Geologic hazards of the Wasatch Front, Utah

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ABSTRACT

The Wasatch Front, Utah’s population center, faces threats from several geologic hazards. These range from hazards that occur somewhere in the region almost every year, such as landslides and debris flows, to potentially catastrophic but infrequent hazards resulting from large earthquakes along the Wasatch fault zone. Study of these hazards is an ongoing process, and recent related research will be discussed on this field trip. We will observe an active landslide in Salt Lake City and hazard-reduction techniques implemented following fire-related debris flows near Farmington. We will examine evidence of recent flooding from increased levels of Great Salt Lake and discuss flooding hazards posed by the lake. We will observe the effects of large prehistoric earthquakes along the Wasatch fault zone, the longest active, normal-slip fault zone in the United States, and will discuss paleoearthquakes of the fault zone and the potential for earthquake ground shaking in the Salt Lake Valley. We will also examine ongoing efforts to seismically retrofit the Utah State Capitol to withstand strong earthquake ground shaking while preserving the historical integrity of its architecture.

Keywords: geologic hazards, landslide, earthquake, debris flow, flooding, Wasatch fault zone, Wasatch Front.

INTRODUCTION

The Wasatch Range extends more than 100 mi (160 km) through north-central Utah. The bulk of Utah’s population resides in rapidly growing metropolitan areas along the range front, and the Wasatch Front is subject to a variety of geologic hazards due to a unique combination of geologic, topographic, and climatic conditions. The Wasatch Front occupies a series of north-trending valleys at the foot of the western slope of the Wasatch Range. The
mountains rise steeply as much as 7100 ft (2200 m) above valley floors, reaching elevations near 12,000 ft (3700 m) above sea level. This impressive relief is the result of ongoing displacement along the Wasatch fault zone. The fault zone separates the Basin and Range Province to the west from the middle Rocky Mountains to the east and is a major intraplate tectonic boundary. The Wasatch fault zone is the longest active normal-slip fault zone in the United States and one of several fault zones in the region considered capable of producing large (M > 7) earthquakes. These earthquakes would generate surface fault rupture and strong ground shaking, perhaps accompanied by seismically induced liquefaction and landslides. The central five of the 10 independent segments of the Wasatch fault zone are considered the most active, having produced at least 16 large earthquakes in the past 6000 yr (McCalpin and Nishenko, 1996). Large earthquakes on the central five segments recur, on average, about every 350 yr, with the last occurring ca. 620 yr ago.

Great Salt Lake, a remnant of the much larger Pleistocene Lake Bonneville, forms the western boundary of the northern Wasatch Front. The well-established chronology of the four major shorelines of the Bonneville Lake cycle is often used to determine the age of movement for landslides and surface faulting along the Wasatch Front, and Lake Bonneville deposits and shorelines have a profound influence on the distribution of Wasatch Front geologic hazards. For example, loose, saturated Bonneville sands are potentially liquefiable, and steep cliffs near the erosion platforms of Bonneville shorelines increase the potential for slope failures in lake sediment. Thick deposits of soft, fine-grained lacustrine sediment from Lake Bonneville and older Pleistocene deepwater lakes that inundated the valleys could amplify earthquake ground motions. Post-Bonneville Great Salt Lake, occupying a closed basin within the internally draining Great Basin, is subject to climate-induced fluctuations. Rising lake levels between 1983 and 1987 caused flooding in areas along and near the gently sloping lake shores.

In the winter, frontal storms traveling east from the Pacific Ocean encounter the Wasatch Range and produce heavy snowfall in the mountains. Snow avalanches are common and present a significant, widespread hazard. Freeze-thaw cycles in steep exposures of fractured rock produce rock falls. Rapid melting of a lingering snowpack periodically results in slope failures, debris flows, and stream and alluvial fan flooding. Convective storms in the spring and late summer also contribute to these hazards.

This field trip (Fig. 1) provides an opportunity to observe and discuss several of the most significant types of geologic hazards of the Wasatch Front and examine the results of recent and ongoing related research. Field trip topics include (1) slow, active landsliding near the mouth of City Creek Canyon in Salt Lake City; (2) site conditions, earthquake ground shaking, and the Utah State Capitol seismic retrofit; (3) prehistoric liquefaction-induced landsliding in lacustrine sediments near Farmington; (4) debris-flood and debris-flow hazard reduction techniques on alluvial fans at the base of the Wasatch Range in Davis County; (5) Pleistocene Lake Bonneville shorelines and evidence of historical Great Salt Lake flooding on Antelope Island; (6) faulted Pleistocene (?) deposits in western Salt Lake Valley; and (7) surface fault rupture and paleoseismology on the active Salt Lake City segment of the Wasatch fault zone.

FIELD TRIP

Directions to Stop 1

From the Salt Palace Convention Center at 100 South West Temple in Salt Lake City (Fig. 1), proceed south on West Temple. Turn left on 200 South and after two blocks turn left on State Street. Proceed north on State Street ~0.3 mi (0.5 km), and turn right on South Temple, continuing ~0.3 mi (0.5 km). Turn left on B Street and proceed ~0.8 mi (1.3 km) to 11th Ave. Continue straight ahead, driving north on East Bonneville Boulevard ~0.8 mi (1.3 km) and park before crossing City Creek.

Stop 1—East Capitol Boulevard–City Creek Landslide, Salt Lake City

The East Capitol Boulevard–City Creek (CBCC) landslide is possibly the best example of a recurrently active landslide...
along the Wasatch Front. The landslide predates the earliest (1937) aerial photographs of the Salt Lake City area (Van Horn et al., 1972). Movement occurred in five of the seven years between 1998 and 2004 (inclusive). About 10.8 ft (3.3 m) of movement occurred in a single year (2002), surprisingly the driest calendar year of a drought that lasted between 1999 and 2004 (Fig. 2).

Retrogressive enlargement of the landslide in 1998 damaged parts of a backyard and threatened a house above the western part of the slide. A drilled pier (caisson) wall was installed by early 1999 to protect the remainder of the property. Offset on the main scarp subsequently continued, requiring stabilization of the scarp face with soil nails and shotcrete in 2004. Minor offset on the main scarp postdates the shotcrete application. At the northern end of the slide, continued offset of the main scarp oversteepens the slope and threatens a tennis court. Progressive enlargement of the landslide, which began in 1999 with the formation of a new frontal toe thrust, and downslope movement of slide debris have resulted in encroachment on an inlet structure that drains a perennial creek that flows along the left (SE) flank of the slide into City Creek.

Research by the Utah Geological Survey (UGS) on landslide movement and precipitation (Ashland, 2003) defined a method for predicting movement of recurrently active landslides based on recognition of cumulative instability-threshold precipitation levels. This method tracks both antecedent precipitation and conditions at the onset of the snowmelt that overlap with triggering of most northern Utah landslides.

**Geology**

The CBCC landslide is on a southeast-facing slope above a tributary drainage that flows into City Creek. The drainage was once ephemeral (Dames & Moore, 1979), but now flows year round. Late Pleistocene Lake Bonneville fine-grained sediments underlie the head and crown of the landslide (Dames & Moore, 1979, 1981; Personius and Scott, 1992). Exposures in the main scarp consist mostly of weakly laminated silt. Test pits and boreholes in the upper part and the crown area of the landslide indicate that soils consist primarily of interbedded silt, silty sand, sand, and minor gravel (Dames & Moore, 1979, 1981). The lacustrine sediments in turn overlie soils derived from Tertiary (Paleogene) sedimentary and volcanic rocks (Personius and Scott, 1992). Locally, coarse-grained fill overlies natural soils near the crest of the slope and head of the slide.

**Landslide Description**

The landslide is funnel shaped in plan view, narrowing in width downslope. At its head, the landslide is nearly 400 ft (122 m) wide, but it is only ~45 ft (14 m) wide at the toe. In 1979, the landslide was ~480 ft (146 m) long, ~360 ft (110 m) wide at the head, and ~160 ft (49 m) wide at its lower part upslope of the toe. By 1998, the landslide area was ~19,000 yd² (16,000 m²) with an estimated volume between 130,000 and 240,000 yd³ (99,000–148,000 m³). Profiling of the landslide in 1999 indicated that it was ~570–580 ft (174–177 m) long, or ~90–100 ft (28–31 m) longer than in 1979. The difference in length suggests an average annual rate of stretching of ~4.5–5 ft/yr (1.4–1.5 m/yr) during the intervening 20 yr. Survey data document ~29.5 ft (9 m) of stretching in the upper part of the landslide between May 1987 and April 2001.

**Movement History**

Between 1998 and 2004, the landslide moved in all but two years (2000 and 2003). Total movement of the landslide toe between June 1998 and December 2004 exceeded 23.6 ft (7.2 m). Movement measurements by the UGS at the toe of the landslide began on 5 June 1998. Although complete movement data are unavailable for 1998, the landslide moved a little over 8 ft (2.4 m) between 5 June 1998, and 5 June 1999 (Fig. 2). Movement typically triggers in early March and suspends between late May and early July (Ashland, 2002) (see inset in Fig. 2). Wetter than normal conditions in late spring or summer can lengthen the duration of movement or cause reactivation of the slide in the latter part of the year.

**Instability-Threshold Precipitation Level**

Movement of many landslides in northern Utah is triggered by a transient rise in groundwater levels or accumulation of ephemeral perched groundwater associated with the snowmelt. Groundwater level monitoring since 1999 has documented that natural seasonal peak groundwater levels are significantly reduced by the lack of a late winter snowpack (Ashland et al., 2005).

Cumulative precipitation is a reasonable basis for estimating the fluctuation in landslide groundwater levels. Ashland (2003) documented that groundwater levels along the Wasatch Front rise during extended long-term periods of excess pre-
cipitation. In addition, Ashland (2003) showed a correlation between cumulative precipitation and landslide movement suggesting that groundwater levels fluctuate with the cumulative precipitation budget (excess versus deficit precipitation conditions) of a site.

Figure 3 shows mean cumulative-precipitation curves for active and inactive periods between 1998 and 2004 that define the limits on the instability-threshold precipitation levels for the CBCC landslide. The instability-threshold precipitation levels fall between the two curves and are inferred to be near the normal cumulative-precipitation curve. The shaded area indicates the critical period for landslide movement (typically March through May in northern Utah). Infiltration of cumulative excess precipitation prior to and during the critical period causes a rise in groundwater levels relative to the previous year.

The exceptionally low instability-threshold precipitation levels of the CBCC landslide suggest that human modifications of the hillslope have significantly reduced the natural stability of the slide. The most important change that impacts slope stability may be a long-term rise in the base groundwater level due to infiltration of landscape irrigation water (lawn watering) in the residential subdivision directly upslope of the landslide and redirected runoff from the residential lots onto the slide. As a result of the elevated base groundwater levels, the transient snowmelt-induced groundwater-level rise needed to reach an instability-threshold level and trigger movement has likely decreased in magnitude in the past few decades, so that it occurs even in years with near-normal cumulative precipitation.

Table 1 shows that a prerequisite for renewed landslide movement of the CBCC landslide is cumulative excess precipitation sometime during the critical period. In the only two years in which the landslide remained dormant (2000 and 2003), cumulative deficit conditions existed during the entire critical period. In 1999, cumulative-excess precipitation conditions occurred only in May. Most (~90%) of the movement at the toe of the landslide in 1999 occurred in May and early June, during the transition from cumulative-deficit to cumulative-excess precipitation conditions.

Total movement at the toe of the landslide shows a roughly inverse correlation with the maximum cumulative excess precipitation during the critical period. The largest measured annual displacements at the toe occurred at the end of successive years in which excess precipitation existed in the critical period (1999 and 2001). The cumulative effects of successive wetter-than-normal years may have more influence on groundwater levels in a landslide and total annual displacement than the excess precipitation in a single year (such as in the 2003–2004 landslide water year). In addition, other factors such as landslide geometry and boundary conditions may have an increasing influence on the magnitude of total annual displacements, particularly near the end of a period of prolonged movement (1998–2004).

### Table 1. Summary of Cumulative-Precipitation Conditions Necessary to Trigger Renewed Movement

<table>
<thead>
<tr>
<th>Months</th>
<th>Cumulative-precipitation budget (inches)</th>
<th>Landslide water year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept.–Feb.</td>
<td>-0.18</td>
<td>0.94</td>
</tr>
<tr>
<td>Sept.–Mar.</td>
<td>-1.29</td>
<td>0.58</td>
</tr>
<tr>
<td>Sept.–Apr.</td>
<td>-0.32</td>
<td>0.92</td>
</tr>
<tr>
<td>Sept.–May</td>
<td>-3.38</td>
<td>-0.66</td>
</tr>
<tr>
<td>Active?</td>
<td>Yes</td>
<td>No</td>
</tr>
</tbody>
</table>

**Note:** Bold indicates cumulative excess conditions; italics indicate near normal conditions; normal print indicates cumulative deficit conditions.
Directions to Stop 2

Cross City Creek and proceed south on West Bonneville Blvd for ~0.8 mi (1.3 km). Turn right on 500 North, continue for ~0.3 mi (0.5 km), and turn left on Columbus Street. After ~0.2 mi (0.3 km), turn left into the State Capitol parking lot.

Stop 2—Site Conditions, Earthquake Ground Shaking, and the Utah State Capitol Seismic Retrofit

Site Conditions and Earthquake Ground Shaking in Salt Lake Valley

Unconsolidated surficial Quaternary deposits in Salt Lake Valley were mostly derived from late Pleistocene Lake Bonneville, but surficial and underlying deposits also include pre-Bonneville alluvial fan deposits, late Pleistocene (and older?) glacial deposits (till and outwash), and Holocene stream alluvium, alluvial fan deposits, and lacustrine and deltaic deposits. Sediments on the northeast margin of Salt Lake Valley occupy the footwall of the active Wasatch fault zone and overlie shallow, older (pre-Bonneville) semi-consolidated valley fill sediments, and/or weathered Tertiary or older rock. Unconsolidated deposits on the west side of Salt Lake Valley possibly overlie a relatively shallow rock bench that extends several miles from the base of the Oquirrh Mountains, where Holocene faulting is absent.

Site conditions reflect the properties of these near-surface geologic materials and can have a significant impact on earthquake ground shaking. Analytical (Martin and Dobry, 1994) and empirical ground motion studies (Borcherdt, 1994) suggest site conditions can be reasonably characterized using the average shear-wave velocity in the upper 100 ft (30 m) (Vs30). The International Building Code (IBC), adopted in Utah in 2002 and updated with a 2003 edition (International Code Council, 2002), uses Vs30 to group sites into five broad site classes that are assigned factors for use in estimating earthquake loads in the design of engineered buildings and structures. The five site classes are designated A through E, and are respectively named hard rock, rock, very dense soil and soft rock, stiff soil profile, and soft soil profile; a sixth IBC site class, F, includes soil that poses engineering problems for building foundations and is not characterized by Vs30.

Site conditions in Salt Lake Valley were mapped by combining 30 new shear-wave velocity profiles with the areal limited, existing shear-wave velocity profile data and previous mapping of Ashland and McDonald (2003). The existing site conditions map was simplified by redefining some previously mapped unit and subunit boundaries and characterized all four previously mapped Quaternary (unconsolidated) site condition units. Some shear-wave velocity profiles encountered buried rock, providing better characterization of Vs30 in the three previously mapped rock site condition units in the mountain areas bordering Salt Lake Valley (Ashland and McDonald, 2003) and of shallow impedance contrasts along local valley margins.

Figure 4 shows the newly revised site conditions map for Quaternary units in Salt Lake Valley, and Table 2 summarizes Vs30 in these units. The Vs30 statistics in Table 2 allow for estimating the potential variation in Vs30 at a site. IBC site classes in Salt Lake Valley Quaternary deposits range from site class E to site class C. A boundary between IBC site classes falls within the range of Vs30 in all but one of the Quaternary site conditions units.

The Utah State Capitol lies within site conditions unit Q02 (Fig. 4), which includes Vs30 values characteristic of IBC site classes D and C. IBC site coefficients for these site classes indicate the potential for significant amplification of strong earthquake ground motions, and such motions may be particularly strong because the Capitol lies near the Wasatch fault zone.

Utah State Capitol Seismic Retrofit

In 1888, Salt Lake City donated land to the State of Utah for the construction of the State Capitol at its present site. Construction began in December 1912, and its cornerstone was set in 1914. The legislature relocated to the new Capitol in February 1915 and in September 2004, the building was closed for seismic renovations (Cooper Roberts Simonson Architecture, 2000).

When constructed, the Utah State Capitol was one of the first concrete frame buildings west of the Mississippi River. While this technological advancement was new and exciting, it did not replace many of the common construction practices of the
day. Those techniques did not account for the strong earthquake ground shaking that recent studies indicate may occur along the Wasatch Front.

The exterior of the building was made of large, cut granitic stone from Little Cottonwood Canyon, ~20 mi (30 km) south of the Capitol. Rather than anchoring the stone to the frame, as is current practice, the stone blocks were stacked one on top of another, as a load-bearing masonry wall would have been built at the time. Use of a concrete frame was a new technique and the need for reinforcing bars was not well understood. Some steel was added to the frame to provide additional strength, but sometimes the steel was used only as a form-tie element to hold the formwork in place.

Despite older construction techniques, much of the Capitol was quite strong and able to withstand significant vertical force. However, studies have indicated that when the structure is subjected to a lateral load of a large M 7.0 earthquake, the building may fail and collapse at the parapet level.

An important architectural feature of the Utah State Capitol, the dome-and-drum assembly on top of the building, was in very poor condition. Engineers estimated that a moderate earthquake could cause its collapse because of the reduced strength of the concrete used in its construction (Reaveley Engineers and Associates, 2003). Compression tests of concrete core samples from the drum indicated that concrete breaks could occur at stresses as low as 900 psi (6000 kPa) (AGRA Earth and Environmental, 2000), perhaps because the concrete was poured during the winter and may have frozen during the cold evening hours. The concrete also had a very high water-to-cement ratio (AGRA Earth and Environmental, 2000), further contributing to its low strength. Because of the nature of the concrete, outdated construction techniques, and potential amplification of earthquake ground shaking by the tall dome-and-drum assembly (Reaveley Engineers and Associates, 2003), the seismic retrofit would have to provide a major reduction in force displacement.

**Seismic Base Isolation and Shear Walls**

The seismic design of the Capitol had to account for the seismicity of the site, the safety of building occupants, and the historic and symbolic nature of the structure. This placed a higher level of demand on the structural engineer and the design team to provide the needed seismic design without altering the building’s architectural integrity.

After studying various structural systems and their impacts, the structural engineers recommended that a combination of base isolation and shear walls be used. Base isolation would decouple the building from the horizontal component of earthquake ground motions, and shear walls would resist and limit inter-story drift. Base isolation also reduced requirements for seismic reinforcement, avoiding damage to the integrity of the architecture of the Capitol (Reaveley Engineers and Associates, 2003).

Base isolation will lengthen the periodic response of the Capitol from ~1 s to 3 or 4 s (Fig. 5). The shorter response results in lateral accelerations of ~1.4 g, whereas the longer response resulting from base isolation lowers lateral accelerations to ~0.3 g. This reduction allows the use of other seismic design elements that are less intrusive than would otherwise be required to resist the effects of greater accelerations.

![Figure 5. Typical seismic-response spectrum illustrating the effects of base isolation and shear-wall reinforcement (Reaveley Engineers and Associates, 2003). Base isolation lengthens the periodic response of the Capitol, decoupling the building from the horizontal component of ground shaking. Shear walls resist and limit inter-story drift.](image-url)
To design the base isolation system, site-specific data were collected from several boreholes drilled to depths of at least 350 ft (110 m). Data included the results of downhole seismic surveys, which were used to develop several synthetic time histories that were then used to develop site-specific response spectra (AMEC Earth and Environmental, 2003).

Once the base isolators were designed, the design of shear walls was developed. Shear walls were placed vertically throughout the exterior and interior of the building (Fig. 6). The amount and length of the shear walls were greatly reduced by using base isolation and site-specific response spectra, and this allowed the architects greater latitude on placement of the shear walls within the building. Shear walls were located without disturbing the historic fabric of the building, and the integrity of the Capitol was preserved (Reaveley Engineers and Associates, 2003).

The drum on top of the Capitol was also reinforced with shear walls, which were extensive because of the poor quality of the concrete (Fig. 6). New shear walls were placed on the interior of the drum wall, and steel rods were inserted from the shear walls through the existing drum wall and into a new exterior concrete wall. This design provided added strength to the drum and use of its existing concrete continued, reducing costs and schedule by eliminating the need to remove the old drum and construct a new one (Reaveley Engineers and Associates, 2003).

Directions to Stop 3

From the State Capitol parking lot, turn right on Columbus Street and return to the intersection with 500 North. From the intersection, bear left on Victory Road for ~1.2 mi (1.9 km) until it merges with Beck Street. Continue northwest on Beck Street ~1.7 mi (2.7 km) to the entrance to I-15 North (toward Ogden). The remainder of this route will travel across the Farmington Siding landslide complex for a better view. For a shorter, more direct route to stop 3, leave I-15 at exit 326 and travel north on U.S. 89, turning on Main Street as described below. If continuing on the longer route, take I-15 north 10 mi (16 km) to exit 325 (Lagoon Drive–Farmington), bear right to continue north on 200 West ~1.0 mi (1.6 km), then turn left (west) on State Street (U.S. 227). Follow State Street ~0.5 mi (0.8 km), crossing over the freeway, and bear right as State Street merges with 100 North. Continue west on 100 North ~1.0 mi (1.6 km); 100 North changes to Clark Lane. Turn right on 1525 West ~0.7 mi (1.1 km). Turn left on Burke Lane ~0.3 mi (0.5 km), continue to the right ~0.3 mi (0.5 km) as Burke Lane changes to 1875 West, and turn left on 900 North. Continue ~0.1 mi (0.2 km), turn right on 2000 West, continue ~0.3 mi (0.5 km) over the railroad crossing and past the pond on the right, and turn right on Shepard Lane (Utah 106). Travel east on Shepard Lane ~1.4 mi (2.3 km), crossing under I-15 and passing the Oakridge Country Club, and turn left on U.S. 89 ~0.9 mi (1.4 km). Turn right on Main Street (Utah 272) ~0.4 mi (0.6 km) and park on Main Street between Somerset Street and Leonard Lane.

Stop 3—Liquefaction-Induced Farmington Siding Landslide Complex

The prehistoric Farmington Siding landslide complex comprises some of the largest landslides triggered by earthquakes in the United States. The landslide complex covers an area of ~7.5 mi² (19.5 km²) in southeastern Davis County, and is one of over a dozen late Pleistocene–Holocene features along the Wasatch Front that have been mapped as possible liquefaction-induced lateral spreads (Van Horn, 1975; Anderson et al., 1982; Nelson and Personius, 1993; Harty and Lowe, 2003). From the location of this field trip stop near the landslide main scarp, the landslide complex extends southwestward to Great Salt Lake. The Weber segment of the Wasatch fault zone lies ~0.5 mi (0.8 km) to the east. Data in this discussion are primarily from Hylland and Lowe (1998) and Harty and Lowe (2003).

Physical Setting and Geology

The Farmington Siding landslide complex is in a gently sloping area underlain at shallow depths primarily by fine-grained, stratified, latest Pleistocene to Holocene lacustrine deposits of Lake Bonneville and Great Salt Lake. Ground slopes within the landslide complex range from ~0.4%–0.8% and adjacent to the complex range from ~1%–2% along the flanks to 6%–11% in the crown area. The deposits involved in landsliding consist of interbedded, laterally discontinuous layers of clayey to sandy silt, well-sorted fine sand to silty sand, and minor clay and gravel. The crown area is underlain by Lake Bonneville sand and silt deposits and is at an elevation of ~4400 ft (1340 m) in the vicinity of the city of Farmington. The toe may have been encountered beneath Great Salt Lake during a drilling project in Farmington Bay to
test foundation conditions for a proposed water storage reservoir (Everitt, 1991).

Soil grain size distribution, standard penetration resistance, and groundwater depth noted on logs of geotechnical boreholes (Anderson et al., 1982) indicate liquefiable deposits in the shallow subsurface beneath the landslide complex. These data, as well as data from three boreholes drilled in an attempt to correlate beds beneath and adjacent to the landslide complex (Miller et al., 1981), indicate a possible landslide failure zone within a depth range of 14–40 ft (4–12 m). This zone locally corresponds to the contact between a relatively dense transgressive lacustrine sequence consisting of nearshore sand and gravel deposited during the early part of the Bonneville paleolake cycle, or possibly pre-Bonneville alluvium, and overlying loose-soft, offshore, fine-grained sediment subsequently deposited in deeper water.

Geomorphic features within the landslide complex include scarps, hummocks, closed depressions, and transverse lineaments. Well-preserved lateral and main scarps in the northern part of the complex range from ~10–40 ft (3–12 m) high. Hummocks and closed depressions are present across most of the complex, but are more common in the northern part (Fig. 7). Hummocks in the northern part are morphologically distinct, having as much as ~20 ft (6 m) of relief and lateral dimensions locally exceeding 1000 ft (300 m). Hummocks in the southern part are morphologically subtle, generally having <~6 ft (2 m) of relief.

**Landslide Characteristics**

Trenches were excavated in several areas to examine the style of deformation and characterize the type of mass movement of the landslide complex (Hyland and Lowe, 1998; Harty and Lowe, 2003). Observed subsurface deformation of lacustrine deposits includes inclined strata, gentle to strong folding, high- and low-angle shear surfaces, and loss of original bedding due to liquefaction. Small sand dikes are present locally, some of which were injected along shear planes.

Mass movement within the landslide complex was probably a combination of lateral spread and flow. By excavating trenches across hummock flanks and adjacent ground in the northern part of the complex, Harty and Lowe (2003) determined the hummocks are relatively intact “islands” of lacustrine strata surrounded by liquefied sand, which resulted from flow failure. Other evidence for flow failure includes the existence of a landslide main scarp up to 40 ft (12 m) high, overall negative relief in the head region of the complex indicating evacuation of a large volume of material, and overall positive relief in the distal region of the complex indicating accumulation of landslide material.

**Landslide Timing and its Relation to Surface-Faulting Earthquakes**

The landslide deposits can be grouped in two age categories relative to the age of the Gilbert shoreline complex of Lake Bonneville, which formed between 11,000 and 10,000 14C yr B.P. (Currey, 1990). The northern part of the landslide complex truncates the Gilbert shoreline (Van Horn, 1975), indicating major post-Gilbert movement. However, the Gilbert shoreline can be traced across the southern part of the landslide complex (Anderson et al., 1982; Harty and Lowe, 2003), indicating pre-Gilbert movement in this area.

Relative timing information and radiocarbon soil ages indicate at least three, and possibly four, episodes of large-scale liquefaction-induced landsliding: the first sometime between 14,500 and 10,900 14C yr B.P., the second just prior to 7310 ± 60 14C yr B.P., the third (?) sometime prior to 5280 ± 60 14C yr B.P., and the fourth between 2340 ± 60 and 2440 ± 70 14C yr B.P. Hylland and Lowe (1998) considered the timing of these landslide events within the context of paleoclimatic and lacustral fluctuations and observed that landsliding was associated with climate-induced highstands of Great Salt Lake. The apparent correspondence between landslide events and lacustral highstands suggests that landsliding may have occurred under conditions of relatively high soil pore-water pressures, and possibly increased artesian pressures, associated with rising lake and groundwater levels.

Many features (for example, evidence of lateral spread, flow failure of gentle slopes, sand dikes, deposits susceptible to liquefaction, and proximity to faults with recurrent Holocene activity) indicate landsliding was likely triggered by strong earthquake ground shaking. Comparison of the timing of surface-faulting earthquakes on the active segments of the Wasatch fault zone with the timing of Farmington Siding landslides indicates a close correspondence between landsliding and certain earthquakes (Fig. 8). Within uncertainty limits, surface-faulting earthquakes on the Brigham City segment coincide with all four possible landslide events. Surface-faulting earthquakes on the Weber, Salt Lake City, and Provo segments also coincide with the more recent landslide events. However, Hylland and Lowe (1998) and Hylland (1999) determined that large (surface-faulting) earthquakes on the nearby Weber segment of the Wasatch fault zone are significantly

![Figure 7](image_url). Aerial view of hummocky landslide terrain on the northern part of the Farmington Siding landslide complex. View is to the northwest, with Interstate 15 near the middle of the picture.
more likely than earthquakes on other segments to trigger the widespread liquefaction-induced ground failure and large lateral displacements characteristic of the Farmington Siding landslide complex. The lack of evidence for liquefaction-induced landsliding triggered by the most recent surface-faulting earthquake on the Weber segment, which occurred during a relative lowstand of the lake, indicates that liquefaction-induced landsliding at the Farmington Siding landslide complex is tied not only to the likelihood of a surface-faulting earthquake on the Weber segment, but also to concurrent hydrogeologic conditions.

Directions to Stop 4

Continue south on Main Street (Utah 272) ~0.4 mi (0.6 km), turn left on 1400 North ~0.5 mi (0.8 km), and turn right on Compton Road. Drive ~0.4 mi (0.6 km) on Compton Road and park.

Stop 4—Compton Bench Fire-Related Debris Flows, Farmington

Debris flows consist of rock, soil, and other debris that mix with water from intense thunderstorms or spring snowmelt and travel downslope at high speeds. Because of their considerable mass and speed, debris flows are a threat to life and can damage anything in their path, particularly buildings, roads, and utility connections. Debris flows are usually much more damaging than flash floods, but are generally more restricted in the area they impact.

The Compton Bench fire-related debris flows are an excellent example of small-volume debris flows produced from small drainage basins. The lower slopes of the Wasatch Range above and east of Farmington City were burned in the Farmington fire of July 2003. Intense thunderstorms on 6 April 2004 triggered five small debris flows in the Compton Bench area of Farmington, a gently sloping area of coalesced alluvial fans.

The Farmington Fire and Post-Fire Hazard Assessment

The Compton Bench debris flows occurred eight months after a wildfire in the mountains above Farmington. The Farmington wildfire was a human-caused fire that burned ~2000 acres (800 ha) of U.S. Forest Service and private land on the lower western flank of the Wasatch Range above Farmington (U.S. Forest Service, 2003). The fire affected drainage basins above Farmington and Rudd Canyons and several smaller drainage basins.

A debris-flow hazard assessment conducted by the UGS following the fire recognized a heightened debris-flow–flooding hazard for the tributaries within the burn area based on evidence of previous debris flows, erodible sediment stored in channels, burned hillslopes capable of generating rapid runoff, and rapid snowmelt and/or thunderstorm-rainfall potential (Giraud, 2003). Since Farmington and Rudd Canyons had previously produced damaging debris flows, debris basins had been constructed at the mouths of these canyons and provided protection for downslope development on alluvial fans. However, small drainage basins with areas of 5–80 acres (2–30 ha) along the lower flank of the Wasatch Range had no protective structures. Due to the heightened debris-flow–flooding hazard, the U.S. Forest Service and U.S. Natural Resources Conservation Service undertook hazard-reduction measures to protect the watersheds. These protective measures, completed in late fall 2003, involved reseeding within the burn area and constructing sediment fences within channels below the small, unprotected drainage basins where houses were at risk.

Physical Setting and Geology

The small drainages that produced the Compton Bench debris flows lie along the foot of the Wasatch Range and rise in elevation from ~4600–4800 ft (1400–1500 m) at their mouths to 6400 ft (2000 m). The bedrock in these small drainages comprises the Precambrian Farmington Canyon Complex (Bryant, 1988), which includes schist, gneiss, and quartzite. Above 5200 ft (1600 m), the elevation of the Bonneville shoreline of Pleistocene Lake Bonneville, the slopes consist of colluvium with minor rock outcrops. Below 5200 ft (1600 m) lacustrine sand and gravel deposits of Lake Bonneville overlie the Farmington Canyon Complex (Nelson

Figure 8. Comparison of the timing of Farmington Siding landslides (shaded areas) with Wasatch fault zone surface-faulting earthquakes (top) and Great Salt Lake fluctuations (bottom) (modified from Hyl-land and Lowe, 1998). Dashed lines for landslide events indicate limiting ages; ages for landslides 2 and 3(?) are minimum limiting ages. Earthquake consensus preferred ages (dashed lines) and uncertainty limits (boxes) from Lund (2005); Great Salt Lake hydrograph after Murchison (1989). Age picks for calendar-calibrated time scale (top) were determined using the calibration program of Stuiver et al. (2005) and are approximate.
and Personius, 1993). Alluvial fans formed at the mouths of drainages following the recession of Lake Bonneville.

The Compton Bench Debris Flows

The Compton Bench debris flows were triggered by intense thunderstorm rainfall between 8:30 and 9:00 p.m. on 6 April 2004. Two rain gauges about a mile from where the debris flows initiated recorded relatively small amounts of rainfall. A Davis County rain gauge at the Farmington Canyon debris basin at 4800 ft (1500 m) elevation recorded 0.17 in. (0.43 cm) in 17 min, and a U.S. Forest Service rain gauge in Rudd Canyon at 5200 ft (1600 m) elevation recorded 0.57 in. (1.4 cm) in 23 min. The debris flows initiated at elevations of ~6000–6200 ft (1800–1900 m), and the rainfall amounts and intensity were likely higher in the initiation areas than measured at the rain gauges. Rainfall measurements in other burned areas indicate that relatively small amounts of intense thunderstorm rainfall in the range of 0.27–0.35 in/h (0.7–0.9 cm/h) are capable of triggering fire-related debris flows (McDonald and Giraud, 2002; Cannon et al., 2003). Other factors contributing to debris flows include steep slopes, ample supply of channel sediment, and increased runoff caused by soils with high moisture content from recent snowmelt.

A traverse up the small drainage basins was conducted the morning after the debris flows to assess initiation processes and sediment bulking characteristics and to assist Farmington City in determining the potential for future debris flows. The hillslopes in the upper parts of the small drainage basins showed evidence indicating the debris flows began as intense runoff, and sheetwash erosion concentrated as rills and quickly flowed into the drainage basin channels. The debris flows evidently began entraining sediment in the upper parts of the drainage basin channels and continued to bulk sediment progressively downstream, through erosion and scour of the main channels. Below the Bonneville shoreline, abundant, loose, easily erodable sediment in the channels was bulked into the flows. At one locality, channel erosion threatened a section of a Weber Basin Water Conservancy District aqueduct running along the mountain front.

Deposit volumes ranged from 200 to 1500 yds$^3$ (150–1100 m$^3$). Debris-flow deposits include thin levees along alluvial fan channel margins and thin slabs and lobes on small alluvial fans. The lateral deposit margins were flat rather than steep, indicating high water content for the flows.

Two of the larger flows converged on Compton Road and flowed west across the road, covering parts of three lots with sediment (Fig. 9). These two flows traveled ~3000 ft (910 m) and dropped 1200 ft (370 m) in elevation. The flows traveled down shallow alluvial fan channels, collapsing the sediment fences in the channels. The high water content and highly fluid character of these flows promoted a longer runout. Sediment deposition was mostly restricted to streets and yards, but damage also occurred to several vehicles, garages, and homes (Fig. 10). Most of the deposits had a maximum thickness of 0.5 ft (0.2 m) unless the sediment flowed up against a building or other flow barrier, where sediment burial depths were up to 3 ft (0.9 m).

Compton Bench Risk-Reduction Structures

A reconnaissance of the small drainage basins following the 6 April 2004 debris flows indicates that ample sediment is available for future debris flows. The potential for future debris flows prompted Farmington City, Davis County, and the U.S. Forest Service to collaborate and construct debris basins east of Compton Road to reduce the risk to lots impacted by the 6 April debris flows. Two small debris basins were constructed on the alluvial fan. An upper basin intercepts water and debris from two channels and connects to a lower basin where a high-level water outlet conveys water to the stormwater system. A berm downslope of the lower basin along the east side of Compton Road prevents additional sediment storage if the capacity of the two debris basins is exceeded.

Directions to Stop 5

Continue south on Compton Road ~0.4 mi (0.6 km), turn right on 1100 North, and follow 1100 North ~0.3 mi (0.5 km) as it changes to Quail Flight, which ends at Quail Circle. Turn right on Quail Circle and take a quick turn left on Quail Run Road at the end of Quail Circle. Take Quail Run Road ~0.1 mi (0.2 km), turn right on Main Street ~0.1 mi (0.2 km), and turn left on Shepard Lane. Continue 0.4 mi (0.6 km) on Shepard Lane, turn left (south) on U.S. 89 ~1.2 mi (1.9 km) and turn right (west) on I-15 northbound ~8.3 mi (13.4 km). Take Exit 332 (Syracuse) from I-15, turn left (west) on Antelope Drive (U.S. 108), and proceed west ~6.8 mi (10.9 km) to the Antelope Island State Park fee station following signs for Syracuse and Antelope Island. Cross the causeway to Antelope Island, ~6.8 mi (10.9 km), and bear left on the paved road. Follow the paved road ~1.2 mi (1.9 km) to an intersection and turn right. Continue on the road ~0.9 mi (1.4 km) to Buffalo Point, and park at the Bridger Bay Campground.

Stop 5—Flood Hazards Related to Great Salt Lake

A substantial portion of the Wasatch Front metropolitan area, from Salt Lake City north to Brigham City, is developed on a relatively narrow plain sandwiched between the eastern shore of Great Salt Lake and the Wasatch Range (Fig. 1). Low-lying coastal areas adjacent to Great Salt Lake’s 60-mi-long (100-km) eastern shore are subject to flooding from three main processes: (1) lake-level rise in response to multiyear wet cycles; (2) lake seiches; and (3) tectonic subsidence associated with surface faulting on the Wasatch fault zone.

The most recent damaging flood took place between 1983 and 1987, when above-normal precipitation caused the lake level to rise to a new twentieth century high of 4212 ft (1284 m) above sea level in 1986 and again in 1987. Flooding associated with this highstand caused an estimated $240 million in damage to shoreline development. As a result of the recent drought (1999–2004), the lake level declined to a lowstand of ~4194 ft (1279 m) at the end of summer 2004. However, the winter of 2004–2005 produced above-normal precipitation and snowpack over most of the lake’s drainage basin, and the lake is rising again.
Geologic Setting

Great Salt Lake is the largest closed-basin lake in the Great Basin and the sixth largest lake in the United States; only the Great Lakes are larger. The complex structural basin occupied by Great Salt Lake was created during the episode of extensional tectonics, dating from the past 17 m.y. (Hintze, 1988), that gave rise to the Basin and Range Province. The lake occupies the lowest spot in a 22,000-mi² (57,000-km²) drainage basin that extends beyond the northeastern Great Basin into the adjacent middle Rocky Mountains. Runoff from the Jordan, Weber, Ogden, and Bear Rivers is the major water input; evaporation from the lake surface is the major output from the system. Salinity of the water varies inversely with lake level from 5% to 28%.

The present lake basin, and the deeper portions of its Pleistocene predecessor Lake Bonneville, coincide with three interconnecting fault-bounded grabens (Stokes, 1980). The intervening horst blocks give rise to the various islands in the lake, of which Antelope Island is the largest. Structurally, the island is the exposed tip of a tilted horst block, bounded on the west by the Antelope Island section of the Great Salt Lake fault zone (Willis et al., 2000).

Lake Bonneville and Its Shorelines

Lake Bonneville was the largest and most recent of the pluvial lakes that occupied the Bonneville Basin during the late Pleistocene (Fig. 11) (Currey, 1990; Oviatt et al., 1992; Oviatt, 1997). The elevations of the prominent shorelines of Lake Bonneville and Great Salt Lake are given in Table 3. Lake Bonneville began to rise from levels close to those of modern-day Great Salt Lake ca. 28,000 ¹⁴C yr B.P. The lake rose gradually, but experienced a major, climatically induced oscillation between 21,000 and 20,000 ¹⁴C yr B.P. that produced the Stansbury shoreline (Oviatt et al., 1990). The lake eventually resumed its rise, but the rise slowed as the lake level approached an external basin overflow threshold in southern Idaho. Lake Bonneville reached this threshold and occupied its highest shoreline, which Gilbert (1875) named the Bonneville beach, after 15,500 ¹⁴C yr B.P. The lake remained at this level perhaps as late as 14,500 ¹⁴C yr B.P., when catastrophic failure at Red Rock Pass, Idaho, an event known as the Bonneville Flood (O’Conner, 1993), lowered the lake level 350 ft (110 m) in less than two months, stabilizing at...
the newly established Provo shoreline. After 14,000 $^{14}$C yr B.P., Lake Bonneville became a reduced, hydrologically closed system as overflow ceased when the climate warmed in response to a northward shift in the mean position of the westerly storm tracks following melting of the Laurentide and Cordilleran ice sheets (Oviatt, 1997). By ca. 12,000 $^{14}$C yr B.P., the lake dropped to levels possibly at or lower than those of modern-day Great Salt Lake (Oviatt et al., 1992), but Lake Bonneville then transgressed to form the Gilbert shoreline between 11,000 and 10,000 $^{14}$C yr B.P. Since the moderate rise to the Gilbert, possibly correlative with the climatic cooling of the Younger Dryas event (Oviatt, 1997), lake levels in the much-reduced Great Salt Lake have fluctuated more modestly during Holocene time.

The Bonneville shoreline is a prominent geomorphic feature of the Wasatch Front. Easily visible from much of Salt Lake Valley, the elevation of the Bonneville shoreline varies from 5161–5216 ft (1573–1590 m) due to a combination of post-lake isostatic rebound and faulting (Hylland et al., 1997). Many shoreline features and other scientifically important “geoantiquities” related to Lake Bonneville have been covered, quarried, or otherwise lost due to population growth and associated urbanization along the Wasatch Front (Chan and Milligan, 1995).

### Historic Lake-Level Fluctuations and Flooding During the 1980s

The following discussion of historic lake levels and flood events is abstracted from Arnow and Stephens (1990), Atwood et al. (1990), Atwood and Mabey (1995), Hylland et al. (1997), and Atwood (2002).

The United States Geological Survey (USGS) has been monitoring water levels in Great Salt Lake since 1875. These data combined with Gilbert’s (1890) pregauge hydrograph (1847–1875) provide a nearly complete record of lake-level fluctuations since the time of Mormon settlement in the area. The historic average level of Great Salt Lake is ~4202 ft (1281 m); at this level, the lake covers an area of ~1700 mi$^2$ (4400 km$^2$).

The 150-year-plus record of lake levels reveals long-term fluctuations that correlate directly with variations in the amount of precipitation within the lake’s drainage basin. Prior to the mid-1980s, the historic high of Great Salt Lake was ~4212 ft (1284 m), which was reached in the early 1870s. Over the next 90 yr, the lake slowly dropped until reaching a historic low of 4191 ft (1277 m) in 1963, covering only 950 mi$^2$ (2460 km$^2$). September 1982 was the wettest month in the history of precipitation measurements at Salt Lake International Airport. This record wet month began a four-year wet cycle, in part related to strong El Niño events in 1982–1983 and 1986–1987 (Alder, 2002), over most of the Great Salt Lake drainage basin. At the beginning of September 1982, the lake was at an elevation of ~4200 ft (1280 m). From September 1982 to June 1983, the lake rose 5.2 ft (1.6 m), a record seasonal fluctuation. The lake continued to rise and reached a highstand of 4212 ft (1284 m) in 1986 and again in 1987, equaling or slightly exceeding the level reached in the 1870s. At this elevation the lake covered ~3300 mi$^2$ (8547 km$^2$), resulting in flooding of low-lying coastal areas. Post-flood mapping of the debris lines on Antelope Island created during this flood indicate significant wind setup and wave runup above the static level of the lake (Atwood, 1994; Atwood and Mabey, 1995; Atwood, 2002). On Antelope Island, the elevation of debris lines associated with the 1980s highstand of 4212 ft (1284 m) ranges from 4212–4218 ft (1284–1286 m). The magnitude and spatial variation in this process of shoreline superelevation needs to be considered when defining the lake’s flood plain (Atwood, 2002).

From 1982 to 1986 the lake rose 12 ft (3.7 m), doubling its volume and flooding ~500,000 acres (202,500 ha) of the lake’s historic flood plain. The flooding and associated wave erosion damaged or destroyed public and private resources and facilities, including damage to the Rose Park industrial area, Interstate 80, the Antelope Island causeway, park pavilions at Antelope Island State Park (Fig. 12), and salt company evaporation ponds and dikes. The overall economy of the State of Utah was dramatically impacted, with total flood-related losses estimated at ~$240 million (Austin, 2002).

In response to the unprecedented flood damage, the State of Utah employed two measures in an attempt to quickly lower the level of Great Salt Lake: (1) breaching the Southern Pacific Railroad (SPRR) causeway in 1984, and (2) pumping lake water into the West Desert in 1987 (Gwynn, 2002). By 1983, the SPRR causeway had created a 3.5-ft (1.1-m) lake-level differential between the north (lower) and south (higher) arms of the lake due to the impermeable nature of the causeway and the abnormally high inflows of water into the south arm of the lake during the early 1980s. The causeway was breached on 1 August 1984 and within several months the lake level differential had been reduced to ~1 ft (0.3 m). However, the south arm of the lake continued to rise. In 1986, the State of Utah decided to pump lake water westward into the Great Salt Lake desert, adjacent to the Bonneville Salt Flats. The West Desert Pumping Project was an attempt to increase evaporation rates and remove 690,000 acre-ft (84,870 ha-m) from the lake. Three large pumps were installed near the south end of the Hogup Mountains that lifted water from the lake into a 4.1-m (6.6-km) canal and transported it westward into a shallow topographic depression known as the West Pond. By the

### Table 3. Elevations and Ages of Major Shorelines and Levels of the Lake Bonneville-Great Salt Lake System

<table>
<thead>
<tr>
<th>Shoreline or lake level</th>
<th>Elevation (ft)</th>
<th>Date or age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Historic high</td>
<td>4212</td>
<td>1873, 1986, and 1987</td>
</tr>
<tr>
<td>Historic average</td>
<td>4202</td>
<td>N.A.</td>
</tr>
<tr>
<td>Historic low</td>
<td>4181</td>
<td>1963</td>
</tr>
<tr>
<td>Holocene high</td>
<td>4221</td>
<td>ca. 3000 $^{14}$C yr B.P.</td>
</tr>
<tr>
<td>Holocene low</td>
<td>4180 (?)</td>
<td>ca. 6000 $^{14}$C yr B.P.</td>
</tr>
<tr>
<td>Gilbert shoreline</td>
<td>4250</td>
<td>ca. 11,000–10,000 $^{14}$C yr B.P.</td>
</tr>
<tr>
<td>Provo shoreline</td>
<td>4740</td>
<td>ca. 14,500–14,000 $^{14}$C yr B.P.</td>
</tr>
<tr>
<td>Bonneville shoreline</td>
<td>5200</td>
<td>ca. 15,500–14,500 $^{14}$C yr B.P.</td>
</tr>
<tr>
<td>Stansbury shoreline</td>
<td>4500</td>
<td>ca. 21,000–20,000 $^{14}$C yr B.P.</td>
</tr>
</tbody>
</table>

*Modified from Currey et al., 1984.
November 1988, the lake level had dropped ~5.4 ft (1.6 m) due to a combination of pumping, evaporation, and decreased inflow resulting from two drier-than-average years (Hylland et al., 1997). Pumping (April 1987 to June 1989) probably lowered the level of the lake ~2 ft (0.6 m), or 36% of the total decline during that period (Gwynn, 2002). The pumping facility is maintained in a state of ready reserve for future high lake levels.

Lake Seiches

A seiche is a free or standing-wave oscillation of the water level in an enclosed or semi-enclosed basin, such as a lake, bay, or harbor. Wind-driven lake setup, landslides, earthquake-induced ground shaking, or surface faulting on the lake floor may create a seiche. The large waves created by surface faulting beneath the body of water are sometimes called surges to differentiate them from the smaller oscillations caused by wind setup or ground shaking (Myers and Hamilton, 1964).

Wind seiches in Great Salt Lake are a fairly common occurrence, are well documented by USGS lake-level monitoring (Atwood, 2002), and have been studied in some detail (Wang, 1978). Southerly winds in advance of a passing cold front will induce lake setup in the north and lake set-down along the southern shoreline. When the wind velocity drops, the water level will oscillate, with a fundamental period of ~6 h, as the lake tries to regain equilibrium (Fig. 13). Wang (1978) found that significant lake setup requires a wind velocity of 10 knots (18.5 km/h) for 12 h duration. However, Atwood (2002) suggests that wind setup in the southern arm of Great Salt Lake may not require steady, strong winds of long duration. The maximum amplitude of wind-driven seiching in Great Salt Lake is ~2 ft (0.6 m) (Wang, 1978). However, this could cause significant wave damage if superimposed on an already high lake level.

The potential for earthquake-induced seiching in Great Salt Lake is poorly understood, but Lowe (1993) makes a strong case, based on historical lake-level data and earthquake accounts, that the 1909 Hansel Valley earthquake (M 6+) generated a 12-ft (3.7-m) wave that overtopped the Lucin Cutoff railroad trestle. Pechmann (1987) notes the exceptional size of this wave and that the larger 1934 M 6.6 Hansel Valley earthquake did not generate seiching in Great Salt Lake. He concludes that the epicenter of the 1909 earthquake may have actually been beneath Great Salt Lake and the reported wave was in fact a surge related to ground rupture beneath the lake, as opposed to ground shaking from a more distant earthquake. A future surge of similar size along the southern or eastern shore of Great Salt Lake would be a significant and potentially damaging event. However, for a scenario M 7 earthquake along the Salt Lake City segment of the Wasatch fault zone, neither seiching related to ground shaking nor a surge related to surface faulting along the Salt Lake City segment would be significant (Solomon et al., 2004).

Tectonic Subsidence

Several papers and reports address potential lake margin flooding caused by earthquakes (e.g., Lowe, 1993). Two major studies on the potential impacts of earthquake-induced tectonic deformation, tilting, and flooding along the Wasatch fault zone were published by Keaton (1986) and Chang and Smith (1998).
Chang and Smith (1998) refined previously published deformation models by inputting more accurate elevation data into an elastic, three-dimensional boundary-element model to simulate ground-surface deformation associated with large, surface-faulting earthquakes on the Weber and Salt Lake City segments of the Wasatch fault zone. The study by Chang and Smith (1998) demonstrates that flooding from Great Salt Lake resulting from tectonic subsidence or ground tilt associated with a large surface-faulting earthquake on the Wasatch fault zone could be a significant hazard for developed areas of Salt Lake and Davis Counties. The flooding hazard associated with such an earthquake is based on the premise that the topography of low-lying areas along the eastern shore of Great Salt Lake can be permanently altered by subsidence of the hanging wall, resulting in an eastward migration of the lake shoreline and inundation of adjacent areas. Superimposing the effects of any accompanying surges and seiches would cause even greater inland flooding (Atwood and Mabey, 1995). A secondary hazard would be the adverse effects of backtilting on the many gravity-driven water and sewer lines that cross the fault zone.

The worst-case scenario modeled by Chang and Smith (1998) (based on a M 7.5 earthquake on the Salt Lake City and Weber segments of the Wasatch fault zone similar to the 1959 Hebgen Lake, Montana, earthquake and Great Salt Lake at a level equal to its historic highstand of 4212 ft [1284 m]) results in a 3.5-mi (5.6-km) southeastward displacement of the shoreline onto land containing commercial developments and major transportation corridors. The degree of flooding that results from their models is very sensitive to the level of Great Salt Lake at the time of the earthquake. Recent GIS-based mapping (Hernandez, 2004) of the potential extent of flooding in southern Davis County demonstrates the vulnerability of the I-15 corridor in that area (Fig. 14). Chang and Smith (1998) state their results should be used with caution because of the lack of site-specific data in their models. In addition, Keaton (1986) indicates that the amount of hanging wall deformation produced by the 1959 Hebgen Lake earthquake (~20 ft [6.1 m] of vertical displacement) may be more than twice what would be expected for a slightly lower magnitude earthquake on either the Weber (Fig. 14) or Salt Lake City (Fig. 4) segments of the Wasatch fault zone (~6.9 ft [2.1 m], based on paleoseismic studies). Recent discussions of the seismic records of the Hebgen Lake earthquake are also raising questions about the number and magnitude of seismic events that actually occurred on that day (J.R. Keaton, 2004, personal commun.). Therefore, the extent of tectonic-related lake-margin flooding as shown in Figure 14 may be overestimated.

Directions to Stop 6

Retrace the route to I-15, turn right (S) on I-15, travel for ~19 mi (31 km) to Exit 316, and turn right (W) to I-215. Take I-215 west and south ~9.0 mi (14 km) to exit 20A, turn right (W) on State Route 201, travel for ~7.5 mi (12 km), and turn left on 8400 West (U.S. 111) in Magna. Continue south on 8400 West for ~5.0 mi (8.0 km), turn right (W) on 5400 South, and park after ~0.5 mi (0.8 km), before the road turns left into the gravel pit.

Stop 6—Faulted Pleistocene (?) Deposits in Western Salt Lake Valley

The Wasatch fault zone has been extensively studied, exhibiting evidence of repeated movement during the Holocene, and its hazards are relatively well known. However, other Quaternary faults are also found in and near the central Wasatch Front. Some have not moved during the past 10,000 yr, the conventional criterion to define an active fault, but Pleistocene faults with long recurrence intervals may pose significant seismic hazards as well. DePolo and Slemmons (1998) suggest that a longer time period, 130,000 yr, is a more appropriate criterion because most earthquake recurrence intervals in the Basin and Range

Figure 14. Tectonic-related flood distribution based on the historic high stand (4212 ft [1284 m]) of Great Salt Lake and hypothesized ground deformation on the southern portion of the Weber segment of the Wasatch fault zone. The mapping assumes the worst-case scenario (the observed hanging wall deformation associated with the 1959 M 7.5 Hebgen Lake, Montana, earthquake) proposed by Chang and Smith (1998) (modified from Hernandez, 2004).
Province exceed 10,000 yr; the larger time period encompasses most recurrence intervals for faults in the province, and at least 50% of historical earthquakes in the province of M 6.5 or greater involved fault traces that lacked prior Holocene faulting. At Stop 6 we examine a possible Pleistocene fault exposure near the west margin of Salt Lake Valley that has no surface expression but may be considered active using the criteria of DePolo and Slemmons (1998).

Possible Pleistocene Faulting in Western Salt Lake Valley

Recent detailed geologic mapping in western Salt Lake Valley revealed faulted deposits of possible Pleistocene age on the east side of the Oquirrh Mountains (Biek et al., 2004) (Fig. 4). These faults include the Harkers fault, extending at least 6 mi (10 km) from the range front to the range interior and separating the Permian and Pennsylvanian Oquirrh Group in the footwall from Miocene (?) to Pleistocene alluvial fan deposits in the hanging wall. Other small faults, exposed at Stop 6, separate the alluvial fan deposits from tuff in the Jordan Narrows unit of the Salt Lake Formation near the northern end of the Harkers fault, 1.0 mi (1.6 km) east of the range front. Bryant et al. (1989) reported a fission-track age of 4.4 ± 1.0 Ma for a rhyolitic tuff in the Jordan Narrows unit, which, with its stratigraphic position above rocks of known Miocene age, indicates a Miocene to Pliocene age for the unit (Biek et al., 2004).

Biek et al. (2004) differentiated the pre-Bonneville alluvial fan deposits in western Salt Lake Valley into two units based on morphology and paleosol development: (1) older alluvial fan deposits form relatively steep, deeply dissected, erosionally resistant remnants commonly overlain by a stage IV calcic paleosol, indicating an early to middle Pleistocene age for the upper part of the unit; and (2) younger pre-Bonneville alluvial fan deposits form relatively gently sloping aprons on piedmont slopes, are commonly truncated by the Bonneville shoreline, and are overlain by stage II or III calcic paleosols, suggesting a middle to late Pleistocene age. No fossil evidence has been found in the pre-Bonneville alluvial fan deposits, but glass shard analyses of tuff samples from the adjacent Copperton quadrangle (Biek et al., 2004) suggest a chemical correlation to the 6.4 ± 0.1 Ma Walcott Tuff, indicating that the oldest beds of the deposits may be as old as late Miocene.

Faults Exposed at Stop 6

Faults are exposed in pre-Bonneville beds in a railroad cut ~3 mi (5 km) south of Magna (Fig. 15). Two of the faults juxtapose tuffaceous lake beds and pre-Bonneville alluvial fan deposits, and several smaller faults occur within each unit. The amount of displacement on the larger faults cannot be determined, but is at least ~20 ft (6 m), the height of the exposed cut slope. The tuffaceous lake beds are part of the Miocene to Pliocene Jordan Narrows unit of the Salt Lake Formation. The alluvial fan deposits are assumed to be of Miocene (?) to middle Pleistocene age because of their proximity to outcrops of that unit, but paleosol preservation is poor at the exposure. The cut slope is ~0.8 mi (1.3 km) northeast of the Lake Bonneville highstand shoreline and is ~120 ft (37 m) lower in elevation. A thin veneer of Bonneville gravel that is not displaced caps the exposure. Because of the uncertainty in the age of the alluvial fan deposits, the faults may have been active during the Pleistocene prior to the Lake Bonneville transgression, but displacement may be as old as Miocene.

The bounding faults trend N10–20°E and form a small graben that has displaced the alluvial fan deposits down with respect to the tuffaceous lake beds. The exposed graben is ~400 ft (120 m) wide. Although the alluvial fan deposits are downdropped, their greater relative erosional resistance resulted in a topographic high, but no linear fault trace is evident on the ground surface. Lake Bonneville gravel accumulated on the high to form a curved spit ~2000 ft (600 m) long on the fan remnant. The spit helped to shelter an adjacent small lagoon ~400 ft (120 m) in diameter, whose fine-grained deposits underlie a small, shallow, grass-covered depression.

Directions to Stop 7

Return to 5400 South and travel east ~7.4 mi (11.9 km), passing under I-215, to Redwood Road. Turn right on Redwood Road, go ~0.5 mi (0.8 km) to I-215, and take I-215 east ~7.0 mi (11.3 km) to Exit 6 (6200 South). Turn right onto 6200 South, proceed uphill, and continue south on Wasatch Blvd for ~4.0 mi (6.4 km). Bear right at the traffic signal on Wasatch Blvd as it diverges from Little Cottonwood Road, and continue ~1.1 mi (1.8 km). Turn right on 9800 South and then immediately turn right into the parking lot for G.K. Gilbert Geologic Interpretive Park.
Stop 7—Surface Faulting on the Salt Lake City Segment of the Wasatch Fault Zone

The Wasatch fault zone is one of the longest and most active normal-slip faults in the world. The Salt Lake City segment of the Wasatch fault zone trends through the densely populated Salt Lake Valley and poses a significant seismic risk to the Salt Lake City metropolitan area (Fig. 16). The segment extends for ~29 mi (46 km) from the Traverse Mountains on the south to the Salt Lake salient on the north. The Salt Lake City segment displays abundant geologic and geomorphic evidence for multiple surface-faulting earthquakes during Holocene time. Personius and Scott (1992) subdivided the Salt Lake City segment into three subsegments (from north to south): (1) the Warm Springs fault, (2) the East Bench fault, and (3) the Cottonwood section. Urbanization within Salt Lake Valley has obscured or modified scarps along the Salt Lake City segment in many places.

Here at Stop 7, the Wasatch fault principally cuts prominent Pinedale-age glacial moraines of Little Cottonwood and Bells Canyons, the only areas in Salt Lake Valley where the glaciers reached the mountain front. The highest shoreline of Lake Bonneville is just to the west, and recent, unpublished work by Elliot Lips of the University of Utah indicates that the maximum glacial advance and occupation of the Bonneville shoreline were contemporaneous (E. Lips, 2005, personal commun.). Scarps and the graben at this locality thus represent surface faulting over about the past 16,000 yr.

Little regard was given to surface-faulting hazards in Salt Lake County land-use planning prior to 1970, when an informal fault investigation and review process was implemented for new buildings. In 1989, Salt Lake County enacted the Natural Hazards Ordinance, now revised and renamed the Geologic Hazards Ordinance, and requires site-specific investigations to locate active faults and establish appropriate building setbacks prior to development (Batatian, 2002). At Stop 7, we will discuss surface-faulting hazards along the Salt Lake City segment and paleoseismic studies to evaluate them and will observe significant geologic and geomorphic features of the segment.

Regional Setting of the Wasatch Fault Zone

The Wasatch fault zone extends for 213 mi (343 km) from southeastern Idaho to north-central Utah (Machette et al., 1992). The fault zone generally trends north-south and, at the surface, can form a zone of deformation up to several hundred feet wide containing many subparallel west-dipping main faults and east-dipping antithetic faults. Schwartz and Coppersmith (1984) originally divided the Wasatch fault zone into six independent, seismogenic segments based on scarp morphology, surface-faulting patterns, range-crest morphology, geophysical evidence, and limited trenching information. Based on additional detailed trenching studies and geologic mapping, Machette et al. (1992) proposed a revised segmentation scheme consisting of 10 segments. The central five segments of the fault zone (Brigham City, Weber, Salt Lake City, Provo, and Nephi), along the western base of the Wasatch Range, each show evidence of two or more surface-faulting earthquakes in the past 6000 yr (Black et al., 2003).

The Wasatch Range is a major north-south-trending mountain range in Utah, and the Wasatch fault zone forms a prominent west-facing escarpment along its base. The fault zone is the easternmost feature of the Basin and Range Province, which is characterized by a series of generally north-trending elongate mountain ranges, separated by predominate alluvial and lacustrine sediment-filled valleys and typically bounded on one or both sides by major normal faults (Stewart, 1978). Late Cenozoic normal faulting, a characteristic of the Basin and Range Province, began between ca. 17 and 10 Ma in the Nevada (Stewart, 1978) and Utah (Anderson, 1989) portions of the province. The faulting is a result of a roughly east-west–directed, regional extensional stress regime that continues to the present (Zoback, 1989).

The Wasatch fault zone is also near the center of the Intermountain seismic belt, a generally north-south–trending zone of historical seismicity extending from northern Arizona to northwestern Montana (Smith and Sbar, 1974). At least 16 earthquakes of M 6.0 or greater have occurred within the Intermountain seismic belt since 1850, but none of them occurred...
along the Wasatch fault zone (Arabasz et al., 1992). The largest of these earthquakes was the M₅ 7.5 event in 1959 near Hebgen Lake, Montana.

**Paleoseismic Studies and Earthquake History**

The Holocene chronology of surface-faulting earthquakes on the Salt Lake City segment has been the subject of paleoseismic studies for more than two decades (Swan et al., 1981; Lund and Schwartz, 1987; Schwartz and Lund, 1988; Lund, 1992). The earthquake chronology was initially based on paleoseismic investigations at Little Cottonwood Canyon in 1979 and South Fork Dry Creek in 1985 (Fig. 16). A subsequent study at Dry Gulch in 1991 discovered a previously unrecognized event on a scarp not trenched at the South Fork Dry Creek site, which suggested the surface-faulting chronology was incomplete. Black et al. (1996) conducted additional trenching at the South Fork Dry Creek site in 1994 to establish a complete surface-faulting earthquake chronology from at least the middle Holocene for the Salt Lake City segment. McCalpin (2002) excavated a megatrench at the original Little Cottonwood Canyon site in 1999 in an attempt to capture the entire post-Bonneville record of paleoearthquakes on the Salt Lake City segment.

In 2003, the UGS, with funding from the National Earthquake Hazards Reduction Program, convened the Utah Quaternary Fault Parameters Working Group, a panel of expert paleoseismologists and seismologists, to make a comprehensive evaluation of the paleoseismic trenching data available for Utah’s Quaternary faults and, where the data permitted, assign consensus-preferred recurrence-interval and vertical slip-rate estimates for the faults and fault sections under review (Lund, 2005). As part of the review, the Working Group considered the paleoseismic trenching data available for the Salt Lake City segment.

**Earthquake timing.** The paleoseismic trenching data include stratigraphic evidence for seven paleoearthquakes (events T through Z) younger than the Bonneville flood, ca. 17.2 ka, and possibly an eighth event (event S) that occurred while Lake Bonneville was at or near its highstand at the Bonneville shoreline. The Working Group’s consensus for the timing of surface faulting on the Salt Lake City segment is shown in Table 4. Timing for earthquakes W, X, Y, and Z is from Black et al. (1996) and comes from the South Fork Dry Creek and Dry Gulch trench sites. The confidence limits for each earthquake were increased to accommodate the full range of limiting ¹⁴C ages used to constrain the timing of the earthquakes (Lund, 2005). The Working Group believes that the resulting ranges account for both the laboratory and geologic uncertainty associated with the timing of each earthquake.

McCalpin (2002) identified earthquakes S, T, U, and V at Little Cottonwood Canyon on the basis of a retrodeformation analysis of stratigraphic and structural relations exposed in the megatrench. No direct evidence (such as colluvial wedges, tectonic crack fills, or fault terminations) was found to document these earthquakes, and consequently their timing is only broadly constrained.

**Surface-faulting recurrence.** Table 4 also shows the inter-event recurrence intervals for the Salt Lake City segment, with associated confidence limits, determined from the earthquake chro-

### Table 4. Timing and Recurrence Intervals of Holocene and Latest Pleistocene Surface-Faulting Earthquakes on the Salt Lake City Segment of the Wasatch Fault Zone

<table>
<thead>
<tr>
<th>Surface-faulting event</th>
<th>Event timing* (cal yr B.P.)</th>
<th>Inter-event recurrence interval† (cal yr)</th>
<th>Post-event elapsed time (cal yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Event Z</td>
<td>1300 ± 650</td>
<td>N.A.</td>
<td>1300 ± 650</td>
</tr>
<tr>
<td>Event Y</td>
<td>2450 ± 550</td>
<td>1150 ± 900</td>
<td>N.A.</td>
</tr>
<tr>
<td>Event X</td>
<td>3950 ± 550</td>
<td>1500 ± 800</td>
<td>N.A.</td>
</tr>
<tr>
<td>Event W</td>
<td>5300 ± 750</td>
<td>1350 ± 900</td>
<td>N.A.</td>
</tr>
<tr>
<td>Event V</td>
<td>ca. 7.5 ka</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>(after 8.8–9.1 ka but before 5.1–5.3 ka)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event U</td>
<td>ca. 9 ka</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>(shortly after 9.5–9.9 ka)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event T</td>
<td>ca. 17 ka</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>Event S (?)</td>
<td>ca. 17–20 ka</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
</tbody>
</table>

†Confidence limits of recurrence intervals equal the square root of the sum of the squares of the individual confidence limits for each bracketing earthquake, rounded to the nearest 100 yr.

**Note:** Based on Lund (2005).
nology. The weighted mean recurrence for the three most recent inter-event intervals (from event W through event Z) is 1300 ± 400 cal yr (Lund, 2005). The timing of earthquakes U and V on the Salt Lake City segment is broadly constrained. McCalpin (2002) reports the U−V and V−W interevent intervals are both ~2 k.y., resulting in a similarly broadly constrained mean recurrence for surface faulting of 2 k.y. for mid- to early Holocene time.

The timing of event T is likewise broadly constrained. McCalpin (2002) reported a range in the interevent interval between events T and U of 7.1−9.6 k.y., with a mean of 8.4 k.y., indicating a long period of surface-faulting quiescence during earliest Holocene and latest Pleistocene time. However, McCalpin (2002) noted that the physical evidence of additional earthquakes in the gap may have been removed by alluvial fan erosion in the interval 9−10 ka and subsequently lost from the stratigraphic record.

Based on currently available information on earthquake timing and variability in the length of individual interevent intervals, the Working Group’s preferred recurrence-interval estimate for the Salt Lake City segment is 1300 yr, with a minimum estimate of 500 yr and a maximum of 2400 yr (Lund, 2005).

**Vertical slip rate.** Trenching investigations at Little Cottonwood Canyon and at the South Fork Dry Creek and Dry Gulch sites did not produce well-constrained net vertical displacement data. Swan et al. (1981) reported 47.6 (+30/−10) ft (14.5 [+10/−3] m) of net vertical displacement across the Wasatch fault zone determined from a scarp profile measured along the crest of the Bells Canyon glacial moraine a few hundred meters south of Little Cottonwood Canyon. Scott (1988) reports the age of the moraine as 18−26 ka. The resulting slip rate using that age is a preferred rate of 0.03 in/yr (0.7 mm/yr), with a minimum estimate of 0.02 in/yr (0.4 mm/yr) and a maximum of 0.06 in/yr (1.4 mm/yr) (Lund, 2005).

The Bells Canyon vertical slip rate is a long-term rate extending from the latest Pleistocene. Well-constrained vertical slip-rate data are lacking elsewhere on the Salt Lake City segment. Because the Bells Canyon long-term slip-rate estimate includes a possible period of seismic quiescence in the late Pleistocene (McCalpin, 2002), the Working Group’s slip-rate estimate for the Salt Lake City segment in the Holocene is higher than the longer-term rate at Bells Canyon. Based on currently available information on earthquake timing and displacement, the Working Group’s preferred vertical slip-rate estimate and confidence limits for the Salt Lake City segment in the Holocene is 0.05 in/yr (1.2 mm/yr), with a minimum estimate of 0.02 in/yr (0.6 mm/yr) and a maximum estimate of 0.16 in/yr (4.0 mm/yr) (Lund, 2005).

**SUMMARY**

This field trip exposes participants to examples of the wide variety of geologic hazards threatening the Wasatch Front urban corridor. Some of these hazards, such as landsliding examined at Stop 1 and debris flows at Stop 4, are geologically frequent—residents are vulnerable to similar hazards somewhere in the region almost every year, with risks increasing during periods of above-average precipitation, rapid spring snowmelt, and wildfires and cloudburst storms. Other hazards, such as the earthquake-related hazards of ground shaking discussed at Stop 2, liquefaction at Stop 3, and surface fault rupture at Stop 7, have longer recurrence intervals of several hundred years. Although infrequent, earthquake hazards can result in catastrophic consequences. One attempt to reduce earthquake hazards is the Utah State Capitol seismic retrofit (Stop 2). Flooding from Great Salt Lake (Stop 5) may be either climatically induced following cycles of several years, or earthquake induced, paralleling the pattern of Wasatch Front seismicity. Although hazards from large earthquakes are largely controlled by activity along the Wasatch fault zone, faults exposed at Stop 6 illustrate the potential for seismic hazards posed by faults with longer recurrence intervals.

Whatever the cause of geologic hazards, residents of the Wasatch Front accept risk regardless of location. Our field trip, and related studies, is part of the effort to enhance awareness of the hazards. Industry and local governments can then use the results of these studies to implement hazard-reduction measures.

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