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Hydrology and simulation of the ground-water flow system in Tooele Valley, Utah

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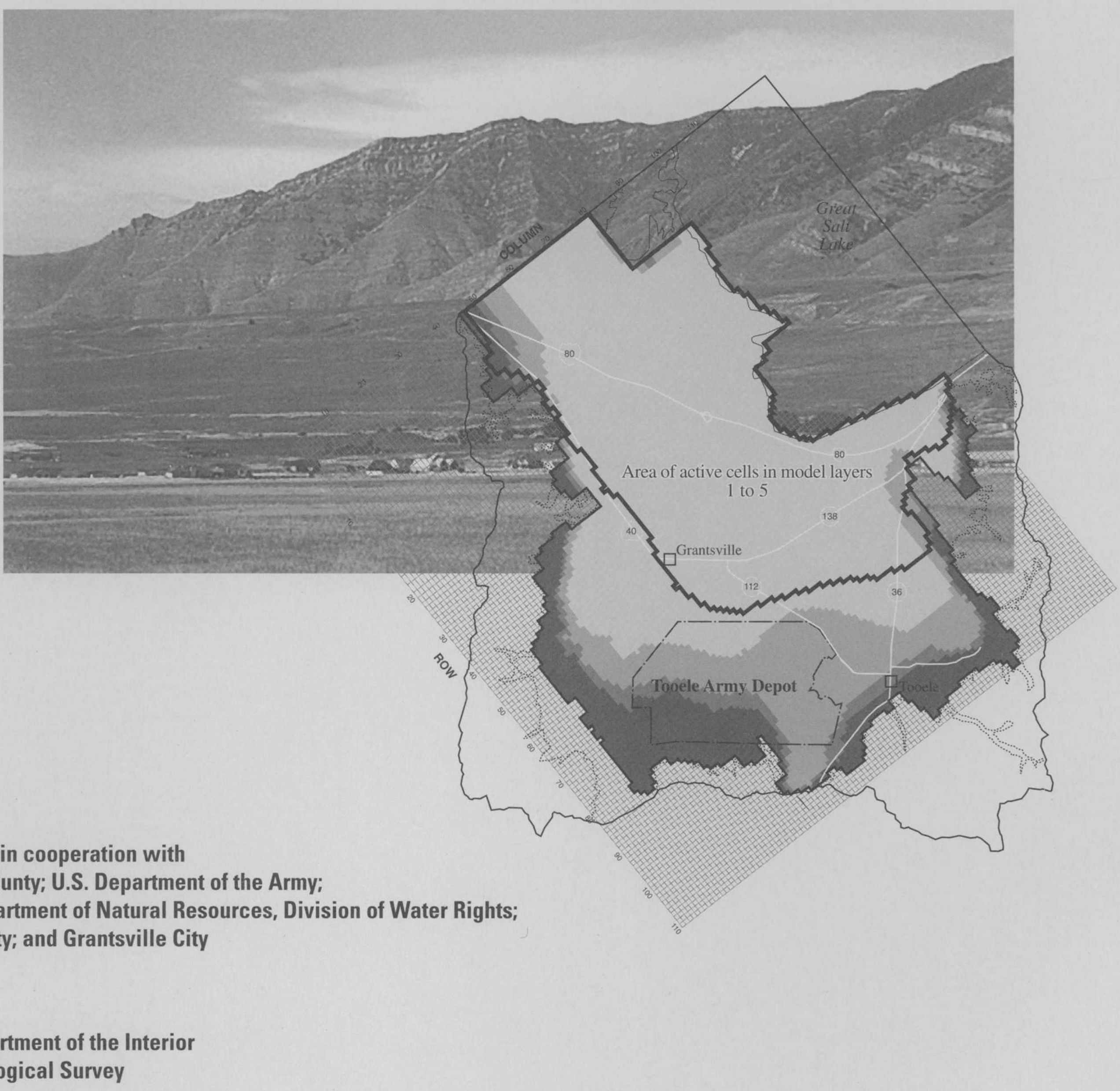
Water-Resources Investigations Report 99-4014



Prepared in cooperation with
Tooele County; U.S. Department of the Army;
Utah Department of Natural Resources, Division of Water Rights;
Tooele City; and Grantsville City

U.S. Department of the Interior
U.S. Geological Survey

HYDROLOGY AND SIMULATION OF THE GROUND-WATER FLOW SYSTEM IN TOOELE VALLEY, UTAH

By P.M. Lambert and B.J. Stolp

U.S. GEOLOGICAL SURVEY

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Salt Lake City, Utah
1999

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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

Multiply	By	To obtain
acre	4,047	square meter
acre-foot (acre-ft)	1,233	cubic meter
	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	0.00003907	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
inch (in.)	25.4	millimeter
	2.54	centimeter
inch per year (in/yr)	2.54	centimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929—a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Chemical concentration is reported only in the metric unit, milligrams per liter (mg/L). Milligrams per liter is a unit expressing the solute per unit volume of water. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentrations in parts per million. The density of water is reported in grams per cubic centimeter (g/cm^3) at 20 degrees Celsius and is obtained by adjusting the specific-gravity measurement with the appropriate conversion factor that corresponds to temperature.

Hydrology and Simulation of the Ground-Water Flow System in Tooele Valley, Utah

By Patrick M. Lambert and Bernard J. Stolp

ABSTRACT

The study described in this report was conducted in cooperation with local, State, and Federal agencies to improve current understanding of the regional ground-water flow system in Tooele Valley. Available data were compiled and analyzed and then used to develop a conceptual model of the flow system. Recharge to the ground-water flow system is mainly from subsurface inflow from consolidated rock and stream-channel deposits, infiltration of precipitation on the valley floor, seepage from irrigated fields, and subsurface inflow from Rush Valley. Long-term average recharge for these sources is 48,000 acre-feet per year, 12,000 acre-feet per year, 10,000 acre-feet per year, and 5,000 acre-feet per year, respectively. Discharge from the ground-water flow system is mainly by pumped and flowing wells, evapotranspiration, springs, subsurface outflow to Great Salt Lake, and shallow drains and ditches. Long-term average discharge to these processes is 26,000 acre-feet per year, 23,000 acre-feet per year, 16,000 acre-feet per year, 3,000 acre-feet per year, and unknown, respectively.

Numerical simulation of the ground-water flow system in Tooele Valley was used to test and refine the conceptual understanding. The numerical simulation was calibrated to match steady-state conditions in 1968 and transient-state conditions during 1969-94. Calibration was achieved by adjusting numerical parameters until a reasonable match between (1) model-computed and measured water levels, (2) model-computed and estimated discharge, and (3) model-computed and measured water-level fluctuation, was attained.

Steady-state calibration resulted in a reasonable match between model-computed and measured water levels throughout most of Tooele Valley. The results of transient-state calibration indicate that the generally observed rising and declining trends in water levels at observation wells are reproduced in many areas of Tooele Valley. The model generally does not simulate the large and abrupt water-level changes from year to year in some areas of Tooele Valley.

INTRODUCTION

Ground water is the most important source of water in Tooele Valley (fig. 1). Ground-water withdrawals are expected to increase in the near future to keep pace with the growing population in the valley. State, county, and city officials, and local water users need information concerning the effects of increased ground-water withdrawals on water levels, flows at natural discharge areas, and movement of poor-quality ground water in the valley. To provide this information, the U.S. Geological Survey, in cooperation with Tooele County; the U.S. Department of the Army; the Utah Department of Natural Resources, Division of Water Rights; Tooele City; and Grantsville City, began a study of the ground-water flow system in Tooele Valley in September 1995. The objectives of this study are to (1) improve current understanding of the regional ground-water flow system in Tooele Valley, and (2) provide information on the effects of regional ground-water flow processes as they relate to ground-water movement in subregional areas of Tooele Valley.

Purpose and Scope

This report describes the current understanding of the regional ground-water flow system in Tooele Valley and the simulation of that system using a three-dimensional, finite-difference, numerical model. Available hydrogeologic data were compiled and analyzed and then used to develop a conceptual model of the ground-water flow system. To define the conceptual model, the following parameters of the ground-water system were estimated: (1) general geometry, (2) hydrologic properties, and (3) recharge/discharge processes and amounts. The conceptual model was then incorporated into a three-dimensional numerical model that simulates ground-water flow in the valley. The

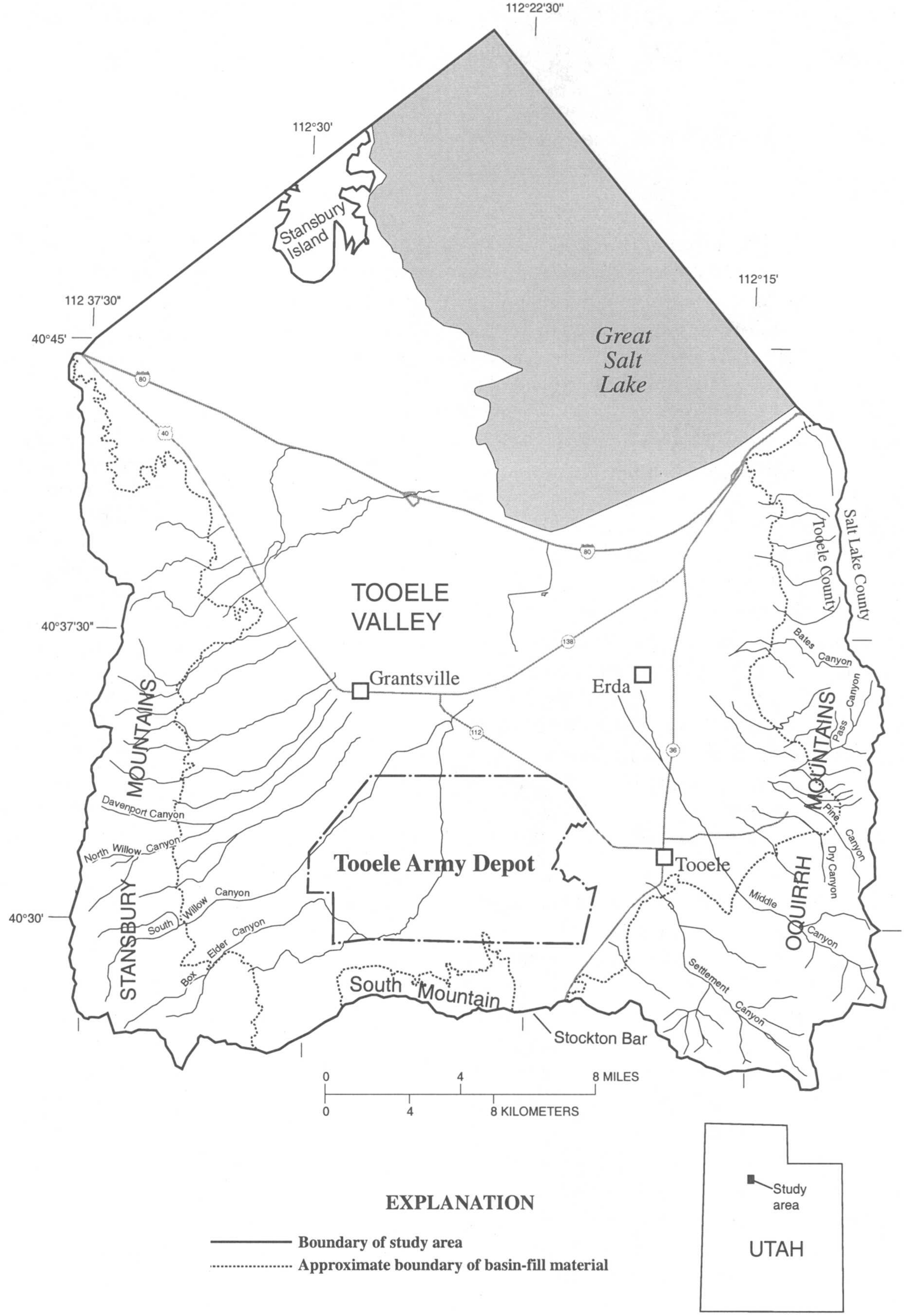


Figure 1. Location of Tooele Valley study area, Utah.

numerical model was used to test and refine conceptual understanding of the system. The conceptual and numerical models described in this report can be used to evaluate regional ground-water movement and to provide information for flow analyses in subregional areas of Tooele Valley.

Previous Work

The first hydrologic reconnaissance of the study area was by Carpenter (1913), who identified and described ground water under confined conditions in alternating layers of coarse- and fine-grained material in the central part of the valley. Carpenter (1913) also identified areas at the margins of the valley, where fine-grained confining sediments generally are absent, as major recharge areas for the confined aquifers in the central part of the valley.

During 1940-42, the U.S. Geological Survey in cooperation with the State of Utah, Office of State Engineer (Thomas, 1946) studied the ground water in Tooele Valley and quantified some of the earlier observations made by Carpenter. Descriptions of sediments penetrated during well drilling were used to identify and correlate water-bearing strata in the basin-fill material of Tooele Valley. A water-level contour map for the northern part of Tooele Valley was constructed, and seasonal changes in water levels were correlated to ground-water withdrawals and precipitation. Streamflow for Settlement and Middle Canyons was estimated. Thomas (1946) also presented information on the occurrence, movement, and discharge of water in the consolidated rocks that surround Tooele Valley.

In a study during 1958-63, Gates (1965) re-evaluated the ground-water resources of Tooele Valley and estimated a water budget for the principal artesian aquifer system. Gates (1965) redefined faults in the area and outlined five faults (fig. 2). Aquifer tests were done to quantify the hydrologic properties of the basin-fill material, and water-quality data were collected to define areas of poor-quality water in the valley.

Razem and Steiger (1981) updated the water budget for the principal aquifer system developed by Gates (1965) on the basis of longer term surface- and ground-water data, additional aquifer tests, and a more detailed analysis of evapotranspiration. Concurrent with the Razem and Steiger (1981) study, test holes were drilled in Tooele Valley to obtain hydrologic and geologic information on basin-fill material (Ryan and others, 1981). Also, a two-dimensional ground-water flow model (Razem and Bartholoma, 1980) was used to project future ground-water conditions on the basis of several water-management alternatives (Razem and Steiger, 1981).

Stolp (1994) studied the surface and ground-water resources of southeastern Tooele Valley and the adjacent Oquirrh Mountains (fig. 1) during 1988-90. Stolp estimated flow in streams and stream-channel deposits in Settlement and Middle Canyons and described ground-water conditions in the basin-fill material of southeastern Tooele Valley.

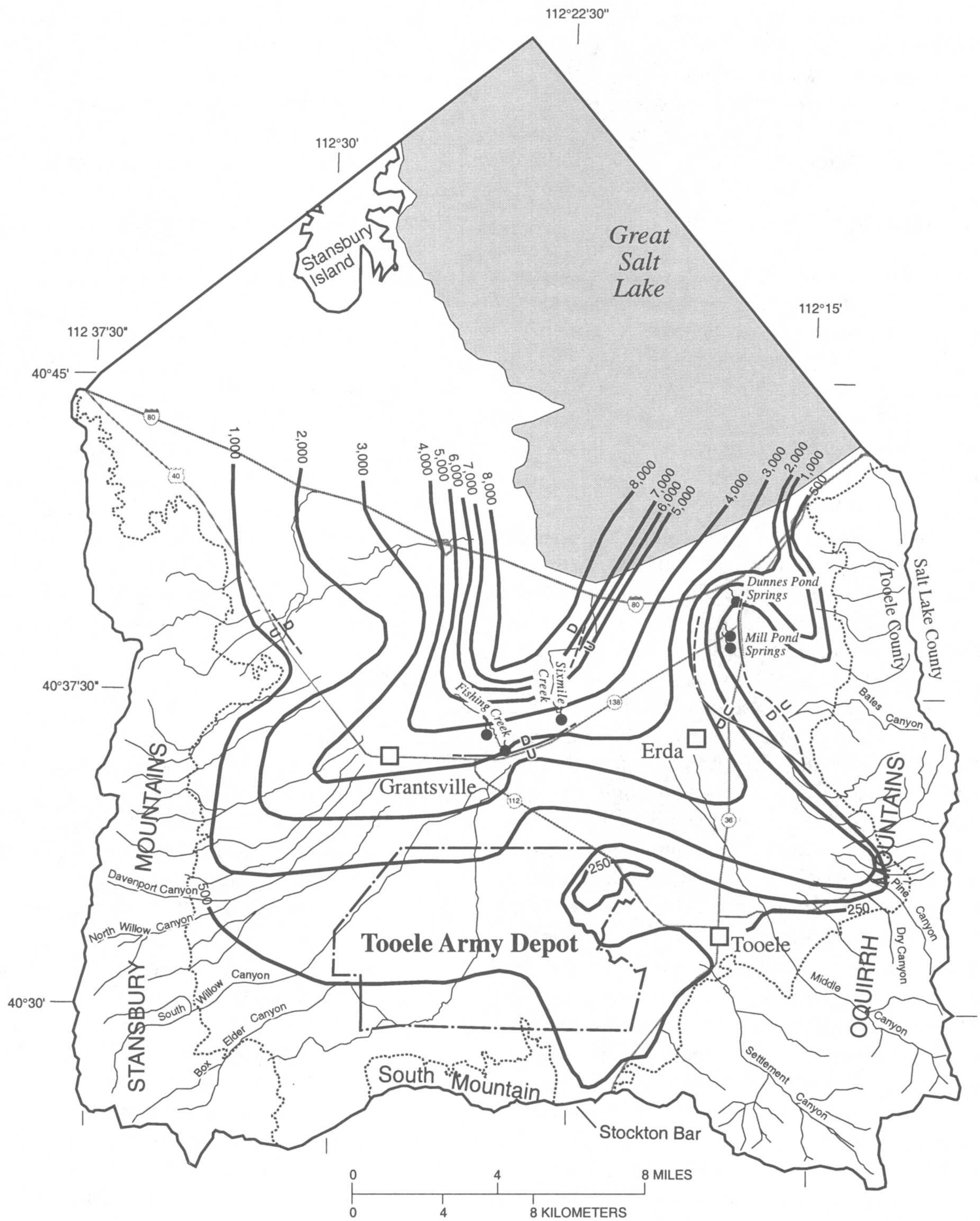
The U.S. Department of the Army and the U.S. Army Corps of Engineers have collected and analyzed geologic and hydrologic data at Tooele Army Depot (fig. 1). The data and analyses are presented in many volumes and appendices of data reports and interpretive reports, some of which are referenced individually in this report. The U.S. Army Corps of Engineers Hydrologic Engineering Center (1994, 1995) developed a three-dimensional ground-water flow model of the eastern part of the depot to evaluate possible approaches for controlling migration of solvents that have seeped from depot facilities into the ground water.

Geology

Tooele Valley covers about 300 mi^2, and altitude ranges from about 4,200 ft near Great Salt Lake to about 5,200 ft at the valley margins (fig. 1). The valley is a structural depression filled with unconsolidated and semiconsolidated basin-fill material and is surrounded by mountains composed of consolidated rock on the west, south, and east, and by Great Salt Lake on the north.

The eastern border of Tooele Valley is formed by the north-trending Oquirrh Mountains. The Oquirrh Mountains are 5 to 12 mi wide and rise abruptly from the valley floor from an altitude of about 5,200 ft to more than 10,000 ft in the southeastern part of the drainage basin. The valley is bordered on the south by South Mountain, a relatively low transverse divide, and Stockton Bar, a unconsolidated-rock bar-like feature deposited by ancient Lake Bonneville. The Oquirrh Mountains and South Mountain are composed mainly of the Oquirrh Formation of Late Mississippian, Pennsylvanian, and Early Permian age. This formation consists predominantly of alternating quartzite and limestone beds.

The western border of Tooele Valley is formed by the Stansbury Mountains. The Stansbury Mountains are relatively narrow and rise abruptly from the valley floor to a maximum altitude of 11,031 ft in the south-



EXPLANATION

4,000 **Line of equal thickness of basin-fill material**—Contour intervals 250, 500, and 1,000 feet

Approximate boundary of basin-fill material

U D **Normal fault (from Gates, 1965)**—Dashed where inferred; U, upthrown side, D, downthrown side

Spring

Figure 2. Thickness of basin-fill material in Tooele Valley, Utah. (Modified from James M. Montgomery, Consulting Engineers, Inc., 1986.)

western part of the drainage basin. Numerous formations crop out in the Stansbury Mountains; the thickest are the Oquirrh Formation and the Tintic Quartzite, which is of Cambrian age. The rocks in all three of the mountain ranges that border the valley have been extensively folded and faulted.

The valley is filled with basin-fill material of Tertiary and Quaternary age that consists mainly of sand, gravel, silt, clay, and volcanic detritus and ash. Subaerial and lacustrine conditions of deposition alternated during the Tertiary and Quaternary history of the valley, producing, correspondingly, alluvial deposits or lake-bottom and lake-shore deposits (Gates, 1965, p. 17). The sediments of Tertiary age in the valley make up the Salt Lake Group (Slentz, 1955), which consists mainly of semiconsolidated and unconsolidated lacustrine and alluvial-fan deposits. The younger Quaternary-age sediments are unconsolidated and are generally composed of lacustrine and alluvial material. The thickness of the basin-fill material ranges from a feather's edge at the margins of the valley to more than 8,000 ft in the north-central part of the valley (fig. 2). Geophysical data indicate that the buried consolidated-rock base of the valley is an irregular surface formed by a complex collection of troughs and ridges caused by several down-faulted blocks (ERTEC, 1982).

HYDROLOGY OF THE GROUND-WATER SYSTEM

The ground-water flow system of Tooele Valley is thought to be contained primarily in the Quaternary-age basin-fill materials. Few wells in Tooele Valley yield water from the Tertiary-age basin-fill material, which is generally assumed to be less permeable than the overlying Quaternary-age material. Thickness of the Quaternary-age basin-fill material is not well known but is estimated to be about 1,000 ft. The underlying Tertiary-age Salt Lake Formation is typically identified by the presence of volcanic rock fragments and sediments with some degree of consolidation. The boundary between the Quaternary-age and Tertiary-age sediments is generally gradational and is not easily distinguished. Well logs from previous studies of Tooele Valley (Razem and Steiger, 1981, and ERTEC, 1982) indicate a general change in basin-fill lithology from 800 to 900 ft below land surface, which possibly represents the top of the Tertiary-age sediments. More data are available for Salt Lake Valley, which is about 25 mi east of and separated from Tooele Valley by the Oquirrh Mountains (fig. 1). Throughout most of Salt Lake Valley, the altitude of the top of the Tertiary-age sediments is estimated to be less than 1,000 ft below land surface (Arnow and others, 1970, and Lambert, 1995, fig. 4). The Tertiary-Quaternary contact may be more shallow near the margins of the valley and in the vicinity of Tooele Army Depot where consolidated rock is present at relatively shallow depths (fig. 2). Tertiary-age sediments in wells west of the mouth of Middle Canyon are present at depths of a few ft to about 100 ft below land surface.

An uplifted block of consolidated rock (quartzite, sandstone, and limestone) located beneath the northeastern corner of Tooele Army Depot (fig. 2) has significant effects on the local ground-water system. The consolidated rock, which has been studied by the U.S. Army Corps of Engineers (James M. Montgomery, Consulting Engineers, Inc., 1986, 1987, 1988, and the U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994), is composed of thin- to massive-bedded sedimentary rocks striking roughly east-northeast and dipping to the north-northwest. The consolidated rock is extensively fractured and ground water flows through the rock under a steep hydraulic gradient to basin-fill material in the north (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994, p. 17). Geophysical and well-log data indicate another consolidated-rock high extending into the northeastern part of the valley from the Bates Canyon area (fig. 2). Although little is known about the lithologic makeup and integrity of this block, its presence is assumed to affect ground-water movement in the area. However, on the basis of water-level data, the block is not considered a part of the Tooele Valley ground-water flow system.

The basin-fill ground-water flow system in Tooele Valley (fig. 3) is conceptualized in this report as consisting of two aquifers separated by a shallow confining layer. The first aquifer is referred to as "the shallow unconfined aquifer" and is present only in the northern and central parts of the valley. The shallow unconfined aquifer is underlain by a shallow confining layer. The shallow confining layer is made up of overlapping and discontinuous lenses of fine-grained material (clay). The overlapping nature of these lenses creates a confining layer that is conceptualized as laterally extensive and continuous. The second and larger aquifer, referred to as "the principal aquifer" underlies both the shallow confining layer and the remainder of Tooele Valley. Where the principal aquifer underlies the shallow confining layer (northern and central parts of the valley), the principal aquifer is considered to be

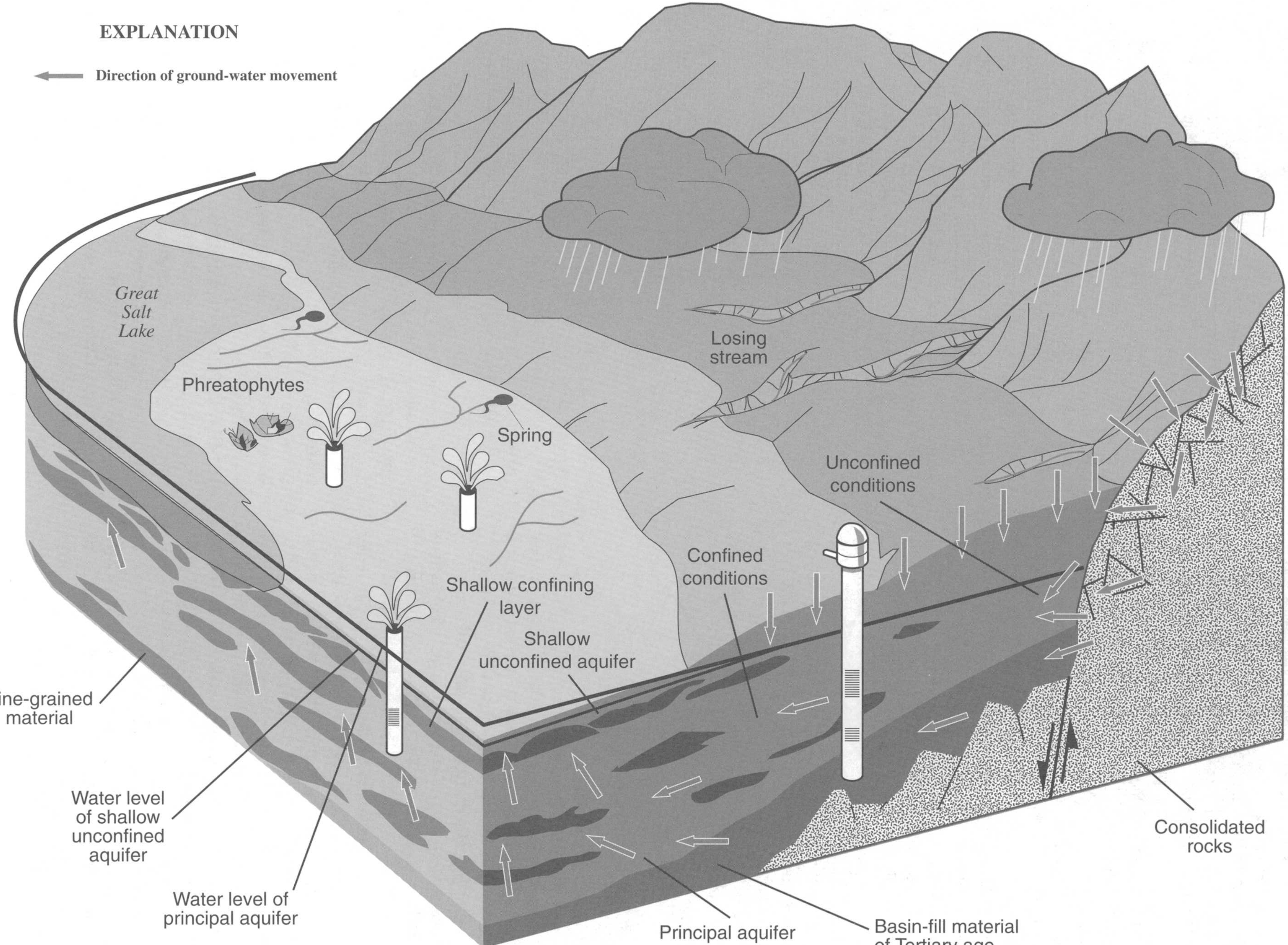


Figure 3. Generalized block diagram showing the basin-fill ground-water flow system in Tooele Valley, Utah.

confined. In the remainder of the valley (southern part and along the valley margins), the principal aquifer is considered to be unconfined.

The shallow unconfined aquifer is generally present within the upper 50 ft of the basin-fill material. Few wells in the valley tap into this aquifer, which typically consists of sand or finer-grained sediments and contains ground water of poor quality (Gates, 1956, p. 20).

The shallow confining layer underlies the shallow unconfined aquifer and consists mainly of fine-grained lake-bottom deposits. These fine-grained deposits consist of interfingering and overlapping layers and lenses of clay, silt, and fine-grained sand and are generally more extensive and continuous than the fine-grained material present within the principal aquifer.

Reports from previous studies and a review of available drillers' logs of wells in the valley done during this study indicate that the altitude of the base of the shallow confining layer varies substantially. Thomas (1946, p. 145-146) reported that the top of the confined zone of the principal aquifer (base of the shallow confining layer) ranged from about 90 to 110 ft below land surface in the Grantsville area and probably exceeded 167 ft below land surface in the Erda area. In a review of more than 180 drillers' logs during this study, the occurrence of the uppermost zone of material that (1) consisted mostly of clay and (or) silt and (2) was more than 20 ft thick was recorded. The depth to the bottom of this zone (relative to land surface) was assumed to be representative of the base of the shallow confining layer. On the basis of that criteria, the base of the shallow confining layer ranged from a minimum of 55 ft to more than 250 ft below land surface and varied substantially from well to well. In the northern part of the valley, data from the drillers' logs indicate sequences of fine-grained sediments that exceed 300 ft. Because of uncertainty in the accuracy of the data from drillers' logs, the lack of log data in some areas, and the extreme variability of the estimated base of the confining layer, the base of the shallow confining layer could not be defined accurately in all areas. On the basis of the review of drillers' logs, however, the base of the shallow confining layer is assumed to be about 100 ft below land surface in most areas. This simplifying assumption represents a qualitative average.

In 1995, the Utah Geological Survey (Mike Lowe, written commun., 1996) reviewed 160 drillers' logs of wells in Tooele Valley to identify the areal extent of zones of clay or silt in the shallow sediments. The criteria and methods for identifying the shallow clay/silt zones were those outlined by Anderson and others (1994). From this analysis, areas of the valley have been delineated that are underlain by shallow layers of fine-grained sediments such as clay, silt, sandy clay, or silt and clay that are more than 20 ft thick (Steiger and Lowe, 1997). These areas, labeled by Steiger and Lowe (1997) as the "discharge" and "secondary recharge" areas of the valley (fig. 4), are assumed to represent maximum possible horizontal extent of the shallow unconfined aquifer and the shallow confining layer.

The principal aquifer underlies all of Tooele Valley. Where it underlies the shallow confining layer (northern and central parts of the valley), the aquifer is considered to be confined. In the remainder of the valley (southern part and along the valley margins), the principal aquifer is considered to be unconfined. In previous literature (Thomas, 1946, and Gates, 1965), the term "principal aquifer" for Tooele Valley has been used to refer only to the confined aquifer in the northern and central parts of the valley.

Where the principal aquifer is confined, the basin-fill material consists of clay, silt, sand, and gravel. In these areas, ground water within the principal aquifer is confined in coarser material that overlies and underlies discontinuous fine-grained beds that are mostly lake-bottom materials (Gates, 1965, p. 20). In areas where the principal aquifer is unconfined, the basin-fill material consists mainly of sands and gravels, and fine-grained sediments generally are absent. In the subsurface near Tooele Army Depot and in the northwestern part of the valley, the principal aquifer includes saturated blocks of fractured consolidated rock that extend into the valley (fig. 2, where the contours indicate thinning of the basin-fill material).

Hydrologic Properties

Few wells have been completed in the shallow unconfined aquifer and shallow confining layer in Tooele Valley (Gates, 1965, p. 20); thus, few data are available to define the hydrologic properties of these materials directly. This lack of data precludes direct analysis of the shallow basin-fill materials, and probable hydrologic properties are based on data pertaining to Salt Lake Valley. Salt Lake Valley is about 25 mi east of and separated from Tooele Valley by the Oquirrh Mountains. The physiographic setting of both valleys is similar; each is surrounded by high-altitude consolidated-rock mountains on the west and east and terminates to the north against Great Salt Lake. The shallow

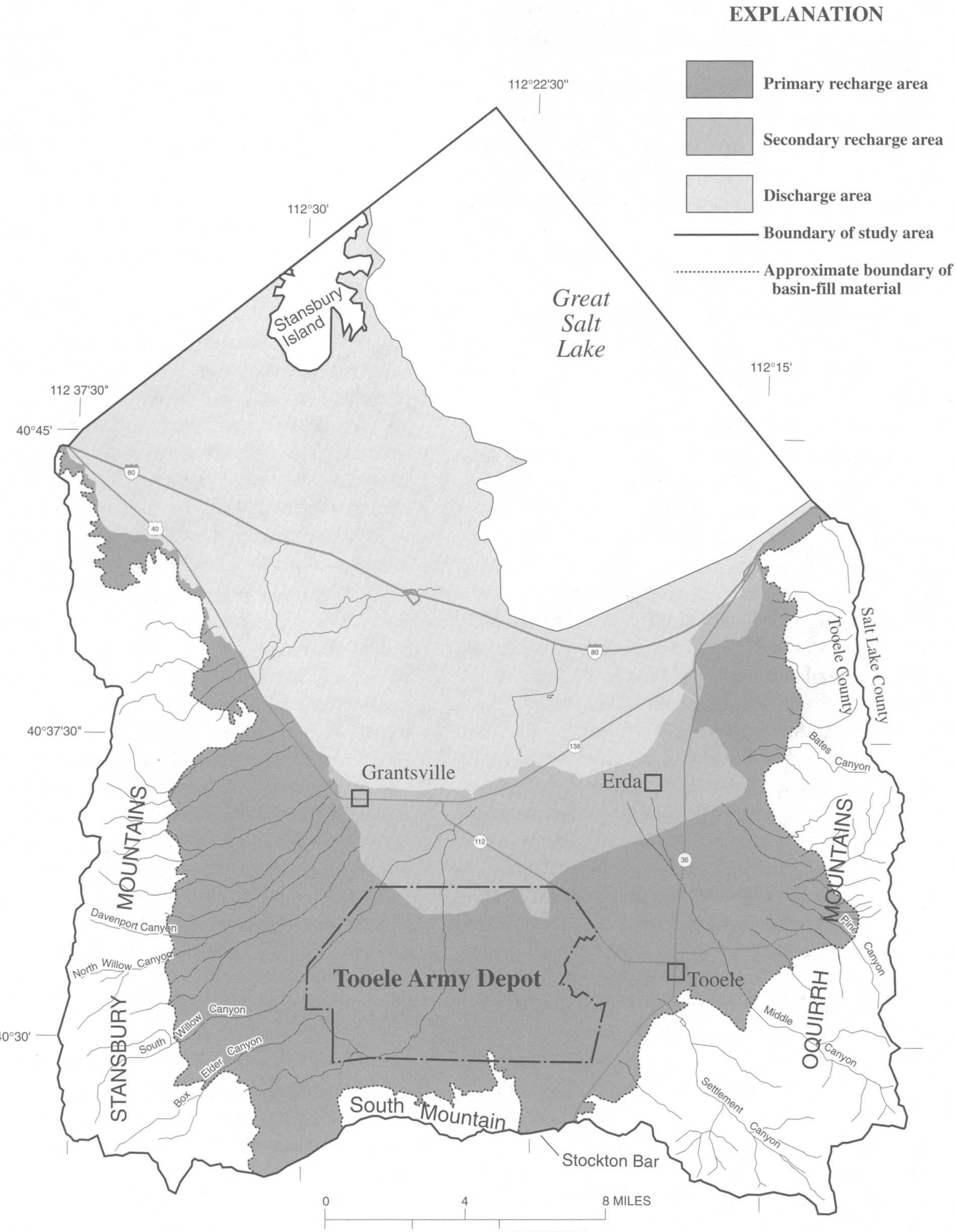


Figure 4. Recharge and discharge areas for the principal aquifer in Tooele Valley, Utah (from Steiger and Lowe, 1997).

lacustrine and alluvial sediments in Salt Lake Valley are the result of the filling and drying of the same ancient lakes that occupied Tooele Valley. Conceptually, the ground-water system in Salt Lake Valley also is thought to have a shallow unconfined aquifer underlain by a shallow confining layer.

Horizontal Hydraulic Conductivity and Transmissivity

The hydrologic properties of shallow basin-fill materials that make up the shallow unconfined aquifer and underlying confining layer in Salt Lake Valley were studied previously by Thiros (1995), and the results of that study were used to define a range of probable values for the horizontal hydraulic conductivity of the shallow unconfined aquifer and confining layer in Tooele Valley. Hydraulic-conductivity values were estimated in slug tests by Thiros (1995) at 32 wells completed in the shallow unconfined aquifer of Salt Lake Valley. Estimated values ranged from 0.003 ft/d to 65.5 ft/d (Thiros, 1995, table 4). The wide range of values typifies the spatial variability of the hydrologic properties of the shallow sediments in both Salt Lake and Tooele Valleys. The results of four slug tests at wells completed in layers of clay and silt of what was assumed to be the shallow confining layer in Salt Lake Valley (Lambert, 1995, p. 14) indicate values of hydraulic conductivity that range from 0.04 ft/d to 2.28 ft/d. These tests were conducted in small intervals of material, and individual test results were not assumed to represent the equivalent hydraulic-conductivity value of the shallow unconfined aquifer or the shallow confining layer. They do indicate the range of values that might be expected for the shallow unconfined aquifer and underlying shallow confining layer in Salt Lake Valley. These same ranges of values also are thought to exist in Tooele Valley.

The hydraulic properties of the basin-fill material that makes up the principal aquifer have been evaluated for some areas in Tooele Valley. The evaluations were done during previous studies by various methods including analyses of specific-capacity data and the results of aquifer tests. Field measurements, however, are not available for all areas of the valley. Also, aquifer tests used to estimate transmissivity values were conducted, in most cases, using wells that do not penetrate the entire thickness of the principal aquifer. Thus, the results of available field tests probably do not accurately represent the transmissivity of the principal aquifer. It was assumed, however, that the results from these tests could be used to estimate probable ranges of hydraulic conductivity for zones of the principal aquifer that represent similar sediment types and to define general ranges of transmissivity for areas of the aquifer.

Estimates of hydraulic conductivity, derived from specific-capacity values and from aquifer tests reported in previous studies (Gates, 1965, table 3; Razem and Steiger, 1981, table 4; James M. Montgomery, Consulting Engineers, Inc., 1986 and 1988; Stolp, 1994, table 8; and U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994) and estimates made from specific-capacity data analyzed during this study indicate a range from 5 ft/d to more than 600 ft/d. The wide range of values typifies the variability of material sampled in the valley. Tests of the properties of the basin-fill material generally were done at production wells screened at permeable intervals of basin-fill material and were assumed to represent the properties of coarse-grained material (sand and gravel). The northern half of the valley (zone 1 in fig. 5) represents an area where permeable intervals of the principal aquifer generally consist of sand with little or no gravel. In the northern half and western margin of the valley (zone 2 in fig. 5), permeable intervals generally consist of coarse sand or sand and gravel. Of the 24 estimated values of hydraulic conductivity from the previously mentioned tests, 7 were determined from tests at wells located in zone 1 (fig. 5) where fine-grained sediments predominate, and 17 were determined from tests at wells located in zone 2. Estimates of hydraulic conductivity for coarse-grained basin-fill material within zone 1 ranged from 17 ft/d to 90 ft/d, with a mean of 56 ft/d. These values are assumed to represent the properties of the sand layers sampled by the wells within zone 1. Estimates of hydraulic conductivity at wells located in zone 2 ranged from 5 ft/d to 620 ft/d, with a mean of 137 ft/d.

Estimates of hydraulic conductivity of coarse-grained basin-fill material and consolidated rock also have been made by the U.S. Army Corps of Engineers for the eastern part of Tooele Army Depot on the basis of various analyses including field tests and contaminant plume migration on the depot (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994) (table 1). The U.S. Army Corps of Engineers reports that the results of 65 short-term aquifer tests conducted in wells penetrating coarse-grained alluvium indicate an average hydraulic-conductivity value of about 90 ft/d. The U.S. Army Corps of Engineers reports that some of the most reliable field-test data included a single long-term

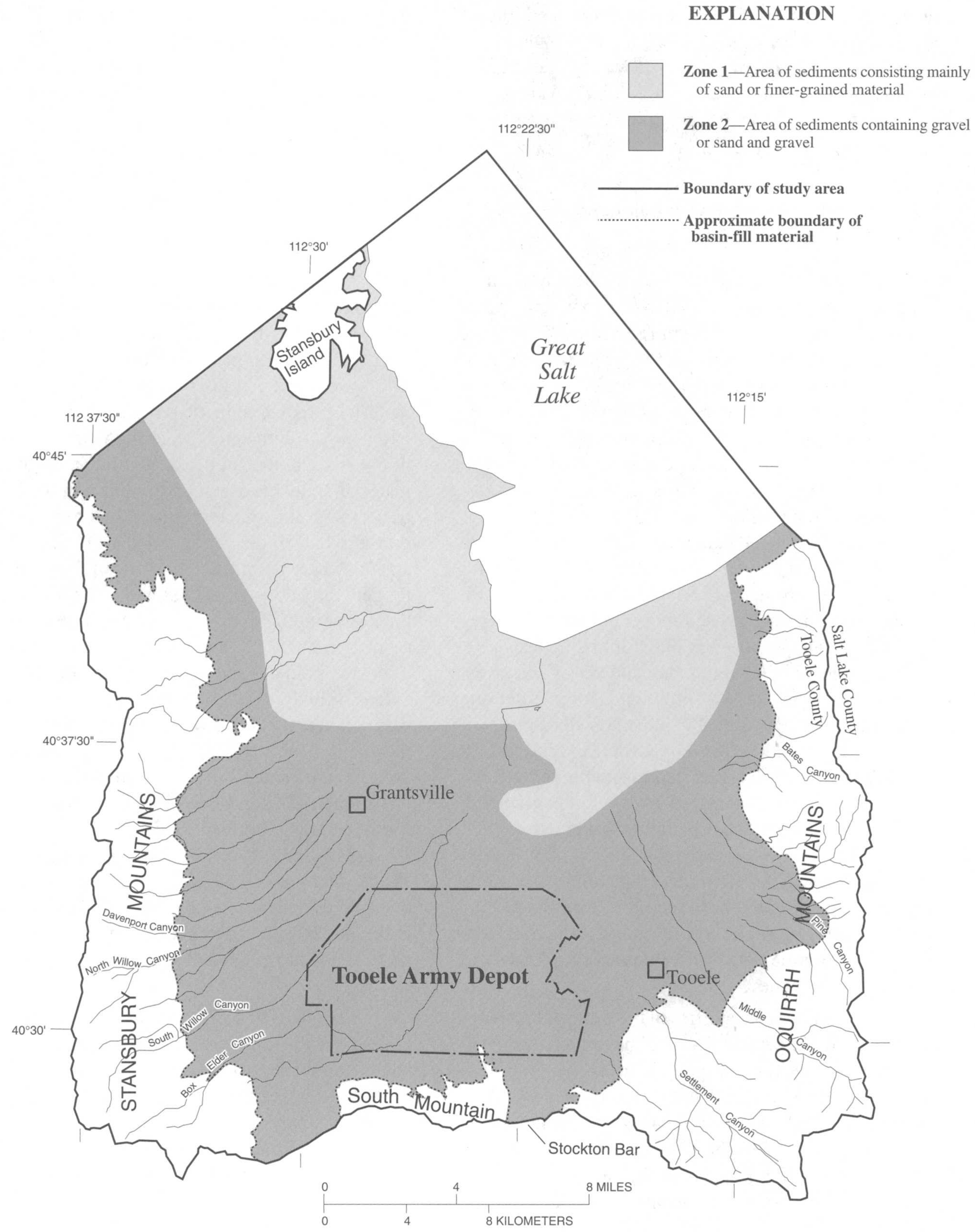


Figure 5. Zones of basin-fill material of similar permeability in the principal aquifer in Tooele Valley, Utah.

alluvial aquifer test (48-hour pumping period) of a well in the northeastern part of the depot. The long-term test yielded a result of 200 ft/d for the upper 300 ft of alluvium in the area. In the area of the consolidated-rock high, which is beneath the southeastern part of the depot, the average hydraulic-conductivity value computed from 32 short-term aquifer tests (less than 1-hour pumping period) was about 30 ft/d. Ranges of values of hydraulic conductivity for basin-fill material and consolidated rock estimated by the U.S. Army Corps of Engineers are reported in table 1.

Transmissivity of the principal aquifer was determined from the estimates of hydraulic conductivity and the thickness of permeable intervals of the aquifer. On the basis of information from drillers' logs, the ratio of sand/gravel (coarse-grained) intervals to the total depth of material in the aquifer was estimated (fig. 6). The ratio represents the percentage of the aquifer that is assumed to be coarse-grained permeable basin-fill material. The drillers' logs indicate that material composed mainly of clay and silt predominates in the northern part of Tooele Valley and, in general, less than 15 percent of the principal aquifer is made up of coarse-grained material. Drillers' logs indicate that this ratio generally increases near the mountain front and in the southern parts of Tooele Valley. In much of southern Tooele Valley, the percentage of sand/gravel intervals is greater than 50 percent (fig. 6), and in some areas, the ratio exceeds 70 percent. In the southwestern corner of Tooele Valley where few wells have been drilled and consequently little information is available, the percentage of coarse-grained material was estimated to be greater than 50 percent. This area lies along the valley margin where, in Tooele Valley, continuous layers of fine-grained material are generally not found and the basin-fill material typically consists of coarse-grained material.

A general range of transmissivity was determined for the principal aquifer by using estimates of hydraulic conductivity and the percentage of the principal aquifer that consists of permeable material (the sand/gravel-bearing intervals) (fig. 6). In the northern part of the valley (zone 1 in fig. 5), the hydraulic-conductivity value of the permeable materials ranges from 17 to 90 ft/d, the permeable interval consists of 15 to 50 percent of the principal aquifer, the estimated thickness of the principal aquifer is about 900 ft, and the hydraulic-conductivity value of the fine-grained basin-fill material is estimated to be 1 ft/d. With these figures, transmissivity may range from about 3,000 to 41,000 ft^2/d. In the southern parts of Tooele Valley and along the valley margins (zone 2 in fig. 5), the hydraulic-conductivity value of the permeable materials ranges from 5 to 620 ft/d, the permeable interval consists of 15 to 70 percent of the principal aquifer, the estimated thickness of the principal aquifer ranges from about 150 to 900 ft, and the hydraulic-conductivity value of the fine-grained basin-fill material is estimated to be 1 ft/d. These amounts result in transmissivity values that range from 200 to 391,000 ft^2/d.

Table 1. Estimated hydraulic-conductivity (K) values for basin-fill material and consolidated rock in Tooele Valley, Utah

[—, no range or value estimated]

Description of material	Range of K estimated at Tooele Army Depot[1] (feet per day)	Values of K incorporated in calibrated U.S. Army Corps of Engineers ground-water flow model of the Tooele Army Depot[1] (feet per day)	Range of K values estimated in other areas of Tooele Valley (feet per day)
Clay/silt	0.04-2.28	—	—
Sand	5-250	60, 200	17-90
Sand/gravel	100-500	280, 385	5-620
Displaced sediments	—	1.5	—
Fractured limestone	0.1-100	35	—
Other fractured consolidated rock	0.01-10	0.1, 3.0	—

[1]From U.S. Army Corps of Engineers Hydrologic Engineering Center (1994, p. 45-49).

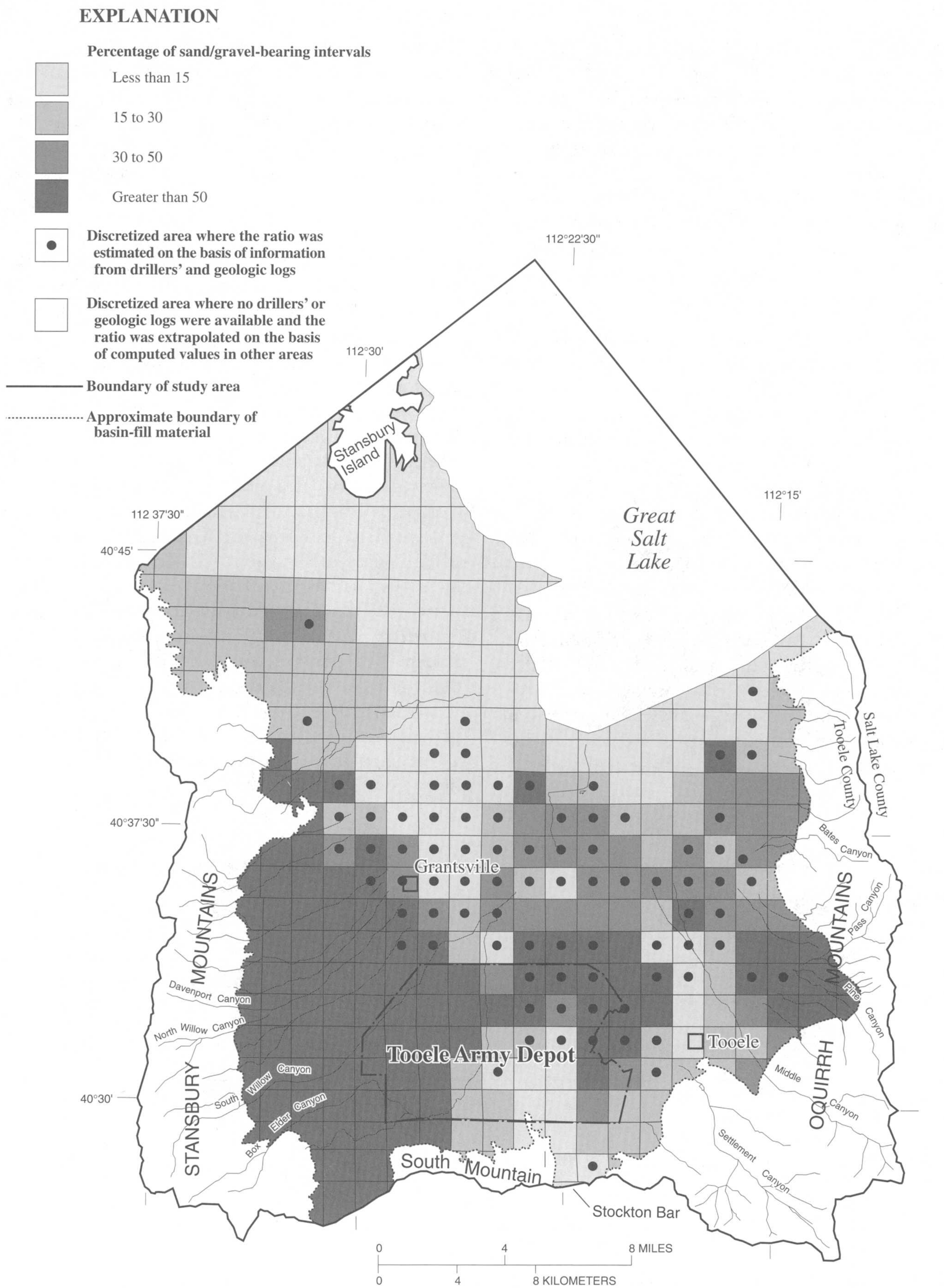


Figure 6. Estimated percentage of the principal aquifer that consists of sand/gravel-bearing intervals, Tooele Valley, Utah.

Vertical Hydraulic Conductivity

No estimates of vertical hydraulic conductivity (K_v) for the shallow unconfined aquifer and the shallow confining layer have been reported for Tooele Valley. Vertical hydraulic-conductivity values of similar aquifer zones in adjacent Salt Lake Valley, however, have been estimated by Waddell and others (1987, p. 30) and by Thiros (1995, p. 33-38) based on the results of aquifer tests. Estimates of K_v from four tests in Salt Lake Valley range from 0.01 to 1.0 ft/d. Thiros (1992, table 12) also reported a range in magnitude for K_v in shallow sediments of 5.1×10^{-5} to 0.02 ft/d determined from laboratory tests of 35 core samples. The cored material typically consisted of silt and clay of lacustrine origin and was assumed to represent sediments of the shallow confining layer or shallow unconfined aquifer in Salt Lake Valley. As mentioned previously, the shallow aquifer system in Salt Lake Valley is considered to be analogous to the shallow aquifer system in Tooele Valley, thus estimates of K_v listed above are considered indicative of values for the shallow system in Tooele Valley.

There are also no reported vertical hydraulic-conductivity (K_v) data from aquifer tests for the sediments of the principal aquifer in Tooele Valley. Estimates for the basin-fill material of the principal aquifer and for consolidated rock have, however, been made by the U.S. Army Corps of Engineers during calibration of a ground-water flow model of part of Tooele Army Depot (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994, p. 54 and 55). Initially, the U.S. Army Corps of Engineers estimated K_v for use in their flow model as ratios of estimated horizontal hydraulic conductivity of a given sediment type at the depot (anisotropy), from studies conducted in similar physiographic domains, and as a function of measured vertical hydraulic gradients (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994, p. 54-57). Final estimates of K_v for basin-fill material resulting from calibration of the U.S. Army Corps of Engineers flow model ranged from 0.2 to 1.35 ft/d. Estimates for K_v for consolidated rock beneath the depot ranged from 1.4 to 7 ft/d. Estimates of K_v also have been derived for similar basin-fill material in the adjacent Salt Lake Valley during calibration of ground-water flow models. Dames and Moore (1988, table 8) estimated K_v of unconsolidated basin-fill material in southwestern Salt Lake Valley to be 0.5 ft/d. In a two-dimensional flow model in the same area, Holdsworth (1985, table 2) estimated K_v for unconsolidated basin-fill material to range from 1 ft/d to 30 ft/d. Lambert (1995, fig. 8) estimated K_v for the principal aquifer in Salt Lake Valley to range from 0.01 to 5.0 ft/d.

Vertical hydraulic conductivity of the principal aquifer may be affected, in areas, by Tertiary- and Quaternary-age faults (fig. 2) identified by Thomas (1946, p. 149-153) and Gates (1965, p. 17-18). No field observations or measurements are available to define the characteristics of these fault zones although it is possible that the faults have produced zones of relatively high K_v. Large springs in the valley rise along presumed faults, including Dunne's Pond Springs and Mill Pond Springs in the northeastern part of the valley and the spring sources of Fishing Creek and Sixmile Creek east of Grantsville (fig. 2). The water of these springs is similar in chemical composition to water yielded by nearby wells (Thomas, 1946), and the faults possibly act as conduits for ground-water flow from the principal aquifer to the surface.

Storage Coefficient and Specific Yield

Data from three aquifer tests evaluated by Gates (1965, table 3, p. 30) indicate a range of 2.0×10^{-4} to 4.2×10^{-3} for storage-coefficient values (S) in confined zones of the principal aquifer. The thickness of the principal aquifer in the vicinity of each of these aquifer tests is estimated to be 900 ft. With this information, a range for specific storage (S_s, which is equal to S divided by aquifer thickness) of 4.7×10^{-6} ft^{-1} to 2.2×10^{-7} ft^{-1} is assumed. Razem and Bartholoma (1980, p. 6) used a value of 2.0×10^{-3} for storage coefficient in their two-dimensional ground-water flow model of the valley.

Specific yield in the unconfined zone of the principal aquifer at a well on Tooele Army Depot was estimated to be 0.30 (James M. Montgomery, Consulting Engineers, Inc., 1988, appendix C). Razem and Steiger (1981, p. 20) estimated the value of average specific yield for the basin-fill material in Tooele Valley to be 0.10. Specific yield of the basin-fill material of the principal aquifer was reported to range from 0.1 to 0.2 based on the calibration of the U.S. Army Corps of Engineers model of Tooele Army Depot (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994, fig. 20). Specific yield of basin-fill material in unconsolidated zones of the principal aquifer in adjacent Salt Lake Valley was estimated to range from 0.05 to 0.30 during calibration of a ground-water flow model (Lambert, 1995, fig. 27).

Recharge

Recharge to the basin-fill ground-water flow system in Tooele Valley is mainly from (1) subsurface inflow from the consolidated-rock aquifers in the surrounding mountains and stream-channel deposits at the mouths of canyons, (2) infiltration of precipitation on the valley floor, (3) seepage of unconsumed irrigation water from irrigated fields and lawns/gardens, and (4) subsurface inflow from Rush Valley through the Stockton Bar. Rush Valley is adjacent to Tooele Valley on the south. Estimated long-term average amounts of recharge from these processes are listed below.

Long-term average recharge in Tooele Valley, in acre-feet per year	
Consolidated rock and stream-channel deposits	**48,000**
Infiltration of precipitation	**12,000**
Seepage of unconsumed irrigation water	**10,000**
Subsurface inflow from Rush Valley	**5,000**
TOTAL	**75,000**

Consolidated Rock and Stream-Channel Deposits

The long-term average inflow from consolidated rock and stream-channel deposits near the mountain fronts is estimated to be about 48,000 acre-ft/yr. This amount is based on the ground-water recharge estimates made by Razem and Bartholoma (1980) and Razem and Steiger (1981).

The approach used by Razem and Steiger (1981, p. 12) to estimate ground-water recharge to the basin-fill material in Tooele Valley is based on methods described in Eakin and others (1951, p. 79-81) and Hood and Waddell (1968, p. 22). Their approach assumed that a fixed percentage of the average annual precipitation that occurs in the surrounding mountains and at the margins (foothills) of the valley recharges the ground-water system. The fixed percentage is determined on the basis of altitude, geology, gradient of land surface, and amount of precipitation. The results of that approach indicate that the average annual recharge to the valley was about 51,000 acre-ft, of which about 32,000 acre-ft comes from the Oquirrh Mountains, about 19,000 acre-ft from the Stansbury Mountains, and less than 500 acre-ft from South Mountain.

Estimates of ground-water recharge to the basin-fill material in Tooele Valley from consolidated rock and stream-channel deposits were evaluated and refined in a two-dimensional numerical model of the ground-water flow system of the valley developed by Razem and Bartholoma (1980). Results from calibration of this model (Razem and Steiger, 1981, p. 12) indicate an average annual recharge to the basin-fill material of about 52,000 acre-ft. This consists of about 40,000 acre-ft from the Oquirrh Mountains, about 12,000 acre-ft from the Stansbury Mountains, and less than 500 acre-ft from South Mountain. This refinement represents only a small change in total recharge from consolidated rock and stream-channel deposits; however, the distribution of recharge was altered substantially. Recharge from the Oquirrh Mountains was increased from about 60 to 80 percent of the recharge from consolidated rock and stream-channel deposits. Recharge from the Stansbury Mountains was likewise reduced from about 40 to 20 percent.

The southeastern and southwestern margins of Tooele Valley were assumed by Razem and Steiger (1981) to be the areas of greatest ground-water recharge because precipitation is greatest and stream-drainage areas are largest in these areas. Results of the Razem and Bartholoma model (1980) indicate that an average of about 18,000 acre-ft/yr of recharge (45 percent of total recharge from the Oquirrh Mountains) occurs near the basin-fill material/consolidated-rock boundary between the east edge of Stockton Bar and the mouth of Middle Canyon. About 15,000 acre-ft/yr of recharge (38 percent) is estimated to occur between the mouth of Middle Canyon and the mouth of Dry Canyon (Razem and Bartholoma, 1980, table 2 and fig. 2). From Dry Canyon north to the boundary of the study area the recharge is estimated to be about 7,000 acre-ft/yr (17 percent). In a study of southeastern Tooele Valley, Stolp (1994) estimated that about 2,800 acre-ft of the annual recharge (7 percent of the total recharge from the Oquirrh Mountains) that occurs along the mountain front from Stockton Bar to Bates Canyon is from adjoining stream-channel deposits. Some of the ground water in these deposits is intercepted by wells and tunnels before it reaches the principal aquifer.

The recharge estimates by Razem and Steiger (1981, p. 12) and Razem and Bartholoma (1980) represent the total recharge from the consolidated-rock mountains, unconsolidated stream-channel deposits, and precipitation at the valley margins. During this study, about 2,600 acre-ft/yr of the recharge estimated for the Oquirrh Mountains and about 1,800 acre-ft/yr of the recharge estimated for the Stansbury Mountains was determined to be from precipitation at the valley margins. Those amounts are subtracted from the amounts estimated above and included as recharge from infiltration of precipitation.

Infiltration of Precipitation

The long-term average ground-water recharge from precipitation on the valley floor is estimated to be about 12,000 acre-ft. This amount of recharge was computed as a function of the distribution of average annual precipitation (fig. 7) by using estimated percentages of total precipitation that recharges the aquifer system (table 2). Estimated percentages of precipitation on the valley floor that become recharge were derived from Hood and Waddell (1969, table 8) for zones of precipitation in Rush Valley, which is located directly to the south of Tooele Valley. These estimates were derived using the methods of Eakin and others (1951) and Gates (1965) as discussed previously. The percentage of precipitation that is considered to be recharge to the ground-water system ranged from 1 percent for the lower altitudes of the valley where annual precipitation generally is less than 12 in., to 25 percent in areas along the valley margins where average annual precipitation exceeds 20 in.

Table 2. Percentage of precipitation on the floor of Tooele Valley, Utah, that is estimated to recharge the ground-water system

[Precipitation shown in fig. 7]

Precipitation amount (inches)	Percentage of precipitation estimated to recharge the ground-water system
More than 20	25
16-20	8
12-16	3
Less than 12	1

Seepage of Unconsumed Irrigation Water

The long-term average ground-water recharge from infiltration of unconsumed irrigation water from fields and lawns/gardens is estimated to be about 10,000 acre-ft/yr. This estimate is based on a percentage of the total amount of irrigation water applied. Stolp (1994, p. 18) estimated that about 25 percent of applied irrigation water in southeastern Tooele Valley recharges the ground-water system. This percentage was assumed for the entire valley to compute the total recharge from irrigated land.

Sources of irrigation water that is delivered to fields in Tooele Valley include pumped and flowing irrigation wells, diversions from valley springs, and diversions from streams that enter the valley. Discharge from pumped and flowing irrigation wells averaged about 21,700 acre-ft/yr during 1963-94 (Utah Department of Natural Resources, Division of Water Resources, written commun., 1996). The average annual diversion from valley springs during 1963-84 was about 3,200 acre-ft as determined from unpublished records of the U.S. Geological Survey (unpub. data, 1965-84). Stolp (1994, p. 18) estimated that diversion from Settlement and Middle Canyons for the irrigation of fields is about 4,900 acre-ft/yr or about 60 percent of the total estimated annual flow in the streams (some water is diverted for other uses). If the same percentage of water is assumed to be diverted from other streams entering the valley, and estimates of flow in these streams from Razem and Steiger (1981, table 1) are used, then an additional 7,200 acre-ft/yr of surface water is estimated to be diverted to irrigated fields annually. Summing the diversions from all of the sources listed above yields an estimated 37,000 acre-ft of water applied annually to irrigated fields. Twenty-five percent of this amount equates to about 9,200 acre-ft/yr of recharge to the ground-water system.

Stolp (1994, p. 18) estimated the amount of irrigation water delivered to lawns/gardens in southeastern Tooele Valley to be about 700 acre-ft/yr on the basis of differences in summer and winter water deliveries by Tooele City. Residential acreage in that study was estimated to be about 2,300 acres (Utah Department of Natural Resources, Division of Water Resources, written commun., 1996). Using these amounts, the rate of water applied to lawns/gardens in southeastern Tooele Valley is assumed to be about 0.3 acre-ft/yr for each acre. There is an additional 4,200 acres of residential and commercial land outside of the study area that also may receive irrigation water for lawns/gardens. Assuming that the computed application rate is valid for all of Tooele Valley, it is estimated that the total amount of irrigation water applied to lawns/gardens is 1,950 acre-ft annually. Twenty-five percent of that water, or about 500 acre-ft, is assumed to recharge the ground-water system.

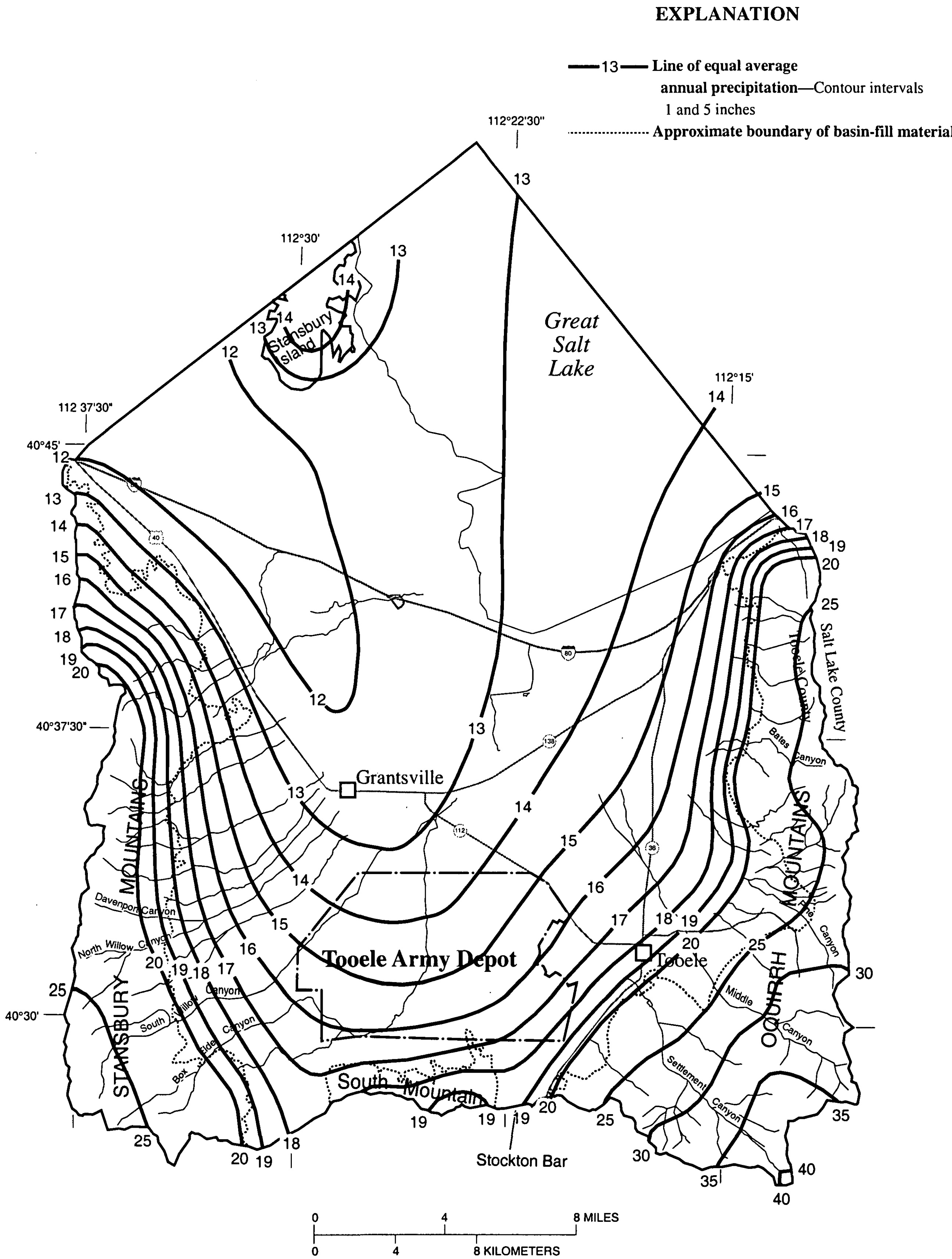


Figure 7. Average annual precipitation in Tooele Valley, Utah, 1961-90. (Data from Utah Climate Center.)

Subsurface Inflow from Rush Valley

Subsurface inflow of ground water from Rush Valley was assumed by Thomas (1946, p. 195), Gates (1965, p. 22) and Hood and Waddell (1969, p. 30-31). However, no field observations are available to estimate the amount of subsurface flow that occurs. Based on the calibration of a two-dimensional ground-water flow model of Tooele Valley, Razem and Bartholoma (1980) estimated subsurface inflow from Rush Valley through the Stockton Bar to be about 5,000 acre-ft/yr. No data are available that could be used to refine this estimate of subsurface inflow and therefore the amount of 5,000 acre-ft/yr is used in this study.

Discharge

Discharge from the basin-fill ground-water flow system in Tooele Valley is mainly by (1) withdrawals from pumped wells and discharge from flowing wells, (2) evapotranspiration, (3) spring discharge at Dunne's Pond Springs, Mill Pond Springs, source of Sixmile Creek, and source of Fishing Creek, (4) subsurface outflow to Great Salt Lake, and (5) flow to shallow drains and ditches.

Long-term average discharge in Tooele Valley, in acre-feet per year

Withdrawal from pumped wells	13,500
Discharge from flowing wells	12,500
Evapotranspiration	23,000
Spring discharge	16,000
Outflow to Great Salt Lake	3,000
Shallow drains and ditches	unknown
TOTAL	68,000

Wells

Ground-water use in Tooele Valley increased from about 7,000 acre-ft in 1939 (Gates, 1965, p. 24) to an average of about 29,000 acre-ft/yr for 1990-94 (Allen and others, 1995, table 2). The rate of withdrawal had stabilized by 1963 and averaged about 26,000 acre-ft/yr during 1964-94 (fig. 8). Average annual withdrawal from pumping wells for 1964-94 is about 13,500 acre-ft. Average-annual discharge from flowing wells for the same period is estimated to be about 12,500 acre-ft.

Evapotranspiration

Discharge by evapotranspiration in Tooele Valley is estimated to average about 23,000 acre-ft/yr (Razem and Steiger, 1981, table 3). This estimate is based on an evaluation of the area and types of phreatophytes, in conjunction with depth to the water table in the phreatophyte areas (Razem and Steiger, 1981, p.16-17). Gates (1965, table 2) previously estimated annual discharge by evapotranspiration to be about 40,000 acre-ft/yr. However, the method used by Razem and Steiger (1981) is thought to be more realistic because it considers evapotranspiration as a function of depth to the water table.

The discharge by evapotranspiration estimated by Razem and Steiger (1981, p. 16) was determined by considering the following five categories: bare ground and four associations of phreatophytes; greasewood, rabbitbrush, salt grass, and pickleweed. Separate evapotranspiration rates were assigned to each of these categories depending on whether the average depth to the water table is 0 to 5 ft or 5 to 10 ft (Razem and Steiger, 1981, table 3).

Spring Discharge

The estimated average annual ground-water discharge at springs in Tooele Valley is about 16,000 acre-ft/yr. This estimate represents discharge from Dunne's Pond Springs, Mill Pond Springs, source of Sixmile Creek, and source of Fishing Creek. Measurements made by Kennecott Copper Corporation during 1962, 1965, and 1979-82 and estimates of annual discharge reported in Gates (1965, p. 16), in Razem and Steiger (1981, p.16), and in unpublished records of the U.S. Geological Survey for 1972-84 indicate the following average annual discharge for each spring: Dunne's Pond Springs, 6,400 acre-ft/yr; Mill Pond Springs, 4,600 acre-ft/yr; source of Sixmile Creek, 2,800 acre-ft/yr; and source of Fishing Creek, 2,200 acre-ft/yr. Estimates of annual discharge from Dunne's Pond Springs exceeded 11,000 acre-ft during 1974-76; which represents a significant increase from other flow data available for the spring. Therefore, the average annual discharge rate computed for Dunne's Pond Springs excluded the discharge data for 1974-76. It was assumed that this results in a better representation of the long-term average.

Dunne's Pond Springs and Mill Pond Springs are located near the consolidated-rock high that extends into Tooele Valley from Bates Canyon (see section

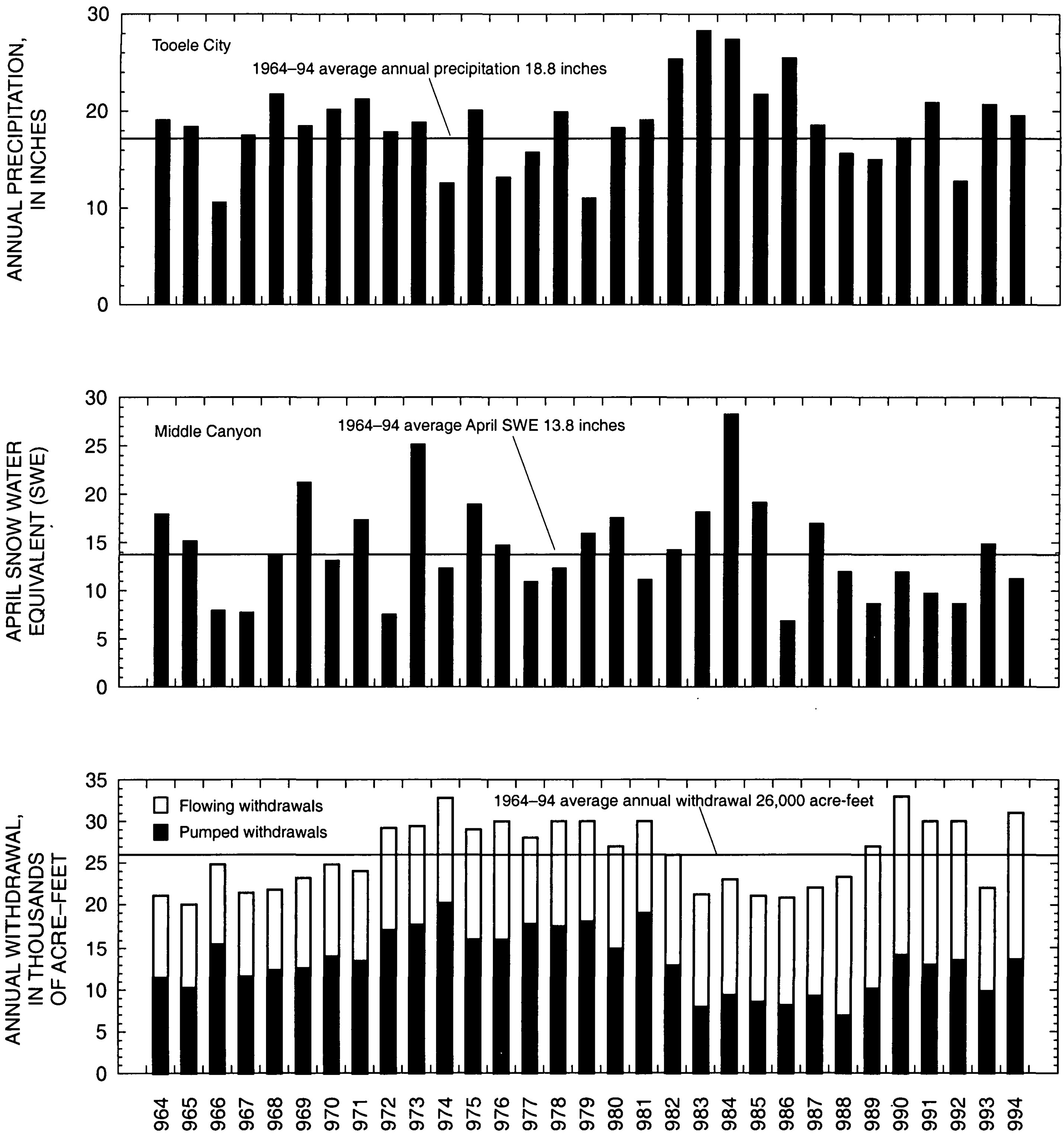


Figure 8. Annual precipitation at Tooele City, Utah, April Snow Water Equivalent (SWE) in Middle Canyon, and annual withdrawals of water by wells in Tooele Valley, Utah.

titled "Hydrology of the ground-water system"). On the basis of location, some of the ground water that discharges from these springs may represent flow that is coming directly from the consolidated rock. However, data are not available to quantify nor confirm that this is occurring.

Flow to Great Salt Lake and Shallow Drains and Ditches

The estimated average subsurface outflow to Great Salt Lake is about 3,000 acre-ft/yr (Razem and Steiger, 1981, p. 17). This amount was determined by applying Darcy's Law across a shoreline length of about 15 mi. A transmissivity of 1,000 ft^2/d and an average hydraulic gradient of about 20 ft/mi were used to calculate outflow to the lake.

Discharge from the shallow unconfined aquifer to surface drains and ditches have not been measured previously and data are not available to make an estimate of average-annual discharge; however, observations verify that base flow does occur to these sinks. Discharge to shallow drains and ditches only occurs in the northern parts of Tooele Valley and during times when the water table is at or near land surface.

Movement

Ground water in Tooele Valley generally moves from primary recharge areas near the margins of the valley (fig. 4) to the central and northern parts of the valley (Razem and Steiger, 1981). Vertical gradients are downward in the recharge areas and upward in the discharge areas, where ground water discharges naturally from the system. In the central and northern parts of the valley, ground water moves upward in the confined aquifer, into and through the overlying confining layer, and into the shallow unconfined aquifer.

Recharge and discharge areas in the valley (fig. 4) have been defined by Steiger and Lowe (1997) on the basis of information gathered from drillers' logs, aerial photographs and topographic maps, ground reconnaissance, and phreatophyte mapping. The primary recharge area of the valley is defined where the principal aquifer and the sediments above the aquifer contain no continuous clay/silt confining layers greater than 20 ft thick. Recharge enters the principal aquifer in this area through the mechanisms discussed previously. A secondary recharge area is defined by areas where ground water may be moving through clay/silt confining layers but where natural pathways of ground-water discharge generally do not exist. The discharge area was delineated by Steiger and Lowe where ground water discharges to the surface through natural means. In the discharge area, the shallow unconfined aquifer receives most of its recharge by upward movement of water from the principal aquifer, through the shallow confining layer.

Water-Level Fluctuations

Water-level data for Tooele Valley indicate that long-term water-level fluctuations result from changes in precipitation and ground-water withdrawals. Initial water-level declines during 1953-63 at wells (C-3-5)7dcc-1 and (C-2-6)36dcc-1 (fig. 9) probably are the result of increased ground-water withdrawals that started in 1939 and stabilized by 1963 (see previous section titled "Wells"). Since the early 1960s, water levels have not shown significant declining or rising trends, with the exception of the mid- to late 1980s, when water levels rose rapidly. This response probably was caused mainly by greater-than-average precipitation and somewhat by decreased ground-water withdrawals (fig. 8). By the early 1990s, water levels in Tooele Valley had declined and were similar to levels observed in the early 1980s.

NUMERICAL SIMULATION OF THE GROUND-WATER SYSTEM

Modeling Approach

The modular, three-dimensional, finite-difference, numerical model known as MODFLOW (McDonald and Harbaugh, 1988) was used to simulate the ground-water flow system in Tooele Valley. The numerical model was constructed to test and refine the conceptual understanding of the ground-water system as defined in the previous sections of this report. Two simulations were developed: a steady-state simulation that represents conditions in 1968; and a transient-state simulation based on hydrologic data for 1969-94 that incorporates annual variations in recharge to and discharge from the ground-water system.

Requirements for numerical modeling include horizontal and vertical discretization of the ground-water system and establishing spatial distributions for the hydrologic properties. Also, mathematical boundaries that realistically depict actual hydrologic boundaries and conditions must be assigned to the model domain. The approach to mathematically simulating

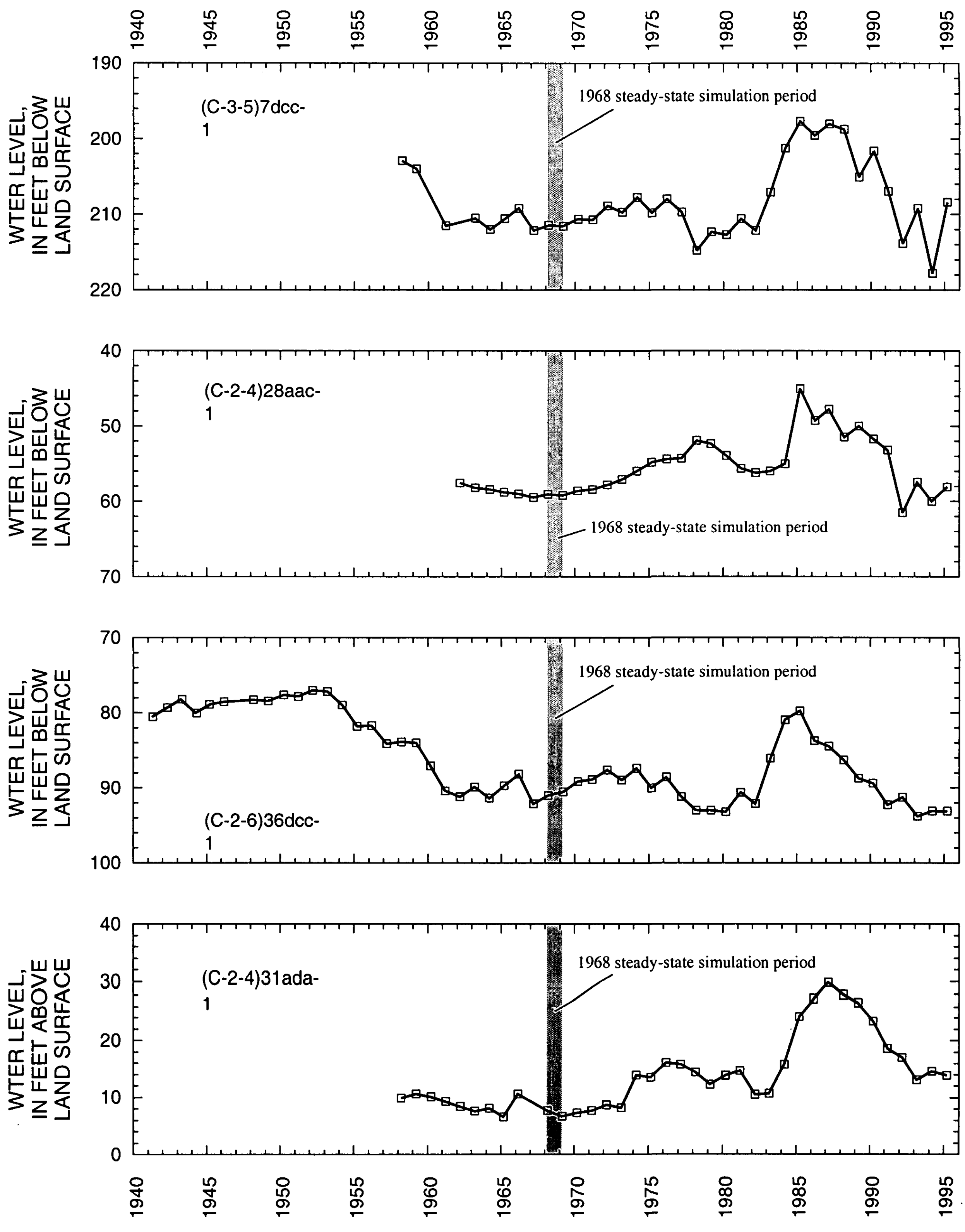


Figure 9. Long-term water-level fluctuations in four wells completed in the basin-fill material in Tooele Valley, Utah.

ground-water flow in Tooele Valley is presented in the following subsections. Where necessary, the requirements of the steady-state and transient-state simulations are distinguished.

Discretization

The ground-water flow system in Tooele Valley was subdivided horizontally into a grid of 110 rows and 118 columns (fig. 10), with grid dimensions varying from 1,000 ft in each direction to 1,000 ft by 2,160 ft. Vertically, the aquifer system is divided into five layers. The shallow unconfined aquifer and the underlying single shallow confining layer were represented by a single layer; model layer 1 (fig. 11). The bottom of model layer 1 corresponds to the base of the shallow confining layer and was set at 100 ft below land surface. The simulated saturated thickness of model layer 1 is variable and corresponds to model-computed water levels in the layer. The principal aquifer, which includes confined zones in the northern part of the valley, was divided into four layers (model layers 2 to 5, fig. 11). Model layers 2 and 3 are each 150 ft thick; where unconfined the simulated saturated thickness of model layer 2 may vary during problem solution. Model layer 4 is 300 ft thick. Model layer 5 ranges in thickness from 50 to 400 ft. The thickness of model layers 3 to 5 was not explicitly incorporated into the model but was implicitly incorporated in model input as transmissivity for those model layers. Vertical discretization of the principal aquifer allowed for improved simulation of the geometry of the basin-fill aquifer system and representation of vertical anisotropy.

In this report, an "i, j, k" indexing convention is used when discussing model cells and their location in the model grid. The value of the row index "i" corresponds to the row number shown in figure 10; the value of the column index "j" corresponds to the column number. The index "k" refers to model layers; layer 1 (k=1) is the top model layer and layer 5 (k=5) is the bottom model layer. The term "vertical column," as used in this report, is the set of model cells with the same row (i) and column (j) index.

Boundary Conditions

The no-flow boundary is used to simulate the assumed bottom of the principal aquifer. The farthest lateral extent of the principal aquifer, in most areas, also is simulated as a no-flow boundary. Exceptions are locations along the valley margins and near Great Salt Lake, where specific recharge and discharge processes occur. The northwestern border of the model area, between the tip of the Stansbury Mountains and Stansbury Island (fig. 1), is simulated as no-flow; ground-water flow in the area is nearly parallel to the boundary.

Specified-flux boundaries are used to simulate recharge entering the ground-water flow system as (1) inflow from consolidated rock at the consolidated-rock/basin-fill boundary at the margins of the valley, (2) underflow in stream-channel material at the mouths of canyons and seepage from streams that enter the valley, (3) subsurface inflow from Rush Valley through the Stockton Bar, (4) infiltration of precipitation on the valley floor, and (5) infiltration of unconsumed irrigation water from fields and lawns/gardens. Specified-flux boundaries also are used to simulate discharge from the ground-water flow system to wells. The specified-flux boundary condition allows the flow rate across a given boundary to be specified as a function of location and time. Flow rates across these boundaries are specified in the steady-state simulation and for each stress period of the transient-state simulation. Flow rates do not deviate at these boundaries during problem solution for each stress period and are not affected by simulated conditions in the ground-water flow system.

Head-dependent flux boundaries are used to simulate (1) discharge to springs, flowing wells, and shallow drains and ditches, and (2) discharge from evapotranspiration. A head-dependent flux boundary allows the flow rate across the boundary surface to change in response to changes in the model-computed water level for the aquifer adjacent to the boundary. Flow rate therefore is computed as a function of the water level in the adjacent aquifer and may vary during problem solution. Constant-head cells, in which flux is also a function of the water level in the aquifer, are used to simulate the interaction between the ground-water system and Great Salt Lake.

A head-dependent drain boundary is used to simulate discharge to springs and flowing wells, and seepage of shallow ground water to shallow drains and ditches in the northern part of the valley. The drain boundary allows for the simulation of ground-water discharge from the aquifer to a drain (or other sink being represented as a drain) but does not simulate seepage to the aquifer. When the model-computed water level (h) in a given cell containing a drain boundary is higher than the specified altitude (d) of the drain in that cell, flow is simulated from the aquifer to the drain (QD) according to the following equation (McDonald and Harbaugh, 1988, p. 9-3):

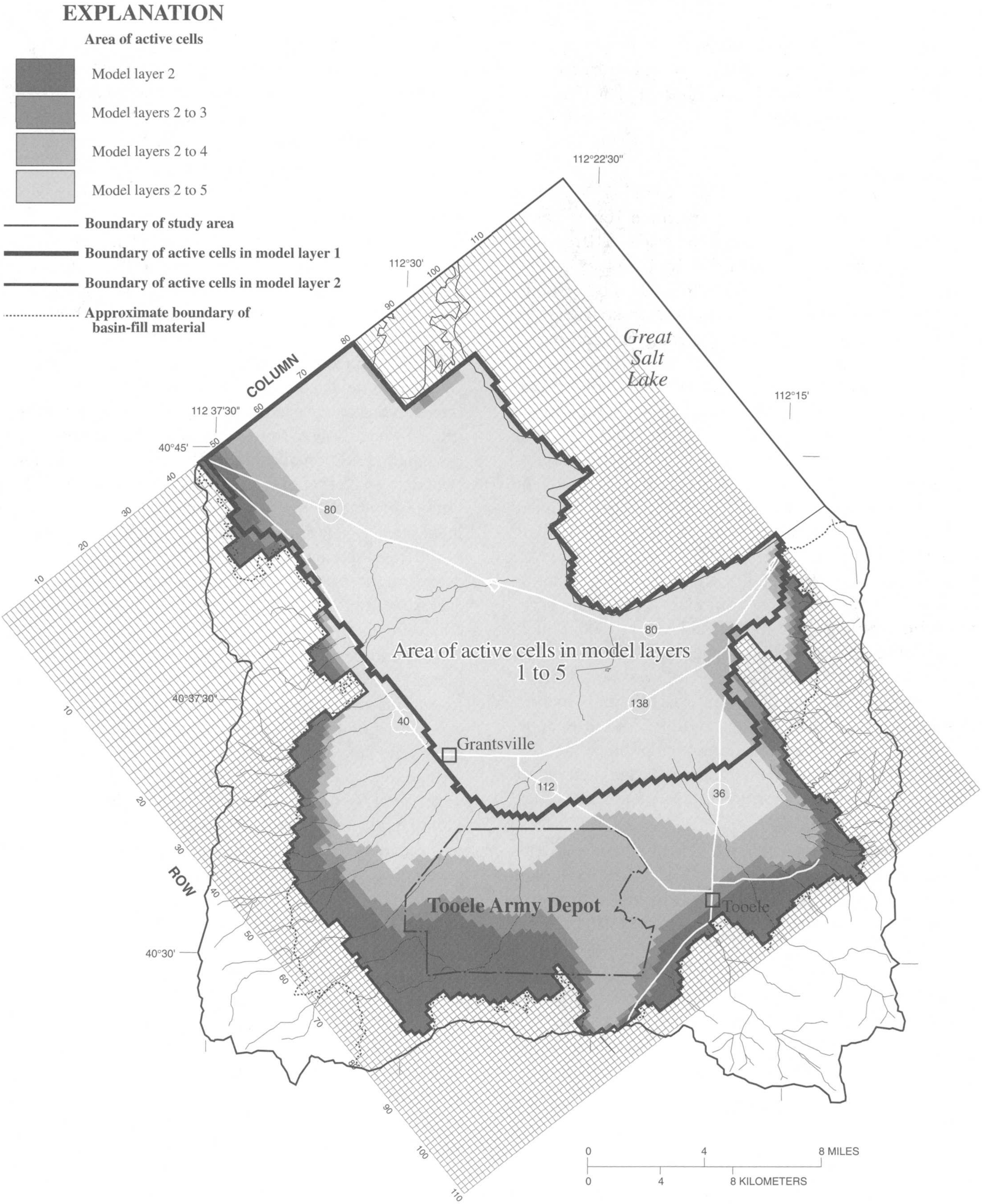


Figure 10. Grid and location of active cells in the ground-water flow model of Tooele Valley, Utah.

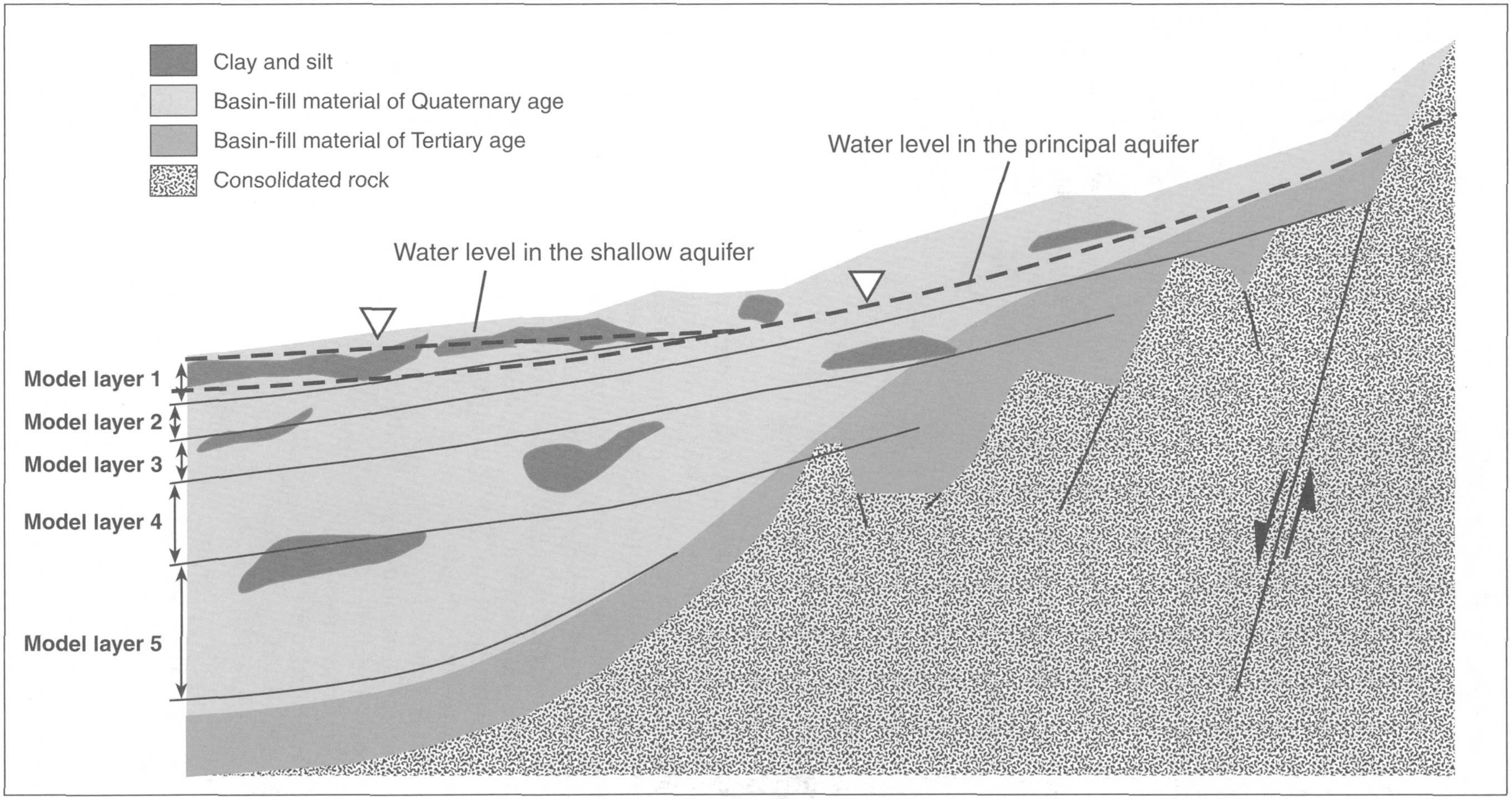


Figure 11. Generalized cross section showing model layers of the ground-water flow model of Tooele Valley, Utah.

$$QD = CD(h-d) \; for \; h>d \qquad (1)$$

where *CD* (drain conductance) is equal to the hydraulic conductance of the drain-aquifer interconnection. When the model-computed water level is lower than the specified drain altitude, no flow to or from the drain is simulated.

The model incorporates a linear head-dependent function to simulate ground-water discharge by evapotranspiration. In a given cell, the evapotranspiration rate will be equal to a specified maximum rate if the model-computed water level in the cell is higher than a specified altitude (typically land surface). If the model-computed water level is at a depth equal to or less than a specified extinction depth, the evapotranspiration rate will be zero. If the model-computed water level is between the land surface and the extinction depth, the evapotranspiration rate will decrease linearly from the maximum rate to zero.

Transmissivity, Storage Coefficient, and Vertical Leakance

In model layers 1 and 2, transmissivity varies spatially as a function of the saturated thickness of the layer and the equivalent horizontal hydraulic conductivity of the material in the layers. In model layers 3 to 5, which represent deep sediments of the principal aquifer, the saturated thickness of the layers was assumed to remain constant, and transmissivity was specified in all simulations and does not vary during problem solution.

In all of the area simulated by model layer 1 and in part of the area simulated by model layer 2 (specifically, the area where model layer 1 is not active, fig. 10), unconfined conditions are always assumed and changes in storage are computed by the model as a function of drainable porosity. Accordingly, the storage coefficient used in the model is equivalent to specific yield. In areas of the model represented in model layer 2 where the shallow unconfined aquifer is simulated (model layer 1 is active), the model allows for a storage coefficient that depends on the relation between the model-computed water level in the layer and the top of the layer. If the model-computed water level in layer 2 is higher than the top of the layer (confined conditions), then the change in storage caused by water-level changes is a function of the elastic properties of the aquifer and the water:

$$S_k = S_s b_k \qquad (2)$$

where

S_k = storage coefficient of layer k,
S_s = specific storage (L^{-1}), and
b_k = thickness of layer k.

If the model-computed water level in layer 2 is lower than the top of the layer (unconfined conditions), then changes in storage are a function of drainable porosity, and the storage coefficient (*S*) used by the model is equivalent to specific yield. The model does not convert from confined-aquifer storage coefficient to specific yield in model layers 3 to 5, where confined conditions are always assumed.

The model computes vertical conductance from a user supplied "vertical leakance" (McDonald and Harbaugh, 1988, p. 5-12). Vertical leakance between model layers varies as a function of equivalent vertical hydraulic conductivity and the thickness of sediments present between the midplanes of adjacent model layers. The vertical leakance (*VL*) between model layers *k* and *k+1* is:

$$VL = K'_{v,k+1/2} / b_{k+1/2} \qquad (3)$$

where

$K'_{v,k+1/2}$ = equivalent vertical hydraulic-conductivity value between the midplanes of model layers *k* and *k+1* (L/t), and

$b_{k+1/2}$ = distance between the midplanes of the two adjacent model layers (L).

Where adjacent model layers are characterized by different vertical hydraulic-conductivity values ($K_{v,k}$,$K_{v,k+1}$), such as layer 1 (shallow unconfined aquifer and shallow confining layer) and layer 2 (upper zone of the principal aquifer), the equivalent vertical hydraulic-conductivity value ($K'_{v,k+1/2}$) can be calculated with the following equation (Freeze and Cherry, 1979, p. 34):

$$K'_{v,k+1/2} = \frac{b_{k+1/2}}{(b_k/2)/K_{v,k} + (b_{k+1}/2)/K_{v,k+1}} \qquad (4)$$

where

b_k = thickness of layer *k* (L), and
b_{k+1} = thickness of layer *k+1* (L).

Parameter Estimation and Model Input

Data from previous work in the valley and data collected concurrent with this study that are described in previous sections of this report were used to evaluate model parameters and derive initial model input. Model parameters, other than those defining ground-water system geometry and discharge from pumped wells, were considered to be calibration variables that could be adjusted within previously defined ranges during model calibration. In the following subsections, the derivation and assignment of initial model-parameter values to active cells and model boundaries is discussed and probable ranges of parameters values used during model calibration are presented. Final estimates of model parameters resulting from model calibration are presented later in subsections of the "Model calibration" section of this report.

System Geometry

The location of active cells and no-flow boundaries in the model grid (figs. 10 and 11) represent the general geometry of the valley and the principal and shallow unconfined aquifers as defined in the "Hydrogeology of the ground-water system" section of this report. Location of no-flow boundaries in each model layer and the specified altitude of the tops and bottoms of model layers were determined mainly from estimates of (1) the horizontal extent and general thickness of the shallow confining layer and the shallow unconfined aquifer, (2) the altitude of the contact between consolidated rock and saturated basin-fill material near the margins of the valley, and (3) the general depth of the Tertiary-age sediments in the central and northern parts of the valley.

The area of active cells in model layer 1 (fig. 10) corresponds approximately with the area where the shallow aquifer and shallow confining layer were identified by Steiger and Lowe (1997) ("discharge" and "secondary recharge" areas in fig. 4) and where sediments above the base of that layer are simulated in the model as being saturated. Active cells in model layers 2 to 5 represent basin-fill material of Quaternary age in the principal aquifer, and in some areas, include sediments assumed to represent the upper zone of Tertiary-age basin-fill material and fractured consolidated rock. Tertiary-age basin-fill material in the valley is believed to be substantially less permeable than the younger Quaternary-age basin-fill material and may not be a significant avenue for ground-water movement in most areas. However, in some areas near the margins of Tooele Valley, saturated Quaternary-age basin-fill material does not exist and Tertiary-age basin-fill material underlying primary recharge areas of the valley plays a significant role in the ground-water flow system. For this reason, sediments assumed to be upper Tertiary-age basin-fill material were simulated as part of the principal aquifer for selected areas of the valley. Ground water flows through the uplifted block of fractured consolidated rock beneath Tooele Army Depot under a steep hydraulic gradient and this feature has significant effects on the local ground-water system. For this reason, the fractured rock in that area was simulated as part of the principal aquifer.

Active cells in model layers 2 to 5 were assigned on the basis of the type of material represented. All cells in layers 2 to 5 that contain saturated basin-fill material were defined as active. Cells below the basin-fill/consolidated-rock contact, other than in the area of Tooele Army Depot consolidated-rock high, were designated inactive. In areas where the thickness of basin-fill material exceeded the combined thickness of model layers 1 to 5 (1,100 ft maximum), the bottom of the principal aquifer was assumed to be the base of model layer 5. Saturated consolidated rock in the area of Tooele Army Depot is represented with active cells in model layers 2 to 4; model layer 5 is not active in the area.

Hydrologic Properties

To construct a numerical model of the Tooele Valley ground-water flow system, estimated values for parameters that describe hydrologic properties of the system were spatially distributed and assigned to all active model cells on the basis of the material represented by the cell. These parameters include horizontal hydraulic conductivity, transmissivity, vertical hydraulic conductivity, and aquifer storage. In many cases, parameter values assigned to a cell are composite (or equivalent value) because a model cell may represent sediment made up of different material types. For example, model layer 1 represents the shallow unconfined aquifer and the less permeable material of the underlying shallow confining layer. Estimates of hydrologic properties for model cells in layer 1, therefore, are equivalent values that are assumed to be representative of the combined properties of the materials in the model layer. The same is true for model cells in layers that represent the principal aquifer, which may represent coarse- and fine-grained basin-fill material, and consolidated rock. Approaches to calculating equivalent parameter values for active cells are discussed in the following subsections where necessary.

Horizontal Hydraulic Conductivity

Initial estimates for equivalent horizontal hydraulic-conductivity values incorporated in model layer 1 were made by dividing the layer into areas that correspond to the discharge and secondary recharge area of the valley (fig. 4). These delineations, made by Steiger and Lowe (1997, fig. 3), are based partly on data from drillers' logs that indicate the predominance of interbedded layers of clay, silt, and fine- to coarse-grained sand. Sediment in the discharge area is relatively finer grained, and horizontal hydraulic conductivity for the part of model layer 1 that corresponds with the discharge area was initially set at 5 ft/d. Sediment in the secondary recharge area is relatively coarser grained, and hydraulic conductivity for that area of model layer 1 was initially set at 30 ft/d.

The horizontal hydraulic conductivity of model layers 2 through 5, which represent the principal aquifer, was determined on the basis of the hydraulic conductivity of the material represented by the cell. Each vertical column of model cells in these layers may contain coarse- or fine-grained basin-fill material, or consolidated rock. The approach used to estimate horizontal hydraulic-conductivity values for active cells in these layers entailed defining estimates of hydraulic conductivity for these three material types for zones of the aquifer system and computing an equivalent hydraulic-conductivity value (K') for model cells with the following equation (Freeze and Cherry, 1979, p. 34):

$$K'_{i,j,k} = \frac{(K_{FINE,i,j,k}b_{FINE,i,j,k}) + (K_{CRSE,i,j,k}b_{CRSE,i,j,k}) + (K_{CON,i,j,k}b_{CON,i,j,k})}{b_{i,j,k}} \quad (5)$$

where

$K_{FINE,i,j,k}$ = hydraulic-conductivity value of fine-grained clay-silt intervals within model cell i,j,k (L/T),

$K_{CRSE,i,j,k}$ = hydraulic-conductivity value of coarse-grained gravel or sand/gravel intervals within model cell i,j,k (L/T),

$K_{CON,i,j,k}$ = hydraulic-conductivity value of consolidated rock within model cell i,j,k (L/T),

$b_{FINE,i,j,k}$ = total thickness of fine-grained clay/silt intervals within model cell i,j,k (L),

$b_{CRSE,i,j,k}$ = total thickness of coarse-grained sand or sand/gravel intervals within model cell i,j,k (L),

$b_{CON,i,j,k}$ = thickness of consolidated rock within model cell i,j,k (L), and

$b_{i,j,k}$ = thickness of model cell i,j,k (L).

The following estimates for the hydraulic conductivity of sediment types were initially used: 1 ft/d for K_{FINE}, 60 ft/d for K_{CRSE} in zone 1 (fig. 5), 120 ft/d for K_{CRSE} in zone 2 (fig. 5), and 1 ft/d for K_{CON}. For use during model calibration it was assumed that the K_{FINE} could range from 0.04 ft/d to 2.28 ft/d; K_{CRSE} for basin-fill material within zone 1 could range from 15 ft/d to 90 ft/d; K_{CRSE} in zone 2 could range from 15 ft/d to 600 ft/d; and K_{CON} could range from 0.01 ft/d to 35 ft/d. Initial estimates and probable ranges are based on data, the discussions presented in previous sections of this report, and values presented in table 1.

The initial thicknesses of coarse- and fine-grained intervals (b_{CRSE} and b_{FINE} in equation 5) within model layers 2 through 5 is based directly on the ratio of sand/gravel intervals to the total depth of material in the aquifer shown in figure 6. The ratio represents the percentage of the aquifer that is assumed to be made up of coarse-grained permeable basin-fill material.

In the southeastern part of Tooele Army Depot (fig. 1) initial values for K' incorporated in the model were derived directly from the results of the U.S. Army Corps of Engineers ground-water flow model of the depot (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994) (table 1).

Vertical Hydraulic Conductivity and Vertical Leakance

Vertical hydraulic conductivity (K_v) is not explicitly input to the model but is implicitly incorporated into the model as part of the vertical leakance (VL) term defined in equation 3. Initial estimates of K_v used to calculate VL were based on the presence and physical characteristics of clay/silt layers within the basin-fill material. In the discharge area of Tooele Valley (fig. 4) where bedded clays and silts predominate, a K_v value of 0.001 ft/d was initially assigned for model layer 1 and 0.01 ft/d for layers 2 to 5. In the secondary recharge area (fig. 4) where the shallow confining layer exists but clay layers are less predominant at depth and are assumed to be less continuous, a K_v value of 0.001 ft/d was initially assigned for model layer 1 and a K_v value of 0.1 ft/d was used for model layers 2 to 5. For the primary recharge area of the valley (fig. 4), where continuous clay layers generally are not present, a K_v value of 1.0 ft/d was initially assumed for model layers 2-5 (model layer 1 does not extend into the primary recharge area). At Tertiary and Quaternary-age faults (fig. 2), the relatively high K_v value of 5.0 ft/d was initially assigned to all model layers.

The probable range of vertical hydraulic-conductivity values (K_v) used during model calibration varies for different areas of the valley. In the discharge and secondary recharge areas of the valley (fig. 4), K_v was assumed to vary from 1.0 x 10^{-5} ft/d to 1.0 ft/d. Near Tertiary and Quaternary-age faults and in the primary recharge area of the valley (figs. 2 and 4), K_v was assumed to vary from 0.2 ft/d to 30.0 ft/d.

Storage Coefficient and Specific Yield

Simulation of a transient-state flow system requires specification of storage coefficient and specific yield. Storage-coefficient values for active model cells ($S_{i,j,k}$) representing the principal aquifer were determined by multiplying an estimate for specific storage (S_s) and the cell thickness ($b_{i,j,k}$) using a form of equation 2. Initial estimates for S were defined using a value of 2.0 x 10^{-6} for S_s ft^{-1}. A probable range of S_s for use during model calibration was assumed to be 2.2 x 10^{-7} ft^{-1} to 4.7 x 10^{-6} ft^{-1}.

An initial estimate of 0.1 was used for the specific yield of the unconfined areas of Tooele Valley. Those areas are simulated by model layer 1 and parts of model layer 2. A probable range for specific yield was assumed to be 0.05 to 0.30.

Recharge Simulated at Specified-Flux Boundaries

The numerical model simulates recharge to the Tooele Valley ground-water system from four major sources using specified-flux boundaries: (1) inflow from consolidated rock and stream-channel deposits, (2) infiltration of precipitation, (3) seepage of unconsumed irrigation water, and (4) subsurface inflow from Rush Valley.

The amount of recharge simulated for steady-state conditions is representative of long-term averages. In the transient-state simulation, recharge rates from these sources of recharge were varied with time. Initial specified recharge rates for the steady-state simulation and the methods used to vary annual recharge for the transient-state simulation are presented in the following sections.

Inflow from Consolidated Rock and Stream-Channel Deposits, and Rush Valley

Initially, 47,700 acre-ft/yr of recharge was specified as inflow from consolidated rock and stream-channel deposits (see section titled "Hydrology of the ground-water system"). Inflow from Rush Valley was initially estimated at 5,000 acre-ft/yr. These initial rates of recharge at the basin-fill material/consolidated-rock boundary represent the long-term average inflows and are simulated as injection wells using the Well Package specified-flux boundary formulation (McDonald and Harbaugh, 1988, p. 8-1). The injection wells are placed, with a few exceptions due to local factors, at model cells in layers 2 and 3 that simulate the edge of the principal aquifer adjacent to the Oquirrh and Stansbury Mountains (fig. 12). Areally, recharge is distributed to closely simulate the conceptual distribution discussed in previous sections of the report.

Initially, recharge along the Oquirrh Mountains was set at about 37,000 acre-ft/yr. To simulate enhanced recharge amounts at canyon mouths, 8,000 acre-ft/yr (about 20 percent of the recharge) was assigned, arbitrarily, to model cells simulating stream-channel deposits near the mountain front. Two areas, 1 represented by 6 model cells at the mouth of Settlement Canyon, and the other represented by 14 cells at the mouth of Middle Canyon (fig. 12), were each assigned 3,000 acre-ft/yr of recharge. Cells at the mouth of Pine Canyon were assigned 1,000 acre-ft/yr. The remaining 1,000 acre-ft/yr of recharge was associated with Bates Canyon and two additional canyon mouths toward the north end of the Oquirrh Mountains. The remaining 29,000 acre-ft/yr was assigned according to the conceptual distribution discussed earlier; 50 percent (about 14,000 acre-ft/yr) was uniformly distributed at wells in 65 model cells in layers 2 and 3 in the area between the east side of Stockton Bar and the mouth of Middle Canyon; 35 percent (about 10,000 acre-ft/yr) was assigned to wells in 28 model cells simulating the area between Middle Canyon and Dry Canyon; and 15 percent (about 5,000 acre-ft/yr) was assigned to wells in 115 model cells simulating the area between Dry Canyon and the northern end of the Oquirrh Mountains (fig. 12).

Recharge along the Stansbury Mountains was initially set at about 11,000 acre-ft/yr. Because a large number of streams originate from the Stansbury Mountains (fig. 12), about 75 percent (chosen arbitrarily) of the total recharge was initially simulated to occur near canyon mouths. This amount represents about 8,000 acre-ft/yr, the same as is concentrated at canyon mouths

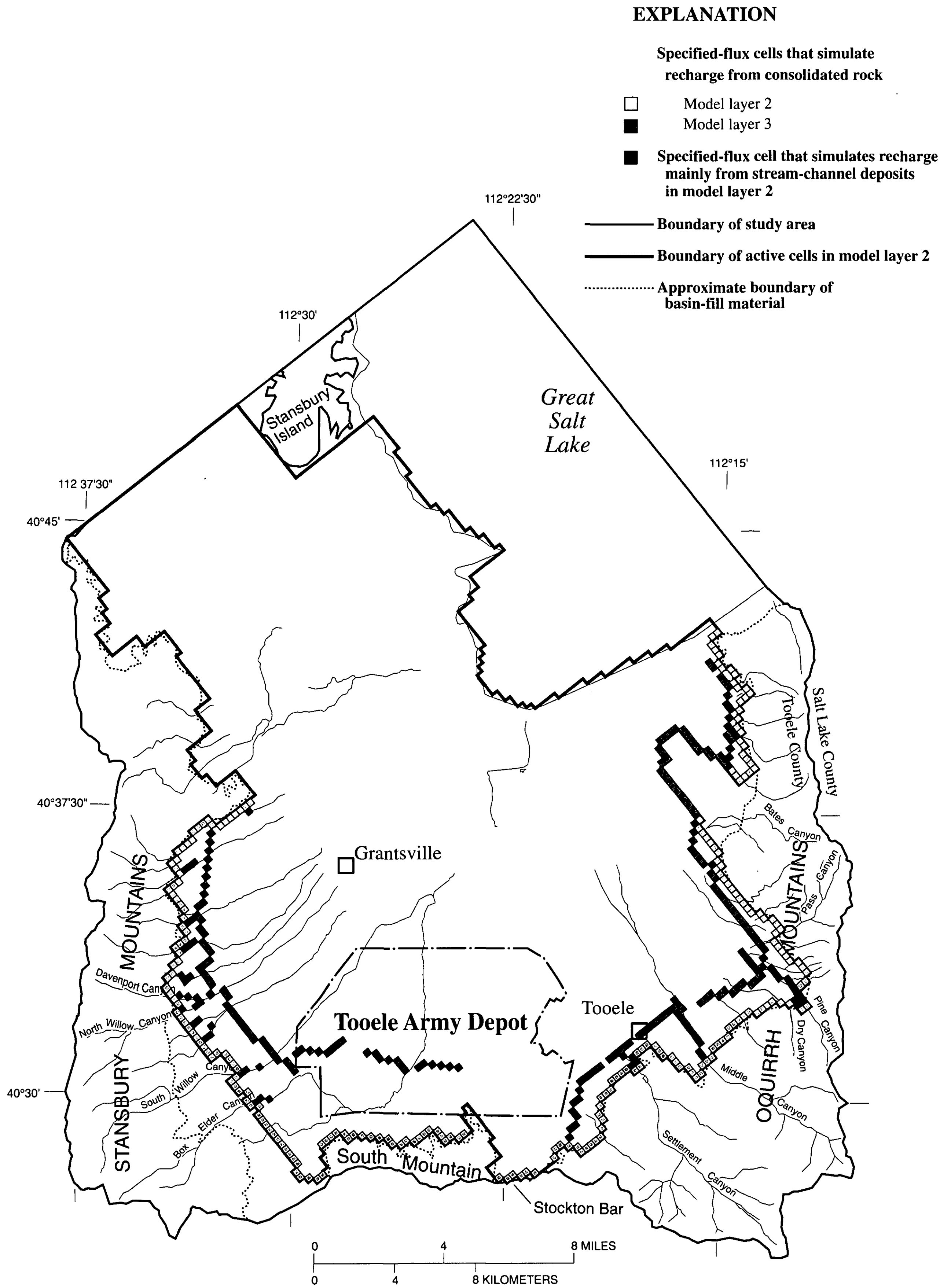


Figure 12. Location of specified-flux cells used to simulate recharge from consolidated rock and stream-channel deposits in the ground-water flow model of Tooele Valley, Utah.

along the Oquirrh Mountains, but the percentage of total recharge is about 4 times as great. Initially, 1,500 acre-ft/yr of recharge was assigned uniformly to nine model cells simulating recharge from the two northern most simulated streams (fig. 12). Another 1,500 acre-ft/yr was evenly distributed to wells in 24 model cells that simulate 7 streams; 3 streams north of Davenport Canyon, Davenport Canyon, North Willow Canyon, and the 2 streams south of North Willow Canyon (fig. 12). Two areas, model cells at the mouth of South Willow Canyon, and model cells at the mouth of Box Elder Canyon (fig. 12), were each assigned 2,500 acre-ft/yr of recharge. The remaining 3,000 acre-ft/yr was distributed uniformly at wells in 131 model cells in layers 2 and 3.

Subsurface inflow from Rush Valley is simulated at six model cells located adjacent to Stockton Bar in layer 2. This recharge was initially set at 5,000 acre-ft/yr. Recharge from South Mountain was initially set at about 500 acre-ft/yr and was uniformly distributed at wells in 37 model cells simulating layers 2 and 3. The cells are located directly north of South Mountain (fig. 12).

Recharge was concentrated near the mouths of canyons to represent inflow from consolidated rock and stream-channel deposits at the base of mountain drainages. As indicated in figure 12, specified-flux cells were placed along stream channels leading out of the canyons to simulate inflow from stream-channel deposits and seepage from streams near the mouths of canyons. Seepage from the bottom of stream channels to the principal aquifer where channels traverse the primary recharge area is probably not substantial during most years because stream water is diverted for irrigation near canyon mouths. Recharge to the principal aquifer from stream channels may be significant, however, during periods of greater-than-average precipitation when excess water is allowed to flow down the channels in the spring.

Specified recharge at the basin-fill material/consolidated-rock boundary and from Rush Valley was varied with time in the transient-state simulation on the basis of the assumption that annual recharge at the boundary varies with changes in annual precipitation in the surrounding mountains. Water stored as snow in the surrounding mountains becomes available to recharge the consolidated-rock aquifers and stream-channel deposits when temperatures rise during the spring. Estimates of annual recharge from consolidated rock ($QCON_{yr}$) and stream-channel deposits and streams ($QSTRM_{yr}$) for yearly stress periods in the transient-state simulation were made by assuming that the amount of recharge determined during steady-state calibration ($QCON_{ss}$ and $QSTRM_{ss}$) varied as a function of the ratio of the amount of water stored as snow (snow-water equivalent or SWE) in Middle Canyon in April for a given year ($MSNO_{yr}$) (fig. 8) to the average SWE in April during 1963-94 in Middle Canyon ($MSNO_{ave}$):

$$QCON_{yr} = QCON_{ss}[(((MSNO_{yr}/MSNO_{ave}) - 1) \times C) + 1] \qquad (6)$$

and $$QSTRM_{yr} = QSTRM_{ss}(((MSNO_{yr}/MSNO_{ave}) - 1) \times C) + 1 \qquad (7)$$

The coefficient C in these equations was used during transient-state calibration as a variable to adjust the simulated effect of fluctuations in the amount of water stored as snow (SWE) in the adjacent mountains on recharge to the principal aquifer at the basin-fill material/consolidated-rock boundary. For example, if C is set at zero, then $QCON_{yr}$ is equal to $QCON_{ss}$ and $QSTRM_{yr}$ is equal to $QSTRM_{ss}$ for all stress periods simulating no effect from annual fluctuation in SWE. If C is set at 1, then recharge as inflow from consolidated rock and stream-channel deposits is simulated as varying proportionately with $MSNO_{yr}/MSNO_{ave}$. Separate equations were used to calculate $QCON_{yr}$ at the basin-fill material/consolidated-rock boundary and $QSTRM_{ss}$ at canyon mouths because different probable ranges for the coefficient C were assumed for the two sources during calibration. Annual recharge from consolidated rock was assumed to change, at the most, proportionately with annual changes in SWE. Thus values for C used in equation 6 during calibration did not exceed 1. It was assumed that the effect of deviation from the average SWE in the mountains on recharge near canyon mouths could be magnified relative to a proportionate relation. This concept is based on the observation that during wet periods, stream channels carry more water across a larger area for a longer time. Not only are water levels in adjoining stream-channel deposits higher during these periods, but streams that do not generally flow across the primary recharge area may be full for a substantial amount of time in the spring and early summer months. For equation 7, C was adjusted during calibration from 0 to 3.

Infiltration of Precipitation

Recharge from precipitation on the valley floor was incorporated into the model as a specified flux at the uppermost active cells of vertical columns. Initial rates and distribution of recharge from precipitation were simulated to approximate the distribution in the valley (table 2 and fig. 7). The total annual recharge from precipitation initially incorporated in the model was about 12,000 acre-ft.

Estimates of recharge from precipitation on the valley floor for yearly stress periods in the transient-state simulation ($QPRE_{yr}$) were made by assuming that the recharge determined during steady-state calibration ($QPRE_{ss}$) varied as a function of the ratio of annual precipitation at Tooele (PRE_{yr}) to average annual precipitation at Tooele (PRE_{ave}) (fig. 8):

$$QPRE_{yr} = QPRE_{ss}[(((PRE_{yr}/PRE_{ave}) - 1) \times C) + 1] \qquad (8)$$

For transient-state calibration, the coefficient C in equation 8 was initially set at 0.75. It was assumed during calibration that C could not exceed 1.

Seepage of Unconsumed Irrigation Water

Recharge to the ground-water system by infiltration of unconsumed irrigation water from fields and lawns/gardens was incorporated in the model as 25 percent of applied irrigation water in southeastern Tooele Valley (see section titled "Hydrology of the ground-water system," "Seepage of unconsumed irrigation water"). This percentage was assumed for the entire valley to compute rates of recharge from irrigated land throughout the model area. Initially, recharge by infiltration of unconsumed irrigation water from fields and lawns/gardens was specified at 9,700 acre-ft/yr.

The model cells where recharge from unconsumed irrigation water was simulated was determined on the basis of land-use data obtained from the Utah Department of Natural Resources, Division of Water Resources (written commun., 1996). Irrigated areas were identified using this data and the corresponding model cells were assigned a specified flux using the Recharge Package boundary formulation (McDonald and Harbaugh, 1988, p. 7-1). Recharge was simulated at the uppermost active cells of vertical columns in the defined areas shown in figure 13. Recharge was distributed to all cells indicated in figure 13 on the basis of the percentage of cell area that represented irrigated fields and (or) the percentage that represented residential or commercial land.

Specified recharge from unconsumed irrigation water was varied with time in the transient-state simulation on the assumption that annual recharge from irrigation water varies with changes in the amount of available irrigation water diverted from streams and springs. Estimates of recharge from irrigation water for yearly stress periods in the transient-state simulation ($QIRR_{yr}$) were made by assuming that the recharge determined during steady-state calibration ($QIRR_{ss}$) varied as a function of the ratio of the quantity of water as snow (SWE) in Middle Canyon in April before spring runoff for a given year ($MSNO_{yr}$) (fig. 8) to the average April SWE in Middle Canyon during 1964-94 ($MSNO_{ave}$):

$$QIRR_{yr} = QIRR_{ss}[(((MSNO_{yr}/MSNO_{ave}) - 1) \times C) + 1] \qquad (9)$$

This concept is based on the assumption that during wet periods, streamflow and spring discharge are greater than average. Consequently, more irrigation water may be distributed to fields for a longer period of time and more field acreage may be irrigated. The coefficient C in this equation allows the effects of SWE on irrigation recharge to be either muted or exaggerated. Initially the coefficient C was set equal to 1 in the transient-state simulation. It was assumed during calibration that the value could range up to 3.

Specified Discharge from Pumped Wells

Withdrawal from pumped wells in the principal aquifer was simulated using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Specific discharge rates at cells that simulate ground-water withdrawal from public-supply, irrigation, and industrial wells (fig. 8) were based on annual withdrawal data reported by water users and on unpublished records of the U.S. Geological Survey for 1964-91. Discharge from pumped wells was distributed vertically mainly on the basis of available well-construction data that described the depth of the wells and the depth of well-casing perforations. If no construction data were available for a well, total discharge of the well was distributed equally to all active cells of model layers 2 to 5 in the vertical column containing that well.

In January 1994, the U.S. Department of Army began extracting ground water contaminated with solvents at Tooele Army Depot, treating the water, and injecting it back into the principal aquifer in the northeastern part of the depot. Rates of extraction and injection reported by the U.S. Army Corps of Engineers Hydrologic Engineering Center (1995, Appendix D; and 1996, tables 5.1-5.3) for 1994 were incorporated into the model during the final stress period of the transient-state simulation.

Head-Dependent and Constant-Head Boundaries

Discharge from valley springs, flowing wells, and shallow drains and ditches was simulated by the ground-water flow model at head-dependent bound-

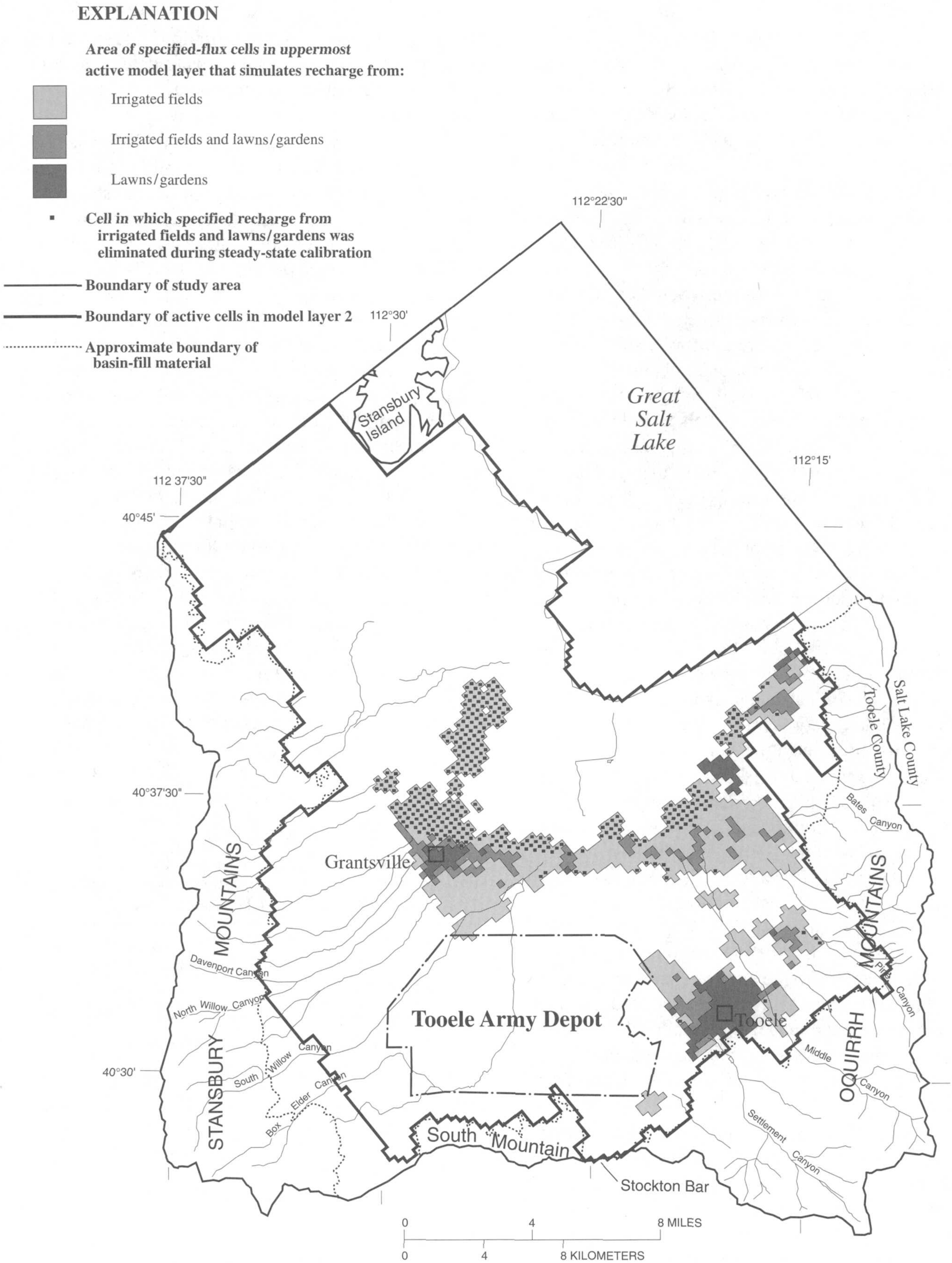


Figure 13. Location of specified-flux cells used to simulate recharge from irrigated fields and lawns/gardens during development and calibration of the ground-water flow model of Tooele Valley, Utah.

aries using the Drain Package (McDonald and Harbaugh, 1988, p. 9-1) (eq. 1). Drain cells were located in model layer 1 (fig. 14) at Mill Pond Springs, Dunne's Pond Springs, and at the sources of Fishing Creek and Sixmile Creek to simulate discharge from those springs.

Altitudes of drain boundaries simulating discharge from springs were specified to approximate the altitude of the orifices of the valley springs from U.S. Geological Survey 7.5-minute-quadrangle topographic maps. Drain conductance for these boundaries is not specified on the basis of measurements; rather, initial values used in the model were arbitrarily selected and adjusted during model calibration to obtain a match between computed and observed spring discharge and computed and measured water levels near the springs.

Head-dependent drain cells also were used to simulate discharge from flowing wells in model layer 2 in the northern part of the valley (fig. 14). By specifying the altitude of land surface as the drain altitude, discharge to flowing wells is not simulated when computed water levels in drain cells are below the land surface. Land-surface altitudes were determined from U. S. Geological Survey 7.5-minute quadrangles. Assigned conductance values for boundary cells were based on the number of flowing wells in the cell: if a given model cell simulates more than one flowing well, the drain conductance was multiplied by the number of flowing wells simulated at the cell.

Head-dependent drain cells are located in model layer 1 (fig. 14) to simulate ground-water discharge from the shallow unconfined aquifer to surface drains and ditches, and to ponds and stream channels in areas where the water table is shallow. Discharge to these sinks has not been previously measured, although observations verify that base flow does occur. Drain altitudes were specified from topographic maps of the area and from depths to the base of drains observed in the field. Drain conductance is not measured, and initial values used in the model were arbitrarily selected and adjusted during model calibration to maintain computed water levels near land surface in the areas of the drains.

Discharge by evapotranspiration was simulated in the ground-water flow model at head-dependent boundaries by using the Evapotranspiration Package (McDonald and Harbaugh, 1988, p. 10-1). Implementation of evapotranspiration in the numerical model varies somewhat from the framework presented in Razem and Steiger (1981, p. 16) and discussed in the "Hydrology of the ground-water system" section. These variations in evapotranspiration classifications, rates, and areas are the result of better delineation of land-use and updated consumptive-use estimates for phreatophytes.

The assignment of evapotranspiration cells in model layer 1 was based on current land-use information. Three major land-use categories were used for the purpose of specifying evapotranspiration rates to boundary cells in the model (fig. 15): (1) bare ground, mud flats, or temporary shallow water, (2) cultivated land, and (3) phreatophytes. Initial maximum evapotranspiration rates assigned to the boundaries are dependent on the land-use category and in the specific case of phreatophytes, predominant plant type, and are listed in table 3. The maximum evapotranspiration rates for non-phreatophyte areas are the same as those estimated by Razem and Steiger (1981, table 3). Maximum evapotranspiration rates for phreatophyte types are based on estimates by Lambert (1995, table 1) for phreatophytes near Great Salt Lake in Salt Lake Valley. These estimates were based on formulas derived by Blaney and Criddle (1962) for consumptive use during an entire growing season and assume 100-percent foliage density. Although 100-percent foliage density does not occur in most of the evapotranspiration area, initial incorporation of these values in the model allowed for the simulation of relative differences in evapotranspiration rates between land categories and plant types. During calibration, evapotranspiration rates were adjusted from initial values in table 3. The extinction depth at which evapotranspiration is assumed to be zero was set at 15 ft.

Table 3. Maximum evapotranspiration rate for three major land-use categories used during construction and calibration of the ground-water flow model of Tooele Valley, Utah

Land-use category	Maximum evapotranspiration rate (feet per year)
Bare ground/mud flat	0.09
Cultivated land	.38
Phreatophytes:	
Predominantly—	
Pickleweed	2.47
Greasewood	2.67
Salt Grass	2.28

Constant-head cells are located in model layer 1 along the northern border of the model area, which rep-

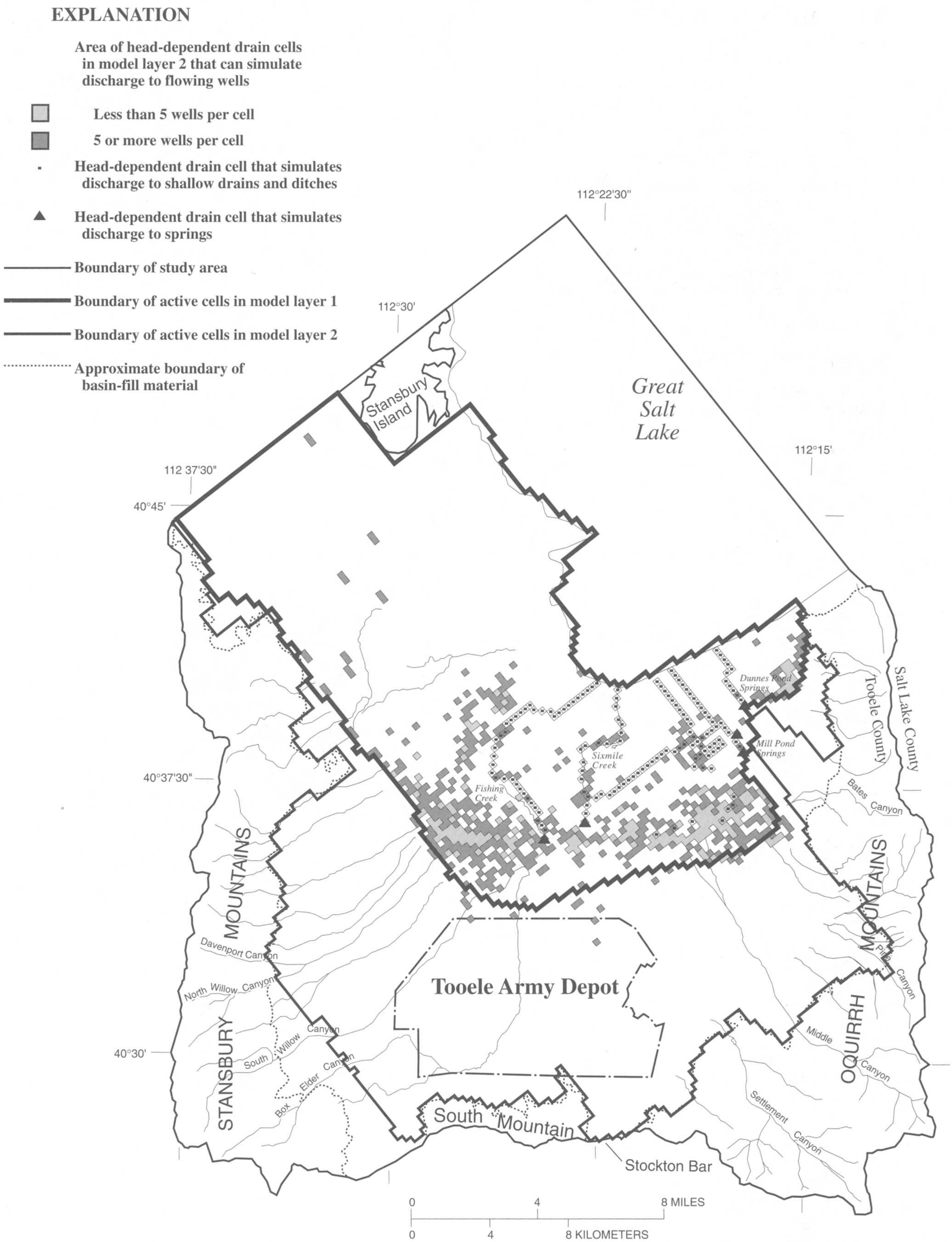


Figure 14. Location of head-dependent drain cells that can simulate discharge to flowing wells, springs, and drains in the ground-water flow model of Tooele Valley, Utah.

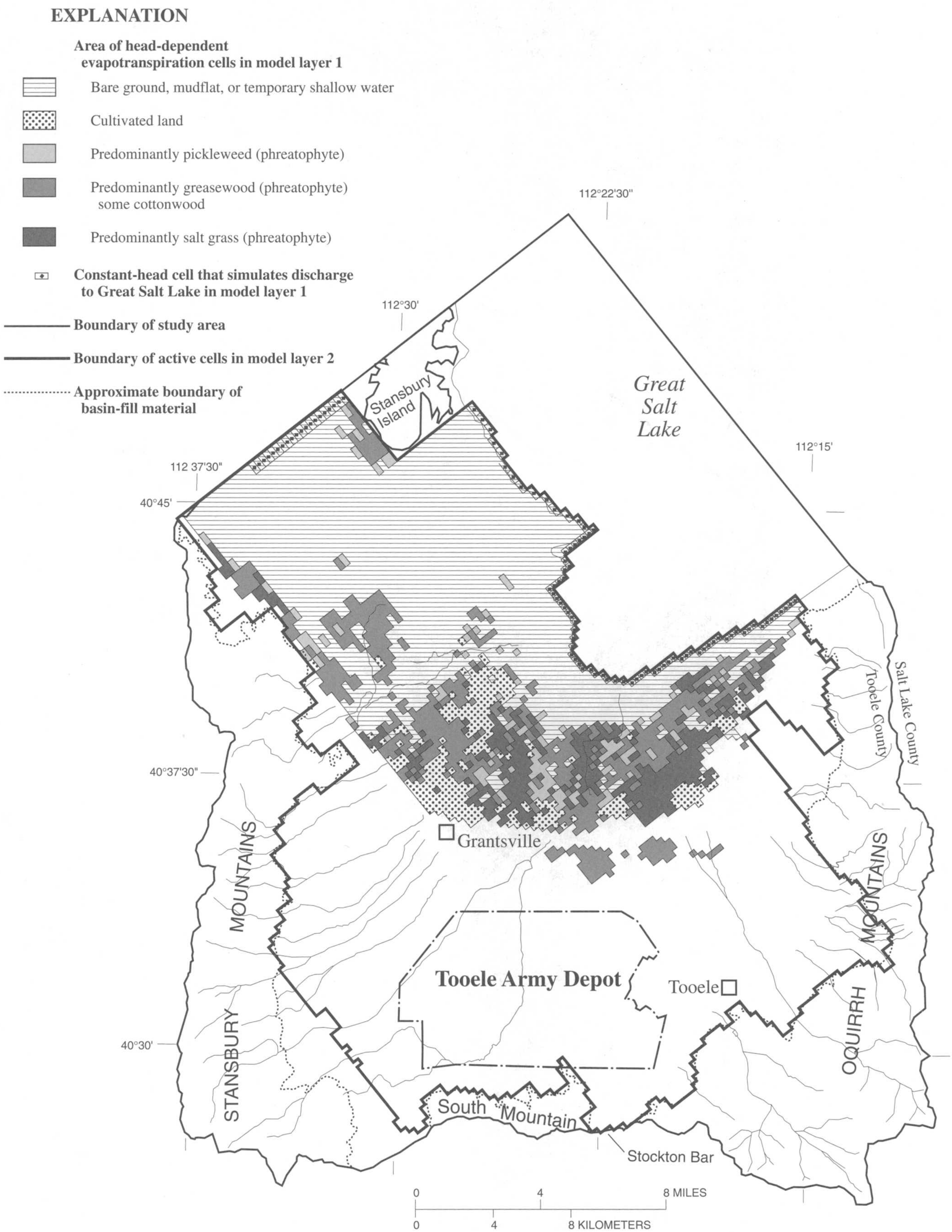


Figure 15. Location of head-dependent evapotranspiration cells and constant-head cells that simulate discharge to Great Salt Lake in the ground-water flow model of Tooele Valley, Utah.

resents the shore of Great Salt Lake (fig. 15). Specified water-level altitude in these cells was set at 4,200 ft, the approximate average historic altitude of Great Salt Lake (Arnow and Stephens, 1990, p. 1). The density of ground water and water in Great Salt Lake varies spatially in this area. The density of salt water in the southern end of Great Salt Lake, which may be as much as 5 ft deep in the area of constant-head cells, is typically about 1.10 g/cm^3 for lake-level altitudes near 4,200 ft (ReMillard and others, 1993, p. 18, and ReMillard and others, 1994, p. 179). Less saline ground water, of lower density, occurs beneath the edge of the lake and to the south of the lake. Density variations in a flow system may create pressure gradients within the system that are not indicated from measured water-level altitudes. Mechanisms used to normalize water-level measurements near the lake shore relative to a constant, however, resulted in an adjustment of less than 1 ft in the lake altitude at the edge of the lake. Therefore, the average value of 4,200 ft was assumed to be a reasonable specified water-level altitude for constant-head cells located at the lake shore.

Model Calibration

Steady-State Calibration

Method

A steady-state flow-model simulation was developed to represent conditions in 1968. Water-level fluctuations for 1964-68 (fig. 9) indicate only small yearly changes throughout the valley during this period. Thus, recharge was assumed to be about the same as discharge and the hydrologic conditions in 1968 were assumed to be near steady-state. The simulation incorporated specified rates of recharge from natural sources that were assumed to represent long-term average rates as defined in the "Hydrology of the ground-water system" section of this report. Discharge from pumping wells incorporated in the 1968 steady-state simulation was specified as the annual average pumpage during 1964-68.

Two measures of the state of the ground-water flow system were used during steady-state calibration: (1) measured water levels in the principal aquifer, and (2) estimated flow at model boundaries including discharge at four valley springs. A comparison of the total simulated ground-water budget and the average long-term ground-water budget derived previously in this report was used during calibration to help evaluate the fit of the model to measured conditions.

Water levels computed during the 1968 steady-state simulation were compared with water levels measured primarily during February 1969 in 56 wells completed in the principal aquifer. For each individual model run, three statistics that describe the difference between model-computed water levels and measured water levels (residuals) were calculated and analyzed to determine the accuracy of the simulation. The three statistics are (1) the mean of the residuals, which indicates the bias in the distribution of positive and negative values, (2) the standard error, which is the mean of the absolute values of residuals, and (3) the standard deviation of the residuals. In addition, the match between model-computed and measured water levels was evaluated at individual observation sites. An observation site was considered to be calibrated when the model-computed water level in the cell containing the site was within a predetermined range of the measured water level. The criteria for the range were determined on the basis of the estimated accuracy of the water level and the observed horizontal hydraulic gradient across the model cell that contained the observation site; therefore, near the margins of the valley and in other areas where the gradient is steep, the acceptable range was large, as much as 50 ft. For most sites in the northern part of the valley, where the horizontal gradient is much smaller, a limit of 10 ft generally was chosen.

Few water-level data were available from wells in the shallow unconfined aquifer for use during calibration. The water table in the discharge area of the valley (fig. 4) is generally known to be within 1 ft to 10 ft of land surface, and during calibration computed water levels were compared to land-surface altitude in an attempt to maintain computed levels near land surface.

During calibration, model-computed flow rates at cells that contain drain boundaries simulating discharge from springs were compared to estimated annual discharge at individual springs. Simulated flow at springs was affected by changes in most calibration variables including specified recharge representing flow from natural sources. Simulated flow at springs was most sensitive to adjustments of drain conductance (*CD*) and vertical hydraulic conductivity (K_v) at cells in model layer 1 that contain the drain boundaries and cells in underlying layers. As stated earlier in this report, it was assumed that flows at four springs, Dunne's Pond Springs, Mill Pond Springs, and the sources of Sixmile and Fishing Creek, are related to faults or fault zones that act as vertical conduits for the flow of ground water to land surface. The simulated discharge at these springs was adjusted during calibration by choosing reasonable values for K_v for the sediments underlying the springs and then varying *CD* at the boundary. The specified elevation of the drain representing the spring was not adjusted during calibration.

Results of Calibration

Final statistics for residuals for the steady-state simulation (table 4) indicate that steady-state calibration resulted in a reasonable match between model-computed and measured water levels throughout most of the modeled area. The model-computed potentiometric surface in model layer 2, which represents the upper zone of the principal aquifer, and residuals for observation wells in the principal aquifer for the steady-state simulation are shown in figure 16. The vertical locations of observation wells were factored into the computation of residuals. The distributions of residuals for individual observation sites generally do not indicate a bias in the distribution of positive and negative values. In the Erda area near a zone of high horizontal gradient, however, computed water levels in the steady-state simulation generally are lower than observed levels. Efforts to improve the match between computed and measured water levels in this area by adjusting rates of simulated recharge from the consolidated rock east of Erda increased water levels in that area but resulted in computed water levels that were too high in the general southeastern region of the model area. Efforts to improve the match by adjusting horizontal hydraulic conductivity from final calibrated values in the Erda area and in the area downgradient of Erda did not result in an improved match between computed and measured water levels in the region.

Residuals at four sites (three of them near Erda) exceeded their prescribed calibration criteria. These sites, however, are in proximity to other observation sites where the match between model-computed and measured water levels is satisfactory. Attempts to improve the match at these sites by adjusting calibration parameters resulted in a worse match between computed and measured water levels at surrounding calibration sites.

Model-computed annual rates of ground-water discharge to springs are compared with estimated average annual rates in table 5. Total model-computed discharge to valley springs in the steady-state simulation was about 14,600 acre-ft. The match between model-computed and the estimated discharge at the sources of Sixmile and Fishing Creeks is reasonably good. Computed discharge at Dunne's Pond Springs is substantially lower than estimated discharge, and computed discharge at Mill Pond Springs is higher than estimated discharge. Total computed discharge at the two springs is 9,300 acre-ft/yr, which is about 85 percent of the total estimated discharge for the two springs. Efforts to improve the match between computed and estimated discharge at Dunne's Pond Springs by adjusting calibration parameters that define recharge at the basin-fill material/consolidated-rock boundary east of the springs, drain conductance at the boundaries representing Dunne's Pond Springs and Mill Pond Springs, and vertical hydraulic conductivity beneath the springs resulted in increased simulated flow to the spring but also resulted in a worse match between computed and measured water levels in the region and between computed and estimated flow at Mill Pond Springs. These springs are located near the hydrologic boundary that exists between the basin-fill material and consolidated rock that extends from the western side of the Oquirrh Mountains. The hydraulic characteristics of this boundary are not well understood and therefore difficult to simulate accurately. As discussed earlier in the report, it is possible that a part of the ground-water discharge from Dunne's and Mill Pond Springs originates directly from the fractured consolidated rock near the springs.

Table 4. Statistical differences between model-computed and measured water levels in the steady-state simulation of the ground-water flow model of Tooele Valley, Utah

[Values calculated as model-computed minus measured water level]

Number of comparisons	56
Mean	2.4 feet
Standard error (mean of absolute value of differences)	8.2 feet
Standard deviation	11.0 feet
Maximum difference (lower than measured)	-26.0 feet
Maximum difference (higher than measured)	33.2 feet

The steady-state ground-water budget (table 6) matches reasonably well with conceptual budget components presented in previous sections of this report. The discrepancy in the conceptual budget estimates

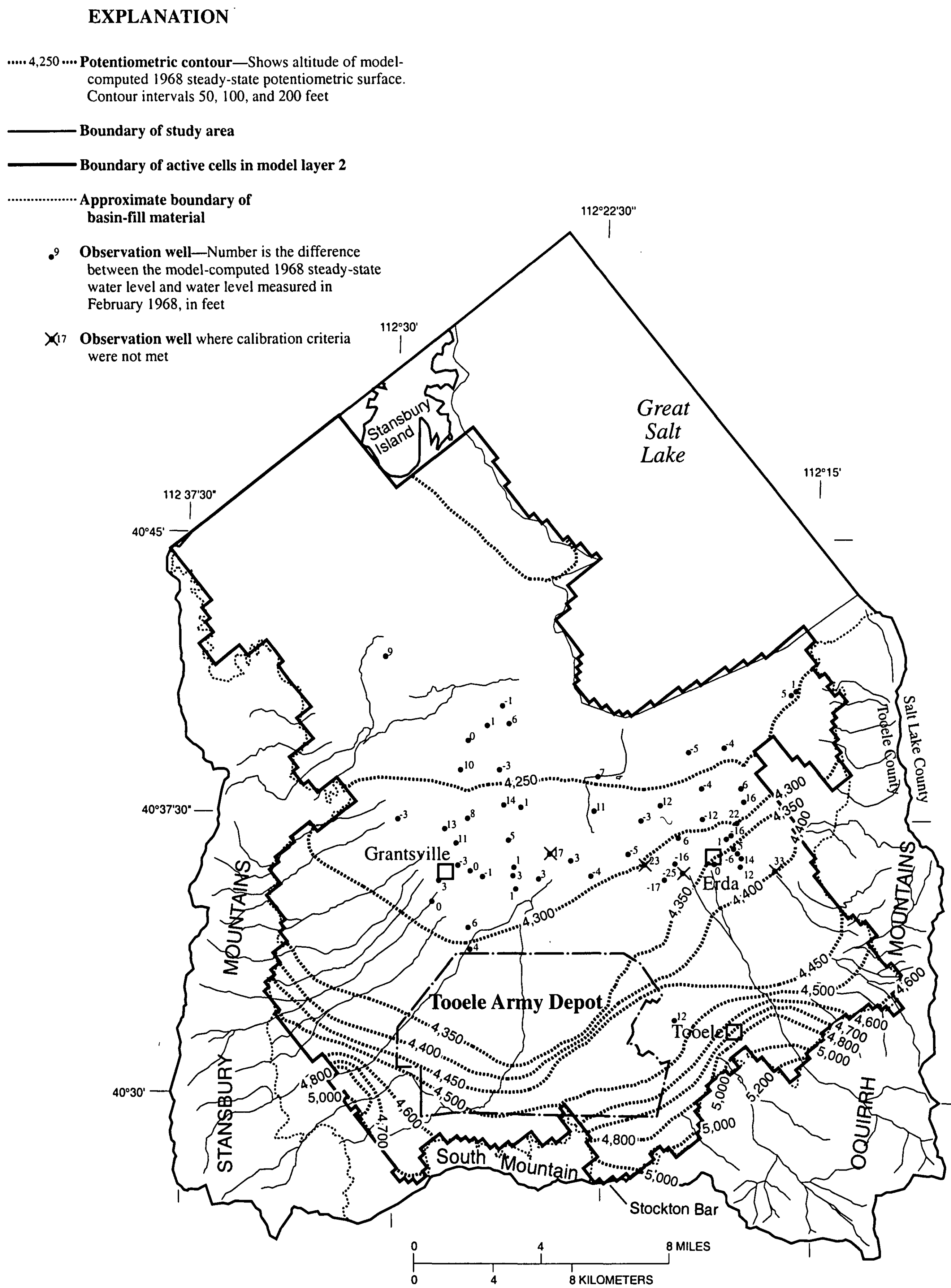


Figure 16. Model-computed potentiometric surface of model layer 2 for the 1968 steady-state simulation of the ground-water flow model of Tooele Valley, Utah, and the difference between model-computed and measured water levels for that simulation period.

Table 5. Model-computed steady-state discharge rates and estimated average annual discharge of ground water at selected springs in Tooele Valley, Utah

Spring	Model-computed discharge in the 1968 steady-state simulation (acre-feet per year)	Estimated average annual discharge (acre-feet per year)
Dunne's Pond Springs	3,700	6,400
Mill Pond Springs	5,600	4,600
Source of Sixmile Creek	2,700	2,800
Source of Fishing Creek	2,600	2,200
Total	14,600	16,000

between total inflow and total outflow results because each budget component was estimated separately and each component has some error. Total flow into the ground-water system computed in steady-state simulation is nearly identical to the amount estimated for the conceptual budget, however, the distribution of recharge is different. During model calibration, recharge due to subsurface inflow from consolidated rock and stream-channel deposits was increased from the rate estimated for the conceptual model. Infiltration of precipitation and seepage of unconsumed irrigation water for the central lower-altitude areas was eliminated to decrease the number of model cells where computed water levels exceeded land-surface altitude. Overall, simulated recharge was increased along the valley margins and decreased in the central parts of the valley. This represents a modification of the conceptualization of recharge distribution in Tooele Valley presented in the "Hydrology of the ground-water system" section of this report.

The largest discrepancy between computed and estimated flows in the valley was in discharge by evapotranspiration. In the conceptual model, discharge by evapotranspiration in the valley is estimated at 23,000 acre-ft/yr. A comparison between conceptual and numerical values is considered valid even though model-computed evapotranspiration is based on parameters that have been slightly modified from those used to determine evapotranspiration conceptually (see section titled "Parameter estimation and model input", "Head-dependent and constant-head dependent boundaries").

Discharge by evapotranspiration as computed by the model is also a function of specified maximum rates of evapotranspiration. Initial values for maximum rates of evapotranspiration were adjusted during calibration to maintain water levels in the discharge area in model layer 1 at a reasonable level relative to land surface and to match the estimated flow budget for the valley (tables 6 and 7). The best results were achieved when maximum rates for phreatophytes were decreased uniformly by 40 percent. Attempts to improve the match between computed and estimated discharge by evapotranspiration by reducing maximum rates even further resulted in computed water levels that exceeded land-surface altitude in a substantial number of cells and worsened the match between computed and measured water levels in the principal aquifer. The maximum rate of discharge by evapotranspiration for cultivated land was not changed during calibration from initial estimates. The final value for maximum rate of evapotranspiration at bare lands or mud flats incorporated in the model was 0.11 ft/yr. The specified extinction depth for evapotranspiration cells was not adjusted during calibration.

During steady-state calibration, values for conductance at head-dependent boundaries that simulate discharge to springs, shallow drains and ditches, and flowing wells were adjusted to match measured water levels and measured or estimated flow rates. Final conductance values for head-dependent drain boundaries that simulate discharge to springs were 260,000 ft^2/d at Dunne's Pond Springs, 90,000 ft^2/d at Mill Pond Springs, 30,000 ft^2/d at the source of Sixmile Creek, and 20,000 ft^2/d at the source of Fishing Creek (values are rounded). Drain-conductance values at cells that simulate discharge to shallow drains and ditches varied from 860 to 8,600 ft^2/d. Final values for drain-conduc-

Table 6. Ground-water budget specified or computed in the steady-state simulation of the ground-water flow model of Tooele Valley, Utah, compared to conceptual budget estimates reported in previous studies or defined during this study

[Data in acre-feet per year; —, no data, budget amounts listed in italics were specified and not computed by the model]

Budget component	Specified or computed in the 1968 steady-state simulation	Conceptual budget estimate
Recharge from		
Consolidated rock and stream-channel deposits	*55,600*	47,700
(Oquirrh Mountain Front)	*(43,400)*	(37,200)
(Stansbury Mountain Front)	*(12,200)*	(10,500)
Subsurface inflow from Rush Valley	*4,600*	5,000
Infiltration of precipitation on the valley floor	*9,700*	12,000
Seepage of unconsumed irrigation water	*5,700*	9,700
Total (rounded)	**75,600**	**74,400**
Discharge		
Wells	23,900	26,000
(Pumped)	[1]*(11,500)*	(13,500)
(Flowing)	(12,400)	(12,500)
Evapotranspiration	30,100	23,000
Springs	14,600	16,000
Drains	5,100	—
Great Salt Lake	2,000	3,000
Total (rounded)	**75,700**	**68,000**

[1]Average annual pumpage during 1964-68.

Table 7. Initial and final maximum evapotranspiration rate for three major land-use categories used during construction and calibration of the ground-water flow model of Tooele Valley, Utah

Land-use category	Initial maximum evapotranspiration rate (feet per year)	Final maximum evapotranspiration rate (feet per year)
Bare ground/mud flat	0.09	0.11
Cultivated land	.38	.38
Phreatophytes:		
Predominantly—		
Pickleweed	2.47	1.48
Greasewood	2.67	1.60
Salt Grass	2.28	1.37

tance at cells that simulate discharge to flowing wells ranged from 180 ft^2/d to 10,100 ft^2/d. Specified drain-cell elevations were not adjusted during calibration.

Steady-state calibration resulted in refined estimates of model parameters that define the hydrologic properties of the aquifer system. During calibration the specified equivalent hydraulic conductivity and transmissivity of the principal aquifer were adjusted by varying hydraulic conductivity in equation 5 for permeable coarse-grained basin-fill material (K_{CRSE}), fine-grained basin-fill material (K_{FINE}), and consolidated rock (K_{CON}). Defined zones and final values for K_{CRSE} used in the calibrated model are shown in figure 17. During calibration, the two sediment-type zones (zones 1 and 2 in fig. 5) that were used to define the initial distribution of K_{CRSE} in the model were further divided to create a total of six zones. The delineation of additional zones was done to create a more detailed representation of the spatial variation in sediment type observed in data from drillers' logs. Zones 1 to 3 represent the change in sediment type from the interbedded fine-grained sands and clays that make up the principal aquifer in the northern part of the valley near the shore of the lake (zone 1 in fig. 17) to the coarse-grained sand and gravel deposits that are present in the central part of the valley (zone 3 in fig. 17). Zones 4 and 5 (fig. 17) represent areas of the principal aquifer near the margins of the valley where basin-fill material consists of coarse-grained sediments that are generally poorly sorted and may contain semiconsolidated sediments of Tertiary age. Zone 6 was defined during calibration to improve the match between model-computed and measured water levels in an area of high horizontal hydraulic gradient in the eastern part of the valley. Zone 6 is associated with areas surrounding the bedrock high near Erda, which is probably related to Tertiary and (or) Quaternary-age faults. It was assumed that the large hydraulic gradient in that region could be the result of a zone of fine-grained sediments near the shallow bedrock or displaced sediments near fault zones.

The hydraulic conductivity of fine-grained silt/clay layers (K_{FINE}) within model cells was assigned a final value of 1 ft/d throughout the model. The final value for hydraulic conductivity of consolidated rock (K_{CON}) represented in some active model cells was 1 ft/d for consolidated rock extending into the valley from the Stansbury Mountains.

The ratio of permeable coarse-grained sediment to total thickness of the principal aquifer that was estimated from drillers' logs (fig. 6) also was considered a calibration variable. Percentages defining the ratio were adjusted, particularly in model areas where few data from wells logs were available, which changes the simulated thickness of permeable coarse-grained material in model cells (b_{CRSE} and b_{FINE} in eq. 5). The final percentages of permeable coarse-grained material (sand/gravel-bearing intervals) in the principal aquifer incorporated in the model are shown in figure 18.

Hydraulic-conductivity values for the basin-fill material and consolidated rock in the area of Tooele Army Depot (fig. 17) were derived directly from the results of calibration of the 1994 U.S. Army Corps of Engineers ground-water flow model (table 1) and were not adjusted during calibration of the regional model.

Equivalent hydraulic-conductivity (K' in eq. 5) and transmissivity values were calculated for model layers on the basis of specified values of K_{CRSE}, K_{FINE}, and K_{CON} for individual model runs during calibration. Final ranges of total transmissivity, which is the aggregation of (1) saturated thickness and hydraulic conductivity of model layer 2 and (2) transmissivity for model layers 3 through 5, are shown in figure 19. The final distribution of hydraulic-conductivity values for model layer 1 is shown in figure 20.

The final distribution of vertical hydraulic-conductivity values (K_v) incorporated in the vertical leakance among model layers is shown in figures 21 and 22. The vertical leakance values incorporated for Tooele Army Depot (figs. 21 and 22) were derived directly from the results of calibration of the 1994 U.S. Army Corps of Engineers ground-water flow model. These values were based on K_v values for sediments in the area that ranged from 0.2 ft/d to 7 ft/d (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994, p. 54-55).

Transient-State Calibration

Method

The transient-state simulation represents estimated hydrologic conditions for 1969-94. The purposes of the simulation were to (1) evaluate the relation between annual water-level fluctuations in the valley and variations in annual ground-water recharge caused by changes in precipitation in and around the valley and (2) estimate storage properties of the aquifer system.

The transient-state simulation period, from January 1969 to December 1994, was divided into 26 stress periods of 1 year in length. During a stress period, external stresses on the simulated system, representing average recharge or discharge for a given year, are held constant. Each stress period was divided into three time steps. The length of the first time step of each stress period was 77 days (rounded) and was increased with advancing time by a time-step multiplier of 1.5. During transient-state calibration, the initial time-step length was reduced to ensure that the accuracy of the simulation was not affected by truncation error resulting from an inappropriately large initial time-step size. The results of simulations using a shorter initial time step did not indicate a significant change in model-computed water levels or flow rates.

The results of the 1968 steady-state simulation were used as the initial conditions for the transient-state simulation. Annual fluctuations in recharge and withdrawal from pumping wells were simulated using yearly stress periods. Calibration involved adjusting calibration variables and comparing model-computed water levels and water-level changes with measured water levels and water-level changes at observation

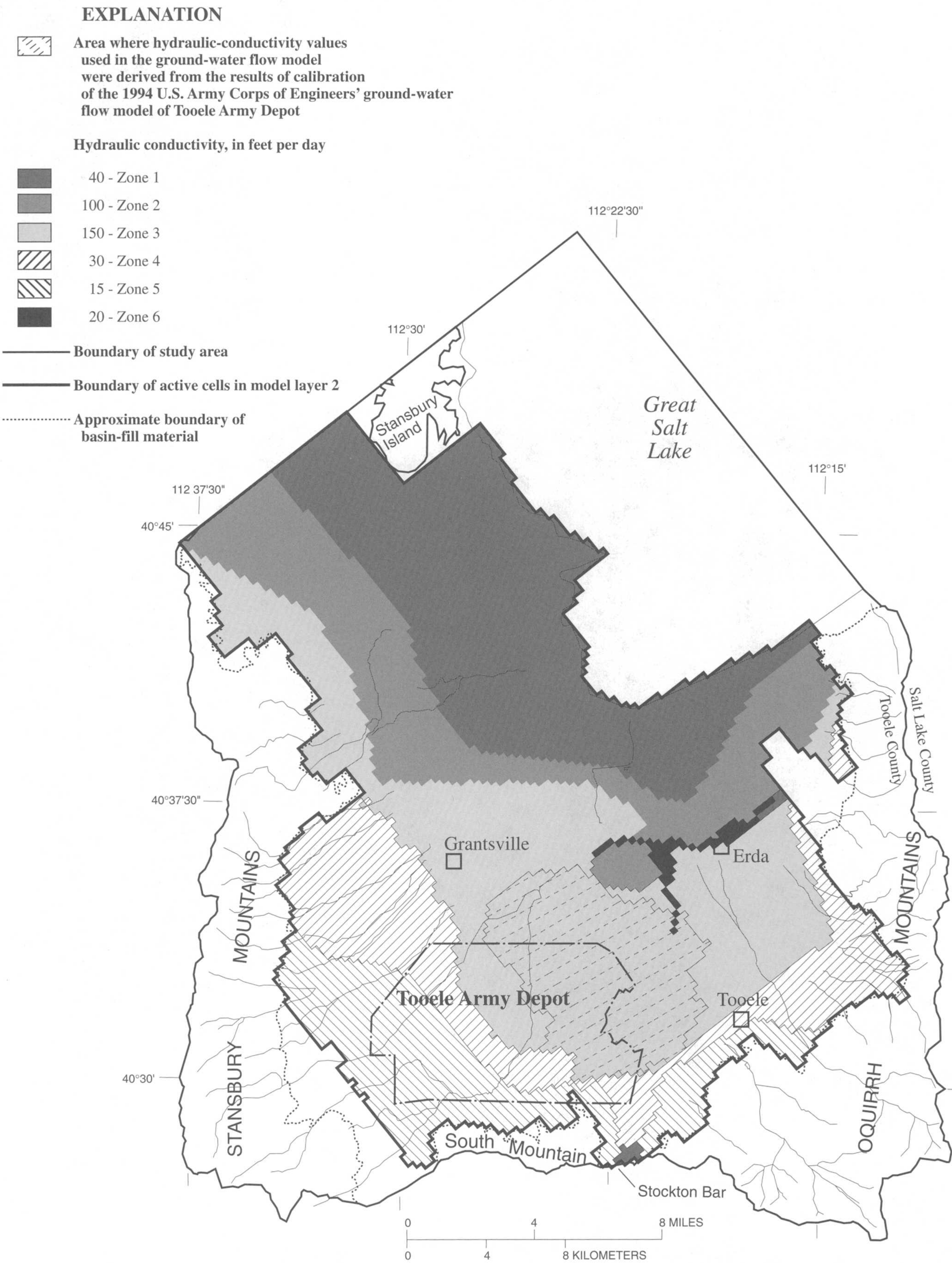


Figure 17. Zones and values of hydraulic conductivity of coarse-grained basin-fill material (K_{CRSE}) used to define final equivalent hydraulic-conductivity and transmissivity values in model layers that represent the principal aquifer in the ground-water flow model of Tooele Valley, Utah.

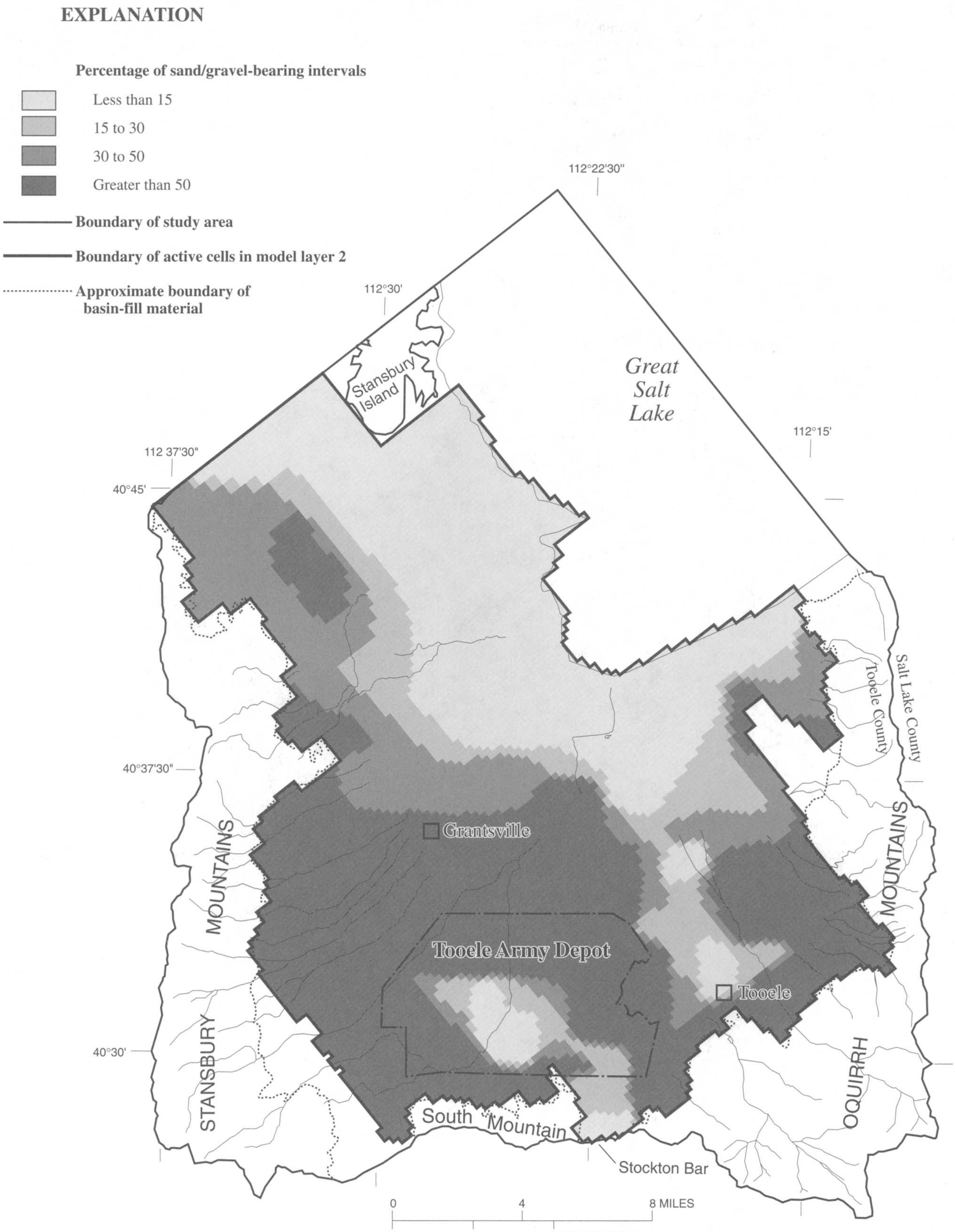


Figure 18. Final percentage of the principal aquifer that consists of sand/gravel-bearing intervals in the ground-water flow model of Tooele Valley, Utah.

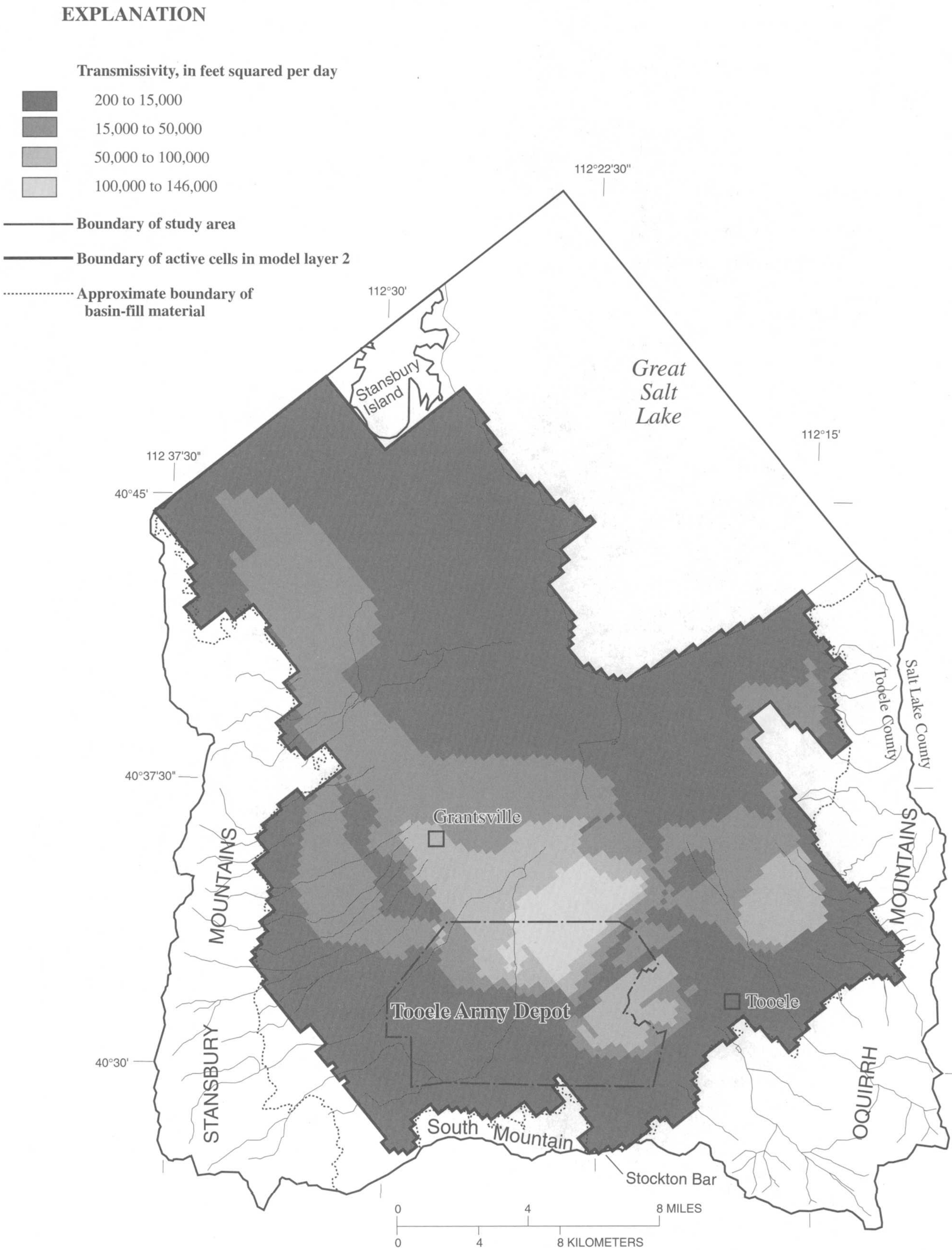


Figure 19. Final distribution of transmissivity for the principal aquifer simulated in model layers 2 through 5 of the ground-water flow model of Tooele Valley, Utah.

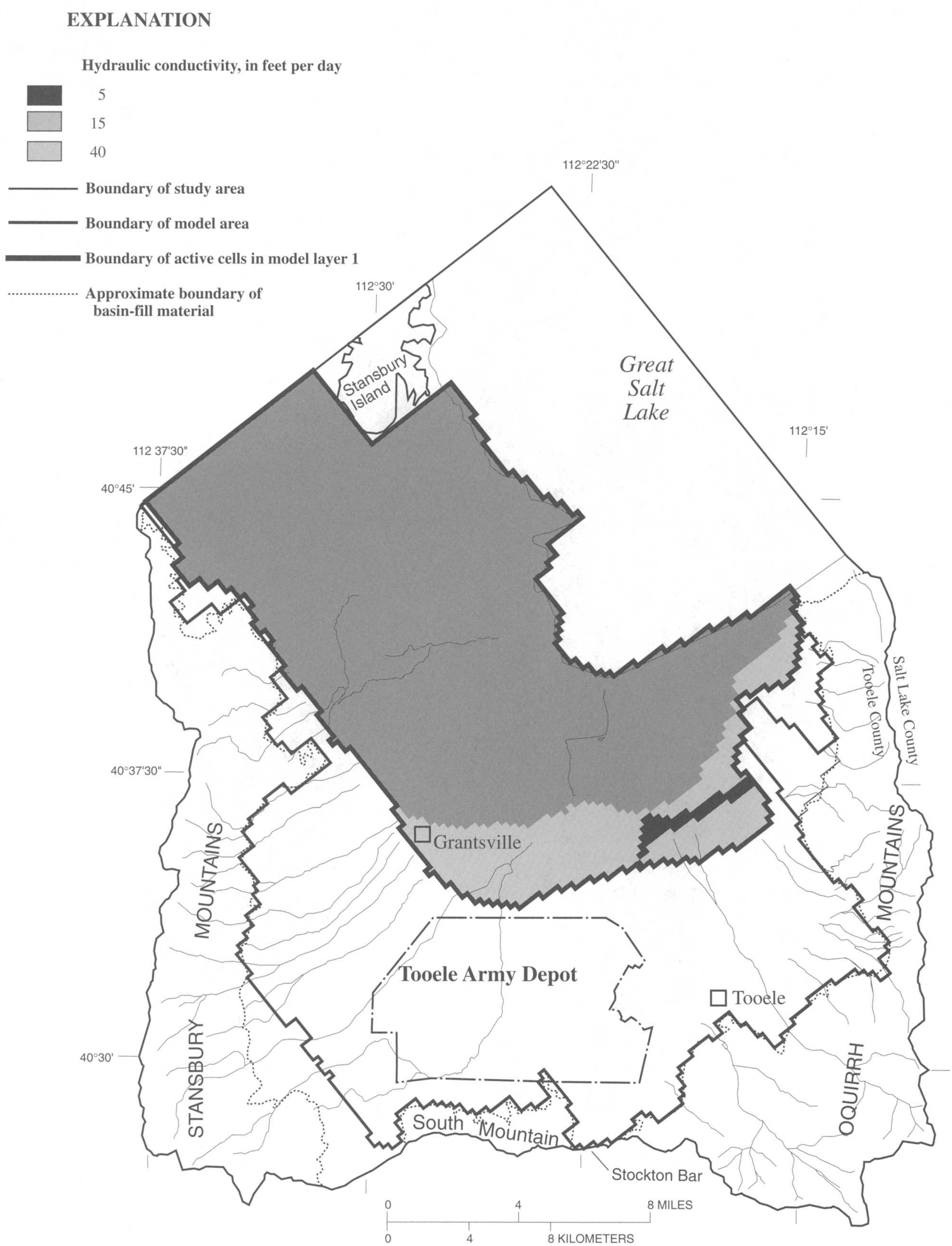


Figure 20. Final distribution of hydraulic-conductivity values for model layer 1 of the ground-water flow model of Tooele Valley, Utah.

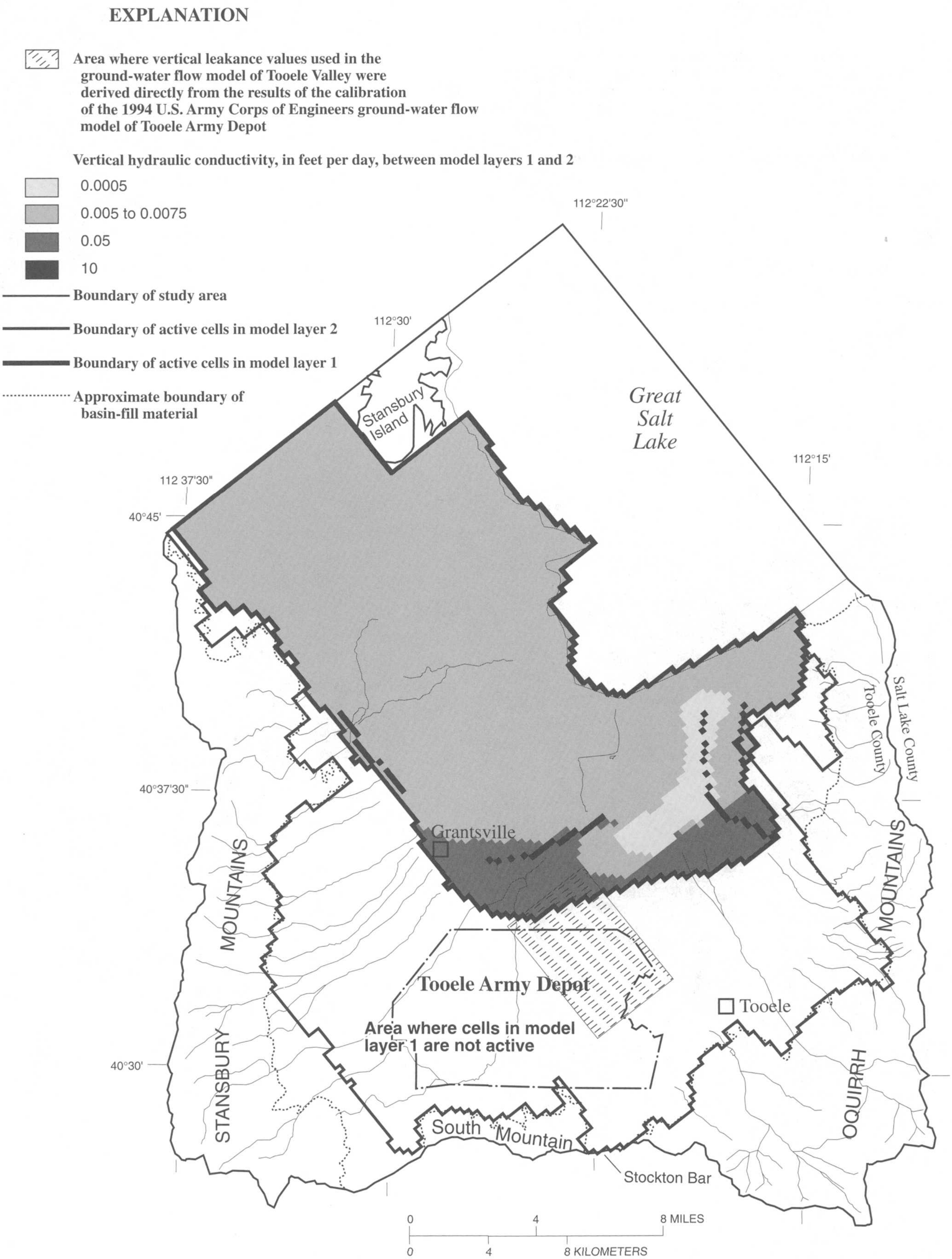


Figure 21. Final distribution of vertical hydraulic-conductivity values for model layer1 incorporated in the vertical leakance between layers 1 and 2 of the ground-water flow model of Tooele Valley, Utah.

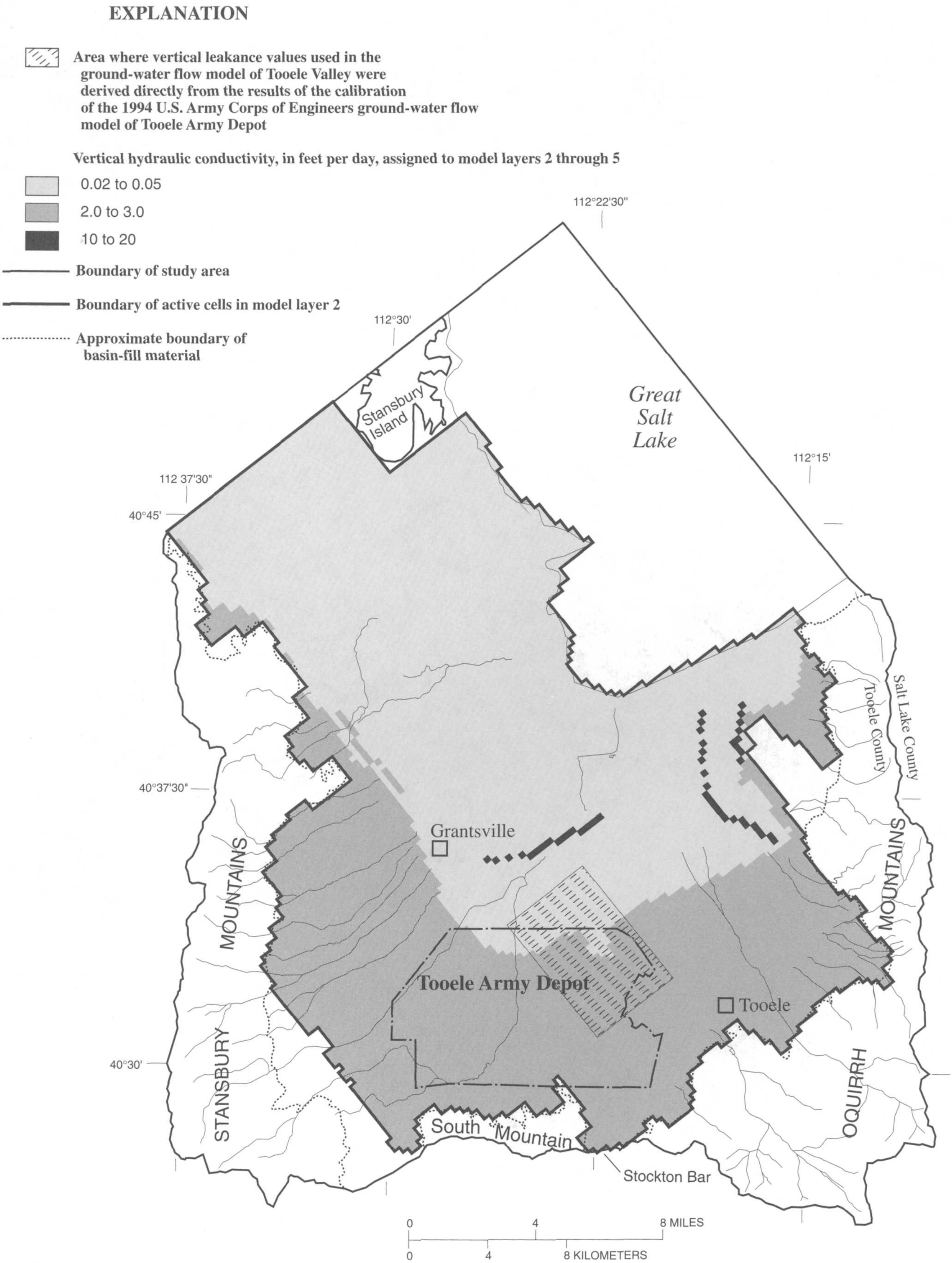


Figure 22. Final distribution of vertical hydraulic-conductivity values for model layers 2 through 5 incorporated in the vertical leakance between layers 2 through 5 of the ground-water flow model of Tooele Valley, Utah.

wells in the principal aquifer. Model parameters considered to be calibration variables during transient-state calibration were (1) storage coefficient of confined zones of the ground-water system, (2) specific yield of unconfined zones of the ground-water system, (3) horizontal hydraulic conductivity and transmissivity of the principal aquifer, and (4) variations from steady-state values of simulated annual recharge to the ground-water flow system simulated at the basin-fill material/consolidated-rock boundary, as precipitation on the valley floor, and as recharge from unconsumed irrigation water.

To evaluate the accuracy of the transient-state simulation, for each individual model run (1) model-computed water-level changes from one stress period to the next were compared with measured water-level fluctuations at observation wells, (2) water levels computed by the model for the end of stress period 24 representing conditions during 1992 were compared to measured 1992 water levels in the south end of the valley near Tooele Army Depot, and (3) model-computed ground-water discharge to valley springs was compared with available estimates of annual discharge at the springs. Comparison of computed water levels for stress period 24 to the Tooele Army Depot 1992 data set allowed for the evaluation of the accuracy of the model in the south end of the valley where water-level data for the steady-state simulation period were sparse.

Results of Calibration

Measured and model-computed water-level changes at selected observation wells in the valley are shown in figure 23. Measured water-level data displayed in the hydrographs indicate moderate changes in water levels at wells in the confined zone of the principal aquifer near Great Salt Lake (fig. 23a and b). The magnitude of fluctuations in wells is generally greater in the vicinity of the primary recharge areas near the margins of the valley, where observed water-level fluctuations of 15 to 20 ft during a multi-year period are not uncommon (fig. 23d). The results of transient-state calibration indicate that the generally observed rising and declining trends in water levels at observation wells are reproduced in many areas of the model (fig. 23). The model, however, generally does not simulate the large and abrupt changes in water levels from year to year that are observed at some wells; that is, the changes in water-level altitude simulated in the model generally are more gradual than those observed in the field. Also, the model did not accurately simulate large rises in water levels in several areas during periods of greater-than-normal precipitation during 1982-84. These discrepancies may indicate that the horizontal and vertical distribution of subsurface recharge and the method of varying recharge annually in the model do not accurately represent actual ground-water flow.

Computed water-level changes at selected wells in the northern part of the valley (fig. 23a, b, and c) match measured changes reasonably well. Simulated and measured water-level changes at the well in cell 59i, 110j, 3k (fig. 23a) in the northeastern part of the valley indicate the effects of fluctuations in recharge at the adjacent mountain front and generally match the observed rises and declines that have resulted from periods of greater-than-average precipitation during 1972-76 and 1982-84. Although water levels are simulated as rising at other observation sites in the northern part of the valley, simulated magnitudes of rises during 1982-84 do not generally match observed magnitudes (fig. 23d). Observed declining and rising water-level trends in wells in the southeastern part of the valley (fig. 23d, cells 84i, 70j, 3k and 94i, 84j, 3k) generally are matched by simulated trends. The match between observed and simulated water-level fluctuations is not good, however, at the well in cell 76i, 51j, 3k (fig. 23d) near shallow consolidated rock at Tooele Army Depot.

Discrepancies between model-computed and measured water-level changes in the west-central and south-central parts of the valley indicate that the distribution of recharge and the method of varying recharge in these areas as a function of the annual supply of recharge water from precipitation may not accurately represent the actual system. As stated previously, rates of recharge to the Tooele Valley ground-water system at the surfaces of the primary recharge area have not been measured directly in the field. Also, how recharge from primary sources varies throughout time is not well understood. It is also possible that model parameters that define aquifer properties in these areas are not accurate. Simulated water levels, and thus annual water-level changes, are computed as a function of aquifer transmissivities and storage. Although the calibrated parameter values that define aquifer properties in the model are within prescribed ranges defined from field tests and other analyses, the inability of the model to accurately simulate observed changes in water levels indicates that these parameter values may be inaccurate and that more field data are required to improve current estimates. This issue is discussed further in the "Limitations of the model" section of this report.

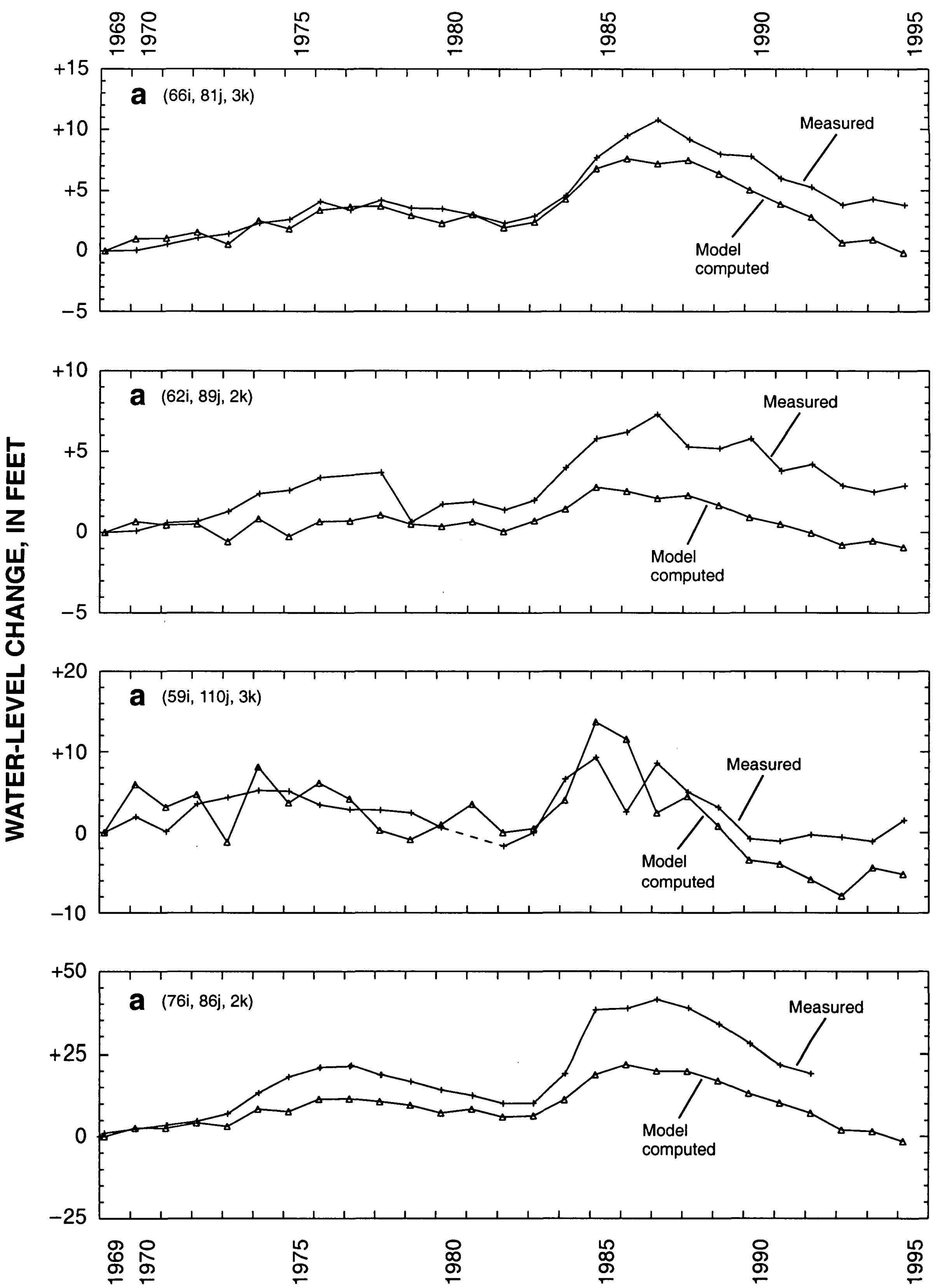


Figure 23. Model-computed and measured water-level changes at observation wells in selected cells in the (a) northeastern, (b) north-central, (c) northwestern, and (d) southern parts of Tooele Valley, Utah, 1969–94. Numbers in parentheses represent row, column, and layer of model cell.

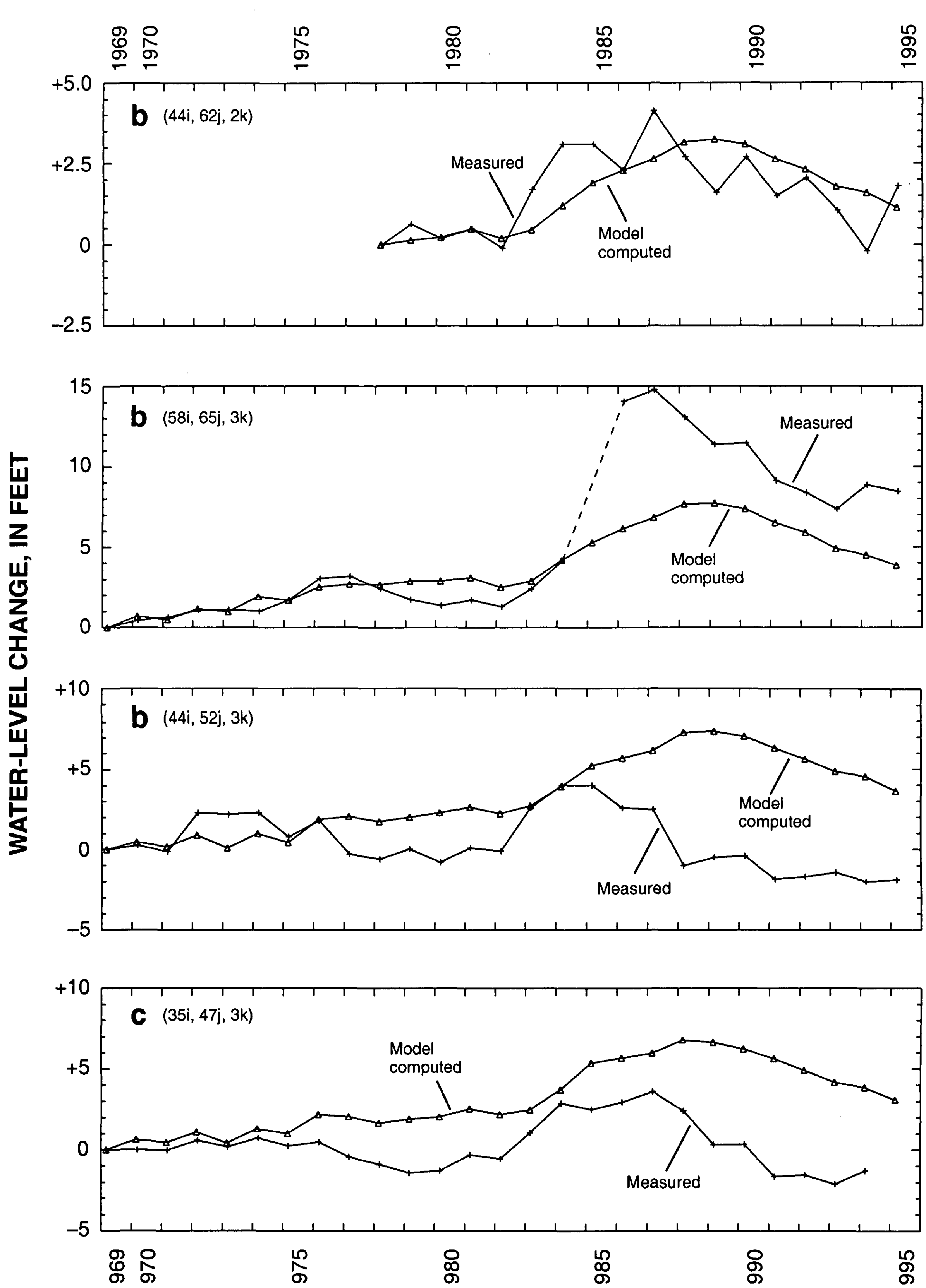


Figure 23. Model-computed and measured water-level changes at observation wells in selected cells in the (a) northeastern, (b) north-central, (c) northwestern, and (d) southern parts of Tooele Valley, Utah, 1969–94—Continued. Numbers in parentheses represent row, column, and layer of model cell

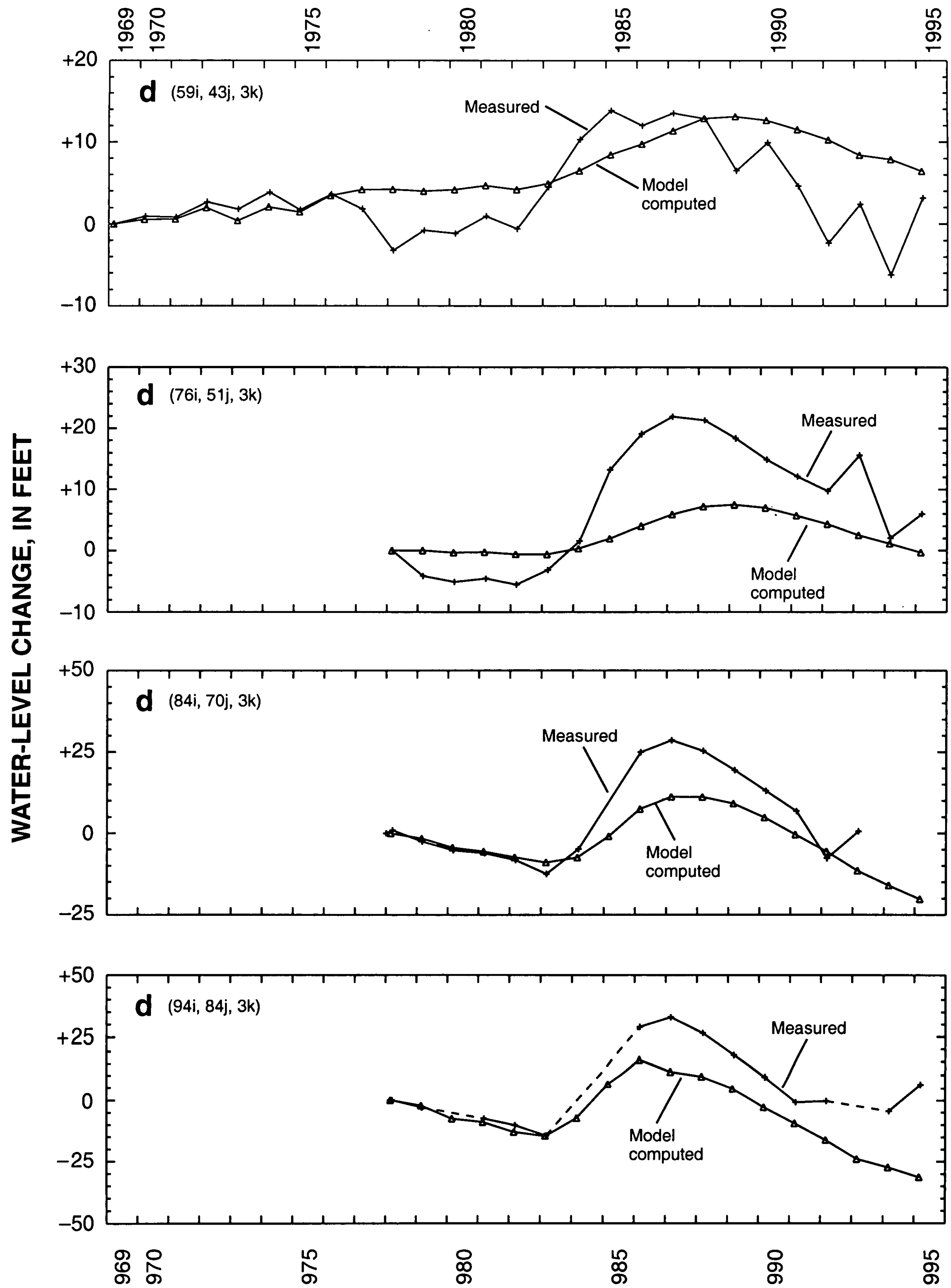


Figure 23. Model-computed and measured water-level changes at observation wells in selected cells in the (a) northeastern, (b) north-central, (c) northwestern, and (d) southern parts of Tooele Valley, Utah, 1969–94—Continued. Numbers in parentheses represent row, column, and layer of model cell

Residuals from the comparison of computed water levels in stress period 24 and 1992 measured water levels near Tooele Army Depot are shown in figure 24. These observation sites are the same sites used by the Army Corps of Engineers Hydrologic Engineering Center (1994) in the calibration of their ground-water flow model of part of the depot. Both the horizontal and vertical locations of observation sites were factored into the computation of residuals. The residuals indicate a reasonable match between model-computed and measured water levels on the depot. The largest residuals in the depot area occur in an area of high horizontal hydraulic gradient related to the shallow consolidated rock underlying the depot. Some inaccuracies are indicated in the northern end of the depot (fig. 24). Measured water levels in that area indicate a substantial upward vertical hydraulic gradient that is not accurately represented in the model.

For each stress period estimates of annual recharge to the ground-water system were calculated (1) at the basin-fill material/consolidated-rock boundary, (2) as precipitation on the valley floor, and (3) as seepage of unconsumed irrigation water from steady-state values using equations 6 to 9. During calibration, the effects of annual precipitation fluctuations in the valley and surrounding mountains on simulated water-level changes were adjusted by varying the coefficient *C* in those equations within defined ranges. Model-computed water-level changes in the principal aquifer near the margins of the valley were substantially affected by varying the coefficient *C*. The best match between model-computed and measured water-level changes was achieved using a value of 1 for the coefficient *C* in equations 6 and 8, which were used to compute recharge from consolidated rock and precipitation on the valley floor, and a value of 3 for coefficient *C* in equations 7 and 9, which were used to compute recharge from stream-channel deposits at the mouths of canyons and from irrigated land. Annual rates of recharge simulated at specified-flux boundaries from consolidated rock and stream-channel deposits at the mountain front, from unconsumed irrigation water from fields and lawns/gardens, and from precipitation on the valley floor are shown in figure 25.

The final value of specific yield assigned to model layer 1 during transient calibration of the ground-water flow model is 0.15. The final distribution of specific-yield values for unconfined zones in layer 2 (zones that correspond to the area where the principal aquifer is considered unconfined) are shown in figure 26. The final value of specific storage assigned to the confined zones in layer 2 and all active zones of model layers 3, 4, and 5 is 2.5×10^{-6} ft^{-1}. This value of specific storage produces a final range for storage coefficient for the principal aquifer of 3.8×10^{-4} to 2.5×10^{-3}.

SENSITIVITY ANALYSIS

Sensitivity analysis is the evaluation of the effects that model parameter values have on model results. The purpose of sensitivity analysis is to determine which model parameters have the greatest influence on model results. Model sensitivity gives an indication of how much an incorrectly estimated model parameter may effect the simulation. Model sensitivity was not tested against all model parameters. Parameters tested in this analysis were chosen either because the processes they represent are not well understood or the parameter values are at the limit of their probable range.

Sensitivity of the model to changes in the following model parameters were analyzed: (1) vertical hydraulic conductivity of shallow sediments represented in model layer 1 and incorporated in the model as vertical leakance between model layers 1 and 2, (2) horizontal hydraulic conductivity of the principal aquifer, and (3) subsurface inflow from consolidated rock. The analysis was made by independently varying each parameter, or the model input derived from the parameter, in the 1968 steady-state simulation. Sensitivity is measured by the degree of change that is initiated in model-computed water levels as a result of the change in the model parameter (table 8).

A 50-percent reduction of the vertical leakance (*VL*) between model layers 1 and 2 caused simulated water levels to increase slightly. Most head-dependent discharge boundaries are located in model layer 1, and to maintain flux at these boundaries, the model-computed hydraulic gradient between model layer 1 and the layers simulating the principal aquifer (model layers 2 to 5) increased somewhat. Statistical differences are comparable with the calibrated solution.

Decreasing horizontal hydraulic conductivity of the principal aquifer caused an average simulated water-level rise of about 6 ft. Simulated water levels are higher at almost all model cells. Opposite and larger effects were noted with increased conductivity. Raising conductivity values more than 5 percent caused de-watering at individual cells and the model was not able to reach a numerical solution.

Model results are also sensitive to changes in specified recharge from consolidated rock. Decreasing

EXPLANATION

—*4,300*— **Potentiometric contour**—Shows altitude of model-computed 1992 steady-state potentiometric surface. Contour interval 50 feet

10/3 • **Observation well or nested observation wells**—Number(s) is the difference between the model-computed water levels for transient stress period 24 and water level measured in 1992. Two numbers separated by a slash indicate two wells in the same location that monitor water levels at different depths below land surface. The first number represents the difference in model layer 2, the second number represents the difference in layer 3 or 4

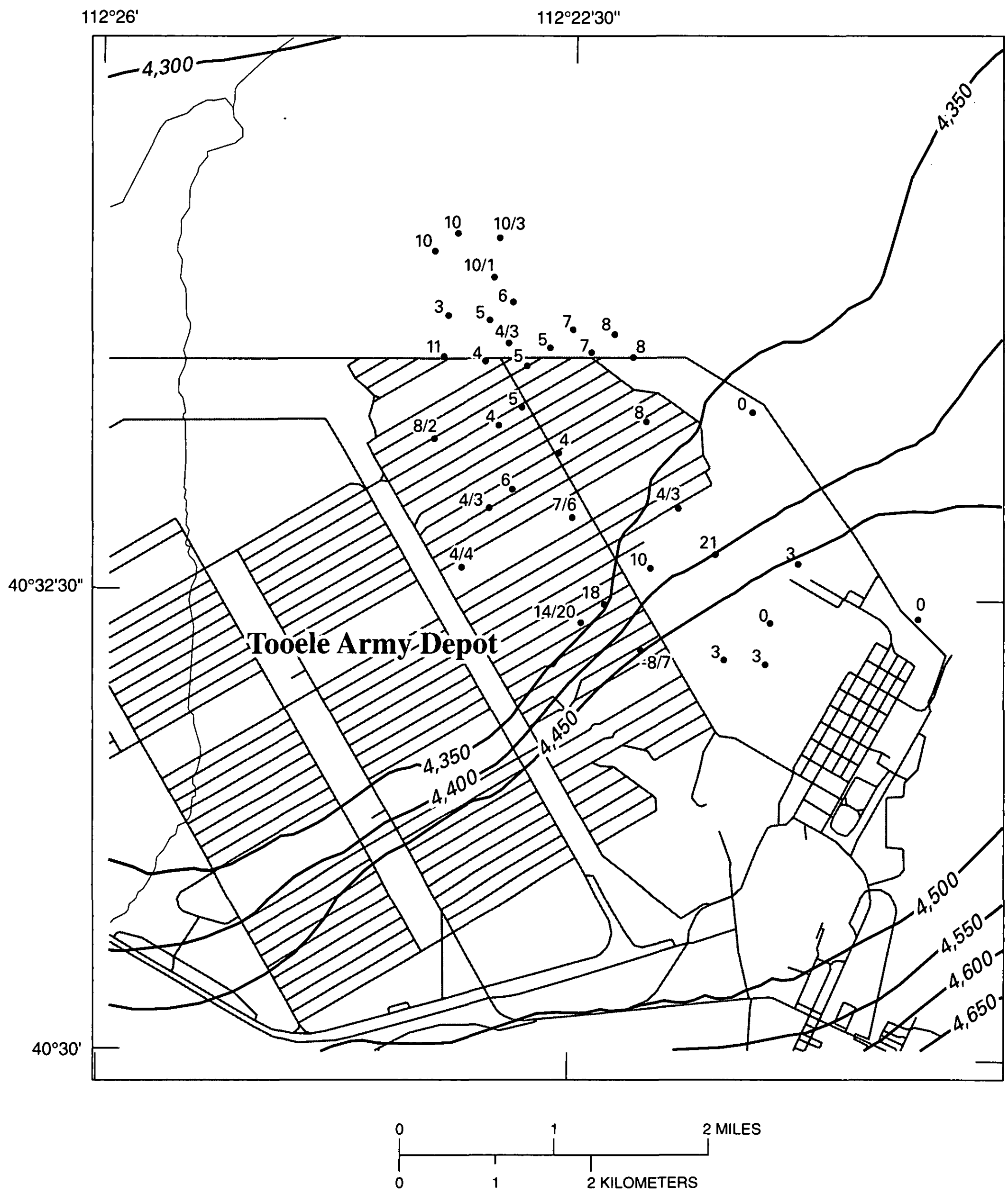


Figure 24. Model-computed potentiometric surface of model layer 2 for stress period 24 (1992) of the transient-state simulation of the ground-water flow model of Tooele Valley, Utah, and the difference between model-computed and measured water levels for that stress period at Tooele Army Depot, Utah.

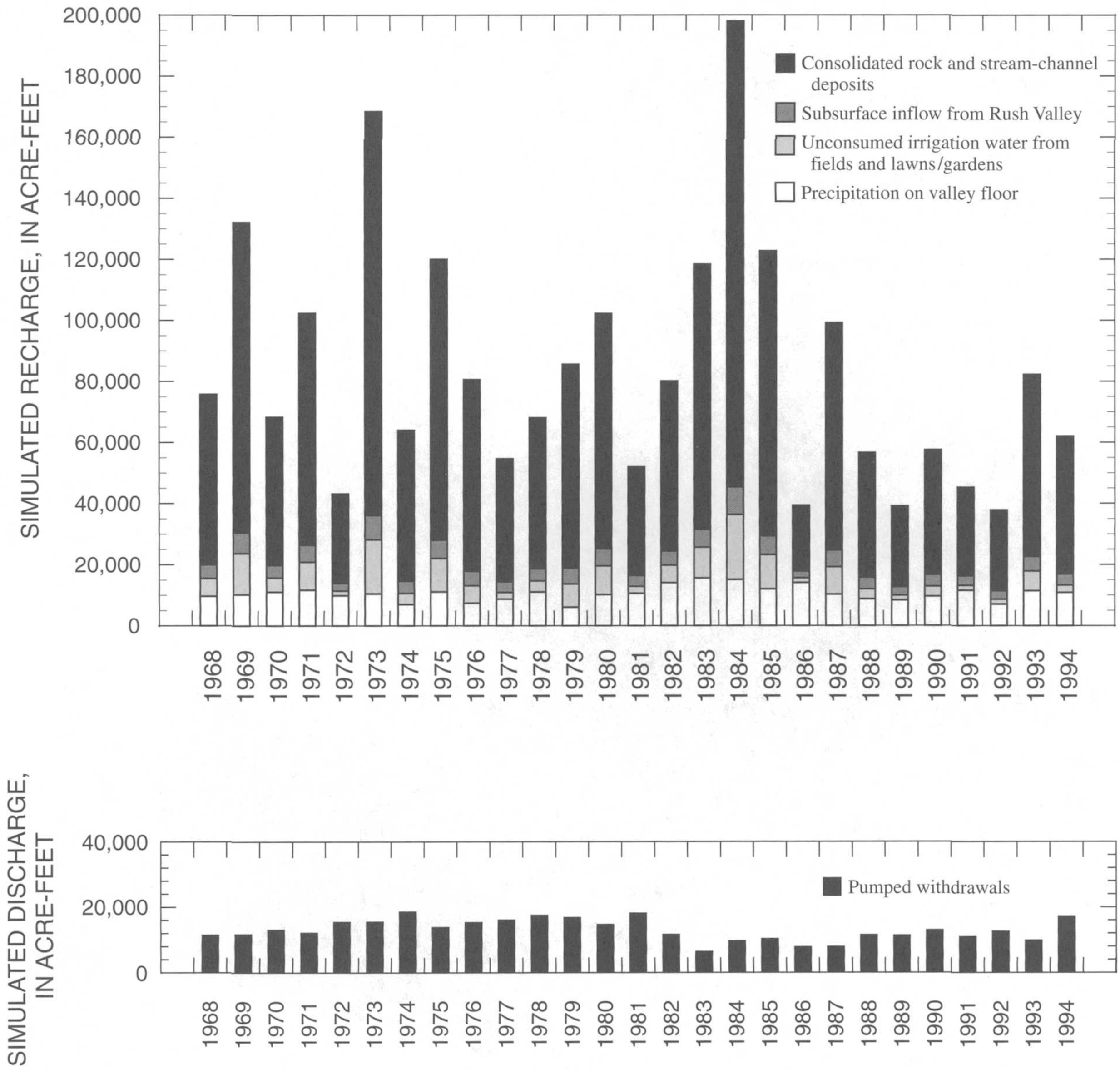


Figure 25. Simulated recharge and discharge at the specified-flux boundaries for the 1968 steady-state simulation and the 1969–94 transient-state simulation of the ground-water flow model of Tooele Valley, Utah.

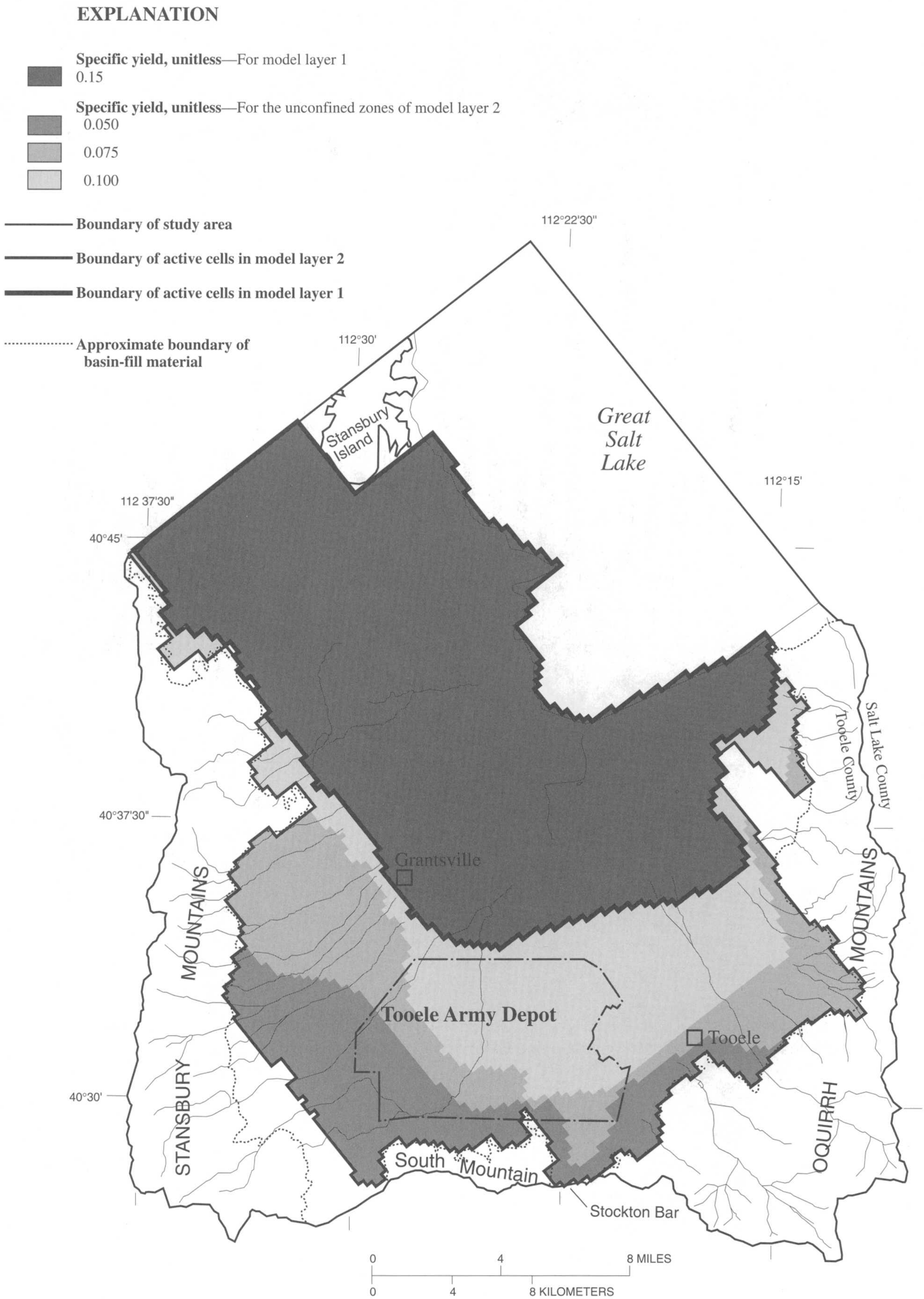


Figure 26. Final distribution of specific-yield values for model layer 1 and for unconfined zones in model layer 2 of the ground-water flow model of Tooele Valley, Utah.

Table 8. Statistical difference between model-computed and measured water levels in the 1968 steady-state simulation and sensitivity-analysis simulations using the ground-water flow model of Tooele Valley, Utah

[Values calculated as model-computed minus measured water level; VL, vertical leakance; HC, hydraulic conductivity; —, not applicable]

	Statistical difference between model-computed and measured water levels in the 1968 steady-state simulation	Statistical difference between model-computed and measured water levels in sensitivity-analysis simulation					
		VL between model layers 1 to 2		Horizontal HC in model layers 2 to 5		Recharge from consolidated rock	
		x0.5	x1.5	x0.75	x1.05	x0.85	x1.25
56 observation sites							
Mean (in feet)	2.4	5.0	1.3	3.3	2.0	-4.9	6.5
Standard error (in feet) (mean of absolute value of differences)	8.2	9.4	7.9	9.1	8.0	9.2	10.2
Standard deviation (in feet)	11.0	12.2	10.7	13.2	10.8	13.0	13.9
Maximum difference lower than measured (in feet)	-26.0	-25.1	-26.9	-24.8	-27.2	-40.1	-21.3
Maximum difference higher than measured (in feet)	33.2	32.9	31.1	50.4	28.5	11.5	45.0
26,987 active model cells							
Percentage of total model grid affected	—	93	88	94	57	84	82
Number of model cells where water levels changed	—	25,113	23,670	25,313	15,396	22,645	22,229
Total change in model-computed water level (absolute value in feet)	—	78,052	34,327	221,713	37,312	224,644	168,944
Average water-level change (in feet)	—	2.9	-1.3	6.2	-1.7	-9.9	7.6

this recharge by 15 percent caused average simulated water levels to decline almost 10 ft. The model will not reach a solution if decreases of greater than 15 percent are simulated. The model is not as sensitive to increasing recharge amounts. This is likely due to head-dependent boundaries where increasing discharge will partially compensate for the additional recharge. For both decreased and increased recharge, statistical differences between model-computed and measured water levels are larger than those for the 1968 steady-state simulation.

LIMITATIONS OF THE MODEL

The hydrologic system in Tooele Valley is complex and cannot be defined completely with available data. The numerical model documented in this report is based on mathematical representations of ground-water flow and on a simplified set of assumptions about the hydrologic system. As a result, the calibrated model has limitations that need to be considered when evaluating simulation results.

This model was constructed and calibrated to simulate the regional scale ground-water flow system of Tooele Valley. Model parameters, including parameters that define aquifer properties, were estimated on a regional scale. This spatial averaging and simplification results in smoothing of local anomalies. These aspects must be considered when evaluating model response for local areas and subregions of Tooele Valley. They may not be as pertinent for Tooele Army Depot, where model parameters are based on the ground-water flow model for the depot (U.S. Army Corps of Engineers Hydrologic Engineering Center, 1994), although they must still be considered. Limitations in time also need to be considered when evaluating simulation results. The transient-state simulation period was discretized into yearly stress periods, and seasonal or shorter time interval changes in water levels and flow at head-dependent boundaries were not simulated. Withdrawals from wells and recharge along the mountain front, which may change substantially within a given year, were simulated using yearly stress periods. If the model were used to simulate changes in hydrologic conditions for time periods shorter than a year, seasonal or monthly changes in recharge and discharge might need to be incorporated into the model and it might be necessary to recalibrate the model.

Few field data were available with which to determine initial estimates for vertical hydraulic conductivity (K_v) of valley sediments, including the shallow unconfined aquifer and the shallow confining layer represented by model layer 1. Vertical gradients and flows simulated in the model are controlled, in part, by the vertical hydraulic conductivity incorporated in model input. The shallow confining layer is represented in the model as a continuous layer and K_v values representative of basin-fill material consisting of clay and silt layers were incorporated in active cells in model layer 1. The representation of the confining layer as a continuous clay/silt layer may not be accurate in all areas of the valley. Data from well logs indicate that the shallow zone of fine-grained sediments is made up of interfingering and overlapping layers and lenses of clay, silt, and sand and that the continuity of the layer may vary greatly. The fact that the heterogeneity of the shallow confining layer of the aquifer system is not represented in the model, and the uncertainty of the final estimates of K_v, should be noted when evaluating model results.

Neither water-level measurements in the consolidated rock that surrounds the valley nor data needed to define the hydrologic connection between the principal aquifer and the surrounding consolidated rock were available. The simulation of subsurface inflow from consolidated rock to the principal aquifer was therefore simplified in the model by using specified-flux boundaries. Simulated recharge from consolidated rock at these boundaries does not change during problem solution in the steady-state simulation and is not affected by model-computed water-level fluctuations during the transient-state simulation. In the physical system, however, subsurface inflow from consolidated rock is head dependent, being controlled by the hydraulic gradient existing across the boundary between the principal aquifer and the consolidated rock. Large declines or rises in water level near the margins of the valley affect inflow at the basin-fill material/consolidated-rock boundary at the mountain front. Because a specified-flux boundary is used to represent inflow from consolidated rock, effects on inflow resulting from water-level fluctuations in the principal aquifer are not simulated.

The match between model-computed and measured water-level fluctuations for the transient-state simulation period was poor in some parts of the valley. The inability of the model to accurately duplicate measured water-level changes may indicate that conceptualization of subsurface inflow from consolidated rock may not accurately represent the actual ground-water flow system. The discrepancies also may be the result of inaccurate estimates of hydrologic properties, particularly for areas near the mountain front where recharge is simulated. Although these estimates were based on available field data and the results of previous analyses, probable ranges of values were large in some cases and different combinations of model input defining recharge and aquifer properties in these areas may improve results. The largest discrepancies between simulated and observed water-level change were for periods of extremely high precipitation and spring runoff. In general, simulated responses in future predictive model runs using the model that incorporates periods of greater-than-average precipitation may be inaccurate in some areas.

Although discrepancies were present between simulated and measured conditions in Tooele Valley, the overall accuracy of the model, particularly in the steady-state simulations, is considered to be reasonably good on the basis of (1) the match between model-computed and measured water levels, and (2) the match between model-computed and independently estimated budget components, including flow to valley springs. The model incorporates the current conceptual model of the flow system as defined from available data and field observations, and analyses of ground-water flow using this model and future simulations to determine the effects of regional changes in recharge and discharge to the ground-water flow system are expected to produce useful results. The accuracy of the model may be improved, however, if estimates of system parameters can be refined on the basis of new field data. Information describing the following aspects of the ground-water flow system would be the most useful in improving future simulations using the model: (1) the physical properties, including transmissivity of the principal aquifer near the margins of the valley, particularly in the southwestern part of the valley, (2) the location and physical properties of shallow consolidated rock in the interior of the valley and information on how these deposits affect flow in the principal aquifer, and (3) the quantity and distribution of recharge that enters the valley as subsurface inflow.

SUMMARY

In September 1995, the U.S. Geological Survey, in cooperation with Tooele County, the U.S. Department of the Army, the Utah Department of Natural Resources, Division of Water Rights, Tooele City, and Grantsville City, began a study of ground-water flow in Tooele Valley. The principal objectives of this study were to (1) improve understanding of the regional ground-water flow system in Tooele Valley, and (2) provide information on the effects of regional ground-water flow processes as they relate to subregional areas of Tooele Valley. This report describes the current understanding of the regional ground-water flow system in Tooele Valley and the simulation of that system using a three-dimensional, finite-difference, numerical model. The numerical model was used to test and refine the conceptual understanding of the system.

The ground-water flow system of Tooele Valley is thought to be contained primarily in the Quaternary-age basin-fill material. Few wells in Tooele Valley yield water from the Tertiary-age basin-fill material, which is generally assumed to be less permeable than the overlying Quaternary-age material. Thickness of the Quaternary-age basin-fill material is not well known but is estimated to be about 1,000 feet. The ground-water flow system is conceptualized as two aquifers separated by a shallow confining layer. The first aquifer, referred to as "the shallow unconfined aquifer," is present only in the northern and central parts of the valley. The shallow unconfined aquifer is underlain by a shallow confining layer. The second and larger aquifer, referred to as "the principal aquifer," underlies both the shallow confining layer and the remainder of Tooele Valley. Where the principal aquifer underlies the shallow confining layer (northern and central parts of the valley), the principal aquifer is considered to be confined. In the remainder of the valley (southern part and along the valley margins), the principal aquifer is considered to be unconfined.

Recharge to the basin-fill ground-water flow system in Tooele Valley is mainly from (1) subsurface inflow from the consolidated-rock aquifers in the surrounding mountains and stream-channel deposits at the mouths of canyons, (2) infiltration of precipitation on the valley floor, (3) seepage of unconsumed irrigation water from irrigated fields and lawns/gardens, and (4) subsurface inflow from Rush Valley through the Stockton Bar. Discharge from the basin-fill ground-water flow system in Tooele Valley is mainly by (1) withdrawals from pumped wells and discharge from flowing wells, (2) evapotranspiration, (3) spring discharge at Dunne's Pond Springs, Mill Pond Springs, source of Sixmile Creek, and source of Fishing Creek, (4) subsurface outflow to Great Salt Lake, and (5) flow to shallow drains and ditches.

Ground water in Tooele Valley generally moves from primary recharge areas near the margins of the valley to the central and northern parts of the valley. Vertical gradients are downward in the recharge areas and upward in the discharge areas, where ground water discharges naturally from the system. In the central and northern parts of the valley, ground water moves upward in the confined aquifer, into and through the overlying confining layer, and into the shallow unconfined aquifer.

To test and refine the conceptual understanding of the ground-water system of Tooele Valley, a numerical model was constructed that formulates separate components of the system and simulates the interactions between them. Requirements for construction of a numerical model include horizontal and vertical discretization, and establishing spatial distributions for the hydrologic properties of the ground-water system. Mathematical boundaries were assigned to the model domain that realistically depict actual hydrologic boundaries and conditions.

The ground-water flow system in Tooele Valley was subdivided horizontally into a grid of 110 rows and 118 columns, with grid dimensions varying from 1,000 ft in each direction to 1,000 ft by 2,160 ft. Vertically, the aquifer system is divided into five layers. The shallow unconfined aquifer and the underlying shallow confining layer are represented by model layer 1; model layers 2 through 5 represent the principal aquifer.

Specified-flux boundaries are used to simulate recharge entering the ground-water flow system as (1) inflow from consolidated rock at the consolidated-rock/basin-fill boundary at the margins of the valley, (2) underflow in stream-channel material at the mouths of canyons and seepage from streams that enter the valley, (3) subsurface inflow from Rush Valley through the Stockton Bar, (4) infiltration of precipitation on the valley floor, and (5) seepage of unconsumed irrigation water from fields and lawns/gardens. Specified-flux boundaries also are used to simulate discharge from the ground-water flow system to wells. Head-dependent flux boundaries are used to simulate (1) discharge to springs, flowing wells, and shallow drains and ditches, and (2) discharge from evapotranspiration.

A steady-state simulation was developed to represent conditions in 1968. The simulation incorporated specified rates of recharge from natural sources that were assumed to represent long-term average rates. Discharge from pumping wells incorporated in the 1968 steady-state simulation was specified as the annual average pumpage during 1964-68.

Final statistics for residuals for the steady-state simulation indicate that steady-state calibration resulted in a reasonable match between model-computed and measured water levels throughout most of the modeled area. The distributions of residuals for individual observation sites generally do not indicate a bias in the distribution of positive and negative values. In the Erda area near a zone of high horizontal gradient, however, computed water levels in the steady-state simulation are generally lower than observed levels. The match between model-computed and the estimated discharge at valley springs at the sources of Sixmile and Fishing Creeks is reasonably good. Computed discharge at Dunne's Pond Springs is substantially lower than estimated discharge, and computed discharge at Mill Pond Springs is higher than estimated discharge. Total computed discharge at the two springs is 9,300 acre-ft/yr, which is about 85 percent of the total estimated discharge for the two springs.

The transient-state simulation represents estimated hydrologic conditions for 1969-94. The purposes of the simulation were to (1) evaluate the relation between annual water-level fluctuations in the valley and variations in annual ground-water recharge caused by changes in precipitation in and around the valley, and (2) to estimate storage properties of the aquifer system.

The accuracy of the transient-state simulation was determined mainly by comparing model-computed water-level changes from one stress period to the next with measured water-level fluctuations at observation wells. Also, computed water levels for stress period 24 representing conditions during 1992 were compared with measured 1992 water levels in the south end of the valley near Tooele Army Depot. The results of transient-state calibration indicate that the generally observed rising and declining trends in water levels at observation wells are reproduced in many areas of the model. The model, however, generally does not simulate the large and abrupt changes in water levels from year to year that are observed at some wells; that is, the changes in water-level altitude simulated in the model are generally more gradual than those observed in the field. Also, in several areas the model did not accurately simulate large rises in water levels during periods of greater-than-normal precipitation during 1982-84. These discrepancies may indicate that the horizontal and vertical distribution of subsurface recharge and the method of varying recharge annually in the model do not accurately represent the actual system.

At the end of model calibration, a sensitivity analysis was done to determine the effects that model parameters have on results of the 1968 steady-state simulation. A 50-percent reduction in the vertical leakance value between model layers 1 and 2 caused simulated water levels to increase slightly. Results are sensitive to increases in horizontal hydraulic conductivity; raising values more than 5 percent caused dewatering at individual cells and the model was not able to reach a numerical solution. Model results are also sensitive to changes in specified recharge from consolidated rock. Decreasing this recharge by 15 percent caused average simulated water levels to decline almost 10 ft.

The hydrologic system in Tooele Valley is complex and cannot be defined completely with available data. The numerical model documented in this report is based on mathematical representations of ground-water flow and on a simplified set of assumptions about the hydrologic system. Model parameters, including parameters that define aquifer properties, were estimated on a regional scale. This spatial averaging and simplification results in smoothing of local anomalies. These aspects must be considered when evaluating model response for local areas and subregions of Tooele Valley. If the model were used to simulate changes in hydrologic conditions for periods shorter than a year, seasonal or monthly changes in recharge and discharge might need to be incorporated and it might be necessary to recalibrate the model. However, the model incorporates the current conceptual model of the valley's flow system as defined from available data and field observations, and analyses of ground-water flow using this model and future simulations to determine the effects of regional changes in recharge and discharge to the ground-water flow system are expected to produce useful results.

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