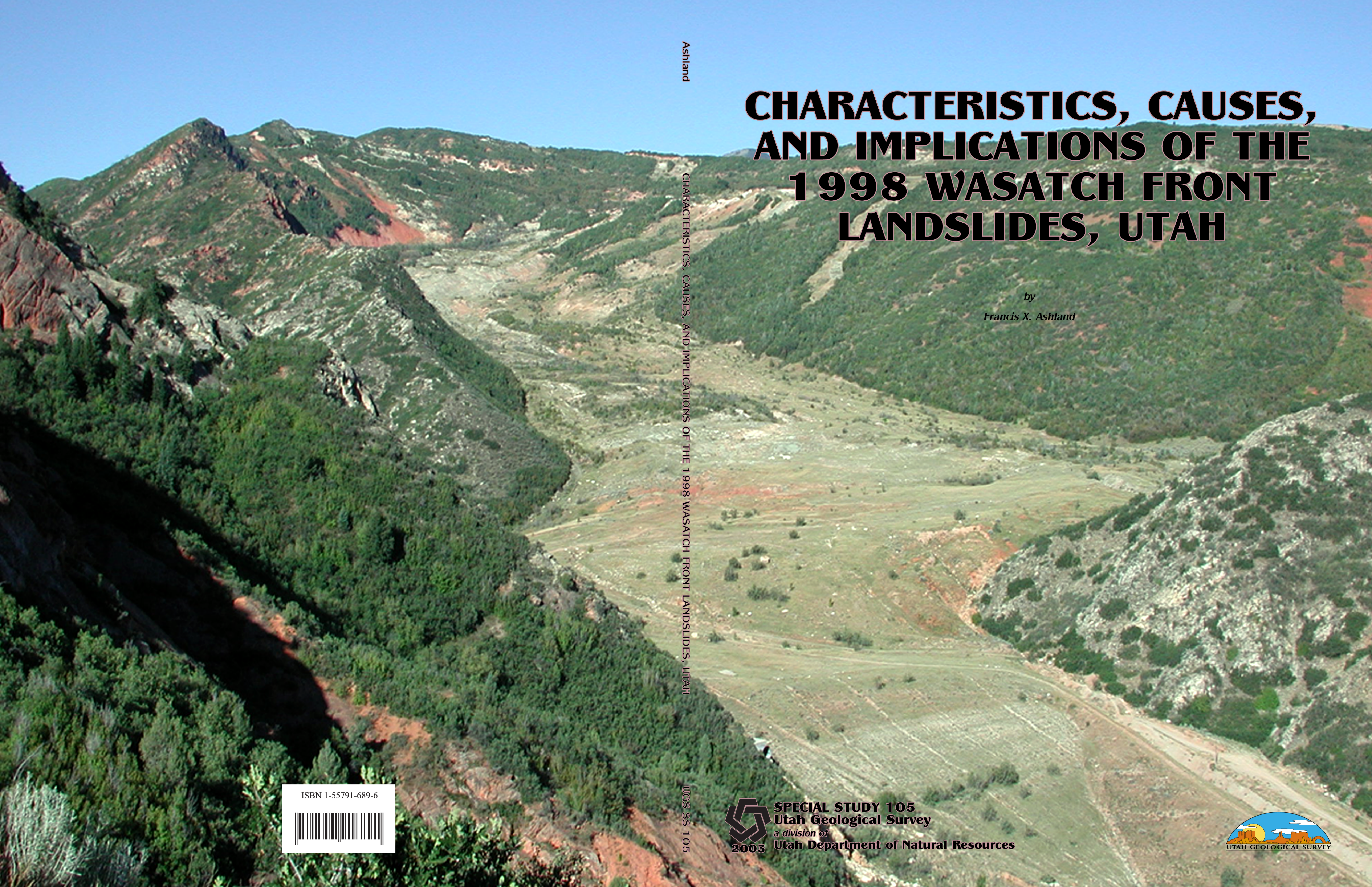

Ashland

CHARACTERISTICS, CAUSES, AND IMPLICATIONS OF THE 1998 WASATCH FRONT LANDSLIDES, UTAH

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by

Francis X. Ashland

SPECIAL STUDY 105
Utah Geological Survey
a division of
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2003

UTAH GEOLOGICAL SURVEY

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CHARACTERISTICS, CAUSES, AND IMPLICATIONS OF THE 1998 WASATCH FRONT LANDSLIDES, UTAH

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Francis X. Ashland

ABSTRACT

Damaging landslide movement in 1998 occurred near the end of a period of four or more successive years of above-normal precipitation, a period referred to as a precipitation period. The precipitation period began in 1995 in the northern and central Wasatch Front where mean annual precipitation for the period was about 116 percent of normal. The precipitation period began in 1993 in the southern Wasatch Front where mean annual precipitation for the period was 120 percent of normal. Landsliding extended from Ogden Valley in the north to Spanish Fork Canyon in the south. The majority of landslides occurred on modified hillsides near the urbanized part of the Wasatch Front. This study characterizes the 10 most significant 1998 Wasatch Front landslides; quantifies the rate, timing, duration, and amount of movement of some of these; defines the relation between precipitation and landslide movement; and examines the effects of the recent and previous precipitation periods on ground-water levels and the stability of pre-existing landslides. In addition, this study examines the use of instability threshold levels for understanding conditions that trigger landsliding along the Wasatch Front and evaluates the impacts of hillside development on these thresholds.

The 1998 Wasatch Front landslides consisted mostly of shallow to deep-seated earth slides. Landsliding consisted primarily of renewed movement of historical (suspended or dormant) and prehistoric (dormant) landslides. Over the entire region the rate of movement of these landslides ranged from very slow to moderate (0.005 to 230 centimeters per day), but ranged from very slow to slow (0.005 to about 5 centimeters per day) in the Wasatch Front urban corridor west of the Wasatch Range. The onset of landslide movement occurred during the spring snowmelt and contemporaneous above-normal precipitation between March and May during which a transient rise in ground-water levels is inferred. The majority of damaging landslide movement in 1998 occurred between March and May, but at several landslides movement continued through most of the year at a very slow rate. Landslide movement ranged from 0.3 inch (0.76 cm) to 159 feet (48 m) and was greatest at the large landslides in Spanish Fork Canyon. Small movement (only a few inches or less) caused severe damage where buildings straddled landslide deformation features.

The enhanced susceptibility of pre-existing landslides to renewed movement during a precipitation period was demonstrated in 1998 by the prevalence of reactivation versus landsliding in previously unfailed hillsides. Renewed movement of pre-existing landslides in 1998 was preceded by similar reactivation of landslides during the previous two precipitation periods. Landsliding in 1998 demonstrated the necessity to reevaluate the stability of pre-existing landslides. In several cases, pre-development slope-stability evaluations overestimated the static stability of the pre-existing landslides. Renewed movement suggests that these landslides were marginally stable to possibly metastable prior to 1998 movement, possibly as a result of hillside modifications. This study proposes a new landslide classification that characterizes the state of activity of the landslide rather than its geomorphic characteristics from which false inferences might be derived regarding the potential for reactivation.

Historical changes in annual precipitation in combination with hillside modifications such as landscape irrigation may have caused ground-water levels to rise and reduce stability at many Wasatch Front landslides. Between 1967 and 1998, three precipitation periods have occurred, each separated by only 8 years, and define a long-term "wet cycle." Each of these precipitation periods caused reactivation of pre-existing landslides in the Wasatch Front. The cumulative effects of the excess precipitation during this most recent "wet cycle" likely contributed to record high ground-water levels by the late 1990s in several wells in and near Salt Lake City that coincided with the onset of movement at two nearby landslides. Seasonal effects of landscape irrigation on ground-water levels have been measured at two of the 1998 Wasatch Front landslides, but at present the data are insufficient to precisely quantify the cumulative effects on ground-water levels and slope stability.

The recognition of instability thresholds at specific landslides might define conditions at which movement is triggered. Data on instability threshold ground-water levels provide direct information of the relation between ground-water-level fluctuations and slope stability, but practical limitations reduce the availability of such data. Instability threshold precipitation levels are more easily recognizable, but considerable uncertainties exist including a lack of understanding of the relation between precipitation and ground-water levels. Data on instability threshold precipita-

tion levels suggest the thresholds decrease with an increase in antecedent precipitation.

Pre-development slope-stability evaluations and land-use planning decisions by local governments did not prevent costly damage by the 1998 landslides. Commonly, the pre-development slope-stability evaluations did not consider and evaluate the effects of hillside modifications, particularly those influencing ground-water levels. At some landslides a lag period, sometimes exceeding a decade, occurred between hillside development and damaging landslide movement. Partly as a result, losses due to landsliding in residential subdivisions were incurred mostly by lot owners and local governments and not by the original developers. Despite post-failure slope-stability investigations at two of the most significant of the 1998 Wasatch Front landslides, effective stabilization has not been completed and continued losses due to ongoing or renewed movement are possible. The implied vulnerability of much of the hillside development in the Wasatch Front to landsliding, as demonstrated in 1998, suggests a more conservative approach to development in and near pre-existing landslides is warranted.

INTRODUCTION

Damaging landslide movement in the Wasatch Front area of Utah (figure 1) began in March 1998 and generally lasted through the end of June. At a few landslides, movement continued through the remainder of the year at very slow (less than 0.2 centimeters per day), imperceptible rates. Landsliding in 1998 consisted primarily of renewed movement of historical (suspended or dormant) and prehistoric (dormant) landslides. Examples of the former include reactivation of the 1997 Shurtz Lake and 1983 Thistle landslides in Spanish Fork Canyon whereas examples of the latter include the Sunset Drive landslide in Layton. The Springhill landslide in North Salt Lake is possibly the only damaging landslide that occurred on a previously unfailed hillside in the Wasatch Front in 1998. Most of the Wasatch Front landslides occurred on hillsides modified for residential development.

Direct and indirect losses in the Wasatch Front area resulting from 1998 landslides exceeded $1 million. As of March 1999, the estimated losses at the Sunset Drive landslide in Layton (figure 2) were $456,000 including the loss of two houses, one destroyed and demolished in 1998, another severely damaged and abandoned by early 1999. In North Salt Lake, an imperceptibly slow-moving landslide caused such severe damage to a house that it was subsequently condemned (figure 3). At least three other houses were also severely damaged and further damage is possible if landslide movement does not suspend. Damaging landslide movement also occurred in many other Wasatch Front communities including Salt Lake City and Provo where expensive attempts to prevent further movement or reduce property damage were made. Most of the landslide damage in 1998 occurred to residential property, but landsliding also damaged roads, a canal, and utility and culinary water lifelines. Most of the losses due to landsliding at residential properties were incurred directly by the homeowners because such losses are typically not covered by homeowner's insurance. However, some of the costs associated with damaging landslide movement in residential subdivisions were incurred by local governments.

In addition to the landslides described in this report, a damaging debris flow also occurred in Joes Canyon, a tributary to Spanish Fork Canyon (Ashland and others, 1999). The debris flow apparently killed large game animals and at least two cows. In 1999 livestock grazing ceased, at least temporarily, in the area.

The majority of the 1998 Wasatch Front landslides were likely triggered following a cumulative rise in ground-water levels resulting from four or more successive years of above-normal precipitation (a period defined in this paper as a precipitation period). Triggering of landslide movement likely coincided with a transient ground-water-level (pore-pressure) rise associated with the spring snowmelt and contemporaneous above-normal precipitation. In most Wasatch Front areas, 1998 was the wettest as well as the last year of the precipitation period. An increase in landslide activity began in 1997, following two to four successive years of above-normal precipitation.

This study examines the relation between the 1998 landslides and the 1995-98 precipitation period (1993-98 in Spanish Fork Canyon). Accordingly, this study investigates the significance of the most recent precipitation period in relation to the historical precipitation record, and compares it with the 1980-86 period. In addition, other causes of the 1998 landsliding are explored, most importantly hillside modification related to residential development. This study also examines several issues, and their implications, related to the 1998 Wasatch Front landslides including the susceptibility to reactivation of pre-existing landslides (Fleming and Schuster, 1985; Godfrey, 1985), consideration of the state of landslide activity, and the possibility of developing landslide-movement prediction tools based on an instability threshold concept (Godfrey, 1985; critical stability threshold of Harp and others, 1998). The majority of the landslides discussed occurred near urbanized areas of the Wasatch Front and consisted of either translational or rotational earth slides in pre-existing landslide areas. The discussion and conclusions are limited to these landslides and locations. The case histories presented provide new data intended to further the understanding of landslide hazards in the Wasatch Front.

PREVIOUS WORK

Landsliding along the Wasatch Front has been previously documented by numerous researchers (Pashley and Wiggins, 1972; Van Horn and others, 1972; Kaliser and Slosson, 1988). Van Horn and others (1972) described the geologic setting of many Wasatch Front landslides as well as some important causes and triggering mechanisms of these landslides. Pashley and Wiggins (1972) described prehistoric rock slides and several large landslide complexes in the northern Wasatch Front. Fleming and Schuster (1985) discussed the relation between landsliding in Utah during 1983 and 1984 and the cumulative rise in ground-water levels caused by long-term surplus (the term excess is used in this report) precipitation. Godfrey (1985) also recognized the role of cumulative surplus (excess) precipitation in causing a rise in ground-water levels in landslides and proposed that triggering of the 1983 landslides in the Wasatch Plateau of

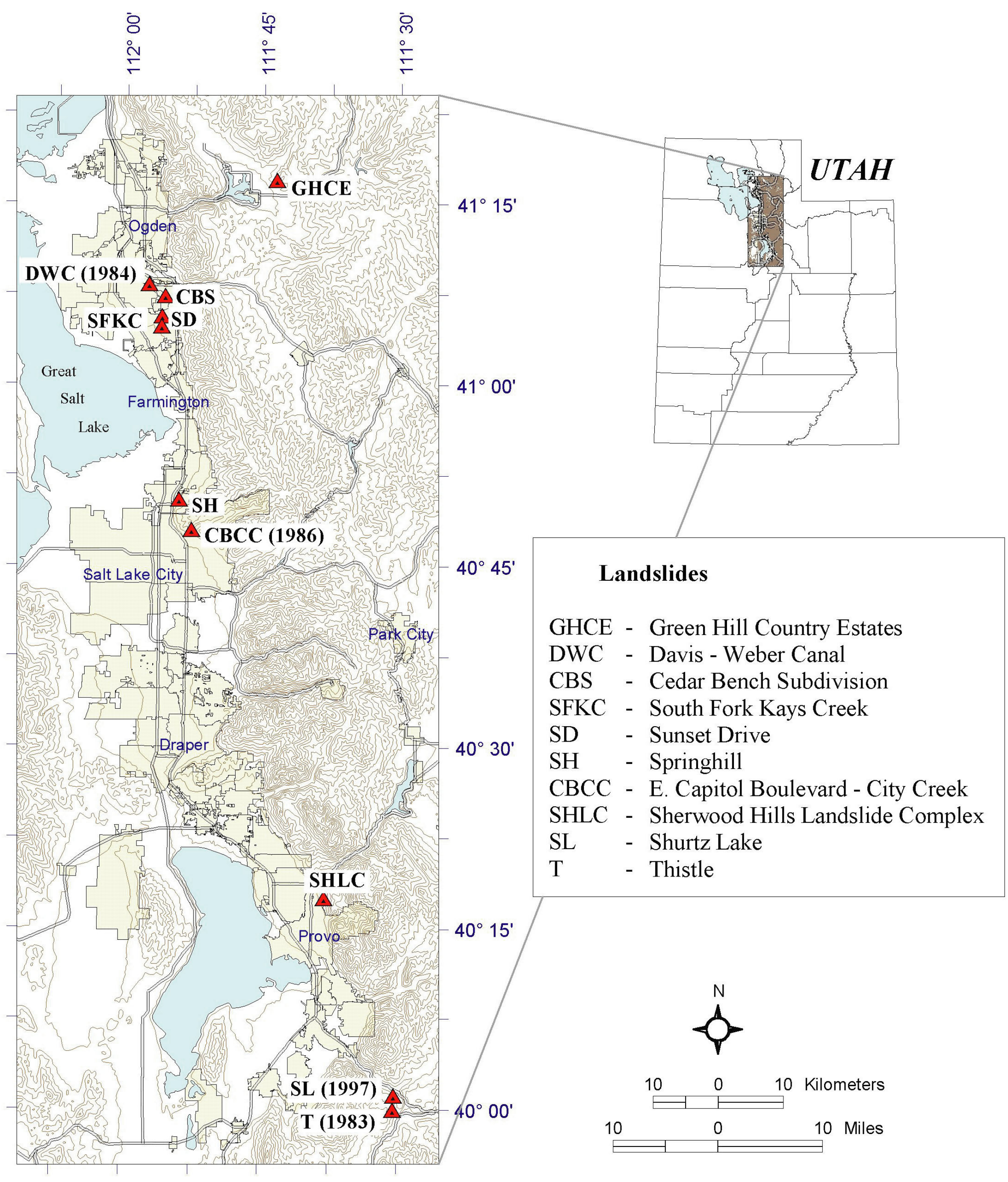


Figure 1. *Location map showing significant landslides of 1998. Year of most recent documented landslide movement prior to 1998 shown in parentheses.*

Figure 2. *Condemned house at the Sunset Drive landslide in Layton prior to demolition. House to left was abandoned in 1999.*

Figure 3. *Condemned house at the Springhill landslide in North Sale Lake prior to demolition. View of northwest corner of house showing severe foundation damage.*

Utah occurred as ground-water levels rose above an instability threshold. Kaliser and Slosson (1988) documented the landslides, debris flows, and other geologic effects associated with the 1983 "wet year." Their study explores the relation of landsliding to the 1980-86 precipitation period, with particular emphasis on 1982 and 1983, and a comparison of its conclusions to this paper is recommended. Harty (1991) summarized landsliding in Utah between 1987 and 1990, a period of below-normal precipitation. Hylland and Harty (1995) discussed landslide susceptibility along the Wasatch Front. Many of these researchers (Van Horn and others, 1972; Pashley and Wiggins, 1972; Kaliser and Slosson, 1988; Hylland and Harty, 1995) discussed the importance of landslide awareness in land-use planning and the use of landslide-hazard reduction measures in hillside development.

The 1998 Wasatch Front landslides were documented by the Utah Geological Survey (UGS) as well as by several consulting geologists and geotechnical engineers. The initial increase in landslide activity in 1997 was documented in several UGS reconnaissance investigations (Ashland, 1997b, 1998a, 1998b, 1998c; Giraud, 1998; Solomon, 1998). UGS reconnaissance investigations (Black, 1999a, 1999b; Giraud, 1999a, 1999b, 1999c; Solomon, 1999) of most of the 1998 Wasatch Front landslides documented landslide types, processes, causes, and triggering mechanisms. Detailed geotechnical slope-stability investigations, as described by Hylland (1996), were conducted at two of the 1998 Wasatch Front landslides (Terracon Consultants, Inc., 1998; Terracon, 1998). Additional landslide research included monitoring and analysis of landslide movement, precipitation, and ground-water-level data (Ashland and Horns, 1998; Ashland, 1999; Ashland and others, 1999; Giraud and Fadling, 1999).

THE USE OF INSTABILITY THRESHOLDS IN UNDERSTANDING SLOPE STABILITY

Slope stability can be described as the delicate balance between the forces acting to move soil and rock downslope and those resisting this movement. The factors contributing to instability, such as weak soil and rock or oversteepened slopes, are considered the *causes* of landsliding. Among these is commonly a single factor, which in some cases may be difficult to discern (Wasowksi, 1998), that initiates landslide movement and is referred to as the *triggering mechanism* or simply *trigger* (Varnes, 1978; Wieczorek, 1996). A trigger is traditionally considered a factor or event of short duration (Cruden and Varnes, 1996). Landsliding can be triggered by the culmination of a long-term process such as a gradual rise of ground-water levels to an instability threshold that reduces resisting forces and initiates movement (Godfrey, 1985; Fleming and Schuster, 1985). Whereas the cumulative rise in ground-water levels may occur over a long period, exceedance of a threshold level occurs during the final fraction of that time period prior to movement initiating. Thus, the triggering mechanism can be considered an event of short duration culminating a long-term process. In some cases, a short-duration precipitation event, such as rapid snowmelt or a cloudburst rainstorm, may cause the final critical rise in ground-water levels above an instability threshold, and thereby is the triggering event.

The recognition of instability thresholds has recently been promoted by landslide researchers (Cotton, 1998; Harp and others, 1998; Ashland and others, 1999) as a possible landslide-movement prediction tool. However, the concept of threshold-based landslide prediction in the western United States dates back to at least the early 1970s (Cleveland, 1971; Nilsen and Turner, 1975). Whereas the application of this potential tool to landslide prediction in Utah is still untested, its use will be explored in this paper. Hillsides are continuously changing complex systems where the transition from stability to instability is controlled by numerous geomorphological, physical, and human-related processes (Wasowski, 1998). Thus, inherent difficulties remain in the use of threshold-based landslide prediction (Wasowski, 1998) because a distinct combination of these processes controls the stability of each landslide.

Ground-Water Level

The ability of soil or rock to resist shear failure, or sliding, decreases with a rise in ground-water level or pore pressure. The upward pressure of ground water on the soil particles or rock reduces the normal stress acting on the surface of rupture along which sliding occurs. As the normal stress decreases, the frictional resistance of the soil or rock is reduced. This frictional resistance is the primary force resisting downslope movement in most pre-existing landslides and in many steep slopes. Thus, as the ground-water level rises, ground-water pressure increases reducing the normal stresses acting on the surface of rupture and the frictional resistance of the soil or rock.

In many recently active landslides in the western United States, a rise in ground-water levels above landslide-specific instability thresholds (Cotton, 1998) has been recognized or inferred as the triggering mechanism for movement (Cotton, 1998; Snell and Meldrum, 1999). Unfortunately, in Utah, detailed ground-water-level data exist for only two landslides (Terracon Consultants, Inc. 1998; Terracon, 1998). At both of these landslides, 1998 ground-water levels are available only for the latter part of the year and do not document the instability threshold ground-water levels that occurred in the spring of 1998.

Precipitation Level

In the absence of ground-water-level data, precipitation levels (Cleveland, 1971; Nilsen and Turner, 1975; Harp and others, 1998; Pasuto and Silvano, 1998; Wasowski, 1998) may substitute as a method of indirectly assessing the potential rise in ground-water levels above instability thresholds and predicting landsliding. The basic assumption of this method is that ground-water level (pore pressure) rises as excess precipitation infiltrates into the ground during periods of above-normal precipitation. Ashland and others (1999) suggested that cumulative precipitation could be used in conjunction with landslide movement data to predict whether movement was likely at a specific landslide. At the Shurtz Lake landslide in Utah County, Ashland and others (1999) used cumulative precipitation during an informal landslide water year (LWY) for Utah, which begins September 1, to identify the instability threshold precipitation level at which landsliding triggered. The September start date of the landslide water year allows for the tracking of antecedent precipitation prior to the months of March through May during which landslide movement typically triggers in the Wasatch Front.

Application of this method is problematic because significant uncertainty exists regarding the duration of precipitation required to cause ground-water levels to rise above an instability threshold and whether other factors, including landslide geometry and boundary conditions, may control stability. Ground-water-levels from the Springhill landslide in North Salt Lake show rapid but transient response to a short-duration rainstorm event. At one well, ground-water levels rose over 1.5 feet (0.5 m) in response to an intense, short-duration rainstorm event (figure 4). Such a rapid rise may trigger movement in some Wasatch Front landslides; however, in others an antecedent rise in ground-water levels, either due to longer term or seasonal periods of above-normal precipitation, may be required before movement initiates. Successive years with above-normal precipitation may cause a cumulative rise in ground-water levels in hillsides, particularly where hillside modifications, such as flattening parts of slopes for home sites, promotes infiltration of the excess precipitation (Kaliser and Slosson, 1988). In other areas, periods of above-normal precipitation of only moder-

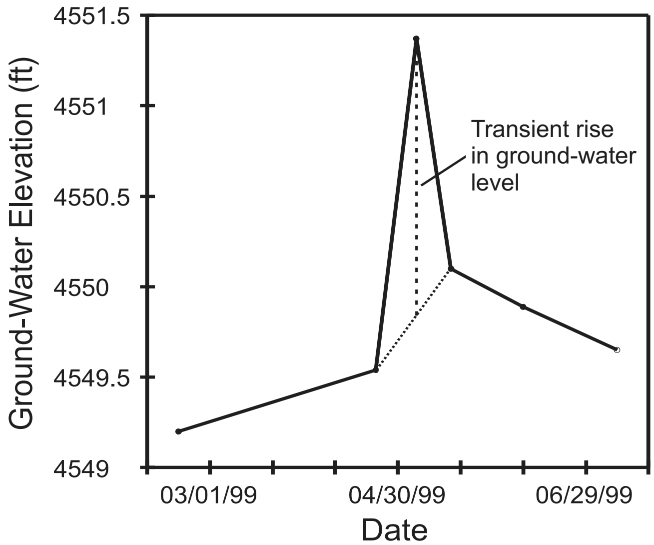


Figure 4. Transient ground-water-level rise associated with a four-day rainstorm in May 1999. Ground-water-level data from well P-4 (see figure 19 for borehole B-4 location) at Springhill landslide, North Salt Lake.

ate (seasonal) duration may trigger landsliding. The failure of one landslide to move when nearby landslide movement occurs is also problematic and may restrict recognition of instability threshold precipitation levels to specific landslides.

THE 1995-98 PRECIPITATION PERIOD

The 1998 Wasatch Front landslides occurred in the latter part of a period of successive years having above-normal precipitation. The period began in 1995 in most areas, but began in 1993 in Spanish Fork Canyon. The 1998 calendar year was the wettest of the period in many Wasatch Front areas. In Ogden (northern Wasatch Front), 1998 was the third wettest year since 1902. In Salt Lake City (central Wasatch Front), 1998 was the second-wettest year on record (since 1875). Annual precipitation was only about a half-inch less than in 1983, the wettest year on record. In Spanish Fork Canyon (southern Wasatch Front), 1998 was the third-wettest year on record (since 1928), surpassed only by 1983 and 1982, the wettest and second-wettest years, respectively.

During the 1995-98 precipitation period, excess precipitation fell during each of the calendar years. Mean annual excess precipitation equaled about 3.9 inches (9.9 cm) in Ogden, 2.6 inches (6.6 cm) in Salt Lake City, 4.3 inches (10.9 cm) at the Deer Creek Dam in Provo Canyon, and 4.1 inches (10.4 cm) in Spanish Fork Canyon (table 1). Although mean annual excess precipitation was lower in the 1995-98 precipitation period than in the 1980-86 precipitation period, landslides were triggered at many Wasatch Front localities in 1997 and 1998.

In several Wasatch Front locations, precipitation periods are unprecedented in the historical precipitation records prior to 1980. In Ogden, no precipitation periods occurred between 1902 and 1979, and in Spanish Fork Canyon no periods occurred between 1928 and 1979. Seven or eight periods have occurred in Salt Lake City since 1875 depending on the mean (or normal) value used (table 2), but only two periods occurred in the 50 years prior to the 1980-86 precipitation period. The 1944-48 precipitation period contributed to a significant increase in mean annual precipitation during the 1940s relative to the previous decades. Mean annual precipitation in the 1940s was 17.18 inches (43.6 cm) and slightly higher than in the 1980s. Unfortunately, little is documented regarding landsliding in the 1940s. A second period occurred between 1967 and 1971, a period during which landslide movement is documented along the Wasatch Front.

DAMAGING LANDSLIDES OF THE 1995-98 PRECIPITATION PERIOD

Damaging landsliding in the Wasatch Front area began in the latter part of the 1995-98 precipitation period and locally continued through 1999. Most of the documented 1997 and 1998 landslides were dormant in 2000. However, high ground-water levels remaining from the 1995-98 precipitation period may have contributed to both new landslide movement and reactivation of a few of the 1998 landslides in 2000 and 2001. In some cases, movement of the 1999 and later landslides may have occurred in 1998, but was sufficiently minor to preclude detection. Indeed, the exact number of active landslides between 1997 and 2001, particularly in remote and undeveloped areas, is unknown.

Damaging landslide movement began in 1997, the third year of the 1995-98 precipitation period. Landslide movement first occurred in January 1997 along the east bank of the Bear River in Honeyville, damaging farm facilities and agricultural land (Ashland, 1998a). Landslide movement suspended, but the landslide reactivated in August causing further damage (Solomon 1998). In early May, the Shurtz Lake landslide in Spanish Fork Canyon moved, disrupting transmission on two sets of high-voltage power lines (Ashland, 1997b). Damaging landsliding also occurred outside the Wasatch Front in 1997 (Ashland, 1998b, 1998c). A landslide on the south slope of Chalk Creek east of Coalville moved in May 1997, diverted the creek, and threatened the county road as the creek eroded the opposite bank (Ashland, 1998b). The potential for the landslide to block the creek or divert it to a low-lying area north of the road created a possible flood hazard to buildings and property downstream until movement suspended later that summer. A landslide in Duchesne County threatened Bluebell Road (Ashland, 1998c), the main transportation corridor connecting the town of Altamont to the city of Roosevelt.

The following section of this report describes the significant landslides of 1998. Additional smaller landslides in 1998, which had only minor impacts or occurred solely in fill soil, are not discussed in this report. Several of the 1998 landslides reactivated in subsequent years. Movement data from 1999 and the early part of 2000 are discussed below. Landslides are not described in this report for which movement triggered in 1999 or later years or that was not documented in 1998.

Table 1. *Comparison of mean and excess precipitation during the 1980-86 and 1995-98 precipitation periods.*

Location	Precipitation Period	Mean Annual Precipitation for Period (percent relative to normal)[1]	Mean Annual Excess Precipitation[2] (inches)	Wettest Year in Period
Ogden	1980-86	133[3]	7.4[3]	1983
(Northern WF)	1995-98	117[3]	3.9[3]	1998
Salt Lake City	1980-86	123	3.7	1983
(Central WF)	1995-98	116	2.6	1998
Deer Creek Dam	1980-86	136	8.2	1982
(South-Central WF)	1995-98	119	4.3	1996
Spanish Fork Cyn.	1980-86	132	6.6	1983
(Southern WF)	1993-98	120	4.1	1998

[1]*Normal annual precipitation calculated using monthly normal precipitation reported by National Weather Service Salt Lake City office. Monthly normal precipitation equals average for thirty-year period from 1961 to 1990. Annual precipitation data from Western Regional Climate Center and National Weather Service Salt Lake City office.*
[2]*Mean annual precipitation for period minus normal annual precipitation.*
[3]*One year of precipitation data missing for period.*
Abbreviation: WF - Wasatch Front.

Table 2. *Summary of historical precipitation periods in Salt Lake City, 1875-1999.*

Period	Number of Years	Mean Annual Precipitation for Period[1] (inches)	Mean Annual Excess Precipitation[2] (inches)	Cumulative Excess Precipitation (inches)
1875-78	4	20.26	4.24	16.97
1896-99	4	17.21	1.18	4.74
1906-09	4	20.26	4.24	16.95
1922-25	4	19.16	3.14	12.54
1944-47	4	18.01	1.99	7.96
1967-71[3]	5	18.48	2.46	12.28
1980-86	7	19.83	3.81	26.68
1995-98	4	18.74	2.72	10.89
Average	—	18.99	2.97	13.63

[1]*Annual precipitation data from Western Regional Climate Center and the National Weather Service Salt Lake City office.*
[2]*Difference between precipitation period mean and mean annual precipitation for period from 1875 to 1999 (16.02 in.), and therefore values vary slightly from table 1.*
[3]*The period between 1967-71 qualifies as a precipitation period if annual precipitation is compared to mean of the historical precipitation record (1875-1999) rather than the normal annual precipitation value reported by the National Weather Service Salt Lake City office.*

THE SIGNIFICANT LANDSLIDES OF 1998

In 1998, damaging landslides in the Wasatch Front occurred from Ogden Valley in the north to Spanish Fork Canyon in the south. The 10 most significant of these (figure 1) are described below. The descriptions include information obtained during detailed post-failure slope-stability studies, and from observations and monitoring subsequent to the initial reconnaissance investigations by the UGS. Information is also included on four landslides, the East Capitol Boulevard-City Creek, Sherwood Hills, Shurtz Lake, and Thistle landslides, for which UGS reconnaissance investigation reports were not completed in 1998.

Sunset Drive Landslide, Layton

Introduction

Renewed movement of the Sunset Drive landslide in Layton began sometime in the early spring of 1998 (Giraud, 1999a). Recent landslide movement had reportedly occurred at this site earlier in the 1990s. Movement at the head of the landslide in 1998 caused building distress to three houses at the top of the landslide and affected landscaped and naturally vegetated areas on four adjacent lots. The most severely damaged house (figure 5), on lot 105 (1851 Sunset Drive) (figure 6), was condemned in April 1998 by Layton City and subsequently demolished later that year. Damage to the house at lot 104 caused the owners to abandon the house in 1999. In 1998, Layton City hired a geotechnical consulting firm to evaluate slope stability and recommend stabilization measures in an effort to protect properties at the crown of the landslide. The city attempted to form a special improvement district to fund the measures recommended by the consultant (Terracon Consultants, Inc., 1998), but affected property owners were unwilling to financially participate in this effort. By the end of 1999, no additional effort had been made to stabilize the landslide and reduce the risk to the houses on the five remaining affected lots. As of March 1999, estimated direct losses resulting from this landslide were about $456,000 (table 3).

Geology

The Sunset Drive landslide is in a mapped prehistoric landslide area (Lowe, 1989) on a northwest-facing bluff that has an average slope of about 30 percent. The bluff was formed by latest Pleistocene to Holocene incision of the North Fork of Kays Creek into lake (lacustrine) sediments that were deposited in late Pleistocene Lake Bonneville as part of the Weber River delta (Giraud, 1999a). Nelson and Personius (1993) mapped lacustrine sands overlying lacustrine silt, clay, and fine sand in the vicinity of the site; however, borehole logs (Terracon Consultants, Inc., 1998) suggest that the overlying lacustrine sands are absent at the site. Erickson and others (1968) mapped the surficial slope soils along the bluff as eolian (wind-blown origin). Terracon Consultants, Inc. (1998) interpreted near-surface natural silty clay deposits on the slope to be of this origin. Alternatively, these soils may be either shallow colluvium or landslide deposits. Fill soil exists at the crest of the bluff. Two boreholes show that the fill is about 12.5 feet (4 m) thick near the crest of the bluff and downslope of the main scarp of the 1998 landslide. Placement of the fill soil at the crest of the bluff is a significant modification to the natural slope and likely decreased slope stability by the addition of a surcharge load at the head of the prehistoric landslide.

Landslide Description

The extent of the landslide is somewhat problematic because of the lack of continuous landslide deformation fea-

Figure 5. Damage to house on lot 105 at the Sunset Drive landslide in Layton. View of west side of house showing severe damage. Main scarp of the landslide is visible in foreground (arrow).

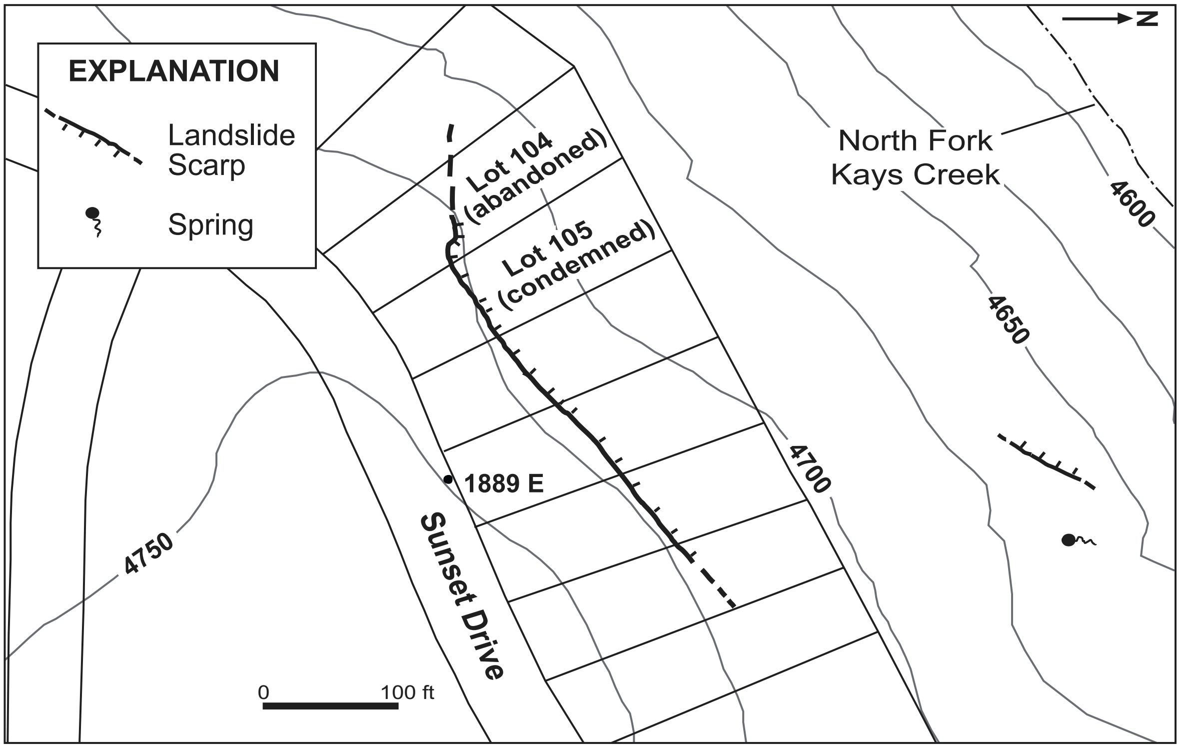


Figure 6. *Sketch map of the Sunset Drive landslide showing residential lots, topography, and approximate scarp locations. Well 1889 E location shown. Well B-2 located near "t" in lot 105. Modified from Giraud (1999a).*

Table 3. *Estimated direct losses at the Sunset Drive landslide.*

Description of Loss	Estimated Value (dollars)	Source of Loss Estimate
House at lot 105 (demolished)	$200,000	*Deseret News*, April 21, 1998
House at lot 104 (abandoned)	$200,000[1]	*Deseret News*, April 21, 1998
Demolition of house at lot 105	$10,000	*Standard Examiner*, October 20, 1998
Geotechnical slope-stability investigation	$40,000	*Deseret News*, October 15, 1998
Relocation fees for family at lot 105	$6,000	*Standard Examiner*, September 10, 1998
Subtotal Value of Direct Losses (as of March 1999)	$456,000	

[1]House subsequently re-occupied and thus, loss may be overestimated.

tures defining its boundaries. Giraud (1999a) identified only the main scarp, one minor scarp, and some minor ground cracks on the bluff related to the landsliding (figure 6). A second minor scarp identified on the lower slope of the bluff was attributed by Terracon Consultants, Inc. (1998) to an earlier episode of movement sometime in the 1990s. Based on the main scarp length, the 1998 landslide was about 400 feet (122 m) wide at its head. The width of the landslide likely exceeded 450 feet (137 m) lower on the slope, but the lack of well-defined flank features prevents a more exact estimate. The position of the toe of the landslide is uncertain, possibly because movement in 1998 was confined (defined by Cruden and Varnes [1996] as landsliding in which the surface of rupture has not emerged from below ground to form the toe).

Two alternatives for landsliding have been proposed and evaluated for the Sunset Drive landslide (Chen & Associates, 1987; Terracon Consultants, Inc., 1998; Giraud, 1999a). Giraud (1999a) suggested that the minor scarps lower on the bluff and the main scarp may be part of one large landslide. The toe of such a landslide would likely be at the base of the bluff near the North Fork of Kays Creek. This large landslide would be about 500 feet (152 m) long and more than 450 feet (137 m) wide, as described above, or 25,000 square yards (20,900 m^2) in area. Alternatively, the minor scarps lower on the slope may be unrelated to the main scarp at the crest of the bluff (Terracon Consultants, Inc., 1998; Giraud, 1999a).

In this case, the lower scarp identified by Giraud (1999a) would be the main scarp of a local landslide lower on the bluff. Terracon Consultants, Inc. (1998) interpreted the upper scarp to be the main scarp of a local slope failure that has its toe near the base of the fill soil and cite a slight pressure ridge below the base of the fill as evidence supporting this interpretation. In addition, Terracon Consultants, Inc. (1998) interpreted inclinometer data as providing unequivocal evidence for movement only in the upper part of the bluff near the base of the fill.

Site Investigations and Development

All of the affected properties at the head of the landslide are part of a second phase of the Heatherglen subdivision (Chen & Associates, 1987). Whereas no geotechnical investigation report has been discovered for this phase, a geotechnical investigation (Chen & Associates, 1987) was performed for adjacent later phases (3 and 4) of the subdivision to the south and east that addressed slope stability prior to completion of phase 2 construction. The geotechnical report did not indicate a prehistoric landslide at the site, but identified four nearby landslides in bluffs similar to the Sunset Drive site. The report also indicated the landslides were either rotational earth slides or lateral spreads, the latter caused by earthquake-induced liquefaction.

Chen & Associates' (1987) assessment of the natural slope stability of the bluff adjacent to the Sunset Drive landslide indicated relative stability under static conditions (a factor of safety of 1.3), but recognized the likelihood for landsliding during earthquake ground shaking. Chen & Associates (1987) also analyzed the potential for localized landsliding in the upper part of the natural bluff and concluded that it was less likely than the potential for deep-seated landsliding of the entire slope under natural conditions. Based on its conclusions, Chen & Associates (1987) characterized proposed lots along the crest of the bluff as "high risk," primarily because of the potential for earthquake-induced landsliding. One of these "high risk" lots, lot 113, is only three lots away, or about 165 feet (50 m), from the northeast tip (end) of the main scarp of the Sunset Drive landslide. Chen & Associates (1987) recommended disclosure of the possibility of earthquake-induced landsliding to potential homebuyers. At the time of these recommendations, Chen & Associates (1987) indicated that houses had been completed or were under construction in phase 2 of the subdivision. Based on Chen & Associates (1987) assessment, it seems likely that lots in phase 2 of the subdivision north of Sunset Drive should have, at a minimum, also been characterized as "high risk."

Causes and Trigger

Hillside modifications, particularly emplacement of fill soil at the crest of the natural slope and the cumulative effects on ground-water levels from landscape irrigation and redirected runoff, likely contributed to landsliding in 1998. Preliminary analysis conducted as part of this study indicated that the additional surcharge load from the soil fill at the crest of the slope decreased, although nominally, the stability of the slope. Further destabilization of the bluff likely resulted from a cumulative ground-water-level rise caused by landscape irrigation, particularly in the summer when ground-water levels would otherwise have lowered, and redirected runoff over the 11 years that the phase 2 portion of the subdivision existed. Atop the bluff, bluegrass-type lawns underlain by permeable fill likely promote excessive watering, particularly in the dry summer months. Locally, runoff from paved areas and rooftops was concentrated and redirected toward the bluff instead of infiltrating into the natural soils atop the bluff. On at least one property (lot 104), plastic drain pipes carried water from downspouts to the upper part of the bluff.

A cumulative rise in ground-water levels is suggested by comparison of ground-water levels near the crest of the bluff reported in the Chen & Associates (1987) report with those in the Terracon Consultants, Inc. (1998) report (figure 7). Ground-water-level measurements from April 1987 indicated ground water ranged from about 22 to 33 feet (7-10 m) below the natural ground surface. The depth to ground water in the closest 1987 borehole to the 1998 Sunset Drive landslide was 22 feet. Ground-water levels in June through September 1998 (Terracon Consultants, Inc., 1998) relative to the base of the fill, a reasonable estimate of the pre-fill (1987) ground surface, were about 5 to 23 feet (1.4-7 m) higher than in 1987.

Consideration of the timing of the ground-water-level measurements is important in comparing the 1987 and 1998 ground-water levels. Seasonally high ground-water levels in natural slopes likely occur in the spring (late March through early June) following the snowmelt. Thus, the April 1987 ground-water levels, which predate the majority of development in the subdivision, should represent the seasonal high ground-water levels for the year. Ground-water levels in the Terracon Consultants, Inc. (1998) report were measured between June and September, 1998, typically the driest months of the year, and thus, a period where ground-water levels would be expected to be lower than in April.

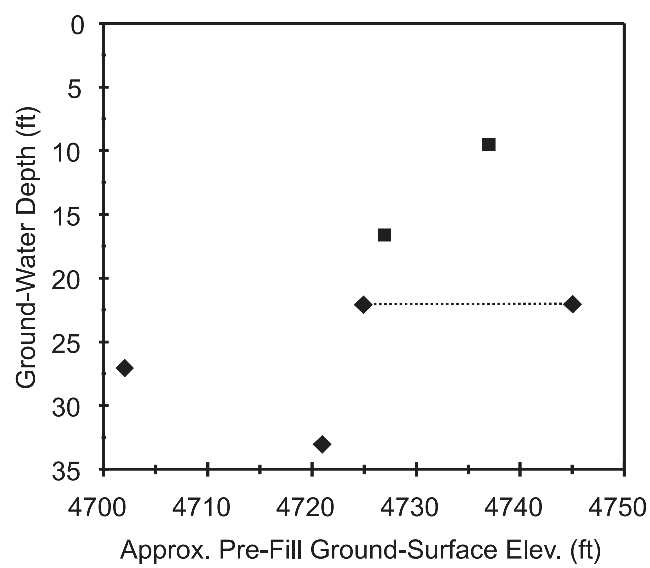


Figure 7. *Comparison of 1987 and 1998 ground-water depths from approximate pre-fill ground surface at the Sunset Drive landslide. Ground-water is significantly shallower (higher) in 1998 (squares) than in 1987 (diamonds) possibly because of a cumulative rise in ground-water levels from landscape irrigation. Seasonal difference in the timing of ground-water-level measurements supports this interpretation because ground water is typically shallower in the late spring (1987 measurements) than in the late summer and early fall (1998 measurements). Dotted line shows uncertainty in borehole elevation.*

Ground-water levels in 1987 and 1998 also likely reflect a natural rise in ground-water levels associated with the precipitation periods that preceded or were occurring at the time of the measurements. Although precipitation in 1987 was below normal in both Salt Lake City and Ogden, the ground-water levels were measured only one year following the 1980-86 precipitation period and therefore ground-water levels in the spring of 1987 were likely higher than normal. Ground-water levels elsewhere in Weber and Davis Counties (Kenney, 1999) in wells where the effects of large water withdrawals are absent, were generally higher in the mid-1980s than in the late 1990s. However, by 1987 ground-water levels were declining from their peak levels. Peak ground-water levels at the Sunset Drive landslide associated with the 1995-98 precipitation period likely occurred in 1998 as indicated by declining water levels since measurements began in 1998 (July 1998 levels are higher than July 1999 levels). Thus, the higher ground-water levels in 1998 were partially caused by the cumulative effects of the 1995-98 precipitation period.

The contribution of the cumulative increase in ground-water levels caused by the 1995-98 precipitation period cannot be neglected as a probable cause of the 1998 landslide. Nevertheless, the component contributed by infiltration of excess precipitation during the precipitation period to the higher ground-water levels in 1998 relative to 1987, as discussed above, is uncertain. Most likely, the combined contribution of infiltration of excess precipitation during the precipitation period and excess landscape irrigation since the late 1980s resulted in the instability threshold ground-water level being exceeded. Whether landslide movement would have been triggered during the 1995-98 precipitation period in the absence of landscape irrigation is uncertain.

East Capitol Boulevard-City Creek Landslide, Salt Lake City

Introduction

Renewed movement of the East Capitol Boulevard-City Creek (CBCC) landslide (figure 8) in Salt Lake City was reported in May 1998, but initiated prior to 1998 (James Nordquist, Applied Geotechnical Engineering Consultants, Inc., verbal communication, May 1998). Landslide movement resulted in local retrogressive failure, created a new main scarp that exceeded 10 feet (3 m) in height by late July 1998 (figure 9), and caused property damage in the backyard of a residential lot on 1000 East Capitol Boulevard. Hairline fractures in the foundation wall of a brick house on the lot suggested minor ground deformation in the crown area as downslope movement of the head of the landslide removed lateral support. Continued movement of the landslide throughout most of 1998 increased the height of the main scarp, potentially destabilizing the crown area. At year's end, following a geotechnical investigation, several deep caissons, some exceeding 75 feet (23 m), were installed in the crown area of the landslide in an attempt to prevent significant damage to the house and the remainder of the lot. Landslide movement continued through the first half of 1999, but suspended by June 1999. The potential for renewed movement still threatens three adjacent residential properties.

Geology

The CBCC landslide predates the earliest (1937) aerial photographs of the Salt Lake City area (Van Horn and others, 1972). The landslide is on a southeast-facing slope that was formed by incision of a small unnamed drainage that flows into City Creek. The drainage was once ephemeral (Dames & Moore, 1979), but now flows year-round (perennial). The head of the landslide is underlain by fine-grained lacustrine sediments (Dames & Moore, 1979, 1981; Personius and Scott, 1992). The lacustrine sediments in turn overlie soils derived from Tertiary (Paleogene) sedimentary and volcanic rocks (Personius and Scott, 1992). A north-striking Tertiary or younger fault is inferred to underlie the landslide and the drainage north of the landslide (Dames & Moore, 1979). Test pits and boreholes in the upper part and the crown area of the landslide indicated soils consist primarily of interbedded silt, silty sand, sand, and minor gravel (Dames & Moore, 1979, 1981). Locally, over 11 feet (3.4 m) of coarse-grained fill soil was exposed in the main scarp of the landslide in 1999.

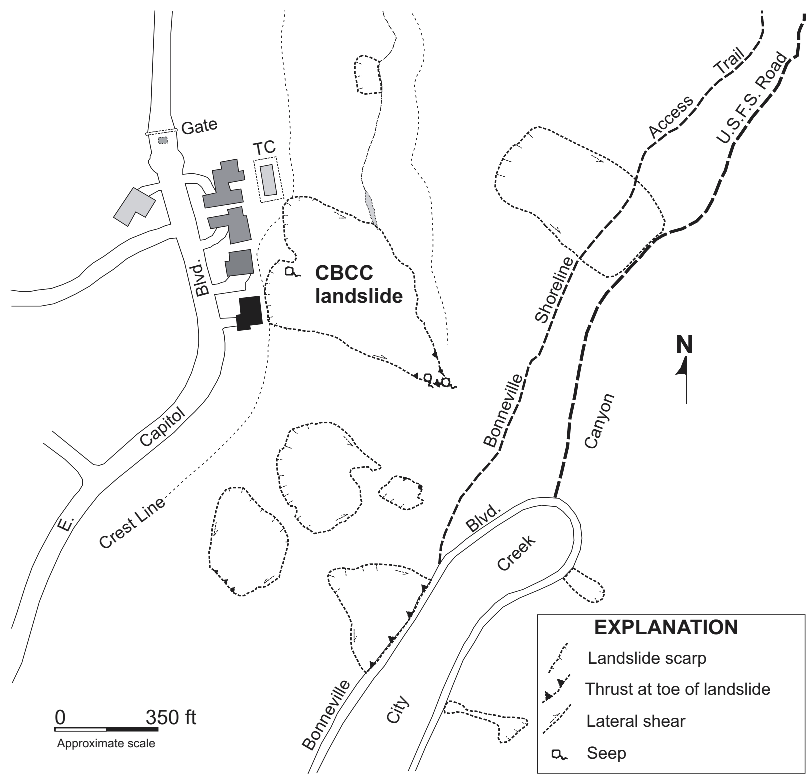
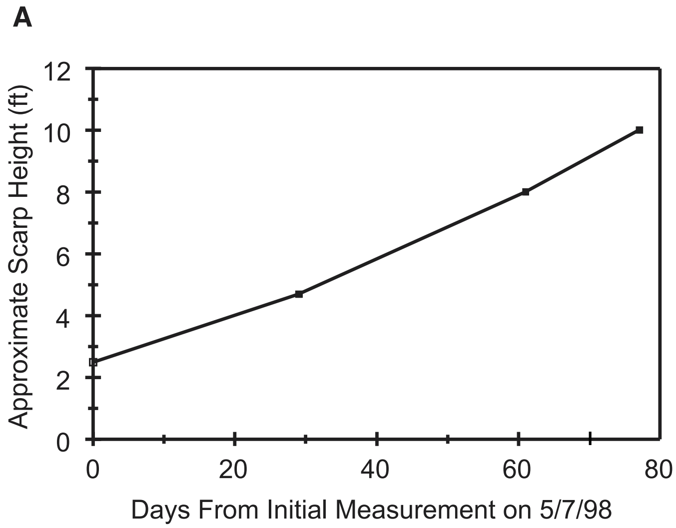


Figure 8. *Sketch map of the East Capitol Boulevard-City Creek (CBCC) landslide and nearby landslides in Salt Lake City. Of the seven landslides on the southeast-facing slope in City Creek Canyon, only the CBCC landslide moved significantly in 1998. Landslide movement caused severe damage to landscaped areas in the rear of the lot at 1000 East Capitol Boulevard (black house) and threatened nearby residential properties (houses in dark gray). Most other nearby landslides are historical including those to the south of the CBCC landslide which moved during the 1980-86 precipitation period. Unnamed tributary drainage (shown to the north of the landslide) has historically been blocked by the landslide as evident from broad floodplain deposits directly upstream of the left flank. Tennis court (TC) along northern head of landslide is threatened by retrogressive failure.*



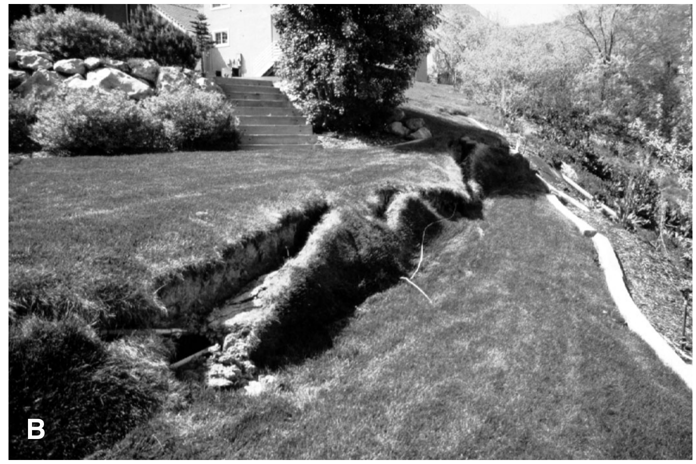


Figure 9 . *Cumulative offset along the main scarp of the CBCC landslide. (A) Plot showing approximate cumulative offset between May 7, 1998, and July 23, 1998. (B) View toward the north of the main scarp on May 7, 1998, and (C) June 5, 1998.*

Landslide Description

The landslide is funnel shaped in plan view (figure 10). At its head, the 1998 landslide was nearly 400 feet (122 m) wide, but was only about 45 feet (14 m) wide at the toe. In 1979, the landslide was about 480 feet (146 m) long, from the toe to the main scarp, about 360 feet (110 m) wide at the head, and about 160 feet (49 m) wide at its lower part upslope of the toe. Profiling of the landslide in 1999 indicated that it was about 570 to 580 feet (174-177 m) long, or about 90 to 100 feet (28-31 m) longer than in 1979. The 1979 landslide area was approximately 14,000 square yards (11,700 m^2) with an estimated volume between 100,000 and 180,000 cubic yards (76,000-138,000 m^3). By 1998, the landslide area was about 19,000 square yards (16,000 m^2) with an estimated volume between 130,000 and 240,000 cubic yards (99,000-184,00 m^3). Dames & Moore (1979) mapped a south-southeast-trending finger-like ridge that separated the uppermost left (north) part of the head from the remainder of the landslide head area farther southwest. By 1998, the main scarp cut across part of the mapped ridge. The average slope of the landslide in 1999 was about 34 percent.

The difference in length between 1999 and 1979 suggests an average annual rate of movement of about 4.5 to 5 feet per year (1.4-1.5 m/yr). In 1999, the landslide moved 5.8 feet (1.8 m) at the toe. Although complete movement data are unavailable for 1998, the landslide moved a little over 8 feet (2.4 m) between June 5, 1998, and June 5, 1999 (figure 11). Total movement at the toe of the landslide likely exceeded 8 feet (2.4 m) in 1998 based on the measured movement for the period between late April and early June in 1999 during which over 5 feet (1.5 m) of movement occurred.

The internal structure of the landslide is complex. In the upper right (west) head of the landslide, rotational blocks or slivers have calved off the former (pre-1998) crown. The lowermost rotational block has a small toe thrust at its base where the block overrides the uppermost main body of the landslide. At the northern part of the head, the main scarp slope has locally failed by earth flow. The earth-flow deposit is relatively thin, less than about 2 feet (0.6 m) thick, at the base of the main scarp. Downslope of the main scarp, the relatively gently sloping upper part of the main body is characterized by extensional (stretching) deformation features including transverse ground cracks, mostly downslope-facing minor scarps, and diagonal shear scarps on the left (east) side of the landslide. Compound thrust-shear zones bound the landslide on the northeast and southwest flanks. The lower part of the landslide consists of both contractional (shortening) and extensional deformation features. A stacked series of thrusts exists at the toe of the landslide. The main thrust is at the base of the steep slope formed by these stacked thrusts. In 1998, a single, thin, frontal thrust sheet extended downslope of the main thrust. Minor scarps and lateral shears deform the landslide mass upslope of the toe area. Flattening of the local slope above the toe by these features occurs in response to the oversteepening of the slope caused by thrusting and ground tilting.

Figure 10. *Oblique view of the CBCC landslide. View is toward the northwest. Dashed line shows approximate boundary. Toe of landslide is concealed by vegetation in foreground.*

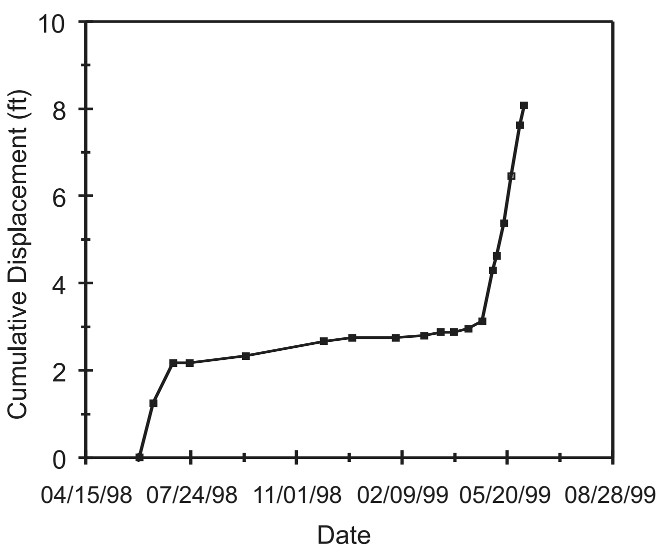


Figure 11. *Cumulative displacement of the CBCC landslide toe from June 5, 1998 to June 5, 1999.*

Site Investigations and Development

The CBCC landslide was mapped in a geotechnical study performed for the design phase of the Ensign Downs subdivision (Dames & Moore, 1979). Dames & Moore (1979) recognized the potential for renewed landslide movement and recommended building setbacks from the slopes surrounding most of the landslide. Recommended building setbacks ranged from about 25 to 50 feet (8-15 m) from the 1979 scarp of the landslide and were based on a 2H:1V projection from the base of the slope. For lots along the abutting slope northeast of the landslide, Dame & Moore (1981) recommended a minimum setback of 25 feet (8 m) from the crest and that downspouts drain directly to the streets rather than onto the slope.

Design recommendations for the four lots at the crown of the landslide appear to have been either loosely followed or overlooked. An adequate building setback was not incorporated into the layout design at 1000 East Capitol Boulevard, the lot most affected by landslide movement in 1998. In June 1998, the distance between the new main scarp and the eastern corner of the house measured only 30 feet (9 m), only 5 feet (1.5 m) more than the setback requirements for lots farther northeast where shallow rock and thus more stable slopes were anticipated. At the rear of the lot, fill was apparently placed over the natural crest of the 1979 landslide scarp. In addition, downspouts from all houses in the crown area of the landslide drain directly onto the upper part of the slide, contrary to Dames & Moore's (1981) drainage recommendations for lots with the more stable slopes.

Causes and Trigger

The CBCC landslide appears to have been triggered when an instability threshold ground-water level was exceeded within the landslide mass and adjacent slopes. Conditions that contributed to this threshold level being reached included a cumulative rise in ground-water levels associated with the 1995-98 precipitation period and infiltration from excess landscape irrigation and redirected runoff. Other sources of water, including water flowing from an apparently abandoned, 16-inch-diameter (41 cm) storm drain that crosses the landslide from head to toe, cannot be ruled out as also having a contributing effect. Other buried utilities, such as the storm drain beneath Capitol Boulevard above the landslide, and their surrounding granular fill, may also be adding water to the ground-water table, but no direct evidence suggests this has or is occurring.

Movement of the CBCC landslide has occurred during the last two precipitation periods suggesting movement is triggered by a natural ground-water-level rise induced by infiltration of excess precipitation. The importance of antecedent precipitation can be inferred from the onset of movement in the latter part of the precipitation periods. The Dames & Moore (1981) report does not describe active landsliding, suggesting no significant movement occurred during the first two years of the 1980-86 precipitation period. An inventory of active landslides in 1983 (Brabb and others, 1989) also does not show the CBCC landslide, but movement, although undetected, likely had initiated by the spring of 1983. Survey data (Lynn Curt, Salt Lake City Surveyor, unpublished data, 2000) indicate movement occurred in 1984 and 1986 at the CBCC landslide and at three other nearby landslides on the southeast-facing slope in City Creek Canyon. In the absence of houses in the crown area of the CBCC landslide during the 1980-86 precipitation period, landscape irrigation water likely did not contribute to the movement.

Unlike during the 1980-86 precipitation period, excess water from landscape irrigation, redirected runoff, and possibly other sources likely contributed significantly to the rise in ground-water levels that triggered landslide movement in 1998. Lots above the landslide all maintain bluegrass-type lawns that likely promote excessive watering, particularly in the dry summer months. In addition, downspouts at the homes are connected to flexible pipes that discharge directly onto the landslide. Landscape-irrigation water is additional to the natural precipitation that falls in Salt Lake City (about 16 inches [41 cm] in an average year at the airport). The level lawns in the crown area underlain by permeable, granular fill soils likely promote infiltration of most of the water. As a result, landscape irrigation contributes to an artificial rise in local ground-water levels during the dry part of the year. The occurrence of standing water in wetland areas on the CBCC landslide in late summer through early winter, a natural period of declining ground-water levels, is evidence of the impact of landscape-irrigation water on ground-water levels. On the CBCC landslide the standing water remains over a long period during which evaporation is at its greatest and despite numerous ground cracks and minor scarps which allow surface water to infiltrate (drain) into the landslide. Thus, new water appears to continually discharge from springs near the base of the main scarp to replenish the water lost by infiltration and evaporation. Residents have indicated the wetland vegetation, which currently occupies a significant percentage of the total area of the landslide, is relatively recent. This suggests the vegetation may have become established only after a sufficient landscape-irrigation-induced rise in the ground-water level occurred to sustain standing surface water on the landslide. A review of a photograph of the landslide, circa the early 1970s, shows the lack of abundant wetland-type vegetation and suggests drier conditions in the landslide prior to development directly upslope.

Landsliding in 1998 involved enlargement of the landslide through retrogressive slope failure as upslope areas in the crown and right flank were incorporated into the landslide. At 1000 East Capitol Boulevard, fill had been placed on the crown and adjacent side slopes. The additional weight of the fill soil added to the driving forces in the main-scarp slope and likely contributed to the retrogressive nature of the failure in this area. Analysis of the stability of the landslide as part of this study indicated that the fill likely did not contribute significantly to the movement of the entire CBCC landslide because the weight of the fill is extremely small compared to the total landslide mass. However, the fill may have been a significant cause of the local retrogressive enlargement of the landslide.

South Fork Kays Creek Landslide, Layton

Introduction

Sometime in April 1998, renewed landslide movement began on a moderate-sloping, north-facing bluff above the South Fork of Kays Creek in Layton (Giraud, 1999b). The head of the landslide crossed the northern landscaped parts of three residential lots. Downslope movement caused displacement of landscaped areas in the rear (north) of the lots (figure 12) and damaged a chain-link fence. A landscaped terrace on one lot showed that landscaping had taken advantage of the natural break in slope associated with a historical (pre-1985) landslide scarp. Elsewhere, the older scarp had been regraded to a smooth slope. Prior to the landslide movement in 1998, a resident indicated some movement occurred in 1997 that was not reported at the time to Layton City officials. Although no houses were directly damaged by the landslide movement, future renewed retrogressive

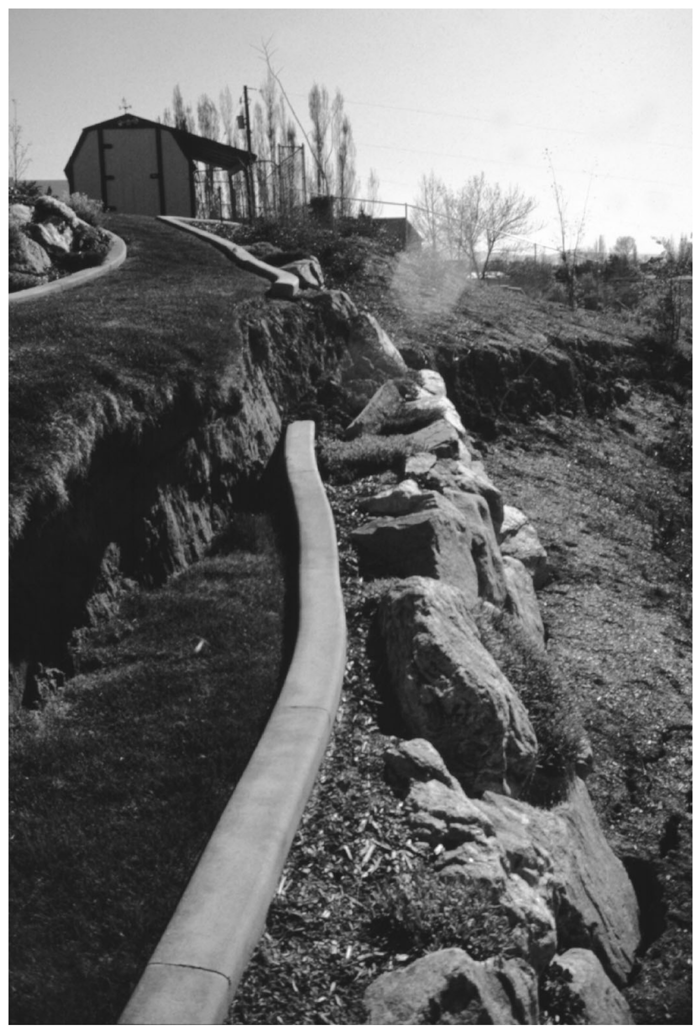


Figure 12. Head of the South Fork Kays Creek landslide in Layton. View is toward the west showing damage to landscaping. Main scarp of the landslide is visible in center of photo.

slope failure could cause building damage (Giraud, 1999b). One landowner pursued a settlement for damages with Layton City related to the loss incurred from landsliding.

Geology

Anderson and others (1982) and Lowe (1989) mapped the north-facing bluff along the South Fork of Kays Creek as a prehistoric landslide. The landslide is underlain by lacustrine (lake) sediments that were deposited in Lake Bonneville as part of the Weber River delta. Nelson and Personius (1993) indicated these sediments consist of interlayered clay, silt, and minor fine sand. The bluff was formed as the creek incised the deltaic deposits. Oversteepening of the bluff occurred as the creek cut into the base of the bluff and periodically caused shallow landsliding (Maughan, 1992). Giraud (1999b) described a well-defined historical landslide scarp visible on 1985 aerial photographs in the vicinity of the 1998 main scarp. The 1985 landslide features likely formed due to local renewed movement in 1982 or 1983 of the larger prehistoric landslide.

Landslide Description

The 1998 landslide features consisted of a roughly east-west main scarp, transverse ground cracks in the head, and a right-lateral shear (Giraud, 1999b). The toe of the landslide was inferred to be at the base of the bluff along the creek, but the only evidence for landsliding in the lower slope was a local, approximately 25-foot-high (7.6 m), rotational earth slide. The main scarp measured about 350 feet (107 m) in length and ranged up to 3 feet (1 m) in height. Survey stake measurements across the main scarp indicated a very slow rate of movement throughout the summer and fall of 1998, accelerating slightly on the east side of the landslide (figure 13). On the west side, the rate of movement consistently slowed throughout 1998 and early 1999, but by 1999, landslide movement had suspended elsewhere. The survey stake measurements also indicated some deformation in the crown (B in figure 13), but no damage occurred to a nearby house. This deformation likely represented a local response of shallow soils in the crown to removal of lateral support adjacent to the main scarp.

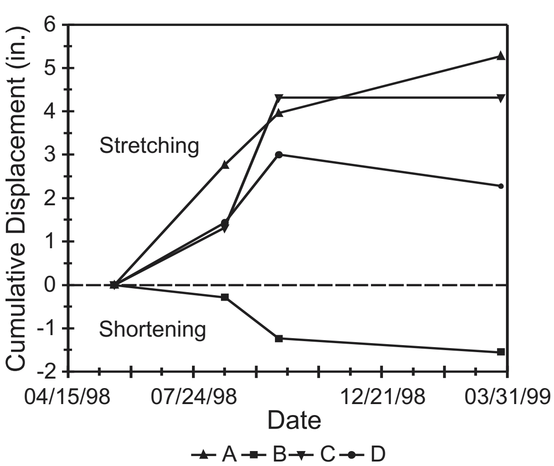


Figure 13. *Cumulative displacement across the main scarp of South Fork Kays Creek landslide between May 22, 1998 and March 26, 1999. Data based on survey stakes and tape measurements. Survey stake series A is on west side of landslide and series D is on east side. Plot indicates movement had likely suspended by October 1998 except at A. Data courtesy of Richard Giraud, Utah Geological Survey.*

The lack of any apparent landslide features at the base of the bluff suggests landslide movement was confined in 1998. The small rotational earth slide in the inferred toe of the landslide may have been caused by local oversteepening as the ground tilted above the emerging surface of rupture. Also, Giraud (1999b) noted that the flank lacked any shear feature, further suggesting incomplete rupture of the landslide.

Site Investigations and Development

Pre-development site investigations included a geologic study by a consulting geologist (Maughan, 1992). Three boreholes were drilled as part of the study. The study generally confirmed the soil types present in the bluff and the presence of relatively shallow (25 feet [7.6 m]) ground water at one location. The Maughan (1992) report concluded the slope was stable under existing moisture conditions, although at the time of the study in 1992 precipitation was below normal and ground-water levels were likely lower than historical high levels (circa mid-1980s). The Maughan (1992) report did indicate that a rise in ground moisture conditions could cause landsliding and made a general recommendation to avoid wetting on-site soils. Whereas the Maughan (1992) report identified a small surficial landslide on the slope along the South Fork of Kays Creek, it did not indicate the presence of a larger prehistoric or historical landslide at the site despite that the prehistoric landslide is shown on available maps (Lowe, 1989).

Giraud (1999b) concluded that landscaping conformed to or took advantage of the break in slope caused by the main scarp of the historical landslide at a minimum of one of the three lots affected by the 1998 landslide. Development of the site included regrading and installing irrigation systems for landscape vegetation, the latter contrary to recommendations by Maughan (1992) to avoid wetting on-site soils.

Causes and Trigger

Giraud (1999b) summarized the causes of the landslide as including a natural oversteepening of the slope from undercutting at the base by the South Fork of Kays Creek, and low-strength soils related, in part, to repeated historical landslide movement. Another important factor was likely a rise in ground-water levels in the bluff caused by the 1995-98 precipitation period and landscape irrigation (which continued even after landslide movement in 1998). Some uncertainty exists as to whether a ground-water-level rise or stream-cutting-induced oversteepening triggered the landslide.

Cedar Bench Subdivision Landslide, South Weber

Introduction

The Cedar Bench Subdivision landslide (Solomon, 1999) was one of two landslides in 1998 along the generally north-facing bluff cut by the Weber River (figure 14). Landslide movement initiated in early April, but appeared to have suspended by the end of the month (Solomon, 1999). Land-

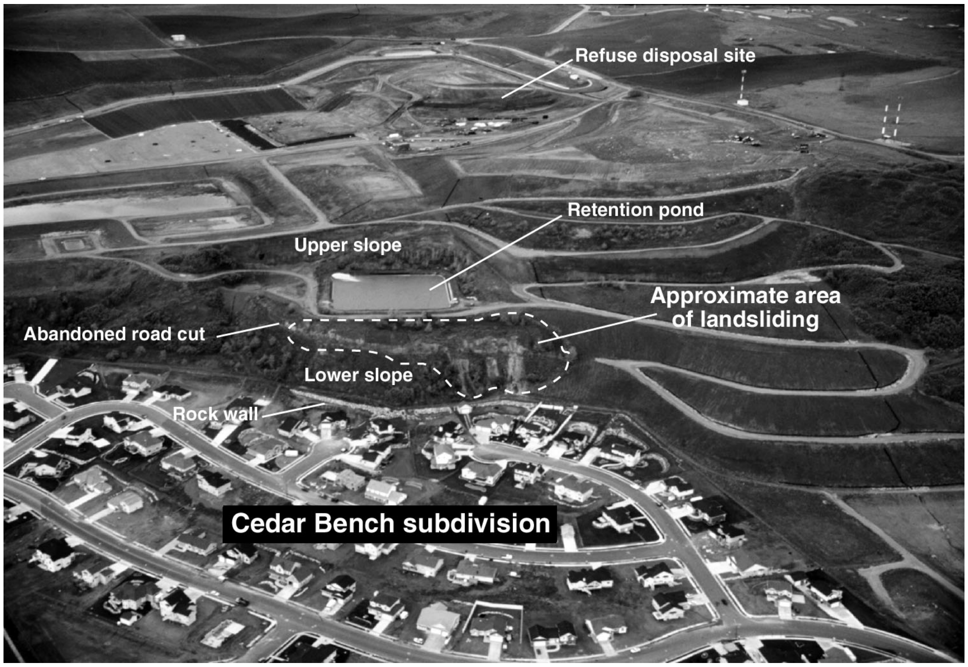


Figure 14. Oblique aerial view toward the south of the 1998 Cedar Bench Subdivision landslide in South Weber. Shallow earth slides occurred on the lower slope below the retention pond and above the abandoned road cut. Earth slides and flows occurred between the road cut and the toe of the slope. Modified from Solomon (1999).

sliding threatened residential properties at the base of the bluff as well as a retention pond directly above the landslide. The 1998 landslide movement was a partial reactivation of a prehistoric landslide following modification of a natural hillside. Whereas the landslide did not cause damage to the residential properties below the slope in 1998, it attracted the attention of residents, local officials, and the media to the vulnerability of lots in the southern part of the Cedar Bench subdivision to landsliding.

Geology

Lowe (1989) and Nelson and Personius (1993) mapped the lower part of the bluff as a prehistoric landslide. The landslide deposits are underlain by late Pleistocene lacustrine sands (upper bluff) and deltaic deposits (lower bluff). Thin, late Holocene alluvial-fan deposits locally overlie the lowermost landslide deposits. Landsliding along the bluff likely initiated in the latest Pleistocene and Holocene, after Lake Bonneville receded and the Weber River incised through the lacustrine-deltaic deposits and formed oversteepened valley slopes. A topographic escarpment exists at the toe of the prehistoric landslide deposits (Nelson and Personius, 1993). The approximate age of the oldest landslide deposits is dependent on the nature of the topographic escarpment. If the escarpment was formed by stream erosion of the landslide deposits then the earliest landsliding is likely latest Pleistocene or early Holocene. If the escarpment is the coalesced toe formed by landsliding that extended downslope onto older (latest Pleistocene to middle Holocene) stream deposits, then landsliding could be relatively young (late Holocene). Late Holocene landsliding is not precluded by the former hypothesis (older escarpment), but, in that scenario, more recent landsliding did not extend out beyond the escarpment.

Landslide Description

The landslide consisted of a complex of shallow earth slides (figure 15) and local earth flows, possibly underlain by a deep-seated rotational slide (Solomon, 1999). The shallow earth slide area was above a road cut in the lower part of the bluff and measured approximately 400 feet (122 m) wide and 50 feet (15 m) long. Solomon (1999) estimated the surface of rupture of these slides to be about 8 feet (2.4 m) deep. Two earth flows originated from the shallow earth-slide area. Deposits of silty sand from these flows extended downslope of the road cut. Transverse ground cracks were present upslope of the road cut. No landslide features were observed in the vicinity of the retention pond. However, Solomon (1999) inferred subtle bulges in the lower slope and a slightly bent chain-link fence as evidence of possible deep-seated rotational sliding. Based on topography shown on a site plan in a Huntingdon Chen-Northern, Inc. (1993) report, the average slope of the lower bluff is about 53 percent.

Site Investigations and Development

A geotechnical report (Huntingdon Chen-Northern, Inc., 1993) for the Cedar Bench subdivision assessed the stability of the bluff prior to development. The report concluded the bluff was stable under static conditions assuming ground water was deeper than 100 feet (30 m). The assumed ground-water depth was unsupported because the only borehole drilled in the bluff was just 20 feet (6 m) deep. In a review of the report by the UGS (Lowe, 1994), data were revealed suggesting shallower ground water to the south and atop the bluff (Gill, 1985). The potential for shallower ground water suggested that the stability of the slope may have been overestimated, particularly for periods of above-normal precipitation that could cause higher ground-water

Figure 15. Shallow earth-slide complex at the Cedar Bench Subdivision landslide. View is toward the southwest.

levels. Lowe (1994) recommended that measures be taken to increase the overall slope stability to acceptable levels and/or delineate adequate building setbacks from the base of the bluff.

Exactly whether hazard-reduction measures were incorporated into the design of the subdivision is unclear. Solomon (1999) described building setbacks, as little as 15 feet (4.6 m), which appear to be inadequate hazard reduction for a slope 80 feet (24 m) high. A cut slope at the base of the bluff is covered with a rock wall constructed of boulders that likely offer little if any significant support. An individual lot owner further excavated below the rock wall in 1998 adding another 10 to 15 feet (3.0-4.6 m) to the cut-slope height. The uncontrolled site modification likely further destabilized the slope.

Causes and Trigger

Solomon (1999) concluded that the shallow landsliding was likely triggered by increased pore pressure caused by recent precipitation and snowmelt in April 1998. Precipitation between January and April in the area was 147 percent of normal. About 3 inches (8 cm) of precipitation fell in April including rainfall the week before landsliding triggered. Shallow soils may have become temporarily saturated or nearly saturated prior to failure as a result of the above-normal precipitation. Solomon (1999) also cited inadequate support for shallow soils above the road cut as a probable cause. The road cut removed lateral support for the uppermost shallow soils and locally increased the slope angle, oversteepening the slope.

Davis-Weber Canal Landslide, South Weber

Introduction

Landslide movement on a northeast-facing slope above the Davis-Weber Canal in South Weber likely began sometime in March. In early to mid-April, the rate of movement accelerated (Black, 1999a), causing the landslide to encroach on the canal and deposit debris in it (figure 16). By late April, a concern existed that flooding could occur if the canal was damaged or blocked by landslide debris. The landslide area consisted of a zone of relatively shallow rotational and translational earth slides that occurred on the east edge of a historical landslide area that had moved in 1984, destroying the upslope concrete canal lining. The landslide occurred along a stretch of the canal that had been replaced and reinforced in 1996, but did not cause damage to the canal lining in 1998. However, the landslide demonstrated the vulnerability of the canal to damage or blockage from landsliding.

Geology

Lowe (1988a) mapped the northeast-facing bluff on which the 1998 landslide occurred as a prehistoric landslide. Localized, renewed movement of this landslide has occurred as recently as 1984 (figure 17) (Lund, 1985). The landslide deposits are underlain by late Pleistocene deltaic deposits (Montgomery-Watson, 1995). The deltaic deposits in the slope above the canal consist of silty clay with thin interbeds of silt and very fine to fine-grained sand. Atop the bluff, sand and gravel deposits overlie the silty clay. The silty clay is underlain by sand and silty sand deposits in the lower slope below the canal.

Landslide Description

The landslide area consisted of a zone of shallow rotational and translational earth slides about 1,200 feet (370 m) wide and extended upslope above the canal about 250 feet (76 m) (Black, 1999a). The average slope above the canal is about 50 percent. By late April 1998, the main scarp of the landslide was about 6 feet (2 m) high. Shallow earth slides

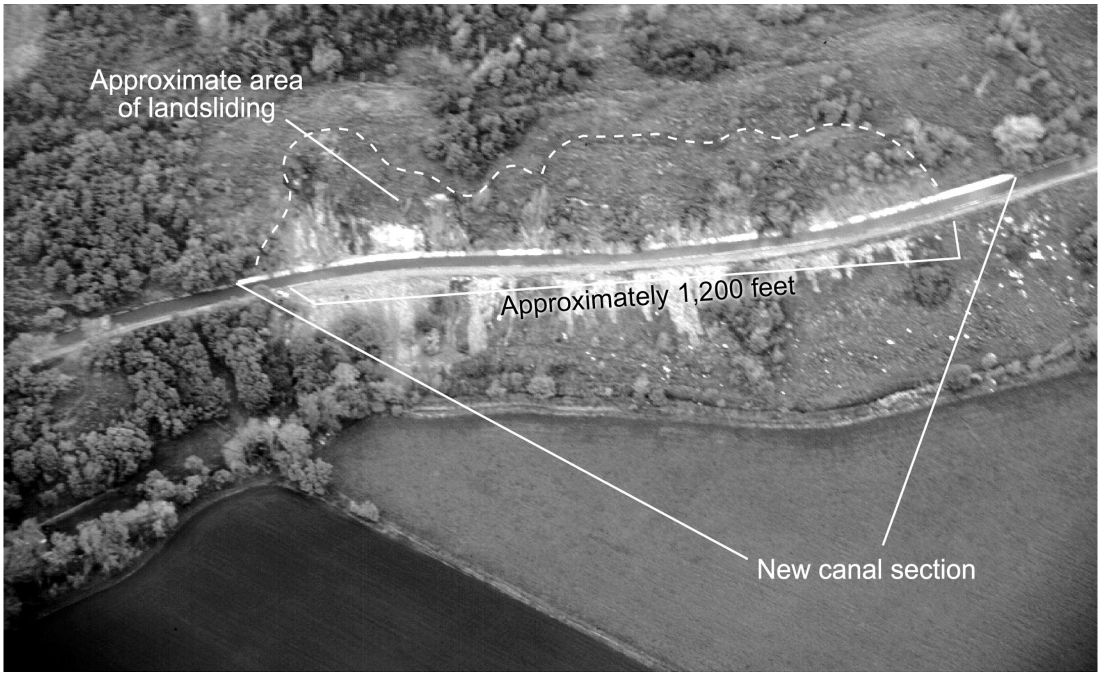


Figure 16. Oblique aerial view toward the southwest of the 1998 landslide along the Davis-Weber Canal in South Weber. A section of the canal was replaced in 1996, and nearly corresponds to the area of landsliding. Modified from Black (1999a).

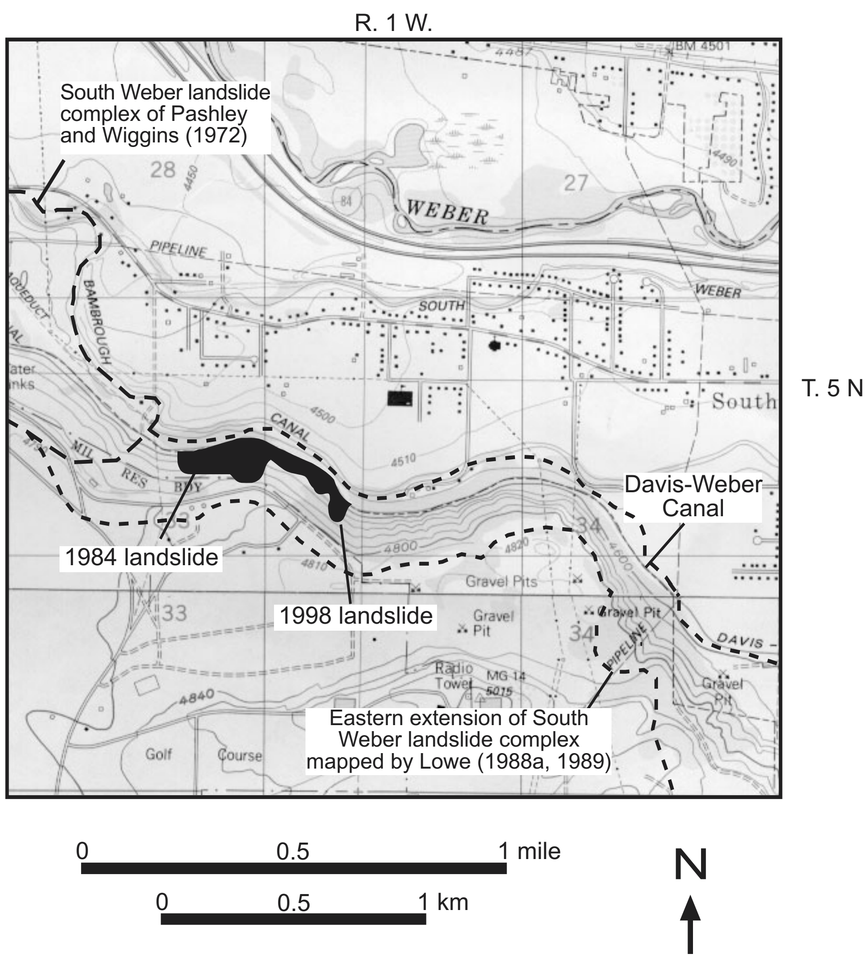


Figure 17. Relation of the 1998 Davis-Weber Canal landslide to a 1984 landslide. Location of latter from Lund (1985). Both landslides are part of the eastern extension of the South Weber landslide complex (Pashley and Wiggins, 1972) mapped by Lowe (1988a, 1989). Lowe (1988a) also mapped the eastern part of the 1998 landslide as a historical landslide (not shown). Modified from Black (1999a).

at the toe of the landslide and directly above the canal were the primary source of debris deposited in the canal. Other landslide deformation features included numerous transverse ground cracks and minor scarps that separated blocks of intact soil.

Site Investigations

The susceptibility of the bluff above the canal to landsliding had been recognized in previous regional (Pashley and Wiggins, 1972) and site-specific investigations (Lund, 1985). A more recent slope-stability analysis (Montgomery-Watson, 1995) demonstrated the marginal stability of the slope. The results of this analysis indicated a factor of safety of only 1.13 and suggested landsliding was possible with only a small increase in ground-water levels. Black (1999a) reported recurrent landslide movement at this site for over a decade prior to the 1998 landslide, further confirming the marginal stability of the slope above the canal.

Causes and Trigger

The probable causes of the 1998 landslide include marginal slope stability of the oversteepened bluff in weak deltaic soils and elevated ground-water levels (Black, 1999a). Montgomery-Watson (1995) concluded that landsliding was likely in the deltaic sediments, particularly the silty clay soils above the canal, when they were saturated. The timing of the landslide activity in 1998, following the snowmelt, suggests that high ground-water levels likely coincided with accelerated landslide movement. Precipitation in January and February 1998 in the area was about 200 percent of normal. As ground-water levels rose, in response to excess precipitation associated with the 1995-98 precipitation period, an instability threshold was likely exceeded that triggered landslide movement in 1998.

Green Hill Country Estates Phase VI Cut-Slope Landslide, Weber County

Introduction

In May 1998, a shallow landslide (figure 18) occurred in a low-angle cut slope in the Green Hill Country Estates Phase VI subdivision in Ogden Valley, Weber County (Black, 1999b). The landslide occurred in a northwest-facing cut slope with an average slope of only 25 percent in clay-rich soils. Four nearby historical landslides had occurred in 1995 in cut slopes that were steeper than 25 percent (Applied Geotechnical Engineering Consultants, Inc. [AGEC], 1996). These four landslides reactivated sometime in 1997 or 1998 by retrogressing (AGEC, 1998a). The five landslides, including the 1998 landslide, demonstrated the susceptibility of flat-lying areas underlain by weak clay soils to landsliding where site modifications, in this case cut slopes, are made.

Geology

The 1998 landslide occurred in an area mapped as clay-rich, mixed alluvium and colluvium (Lowe, UGS, unpublished mapping). These soils are underlain by the argillite member of the Precambrian Maple Canyon Formation (Crittenden, 1972). The surface of rupture of the 1998 landslide was between a shallow high-plasticity clay layer and an underlying clayey gravel deposit. All five of the historical landslides in the area occurred where the shallow clay layer is the upper soil (AGEC, 1996).

Landslide Description

The 1998 landslide was a small, shallow landslide in an approximately 15-foot (4.6-m) high cut slope on the east side of Maple Canyon (Black, 1999b). Although Black (1999b) did not describe the dimensions of the landslide, photographs

Figure 18. *Shallow landslide at Green Hill Country Estates in Weber County. View is toward the southwest.*

suggest that it was about 100 feet (30 m) wide and 60 to 80 feet (18-24 m) long. The surface of rupture of the landslide was only about 3 feet (1 m) below the ground surface and the main scarp of the landslide appeared to be less than 3 feet (1 m) high. Landslide movement deposited debris onto the edge of an unpaved road in the subdivision, but did not affect any structures. AGEC (1998a) indicated that a spring discharged upslope of the landslide.

Site Investigations

A comprehensive landslide investigation (AGEC, 1996) preceded site modifications at this site. The investigation included trenching and profiling of the four on-site historical landslides and laboratory soil-strength testing of samples collected in and adjacent to the landslides. Soil-strength tests demonstrated a significant loss in shear strength of the high-plasticity clay soils upon wetting. Friction angles of the clay soils generally bracketed the range of pre- and post-failure slope angles, suggesting that both the historical landslides and the adjacent slope soils were likely marginally stable, particularly if the shallow soils became wet. Based on these results, AGEC (1996) recommended final cut-slope angles in the clay soils of only 25 percent (a slope flatter than the average pre-failure slope of 75 percent of the nearby historical landslides) and adequate surface drainage above the cut slopes to prevent the shallow soils from becoming saturated. The report also included recommendations for stabilizing the four historical landslides. The UGS (Ashland, 1997a) reviewed the AGEC (1996) report and concurred with its conclusions and recommendations.

Causes and Trigger

All five of the historical landslides, including the May 1998 landside, occurred in the spring following the snowmelt and during the wettest part of the year. On average, April and May are the wettest months of the calendar year in nearby Ogden, Utah. AGEC (1996) concluded that the historical landslides had been triggered by a reduction of shear strength upon becoming saturated. This likely also occurred in 1998. Site modification also apparently contributed to landsliding because none of the landslides occurred on natural slopes. The cut slopes increased the average slope and the driving forces acting on the shallow soils and reduced the threshold for instability. The low-strength, high-plasticity soils remained stable in natural slopes on the site, but failed upon becoming saturated in the cut slopes. Photographs indicate that surface drainage recommendations had not been implemented prior to the May 1998 landslide. This likely contributed to shallow soils in the cut slope becoming wet, a condition that AGEC (1996) had recommended be avoided to prevent landsliding.

Springhill Landslide, North Salt Lake

Introduction

In early July 1998, an area of building distress in a residential subdivision in North Salt Lake was identified as a possible active landslide (Giraud, 1999c). A subsequent geotechnical slope-stability investigation (Terracon, 1998) confirmed that landsliding was the cause of the building distress. The Springhill landslide threatened approximately twenty-three houses, destroying one house (figure 3) and causing severe damage (figures 19 and 20) to four others in 1998. Inclinometer data (Giraud and Fadling, 1999; Terracon, 1999) confirmed a very slow rate of movement throughout 1998 and 1999. The lack of well-defined landslide deformation features suggested incipient landsliding (Giraud and Fadling, 1999). This state of activity made it difficult to discern the extent of landsliding. Homeowners indicated building distress had initiated in the spring of 1997 and had suspended by that summer, but resumed and accelerated in 1998.

Geology

The Springhill landslide is underlain by Tertiary clay-rich tufaceous sediments and volcanic breccia (Van Horn, 1981; Giraud and Fadling, 1999). Surficial lacustrine gravel deposits at the site were previously removed by mining prior to development of the area. As a result of site grading, fill ranges locally up to 19 feet (6 m) thick at the site (Terracon, 1998), but is generally absent, particularly in the upper part of the landslide. The tufaceous sediments are variably weathered but may be severely to completely weathered to clay soils to depths exceeding 70 feet (21 m). The Springhill area is characterized by shallow ground water, springs, and seeps. Soils are thus perennially saturated at depths ranging from 1 to 23 feet (0.3-7 m) (Giraud and Fadling, 1999). The saturated, high-plasticity clay soils are low strength, with average friction angles ranging from 10 to 15 degrees (Terracon, 1998; Giraud and Fadling, 1999). Preliminary inclinometer data (figure 21) suggest these soils exhibit a strain-weakening behavior where shear strength decreases with increased displacement or movement along rupture surfaces.

A trace of the Warm Springs section of the Salt Lake City segment of the Wasatch fault zone crosses the lower part of the landslide. A steep slope about 30 feet (9 m) high is a remnant of the fault scarp and has been altered by past gravel-mining operations. The relation of the fault to the landslide is unclear, but faulting may have contributed to the deep weathering of the adjacent Tertiary rocks and may also control spring locations and thereby ground-water levels.

Landslide Description

Due to the apparent incipient nature of the landslide, its extent is not well defined. Only two landslide deformation features were recognized during a reconnaissance of the landslide in 1998 (Giraud, 1999c); a right-lateral shear zone on the right flank and a possible toe thrust at the base of the remnant fault scarp (figure 19). Terracon (1998) estimated the landslide is about 650 feet (200 m) long and ranges from about 160 to 300 feet (49-91 m) wide. The average slope of the landslide is about 14 percent; however, locally the slope reaches 40 percent at the remnant fault scarp (Giraud, 1999c). Inclinometer data (Terracon, 1998, 2000) indicated the landslide is between 12 and 50 feet (3.7-15 m) deep. This range in depth suggests either a single irregular rupture surface or multiple rupture surfaces in a more complex landslide.

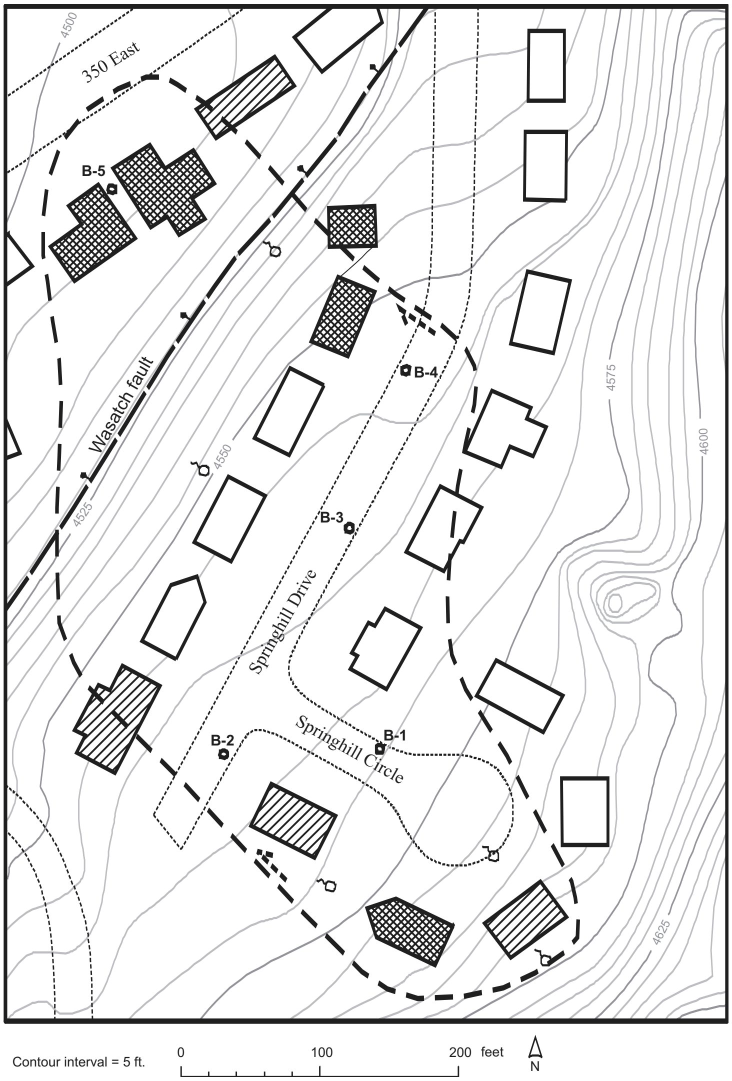


Figure 19. Sketch map of the Springhill landslide in North Salt Lake. Boundary of landslide (dashed line) inferred from building distress and sparse landslide deformation features. Severely distressed houses indicated by crosshatched pattern. Other building distress indicated by diagonal pattern. Map shows locations of springs and seeps. Trace of Wasatch fault inferred in lower part of scarp. Boreholes from Terracon (1998) investigation shown. Inclinometer plot (figure 21) from borehole B-1. Modified from Giraud and Fadling (1999).

Figure 20. Foundation damage to a house at the Springhill landslide. View of southwest corner of house.

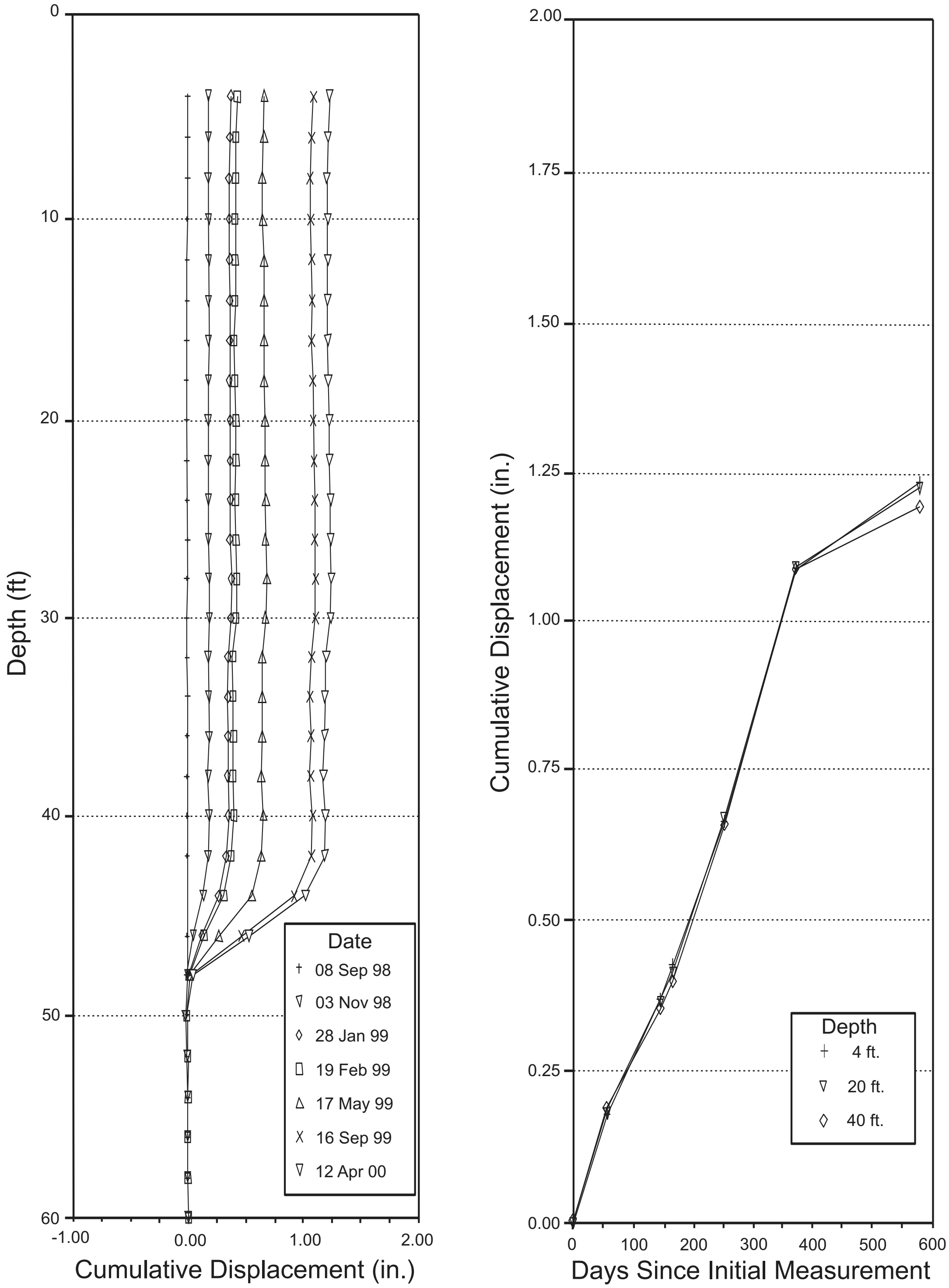


Figure 21. *Inclinometer plot documenting movement in the upper part of the Springhill landslide. Surface of rupture is about 48 feet deep. Rate of movement increased slightly in latter part of 1999, possibly due to displacement (strain)-weakening behavior of soils near the surface of rupture. Subsequent decrease in movement rate may be caused by slight decline in ground-water levels in 2000 or reflect undetected suspension of movement sometime during the measurement period. Plot modified from Terracon (2000).*

Site Investigations and Development

Apparently no site investigations preceded residential development in the Springhill area. Upon exhaustion of the gravel resources, the area was regraded and the subdivision built. Residents indicated the houses were built before the early 1980s, the wettest period on record in the Springhill area, and thus were surprised when building distress initiated in 1997. Subsequent to the 1998 building distress, the Springhill landslide became one of the most thoroughly studied of the 1998 Wasatch Front landslides.

Site development consisted primarily of regrading the abandoned gravel pit and construction of single-family houses. Thick fill underlies the part of Springhill Drive north of Springhill Circle (figure 19) (Terracon, 1998). In addition, cuts were made in the remnant fault scarp to level building lots on the east side of 350 East Street (Terracon, 1998). Rupture of the foundation wall at a severely damaged house at 160 Springhill Drive indicated substandard reinforcement of concrete with steel rebar, rendering the foundations sensitive to minor deformations. Most lots have traditional water-consuming vegetation and generous landscape irrigation is employed by many homeowners.

A variety of means are employed to handle discharge from the numerous springs and seeps in the area (Terracon, 1998). Spring-water collection systems are present on at least two properties (Giraud, 1999c). On one property, a flexible drain pipe carries the collected spring water to the road. Several systems discharge to the local storm drains. Ground-water discharged from other springs and seeps is allowed to infiltrate back into the ground farther downslope (Giraud, 1999c).

Causes and Trigger

Elevated ground-water levels are the most likely trigger of the 1998 landslide movement (Terracon, 1998; Giraud, 1999c); however, other causes may have contributed to the instability. Laboratory soil-strength testing (Terracon, 1998) indicated the high-plasticity clay soils have very low shear strengths. Residual friction angles of the clay soils range from 10 to 15 degrees and are lower than back-calculated friction angles required to initiate landsliding (Giraud and Fadling, 1999). The reason for residual shear-strength conditions is unclear, but may include (1) recent landsliding since 1997 of severely weathered soils, (2) historical landsliding pre-dating the recent movement, (3) unloading-induced failure due to gravel extraction and subsequent landsliding, and (4) gradual, long-term loss of shear strength due to chemical and physical weathering of the clay soils.

Elevated ground-water levels also likely occurred in 1983, the wettest year in the historical record; however, building distress indicative of landsliding did not occur then. The reliance on building distress to determine whether landslide movement occurred, given the very slow rate of movement and lack of landslide deformation features, may be misleading. A minimum amount of movement may have been required to cause building distress (R.E. Giraud, Utah Geological Survey, verbal communication, 2000); thus, some movement may have occurred in 1983 that did not affect the houses. The apparent absence of landslide movement in 1983, however, may suggest ground-water levels were higher in 1998 than 1983, possibly because of the contribution of excess landscape-irrigation water. In 1983, the possible cumulative increase in ground-water levels was limited by the short period of time the subdivision had been in existence. By 1998, fifteen additional years of excess landscape-irrigation water could have contributed significantly to a rise in ground-water levels. Two piezometers in the Springhill landslide show a rise in ground-water levels in the late summer and fall of 1999 (figure 22). Ground-water levels rose between 10 inches (25 cm) to slightly over 14 inches (36 cm) in the wells during a period of below-normal precipitation. Precipitation between June and October 1999 was only 42 percent of normal. Thus, ground-water levels rose in the two wells despite a deficit of about 3 inches (8 cm) of natural precipitation for the period. Whereas the cumulative effects of landscape irrigation on ground-water levels in the Springhill subdivision have yet to be quantified, infiltration of excess

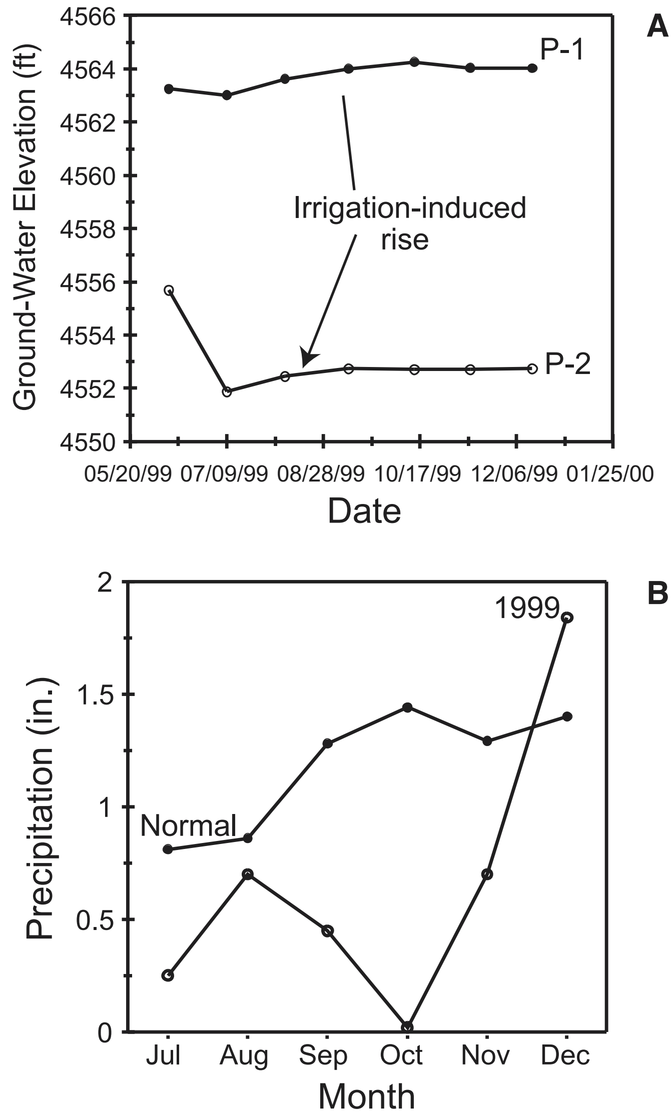


Figure 22. *Irrigation-induced rise in ground-water levels at the Springhill landslide. During the late summer and early fall of 1999, ground-water levels rose in wells P-1 and P-2 at the Springhill landslide in North Salt Lake (A) despite below-normal precipitation during the same period (B). Precipitation data from Salt Lake City International Airport. Irrigation-induced rise reverses natural decline in ground-water levels in early summer.*

landscape-irrigation water likely contributed to a rise in ground-water levels over the past two decades.

Sherwood Hills Landslide Complex, Provo

Introduction

Local landslide movement occurred in 1998 at the Sherwood Hills subdivision in Provo. Near the southern intersection of Windsor Drive and Foothill Drive, a landslide caused building and landscaping distress affecting two residential properties and the abutting sidewalk (figure 23). In 1999, steel sheeting and tiebacks were used to stabilize the landslide area. Minor, unrecognized landslide movement also likely occurred in the area between Mile High Drive and Windsor Drive in the upper part of the landslide complex based on subsequent survey data and building and utility distress reports (Nicholas Jones, Provo City engineer, verbal communication, 2000). In 1998, a house estimated to be worth approximately $500,000 was demolished after being severely damaged by landslide damage sometime during the previous three years (1995-97) (Haddock, 1998).

Geology

Machette (1992) mapped the Sherwood Hills subdivision area as prehistoric landslide and alluvial-fan deposits. Baker (1964) indicated that Mississippian Manning Canyon Shale underlies the landslide deposits (the latter not shown on his map). These units are west of and on the downthrown side of the Provo segment of the Wasatch fault zone (Machette, 1992). Recent mapping by URS/Dames & Moore (2001) shows the area to consist of a complex of landslides of varying relative ages on the north and a combination of alluvial-fan, colluvial, and landslide deposits on the south. Provo City geologic-hazard maps (International Engineering Company, Inc., 1984) show the landslide complex contains several local scarps and distinct landslide areas, including the scarp of the landslide near the intersection of Windsor and Foothill Drives.

Landslide Description

URS/Dames & Moore (2001) mapped the main part of the landslide complex as being about 4,500 feet (1,400 m) wide near its toe along Foothill Drive and about 1,500 to 2,100 feet (460-640 m) long west of the Wasatch fault zone. URS/Dames & Moore (2001) also mapped additional alluvial-fan, colluvial, and landslide deposits that extend at least 1,400 feet south of the landslide deposits. Machette (1992) also mapped prehistoric landslide deposits extending up the steep mountain slope more than 800 feet (240 m) east of the Wasatch fault zone and the subdivision boundary. A lobate deposit mapped by URS/Dames & Moore (2001) in the northernmost part of the complex is continuous with the prehistoric landslide deposits on the mountain slope. Scarps and local landslides mapped by International Engineering Company, Inc. (1984) and URS/Dames & Moore (2001) suggest local reactivation of the prehistoric landslide deposits. Landslide movement data obtained by Provo City using Global Positioning System (GPS) surveying techniques indicate an approximately 1,200-foot-wide (370 m) area in the upper part of the landslide complex with a relatively consistent direction of movement (figure 24). The average slope in the upper part of the landslide complex is about 24 percent.

Site Investigations and Development

Landsliding was recognized in the vicinity of the Sherwood Hills area as early as 1971 (Van Horn and others, 1972) prior to any development. By the mid-1980s residential development of the area was underway and Provo City began requiring lot-specific geologic investigations prior to obtaining a building permit for home construction. In most cases,

Figure 23. *Landslide near the intersection of Windsor Drive and Foothill Drive in the lower part of the Sherwood Hills landslide complex, Provo. View is toward the east.*

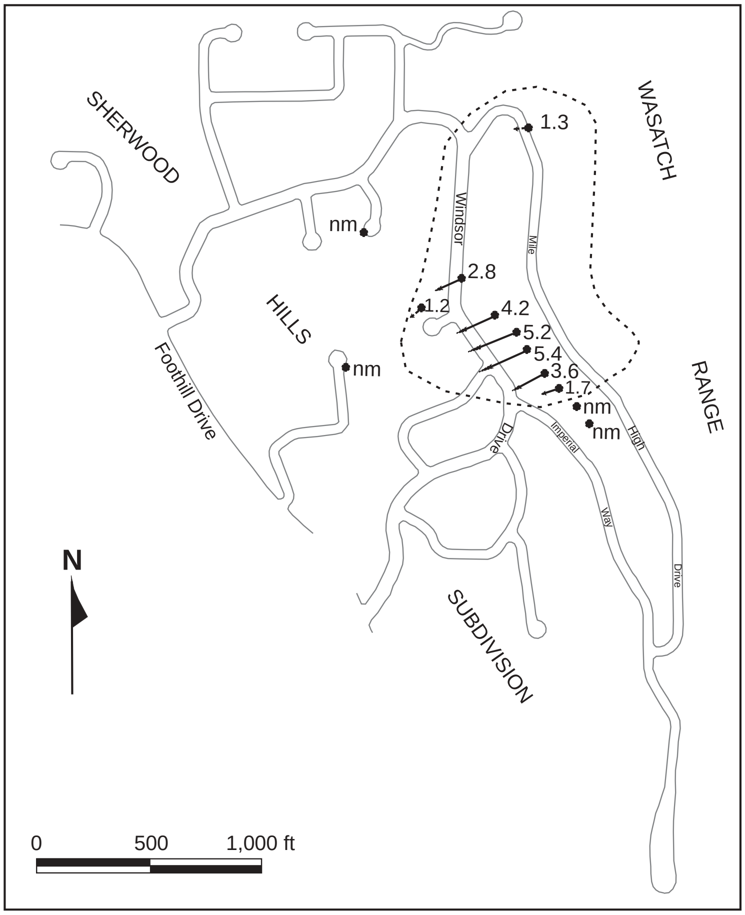


Figure 24. *Area of landslide movement defined by GPS surveying in the upper part of the Sherwood Hills subdivision, Provo. Arrows show direction of movement and are scaled to the magnitude of the displacement. Values indicate cumulative displacement in inches between May 11, 1999 and March 31, 2000. Dashed line shows inferred boundary of active landsliding. No movement indicated by nm.*

these studies were limited in scope and restricted to individual lots. Slope movement occurred, at least locally, in the upper part of the Sherwood Hills subdivision and physical measures, including drilled shaft walls, were used in attempts to stabilize some lots (Rollins and Rollins, 1992). Lot-specific stabilization measures appear to have been only partly effective and some properties suffered significant distress as a result of landslide movement.

Causes and Trigger

The Sherwood Hills landslide complex is underlain by the low-strength soils derived from the Manning Canyon Shale. Limited undrained shear-strength data exist for clay soils in the Sherwood Hills landslide complex. These values (table 4) provide an understanding of the short-term slope stability, but are inappropriate for understanding long-term slope stability. The low shear strength of clay soils derived from the Manning Canyon Shale is a primary cause of ubiquitous landsliding in hillsides underlain by these soils (Rollins and Rollins, 1992). Rollins and Rollins (1992) also indicate that these soils undergo a significant reduction in shear strength upon becoming wet.

Some laboratory soil-strength testing (U.S. Geological Survey, 2001; URS/Dames & Moore, 2001; and Terracon, 2001) defines the ranges in peak and residual friction angles for weathered Manning Canyon Shale at the Sherwood Hills landslide complex. Table 5 summarizes the available peak and residual friction angle data.

The local landslides in the complex in 1998 were also caused, in part, by slope modifications. The landslide near the intersection of Windsor Drive and Foothill Drive had its toe at the base of a steep cut slope. This historical landslide was mapped (In-

Table 4. *Summary of undrained shear-strength data for clay soils in the upper part of the Sherwood Hills landslide complex.*

Site	Range of Undrained Shear Strength (s_u) (psf)	Back-calculated s_u (FS = 1.0) (psf)	Soil Description	Plasticity Index (percent)
Windsor Drive - no. 1	200-2,000	270	Brown to black clay (CL)	15-30
Windsor Drive - no. 2	400-1,200 300 (near surface of rupture)	300	Gray to black clay	—
Mile High Drive - Imperial Way	—	750	Residual clay	—
Borehole SH-1	2,190	—	Silty clay (CL)	13

Source: Rollins and Rollins (1992) except borehole SH-1 (URS/Dames & Moore, 2001).
Abbreviations: FS = factor of safety, CL = inorganic low to medium plasticity clay.

Table 5. Summary of peak and residual shear-strength parameters for Manning Canyon Shale in the Sherwood Hills landslide complex.

Peak Friction Angle (degrees) (range)	Residual Friction Angle (degrees) (range)	Cohesion (psf) (range)	Test Type	Test Conditions	Plasticity Index (percent) (range)	Study
—	15	600	DS	CD	27	URS/Dames & Moore, 2001
—	14	160	RS	CD	—	USGS, 2001
—	16 (3-31)	504 (130-1,060)	DS	CD	30 (23-34)	Terracon, 2001
28 (24-31)	—	1,354 (58-2,650)	DS TR	CD CU	27 (23-31)	Terracon, 2001
20	—	2,826 (302-5,350)	DS TR	CU CU	32 (23-40)	Terracon, 2001

Test types: DS = direct shear, RS = ring (torsional) shear, TR = triaxial.
Test conditions: CD = consolidated-drained, CU = consolidated-undrained.

ternational Engineering Company, Inc., 1984) prior to construction of the house atop the main scarp of the landslide along Churchill Drive and was likely initially triggered after the cut slope was excavated during construction of Windsor Drive. In the upper part of the Sherwood Hills landslide complex, construction of Mile High Drive as well as other site grading likely contributed to subsequent landsliding in the mid-1980s (Rollins and Rollins, 1992) and renewed movement in 1998. Rollins and Rollins (1992) indicated that two drainages were blocked with fill during construction of the road, locally adding a surcharge load in the head of the landslide. The blockage of the drainages also likely increased the seasonal saturation of landslide debris.

Landsliding in 1998 was likely triggered by a cumulative rise in ground-water levels caused by the above-normal precipitation in the same year and the longer-term effects of landscape irrigation and site drainage modification in the subdivision. The 1995(1993)-98 precipitation period did not occur in the eastern part of Utah Valley; however, precipitation was above normal in four out of the six years from 1993 to 1998 in Provo. In nearby Orem, precipitation in four out of the first six months of 1998 exceeded 150 percent of normal. Precipitation in June 1998 was 781 percent of normal with more than 5 inches (13 cm) of excess precipitation falling in that month. Other factors including landscape irrigation and site drainage modification also likely contributed to a cumulative rise in ground-water levels. Many homeowners in the Sherwood Hills subdivision generously irrigate their water-consumptive landscape vegetation as evident by seepage of water at the curbs of lots during the summer. Rollins and Rollins (1992) indicated poor site drainage also contributes to clay soils becoming wet. At one local landslide within the Sherwood Hills area, Rollins and Rollins (1992) indicated that site grading had filled pre-existing drainages, interrupting flow and promoting infiltration of surface runoff into the subsurface soils.

Spanish Fork Canyon Landslides, Utah County

The two most spectacular landslides of 1998 in terms of their size and the total amount of movement occurred in Spanish Fork Canyon. Both landslides, the Shurtz Lake and Thistle landslides, threatened or disrupted utility lifelines. Landsliding was preceded in both areas by historical landsliding in the previous 15 years. The 1998 movement of the Shurtz Lake landslide involved reactivation of a 1997 landslide and enlargement by reactivation of pre-existing landslide (earth-flow) deposits downslope. Renewed movement of the Thistle landslide in 1998 involved most of the landslide above the 1983 blockage (Ashland, 1999), the landslide debris that had filled Spanish Fork Canyon and dammed the river forming Thistle Lake (Anderson and others, 1984).

Shurtz Lake Landslide

Introduction: In March 1998, the Shurtz Lake landslide (figure 25) (Ashland, 1997b) reactivated and the rate of movement in the lower part of the landslide accelerated until about mid- to late-April (Ashland and Horns, 1998). The landslide first became active in early May 1997 shortly following the spring snowmelt. Reactivation in 1998 involved downslope enlargement of the landslide, which began in an incipient phase in late May 1997. A new toe formed in prehistoric landslide deposits downslope of the 1997 landslide directly above the railroad grade abandoned in 1983 following the nearby Thistle landslide, a newly constructed municipal water line, and the Spanish Fork River. As it had in 1997, landslide movement displaced power-line transmission poles on the slide. Movement of the high-voltage lines over nearby trees caused electrical arcing and initiated small fires. Following the 1998 movement the power lines were realigned and the water line relocated to the opposite side of Spanish Fork Canyon.

Geology: Witkind and Page (1983) mapped the northeast-

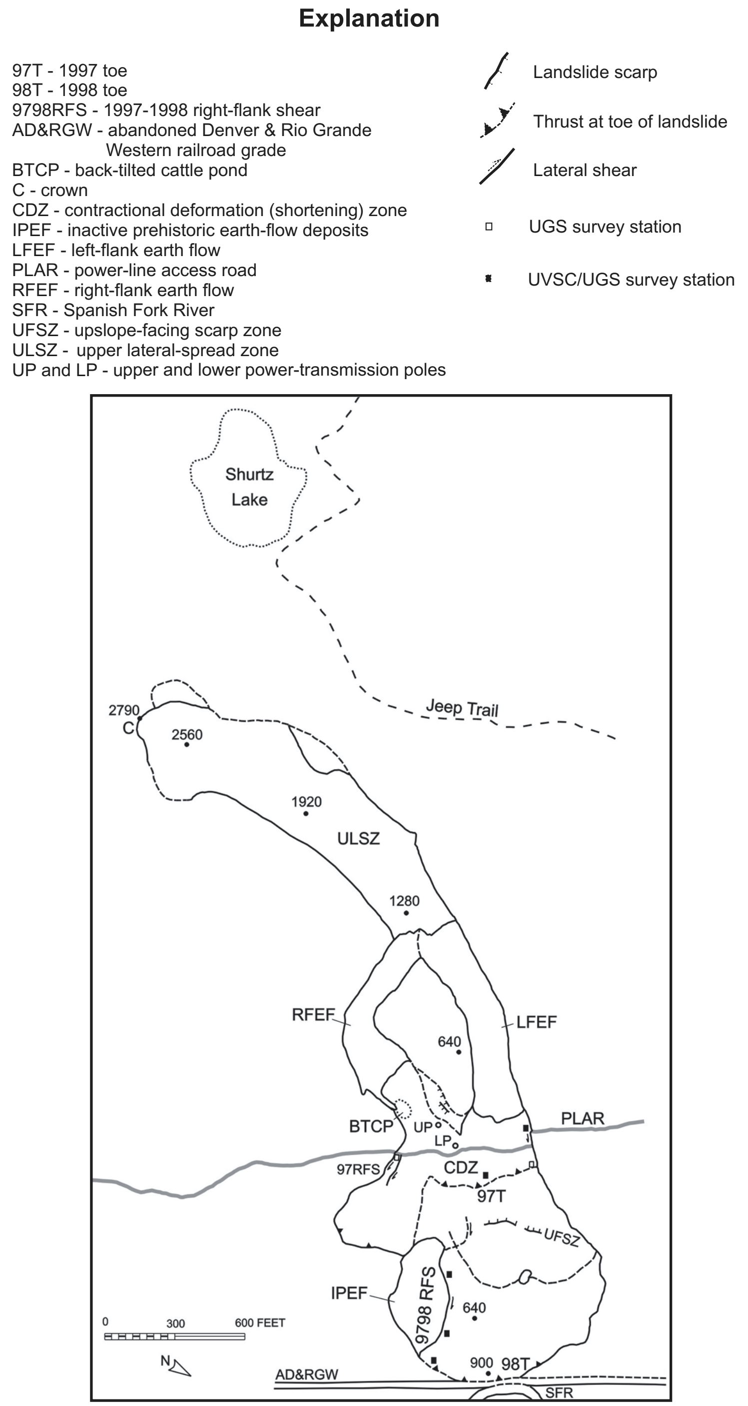


Figure 25. Sketch map of the Shurtz Lake landslide in Utah County. Numbers are approximate distances from the 1997 toe in feet. Mapped using aerial photographs dated October 14, 1998, provided by U.S. Geological Survey. Dashed lines show less well-defined landslide boundaries. Modified from Ashland and others (1999).

facing slope on which the Shurtz Lake landslide occurred as Tertiary North Horn Formation unconformably overlying Triassic Ankareh Formation. Recent surficial geologic mapping (Ashland and Horns, 1998) shows the hillside consists primarily of prehistoric landslide deposits overlying these units. Harty (1992) mapped the hillside surrounding the Shurtz Lake landslide as undifferentiated landslide, colluvial, and slope-creep deposits and indicated small historical landslides west of the slide. Harty's mapping also shows historical landsliding in the vicinity of the lowermost slope along the hillside. Soils in the landslide consist of clayey silt to silty clay (Arnal and Seigneur, 1998).

Landslide description: The Shurtz Lake landslide is a composite slide (figure 25), and consists of four areas having distinct landslide features (Ashland, 1997b; Ashland and Horns, 1998; Ashland and others, 1999). Two separate earth-flow areas exist on the right and left flanks in the steepest part of the landslide and are separated by a relatively intact, triangular-shaped area in the center of the slide. The average slope near the right-flank earth flow is about 50 percent. The heads of the earth flows coincide with an abrupt change in slope. Upslope of the earth flows, the average slope is about 14 percent. In this area, a lateral-spread zone extends for over 1,600 feet (490 m) upslope and to the south of the heads of the earth flows. Contractional (shortening) landslide-deformation features below the earth flows consist of a series of downslope-directed, shallow thrusts. By 1998, the landslide enlarged to include prehistoric landslide deposits between the Spanish Fork River flood plain and the 1997 toe. These deposits were bounded on the right flank by a left-stepping right-lateral shear zone which had initially appeared in May 1997. On the left flank, the prehistoric deposits were bounded by a left-lateral shear that connected a system of thrusts. Both bounding shears occurred at the crests of slopes abutting drainages that bounded the prehistoric deposits rather than in the drainage bottoms. The average slope in the lower part of the 1998 landslide is about 18 percent. The initial landslide volume, in early May 1997, was about 3.3 million cubic yards (2.5 million m^3). By 1998, the landslide volume had enlarged to about 4.5 million cubic yards (3.4 million m^3).

Causes and trigger: Movement in 1998, as well as initially in 1997, coincided with the spring snowmelt. Measurements by the UGS on March 10, 1998, indicated snow depths near the 1997 toe ranged between 16 to 23 inches (41-59 cm) with a snow-water equivalent of about 6 to 9 inches (15-22 cm). Thus, roughly 29 to 44 percent of the normal annual precipitation, about 21 inches (53 cm), was present in the form of snow on the landslide mass prior to landslide movement. Figure 26 shows that most of the movement in 1998 occurred by May 1. Only minor movement occurred in the remainder of the year in both 1998 and 1997.

The relation between landslide activity and annual precipitation is somewhat problematic given that the Shurtz Lake landslide did not move in 1983 or 1982, but did in 1997 and again in 1998. The 1993-98 precipitation period likely contributed to rising ground-water levels in the prehistoric landslide mass on which the Shurtz Lake landslide formed. By the end of 1982, about 16 inches (40 cm) of excess precipitation had fallen in the previous three years. However, by the end of 1996, only 12 inches (31 cm) of excess precipitation had fallen. Thus, ground-water levels may have been higher in the area in the early 1980s than they were in the late 1990s when movement initiated. Thus, the cumulative effects of the 1993-98 precipitation period alone on ground-water levels may not have been enough to exceed an instability threshold level and trigger

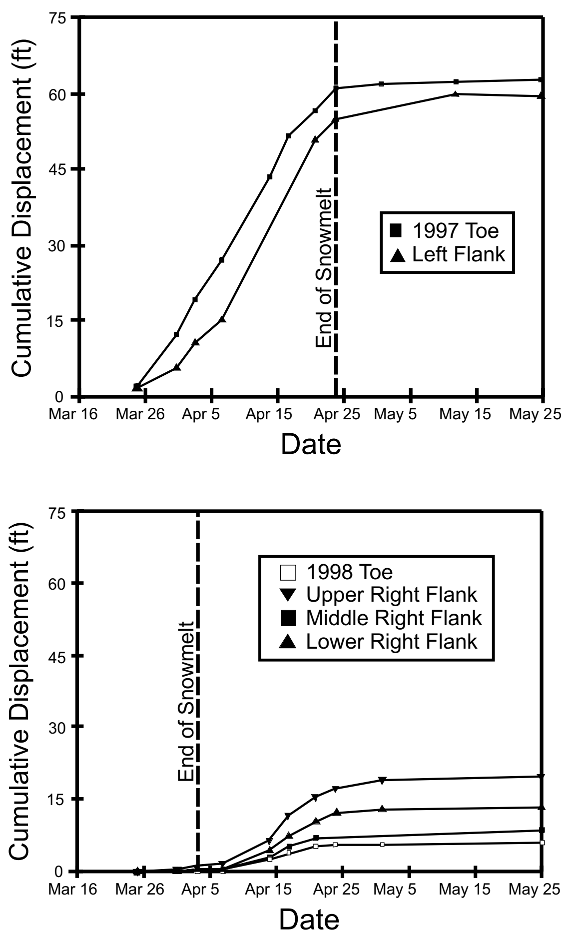


Figure 26. *Cumulative displacement of the lower part of the Shurtz Lake landslide in 1998. The majority of the movement occurred in the vicinity of the 1997 toe. Displacements based on tape measurements of survey stakes across various landslide deformation features. Later end of snowmelt date in upper plot is due to higher elevation of 1997 toe and adjacent stations compared to other stations (lower plot). Data from Ashland and Horns (1998). Modified from Ashland and others (1999).*

movement. Instead, short-duration fluctuations in ground-water levels that coincided with the snowmelt may have triggered landslide movement in the late 1990s, but not in the 1980s. During 1998, four months were wetter than in 1983: January, February, April, and June. As a result, the presence of 29 to 44 percent of the total mean annual precipitation on the ground surface in the form of wet snow in March 1998 may have been anomalous and resulted in a rapid transient rise in ground-water levels that triggered landslide movement.

Other causes of landslide movement in 1998 include residual shear-strength conditions following movement in 1997, loading of the downslope prehistoric landslide deposits by the 1997 landslide, and ground-surface-deformation features that promoted infiltration and a rapid rise in ground-water levels during the snowmelt. Arnal and Seigneur (1998) reported residual friction angles for clay and silt soils collected from the 1997 landslide ranging between 7 and 23 degrees. The presence of numerous slickensides in the 1997 landslide mass suggests cohesion was near zero. Thus, the landslide soils had limited ability to resist renewed movement in 1998 as ground-water levels rose with the snowmelt. The downslope enlargement of the landslide in 1998 was likely caused by loading of the lower slope by the 1997 landslide. Based on the inferred thrust angle of the 1997 Shurtz Lake landslide, both horizontal and vertical loads were likely exerted atop the downslope prehistorical landslide deposits. This loading likely caused a metastable state (factor of safety approximately or slightly above 1.0) to develop in the downslope prehistoric landslide deposits. Movement of these deposits was triggered once ground-water levels rose above an instability threshold in the spring of 1998. The importance of the loading to the failure of these deposits is indicated by their stability in early May 1997. Renewed movement in 1998 was also made more likely because of the presence of ground-surface-deformation features which promoted infiltration during the snowmelt and subsequent rainstorms in the spring. Ground cracks, shears, grabens, and upslope-facing minor scarps (figure 27) collected surface runoff and likely increased the percentage of water that infiltrated the landslide mass relative to the pre-failure infiltration rate in 1997. Such features likely lower the threshold for renewed movement by allowing for a more rapid rise in ground-water levels during the spring despite other stabilizing factors such as the reduced driving force resulting from a flatter post-failure landslide geometry.

Thistle Landslide

Introduction: Renewed movement of the Thistle landslide (figure 28) initiated sometime in May 1998 and, unlike any of the other 1998 Wasatch Front landslides, was triggered by crown instability. Crown instability consisted of detachment and downslope movement of a crescent-shaped part of the steep slope above the main scarp of the 1983 landslide. Downslope movement of the detached crown block caused reactivation of most of the Thistle landslide above the 1983 blockage, the landslide debris that filled Spanish Fork Canyon 15 years earlier. On the extreme right flank, movement extended down onto the steep slope adjacent to the 1983 blockage. Movement of a local lobe in this area offset a fence bounding the Utah County Sheriff's Department shooting range about 23 feet (7 m) during 1998 (figure 29). Landslide movement also distorted another property boundary fence that crossed the toe of the 1998 landslide. The 1998 movement threatened a recently constructed municipal culinary-water pipeline that crossed the eastern part of the 1983 blockage. Partial reactivation of the landslide had also occurred in 1997 (Ashland and others, 1999) primarily in the head and right (southeast) side of the slide.

Geology: The geology of the Thistle landslide is similar to that of the nearby Shurtz Lake landslide about a mile to the north. Witkind and Page (1983) mapped the landslide, which is in a northeast-trending tributary to Spanish Fork Canyon, as overlying Triassic Ankareh Formation and showed it flanked by colluvium on the west and northwest. The ridge abutting the right (southeast) flank of the landslide is Jurassic Nugget Sandstone. Tertiary units, the North Horn Formation and Flagstaff Limestone, unconformably overlie the Mesozoic rocks upslope and west of the landslide. Debris and soil derived from these units comprises a significant component of the landslide material. Duncan and others (1986) indicated that these soils consist primarily of low-

Figure 27. *Upslope-facing minor scarps in the lower part of the Shurtz Lake landslide. Scarps intercept overland flow, increasing infiltration and ground-water levels. View is toward the northwest.*

Figure 28. *View toward the southwest of the Thistle landslide in Utah County. Most of the landslide above the 1983 blockage reactivated in 1998. Arrow points to new main scarp which formed in 1998.*

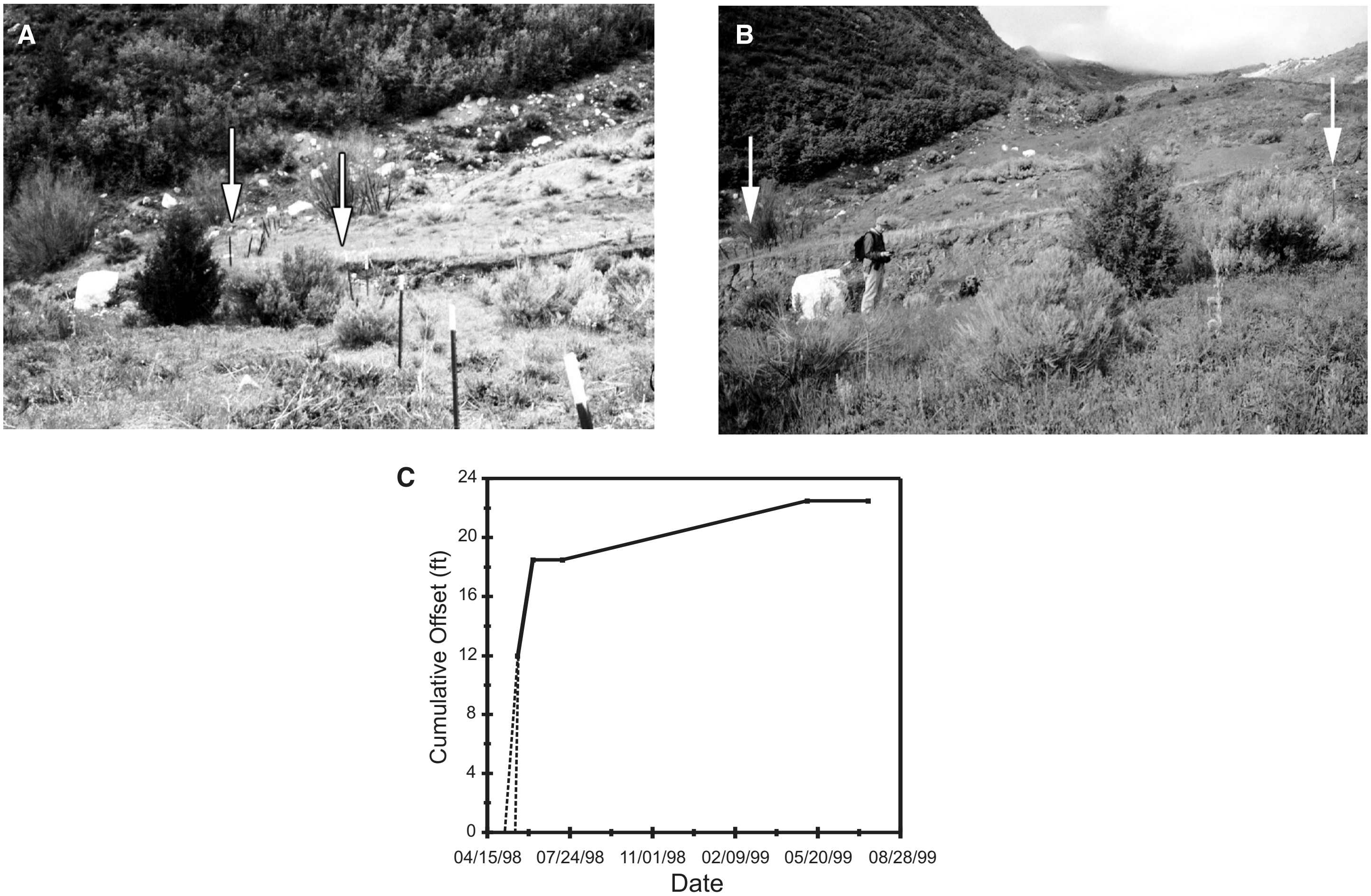


Figure 29. *Cumulative offset of a fence by a left-lateral shear in the right toe area of the Thistle landslide. Photographs show offset of fence on May 22, 1998 (A), and June 9, 1998 (B). Plot (C) shows cumulative offset of fence along shear between May 22, 1998 and July 20, 1999. Dashed lines show probable offset prior to May 22, 1998. Onset of movement constrained by field observations to have occurred between May 5 and May 20, 1998.*

strength clays (CL and CH) that were deposited in the tributary canyon by successive, retrogressive earth flows. Duncan and others (1986) recognized that the pre-1983 landslide mass consisted of five superimposed earth flows. Shroder (1971) mapped the steep slopes west of the pre-1983 landslide as the source area of these earth flows. The steep slopes are underlain by North Horn Formation (Duncan and others, 1986). Duncan and others (1986) speculated that the pre-1983 earth flows overlaid prehistoric landslide deposits.

Landslide description: The 1983 Thistle landslide was described in detail by Duncan and others (1986), but other descriptions of the landslide exist in the literature (Anderson and others, 1984; Slosson and others, 1992). Ashland (1999) indicated that the 1998 landslide involved the most of the 1983 landslide above the blockage (zone of accumulation). Landslide movement in 1998 was complex and consisted of earth flow (figure 30) along the right (southeast) side and translational sliding along the left (northwest) side of the slide. Most of the internal 1983 landslide structures identified by Duncan and others (1986) reactivated in 1998 and new thrusts and folds formed above the 1983 blockage. The thrusts appeared to propagate downslope and were preceded locally by ground tilting as indicated in one area by a raised upslope shoreline of a small pond on the landslide (figure 31). Other landslide deformation included lateral shears, folds, and upslope-facing scarps. The landslide enlarged significantly in 1998 as a large block of the 1983 crown area, estimated to be about 1.2 to 3.5 million cubic yards (0.9-2.7 million m^3) in volume, detached and rotated, displacing the 1983 landslide mass downslope.

Site investigations: Geologic investigations preceded both the 1983 and 1998 movement. Shroder (1971) mapped the pre-1983 landslide and recognized evidence of repeated movement. Duncan and others (1986) provided a detailed description of geologic and geotechnical conditions of the 1983 landslide and assessed the post-1983 stability. Duncan and others (1986) speculated that hundreds of years might be required following the 1983 landslide before enough debris collected in the head for renewed movement to occur. However, by late 1996, a sufficient amount of landslide debris had accumulated in the head of the landslide that driving forces locally exceeded resisting forces within the slide and partial reactivation occurred during the late winter and early spring in 1997. Duncan and others (1986) correctly speculated that crown instability, such as occurred in 1998, was the principal threat for any large-scale reactivation of the landslide.

Causes and trigger: As at the nearby Shurtz Lake landslide, renewed landslide movement in 1998 was preceded by most

Figure 30. *Earth flow in the upper right side of the Thistle landslide. View is toward the southwest.*

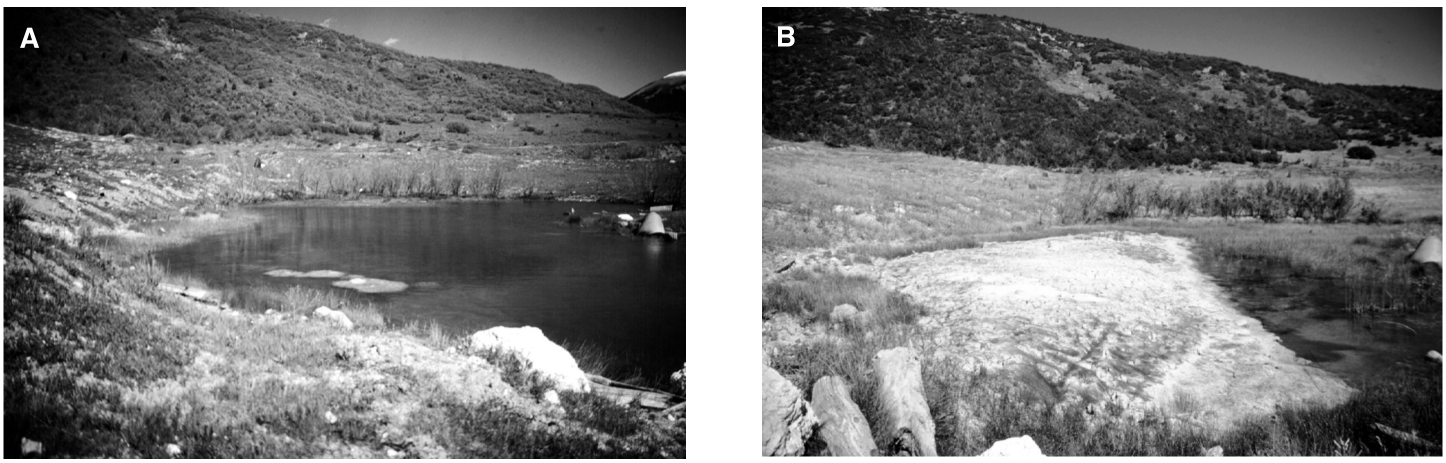


Figure 31. *Uplift and tilting of the upslope (left) shore of a small pond on the Thistle landslide. Photographs show pond on May 20, 1998 (A), and July 15, 1998 (B). View is toward the west.*

of the 1993-98 precipitation period. Three out of eight months between September 1997 and April 1998 were wetter than during the same period between 1982 and 1983; however, total precipitation for the entire period was about 85 percent of that in 1982-83, or 148 percent of normal. Uncertainty exists, however, regarding whether a gradual increase in ground-water levels between 1993 and 1998 alone triggered the crown instability that caused the 1998 movement. Renewed movement of part of the landslide in 1997 suggests that a rise in ground-water levels had reduced the stability of the landslide by the late 1990s. Thus, in 1998, the Thistle landslide was likely metastable.

Crown instability may have been caused by a combination of factors including removal of lateral support and a rise in ground-water levels in the crown of the 1983 landslide during the 1993-98 precipitation period. Significant removal of lateral support at the base of the crown slope occurred in 1983 as the head area of the landslide was evacuated due to downslope movement of the landslide. Duncan and others (1986) estimated local evacuation in the head area of the 1983 landslide in excess of 90 feet (27 meters). Evacuation of the head of the landslide occurred again in 1997 following 14 years of gradual accumulation as a result of earth-flow deposition from source areas on the steep slopes above the landslide. Thus, in both 1983 and 1997, lateral support was removed at the base of the steep slopes forming the crown area of the 1983 landslide.

Summary of Common Geologic and Physiographic Characteristics

Many of the 1998 Wasatch Front landslides shared several common geologic and physiographic characteristics (table 6). At most of the landslides, movement in 1998 involved reactivation, in some cases localized, of pre-existing landslides in which recurrent historical movement had previously been documented. These landslides consist of mostly shallow to deep-seated earth slides, but the larger Spanish Fork Canyon landslides are composite slides that moved by

Table 6. Comparison of geologic and physiographic characteristics of the 1998 Wasatch Front landslides.

Landslide	Type	Surficial Geology	Aspect	Pre-existing Landslide Type	Recurrent Historical Movement	Drainage
SD	Earth slide	Qms-Ql-Qd	NW	Prehistoric	Reported	At toe
CBCC	Earth slide	Qms-Ql-Qr	SE	Historical	Yes	Along left flank
SFKC	Earth slide	Qms-Ql-Qd	N	Historical	Yes	At toe
CBS	Earth slide complex	Qms-Ql-Qd	N	Prehistoric	Unknown	—
DWC	Earth slide	Qms-Ql-Qd	NE	Historical	Yes	—
GHCE	Earth slide	Qms-Qal-Qc	NW	Complex of historical ane prehistoric slides	—	—
Springhill	Earth slide	Qr-Tv	WNW	—	—	—
SHLC	Earth-debris slide/flow?	Qms-Qr	W NW	Complex of historical and prehistoric slides	Yes	Several cross complex
Shurtz Lake	CESEF	Qms-Qr	NE	Historical (98) Prehistoric (97)	Yes	Along left flank
Thistle	CESEF	Qms-Qr	NE	Historical	Yes	Along lower left flank; cross slide

Landslides include Sunset Drive (SD), E. Capitol Blvd-City Creek (CBCC), South Fork Kays Creek (SFKC), Cedar Bench Subdivision (CBS), Davis-Weber Canal (DWC), Green Hill Country Estates (GHCE), and Sherwood Hills landslide complex (SHLC).
Thistle and Shurtz Lake landslides are composite earth slides/earth flows (CESEF) after Cruden and Varnes (1996).
Surficial deposits include lacustrine deposits (Ql), deltaic deposits (Qd), alluvium (Qal), colluvium (Qc), landslide deposits (Qms), residual soils (Qr). The latter are derived from Tertiary volcanics (Tv) at the Springhill landslide.
Dashes indicate not present or documented.

a combination of sliding, flow, and lateral spreading. Seven out of ten of the landslides have a north-facing aspect. Six out of the ten were either bounded or crossed by flowing drainages. All the landslides in northern Davis County occurred in slopes cut into lacustrine deposits in the Weber River delta of Lake Bonneville. All the Utah County landslides occurred in pre-existing landslide deposits which possibly formed initially in residual/colluvial soils.

CHARACTERIZATION OF LANDSLIDING IN 1998

Detailed investigations of several of the 1998 Wasatch Front landslides (Ashland and Horns, 1998; Terracon Consultants, Inc., 1998; Terracon, 1998; Ashland and others, 1999; Giraud and Fadling, 1999) provide data on movement rates, the duration of movement, and total displacements. These parameters, as well as the relation among landslide movement, precipitation, and ground-water levels (where available), are further examined below.

The Rate of Landslide Movement

The rate of landslide movement measured or estimated at Wasatch Front landslides in 1998 and 1999 varied from very slow to moderate (0.005 to 230 centimeters per day) based on the movement rate descriptions of Cruden and Varnes (1996) (table 7). In the Wasatch Front urban corridor, west of the Wasatch Range, the rate of movement was very slow to slow (0.005 to about 5 centimeters per day). At these rates, landslide movement was imperceptible, or nearly so, but caused significant landslide damage where buildings straddled landslide deformation features. Unfortunately, the only complete record of movement in 1998 is from the Shurtz Lake landslide (Ashland and Horns, 1998; Ashland and others, 1999) in Spanish Fork Canyon. Continuous monitoring of the CBCC landslide began in early June 1998, only after the rate of movement at the toe of the landslide began to decrease. Inclinometer measurements at two landslides (Terracon, 1999, 2000) provided only the rate of movement in the latter part of 1998 because baseline measurements were recorded only after the inferred maximum rate of movement in the spring of 1998. Monitoring of the Thistle landslide began as much as several weeks after movement likely triggered; however, the average rate of movement before monitoring could be estimated from several measurements of landslide displacement. At four of the instrumented landslides, movement continued or renewed in 1999. These data are also analyzed in this report because they provide valuable insights concerning movement in 1998.

Duration and Timing of Landslide Movement

Monitoring of landslide movement at the Shurtz Lake landslide in 1998 (Ashland and Horns, 1998) and the CBCC landslide in 1999 provided data on the duration and timing of

Table 7. *Summary of movement rate data, Wasatch Front landslides, 1998-1999.*

Landslide	Measured or Estimated Rate (cm/day)	Rate Description[1]	Date of Measurements	Location/ Measurement Type	Notes	Source
Sunset Drive	0.02 0.006 0.0005 - 0.012	Very Slow Extremely Slow to Very Slow	6/10/98 - 7/22/98 7/22/98 - 9/9/98 9/9/98 - 10/1/99	H/I(B4) H/I(B2, B4)	— Higher rate if movement occurred only in Mar-Apr 1999	This study, based on plots in Terracon Consultants, Inc. (1998) and Terracon (1999)
South Fork Kays Creek	0.08 0.18 0.02	Very Slow	5/22/98 - 8/18/98 8/18/98 - 9/30/98 9/30/98 - 3/26/99	MS/SS	—	This study
Springhill	0.008 0.005 0.007 0.009	Very Slow	9/8/98 - 11/3/98 11/3/98 - 2/19/99 2/19/99 - 5/17/99 5/17/99 - 9/16/99	MB/I(B1)	Displacement weakening behavior (11/3/98 - 9/16/99)?	This study, based on plots in Terracon (1998, 2000)
E. Capitol Blvd.- City Creek	3.8 2.9 4.7	Slow	5/7/98 - 7/23/98 6/5/98 - 7/7/98 3/30/99 - 7/15/99	MS/SS T/SS	Maximum rate	This study
Sherwood Hills - Upper	0.18	Very Slow	5/11/99 - 5/31/99	MB/GPS	Higher rate may have preceded initial measurement	This study
Shurtz Lake	84	Moderate	3/24/98 - 6/09/98	T/SS	Maximum rate; 1997 toe	Ashland and Horns (1998)
Thistle	22 - 183 130 - 230 2.6	Slow to Moderate Moderate Slow	5/5-20?/98 - 5/22/98 5/5-20?/98 - 6/9/98 5/27/98 - 6/25/98	T/E MB/E T/O	Estimated average rate Estimated average rate Higher rate may have preceded initial measurement	This study, based on Ashland and others (1999) This study Ashland (1999)

[1]Rate description from Cruden and Varnes (1996).
Location on landslides: head (H), main body (MB), main scarp (MS), and toe (T).
Measurement type: inclinometer (I) with borehole number indicated in parentheses, survey stakes using measuring tape (SS), Global Positioning System surveying (GPS), other surveying (O), and estimation using other surveying method (E).

movement. Most of the movement at the Shurtz Lake landslide in 1998 occurred between March 25 and April 25; however, movement continued through late May. Thus, most of the movement occurred over a period of about 30 days, and nearly all movement occurred within a 90-day period (mid-March to mid-June) (figure 26). In 1999, most of the movement at the CBCC landslide occurred between about May 1 and June 7, and almost all the movement occurred between March 1 and June 15. In 1998, movement occurred nearly continuously between May 7 and the end of the year; suspending only for a brief period (about two weeks) in July (figure 32). The 1999 movement data suggest that movement in 1998 likely occurred at least two months prior to when monitoring began.

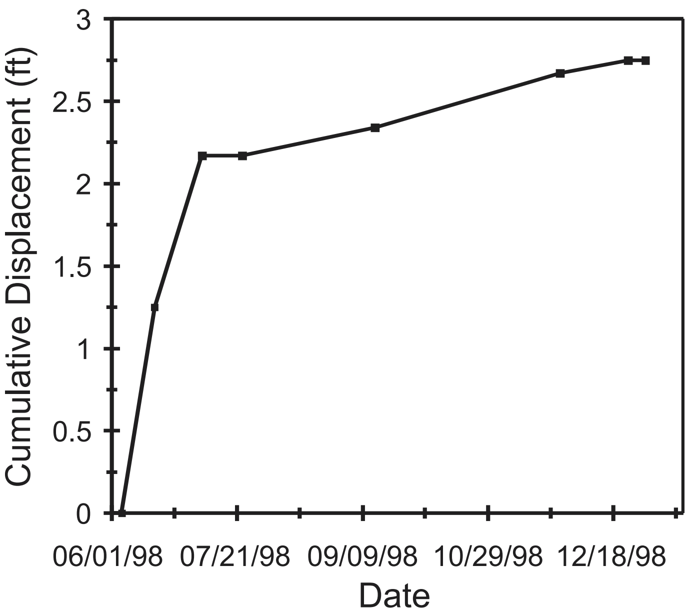


Figure 32. Cumulative displacement of the CBCC landslide toe in 1998. Initial measurement on June 5, 1998 was likely preceded by as much as 5.3 feet of displacement based on measurements in the spring of 1999. Movement suspended twice in 1998, once briefly in July and a second time at the end of December.

Movement also occurred during the latter part of 1998 at the Springhill landslide in North Salt Lake and the South Fork Kays Creek landslide in Layton. Inclinometer data (Terracon, 2000) from the Springhill landslide indicate continuous movement at a very slow rate since September 8, 1998 (figure 21). Similarly, measurements across the main scarp of the South Fork Kays Creek landslide in Layton (figure 13) show that continuous movement occurred at a very slow rate through September 1998.

At the remainder of the 1998 Wasatch Front landslides, no measurements on the duration or timing of damaging landslide movement exist. At four landslides (Sunset Drive, South Fork Kays Creek, Davis-Weber Canal, and Cedar Bench Subdivision), all in Davis County, observations suggest most of the movement occurred in late March and April. Thus, at most of the Wasatch Front landslides, significant movement appears to suspend by about mid-June.

Total Displacement Amounts

Total displacements of most of the 1998 Wasatch Front landslides are unknown. In Spanish Fork Canyon, the total displacements of the Shurtz Lake and Thistle landslides are reasonably well documented (Ashland and Horns, 1998; Ashland, 1999; Ashland and others, 1999). At most other landslides, total displacements in 1998 are unknown because monitoring began only after the initiation of damaging movement. At the CBCC landslide in Salt Lake City, only a fraction of the total displacement at the toe of the landslide was measured in 1998. Similarly, at the Sunset Drive, Springhill, and South Fork Kays Creek landslides, inclinometer and survey stake measurements record only a part of the total displacement at each slide in 1998. Table 8 summarizes the total displacements of some of the Wasatch Front landslides in 1998 and 1999.

Relation Between Landslide Movement and Precipitation

Figure 33 shows the relation between landslide movement and cumulative precipitation at both the CBCC landslide and upper part of the Sherwood Hills landslide com-

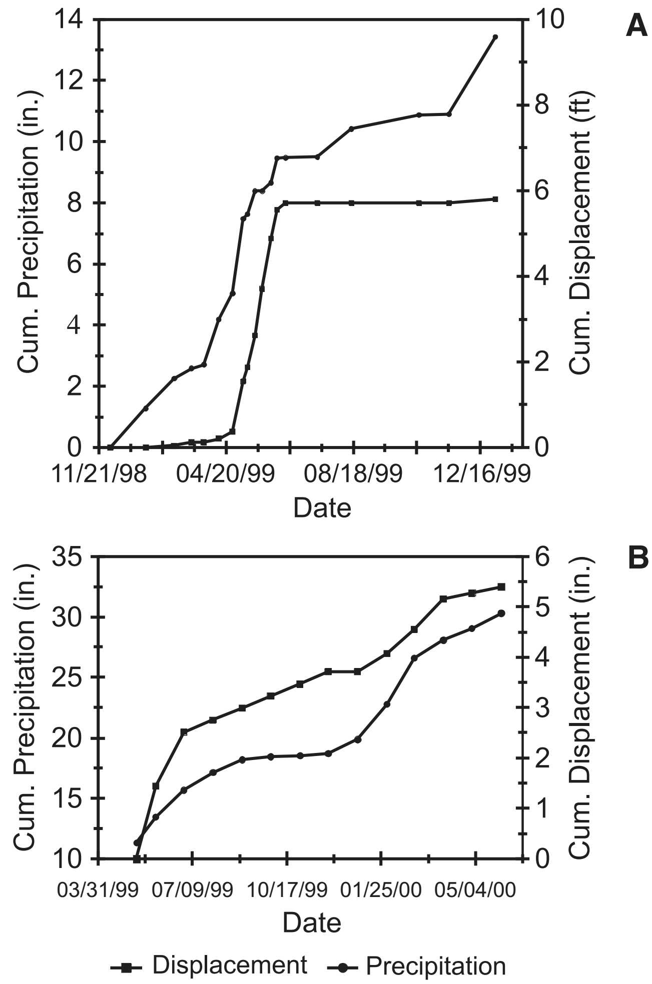


Figure 33. Relation between cumulative displacement and cumulative precipitation at two Wasatch Front landslides. Displacement shown for toe of CBCC landslide in Salt Lake City (A) and upper part of Sherwood Hills landslide complex in Provo (B). Both plots demonstrate parallelism between displacement and precipitation (see text).

***Table 8.** Summary of total landslide displacement of selected Wasatch Front landslides, 1998-1999.*

Landslide	Year (period of measurement)	Total Displacement feet (m)	Notes
Springhill	1998 (9/8/98 - 12/31/98)	0.025 (0.008)	Majority of movement inferred prior to 9/8/98
E. Capitol Blvd. - City Creek	1998 (6/5/98 - 12/31/98) 1999	2.75 (.84) 5.80 (1.8)	Significant movement inferred prior to 6/5/98
Sherwood Hills - Upper	1999 (5/11/99 - 12/31/99)	0.31 (0.09)	Some movement likely prior to 5/11/99
Shurtz Lake	1998 (3/24/98 - 5/25/98)	65 (20)	Minor (nominal) movement possible preceding and after measurement period
Thistle	1998	159 (48)	Estimated total displacement of lower main body

plex. The plots show that a considerable parallelism exists between cumulative displacement (landslide movement) and cumulative precipitation suggesting the rate of movement is influenced, at least indirectly, by the precipitation rate. In 1999, an increase in the rate of precipitation on March 30 preceded, by about four weeks, an increase in the rate of movement at the CBCC landslide. However, a decrease in the rate of movement on June 7 coincided with a decrease in the precipitation rate without any significant lag period. The data for the upper part of the Sherwood Hills landslide complex postdate the onset of movement in 1999; however, an increase in the precipitation rate preceded an increase in the rate of movement in January 2000 by about a month (the minimum resolution of the plot). The plot shows a lag period of about one month between a reduction in the rate of movement and a decrease in the precipitation rate in March 2000. However, the rate of movement decreased nearly simultaneously with a gradual reduction in the precipitation rate between May and July 1999.

DISCUSSION

The 1998 Wasatch Front landslides demonstrated the vulnerability of modified hillsides to landsliding, the potential for reactivation of pre-existing landslides, and the need to reassess the stability and state of activity of many of the area's landslides.

Implications of Historical Renewed Movement of Pre-Existing Landslides

Whereas this study documents renewed movement of pre-existing landslides during 1998, other researchers have documented similar renewed movement during the last three precipitation periods. Of the documented 1997 landslides in northern Utah (Ashland, 1997b, 1998a,b,c; Giraud, 1998; Solomon, 1998; Ashland and others, 1999) all three in the Wasatch Front occurred in pre-existing landslide areas. Kaliser and Slosson (1988) also described reactivation of some prehistoric landslides during the 1980-86 precipitation period; however, inadequate mapping of Wasatch Front landslides prior to 1983 prevented assessment of the relation between landsliding and pre-existing landslides. Pashley and Wiggins (1972) documented landslide movement at the South Weber landslide complex in 1970 and 1971 near the end of the 1967-71 precipitation period. One of the 1998 landslides also occurred at the edge of this complex (Black, 1999a).

Movement in South Weber landslide complex near the end of the 1967-71 precipitation period indicates that threshold levels were locally exceeded. Table 9 shows that the 1967-71 and 1995-98 precipitation periods in Salt Lake City are comparably ranked. Whereas mean precipitation is slightly higher during the latter period, the former extended for an additional year and thus cumulative excess precipitation was higher. Interestingly, the 1967-71 precipitation period is not recognized in the historical record in nearby Ogden, but above-normal precipitation occurred in 1970 and 1971. Thus, movement in the South Weber landslide complex during these years demonstrates either that (1) a boundary separating precipitation conditions in Salt Lake City and Ogden exists north of the complex, or (2) slopes in the complex are so marginally stable that landslide movement can be triggered by periods of above-normal precipitation shorter than the precipitation period defined in this report.

Table 9 shows that the 1995-98 precipitation period ranks near the bottom of the list of significant precipitation periods in Salt Lake City. Thus, renewed movement may have been triggered during other previous precipitation periods in the past 125 years, but may not have necessarily been recognized at many of the Wasatch Front landslides, particularly in the central Wasatch Front.

Despite the increased frequency of landsliding throughout the Wasatch Front in 1998, no significant movement occurred at the majority of pre-existing landslides. The number of known active landslides in 1998 represented only a

Table 9. Ranking of historical precipitation periods in Salt Lake City, 1875-1999.

Ranking Based on Mean Annual Precip.[1]	Ranking Based on Cum. Excess Precip.
1875-78	1980-86
1906-09	1875-78
1980-86	1906-09
1922-25	1922-25
1995-98	1967-71
1967-71	**1995–98**
1944-47	1944-47
1896-99	1896-99

[1]*Mean annual precipitation for period.*

small fraction of the landslide inventory in the Wasatch Front. At the Chalk Creek landslide near Coalville (Ashland, 1998b), a slide that had been active in 1997, no measurable movement was detected in 1998 despite significantly above-normal precipitation in the area during the year. However, other landslides that were active in 1997, such as the Thistle and Shurtz Lake landslides in Spanish Fork Canyon, experienced renewed movement in 1998. Similarly, in 1983 not all historical landslides reactivated.

Excluding the Springhill landslide and shallow debris slides on local steep mountain slopes, the thresholds for landslide movement outside of pre-existing landslide areas apparently were not exceeded during the 1995-98 precipitation period. The threshold levels, however, were apparently exceeded at several sites during the 1980-86 precipitation period (Lowe, 1988b,c,d).

The reactivation of some pre-existing landslides during precipitation periods of only four or more years suggests many slides may be only marginally stable. The apparent susceptibility of some pre-existing landslides to reactivation suggests that factors of safety, the ratio between the forces resisting movement and those driving it, likely range between 1.0 and 1.3 (marginally stable) and are less than values (greater than 1.5) commonly estimated in slope-stability analyses. Back-calculation of the stability of the CBCC landslide in Salt Lake City (this study) shows only marginal stability (FS = 1.2) if the ground-water level is 10 feet (3 m) lower than in 1998. Although the ground-water level at the landslide during dry periods (that is, circa 1966) is undocumented, the ground-water level is unlikely to have recently been more than 20 feet (6 m) below the 1998 level, the condition required for the factor of safety to be equal to 1.5.

A Proposed Classification for Wasatch Front Landslides

One implication of the discussion above is that many of the landslides formerly classified as prehistoric (last movement occurred over 100 years ago) may actually have experienced historical landslide movement. Factors responsible for the incorrect classification of some of these landslides include remoteness at the time of historical movement (no one observed the movement) and the lack of easily recognizable landslide deformation features.

Until recently, many Wasatch Front hillsides remained undeveloped. Photographs in Kaliser and Slosson (1988) show that some of the documented 1983 Wasatch Front landslides occurred in undeveloped areas or on undeveloped hillsides abutting residential areas. Many hillsides that historically experienced only minor movement may have gone undetected because they only affected undeveloped areas that were relatively remote. Prior to 1980, much of the Wasatch Front was sparsely developed and most of the Wasatch Front landslides were unmapped. Thus, documentation of historical landslide movement is rare (Kaliser and Slosson, 1988).

The absence or limited size of landslide deformation features that reveal historical movement is commonly a result of the rate and duration of landslide movement, and the dominance of other surficial processes. Several of the 1998 Wasatch Front landslides demonstrated that because of the generally slow rates and short duration of movement (as described above), landslide features indicative of recent activity may measure only inches in height. At the Sunset Drive landslide, the maximum scarp height in 1998 was only about 12 inches (30 cm). A minor scarp lower in the slope was less than the height of the ground cover, making it difficult to relocate during numerous field visits by UGS geologists. Such features would not be recognizable on aerial photographs typically used by geologists to identify and classify landslides. Surficial processes may quickly erode, cover, or obscure these subtle landslide features. Slope-wash deposits and colluvium may fill in ground cracks and bury small scarps. In less than a decade, surficial evidence of recent landslide movement may be difficult to detect even by an investigator in the field unless subsurface explorations, such as trenching, are conducted. Thus, historical movement of some pre-existing landslides in undeveloped areas may be unrecognized because the slides experienced only minor movement when most recently active. Reconnaissance investigations that lack subsurface explorations and rely solely on geomorphic classifications of landslides such as proposed by McCalpin (1984) may incorrectly conclude that "no evidence of recent instability exists" in some cases where historical, but minor movement has occurred.

As a substitute to the geomorphic landslide classification of McCalpin (1984), an activity-based classification is proposed (table 10). The proposed classification is modified from Cruden and Varnes (1996) and more reasonably addresses the potential marginal stability of many pre-existing landslides in the Wasatch Front. Implementation of this classification may require a re-inventory of mapped Wasatch Front landslides with special attention paid to determining whether historical reactivation has occurred.

Prehistoric-dormant is the preferred default classification of any landslide lacking information regarding the state of activity and evidence of historical movement (occurring in the last 100 years). Previous researchers (McCalpin, 1984; Keaton and DeGraff, 1996) have proposed classification of prehistoric landslides using geomorphic relative age criteria. Landslides with more subtle or weakly defined features are assumed to be older than landslides with sharply defined and distinct features. However, several of the landslide case histories in this study have demonstrated that small displace-

Table 10. Proposed landslide activity classification.

State of Activity	Description	Limitations/Implications
1a. Historical - active (includes reactivated landslides)	Measurements or observations indicate current movement	Most Wasatch Front landslides move only between March and June; limited data available
1b. Historical - suspended	Recent movement ended in past 12 months	Renewed movement likely unless new geometry and subsurface conditions indicate stability
2. Historical - dormant	No movement in past 12 months; probable causes of movement remain or likely to occur again	Renewed movement probable; likely if slope modifications[1] or precipitation periods occur; limited data exist on historical landslides prior to 1983 (Kaliser and Slosson, 1988)
3. Prehistoric - dormant	No record of historical movement (last 100 years); probable causes of movement remain or likely to occur again; likely late Holocene	Renewed movement possible; likely if slope modifications or precipitation periods occur; lower probability of movement than historical landslides
4. Prehistoric - relict	Landslides which formed under geomorphic or climatic conditions different from present conditions; latest Pleistocene to early Holocene	Renewed movement unlikely unless significant slope modifications occur; unlikely to reactivate due to natural ground-water-level fluctuations
5. Stabilized	Landslides stabilized by physical measures (caissons; tiebacks; maintained drain systems)	Preferably instrumentation demonstrates effectiveness of stabilization measures

[1]Includes cuts, fills, landscape irrigation, and drainage modification.

ments may preclude the development of sharply defined landslide deformation features that would be obvious on aerial photographs or during a cursory field reconnaissance. Thus, careful field investigation is required to recognize geologically recent movement in a landslide characterized by "mature" or subdued features. The use of the term prehistoric in this proposed classification purposely includes the possibility of recurrent and geologically recent (late Holocene) movement even in otherwise apparently "mature" landslides and counters the inference by many that prehistoric is equivalent to late Pleistocene. The term dormant indicates that the causes of movement remain apparent (Cruden and Varnes, 1996) and implies a potential for reactivation. Several of the 1998 Wasatch Front landslides described in this report (Shurtz Lake landslide, Sherwood Hills landslides, Sunset Drive landslide) involved the partial reactivation of prehistoric landslides. Partial reactivation of prehistoric Wasatch Front landslides was also documented during the 1980-86 precipitation period (Kaliser and Slosson, 1988).

Recognition of Instability Threshold Ground-Water Levels

Recognition of instability threshold ground-water levels requires a combination of continuous ground-water-level and landslide-movement monitoring. Although such data are rarely available, ground-water-level and inclinometer data from 1999 provide some insight into what the instability threshold level might be at the Sunset Drive landslide in Layton. Figure 34 shows the fluctuation in ground-water level of a well near the crest of the slope beginning in June 1998. The ground-water level generally declines until September 1998 and then rises through the fall and winter until April 1999. Inclinometer data (Terracon, 1999) suggest minor movement between September 1998 and October 1999. Since a significant drop in ground-water level occurred after April 20, 1999, the landslide movement is inferred to have coincided with the high levels prior to that date. Unfortunately, the inclinometer data do not define the exact timing of the landslide movement. Landslide activity during this time period could involve any of the following scenarios:

1. Continuous, extremely slow movement occurs between September 1998 and October 1999.
2. Continuous, extremely slow movement suspends in early May 1999 corresponding to a significant ground-water-level decline.
3. Renewed, very slow movement occurs for a short duration in association with the rise in ground-water level during the spring of 1999.

Under scenario 1, an instability threshold ground-water level is not identifiable, because continued movement is inferred even at the low ground-water levels in the latter part of 1999. The instability threshold ground-water level in this scenario would be below the lowest ground-water level (about elevation 4,696.5 feet) recorded on August 8, 1999. Under scenario 2, an instability threshold ground-water level can be inferred between elevation 4,699 and 4,700 feet. Under scenario 3, an instability threshold ground-water level can be inferred between elevation 4,700.5 and 4,701 feet. One inference from this scenario is the possibility that movement had suspended for a short period when the ground-water level fell below the threshold between late July and October 1998. More frequent inclinometer data are needed to test the validity of the three scenarios. However, scenario 3 is favored because it is both geologically reasonable and the high instability threshold ground-water level explains the onset of landslide movement in 1998 rather than in previous years.

Recognition of Instability Threshold Precipitation Levels

In the absence of ground-water-level data at most of the Wasatch Front landslides, precipitation data were also examined in an attempt to recognize instability threshold levels. An informal landslide water year (LWY) was adopted, which begins September 1, and used to plot cumulative precipitation versus movement and to monitor antecedent precipitation during the months preceding the onset of movement. Exceptionally above-normal precipitation began in September 1982 in Spanish Fork Canyon prior to the onset of movement of the Thistle landslide in early April 1983. The excess precipitation in September equaled more than half the cumulative excess precipitation by the end of March 1983. In most Wasatch Front areas, June through August are the driest months of the year during which landslide movement suspends.

A lack of significant movement in 1999 or 2000 at some landslides allowed determination of preliminary instability threshold precipitation levels (figure 35). At the Sunset Drive landslide in Layton (figure 35 a), the instability threshold precipitation level was exceeded during the 1997-98 LWY suggesting that the level was above the normal precipitation level from March to May. In the following two LWYs, cumulative precipitation from March to May was below normal and no significant movement occurred (also see figure 34). At the CBCC landslide in Salt Lake City (figure 35b), the instability threshold precipitation level during the 1997-98 LWY was likely above the normal precipitation level (shaded area). However, the instability threshold precipitation level for March and April fell below the normal precipitation level during the following LWY (1998-99) subsequent to about 9 inches (23 cm) of antecedent excess precipitation during the 1995-98 precipitation period. The lack of movement during the 1999-2000 LWY provided a lower limit for the new threshold level. At the Shurtz Lake landslide (figure 35c), 1997-98 LWY precipitation levels from March to May exceeded the instability threshold precipitation level which was between the 1996-97 LWY and the normal precipitation levels (shaded area). Future renewed movement of these landslides, or the lack thereof, should better define the instability threshold precipitation levels as well as provide information on the duration of above-normal precipitation required to trigger movement, and the influence of landslide geometry, boundary conditions, and other factors on slope stability.

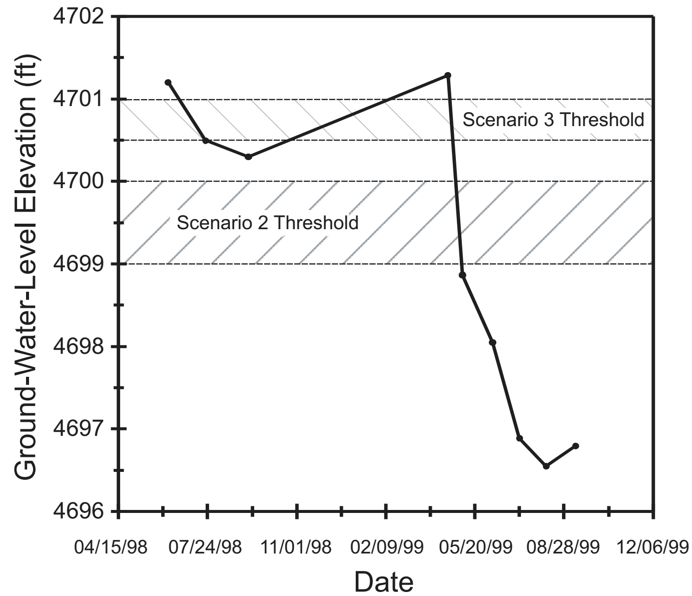


Figure 34. *Possible instability threshold ground-water levels for the upper part of the Sunset Drive landslide in Layton. Curve shows ground-water-level flucuation in well B-2 (see figure 6 for well location description) between June 8, 1998 and September 10, 1999. An adjacent inclinometer indicated some movement occurred between September 9, 1998 and October 1, 1999. Movement most likely occurred prior to rapid decline in ground-water level between May and August. See text for explanation of two possible scenarios which constrain instability threshold ground-water levels. Scenario 3 is the favored scenario and yields higher instability threshold ground-water level.*

The Probability of Future Precipitation Periods and a Continuing Wet Cycle

Precipitation periods have been rare throughout most of the historical precipitation record in the northern (Ogden) and southern (Spanish Fork Canyon) Wasatch Front, but have been more frequent in the central Wasatch Front (Salt Lake City). The Wasatch Front historical precipitation record was examined to characterize climatic variability and provide some insights into the probability of past and future landslide activity. The discussion that follows is limited by the shortness of the historical precipitation record (1875-present in Salt Lake City) and the non-uniformity of the initial record dates (the first year precipitation records were made at a site). The opinions and conclusions made herein are preliminary, and some might change or be abandoned given a more complete record.

Variation in Historical Precipitation

Mean annual precipitation in the Wasatch Front area, during the period of the historical record, has fluctuated significantly. In Ogden (northern Wasatch Front), mean annual precipitation for the decades between the 1900s and the present has varied by nearly 9 inches (23 cm) (figure 36a). In the latter half of the twentieth century, mean annual precipitation rose more than 8 inches (20 cm) between the 1950s and the 1980s and in the 1990s remained about 6 inches (15 cm) higher than in the 1950s. In Salt Lake City (central Wasatch Front), mean annual precipitation for the decades between

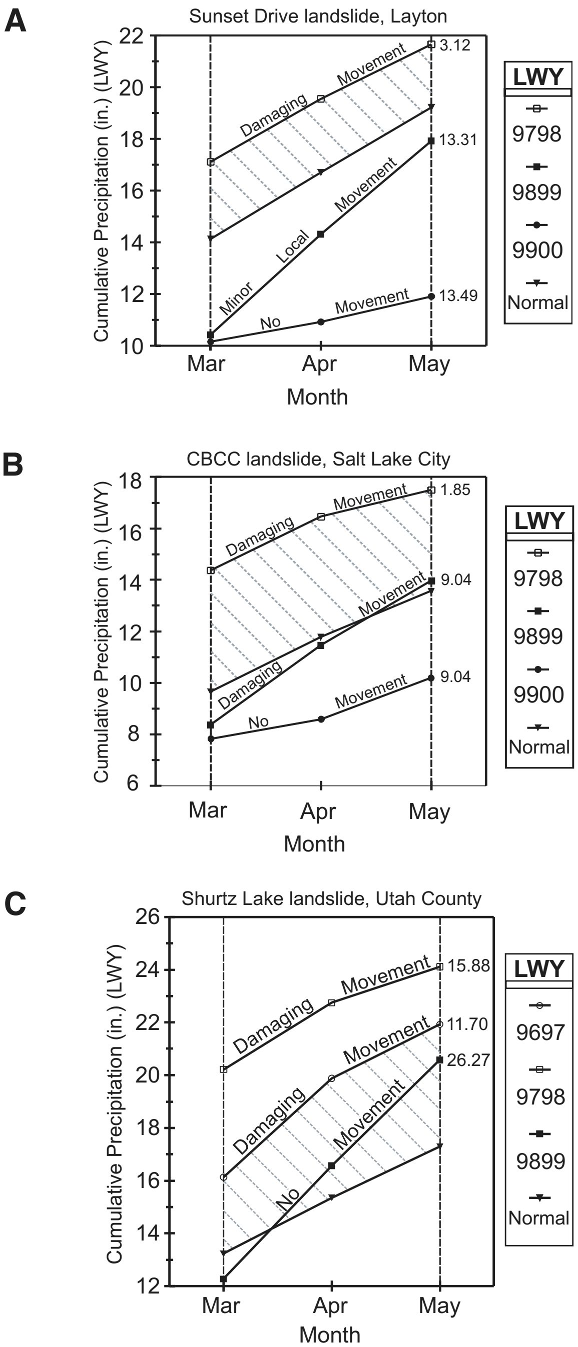


__Figure 35.__ Limits on instability threshold precipitation levels at three Wasatch Front landslides. Plots show cumulative precipitation for three months of the informal landslide water year (LWY) which begins on September 1. The months of March through May coincide with the snowmelt, depending on elevation, during which a transient ground-water-level rise or pulse may trigger landslide movement. During the 1997-98 LWY, damaging movement occurred at the Sunset Drive landslide in Layton (A); the CBCC landslide in Salt Lake City (B); and the Shurtz Lake landslide in Spanish Fork Canyon, Utah County (C). Damaging movement continued during the 1998-99 LWY at the CBCC landslide (B) but suspended at the other two. Antecedent excess precipitation (in inches) for the preceding LWYs is shown to right of the curves. Shaded areas show limits on the threshold levels for initial year of movement and for the antecedent excess precipitation shown (i.e., the instability threshold levels are within the shaded area). At the CBCC landslide (B), the threshold level dropped below the normal precipitation curve in the 1998-99 LWY subsequent to about 9 inches of antecedent precipitation. The threshold level at the Shurtz Lake landslide remained above the normal precipitation level during the second year of movement in the 1997-98 LWY.

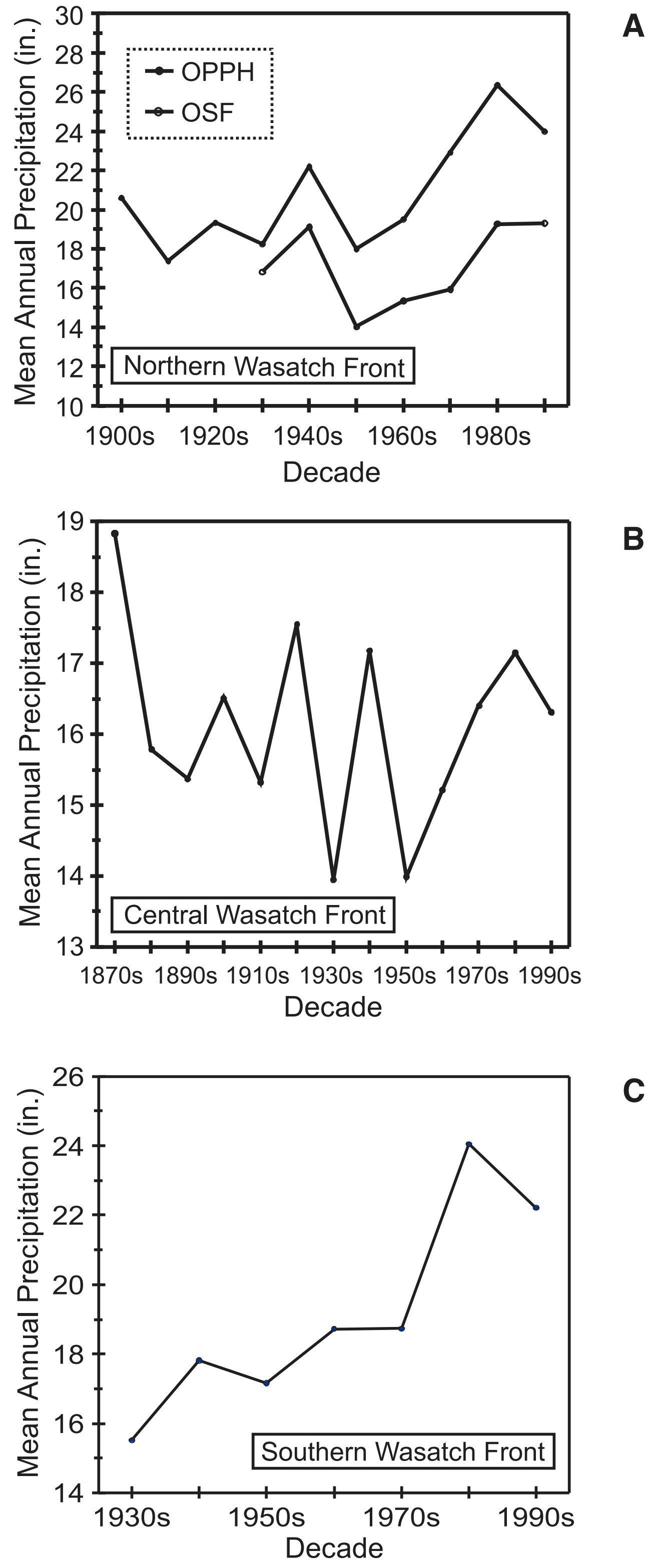


__Figure 36.__ Comparison of mean annual precipitation by decade along the Wasatch Front. Plots show precipitation has varied significantly during the period of historical record (see text). In the northern (A) and southern (C) Wasatch Front, the last two decades were the two wettest on record. In the central Wasatch Front (B), the 1870s, 1920s, and 1940s were wetter than either of the last two decades. Only a nominal difference exists between the 1940s and either of the last two decades at one northern Wasatch Front weather station (OSF - Ogden sugar factory). The last two decades were significantly wetter than the 1940s at the Ogden Pioneer powerhouse (OPPH) station.

the 1870s and the present has varied by more than 4 inches (10 cm) (figure 36b). Mean annual precipitation rose more than 3 inches (8 cm) between the 1950s and the 1980s, and in the 1990s remained more than 2 inches (5 cm) higher than in the 1950s. In Spanish Fork Canyon (southern Wasatch Front), mean annual precipitation rose more than 8 inches (20 cm) between the 1930s and the 1980s (figure 36c) and in the 1990s remained more than 6 inches (15 cm) higher than in the 1930s.

Another aspect of the historical precipitation record is increasing variability in annual precipitation in the latter part of the record. Between 1876 and 1965, annual precipitation in Salt Lake City fluctuated between 10 and 22 inches (25-56 cm) (figure 37b). From 1966 to 1999, annual precipitation fluctuated between 8 and 25 inches (20-64 cm). Three of the wettest years in the last 20 years of the record exceeded the limits of the previous part of the record. Precipitation in 1875, the third-wettest year on record, also exceeded the upper bound of the limit, possibly suggesting that a period of increased variability occurred prior to and ending in 1875. Elsewhere along the Wasatch Front, the available historical precipitation record is shorter in length than in Salt Lake City. In Ogden, with the exception of the 1983 "wet year," increasing variability in annual precipitation is not apparent in the last few decades (figure 37a). In Spanish Fork Canyon, annual precipitation (figure 37c) shows increasing variability beginning in the mid-1970s. The six wettest years in the 72-year record occurred during this period of increased variability.

The Recent Wet Cycle

The duration, amount of annual excess precipitation, and relatively close spacing of the last three precipitation periods has resulted in a period of cumulative excess precipitation referred to as a "wet cycle" (Fleming and Schuster,1985; this study) which may be ongoing (or continued at least through 1998). Prior to the "wet cycle," a cumulative deficit from mean annual precipitation had accumulated (figure 38). This deficit was greatest in 1966, but vanished by 1983. Figure 38 shows that during the "wet cycle," excess precipitation from three precipitation periods (1980-86, 1967-71, and 1995-98; ranked in order of their magnitude) eliminated the deficit. The 1995-98 precipitation period negates the temporary return to a cumulative deficit condition caused by drier years in the late 1980s and early 1990s. The recent "wet cycle" contains both precipitation periods (1967-71 and 1980-86) which exceeded four years in duration. In addition, the gap in time between the three periods was a uniform 8 years, 6 years less than the average time gap (14 years) between all previous precipitation periods, which ranged between 6 and 19 years.

During the recent wet cycle, a disproportionate number of wet years and months occurred as compared to what occurred in the previous part of the historical precipitation record. Precipitation was above normal in 13 of the 20 years from 1980 to 1999 and from 1970 to 1989 in Salt Lake City. At no time between 1880 and 1969 have this many years of above-normal precipitation occurred in two consecutive decades. In addition, 5 of the 12 wettest months on record in Salt Lake City occurred in the 1980s and 1990s. Record precipitation occurred in both February and June of 1998.

One other "wet cycle" occurred between 1906 and 1927

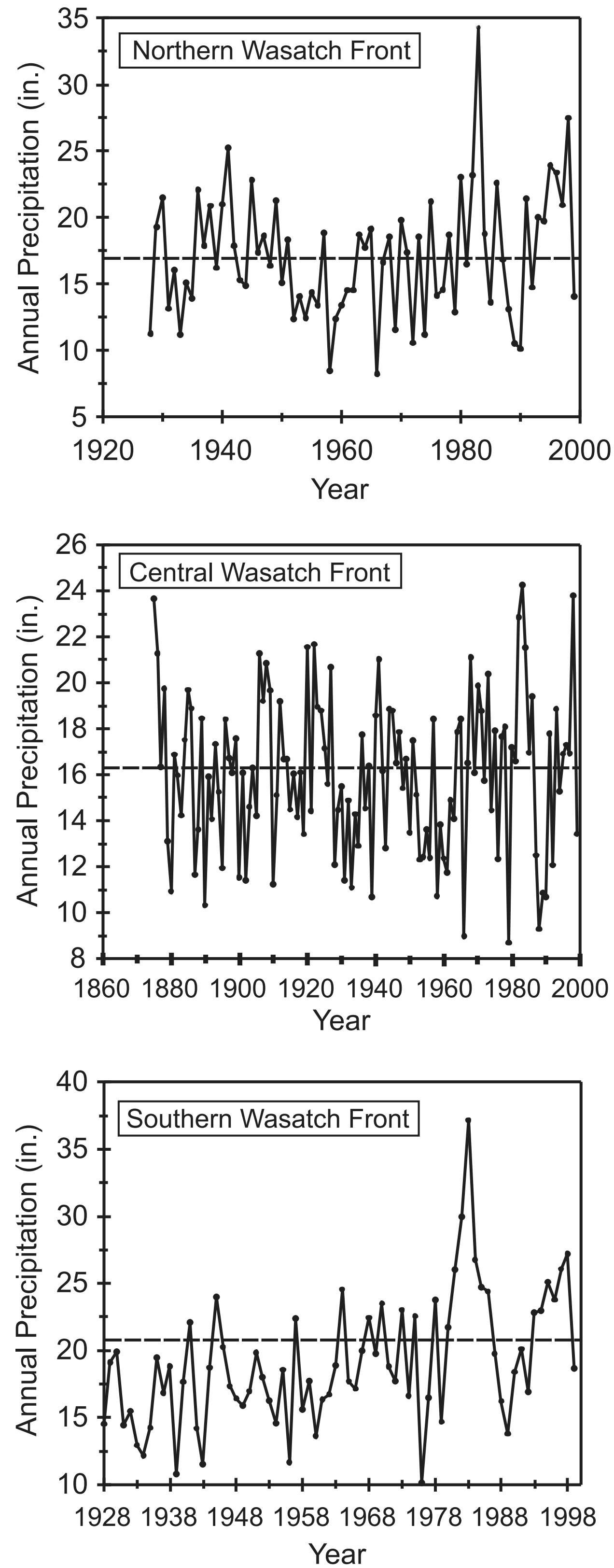


Figure 37. *Comparison of annual precipitation along the Wasatch Front. Data from National Weather Service and Western Regional Climate Center. Mean annual precipitation (thick dashed line) for northern Wasatch Front calculated from the historical precipitation record for the years between 1928 and 1999, inclusive. At other stations, thick dashed line shows normal annual precipitation reported by National Weather Service.*

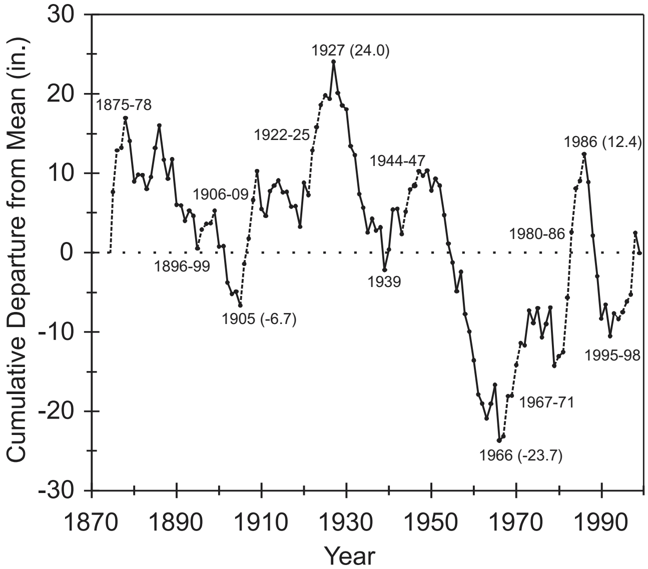


Figure 38. Cumulative departure plot for Salt Lake City (modified from Arnow, 1984). Plot shows variation from mean annual precipitation (16.02 inches for 1875-1999). Positive values represent cumulative excess conditions while negative values represent cumulative deficit conditions. The eight precipitation periods defined by four or more successive years of above-normal precipitation are indicated. Other significant years, including those representing maximum cumulative excess or deficit conditions, or bounding precipitation periods or wet cycles, are also shown. The two wet cycles in the historical record occur between 1906 and 1927, and 1967 to present (or at least to 1998). Values in parentheses indicate cumulative excess or deficit amounts in inches.

(figure 38) and allows comparison with the recent "wet cycle." During the earlier "wet cycle" a net excess of 30.7 inches (78 cm) of precipitation occurred, resulting in a change from a cumulative deficit to a cumulative excess condition. In comparison, the recent "wet cycle" resulted in a net excess of 36.1 inches (92 cm) between 1967 and 1986, inclusively, eliminating the cumulative deficit condition that existed by 1966. By 1998, the net excess precipitation was reduced to 26.2 inches (67 cm) as a result of an interim dry period between 1987 and 1992.

Whether the recent cycle continues or ceased in 1998 is uncertain. The possibility of a sustained "wet cycle" or future significant precipitation periods is difficult to assess; however, some data suggest both are likely. These include:

1. a general increase in mean annual precipitation since about mid-century,
2. the occurrence of wetter periods than the 1980s in the historical record (central Wasatch Front), and
3. the occurrence of many of the wettest months and years in the latter part of the historical record possibly associated with the increasing variability in annual precipitation (central and southern Wasatch Front).

Data supporting a return to drier conditions in the near future include: (1) the slight decline in mean annual precipitation in the 1990s as compared to the 1980s (central and southern Wasatch Front), and (2) the unprecedented length of the current "wet cycle" in Salt Lake City (defined both by variability in annual precipitation and cumulative departure from mean annual precipitation).

The Relation Between the Wet Cycle and Ground-Water Levels

The recent "wet cycle" has caused a gradual rise in ground-water levels in some Wasatch Front wells. Water-well data (Johnson, 1999; Kenney, 1999) near Salt Lake City show ground-water-level fluctuations associated with the 1980-86 and 1995-98 precipitation periods and intervening drier period in the Salt Lake City area. Ground-water levels in several wells were actually higher in 1998 or 1999 than in the 1980s (figure 39). Interestingly, these wells are the closest to the Springhill and CBCC landslides, which moved in

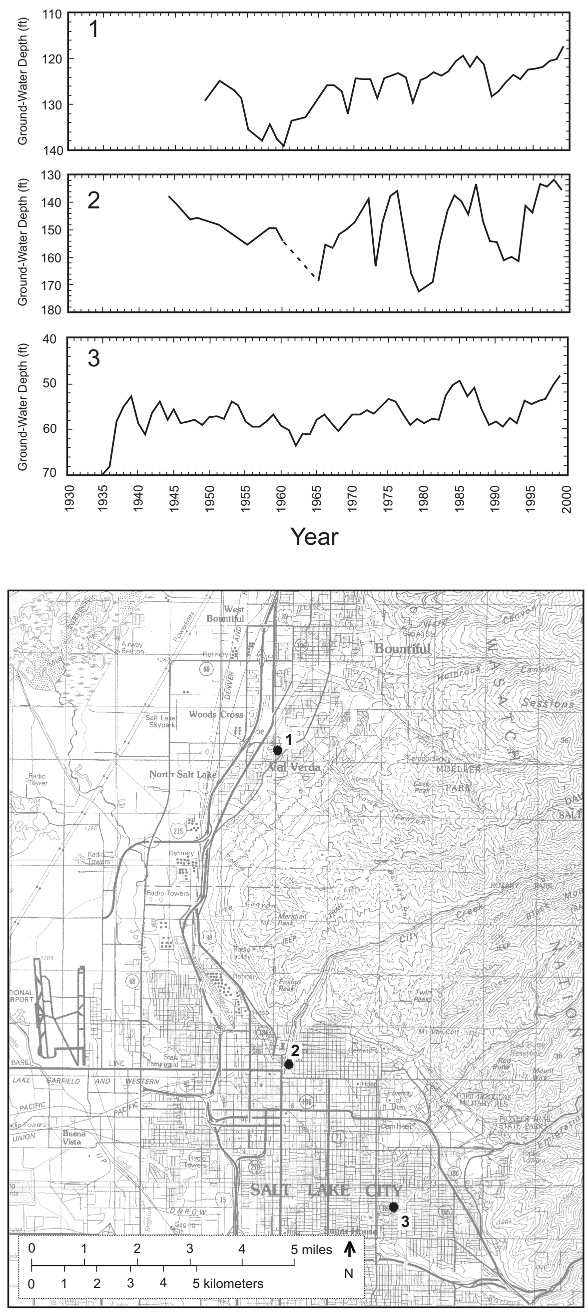


Figure 39. Ground-water-level fluctuations at three wells in Salt Lake City and southern Davis County. In all three wells the highest (shallowest) ground-water levels occurred in the late 1990s. Dashed line indicates gap in data. Ground-water-level plots modified from Johnson (1999) and Kenny (1999).

the late 1990s, but in the case of the Springhill landslide, lacked significant movement in the 1980s. Although these wells do not show the actual ground-water-level fluctuations in the landslides, the long-term increase in ground-water levels may have also occurred at the slides. Thus, the ground-water levels in the landslides may have been lower in the 1980s than in the late 1990s. A sustained wet cycle or future significant precipitation periods could cause ground-water levels to rise above historical high levels in some landslide areas in the Wasatch Front.

Implications to Slope-Stability Assessments

Many pre-development slope stability assessments evaluate only "present conditions." As discussed above, consideration of both seasonal and long-term fluctuations in ground-water levels is critical to adequate assessment of slope stability. Consideration of seasonal fluctuations is necessary particularly for studies performed during drier months of the year that might coincide with seasonal, low ground-water levels. Comparison of the precipitation for the year a site investigation is performed with that of the historical record is needed to estimate long-term fluctuations in ground-water levels. In many cases, the "present" ground-water level is not the proper design level. A complete assessment should include evaluation of the sensitivity of a natural hillside's stability to a ground-water-level rise of the magnitude documented in this study.

Commonly, investigators infer some level of implied stability if a hillside exhibited no movement during the 1980-86 precipitation period. However, as documented above, ground-water levels in some areas reached historical high levels in the late 1990s as a result of a cumulative increase throughout the recent "wet cycle." Thus, the importance of evaluation of the available long-term ground-water-level data in the vicinity of a hillside cannot be overstated.

Hillside Modifications - Reducing the Threshold for Landsliding?

Excluding the Spanish Fork Canyon landslides, the majority of the 1998 landslides occurred on hillsides modified by human activities. Such modifications included grading, cuts and fills, drainage modification, and residential landscape irrigation. The absence of a significant number of landslides on natural slopes in the Wasatch Front area suggests these hillside modifications reduced the instability threshold levels in the affected slopes. Table 11 summarizes hillside modifications at the significant Wasatch Front land-

Table 11. *Summary of hillside modifications.*

	Slope Modification or Human-Induced Factor				
Landslide	**Crest Fill**	**Cut Slope**	**Landscape Irrigation**	**Redirected Drainage/Runoff**	**Other**
Sunset Drive	x	—	x	x	Enhanced erosion of left flank from irrigation-induced transformation of drainage to perennial creek
E. Capitol Blvd. - City Creek	x	—	x	x	
South Fork Kays Creek	L[1]	—	x	—	
Cedar Bench Subdivision	—	x	—	—	
Davis-Weber Canal	—	x	—	—	
Green Hill Country Estates	—	x	—	—	
Springhill	L	L	x	x	1. Regraded slope 2. Unloading due to gravel extraction
Sherwood Hills - Windsor Dr.	—	x	x	x	
Sherwood Hills - Upper	L	—	L	—	Regrading interrupted natural drainages

[1]Presence of fill at South Fork Kays Creek landslide uncertain.
Abbreviation: L = local cuts, fills, or irrigation.

slides. At five out of the seven 1998 landslides in or abutting residential subdivisions, landscape irrigation may have caused a rise in ground-water levels (figure 40). Site grading, including cuts or fills, also occurred at each landslide. In most cases, a combination of hillside modifications occurred prior to landsliding. The effects of these modifications on stability were generally not considered in the pre-development slope-stability assessments at these sites.

Reactivation of the Spanish Fork Canyon landslides was also partly influenced by hillside modifications; however, significant uncertainty exists regarding the importance of the modifications relative to landslide activity in 1998. At Thistle, the lower part of the landslide was significantly regraded in 1983. Excavations left cut slopes in the lower part and likely destabilized the shallowest deposits in the upper two-thirds of the landslide mass. The localized reactivation of the landslide in 1997 occurred entirely upslope of the 1983 cut slopes and appeared to be relatively shallow in nature. Whether this shallow, partly human-induced movement had any influence on the 1998 movement is unclear.

At the Shurtz Lake landslide, the most obvious modification was the very localized leveling of the hillside for a power-line access road. This road may have locally enhanced infiltration of precipitation, but the fractional area of where the access road crossed the 1997 landslide is nominal with respect to the total area of the landslide mass. The initial 1997 toe was directly downslope of this area and thus enhanced infiltration may have contributed to the toe location. By 1998, landslide deformation of the ground surface likely reduced the influence, if any, of the slight slope modification caused by the road. Another modification of the hillside appears to have been collection and redistribution of spring discharge or surface water in the left (northwest) flank drainage. The 1997 landslide debris contained old and clogged steel drain pipe that may have supplied water to a small cattle pond in the lower part of the landslide. The presence of the pipe in the two earth-flow areas implies some control of where the landslide occurred. However, it is equally possible the landslide disrupted the abandoned water line without the latter having any influence on landsliding.

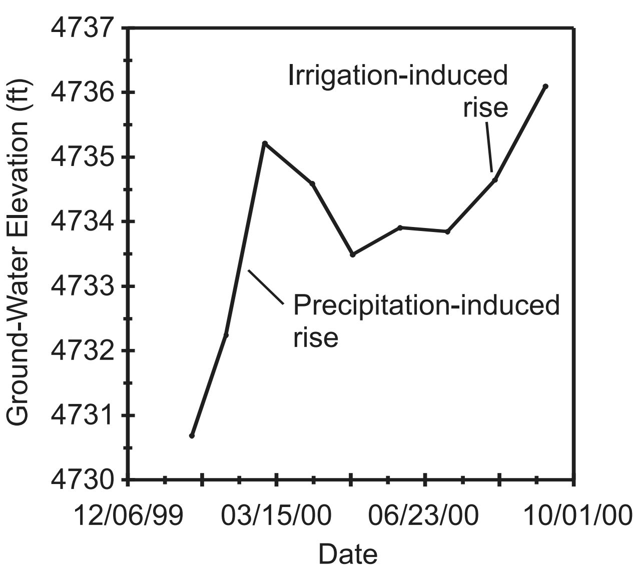


Figure 40. Ground-water-level fluctuation in the shallow ground-water table at the Sunset Drive landslide, Layton in 2000. Rise in ground-water level in early 2000 was associated with infiltration of precipitation during the late winter and early spring, most of which fell as rain. Subsequent rise was induced by landscape irrigation. Drier than normal conditions occurred between March and July, inclusively, in 2000. Ground-water-level data from well 1889 E (see figure 6 for location).

Protecting the Public from Landslides

Despite access to landslide-hazard information by land-use planners, subdivisions continued to be built in pre-existing landslide areas without adequate hazard-reduction measures between 1984 and 1997. Renewed movement of several of these landslides in 1998 caused significant damage to both residential property and utility lifelines. One conclusion following the landslides in 1983 was that much of the damage that occurred in that year could have been avoided if the available geologic and geotechnical data had been used in land-use planning decisions by local governments (Kaliser and Slosson, 1988). Kaliser and Slosson (1988) also concluded that inadequate recognition and mitigation of landslide hazards had occurred prior to 1983. Since 1983, landslides have been inventoried (for example, Lowe, 1988a, 1989) and landslide-hazard maps developed in the four most populated Wasatch Front counties. Thus, land-use planners have had the necessary information to recognize when development was proposed in pre-existing landslide or landslide-hazard areas since the late 1980s. By the mid-1980s, many local governments had adopted site-investigation requirements in hazard areas (Christenson, 1987). However, this process did not always effectively reduce the landslide hazard because of the following:

1. Consistent technical review of geologic-hazard reports was lacking.
2. Report conclusions were vague, often non-conservative, and their implications were overlooked or not recognized by local governments.
3. The standard-of-practice and level of conservatism of consultants performing slope-stability investigations was highly variable.
4. Final development plans were not necessarily reviewed for compliance with recommendations in the site-investigation reports.
5. Consulting engineers and geologists were not required to verify that site work (hillside modification) was performed or constructed in accordance with recommendations.
6. In no instance, was the option to avoid the hazard selected.

Pre-development site investigations were performed at several of the 1998 Wasatch Front landslide sites, but these studies did not prevent damage from landslide movement. Table 12 summarizes the conclusions and recommendations of pre-development site investigations at the 1998 Wasatch Front landslides.

The Cost Benefits of Landslide Avoidance

The potential for renewed movement of many pre-existing landslides that have been modified by residential development exists along the Wasatch Front. Precipitation periods that can trigger landslide movement have occurred twice in

Table 12. Summary of conclusions and recommendations made in pre-development site-investigation reports.

Landslide	Report(s)	Stability Conclusions	Hazard-Reduction Recommendations	Notes on Implementation	1998 Damage
Sunset Drive	Chen & Associates[1,2] (1987)	Crest-side lots were high risk	1. Disclosure of hazard 2. Maximum crest fill thickness	Fill thickness exceeded; disclosure not enforced?	Severe to moderate property damage
E. Capitol Blvd. - City Creek	Dames & Moore[2] (1979, 1981)	Recent activity recognized	1. Building setback 2. Drainage/runoff directed to storm drains	Recommendations not implemented	Severe lot damage requiring expensive repair and stabilization
South Fork Kays Creek	Maughan (1992)	Landslide inferred stable	None	Not applicable	Severe to moderate damage to landscaping
Cedar Bench	Huntingdon Chen-Northern, Inc., (1993)	Landslide inferred stable	None	Not applicable	None
Green Hill Country Estates	AGEC (1996)	Marginal stability inferred	1. Flat (4H:1V) cut slopes 2. Surface drainage	Surface drainage recommendations not implemented	Cut-slope failure
Sherwood Hills - Upper	Rollins and Rollins (1992)	Active movement detected	1. Drains 2. Drilled shaft walls (caissons)	Caisson depth or number inadequate; poor geologic constraints	Damaging movement despite stabilization measures

[1]*Study performed for abutting phase of subdivision.*
[2]*Study predates landslide inventory maps.*

the 20 years between 1980 and 1999 and eight times since 1875 in the Salt Lake City area. Until the 1995-98 precipitation period, little development had occurred in most of the pre-existing landslide areas in the Wasatch Front. As development in these areas continues in response to increasing population and economic growth in the Wasatch Front, the risk associated with renewed movement of pre-existing landslides increases. Land-use planners have the option to reduce the risk to lives and property by eliminating or restricting future development in pre-existing landslides. Whereas an avoidance approach imposes restrictions on land use on private property owners, potentially lowering property values, it could allow for some other desired uses such as open space, agriculture, and possibly low-density development (Jochim and others, 1988).

Economic justification for avoidance of pre-existing landslides can be demonstrated by the costs associated with several of the 1998 Wasatch Front landslides. The present cost of the 1998 Sunset Drive landslide ($456,000) likely exceeds the initial lot price or the profit derived from construction and sale of the seven affected houses and does not include future stabilization costs necessary to prevent additional losses from repeated movement. The current cost of the 1998 Springhill landslide likely exceeds $300,000, or about the initial construction cost of four to five of the affected houses. In addition, subsurface conditions appear unfavorable for cost-effective stabilization (Terracon, 1998) and proposed stabilization measures offer no guarantee that future damaging movement will not occur. The cost of recurrent landslide movement in the Sherwood Hills landslide complex in Provo is undocumented, but likely exceeds several million dollars. In these three landslide areas, the financial responsibility for the costs of landslide movement has been assumed by the current homeowners and local governments rather than the original subdivision developer. Thus, avoidance benefits the parties most impacted by landslide movement while restricting the initial landowner and/or developer who has commonly been unaffected by losses from landsliding.

CONCLUSIONS

An increase in landslide activity in 1998 occurred during the latter part of the 1995-98 precipitation period, a period of successive years with above-normal precipitation. Landsliding consisted mostly of reactivation of pre-existing landslides in modified hillside areas. An inferred cumulative rise in ground-water levels was the likely cause of the 1998 landsliding. However, movement was triggered by a natural, but transient, rise in ground-water levels which coincided with the spring snowmelt and contemporaneous above-normal precipitation.

Monitoring data from several of the 1998 Wasatch Front landslides defined the rate, timing, and duration of movement and demonstrated the relation between movement and precipitation. The rate of movement ranged from very slow to moderate; however, the rate ranged from very slow to slow for landslides in the Wasatch Front urban corridor west of the

Wasatch Range. The majority of damaging landslide movement occurred between March and May. Monitoring data from 1998 and 1999 suggest that movement generally suspends at most landslides by mid-June. However, movement occurred at a very slow rate during the latter part of 1998 at several landslides. At two continuously monitored landslides, the majority of movement occurred over a period of about 30 to 40 days and nearly all movement occurred within 45 to 90 days. Measured or estimated total displacements in 1998 ranged from 0.3 inches (0.8 cm) at the Springhill landslide to 159 feet (48 m) at the Thistle landslide. Displacements were greatest at the two large landslides in Spanish Fork Canyon. Small displacements caused severe damage where buildings straddled landslide deformation features.

The enhanced susceptibility of pre-existing landslides to renewed movement during a precipitation period was demonstrated by the prevalence of reactivation versus landsliding in previously unfailed hillsides in 1998. Renewed movement of pre-existing landslides in 1998 was preceded by similar reactivation of landslides during the previous two precipitation periods (1967-71 and 1980-86). Exceedance of the thresholds for landsliding in unfailed hillsides is documented only during the 1980-86 precipitation period, but has likely occurred locally during other precipitation periods including the most recent 1995-98 period.

The 1998 Wasatch Front landslides demonstrated the necessity to reevaluate the stability of pre-existing landslides. In several cases, pre-development slope-stability evaluations overestimated the stability of the landslides under static conditions. Static slope stability was most likely marginal to metastable prior to the onset of movement in 1998. A new classification is proposed that attempts to characterize the state of activity and the potential for reactivation of the Wasatch Front landslides, recognizing that movement may occur in small increments rather than in significant amounts during a single event. Thorough investigation for possibly subtle or obscure evidence of historical or prehistoric reactivation, which in some cases necessitates subsurface explorations, will be required to reduce uncertainties when using this classification.

Identification of possible instability threshold levels may provide a means for understanding conditions that trigger landslide movement in specific landslides. Only at the Sunset Drive landslide in Layton was ground-water-level data sufficient to estimate the instability threshold level at which movement may have triggered in 1998. Unfortunately, the interval between inclinometer measurements was insufficient to define the initiation and duration of movement, demonstrating the need for more frequent inclinometer monitoring to identify instability threshold ground-water levels. As shown in this study, actual ground-water-level data are invaluable for documenting the relation between precipitation and ground-water-level fluctuations and the relation between the latter and instability. The need to install piezometers prior to movement and the associated costs and short-lived nature of instrumentation in active landslides is a significant limitation to this approach.

Instability threshold precipitation levels provide an alternative, although indirect, means to define conditions that trigger landslide movement. This study demonstrates an approach in which cumulative precipitation can be tracked during an informal landslide water year (LWY), which begins September 1. The LWY approach tracks precipitation antecedent to the spring snowmelt and which follows the dry summer months. Limits on instability threshold precipitation levels now exist at several of the 1998 Wasatch Front landslides where movement suspended in 1999 or 2000. Preliminary results suggest that threshold levels decrease with an increase in the antecedent excess precipitation that occurs during a precipitation period. At the CBCC landslide, the threshold level fell below the normal precipitation level during the 1998-99 LWY following about 9 inches (23 cm) of antecedent excess precipitation. At the Shurtz Lake landslide, the threshold level remained above normal as indicated by the lack of significant movement during 1998-99 LWY despite above-normal precipitation. The influence of landslide geometry, boundary conditions, and other factors on threshold levels and landslide stability require further study.

Historical changes in annual precipitation may have caused ground-water-level fluctuations in many Wasatch Front landslides that affected stability. The last two decades of the historical precipitation record (specifically, the 1980s and 1990s) have been significantly wetter than normal. Excluding the central Wasatch Front (Salt Lake City), the last two decades have been the wettest on record. In the central Wasatch Front, the historical precipitation record reveals increasing variability in annual precipitation since 1967. Fleming and Schuster (1985) recognized these anomalous conditions and named the period between 1967 and 1985 as the "wet cycle." The recent precipitation record suggests that the "wet cycle" extended to at least 1998. During the "wet cycle," ground-water levels have risen in some wells in the central Wasatch Front, reaching historical high levels in the late 1990s coincident with the onset of movement at two landslides near Salt Lake City.

Hillside modifications reduced the threshold for landsliding at the majority of 1998 landslides. However, the onset of landsliding in 1998 coincided with the snowmelt between March and May, suggesting that these landslides were mostly triggered by a transient ground-water-level rise. At several landslides, hillside modifications that cause a rise in ground-water levels, particularly landscape irrigation, significantly reduced hillside stability. The artificial ground-water-level rise is difficult to quantify precisely at most of these landslides; however, seasonal effects of landscape irrigation have been observed in recent ground-water-level data.

Pre-development slope-stability evaluations at sites of the 1998 landslides did not prevent damaging movement or adequately reduce the risk from such movement. In no instance did a consulting geologist or geotechnical engineer recommend not building or complete avoidance of the potential landslide hazard. For various reasons, strict adherence to the consultant's conclusions and risk-reduction recommendations did not occur. Damaging landsliding in 1998 occurred, in some cases, more than a decade after development. The lag time between hillside modifications and site development and damaging landslide movement implies that changes in conditions, particularly those affecting ground-water levels, occurred that were not predicted and/or evaluated by the consultant in the pre-development site investigations.

Losses resulting from damaging landslide movement in 1998 were incurred primarily by property owners and local governments. In most cases, developers of the residential subdivisions affected by damaging landslide movement assumed no financial responsibility for the losses in 1998, possibly because of the amount of time between movement and the completion of development. Thus, developers received the financial benefits of development in landslide hazard areas with no consequences while the risks and the resulting losses were effectively transferred to lot owners and local governments.

In residential subdivisions, damaging landslide movement in 1998 generally affected multiple lots. Thus, the landslide hazard could not be effectively reduced on a lot-specific scale. Regardless, in two cases, lot-specific stabilization was performed to reduce the risk to a house in the crown of a 1998 landslide. Despite post-failure slope-stability investigations which included possible stabilization recommendations, the variability of landslide damage at each of the landslides made it difficult for local governments to convince the majority of property owners to financially contribute to the proposed landslide stabilization efforts.

The landslides of 1998 indicated that an effective local government landslide-hazard reduction program must include:

(1) adoption of an ordinance requiring site-specific investigations in hazard areas,
(2) consideration of the level of risk acceptable to local governments and homeowners,
(3) objective third-party review of geologic-hazard site-investigation reports, and
(4) review of final site-design plans for compliance with report and review recommendations.

In addition, consultants performing site-investigation studies in landslide areas must:

(1) use prudently conservative assumptions, particularly regarding effects of hillside modifications on instability thresholds, and
(2) provide written verification that site conditions did not vary from assumptions and that design recommendations were implemented.

RECOMMENDATION

A possibly more effective approach to hillside development in landslide hazard areas is the establishment of geologic hazard abatement districts (GHADs) similar to those used in California. The GHADs provide a financial mechanism to implement mitigation or, as was needed in 1998, stabilization. Funding for these efforts would be obtained by collecting annual assessments from the property owners in the district. GHADs would have provided a means of financing the proposed stabilization measures at the Sunset Drive and Springhill landslides. Currently, a GHAD-like special service district is under consideration by Provo City for the Sherwood Hills subdivision which continues, as of the date of publication, to be affected by local landslide movement.

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