GEOLOGY AND GEOLOGIC HAZARDS OF TOOELE VALLEY AND THE WEST DESERT HAZARDOUS INDUSTRY AREA, TOOELE COUNTY, UTAH

by

Bill D. Black, Barry J. Solomon, and Kimm M. Harty

SPECIAL STUDY 96
UTAH GEOLOGICAL SURVEY
a division of
Utah Department of Natural Resources
1999

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Utah Geological Survey

ABSTRACT

Geologic hazards are naturally occurring geologic processes that present a risk to life and property. Tooele County is subject to many geologic hazards that need to be identified and considered prior to development. This report provides hazards information for two areas in Tooele County: (1) Tooele Valley, the most populous part of the county and a rapidly developing area; and (2) the West Desert Hazardous Industry Area (WDHIA), an administrative unit established for waste treatment and storage.

Sediments deposited by Pleistocene Lake Bonneville dominate the surficial geology of Tooele Valley and the WDHIA. Lake Bonneville was a large ice-age lake that covered much of northwestern Utah between about 25,000 and 15,000 years ago. Unconsolidated material in the study areas includes gravel, sand, and fine-grained sediment deposited during various stages of the Bonneville lake cycle, alluvium deposited prior to and during lake advance, and alluvium deposited after lake retreat. Bedrock geology consists mainly of Paleozoic sedimentary rock and Tertiary volcanic rock.

Earthquake hazards in Tooele Valley and the WDHIA include ground shaking, surface-fault rupture, tectonic subsidence, and liquefaction. Ground shaking is the most widespread and damaging earthquake hazard, and both study areas may experience ground shaking from a nearby earthquake. A large-magnitude earthquake generated by slip on a fault in Tooele Valley, such as the Oquirrh fault zone, could also produce surface rupture, tectonic subsidence, and liquefaction along the fault. Liquefaction susceptibility is highest in low-lying areas of northern Tooele Valley and the WDHIA where soil and ground-water conditions conducive to liquefaction are found. Lesser earthquake-related hazards in the study areas include earthquake-induced ground failure, subsidence, and flooding, though the extent of these hazards is uncertain. Earthquakes may also induce landslides and rock falls.

Other geologic hazards in Tooele Valley and the WDHIA include landslides; alluvial-fan flooding; lake flooding, ponding, and sheet flooding; shallow ground water; rock fall; problem soils; and radon. Landslide hazards are mainly in Tooele Valley, and the hazard is highest in the southeastern part of the study area. Alluvial-fan flooding is a hazard in both study areas, but the potential for flooding and debris deposition is highest at canyon mouths in eastern Tooele Valley. The hazard from lake flooding is highest along the shore of Great Salt Lake in northern Tooele Valley, and ponding and sheet flooding may occur in low-lying areas in both study areas. Shallow ground water is a hazard in low-lying areas in northern Tooele Valley and throughout most of the WDHIA. Potential hazards from rock fall, problem soils, and radon are also found in both study areas.

The geologic-hazards maps show where hazards may exist, to inform citizens of potential risk and enable local government officials to make responsible planning decisions. The maps are general, and site-specific studies are needed to demonstrate site suitability. Commonly used methods for hazard reduction include avoidance or engineering solutions. Acceptance of a hazard is also an option where it poses a low level of risk, although this method does not reduce the hazard.

INTRODUCTION

Purpose and Scope

Geologic studies have been conducted in Tooele County for more than a century. In the first study, an 1854 expedition across the Great Basin of the western U.S., Beckwith (1855) was impressed by the ancient shorelines of "Tuilla Valley" which "will perhaps afford....the means of determining the character of the sea by which they were formed...." Later, the great American geomorphologist G.K. Gilbert (1890) recognized that the landscape of the region had been shaped to a great extent by a large lake, rather than a "sea," and said of the Great Salt Lake Desert, "The area formerly covered by the main body of Lake Bonneville is now a plain, conspicuous for its flatness." He described the "'lost mountains of Great Salt Lake Desert" as "circled by rocky and inhospitable coasts" during the Lake Bonneville highstand, but the "Cedar Range....bleak and barren as it now is, we may picture as then mantled with verdure" (Gilbert, 1890).

Today, geologic studies determine more than just the nature of ancient processes which formed the landscape. The study of geology provides information to evaluate geologic hazards that must be considered for safe and responsible development. To aid such development, the Utah Geological Survey (UGS) has undertaken a program of geologic-hazards mapping throughout the state. The purpose of these studies is to provide hazards information to planners, local government officials, and concerned citizens. We assessed geologic hazards in two areas in Tooele County (figures 1 and 2): (1) Tooele Valley in eastern Tooele County, and (2) the West Desert Hazardous Industry Area (WDHIA) in north-central Tooele County. Tooele Valley contains most of the county's population, and is on the western margin of expanding metropolitan Wasatch Front communities. The WDHIA is an administrative unit established in 1987 by Tooele County to coordinate the development of hazardous-waste treatment, storage, and disposal facilities.

This report is a compilation of UGS Open-File Reports 296 and 318 (Solomon, 1993; Solomon and Black, 1995), which address geology and geologic hazards of Tooele Valley and the WDHIA. These open-file reports provide detailed information on geology and geologic hazards of Tooele Valley and the WDHIA for land-use planning, and resulted from a multi-year team effort involving surficial-geologic mapping by B.J. Solomon, mapping and compilation of geologic hazards and synthesis of hazards maps by B.D. Black and K.M. Harty, and discussions with local government officials. K.M. Harty mapped landslide, debris-slide, debris-flood, and rock-fall hazards in Tooele Valley; B.D. Black mapped the remaining hazards in Tooele Valley and the hazards in the WDHIA. For this report, we reduced the geology and hazards maps in the open-file reports to a smaller scale (1:100,000) and simplified both the maps and text. This provides geologic information to a wider audience in a format that is easier to use.

Setting

Tooele Valley is in eastern Tooele County (figure 2),

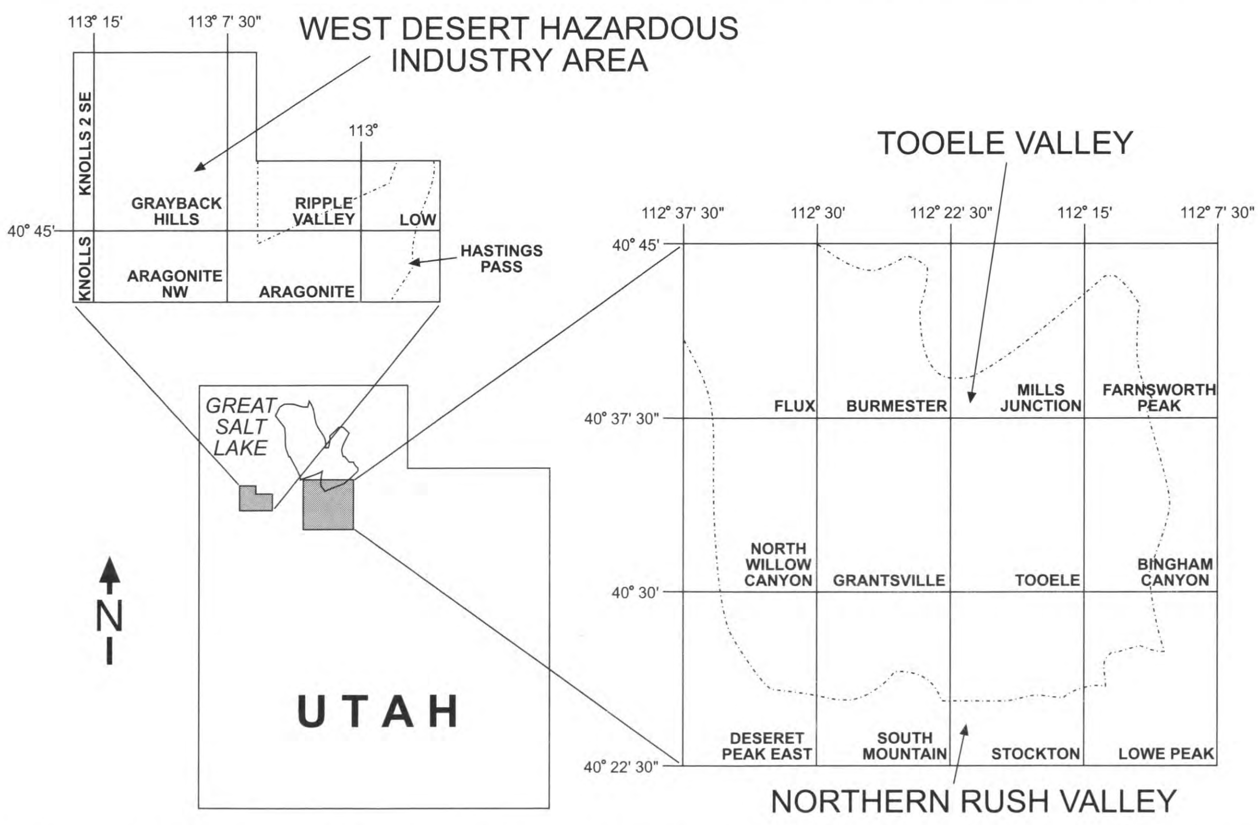


Figure 1. *Index map of Tooele Valley and the West Desert Hazardous Industry Area, and names of U.S.G.S. 7.5-minute topographic quadrangles covered. Approximate study-area boundaries are shown by dashed lines.*

a rural county with a 1990 population density of 3.8 persons per square mile (1.5 persons/km^2) and population of 26,601 (U.S. Census Bureau, 1990). The Oquirrh Mountains form the eastern border of Tooele Valley, and the Stansbury Mountains form the western border (figure 2). Great Salt Lake lies to the north of Tooele Valley, which is separated from Rush Valley to the south by South Mountain. Drainage is northward toward Great Salt Lake.

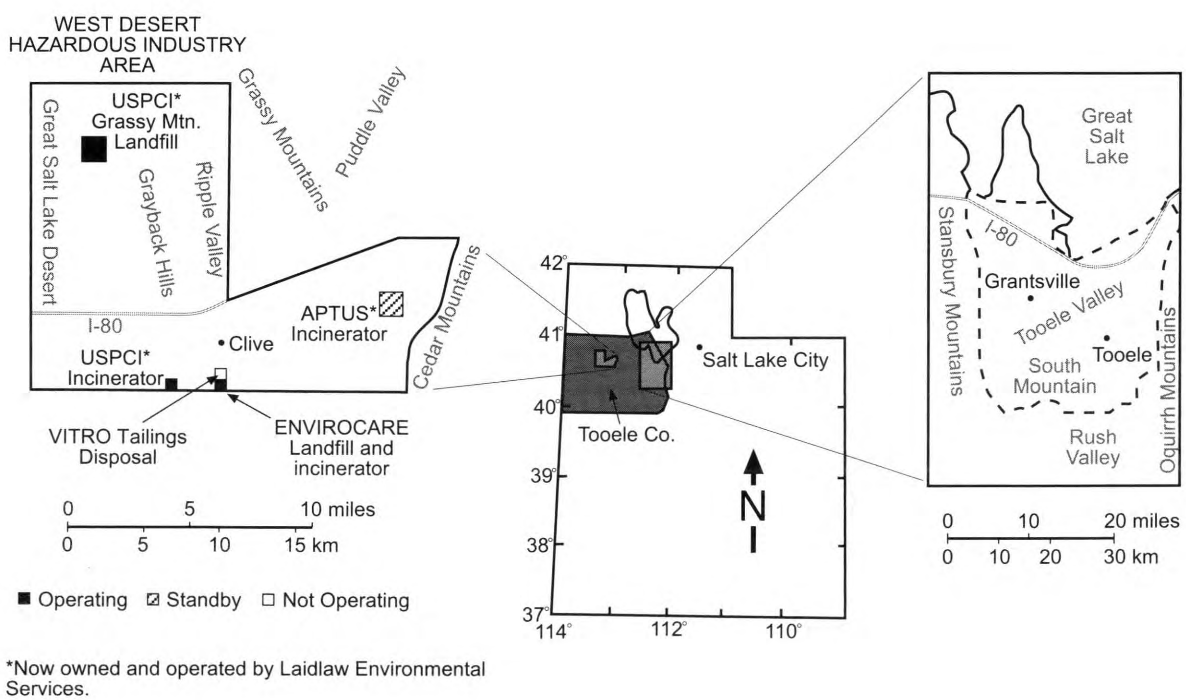
The Tooele Valley study area is bounded by the crest of the Stansbury Mountains to the west, the county line between Tooele and Salt Lake Counties in the Oquirrh Mountains to the east, and the lake shore to the north, and includes the northernmost margin of Rush Valley to the south. The study area has a north-south dimension of about 17 miles (27 km), an east-west dimension of about 22 miles (35 km), and covers about 375 square miles (971 km2). Elevations range from about 4,200 feet (1,280 m) at the Great Salt Lake shore to 11,030 feet (3,360 m) at Deseret Peak in the Stansbury Mountains. The study area includes portions of 12 U.S. Geological Survey 7.5-minute topographic quadrangles (figure 1). Tooele Army Depot, formerly a large army maintenance and storage facility, occupies much of the southern half of Tooele Valley. This facility is currently being privatized as a result of military downsizing in the mid-1990s.

Tooele City, in the southeastern corner of Tooele Valley, is 30 miles (50 km) southwest of Salt Lake City. Tooele City is the county seat and largest community in the county, having a population of 14,797 in 1994 and more than 50 percent of the county total. Grantsville, in northwestern Tooele Valley, is the second-largest community having an estimated population of 4,993 in 1994 (U.S. Census Bureau, 1994).

Tooele Valley has a semi-arid climate with wide seasonal and diurnal temperature variability typical of mid-latitude continental regions (National Oceanic and Atmospheric Administration, 1990). Tooele City has an approximate mean annual temperature of 50.7°F (10.4°C); mean monthly temperatures are lowest in January (28.8°F [-1.8°C]) and highest in July (75.4°F [24.1°C]). Annual precipitation is 16.5 inches (42.0 cm).

The WDHIA, located in north-central Tooele County (figure 2), is essentially unpopulated. The Great Salt Lake Desert bounds the WDHIA to the north, west, and south. The Grassy Mountains and Puddle Valley lie to the northeast, and the Cedar Mountains to the southeast (figure 2). Ripple Valley is in the center of the WDHIA, and is separated from the Great Salt Lake Desert by the Grayback Hills. Drainage of the WDHIA is westward into the Great Salt Lake Desert.

A zoning district established by the Tooele County Commissioners Board as "Hazardous Industrial District MG-H" defines the perimeter of the WDHIA. The district is about 20 miles (32 km) long, has a maximum width of about 15 miles (24 km), and covers about 140



*Now owned and operated by Laidlaw Environmental Services.

Figure 2. Location of Tooele Valley and the West Desert Hazardous Industry Area, Tooele County, Utah. The boundary of the Tooele Valley study area is shown by a dashed line.

square miles (363 km2). Elevations range from 4,225 feet (1,288 m) in the western mudflats to 5,000 feet (1,524 m) in the foothills of the Cedar Mountains. The WDHIA includes portions of eight U.S. Geological Survey 7.5-minute topographic quadrangles (figure 1).

The WDHIA is 65 miles (105 km) west of Salt Lake City. Four hazardous-waste facilities are operating or on standby in the area (Bill Sinclair, Utah Division of Solid and Hazardous Waste [UDSHW], written communication, 1997; figure 2). U.S. Pollution Control, Inc. (USPCI) established the first facility in 1981 when it opened the Grassy Mountain hazardous-waste landfill and an associated incinerator to the south. The USPCI landfill contains several lined pits for the disposal of hazardous, industrial, and PCB wastes (Bill Sinclair, UDSHW, written communication, 1997). Some PCB wastes are also chemically treated and the oils recycled. In 1984, the Utah Department of Health opened a facility for the disposal of low-level radioactive mill tailings and associated contaminated residues and soil removed from the Vitro uranium mill in South Salt Lake City; however, this facility is no longer in operation. The Vitro project encouraged Envirocare of Utah to open in 1988. Envirocare has incinerators and landfills for disposal of naturally occurring radioactive material wastes, low-level radioactive waste, and uranium and thorium mill tailings. Envirocare also treats and disposes of mixed (radioactive and hazardous) waste. Aptus is a similar facility constructed originally by USPCI. Aptus began operation in 1990 and is now on standby status. The USPCI incinerator and landfill, and the Aptus incinerator, are now owned and operated by Laidlaw Environmental Services (Bill Sinclair, UDSHW, written communication, 1997). The USPCI and Aptus incinerators are designed to thermally destruct both "hazardous" chemical waste materials, as defined under the Resource Conservation and Recovery Act, and "toxic" chemical waste materials, as defined under the Toxic Substance Control Act.

The WDHIA has an arid climate, unlike Tooele Valley, but both areas have in common wide seasonal and diurnal temperature variability. The WDHIA has an approximate mean annual temperature of 46.6°F (8.1°C); mean monthly temperatures are lowest in January (19.2°F [-7.1°C]) and highest in July (79.0°F [26.1°C]). Annual precipitation is 6.6 inches (16.8 cm).

Previous Work

A few geologic maps broadly portray Quaternary deposits in Tooele and northern Rush Valleys. Rigby (1958) categorized the Stansbury Mountains piedmont as either lacustrine, alluvial, or eolian deposits, and described pre-Lake Bonneville geomorphic features. Everitt and Kaliser (1980) supplemented existing geologic maps with aerial-photograph interpretations, and mapped the valleys at a scale of 1:50,000. Several geologic quadrangle maps (Tooker, 1980; Tooker and Roberts, 1971a, 1971c, 1988a, 1988b, 1992) show Quaternary deposits and faults in the Oquirrh Mountains piedmont. These maps were consulted and revised in mapping by Solomon (1993).

Previous research has emphasized various geological characteristics of the area. Smith and others (1968) relate Lake Bonneville stratigraphy to Pleistocene fish fossils from several locations within the Bonneville Basin, one of which is at Black Rock Canyon at the northern end of the Oquirrh Mountains. McCoy (1987) sampled gastropod fossils from the same canyon as part of an investigation of Quaternary aminostratigraphy. The Quaternary stratigraphic record was studied in core samples collected near Burmester, in northern Tooele Valley, by Eardley and others (1973). Regional gravity surveys that cover Tooele and northern Rush Valleys were conducted by Johnson and Cook (1957), Johnson (1958), Cook and Berg (1961), Tanis (1963), Zimbeck (1965), and Cook and others (1975, 1989).

Other investigators (Bucknam, 1977; Everitt and Kaliser, 1980; Barnhard, 1988; Barnhard and Dodge, 1988) mapped fault scarps on unconsolidated sediments within the valleys, and scarps of erosional or undetermined origin were also mapped by Everitt and Kaliser (1980). Krinitsky (1989) evaluated earthquake hazards for the Tooele Army Depot by estimating earthquake ground motions. Olig and others (1996) studied the timing and magnitude of earthquakes associated with faults along the western flank of the northern Oquirrh Mountains adjacent to Tooele Valley, and Wu and Bruhn (1990) studied the geometry and kinematics of similar faults along the western flank of the southern Oquirrh Mountains in Rush Valley. Gilluly (1928, 1932) also studied these faults in the vicinity of Stockton, and postulated a process for the integration of the drainage of Rush Valley with Tooele Valley prior to the highstand of Lake Bonneville (Gilluly, 1929).

Burr and Currey (1988, 1992) and Burr (1989) investigated shore features formed at the pass between Rush and Tooele Valleys near Stockton during the last deep-lake cycle of Lake Bonneville. The UGS conducted investigations of proposed construction sites in Tooele Valley (Kaliser, 1971; Lund, 1985b, 1986; Case, 1987b; Solomon, 1992), damage from flooding in canyons in the Oquirrh Mountains (Lund, 1985a), and rock-fall hazards in surrounding mountains (Case, 1987c). The U.S. Department of Agriculture Soil Conservation Service (now Natural Resources Conservation Service) mapped soils in the area (unpublished data, 1989) and made engineering interpretations regarding erosion hazard, permeability, and land use. The Utah State Department of Highways (1963) generated data on Quaternary geology and engineering properties of soils for highway construction. Solomon and Black (1995) summarized the Quaternary geology and geologic hazards of Tooele and northern Rush Valleys.

Several ground-water studies provide information on Tooele and Rush Valleys. Carpenter (1913) included the valleys in his comprehensive northwestern Utah groundwater investigation. He located streams, springs, and wells, and attributed recharge of the valley floor to precipitation on mountain slopes and infiltration in the piedmont zone. Thomas (1946) differentiated eolian and lacustrine deposits on the floor of Tooele Valley, and recognized the presence of fault blocks in the basin from geologic and hydrologic data.

Early investigators of the Great Basin (Stansbury, 1852; Gilbert, 1890) provided regional observations of the area now referred to as the WDHIA, but detailed geologic studies were not undertaken for many years. Academic research included studies of the northwest portion of the WDHIA (Doelling, 1964), the Grayback Hills in the central part of the area (Davies, 1980), and the foothills of the Cedar Mountains to the east (Maurer, 1970), but Quaternary geology was dealt with only peripherally. Doelling and others (1994) remapped the geology of the Grayback Hills and vicinity; volcanic rocks exposed there are discussed by Hogg (1972). Jones (1953), Eardley (1962), and Dean (1978) described eolian deposits of the region, but only at a reconnaissance scale. Halverson (1961), Stepp (1961), Tanis (1963), Cook and others (1975, 1989), and Baer and Benson (1987) conducted regional gravity surveys covering the WDHIA. Jensen (1958) conducted a regional gravity survey covering Tooele Valley.

Detailed geotechnical investigations have been performed by consultants and reviewed by the UGS for the Envirocare low-level radioactive waste disposal site and other facilities in the WDHIA. Limited data on seismicity and the engineering properties of soils were generated for highway construction (Utah State University Engineering Experiment Station, 1962; Utah State Department of Highways, 1963) and for a proposed Superconducting Super Collider site (Dames & Moore and others, 1987a, 1987b; Arabasz and others, 1989), as well as for the siting of hazardous-waste facilities previously noted. The U.S. Department of Agriculture Soil Conservation Service has mapped soils in the area (unpublished data, 1989) and has made engineering interpretations regarding erosion hazard, permeability, and land use. Quaternary geology and geologic hazards of the WDHIA are summarized in Solomon and Black (1990, 1995). Carpenter (1913), Stephens (1974), and Gates and Kruer (1981) summarized the hydrology of the region.

GEOLOGIC SETTING

Tooele and Rush Valleys occupy structural basins in the Basin and Range physiographic province (Hunt, 1967); the WDHIA is on the eastern edge of a basin. Gravity anomalies (Johnson, 1958; Baer and Benson, 1987) indicate that the floors of these basins are complex collections of troughs and ridges rather than single downfaulted grabens. The deepest portion of Tooele Valley is on its north-central margin. The Walker-Wilson No. 1 oil test well penetrated 7,100 feet (2,100 m) of basin fill in this area (Heylmun, 1965), and Everitt and Kaliser (1980) estimate in excess of 8,000 feet (2,400 m) of fill as the basin thickens northward under Great Salt Lake. The deepest portion of the basin occupied by the WDHIA is on the western margin of the study area and in Ripple Valley, where up to 3,000 feet (915 m) of basin fill exists (Baer and Benson, 1987). Fill in these basins is dominated by sediments deposited during the late Pleistocene Bonneville lake cycle.

Tooele and Rush Valleys are geomorphic subbasins of the Bonneville Basin, as a consequence of their integration with Lake Bonneville for part of the Bonneville lake cycle (Gilbert, 1890; Eardley and others, 1957; Currey and others, 1984b; Currey and Oviatt, 1985). The Bonneville lake cycle was essentially coincident with the last global ice age of marine isotope stage 2, and lasted from about 28,000 to 12,000 radiocarbon years B.P. (Currey, 1990; Oviatt and others, 1992; Dr. Donald R. Currey, University of Utah, written communication, 1995). Lake Bonneville began to rise from levels close to those of Holocene Great Salt Lake about 28,000 radiocarbon years B.P. (Oviatt and others, 1992; Dr. Donald R. Currey, University of Utah, written communication, 1995), and transgression was well underway about 26,000 radiocarbon years B.P. (Currey and Oviatt, 1985). The Pilot Valley shoreline in northwest Utah indicates a lacustrine oscillation during the early transgressive phase of Lake Bonneville (Miller and others, 1990), but this shoreline has not been identified in the study areas. The lake experienced a major, climatically induced oscillation between 21,000 and 20,000 radiocarbon years B.P. that resulted in the formation of the Stansbury shoreline (Oviatt and others, 1990; Don Currey, written communication, 1995), the oldest shoreline recognized in the study areas. Shoreline deposits at the south end of Tooele Valley (the unnamed shoreline complex of Burr and Currey, 1988, 1992), and the equivalent shoreline elsewhere in the valley, mark what seems to be one or more important stillstands or moderate oscillations during the transgressive phase of the Bonneville lake cycle as the lake rose above the Stansbury level. Sack (1990), who notes casual references to similar deposits elsewhere in Utah by previous researchers, applied the term "sub-Provo" to the deposits where they are found in Tule Valley of west-central Utah. This sub-Provo lake level formed between 20,000 and 17,700 radiocarbon years B.P., and was so named because it lies just below the later Provo shoreline.

Lake Bonneville occupied its highest shoreline, which Gilbert (1875) named the Bonneville beach, after 15,500 radiocarbon years B.P., and perhaps as late as 14,500 radiocarbon years B.P. (Currey and Oviatt, 1985; Don Currey, written communication, 1995). This shoreline was established by a basin-hypsometric factor, the stabilization of the lake level at an external overflow threshold. Prior to the lake transgression, the drainage of Rush Valley had been integrated with that of Tooele Valley (Gilluly, 1929). During the highest stage of Lake Bonneville, Rush Valley was an embayment separated from Tooele Valley by a strait at the pass between the two valleys near Stockton. Headward erosion of the Snake River-Bonneville Basin drainage divide caused the catastrophic incision of the Zenda threshold in southern Idaho, which lowered the threshold and lake level 340 feet (105 m) in less than one year (Malde, 1968; Currey and others, 1983; Jarrett and Malde, 1987). After this rapid drawdown of Lake Bonneville, Rush Valley was isolated when Lake Bonneville receded below the Stockton Bar barrier between Rush and Tooele Valleys. Rush Valley then became the site of a succession of independent pluvial lakes that include Lake Shambip, Lake Smelter, and Rush Lake (Burr and Currey, 1988, 1992).

In Tooele Valley, and in the remainder of the Bonneville Basin, Lake Bonneville stabilized at a lower level controlled by the Red Rock Pass threshold. The very prominent Provo shoreline was formed at this level (Gilbert, 1875, 1890). Persistent landsliding in the flood-scoured threshold area formed gravel beach ridges at different lake levels of the Provo shoreline complex (Burr and Currey, 1988, 1992; Burr, 1989). About 14,000 radiocarbon years B.P., climatic factors induced regression from the Provo level (Currey and Oviatt, 1985; Don Currey, written communication, 1995). In less than 2,000 years the lake level was below the elevation of the present Great Salt Lake. Transgression was subsequently renewed and the earliest post- Bonneville oscillation, known as the Gilbert, began about 11,000 radiocarbon years B.P. (Murchison, 1989; Don Currey, written communication, 1995). The lake finally regressed sometime between 9,400 and 9,700 radiocarbon years B.P. A late Holocene lake rise, between 3,440 and 1,400 radiocarbon years B.P., resulted in the highest static lake level reached during the Holocene, and is commonly referred to as the Holocene high (Murchison, 1989).

Shorelines within the WDHIA reflect a lacustrine history similar to that of Tooele Valley. During the highest stage of Lake Bonneville, the WDHIA occupied the littoral zone adjacent to the emergent islands of the Cedar and Grassy Mountains.

GEOLOGIC MAPPING

Geologic mapping provides basic data from which geologic hazards can be identified. Plate 1 shows geology (at a scale of 1:100,000) in Tooele and northern Rush Valleys, and the WDHIA. Solomon (1993) mapped the geology of Tooele and northern Rush Valleys on 1:20,000-scale aerial photographs, and the geology of the WDHIA on 1:24,000-scale aerial photographs. This mapping was field checked and transferred onto 1:24,000-scale base maps. Surficial geology on plate 1 was compiled from Solomon (1993) and transferred to 1:100,000-scale base maps; bedrock geology was compiled from mapping by Tooker and Roberts (1971a-c; 1988a-b), Moore and Sorensen (1979), Armin and Moore (1981), Sorensen (1982), and Doelling and others (1994). Solomon (1993) showed surficial deposits, piedmont fault scarps, six regional shorelines, and three local shorelines. Plate 1 shows general bedrock units, surficial deposits, and piedmont fault scarps, but regional and local shorelines were omitted for clarity. Surficial geologic text was adapted from Solomon (1993).

Description of Map Units

Map units shown on plate 1 are designated with several symbols (table 1). Upper-case letters represent the geologic age of the unit, and units are subdivided into Paleozoic undifferentiated bedrock (Pu), Tertiary undifferentiated bedrock (Tu), unconsolidated or semi-consolidated deposits of both the Quaternary and Tertiary (QT), and Quaternary unconsolidated deposits (Q). A lower-case letter follows the age designation and describes the general environment of deposition: alluvial (a), eolian (e), fill (f), lacustrine (l), mass movement (m), playa (p), or spring (s). Another lower-case letter either describes the subenvironment of deposition, such as dune (d) or

Table 1. *Geologic units and letter designations on plate 1.*

UPPER-CASE LETTER: geologic age designations					
Q	Quaternary			T	Tertiary
QT	Quaternary and Tertiary			P	Paleozoic
FIRST LOWER-CASE LETTER: general depositional environment		**SECOND LOWER-CASE LETTER:** depositional subenvironment or material modifier			
a	alluvial	c	colluvial component	m	mud
		f	fan	t	terrace
		l	low-level floodplain and channel deposits		
e	eolian	g	gypsum	o	oolite
		i	silt	s	silica
f	fill	d	mine dumps	t	tailings
l	lacustrine	a	alluvial component	l	lagoonal
		c	clay	m	marl
		f	fine-grained	s	sand
		g	gravel		
m	mass movement	f	flows	t	talus
		s	slides		
p	playa	m	mud		
s	spring	m	marsh		
u	undivided bedrock				
NUMERICAL SUBSCRIPT: relative age indicator					
1		youngest deposit			
2		oldest deposit			

lagoon (l), or serves as a material modifier, such as gravel (g) or marl (m). These material modifiers also supply information about the subenvironment; in the case of Quaternary lacustrine deposits, for example, gravel (Qlg) is deposited in a relatively high-energy environment and marl (Qlm) in a low-energy environment. Some map units are represented by a numerical subscript which indicates the relative age of similar deposits, such as younger alluvial fans (Qaf$_1$) and older alluvial fans (Qaf$_2$). The approximate correlation of map units is given in figure 3.

Bedrock

Bedrock in the study areas is undifferentiated, except for a generalized division into two age groups, Paleozoic (Pu) and Tertiary (Tu). An asterisk denotes units susceptible to landsliding (Pu*). In the Oquirrh Mountains and South Mountain, Pu is primarily limestone and quartzite of the Pennsylvanian to Permian Oquirrh Formation (Tooker and Roberts, 1971a, 1971c, 1988a, 1988b, 1992; Moore and Sorensen, 1979; Tooker, 1980). Numerous formations crop out in the Stansbury Mountains to the west of Tooele and Rush Valleys, but the thickest is the Cambrian Tintic Quartzite (Moore and Sorensen, 1979). Paleozoic rock susceptible to landsliding is commonly the lower Pennsylvanian to upper Mississippian Manning Canyon Shale. Tu commonly includes trachyandesite lava flows of Eocene age covering the surface of most of the Grayback Hills (Doelling and others, 1994), and andesite, dacite, and quartz latite flows and breccias in the Stansbury Mountains and South Mountain.

Alluvial Deposits

Undifferentiated alluvium (Qa): Deposits mapped as undifferentiated alluvium (Qa) include coarse- to fine-grained alluvium on gentle slopes near piedmont toes.

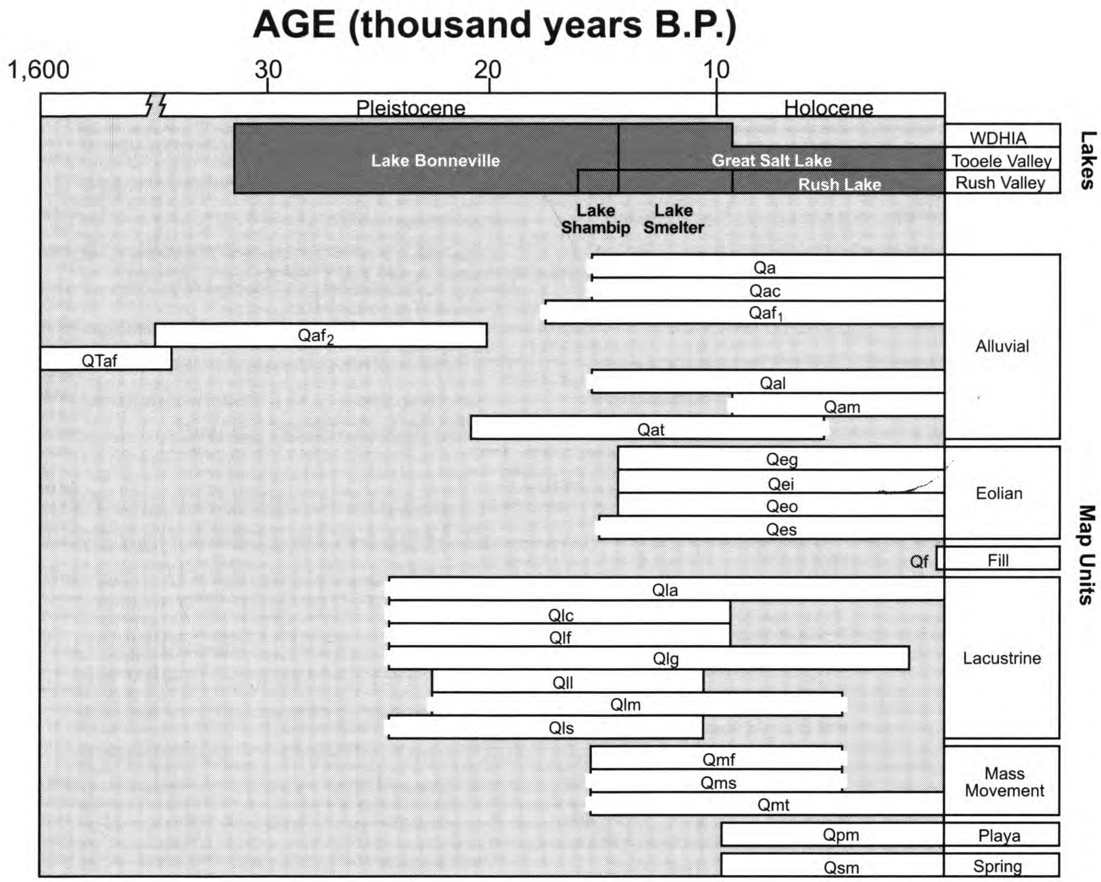


Figure 3. Approximate correlation of Quaternary geologic map units and relation to Bonneville Basin lakes (modified from Solomon, 1993).

Sediment is primarily sandy, with lesser amounts of boulders, gravel, silt, and clay. Deposits are generally less than 10 feet (3 m) thick. Qa is commonly associated with low-order stream channels lacking well-defined floodplains. These deposits occur below the Bonneville level and are of post-Bonneville shoreline age, thus they are as old as latest Pleistocene. The deposits commonly obscure shorelines etched into the adjacent piedmont slope, and may be as young as latest Holocene in age.

Alluvium and colluvium (Qac): Alluvium with a significant colluvial component (Qac) consists of poorly sorted clay, silt, sand, gravel, cobbles, and boulders, and includes alluvially reworked colluvium. Deposits are coarser on steeper slopes, and are generally less than 10 feet (3 m) thick. This unit is present in first-order drainages and wash slopes on mountain fronts at or above the piedmont junction. These deposits are latest Pleistocene to latest Holocene in age.

Younger alluvial-fan deposits (Qaf$_1$): Alluvial-fan deposits that postdate the highstand of Lake Bonneville (Qaf$_1$) include coarse- to fine-grained alluvium and debris-flow sediments deposited on piedmont slopes primarily after regression of the lake from the Bonneville shoreline. Most Qaf$_1$ deposits are found below the Bonneville level, with fan apices at either the Bonneville or Provo shoreline scarps. The small areas of Qaf$_1$ deposits found above the Bonneville level along alluvial-channel slopes may partially consist of sediments deposited before Bonneville shoreline time. The texture of Qaf$_1$ deposits generally becomes finer down the fan slope, and the deposits contain mostly sand and gravel reworked from underlying lake beds. Qaf$_1$ deposits are thickest near fan apices and thin to a feather edge into the basin, but are generally less than 10 feet (3 m) thick. The deposits have been accumulating from latest Pleistocene time through the Holocene to the present.

Older alluvial-fan deposits (Qaf$_2$ and QTaf): Older alluvial-fan deposits that predate the highstand of Lake Bonneville include coarse- to fine-grained alluvium and debris-flow sediments deposited above the Bonneville shoreline on the margins of Tooele and Rush Valleys. These deposits were formed by coalescing alluvial fans which developed bajadas that slope gently away from the mountain front. Qaf$_2$ deposits are older alluvial fans with relatively smooth surfaces truncated by the Bonneville shoreline. Oldest alluvial-fan deposits (QTaf) are typically unconsolidated and poorly stratified, but are sometimes partly consolidated, and are locally carbonate cemented. Extreme dissection is found in QTaf deposits. In mountain-front locations, deposits are generally less than 20 feet (6 m) thick, but they thicken away from the front and probably underlie lacustrine sediments in the valley. Older alluvial-fan deposits do not occur below the Bonneville shoreline and were abandoned when, or before, Lake Bonneville regressed from its highest level.

Considerable confusion exists regarding the age of abandoned alluvial-fan deposits. Thomas (1946) originally included them, where present in Tooele Valley, in the Salt Lake Formation of Pliocene and Pleistocene (?) age. This age was based on the stratigraphic position of the deposits between Miocene (?) tuff and the Pleistocene Bonneville lake beds. Slentz (1955a) redefined deposits in Tooele and Jordan (Salt Lake) Valleys that are younger than the Eocene Wasatch Formation but older than the Pleistocene Lake Bonneville deposits as the Salt Lake Group of Tertiary age, and restricted the youngest alluvial-fan deposits in the group to the Harkers Fanglomerate of Pliocene age. The age of the Harkers Fanglomerate was based on stratigraphic position, geomorphic expression, and lithologic characteristics (Slentz, 1955a, 1955b). Tooker and Roberts (1971b) renamed the unit, where it occurs at its type section in Harkers Canyon on the eastern margin of the Oquirrh Mountains in Jordan Valley, the Harkers Alluvium "because of its great size distribution and unindurated nature." They also assigned an early Pleistocene age to the unit because of its unconsolidated nature and stratigraphic position. This nomenclature and age were extended to similar deposits on the western margin of the Oquirrh Mountains in Tooele Valley (Tooker, 1980; Tooker and Roberts, 1971a, 1988a, 1988b, 1992). However, in western Tooele Valley on the eastern margin of the Stansbury Mountains, Rigby (1958) variously described similar deposits as Quaternary alluvium, Quaternary alluvial fanglomerate and gravel, Quaternary and Tertiary alluvial gravel, and Tertiary Salt Lake Formation, the latter unit consisting of the consolidated portion of the deposits. Rigby (1958) also described the bajadas in the vicinity of North Willow, South Willow, and Box Elder Canyons, but described the surface between Box Elder and East Hickman Canyons as a pediment, attributing its formation to erosion rather than deposition.

No fossil evidence has ever been found in these deposits to provide a definitive age. Because they occupy a stratigraphic position between consolidated sedimentary and volcanic deposits of known Tertiary age and unconsolidated Quaternary lake beds, we assign a Pleistocene age to the younger, less incised portions of these deposits (Qaf_2), whereas older, more incised portions may be as old as late Tertiary.

Stream alluvium (Qal): Alluvium in washover channels is found on mud flats in northern Tooele Valley and the southwestern WDHIA. These alluvial deposits appear on aerial photographs as anastomosing patterns of alternating light and dark stripes interpreted as braided-stream deposits. Thin beds of silt and clay, less than 1 foot (0.3 m) thick, were probably deposited by ephemeral streams in washover channels emanating from locally channelized sheet wash from adjacent slopes during periods of intense rainfall. Alluvium in washover channels is of latest Holocene age.

Other alluvial deposits consist of fine-grained sediment having thin gravel layers and lenses. These deposits are generally less than 10 feet (3 m) thick, and typically are found in channels and associated floodplains on piedmont slopes of mountains within both study areas. These deposits are also present in isolated alluvial channels in adjacent bedrock. Alluvial-channel and floodplain deposits are primarily below the level of the Bonneville shoreline, and are found in channels that dissect the Bonneville abrasion platform. They are predominantly of post-Bonneville shoreline age, but range in age from latest Pleistocene to latest Holocene.

Alluvial-mud deposits (Qam): Alluvial-mud deposits (Qam) consist of fine-grained sediment that has been deposited in low-order stream channels which lack well-defined floodplains. These deposits are generally less than 10 feet (3 m) thick, and occur near Fishing and Sixmile Creeks at the north end of Tooele Valley (plate 1). The deposits obscure Lake Bonneville shorelines etched into the adjacent slope and are Holocene in age.

Alluvial-terrace deposits (Qat): Alluvial-terrace deposits consist of a thin veneer, less than 10 feet (3 m) thick, of coarse- to fine-grained alluvium. The deposits are present at several terrace levels along major drainages in the Stansbury and Oquirrh Mountains, on the margins of Tooele and northern Rush Valleys (plate 1). The terraces are long, narrow, gently inclined surfaces that are elevated above active channels and floodplains, and are bounded along their edges by steep slopes, or risers.

The youngest alluvial-terrace deposits transect the Bonneville shoreline and extend into Tooele Valley near the mouth of Pope Canyon, west of Grantsville (plate 1). Youngest deposits are present below the Bonneville level and are of post-Bonneville shoreline age, but may be as old as latest Pleistocene. The deposits obscure shorelines etched into the adjacent piedmont slope, and may be as young as Holocene in age, but more recent alluvial deposits (Qal) truncate them.

Alluvial-terrace deposits truncated by the Bonneville shoreline occur near the mouths of several canyons on the margins of Tooele and Rush Valleys. Progressively older alluvial-terrace deposits are found elsewhere along the mountain fronts, and are particularly well developed in the vicinity of East Hickman Canyon, where several pre-Bonneville shoreline terrace levels are present (plate 1). Between South Willow and Box Elder Canyons, two pre-Bonneville shoreline terrace levels were mapped by Rigby (1958), but Solomon (1993) shows that three levels are present. Near Middle Canyon, Everitt and Kaliser (1980) mapped three linear scarps "of erosional or undetermined origin," but these are actually risers along the edges of three adjacent terrace levels. All pre-Bonneville shoreline alluvial-terrace deposits predate the regression of the lake from the Bonneville level and are of latest Pleistocene age. The terraces, however, are cut into older alluvial-fan deposits (Qaf_2), and associated alluvial-terrace deposits are therefore younger than Qaf_2.

Eolian Deposits

Gypsum dunes (Qeg): Morphologically well-developed dunes dominated by sand-sized gypsum particles are mapped as Qeg. These gypsum dunes are present only on the western edge of the WDHIA. They form a pair of north-south linear ridges, up to 30 feet (9 m) high, on the eastern margin of the Great Salt Lake Desert (plate 1). The dunes are active in general, but in places are stabilized by vegetation and are capped by shrub-coppice dunes. The gypsiferous material may be derived from the erosion of efflorescent mud-flat salts (Eardley, 1962). A significant oolitic component is also present in parts of Qeg. This fraction, as well as oolite dunes described below, may represent almost in-place reworking of lacustrine beach deposits associated with the final static level of the Gilbert beach cycle of Murchison (1989). A 4,230-foot (1,290-m) lake level formed between 9,400 and 9,700 radiocarbon years B.P., and this is the approximate elevation of the base of Qeg. This phenomenon of eolian sand preferentially deposited along former shorelines has been noted elsewhere in the Bonneville Basin (Dennis, 1944; Ross, 1973; Currey, 1980; Sack, 1990). The results of parametric tests by Dean (1978) also indicate that dunes of the WDHIA are of very local origin and may not have moved far to attain their present locations. The gypsum dunes are largely of Holocene age, but some may have started accumulating in latest Pleistocene time.

Silt dunes (Qei): Parabolic dunes with predominantly silt-sized grains (Qei), and with lesser amounts of clay, fine sand, and sodic material, are present in northwestern Tooele Valley (plate 1). The sand grains are commonly oolitic with a core of silt. The silt dunes occur on the margin of, and overlie, erosional remnants of Lake Bonneville lacustrine sediments; the remnants are commonly surrounded by younger deposits of playa mud. These dunes are generally less than 15 feet (5 m) thick. They were mapped by Everitt and Kaliser (1980), who included the lacustrine-sediment remnants with the eolian material; the remnants are excluded here. The silt dunes appear to have accumulated, in part, from eroded, fine-grained lake beds, and serve as a protective cover for the lake-bed remnants. Many of these dunes, however, coincide with the elevation of the Holocene highstand of Great Salt Lake at approximately 4,221 feet (1,287 m), and may represent eolian reworking and accumulation along the former shoreline; the arcuate pattern of some outcrops forms small embayments and associated spits. Silt dunes are younger than the Bonneville lake beds, but may be as old as latest Pleistocene. Deposition continued during the Holocene.

Oolite dunes (Qeo): Longitudinal and transverse dunes, predominantly composed of sand-sized oolites but with a large component of gypsum, are mapped as Qeo in the southwestern portion of the WDHIA (plate 1). The oolite dunes may be as much as 30 feet (9 m) thick. They differ from adjacent gypsum dunes (Qeg) only in the relative proportion of oolites and gypsum. Both Qeo and Qeg form a continuous outcrop pattern on the edge of Great Salt Lake Desert mud flats, and the proportion of material in the dunes is governed by the relative contribution of material from both Gilbert-level beach deposits (oolites) and efflorescing salts in the mud flats (gypsum). As with the gypsum dunes, the oolite dunes are largely of Holocene age, but some may have started accumulating in latest Pleistocene time.

Silica dunes (Qes): Siliceous sand deposited in dunes (Qes) is found in several areas of both Tooele Valley and the WDHIA. On the Tooele Army Depot, west of the city of Tooele, linear dunes that trend north-south occupy an area of several square miles (plate 1). Most dunes are less than 3 feet (1 m) high, but some are as high as 10 feet (3 m). The material is predominantly medium- to fine-grained quartz sand with considerable silt, and with some secondary calcium carbonate grains. Similar material is present in thinly bedded Lake Bonneville sediments visible in ephemeral stream channels in the vicinity, but in outcrops too small to be mapped. These lake beds likely underlie the dunes and serve as their source material. These dunes were mapped by Thomas (1946), but not by Everitt and Kaliser (1980). Thomas (1946) also mapped similar deposits in the vicinity of Grantsville, but Solomon (1993) found no evidence of eolian material near Grantsville. The dunes near Grantsville were said to have been smaller than those at the Tooele Army Depot, and their development was attributed by Thomas (1946) to the drought years of 1934 and 1935, "when the vegetative cover was destroyed by desiccation over large areas.....the territory around Grantsville became a miniature 'dust bowl' area." Dunes near the army depot, however, are larger and, in part, stabilized by vegetative cover, and are believed to have originated long before the drought years of this century.

Siliceous dunes have also formed west of Silcox Canyon (plate 1), southeast of the city of Tooele, on the eastern edge of a gravel spit associated with the Provo shoreline. These eolian sands are fine- to medium-grained, but with little silt or carbonate, and are stabilized by a vegetative cover. The dunes likely were deposited shortly after the regression of Lake Bonneville from the Provo level. Wind velocity decreased as wind traversed the topographic barrier of the spit, and sand was deposited on the lee side. Sand was also locally derived from the spit.

A small group of siliceous dunes is present on the east-central flank of the Grayback Hills in the WDHIA, but the majority of siliceous dunes in the WDHIA occurs in the eastern portion as small, localized deposits, and in a larger outcrop area which covers several square miles in the northeastern WDHIA and adjacent region (plate 1). The deposits contain longitudinal and parabolic dunes that are, in part, active, and contain silty, fine quartz sand. A peculiar feature of the small, localized deposits to the east is their outcrop pattern. The sand commonly is in linear, narrow deposits up to 1.5 miles (2.4 km) long but only 200 to 400 feet (60 to 120 m) wide that parallel ephemeral stream channels up to 30 feet (9 m) deep. The mechanism that created the dunes is probably similar to that which created the dunes adjacent to the Provo-level spit southeast of Tooele. Changes in wind velocity over negative topographic features such as channels causes sand to accumulate in narrow, linear areas adjacent to the features.

A number of potential sources for sand in siliceous dunes of the WDHIA exist, but the primary source is likely pre-Bonneville alluvial fans. This unit underlies younger deposits on the Cedar Mountains piedmont slope, and was derived from Paleozoic quartzite in the mountains. Older siliceous dunes present at elevations higher than, and truncated by, the Bonneville shoreline are present 1 mile (1.6 km) southeast of the southeastern WDHIA corner (Maurer, 1970). These older dunes were also derived from alluvial fans composed of Paleozoic quartzite from the Cedar Mountains, and are likely a secondary source of quartz sand for younger eolian deposits. The older dunes were reworked below the Bonneville shoreline where they once occurred, and were redeposited as sandy lake beds on the southeastern margin of Ripple Valley; these lake beds also served as a secondary source of sand for the younger siliceous dunes within the WDHIA. The Qes deposits in both Tooele Valley and the WDHIA are largely of Holocene age, but some may have started accumulating in latest Pleistocene time.

Fill Deposits

Undifferentiated fill (Qf): Historical fill deposits are mapped in several locations. Qf in the northeastern corner of Tooele Valley (plate 1) consists of salt evaporated from ponds of water from Great Salt Lake, isolated from the lake by dikes. The Stansbury Park residential development, west of Mills Junction in northeast Tooele Valley, includes fill for foundation soil, a golf course, and a human-engineered lake (plate 1). Earthen flood-control dams are present across Black Rock Canyon at the north end of the Oquirrh Mountains, Settlement Canyon south of the city of Tooele, and Box Elder Wash north of South Mountain (plate 1). In the WDHIA, Lake Bonneville gravel has been used as a landing-strip base west of the Grayback Hills (plate 1). These Qf deposits are probably less than 100 years old.

Mine dumps are present in both Tooele Valley and the WDHIA. Coarse rock fragments mapped as Qf are present on the west side of Tooele Valley in mine dumps near Flux, where high-calcium limestone was once mined, and near Dolomite, where dolomite extraction is ongoing (Chemstar Inc., 1990) (plate 1). Rounded gravel is also mapped as Qf along the walls of a gravel pit on the southern tip of the Grayback Hills in the WDHIA (plate 1). The gravel is used as road base in local highway construction.

Tailings are present only in Tooele Valley, and occur both as silty, fine sand associated with tailings ponds, and as coarse rock fragments. At the northern tip of the Oquirrh Mountains a small tailings pond was possibly associated either with the Calera Mill, a facility in operation during World War II to process cobalt ore from Idaho (Bryce Tripp, UGS, verbal communication, 1991), or with the processing of sodium sulfate found along the southeastern shore of Great Salt Lake (Wilson and Wideman, 1957). Two tailings ponds are mapped northeast of Tooele near the International Smelter, a copper smelter constructed in 1910 and operated into the 1970s (Hansen, 1963), and coarse tailings are present nearby along the margins of Pine Creek (plate 1). An extensive tailings pond at Bauer (plate 1), north of Stockton, is related to a selective flotation plant. The plant, now inactive, processed metallic ore that was transported from the Honerine Mine in the adjacent Oquirrh Mountains through a long adit excavated in Quaternary lacustrine and alluvial deposits (Gilluly, 1932).

Lacustrine Deposits

Undifferentiated lacustrine and alluvial deposits (Qla): Much of the surficial material within the piedmont zone below the Bonneville shoreline, both in Tooele and northern Rush Valleys and in the WDHIA, is mapped as Qla. In the extensive piedmont zones of the Oquirrh, Stansbury, and Cedar Mountains, this unit consists mainly of alluvial-fan deposits of pre-Bonneville lake-cycle age that were only moderately reworked by lacustrine processes. As a result, the mapped area contains alluvial-fan deposits overlain by a thin cover of lacustrine sediment, generally less than 10 feet (3 m) thick. The lacustrine sediment is coarser grained in the proximal piedmont sector, and finer grained in the distal sector, because the pre-lake fan material was finer near the distal end of the fan. In these places the unit is commonly expressed geomorphically as pre-Bonneville alluvial fans etched by Lake Bonneville shorelines. Smaller areas of Qla, notably on the margins of the Grayback Hills, are found where Lake Bonneville lacustrine deposits were slightly reworked by post-lake alluvial-fan processes, or where lacustrine and alluvial gravels intertongue. This unit is of latest Pleistocene through latest Holocene age.

Lacustrine clay (Qlc): Qlc consists primarily of water-saturated, thinly bedded to laminated, lake-deposited clay, but also includes small amounts of marl and non-lacustrine sand and silt. The unit is generally less than 10 feet (3 m) thick, and is commonly overlain by a thin crust of efflorescing salts or thin layer of sand- and silt-sized gypsum particles, particularly where adjacent to coarser-grained deposits. Qlc is found only in mud flats on the western margin of the WDHIA, and the deposits continue westward into the Great Salt Lake Desert (plate 1). Qlc is of latest Pleistocene and earliest Holocene age, but may include surficial layers of younger alluvial mud or playa-lake deposits that are not differentiated.

Lacustrine mud (Qlf): Fine-grained lacustrine deposits (Qlf) are present on the distal piedmont sector within northern and central Tooele Valley. They also occur within the WDHIA at the lower elevations of Ripple Valley and between the Grayback Hills and the mud flats. Qlf is generally less than 30 feet (9 m) thick, and consists of lake-deposited sediment in which silt and clay predominate, and sand and marl are subordinate. In some places this unit is reworked into shrub-coppice dunes; elsewhere, eolian deflation has removed fines down to the level of underlying water-saturated mud, leaving isolated remnant buttes. In northern Tooele Valley, northeast of Grantsville, this unit contains elliptical depressions from 5 to 50 feet (2 to 15 m) in diameter. Many of the depressions are surrounded by a raised rim of sand, and springs emanate from within some of them. Depressions without springs are commonly plugged with dark, peaty clay. These features superficially resemble liquefaction-induced sand blows, but they are likely due to artesian pressure and flowage upward through a leaky confining layer (Obermeier and others, 1990). In the WDHIA, however, this unit is characterized by 1- to 2-foot (0.3-0.6-m) wide crescentic, unvegetated features first named "desert ripples" by Ives (1946), who attributed their formation to a process of precipitation, evaporation, and wind transport of sediment. Qlf is of latest Pleistocene to earliest Holocene age.

Lacustrine gravel (Qlg): Shoreline gravel deposited by Lake Bonneville and Great Salt Lake is mapped as Qlg. It is generally thin, but may be up to 30 feet (9 m) thick in some barrier beaches, and is up to 200 feet (60 m) thick in spits of southern Tooele Valley. Tufa encrustations on gravel beach ridges at the northern end of Tooele Valley, remnants of a tufa drapery on the north-facing slope of the Stockton Bar, and tufa masses on the western slope of the Grayback Hills near the Stansbury shoreline are too small to be mapped individually and are included in Qlg. Small, unmappable tufa encrustations near the Stansbury and Provo shorelines at the north end of the Oquirrh Mountains are included as part of the adjacent Qla and bedrock map units.

Alluvial fans constitute the immediate source of material for most lacustrine gravel depositional features, although Tertiary volcanic rocks are a significant source of lacustrine gravel in the Grayback Hills. Geomorphically, the gravel deposits are found as beaches, spits, tombolos, bayhead barriers, and cuspate barriers (V-bars). Outcrops mapped as Qlg typically contain significant layers of interbedded sand; the proportion of sand increases in the north-central portion of Tooele Valley, away from the mountain fronts. This reflects distance from bedrock source areas, as well as the finer grain size of the distal portions of underlying alluvial-fan deposits. Qlg deposits are primarily latest Pleistocene in age, but some lacustrine gravel in northeastern Tooele Valley was deposited in the Holocene.

Most Lake Bonneville shoreline gravel, and the largest associated depositional features, were deposited during the lake's transgressive phase. Significant transgressive gravel accumulations in Tooele and northern Rush Valleys are located on the east side of Tooele Valley near Mills Junction (plate 1), where bedrock outcrops and gravity contours (Johnson, 1958; Cook and Berg, 1961) suggest that Qlg was deposited on a preexisting bedrock high; at the pass between Tooele and Rush Valleys near the town of Stockton (plate 1), where a cross-valley baymouth barrier and associated spit was constructed (Burr and Currey, 1988, 1992; Burr, 1989); and near the mouth of South Willow Canyon (Currey, 1980, 1982) (plate 1), where a spit indicates a southeastward direction of longshore transport. In northeastern Tooele Valley, parallel beach ridges form the Holocene highstand shoreline at an elevation of 4,221 feet (1,287 m). These ridges extend eastward from the silt dunes at the Holocene highstand level, and oolitic sand is an important constituent of the gravel beaches. In the Grayback Hills of the WDHIA, significant transgressive gravel accumulations are mapped as Qlg (plate 1), but here this unit contains large quantities of volcanic talus (Davies, 1980) that cannot be differentiated at the map scale. A large, transgressive gravel spit extends from the piedmont slope of the Grassy Mountains into the northern part of the WDHIA in Ripple Valley (Doelling, 1964).

Lake Bonneville shoreline gravel was also deposited during the lake's regressive phase. Whereas Bonneville regressive gravel accumulations are volumetrically smaller than transgressive gravel accumulations, significant regressive depositional Qlg features are present. One such feature is the Provo shoreline in southern Tooele Valley, north of the Stockton Bar. Here, the shoreline is a ramp of progradational and aggradational gravel beach ridges that formed at different lake levels caused by persistent landsliding in the flood-scoured Red Rock Pass threshold area (Burr and Currey, 1988, 1992; Burr, 1989).

A conspicuous feature of Great Salt Lake's post-Bonneville transgressive phase is a spit in northern Tooele Valley, extending to the southwest from Mills Junction. The road to Grantsville lies atop the crest of this well-drained spit at an elevation of 4,262 feet (1,299 m), and the community of Stansbury Park occupies an area on the lee side of the spit that was reported to have been a broad, marshy lagoon (Eardley and others, 1957). Eardley and others (1957) recognized the association of this and other depositional features with the transgressive Gilbert shoreline complex of Great Salt Lake.

Lacustrine-lagoonal deposits (Qll): Qll consists of silt, clay, and marl, with small amounts of fine sand-sized material, deposited in lagoons behind Lake Bonneville gravel barriers. A minor amount of post-lacustrine, Holocene sediment may have washed into the depressions by slope wash. The lagoon deposits are generally less than 10 feet (3 m) thick, and latest Pleistocene in age. The Qll unit is present in two locations in Tooele Valley. In the vicinity of Stansbury Park, a lagoon occurs on the lee side of the Mills Junction spit (plate 1). Lagoon deposits are found on the northern edge of South Mountain, but a large portion of the lagoon is overlain by Holocene alluvial-fan deposits (plate 1). The Qll unit is also in several locations along the flanks of the Grayback Hills in the WDHIA (plate 1). Much of the adjacent gravel barriers have been excavated near the southernmost lagoon at the tip of the Grayback Hills, exposing the fine-grained lagoonal deposits along the lagoon margin.

Lacustrine marl (Qlm): Qlm includes Gilbert's (1890) pelagic Lake Bonneville white marl, as well as younger, alluvially reworked white sandy marl and marly sand. Qlm has been mapped only in the vicinity of the Grayback Hills in the WDHIA (plate 1), where it is generally less than 6 feet (2 m) thick. Here, marl is found topographically below the Stansbury shoreline as isolated erosional remnants which typically contain ostracodes and rare gastropods. Marl was also observed on the northeastern margin of Tooele Valley, interbedded with other lacustrine deposits, but outcrops were too small to be mapped. Qlm is latest Pleistocene to middle Holocene in age.

Lacustrine sand (Qls): Qls is sand, marly sand, or pebbly sand in both Tooele Valley and the WDHIA. The sand is typically quartzitic and feldspathic, very fine to fine grained, silty, and locally clayey. The deposits may be as much as 30 feet (9 m) thick. In Tooele Valley, deposits mapped as Qls lie just above the Provo shoreline northeast of the mouth of Box Elder Canyon (plate 1), near the base of a series of progradational spits. Most of the spits have been mapped as Qlg, but considerable interbedded lacustrine sand exists here as elsewhere in the Qlg deposits of Tooele Valley. In the WDHIA, deposits mapped as Qls are found in three distinct areas. In the north-central portion of the WDHIA, Qls is common on the lower piedmont slope of the Grassy Mountains (plate 1), and forms the lower portion of a gravel-capped spit at the Gilbert shoreline level. Qls is on the upper piedmont slope of the Cedar Mountains in the eastern portion of the WDHIA (plate 1), where lacustrine sand extends from the Bonneville shoreline level to below the Provo shoreline level. Qls also is present in the vicinity of the Grayback Hills (plate 1), and is common as isolated beach and barrier deposits near the Gilbert shoreline level. A probable source for the sand in all areas is the pre-Bonneville alluvial fans, originally derived from Paleozoic quartzites, that underlie the piedmont slopes of the mountain ranges. An important local source for Qls near the Cedar Mountains, however, appears to be pre-Bonneville eolian sand in the foothills of the Cedar Mountains to the southeast of the WDHIA (Maurer, 1970). These dunes may once have extended northward, but have since been eroded by the transgressing Bonneville shoreline. Qls is of latest Pleistocene age.

Mass-Movement Deposits

Debris-flow deposits (Qmf): Debris-flow deposits (Qmf), consisting of a mixture of fine- to coarse-grained material, are found at two sites near the western end of the Stockton Bar near South Mountain (plate 1). One site, first identified by Currey and others (1983), consists of a lobe of material derived from a northeast facing alcove on the northern side of the bar. Its origin is attributed to sapping of the bar by underflow from Rush Bay during and following the Bonneville flood drawdown north of the bar. The other site, identified in Solomon (1993), consists of a more subtle lobe derived from the southeast face of the bar and associated small spits on the southern side of the bar. This material probably originated somewhat later than Qmf at the first site, and is likely associated with a decrease in the lake level south of the bar following the isolation of Rush Valley as a separate basin after the Bonneville flood. Younger alluvial-fan deposits, however, overlie parts of Qmf at the second site. Qmf deposits are therefore as old as latest Pleistocene in age, and may be as young as earliest Holocene.

Landslide deposits (Qms): Several slide blocks in the vicinity of Lake Point Junction and Black Rock Canyon, at the northern end of the Oquirrh Mountains, were mapped by Tooker and Roberts (1971a) as landslide debris (Qms). However, they are not shown on plate 1 because of the map scale. These blocks are composed of interlayered limestone, shale, quartzite, and sandstone derived from the Erda Formation of Pennsylvanian age. The blocks were undercut on steep slopes by erosion and Lake Bonneville wave action.

Irregular, unconformable slide blocks are also present on the south side of Soldier Canyon, southeast of Stockton (Tooker and Roberts, 1992). These blocks are composed of detached and rotated Great Blue Limestone of upper Mississippian age, and Manning Canyon Shale of lower Pennsylvanian and upper Mississippian age. Qms is latest Pleistocene or Holocene in age.

Talus (Qmt): Talus (Qmt), or rock-fall debris, is found at scattered locations along the western slope of the Oquirrh Mountains from the city of Tooele northward. Qmt has accumulated where the Bonneville shoreline abrasion platform has undercut bedrock outcrops. Qmt is also along the northern edge of the Stockton Bar south of Bauer, where erosion has undercut the steep bar margin. Talus is as much as 30 feet (9 m) thick, and is latest Pleistocene to latest Holocene in age.

Playa Deposits

Playa mud (Qpm): Playa mud (Qpm) consists of poorly sorted clay, silt, and small amounts of sand. Local accumulations of gypsum, halite, and other salts form on the playa surface. Deposits of playa mud are generally less than 10 feet (3 m) thick, and are present in three areas. The largest area of playa mud is along the shores of Great Salt Lake in northern Tooele Valley (plate 1). In northern Rush Valley, Qpm is found within the playa lake bed of Rush Lake (plate 1). In the WDHIA, small mud-filled playas have formed within, and adjacent to, gypsum dunes on the eastern margin of the Great Salt Lake Desert west of the Grayback Hills (plate 1). Playa muds are Holocene-age deposits.

Spring Deposits

Marsh deposits (Qsm): A small area near Dolomite, in northwestern Tooele Valley, is underlain by fine-grained marsh deposits (Qsm) (plate 1). This area is characterized by the presence of a shallow water table and basin-floor springs. These deposits are organic-rich, saline, and less than 6 feet (2 m) thick. They are Holocene-age deposits.

Quaternary Faults

Piedmont fault scarps are in three areas of Tooele and northern Rush Valleys. The first area is on the west side of the northern end of the Oquirrh Mountains, on the eastern edge of Tooele Valley, where a series of down-to-the-west faults displace Quaternary alluvial and lacustrine deposits and form prominent scarps. This zone of scarps in unconsolidated material, designated the Oquirrh marginal fault by Everitt and Kaliser (1980), northern Oquirrh fault zone by Barnhard and Dodge (1988), and Oquirrh fault zone by Olig and others (1996), extends at least 11 miles (18 km) discontinuously from north of Lake Point to just north of Flood Canyon; Barnhard and Dodge (1988) and Solomon (1993, 1996) also extend the fault zone south to the vicinity of Middle Canyon, where it is a bedrock-alluvium contact. An additional segment near Silcox Canyon southwest of Tooele, identified by Everitt and Kaliser (1980) as a scarp of erosional or undetermined origin but herein identified as a fault, consists of a scarp that displaces pre-Lake Bonneville alluvial-fan deposits. Barnhard and Dodge (1988) estimated the most recent faulting event on this fault zone was close to, but not more than, 13,500 years ago. More recently, Olig and others (1996) found evidence that surface fault rupture occurred during the last 4,000-7,000 years, and a potential exists for it to recur.

The second area, south of the town of Stockton, includes a small fault scarp that displaces late Pleistocene Lake Bonneville sediments. The third area, south of East Hickman Canyon in the northwestern corner of Rush Valley, includes a prominent scarp that offsets pre-Lake Bonneville alluvial-fan deposits. The age of most recent surface-fault rupture is no more than about 15,000 years on the former scarp, and the age of most recent surface faulting on the latter scarp is uncertain.

Thomas (1946) described several faults in Tooele Valley, including the Box Elder Canyon fault in the southwestern part of the valley and the Erda, Mill Pond, and unnamed faults near Lake Point in the northeastern part of the valley. He cited topographic evidence for the Box Elder Canyon and Mill Pond faults, hydrologic evidence for the Erda fault, and exposure of an unnamed fault near Lake Point. Gates (1962, 1965) agreed with the existence of the Mill Pond fault, but reinterpreted its trend. Gates (1962) also postulated an extension of the Occidental fault from the Oquirrh Mountains into Tooele Valley near Erda, based on differences in water-level fluctuations, chemical quality of ground water, altitude of the piezometric surface, and gravity on either side of a discontinuous ridge that crosses the Erda area. He thought that the ridge was probably a surface expression of the fault. Gates (1965) described several additional faults in northern Tooele Valley, including the Warm Springs, Fishing Creek, and Sixmile Creek faults. Anomalously high chloride concentrations in, and temperature of, ground water were cited as evidence for the existence of these faults.

Solomon (1993) found no evidence of surface faulting in the vicinity of the supposed faults of Thomas (1946) and Gates (1962, 1965). Gravity (Johnson, 1958) and hydrologic data (Gates, 1962) suggest that the Box Elder Canyon fault does not cross Tooele Valley along the projection interpreted by Thomas (1946). The ridge in the vicinity of the postulated Occidental fault is likely part of a preexisting bedrock high in depositional contact with younger deposits. Although no evidence for surficial faulting was found in this study for any of these faults, hydrologic data suggest that some of them may exist, but their most recent movement may predate the age of surficial deposits.

Everitt and Kaliser (1980) mapped faults in three additional areas, but Solomon (1993) shows that these are lineaments related to depositional features, rather than the result of faulting. In the northern end of Tooele Valley, west of Mills Junction, linear features mapped as the Sixmile Creek fault by Everitt and Kaliser (1980) actually result from the alignment of the ends of several small Lake Bonneville beach ridges. Southeast of Mills Junction, similar linear features, part of the Oquirrh marginal fault of Everitt and Kaliser (1980), result from Lake Bonneville shoreline features in gravel deposited on the margin of a shallow bedrock high. On the west end of South Mountain, a fault mapped by Gilluly (1928, 1932), the South Mountain marginal fault of Everitt and Kaliser (1980), is interpreted as a depositional contact between bedrock and alluvium.

Tooker and Roberts (1992) mapped several faults in alluvium and along the bedrock-alluvium contact in the vicinity of Stockton, some of which were mapped earlier by Gilluly (1932) as bedrock faults extending into alluvium. Solomon (1993) indicates that, for the most part, these are either Lake Bonneville shorelines or stream-terrace edges in pre-Lake Bonneville alluvium. The one exception is the small scarp near Stockton, noted above.

GEOLOGIC HAZARDS

Tooele Valley and the WDHIA were severely impacted by geologic hazards (primarily flooding) in the 1980s, and various other geologic hazards potentially exist. Above-average precipitation in the early 1980s resulted in basement flooding in Erda from shallow ground water, surface flooding in Tooele City from rapid snowmelt and an uncontrolled release of water over the spillway from Settlement Canyon Dam, and landslides and debris flows in canyons in the Oquirrh Mountains on the east side of Tooele Valley. Potential geologic and related environmental hazards include rock falls, debris flows, and flash floods in canyons and along valley margins; earthquake-related hazards; and contamination of ground water in basin-fill aquifers. Adverse foundation conditions also may be present. Silty and sandy sediments subject to liquefaction or hydrocompaction, clayey sediments and mudflats subject to shrinking or swelling, and gypsiferous dunes and mudflats subject to subsidence due to dissolution are all present in Tooele Valley and the WDHIA. A knowledge of these conditions and related hazard potential will provide decision makers with valuable tools to undertake responsible action.

This report defines and describes geologic hazards that are present in Tooele Valley and the WDHIA, and delineates areas in which hazards are likely to occur. Hazards are described individually in subsequent sections of this report. Subsections define the general nature of the hazard (Introduction), and type of damages caused, extent of the hazard, and methods that can be used to reduce the hazard (Effects, Distribution, and Reduction). The hazards maps (plates 2 though 6) delineate the areas subject to geologic hazards and are at the same scale as plate 1. Plates 2 through 6 show the distribution of mapped hazards, and were compiled from 1:24,000-scale derivative maps in Solomon and Black (1995). Not all hazards described in the text were mapped. The maps are only to be used to determine potential hazards that might be encountered. Once potential hazards at a site have been identified using these maps, site suitability must be demonstrated by detailed site characterization. More detailed recommendations for studies and hazard reduction are given in Solomon and Black (1995).

Ground Shaking

Introduction

Ground shaking is the most widespread and frequently occurring earthquake hazard. The Tooele Valley study area is located in the Intermountain seismic belt, a generally north-south-trending zone of earthquake activity bisecting Utah (figure 4). The WDHIA is west of the Intermountain seismic belt. Many active faults capable of producing earthquakes are in this zone. Both Tooele Valley and the WDHIA could be susceptible to ground shaking from a surface-faulting earthquake centered on a nearby fault or distant fault. In addition, earthquakes large enough to cause damage, but don't cause surface fault rupture (up to magnitude 6.5) and thus may not be attributable to a mapped fault, may occur anywhere in the area (Smith and Arabasz, 1991).

Ground shaking is caused by seismic waves generated during an earthquake. The waves originate at the source of the earthquake (or focus) and radiate out in all directions (figure 5). The extent of property damage and loss of life due to ground shaking depends on factors such as: (1) proximity of the earthquake and strength of seismic waves at the surface (horizontal motions are the most damaging); (2) amplitude, duration, and frequency of ground motions; (3) nature of foundation materials; and (4) building design (Costa and Baker, 1981).

A building need only withstand the force of gravity (1 g) to support its own weight. However, during an earthquake, a structure is also subjected to horizontal accelerations that may be greater than that of gravity. Accelerations are normally expressed in decimal fractions of the acceleration due to gravity (g) (32 feet/second^2 [9.8m/s^2]).

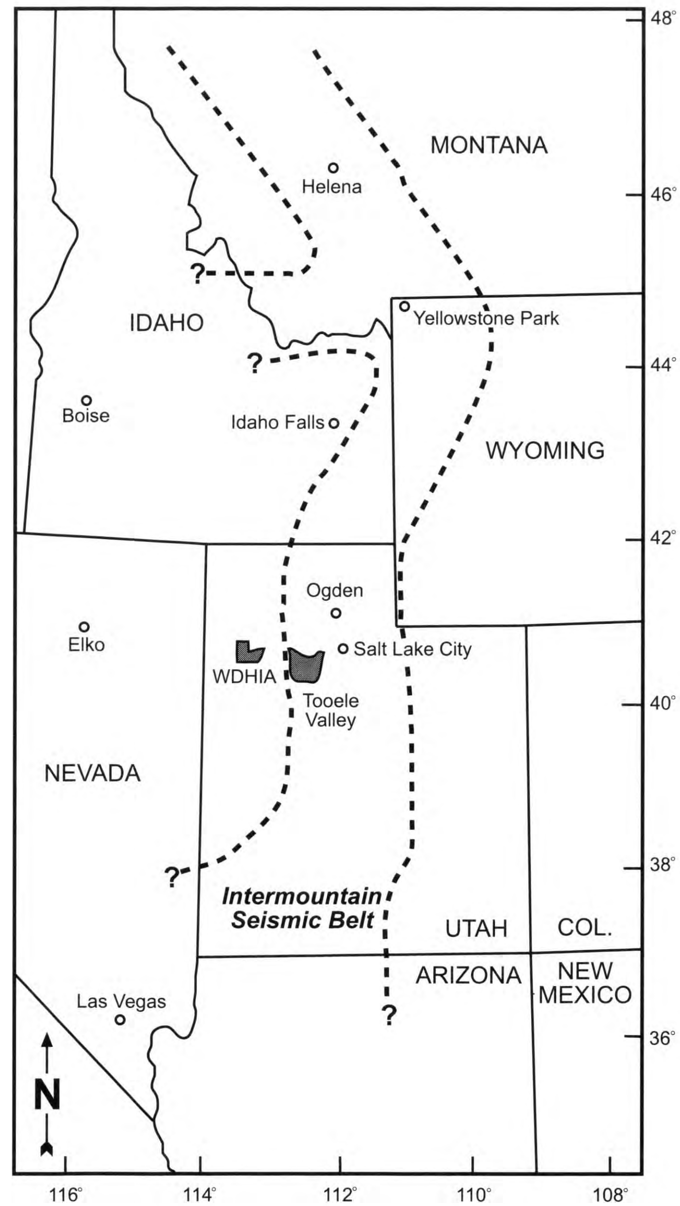


Figure 4. Tooele Valley and the West Desert Hazardous Industry Area (WDHIA) with respect to the Intermountain seismic belt (modified from Smith and Arabasz, 1991).

The threshold for damage to weak structures (buildings not specifically designed to resist earthquakes) is roughly 0.1 g (Richter, 1958).

Larger magnitude earthquakes typically cause more damage because they result in larger amplitudes of ground motion for longer periods of time. Because energy is dissipated as seismic waves travel through the earth, ground shaking generally decreases with increasing distance from the epicenter. Seismic waves can travel long distances, as shown in the September 19, 1985, magnitude 8.1 Michoacan, Mexico earthquake that devastated portions of Mexico City, 240 miles (386 km) from an epicenter off the Pacific coast of Mexico (Ghosh and Kluver, 1986).

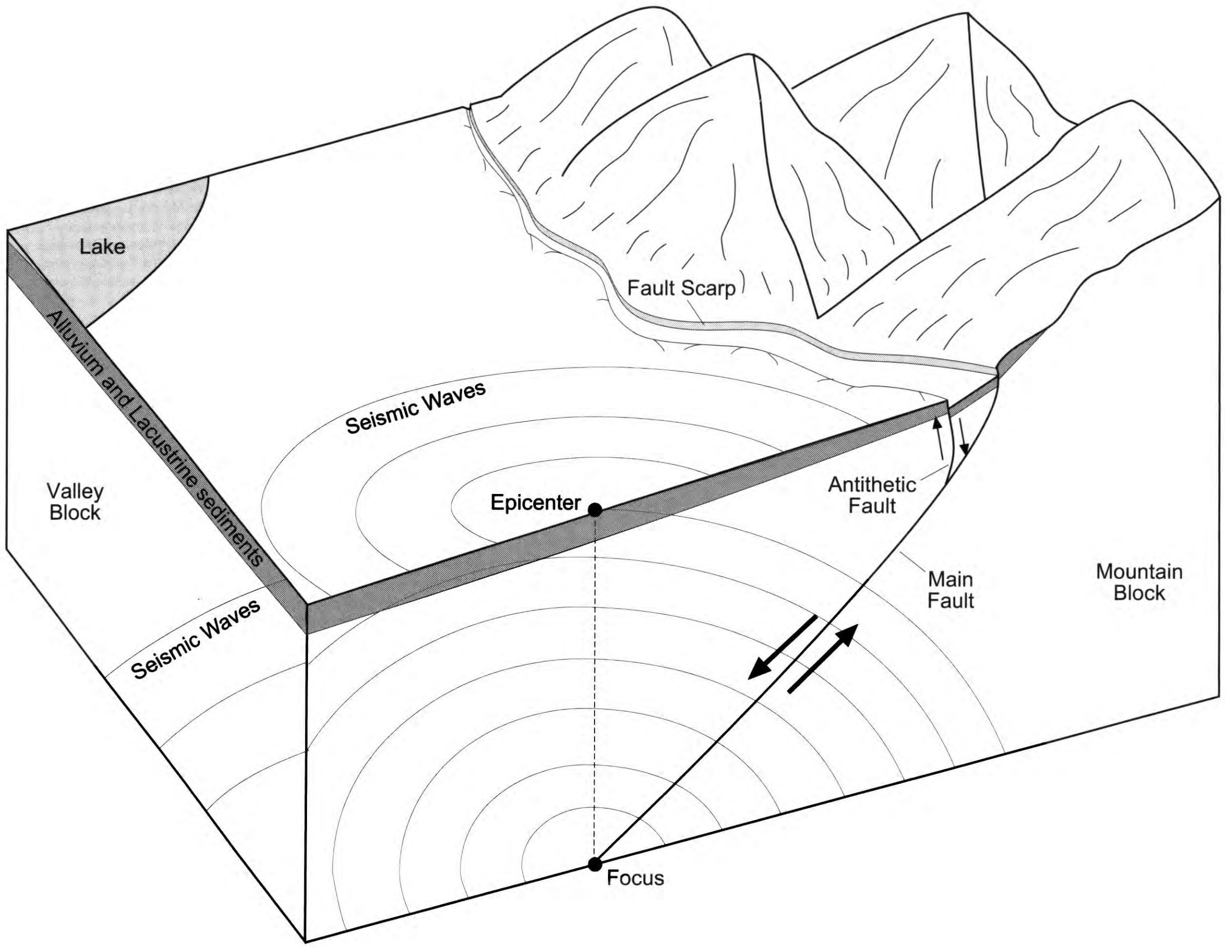


Figure 5. *Factors affecting ground shaking, including: fault location, earthquake focus and epicenter, surficial deposits, and propagation of seismic waves (modified from Robison, 1993a).*

In certain cases, earthquake ground motions can be amplified and shaking duration prolonged by local site conditions (Hays and King, 1982). The degree of amplification depends on factors such as thickness of the sediments and their physical characteristics such as "stiffness" or "softness." "Soft" sediments are generally clays with low shear-wave velocities. Studies along the Wasatch Front of weak ground motions produced by distant explosions at the Nevada Test Site indicate that certain ground motions are amplified on soft-soil sites by as much as 10 to 13 times relative to rock sites (Hays and King, 1982). Studies of earthquakes worldwide have demonstrated that near-surface "soft" sediments amplify ground motions (Gutenberg, 1957; Seed and others, 1987; Borcherdt and others, 1989; Jarpe and others, 1989). These "soft" sediments include fine-grained fluvial or lake deposits, which are extensive throughout Tooele Valley and the WDHIA. Recent theoretical studies by Adan and Rollins (1993) and Wong and Silva (1993) indicate that amplification may also occur in shallow stiff (sandy and gravelly) soils. These conditions may be found around the periphery of Tooele Valley along mountain fronts and around the Grayback Hills in the WDHIA.

Effects, Distribution, and Reduction

Failure of human-engineered structures from ground shaking is responsible for most earthquake losses. Proper building design can reduce damage. Older unreinforced-masonry buildings are at a higher risk than newer earthquake-resistant designs. Studies have cited a high risk from ground shaking along the Wasatch Front because of the large number of older buildings (Algermissen and others, 1988).

Horizontal motions are typically the most damaging type of ground shaking. In addition, different types of structures are affected by different frequencies of vibration. When the dominant frequency of ground shaking matches the natural frequency of vibration of a structure (a function of building height and construction type), resonance can occur that may result in severe damage or collapse. Proximity to the source of the earthquake also influences the damage caused by ground shaking. Ground motion maps prepared by Frankel and others (1996) show the expected peak horizontal acceleration on bedrock with a 10 percent and 2 percent chance of being exceeded in 50 years (figure 6). Horizontal accelerations on the 10 percent in 50-year map are typically used in building design. These accelerations range from 0.10 to 0.20 g in Tooele Valley and from 0.07 to 0.10 g in the WDHIA (figure 6). As an example of damaging ground motions, accelerations of 0.26 and 0.29 g were recorded close to the I-880 freeway overpass that collapsed during the 1989 Loma Prieta earthquake in California (Shakal and others, 1989).

Bolt (1993) relates peak horizontal acceleration to the Modified Mercalli intensity scale. The Modified Mercalli intensity scale measures the intensity of ground shaking through a ranking based on observed effects and damage (table 2). A peak horizontal acceleration of 0.12 g, equivalent to Modified Mercalli intensity VII, was recorded 16 miles (25 km) from the epicenter of the ML 5.7 1962 Cache Valley earthquake (Smith and Lehman, 1979). Despite the relatively modest ground motions, this earthquake caused nearly $1 million of damage (1962 dollars; Lander and Cloud, 1964) and illustrates the power of even moderate-sized earthquakes to cause considerable damage. By comparison, estimated damage from the 1993 magnitude 5.6 Scotts Mills earthquake in Oregon is at least $30 million (Madin and others, 1993).

Both Tooele Valley and the WDHIA are susceptible to ground shaking from earthquakes on mapped faults and faults not evident at the surface. Although the principal active fault mapped in Tooele Valley is the Oquirrh fault zone, several other potentially active faults are within 30 miles (48 km) of Tooele Valley: (1) the Wasatch fault zone, at the base of the Wasatch Range east of Tooele Valley; (2) faults such as the Mercur, St. John Station, and Clover fault zones in Rush Valley to the south (Barnhard and Dodge, 1988), and other lesser-known faults in northern Rush Valley (Tooker and Roberts, 1992; Solomon, 1993); (3) the Stansbury fault zone, on the east side of Skull Valley west of Tooele Valley (Barnhard and Dodge, 1988; Hecker, 1993; Helm, 1995); and (4) the East Great Salt Lake fault zone, beneath Great Salt Lake west of Antelope Island (Pechmann and others, 1987; Arabasz and others, 1992). No active faults are mapped within the WDHIA, but two potentially active faults are within 30 miles (48 km) of the WDHIA: (1) the Puddle Valley fault zone, on the

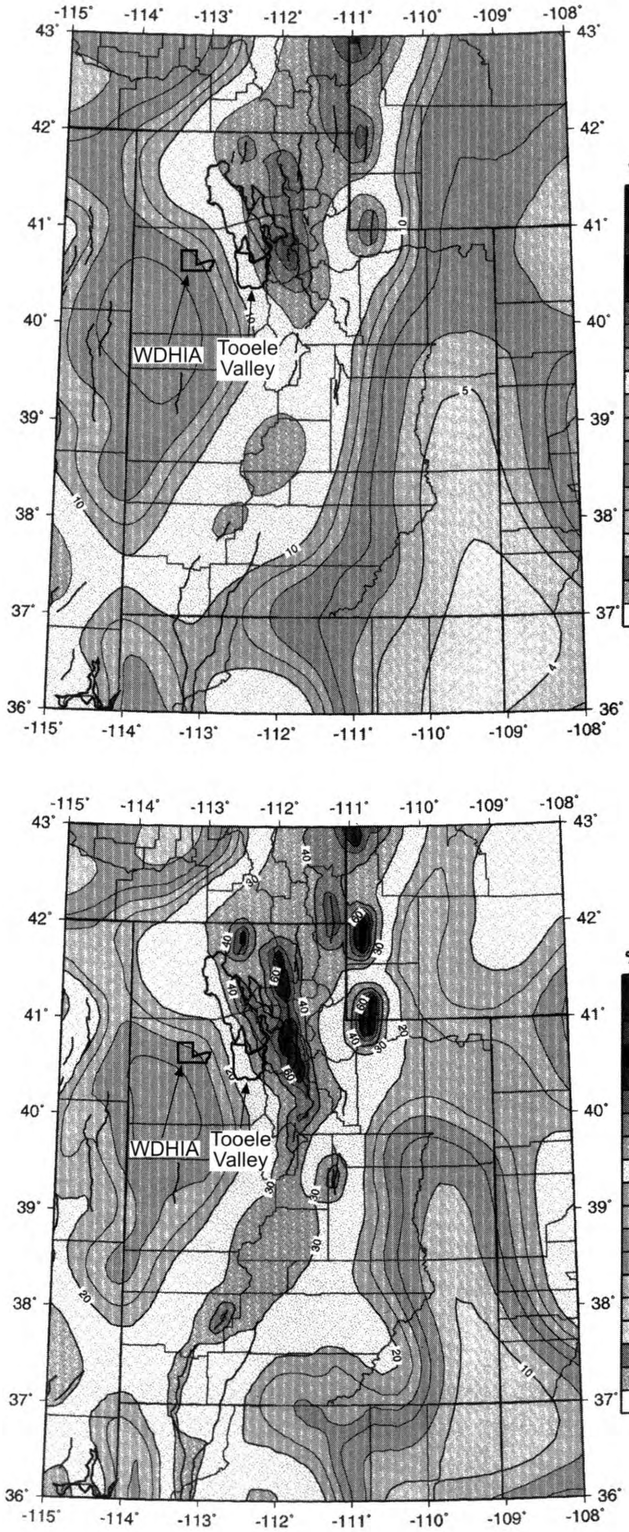


Figure 6. Peak horizontal acceleration on bedrock with a 10 percent (top) and 2 percent (bottom) chance of being exceeded in 50 years (after Frankel and others, 1996).

west side of Puddle Valley to the northeast (Barnhard and Dodge, 1988); and (2) the Stansbury fault zone.

Ground shaking cannot be avoided because it is so widespread, and the best alternative to reduce the poten-

Table 2. Modified Mercalli intensity scale (modified from Bolt, 1993).

	Intensity value and description	Peak horizontal acceleration
I.	Felt only by a very few under especially favorable circumstances.	
II.	Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.	
III.	Felt quite noticeably indoors, especially on upper floors of buildings. However, many do not recognize it as an earthquake. Standing automobiles may rock slightly. Vibration is like a passing truck. Duration estimated.	
IV.	Felt indoors by many during the day, outdoors by only a few. At night some people awakened. Dishes, windows, and doors disturbed; walls make creaking sounds. Sensation is like a heavy truck striking the building. Standing automobiles rocked noticeably.	0.015g - 0.02g
V.	Felt by nearly everyone; many people awakened at night. Some dishes and windows broken; cracked plaster in a few places; unstable objects overturned. Disturbance of trees, poles, and other tall objects is sometimes noticed. Pendulum clocks may stop.	0.03g - 0.04g
VI.	Felt by all, many frightened and run outdoors. Some heavy furniture is moved; a few instances of fallen plaster and damaged chimneys. Damage is slight.	0.06g - 0.07g
VII.	Everybody runs outdoors. Damage is: (1) negligible in buildings of good design and construction; (2) slight to moderate in well-built ordinary structures; (3) considerable in poorly built or badly designed structures. Some chimneys are broken. Noticed by people driving cars.	0.10g - 0.15g
VIII.	Damage is: (1) slight in specially-designed structures; (2) considerable in ordinary buildings, with partial collapse; and (3) great in poorly built structures. Panel walls thrown out of frame structures. Chimneys, factory stacks, columns, monuments, and walls fall down. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well water. People driving cars disturbed.	0.25g - 0.30g
IX.	Damage is considerable in specially designed structures; well-designed frame structures thrown out of plumb. Damage is great in ordinary buildings, with partial collapse. Buildings shifted off of foundations. Ground conspicuously cracked.	0.50g - 0.55g
X.	Some well-built wooden structures destroyed; most masonry and frame structures with foundations destroyed; ground is badly cracked. Rails bent. Numerous landslides from river banks and steep slopes. Sand and mud shifted. Water splashed and slopped over river banks.	more than 0.60g
XI.	Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in the ground. Underground pipelines completely out of service. Earth slumps in soft ground. Rails bent greatly.	
XII.	Damage total. Waves seen on ground surface. Lines of sight and level distorted. Objects thrown into the air.	

tial effects of the hazard is to strengthen structures. Because failure of human-engineered structures is the cause of most earthquake losses, engineers, building officials, and architects play a key role in reducing losses by implementing improved design and construction practices.

The Uniform Building Code (UBC), which was adopted statewide in 1987, specifies minimum requirements for earthquake-resistant design and construction to minimize structural damage and loss of life from earthquakes (International Conference of Building Officials, 1997). It applies to all new building construction, including schools, hospitals, commercial and residential buildings, fire and police stations, and power plants. The "Earthquake Regulations" in the code were extensively revised for the 1988 and 1997 editions, but the basic philosophy to reduce potential structural damage and protect lives during earthquakes remained the same. In any case, the regulations do not ensure that the structure or its contents will not be damaged during an earthquake, a painful lesson learned by many building owners since adoption of the first earthquake-resistant design provisions in 1961.

Two factors, Z and C, are defined in the 1997 UBC to quantify the minimum level of ground shaking that structures must be designed to withstand without collapse. In seismic zone 4, each site is also assigned a near-source factor (N), based on the distance to known seismic sources. Z is the seismic zone factor, which attempts to quantify ground motions on rock. Specifically, Z is tied to accelerations on rock with a 10 percent chance of being exceeded in 50 years (figure 6; top). C is the seismic coefficient, which attempts to quantify the effects of near-surface sediments on the ground motions. C is divided into acceleration (C_a) and velocity (C_v) coefficients. In seismic zones 1 through 3, C_a ranges from 0.06 to 0.36 and C_v ranges from 0.06 to 0.84 based on the type and thickness of sediments underlying a site; larger seismic coefficients attempt to account for larger amplifications of ground motions by near-surface "soft" sediments (table 3). Tooele Valley is in seismic zone 3, whereas the WDHIA is in zone 2B near the edge of zone 3.

Surface Fault Rupture

Introduction

Movement along faults at depth generates earthquakes. During earthquakes larger than Richter magnitude 6.5, ruptures along normal faults in the intermountain region generally propagate to the surface (Smith and Arabasz, 1991) as one side of the fault is uplifted and the other side downdropped (figure 7). The resulting fault scarp has a near-vertical slope. Faults that show evidence of recurrent movement during Quaternary time (last 1.6 million years) have a potential to generate earthquakes that could cause surface fault rupture; the potential is highest along those faults that show evidence of recurrent movement during the Holocene (last 10,000 years).

Surface fault rupture is a potential hazard on faults in the Tooele Valley study area, but no known active faults (and therefore little potential for surface fault rupture) are in the WDHIA. Tooele Valley is the result of millions of years of faulting, which has uplifted the Oquirrh and Stansbury Mountains on the east and west, and downdropped the basin between them (Everitt and Kaliser, 1980; Barnhard and Dodge, 1988). Although no surface faulting has occurred in Tooele Valley in historical time, the Oquirrh fault zone along the base of the Oquirrh Mountains has had a large-magnitude earthquake accompanied by surface faulting within the last 7,000 years (Olig and others, 1996). Other faults in Tooele Valley and northern Rush Valley show evidence for activity during Quaternary time. A potential exists for surface rupture to recur along these faults, and structures which straddle them may be damaged or destroyed by surface fault rupture.

Effects, Distribution, and Reduction

During surface-faulting earthquakes, displacement typically occurs on the main surface trace of the fault zone (Schwartz and Coppersmith, 1984). This displacement forms a near-vertical scarp, commonly in unconsolidated surficial deposits, that begins to ravel and erode back to the material's angle of repose (33-35 degrees). Antithetic faults (faults with an opposite sense of movement from the main fault) on the downthrown side of the main trace may also form, generally exhibiting a lesser amount of offset, but sometimes as much as several feet (figure 8). The zone between these two faults may be faulted and tilted in a complex manner. In some cases, a broad zone of flexure may form on the downthrown side of the main fault in which the surface is tilted downward toward the fault zone. Deformation associated with surface fault rupture can damage or destroy structures and sever lifelines.

Plate 2 shows main fault traces with a potential for future movement in the Tooele Valley study area. These faults include the Oquirrh fault zone and unnamed faults in southern Tooele Valley and northwestern Rush Valley. Plate 2 also shows special study areas where surface fault rupture is a possible hazard and should be considered. The special study areas, which follow fault traces mapped by Solomon (1993), are about 500 feet (152 m)

Table 3. Soil-profile types (based on geotechnical data) and seismic coefficients Ca (acceleration) and Cv (velocity). Z is the seismic zone factor, N is a near-source factor (seismic zone 4 only). In locations where soil properties are not known in sufficient detail to determine the soil profile type, use soil profile SD (modified from International Conference of Building Officials, 1997).

TYPE	DESCRIPTION	Shear Wave Velocity of top 100 feet (30.5 m) of soil profile, feet/second (m/s)
S_A	Hard Rock	> 5,000 (1,500)
S_B	Rock	2,500 to 5,000 (760-1,500)
S_C	Very Dense Soil and Soft Rock	1,200 to 2,500 (360-760)
S_D	Stiff Soil Profile	600 to 1,200 (180-360)
S_E	Soft Soil Profile	< 600 (180)
S_F	Soil requiring site-specific evaluation.	

SOIL TYPE	SEISMIC ZONE (SEISMIC ZONE FACTOR, Z) 1 (Z=0.075)	2A (Z=0.15)	2B (Z=0.2)	3 (Z=0.3)	4 (Z=0.4)
S_A	C_a=0.06 C_v=0.06	C_a=0.12 C_v=0.12	C_a=0.16 C_v=0.16	C_a=0.24 C_v=0.24	C_a=0.32N_a C_v=0.32N_v
S_B	C_a=0.08 C_v=0.08	C_a=0.15 C_v=0.15	C_a=0.20 C_v=0.20	C_a=0.30 C_v=0.30	C_a=0.40N_a C_v=0.40N_v
S_C	C_a=0.09 C_v=0.13	C_a=0.18 C_v=0.25	C_a=0.24 C_v=0.32	C_a=0.33 C_v=0.45	C_a=0.40N_a C_v=0.56N_v
S_D	C_a=0.12 C_v=0.18	C_a=0.22 C_v=0.32	C_a=0.28 C_v=0.40	C_a=0.36 C_v=0.54	C_a=0.44N_a C_v=0.64N_v
S_E	C_a=0.19 C_v=0.26	C_a=0.30 C_v=0.50	C_a=0.34 C_v=0.64	C_a=0.36 C_v=0.84	C_a=0.36N_a C_v=0.96N_v
S_F	Site-specific geotechnical investigations and dynamic site response analyses required.				

wide on both the upthrown and downthrown sides of the main fault scarp. Site-specific investigations addressing surface-fault-rupture hazards are recommended in the special study areas because the fault maps are not detailed enough to include all fault traces and delineate zones of deformation at a particular location.

The Oquirrh fault zone is evident as a series of west-facing normal fault scarps 9.5 to 35.4 feet (2.9 - 10.8 m) high, which displace Quaternary alluvial deposits (Barnhard and Dodge, 1988). The scarps extend discontinuously 11 miles (17 km) north-south along the Oquirrh Mountains, from east of Lake Point to south of Middle Canyon.

Studies have indicated evidence for active faulting on the Oquirrh fault zone. Evidence from trenches excavated across scarps near the mouths of Big Canyon and

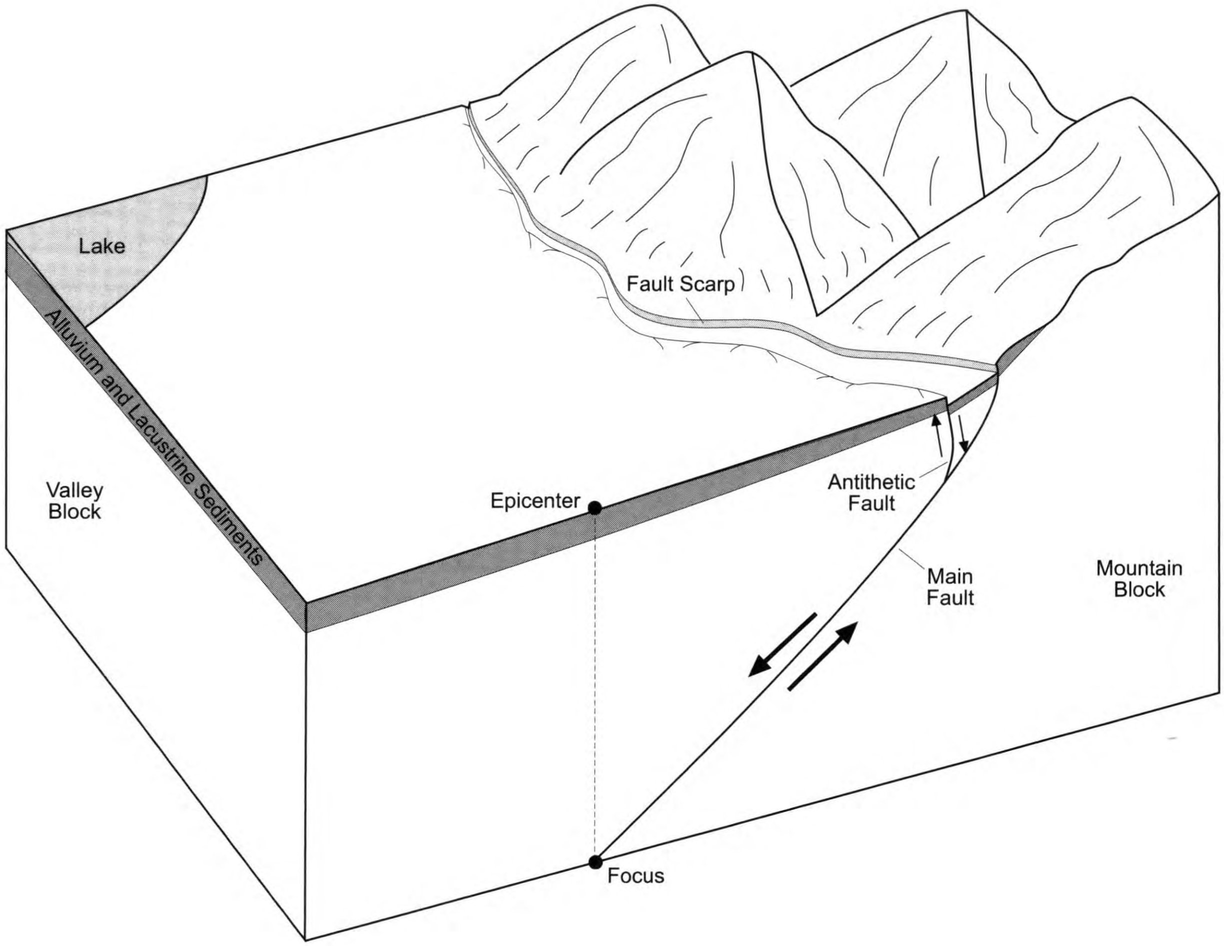


Figure 7. *Normal fault characteristics. The main fault plane likely dips 45-60 degrees toward the valley. Note that the focus of the earthquake is beneath the valley (downdropped) block, not on the trace of surface rupture (fault scarp) (modified from Robison, 1993a).*

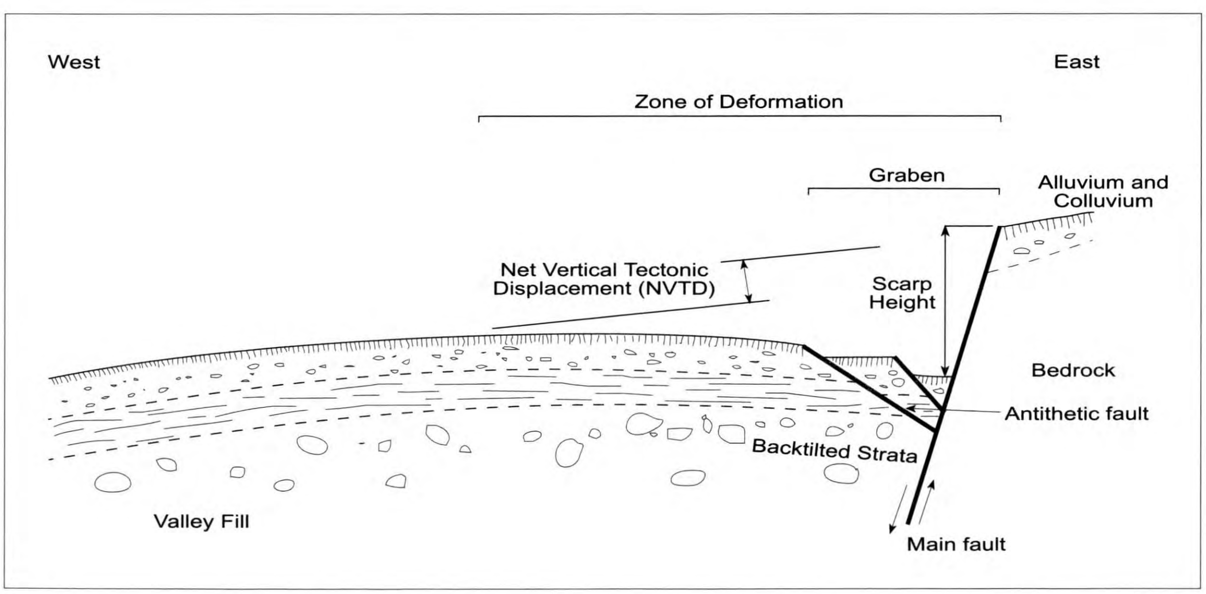


Figure 8. *Normal fault zone features typically found near the ground surface. Although the sketch is not to scale, NVTD is usually 6-9 feet (2-3 m) (modified from Robison, 1993a).*

Pole Canyon show three large-magnitude earthquakes on the Oquirrh fault zone since middle Pleistocene time: (1) a most recent surface-faulting earthquake (MRE) between 4,300 and 6,900 years ago, (2) an event between 20,300 and 26,400 years ago, and (3) an event sometime before 32,800 years ago (Olig and others, 1996). Olig and others (1996) indicate the Bonneville shoreline was displaced 8 to 10 feet (2.5-3.0 m) during the MRE, a large amount considering the relatively short length (7.5 miles [12 km]) of the fault. Geomorphic evidence also indicates recurrent faulting near the northern end of the Oquirrh fault zone, where the scarp of the MRE diverges from an older scarp (Barnhard and Dodge, 1988). The compound scarps, representing both the MRE and older surface-faulting events, are up to twice as high as the single-event scarp and have surface displacements of up to 24 feet (7.3 m) (Barnhard and Dodge, 1988; Hecker, 1993).

Other faults in the Tooele Valley study area also have evidence for Quaternary movement. These include: (1) a discontinuous set of west-facing normal fault scarps south of Tooele, which displace late Pleistocene alluvial-fan deposits topographically above the Bonneville shoreline (Tooker and Roberts, 1992; Solomon, 1993); (2) a 0.2-mile (0.3-km) long west-facing normal fault scarp south of Stockton, which displaces Holocene to late Pleistocene Lake Bonneville deposits (Tooker and Roberts, 1992; Solomon, 1993); and (3) a 0.8-mile (1.3-km) long east-facing normal fault scarp in northwestern Rush Valley near East Hickman Canyon, which displaces Pleistocene alluvial-fan deposits topographically above the Bonneville shoreline (Solomon, 1993). A Pleistocene-age fault not evident at the surface was also found in a gravel pit roughly 2 miles (3 km) northwest of Tooele (Solomon and others, 1994), and similar faults may exist elsewhere. No detailed investigations have been conducted on these faults and no paleoseismic data are available.

Designing a structure to withstand several feet of displacement through its foundation is nearly impossible, both technically and economically. Structural damage may be great, and buildings in the zone of deformation may not be safe for occupants following a large earthquake. Because surface fault rupture occurs without warning and is a life-threatening hazard, avoidance of the main trace of the fault is the most effective hazard-reduction technique. However, in some areas adjacent to the main trace within the zone of deformation, avoidance of small-displacement subsidiary faults may not be necessary. Less damaging smaller displacements and tilting may occur, and structural measures may be taken to reduce damage and threat to life. Youd (1980) suggests displacements less than 4 inches (10 cm) will probably cause damage that is repairable.

Tectonic Subsidence

Introduction

Tectonic subsidence is the warping, lowering, and tilting of a valley floor that accompanies surface-faulting earthquakes on normal (dip-slip) faults, such as the Oquirrh fault zone. Subsidence occurred during the 1959 Hebgen Lake earthquake in Montana and 1983 Borah Peak earthquake in Idaho, and geologic evidence indicates tectonic subsidence also occurred during prehistoric earthquakes along the Wasatch Front (Keaton, 1987). Inundation along lake and reservoir shores, and ponding of water in areas with a shallow water table, may be caused by tectonic subsidence. Also, tectonic subsidence may adversely affect certain structures which require gentle gradients or horizontal floors, particularly wastewater-treatment facilities and sewer lines (Keaton, 1987). The extent of seismic tilting is controlled chiefly by the amount and length of surface displacement. Subsidence typically extends only a short distance beyond the ends of the fault rupture. The maximum amount of subsidence should occur at the fault and decrease gradually away on the downdropped valley block.

Tectonic subsidence could be a hazard in Tooele Valley, along known faults with evidence of surface faulting, particularly those having evidence of movement during the last 10,000 years. However, the potential for tectonic subsidence in Tooele Valley has not been studied. The WDHIA has no active faults, and thus the hazard from tectonic subsidence is very low.

Effects, Distribution, and Reduction

The two major types of hazards associated with tectonic subsidence are tilting of the ground surface and flooding from lakes, reservoirs, or shallow ground water (figure 9) (Smith and Richins, 1984). Tilting of the ground surface may compromise gravity-flow structures such as wastewater-treatment plants and sewer lines, and thus prevent them from working properly. Flooding from lakes and reservoirs may damage structures along shorelines and result in injury or loss of life. Subsidence may also cause ground-water levels to rise, causing water to pond and flood basements and buried facilities.

The probability of tectonic subsidence accompanying an earthquake on a specific fault is the same as that for a surface-faulting earthquake, although the extent of subsidence varies. Because no detailed studies have been made of subsidence characteristics of the Oquirrh fault

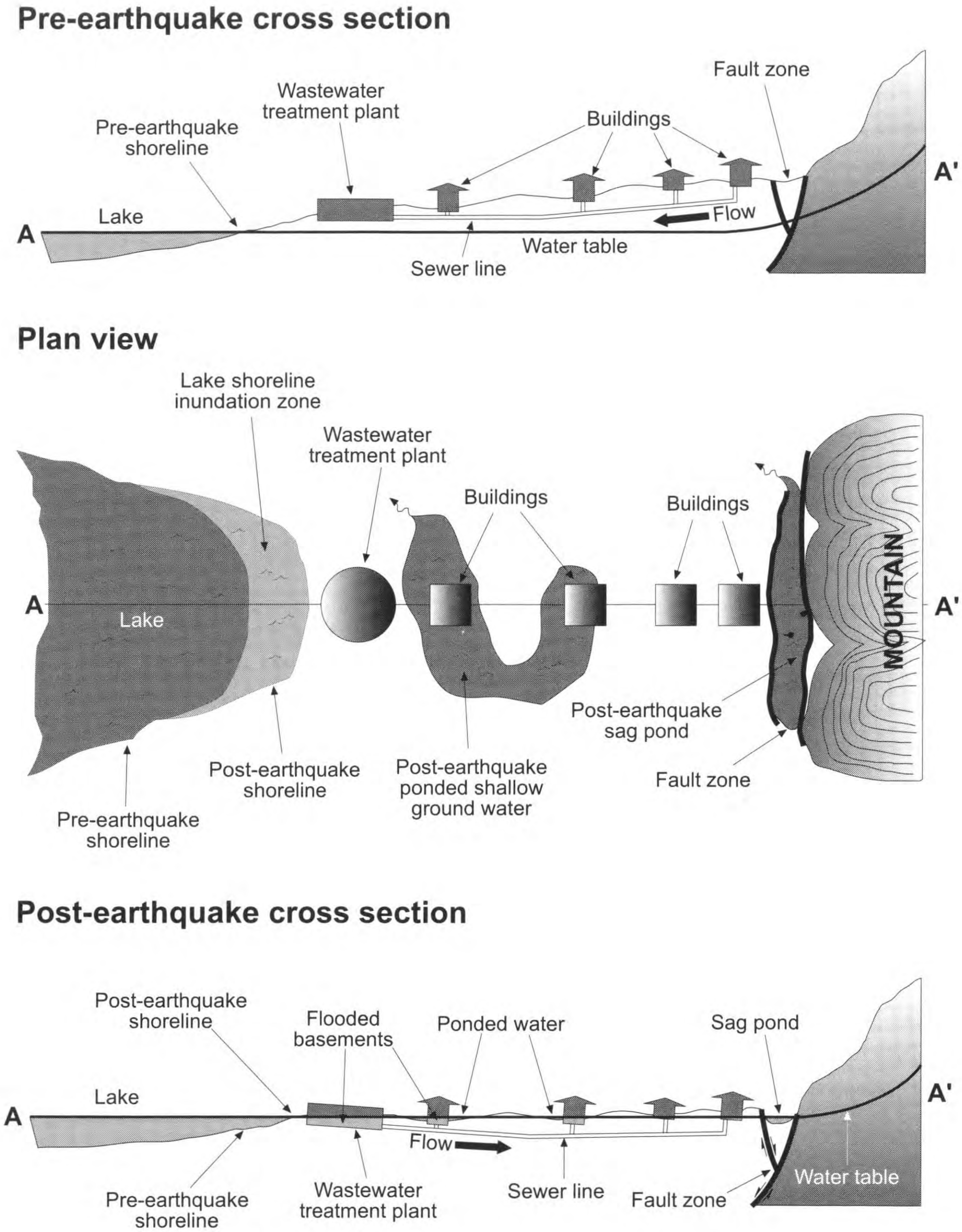


Figure 9. *Hypothetical plan view and cross sections showing tectonic subsidence accompanying a surface-faulting earthquake. Top cross section shows the lake shoreline and structures on the plan view (below) in their pre-earthquake position. Bottom cross section shows the possible effects of tectonic subsidence and their extent on the plan view (above). These effects include inundation along the lake shoreline (lake shoreline inundation zone); post-earthquake flooding, ponded water, and sag ponds (produced by backtilting along the fault zone) due to the rising water table; and changes in gradient from backtilting causing a reversal of flow in sewer lines (modified from Robison, 1993b).*

zone, the effects of subsidence are not known. However, the 1983 Borah Peak earthquake in Idaho may provide a model for subsidence associated with the Oquirrh fault zone. Up to 4.3 feet (1.3 m) of subsidence at the fault was observed following this earthquake, with subsidence extending up to 9.3 miles (15 km) from the fault on the downdropped side (Keaton, 1987).

Tectonic subsidence from an earthquake on the Oquirrh fault zone would be greatest in the eastern part of Tooele Valley on the western (downdropped) side of the fault, where the maximum amount of potential subsidence may occur. Flooding related to tectonic subsidence on the Oquirrh fault zone, as well as ponding of water and disruption of buried facilities, would be greatest in the northeastern part of the valley due to shallow ground-water levels and proximity to the shore of Great Salt Lake.

Flooding problems along the Great Salt Lake shoreline from tectonic subsidence depend on lake levels at the time of the earthquake. The greatest effects would result from high lake levels. At the historical average lake level of about 4,200 feet (1,280 m), flooding due to subsidence is likely within the zone of normal lake flooding. If the potential for an earthquake on the Oquirrh fault zone when lake levels are high is determined to be sufficient to merit hazard reduction, methods such as raising structures above expected flood levels or building dikes should be considered to reduce flooding effects. The magnitude and extent of tectonic subsidence along the Oquirrh fault zone is unclear, and a study similar to Keaton (1987) is required to better define the amount and extent of potential subsidence. Without such a study, estimates of the amount of subsidence can be made based on the amount of fault displacement per earthquake event (from paleoseismic data) and the extent of subsidence from similar historical events.

Because subsidence may occur over a large area, avoidance is generally not practical, except in low-lying lake shoreline areas. Gravity-flow structures (such as wastewater-treatment plants) should be designed to tolerate slight changes in gradient in areas of potential subsidence. Some structures may need to be releveled after tectonic subsidence occurs.

Liquefaction

Introduction

Earthquake-induced liquefaction occurs when ground shaking increases the pressure in the pore water between soil grains, which decreases the stresses between the grains. The loss of intergranular stress can cause the strength of some soils to decrease to nearly zero. When this happens, the soil behaves like a liquid, and therefore is said to have liquefied. Liquefaction of a soil can have four major adverse effects: (1) foundations may crack; (2) buildings may tip; (3) buoyant buried structures, such as septic tanks and storage tanks, may rise; and (4) gentle slopes may fail as liquefied soils and overlying materials move downslope.

Liquefaction potential depends on soil and ground-water conditions and the severity and duration of ground shaking. Liquefaction most commonly occurs in areas of shallow ground water (less than 30 feet [9 m]) and loose sandy soils. In general, an earthquake of Richter magnitude 5 or greater is necessary to induce liquefaction (Kuribayashi and Tatsuoka, 1975, 1977; Youd, 1977). For larger earthquakes, liquefaction has a greater likelihood of occurrence and will be found at greater distances from the epicenter. Liquefaction has been documented up to 170 miles (274 km) from the epicenter of an earthquake (1977 Romanian earthquake, magnitude 7.2) (Youd and Perkins, 1987).

Liquefaction is a hazard that can affect Tooele Valley and the WDHIA. Soil and ground-water conditions are conducive to liquefaction in both areas, although the likelihood of sufficient ground shaking is greater in Tooele Valley.

Liquefaction itself does not necessarily cause damage, but may induce damaging ground failures. Four types of ground failure commonly result from liquefaction: (1) loss of bearing strength, (2) ground oscillation, (3) lateral-spread landslides, and (4) flow landslides (Youd, 1978a, 1978b; Tinsley and others, 1985). Youd (1978a) relates these types of ground failure to the slope of the ground surface (table 4).

Table 4. *Ground slope and expected failure mode resulting from liquefaction (modified from National Research Council, 1985).*

GROUND SURFACE SLOPE	FAILURE MODE
Less than 0.5 percent	Bearing capacity
Less than 0.5 percent, liquefaction at depth	Ground oscillation
0.5 to 5.0 percent	Lateral-spread landslides
Greater than 5.0 percent	Flow landslides

Loss of bearing strength and resulting deformation of a soil mass beneath a structure are the principal effects of liquefaction in areas where slopes are generally less than about 0.5 percent (Youd, 1978a, 1984; National Research Council, 1985). Liquefaction reduces shear strength of the soil which provides foundation support, allowing structures to settle and tilt (Youd, 1984; National Research Council, 1985; figure 10).

Ground oscillation takes place when liquefaction occurs beneath the ground surface, below soil layers that do not liquefy, and where slopes are too gentle for lateral displacement to occur (Tinsley and others, 1985). Under these conditions, liquefaction at depth commonly causes overlying soil blocks to detach from each other and jostle back and forth on the liquefied layer during an earthquake (National Research Council, 1985; figure 11). The detached soil blocks vibrate differently from the underlying and surrounding firm ground, causing fissures to form and impacts to occur between oscillating blocks and adjacent firm ground (National Research Council, 1985; Tinsley and others, 1985).

Where the ground-surface slope ranges between 0.5

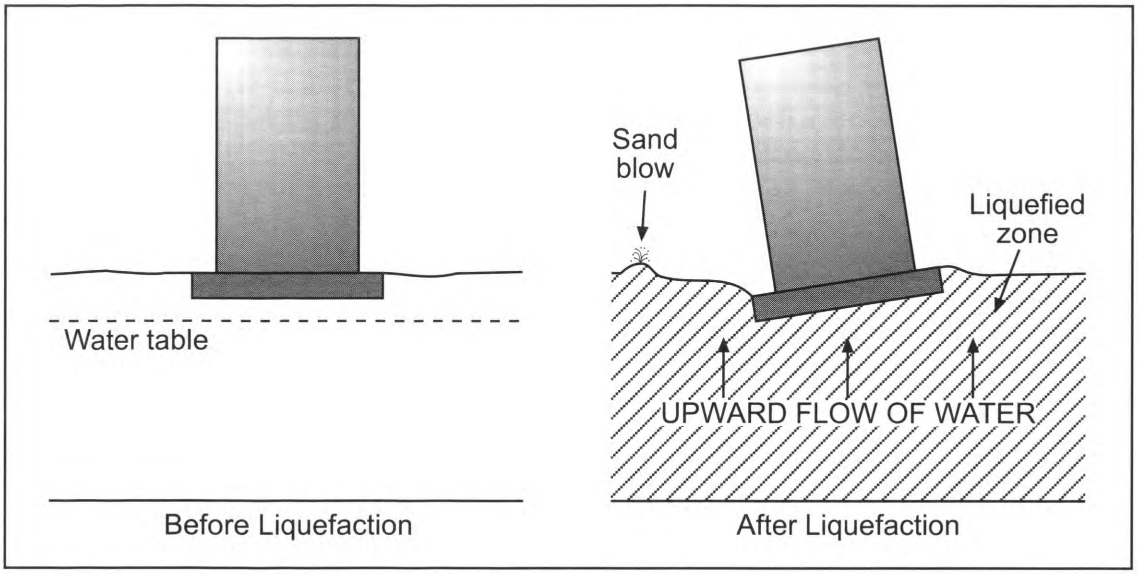


Figure 10. Tilting of a building due to liquefaction and loss of bearing strength in the underlying soil, allowing the building to settle and tilt (after Youd, 1984; National Research Council, 1985).

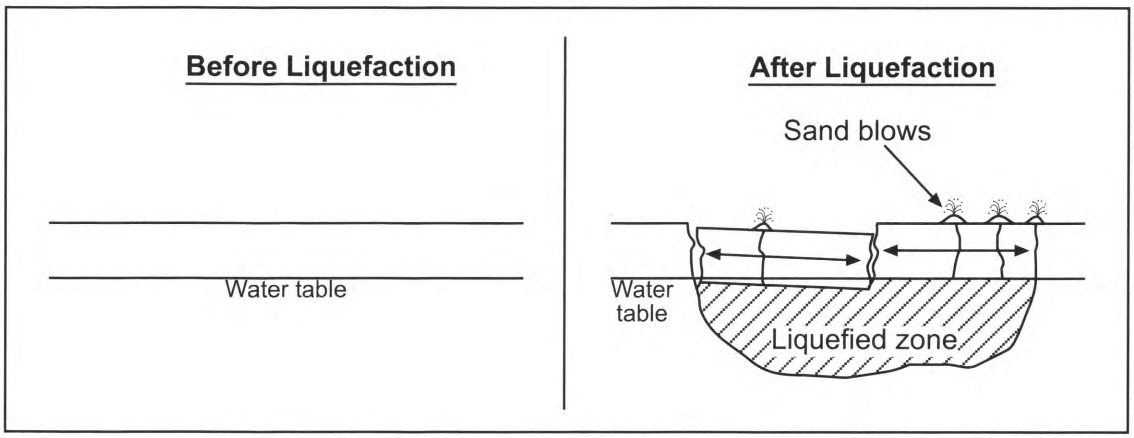


Figure 11. Liquefaction-induced ground oscillation effects. Liquefaction occurs in the crosshatched zone and causes ground settlement, opening and closing of fissures, and sand blows as the surface layer detaches from the surrounding firm ground (after Youd, 1984; National Research Council, 1985).

and 5.0 percent, failure by lateral spreading may occur (Youd, 1984; Bartlett and Youd, 1992). Lateral spreads are characterized by surficial blocks of sediment which are displaced laterally downslope as a result of liquefaction in a subsurface layer (National Research Council, 1985; figure 12). The surface layer commonly breaks up into blocks, bounded by fissures, which may tilt and settle differentially (National Research Council, 1985). The amount of lateral displacement depends on soil and ground-water conditions, slope, and the strength and duration of ground shaking (Tinsley and others, 1985).

Where ground-surface slopes are steeper than about 5.0 percent, slope failure may occur in the form of flow landslides (Youd, 1984; figure 13). Flow landslides are composed chiefly of liquefied soil or blocks of intact material riding on a liquefied layer (National Research Council, 1985). Flow landslides can cause soil masses to be displaced several miles (Tinsley and others, 1985).

Effects, Distribution, and Reduction

Earthquake-induced liquefaction and ground failures have the potential to cause damage to most types of structures. Structures that are particularly sensitive to liquefaction-induced ground failure include: buildings with shallow foundations, railway lines, highways and bridges, buried structures, dams, canals, retaining walls, shoreline structures, utility poles, and towers (National Research Council, 1985).

The expected mode of ground failure for liquefaction at a given site may be evaluated by determining the approximate ground surface slope at the site and referring to table 4. To differentiate between bearing capacity and ground oscillation failure modes in areas of less than

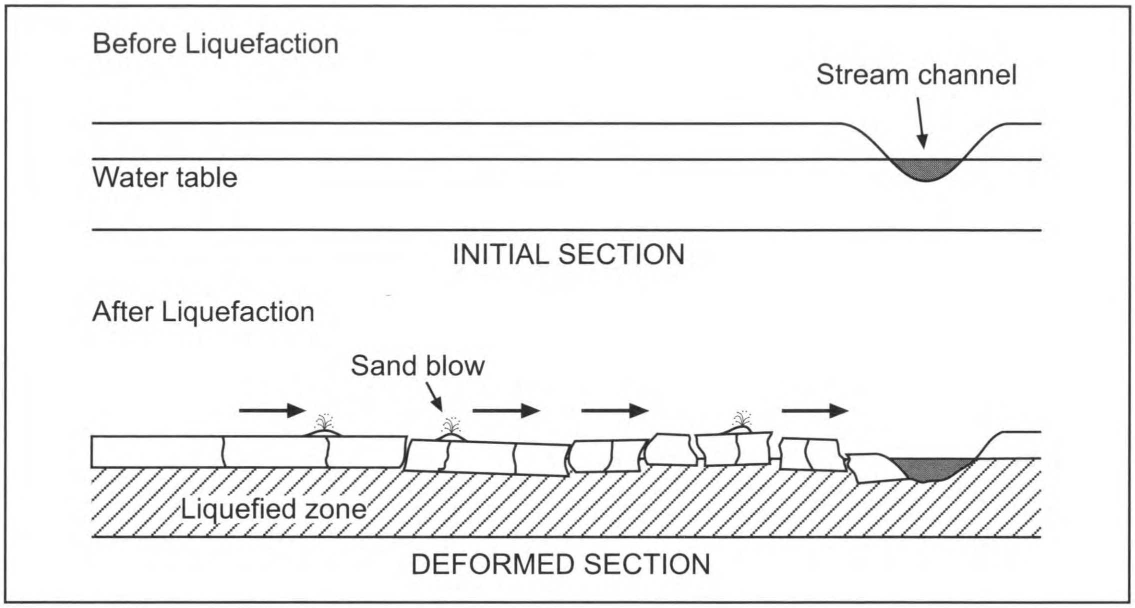


Figure 12. *Lateral spread effects. Liquefaction occurs in the crosshatched zone, causing the surface layer to detach from surrounding firm ground and move downslope (after Youd, 1984; National Research Council, 1985).*

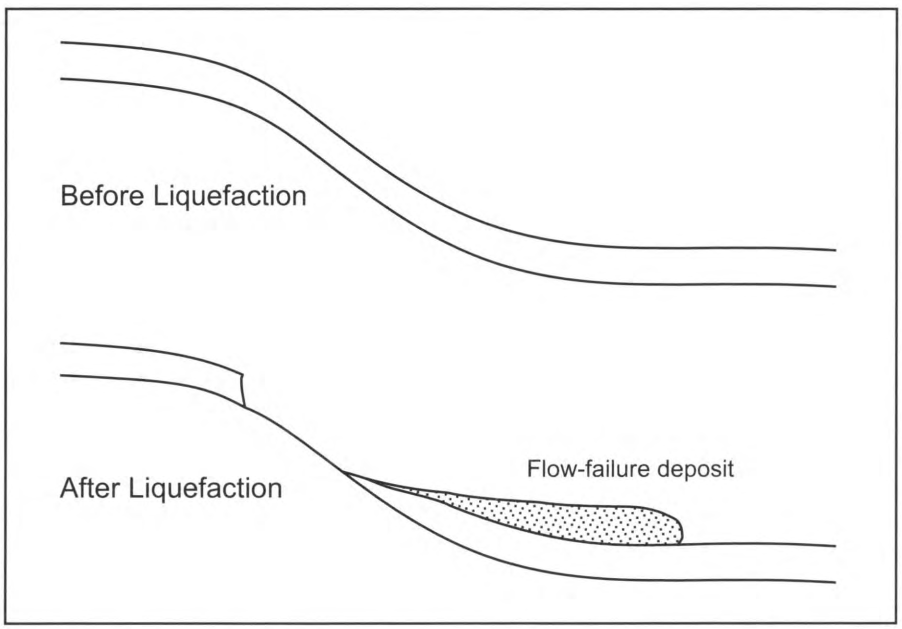


Figure 13. *Flow failure effects. Liquefaction beneath the ground surface causes a loss of shear strength, allowing the soil mass to flow down the steep slope (after Youd, 1984; National Research Council, 1985).*

0.5 percent slope, the depth to the liquefiable layer(s) at the site must be known. Ground oscillation is likely if the liquefiable layer(s) are relatively deep.

Loss of bearing strength in foundation soils causes structures to settle and/or tilt. Buoyant buried structures, such as gasoline storage or septic tanks, may also float upward in liquefied soils (Tinsley and others, 1985). Among the more spectacular examples of a bearing-capacity failure was the tilting of four 4-story buildings, some as much as 60 degrees, in the 1964 magnitude 7.3 earthquake in Niigata, Japan (National Research Council, 1985). Buried septic tanks rose by as much as 3 feet (1 m) during the same earthquake (Tinsley and others, 1985).

Ground oscillation can also cause damage to structures and buried facilities. Damage is caused by differential settlement, opening and closing of fissures, and formation of sand blows which commonly accompany the oscillations (Tinsley and others, 1985).

Lateral-spread landsliding can cause significant damage to structures (table 5) and may be especially destruc-

Table 5. Relationship between ground displacement and damage to structures (modified from Youd, 1980).

GROUND DISPLACEMENT	LEVEL OF EXPECTED DAMAGE
Less than 4 inches (0.1 m)	Little damage, repairable
4 inches (0.1 m) to 1 foot (0.3 m)	Severe damage, repairable
1 foot (0.3 m) to 2 feet (0.6 m)	Severe damage, non-repairable
More than 2 feet (0.6 m)	Collapse, non-repairable

tive to pipelines, utilities, bridge piers, and structures with shallow foundations (Tinsley and others, 1985). Lateral-spread landslides with ground displacements of only a few feet caused every major pipeline break in San Francisco during the 1906 earthquake (Youd, 1978a), and thus were indirectly responsible for the inability to control the fires that damaged the city (Tinsley and others, 1985).

Flow landslides are the most catastrophic mode of liquefaction-induced ground failure (Tinsley and others, 1985). Extensive damage due to flow landslides occurred in the cities of Seward and Valdez, Alaska, during the 1964 Alaska earthquake (Tinsley and others, 1985). A flow landslide near the Mount Olivet Cemetery during the 1906 San Francisco earthquake knocked a powerhouse off its foundation (Youd, 1973).

As originally proposed by Youd and others (1978), a liquefaction potential map is derived by superimposing a liquefaction susceptibility map and liquefaction opportunity map. Liquefaction susceptibility represents properties of near-surface earth materials, whereas liquefaction opportunity represents the seismic potential of a region. Plate 3 shows areas in Tooele Valley and the WDHIA where soil and ground-water conditions may be conducive to liquefaction. Although the probability of earthquake ground shaking sufficient to cause liquefaction was not considered, a significant potential exists for liquefaction-induced ground failure to cause severe damage in areas of high susceptibility in Tooele Valley (Mabey and Youd, 1989). Because of a lesser earthquake potential, the hazard is substantially lower in the WDHIA.

Liquefaction susceptibility on plate 3 was determined primarily from geologic and ground-water data. In areas that may have sediments susceptible to liquefaction where the depth to ground water is less than 50 feet (15 m), susceptibility was mapped as: (1) high, if the depth to ground water is less than 10 feet (3 m); (2) moderate, if the depth to ground water is from 10 to 30 feet (3-9 m); or (3) low, if the depth to ground water was from 30 to 50 feet (9-15 m). Areas with a very low liquefaction susceptibility do not have susceptible sediments, or have ground-water depths greater than 50 feet (15 m). Seasonal and long-term fluctuations in ground-water levels can affect the susceptibility at a given site. Plate 3 is also at a regional scale and, although it can be used to gain an understanding of the susceptibility of a given area for liquefaction-induced ground failure, it was not designed to replace site-specific evaluations. Mapped areas classified with a particular liquefaction susceptibility may contain isolated areas with other classifications, and site-specific geotechnical studies are still recommended.

Areas of moderate to high liquefaction susceptibility need not be avoided; structural measures and site modification techniques are available to reduce hazards. The cost of reducing liquefaction hazards may be high relative to the value of the structure for single-family dwellings, and liquefaction is generally not a life-threatening hazard in such structures. However, hazard reduction may be recommended for larger critical facilities (Anderson and others, 1987).

The National Research Council (1985) identifies several alternative approaches for existing structures threatened by earthquake-induced liquefaction. The choices include: (1) retrofitting the structure and/or site to reduce the potential for liquefaction-induced damage; (2) abandoning the structure if the retrofit costs exceed potential benefits derived from maintaining the structure; or (3) accepting the risk.

Possible actions that may be taken if a liquefaction hazard exists at the site of a proposed structure include: (1) improving site conditions to lower the liquefaction potential; (2) designing the structure to withstand liquefaction effects; (3) avoiding the risk by moving the proposed development to a less hazardous site; (4) insuring the development so that if liquefaction-induced damage occurs, funds will be available to repair the damage; or (5) accepting the risk if the liquefaction potential and consequences are clearly understood.

Structural solutions to reduce the effects of liquefaction for buildings include using end-bearing piles, caissons, or fully compensated mat foundations, designed for the predicted liquefaction phenomena at the site (National Research Council, 1985). Methods of improving liquefiable soil-foundation conditions are: (1) densification of soils through vibration or compaction, (2) grouting, (3) dewatering with drains or wells, and (4) loading or buttressing to increase confining pressures (National Research Council, 1985). Costs of site improvement techniques range from less than $0.50 to more than $500.00 per cubic yard (0.76 m^3) of soil- foundation material treated (National Research Council, 1985).

Other Earthquake Hazards

A variety of phenonema that can damage property and/or threaten lives may accompany earthquakes. The principal hazards are addressed elsewhere in this report, such as surface fault rupture, ground shaking, liquefaction, tectonic subsidence, landslide, and rock falls. Other potentially damaging phenonema associated with earthquakes include: (1) ground failure due to loss of strength in sensitive clays, (2) subsidence in granular materials due to ground shaking, (3) flooding caused by seiches in Great Salt Lake, (4) flooding caused by surface drainage disruptions, and (5) flooding caused by increased ground-water discharge.

Ground Failure Due to Loss of Strength in Sensitive Clays

Most clays lose strength when disturbed; sensitive clays experience a particularly large loss of strength. Sensitive clays are wet clays whose undisturbed shear strength is lost abruptly following a shock or disturbance (Parry, 1974). The sensitivity of clays is defined as the ratio of shear strength in an undisturbed condition to shear strength after being severely disturbed (Costa and Baker, 1981). Rosenqvist (1953, 1966) proposes that these clays originate as platy clay particles deposited in an edge-to-edge "house of cards" (flocculated) structure in saline environments, in which sodium and other cations in water provide bonding strength. Later, when this saline water is leached out by fresh ground water, the clays are left in an unstable arrangement subject to collapse or liquefaction when disturbed or shaken. One triggering mechanism for ground failure is ground shaking generated by earthquakes. During and after disturbance, the clays may revert from a flocculated soil structure in which ground water fills the interstitial pore spaces, to a dispersed soil structure in which the interstitial water is expelled, liquefying the clay (Costa and Baker, 1981).

The potential for ground failure in sensitive clays is related to the intensity and duration of ground shaking, and sensitivity of the clays. Clays with high sensitivities (ratio of undisturbed shear strength to disturbed shear strength of 10 or more) may be prone to failure during earthquake-induced ground shaking (Earthquake Engineering Research Institute, 1986). The existence and distribution of such clays, as well as the intensity and duration of ground shaking needed to induce failure, have not been investigated in Tooele Valley and the WDHIA and are unknown. Maps which show their extent have not been produced and the probability of this type of failure has not been determined. Fine-grained lake sediments underlie much of both areas, deposited by lakes occupying the Great Salt Lake Basin during the last 15 million years (Currey and others, 1984b). Many of these lake sediments are silicate clays, some of which have been classified as sensitive in the Wasatch Front area (Parry, 1974). Assessment of this hazard should be undertaken at the site-specific level, as part of a standard geotechnical investigation, in areas where the depth to shallow ground water is less than 30 feet (9 m)

The principal effect of disturbance of sensitive clays is ground failure. The kinds of ground failure associated with sensitive clays are similar to those accompanying liquefaction, including flow failures, slump-type landslides, and lateral-spread or translational landslides (Costa and Baker, 1981; Earthquake Engineering Research Institute, 1986). Liquefied sensitive clays may flow downhill on slopes as low as 2 percent or less (Costa and Baker, 1981). The most devastating damage resulting from the 1964 Alaska earthquake (magnitude 8.6) was due to translational landslides partly from failure of sensitive clays. The largest of these landslides damaged 75 homes in the Turnagain Heights residential area in Anchorage (Hansen, 1966).

Ground failure due to sensitive clays has the potential to cause damage to most types of structures. Possible actions which may be taken if sensitive clays are present include: (1) improving site conditions by converting the clays from a flocculated soil structure to a dispersed structure using preconstruction vibration techniques, and/or dewatering the site; and (2) designing the structure to withstand the potential effects of ground failure using structural solutions such as end-bearing piles placed below the sensitive clays, caissons, or fully compensated mat foundations designed for the anticipated failure type.

Subsidence in Granular Materials Caused by Ground Shaking

Loose granular materials such as sand and gravel may be prone to subsidence when shaken. Earthquake ground shaking can effectively compact these materials as individual particles move closer together. This rearrangement decreases the volume of the material, causing subsidence. During the 1964 Alaska earthquake, ground shaking caused as much as 5.9 feet (1.8 m) of such subsidence at some locations (Costa and Baker, 1981).

Differential settlement can occur in deposits that are susceptible to vibratory subsidence. This may result in building damage or foundation cracking as one part of a foundation settles more than another (Costa and Baker, 1981). Structural failure of building members may also be caused by excessive settlement (Dunn and others, 1980). Even minor differential settlement can cause extensive damage to earthen-fill structures such as railway embankments, highway foundations, bridge abutments, and dikes and levees. Buried utility lines and connections may also be severed by settlement. The rate of subsidence is an important factor that must be considered in evaluating the potential for damage (Dunn and others, 1980). Subsidence due to earthquake ground shaking would be virtually instantaneous.

Maps delineating areas susceptible to vibratory subsidence in granular soils have not been prepared for Tooele Valley and the WDHIA, and the extent of soils subject to subsidence is unknown. However, areas of Tooele Valley and the WDHIA are underlain by deposits that may be prone to vibratory subsidence, such as clean sand and gravel deposited in Pleistocene Lake Bonneville. If not adequately compacted during placement, artificial fill may also be susceptible to vibratory subsidence (Schmidt, 1986).

Levels of ground shaking necessary for subsidence vary with conditions, and assessment of this hazard must be undertaken on a site-specific basis as part of geotechnical investigations. Standard penetration and cone penetrometer tests are commonly used to evaluate the potential for subsidence (Dunn and others, 1980). The potential for subsidence should be considered during soil-foundation investigations for all major construction, especially for critical facilities.

Structural methods to reduce settlement damage include supporting structures on piles, piers, caissons, or walls founded below the susceptible material (U.S. Bureau of Reclamation, 1985). Where structural measures to reduce settlement in granular soils are not possible, other actions to reduce the hazard include: (1) improving site conditions by removing or compacting in-place granular materials prior to construction, and (2) properly engineering and compacting fill materials.

Flooding Caused by Seiches in Great Salt Lake

Oscillations in the surface of a landlocked body of water can produce unusually large waves, or seiches, similar to oscillations produced by sloshing water in a bowl when shaken or jarred (Nichols and Buchanan-Banks, 1974). Seiches may be generated by wind, landslides, and/or earthquake effects such as ground shaking or surface fault rupture. The magnitude of seiches caused by landslides or surface fault rupture depends on the amount of water and ground displacement. For wind- and ground-shaking-induced seiches, the magnitude is determined by the degree of resonance between the water body and the periodic driving force. The magnitude is greatest when this driving force is oscillating at the same frequency at which the body of water naturally oscillates (Costa and Baker, 1981). A lake's natural oscillation period is determined by parameters such as water depth, lake size and shape, and shoreline configuration, much as the natural frequency of a pendulum is determined by its physical characteristics (Lin and Wang, 1978).

Studies of wind seiches in Great Salt Lake conclude that the maximum wave amplitude is expected to be about 2 feet (0.6 m) (Lin and Wang, 1978); no systematic or theoretical studies of landslide or earthquake-induced seiches have been made. However, seiches were reported along the southern shoreline of Great Salt Lake at Saltair and the Lucin trestle during the 1909 Hansel Valley earthquake (magnitude 6) (Williams and Tapper, 1953). The elevation of the lake was 4202.0 feet (1280.7 m) at this time (U.S. Geological Survey lake elevation records). The seiche generated by this earthquake overtopped the Lucin cutoff railroad trestle at an elevation of 4214.85 feet (1284.69 m) (Southern Pacific Transportation Company records). Assuming lake and trestle elevation records and reports of the seiche are accurate, the seiche was more than 12 feet (3.7 m) high (Lowe, 1993).

Studies from other areas have shown that seiches may raise or lower a water surface from a few inches to several yards (Blair and Spangle, 1979). Seiches may cause damage from flooding and erosion in areas around the margins of lakes, and are a hazard to shoreline development, dams, and in-lake structures. The principal area at risk in Tooele County is along the shore of Great Salt Lake.

Seiches should be taken into consideration when planning development in Great Salt Lake and within the proposed Great Salt Lake Beneficial Development Area

(Utah Division of Comprehensive Emergency Management, 1985). Because no comprehensive studies have been completed for Great Salt Lake, maps have not been produced that show the likely area to be affected by seiches in Tooele Valley. Accounts of the seiche generated by the 1909 Hansel Valley earthquake suggest that maximum wave amplitudes generated by earthquakes may far exceed those associated with wind, and that areas above 4,217 feet (1,285 m) could be affected by seiches during high lake levels.

Dikes protected against erosion on the lakeward side and engineered breakwaters can be used to protect development or dissipate wave energy. Shoreline buildings can also be floodproofed, elevated, and constructed or reinforced to withstand the lateral forces of seiches (Costa and Baker, 1981).

Flooding Due to Surface-Drainage Disruptions

During earthquakes, ground shaking, surface fault rupture, ground tilting, and landsliding can cause flooding if water tanks, reservoirs, pipelines, or aqueducts are ruptured, or if stream courses are blocked or diverted. Areas where such flooding may occur can be predicted to some extent by defining known active faults, active landslides, and potentially unstable slopes. Damming of streams by landslides can cause upstream inundation and, if the dam subsequently fails, cause catastrophic downstream flooding (Schuster, 1987).

Site-specific studies that address earthquake and slope-failure hazards should be completed prior to construction for all major water-retention structures or conveyance systems so that hazard-reduction measures can be recommended. To prepare for water-system breaks, shut-off valves and emergency response/repair plans should be in place. For existing facilities, studies can evaluate the possible locations and extent of flooding and recommend drainage modifications to prevent floods or divert flood waters. Potential flooding from diversion of stream courses is more difficult to evaluate, but should be considered in hazards evaluations for critical facilities.

Increased Ground-Water Discharge

The effects of earthquakes on ground-water systems have not been extensively studied and, consequently, are not well understood. Increases in spring flow and expulsion of water from shallow bedrock aquifers caused surface flooding during the 1983 Borah Peak, Idaho, earthquake. Stream flow increased by more than 100 percent following the earthquake, and flow remained high for two weeks before declining to nearly original levels (Whitehead, 1985). Although this earthquake appeared to cause a more profound effect on ground water than other earthquakes, similar effects may occur during large-magnitude earthquakes in Tooele Valley. Flooding from increased spring flow in mountain drainages will be confined to stream channels, and adherence to Federal Emergency Management Agency floodplain regulations should effectively reduce the risk. However, increased spring flow on valley floors could result in ponded water and basement flooding.

Landslides

Introduction

Landslides are downslope movements of rock or soil under the influence of gravity, including both deep-seated and shallow slope failures. Deep-seated slope failures have surfaces of rupture generally greater than 10 feet (3 m) deep, and include rotational and translational slides and associated earth flows (Varnes, 1978; Cruden and Varnes, 1996). This portion of the text addresses the landslide hazard posed by deep-seated slope failures. Rock falls and shallow earth movements (surface of rupture generally less than 10 feet [3.0 m] deep), such as debris flows, are addressed elsewhere in this report.

Landslides may be caused by oversteepening of slopes, loss of lateral support, weighting of the head, and increased pore pressure (static conditions). In addition, landslides may also be induced by earthquakes (dynamic conditions). Older slope failures may reactivate due to conditions in the landslide such as increased permeability and established surfaces of rupture. Landslides can damage buildings, transportation routes, and utility lines by displacement of the ground, and cause flooding due to discharge of springs and damming of streams.

Because of the predominance of relatively slide-resistant rock, deep-seated landsliding historically has caused little damage in Tooele Valley. The landslide hazard is greatest in the Oquirrh and Stansbury Mountains at the southern end of Tooele Valley, where the slide-prone Mississippian-age Manning Canyon Shale crops out or lies just below the ground surface. No significant landslide hazard exists in the WDHIA. The areas most susceptible to dynamic landsliding are generally those areas most susceptible to static (non-earthquake) landsliding.

Two types of landslides are common. Rotational slides generally have a curved surface of rupture and are called slumps. The head of a slump is back-tilted compared to the original slope. Many slumps include an earth flow at the toe where material moves onto the land surface below the slump (figure 14). Translational slides

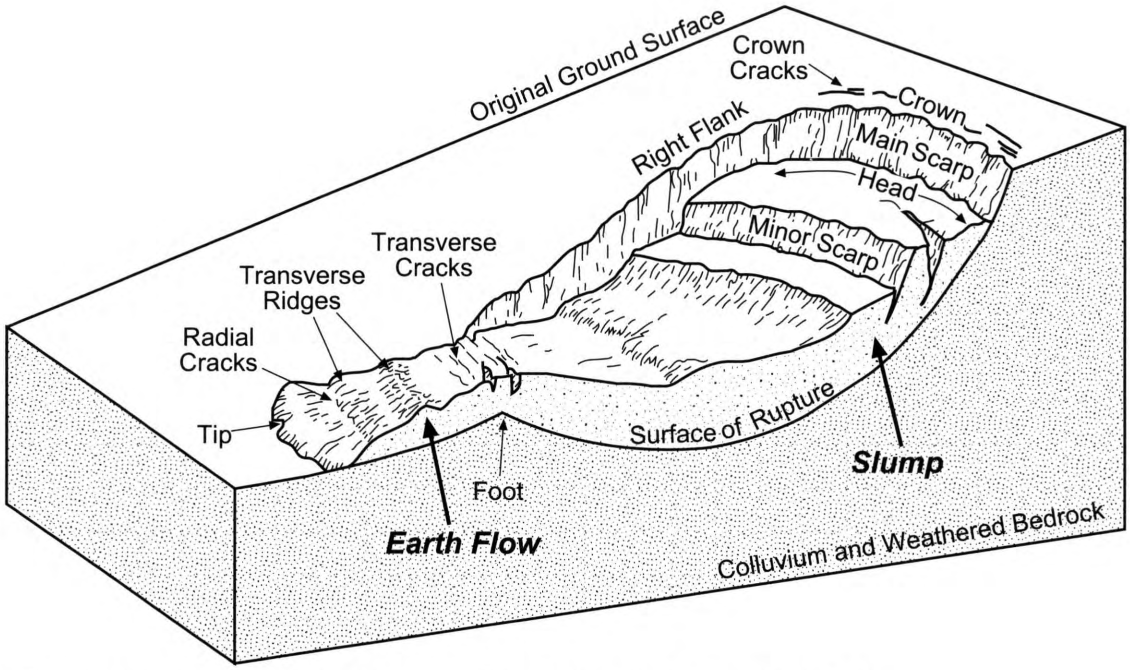


Figure 14. *Block diagram of a rotational landslide (slump) and earth flow (modified from Varnes, 1978).*

generally have a planar surface of rupture, and may be broken into discrete blocks. Slumps and translational slides may move slowly and progressively over periods of years, or rapidly in a matter of a few seconds.

Landslides may occur in Tooele Valley in a moderate to strong earthquake. Slopes considered unstable under static conditions will be even less stable during an earthquake, and some slopes that are stable under static conditions may also fail as a result of earthquake ground shaking, particularly if wet. Most landslides caused by earthquakes are new slope failures, not reactivated older landslides (Keefer, 1984). Deep-seated slumps and translational slides commonly accompany earthquakes with Richter magnitudes greater than 4.5 (Keefer, 1984). Earthquakes have produced slope failures (predominantly rock falls and rock slides) in Utah in historical time (Keaton and others, 1987). One such earthquake, the September 2, 1992 magnitude 5.8 St. George earthquake, caused a large destructive translational landslide near Springdale, Utah, 27 miles (44 km) from the epicenter (Black and others, 1995; Jibson and Harp, 1995). Earthquakes of magnitude 7.0 could cause deep-seated slope failures as far as 100 miles (161 km) from the epicenter (Keefer, 1984).

Landslides are also likely in years of above-average precipitation, such as during the wet years of the 1980s (1982-1986). However, few deep-seated landslides occurred during this time in Tooele Valley because most rock and sediment in the valley and surrounding mountains is not susceptible to landsliding.

Factors such as slope steepness, precipitation, ground-water regime, and bedrock structure are important in determining landslide susceptibility, but the most important factor is rock type. Rock units containing low-strength, moisture-sensitive shale or clay are usually the most susceptible to landsliding. Landslides are not numerous in the mountains in the WDHIA or those surrounding Tooele Valley, and only one geologic unit, the Mississippian-age, clay-rich Manning Canyon Shale, is particularly susceptible to landsliding. The general lack of landslides in the northern Stansbury and Oquirrh Mountains, and in the Cedar Mountains, is due mainly to a lack of susceptible geologic units (Harty, 1990).

The Manning Canyon Shale has been involved in many damaging landslides in northern Utah, particularly in the foothill slopes of the Wasatch Range in and east of Provo (Harty, 1991). In the Stansbury Mountains, the Manning Canyon Shale mainly crops out south of South Willow Canyon (Sorenson, 1982), and from Magpie Canyon north to about Miners Canyon (Rigby, 1958; Sorenson, 1982). In 1983, a large landslide in Manning Canyon Shale occurred at the confluence of Morgan and East Hickman Canyons about 2.5 miles (4.0 km) south of the study area boundary. The slide took out the East Hickman Canyon road, which was restored in 1990 (Paul Dart, Range Technician, U.S. Forest Service, verbal communication, December, 1990). In the Oquirrh Mountains, the Manning Canyon Shale crops out in the southeastern part of the study area, in Soldier Canyon. Here, it was involved in the large, middle-to-early Holocene-age Soldier Canyon landslide about 3.5 miles (5.6 km) up the canyon on its southern flank.

Landslides may also occur in the unconsolidated sediments of Pleistocene Lake Bonneville. Many landslides along the Wasatch Front have occurred in Lake Bonneville sediments, especially in the highest shoreline and delta deposits that typically form steep slopes. However, we identified no landslides in unconsolidated sediments in Tooele Valley and the WDHIA.

Effects, Distribution, and Reduction

Landslide movement may be preceded by cracks at the landslide head and a bulge at the toe. Damage from a landslide can occur either on or adjacent to the slide mass. The top of most landslides is characterized by an arcuate downslope-facing scarp (main scarp) created by downward displacement (figure 14). A building that straddles the main scarp loses foundation support and may collapse. Structures upslope from the landslide head are at risk because the newly formed main scarp is commonly unstable and may fail retrogressively, forming new scarps upslope. Buildings within the central mass of the landslide may experience differential displacement on minor scarps and movement in both vertical and horizontal directions. The toe of a landslide will normally move horizontally and upward and may proceed downslope causing extensive damage. Table 5 shows the relationship between ground displacement and expected levels of damage to structures.

Plate 2 shows landslide susceptibility of natural slopes under static (non-earthquake) conditions for Tooele Valley and the WDHIA. We estimated landslide susceptibility from geologic units, slope steepness, and the presence of existing landslides. We used four landslide-susceptibility categories: high, moderate, low, and very low. Plate 2 thus provides a general indication of where landslide hazards may exist. However, this map is at a regional scale and, although it can be used to gain an understanding of the potential for landslides occurring in a given area, it is not designed to replace site-specific evaluations. Areas we mapped as having high or moderate landslide susceptibilities may contain areas that are not prone to landsliding, even during earthquake ground shaking, and areas in the low or very low hazard category may contain areas that are susceptible to landsliding.

Included on plate 2 are existing landslides in Tooele Valley determined from geologic maps and aerial photographs; we found no existing landslides in the WDHIA. The Soldier Canyon landslide, originally mapped by Tooker and Roberts (1988a), is the only major slump-type failure identified in the Tooele Valley area. The Bear Trap Flat area in Settlement Canyon was mapped as a possible landslide by Colton (1988); however, subdued topography and heavy forest cover make assessment difficult. Two large rock slides in upper Settlement Canyon are shown on both the landslide map (plate 2) and debris-slide/flow map (plate 4) because the slopes on which they occurred may be subject to different types of failure.

Slopes included in the high-susceptibility category are slopes on or adjacent to existing landslides. Existing landslides pose a particular problem for development because they may reactivate. The only areas where a high hazard was assigned is in the vicinity of the Soldier Canyon landslide and on the rock-slide slopes in Settlement Canyon. A high hazard was not assigned to the Bear Trap Flat area because of the uncertainties discussed above.

The moderate-susceptibility category includes slopes greater than 15 percent (9 degrees) that also meet one of the following criteria: (1) slopes underlain by slide-prone material; (2) slopes composed of unconsolidated Lake Bonneville sediments; or (3) slopes that show evidence of sloughing, such as those along some stream-channel banks. In eastern Tooele Valley, most of the moderate designations are slopes in Lake Bonneville deposits.

Low-susceptibility areas include slopes that are equal to or greater than 15 percent, and underlain by slide-resistant material. Most slopes in the Oquirrh and Stansbury Mountains, and in the Grayback Hills, are in this hazard class. Unlike debris slides and flows, which commonly occur on steep slopes, deep-seated landslides such as slumps also occur on moderate slopes; many deep-seated landslides in Utah have initiated on slopes of about 15 percent. A statewide survey shows that the lower limits of slope for rotational slumps range from 7 to 18 degrees (12-32.5 percent), and the lower limits for earth flows range from 4 to 20 degrees (7-36 percent) (Sidle and others, 1985). The lower limit of 15 percent (9 degrees) is a conservative choice for the hazard maps. Landslide susceptibility is designated "very low" where slopes are less than 15 percent. Most of Tooele Valley and the WDHIA is in this category.

In areas where potentially unstable slopes are bounded by flat, stable surfaces, landslide-susceptibility boundaries extend beyond the base and top of the unstable slope. Such areas found along the stream banks of Settlement and Middle Canyons, and at the Stockton Bar, where potential instability in the steep portion of the slope may affect areas both above and below. The width of the susceptiblity zones in these areas depends on the height, steepness, ground-water conditions, and strength of the material underlying the slope. In these areas, an estimated stable slope angle through the center of the steep slope was taken to determine the area potentially affected. We used estimated slope angles of 2 horizontal to 1 vertical (2:1; 50 percent) for dry granular soils, and 2.5:1 (40 percent) for moist fine-grained material.

Many methods have been developed for reducing landslide hazards. Proper planning or avoidance are made possible if slide-prone areas are identified early in the planning process. Where avoidance is not feasible, various engineering techniques are available to stabilize slopes. Care in site grading, with proper compaction of fills and engineering of cut slopes, is necessary for hillside development. De-watering (draining) can stabilize slopes and existing landslides. Retaining structures built at the toe of a landslide may help stabilize the slide and reduce the possibility of smaller landslides. In some cases, piles may be driven through the landslide mass into stable material beneath the slide. If the dimensions of the landslide are known, and the landslide is not excessively large, removing the landslide may be effective. Diversion of drainage away from a slide reduces the destabilizing effects of infiltrating ground water. Other techniques used to reduce landslide hazards include bridging, weighting, or buttressing slopes with compacted earth fills. A more complete list of landslide-hazard-reduction techniques can be found in Costa and Baker (1981) and Kockelman (1986). Chapter 23 (appendix) of the Uniform Building Code (UBC) (International Conference of Building Officials, 1997) includes specifications for site grading and slope design.

Alluvial-Fan Flooding

Introduction

The continuum of debris slides, debris flows, debris floods, and stream flooding on alluvial fans is termed "alluvial-fan flooding" (National Research Council, 1996). A wide range in debris deposition and flooding types is found on poorly defined alluvial landforms in Tooele Valley and the WDHIA. Tooele Valley is susceptible to debris flows, debris floods, and stream flooding from the steep mountains that border the valley. Susceptibility to these hazards is lower in the WDHIA because of subdued topography. Unintentional releases of water from dam failures may also cause flooding in Tooele Valley.

Debris slides, debris flows, and debris floods consist of mixtures of soil, rock, water, and organic material, whereas stream floods are primarily water. These slides, flows, and floods can present a hazard to life and property as they move downslope. Debris slides are generally shallow slope failures, with surfaces of rupture less than about 10 feet (3 m) deep. They form on steep slopes and usually lack sufficient water (less than 10-30 percent by weight) to travel far from their source areas. Debris slides thus present a hazard primarily on and adjacent to steep slopes, such as mountainous areas in and adjacent to valleys. Debris flows are a muddy slurry (70-90 percent solids by weight; Costa, 1984) much like wet concrete, that flow downslope usually in surges or pulses. They generally are confined to slopes and stream channels in mountains, but may deposit debris over large areas on alluvial fans at and beyond canyon mouths. Debris floods, also called hyperconcentrated floods, are mixtures of soil, organic material, and rock debris that are transported by fast-moving flood waters (Wieczorek and others, 1983). Solids account for 40 to 70 percent of the material by weight (Costa, 1984). Like debris flows, debris floods can transport material great distances from their source areas. Stream floods are mostly water, and normally contain less than 40 percent solids by weight (Costa, 1984). Stream floods occur when the stage or height of water exceeds some given datum, such as the banks of the normal stream channel (Costa and Baker, 1981). Dam-failure floods are similar, and are caused by unintentional releases of impounded water.

Debris slides, flows, and floods, and normal stream flow form a continuum of sediment/water mixtures that grade into each other with changes in the relative proportion of sediment to water, and stream gradient (Pierson and Costa, 1987). Debris flows and debris floods present a greater hazard to valley areas than debris slides. Deposition of sediment transported by debris flows and debris floods may take place on alluvial fans at and beyond canyon mouths. Deposition on alluvial fans is caused by the decrease in channel gradient and increase in channel width, resulting in a decrease in depth and velocity of flow and an increase in internal friction of the flowing debris as the stream leaves its constricted channel and enters the main valley floor (Jochim, 1986).

Debris flows can form in at least two different ways. In the Oquirrh and Stansbury Mountains, where cloudburst rainstorms are common, overland flow and flood waters can scour materials from the ground surface and stream channels, thereby increasing the proportion of soil materials to water until the mixture becomes a debris flow (Wieczorek and others, 1983). The size and frequency of debris flows generated by rainfall are dependent upon several factors including the amount of loose material available for transport, the magnitude and frequency of the storms, the density and type of vegetative cover, and the moisture content of the soil (Campbell, 1975; Pack, 1985; Wieczorek, 1987). Debris flows can also mobilize directly from debris slides once the slide reaches a stream, or when the water content in the slide increases by some other means until sufficient to permit flow. As the relative proportion of water to sediment increases with either the addition of water or removal of sediment by deposition, debris flows become debris floods. Debris floods can also originate through progressive incorporation of materials into flood waters (Waitt and others, 1983; Wieczorek and others, 1983).

Many of the debris slides and flows that occurred in the Oquirrh Mountains in 1983-1984 were generated by rapid melting of an unusually thick snowpack. Of the 104 debris slides and flows identified, over 70 percent occurred on south-facing slopes. The high percentage on south-facing slopes was due in part to weather conditions. During the winters of 1983 and 1984, the greater-than-average snowpack was preserved by cool early-spring temperatures (Wieczorek and others, 1989). The more intense solar radiation received by south-facing slopes, combined with sudden, sustained high temperatures in late spring, caused rapid melting of the snowpack. Kaliser and Slosson (1988) report that landslides occurring in 1983 generally followed the melting snowline, generating debris slides and flows at progressively higher elevations. Infiltration of meltwater into porous colluvium on steep mountain slopes probably exceeded the rate of drainage into the underlying bedrock, causing a rapid rise in pore-water pressure in the colluvium, resulting in loss of frictional resistance and sudden failure of the shallow colluvial layer.

Pore-water pressure in colluvium may increase with draining of bedrock aquifers into the colluvium. Mathewson and others (1990) found evidence of this in Davis County by observing sustained spring flow from debris-slide scars. We are uncertain whether such flow occurred following debris slides in the Oquirrh Mountains. However, because south-facing slopes in the Oquirrh Mountains produced more than twice the number of shallow failures than north-facing slopes, accelerated snowmelt on southern slopes was likely the dominant process creating debris slides.

Stream floods may be caused by direct precipitation, melting snow, or a combination of both. In Tooele Valley, floods are most common in April through June during spring snowmelt. High flows are sustained from a few days to several weeks (FEMA, 1989a). Snowmelt floods are somewhat predictable because flood levels depend on the volume of snow in the mountains and the rate of temperature increase. Localized cloudburst storms centered over the mountains are also effective in causing floods. These storms typically last from a few minutes to several hours, and generally occur between mid-April and September.

The flooding potential of cloudburst rainstorms is dependent upon many factors including: (1) the rate of rainfall, (2) the duration of rainfall, (3) the distribution of rainfall and direction of storm movement, (4) soil characteristics, (5) antecedent soil-moisture conditions, (6) vegetation conditions, (7) topography, and (8) drainage pattern. Because many of these conditions are generally not known until rain is actually falling on critical areas, the magnitude of flooding from a given cloudburst storm is difficult to predict. Summer cloudburst floods account for localized but often very destructive flooding and can occur with little warning. Tooele Valley communities have experienced many cloudburst floods in historical times; those occurring between 1850 and 1969 are shown in table 6.

Table 6. *Historical cloudburst floods, Tooele Valley, 1850-1969 (Woolley, 1946; Butler and Marsell, 1972).*

CITY	YEAR
Grantsville	1881, 1887, 1913, 1930 (2), 1955, 1961
Erda	1957
Lake Point	1927
Tooele	1881, 1934 (2), 1954 (2), 1957, 1961, 1963, 1965, 1967, 1968, 1969
Stockton	1936

Flooding can also result from the failure of dams, and may occur with little warning. The severity of flooding depends on the size of the reservoir and the extent of failure. The term dam failure includes all unintentional releases of water from the dam, including complete failure and release of all impounded water (Harty and Christenson, 1988). Only eight of 33 dam failures documented in Utah prior to 1984 were complete failures; most were due to overtopping and/or erosion around spillways and outlets during flood events (Harty and Christenson, 1988). Although dam failures have many causes, the most common cause is structural and foundation failures resulting from piping (Dewsnup, 1987). Uncontrolled release of water over the spillway caused repeated flooding from Settlement Canyon Reservoir (located roughly 1 mile [1.6 km] south of Tooele) in 1983 and 1984. Most historical dam failures in Utah have been small dams in rural areas; larger dams are less prone to failure because of more rigorous design, construction, and inspection practices (Harty and Christenson, 1988). Earthquake-induced ground shaking, liquefaction, landslides, and seiches (flood waves) may occur in Tooele Valley and could cause a dam failure. Settlement Canyon Reservoir is the only dam found in the study areas, and this dam has a high flood-hazard rating (unpublished Utah Division of Water Rights data).

Effects, Distribution, and Reduction

Debris flows, debris floods, and stream flooding have occurred in Tooele Valley during historical time and have caused significant damage to engineered structures and property. Early accounts usually did not distinguish debris flows or debris floods from clear-water stream floods, making it difficult to separate these events. From 1881 to 1969, 12 cloudburst floods affecting the city of Tooele were reported in local newspapers (Woolley, 1946; Butler and Marsell, 1972). At least five of these events deposited debris on roads, or in ditches and houses. Of the seven cloudburst storms reported to have affected Grantsville during this period, three deposited debris. In late July, 1887, a severe rainstorm in the Stansbury Mountains generated a debris flow that covered 0.5 acres (0.2 ha) of crop land in Grantsville to a depth of 2.5 feet (0.8 m) (Deseret News, July 28, 1887, in Woolley, 1946).

Flooding from spring snowmelt and summer cloudburst rainstorms may also contribute to dam-failure flooding. Federal Emergency Management Agency (FEMA) (1989a) and the local newspaper (Tooele Transcript-Bulletin, June 14, 1983) report that major flooding occurred in Tooele City during the spring of 1983, when snowmelt runoff from an above-average snowpack rapidly filled the Settlement Canyon Reservoir, causing an uncontrolled release of water over the spillway.

Snowmelt flooding caused about $4.5 million in damage in Tooele County during the abnormally high precipitation years of 1983 and 1984 (FEMA, in Transcript-Bulletin, July 24, 1984). Most of the major canyons in the Oquirrh and Stansbury Mountains, including Middle, Settlement, Soldier, North Willow, and South Willow Canyons carried floodwaters onto farm and grazing land, and into populated areas. In 1983, stream inflow exceeded that which could be safely released from the Settlement Canyon Reservoir, and on May 30th, the overflow outlet began releasing floodwaters into Tooele Valley. In May 1984, Settlement Canyon Reservoir again released floodwaters. During both events, floodwaters inundated streets in Tooele City, and house and property damage occurred when floodwaters breached a dike (Tooele Transcript-Bulletin, May 31, 1983; May 15, 1984). Major damage caused by the flooding included road destruction in Middle, Settlement, and Soldier Canyons; rupture of the main culinary water line in Middle Canyon; deposition of sediment on farmland; and inundation of roads, farm and grazing land, residential property, and houses in Stockton, Erda, Grantsville, Tooele, and surrounding areas.

Many debris slides occurred in the Oquirrh Mountains during the spring and summer of 1983 and 1984. Although most of the damage sustained in Tooele Valley during these years was related to stream flooding, a number of debris flows and debris floods also caused damage. A rainstorm on July 31, 1983 generated a debris flow about 7 miles (11.3 km) up Settlement Canyon that buried a large part of the canyon road (Tooele Transcript-Bulletin, August 2 and 9, 1983). Kaliser (1989) reports that a debris flow or debris flood that occurred sometime between July 31 and August 19, 1983 destroyed four sections of a main culinary water line in Soldier Canyon. On May 14, 1984, a series of debris flows and floods from an unnamed tributary channel in Settlement Canyon trapped three men in the canyon for seven hours. A truck parked in the canyon washed away during these events (Tooele Transcript-Bulletin, May 15, 1984). Debris flows and floods flowed into Settlement Canyon Reservoir and covered an irrigation intake pipe 60 feet (18.3 m) below the water level with about 6 to 7 feet (1.8 to 2.1 m) of sediment (Tooele Transcript-Bulletin, May 22, June 12, 1984). Also, on May 14, 1984, a debris flow from Baltimore Gulch near the head of Pine Canyon in the Oquirrh Mountains struck and killed a man operating a bulldozer at the Carr Fork mine.

Debris-slide, debris-flow, debris flood, and stream-flood hazards are shown on plate 4. Plate 4 gives a relative rating of slope-failure susceptibility to indicate slopes expected to generate debris slides and debris flows. It also shows areas that may experience flooding and deposition of sediment from debris flows, debris floods, or stream floods. To date, no dam-failure inundation studies have been performed on dams in Tooele County, thus no inundation maps are available.

Over 100 debris slides and flows were identified in the Oquirrh Mountains which likely occurred during 1983 and 1984, shown as dots on plate 4. No debris slides or debris flows were identified in the Stansbury Mountains or the WDHIA. Two probable prehistoric debris flows originated near Stockton. Debris slides and flows in the Oquirrh Mountains occurred only in the southern half of the study area, between Soldier Canyon on the south and Flood Canyon on the north. All but a few are in Settlement, Middle, Pass, and Flood Canyons. No debris flows and only four debris floods deposited material beyond canyon mouths during 1983 and 1984, when debris floods in Pass and Swensons Canyons in the Oquirrh Mountains deposited sand and gravel on alluvial fans east of the town of Lincoln up to 1.5 miles (2.4 km) from the base of the mountains (figure 15). Two small, unnamed canyons in the northern Stansbury Mountains

Figure 15. West view of debris levee near Lincoln, Utah, formed by debris flows originating in the Oquirrh Mountains to the east during 1983-84.

northwest of Timpie Valley also yielded debris floods that deposited material on alluvial fans beyond the canyon mouths.

Slope-failure susceptibility in debris-flow source areas (source-area susceptibility on plate 4) provides a relative rating of susceptibility to failure, but does not estimate probability or likelihood of failure for a given time period. The frequency of occurrence (recurrence) of debris-slide and debris-flow events in a drainage basin depends upon climatic factors as well as the availability of debris. The map ratings are based mainly on the presence of pre-existing slope failures and slope angle. Other factors considered include vegetation type and density, rock and soil type, geologic structure, slope aspect, and elevation.

With a few exceptions, areas with a "high" susceptibility rating are generally slopes that produced debris slides and flows during 1983 and 1984. The "moderate" susceptibility category includes slopes that are steeper than 30 percent (17 degrees) that did not experience debris slides or flows during the wet years. All areas with slopes greater than 30 percent are considered potential debris-flow sources. The "low" susceptibility category includes slopes less than 30 percent. Few debris slides or flows occur on slopes less than 30 percent, and no such slope failures have occurred on these slopes in the WDHIA or Tooele Valley study area. Site investigations addressing slope stability should be performed prior to development in all areas of high and moderate susceptibility.

Plate 4 also shows areas of potential debris deposition and flooding (DFF). Hazard areas were defined from surficial geologic mapping by Solomon (1993), and show active (and potentially active) alluvial fans and stream channels where debris-flow, debris-flood, and stream-flood hazards may occur. Debris flows that reach canyon mouths generally deposit sediment on the heads of alluvial fans at canyon mouths close to mountain fronts. Therefore, areas along the fronts of the Oquirrh, Stansbury, and Cedar Mountains, and the Grayback Hills, have the greatest debris-flow hazard. Site investigations addressing the potential for sediment deposition and flooding from debris flows should be performed in DFF areas in canyon bottoms and at canyon mouths along mountain fronts, where no debris basin or other flood control structure exists above the site. However, debris floods and stream floods can affect areas farther away from mountain fronts than debris flows. Therefore, site investigations addressing these hazards (or disclosure of the hazards) should also be performed for DFF areas in the valley. Because of the scale of the maps, some small hazard areas are not shown. In addition, boundaries of DFF areas could change depending on activities such as road construction and residential development (which can change drainage patterns).

The adequacy of existing dams, debris basins, or structures built to divert debris flows or minimize flooding was not considered during preparation of the hazard maps. Such structures, where properly placed and of sufficient size, may limit the extent of deposition and flooding and reduce the potential hazard. Estimates of flooding and potential sediment yields from large events are necessary in evaluating the adequacy of these structures.

Loss of life during debris slides, flows, and floods may result from drowning, high-velocity impact, or burial. Damage associated with debris flows has been described by Campbell (1975), and is summarized here. Damage to residential structures ranges from simple inundation to complete destruction by high-velocity impact. The velocity of a debris flow is an important consideration in determining the level of damage to structures. Many debris flows move with speeds on the order of 40 feet/second (12.2 m/sec), but others move as slowly as 1 foot/second (0.3 m/sec) as they flow down relatively gentle slopes. Debris flows of sufficient volume and momentum have destroyed residential structures and removed the remains from their foundations. Debris flows of relatively small volume but high momentum have broken through walls and passed completely through structures. Low-velocity debris flows may enter dwellings through doors and windows. Debris flows and floods may fill basements with mud, water, and debris, or pile debris around structures. Debris may also bury yards, streets, parks, driveways, parking lots, and other ground-level structures. In the distal parts of alluvial fans, damage is usually comparatively minor, consisting primarily of mud and water damage to outer walls of buildings, basements, and yards.

Methods for reducing debris-related hazards include: (1) avoidance, (2) source-area stabilization, (3) transportation-zone modification, and (4) defensive measures in the depositional zone (Hungr and others, 1987). Different methods or combinations of methods may be appropriate for different drainages or types of development.

Debris-flow hazards may be reduced by avoiding, either permanently or at the time of imminent danger, areas at risk (source areas, transportation zone, and depositional zones). The source area is the origination point of the debris flow. The transportation zone is the debris-flow track between the source area and depositional zone. The depositional zone is the area where most of the sediment mobilized by the debris flow is deposited and where damage from debris flows may occur. Permanent avoidance is not possible in all areas because some Tooele Valley communities are on active alluvial fans (potential depositional zones). Reduction of debris-flow hazards could be required for proposed new development through creation and enforcement of foothill (zoning) ordinances that prohibit or regulate development in depositional zones.

Warning systems may be used to avoid life threats from debris flows at the time of imminent danger, generally through evacuation of threatened areas. Hungr and others (1987) identify three categories of debris-flow warning systems: pre-event, event, and post-event. Pre-event warning systems identify when climatic conditions have increased the potential for debris-flow occurrence. Event warning systems provide an alarm when a debris-flow event is occurring (Hungr and others, 1987). Post-event warning systems, such as slide-warning fences, are usually designed to warn of disruption of transportation routes (Hungr and others, 1987).

Source-area stabilization reduces the amount of hillside material available for incorporation into debris slides, flows, or floods. Improving drainage-basin vegetation is one method of source-area stabilization. Prevention of wildfires, overlogging, and overgrazing will protect existing vegetation. Terracing of mountain slopes, such as that done in the 1930s in Davis County by the Civilian Conservation Corps (Bailey and Croft, 1937), may be useful in preventing debris flows caused by erosion during cloudburst storms. Additional hazard-reduction techniques used near the source area include: (1) control of subsurface drainage, (2) diversion of surface drainage, (3) grading of source areas to a uniform slope, (4) riprap repair of source areas, and (5) retaining walls (Baldwin and others, 1987).

Transportation-zone modifications are generally designed to reduce incorporation of channel material into debris flows and floods, and to improve the ability of the channel to pass debris downstream. Scour of material in stream beds and undercutting of unstable stream banks are two of the most important processes contributing to the growth of debris flows (Hungr and others, 1987). Check dams (small debris-retention structures placed in unstable channels) are used to arrest or retard debris flows, and prevent incorporation of channel material (Hungr and others, 1987). Stream-bed stabilization is also achieved by lining the channel. The ability of channels to pass debris surges downstream may be improved through: (1) removal of channel irregularities; (2) enlargement of culverts with upstream removable grates to prevent blockage; and (3) flumes, baffles, deflection walls, and dikes (Jochim, 1986; Baldwin and others, 1987). Structures crossing potential debris-flow channels may be protected by: (1) bridging the channels to allow debris to pass under structures; (2) constructing debris sheds designed to allow debris flows to pass over structures; and (3) designing structures to withstand debris-flow impact, burial, and re-excavation (Hungr and others, 1987).

Defensive measures in depositional zones are designed to control the extent of deposition and prevent damage to structures (Hungr and others, 1987). Defensive measures include deflection devices, impact walls, and debris basins. Deflection devices are used to control the direction and reduce the velocity of debris flows (Baldwin and others, 1987). Impact walls are designed to sustain the force of impact from debris flows while containing the soil and vegetation debris until it can be removed (Baldwin and others, 1987). Debris basins are used to constrain the area of debris deposition, but require access for maintenance and removal of entrapped debris.

Loss of life during stream and dam-failure floods may occur by drowning where floodwaters are deep or flowing swiftly. Water damage depends largely on depth of inundation, and damage potential increases dramatically with increases in floodwater velocity (Federal Emergency Management Agency, 1985). High-velocity floodwaters can cause structures to collapse due to pressures applied by fast-moving water. Flowing water can also induce erosion and undermine structures. Areas subject to rapid inundation by flash floods pose special threats to life and property because of insufficient time for evacuation, emergency floodproofing, or other protective measures (Federal Emergency Management Agency, 1985). Methods for reducing stream-flood hazards and risk include: (1) avoidance, (2) drainage-basin improvement, (3) flow modification and detention, (4) flood warning and evacuation, and (5) floodproofing.

Requiring flood insurance in areas of frequent flooding is another means of dealing with flood hazards. In Tooele Valley, Flood Insurance Rate Maps are only available for major drainages in Tooele City (Federal Emergency Management Agency, 1989b); no such maps exist for the WDHIA. The Utah Division of Comprehensive Emergency Management may be contacted for information regarding the National Flood Insurance Program. County and city planning offices can provide information regarding zonation on the FEMA Flood Insurance Rate Maps.

Little can practically be done through land-use planning to reduce hazards from dam-failure floods. Methods used to reduce hazards from stream flooding, such as proper land use along floodplains, will help decrease damage due to dam-failure flooding to some extent. Emergency evacuation based on dam-failure-inundation maps is the principal means of reducing hazards due to dam-failure flooding. The Utah Division of Water Rights, Dam Safety Section, is the agency regulating dam safety in Tooele County.

Rock Fall

Introduction

Rock fall is a natural erosional process in mountainous areas of Tooele Valley and the WDHIA. As urban development advances towards the mountains, the risk from falling rocks increases. Rock falls can damage structures, roadways, and vehicles and may pose a significant safety hazard. The potential for rock-fall hazards is greatest along the Oquirrh Mountains in eastern Tooele Valley; however, a lesser rock-fall hazard also exists along the Stansbury Mountains and South Mountain in Tooele Valley, and along the Grayback Hills in the WDHIA.

Rock falls originate when weathering and erosion of supporting rock and sediment destabilize and eventually dislodge rocks from slopes. The most susceptible slopes are those with outcrops broken by bedding surfaces, joints, or other discontinuities into abundant, loose individual rock fragments called clasts. Shoreline benches eroded by Lake Bonneville and alluvium also contain clasts that may dislodge and fall. When the clast falls or rolls from the slope, it may travel great distances by sliding, rolling, and bouncing.

A primary mechanism responsible for triggering rock falls is water in outcrop discontinuities. In Norway, for example, 60 percent of all rock falls occur in April and May during maximum snowmelt and October and November during periods of heavy rainfall (Costa and Baker, 1981). In addition, rock falls are also the most common type of slope instability initiated by earthquakes. Case (1987d) estimates that a major Wasatch Front earthquake (magnitude 7-7.5) could produce thousands of rock falls along the Wasatch Front, including Tooele Valley. Keefer (1984) indicates that rock falls may occur in earthquakes as small as magnitude 4.0. In August 1988, the San Rafael Swell earthquake (magnitude 5.3) in central Utah produced hundreds of rock falls, temporarily obscuring the surrounding cliffs in clouds of dust (Case, 1988a). The September 1992, ML 5.8 St. George earthquake produced numerous rock falls that caused minor damage (Black and others, 1995).

Effects, Distribution, and Reduction

Rock falls are hazardous because a large rock mass traveling at high speed can damage structures and increase risk to personal safety. Rock falls that occur in remote or uninhabited regions often go unnoticed. A 1987 rock fall near Dead Horse Point, Utah, was large enough to register on seismographs as far away as Blanding (Case, 1987a). Along the Wasatch Front, rock falls have historically caused problems along canyon roads by damaging paved surfaces, blocking traffic, or striking vehicles. Structures most often affected by rock falls in canyons are roads and above-ground aqueducts. On January 14, 1995, a rock fall in Big Cottonwood Canyon crushed a car, fatally injuring one occupant and seriously injuring a second (Hylland, 1995). Water service in both Big Cottonwood and Provo Canyons has also been suspended due to aqueduct damage by impact and puncture from falling rocks. Homes built along the mountain front are also subject to rock falls. On March 10, 1994, a rock fall in Olympus Cove southeast of Salt Lake damaged a fence, retaining wall, and tennis court at a private residence (Black, 1994).

Plate 5 shows areas that may be susceptible to a rock-fall hazard in Tooele Valley and the WDHIA. The primary factor in determining these areas is the presence of a source for rock-fall clasts. If there are no rocks on a slope, the rock-fall hazard is low. Case (1987c, 1988b) identified some of the range-front slopes, called spurs, along the Oquirrh Mountains in Tooele Valley on which a rock-fall source was found. Additional source areas along the Oquirrh, Stansbury, and Cedar Mountains, South Mountain, and the Grayback Hills, were identified during this study.

The hazard area for each susceptible spur was determined using a computer model called the Colorado Rock-fall Simulation Program (CRSP) (Pfeiffer and Higgins, 1988). This program was primarily designed to predict rock bounce heights, but was used here to simulate maximum travel distances of rock clasts. The program incorporates factors such as velocity, rock size and shape, roughness of the travel surface, and topography of the slope. Rock-fall events were simulated using the highest and steepest potential rock-fall source areas. Rocks were started with an initial velocity (throw) of 1 foot/second (0.30 m/sec). The size of rock-fall clasts used in the simulation was based on the largest clast observed on the slopes below the rock-fall source area.

The program simulates 100 rock falls for each source area; the clast traveling the longest distance from its source was used to delineate rock-fall-hazard areas. Possible deceleration of rock clasts by existing structures, such as roads, railroad tracks, and fences, was not used in the analysis. Thus, the hazard areas represent conservative, worst-case rock-fall events.

Rock-fall simulations were run only on susceptible slopes along mountain-front areas; mountain interiors generally contain numerous rock-fall source areas and all canyons were included in the hazard areas shown on plate 4. Using a conservative approach, mountain-front slopes greater than 30 percent were also generally included in the hazard areas. Exceptions to this rule were mainly steeper areas where the rock-fall hazard is lessened by dense vegetation, such as in the southwestern South Mountain area, and in the Stansbury Mountain foothills between Box Elder and North Willow Canyons. Rock-fall hazard was not evaluated in the Flux vicinity due to active quarry operations that continuously alter the natural slopes.

Rock-fall hazard areas are numerous along the base of the Oquirrh Mountains in Tooele Valley due to steep slopes created by active mountain uplift and valley down-drop along the Oquirrh fault zone, and by erosion along the Lake Bonneville shoreline bench. In contrast, slopes along the eastern base of the Stansbury Mountains are generally gradual and more heavily vegetated than those along the Oquirrh Mountains. Thus, the rock-fall hazard is lower along the front of the Stansbury Mountains. Rounded basalt boulders and short, steep slopes contribute to the rock-fall hazard along the Grayback Hills in the WDHIA.

Techniques for reducing rock-fall hazards include rock stabilization or modification of exposed structures or facilities. Physical methods of reducing the hazard include rock-stabilization techniques such as bolting, cable lashing, burying, and grouting discontinuities, and removal or break-up of potential rock clasts. Deflection berms, slope benches, and rock-catch fences may stop or at least retard falling rocks. Strengthening a structure to withstand impact is an example of modifying structures at risk. Twenty-seven techniques for reducing landslide hazards, including rock falls, are described by Kockelman (1986). Hazard-reduction problems can arise when rock-fall source areas are on land not owned by those in the rock-fall runout zone.

In areas where the rock-fall hazard is present but is determined through site-specific investigation to be low, disclosure of potential hazards to land owners and residents may be an acceptable alternative to avoidance or mitigation, at least for single-family residences. Disclosure ensures that buyers are informed of the hazard, acknowledge the risks, and willingly accept them.

Lake Flooding, Ponding, and Sheet Flooding

Introduction

A flood is the stage or height of water above some given datum, such as a commonly occupied lake shoreline. Floods are recurrent natural events that become a hazard to residents of a floodplain or shoreline whenever water rises to the extent that life and property are threatened. Tooele Valley is subject to flooding from rises in Great Salt Lake, and both Tooele Valley and the WDHIA are subject to local ponding and sheet flooding.

Although fluctuating water levels are a problem in lakes, they are especially acute on lakes which, like Great Salt Lake, have no outlet. Natural factors causing fluctuations include precipitation, evaporation, runoff, ground water, ice, aquatic growth, and wind; human factors include dredging, diversions, consumptive use, and regulation by engineered works (Federal Emergency Management Agency, 1985). Lake-level fluctuations may be grouped into three categories: (1) long term, (2) seasonal, and (3) short term. Fluctuations of Great Salt Lake have occurred in prehistoric and historical time, and flooding due to rising water levels is a hazard in Tooele Valley.

Long-term fluctuations are the result of persistent low or high water-supply conditions for more than one year. Figure 16 shows the effects of long-term excess precipitation on Great Salt Lake elevation during the 1980s. Long-term climatic trends play a major role in determining lake levels, as do diversions of water sources by man. The intervals between periods of high and low lake levels, as well as the length of such periods during long-term fluctuations, vary widely and erratically (Federal Emergency Management Agency, 1985). Extreme lake levels are likely to persist even after the factors which caused them have changed.

Seasonal fluctuations reflect the annual hydrologic cycle. Lake levels are lowest in winter and generally rise in the spring due to melting snow, heavier rains, and cooler temperatures, until the lake peaks in early summer (Federal Emergency Management Agency, 1985). During the summer, more persistent winds, drier air, and warmer temperatures intensify evaporation; runoff and ground-water flow to the lake decrease significantly. As the amount of water supplied to the lake becomes less than that removed by evaporation, the water level drops to winter minima (Federal Emergency Management Agency, 1985). Great Salt Lake elevations fluctuate approximately 2 feet (0.6 m) between winter low and summer high lake levels.

Short-term fluctuations are caused by strong winds and sharp differences in barometric pressure (Federal Emergency Management Agency, 1985). These fluctuations usually last less than one day and do not represent any changes in the amount of water in the lake.

Ponding and sheet flooding are flood hazards result-

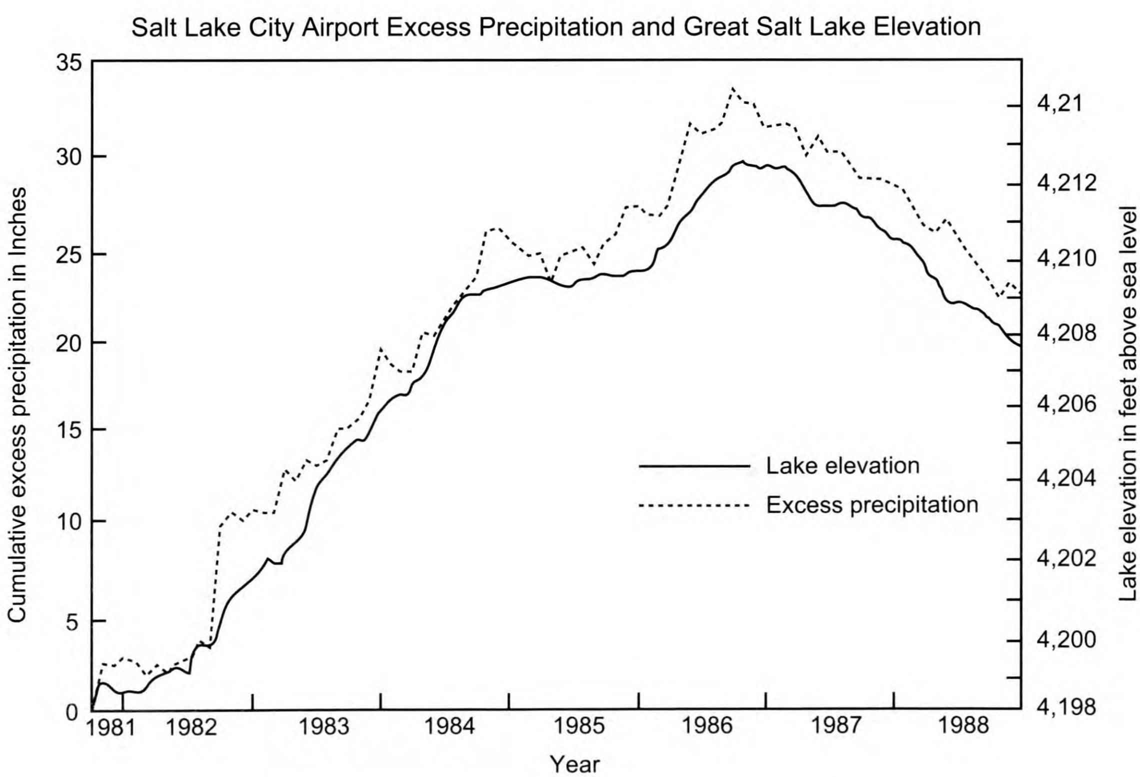


Figure 16. Effects of excess precipitation on Great Salt Lake levels (modified from Atwood and Mabey, written communication in Lowe and others, 1992).

ing from runoff or precipitation collecting in flat low-lying areas following periods of intense, cloudburst rainfall, or rapid melting of snow. Localized, high-intensity, cloudburst rainstorms, which last from a few minutes to a few hours, are unpredictable and likely cause most of the ponding and sheet flooding. These rainstorms are characterized by high peak, high velocity, short duration, and small volume runoff. Snowmelt floods may also cause ponding and sheet flooding. These floods are generally predictable, and are characterized by large volume runoff, moderately high peak flows, and marked diurnal fluctuation in flow.

Effects, Distribution, and Reduction

In prehistoric time, water levels in lakes occupying the Great Salt Lake basin, such as Lake Bonneville, fluctuated with great elevation differences between high and low stands (figure 17). Geologic evidence indicates that Great Salt Lake reached a post-Lake Bonneville high of approximately 4,221 feet (1286 m) about 2,000 years before present (Murchison, 1989). Archaeological evidence indicates that the most recent high stand of Great Salt Lake was at 4,217 feet (1285 m) sometime during the 1600s (Utah Division of Comprehensive Emergency Management, 1985; Murchison, 1989).

Water levels in Great Salt Lake have also fluctuated in historical time. Mean historical elevation of the lake is about 4,200 feet (1,280 m) (J. Wallace Gwynn, UGS, verbal communication, September 1998). Until mid-1986, the historical high of Great Salt Lake was about 4,211.5 feet (1283.6 m) (Arnow and Stephens, 1990). This level was reached in the early 1870s and is based on a relative elevation estimate of water depth over the Stansbury bar (Gilbert, 1890). Direct measurements of the lake's elevation began in 1875 (Currey and others, 1984a). The lake dropped slowly from its high in the 1870s, reaching a historical low of 4,191.35 feet (1277.46 m) in 1963. Above-average precipitation in the 1980s caused Great Salt Lake to attain a new historic high of 4,211.85 feet (1283.71 m) in June, 1986 (Arnow and Stephens, 1990) and April, 1987 (U.S. Geological Survey records). These rises in lake level caused damage to structures and property along the shoreline and within the lake (power lines, causeways, dikes, buildings, and refuse dumps). Figure 18 summarizes historical levels of Great Salt Lake and illustrates that significant lake fluctuations can occur within a relatively short time.

Rush Lake has fluctuated from the size of a "small pond" in the early 1860s (Gilbert, 1890) to marsh-like and dry in the late 1950s to mid-1970s (Harty and Christenson, 1988). The lake was at or near 4,979 feet (1,518 m) when measured in 1872 (Gilbert, 1890), and reached its highest elevation in 1876 or 1877, although no measurements were made at that time (Harty and Christenson, 1988). Like Great Salt Lake, water levels in Rush Lake also rose in the 1980s; between 1983 and 1985 Rush Lake rose nearly 10 feet (3 m), damaging powerlines and crop lands surrounding the lake (Harty and Christenson, 1988).

Ponding and sheet flooding may occur in mudflats of the western WDHIA and in northern Tooele Valley. Any runoff or precipitation that reaches the mudflats usually evaporates, but ponding often occurs in the winter and early spring.

Water damage accompanies flooding and ponding, and the amount of damage largely depends upon depth of inundation and duration. Along the shore of Great Salt Lake, the problems associated with water damage are also compounded by the presence of salt in the water. In areas where flooding is deep and of long duration, such as along the shoreline of Great Salt Lake, water damage to structures is especially serious. Although this flooding generally is not life-threatening, it will likely cause permanent property loss or damage.

Hazard-reduction methods for lake flooding include avoidance, diking, diverting inflow to the lake, and increasing outflow and/or evaporation through pumping (Utah Division of Water Resources, 1977). Avoidance, floodproofing, and site grading can reduce ponding and sheet-flooding hazards. Different methods or combinations of methods may be appropriate for different types of flooding or development.

Using the best available historical and scientific data on Great Salt Lake, government policy makers and lake experts have recommended that a beneficial development strategy should exist for lake-shore areas up to 4,217 feet (1,285 m) in elevation (Utah Division of Comprehensive Emergency Management, 1985). This strategy establishes a "Beneficial Development Area" along the shore of Great Salt Lake between 4,191.4 feet (1,277.5 m) (historical low stand, 1963) and 4,217 feet (1,285 m). Within this area, it is recommended that development take place in a manner that will encourage the maximum use of the land for the people of Utah, while avoiding unnecessary disaster losses (Utah Division of Comprehensive Emergency Management, 1985). The most effective way to reduce hazards would be to adopt this beneficial development strategy and ensure that development within this area is either compatible with or protected from the flood hazard.

Recent shoreline flooding around Great Salt Lake has been locally controlled by dikes. However, this is not a long-term solution. Stabilization of rising water levels is difficult to accomplish. Diversion of inflow from rivers that flow into the lake is not generally con-

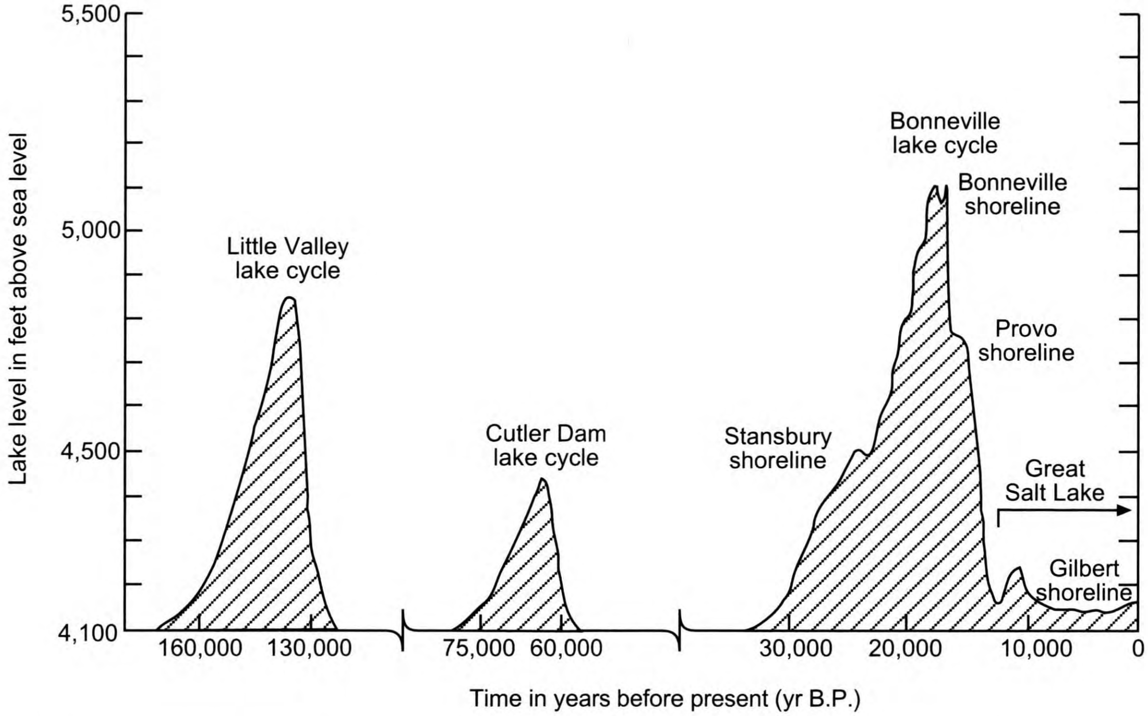


Figure 17. Probable lake levels in the Lake Bonneville (Great Salt Lake) Basin for the past 150,000 years (modified from Currey and Oviatt, 1985; Machette and others, 1992).

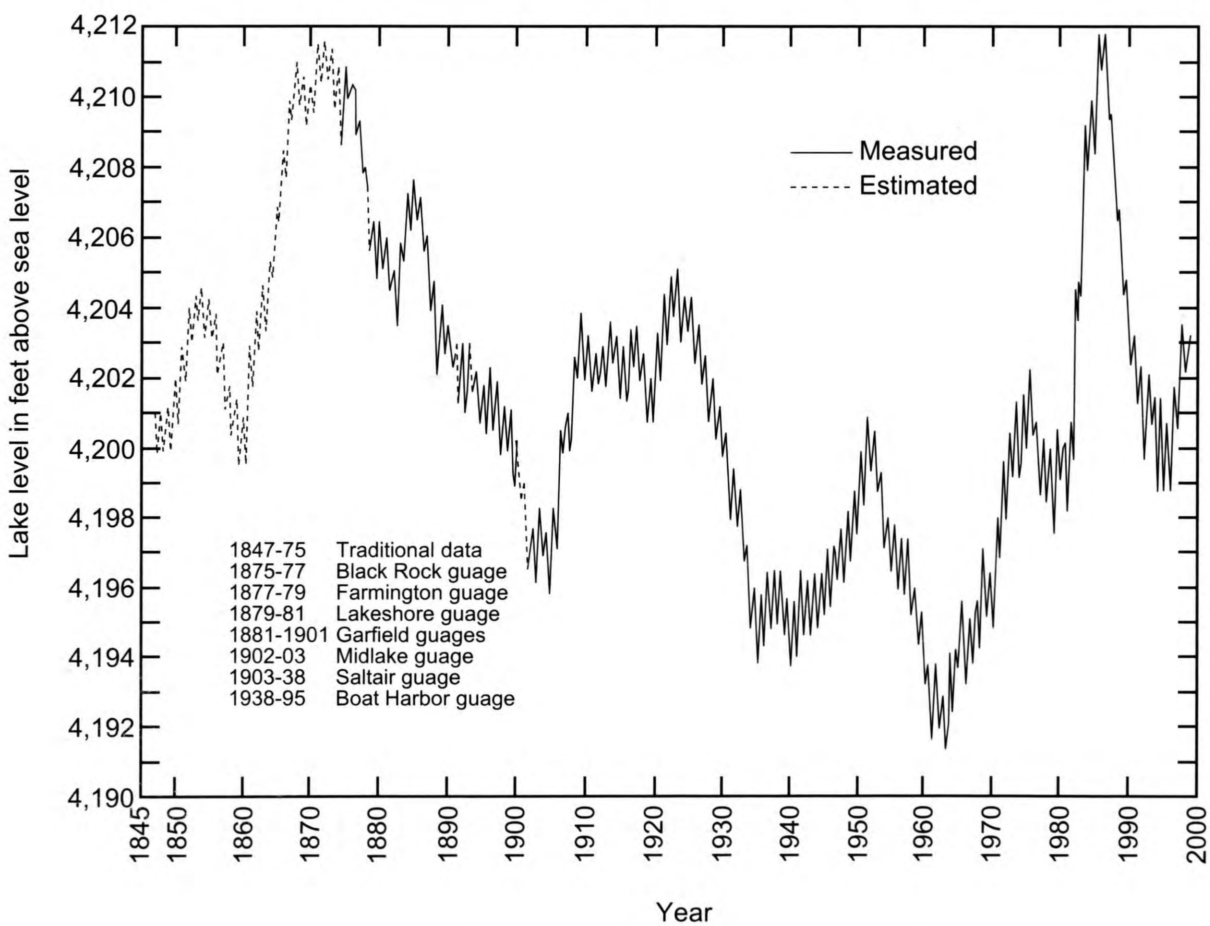


Figure 18. Historical Great Salt Lake hydrograph (J. Wallace Gwynn, UGS, written communication, 1999).

sidered an effective option. Pumping to adjacent basins to increase evaporation can reduce flooding, but also cannot control water levels. Flooding around the margins of Great Salt Lake has been reduced by such means. In the late-1980s, lake water was pumped into a shallow desert basin west of the lake to increase surface area subject to evaporation. Although the pumps are effective in reducing flooding, precipitation during a very wet period may exceed the capabilities of pumping and evaporation.

Avoidance is one method of dealing with ponding and sheet flooding, although it may not be possible where population centers are on relatively flat valley floors. Floodproofing is also an effective way of reducing flood damage in areas where floods are of short duration and have low stages and velocity. Floodproofing measures include the use of special cements for flooring, adequate electrical fuse protection, anchors for buoyant tanks, sealed outside walls and basements, wire-reinforced glass, automatic sump pumps, sewer check valves, sealed windows and doors, and window and door flood shields (Kockelman, 1977). Modifications of site grade, such as elevating structures and access roads, may also be needed.

Plate 2 depicts areas that may be subject to lake flooding, ponding, and sheet flooding. Areas subject to ponding and sheet flooding are restricted to mudflats in the western WDHIA, northern Tooele Valley, and Rush Lake. Areas in Tooele Valley along the southern shoreline of Great Salt Lake, where the proposed lake flooding beneficial development strategy is recommended, include all areas below an elevation of 4,217 feet (1,285 m). The location of the 4,217-foot (1,285-m) contour has been interpolated from 1:24,000-scale U.S. Geological Survey topographic quadrangle maps. Hazard areas in northern Rush Valley are adjacent to Rush Lake and below an elevation of 4,979 feet (1,518 m), which was defined as the potential flood boundary for Rush Lake in Harty and Christenson (1988). However, these lines are only approximate and accurate field surveys should be performed prior to development.

Shallow Ground Water

Introduction

Ground water is water in saturated zones beneath the land surface. Ground water fills fractures and pore spaces in rocks and voids between grains in unconsolidated deposits (clay, silt, sand, and gravel). Ground water is considered shallow where the water table is within 30 feet (9 m) of the ground surface (Hecker and others, 1988).

Shallow ground water in rock is not considered here because it poses a relatively insignificant geotechnical hazard. Foundations and conventional waste-water disposal systems in rock are uncommon, and foundation stability is not appreciably reduced by saturated conditions (Hecker and others, 1988). However, most construction takes place in areas of unconsolidated sediments subject to various hazards associated with shallow ground water. Such hazards include flooding of basements and buried facilities, destabilization of foundations or excavations, surface flooding, and liquefaction of soils during earthquakes. Shallow ground water is found in northern Tooele Valley and in much of the WDHIA, and must be taken into consideration when siting waste-disposal facilities and septic-tank soil-absorption systems.

Flooding due to shallow ground water in basements, foundations, and excavations generally only occurs when the saturated zone is within the depth to which most building foundations are excavated. Surface flooding due to shallow ground water can occur anytime ground water rises to the surface. Liquefaction during earthquakes, and potential ground failure, may occur in saturated sandy soils where the depth to ground water is less than 30 feet (9 m) (Youd and others, 1978b). Earthquakes may also cause rises in water tables and increased ground-water discharge.

Ground water in unconsolidated deposits, chiefly stream, alluvial-fan, and lacustrine sediments, occurs under unconfined and confined conditions in geologic units known as aquifers. These units are permeable enough to yield water in usable quantities to wells and springs (Heath, 1983). The principal water-bearing aquifer in Tooele Valley is in unconsolidated alluvial and lacustrine basin-fill deposits (Steiger and Lowe, 1997). The principal aquifer in the WDHIA is in unconsolidated lacustrine deposits (Stephens, 1974).

An unconfined aquifer is generally not saturated throughout its entire thickness; the top of the saturated zone in unconsolidated sediments is termed the water table (figure 19). Localized occurrences of unconfined ground water above the water table are called "perched zones" (figure 18). Perched ground water commonly is above localized layers of low-permeability sediments, such as clay.

Where ground water saturates the entire thickness of an aquifer below an areally extensive low-permeability layer, termed a confining bed, the aquifer is said to be under confined conditions. Ground water under confined conditions (artesian water) is usually under hydrostatic pressure exerted by higher water levels in recharge areas. Water in wells which penetrate a confined aquifer usually rises above the top of the aquifer to the potentiometric

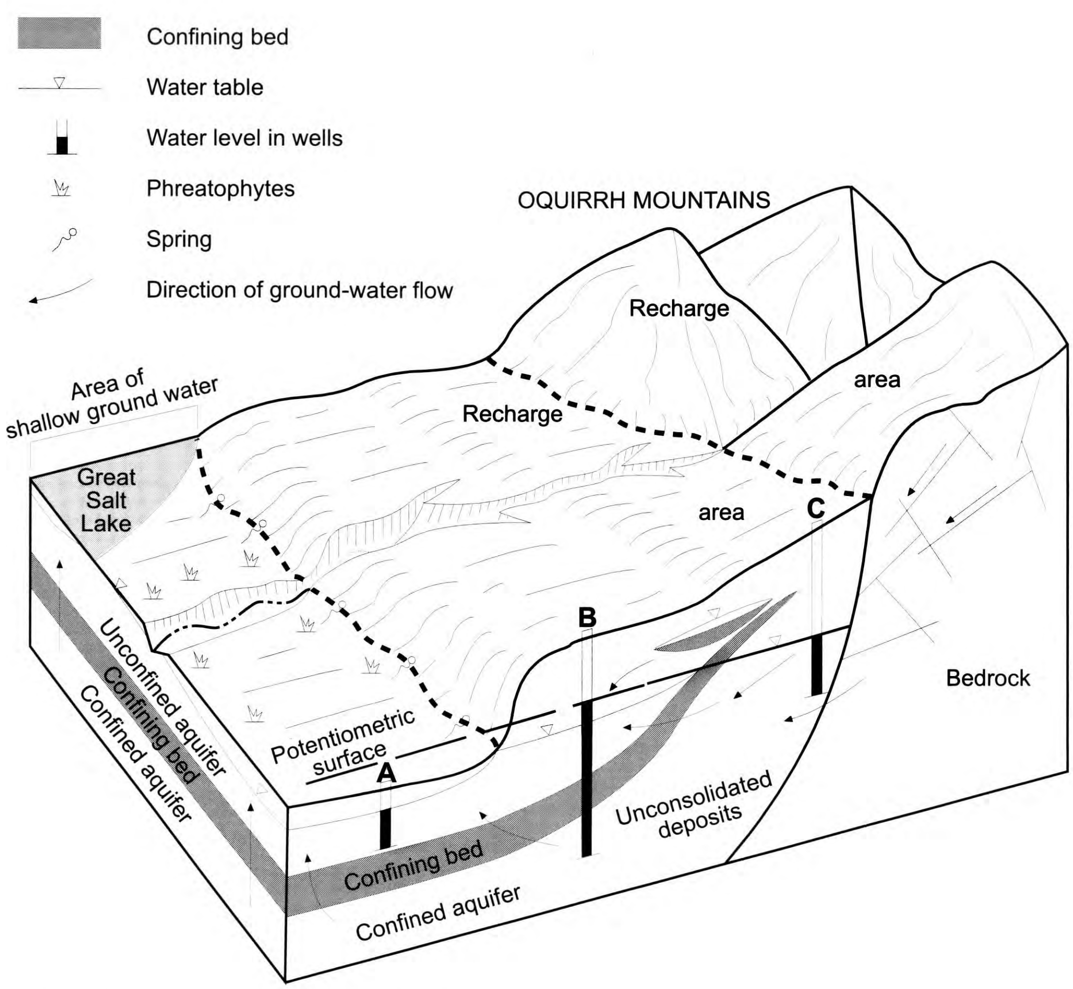


Figure 19. *Ground-water conditions in Tooele Valley showing: unconfined and confined aquifer, confining bed, perched water, water table, potentiometric surface, recharge area, and area of shallow ground water. Note level of water in well B rises above the water table due to artesian (confined) conditions (modified from Hely and others, 1971).*

surface (well B, figure 18), which is determined by hydrostatic pressure in the aquifer. However, confining beds in unconsolidated deposits are generally semi-permeable and may allow water to leak upward and help maintain the water table above the confined aquifer (Hely and others, 1971; Razem and Steiger, 1981) (figure 18).

Throughout much of northern Tooele Valley, discontinuous lacustrine clay layers separate the basin-fill aquifer into multiple water-bearing zones under both confined and unconfined conditions (Razem and Steiger, 1981; Bishop, in preparation). In southern Tooele Valley and along the margins of the valley, the clay layers are absent and only one unconfined aquifer is present (Razem and Steiger, 1981; Bishop, in preparation).

Ground water in the WDHIA is chiefly under unconfined conditions (Stephens, 1974). Studies of the hydrogeology of the Bonneville Salt Flats to the west of the WDHIA (Nolan, 1928; Turk, 1973) suggested to Stephens (1974) a model in the Great Salt Lake Desert of a ground-water system divided into three distinct segments: a surficial-brine aquifer composed of surficial lake beds and crystalline salt, an alluvial-fan aquifer on piedmont slopes that yields moderately saline water, and a valley-fill aquifer which underlies lake beds and also yields brine. This model is applicable to the WDHIA, where the surficial-brine aquifer occurs in the mud flats west of the Grayback Hills, the alluvial-fan aquifer occurs in the piedmont zone of the Cedar Mountains, and the valley-fill aquifer underlies the entire area.

Hydrologic investigations of Tooele and Rush Valleys demonstrate that ground water in Tooele Valley flows northward into Great Salt Lake (Gates, 1963, 1965; Gates and Keller, 1970; Razem and Steiger, 1981; Ryan and others, 1981), whereas northern Rush Valley is essentially a closed hydrologic basin in which recharge flows from surrounding mountains into Rush Lake (Hood and others, 1969). Thomas (1946), Gates (1962, 1965), and Razem and Steiger (1981) suggest some faults in Tooele Valley may serve as ground-water barriers and complicate ground-water flow.

Shallow ground water is replenished by infiltration from streams, lakes, precipitation, lateral subsurface flow from adjacent higher ground-water areas, and upward leakage from underlying confined aquifers (Heath, 1983). The shallowest water tables are generally found in the central parts of valleys, where leakage from underlying artesian aquifers is greatest and potentiometric surfaces are commonly above the ground surface (figure 18). Man influences local water levels through irrigation, pumping from wells, and surface-drainage diversions and reservoirs (Hecker and others, 1988).

The shallow water table is dynamic and fluctuates in response to a variety of conditions. Ground-water levels may rise and fall with seasonal variations in precipitation, long-term climatic changes, or changes in rates of irrigation or pumping. A series of years with greater-than-average precipitation beginning in the late 1960s, but particularly between 1982 and 1986, increased ground-water recharge to basins and elevated ground-water levels statewide. Drought conditions in the late 1980s caused a general decline.

Effects, Distribution, and Reduction

The most significant hazards associated with shallow ground water are flooding of subsurface facilities (such as basements) and damage to underground utility lines; inundation of landfills and waste dumps and effects on septic-tank soil-absorption fields; and possible damage to foundations, roads, and airport runways from soils affected by moisture. Structures extending below the water table may experience water damage to their foundations and/or contents; underground utilities may also experience water damage. Landfills and waste dumps may become inundated and contaminate aquifers, and septic-tank soil-absorption fields can become flooded and cause ground-water contamination as well as system failure. In addition, certain foundation soils can settle or expand when wet, causing damage to foundations and structures. Roads and airport runways may buckle or settle as bearing strength of foundation soils is reduced by saturation.

Shallow ground water may also erode and dissolve subsurface materials, resulting in soil piping and settlement. Water flowing through bedrock fissures in limestone or gypsiferous rocks can dissolve the rock and create holes which may also collapse.

Contaminants are easily introduced into shallow ground water because it is readily accessible from the surface. Pollutants will flow with the ground water and may enter deeper aquifers or seep into wells. About 85 percent of Utah's wells are located within basin-fill aquifers; some are becoming increasingly contaminated (Waddell and Maxell, 1987). Some wells in the Erda area in eastern Tooele Valley show evidence of nitrate contamination (Bishop, in preparation).

Plate 5 shows areas where a shallow ground-water hazard may be found in the WDHIA and Tooele Valley. Ground-water depths are grouped into four zones on the maps: (1) less than 10 feet (3 m; zone A), (2) 10 to 30 feet (3 to 9 m; zone B), (3) 30 to 50 feet (9 to 15 m; zone C), and (4) greater than 50 feet (15 m; zone D). Information on Tooele and Rush Valleys is from Razem and Steiger (1981), Hood and others (1969), and well-log data from the Utah Division of Water Rights. Information on the WDHIA is from Dames & Moore and others (1987b), Stephens (1974), and U.S. Department of Energy (1983). We also used distribution of springs and phreatophytes (plants whose roots intersect the water table) as an indication of the presence of shallow ground water.

Most problems associated with shallow ground water occur where the water table is within about 10 feet (3 m) of the ground surface. Ground water at this depth is found in both the WDHIA and Tooele Valley. Site-specific shallow ground-water studies are recommended for all types of construction with subsurface facilities in areas where the water table is likely to be within 10 feet (3 m) of the ground surface.

Avoidance, although not always possible, is one method of reducing shallow ground-water problems. Construction techniques may be employed which reduce or eliminate the adverse effects of ground-water flooding. Waterproofing of subsurface structures may be the most common technique, and may include drainage systems around basements. Waterproofing requirements are given in the Uniform Building Code (International Conference of Building Officials, 1997). Slab-on-grade buildings, which have no basement, are common in areas with a shallow water table. Pile foundations may also be used to increase foundation stability. Fill may be added to raise building elevations.

Pumping to lower the water table is also possible, but is typically used only during the construction phase. Pumping is an expensive and unreliable technique for permanently lowering a water table. Basement sump pumps are usually effective for individual homes.

Septic-tank soil-absorption fields do not function properly if inundated by shallow ground water. Utah State Health Department regulations therefore require that the base of the drain lines be at least 2 feet (0.6 m) above the highest expected ground-water table. Wisconsin mound septic-tank soil-absorption systems are currently experimental in Utah, but may be an alternative system that could be used in shallow ground-water areas. The drain lines in this type of system are buried in a mound above the natural ground surface to increase evaporation and the soil thickness between the lines and the water table.

Problem Soils and Subsidence

Introduction

Problem soil is a broad category of geologic hazards that involve unconsolidated surficial geologic materials having characteristics that make them susceptible to volumetric changes, collapse, subsidence, or other engineering-geologic problems. These hazards include expansive soil, gypsiferous soil, piping, and mine subsidence. Expansive soil is a hazard in both Tooele Valley and the WDHIA; deposits susceptible to piping may also occur in these areas. Gypsiferous soil may be found in mudflats of northern Tooele Valley and western portions of the WDHIA, whereas mine subsidence is generally only a hazard in the Oquirrh Mountains east of Tooele Valley.

Geology is the main factor influencing the extent of problem soil, and the geologic parent material largely determines the type of hazard. For example, expansive soil is most often associated with clay and shale, whereas dissolution features commonly form in limestone and gypsiferous material. Climate is an additional factor for soils subject to dissolution and collapse. However, one subsidence problem is not soil related; mine subsidence is due to the collapse of underground mines and is solely related to the activities of man.

Expansive soil is clay rich. Clay minerals cause the soil to expand and contract with changes in moisture content. All clay minerals expand to some degree with the addition of water, but some varieties such as montmorillonite (the most common variety of clay in Utah) can swell to 2,000 times their original dry volume (Tourtelot, 1974). Expansive soil may be found in fine-grained lake deposits in northern Tooele Valley and the western half of the WDHIA.

Clays may swell in two ways when wetted, either by absorption of water between clay particles or by absorption of water into the crystal lattice that makes up individual particles (figure 20). In both processes, the absorbed water causes the clay to expand. Montmorillonite commonly swells by absorption of water between individual crystals. As the material dries, the loss of water causes it to shrink. The processes of wetting, drying, freezing, and thawing churn and disturb the surface of expansive deposits, giving some of them a characteristic "popcorn" texture. This texture is a good indicator of the presence of expansive soil.

Gypsum is soluble, and gypsiferous soil may be sub-

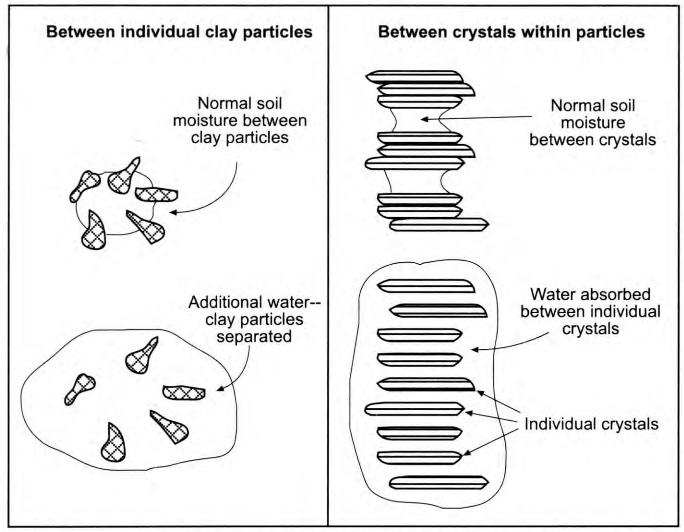


Figure 20. *Water-absorption processes in clay minerals (after Mulvey, 1992).*

ject to dissolution. Settlement may occur due to loss of internal structure and volume from dissolution. Gypsum-rich soil may be formed as a secondary mineral leached from surficial layers and concentrated lower in the soil profile, or may be transported by wind or water from outside sources. The most common sources for airborne gypsum are playas, on which crusts of gypsum salts are formed as the wetted playa surface dries during warmer months. These gypsum crusts are easily eroded and transported by wind. Gypsiferous soil may occur in wind-blown deposits in the western half of the WDHIA.

Piping is a common process in arid climates where fine-grained, uncemented, unconsolidated deposits are incised by streams. Piping occurs when ground water, moving along permeable, noncohesive layers in unconsolidated materials and exiting at a free face that intersects the layer, causes subsurface erosion (Cooke and Warren, 1973; Costa and Baker, 1981). Removal of fine-grained particles (silt and clay) by this process creates voids that act as minute channels which direct the movement of water (figure 21). As channels enlarge, water moving through the conduit increases velocity and removes more material, forming a "pipe." The "pipe" becomes a preferred avenue for ground-water flow, growing in size as larger volumes of water are intercepted. Increasing the pipe size removes support for its walls and roof, causing eventual collapse. Collapse features form on the surface above the pipes, directing even more surface water into them. Eventually, total collapse forms a gully that concentrates erosion along a line of interconnected collapse features.

Deposits susceptible to piping in Tooele County include fine-grained marl and silt deposited by Lake Bonneville (Mulvey, 1992). Several conditions are nec-

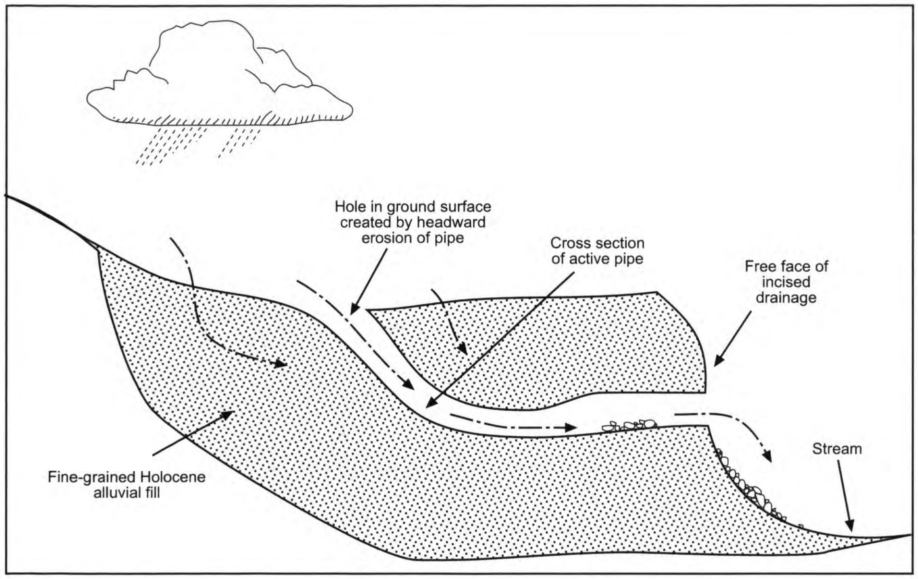


Figure 21. *Cross section of a pipe in Holocene alluvium (after Mulvey, 1992).*

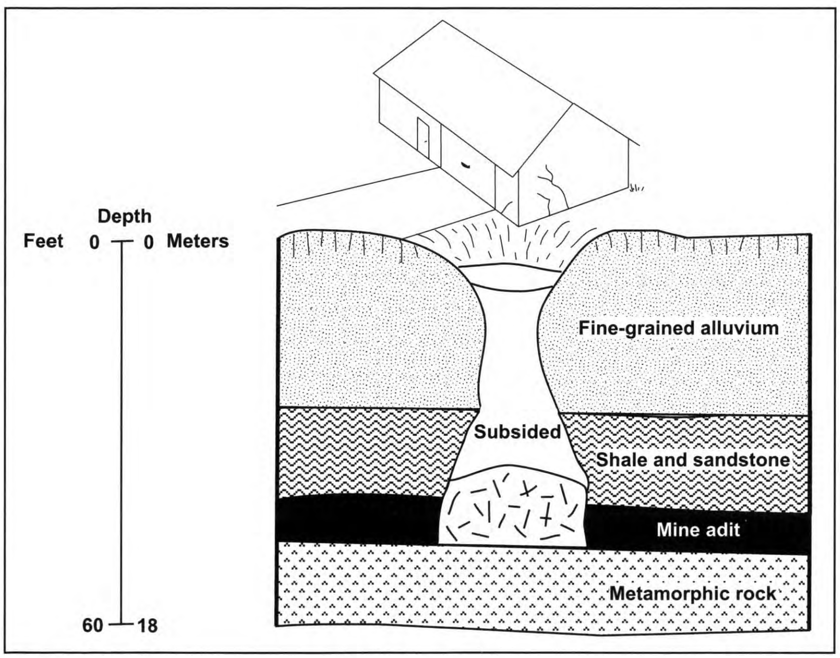


Figure 22. *Cross section of a mine-collapse subsidence pit, under a house, in an area of thick soil cover (modified from Turney, 1985).*

essary for piping. Water must be present in volumes large enough to soak into the subsurface and reach layers or zones (animal burrows, decayed plant roots) which conduct the water to a free face. The local surface topography must also have enough relief to create a hydraulic head, and move water through the subsurface. Deposits susceptible to piping must be fine grained and uncemented, but permeable enough to allow subsurface movement of water. Finally, a free face or cliff is necessary for water and sediment to exit the deposit (Costa and Baker, 1981).

Mine subsidence occurs above both active and abandoned mines. The removal of rock from the subsurface can cause subsidence of the land surface above, as the void left by mining is filled by collapse of overlying material (figure 22). The long history of mining in Utah has created many areas with surface subsidence or sinkholes. Companies removing rock from the subsurface are now required by law to devise a mining method that reduces the potential for surface subsidence, monitor subsidence, and file a report with the Utah Division of Oil Gas and Mining each year. The subsidence investigations are based on surveyed grids laid out over mining areas. If subsidence occurs, the mine is required to alter their mining methods to prevent further subsidence (A.C. Keith, Utah Geological and Mineral Survey, personal communication, January, 1990). The Bingham mining district, in the Oquirrh Mountains on the eastern edge of Tooele Valley, may be subject to this hazard, although there are no documented occurrences of mine subsidence.

Effects, Distribution, and Reduction

Problems commonly associated with expansive soil are cracked foundations (figure 23), heaving and cracking of road surfaces, and failure due to plugging of septic-tank soil-absorption systems. Single-family homes are particularly susceptible to damage from expansive soil because foundation loads (1,500 to 2,500 lbs/ft^2) [7,323 to 12,205 kg/m^2] may be less than the expansive pressures (3,000 to 11,200 lbs/ft^2) [14,646 to 54,678 kg/m^2] caused by the swelling material, making them subject to heave (Costa and Baker, 1981). Larger, heavier buildings are better able to withstand the expansive pressure, and are less susceptible to damage. Sidewalks, roads, buried utilities, and slabs-on-grade are also susceptible to cracking and damage due to differential expansion and contraction of underlying material.

Wastewater disposal systems using soil-absorption fields can also be affected by expansive soil. Clay-rich deposits develop cracks when dry, leaving voids which allow large volumes of water to infiltrate initially. Once

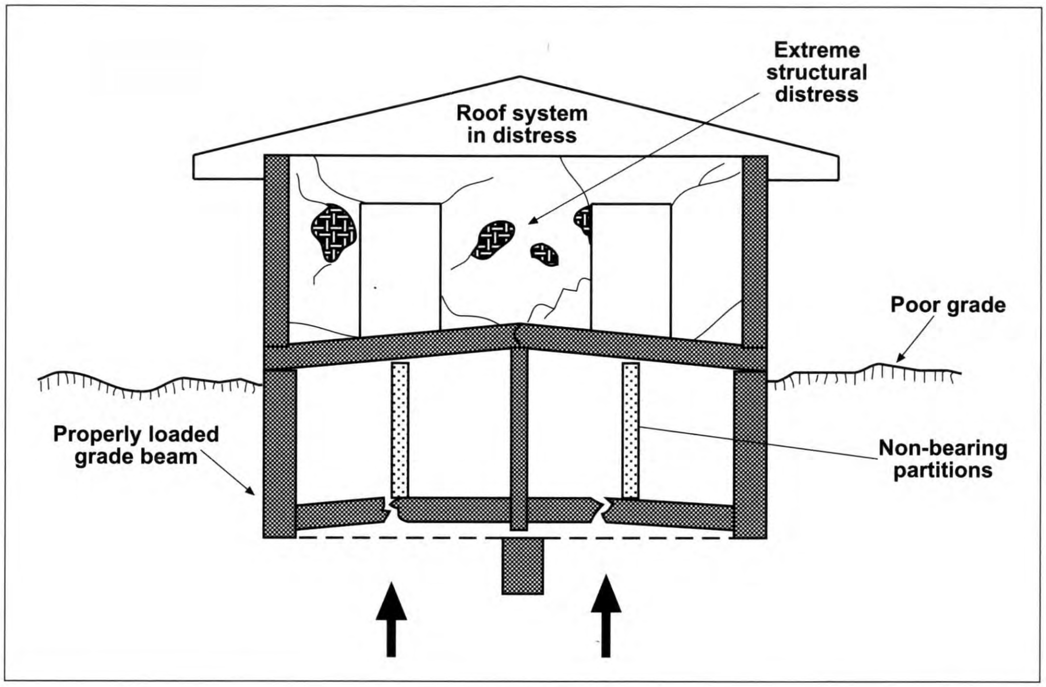


Figure 23. *Typical major house damage from expansive soil (from Holtz and Hart, 1978).*

saturated, the clay minerals swell, closing the voids. Soil-absorption systems installed in expansive soil work until the soil becomes saturated and swells. The soil quickly becomes impermeable and the systems clog and fail, causing wastewater to flow to the surface creating a health hazard.

Gypsiferous soil has the potential to cause damage to foundations and/or cause land subsidence and sinkholes. When wetted by irrigation for crops or landscaping, or by water from wastewater disposal systems, gypsiferous soil may subside due to dissolution. In some cases, large underground solution cavities may form and then collapse. Gypsum is also a weak material with low bearing strength. When gypsum weathers it forms sulfuric acid and sulfate (Bell, 1983). These compounds may react with certain types of cement, weakening foundations by damaging the exterior surface.

Piping and mine subsidence can cause damage to any overlying structure. Earthfill structures such as dams may be susceptible to piping, and piping of fine-grained embankment materials at the base of the Quail Creek dike, near St. George, contributed to its failure in 1989 (James and others, 1989). In the Uinta Basin, irrigation of crop land adjacent to incised drainages has caused extensive piping. In areas where piping is common, roads are most frequently damaged because they commonly parallel stream drainages and cross-cut numerous pipes. In addition, their construction commonly disturbs natural runoff, concentrating it near the roads. Collapse of underground mine adits may damage overlying structures and alter local surface topography. Mine subsidence is affected by factors such as depth of the mine, size and orientation of adits, and subsurface geology. Unlike other problem soil hazards, mine subsidence is related to human activity and is only a hazard in areas of underground mining.

Plate 6 shows the likely extent of expansive and gypsiferous soils, based on surficial geology, in Tooele Valley and the WDHIA. The map is designed to highlight areas where these soils may be present and should be evaluated in standard soil-foundation investigations prior to development. In hazard areas, improperly designed roads and structures can be susceptible to damage. The maps are generalized and other localized areas may occur outside of mapped problem-soil areas. Areas of possible piping or mine subsidence were not mapped.

The best method to reduce the hazard from expansive soil is to restrict changes in water content. Drainage conditions affecting soil moisture are important in areas of expansive soil. When water from sprinkler systems or runoff from roofs and roads reaches deposits beneath the structure, damage may occur as the material expands.

To reduce damage from expansive soil, several techniques can be used. For structures, these include: (1) using gutters and downspouts to direct water at least 10 feet (3 m) away from foundation slabs; (2) avoiding vegetation that concentrates or draws large amounts of water from the soil near foundations; (3) insulating floors or walls near heating or cooling units, which prevents evaporation and local changes in soil moisture; (4) strengthening house foundations by reinforcing concrete with steel bars; and (5) driving pilings into the soil to a depth below the active zone to support walls (Costa and Baker, 1981). Wide shoulders and good drainage along highways can prevent road damage. In highway foundations, a combination of hydrated lime, cement, and organic compounds can be added to road subgrade materials to stabilize the underlying soil (Costa and Baker, 1981). For wastewater-disposal systems, a 24-hour "presoak" of the material (prior to determining percolation rates) may yield a more reliable percolation rate on which to base system design and approval.

Because gypsum is dissolved by contact with water, runoff from roofs and gutters should be directed away from the structure. Landscaping close to the house should not include plants which require regular watering. Damage to structures from gypsiferous soil can be limited by several methods. The outer walls of structures can be coated with impermeable membranes or bituminous coatings to protect them from deterioration. Special sulfate-resistant concrete can also be used.

Damage caused by piping can be reduced by controlling drainage in susceptible soil. Runoff concentrated or ponded along paved surfaces allows greater infiltration and encourages piping. Culverts to collect runoff, and closed conduits to carry the water away from the road, will prevent damage. Concrete-lined drainage ditches, and concrete or asphalt around culvert inlets and outlets, can also limit damage. Damage to crop land can be reduced by limiting the amount of irrigation along incised stream drainages. Avoidance is the easiest and most cost-effective hazard-reduction technique for mine subsidence. In areas above mines, assessment of the potential for collapse should be made prior to development.

Indoor Radon

Introduction

Most geologic hazards are natural, dynamic, earth processes that alter the landscape and adversely affect the works of society. The occurrence of high radon concentrations in buildings, although not a process of landscape alteration like most geologic hazards, is nonetheless recognized as a geologic hazard.

Radon is an odorless, tasteless, radioactive gas. When inhaled, radon can be a significant cause of lung cancer. Whereas high levels of radon gas in uranium mines have long been recognized as a health hazard to miners, the hazard from indoor radon at lower levels has only recently been recognized. Radon has been found in many buildings throughout the United States in sufficient concentrations to represent a health hazard to building occupants. Concern for the health consequences associated with long-term indoor-radon exposure has prompted scientists and health officials, at both the national and state levels, to assess the radon hazard and determine the extent of the problem.

Radon forms as a product of radioactive decay. The most common source of radon is decay of uranium (^{238}U) to stable lead (^{206}Pb) (figure 24). During this decay sequence, new isotopes form which emit radiation. One such isotope, radon (^{222}Rn), forms directly from decay of radium (^{226}Ra). Two other isotopes of radon (^{219}Rn and ^{220}Rn) also occur in nature and may contribute to the indoor-radon problem. However, ^{222}Rn is the most abundant of the radioactive radon isotopes, has the longest half-life at 3.825 days, and is considered to be the most significant contributor to the indoor-radon hazard. Subsequent references to radon imply ^{222}Rn derived from the ^{238}U decay chain.

In nature, radon is found in small concentrations in nearly all rocks and soils. The exposure to the hazard, in most cases, depends on factors such as geology, foundation condition, building ventilation, construction material, and occupant lifestyle. Tanner (1986) suggests four prerequisites for elevated indoor-radon concentrations. The home must: (1) be built on ground that contains a radon source material, (2) have underlying soils that promote easy movement of radon, (3) have porous building materials or openings below grade, and (4) have a lower atmospheric pressure inside than outside.

Several geologic factors affect the radon hazard. The first is the distribution of uranium-enriched rock and soil. Granite, metamorphic rocks, some volcanic rocks, and black, organic-rich shales are generally associated with indoor-radon hazards. Once uranium is present in a rock or soil, other factors can enhance or impede radon production and movement, including permeability and water saturation (Tanner, 1964, 1980; Barretto, 1975). A high permeability enhances radon movement by allowing the gas to diffuse through the rock or soil. Water saturation inhibits radon migration by filling pore spaces and restricting the flow of soil gas (Tanner, 1980). Although radon may move with the water, the flow of water through geologic materials is usually much slower. However, water does provide an effective means to carry radon from its rock source (Tanner, 1980). Where domestic water sources contain high levels of radon, they may contribute to indoor-radon levels (Vitz, 1989).

Radon is highly mobile and can find its way into buildings through small basement cracks or other foundation penetrations such as utility pipes (figure 25).

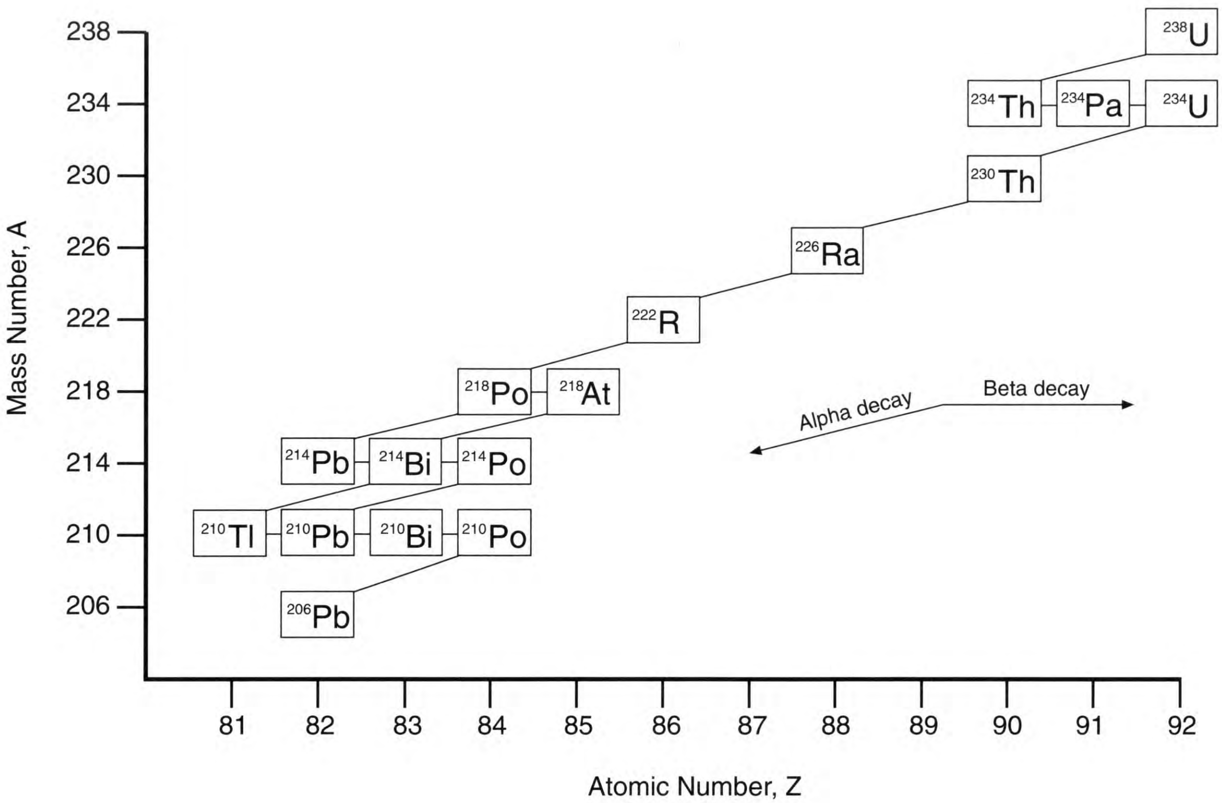


Figure 24. Uranium (^{238}U) decay series. Radon (^{222}Rn) is derived from Radium (^{226}Ra) and is the only isotope in the series that is a gas. Because it is inert, radon has the ability to move with air or water without participating in chemical reactions (modified from Durrance, 1986).

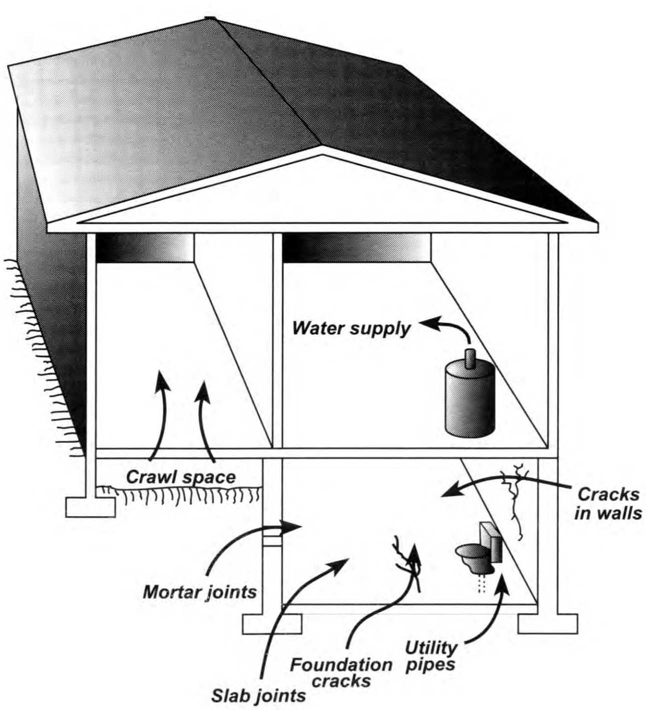


Figure 25. Various pathways for radon to enter a home. Most of the entry routes are in the basement, because that is the part of the house with the greatest surface area exposed to the surrounding soil (modified from U.S. Environmental Protection Agency, 1992).

Although outdoor radon concentrations never reach dangerous levels because air movement dissipates the gas, people can be subject to a radon hazard in buildings that have poor air circulation. Maximum radon concentrations are often found in basements or low crawl spaces (Fleischer and others, 1982), which are in contact with the ground and usually poorly ventilated.

Radon concentration is measured in picocuries per liter of air (pCi/L), which represents a decay of 2 radon atoms per minute per liter of air. Most buildings throughout the United States usually have concentrations less than 3 pCi/L (Nero and others, 1986). The U.S. Environmental Protection Agency (EPA) (1992) recommends that action be taken to reduce indoor levels when they exceed 4 pCi/L.

Changes in building practices over the past 15 years have contributed to the radon problem. Since the 1973 oil embargo, conservation of non-renewable energy resources has been a national goal through energy-efficient practices. Although the building industry has made structures more energy efficient, they have not improved ventilation systems to accommodate restricted natural air flow. Buildings constructed before 1973, including single-family homes, often did not use energy-efficient measures and allowed indoor air to escape through above-grade joints and uninsulated walls and attics. Energy-efficient homes and buildings prevent the loss of indoor air to the outside. Studies have shown that newer, energy-efficient buildings with under-designed ventilation systems generally have higher indoor-radon levels compared with older, conventional buildings (Fleischer and others, 1982; Nero and others, 1982). Although many buildings in Tooele Valley are older than 25 years, the valley is experiencing growth and new energy-efficient buildings are being constructed. Thus, the radon problem in Tooele Valley will likely worsen. The risk from radon is lower in the WDHIA due to lack of development.

Effects, Distribution, and Reduction

Radon and other sources of natural radiation are widespread in low levels, but most natural background radiation is not a health threat. Most buildings throughout the United States contain some radon, but concentrations are usually less than 3 pCi/L. Long-term exposure to these levels is generally considered a small health risk. However, health officials believe breathing elevated levels of radon over time increases a person's risk of lung cancer because of internal radiation damage to the lungs from decaying radon and radon progeny (Jacobi and Eisfeld, 1982; National Council on Radiation Protection and Measurements, 1984a, 1984b; Samet, 1989; figure 26).

The greater your exposure to radon, the greater your risk of developing lung cancer. The EPA estimates that from 8,000 to 40,000 Americans will die each year from lung cancer caused by long-term radon inhalation (Schmidt and others, 1990). If you regularly drink household water containing radon, it is not considered a health risk. Waterborne radon is a problem only when the radon is released from the water and enters the household air. Estimates of the contribution of radon in water to airborne radon range from 1 to 2.5 pCi/L in air for every 10,000 pCi/L in water.

Inhalation of radon is not thought to be the primary source of internal radiation because radon atoms are inert and do not attach themselves to the lining of the lungs. In addition, most radon atoms are exhaled before they can decay and emit dangerous alpha particles to lung tissue. The radioactive isotopes formed from radon decay are of more concern because they are not inert and readily attach themselves to the first charged surface they come in contact with (typically dust or smoke in the air). People who smoke place the occupants of a building at greater risk because the smoke increases the number of airborne particles, to which radon progeny then become attached and are inhaled into the lungs. Once dust or smoke particles with attached radon progeny become lodged in the lungs, these particles allow tissue to be directly bombarded and damaged by energetic alpha particles as radioactive decay occurs.

Previous investigators mapped the extent of radon hazards statewide (Black, 1993) and in Tooele Valley (Black and Solomon, 1996). Figure 27 shows results of a detailed radon-hazard-potential study in Tooele Valley. Figure 28 is a portion of the statewide radon-hazard-potential map for Tooele County showing the WDHIA; no detailed study was done for the WDHIA. Hazard potential on these maps was determined from geologic factors such as uranium concentration, soil permeability, and depth to shallow ground water (Black, 1993; Black and Solomon, 1996). Three categories of hazard potential are mapped: (1) high, areas where all geologic factors contribute to elevated indoor-radon levels; (2) moderate, areas where some geologic factors contribute to elevated indoor-radon levels; and (3) low, areas where no geologic factors contribute to elevated indoor-radon levels.

Detailed studies by the UGS show the radon-hazard potential of Tooele Valley is mostly moderate (figure 26). Scattered areas of high hazard potential are where deep ground water and highly permeable soils with moderate-to-high uranium levels are found. Areas of low hazard potential are in the northern part of the valley in low-lying areas surrounding Great Salt Lake, due to shallow ground water and impermeable, clay-rich soils. The Utah Division of Radiation Control (UDRC) measured indoor-radon concentrations in 70 homes in Tooele Valley, most of which were in moderate-hazard areas (Black and Solomon, 1996). Mean concentration of these measurements was 2.2 pCi/L (81 Bq/m^3) (Black and Solomon, 1996). The highest measured indoor-radon concentration in Tooele Valley was 8.0 pCi/L (296 Bq/m^3), with 18.6 percent of the measurements greater than or equal to 4 pCi/L (148 Bq/m^3) (Black and Solomon, 1996).

The radon-hazard potential of the WDHIA is also mostly moderate (figure 27). Isolated areas of high hazard potential are found in the Cedar Mountains, on the eastern edge of the WDHIA, and in the Grayback Hills. Deep ground water and highly permeable soils with moderate to high uranium levels are found in these areas. The hazard potential is low in the Great Salt Lake Desert, on the western edge of the WDHIA, where shallow ground water and impermeable, clay-rich soils are found. No indoor-radon concentrations have been measured in the WDHIA. Although radon emanation from low-level nuclear waste repositories such as Vitro and Envirocare is unknown, high on-site levels have been found at similar facilities (Tomczak and others, 1993).

RADON RISK IF YOU SMOKE

(If you are a former smoker, your risk may be lower)

Radon Level	*If 1,000 people who smoked were exposed to this level over a lifetime...*	*The risk of cancer from radon exposure compares to...*	*WHAT TO DO: Stop smoking and...*
20 pCi/L	*About 135 people could get lung cancer*	← *100 times the risk of drowning*	*Fix your home*
		← *100 times the risk of dying in a home fire*	
10pCi/L	*About 71 people could get lung cancer*		*Fix your home*
8 pCi/L	*About 57 people could get lung cancer*		*Fix your home*
4 pCi/L	*About 29 people could get lung cancer*	← *100 times the risk of dying in an airplane crash*	*Fix your home*
2 pCi/L	*About 15 people could get lung cancer*	← *2 times the risk of dying in a car crash*	*Consider fixing between 2 and 4 pCi/L*
1.3 pCi/L	*About 9 people could get lung cancer*	*(Average indoor radon level)*	*(Reducing radon levels below 2 pCi/L is difficult)*
0.4 pCi/L	*About 3 people could get lung cancer*	*(Average outdoor radon level)*	

RADON RISK IF YOU DON'T SMOKE

(If you are a former smoker, your risk may be higher)

Radon Level	*If 1,000 people who never smoked were exposed to this level over a lifetime...*	*The risk of cancer from radon exposure compares to...*	*WHAT TO DO:*
20 pCi/L	*About 8 people could get lung cancer*	← *The risk of being killed in a violent crime*	*Fix your home*
10pCi/L	*About 4 people could get lung cancer*		*Fix your home*
8 pCi/L	*About 3 people could get lung cancer*	← *10 times the risk of dying in an airplane crash*	*Fix your home*
4 pCi/L	*About 2 people could get lung cancer*	← *The risk of drowning*	*Fix your home*
2 pCi/L	*About 1 person could get lung cancer*	← *The risk of dying in a home fire*	*Consider fixing between 2 and 4 pCi/L*
1.3 pCi/L	*Less than 1 person could get lung cancer*	*(Average indoor radon level)*	*(Reducing radon levels below 2 pCi/L is difficult)*
0.4 pCi/L	*Less than 1 person could get lung cancer*	*(Average outdoor radon level)*	

Figure 26. *Radon risk evaluation chart. The U.S. Environmental Protection Agency (1992) has developed this chart to provide comparable risks for people to evaluate their personal risk from radon.*

Figures 26 and 27 are generalized and show only the relative geologic potential for radon hazards. Actual indoor-radon levels may vary, and the map should not be used to predict indoor-radon levels. Indoor testing is the only reliable way to determine if a radon hazard exists, and is recommended in all areas regardless of radon-hazard potential. New construction in high hazard-potential areas may also wish to incorporate radon-reduction techniques.

If elevated indoor-radon levels are discovered in a

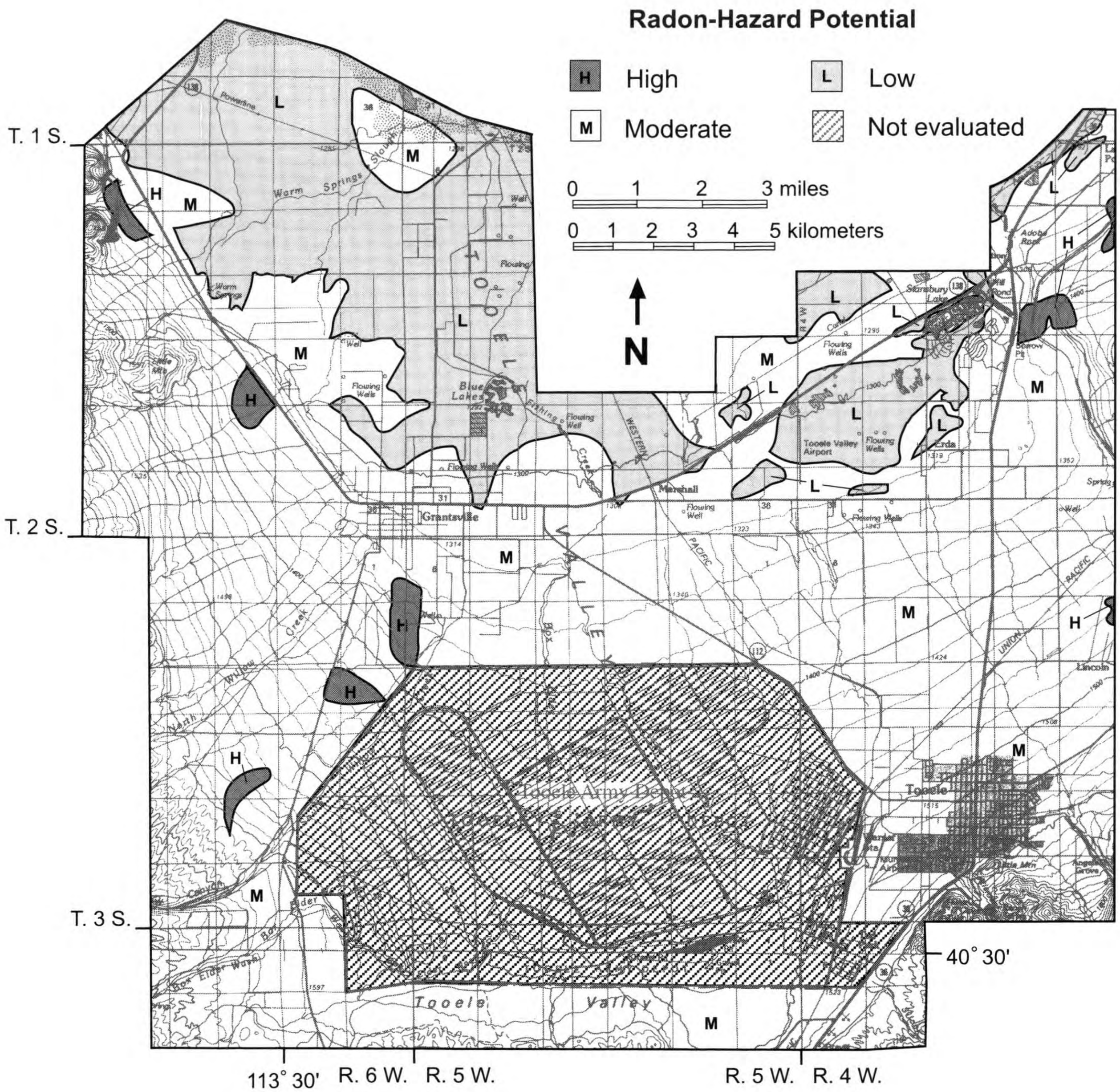


Figure 27. Radon-hazard potential of Tooele Valley based on geologic factors (modified from Black and Solomon, 1996).

home, a number of methods can be considered for reducing levels. These methods fall into two categories: (1) preventing radon from entering the house, and (2) removing radon (or decay products) after entry. The specific method chosen will depend upon the initial radon concentration, house design, and type of construction.

Some actions may be taken immediately, and can be done quickly with a minimum of expense. Discourage smoking inside a home; this not only reduces the risk from radon exposure but also the overall chance of developing lung cancer. Radon collects in the basement and low areas of a home; spending less time in these areas of higher radon concentrations will reduce the risk. Ventilation can be improved by opening windows and turning on fans, but is not always desirable during cold winter months.

Although immediate actions are effective, they are not long-term solutions. The selection of permanent radon-reduction methods requires identification of radon entry routes and driving forces, and diagnostic testing to aid in the selection of the most effective method. Professional assistance is often required. Five classes of permanent methods exist: (1) increased ventilation through natural means (such as opening windows) or ventilators; (2) sealing to restrict movement of radon from soil into the house and gas flow through entry routes (known as "closure"); (3) soil ventilation to withdraw radon-contaminated soil gas and divert it outdoors; (4) house pres-

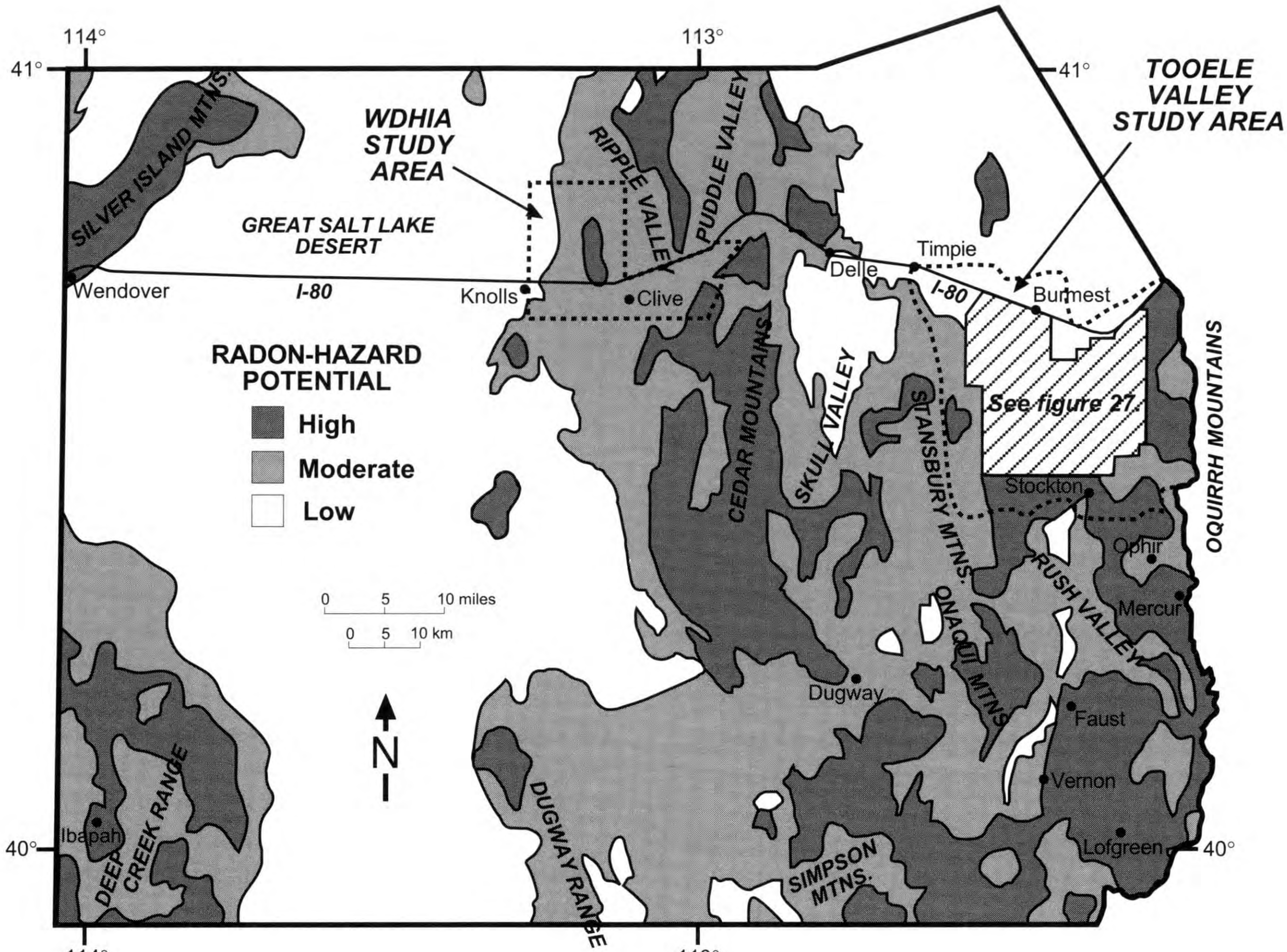


Figure 28. *Radon-hazard potential of Tooele County from geologic factors (modified from Black, 1993).*

sure adjustments to restrict flow of soil gas into the house by altering pressure differentials between the house and soil; and (5) air cleaning to remove radon decay products (which are solid particles) from the air after radon entry (U.S. Environmental Protection Agency, 1989). Once appropriate radon-reduction methods are chosen and implemented, diagnostic tests should also be conducted to ensure that radon levels have been reduced.

An effective method of hazard reduction is to prevent radon from entering the structure. Prevention is advisable in new construction, particularly in high hazard areas. New design and construction may incorporate methods to restrict radon entry by minimizing: (1) soil gas entry pathways; and (2) indoor-outdoor pressure differences, because these differences are the driving force for soil gas to enter a home (Osborne, 1988). Features can also be incorporated during construction that facilitate radon removal, using methods similar to the radon-reduction methods discussed above.

If no measured problem with airborne radon is found in a home, homeowners generally do not need to test household water for radon. If indoor levels are high, low-cost water test kits are available from commercial laboratories. Testing of water from municipal water supplies is generally not necessary; radon contamination usually only occurs in well water and is not common.

If a water test indicates radon problems, the radon may be either removed from the air after it has left the water or from the water before it reaches indoor air (U.S. Environmental Protection Agency, 1987). Good ventilation of bathrooms, laundry rooms, and kitchens, particularly during periods of water use, may be adequate to remove radon from indoor air. Methods to remove radon from water include: (1) storing water for several days to allow radon time to decay, which may require large storage tanks; (2) home aeration systems that spray water through an air-filled chamber and use fans to remove the contaminated air; and (3) devices which use granular activated charcoal to remove radon from water. Activated charcoal devices are presently the least costly alternative for homes using their own wells and, to date, the most extensively tested and used method.

Radon can be measured with both short-term and long-term passive detectors and electronic instruments. Some detectors can be placed by homeowners, whereas others require professional installation. Because most people want information quickly, they often select short-term monitoring methods. A short-term measurement is one conducted for a period less than three months. However, long-term monitoring, typically for a twelve-month period, provides more reliable information.

Measurements taken over a few days or on a single day provide only a snapshot of indoor-radon levels for that particular time. Radon emissions from the ground, and resultant indoor-radon levels, can fluctuate daily, weekly, and monthly because of atmospheric changes. In addition, concentrations fluctuate seasonally because building ventilation is less in winter than summer, and indoor heating and air conditioning affect concentrations. A longer period of monitoring is recommended to smooth out short-term fluctuations. This provides a reliable picture of the yearly average concentration. The UDRC provides information on types of radon detectors available, their advantages and disadvantages, and comparative cost.

Radon measurement protocols suggested by the EPA (U.S. Environmental Protection Agency, 1992) attempt to assure accuracy and consistency of data. The protocols were developed to balance the need for quick results with measurements that best reflect long-term indoor-radon levels. To accurately determine indoor-radon levels throughout a home, long-term monitoring is needed on each floor. However, short-term screening measurement which follows EPA protocol (closed-house conditions) may be conducted in the lowest living area to determine if additional testing is required. Charcoal canisters are commonly used for short-term measurements; alpha-track detectors are commonly used for long-term measurements.

EPA protocols emphasize immediate follow-up testing in homes with screening measurements exceeding 4 pCi/L (U.S. Environmental Protection Agency, 1992). Occupants of homes with radon levels exceeding 4 pCi/L should take action to reduce radon concentrations. Additional testing is not needed if a short-term screening measurement is less than 4 pCi/L and, although a small health risk is present, remediation is unnecessary. If a result is greater than 4 pCi/L and less than 20 pCi/L, a 12-month follow-up measurement is recommended. If retesting confirms screening measurements, remediation should be done within the next few years. If a screening measurement is from 20 to 200 pCi/L, a 3-month follow-up measurement is recommended. If the measurement is confirmed, remediation should take place within a few months. If a screening measurement exceeds 200 pCi/L, retest immediately. If confirmed, remediation should take place within weeks.

SUMMARY

Geologic mapping provides basic data on which geologic hazards can be identified. Surficial geology of Tooele Valley and the WDHIA has been compiled for this study (at a scale of 1:100,000) from existing maps. Unconsolidated material in these study areas consists primarily of sediments deposited during various stages of Pleistocene Lake Bonneville, and alluvium deposited after lake retreat. Bedrock is primarily Paleozoic sedimentary and Tertiary volcanic rocks. Based on the geologic mapping, derivative hazard maps were produced at the same scale as the geologic map. The hazard maps are intended to inform citizens of their risk and provide a tool for responsible development. However, the maps are general and only indicate potential hazards that may be encountered. The risk posed by these hazards should be evaluated in more detailed, site-specific studies.

Mapped geologic hazards in the two study areas include surface fault rupture; liquefaction susceptibility; landslide susceptibility; lake flooding, ponding, and sheet flooding; susceptibility of slopes to generate debris slides and debris flows; debris-flow deposition, debris flooding, and stream flooding; rock fall; shallow ground water; problem soils; and radon. These hazards are shown on plates 2 through 6 and figures 26 and 27. Hazards not mapped are mainly earthquake hazards that were difficult to quantify or that lack sufficient data, including ground shaking, tectonic subsidence, ground failure due to sensitive clays, subsidence in granular materials, and earthquake-related flood hazards. Previous studies evaluated radon hazards in Tooele Valley, but no such studies have been done for the WDHIA. Text in the geologic hazards section summarizes the characteristics and effects of all potential geologic hazards, and includes commonly used hazard-reduction techniques.

Tooele Valley and the WDHIA are prone to many earthquake hazards. Both areas are subject to ground shaking, which is typically the most widespread and damaging earthquake hazard. However, the likelihood of strong ground shaking (and hazards resulting from ground shaking) is greater in Tooele Valley. Surface fault rupture is a hazard in Tooele Valley, but there are no known active faults (and therefore little potential for surface rupture) in the WDHIA. Both areas also have geologic conditions susceptible to liquefaction, and lesser understood earthquake hazards such as ground failure in sensitive clays and subsidence in granular materials. Tooele Valley could also be subject to hazards from tectonic subsidence, flooding from seiches in Great Salt Lake, surface drainage disruptions, and increased ground-water discharge.

Tooele Valley and the WDHIA are also prone to other geologic hazards. Although landslides are not numerous in the mountains surrounding Tooele Valley or in the WDHIA, the clay-rich Mississippian Manning Canyon Shale (which is particularly susceptible to landsliding) is found in the Stansbury and Oquirrh Mountains in southern Tooele Valley. Existing landslides are in this unit in Tooele Valley; there are no existing landslides in the WDHIA. Tooele Valley is susceptible to debris flows, debris floods, and stream flooding from the steep mountains surrounding the valley, but susceptibility to these hazards is lower in the WDHIA due to subdued topography. The potential for rock-fall hazards is greatest along the Oquirrh Mountains in eastern Tooele Valley, but a lesser hazard also exists along the Stansbury Mountains and South Mountain in Tooele Valley, and along the Grayback Hills in the WDHIA. Tooele Valley is subject to flooding along the shores of Great Salt Lake and Rush Lake, and both areas are subject to ponding and sheet-flooding hazards. Problem soils are also found in both study areas. The hazard potential from radon is mostly moderate in both study areas.

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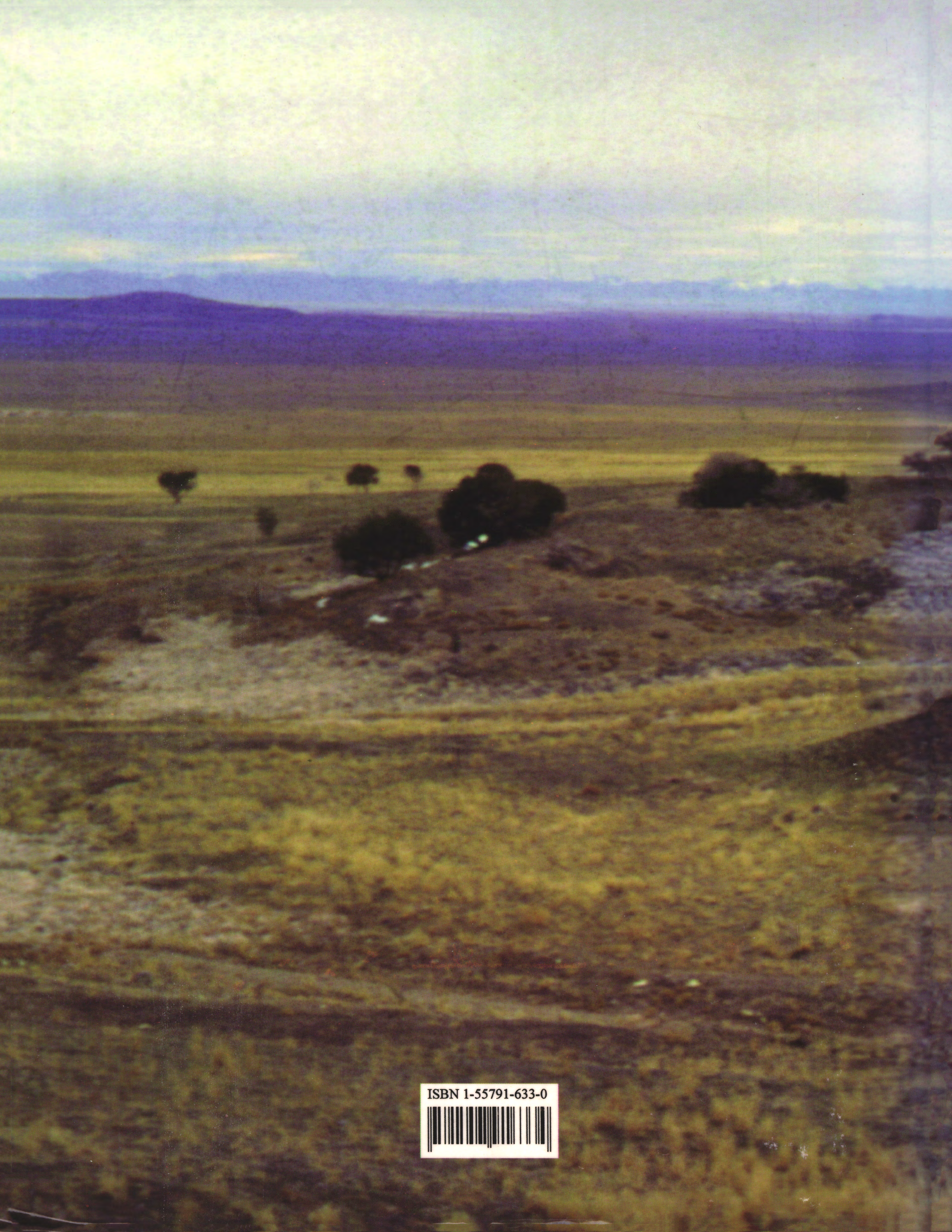

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Geology and geologic hazards of Tooele Valley and the West Desert Hazardous Industry Area, Tooele County, Utah

Special Study 96
1999
UTAH GEOLOGICAL SURVEY
a division of
UTAH DEPARTMENT OF NATURAL RESOURCES

SCALE 1:100,000

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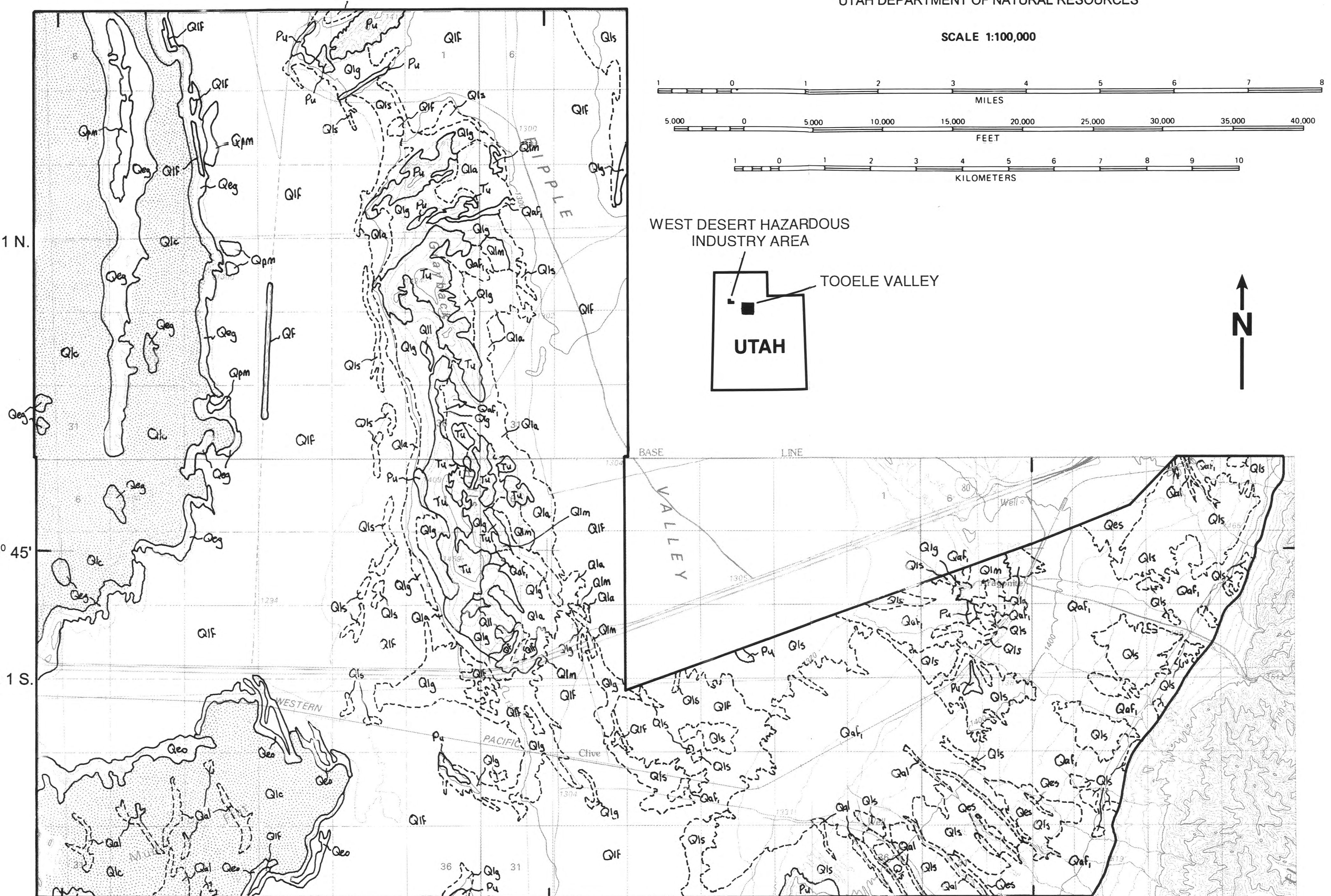
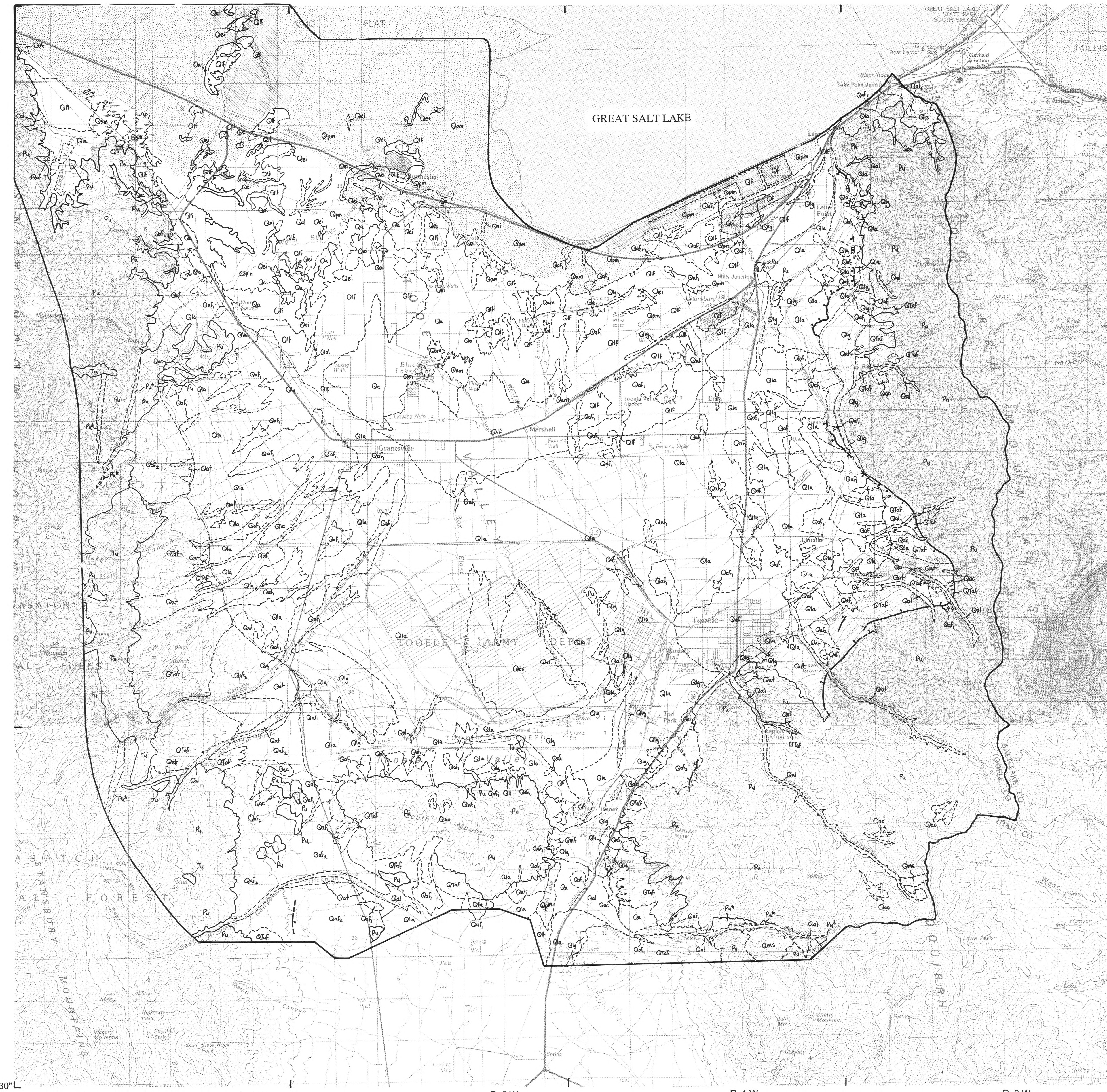




EXPLANATION

DESCRIPTION OF MAP UNITS

Qf	Fill (Holocene).
Qal	Stream alluvium (Holocene and latest Pleistocene).
Qam	Alluvial-mud deposits (Holocene).
Qpm	Playa mud (Holocene).
Qsm	Spring deposits (Holocene).
Qeg	Gypsum dunes (Holocene and latest Pleistocene).
Qei	Silt dunes (Holocene and latest Pleistocene).
Qeo	Oolite dunes (Holocene and latest Pleistocene).
Qes	Silica dunes (Holocene and latest Pleistocene).
Qa	Undifferentiated alluvium (Holocene and latest Pleistocene).
Qac	Alluvium and colluvium (Holocene and latest Pleistocene).
Qmt	Talus (Holocene and latest Pleistocene).
Qaf_1	Younger alluvial-fan deposits (Holocene and latest Pleistocene).
Qla	Undifferentiated lacustrine and alluvial deposits (Holocene and latest Pleistocene).
Qlg	Lacustrine gravel (Holocene and latest Pleistocene).
Qat	Alluvial-terrace deposits (Holocene and latest Pleistocene).
Qmf	Debris-flow deposits (Holocene and latest Pleistocene).
Qms	Landslide deposits (Holocene and latest Pleistocene).
Qlm	Lacustrine marl (Holocene and latest Pleistocene).
Qlc	Lacustrine clay (Holocene and latest Pleistocene).
Qlf	Lacustrine mud (Holocene and latest Pleistocene).
Qll	Lacustrine lagoonal deposits (latest Pleistocene).
Qls	Lacustrine sand (latest Pleistocene).
Qaf_2	Intermediate-age alluvial-fan deposits (latest to middle Pleistocene).
QTaf	Older alluvial-fan deposits (early Pleistocene and Pliocene).
Tu	Undifferentiated Tertiary bedrock.
Pu	Undifferentiated Paleozoic bedrock.
Pu*	Undifferentiated Paleozoic bedrock susceptible to landsliding (primarily the Mississippian Manning Canyon Shale).

MAP SYMBOLS

CONTACT, dashed where location approximate.

HIGH-ANGLE FAULT, dashed where location approximate, bar and ball on downthrown side.

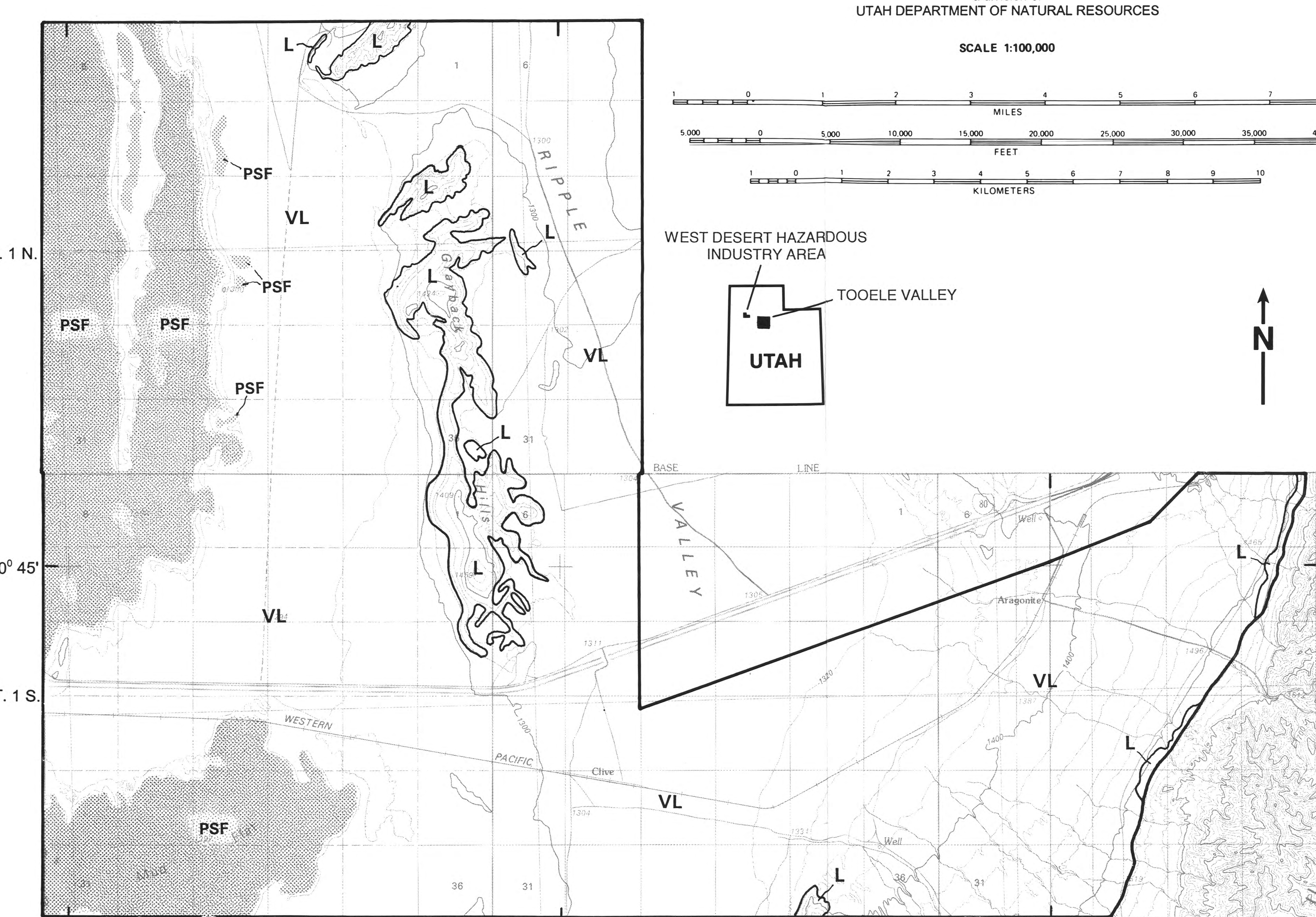
STUDY AREA BOUNDARY

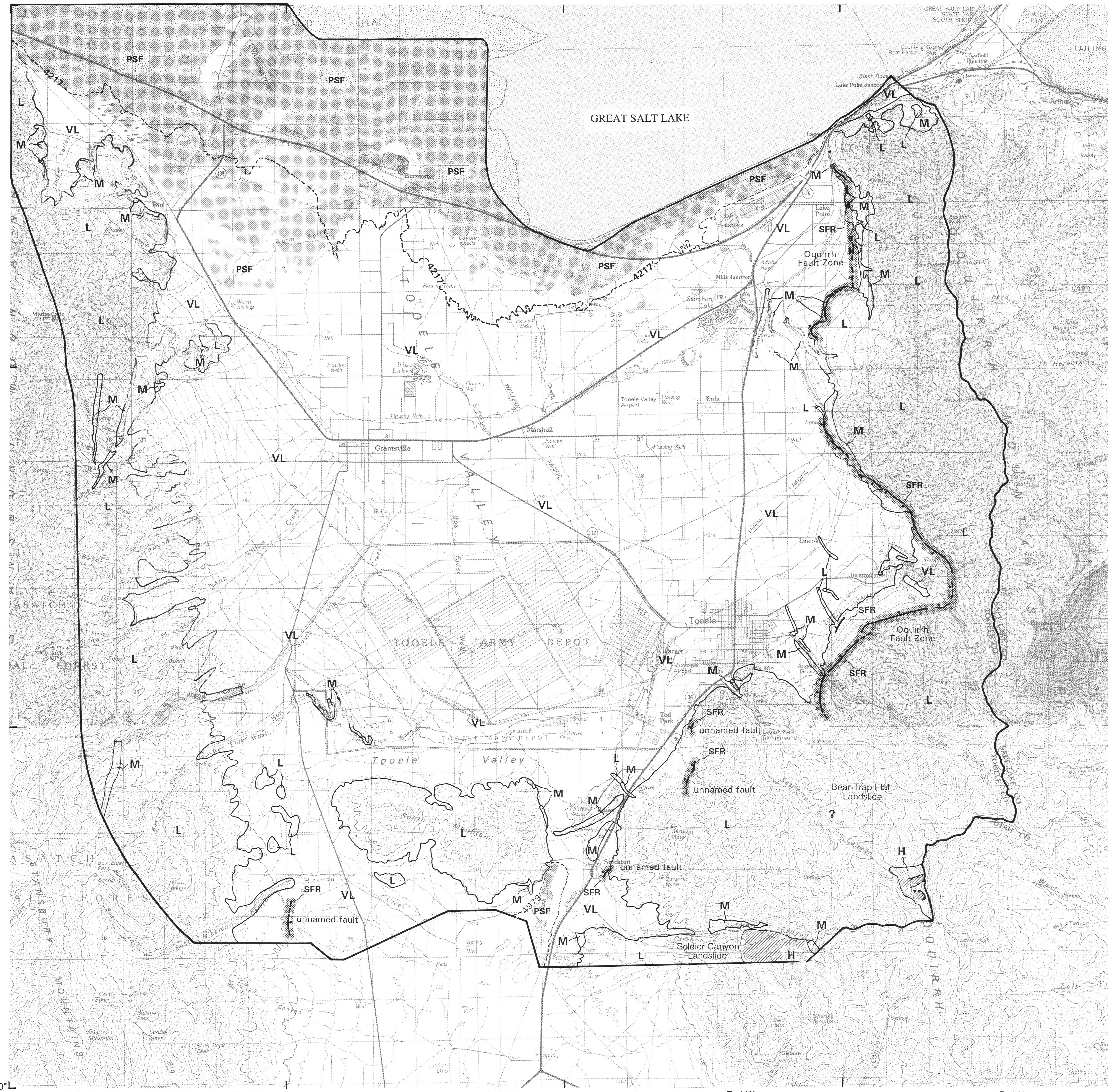
Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 1. Geology.

Geology and geologic hazards of Tooele Valley and the West Desert Hazardous Industry Area, Tooele County, Utah
Special Study 96
1999
UTAH GEOLOGICAL SURVEY
a division of
UTAH DEPARTMENT OF NATURAL RESOURCES

SCALE 1:100,000





EXPLANATION

Deep-seated landslide susceptibility

- H — High; includes existing landslides (crosshatched).
- M — Moderate.
- L — Low.
- VL — Very low.
- Landslide scarps.

Surface fault rupture

- Main trace of active faults with evidence of possible Quaternary movement (within the past 1.6 million years). Dashed where location approximate, bar and ball on downthrown side.
- SFR — Subject to possible surface fault rupture.

Lake flooding, and ponding and sheet flooding

- 4217' — Boundary of the "Beneficial Development Area" (4,217-foot [1,285-m] contour). Areas below this elevation around Great Salt Lake are subject to lake flooding.
- 4979' — Measured historic high of Rush Lake (4,979-foot [1,518-m] contour. Areas below this elevation around Rush Lake are subject to lake flooding.
- PSF — Subject to ponding and sheet flooding.

This map involves a general evaluation on a regional scale and does not preclude the neccessity for site-specific investigations.

Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 2. Landslide, surface-fault-rupture, lake-flooding, and ponding and sheet-flooding hazards.

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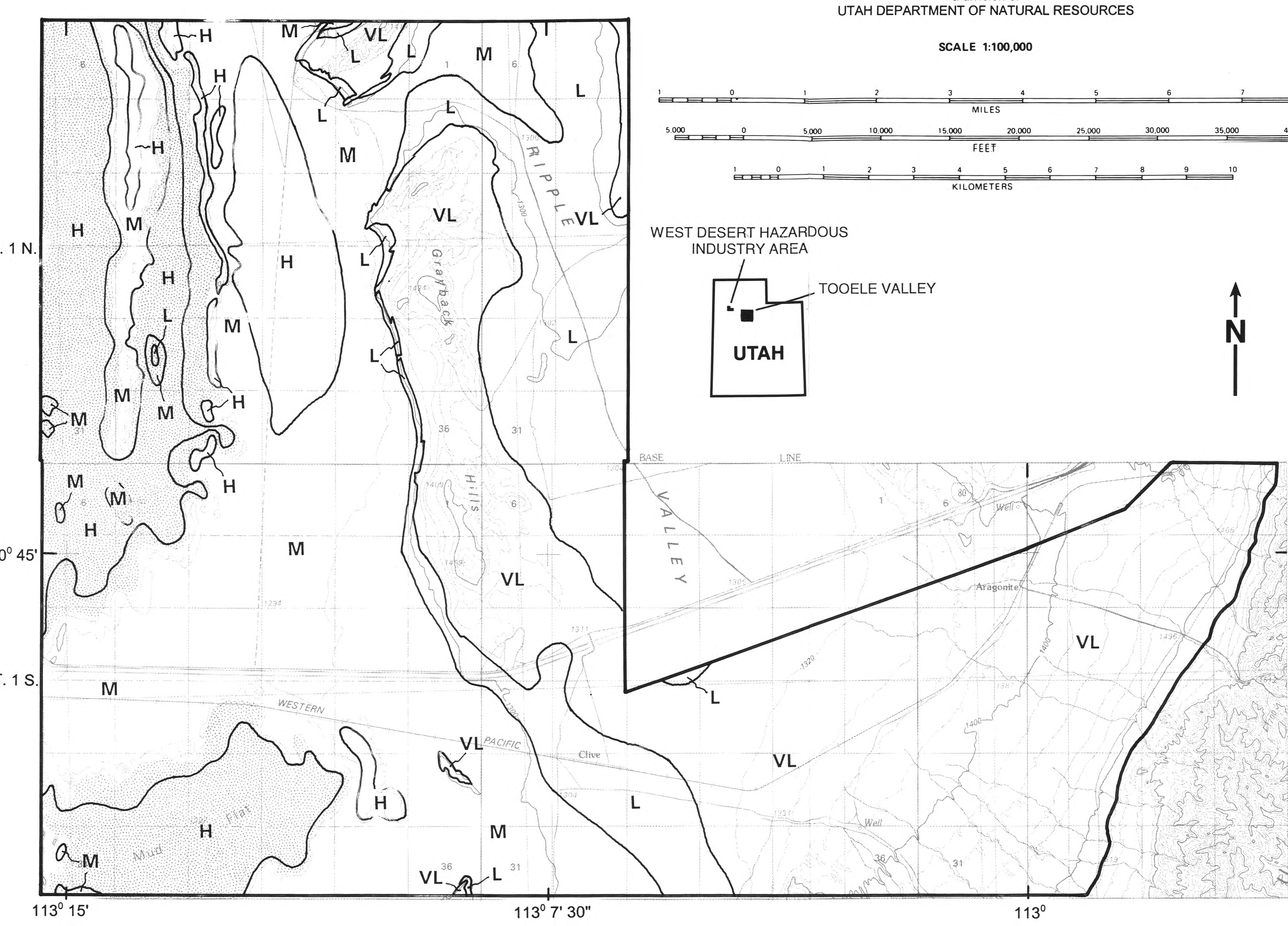

EXPLANATION

H	High; possible susceptible soil conditions and depth to ground water less than 10 feet (3 m).
M	Moderate; possible susceptible soil conditions and depth to ground water from 10 to 30 feet (3-9 m).
L	Low; possible susceptible soil conditions and depth to ground water from 30 to 50 feet (9-15 m).
VL	Very low; rock, unsusceptible soil conditions, or depth to ground water greater than 50 feet (15 m).

Note: this map is not a liquefaction potential map, because it does not consider the probability of earthquake ground shaking needed to cause liquefaction in areas of susceptible soil and ground-water conditions. This map involves a general evaluation on a regional scale and does not preclude the neccessity for site-specific investigations.

Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 3. Liquefaction susceptibility.

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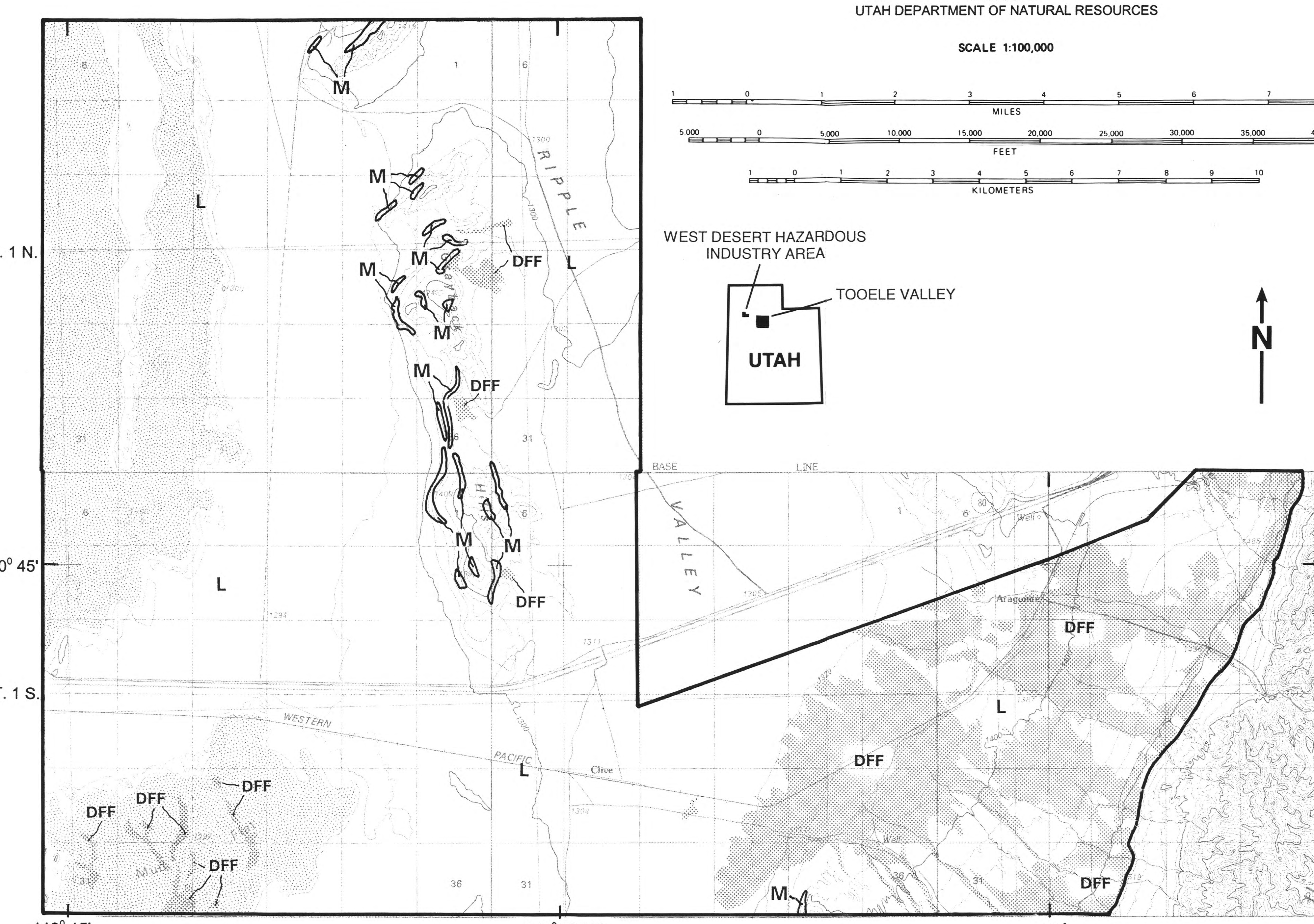

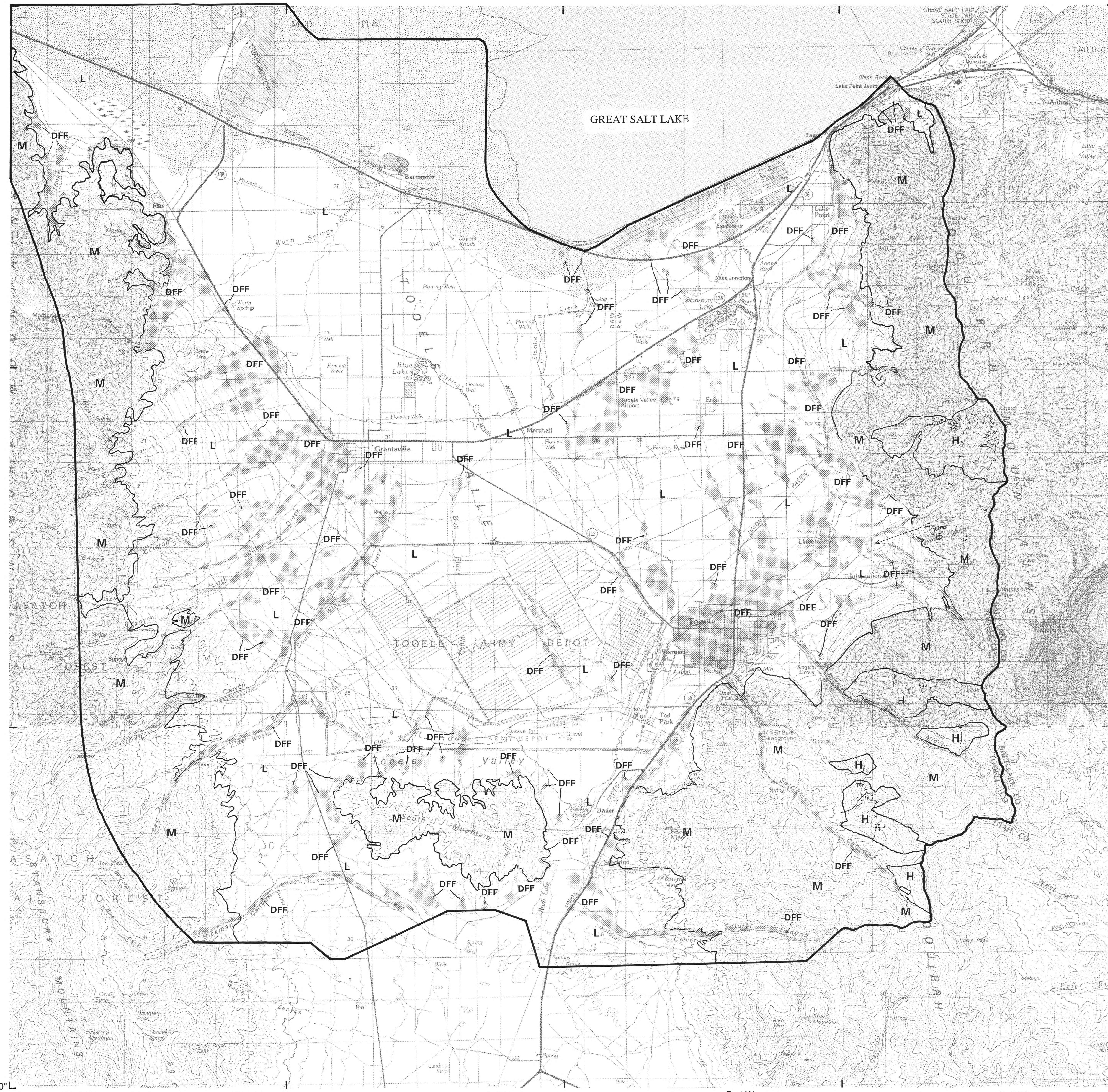

EXPLANATION

Source-area slope-failure susceptibility*

H	High; includes slopes that failed during the 1983-84 wet years.
M	Moderate.
L	Low.

Debris deposition and flood hazard

DFF Possible sediment deposition and flooding from debris flows, debris floods, and stream floods; includes 1983-84 debris deposits.

1983-84 debris flow source scar and travel path

*This map shows susceptibility to shallow slope failures (debris slides and debris flows). See plate 2 for deep-seated landslide susceptibility. This map involves a general evaluation on a regional scale and does not preclude the need for site-specific investigations.

Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 4. Debris-slide, debris-flow, debris-flood, and stream-flood hazards.

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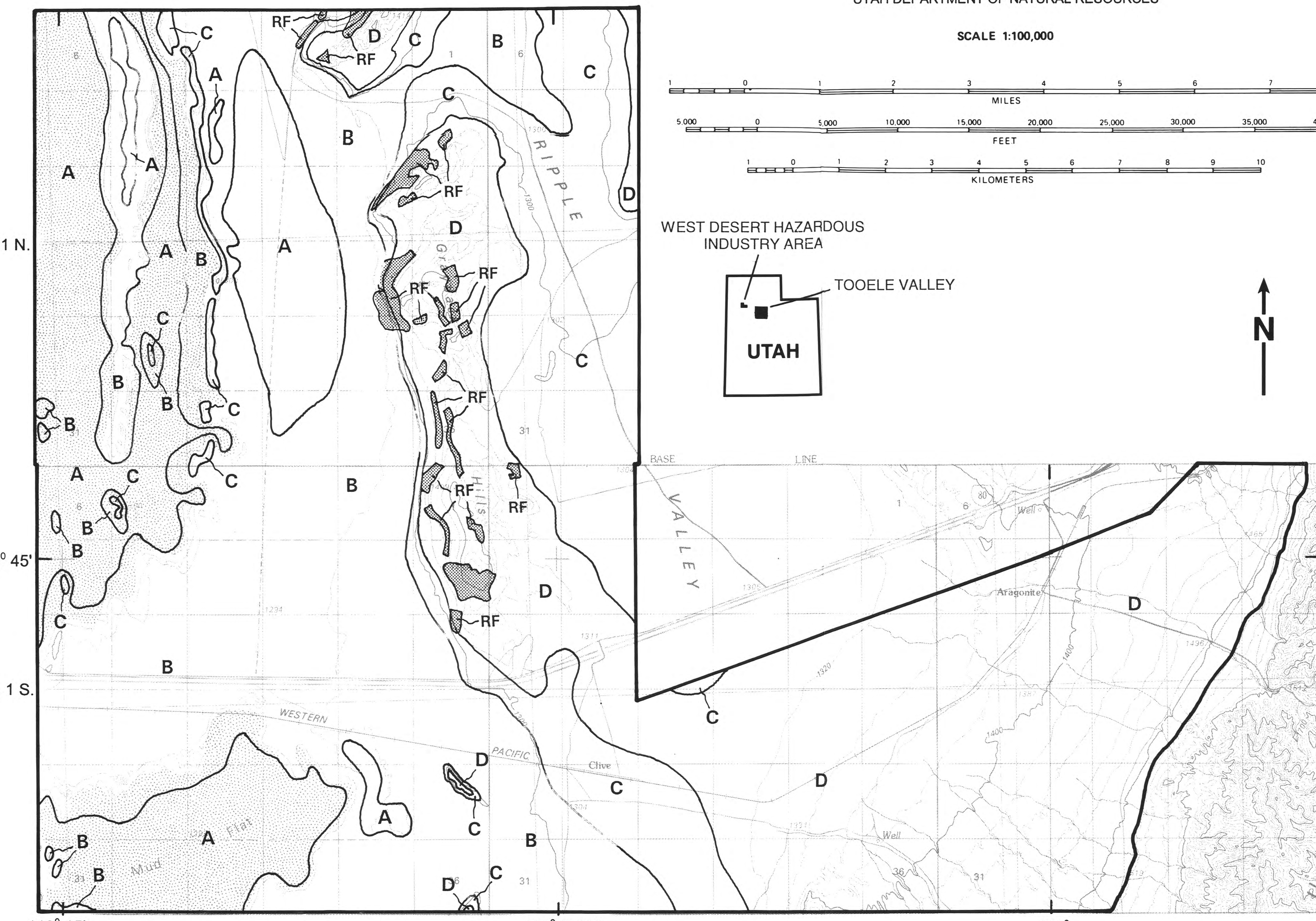

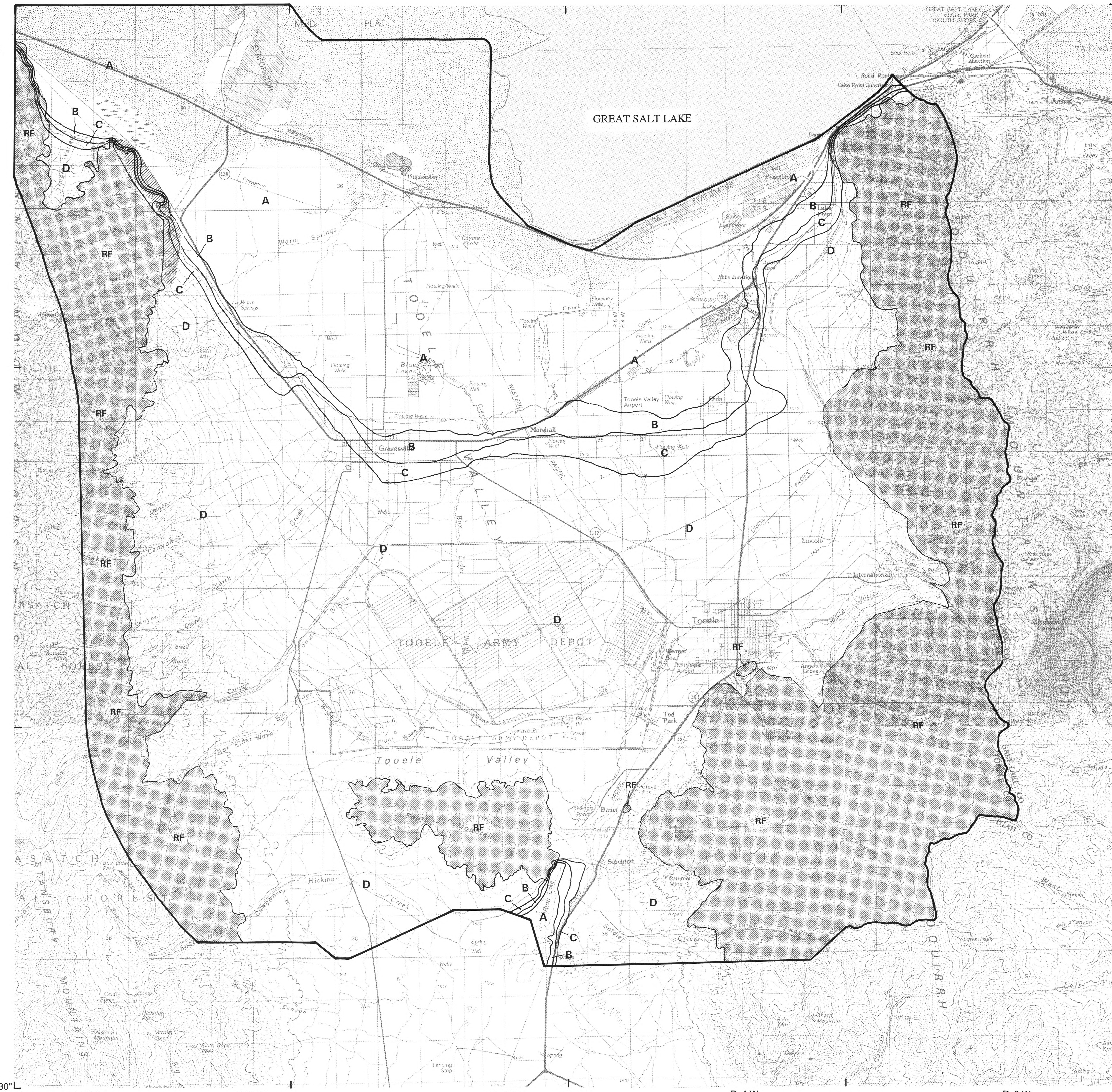

EXPLANATION

Rock Fall

RF — Potentially subject to impact by rock fall (crosshatched area not studied due to mining slope modifications)

Depth to ground water

A — Less than 10 feet (3 m).

B — 10 to 30 feet (3-9 m).

C — 30 to 50 feet (9-15 m).

D — Greater than 50 feet (15 m).

This map involves a general evaluation on a regional scale and does not preclude the neccessity for site-specific investigations.

Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 5. Rock-fall hazard and depth to ground water.

Geology and geologic hazards of Tooele Valley and the West Desert Hazardous Industry Area, Tooele County, Utah

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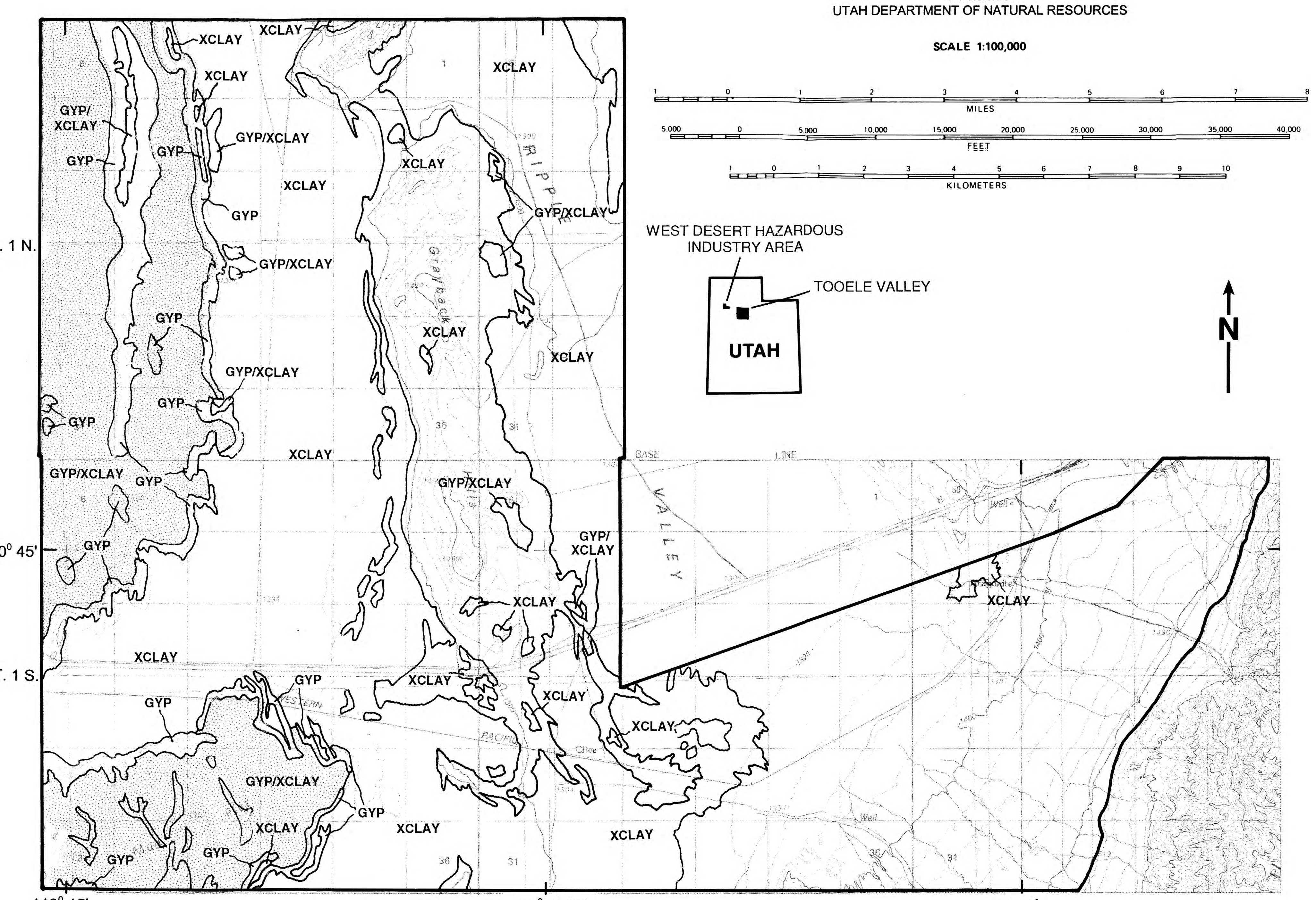

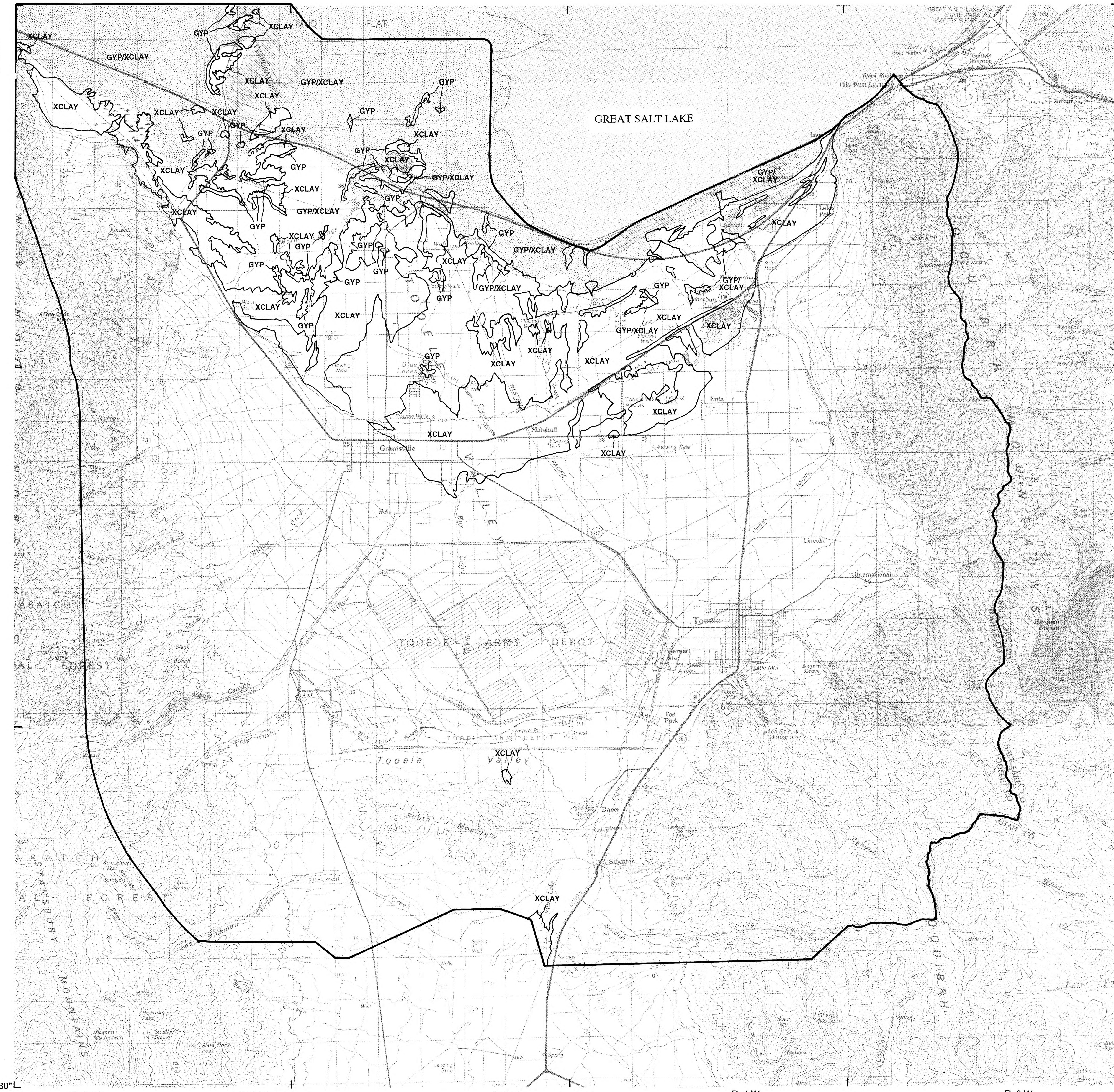

EXPLANATION

GYP	Possible gypsiferous soil.
XCLAY	Possible expansive soil.
GYP/XCLAY	Possible gypsiferous and expansive soils.

This map involves a general evaluation on a regional scale and does not preclude the neccessity for site-specific investigations.

Base map from TOOELE, RUSH VALLEY, and BONNEVILLE SALT FLATS, U.S. Geological Survey 30 x 60 minute topographic quadrangles.

Plate 6. Problem soils.