169-174) record 252 leveling stations, include several sketches, and clearly indicate that by the end of a week's field work he had a firm understanding of the coastal geomorphology of the Stockton area. Work in recent years has added greatly to the detail, but has invalidated very little of the substance of Gilbert's original interpretations.

To place the littoral deposits of the Stockton Bar and nearby area in context, it is helpful to review briefly the hydrographic history of the Bonneville lake cycle (figure 3). The Bonneville cycle spanned the interval from about 32 to 13 ka, which is essentially coincident with the last global ice age, i.e., marine isotope stage 2. In broad outline, the cycle comprised three



THE GREAT BAR AT STOCKTON UTAH

Figure 1. View southeastward from South Mountain, showing The Great Bar at Stockton, Utah (Gilbert, 1890, plate IX). Drawn by Gilbert and retouched by Holmes (Hunt, 1982a, p. 170).

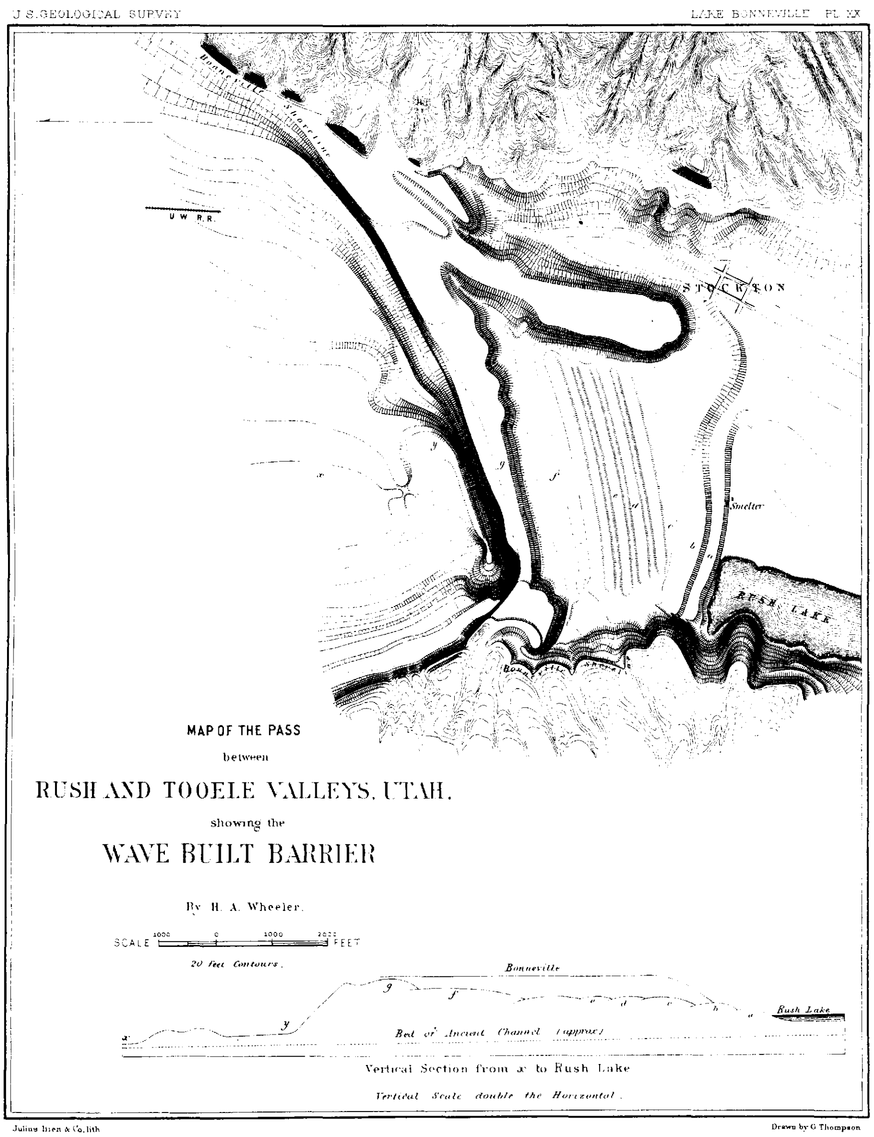


Figure 2. *Contour map and vertical section of the Stockton Bar (Gilbert, 1890, plate XX).*

major phases: a protracted phase of closed-basin, predominantly transgressive stages until about 15.5 ka; a phase of intermittently open-basin, threshold-controlled stages beween about 15.5 and 14.2 ka (see Currey and Burr, this volume); and a brief phase of closed-basin, rapidly regressive stages after about 14.2 ka. The portion of the Bonneville lake cycle that is represented by littoral deposits at and near the Stockton Bar is indicated by the heavy line in figure 3. The purpose of the Stockton Bar segment of this trip is to examine and discuss morphostratigraphic evidence pertaining to the unnamed shoreline complex (USC), late transgressive stages (LTS), Bonneville shoreline complex (BSC), Bonneville Flood (BF), and Provo shoreline complex (PSC), as well as post-flood shorelines that are unique to Rush Valley. The physical nature and significance of this evidence is summarized below.

UNNAMED SHORELINE COMPLEX (USC)

Approximately 20,000 yr B.P. the transgressing body of water began to attack pre-Lake Bonneville alluvial fans at 4800 feet (1463 m) a.s.l. Longshore currents primarily from the northeast started to build a series of retrograding coastal landforms at the southern end of Tooele Valley. The USC is a large cross-valley barrier (figure 4, [U]) and, as at many other localities in the Bonneville basin, is partly concealed by overlying PSC deposits. The USC marks what seems to be one or more important stillstands or moderate oscillations during the transgressive phase of the Bonneville lake cycle.

LATE TRANSGRESSIVE STAGES (LTS)

The transgressing lake continued to develop spits and barriers in Tooele Valley, including a bayhead barrier constructed at an altitude about 4900 feet (1493 m)(figure 4, [L_2]). Chemical and mine tailings reclamation activities at Bauer have used the lagoon behind the barrier at this altitude as a settling pond.

As Lake Bonneville transgressed into Rush Valley (figure 4, [L_5]) the retrograding and aggrading pattern of spit-barrier development shifted to a prograding and aggrading pattern. This may have been due to an increasing rate of local hydro-isostatic subsidence. Surface drainage from Rush Valley was blocked as the spit-barrier development continued to prograde and aggrade (Gilbert, 1890, Plate XX; Gilluly, 1929). The spit-barrier construction continued to about 5160 feet (1572 m)(figure 5, B_0), at which altitude in the Stockton Bar area the lake became a threshold-controlled water body, under the influence of the Zenda threshold far to the north; this marks the beginning of the BSC.

Figure 3. *Schematic hydrograph of the Bonneville basin, showing generalized paleolake stages during the Bonneville lake cycle, A, and in early post-Bonneville time, B. Segments of the hydrograph are: PBL, pre-Bonneville low stages; ETS, early transgressive stages; SSC, Stansbury shoreline complex; MTS, middle transgressive stages; USC, unnamed shoreline complex; LTS, late transgressive stages; BSC, Bonneville shoreline complex; BF, Bonneville Flood; PSC, Provo shoreline complex; LRS, late regressive stages; PGS, pre-Gilbert low stages; GSC, Gilbert shoreline complex; and HS, Holocene stages. Stages USC through PSC (heavy line) are exceptionally well represented by littoral morphostratigraphic units at or near the Stockton Bar.*

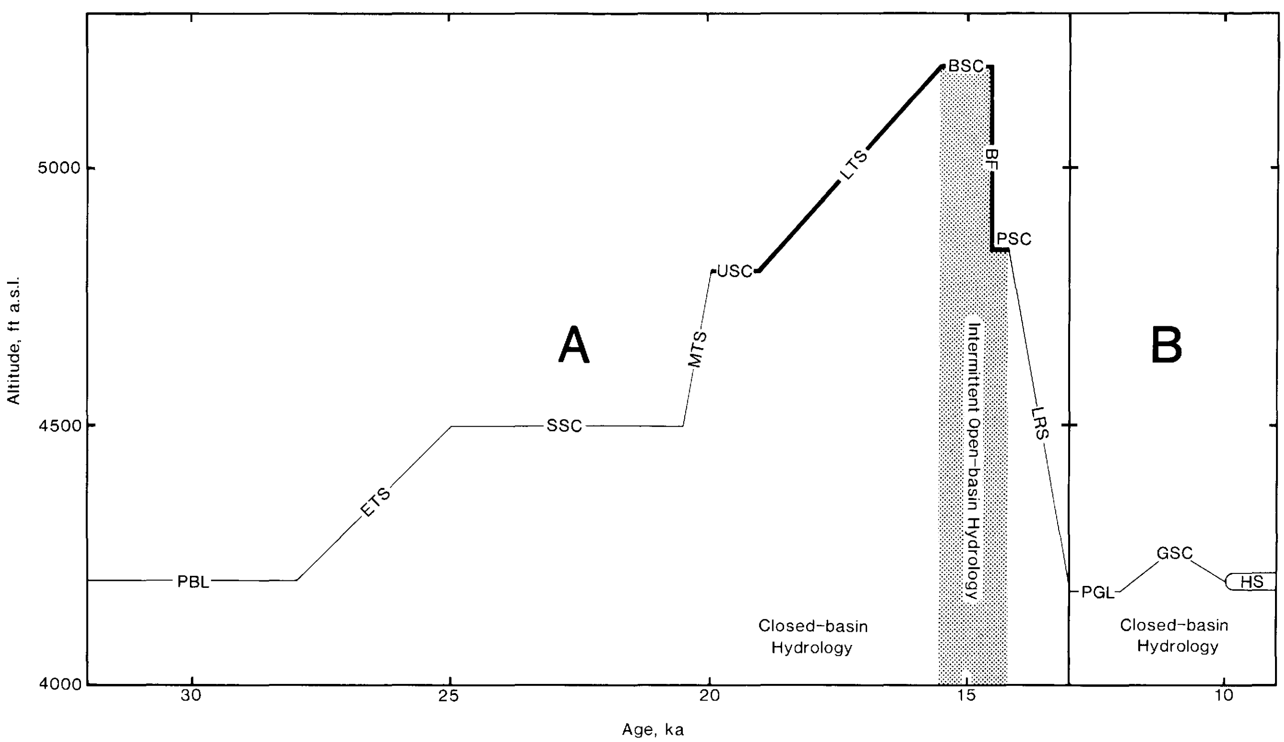
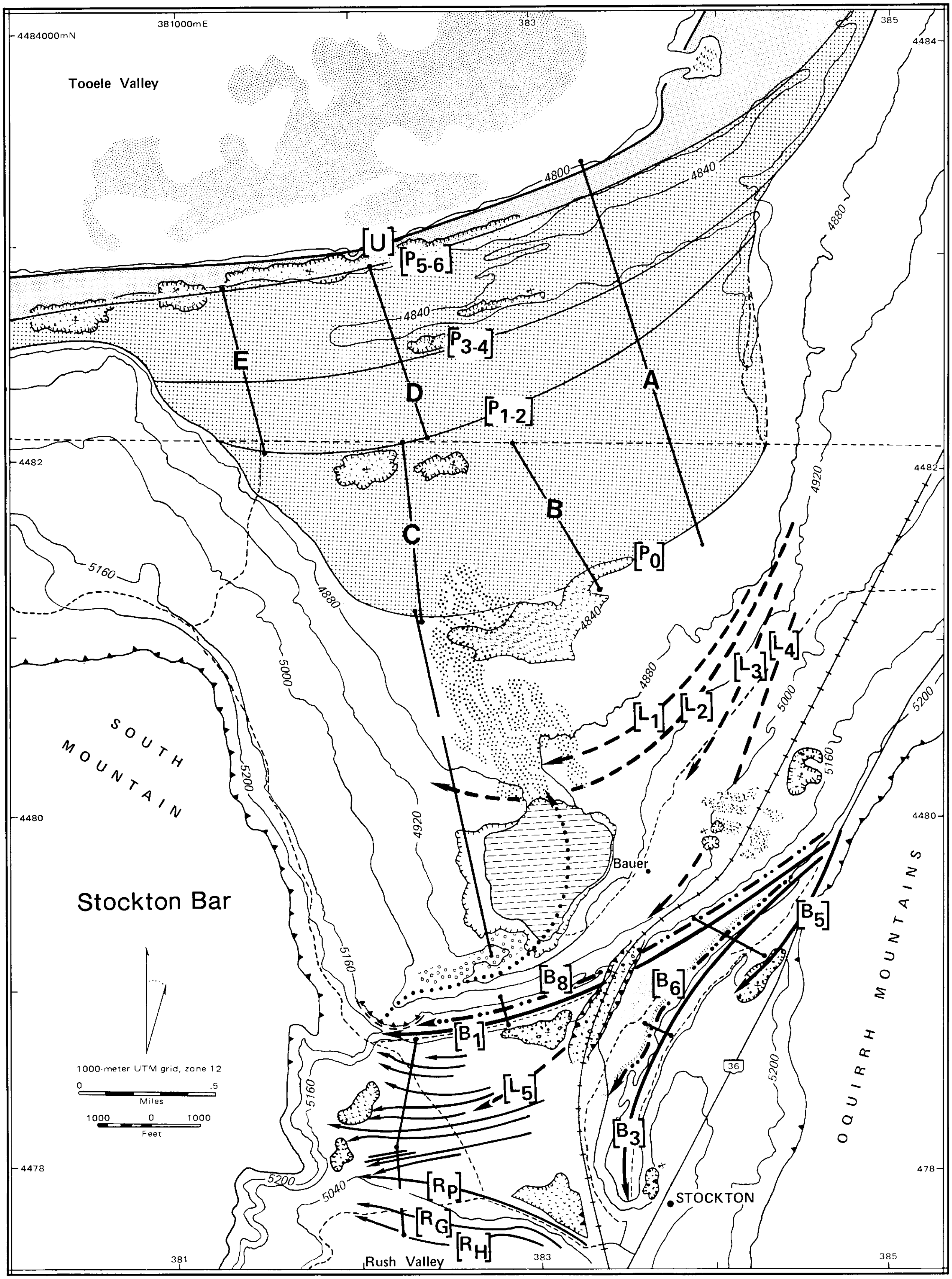


Figure 4. Map of the Stockton Bar area, showing the morphostratigraphic units of the unnamed shoreline complex [U], late transgressive stages [L_1-L_5], Bonneville shoreline complex [B_0-B_8], and Provo shoreline complex [P_0-P_6], and Rush Valley post-flood shorelines (R_P, R_G, R_H).



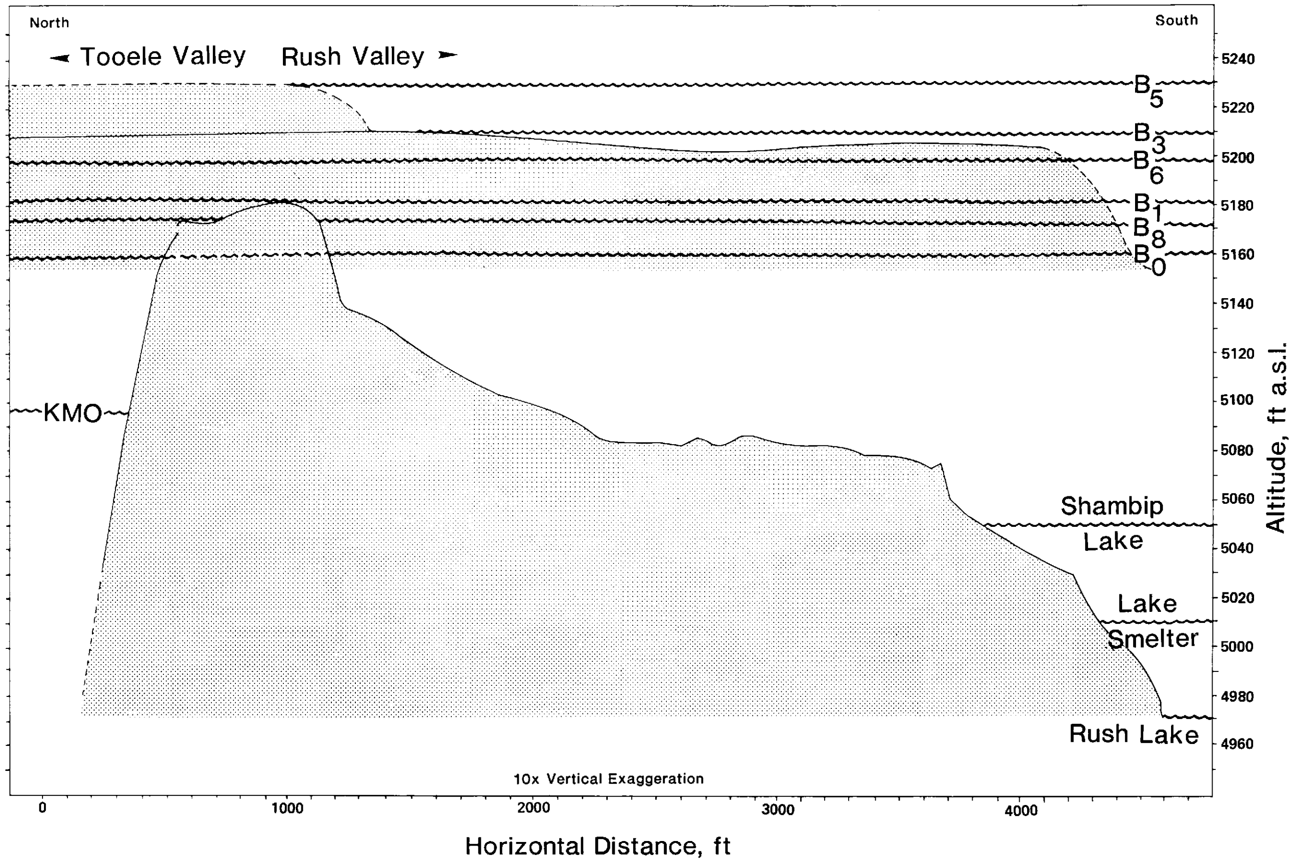


Figure 5. North-south composite profile of the Stockton Bar, showing hypsometric relations among several of the morphostratigraphic units that are mapped in figure 4.

BONNEVILLE SHORELINE COMPLEX (BSC)

At many localities morphostratigraphic evidence of the culminating, open-basin phase of the Bonneville lake cycle suggests that four stages of non-catastrophic discharge at the Zenda threshold (5090 feet, 1552 m) were interrupted by three sub-threshold lake stages, all of which were complicated by an isostatically subsiding basin (Currey and Burr, this volume). In the Stockton Bar area, as well, morphostratigraphic features B_1, B_3, B_5, and B_8 show evidence of four periods of threshold-controlled deposition that occurred during non-catastrophic discharge at Zenda.

The first threshold-controlled morphostratigraphic unit ([B_1] in figure 4 and B_1 in figure 5), a cross-valley baymouth barrier at about 5180 feet (1579 m), was built approximately 15,350 yr B.P. by vertical accretion that occurred at essentially the same rate that local hydro-isostatic subsidence was occurring relative to the threshold-controlled water plane. A long, lakeward-dipping abrasion platform extends from Tooele southwestward to the proximal end of BSC depositional features near Stockton (figure 6). It marks the avenue along which material travelled during longshore transport. As in the late transgressive phase, unconsolidated material from pre-Lake Bonneville alluvial fans, mainly readily rounded clasts of Pennsylvanian quartzite, provided the sediments that comprise the BSC barrier and spits. Baymouth barrier B_1 formed by prolongation and aggradation of a spit that eventually extended across the Tooele-Rush strait to South Mountain.

The area continued to subside during a sub-threshold interval. After threshold control resumed, a massive spit ([B_3] in figure 4 and B_3 in figure 5) aggraded and extended southward into Rush Valley about 1.5 miles (2.4 km), attaining a maximum altitude of 5212 feet (1588 m) approximately 15,150 yr B.P. After a second sub-threshold interval, the massive B_3 spit was followed by the construction of a smaller spit ([B_5] in figure 4 and B_5 in figure 5) that aggraded at the rate of local subsidence and attained an altitude of 5231 feet (1594 m) approximately 15,000 yr B.P. The B_5 spit marks the maximum elevation of the BSC near the Stockton Bar, and correlative features mark the Bonneville maximum elsewhere throughout the basin. Direct evidence of the two sub-threshold lake stages that occurred between B_1, B_3, and B_5 has not yet been found in the Stockton Bar area, although those stages can be identified elsewhere in the Bonneville basin (e.g., lower beach ridges of Gilbert, 1890, plate XI).

After the development of the highest, B_5 component of the BSC, the Zenda threshold underwent about 40 feet (12 m) of non-catastrophic incision approximately 15,000 yr B.P. (Currey and Burr, this volume). The conclusion of this early Zenda incision is marked in the Stockton Bar area by a boulder beach ([B_6] in figure 4, and B_6 in figures 5 and 7) about 5191 feet (1582 m), where the regressing water plane lingered only long enough to winnow the pebbly sand matrix from between cobbles and boulders that had been deposited approximately 200 years earlier in the construction of the massive B_3 spit.

Figure 6. View southwestward from the south edge of Tooele, showing the Stockton Bar in middle distance and the B_5 shoreline angle and shore platform at the toe of the Oquirrh Mountains (to the left of what is now Utah Highway 36). The Stockton Bar was the sediment sink for longshore transport from sediment sources that included the now scree-mantled quartzite cliffs in the left foreground and the arcuate bluff in highly erodible pre-Bonneville fan gravels in the left middle ground. Photo by Barnum Brown, 1934.

The short-lived development of the B_6 shoreline was followed by a major sub-threshold cycle between approximately 15,000 and 14,550 yr B.P. This Keg Mountain oscillation (KMO) of Currey and others (1983; see Currey and Burr, this volume; see also day 3-stop 3) clearly reflects significant hydroclimatic forcing. During the regressive phase of the KMO, the lake receded about 100 feet (30 m) (figure 5, KMO) below the previous B_6 stage. The basin underwent partial hydro-isostatic unloading and reloading during the KMO with the net result being measureable isostatic rebound. At the conclusion of the KMO, after having deposited a tufa drapery on the north-facing slope of the Stockton Bar, the lake returned to threshold control at an elevation of about 5170 feet (1575 m). Between approximately 14,550 and 14,500 yr B.P. the lake transgressed at the local rate of hydro-isostatic subsidence to an elevation of 5177 feet (1578 m), where the final tufa-encrusted BSC beach ridge ([B_8] in figure 4 and B_8 in figure 5) was deposited.

Further development of the Bonneville shoreline complex was prevented by catastrophic incision of the Zenda threshold and the rapid drawdown of the lake that occurred as a consequence. The early Zenda incision and the short duration of Zenda threshold control after the KMO prevented the basin from subsiding to its previous maximum, and thereby reduced somewhat the volume of water that otherwise would have been released.

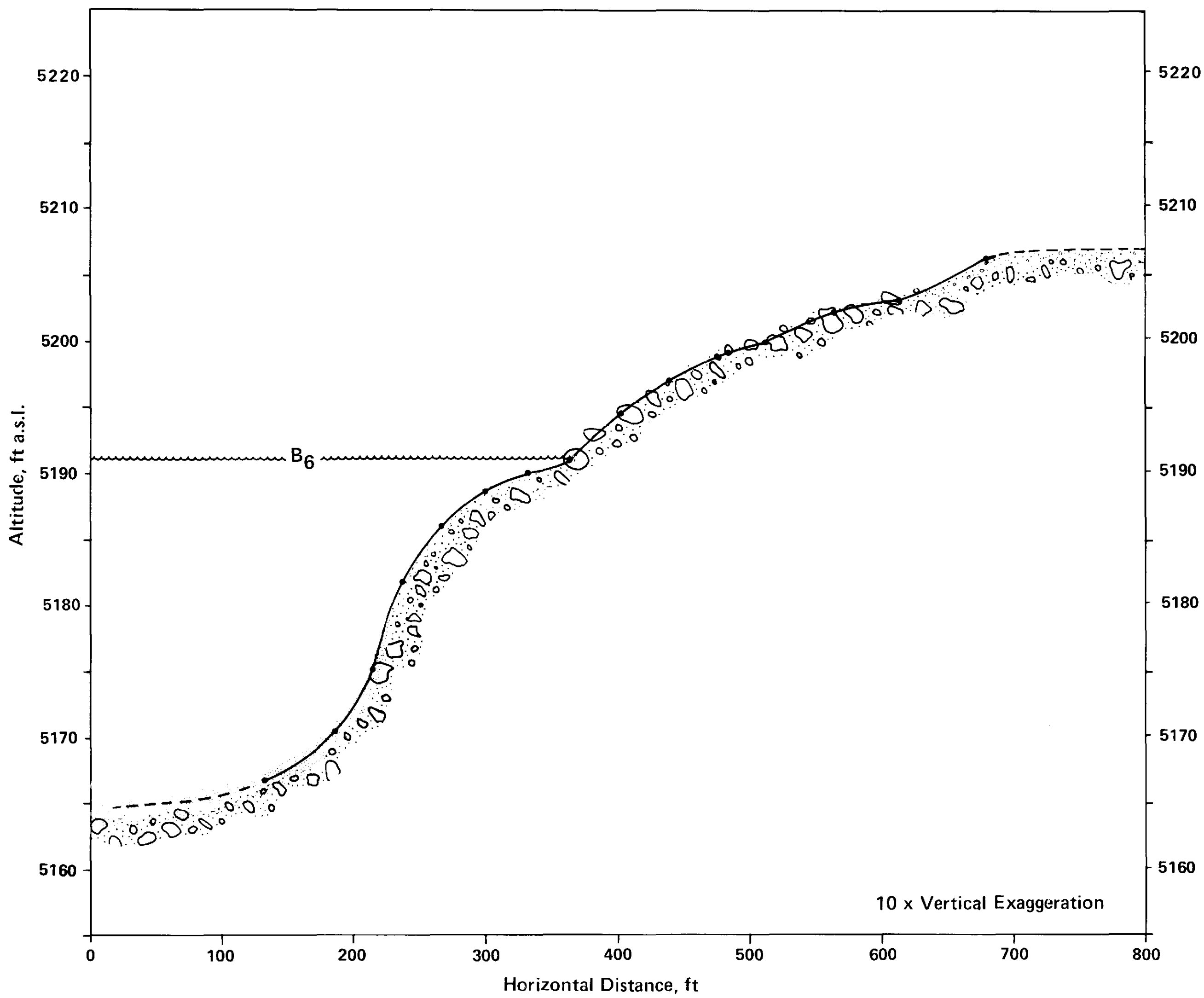
BONNEVILLE FLOOD (BF)

The Bonneville Flood (Malde, 1968), which occurred about 14,500 yr B.P., rapidly released the upper 340 feet (104 m) of Lake Bonneville into the Snake River system. The floodwaters headwardly eroded the Snake River-Bonneville basin (Pacific Ocean-Great Basin) drainage divide southeastward about 2 miles (3.2 km)(Currey, Oviatt and Plyler, 1983) and lowered it 340 feet (104 m). The current consensus is that the duration of the flood exceeded 8 weeks (e.g., Jarrett and Malde, 1987), but probably did not exceed one year. As measured at and near the Stockton Bar, the stage change that occurred during the flood is the vertical difference between the B_8 shoreline, which was the highest that formed during post-KMO time, and the base of the small bluff ([P_0] in figure 4 and P_0 in figure 8) that began to form on day one of control by the newly excavated threshold at Red Rock Pass. Total threshold lowering, including the early Zenda incision, was about 380 feet (116 m).

PROVO SHORELINE COMPLEX (PSC)

The Provo shoreline north of the Stockton Bar is, in essence, an enormous ramp of prograding and aggrading beach ridges; continuity of the ramp is interrupted by small downward steps (P_2, P_4, and P_6 in figure 8). This morphostratigraphic signature, which is seen basin wide, resulted from persistent landsliding in the flood-scoured threshold area, and from intermittent incision of the post-flood outlet channel across the 5-mile-long (8 km) toe of the landslide (Currey and Burr, this volume). After the Provo water plane stabilized initially at P_0, landsliding gradually raised the Red Rock Pass threshold and caused the first beach ramp to prograde and aggrade from P_0 to P_1. A threshold incision of about 5 feet (1.5 m) shows up in the figure 8 profiles at P_2. Subsequent landsliding resulted in continuing beach ramp construction to P_3, after which 15 feet (4.5 m) of threshold downcutting is indicated in the figure 8 profiles at P_4. Threshold incision was again followed by landsliding and beach ramping to P_5. A final threshold incision of 10 feet (3 m) appears in the figure 8 profiles at P_6. Threshold-controlled littoral deposition continued briefly, until approximately 14,200 yr B.P., when the lake finally regressed below Red Rock Pass. Shore-face accretion on the well-developed beaches of the PSC was then replaced by incipient development of littoral features during the transient regressive stages that characterized LRS time.

Figure 7. *Profile of the B_6 boulder beach, which formed after the Zenda threshold was lowered 40 feet during the non-catastrophic early Zenda incision. The water plane lingered only long enough to winnow the pebbly sand matrix from between cobbles and boulders that had been deposited approximately 200 years earlier in the construction of the massive B_3 spit.*



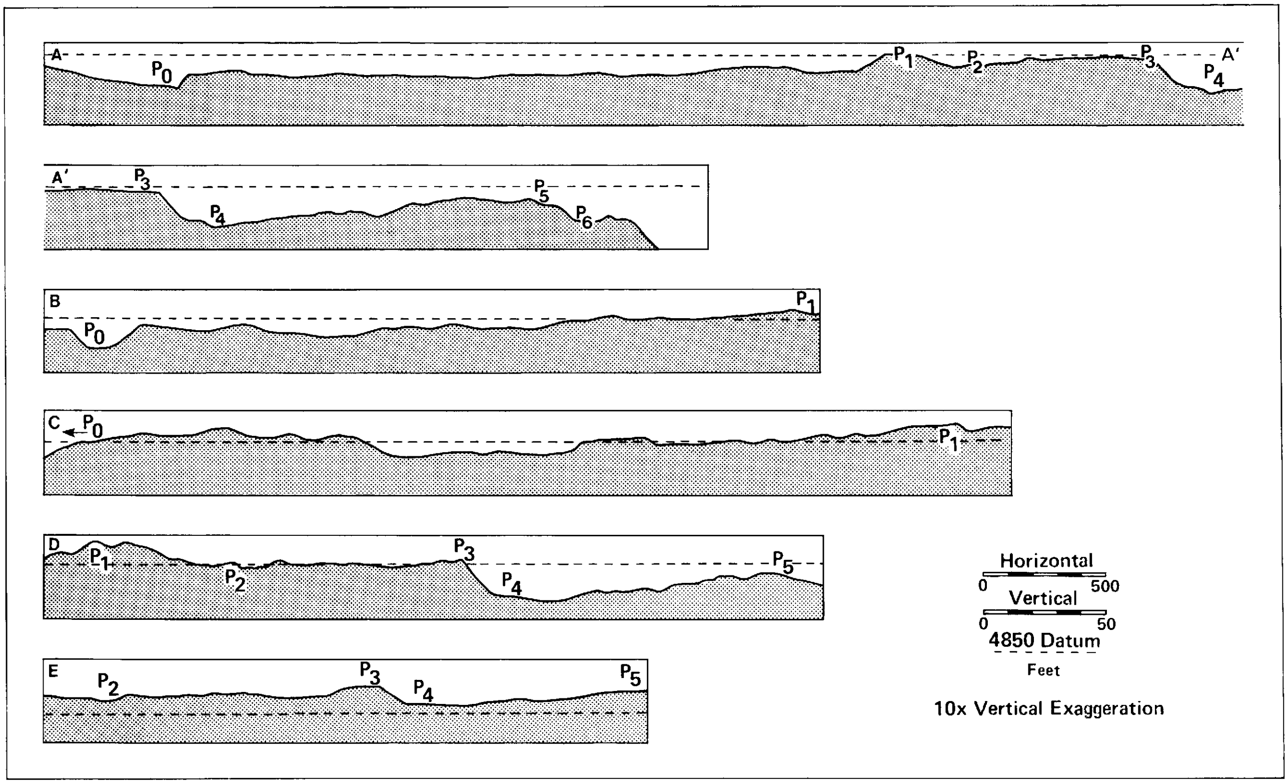


Figure 8. *Profiles of the Provo shoreline complex north of the Stockton Bar, depicting the shore-face accretion that prograded and aggraded ramp-like to the north (right), and which was interrupted a few times by small downward steps in the locus of littoral deposition. See figure 4 for locations of profiles.*

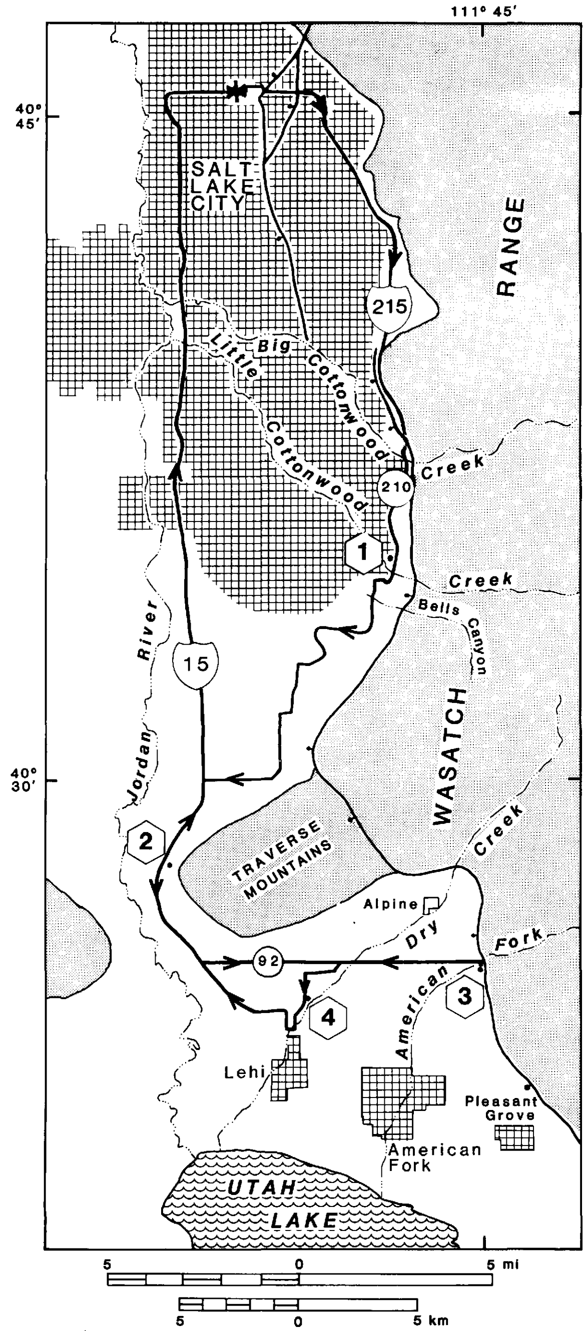
RUSH VALLEY SHORELINES

The post-flood shorelines on the Rush Valley side of the Stockton Bar (figures 4 and 5) comprise the Provo shoreline equivalent, R_P, which enclosed Lake Shambip at about 5050 feet (1539 m) and the Gilbert shoreline equivalent, R_G, which enclosed Lake Smelter, about 5010 feet (1527 m). The 4970-foot (1515 m) level of Rush Lake during the historic high, R_H, of the 1980s seems to have been essentially coincident with the level of the Holocene high.

Shorelines that formed on the Rush Valley side of the Stockton Bar during and subsequent to PSC time were not addressed in detail by Gilbert, but he did note historic fluctuations of Rush Lake (Gilbert, 1890, p. 228-229). He also recognized that Rush Valley retained a separate body of water (Gilbert, 1890, p. 149) when Lake Bonneville receded below the Stockton Bar barrier between Rush and Tooele Valleys. However, he mistakenly concluded that an isolated water body in Rush Valley would not be large enough to leave a geomorphic imprint on surrounding topography. On the contrary, the morphostratigraphic record in Rush Valley at the elevations mentioned above clearly indicates that post-flood coastal processes did, indeed, rework littoral sediments that earlier had been deposited as Lake Bonneville transgressed into Rush Valley.

ACKNOWLEDGMENTS

Special thanks go to Robert M. St. Louis and the Tooele Army Depot for allowing entrance onto private and restricted property. The research that led to this paper was supported in part by NSF Grant EAR-8721114 and USGS Contracts 14-08-0001-18289 and 14-08-0001-G1536.



111° 45'
40° 45'
40° 30'
SALT LAKE CITY
RANGE
215
Big
Little
Cottonwood
Cottonwood
Creek
210
1
Creek
Bells
Canyon
Jordan
River
15
WASATCH
2
TRAVERSE MOUNTAINS
Creek
Alpine
Dry
Fork
92
American
3
4
Lehi
Pleasant Grove
American Fork
UTAH
LAKE
5 0 5 mi
5 0 5 km

DAY 3
Road Log from Salt Lake City to the Northern Part of Utah Valley and Return

by
Michael N. Machette and Donald R. Currey

Field trip leaders: William Scott, William Lund, David Schwartz, and Michael Machette

The final day of our field trip takes us south along the Wasatch Front and into the Utah Valley (figure 1). Our first stop is near the mouth of Little Cottonwood Canyon, a classic site for observing the relations between glacial, fluvial, and lacustrine deposits of the Bonneville and older lake cycles (stop 1A). In addition, we will discuss some of the tectonic features of the Wasatch fault zone (stop 1B), which are spectacularly displayed in this area. From here, we continue south away from the mountain front and traverse deltas and spits graded to the Bonneville and Provo levels. Stop 2 is at a large gravel pit at the Point of the Mountain, at the northwest end of the Traverse Range. Here we will see thick near-shore deposits of the Bonneville lake cycle, and evidence for the older Little Valley lake cycle.

Proceeding south around Point of the Mountain and into the northern part of the Utah Valley, we will visit a trench site (stop 3) on the Wasatch fault zone at American Fork Canyon. In addition, nearby gravel pits reveal a two-fold transgressive sequence which is interpreted as evidence for the Keg Mountain oscillation. After lunch, we will head west, crossing the large fan-delta complexes graded to the Provo shorelines at American Fork and Lehi. Our final stop (4) for the day will be in a deep sand and gravel pit along the north bank of Dry Creek. This pit exposes sediments of the transgressive and regressive phase of the Bonneville lake cycle and appears to be one of the more complete sections now exposed in the Utah Valley. From here we will return to Salt Lake, via I-15.

Mileage	Description of features along route.
0.0	Intersection of 4th South and 7th East (Residence Inn). Proceed east on 4th South crossing the East Bench fault at 0.5 mi.
1.4	Cross a branch of the East Bench fault that goes through the University of Utah campus. Surface evidence of this fault scarp has been removed, but Gilbert's 1901 photograph (USGS archive no. 1793) shows the fault in an arroyo.
2.0	Intersection with Wasatch Boulevard. Continue south (right) on Foothill Boulevard (Utah Hwy 186). Research Park (University of Utah) is built on the Provo shoreline. According to Richard Van Horn, the Bonneville shoreline to the east has been bulged upward for a short distance at Georges Hollow.
4.3	Bonneville shoreline at base of hill to the east (note "H" on hillsope).
5.0	Proceed south on I-215 (Belt Route extension of Wasatch Boulevard).
5.5	24th South offramp. From here you get a good view to the south of the Parleys Creek paleodelta formed during the Provo stillstand and, northeast of the highway interchange, a transgressive boulder beach of the Bonneville lake cycle.
6.1	The Promontory Soil, developed on pre-Bonneville lake cycle deposits, is exposed on the northeast side of I-80 and I-215.
7.5	3900 (39th) South. Crossing a young lobe of the Neff's Canyon alluvial fan. Two faults to the southeast cut older parts of this fan complex between here and 4800 (48th) South.
8.4	4500 South offramp is at Provo shoreline level.
8.9	End of I-215 (temporary). Continue south on Wasatch Boulevard.
9.7	Steep cliffs on left (east) for the next mile are erosional shorelines.
10.2	Road bends to left (east) and deltas of Big Cottonwood Canyon come into view. The large gravel pit to the south is in the Provo delta, whereas the Bonneville delta is to the southeast.
11.1	Several splays of the Wasatch fault zone form small scarps west of the road.
11.7	Holladay Gun Club, site of good exposures of Bonneville and Little Valley Alloformations (Alpine Formation of some workers) and associated proglacial deposits. Cross main trace of Wasatch fault zone.
12.8	Mouth of Big Cottonwood Canyon and intersection with Utah Highway 152. Proceed south on Wasatch Boulevard. As you approach the crest of the hill (Bonneville delta), notice the multistory homes built on fill in the broad graben of the Wasatch fault zone.
15.1	Bear right on Wasatch Boulevard (3590 East); left fork leads to Little Cottonwood Canyon Road. As we enter the deep, well-developed graben of the Wasatch fault zone notice the Christmas tree farm on left--site of Woodward Clyde and Associates 1978 exploratory trench.
15.8	As the road bends to left and descends to base of graben, pull off to right on small dirt road. Stop 1 will be near large graffiti-covered boulder on west side of road.

DAY 3-STOP 1 discussion by Scott (1A) and Schwartz and Lund (1B).

Mileage	Description of features along route.
0.0	Leave stop 1. Proceed south along graben. Note housing development within graben of Wasatch fault zone.
0.6	Turn right (west) onto Little Cottonwood Road.
0.7	Turn left (south) on first paved road which is marked 9710 South and becomes Ruskin Circle (3315 East). Vehicles should stay close for the next mile to avoid getting separated.
0.9	Road turns sharply to right (west) and becomes 9800 South. At 1.0 mile, road drops off end moraine and onto outwash that partly buries the moraine and forms upper part of large Bonneville-level fan delta of Little Cottonwood and Dry Creeks.

1.1 3100 East. Turn left (south). Low gravel bar of Bonneville shoreline is 500 feet (150 m) to west. Just past the LDS Church, vacant lot to east affords an excellent view of the left-lateral moraine of Little Cottonwood Canyon and the apparent tectonic offset of the moraine crest.

1.3 Turn right (west) onto 10000 South, then immediately left (south) on Dimple Dell Road. Road drops off delta and follows drainage channel of Lower Bells Canyon.

1.5 Steep slope with water tank to left (east) is terminal moraine of the Bells Canyon advance (late Pinedale).

1.9 Roadcuts expose till of both Bells Canyon and Dry Creek (Bull Lake) advances.

2.3 Road swings to right (west) and crosses a fluvial terrace graded to the Provo shoreline and delta.

3.1 Type localities of the Draper Formation and Dimple Dell Soil of Morrison (1965b) are in the valley of Dry Creek on the right. Morrison interprets the Draper as the youngest part of the Bonneville Group and the Dimple Dell Soil as being older than the Alpine Formation of the Bonneville Group (see Introduction to this volume). The Draper Formation consists of very poorly sorted and poorly bedded deposits of mostly pebbly sand that dip gently toward the axis of the stream valley. Scott and others (1983) interpret the Draper as valley-fill alluvium and colluvium that postdate the Bonneville lake cycle; locally a soil (Graniteville Soil of Morrison) is formed in the lake deposits and is buried by Draper sediment. An age of 3520 ± 80 yr B.P. (W-4566) from charcoal indicates that the Draper is late Holocene (in part). The type Dimple Dell Soil is similar in degree of development to the type Promontory Soil, which Morrison (1965b) believes is older than the Alpine Formation in Little Valley, Utah. Scott and others (1983) regard both soils as having formed during the same time inveral--the interval between the Little Valley and Bonneville lake cycles.

3.7 Turn left (south) onto 2000 East and cross upper edge of extensive Provo delta. Note eroded edge of Bonneville delta to east and the south-trending Draper (Provo) spit to southwest.

4.5 Road turns to west and becomes 11270 South. Proceed along north side of Willow Creek. Morrison described several exposures of the Draper Formation in this area.

5.3 Turn left and proceed south on 1700 east. Road descends from Provo delta and skirts the east edge of the Draper Spit (a southward extension of the Provo delta).

6.6 Turn right (west) onto Pioneer Avenue (12300 South). Note the foreset beds at the south end of the Draper Spit.

7.0 Turn left (south) onto 1300 East and cross railroad tracks. As we approach the Traverse Mountains, look to the southeast. Corner Canyon is the deeply incised canyon between the Traverse Mountains to the west and the Cottonwood Stock (Tertiary intrusives) of the Wasatch Range to the northeast. Gilbert (1928, p. 29) noted 37 feet (11.2 m) of displacement of the Bonneville shoreline at Corner Canyon, but only 15 feet (4.6 m) of throw across the alluvial fan that is graded to the Provo level (W.E. Scott, written commun., 1988, suspects that this fan may be of Holocene age). This relation led Gilbert to conclude that the remaining 25 feet (6.6 m) of throw "took place during the presence of the lake." Alternatively, one could interpret the 25 feet of throw as related to catastrophic faulting during the regression of the lake (see Machette and others, 1987) or, if the fan is Holocene, reflects a linear time-displacement relation.

8.4 Road makes two 45-degree right turns and becomes 13800 South. Proceed west toward I-15. Note Bingham open-pit copper mine on east face of Oquirrh Mountains west of the Jordan River. Hot springs on the south side of road are in bedrock.

10.5 Turn left (south) onto Minuteman Drive, the frontage road east of I-15.

11.9 Overpass and onramps to I-15. Note bedrock exposures.

12.2 To the south and west, the prominent bluff with a band of vegetation marks a wave-cut abrasion platform. The upper 1/3 of the bluff is lake sediment. Gilbert thought that this platform was a pediment (that is, created by fluvial rather than lacustral erosion).

13.2 Entrance to Geneva Rock Products quarry. Proceed to southwest (upper) edge of quarry and park near screening operation (mileage 13.6) for stop 2.

DAY 3-STOP 2 discussion by Scott.

Mileage **Description of features along route.**

0.0 Entrance to Geneva Rock Products. Turn left (west) onto frontage road and proceed along Point of the Mountain.

1.1 Salt Lake-Utah County line, near crest of Provo spit. Note foreset beds exposed in long road cut through constructional spit at Bonneville level. After passing Jordan Narrows (to west), road descends to slightly below Provo shoreline level.

3.4 Intersection with road to Alpine (Utah Hwy 92). Turn left (east).

7.3 Cross Dry Creek, a major stream that flows from the Wasatch Range past Alpine. As mapped by Hunt and others (1953), the exposures along Dry Creek in this area are mainly Alpine Formation, covered by gravel of the Provo Formation. Machette has mapped these same deposits as transgressional shallow to moderately deep-water facies of the Bonneville lake cycle, capped by fluvial gravels graded to and interfingered with the regressive deltas at the Provo level (see day 3-stop 4).

8.0 Small hill to south of road intersection (6000 West) is an outlier of moderately deep-water sediment (fine sand to silt) deposited during the transgressive phase of the Bonneville lake cycle, surrounded by coarse pebble to cobble gravels of the Provo fan-delta complex.

8.9 Intersection with Utah Hwy 74 (to American Fork and Alpine). Proceed east on fan-delta surface. Note the small (3-m-high) scarp that the road crosses. Although this scarp is not exposed in cross section, Machette suspects that it is probably the result of down-to-the-west faulting that occurred during deposition of the Provo level fan-delta complex about 14 ka. As you enter the mouth of American Fork Canyon, watch for Utah Hwy 146 which descends into canyon from south.

10.7 Turn right (south) onto Utah Hwy 146. Road cut near crest of hill exposes the main splay of the Wasatch fault zone (note clean gray fluvial gravel on east side and poorly sorted light-reddish brown gravelly colluvium on west side). At crest of hill, road is in wide graben of the fault zone.

11.1 Turn left onto dirt road and park at base of faulted alluvial fans for stop 3.

DAY 3-STOP 3 discussion by Machette.

Mileage	Description of features along route.
0.0	Retrace route north to intersection of Utah Hwys 146 and 92.
0.4	Turn left (west) onto Utah Hwy 92. Road traverses Holocene flood-plain gravels of the American Fork River, then climbs up onto the Provo fan-delta complex.
0.9	On the north, notice entrance to the Highlands Westroc gravel pit (see day 3-stop 3).
2.2	Intersection with Utah Hwy 74. Proceed west and cross channel of Dry Creek.
4.2	Turn (left) south on first road west of Dry Creek. This road affords a typical view of exposures along banks of Dry Creek. Proceed southwest for ¼ mile along north bank of Dry Creek, cross canal, and turn (right) west onto 10400 North.
4.5	Turn left (south) on 7600 West. Proceed south across surface of Provo delta and descend into valley of Dry Creek.
5.2	Just before Dry Creek, turn left (east) into Lehi Mortar Sand pit for stop 4. Park on south side of pit.

DAY 3-STOP 4 discussion by Machette and Currey.

Mileage	Description of features along route.
0.0	Leave gravel pit, turn left (south) onto 7600 West. Cross Dry Creek and climb back onto Provo fan-delta complex (here the road is Dry Creek Way).
0.7	Descend front of delta. On left (southeast), note exposures of foreset beds as well as backset beds related to construction of a northwest-trending spit.
0.9	Turn right (west) on 9600 N. Proceed ¼ mile.
1.1	Turn right (north) on 8000 West.
1.6	Cross floodplain of Dry Creek and turn left (west) onto 10000 North. The road crosses a series of regressional shorelines etched across the Provo delta.
2.5	After crossing the railroad tracks, the road swings north parallel to I-15.
2.8	Enter onramp to northbound I-15. Proceed north toward Salt Lake City on I-15.
4.6	Intersection with Utah Hwy 93.
8.8	I-15 offramp to Utah State Prison (on left) and field trip stop 2.
15.0	Between 106th South (Southtown Shopping Center) and 90th South, the low hill on the right (east) is a constructional spit of the Stansbury level (analogous to the larger and younger Draper Spit), which yielded a date of 20.4 ka (see day 2-stop 2). The spit is draped with younger deep-water pelagial sediment and shallow-water regressive sediment of the Bonneville lake cycle.
16.8	5400 South overpass.
25.2	Take exit for 6th South (east). Proceed into Salt Lake City.
27.5	Turn left (north) onto 7th East.
27.6	7th East and 4th South, Salt Lake City; end of road log for day 3.

TEMPORAL RELATIONS OF LACUSTRINE AND GLACIAL EVENTS AT LITTLE COTTONWOOD AND BELLS CANYONS, UTAH

by

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INTRODUCTION

Pleistocene glaciers descended below the shorelines of pluvial lakes in the western United States in only two localities--Mono Basin in the eastern Sierra Nevada and at Little Cottonwood and Bells Canyons in the Wasatch Mountains. The possibility of determining the temporal relations between lake and glacier fluctuations has led many investigators to these localities. G.K. Gilbert (1890), greatly influenced by I.C. Russell's (1889) work in Mono Basin, was the first scientist to address seriously this topic at Little Cottonwood and Bells Canyons:

> *The moraines of three Pleistocene glaciers descend from the Wasatch Mountains to the level of the Bonneville shore-line; the moraines of four glaciers descend from the Sierra Nevada to the level of the old shore-line of Mono Lake; and the relations of these moraines to the shores of the lakes and the associated deposits indicate that the maximum stage of the lakes coincided closely with the epoch of maximum glaciation (p. 318).*

The purpose of this paper is to trace the development of ideas about the relations between glacier and lake fluctuations and to discuss some paleoclimatic implications of current interpretations.

GLACIER AND LAKE RELATIONS

Gilbert (1890, p. 310) did not see evidence of the Bonneville shoreline eroded into the moraines and concluded that the glaciers advanced to their maximum extents after lake-level fell from the Bonneville to the Provo shoreline. However, he noted that the lake-level fall was caused by outlet erosion, and not by climatic change, and thus he reasoned that both events were broadly synchronous and had responded to the same climate change. Russell (1889) showed that Mono Lake, which did not overflow during its last deep-lake cycle, reached its maximum after Sierran glaciers had begun to retreat. Gilbert (1890, p. 313) inferred a similar relation in the Bonneville basin; that is, optimum conditions for supporting the lake lagged behind those for supporting glaciers. Gilbert (1890, p. 314-315) called for storage of precipitation during glacier advance and release of this water during glacier retreat to account for the lag in the lake maximum. He refuted the conclusion of Whitney (1882) that the rises of Great Basin lakes were caused directly by the melting of glaciers, noting that the volumes of ice and snow, even if melted instantaneously, would not fill the lakes. Gilbert's liberal ice-volume estimate in the Bonneville basin equals about 5 percent of the maximum lake volume.

In his notes, Gilbert (Hunt, 1982a, p. 164) described the greater fault offset of the "abruptly ended problematic moraine between the can(y)ons" (A on figure 1) compared with that of the conspicuous moraines and concluded that there were moraines of two glacial advances. However, no mention is made of evidence for multiple glaciation in Monograph 1. Rather, Gilbert (1890, p. 318) hypothesizes a bipartite glacial history as a corollary of the two-fold lake history. Beginning with Atwood's (1909) study, all subsequent investigators have agreed that glacial drift of at least two advances is present at the canyon mouths (figure 1; table 1). The difference in weathering of the drifts is conspicuous (Atwood, 1909; Ives, 1950; McCoy, 1977). The older drift has a greater degree of surface-stone weathering and boulders of quartz monzonite lying as much as several meters below the surface are partly to totally distintegrated. Soils developed in the older drift have thick, clay-rich B-horizons and depths of oxidation of several meters (150-ka soil in figure 2), whereas those in the younger drift have much weaker argillic B-horizons and depths of oxidation of about 3.3 feet (1.5 m) (similar to 13-ka soil in figure 2). On the basis of these differences, most workers have considered the drifts to represent separate glaciations; however, Richmond (1964) and Morrison (1965b, c) concluded that the drifts represented stades of a single (Bull Lake) glaciation (table 1).

Blackwelder (1931) and numerous subsequent workers demonstrated that the Bonneville shoreline is locally cut on the moraines and that lacustrine sediments overlie the younger drift (figure 1). A conspicuous shoreline is cut on the southwest flank of the terminal moraine of Bells Canyon. Relations at Little Cottonwood Canyon are different as the moraines there are partly buried by outwash deposited during glacier retreat. In contrast to the incision of moraine belts that typically accompanies retreat of valley glaciers, aggradation continued at Little Cottonwood in response to the rising lake. A large fan-delta (figure 1) was constructed that provided a broad, shallow area that lessened the energy of waves, and only low inconspicuous bars of gravel and sand mark the Bonneville shoreline on the outwash apron fronting the moraines.

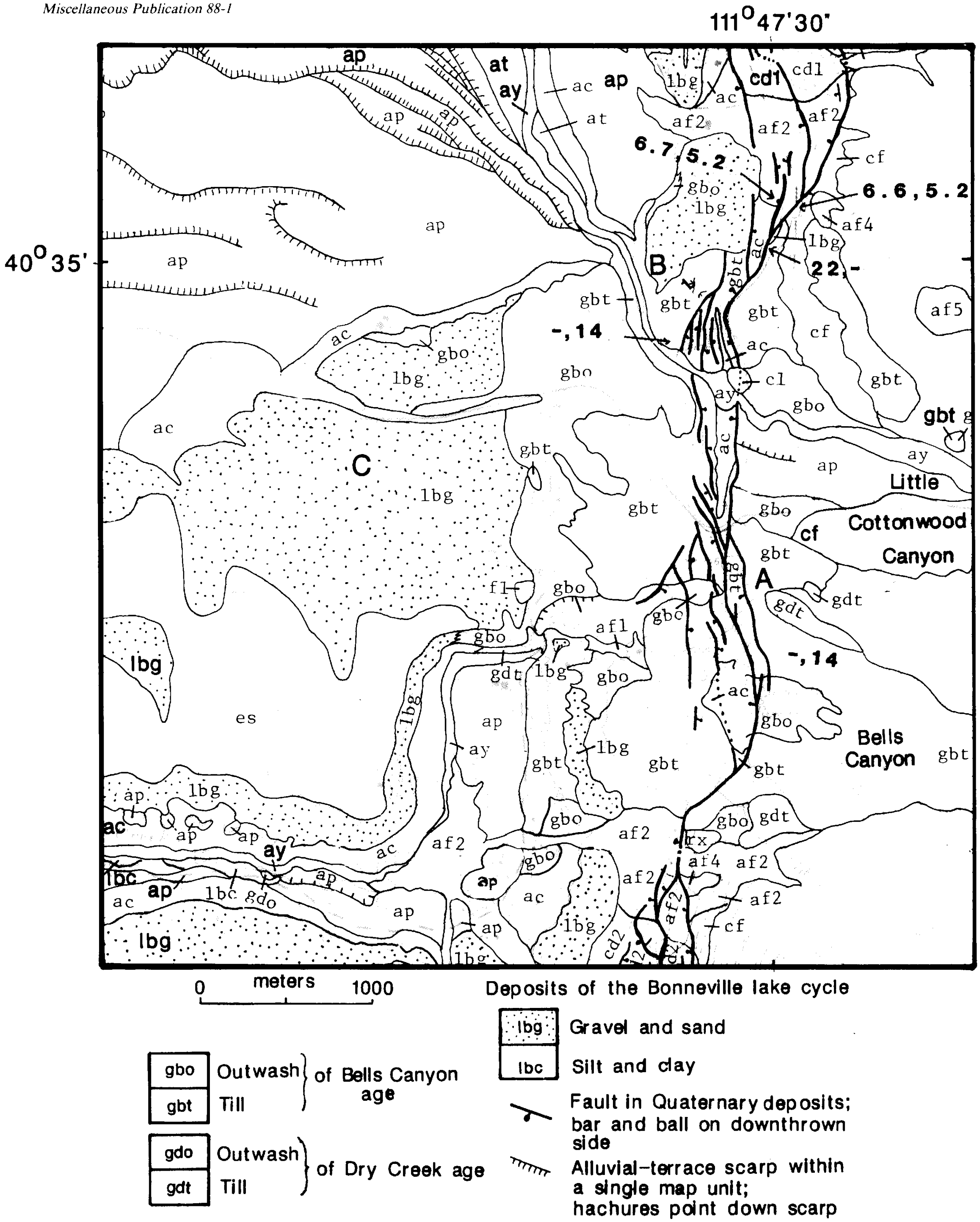

*Figure 1. Surficial geologic map of the Little Cottonwood-Bells Canyon area (from Scott and Shroba, 1985). Deposits other than those in explanation: Alluvial deposits--ay, upper Holocene flood-plain alluvium; at, Holocene alluvium of low terraces; ap, upper Pleistocene gravelly alluvium of terraces graded to Provo and high recessional shorelines; ac, Holocene alluvium and colluvium derived from unconsolidated deposits; af1, upper Holocene, af2, lower Holocene and upper Pleistocene, af4, middle Pleistocene, and af5, lower? Pleistocene alluvial-fan deposits (unit af5 may include some till). Colluvial deposits--cd1, upper Holocene, and cd2, lower Holocene and upper Pleistocene debris-flow deposits; cf, colluvium derived from bedrock; cl, landslide deposits. es, eolian sand; fl, artificial fill; rx, bedrock. Bold numerals indicate scarp height and net vertical displacement (in meters) across fault scarps or zones of fault scarps. A, older moraines referred to in text; B, site with buried soil between units gbt and lbg; C, fan delta of Little Cottonwood Creek. *1 is stop site.*

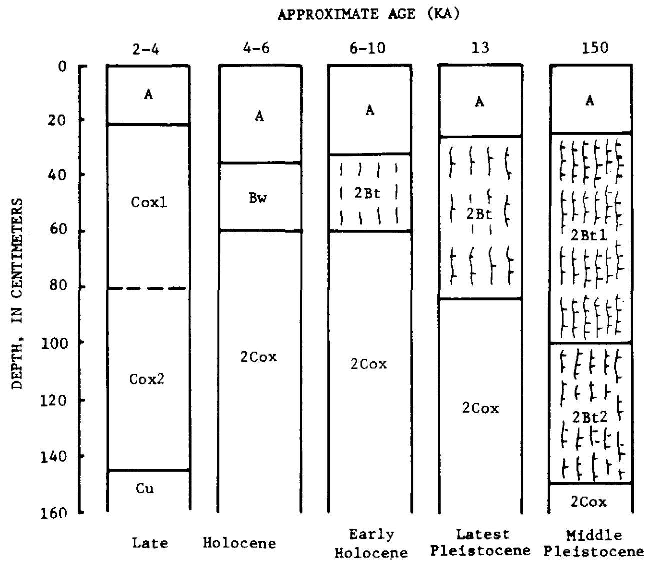

__Figure 2.__ Generalized soil profiles developed in unconsolidated deposits of various ages along the Wasatch front (from Shroba, 1982, 1984; Scott, McCoy, and Shroba, 1982; Scott and Shroba, 1985). Bw, cambic B horizon; Bt, argillic B horizon; Cox, oxidized C horizon; Cu, unoxidized C horizon. Numerals preceding horizon designations indicate changes in soil parent material; numerals following indicate subdivisions of horizon. For all of the soils, except the youngest one, the upper parent material is loess that is mixed with sand and gravel from the underlying alluvium or till. The relative density of the pattern in the Bt horizons corresponds to increased clay content, color, structure, and clay-skin development.

Table 1.
Interpretations of glacial and lacustrine deposits at the mouths of Little Cottonwood and Bells Canyons

Author	Which younger, moraines or Bonneville SL[a]	Age of event glacier	lake	Number of glaciations	Number of lake cycles
Gilbert (1890); *in* Hunt (1982a)	moraines, but both close	—	—	1 (hint of older)	1
Atwood (1909)	both close	Pinedale[b]	Pinedale[b]	2	1
Blackwelder (1931)	Bonneville SL[a]	Tahoe	Tioga	2	1
Ives (1950)	Bonneville SL[a] but both close	Pinedale[b]	Pinedale[b]	3	2
Eardley and others (1957)	moraines	late Wisconsin	early Wisconsin	3	2
Morrison (1965b)	Bonneville SL[a] but both close	late Bull Lake	late Bull Lake	2	2(3)[c]
Richmond (1964)	Bonneville SL[a]	late Bull Lake	Pinedale	2	3
Morrison (1965c)	Bonneville SL[a]	late Bull Lake	Pinedale	3(5)[c]	3(5)[c]
McCoy (1977)	Bonneville SL[a] but both close	Pinedale	Pinedale	2	1
Madsen and Currey (1979)	Bonneville SL[a], but both close	Pinedale	Pinedale	2	1
Scott and others (1983); Scott and Shroba (1985)	Bonneville SL[a] a few thousand yr after retreat from moraine	Pinedale	Pinedale	2	1

[a]Shoreline
[b]Author didn't use term, but their correlations indicate Pinedale age.
[c]Number in parentheses includes second-order lake or glacial advances rather than only first-order events.

The ages of the younger drift units and the overlying lake sediments have been contentious points. Blackwelder (1931), Eardley and others (1957), Richmond (1964), and Morrison (1965b, c) thought that the younger canyon-mouth moraines were of Tahoe or Bull Lake age (Iowan or early Wisconsin in prevailing terminology); Ives (1950) thought they were younger. Also debated was whether the overlying lake sediments are slightly (Ives, 1950; Morrison, 1965b) or substantially (Blackwelder, 1931; Richmond, 1964; Morrison, 1965c) younger than the moraines. After radiocarbon ages of deposits of the Bonneville lake cycle appeared (Broecker and Orr, 1958; Broecker and Kaufman, 1965) everyone agreed that the Bonneville shoreline was occupied during late Wisconsin, or Pinedale, time. Thus, by the mid-1960s the most widely referenced works (Richmond, 1964; Morrison, 1965c) asserted that the younger moraines were of early Wisconsin age and that the lake features post-dating the moraines were of late Wisconsin age (table 1).

Relative-age and radiocarbon-dating studies by McCoy (1977) and Madsen and Currey (1979) supported the interpretations of Atwood (1909) and Ives (1950) that the younger moraines (composed of till of Bells Canyon age) are of Pinedale age, as they are bracketed by a radiocarbon age of soil below the till of 26,080±1100/1200 yr B.P. and the overlying sediments of the Bonneville lake cycle. Madsen and Currey (1979) correlated the older moraines (till of Dry Creek age) with moraines of Bull Lake age in the Rocky Mountains, dated about 150 ka. Thus, the conclusions of Gilbert (1890) and Atwood (1909) that the last glacial and lake maxima were broadly synchronous in age were upheld, as were the views of Blackwelder (1931), Ives (1950), and others that Bonneville shoreline postdates the canyon-mouth moraines.

Evidence of the time interval between glacier and lake maxima can be found in an exposure along Little Cottonwood Creek about 1650 feet (500 m) northwest of the field-trip stop (B on figure 1). Here a weak soil developed in till of the younger moraines is buried by sediments of the Bonneville lake cycle. The soil consists of a partly eroded, oxidized C horizon about 20 inches (50 cm) thick and, compared with soils formed during the Holocene, indicates that a few thousand years elapsed between deglaciation and submergence by the lake (figure 2; Scott and others, 1982, 1983). Lake Bonneville reached the altitude of the buried soil (essentially the Bonneville shoreline) about 16,000-16,500 years ago (Scott and others, 1983; Currey and Oviatt, 1985) suggesting that deglaciation of this site, which is close to the maximum ice-front position, commenced about 18,000-20,000 years ago.

In summary, the Little Cottonwood and Bells glaciers reached their maxima of Pinedale age after 26,000 years ago, and began to retreat between 20,000-18,000 years ago as evidenced by the weak buried soil developed in the till and overlain by lake deposits. The lake first reached the moraine area about 16,500 years ago; optimal climatic conditions for supporting the lake occurred after 16,500 but before 14,000 years ago, excluding some time for lower lake levels during the Keg Mountain oscillation (Currey and Oviatt, 1985). Thus, a lag of several thousand years occurred between the times of optimal conditions for supporting the glaciers and the maximum stand of the lake.

PALEOCLIMATIC IMPLICATIONS

As suggested in the previous discussion, the Little Cottonwood and Bells glaciers reached their maxima between 26,000 and 18,000 years ago, a time interval during which the surface area of Lake Bonneville increased from 35 to 80 percent of its maximum extent (areas figured using time/lake-surface altitude and hypsometric data of Currey and Oviatt, 1985). For the lake to have been at an intermediate level when glaciers were close to their maximum extents suggests that the full-glacial climate in the Great Basin was cold and dry. Estimates of paleotemperatures based on a snowmelt model of the Little Cottonwood glacier and on rates of amino-acid epimerization (McCoy, 1981), and on the differences between glacial and present snowline altitudes (Porter and others, 1983), suggest that the full-glacial climate was 12-16 degrees C colder (mean annual temperature) than at present. As the lake was only at intermediate levels during such a low-temperature interval, and as it would fill to its highest shoreline under present precipitation conditions if mean annual temperature were decreased by 7 degrees C (McCoy, 1981), the precipitiation must have been substantially less than at present. Therefore, for the lake maximum to coincide with the retreat of the Little Cottonwood and other Rocky Mountain glaciers (Porter and others, 1983), the climate must have become relatively wetter and warmer. The increase in precipitation rate had to have been great enough to offset the effects of higher temperatures in increasing lake evaporation rates. Perhaps by the time of lake maximum, precipitation rates were close to present values and mean annual temperature was about 7 degrees C less than at present.

PALEOSEISMICITY AND EARTHQUAKE RECURRENCE AT LITTLE COTTONWOOD CANYON, WASATCH FAULT ZONE, UTAH

by

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At the mouth of Little Cottonwood Canyon, the Wasatch fault zone is expressed by some of the most spectacular and complex scarps observed anywhere along its length. The fault zone is as wide as 1320 feet (400 m). It consists of prominent curvilinear and *en echelon* west-facing and antithetic scarps that form a major graben. Individual west-facing scarps within the zone reach a height of 115 to 130 feet (35-40 m), and antithetic scarps range in height from several meters to 65 feet (20 m).

The fault at this location occupies a prominent place in Gilbert's field notes (Hunt, 1982a), where he included numerous sketches of scarps and moraines. It is likely that his observations here helped form many of his ideas about Basin and Range faulting. In fact, he visited this locality twice in 1877 to make notes on geology and surface-water resources, and in 1880 to spend several days mapping the geology. Gilbert (1890) mapped and profiled the glacial moraines, and he mapped fault scarps where they crossed the moraines and fan deposits and defined the major graben (figure 1). He noted that the fault zone was complicated and contained opposing scarps. Based on his observation that scarps in moraines were higher than scarps in alluvial deposits, he concluded that "It is evident that the total displacement was accomplished by a series of efforts" (Gilbert, 1890, p. 347). Gilbert's mapping compares favorably with detailed 1979 field and aerial photo mapping of the same location (figure 2).

The first paleoseismicity studies to quantify the past behavior and earthquake potential of this part of the Wasatch fault were made in 1979. These included excavation of trenches to define earthquake recurrence intervals and the amount of displacement per event, and topographic profiling of the faulted Bells Canyon lateral moraine to calculate the late Pleistocene slip rate. Trench and profile locations are shown on figure 2. Results of these investigations are discussed by Swan and others (1981) and Schwartz and Coppersmith (1984). One of the most important results was the recognition of a distinct Salt Lake segment of the Wasatch fault zone.

Estimates of recurrence and displacement per event at Little Cottonwood Canyon are complicated by multiple fault traces and the wide, complex zone of deformation. Trenches were excavated across the major graben, the main east-facing antithetic scarp, and the westernmost of three west-facing scarps that define the main fault zone. The western scarp on the main fault is 13 to 14 feet (4-4.5 m) high, and the two scarps to the east have heights of 6.5 and 11 feet (2 and 3.5 m). The trenches exposed till of the Bells Canyon advance, sediments of the Bonneville lake cycle, post-Bonneville alluvial-fan and graben-fill deposits, and scarp-derived colluvium. Within the trenches, two surface-faulting events were recognized. An early use of accelerator-mass-spectrometry radiocarbon dating on charcoal yielded an age of 9000 (+400; -600) ^{14}C yr B.P. for alluvium that graded to scarp-derived colluvium from the antithetic scarp (trench LC-3). Accelerator dates of 7800 (+400; -600) and 8600 (+500; -400) ^{14}C yr B.P. were obtained on charcoal from graben-fill deposits in a trench (LC-1) across the main scarp. This alluvium was correlative with the alluvium near the antithetic scarp. At both trenches the dated alluvium has been displaced by one surface-faulting event. In summary, the trenching suggested two surface-faulting events (earthquakes) within the main graben during the past 8000-9000 years. One occurred shortly before 8000-9000 years ago; the timing of the most recent was not constrained. These observations were used to calculate a maximum average recurrence interval of $\geq$ 4000-4600 years. An alternative recurrence interval of 2400-3000 years was calculated using a net tectonic displacement of 47 feet (14.5 m) across the Bells Canyon moraine, an age for the moraine of 19,000 $\pm$ 2000 years, and a displacement per event of 6.5 feet (2 m) (see below).

An average displacement per event was calculated using the depth to deposits of the Bonneville lake cycle across the graben. The top of these deposits is displaced approximately 42 feet (13 m) down to the west across the main scarp, and 30 feet (9 m) down to the east across the antithetic scarp. This

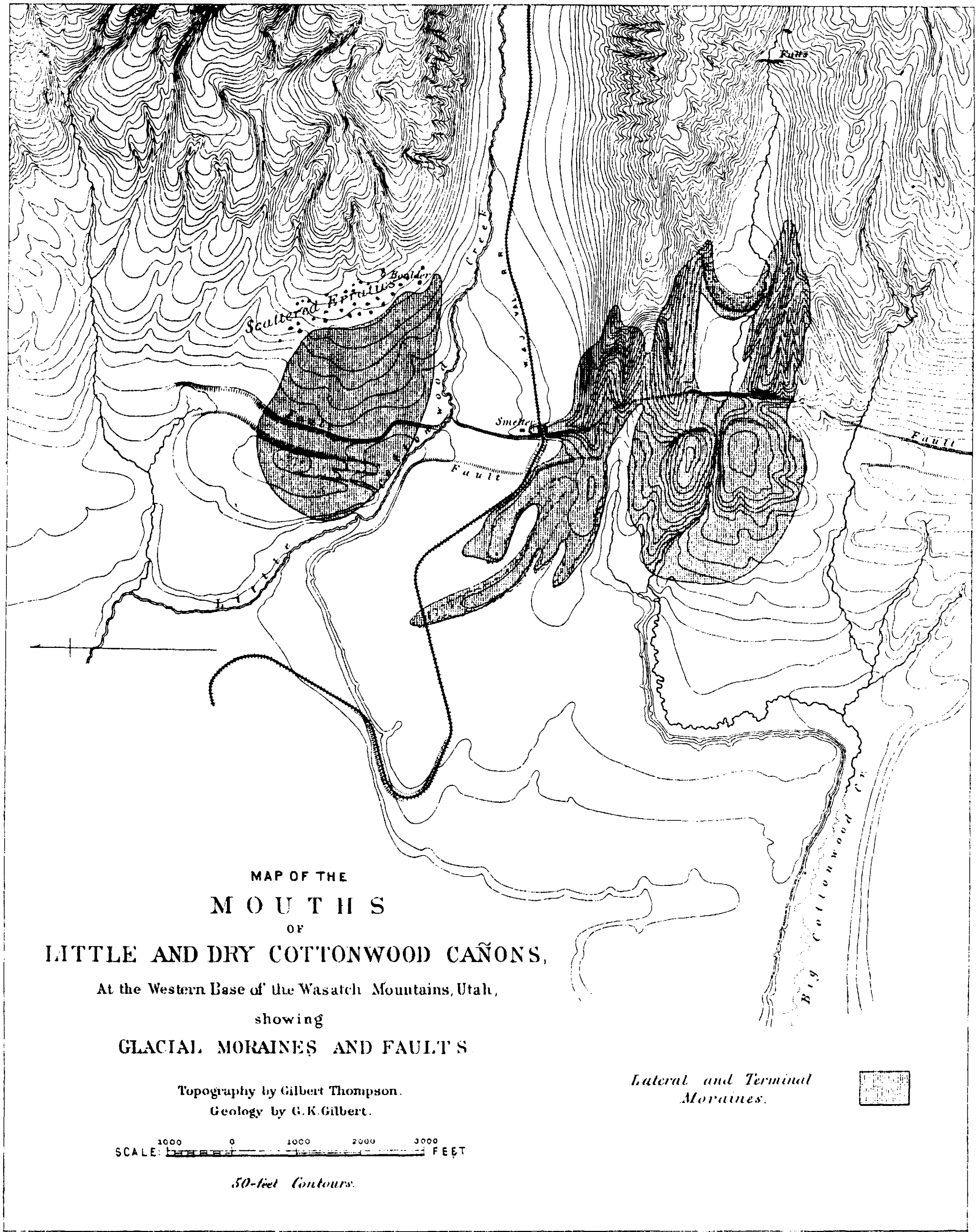

Figure 1. *Gilbert's (1890), plate XLII, mapping of moraines and fault scarps at Little Cottonwood Canyon, Wasatch fault zone.*

Figure 2. Low-sun-angle photograph showing the complexity and width of the Wasatch fault zone at the mouths of Little Cottonwood and Bells Canyons (from Swan and others, 1981).

yielded a net tectonic displacement across the graben of 13 feet (4 m), and an average displacement for the two events in the trenches of 6.5 feet (2 m) per event. However, there are two other splays of the main scarp for which there are no subsurface data. Both Swan and others (1981) and Schwartz and Coppersmith (1984) note that it is uncertain as to what degree the parallel scarps represent additional events or contribute to displacement per event.

The slip rate at Little Cottonwood Canyon is 0.76 (+0.6, -0.2) mm/yr. This is based on 47.5 feet (+32, -10) or 14.5 m (+10, -3) of net slip measured from topographic profiles across the left lateral moraine at Bells Canyon (profile A-A', figure 2) during the past 19,000 ± 2000 years. The slip rate here is similar to late Pleistocene-Holocene rates calculated elsewhere along the fault zone.

In retrospect, the uncertainty regarding the role of parallel fault traces during individual events appears to have been warranted. Recent investigations at Dry Creek Canyon, 1¼ miles (2 km) south of Little Cottonwood Canyon (Lund and Schwartz, 1987; Schwartz and others, 1988) clearly demonstrate the occurrence of two post-middle-Holocene events. Mean-residence-time radiocarbon dates of A-horizon soils buried by scarp-derived colluvium show that the event prior to the most recent occurred shortly after 5545-5975 yr B.P. The most recent event is less well-constrained; it occurred shortly after 1130-1830 yr B.P. Both of these events should also have occurred at Little Cottonwood Canyon. The fault at Dry Creek contains five sub-parallel to *en echelon* scarps in a zone 1312 feet (400 m) wide, comparable to Little Cottonwood Canyon. Displacement occurred along each scarp during the past two surface-faulting earthquakes. Estimates of slip per event based on topographic profiling of a levee-forming debris flow displaced by only the most recent event and of alluvial-fan deposits displaced by two events, and estimates based on the thickness of colluvial wedges and displaced marker horizons exposed in trenches, indicate net tectonic slip of 13 to 16 feet (4.5-5 m) per event during each of the past two earthquakes. This is the largest displacement value for a single event measured along the Wasatch fault zone. This net slip could also define the slip per event at Little Cottonwood Canyon and would be consistent with the broad fault zone and high scarps. The style of faulting at Dry Creek suggests that the parallel scarps at the Little Cottonwood Canyon trench site may have all slipped simultaneously during individual events. The difference in fault behavior interpreted at Little Cottonwood Canyon and Dry Creek Canyon emphasizes that the true paleoseismic history can only be developed when information on timing and slip per event is obtained for every scarp in a broad, complex fault zone.

By combining observations from Little Cottonwood Canyon and Dry Creek Canyon, it appears that there have been three large-magnitude surface-faulting earthquakes during the past 8000-9000 years. One occurred shortly before about 8000-9000 years ago, one shortly after 5500-6000 years, and the most recent shortly after 1100 to 1800 years ago. Taking into account the uncertainties in timing of events, intervals of 4000±1000 years may characterize recurrence along this segment of the Wasatch fault zone.

In 1883, after studying Little Cottonwood Canyon and other Wasatch sites, and confident that his theories about mountain building and earthquakes were correct, Gilbert issued an earthquake hazard warning to the residents of Salt Lake City. In an article in the Salt Lake City Tribune (Sept. 20, 1883), reprinted in the American Journal of Science (1884), he summarized his ideas and emphasized their practical application in Utah. He stated that the mountains in the Basin and Range Province are uplifted along faults, that they rise in small increments following the release of stress that has accumulated slowly over long periods of time, and that the "instant of yielding is so swift and abruptly terminated as to constitute a shock" (Gilbert, 1884, p. 50). Wallace (1980, p. 38) points out that given the current understanding of how earthquakes are generated, Gilbert's reasoning relating mountain building to earthquakes "is so modern that, in 1980, it is difficult to understand why, once stated, the concept would not have been generally accepted and become a firm part of the working base of geologists and seismologists."

One hundred five years have passed since Gilbert's earthquake warning to Salt Lake City. That warning is still in effect. Present-day investigators of the Wasatch fault zone are attempting to quantify its earthquake potential by developing site-specific information on the fault's past behavior. These studies seek to quantify fault behavior in terms of timing of individual events, recurrence intervals, slip per event, slip rate, and segmentation models. It was Gilbert's insights into faulting, earthquakes, and mountain building processes along the Wasatch fault zone, and particularly at Little Cottonwood Canyon, that laid the foundation for the modern paleoseismic investigations.

DEPOSITS OF THE LAST TWO DEEP-LAKE CYCLES AT POINT OF THE MOUNTAIN, UTAH

by

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INTRODUCTION

Point of the Mountain is a compound, V-shaped spit that lies at the west end of the eastern Traverse Mountains between Salt Lake and Utah Valleys. Gilbert (1890) described the Bonneville shore embankment on the northwest face of the range and a great 1000-feet-high (300 m) shoreline cliff towering above the embankment that provided most of the debris in the spit. A lesser volume of debris was transported from the south flank of the range by currents in the bay that occupied Utah Valley. The Provo shoreline is marked by a cliff and broad terrace eroded into the Bonneville embankment and spit. A marked lower limit of juniper trees above the Provo shoreline along all but the southern part of the embankment coincides roughly with the contact between Lake Bonneville deposits and an abrasion platform cut on highly fractured quartzite bedrock of the Pennsylvanian Oquirrh Formation. The high degree of fracturing of the bedrock was a major factor leading to formation of the high cliffs behind both shorelines and deposition of voluminous spits.

Our field-trip stop is at the pit operated by Geneva Rock Products Corporation, the southern of two gravel quarries excavated into the cliff at the back of the Provo shoreline near the end of the spit. Both quarries expose deposits of the Bonneville and Little Valley lake cycles (figure 1; Scott and others, 1983); the northern one also exposes quartzite bedrock. The informal terminology of Scott and others (1983) was replaced by McCoy (1987), who designated formal allostratigraphic units (North American Commission on Stratigraphic Nomenclature, 1983), the Bonneville and Little Valley Alloformations. This area is not part of a steep range front with deeply incised stream valleys, so the disconformity between the alloformations does not have the great (tens to more than 100 m) relief that is found elsewhere along the Wasatch front. The alloformations are also separated by subaerial deposits and a well-developed soil, the Promontory Geosol (Promontory Soil of Morrison, 1965; Scott and others, 1983; McCoy, 1987).

BONNEVILLE AND LITTLE VALLEY ALLOFORMATIONS

The Bonneville Alloformation exposed in the Geneva Rock Products quarry consists of generally northwest-dipping, well-sorted beds of gravel, sand, and minor silt and sand that were deposited in a complex of bars and spits during the transgressive phase and maximum stand of the Bonneville lake cycle. It is identified on the basis of its stratigraphic position and a radiocarbon age of 18,600 ± 150 yr B.P. (W-4693) on wood from a thin lagoonal mud at its base (Scott and others, 1983; McCoy, 1987). This age and another of 20,900 ± 250 yr B.P. (W-4897) on wood from a lower altitude in the quarry operated by Salt Lake Sand Co. (west of Highway I-15) are consistent with other dated transgressive deposits of the Bonneville lake cycle.

The Little Valley Alloformation consists of mostly northeast- to east-dipping, well-sorted beds of gravel, sand, and minor marly silt and fine sand. The northeast dips indicate that the deposits represent the backset (shoreward-dipping) beds of a bar or spit whose lakeward portion was eroded during lake stands at or near the Provo shoreline. The Little Valley deposits are identified on the basis of their position below the Promontory Geosol and the ratio of alloisoleucine to isoleucine (aIle/Ile) in fossil gastropod shells. Shells of the gastropod *Amnicola* have aIle/Ile ratios of 0.47 ± 0.02 compared with an average of 0.43 ± 0.05 for all samples from the Little Valley Alloformation and an average of 0.19 ± 0.04 for samples from the Bonneville Alloformation (McCoy, 1987). Shells of marginal-lacustrine snails suitable for amino-acid analysis are markedly more abundant in deposits in areas like Point of the Mountain than in areas close to major streams draining the Wasatch Mountains. Presumably the cold, sediment-laden, glacial-outwash streams provided an inhospitable environment for snails.

PROMONTORY GEOSOL

The Promontory Geosol exposed in the Geneva quarry has a well-developed, cumulic Bw/Bt/K/Ck profile, but it has not been analyzed in detail. The upper part of the soil is formed in silty clay loam with scattered pebbles, which is probably a colluvial deposit derived from sand and gravel of Little Valley age and loess. In descending order the most complete soil profile consists of a color B or weak argillic B horizon 21 inches (55 cm) thick, an argillic B horizon 16 inches (40 cm) thick that has clay films on stones and ped faces and has calcium carbonate nodules and coatings on stones (stage I-II carbonate morphology), a K horizon 16 inches (40 cm) thick that has continuous calcium carbonate accumulation in the matrix and continuous coatings on stones (stage III carbonate morphology), and a Ck horizon 14 inches (35 cm) or more thick that displays stage I and minor stage II carbonate morphology. The soil is variably eroded and locally has been entirely removed The buried soil is similar in degree of development to other examples of the Promontory Geosol that have been analyzed in detail by R.R. Shroba (Scott and others, 1982, 1983). Calcic horizons of these soils contain 50 ± 10 g of secondary calcium carbonate per cm^2-column of soil. Based on regional rates of secondary calcium carbonate accumulation in soils during post-Bonneville time (0.5 $g/cm^2/10^3$ yr), these buried soils are estimated to have formed over an interval of 100,000 ± 20,000 years. Most were buried about 20,000 years ago by transgressive deposits of the Bonneville lake cycle, and therefore the estimated age of the parent material is 120±20 ka (figures rounded from those in Scott and others, 1983, table 5).

AGE OF THE LITTLE VALLEY ALLOFORMATION

Scott and others (1983) concluded that the Little Valley lake cycle probably is contemporaneous with the younger part of oxygen isotope stage 6 of the marine record (Shackelton and Opdyke, 1973; Imbrie and Imbrie, 1980), which ended about 130 ka, based on the following lines of reasoning.

1) The estimated age of the Little Valley Alloformation, based on the degree of development of the Promontory Geosol as discussed above, is about 120 ka.

2) Shells from deposits of Little Valley age have ^{230}Th ages of 93±10 ka and greater than 105 ka (Kaufman and Broecker, 1965) and probably should be regarded as minimum ages (Kaufman and others, 1971).

3) The Bonneville lake cycle culminated late in the last glaciation, which is broadly equivalent to oxygen-isotope stage 2. The lake apparently did not reach the Bonneville shoreline until several thousand years after the Little Cottonwood glacier had retreated from its late Wisconsin maximum (see day 3-stop 1). By analogy, the Little Valley lake cycle may also date from late in a major glaciation. The one that best fits the above estimates is the Bull Lake glaciation in the Yellowstone area, which occurred during oxygen-isotope stage 6 (Pierce and others, 1976).

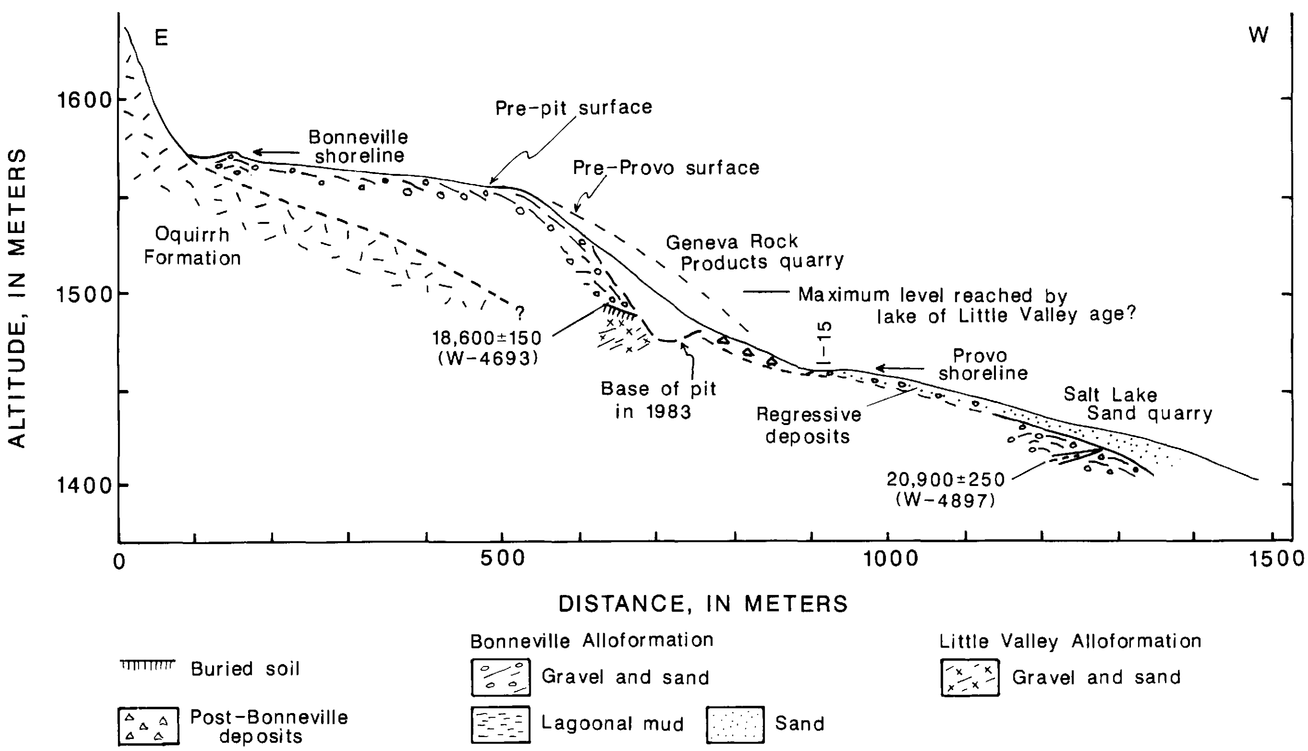

Figure 1. *Sketch of stratigraphic relations in the Geneva Rock Products and Salt Lake Sand Co. quarries at Point of the Mountain.*

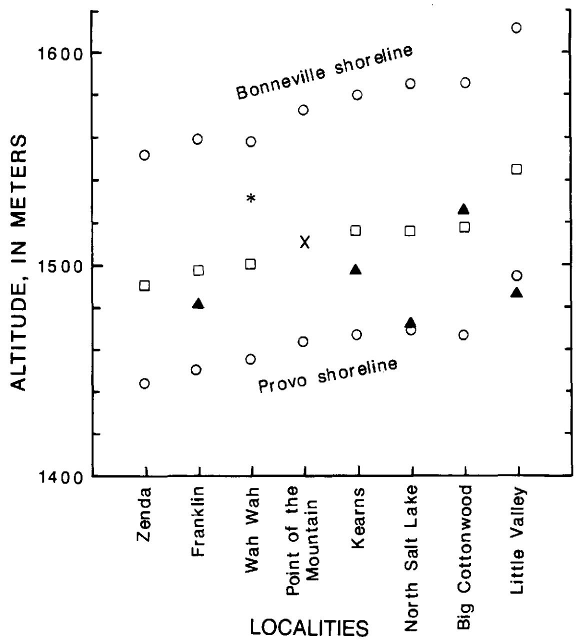

*Figure 2. Present (rebounded) altitudes of deposits of the Little Valley lake cycle (triangles; X, Point of the Mountain) plotted with local altitudes of the Bonneville and Provo shorelines (circles). *, highest Intermediate shorelines of Gilbert (1890) in Preuss (Wah Wah) Valley, which he interpreted as being of Yellow Clay age. Squares denote estimated maximum local altitude of lake of Little Valley age. These estimates are based on the assumptions that (1) the top of the Little Valley Alloformation at Point of the Mountain represents the culmination of the Little Valley lake cycle and (2) the highest altitude reached by the lake of Little Valley age at a locality is about 60 percent of the vertical distance down from the Bonneville to Provo shorelines. As discussed in text, relations at Big Cottonwood are complicated by faulting.*

MAXIMUM LEVEL REACHED BY LAKE DURING LITTLE VALLEY CYCLE

As quarrying progressed during 1980-1983, the northwest-sloping top of the Little Valley Alloformation in the Geneva quarry was found to flatten at an altitude of about 5000 feet (1510 m), which may mark the highest level reached during the Little Valley lake cycle (figures 1, 2). This level is about 200 feet (63 m) below the local altitude of the Bonneville shoreline, which represents about 60 percent of the altitudinal drop between the Bonneville and Provo shorelines (350 feet; 109 m). Isostatic deformation makes altitudinal comparisons difficult between features in different parts of the basin, but this problem can be minimized by determining the position of features relative to local altitudes of the Bonneville and Provo shorelines. By using this approach, the only higher locality at which deposits of probable Little Valley age have been found is about 15 miles (25 km) north of Point of the Mountain, near the mouth of Big Cottonwood Canyon. There the contact between foreset deltaic gravel and interbedded and overlying outwash of Bull Lake age is about 5000 feet (1525 m), or about one-half the vertical distance between the local altitudes of the Bonneville and Provo shorelines (figure 2). However, the contact has been displaced about 130 feet (40 m) by faulting so that its original altitude is poorly controlled (Scott and others, 1982). Within the uncertainties related to faulting and isostatic deformation, the deposits at both localities could relate to the same lake level. Exploration of numerous areas at altitudes closer to the Bonneville shoreline have failed to find any convincing evidence of Little Valley deposits (see also day 3-stop 3). Thus Gilbert's (1890) conclusion from study of shorelines in the Wah Wah (Preuss) Valley that the highest level reached by the earlier lake lies about 90 feet (27 m) below the level of the Bonneville shoreline (see Introduction to this volume) appears in error; it probably got no closer to the Bonneville shoreline than about 200 feet (60 m).

The lake level of 200 feet (60 m) during Little Valley time corresponds to a lake area about 15-20 percent smaller than that during the Bonneville lake cycle. Because the lake overflowed before reaching its highest potential level during the Bonneville lake cycle, this difference represents a minimum estimate of the contrast between the two cycles. The cause of the difference may have been the diversion of the Bear River, which presently contributes about 20 percent of the inflow to Great Salt Lake, into the Bonneville basin late in the interlacustral episode between the Little Valley and Bonneville lake cycles (Bright, 1963; McCoy, 1987; Oviatt and others, 1987).

AMERICAN FORK CANYON, UTAH: HOLOCENE FAULTING, THE BONNEVILLE FAN-DELTA COMPLEX, AND EVIDENCE FOR THE KEG MOUNTAIN OSCILLATION

by

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INTRODUCTION

We have stopped here in the northeastern corner of the Utah Valley to review evidence for Holocene movement along the Wasatch fault zone, observe Gilbert's type locality for a Bonneville-level fan-delta complex, and discuss evidence for interpretating a two-fold high-stand of Lake Bonneville that is punctuated by the Keg Mountain oscillation.

G.K. Gilbert passed through the Utah Valley several times in the 1870s, first in 1872 as a member of the Wheeler Survey, again in 1876 on his way south to the Henry Mountains, and lastly in 1879 on his way west to explore the Bonneville basin. On August 20, 1876, he travelled from Salt Lake City to York, Utah (near Nephi) by train, and noted the youthfulness of faulting along the front of the Wasatch Range in the Juab Valley. However, on the return trip to Salt Lake City (November 22, 1876) he recognized some of the gross geomorphic features of the Utah Valley. Gilbert was an astute observer and, aided by field glasses, he saw evidence of young faulting along the Wasatch Front of the Utah Valley. Gilbert was still struggling with several working hypotheses to explain the morphology of mountain fronts, as evidenced by this entry in his notebook:

> *The fault lines in this locality (sic, 2 miles north of Springville, Utah), are crooked and the mountainside is so steep that if there were no other terms to the argument it might not be easyto †ntertain (sic, entertain) an hypothesis of faults against one of superficial slides (Hunt, 1982a, p. 19).*

One of the "other terms to the argument" was the continuation of these scarps across river floodplains, where they clearly are not related to landslides.

From the train (4 miles southwest of American Fork Canyon) Gilbert observed that

> *near American Fork Cañon one of these faults comes down to the Bonneville Beach but is of small throw and the spit does not seem good for measurement.*

In fact, on closer examination Gilbert would have found that the mouth of American Fork Canyon is one of the best places to measure the throw across the Wasatch fault zone.

Gilbert made a second and more productive trip through the Utah Valley three years later. He and his assistant, Israel C. Russell, stayed in Provo for two nights and spent three days (November 9-11, 1879) exploring the eastern margin of the Utah Valley, presumably on horseback.

HOLOCENE FAULTING AT AMERICAN FORK CANYON

Gilbert's main interest at American Fork Canyon was the Bonneville- and Provo-level deltas, rather than the evidence for faulting. However, his observations in this area, which, coupled with his study of faulting at Rock Creek (northeast of Provo), influenced his thinking about earthquake processes and faulting mechanisms in the eastern Basin and Range Province. For example, he states that the northern and southern parts of

> *...the delta are traversed close to the mountain base by a fault scarp 60-70 feet high. The same displacement (sic, fault) traverses the flood plain of the stream, but its throw there is only 15 feet, showing that the entire displacement of the delta was not accomplished in a single movement (Gilbert, 1980, p. 346).*

Gilbert's notes also mention 30-40 feet (9-12 m) of throw on the modern flood plain (Hunt, 1982a, p. 123), although I suspect that this measurement was made on a fault scarp on terrace deposits of the American Fork River that grade into the Provo fan-delta complex. In 1985, I measured a 27-feet-high (8.2 m) scarp on this same terrace but found no scarp in the modern floodplain (which has been regraded for a flood-catchment basin). Nevertheless, Gilbert was comfortable with the concept of episodic, repeated movement on the Wasatch fault zone.

Gilbert was impressed by the apparent youthfulness of faulting, both here and elsewhere along the Wasatch fault zone. Concerning faulting across the American Fork River, he states "The last disturbance of the floodplain was so recent that a rapid still marks the acclivity it produced in the bowlder-paved stream" (Gilbert, 1890, p. 346). Obviously, Gilbert had an intuitive feeling for the interdependence of stream processes and mountain-building processes but had no way to ascertain the absolute timing of geologic events. However, many of his impressions of relative timing, derived largely from estimates of volumes of materials and rates of processes, would prove to be accurate on the basis of absolute dating techniques a century later.

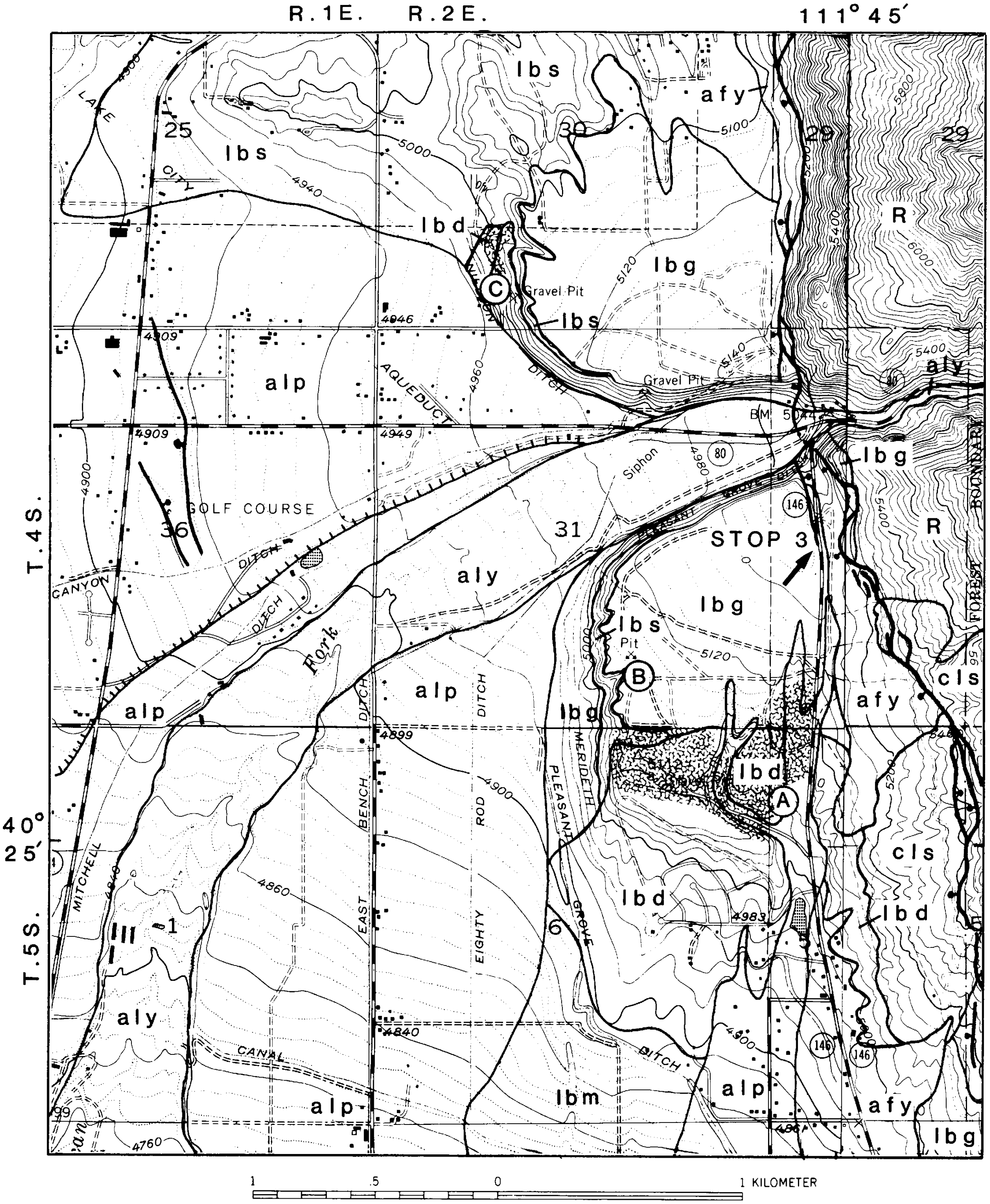


Figure 1. Index map showing surficial geology and scarps of the Wasatch fault zone at field trip stop 3, just south of the mouth of American Fork Canyon. Map units are from young to old: afy, Holocene alluvial and debris-flow fans; aly, Holocene stream allulvium; alp, gravelly alluvium graded to Provo-level deltas; lbg, lbs, and lbm--gravel, sand, and silt/clay (respectively) of the transgressive phase of the Bonneville lake cycle; lbd, sediment (sand and gravel) of Bonneville-level delta; cls, pre-Holocene landslide deposits; R, Paleozoic bedrock undivided. Faults are heavy solid line with bar and ball symbol on downdropped block. Hachured line indicates fluvial scarp on Provo fan-delta complex. Arrow indicates camera station for photograph of fault scarps (figure 2). Circled letters A, B, and C are locations of sections shown in figure 3.

The Wasatch fault zone is marked by a set of prominent fault scarps at the foot of the Wasatch Range, both north and south of American Fork Canyon (figure 1). As Gilbert noted, repeated movement on the fault zone has produced progressively more net displacement of increasingly older deposits. Just to the north of this field trip stop, gravels of the highest Bonneville shoreline are displaced 50-65 feet (15-20 m) across a wide graben that has a synthetic range-bounding fault. On the north bank of American Fork Canyon there is minor backtilting but no extensive graben, yet gravels of the same stratigraphic level are displaced 75 feet (23 m) (net). If one uses 50-75 feet (15-23 m) as the permissible range in net displacement and 15 ka for the youngest probable age of the gravel, then the average slip rate on the Wasatch fault zone has been 1.0-1.5 mm/year at this site. In contrast, the long-term slip rates recorded by 130- to 250-ka alluvium along this and other segments of the central Wasatch fault zone have been only 0.1-0.2 mm/year (Machette and others, 1987, fig. 5).

In order to document the episodic Holocene movement of the Wasatch fault zone, Bill Lund (UGMS) and I excavated three trenches 80-165 feet (25-50 m) across several strands of the fault zone in 1986. All three trenches were placed on Holocene alluvial fans that consist primarily of debris-flow deposits. An age of 7290 ± 100 ^{14}C yr B.P. was obtained from charcoal in loess directly beneath the debris flows and an age of 4740 ± ^{14}C yr B.P. was obtained from charcoal in the upper part of the faulted debris flows. These two dates correspond to dendrochronologically calibrated ages of 8061 cal yr B.P. and 5518 cal yr B.P., respectively (note that the corrections result in ages 771-778 years (10-16 percent) older than the ^{14}C dates). The younger dated charcoal comes from about 3 feet (1 m) below the surface of the debris flow sequence, is not associated with a soil, and thus probably predates the pre-fault surface of the alluvial fan by 100 years or less. Therefore, the ^{14}C ages constrain a major phase of debris-flow deposition and associated fan building to the period from less than 8.1 ka to about 5.4 ka--that is, from early to middle Holocene.

Figure 2. *Photograph of the main scarp of the Wasatch fault zone at the site of trench AF-1 (labeled AF-1). Scarp on left side of photo is about 65 feet (20 m) high. View to northeast; see arrow in figure 1 for camera station.*

The surface of the debris-flow deposits commonly is displaced 23-26 feet (7-8 m) by three faulting events where the fault zone consists of a single fault or closely spaced parallel normal faults, such as at the site of trench AF-1 (figure 2). The Holocene events each produced discrete colluvial wedges, although only trench AF-1 contained evidence for all three events. Trench AF-2 recorded the most recent and probably the second faulting events. Trench AF-3, which is not illustrated in figure 2, recorded the second and an older (fourth) event that occurred pre-5.4 ka and post-8.1 ka.

Trench AF-1, the deepest and longest of the three trenches, was placed across the main fault zone and penetrated a relatively simple sequence of three colluvial wedges, each of which were derived locally from the erosion of material exposed during faulting. In turn, each of the wedges is separated by 1.6-3.3 feet (0.5-1 m) of well-developed A horizon and a weakly developed calcic A horizon and (or) calcic C horizon. Machette and Lund (1987) had estimated that these soils probably required 1000-2000 years to form on the basis of their development and thickness.

Trench AF-2 crossed an east-facing antithetic scarp that forms an 80-foot-wide (25 m) graben adjacent to the main scarp of the fault zone. The third trench (AF-3) crossed a small synthetic fault scarp that is the northern end of a left-stepping *en echelon* splay of the main fault zone. Thermoluminescence (TL) and radiocarbon (^{14}C) dating of samples from the three trenches yielded age estimates that establish relatively tight constraints on the time of the three most recent faulting events (figure 3).

Two radiocarbon dates (980 ± 70 and 620 ± 150 yr B.P.) and two TL age estimates (400 ± 100 and 500 ± 200 yr ago) from trenches AF-1 and AF-2 provide relatively firm constraints on the time of the most recent faulting at the American Fork Canyon site. The calendar-corrected AMRT (apparent mean resident time) dates from the soil yield ages of 714 and 612 cal yr B.P. which I interpret as a maximum time for the most recent event. In addition, I subtract 100 years from these dates to allow time for accumulation of the dated carbon prior to burial; thus, the final AMRT-derived ages are 0.6 and 0.5 ka. These ages and the two TL age estimates (figure 3) suggest that the most recent faulting occurred between 400 and 600 years ago (the window for all four age estimates). However, the accuracy of TL age estimates of less than 1,000 years is suspect because of low light levels and potential errors in determining dose rates (S.L. Forman, oral communication, 1988). Therefore, we place less emphasis on the TL age estimates of 400 and 500 years and more emphasis on the AMRT-derived ages of 0.5 and 0.6 ka. On this basis, we suggest that the most recent faulting at American Fork Canyon occurred 550±100 years ago.

I suspect that Gilbert would be surprised at our estimate of the antiquity of the most recent faulting event. He was impressed by the fresh character of the fault scarps that cross alluvial fans which debouch from the Wasatch Range, here and near Nephi (where many of the scarps are on bedrock). He suggested that "its (sic, the faults) antiquity is measured perhaps by years instead of centuries. It must be far more recent than the Bonneville Beach" (Hunt, 1982a, p. 19). If the most recent faulting is 550 yr BP., it is but 1/30th the age of the "Bonneville Beach."

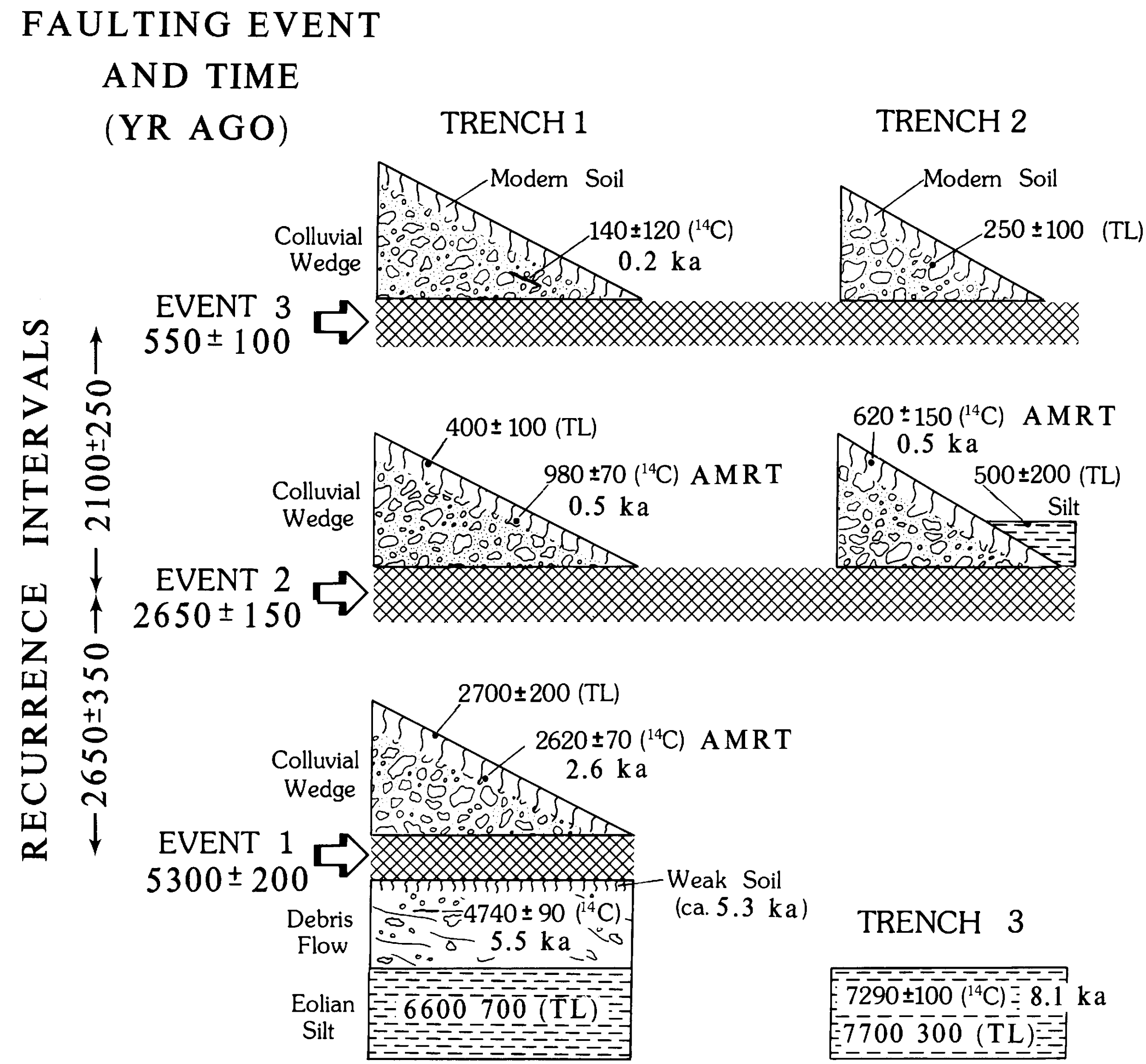


Figure 3. Schematic interpretation of timing of Holocene faulting events derived from trenches at American Fork Canyon. Bold arrows indicate times of major faulting events since 5.5 ka. Values in ka are estimated age of the upper soil horizon (buried contact) based on calendar-corrected radiocarbon dates (^{14}C) and adjustments for apparent mean-residence times (AMRT) of soil carbon at time of burial.

The evidence for and dating of the second-to-most-recent faulting event is clearest in trench AF-1 (figure 3). The strategy for dating this event was the same as the most recent event. Samples for TL and ^{14}C analysis were collected from the upper part of the buried soil on the third (and oldest exposed) colluvial wedge in the trench. These samples provide maximum limiting dates on the time of burial during the second faulting event. The A horizon yielded two ages: (1) a TL age estimate of 2.7 ka and (2) an AMRT date of 2620±70 yr B.P. that corresponds to a calendar-corrected date of 2777 cal yr B.P. I subtracted 200 years for the estimated ^{14}C age of the carbon in the soil at the time of burial which gives an estimate of 2.6 ka for the time of burial of the soil on the third (lowest) colluvial wedge. Because the TL- and ^{14}C-derived age estimates are in stratigraphic accord and have concordant error limits (about 150-200 years each), I used the average of 2.6 and 2.7 ka to date the second faulting event at 2650±150 yr B.P. (see figure 3). Thus, the interval of quiescence (recurrence interval) between the most recent (550±100 years B.P.) and second (2650 ± 150 years B.P.) faulting is 2100 ± 250 years.

The first fault event recorded in trench AF-1 must have occurred after about 5.4 ka as determined from a ^{14}C date of 4740±90 yr B.P. from charcoal within debris-flow deposits that are buried by the third and stratigraphically lowest colluvial wedge. The surface of the debris flow has a very weak A horizon on it, one that might form in 100-200 years. Calendar correction of the ^{14}C date resulted in three possible calendar ages whose mean is 5518 cal yr B.P. Assuming that 5.5 ka is an accurate age for the charcoal, I suspect that the surface of the overlying debris flow was buried by a wedge of fault-scarp colluvium about 5.3±0.2 ka. Therefore, the recurrence interval between the first and the second events is about 2650±350 years (if the previous assumptions about timing and error limits are correct).

In summary, the analysis of paleosismicity at the American Fork Canyon site relies on conventional- and accelerator-method radiocarbon dating of charcoal and soil carbon (AMRT dates), TL dating analysis, and detailed mapping of lithologic and stratigraphic units. The results form a coherent basis for interpretating three Holocene fault events about 550, 2650, and 5300 years ago. If the most recent surface faulting event is correctly dated at about 550 yr B.P. and the oldest at 5.3 ka, then major faulting events with 16-20 feet (5-6 m) of vertical slip (displacement) occurred during a time span of 4,750 years (±350 yr). The resulting slip rate for the middle to late Holocene is 1.1-1.4 mm/year (5-6 m/4, 750±350), which is consistent with the slip rates of 1.0-1.5 mm/year determined for the past 15 ka from faulted beach gravels of the Bonneville lake cycle. These values confirm that the rate of slip on the Wasatch fault zone has remained high relative to other normal faults in the Basin and Range province during the past 15 ka and especially during the latter half of the Holocene.

THE BONNEVILLE FAN-DELTA COMPLEX: GILBERT'S CONCEPTS AND EVIDENCE FOR THE KEG MOUNTAIN OSCILLATION

The mouth of American Fork Canyon is dominated by two well-preserved fan-delta complexes; a small, high one that was constructed during the deepest phase of the Bonneville lake cycle (about 18-15 ka from Currey and Oviatt, 1985, fig. 2) and a second, larger one constructed during and after the catastrophic fall to the Provo level of the Bonneville lake cycle (about 14 ka). Russell (*in* Gilbert, 1890, p. 158) calculated that 1/10th of the material in the Provo level fan-delta was derived from erosion of the older fan-delta and from material stored in the canyon of the American Fork River. Although both of these features are of great size and volume, neither took long to construct.

By about 13 ka, Lake Bonneville had shrunk to an altitude below 4500 feet (1371 m), which left Utah Lake near its present level, permanently isolated from the progressively more restricted main body of Lake Bonneville (Van Horn and Varnes, this volume, would argue for a younger major expansion of the lake to form the Draper Formation). Utah Lake was controlled by a natural sill at an altitude of about 4490 feet (1368 m). After Lake Bonneville dropped below the sill, subsequent rises of Lake Bonneville due to minor climatic variations (such as the Gilbert expansion of Currey and Oviatt, 1985) could only have been accompanied by overflow of Utah Lake rather than a rise in its level. This relation suggests that none of the regressive shorelines above 4,500 feet (1371 m) in the Utah Valley can be much younger than 13 ka. Some of the low-altitude alluvial-fan complexes mapped as Holocene by Hunt and others (1953) and Miller (1982), such as the ones at Lehi and American Fork, have shorelines cut across their distal toes at altitudes of 4520 to 4560 feet (1377-1390 m); these shorelines are related to the regression of Lake Bonneville and not to Holocene expansions of Utah Lake.

Gilbert thought that American Fork Canyon was the best locality to study a Bonneville-level delta (Gilbert, 1890, p. 155), and dispatched I.C. Russell to examine it for him. Russell's observations, which were based on a single visit, are enlightening especially considering the relatively poor natural exposures that he had compared to the artificial exposures we have today.

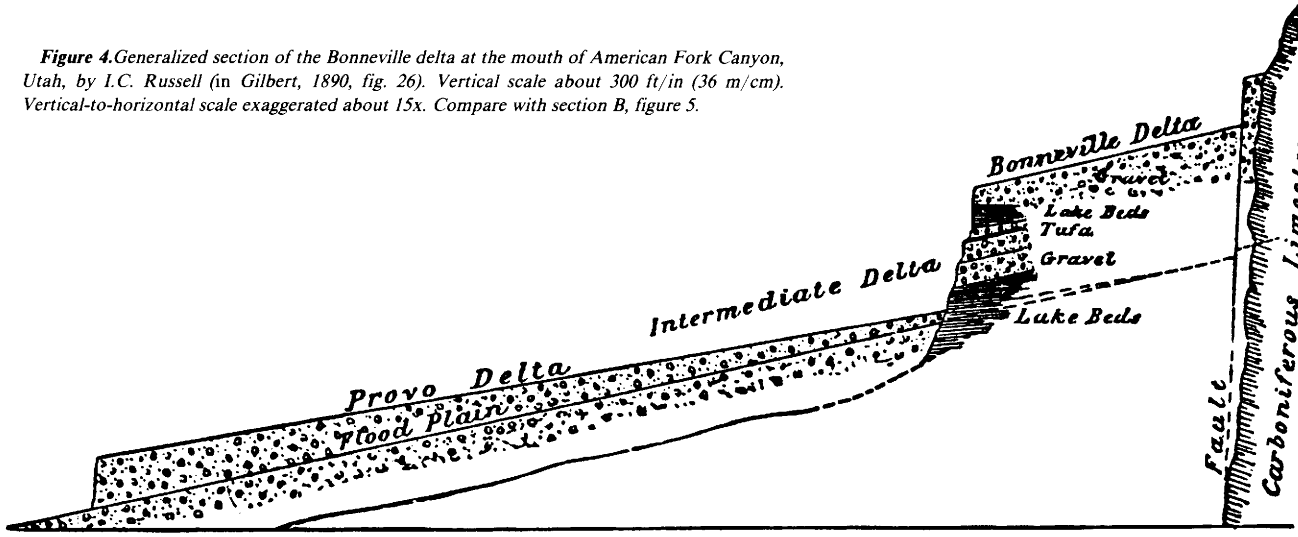
Russell found evidence for a two-fold depositional sequence within the high delta at American Fork Canyon, that is, an intermediate-level delta that was subsequently covered by "the Bonneville delta" (Gilbert, 1890, p. 157). Separating these two deltas is a tufa-encrusted beach gravel (bed 3, figure 4) that, according to Russell, forms a prominent bench around three-fourths of the terrace (i.e., the eroded edge of the delta). There is no evidence of the tufa in the present exposures around or in the delta. On the basis of Russell's section, one could interpret the tufa (figure 4) in two grossly different stratigraphic contexts: (1) the tufa of the intermediate-level beach represents the high stand of an old (pre-Bonneville) lake cycle, or (2) it represents a still stand or a minor regression within the predominantly transgressive phase of the Bonneville lake cycle. Russell found the tufa 65 feet (20 m) below the top of the Bonneville fan-delta surface, which places it at an altitude of 5000 to 5020 feet (1524-1530 m) or about 140-160 feet (43-49 m) below the highest deposits of the Bonneville lake cycle on the downdropped side of the fault zone. According to Scott and others (1983, p. 280) the "highest known deposits of the little Valley lake cycle ...range from 75-120 m below the local altitude of the Bonneville shoreline, the highest level reached during the Bonneville lake cycle." Thus, based solely on altitude, Gilbert's Intermediate delta must relate to an oscillation in the late (high) stage of the Bonneville lake cycle rather than to a pre-Bonneville lake cycle.

Currey and others (1983, p. 77, fig. 11) and Currey and Oviatt (1985, p. 12-13) present evidence from the Keg Mountain area of west-central Utah for a minor regression (oscillation) during the final transgressive phase of the Bonneville lake cycle. They suspect that the lake may have dropped 130-165 feet (40-50 m) below its previous high during the Keg Mountain oscillation. These values agree with the position of the tufa in the American Fork delta (140-160 feet or 43-49 m below the Bonneville shoreline) and lend credence to such an interpretation.

The general sedimentology of the American Fork fan-delta complex is marked by a coarse-fine-coarse sequence that requires a minor regression (oscillation) during the final transgressive phase of the Bonneville lake cycle. Figure 5 shows schematic sections of the coarse-fine-coarse sequence of the American Fork fan-delta complex as exposed in abandoned gravel pits in the southern lobe of the delta (section A), in natural exposures on the west face of the delta (section B), and in the 180-foot-deep (55 m) Highlands Gravel Pit north of the American Fork River (section C). Although deep enough, tufa has not been found in either section C or B. Section A is not deep enough to penetrate the tufa as depicted by Gilbert (figure 4).

I interpret the sections in figure 5 to represent a two-fold transgression of Lake Bonneville, resulting in two sequences of

Figure 4. Generalized section of the Bonneville delta at the mouth of American Fork Canyon, Utah, by I.C. Russell (in Gilbert, 1890, fig. 26). Vertical scale about 300 ft/in (36 m/cm). Vertical-to-horizontal scale exaggerated about 15x. Compare with section B, figure 5.



sediment. The first (lower) transgressive sequence consists of coarse beach gravels and near-shore sediments (mainly silty sand to sandy gravels) that were deposited as the water depth increased. In my interpretation, the lower sequence is separated from the incomplete upper one by a change from fine to coarse sediment, whereas Russell interpreted the tufa (bed 3, figure 4) as marking an unconformity (the top of the intermediate delta). Regressive deposits (gravels) related to the fall in lake level during the Keg Mountain oscillation are not seen in these sections, probably because the American Fork River would have cut a channel through the delta and transferred its depocenter westward into the basin. In addition, the low stand during the Keg Mountain oscillation may have trimmed the toe of the pre-existing fan-delta complex at American Fork Canyon. The second transgression would then have overlapped the first delta, thus resulting in a two-fold delta sequence. The second regression, from the Bonneville high stand to the Provo level during the catastrophic Bonneville Flood, caused the American Fork River to cut 165±6.6 feet (50±2 m) into the Bonneville fan delta, thereby eroding its central part. Although this regression is not preserved on or in the Bonneville delta, it is recorded by the construction of large Provo-level fan-delta complexes, such as the one we will see next at stop 4.

The sediment exposed in the Highland Gravel Pit (section C, figure 5) is particularly interesting for three reasons. First, the gravel pit provides a unique opportunity to look deep into a Bonneville fan-delta complex. The lower two-thirds of the pit is composed of monotonous, non-bedded to uniformly bedded sandy pebble to small cobble gravel that fines noticeably to the northwest (away from the source area). The poorly bedded gravel in the lower one-third of the pit could be glacial outwash (graded to a transgressive delta) that formed a large fan before Lake Bonneville rose to this altitude (about 17 ka; from Currey and Oviatt, 1985, fig. 2). Beneath the delta in the adjacent natural exposures, Hunt and others (1953) mapped gravel and sand of the Alpine Formation, whereas Miller (1982) mapped Pinedale glacial outwash: the material found in the current pit supports the latter interpretation, which appears to have been made on a conceptual rather than factual basis. Along Dry Creek (about 4 miles or 6 km to the north), Pinedale glacial outwash is composed of rounded boulders as large as 3-6 feet (1-2 m) in diameter at the mouth of the canyon and about 1 foot (0.3 m) in diameter near the Bonneville shoreline (see Gilbert; 1890, Hunt and others, 1953; Miller, 1982). One would expect the glacial outwash to be finer-grained at the mouth of American Fork Canyon because the Pinedale glaciers terminated at least 6 miles (10 km) farther upstream. The lake beds beneath the intermediate delta noted by Russell (figure 4), however, are not exposed in this rather deep pit.

Secondly, the deltaic foreset beds exposed in the northwest corner of the Highlands pit (section C, figure 5) are most likely part of the second transgression (and high stand) of the lake. Evidence for this interpretation is two-fold: (1) the foresets contain chaotic blocks of sandy silt that only could have been derived from erosion of the finer-grained part of the first transgression (tl, figure 5) and (2) the foresets project to or above the fine-grained section, and thus probably relate to the topset beds that are exposed at the surface of the delta (t2, figure 5).

Thirdly, the foreset beds form rhythmically bedded packets that were apparently deposited in response to annual variations in stream discharge and winter freezing of the Utah Valley arm of Lake Bonneville. These annual packets are 4-8 inches (10-20 cm) thick and generally consist of the following subunits:

1. Sandy silt to silt, finely bedded with sparse, coarse pebbles (dropstones?). Deposited in low-flow regime; interpreted as mainly glacial flour that settled out in quiet-water conditions (winter).
2. Fine-to-medium-grain sand, slightly pebbly, moderately well sorted in well-layered planar beds. Deposited in moderate- to declining-flow regime (summer and fall).
3. Basal granule to coarse pebble gravel with a silty sand matrix, crudely bedded, poorly sorted to unsorted. Deposited under high-flow regime (spring and summer).

This section, plus foreset beds of the first transgression (tl) exposed at locality A (figure 5) indicate that the Bonneville fan-delta complex had a pre-incision radius of 0.9 to 1.2 mile (1.5-2 km).

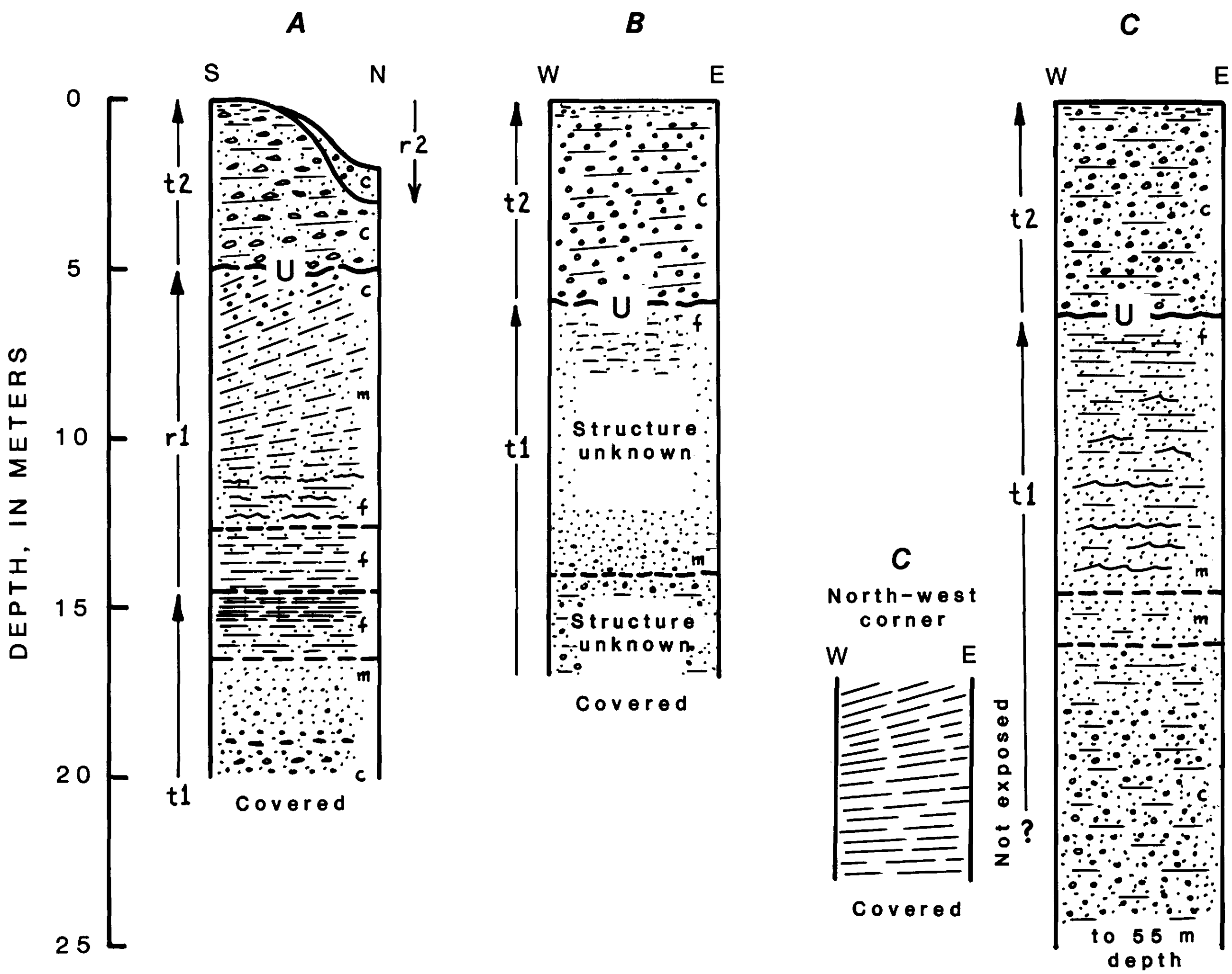


__Figure 5.__ Schematic sections of Bonneville fan-delta complex of American Fork Canyon, Utah. Section A, composite from gravel pits on south end of southern lobe of delta; section B, bluff exposure on west side of southern lobe of delta (compare with figure 4); two sections at C, from Highlands gravel pit on western edge of northern lobe of delta. See figure 1 for location of sections. Symbols: U, major unconfomiry (wavy line); t, transgressive sequence; r, regressive sequence; gravel, open circles; sand, fine stipple; silt/clay, dashed lines. Predominant grain size of units shown in right-hand part of columns: c, sandy pebble gravel to pebbly sand; m, fine to medium sand; f, silty sand to silt. Inclined bedding indicates deltaic foreset beds.

SUMMARY

Although G.K. Gilbert and I.C. Russell only spent three field days in the Utah Valley (and only one near this stop), their conclusions about the Bonneville deltas and recency of movement on the Wasatch fault zone amaze me. Again, we are presented with clear evidence that "G.K. Gilbert should be rated the greatest geologist this country has ever produced" (Hunt, 1982a, p. vii).

EXPOSURES OF TRANSGRESSIVE AND REGRESSIVE SEDIMENTS OF THE BONNEVILLE LAKE CYCLE ALONG DRY CREEK NEAR LEHI, UTAH

by

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INTRODUCTION

This field trip stop, known informally as the Lehi Mortar Sand pit, is the last stop on the southern leg of our field trip. The pit (and others along Dry Creek that have been largely regraded) exposes a nearly complete section of the sediment of the Bonneville lake cycle (the Bonneville Alloformation). During the past five years mining in this pit has exposed an upper section of deltaic gravelly sands, silts and clays, and a lower section of pebbly sand and gravel. Elsewhere in Utah Valley, one can see only the upper part of this section owing to the thickness of sediment and a general paucity of deep excavations.

Gilbert probably did not travel along Dry Creek, but instead stood upon the Provo-level delta just north of American Fork and cast his eyes both to the west (Lehi) and south (Orem and Provo) at deltas graded to the Provo level. Hunt and others (1953) were the first to examine carefully the internal structures of these deltas in the Utah Valley, and a number of subsequent workers (e.g., see Jones, 1965 and Fouch and Dean, 1982) studied the sedimentology of these classic deltas. In 1963, Bissell published his mapping of the southern part of the Utah Valley, an area where the Spanish Fork River had constructed a large Provo-level delta. More recently, Miller (1982) compiled a reconnaissance map of the surficial geology of the Utah Valley.

While exploring the Utah Valley, Gilbert had a topographic map of the delta at Provo, which J.H. Renshawe had made for E.E. Howell during the 1872 Wheeler Survey (in Hunt, 1982a, p. 123). Howell recognized the deltas as fluvial-lacustrine features and was the first to apply the name Provo to the shorelines and deltas at that level (Gilbert, 1890, p. 153). Two-thirds of a century later Bates (1953, figure 3) applied the term "Gilbert-type delta" to the large coarse-grained deltas with topset, foreset, and bottomset beds that result from homopycnal flow, typically at the front of mountains and build out into deep lakes. These features have become widely known as Gilbert deltas in recognition of his concepts of delta formation. Most of the Gilbert deltas of the Bonneville basin are associated with the Provo level of the lake; the Bonneville-level deltas often provided material for the Provo-level features.

Surficial geology along Dry Creek

The surficial geology along Dry Creek was mapped by Hunt and others (1953; 1:62,500 scale), compiled by Miller (1982 1:100,000 scale), and remapped by me at 1:50,000 scale. Each of our maps show a cover of Provo-level deltaic deposits (figure 1), but Hunt and others (1953) map the underlying silt and clay as Alpine Formation, the oldest of three formations they related to Wisconsin-age lakes in the Bonneville basin. They envisioned an early, deep lake that deposited most of the lacustrine deposits (Alpine Formation) in the Utah Valley and a later equally large lake that deposited near-shore sand and gravel (but, inexplicably, no deep-water silt or clay) of the Bonneville Formation, and all of the Provo Formation. In this part of Utah Valley, I interpret their Alpine and Bonneville Formations as the transgressive part of the Bonneville Alloformation (see McCoy, 1987) and their Provo Formation as the regressive part of the Bonneville Alloformation, all of which was deposited during the Bonneville lake cycle as defined by Scott and others (1983). Hunt and others (1953) correlated the Alpine and Bonneville Formations with glacial outwash from Wisconsin (Bells Canyon advance) moraines and from older Wisconsin (?) (Dry Creek advance) moraines, respectively.

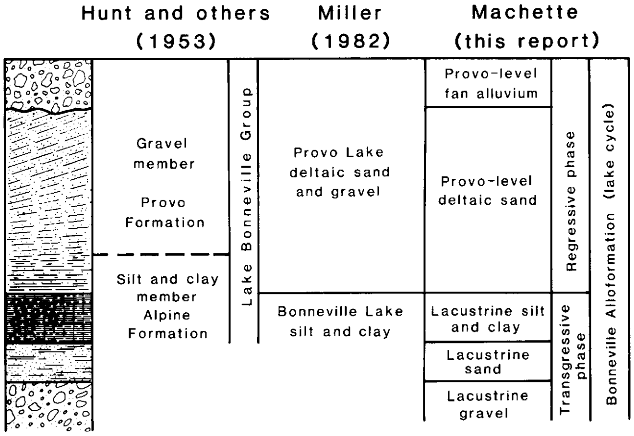


Figure 1. *Schematic section of surficial deposits along Dry Creek as showing nomenclature.*

Neither Miller (1982) nor I found evidence for an earlier Wisconsin (i.e., pre-Bonneville, post-Little Valley) lake in the Utah Valley. I suspect that there are subsurface deposits of the Little Valley lake cycle (Little Valley Alloformation) in the Utah Valley. However, none of the exposures that I have studied show the characteristic, well-developed soils (such as the Promontory or Dimple Dell Geosols of Morrison, 1965a) that indicate a significant lacuna between the Alpine and Bonneville Formations. Oviatt and others (1987) suggest that the lake of the mid- to early Wisconsin Cutler Dam cycle rose to a maximum altitude of about 4400 feet, (1340 m) which is 90 ft (27 m) below the natural threshold between Utah Lake and Great Salt lake. Thus, it appears that there should be no record of a mid- to early Wisconsin lake in Utah Valley, and that any significant record of the older (pre-Wisconsin) Little Valley lake cycle still is buried beneath the thick cover of sediments of the Bonneville Alloformation.

Stratigraphic section at Lehi Mortar Sand pit

Although there have been no published accounts of this nearly complete sedimentary section, the beds exposed in the Lehi Mortar Sand pit have been briefly studied by W.E. Scott during his 1977-1982 reconnaissance study of Bonneville lake cycle deposits, by Miller (1982) during his mapping of the Utah Valley, and by D.R. Currey during several forays into the valley. I measured a section in the pit in 1986 and revisited the pit with D.R. Currey and T.D. Fouch on two separate occasions during the spring of 1988. The following description (table 1) is compiled from notes taken during my three brief visits.

Interpretation of depositional environments

The beds (units A-F) described in table 1 are interpreted as recording a simple transgression and regression of Lake Bonneville during the last deep lake cycle (the Bonneville lake cycle). Deposition at this site was as much as 400 feet (125 m) below the high stand of the lake, thus there should not have been a marked change in sedimentation rate, texture, or chemistry during the Keg Mountain oscillation of Currey and others (1983; compare with evidence for a major oscillation at the previous field trip stop).

In some of the formerly deep-water parts of the Bonneville basin (such as near Salt Lake City), the rapid drop in lake level during the catastrophic fall to the Provo level (-353 feet, -108 m) may have caused deposition of anomalously high concentrations of microfossils (such as ostacode coquina beds; see Keaton and others, 1987, p. 4) or produced subtle sedimentologic changes such as a sandy bed within deep-water clays. At this site, the drop in lake level was recorded by a subtle shift from deep-water silty clays (unit D) to a fining upward sequence of sandy silt (lake bottom) and silty sand (bottomset deltaic sediments, unit C) as the Dry Creek delta shifted basinward. Therefore, the contact between units D and C is the least well-established in the section.

The foreset beds of unit B, which core the regressive Provo-level delta, are in the coarsest part of the deltaic sequence. Within the delta, individual beds commonly are trough cross-stratified and have abundant evidence of gravity-induced downslope slumping, scouring, and sediment flow (Fouch and Dean, 1982, figures 48-50). Periodic floods (during the waning Pinedale glaciation) and seismic events could have induced foreset avalanches and sedimentary gravity flows that ultimately resulted in multiple-graded bottomset sand and silt beds (T.D. Fouch, written communication, 1988).

The uppermost sedimentary unit (A) of the section is comprised of topset beds of the Provo-level fan delta, which rises 200 feet (60 m) to the northeast in about 3 miles (5 km) (see figure 1, day 3-stop 3). These beds are composed of poorly sorted, trough cross-stratified gravel to sandy gravel. About 0.6 miles (1 km) south of this field trip stop, T.D. Fouch (written communication, 1988) found a rich assemblage of pulmonate molluscs that are indicative of a fluvial environment. The fine-silt component of the soil (table 1) that has formed on the fan-delta is interpreted as an eolian component rather than as a lacustrine unit (the Draper Formation) related to a reoccupation of the Provo level late in the Bonneville lake cycle (see Van Horn and Varnes, this volume).

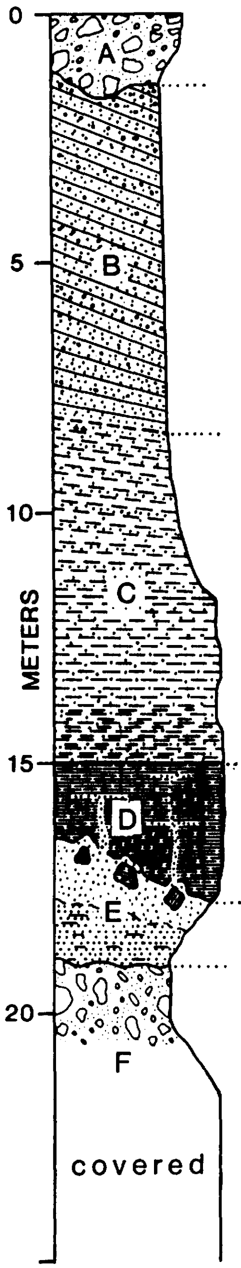
The sediment of the Provo-level delta in this area is fairly restricted in age owing to the relatively high base level (4700 feet; 1432 m) of the sub-delta platform. From Currey and Oviatt's (1985) hydrograph of Lake Bonneville, I suspect that the transgressive sediment is about 20 ka (unit F) to about 15 ka (unit D), whereas the regressive sediment is 14.5-13 ka (units C-A).

Tectonic implications of injected sand in unit D

Clastic sand dikes and sills in unit D and nonabraded chunks of unit D mixed in the source sand (unit E, figure 2) suggest a period of strong ground shaking (paleoliquefaction) that is probably related to a major surface-rupturing earthquake along the Wasatch fault zone. The sand was injected as a fluid in unit D but not higher in the section (units C-A), and I found no evidence of surface venting of the sand (sand cones or wedges) in unit D or above. This relation suggests that the injection occurred when there was an overlying section of sediment (i.e., load) to constrain the injection. In addition, because the sediment was water saturated, the water table had to have been high (at or above the base of the pit) during this event. As Lake Bonneville retreated from the Utah Valley, streams incised below the deltas and effectively drained them. Currey and Oviatt's (1985) hydrograph of Lake Bonneville suggest that the lake level dropped below Utah Valley (4400 feet; 1340 m) by about 13 ka, thereby leaving a narrow time window for the injection of the sand dikes and sills. Thus, the injection event must be contemporanous with or soon after the regressive Provo phase of the lake cycle (14.5-13 ka here).

TABLE 1. MEASURED SECTION THROUGH THE BONNEVILLE ALLOFORMATION (SEDIMENT OF THE BONNEVILLE LAKE CYCLE)

[Lehi Mortar Sand pit, north bank of Dry Creek, 1 1/4 mi (2 km) NE of Interstate Highway 15 at Lehi, Utah. W1/2 W1/2 NW1/4 NE1/4 Sec. 4, T.5S., R.1E., Lehi 7.5-minute topographic quadrangle. Altitude of top of pit (surface of unit A) is 4784±3 ft (1458±1 m); altitude of working base of pit (unit F) is 4700±10 ft (1432±3 m). Measured section is 60-70 ft (18.3-21.3 m) thick]



Unit A. Topset beds of delta. Fluvial sandy pebble to small cobble gravel: light gray to light brown (where silty or clayey), crudely bedded in troughs, dip parallel to surface (1-2° west). Has soil with light-reddish brown argillic B horizon (mainly from loess) and whitened stage I-II calcic C horizon). Thickness variable, commonly only 3-6 ft (1-2 m) with angular unconformity at base.

Unit B. Foreset beds of delta: Lacustrine sand and pebbly sand; light gray to light brown, well bedded, dip west about 20° (beds strike N.10-20°E.). Contains abundant large- and medium-scale subparallel trough bedsets (some graded) with sharp bases. Comprised of 70 percent sand beds and 30 percent pebbly sand beds; coarsest material is pea to small pebble-size gravel (1 in., 2.5 cm diameter). Best exposed in north and west walls of pit. Thickness 21-25 ft (6.5-7.5 m), base is covered by spoil.

Unit C. Bottomset beds. Upper part (10-13 ft; 3-4 m): proximal beds of delta. Well bedded sand, pebbly sand, and sandy silt; concave upward, dip 2-5° west. Abundant convoluted bedding from slumping and (or) dewatering of sediment. Lower part (10-13 ft; 3-4 m): distal beds of delta and deep-water sediment of regressive phase of lake cycle (basal contact obscure), well bedded, dip 1-2° west. Light brown sand (with finely laminated ripple marks) to sandy silt, regular-planar bedded to slightly cross-bedded in shallow channels. Many beds show down-dip slump structures. West side of pit has dune-form beds created by west-trending oscillatory (standing) wave scour and fill. Beds are commonly 0.25- to 0.5-inch (0.7- to 1.2-cm) thick and fine upwards (annual couplets?). Contains load structures, but no injected sand (as in unit D). Thickness 23 ft (7 m), laterally gradational with unit B. Base of regressive phase.

Unit D. Distal (upper) to proximal (lower) deep-water sediment of transgressive phase of Bonneville lake cycle (the "White Marl" of Gilbert, 1890). Consists of brown to olive green silt, olive to dark gray (organic-rich) silty clay, and minor beds of light brown fine sand to silty sand. Calcareous, forms blocky outcrop. Silts finely laminated in planar beds; locally has penecontemporaneous deformation structure. Fine- to medium-grained sand has been injected as dikes and as sills (source of sand is unit E). Thickness 5 to 8 ft (1.5 to 2.5 m).

Unit E. Distal near-shore sediment. Light gray medium- to coarse-grained sand with sparse granules. Upper 1.5 ft (0.5 m) is unbedded (liquified?); contains detached, angular blocks of unit D that are rotated but not abraded. Lower part is crudely bedded in horizontal layers. Thickness 5-8 ft (1.0-1.5 m).

Unit F. Proximal near-shore sediment. Light gray sandy pebble to small cobble gravel. Unit covered during visit in June 1988. Lower part probably contains 3-6 ft (1-2 m) of beach gravel deposited during transgression of lake. Lake gravel may lie unconformably on well developed calcic soil (the Promontory or Dimple Del Geosols of Morrison, 1965) that is on either alluvial fan sediment or lacustrine sediment of the Little Valley Alloformation.

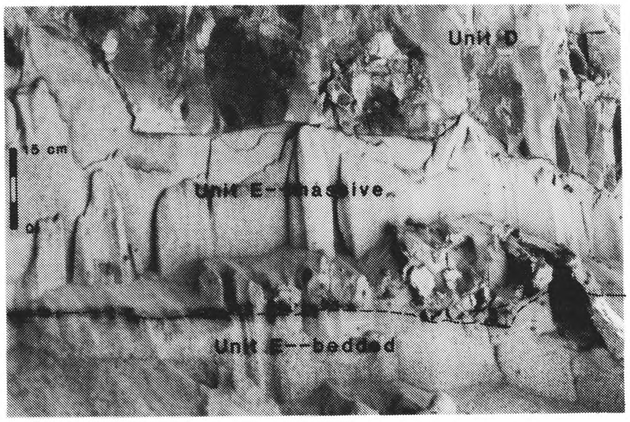


Figure 2. Photograph of deformed sand (unit E, table 1) directly beneath silty clay (unit D, table 1) Note (1) large blocks of unit D that have been detached and moved into underlying sand (unit E), (2) lack of bedding in upper half meter of unit E that is probably due to liquefaction, and (3) disturbed but nearly horizontal base of unit D.

Machette and others (1987, p. 44) have found sparse evidence of rapid slip along the Wasatch fault zone between the high stand of Lake Bonneville and the stabilization of the lake at the Provo level. For example, at Hobble Creek (southeast of Provo, Utah), the 15- to 17-ka Bonneville shoreline is offset about 130-150 feet (40-45 m) along the Wasatch fault zone, whereas the slightly younger Provo-level fan-delta complex is only offset 37.7 to 44.3 feet (11.5-13.5 m) (Swan and others, 1980). This relation demands a short-lived period of rapid tectonic offset on the Wasatch fault zone. Machette and others (1987) have proposed that this paleoseismic episode is associated with the rapid crustal rebound (and extension) that accompanied draining of Lake Bonneville. If one projects the Wasatch fault zone at dips of 45-60° in the subsurface, earthquakes nucleating 9 miles (15 km) deep (a common depth for normal faults in the Basin and Range Province) should have hypocenters about 5.5-9 miles (9-15 km) west of the mountain front; that is, along the central axis of Utah Valley. The Lehi Mortar Sand pit is about 4 miles (6.5 km) west of the mountain front at American Fork Canyon. Thus, sandy lacustrine sediment at this pit lies close to the potential epicentral zone for earthquakes along this part of Wasatch fault zone and, thus, probably experienced strong ground acceleration and liquefaction during paleoseismic events.

THE DRAPER FORMATION (LAKE BONNEVILLE GROUP) IN SOUTHERN UTAH

by

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ABSTRACT

The existence of the Draper Formation of the Lake Bonneville Group in the northern part of the Bonneville basin is discredited by some workers. The formation, however, is well displayed in the southern part of the basin. Here, the Alpine, Bonneville, and Draper Formations, complete with intervening unconformities, are recognizable in the bluffs of the Sevier River between Leamington and Delta, Utah. The history of Lake Bonneville is, we believe, one of repeated filling and lowering during the Pleistocene. On its second major transgression, Lake Bonneville reached its highest level, overflowed, eroded its outlet, and paused at the Provo level. The lake level then fell at least 110 feet (33 m); deposits of the Bonneville stage were then eroded and gravels deposited on the Sevier River delta. Subsequently, the lake rose almost to the Provo level. During this cycle, lake sediments of the Draper Formation were deposited on top of the post-Bonneville Formation gravel and unconformity.

The Draper Formation (Morrison, 1965b) of the Lake Bonneville Group was named for exposures 8 miles (13 km) south of Salt Lake City, Utah, near the city of Draper, Utah. It originally was believed to have three tongues or members, each keyed to a separate rise and fall of Lake Bonneville. Later, it was determined that the oldest, and highest, member in the type area was not lacustrine (Scott, 1980). In the same paper, Scott proposed that Lake Bonneville rose and fell but one time and, therefore, was composed of a single formation, the Bonneville Formation. Scott did not mention the younger and topographically lower two members of the Draper; they had been temporarily exposed in excavations and presumably were covered by structures of urban development when Scott conducted his study.

Morrison's type locality for the Draper Formation, in the area of dispute, lies in the northern part of the Lake Bonneville basin; our area of study is in the southern part of the basin. The threshold between the northern and southern parts is about 4600 feet (1400 m) altitude; therefore, we believe that the geologic history of the two basins above this altitude should be similar. New evidence in support of the existence of the Draper Formation is extant in the bluffs of the Sevier River in the Leamington-Lynndyl-Delta area, Utah (Varnes and Van Horn, 1984).

The Sevier River was the major contributor of water and sediment to Lake Bonneville in southern Utah (figure 1). The delta formed by the Sevier in earliest Lake Bonneville time extended southwestward to the town of Delta from a point several miles up the Sevier Canyon east of Leamington. This delta, deeply incised and exposed by the present Sevier River, is composed principally of silt and clay of the Alpine Formation (figure 2). The Alpine unconformably overlies fluvial deposits on which occur the Cca horizon of a strongly developed and eroded soil. An unconformity overlying the Alpine is marked at various places either by terrestrial sand and gravel, pebbles and cobbles (some with coatings of algal tufa), the eroded Cca horizon of a moderately to strongly developed soil, or by erosion of beds of the Alpine Formation.

Overlying the post-Alpine unconformity is the white marl of Gilbert (1890), composed of sand, silt, clay, and marl beds of the Bonneville Formation. These beds have been recognized at different altitudes from as high as 5040 feet (1535 m), just below the Bonneville shoreline at Gilbert's classic section near Leamington, to as low as about 4680 feet (1425 m), near the Intermountain Power Plant west of Lynndyl. There being no evidence to the contrary, it is believed that the Bonneville Formation extended completely across the present valley of the Sevier River all the way to Delta. Following deposition of the Bonneville, the lake overflowed, rapidly cut its outlet to the Provo level (about 4810 feet (1465 m) in this area), and the newly exposed lake beds were eroded. The lake stabilized at the Provo level while additional Bonneville Formation was deposited, then receded slowly by evaporation to some unknown level. Erosion at this time cut a wide valley into the deposits between Leamington and Delta. The valley was cut completely through the Bonneville Formation and into the Alpine Formation to an altitude of at least 4700 feet (1430 m). The lowest part of this unconformity is marked by a fluvial sand and gravel deposit, informally named the gravel near Lynndyl. A finer grained fluvial deposit overlying the Lynndyl bears a weakly developed soil.

The lake again rose and covered the gravel near Lynndyl and the weak soil with as much as 10 feet (3 m) of lacustrine sediment of the Draper Formation. The deposit consists of grayish-tan to reddish-brown, thin-bedded to massive, coherent silt and clay, and thin layers of silty sand and sand. The

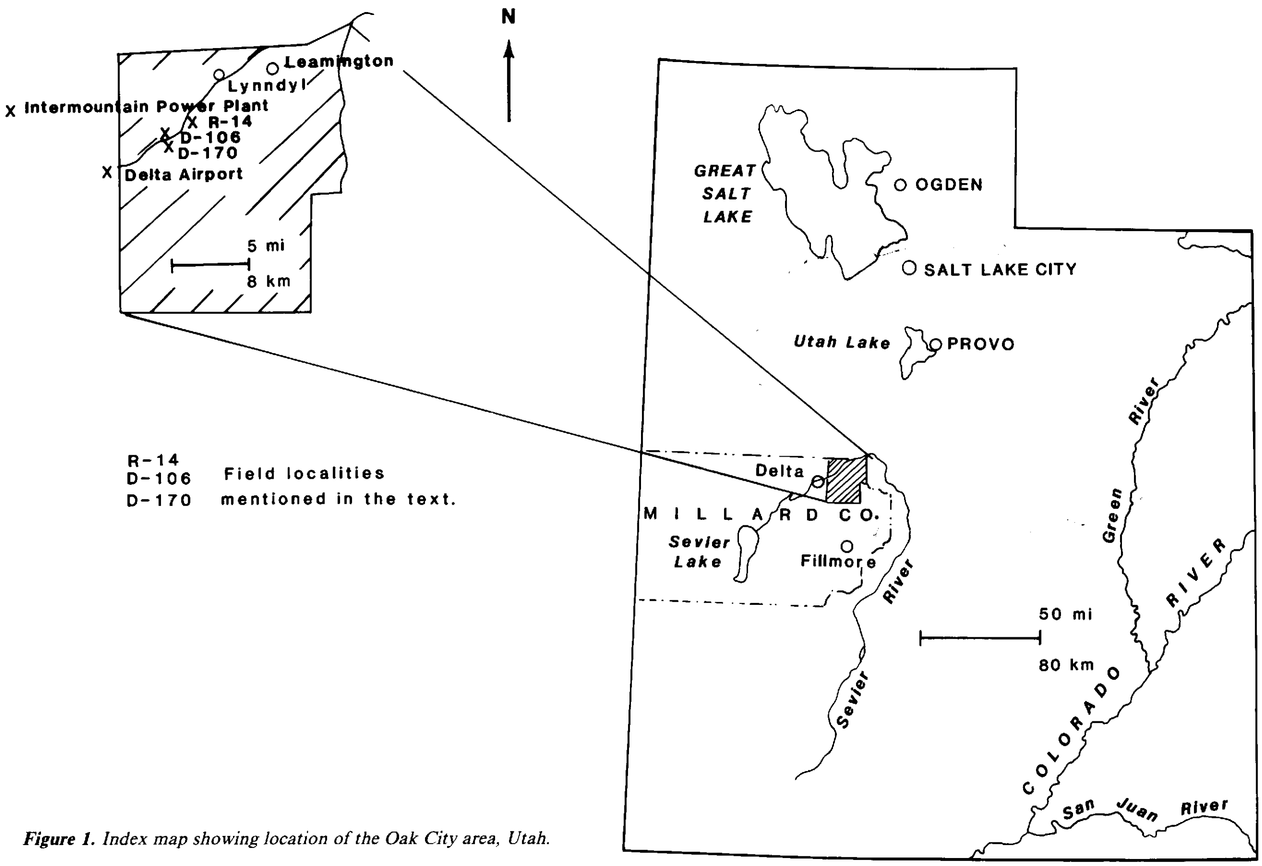

Figure 1. Index map showing location of the Oak City area, Utah.

beds range in thickness from 0.05 foot (2 cm) to 3 feet (90 cm). East of the Delta airport, the lacustrine origin of these beds is demonstrated by the gastropods *Valvata humeralis, Stagnicola elodes, Gyraulus parvus,* and *Heliosoma* or *Planorbella,* which were identified by Steven C. Good and Emmett Evanoff of the University of Colorado (oral communication, October 12, 1987). Evanoff and Good indicate that these forms lived in a lake that was probably large and relatively long-lasting. Evanoff and Good saw no terrestrial or fluvial forms in the collection. The lacustrine fossils were collected from deposits at station D-170 from an altitude of 4740 feet (1445 m) but we have seen similar deposits as high as 4780 feet (1455 m) at locality R-14. Gastropods from the Draper Formation collected 1 mile (1.6 km) southwest of the following measured section (table 1) gave an age of about 13,060 yr B.P. A weak, relict soil is present on the Draper Formation, which is locally much eroded by wind and overlain by modern dunes.

We conclude that after the stillstand at the Provo shoreline, Lake Bonneville receded to below 4700 feet (1430 m), then transgressed to at least 4740 feet (1445 m) and probably to 4780 feet (1455 m) while the Draper Formation was deposited.

Table 1

Section at station D-106; Draper Formation

(Measured August 28, 1957, by D.J. Varnes on the west bluff of Sevier River in the SE ¼, NW ¼, Sec. 8, T 16 S, R 5 W, Salt Lake Meridian)

	Thickness: Feet	Thickness: Centimeters
Draper Formation:		
9. Covered from top of bluff	0.5	15
8. Silt, coarse-grained; and clay, pale yellowish-brown to very pale orange, very crudely bedded; grayish-pink silt and clay "red bed" 0.2 ft (6 cm) thick about 0.2 ft (6 cm) above base	2.6	78
7. Silt, coarse-grained, very pale orange; thin-bedded with some crossbedding	0.8	24
6. Sand, fine-grained, grayish-orange; not coherent	0.4	12
5. Silt, coarse-grained, very pale orange; contains yellowish-brown limonite stains; moderately coherent	2.3	69
4. Silt, very coarse grained, very pale orange; contains several 0.05-ft-(2-cm) thick beds of very fine grained silt in upper half	2.4	72
3. Silty clay, pale yellowish-brown, slightly plastic; tastes salty	0.3	9
Gravel near Lynndyl:		
2. Pebbly sand, gray, not coherent; contains pebbles as much as 1 in. (3 cm) in diameter and some clay balls in upper half	8.7	261
Alpine Formation:		
1. Sand, interbedded coarse- and fine-grained in beds 1-6 in. (3 -15 cm), base covered	0.8	24

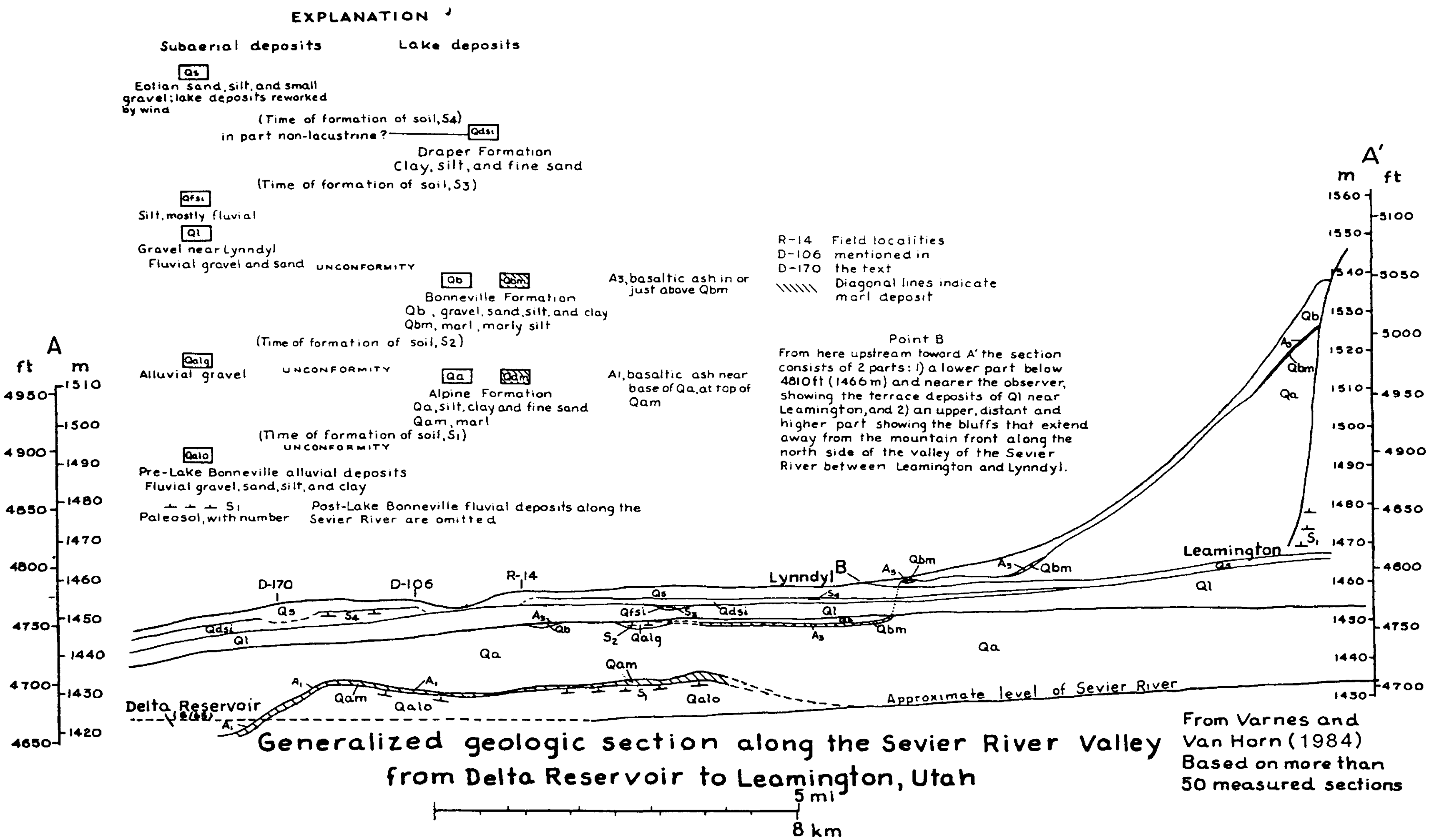

Figure 2. Generalized geologic section along the Sevier River Valley from Delta Reservoir to Leamington.

LINEAR MODEL OF THRESHOLD-CONTROLLED SHORELINES OF LAKE BONNEVILLE

by
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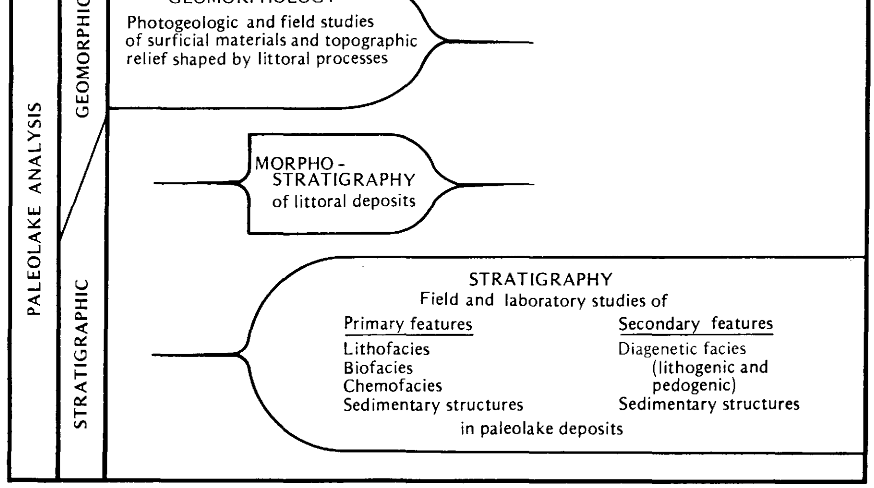
INTRODUCTION

The littoral morphostratigraphic record (figure 1) of the last deep-lake cycle in the Bonneville basin includes hundreds of localities with multiple shorelines. The variability of those shorelines can seem staggering--erosional and depositional, incipient and well developed, well preserved and poorly preserved, frequently transgressive and occasionally regressive, solitary shorelines and sets of shorelines, and so on. More importantly, the details of their patterns in space and time have often seemed baffling (e.g., Gilbert, 1890, p. 133).

Shorelines in pluvial lake basins generally form under closed-basin conditions. However, a special case prevailed during the culminating phase of the last deep-lake cycle in the Bonneville basin, when shorelines formed under open-basin, threshold-controlled conditions. Work by many persons in the past decade has contributed to substantial clarification of the spatiotemporal patterns of Lake Bonneville shorelines, especially of those that were controlled by the Zenda threshold in southern Idaho prior to the Bonneville Flood and by the Red Rock Pass threshold 2 miles (3 km) south of Zenda subsequent to the flood (Currey, 1982; Currey and others, 1983; Currey and others, 1984, fig. 1).

The purpose of this paper is to present the simplest model that describes what is now known about the spatiotemporal patterns of the threshold-controlled shorelines of Lake Bonneville. All of the main morphostratigraphic components of the Bonneville shoreline complex (BSC), which predates the flood, and the Provo shoreline complex (PSC), which postdates the flood, are modeled through the full range of their basin-wide hypsometry. Although all of the hypsometric and kinematic information in this initial iteration of model building has been

Figure 1. *The morphostratigraphy of littoral deposits, which combines the methods of geomorphology and stratigraphy, is a powerful tool in the analysis of paleolake records.*



hand fit under a simplifying assumption of linearity, the model provides useful insight into the interaction of hydrographic and isostatic kinematics within the basin during the period of intermittent threshold control. The model also provides insight into the geomorphic kinematics of the threshold region during threshold control.

THE LINEAR MODEL

As modeled here, the BSC comprises nine threshold-controlled stages and three sub-threshold stages of Lake Bonneville (figure 2). At all BSC localities except those near the basin perimeter, the B_5 shoreline is now the highest BSC component (figure 2, A-A' and B-B') because it formed under conditions of maximum hydro-isostatic subsidence relative to the Zenda threshold and subsequently has undergone the most isostatic rebound. Deflection of BSC shorelines relative to the threshold and to each other has been minimal near Zenda (figure 2, C-C'). Because the B_5 shoreline is the highest paleolake datum throughout the basin interior, it is a convenient starting point in studies of comparative shoreline hypsometry (tables 1 and 2). As modeled here, the PSC comprises seven threshold-controlled stages of Lake Bonneville (figure 2).

The numerical framework of the linear model is presented here in seven tables. Modern altitudes of the earlier BSC components, from earliest threshold control (B_0) to the highest shoreline (B_5), are in table 1a. Modern altitudes of the later BSC components, from B_5 through the Bonneville Flood, are in table 1b. Modern altitudes of PSC components, from the end of the flood (P_0) to the last stage of threshold control (P_6), are in table 1c. Differences in modern altitude between one BSC or PSC component and the next, in chronological order, are in tables 2a, 2b, and 2c. Tabulated modern altitudes are in feet above sea level (North American Vertical Datum of 1929) and altitude differences are in feet, rather than in metric units, because altitudes are expressed in feet on almost all of the largest scale U.S. Geological Survey topographic maps of the region. Estimated ages of the paleolake stages in tables 1 and 2 are shown in table 3.

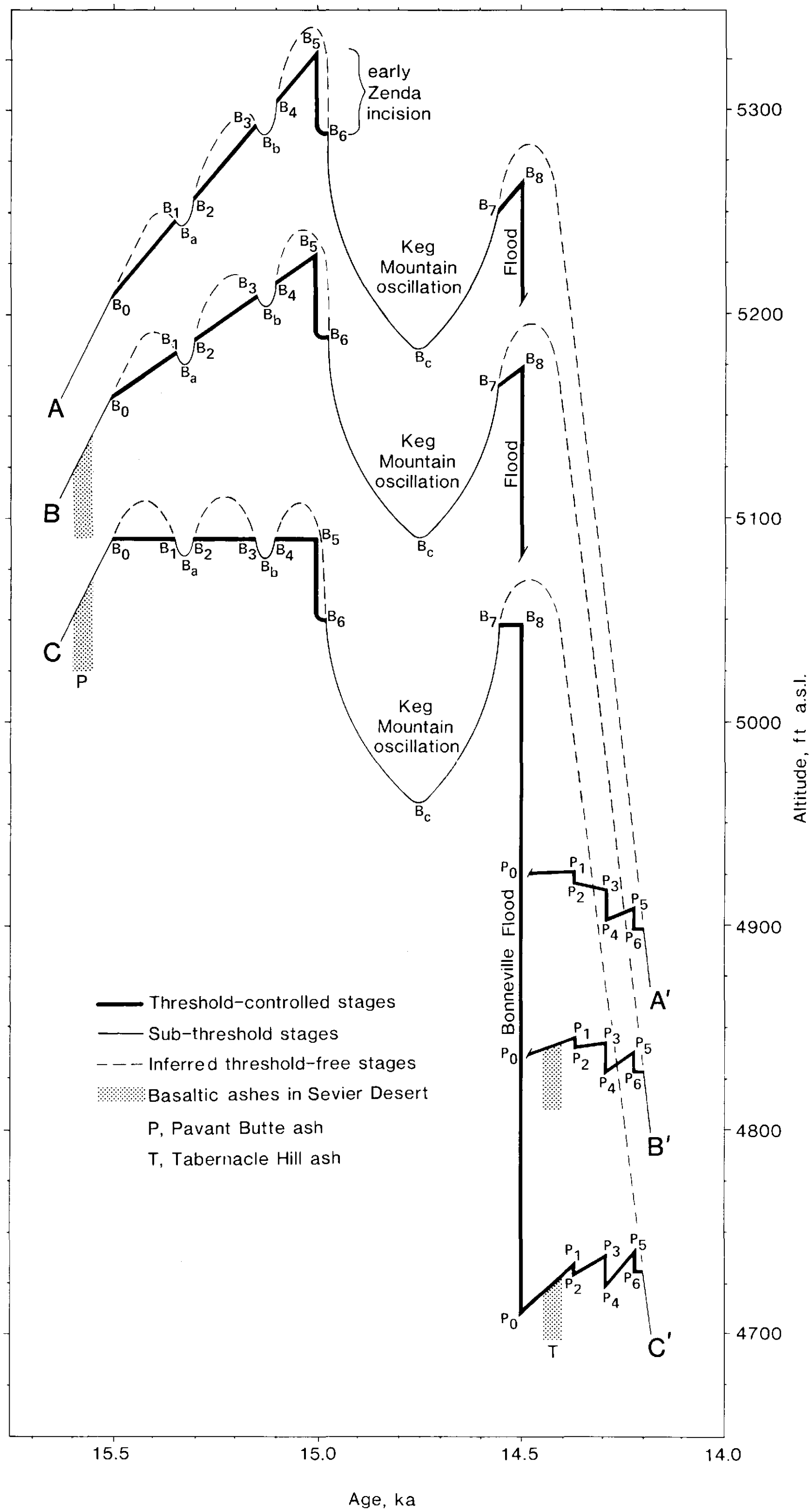


Figure 2. Threshold-controlled stages of Lake Bonneville in the region of maximum hydro-isostatic deflection, A-A'; at a typical basin-interior locality (Stockton Bar), B-B'; and in the threshold region and other basin perimeter regions of minimum deflection, C-C'. Stages B_0-B_8 comprise the Bonneville shoreline complex (BSC) and stages P_0-P_6 comprise the Provo shoreline complex (PSC).

Age estimates reflect the latest refinements of the chronology of Currey and Oviatt (1985), and have an average error that probably does not exceed 300 ^{14}C years.

Table 1a. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

Modern Altitudes in Feet

B_5	B_0	B_1	B_a	B_2	B_3	B_b	B_4	B_5
5330	5210	5246.0	5242.0	5258	5294.0	5290.0	5306	5330
5320	5205	5239.5		5251	5285.5		5297	5320
5310	5200	5233.0	5228.5	5244	5277.0	5272.5	5288	5310
5300	5195	5226.5		5237	5268.5		5279	5300
5290	5190	5220.0	5215.0	5230	5260.0	5255.0	5270	5290
5280	5185	5213.5		5223	5251.5		5261	5280
5270	5180	5207.0	5201.5	5216	5243.0	5237.5	5252	5270
5260	5175	5200.5		5209	5234.5		5243	5260
5250	5170	5194.0	5188.0	5202	5226.0	5220.0	5234	5250
5240	5165	5187.5		5195	5217.5		5225	5240
5230	5160	5181.0	5174.5	5188	5209.0	5202.5	5216	5230
5220	5155	5174.5		5181	5200.5		5207	5220
5210	5150	5168.0	5161.0	5174	5192.0	5185.0	5198	5210
5200	5145	5161.5		5167	5183.5		5189	5200
5190	5140	5155.0	5147.5	5160	5175.0	5167.5	5180	5190
5180	5135	5148.5		5153	5166.5		5171	5180
5170	5130	5142.0	5134.0	5146	5158.0	5150.0	5162	5170
5160	5125	5135.5		5139	5149.5		5153	5160
5150	5120	5129.0	5120.5	5132	5141.0	5132.5	5144	5150
5140	5115	5122.5		5125	5132.5		5135	5140
5130	5110	5116.0	5107.0	5118	5124.0	5115.0	5126	5130
5120	5105	5109.5		5111	5115.5		5117	5120
5110	5100	5103.0	5093.5	5104	5107.0	5097.5	5108	5110
5100	5095	5096.5		5097	5098.5		5099	5100
5090	5090	5090.0	5080.0	5090	5090.0	5080.0	5090	5090

Table 1b. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

Modern Altitudes in Feet

B_5	B_6	B_c	B_7	B_8	P_0
5330	5290	5184.5	5254.0	5266	4926
5320	5280		5245.5	5257	4917
5310	5270	5166.0	5237.0	5248	4908
5300	5260		5228.5	5239	4899
5290	5250	5147.5	5220.0	5230	4890
5280	5240		5211.5	5221	4881
5270	5230	5129.0	5203.0	5212	4872
5260	5220		5194.5	5203	4863
5250	5210	5110.5	5186.0	5194	4854
5240	5200		5177.5	5185	4845
5230	5190	5092.0	5169.0	5176	4836
5220	5180		5160.5	5167	4827
5210	5170	5073.5	5152.0	5158	4818
5200	5160		5143.5	5149	4809
5190	5150	5055.0	5135.0	5140	4800
5180	5140		5126.5	5131	4791
5170	5130	5036.5	5118.0	5122	4782
5160	5120		5109.5	5113	4773
5150	5110	5018.0	5101.0	5104	4764
5140	5100		5092.5	5095	4755
5130	5090	4999.5	5084.0	5086	4746
5120	5080		5075.5	5077	4737
5110	5070	4981.0	5067.0	5068	4728
5100	5060		5058.5	5059	4719
5090	5050	4962.5	5050.0	5050	4710

Refer to figure 2 for spatiotemporal relations of all alphanumeric stage symbols.

Table 1c. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

			Modern Altitudes in Feet				
B_5	P_0	P_1	P_2	P_3	P_4	P_5	P_6
5330	4926	4927	4922	4918.0	4903.0	4909	4899
5320	4917	4919	4914	4910.5	4895.5	4902	4892
5310	4908	4911	4906	4903.0	4888.0	4895	4885
5300	4899	4903	4898	4895.5	4880.5	4888	4878
5290	4890	4895	4890	4888.0	4873.0	4881	4871
5280	4881	4887	4882	4880.5	4865.5	4874	4864
5270	4872	4879	4874	4873.0	4858.0	4867	4857
5260	4863	4871	4866	4865.5	4850.5	4860	4850
5250	4854	4863	4858	4858.0	4843.0	4853	4843
5240	4845	4855	4850	4850.5	4835.5	4846	4836
5230	4836	4847	4842	4843.0	4828.0	4839	4829
5220	4827	4839	4834	4835.5	4820.5	4832	4822
5210	4818	4831	4826	4828.0	4813.0	4825	4815
5200	4809	4823	4818	4820.5	4805.5	4818	4808
5190	4800	4815	4810	4813.0	4798.0	4811	4801
5180	4791	4807	4802	4805.5	4790.5	4804	4794
5170	4782	4799	4794	4798.0	4783.0	4797	4787
5160	4773	4791	4786	4790.5	4775.5	4790	4780
5150	4767	4783	4778	4783.0	4768.0	4783	4773
5140	4755	4775	4770	4775.5	4760.5	4776	4766
5130	4746	4767	4762	4768.0	4753.0	4769	4759
5120	4737	4759	4754	4760.5	4745.5	4762	4752
5110	4728	4751	4746	4753.0	4738.0	4755	4745
5100	4719	4743	4738	4745.5	4730.5	4748	4738
5090	4710	4735	4730	4738.0	4723.0	4741	4731

Table 2a. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

Alt, Ft				Altitude Differences in Feet			
B_5	$B_1 - B_0$	$B_1 - B_a$	$B_2 - B_a$	$B_3 - B_2$	$B_3 - B_b$	$B_4 - B_b$	$B_5 - B_4$
5330	36.0	4.0	16.0	36.0	4.0	16.0	24
5320	34.5			34.5			23
5310	33.0	4.5	15.5	33.0	4.5	15.5	22
5300	31.5			31.5			21
5290	30.0	5.0	15.0	30.0	5.0	15.0	20
5280	28.5			28.5			19
5270	27.0	5.5	14.5	27.0	5.5	14.5	18
5260	25.5			25.5			17
5250	24.0	6.0	14.0	24.0	6.0	14.0	16
5240	22.5			22.5			15
5230	21.0	6.5	13.5	21.0	6.5	13.5	14
5220	19.5			19.5			13
5210	18.0	7.0	13.0	18.0	7.0	13.0	12
5200	16.5			16.5			11
5190	15.0	7.5	12.5	15.0	7.5	12.5	10
5180	13.5			13.5			9
5170	12.0	8.0	12.0	12.0	8.0	12.0	8
5160	10.5			10.5			7
5150	9.0	8.5	11.5	9.0	8.5	11.5	6
5140	7.5			7.5			5
5130	6.0	9.0	11.0	6.0	9.0	11.0	4
5120	4.5			4.5			3
5110	3.0	9.5	10.5	3.0	9.5	10.5	2
5100	1.5			1.5			1
5090	0.0	10.0	10.0	0.0	10.0	10.0	0

Refer to figure 2 for spatiotemporal relations of all alphanumeric stage symbols.

Table 2b. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

Alt, Ft	Altitude Differences in Feet				
B_5	$B_5 - B_6$	$B_6 - B_c$	$B_7 - B_c$	$B_8 - B_7$	$B_8 - P_0$
5330	40	105.5	69.5	12.0	340
5320	40			11.5	340
5310	40	104.0	71.0	11.0	340
5300	40			10.5	340
5290	40	102.5	72.5	10.0	340
5280	40			9.5	340
5270	40	101.0	74.0	9.0	340
5260	40			8.5	340
5250	40	99.5	75.5	8.0	340
5240	40			7.5	340
5230	40	98.0	77.0	7.0	340
5220	40			6.5	340
5210	40	96.5	78.5	6.0	340
5200	40			5.5	340
5190	40	95.0	80.0	5.0	340
5180	40			4.5	340
5170	40	93.5	81.5	4.0	340
5160	40			3.5	340
5150	40	92.0	83.0	3.0	340
5140	40			2.5	340
5130	40	90.5	84.5	2.0	340
5120	40			1.5	340
5110	40	89.0	86.0	1.0	340
5100	40			0.5	340
5090	40	87.5	87.5	0.0	340

Table 2c. *Linear model of basin-wide hypsometric relations among selected threshold-controlled stages of Lake Bonneville.*

Alt, Ft	Altitude Differences in Feet					
B_6	$P_1 - P_0$	$P_1 - P_2$	$P_3 - P_2$	$P_3 - P_4$	$P_5 - P_4$	$P_5 - P_6$
5330	1	5	-4.0	15	6.0	10
5320	2	5	-3.5	15	6.5	10
5310	3	5	-3.0	15	7.0	10
5300	4	5	-2.5	15	7.5	10
5290	5	5	-2.0	15	8.0	10
5280	6	5	-1.5	15	8.5	10
5270	7	5	-1.0	15	9.0	10
5260	8	5	-0.5	15	9.5	10
5250	9	5	-0.0	15	10.0	10
5240	10	5	0.5	15	10.5	10
5230	11	5	1.0	15	11.0	10
5220	12	5	1.5	15	11.5	10
5210	13	5	2.0	15	12.0	10
5200	14	5	2.5	15	12.5	10
5190	15	5	3.0	15	13.0	10
5180	16	5	3.5	15	13.5	10
5170	17	5	4.0	15	14.0	10
5160	18	5	4.5	15	14.5	10
5150	19	5	5.0	15	15.0	10
5140	20	5	5.5	15	15.5	10
5130	21	5	6.0	15	16.0	10
5120	22	5	6.5	15	16.5	10
5110	23	5	7.0	15	17.0	10
5100	24	5	7.5	15	17.5	10
5090	25	5	8.0	15	18.0	10

Refer to figure 2 for spatiotemporal relations of all alphanumeric stage symbols.

Table 3. *Model of chronometric relations among threshold-controlled stages of Lake Bonneville.*

Threshold-Controlled Stages	Sub-Threshold Stages	Estimated Age (^{14}C yr B.P.)
	pre-B	>15,500
B_0		15,500
B_1		15,350
	B_a	15,325
B_2		15,300
B_3		15,150
	B_b	15,125
B_4		15,100
B_5		15,000
B_6		14,975
	B_c	14,750
B_7		14,550
B_8		14,500
Bonneville Flood		14,500
P_0		14,500
P_1		14,360
P_2		14,360
P_3		14,290
P_4		14,290
P_5		14,220
P_6		14,220
	post-P	<14,200

Refer to figure 2 for spatiotemporal relations of all alphanumeric stage symbols.

THE MODEL SCENARIO

Morphostratigraphic evidence from numerous localities, including most of the sites that are listed in Currey (1982) and Currey and Oviatt (1985), constrains many of the attributes of the linear model presented here. Furthermore, the modeling process itself--the process of coherent synthesis that is not only consistent with observable field relations but is also internally consistent in its own basin-wide numerical structure--provides additional constraints. From the empirical constraints and the modeling constraints, a probable scenario of hydrographic, isostatic, and threshold geomorphic events can be postulated for the open-basin phase of Lake Bonneville history.

Lake Bonneville first became a threshold-controlled water body at B_0 (figure 2) and transgressed to B_1 at the local rate of hydro-isostatic subsidence, as the Zenda threshold remained essentially undissected. Hydroclimatic factors caused the lake to fall slightly below threshold control between B_1 and B_2, while rates of isostatic subsidence remained relatively constant. Threshold control resumed at B_2 and the lake transgressed to B_3, again at the local rate of isostatic subsidence and under the control of an essentially undissected threshold. Hydroclimatic factors again caused the lake to fall slightly below threshold control between B_3 and B_4, while isostatic subsidence continued. Threshold control resumed at B_4 and the lake transgressed to B_5, again at the local rate of isostatic subsidence and under the control of a still essentially undissected threshold. Incision of the Zenda threshold then became significant, initially causing the lake to regress about 40 feet to B_6 from its all-time high at B_5. Hydroclimatic factors caused the lake to fall many tens of feet below threshold control between B_6 and B_7; this sub-threshold cycle has been termed the Keg Mountain oscillation (Currey and others, 1983; see also day 3-stop 3). During the oscillation, hydro-isostatic deflection in the basin interior changed from subsidence to rebound and back to subsidence, with net deflection during the oscillation being rebound. Threshold control resumed at B_7 and the lake briefly transgressed to B_8, again at the local rate of isostatic subsidence but with negligible threshold incision. However, threshold incision soon resumed--catastrophically this time--and the resulting Bonneville Flood (Jarrett and Malde, 1987) caused rapid additional lowering of lacustrine base level by about 340 feet (104 m), from the Zenda threshold to the Red Rock Pass threshold. Total threshold lowering between B_5 and the end of the flood was about 380 feet (116 m); net isostatic rebound during that interval locally ranged from negligible near the perimeter of the basin to about 24 feet (7 m) in the region of greatest water depth (between 1200 and 1300 feet; 365-400 m).

Not only did the Bonneville Flood cause deep incision of the Zenda threshold, it also oversteepened the east flank of the adjacent Malad Range and thereby triggered recurrent land-sliding of major proportions. The earliest landslide occurred during the flood, briefly deflecting the axis of incision to the east. In the Red Rock Pass-Zenda area, landslide activity continued along a 5-mile (8 km) segment of the range front long after the earliest post-flood level of the lake stabilized at P_0. Landsliding steadily elevated the Red Rock Pass threshold about 25 feet (7.5 m), but the net transgression from P_0 to P_1 was less than that over most of the basin because local isostatic rebound partly counteracted the threshold rise. Then about 5 feet (1.5 m) of incision of the landslide at the Red Rock Pass threshold caused the lake to regress from P_1 to P_2. Continued landsliding steadily elevated the Red Rock Pass threshold about 8 feet (2.5 m), but the net transgression from P_2 to P_3 in the basin interior was slight because local isostatic rebound largely counteracted the threshold rise. Then about 15 feet (4.5 m) of incision of the landslide at Red Rock Pass caused the lake to regress from P_3 to P_4. Continued landsliding steadily elevated the Red Rock Pass threshold about 18 feet (5.5 m), but the net transgression from P_4 to P_5 was less than that over most of the basin because local isostatic rebound partly counteracted the threshold rise. Then about 10 feet (3 m) of incision of the landslide at Red Rock Pass caused the lake to regress from P_5 to P_6, shortly before the lake reverted to closed-basin conditions by finally regressing below its threshold.

IMPLICATIONS OF THE MODEL

The hypsometric attributes of the linear model presented here have several hydro-isostatic and geomorphic implications that should be noted.

1. About half of the total isostatic subsidence at any basin-interior locality occurred during threshold-controlled stages B_0 to B_5.
2. About 10 percent of the net isostatic rebound at any basin-interior locality occurred during the Keg Mountain oscillation, before the Bonneville Flood; about 90 percent occurred after the flood.
3. The Zenda threshold was incised twice, with about 10 percent of the total downcutting occurring non-catastrophically during a pre-Keg Mountain oscillation event and about 90 percent of the downcutting occurring catastrophically during the Bonneville Flood.
4. The Bonneville Flood incised the threshold topography to at least 70 feet (21 m) below modern grade at Red Rock Pass, where the flood channel has been partly back-filled by the Holocene alluvial fan of Marsh Creek. If the flood scoured a kolk into the bed of its channel, as seems probable, the lowest isolated hole scoured by the flood was substantially lower than 70 feet (21 m) below the modern level of the pass.
5. During PSC time, sustained landsliding caused about 50 feet (15 m) of total aggradation by mass movement at the Red Rock Pass threshold, and threshold-controlled discharge from Lake Bonneville caused about 30 feet (9 m) of total degradation by fluvial incision at that threshold, meaning that the local base level underwent a net rise of about 20 feet (6 m). Alluviation in response to that net rise may explain the prism of fine sand that seems to have been reworked from the Bear River delta in the northeast part of Cache Valley and deposited in the northwest part of the valley during PSC time.

The hypsometric and chronometric co-attributes of the linear model presented here imply several interesting kinematic relations.

1. The average rate of change of lake stage during the Keg Mountain oscillation, from the beginning of the regression to the end of the transgression, was about 0.4 feet (12 cm) per year.
2. The rate of isostatic subsidence during stages B_0 to B_5 varied basin wide, from an essentially negligible minimum to a maximum of about 0.24 feet (7.3 cm) per year.
3. The rate of isostatic rebound during stages P_0 to P_5 varied basin wide, from an essentially negligible minimum to a maximum of about 0.17 feet (5.2 cm) per year.

The linear model presented here is eminently testable by at least three lines of evidence--hypsometric, chronometric, and geomorphic. In its present form, the model suggests constraints that may be applied in current efforts to elucidate basic properties of the lithosphere and asthenosphere (e.g., Bills and May, 1987). With the refinements in hypsometric and chronometric inputs that are inevitable, and with inevitable refinements in its numerical structure to better reflect the reality of kinematic and kinetic nonlinearities, the model can evolve into an even more versatile and robust tool in the future.

ACKNOWLEDGMENTS

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THE HISTORY OF LAKE BONNEVILLE IN CACHE VALLEY, UTAH: UPDATING GILBERT'S OBSERVATIONS

by

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INTRODUCTION

Cache Valley is an elongate north-south-trending structural basin 48 miles (77 km) long and 7 to 13 miles (12-20 km) wide that straddles the Utah-Idaho border (figure 1). During the most recent high stand of Lake Bonneville the valley was occupied by what Gilbert (1890, p. 102) called "Cache Valley bay," the northernmost of the large appendages of Lake Bonneville. Geologic investigations of the lake's morphologic and stratigraphic record in Cache Valley subsequent to Gilbert's initial studies have lagged behind those of the central parts of the Bonneville basin, partly because of the valley's isolation and lack of good exposures. J.S. Williams produced a 1:62,500-scale surficial geologic map of the Utah portion of the valley (Williams, 1962) that utilized Lake Bonneville stratigraphic concepts as developed in the 1953-1965 period (for a discussion of the concepts of the "Hunt era", see the Introduction to this volume). Between 1962 and the early 1980s, only two 1:24,000-scale maps have depicted Quaternary deposits in more detail than Williams (Paradise quadrangle, Mullens and Izett, 1964; Cutler Dam quadrangle, Maw, 1968), but these retained the old stratigraphic framework and contributed no new dating control. Beginning in 1983, a series of new maps have been prepared with support from the U.S. Geological Survey National Earthquake Hazards Reduction Program and Utah Geological and Mineral Survey. The new products include a map of Quaternary geology of the Smithfield quadrangle (Lowe, 1987) and a 1:50,000-scale map of the surficial geology along the East Cache fault zone (McCalpin, 1987a). These maps use the stratigraphic model proposed by Scott and others (1983) and Currey and Oviatt (1985). Our recent investigations both confirm and extend some of Gilbert's early observations in three topical areas: 1) geomorphology and chronology of the Bonneville lake cycle in Cache Valley, 2) Quaternary activity on the East Cache fault, and 3) deformation of the Bonneville shoreline.

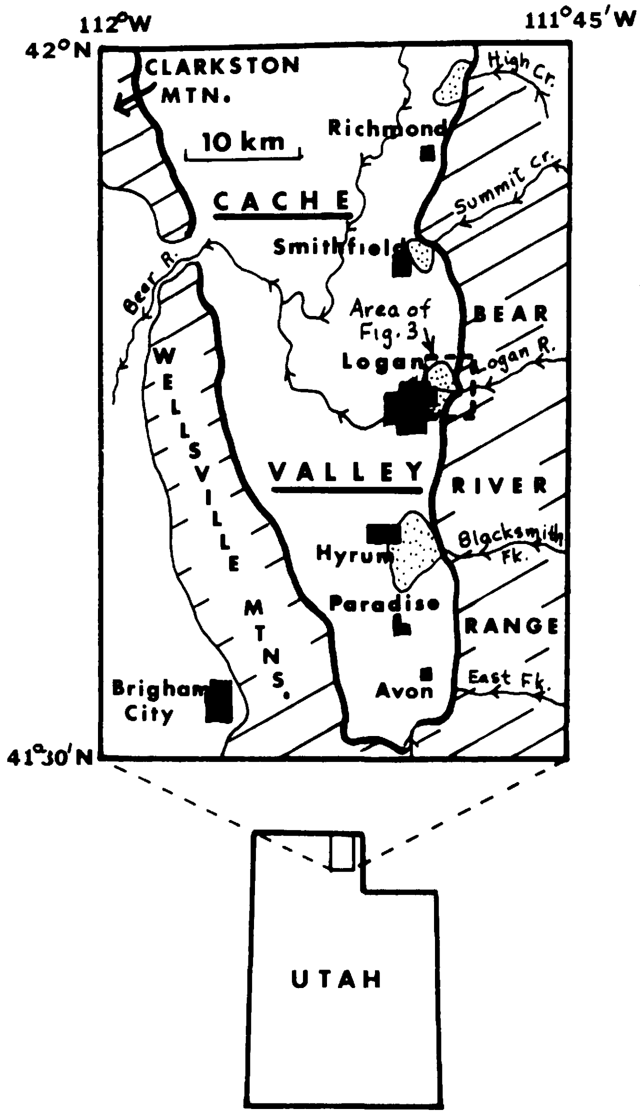


Figure 1. Map of Cache Valley, Utah, showing geographic locations and Provo-level deltas (fine stipple). Area covered by figure 3 is outlined by heavy dashes.

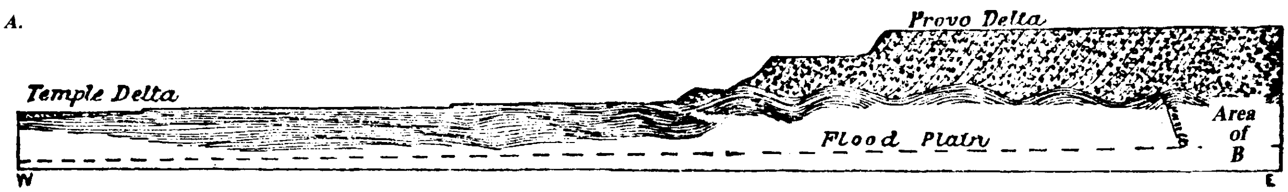
TABLE 1

Depth	Lithology	Interpretation
0-22m	cobble & pebble gravel	Provo level deltaic topset beds
22-32m	gravelly sand	Provo level foreset beds
32-51m	silt and fine sand	Bonneville level prodelta deposit
51-84m	clayey silt and organics	Basal prodelta deposit
84-88m	pebbles and boulders	Pre-Bonneville lake cycle, alluvial fan deposit

Location: Utah State University campus, Logan, UT. 32 feet (10 m) SE of the SE corner of the Old Main building. Source: Shannon and Wilson and Agbabian Associates, 1980.

Figure 2 A. Partial section of deltas at Logan, Utah, by I.C. Russell (from Gilbert, 1890, fig. 27, p. 162). Vertical scale greater than horizontal. The fault indicated by Russell is no longer exposed on the canal bank; it may be an antithetic feature to the fault shown in B. B. Sketch of fault on the north side of Logan River, about 650 feet (200 m) SE of the Education Building, Utah State University (from McCalpin, 1987a, fig. 9, p. 49). Units are informally numbered from 1 (oldest) to 13 (youngest). Units 1-12 are lacustrine sands (fine stipple) and pebble gravels (open circles) of the Bonneville lake cycle; unit 13 is Holocene colluvium that mantles the 35° slope. Fault zone shown by opposing arrows.

A.



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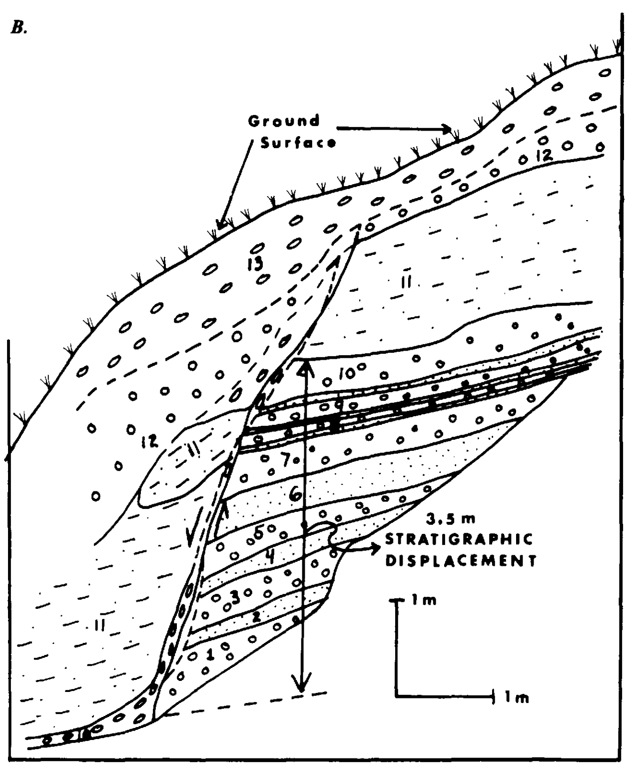

GEOMORPHOLOGY OF LAKE BONNEVILLE DEPOSITS

In describing "Cache Valley bay" Gilbert drew special attention to the most prominent landforms in the valley -- the Provo-level deltas at the mouths of the major canyons draining the Bear River Range (figure 1.). He called the Logan River delta "one of the most beautiful and symmetrical of all the deltas" (Gilbert, 1890, p. 159). These Provo-level deltas formed after the Bonneville shoreline was catastrophically lowered 330 feet (100 m) to the Provo shoreline. Prior to the drop, pluvial lake waters extended roughly 5 miles (8 km) up into Logan Canyon in a large embayment. Alluvium and glacial outwash from upper Logan Canyon glaciers were stored at the head of this embayment. With the sudden lowering of base level 330 feet (100 m) after the Bonneville Flood, the Logan River incised into this gravelly fill and retransported the debris down to the Provo shoreline at the range front. Here the gravel was deposited by braided streams across the top of an earlier lake bottom deposit of prodelta sands, silts, and clays. These finer sediments represent glacial flour which settled out of suspension when the Bonneville shoreline was occupied 5 miles (8 km) upvalley. Thus, the Provo-level gravels which define the morphology of these classic deltas are only a veneer of 65 to 100 feet (20-30 m) thick on a much more massive prodelta deposit. A well log through the Logan River delta illustrates the compound nature of that deposit (Table 1):

The prodelta deposits are often deformed by folds and diapir structures and, more rarely, by faults (Feth, 1955; figure 2). Such deformation could have been induced by: 1) rapid pore pressure changes during the Bonneville Flood, 2) rapid loading of the saturated silts by Provo-level deltaic gravels, or 3) earthquake shaking. Occasionally faults occur within the prodelta silts that do not extend through the overlying Provo deltaic gravels (figure 2). These faults could represent either seismogenic faults up to 1¼ miles (2 km) valleyward of the main range front fault, or could be secondary features accommodating lateral spreading within the delta.

Lake dessication after the fall to the Provo shoreline led to incision of each delta and construction of Holocene alluvial fans valleyward of the entrenched deltas. This geomorphology controlled pioneer settlement patterns in the valley; towns were built on the Holocene alluvial fans, with the wide Provo deltas serving as irrigated fields and orchards. Since the late 1950s suburban expansion onto the Provo deltas has rapidly displaced the previous agricultural uses.

POST-BONNEVILLE SURFACE FAULTING

Gilbert noted the active nature of the East Cache fault in 1890 (p. 351):

> *The eastern wall of the valley is an important mountain range (sic, the Bear River Range), whose bold western front has the topographic configuration of a worn fault cliff. At its base there are obscure indications of late movements, either during or just after the lake epoch, and at one point, near Logan, a post-lacustrine fault scarp crosses a delta of Provo date. The displacement is about six feet.*

Not surprisingly, his data have not been improved on significantly in the ensuing 100 years. Geologic mapping and trenching across these fault scarps near Logan by Woodward-Clyde Associates (Swan and others, 1983) and by McCalpin (1987a) have confirmed Gilbert's early estimates of fault displacement and timing (figure 3). The scarp across the "delta of Provo date" additionally extends across a post-Provo river terrace that is incised 40 feet (12 m) below the surface of the Provo delta. Peterson (1936) described the causative fault which was exposed in the 1930s in a fresh south-facing roadcut of U.S. Highway 89, and concluded that prodelta silts and sands were offset more than 16 feet (5 m), whereas the terrace gravels and ground surface were only offset 5 to 6 feet (1.5-1.8 m). If this displacement represents a typical faulting event, then 16 feet (5 m) could represent three faulting events. The earlier two events must post-date the deposition of prodelta silts (15-20 ka), but predate the deposition of the post-Provo-level terrace (12-13 ka?).

Such faulting after the Bonneville transgression has been demonstrated by trenching in two other localilties in Cache Valley. Immediately south of Green Canyon (northern part of figure 3), Swan and others (1983, p. 6) exposed a "20 m-wide zone containing 6 to 7 faults having down-to-the-west displacement" in their 1983 trenches. No data for displacement were given because the scarp parallels irregular Bonneville transgressive shorelines and the net offset cannot be simply reconstructed by surface profiling. However, multiple faulting events were inferred on the basis of faulted scarp-derived colluvium. This evidence suggests (1) that some low escarpments between the Bonneville- and Provo-level shorelines are actually old fault scarps, not transgressive shorelines as previously assumed, and (2) multiple surface faulting had to occur within the relatively narrow time interval between the initial Bonneville transgression (ca. 20 ka at this elevation) and abandonment of the Provo delta surface (13-14 ka).

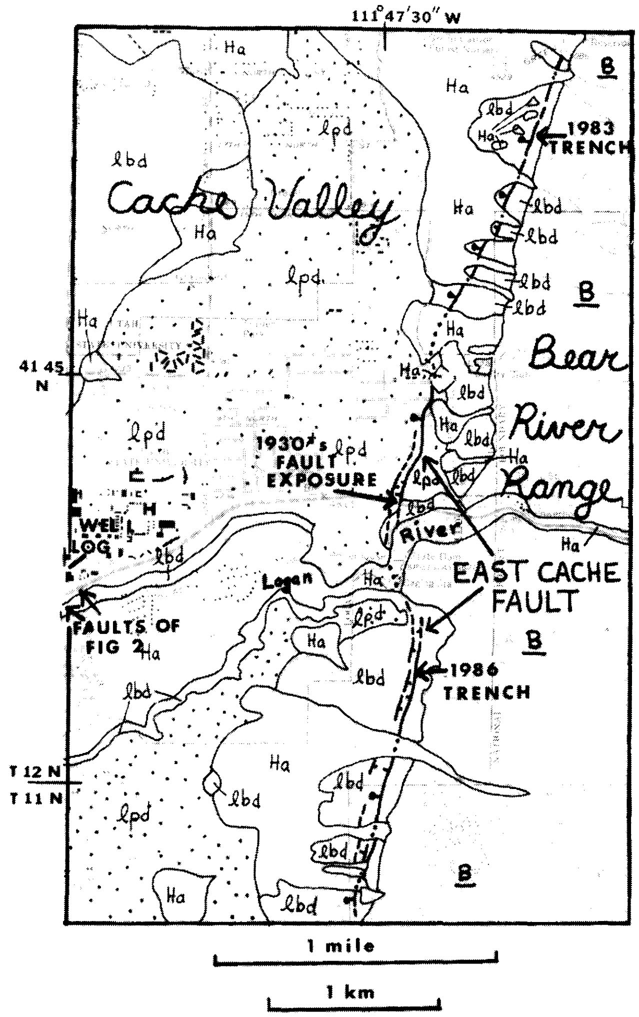

Figure 3. Geologic map of the area east of Logan, Utah, showing post-Bonneville lake-cycle fault scarps, trench sites, and natural exposures along the central segment of the East Cache fault zone. Map units: Ha, Holocene stream channel and fan alluvium; lpd, Provo-level delta (stippled pattern); lbd, Bonneville highstand gravel, sand, silt, and clay; B, Paleozoic bedrock, undifferentiated (from McCalpin, 1987a, fig. 6, p. 30).

A second trench south of the Logan River (figure 3) sheds some light on timing of fault events. The dissected remnants of a Bonneville shoreline embankment preserve a degraded scarp 20 feet (6 m) high that was trenched in October, 1986. In the trench, Bonneville highstand shoreline sands and gravels are faulted against a loess-derived colluvial wedge--faulted wedges are typically taken as proof of recurrent faulting (McCalpin, 1987b). Beneath the wedge is a tectonic melange of intact sand blocks interstratified with more lacustrine sands. Interpretation of our detailed trench log and 13 thermoluminescence (TL) dates on silty sediments yield the following sequence of events (figure 4):

1) Deposition of transgressive shoreline gravels of the Bonneville lake cycle from roughly 25 ka to 17 ka.
2) The first faulting event recorded at this trench ruptures saturated sediments on lake floor between 17.3 and 15 ka, momentarily creates a subaqueous scarp, which then slumps basinward.
3) Minor lacustrine deposition continues until 15 ka.
4) Water recedes over the scarp as the lake drops catastrophically to the Provo level, about 14 ka.
5) A loess-rich colluvial wedge accumulates at the base of the scarp, derived mainly from local silt blown off the newly-exposed floor of Cache Bay.
6) A soil develops on the colluvial wedge. This soil represents about 8 ka of soil formation based on pedogenic clay accumulation rates.
7) A second faulting event occurs, offsetting the colluvial wedge by at least 3 feet (1.0 m), probably between 4 and 7 ka.
8) Colluvium is deposited across and buries the new face of the fault scarp.
9) A modern soil develops across the scarp, requiring roughly 4 ka based on clay accumulation rates.

The combined data from the two trenches and one roadcut suggest three surface faulting events: 1) a subaqueous event late in the Bonneville transgression, between 20 (?) ka and 17 ka, but which is not exposed in either trench 2) another subaqueous event at about the Bonneville highstand, between 17 ka and 15 ka, and 3) a much later event in the Holocene, between 4 and 7 ka. The apparent temporal association of faulting with final filling of the lake basin recalls Gilbert's earlier speculations:

> *It is therefore theoretically conceivable that during the presence of the lake the process of faulting along the mountain bases was stimulated, and that after the evaporation of the water the process was correspondingly retarded (Gilbert, 1890, p. 357).*

Interestingly, other workers on the Wasatch fault zone have correlated increased faulting activity with the rapid regression during the Bonneville Flood (see day 3-stop 3). With the broad dating control available for the second event on the East Cache fault, it is possible that this faulting was triggered by the regression but that sediments were still saturated by shallow ground water. In summary, interpretation of recent data collected with state-of-the-art trenching and experimental dating techniques developed almost 100 years after Gilbert's observations suggest that he was correct in his early inference.

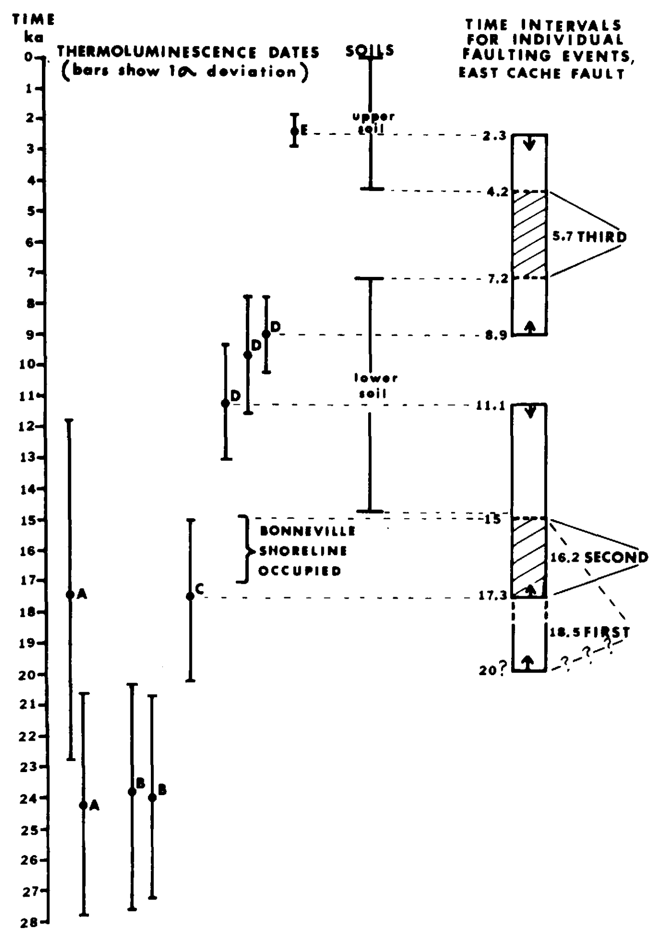

Figure 4. Timing of surface-faulting events recorded at the 1986 East Cache trench. Explanation for TL dates and samples: A, in-situ *Bonneville lacustrine sands, faulted by earlier event; B, blocks of lacustrine sand in a tectonic melange formed by the earlier event; C, lacustrine sands which grade into the melange laterally; D, loess-rich colluvium deposited between the earlier and later events; E, colluvium deposited after the later event. Length of soil development (in ka) was estimated from total pedogenic clay in the soil divided by clay accumulation rates from elsewhere on the Wasatch Front (see McCalpin, 1987a, p. 44-48 for detailed discussion). Windows at right indicate probable times of surface-faulting events as constrained by TL dates; cross-hatched portions depict more likely times further constrained by soil development intervals. Modified from McCalpin, 1987a, fig. 8, p. 45.*

DEFORMATION OF THE BONNEVILLE SHORELINE

Gilbert recognized that the present Bonneville, Provo, and other shorelines occur at different elevations throughout the basin, and he worked through a series of hypotheses to conclude that isostatic rebound had caused differential uplift. He stated

> *The principal recent displacements of the basin have been of the nature of broad, gentle undulations, not affecting the horizontality of the shorelines, so far as that is distinguishable by the eye. The region including each group of localities may properly be assumed to have risen or fallen in consequence of such earth movements without important internal change ...*
> (Gilbert, 1890, p. 140).

His assumption that individual mountain blocks rose or fell as units somewhat ignored the complicating effect of post-Bonneville surface-faulting, especially if such local uplift was restricted to individual segments of a range-front fault.

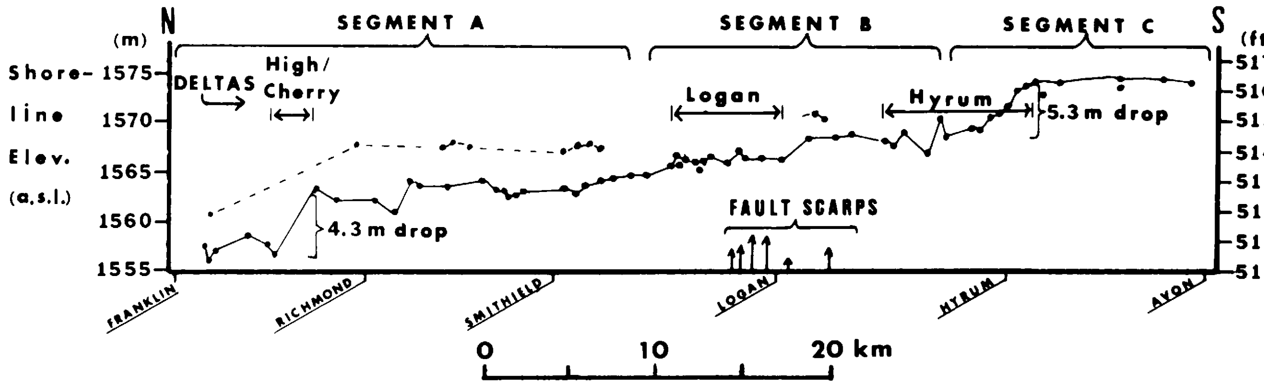

Figure 5. *Elevation profile of the highest Bonneville shoreline along the eastern side of Cache Valley from the Utah-Idaho border (N) to Avon, Utah (S). Elevation measurements (dots) have an uncertainty of ±4 feet (±1.2 m). Heights of arrows under "FAULT SCARPS" indicates the true-scale net surface offset of 4.6 feet (1.4 m) to 13.4 feet (4.1 m) measured from fault scarp profiles in the central segment. Names of towns are shown at base of horizontal axis. Solid line connects erosional shorelines, dashed lines connect bay-mouth bars. Modified from McCalpin, 1987a, fig. 11, p. 65.*

In Cache Valley, fault scarps younger than the Bonneville lake cycle are only present on the central 6 miles (10 km) of an active-appearing range front fault that is over 35 miles (55 km) long. Accordingly, the fault has been divided into "segments" of 19 miles (26 km), 9 miles (15 km), and 8 miles (14 km) from north to south based on recency of rupture (McCalpin, 1987a). The Bonneville shoreline exists as a paleo-datum plane common to all three fault segments and could be expected to record any inter-segment variations in post—Bonneville shoreline uplift along the range front. So, despite Gilbert's assumption that mountains act as coherent blocks, evidence of surface faulting suggests that we should find the Bonneville shoreline as much as 13 feet (4 m) higher today in the central segment than in the end segments, if fault scarps adequately portrayed all post—Bonneville tectonic movements.

To test this hypothesis, the elevation of the Bonneville Shoreline was measured to an accuracy of ± 4 feet (±1.2 m) by theodolite and electronic distance meter at 82 locations along a 33.5 mile (54 km) distance at the base of the Bear River Range. Determination of the highest paleo-water level of the Bonneville lake cycle from degraded 15 ka shoreline features is not simple, as Gilbert illustrated (Gilbert, 1890, p. 122-125), but corrections were made for a variety of complicating factors (see McCalpin, 1987a, for full discussion of methods). The shoreline elevations shown in figure 5 exhibit a general decrease from 5168 feet (1575 m) elevation at the southern end, to 5109 feet (1557 m) elevation at the northern end of the valley. This change in elevation reflects a slope of 0.33 m/km, which is low because the traverse is not perpendiucular to regional contours of isostatic rebound (see Crittenden, 1963). However, the 59 feet (18 m) of drop in elevation is not uniformly distributed along the traverse—60 percent of it occurs in two discrete drops of 14 feet (4.3 m) and 17 feet (5.3 m) (figure 5). At both locations the shoreline has similar geomorphology across the anomaly, ruling out obvious geomorphic causes such as superelevation. If the elevation anomalies were the result of differential tectonic movement across segment boundaries, we would expect the shoreline to be roughly 13 feet (4 m) higher in the central segment than on the end segments -- clearly this is not the case. In fact, there is no detectable difference in shoreline elevation as it traverses the short area of post-Bonneville fault scarps. What then is the cause of the anomalies?

The two large anomalies are coincident with two thick Provo-level deltas that, when deposited, added a rapid depositional load on the lake floor in addition to the weight of the lake water. The deltas have an estimated original mass (excluding Holocene stream entrenchment) of 1 x 10^{12} to 4 x 10^{12} kg. This mass can be modelled as a point load applied to a thin, brittle crust overlying a more ductile half space. A simple bending-beam analogy predicts local crustal subsidence of 11.5 to 23 feet (3.5-7.0 m) (similar to that observed) if crustal thickness of 4.4 miles (7 km), Young's modulus of 1 x 10^{10} Nm, and Poisson's ratio of 0.1 are assumed. This thickness of brittle crust is near the minimum estimate of 5 miles (8 km) proposed by Arabasz and others (1987) based on seismologic evidence for the brittle/ductile transition.

One problem with this model is that no shoreline deflection occurs near the Logan River delta, which has enough mass to cause 11.5 feet (3.5 m) of calculated subsidence. Coincidentally, the only post-Bonneville fault scarps in the valley separate this delta (a load placed on the downthrown fault block) from the Bonneville shoreline (which is here carved on the upthrown fault block). These relations suggest that post-15 ka surface faulting events have effectively decoupled the loaded downthrown block from the upthrown block on which the shoreline is preserved. In the northern and southern fault segments no such faulting and decoupling has occurred, and loads applied to the downthrown (valley) block have also been transmitted to the near edge of the upthrown block (mountain front). It appears that the shoreline elevation curve of figure 5 is the result of localized depositional loads being superimposed on the regional-scale isostatic rebound.

SUMMARY

Work in the past 5 years has confirmed and extended some of Gilbert's early observations on Lake Bonneville in Cache Valley. Deltas at the Provo level which built out rapidly onto the lake floor following the Bonneville Flood, caused soft-sediment deformation, perhaps induced localized faulting, and loaded the valley floor so heavily as to warp adjacent shorelines. The young faulting events deduced by Gilbert have been dated at between 15 and 17.3 ka and between 4 and 7 ka, based on trenching and TL dating. In addition, these faulting events on the central segment may have decoupled the mountain block from local depositional loads on the valley block, and even disturbed the block's reponse to regional isostatic rebound. Certainly more work remains to be done in confirming Gilbert's many fruitful hypotheses in the Bonneville basin.

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