DETERMINATION OF HYDROLOGIC PROPERTIES NEEDED TO CALCULATE AVERAGE LINEAR VELOCITY AND TRAVEL TIME OF GROUND WATER IN THE PRINCIPAL AQUIFER UNDERLYING THE SOUTH-EASTERN PART OF SALT LAKE VALLEY, UTAH

By G.W. Freethey, L.E. Spangler, and W.J. Monheiser[1]

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
foot	0.3048	meter
foot per day	0.3048	meter per day
foot per mile	0.1894	meter per kilometer
gallon per minute	0.06308	liter per second
mile	1.609	kilometer
square mile	2.59	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.

DETERMINATION OF HYDROLOGIC PROPERTIES NEEDED TO CALCULATE AVERAGE LINEAR VELOCITY AND TRAVEL TIME OF GROUND WATER IN THE PRINCIPAL AQUIFER UNDERLYING THE SOUTHEASTERN PART OF SALT LAKE VALLEY, UTAH

by G.W. Freethey, L.E. Spangler, and W.J. Monheiser[1]

ABSTRACT

A 48-square-mile area in the southeastern part of the Salt Lake Valley, Utah, was studied to determine if generalized information obtained from geologic maps, water-level maps, and drillers' logs could be used to estimate hydraulic conductivity, porosity, and slope of the potentiometric surface: the three properties needed to calculate average linear velocity of ground water. Estimated values of these properties could be used by water-management and regulatory agencies to compute values of average linear velocity, which could be further used to estimate travel time of ground water along selected flow lines, and thus to determine wellhead protection areas around public-supply wells.

The methods used to estimate the three properties are based on assumptions about the drillers' descriptions, the depositional history of the sediments, and the boundary conditions of the hydrologic system. These assumptions were based on geologic and hydrologic information determined from previous investigations. The reliability of the estimated values for hydrologic properties and average linear velocity depends on the accuracy of these assumptions.

Hydraulic conductivity of the principal aquifer was estimated by calculating the thickness-weighted average of values assigned to different drillers' descriptions of material penetrated during the construction of 98 wells. Using these 98 control points, the study area was divided into zones representing approximate hydraulic-conductivity values of 20, 60, 100, 140, 180, 220, and 250 feet per day. This range of values is about the same range of values used in developing a ground-water flow model of the principal aquifer in the early 1980s.

Porosity of the principal aquifer was estimated by compiling the range of porosity values determined or estimated during previous investigations of basin-fill sediments, and then using five different values ranging from 15 to 35 percent to delineate zones in the study area that were assumed to be underlain by similar deposits. Delineation of the zones was based on depositional history of the area and the distribution of sediments shown on a surficial geologic map.

Water levels in wells were measured twice in 1990: during late winter when ground-water withdrawals were the least and water levels the highest, and again in late summer, when ground-water withdrawals were the greatest and water levels the lowest. These water levels were used to construct potentiometric-contour maps and subsequently to determine the variability of the slope in the potentiometric surface in the area.

Values for the three properties, derived from the described sources of information, were used to produce a map showing the general distribution of average linear velocity of ground water moving through the principal aquifer of the study area. Velocity derived ranged from 0.06 to 144 feet per day with a median of about 3 feet per day. Values were slightly faster for late summer 1990 than for late winter 1990, mainly because increased withdrawal of water during the summer created slightly steeper hydraulic-head gradients between the recharge area near the mountain front and the well fields farther to the west. The fastest average linear-velocity values were located at the mouth of

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Little Cottonwood Canyon and south of Dry Creek near the mountain front, where the hydraulic conductivity was estimated to be the largest because the drillers described the sediments to be predominantly clean and coarse grained. Both of these areas also had steep slopes in the potentiometric surface. Other areas where average linear velocity was fast included small areas near pumping wells where the slope in the potentiometric surface was locally steepened. No apparent relation between average linear velocity and porosity could be seen in the mapped distributions of these two properties. Calculation of travel time along a flow line to a well in the southwestern part of the study area during the summer of 1990 indicated that it takes about 11 years for ground water to move about 2 miles under these pumping conditions.

INTRODUCTION

Protecting ground water from organic, inorganic, radioactive, and biological contaminants, many of which can be classified as hazardous waste, is of great concern to cities that derive part or all of their public water supply from aquifers. The 1986 Amendments to the Safe Drinking Water Act (SDWA) established the Wellhead Protection (WHP) Program. This program is designed to assist States in protecting areas surrounding public water-supply wells against contaminants that could have an adverse effect on human health. The U.S. Environmental Protection Agency (EPA) assists the States in the development of State WHP Programs. One of the major elements of WHP is the determination of zones within which contaminant assessment and management is addressed. These zones, called Wellhead Protection Areas (WHPAs), are defined in the SDWA as the surface and subsurface area surrounding a water well or wellfield, supplying a public water system, through which contaminants are reasonably likely to move toward and reach such water well or wellfield.

Criteria for delineating a protection area have been based on distance, water-level declines caused by pumping, travel time, flow-system boundaries, and the capacity of an aquifer to absorb contaminants. These criteria have led to the development of methods as uncomplicated as defining an arbitrary fixed radius around a well and as complicated as using analytical solutions to flow equations to determine zones of capture of a pumping well. This report describes and demonstrates a method of estimating horizontal ground-water travel time along flow lines, which could be used in defining a WHPA.

The basic methods of estimating hydraulic conductivity, effective porosity, and slope of the potentiometric surface, and calculating average linear velocity of ground water along a flow line described in this report can be used by water-supply agencies to obtain a preliminary estimate of ground-water travel time to a specific well at one point in time, and to obtain a general idea of how large the wellhead protection area might have to be.

The State of Utah is developing a WHP program that will require defining areas of protection on the basis of various time periods necessary for contaminated ground water to move to a public supply well. The first time period specified is 250 days, which is defined as the minimum time necessary to decrease the risk of contamination from pathogenic microorganisms and some organic chemicals to an acceptable level and/or to complete a suitable remediation process (M. Jensen, Utah Department of Environmental Quality, Division of Drinking Water, written commun., 1991). The second time period specified is 15 years. The zone defined by a 15-year travel time was designed to decrease the risk of inorganic chemical contamination to an acceptable level.

At the request of EPA, the U.S. Geological Survey (USGS) investigated methods to derive or estimate the three hydrologic properties necessary to calculate average linear velocity of ground water moving horizontally through the principal aquifer in the southeastern part of Salt Lake Valley. These properties are hydraulic conductivity, effective porosity, and slope of the potentiometric surface.

Purpose and Scope

The purpose of this report is to summarize and describe the method used to derive or estimate hydrologic properties needed to calculate horizontal average linear velocity of ground water through an aquifer system similar to that found along the Wasatch Front in Utah, and to demonstrate how average linear velocity can be used to obtain travel time of ground water along a flow line.

The scope of the investigation was governed by the need to use simple methods of determining the three properties. Many complicating factors were not considered as part of this investigation because of these limitations. As a result, only one part of a ground-water flow line was considered: the horizontal flow through the principal aquifer. Other parts of a ground-water flow line that could not be considered would include those through unsaturated zones, vertical flow through aquifers and confining layers, and vertical and horizontal flow through the consolidated rocks of the Wasatch Range. Limitations of using this methodology will be discussed further in "Limitations in application of methods and use of results" (at end of report).

The area used to demonstrate the methods is a typical part of the basin-fill aquifer along the Wasatch Front in Utah (fig. 1). The area includes a recharge area to the east along the front of the Wasatch Range where the principal aquifer is mostly unconfined and a discharge area to the west along the Jordan River where the principal aquifer is mostly confined.

Geographic and Geologic Setting

The study area includes 48 square miles in the southeastern part of Salt Lake Valley, northwestern Utah. The Wasatch Range is at the eastern boundary of the area, and the Jordan River is at the western boundary (fig. 1). The surface drainage over the study area is from east to west with about 1,000 feet of topographic relief in a distance of about 7 miles. The study area includes the smaller communities of Sandy City and Midvale, south of Salt Lake City, that have their own public water-supply systems. In the study area, the consolidated rocks of the Wasatch Range consist of faulted quartz monzonite of Tertiary age and quartzite and shale of Precambrian age (Davis, 1983). Surficial deposits on the valley floor are of Quaternary age and consist of flood-plain deposits along the Jordan River, Big Cottonwood Creek, and Little Cottonwood Creek; glacial moraines and talus deposits near the front of the Wasatch Range; narrow bands of stream alluvium in the channels of Little Cottonwood and Dry Creeks; abandoned flood-plain and stream deposits along Dry Creek near the mountains and south of Little Cottonwood Creek; and Lake Bonneville deposits grading from coarse-grained beach deposits near the mountains to fine-grained lake-bottom sediments near the Jordan River (fig. 2).

Conceptualized Ground-Water System

The ground-water system in the Salt Lake Valley consists of a discontinuous shallow unconfined aquifer; a deep, confined principal aquifer system made up of several sand and gravel beds generally separated from one another by silt or clay confining beds; and a deep, unconfined or semi-confined aquifer near the mountain front in which fine-grained deposits that can form confining units are absent or less numerous (fig. 3). All of the saturated sand and gravel beds are hydraulically connected to a greater or lesser degree, either horizontally through stringers of sand and gravel, or vertically through layers of silt and clay. In addition, the fractured consolidated rocks of the Wasatch Range, at least near the mountain front, also are part of the ground-water system. These rocks apparently are an important avenue of subsurface movement of ground water from the areas in the mountains where snow accumulates, melts, and recharges the rocks, to the aquifers in the basin fill (fig. 3).

The unconsolidated basin fill in the valley is recharged by downward percolation of precipitation, by seepage from streams and canals, by infiltration of water downward from irrigated fields, lawns, and gardens, and by channel underflow from canyons. Recharge also occurs by movement of water downward from the land surface into interconnected fractures in the consolidated rocks of the mountains and then laterally in the subsurface into the deeper part of the unconsolidated and semiconsolidated basin fill (fig. 3).

Three recharge areas have been conceptualized (Gates and Freethey, 1989) on the basis of relative differences in the surface and subsurface character of the porous or fractured media through which infiltrating water must move in order to reach the saturated zone. These three recharge areas have been termed the consolidated-rock recharge area, the primary recharge area, and the secondary recharge area (fig. 3). The consolidated-rock recharge area is defined as that part of the Wasatch Range where precipitation percolates down through interconnected fractures in the rocks and moves laterally toward and into the adjacent basin fill (fig. 3). The primary recharge area is defined as that part of the valley, usually immediately adjacent to the mountain front, where precipitation and runoff from the mountains can percolate downward to the water table with little or no impedance because few, if any, silt and clay layers are present in the basin fill. The secondary recharge area is defined as that part of the valley, usually immediately downslope from the primary recharge area, where discontinuous silt and clay layers exist, but where recharge is still possible because hydraulic heads decrease with depth implying downward ground-water movement.

After water reaches the unconfined part of the principal aquifer, it moves horizontally toward the center of the valley through the coarse-grained deposits of the basin fill, and then vertically upward toward the land surface where it is discharged by evapotranspiration, springs, or seepage to the Jordan River.

The horizontal and vertical distribution of the various types of unconsolidated deposits, in terms of type of depositional process and lithology, influence the variability in the rate of movement of ground water in the study area. Quaternary sediments at depth probably were deposited under conditions related to lacustrine, fluvial, alluvial, and glacial depositional processes. These depositional conditions were also

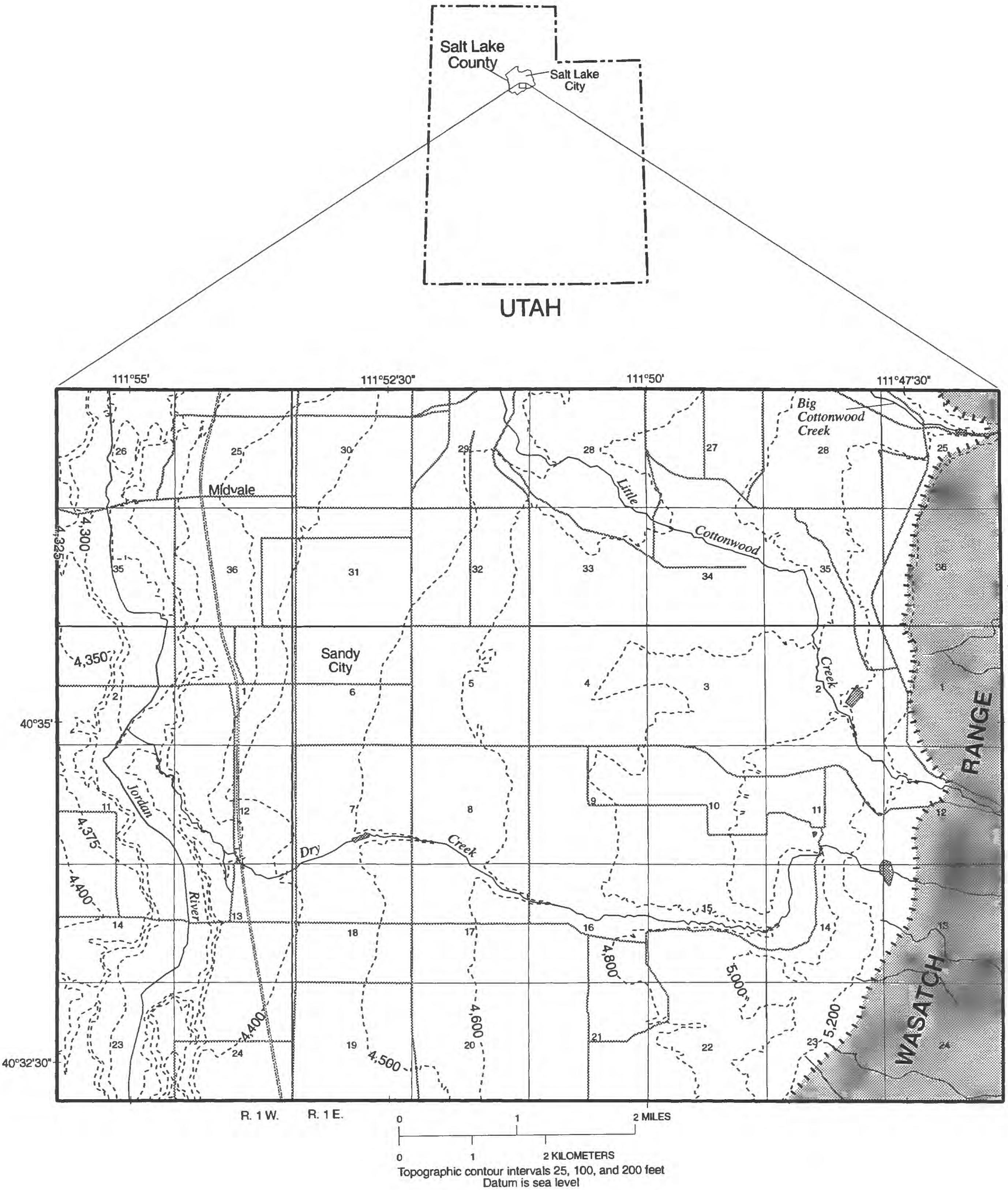

Figure 1. Location and setting of the Salt Lake Valley study area.

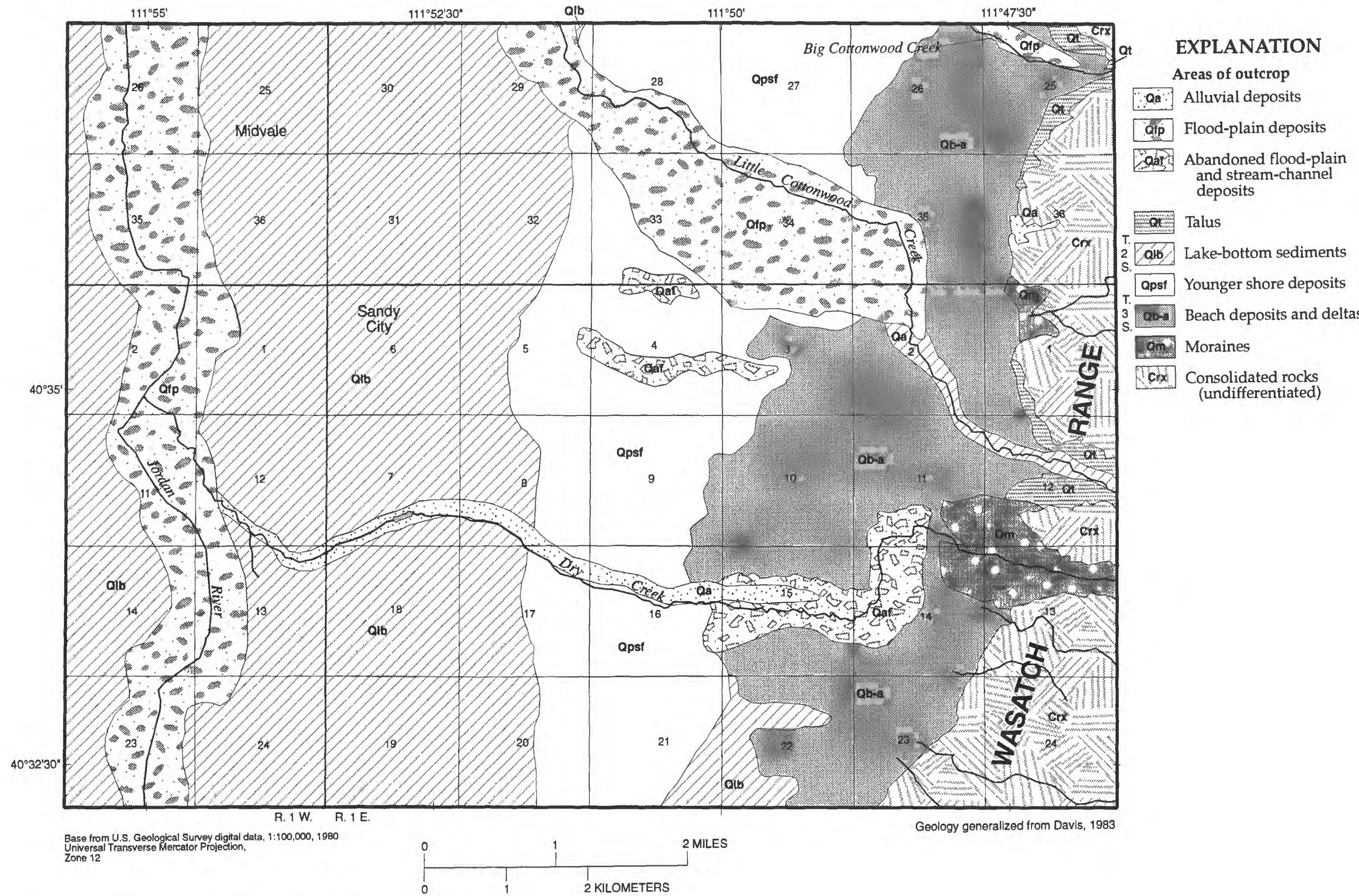

Figure 2. Generalized surficial geology of Salt Lake Valley.

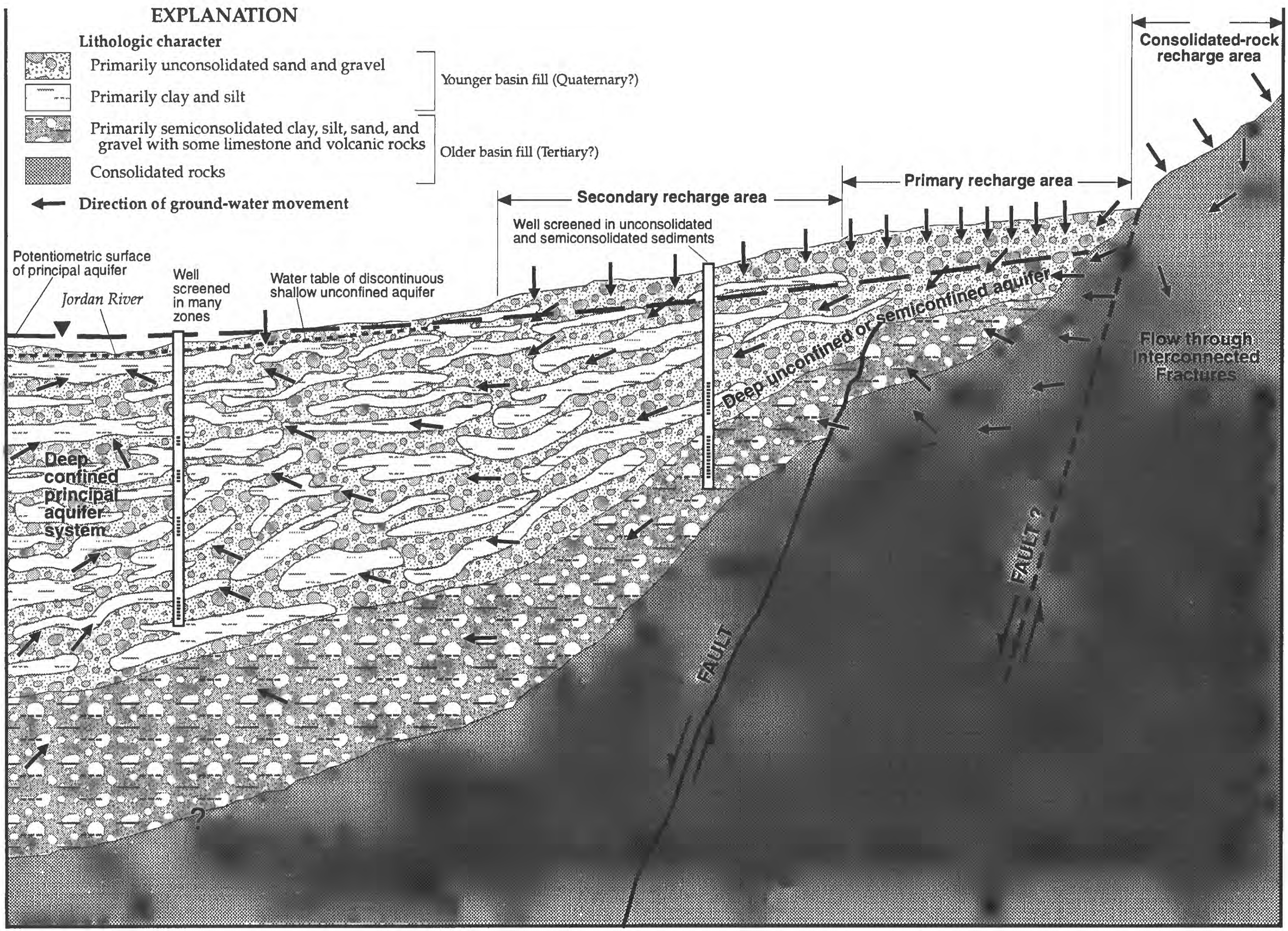

Figure 3. Schematic diagram of recharge to and movement of ground water in the basin-fill aquifers of Salt Lake Valley.

prevalent during deposition of the near-surface sediments; thus, values for hydraulic conductivity or effective porosity for sediments at depth are probably similar to the values for sediments near land surface. The main difference between hydrologic properties of deposits mapped at land surface and those at depth is in the lateral distribution rather than the values. The contacts shown on the surficial geologic map would therefore give only a general indication of where the types of deposits might change from one to another at depth.

Availability of Hydrologic Data

Information about hydraulic-head values and aquifer properties is available for selected locations where public water-supply companies and private land owners have drilled wells. Aquifer properties are best defined in the northern and western parts of the study area where population is largest and where pumping levels of wells are near land surface, thus decreasing pumping costs (fig. 4). Some information is available for other parts of the study area because water suppliers have had to explore the subsurface in outlying areas in order to meet the increasing demand for water. Aquifer tests of limited scope have been done on nine wells, mainly located in a relatively small area east of Midvale and Sandy City, and the results were used to estimate hydrologic properties of aquifers and confining layers.

Numbering System for Wells

The system of numbering wells in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the well, describes its position in the land net. By the land-survey system, the State is divided into four quadrants by the Salt Lake Base Line and the Salt Lake Meridian. These quadrants are designated by the uppercase letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range, in that order, follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section, and is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section, generally 10 acres[1] for regular sections. The letters a, b, c, and d indicate, respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the well within the 10-acre tract; thus (D-2-1)28cdd-2 designates the second well constructed or visited in the southeast quarter of the southeast quarter of the southwest quarter, section 28, T. 2 S., R. 1 E. The numbering system is illustrated in figure 5.

DETERMINATION OF HYDROLOGIC PROPERTIES

The methods used to estimate the three properties needed to calculate average linear velocity and travel time are based on assumptions about the drillers' descriptions, the depositional history of the sediments, and the boundary conditions of the hydrologic system. These assumptions were based on geologic and hydrologic information determined from previous investigations. The reliability of the estimated values for hydrologic properties and average linear velocity depends on the accuracy of these assumptions.

The equation used to obtain average linear velocity of ground water moving through a porous media (Lohman, 1972, p. 10) is:

$$\text{Average linear velocity} = -\frac{\text{hydraulic conductivity}}{\text{effective porosity}} \times \text{hydraulic head gradient} \tag{1}$$

The negative sign indicates that the direction of movement is downgradient. Methods used to determine an areal distribution of values for hydraulic conductivity, effective porosity, and slope of the potentiometric surface included obtaining values from textbooks and previously published reports, using descriptions of sediments by drillers to estimate an average hydraulic-conductivity value, estimating effective porosity on the basis of character of sediments and assumed depositional environment, and collecting water-level data from well owners and measuring water levels in some wells.

Travel time along any given segment of a flow line is obtained from:

$$\text{Travel time} = \frac{\text{Length of flow-line segment}}{\text{Average linear velocity along flow-line segment}} \tag{2}$$

Travel time along a specific flow line is a summation of travel times along all flow-line segments.

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[1]Although the basic land unit, the section, is theoretically 1 square mile, many sections are irregular. Such sections are divided into 10-acre tracts, generally beginning at the southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

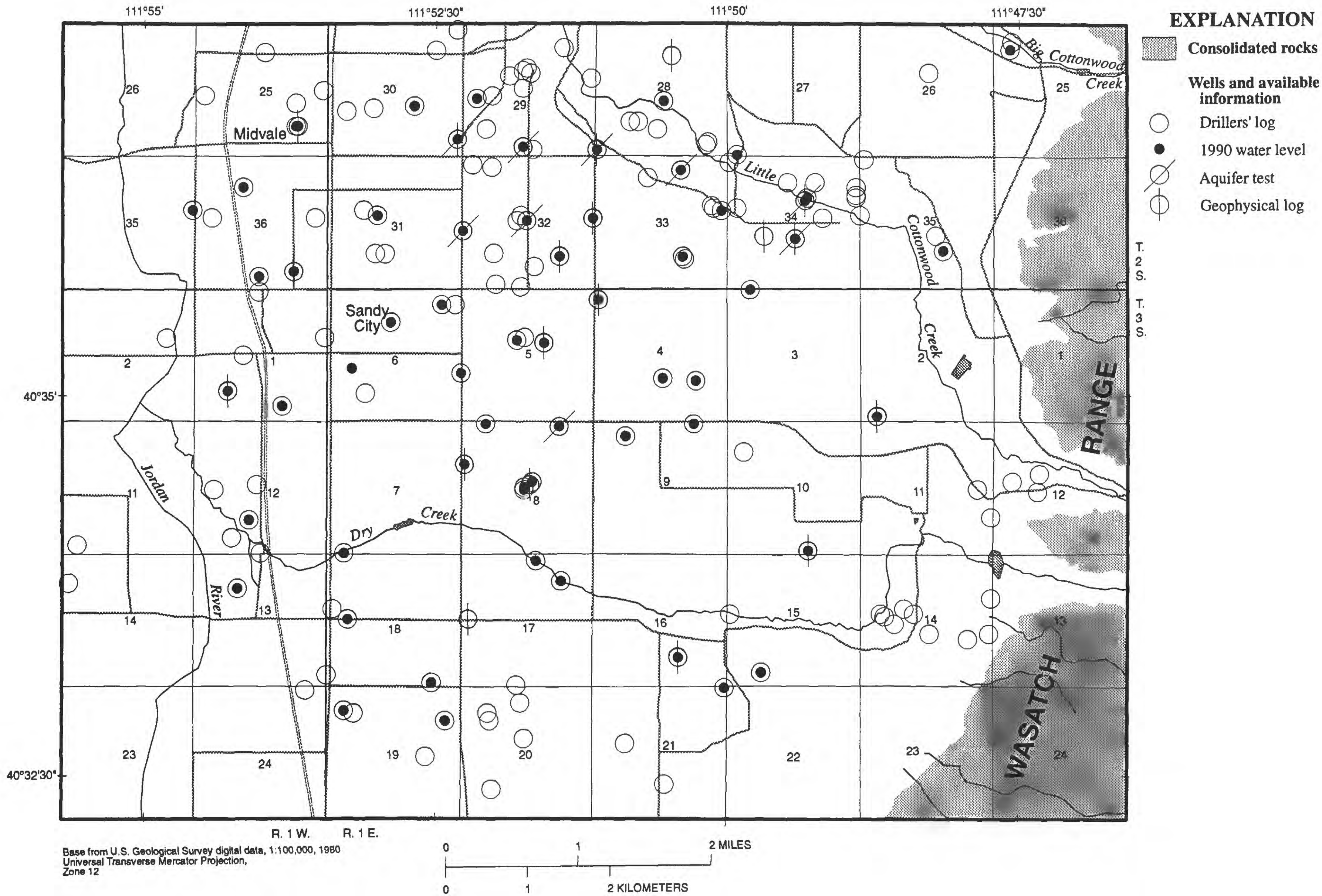


Figure 4. Distribution of wells in Salt Lake Valley and available information.

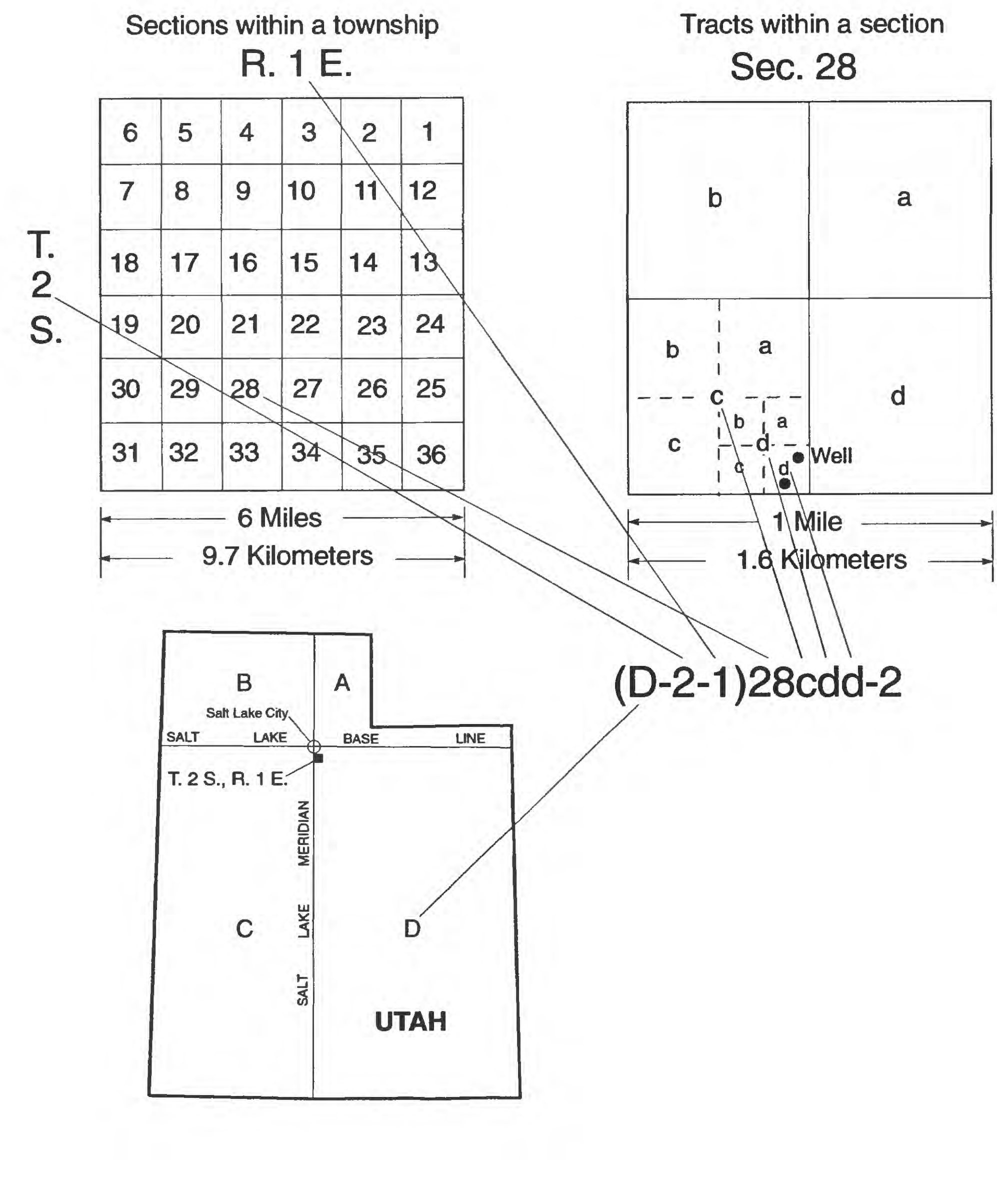


Figure 5. Data-site numbering system used in Utah.

Travel time along a given flow line is determined by accumulating the individual travel times for smaller flow-line segments radiating from a well until the given travel time is exceeded. The length of the flow line is equal to the sum of the lengths of the complete flow-line segments that were required plus the length of the last partial flow line within which the given travel time was exceeded. The length of that partial flow-line segment to be included in the total length of the flow line is determined by multiplying the average linear velocity for that last segment by the travel time remaining after the travel time through the complete flow-line segments is subtracted from the given travel time.

Hydraulic Conductivity

Ground water moving from the recharge area to the discharge area of the principal aquifer underlying the Salt Lake Valley would flow preferentially through the coarse-grained basin fill. Hydraulic-conductivity values for these basin-fill sediments have a large range. Values compiled by Bedinger and others (1986) for coarse-grained basin-fill sediments typical in the Basin and Range province (Fenneman, 1931) had a range of four orders of magnitude. Most hydraulic-conductivity values compiled, those within one standard deviation of the mean value, ranged from about 3 to 230 feet per day. The mean value was about 30 feet per day.

Hydraulic-conductivity values were estimated for saturated sediments penetrated by five of the nine wells for which some type of aquifer testing was done. Four of the wells on which aquifer tests had been done could not be positively matched with a drillers' log and were not used in the comparison. The five single-well tests used for this study were analyzed in the mid-1960s using the Cooper and Jacob (1946) straight-line solution for drawdown and recovery in the pumped well; no observation wells were used. These values, along with values derived from numerous other aquifer tests and specific capacity values, were used by Hely and others (1971, fig. 59) to create a transmissivity distribution for the Salt Lake Valley. The hydraulic-conductivity values from the aquifer tests were calculated by dividing the test-derived value of transmissivity by the thickness of the coarse-grained unit or units from which it was assumed water was being pumped during the aquifer test. Average hydraulic-conductivity values for the five aquifer tests ranged from about 7 to 95 feet per day. Because the analyses used to calculate transmissivity from the test data did not consider leakage through or drainage from the overlying confining layers, the values of transmissivity, and thus average hydraulic conductivity determined for the aquifer at these wells are probably larger than actual values.

The estimated hydraulic-conductivity values derived from aquifer testing were used to assign hydraulic-conductivity values to individual layers of sediments described in the drillers' logs of wells used for the aquifer tests. The sediments described by drillers opposite the perforated intervals in the tested wells were assigned initial average hydraulic-conductivity values on the basis of typical hydraulic-conductivity values from textbooks such as Freeze and Cherry (1969, p. 29), Davis and DeWiest (1966, p. 375), and Bouwer (1978, p. 38). These values were arbitrarily adjusted until an average hydraulic conductivity derived from assigned values for the different drillers' descriptions was about the same as the values derived from aquifer tests at the five aquifer-test sites.

The first step in making hydraulic-conductivity estimates for the wells where no aquifer testing was done was to determine the top of the principal aquifer in each drillers' log. This determination depended on the location of the well. For wells in the eastern part of the study area where the principal aquifer is unconfined, the top of the aquifer was defined as the depth to water at the time the well was drilled. Farther to the west, where the principal aquifer is confined and the potentiometric surface is near land surface, the top of the aquifer is more difficult to identify. Generally, the top was defined as the bottom of the first clay or silt layer. This layer was below the potentiometric surface and below the water table of the overlying shallow unconfined aquifer, which was more than 20 feet thick. A fine-grained layer was not always evident in drillers' descriptions, and occasionally it was necessary to estimate the top of the principal aquifer on the basis of the depth of fine-grained layers that may have also contained some sand or gravel, such as those described for example, as gravelly clay, sandy clay, or sand, gravel, and clay.

The total thickness of basin fill penetrated by drilling also varied. As a result, average hydraulic-conductivity values were based on different thicknesses of coarse-grained basin fill. Hydraulic conductivity estimated from a drillers' log of a well that penetrated only 250 feet of aquifer material may differ substantially from a value estimated from a drillers' log of a well that penetrated 750 feet of material, especially if the deeper sediments are, on the average, more coarse or fine grained than the upper 250 feet of aquifer sediments. To decrease inconsistencies because of depth, wells less than 200 feet deep were not used, and wells between 200 and 300 feet deep were selectively omitted if the drillers' log was not particularly detailed in the lithologic description. On the basis of these arbitrary criteria, about one-third of the wells with drillers' logs were not used to estimate an average hydraulic-conductivity value to define the distribution for the study area.

The second step needed for the hydraulic-conductivity estimates was to assign the hydraulic-conductivity values determined from the drillers' logs of wells where the five aquifer tests had been done (table 1) to the various types of basin-fill sediments described in drillers' logs of wells where no testing was done. A typical drillers' log submitted to the State of Utah includes a description of materials penetrated during drilling and the associated depth interval. The values were applied to sediments described in drillers' logs of wells that were not used for an aquifer test and an equivalent horizontal hydraulic conductivity was calculated (Freeze and Cherry, 1979, p. 33-34) as the thickness-weighted arithmetic mean of the coarse-grained basin-fill sediments. In this way, hydraulic-conductivity values were estimated from the logs of 98 wells in the study area. Layers of silt and clay were not included as part of the overall thickness. Because the values presented in table 1 are arbitrary and have no analytical or laboratory basis, they should not be used if values determined from testing are available. They represent one set of many possible sets of hydraulic-conductivity values that could be assigned to basin-fill materials.

The distribution of hydraulic-conductivity values (fig. 6) that resulted from the arbitrary assignments is probably typical of aquifers in basin-fill sediments in that the largest values are concentrated close to stream channels near the mountain front where the coarsest grained deposits would have accumulated. The distribution at a distance from the Wasatch Range appears random, but is probably related, in some way, to depositional patterns. It is likely that the sediments associated with mass wasting deposits near the mountain front have a small hydraulic conductivity. This pattern is not shown on the map because no control points were located in these sediments. The magnitude of the hydraulic-conductivity values shown in figure 6, multiplied by the saturated thickness of coarse-grained material, is consistent with the magnitude of transmissivity values used in a ground-water flow model of the area (Waddell and others, 1987, p. 34). Comparison of values in the model with these arbitrary assignments from drillers' descriptions indicate that the assigned values are of the correct order of magnitude to allow an appropriate quantity of ground water to recharge, to move horizontally through the aquifer, and to discharge, as conceptualized by Waddell and others (1987).

Porosity

Effective porosity as used in the average linear-velocity equation, for the purposes of this report termed porosity, refers to the percentage of interconnected pore space available for fluid movement (Lohman and others, 1972, p. 10). For this investigation, all pores in the basin-fill sediments of the study area are assumed to be interconnected. The percentage of pore space assumed to be interconnected in sediments that include clay- or silt-size particles is usually greater than that of sediments that are all sand- or gravel-size particles; however, the size of pores in sediments containing clay and silt is small, the hydraulic conductivity is small, and fluid movement is much slower than through the pores of sand and gravel deposits. The slower movement is attributed to increased molecular attraction between the porous medium and the fluid as pores become small and as the surface area common to the particles and the fluid increases. Thus, in the equation used to obtain average linear velocity, the larger values of porosity and the smaller values of hydraulic conductivity associated with basin fill that contains a greater percentage of clay and silt would yield the slowest average linear-velocity values unless hydraulic-head gradients were extremely large.

Porosity values of unconsolidated sediments, reported in the literature, range from 10 to 70 percent (table 2).The sediments through which most of the ground water in the study area moves are mixtures of gravel and sand containing some silt and clay. A common range of values for these types of sediment is about 15 to 35 percent. Davis and DeWiest (1966, p. 375) indicated that values of porosity depend on the mode of deposition, which controls the degree of sediment sorting. A deposit that includes grains of many different sizes (poorly sorted) will have a smaller porosity than a deposit that contains only a few different grain sizes (well sorted).

The depositional environment of the sediments that make up the aquifers in the basin fill in the study area varies from predominantly alluvial-fan, stream-channel, beach, moraine, and mass-wasting environments near the Wasatch Range to mainly flood-plain and lacustrine environments on the west side near the Jordan River. Although overall depositional changes from Wasatch Range to the Jordan River are probably gradational, changes between individual deposits could be distinct. Coarse- to fine-grained, poorly sorted landslide, mudslide, and moraine deposits occur closest to the mountains. Poorly to moderately sorted coarse-grained alluvial fan material occurs near the mouths of existing drainages and becomes finer grained with increasing distance from the Wasatch Range. Moderately well-sorted, coarse-grained stream-channel

Table 1. Assignments of hydraulic-conductivity values to sediments described by drillers

Material described	Assigned hydraulic-conductivity value (feet per day)
Fines (clay and silt)	0
Fines and sand	10
Fines, sand, and cobbles	20
Fines and gravel or cobbles	30
Fines, sand, and gravel	50
Sand	80
Sand and cobbles	100
Sand and gravel	120
Sand, gravel, and cobbles	170
Gravel	200
Gravel and cobbles	230
Cobbles	250
(If any of the above were described as cemented or hard, a hydraulic conductivity of 10 percent of the assigned value was used.)	
Conglomerate	10
(assumed to be less permeable than cemented or hard sand and gravel; sand and cobbles; sand, gravel, and cobbles; gravel; gravel and cobbles; or cobbles)	
Hardpan	5

deposits occur downgradient from the canyons where the large streams flow from the Wasatch Range. Finer grained flood-plain deposits would probably occur adjacent to these channels and along the axis of the valley. Well-sorted sand and gravel occurs as lake-shore deposits parallel to the mountain front and at different depths and distances from the Wasatch Range. Lake deposits are interbedded with alluvial and stream-channel deposits and might not occur at all depths. In a general way, the surficial geology map (fig. 2) indicates the distribution of different deposits at depth, but surficial geology gives no indication of the extent of interlayering at depth.

During a previous study (P.B. Anderson, K.R. Thompson, S.R. Wold, V.M. Heilweil, and R.L. Baskin, U.S. Geological Survey, written commun., 1990), which delineated different types of recharge areas along the Wasatch Front, investigators studied well logs to find the approximate limit of the fine-grained unit that confines the principal aquifer. This limit indicates where the confining unit can be identified by the increased prevalence of thicker clay lenses. This limit of the confining unit also may be a good approximation of the area in which porosity values probably change because the lithologic character changes from poorly sorted alluvial-fan deposits to better sorted lacustrine and stream deposits.

To determine the distribution of porosity values to be used in the calculation of average linear velocity of ground water, the geologic map, the surface-drainage system, the location of the limit of the confining unit for the principal aquifer, and the location of valleys cut into the consolidated rocks of the Wasatch Range were used to delineate five areas in which porosity values might differ substantially (fig. 7). An area along the front of the Wasatch Range where deposition is mainly from landslides, mudslides, talus, and glacial moraines was assigned a porosity value of 15 percent because of the extremely heterogeneous nature and poor sorting of these types of deposits. Areas farther from the mountain front, where alluvial fan deposits are likely, were assigned a porosity of 20 percent. Sorting in these

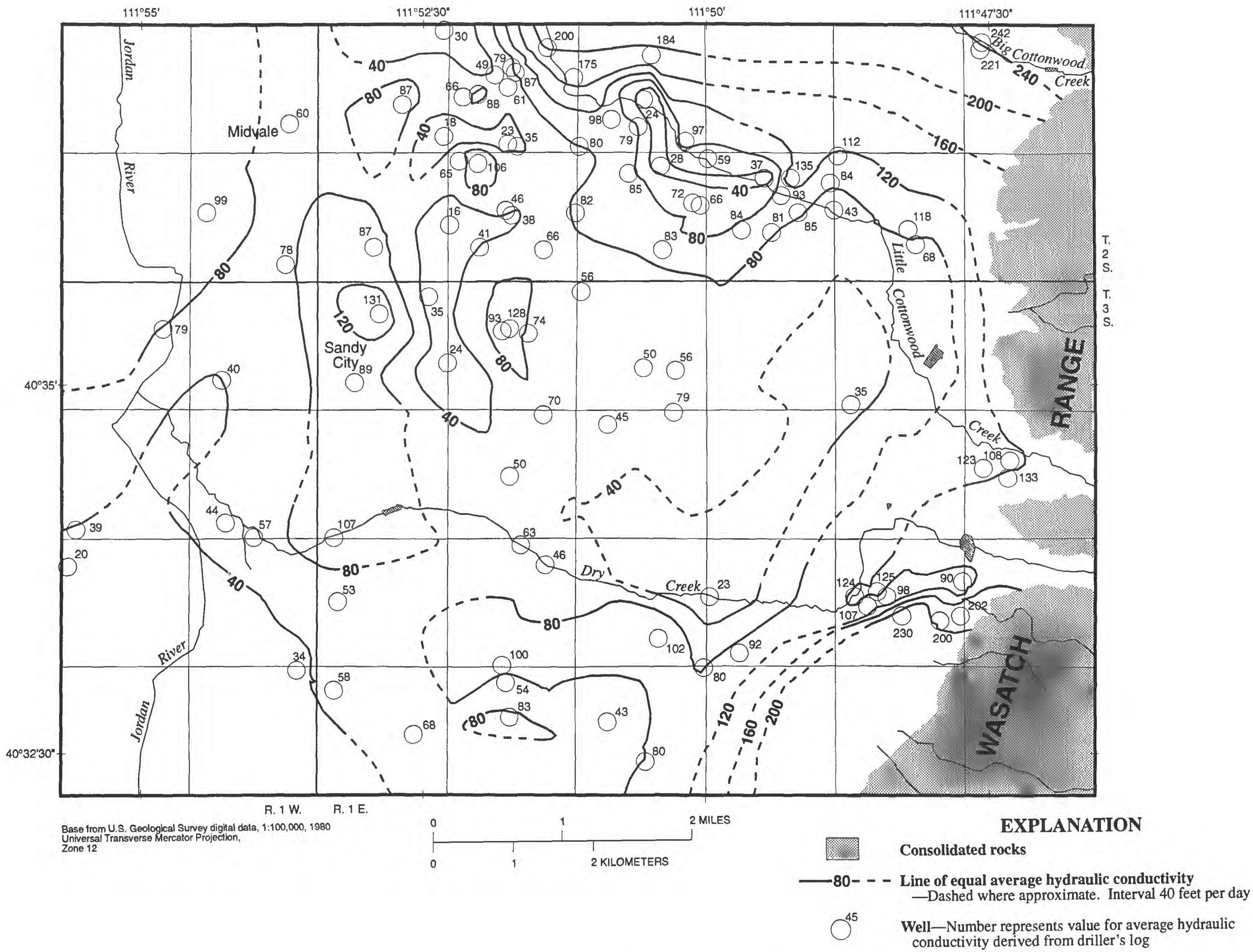


Figure 6. Distribution of average hydraulic conductivity for the principal aquifer in Salt Lake Valley on the basis of drillers' descriptions of sediments.

Table 2. Compilation of porosity values for unconsolidated sediment

Sediment description	Porosity value (in percent)	Source
GRAVEL	25 to 40	Freeze and Cherry, 1979
	30 to 40	Walton, 1970
	20 to 30	Davis, 1969
coarse	24 to 36	Morris and Johnson, 1967
medium	24 to 44	Do.
fine	25 to 39	Do.
alluvial	25	Wenzel and Fischel, 1942
SAND AND GRAVEL	10 to 35	Driscoll, 1986
	20 to 35	Walton, 1970
	10 to 30	Davis, 1969
mixed	20 to 35	Fetter, 1980
well sorted	25 to 50	Do.
coarse-grained basin fill	12 to 23	Bedinger and others, 1986
GLACIAL TILL	10 to 25	Driscoll, 1986
	25 to 45	Davis, 1969
	22 to 41	Morris and Johnson, 1967
	10 to 20	Fetter, 1980
SAND	25 to 40	Driscoll, 1986
	25 to 50	Freeze and Cherry, 1979
	35 to 40	Walton, 1970
coarse	25 to 35	Davis, 1969
coarse	31 to 46	Morris and Johnson, 1967
coarse alluvial	33	Wenzel and Fischel, 1942
medium	35 to 40	Davis, 1969
medium	29 to 49	Morris and Johnson, 1967
medium marine	40 to 42	MacCary and Lambert, 1962
fine	40 to 50	Davis, 1969
fine	26 to 53	Morris and Johnson, 1967
fine alluvial	46 to 52	MacCary and Lambert, 1962
fine alluvial	52	Wenzel and Fischel, 1942
aeolian	40 to 51	Morris and Johnson, 1967
dune	36	Brown and Newcomb, 1963
SAND AND SILT, aeolian	30 to 45	Freeze and Cherry, 1979
SAND, SILT, AND CLAY	38 to 42	K.M. Waddell, USGS, written commun., 1990
WASHED DRIFT	35 to 59	Morris and Johnson, 1967
SILT	35 to 50	Driscoll, 1986
	35 to 50	Freeze and Cherry, 1979
	34 to 61	Morris and Johnson, 1967
	35 to 50	Fetter, 1980
loess	44 to 57	Morris and Johnson, 1967
loess	49 to 51	MacCary and Lambert, 1962
dune	36	Brown and Newcomb, 1963

Table 2. Compilation of porosity values for unconsolidated sediment—Continued

Sediment description	Porosity value (in percent)	Source
SILT AND CLAY	50 to 60	Davis, 1969
fine-grained basin fill	29 to 36	Bedinger and others, 1986
SILT, CLAY, AND SAND	36 to 58	K.M. Waddell, USGS, written commun., 1990
CLAY	45 to 55	Driscoll, 1986
	40 to 70	Freeze and Cherry, 1979
	45 to 55	Walton, 1970
	34 to 57	Morris and Johnson, 1967
	33 to 60	Fetter, 1980
marine	49	MacCary and Lambert, 1962
kaolinite	50	Barber, 1955
montmorillonite	67	Do.
SOIL	50 to 60	Todd, 1959

deposits would tend to be poor and more of the fine sediments were removed and deposited farther downgradient. Stream-channel deposits occur along old stream courses, but these courses cannot be delineated from drillers' logs. Separate porosity values were not assigned to these kinds of deposits.

Flood-plain deposits on either side of the main stream channels contain fewer gravel-size particles than the stream alluvium, but are better sorted than alluvial fan deposits, talus, and moraines. The area about one-half mile on each side of Little and Big Cottonwood Creeks was assigned a porosity value of 25 percent. The western one-half of the study area is underlain by predominantly mixed lacustrine and alluvial deposits that probably are finer grained to the west. The area about 2 to 4 miles from the Jordan River is underlain by relatively well-sorted sediments that have been reworked by streams and nearshore wave action. A porosity value of 35 percent was assigned to these areas (fig. 7). The area adjacent to and underlying the Jordan River probably is underlain by a combination of fine-grained deposits and alluvial deposits. They probably are not as well sorted as the reworked sediments, and as a result, this area was assigned a porosity value of 30 percent.

The assignment of porosity values is not based on subsurface lithologic data collected in the study area, and the values may be inaccurate; however, the magnitude of possible inaccuracy in porosity is of least importance in the calculation of average linear velocity because of the relatively small range that is likely for these values. Porosity of the fine-grained confining unit was not considered because the confining sediments are not the primary sediments through which ground water moves.

Hydraulic-Head Gradient

The third factor needed to calculate average linear velocity of ground water is the hydraulic-head gradient. In order to obtain hydraulic-head gradient, water levels were measured in wells during late winter 1990 and again in late summer 1990. Late winter represents the time when ground-water withdrawals are the least and water levels the highest (January, February, and March). Late summer represents the time when ground-water withdrawals are the greatest and water levels the lowest (July, August, and September).

During late winter 1990, water levels were measured in 54 wells by personnel from the public water-supply companies and the USGS. During late summer 1990, water levels were measured in 52 wells (table 3). Measurements were made using calibrated air lines, pressure transducers, electric tape, and steel tape. Because the different methods have different measurement errors, the overall accuracy of the measurements is considered to be plus or minus 1 foot. The altitude of the measuring point for each well, in some cases, is known to within 0.1 foot from land surveys, and in other cases, it was estimated from topographic maps with an accuracy for land-surface altitudes of only plus or minus 5 feet. This could introduce some error into the altitude of the water levels shown in table 3; how-

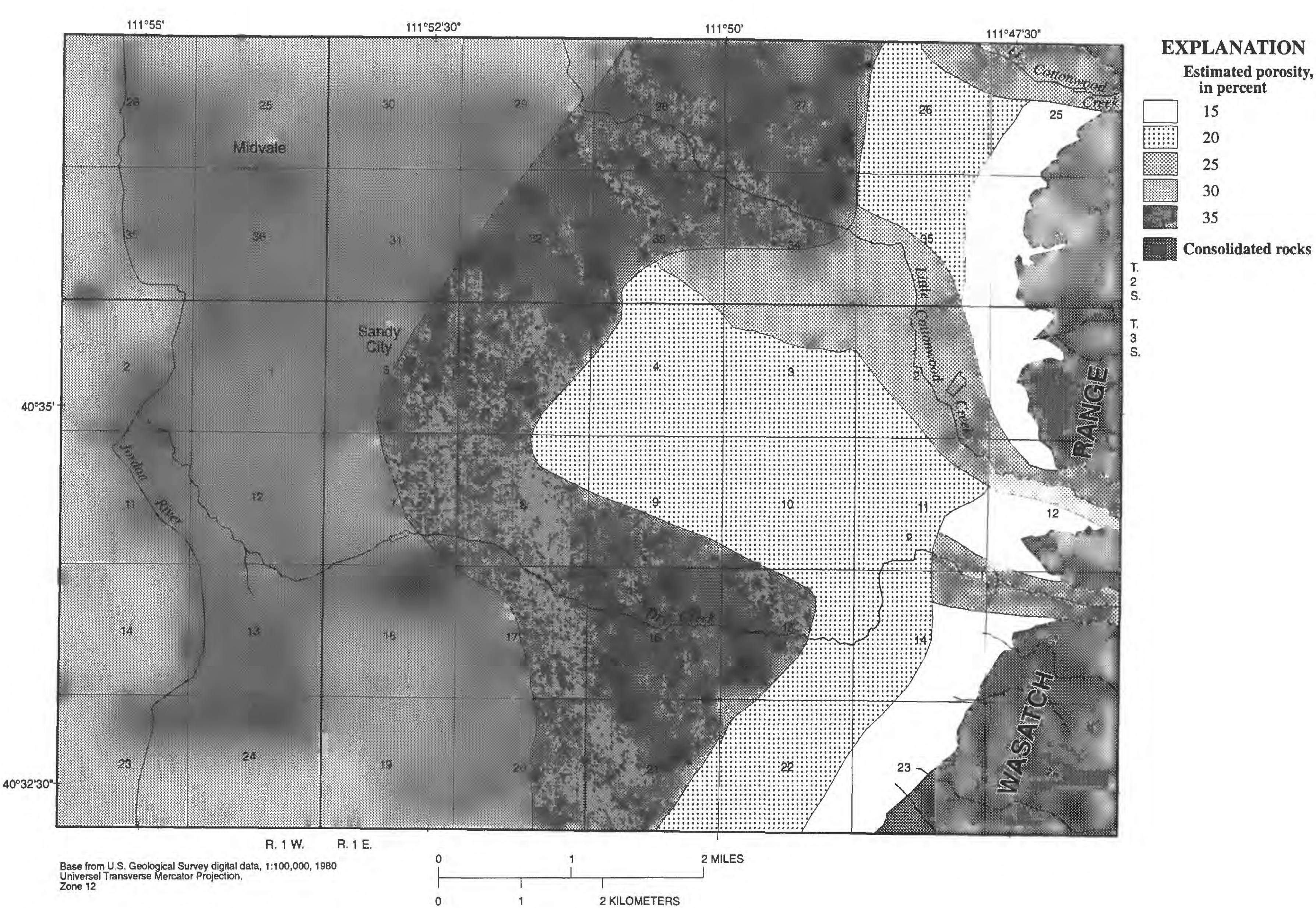


Figure 7. Generalized distribution of estimated porosity for the principal aquifer in Salt Lake Valley.

Table 3. Wells used to measure water levels in Salt Lake Valley

[—, no data]

Map number: See figs. 8 and 9.
Local well number: See text for explanation of numbering system for wells.
Land-surface altitude: In feet above sea level.
Well depth: Depth of casing below land surface.
Water use: P, public supply; I, irrigation; U, unused; H, domestic; O, observation.
Perforated interval: In feet below land surface; numbers refer to gross interval.
Water levels: Altitudes rounded to nearest foot.
Other available information: G, gamma-ray log, R, resistivity log, P, aquifer test (pumped well), Rp, aquifer test (recovery on pumped well), Rm, aquifer test (recovery on observation well), D, well development, L, log suite.

Map number	Local well number	Land surface altitude (feet)	Well depth (feet)	Water use	Perforated interval (feet)	Water levels Winter 1990 Below land surface (feet)	Water levels Winter 1990 Altitude (feet)	Water levels Summer 1990 Below land surface (feet)	Water levels Summer 1990 Altitude (feet)	Other available information
1	(C-2-1)25ddb-1	4,386	475	P,I	150-315	85.0	4,301	137.9	4,248	
2	(C-2-1)25ddb-2	4,386	782	P	588-767	92.0	4,294	113.0	4,273	G,R
3	(C-2-1)35add-1	4,347	244	U	230-244	—	—	65.6	4,281	
4	(C-2-1)36bac-1	4,362	701	P	200-545	72.0	4,290	89.0	4,273	
5	(C-2-1)36cdd-1	4,356	240	P	155-230	81.0	4,275	93.0	4,263	
6	(C-2-1)36ddb-1	4,388	605	P	200-600	105.5	4,283	128.0	4,260	G
7	(C-3-1) 1cca-1	4,360	800	P	319-750	51.0	4,309	56.8	4,303	G,R
8	(C-3-1) 1dca-1	4,395	165	I	125-135	61.0	4,334	65.5	4,330	
9	(C-3-1)12cad-1	4,375	178	H	155-178	58.6	4,316	61.2	4,314	
10	(C-3-1)13bac-1	4,377	124	U	100-124	66.0	4,311	66.9	4,310	
11	(D-2-1)25bbd-2	4,800	433	P	278-422	277.0	4,523	271.0	4,529	
12	(D-2-1)27ccc-1	4,590	445	U	268-445	240.0	4,350	251.6	4,338	
13	(D-2-1)28ccc-1	4,560	691	P	515-678	235.0	4,325	265.0	4,295	P,G
14	(D-2-1)28dbb-1	4,525	671	P,U	430-670	188.2	4,337	220.8	4,304	
15	(D-2-1)29cbb-1	4,460	525	I	442-522	133.0	4,327	173.9	4,286	
16	(D-2-1)29cdd-1	4,497	900	P	470-896	177.0	4,320	243.0	4,254	Rp
17	(D-2-1)30dbd-1	4,412	537	P	242-532	100.0	4,312	129.5	4,283	
18	(D-2-1)30dda-1	4,456	1,002	P	560-990	140.0	4,316	175.0	4,281	Rp,G
19	(D-2-1)31bdc-2	4,425	130	H	120-130	99.5	4,326	104.8	4,320	
20	(D-2-1)32dbd-1	4,560	775	P	290-740	229.0	4,331	254.5	4,306	
21	(D-2-1)32add-1	4,623	685	P	382-635	298.0	4,325	323.0	4,300	G
22	(D-2-1)32bdd-3	4,517	881	P	475-870	194.0	4,323	234.0	4,283	P,Rp
23	(D-2-1)32cbb-1	4,488	1,007	P	475-996	165.0	4,323	208.0	4,280	P,Rp
24	(D-2-1)33aba-1	4,569	904	P	405-904	251.0	4,318	281.0	4,288	Rm,P
25	(D-2-1)33add-3	4,640	585	P	300-560	285.0	4,355	—	—	

Table 3. Wells used to measure water levels in Salt Lake Valley—Continued

Map number	Local well number	Land surface altitude (feet)	Well depth (feet)	Water use	Perforated interval (feet)	Water levels: Winter 1990: Below land surface (feet)	Winter 1990: Altitude (feet)	Summer 1990: Below land surface (feet)	Summer 1990: Altitude (feet)	Other available information
26	(D-2-1)33dca-2	4,740	724	P	396-715	378.0	4,362	394.5	4,346	
27	(D-2-1)34acb-1	4,640	326	U,O	286-306	292.0	4,348	—	—	G,Rm
28	(D-2-1)34acb-2	4,650	802	U	480-730	—	—	328.5	4,322	
29	(D-2-1)34dbb-1	4,677	700	P	285-365	319.0	4,358	344.0	4,333	P,Rp,G,R
30	(D-2-1)35dbc-1	4,867	585	H,I	385-585	520.0	4,347	527.5	4,340	
31	(D-3-1) 2ccc-1	4,960	1,007	U,O	525-990	567.0	4,393	580.5	4,380	G,R
32	(D-3-1) 3bba-1	4,738	875	P	435-762	386.0	4,352	418.0	4,320	
33	(D-3-1) 4bbb-1	4,680	904	P	406-840	344.0	4,336	384.0	4,296	G,R
34	(D-3-1) 4dbc-1	4,720	938	P	500-931	390.0	4,330	426.0	4,294	
35	(D-3-1) 4dac-1	4,747	998	P	540-998	415.5	4,332	449.0	4,298	D
36	(D-3-1) 5acd-1	4,600	656	P	518-650	316.0	4,284	349.0	4,251	G,R
37	(D-3-1) 5bda-1	4,570	600	P,U	486-600	246.3	4,324	276.1	4,294	
38	(D-3-1) 6aab-1	4,500	915	P	390-895	191.5	4,309	221.0	4,279	
39	(D-3-1) 6bad-1	4,475	442	P	280-410	140.0	4,335	163.0	4,312	
40	(D-3-1) 6cba-1	4,460	430	P	240-420	—	—	160.0	4,300	
41	(D-3-1) 6dad-1	4,516	1,000	P	473-1,000	233.5	4,283	244.0	4,272	D
42	(D-3-1) 7ccc-1	4,420	461	P	179-461	162.0	4,258	164.5	4,256	
43	(D-3-1) 8aba-1	4,615	531	P	243-531	285.0	4,330	316.0	4,299	P
44	(D-3-1) 8acc-2	4,550	723	P	369-721	231.0	4,319	267.5	4,283	
45	(D-3-1) 8bba-1	4,538	915	P	780-880	253.0	4,285	—	—	
46	(D-3-1) 8bcb-1	4,520	598	P,U	381-586	189.2	4,331	220.4	4,300	G
47	(D-3-1) 8bdd-1	4,548	657	P	588-598	230.0	4,318	268.5	4,280	
48	(D-3-1) 9aab-1	4,782	950	P	570-925	447.4	4,335	—	—	
49	(D-3-1) 9bba-1	4,792	1,015	P	627-995	417.5	4,375	438.0	4,354	D
50	(D-3-1)10dcc-1	4,970	1,510	P,U	543-1,136	—	—	587.1	4,383	D,L
51	(D-3-1)15cca-1	4,940	640	P	520-640	515.0	4,425	516.0	4,424	
52	(D-3-1)16dca-1	4,797	701	I	415-700	402.0	4,395	—	—	G,R
53	(D-3-1)17aac-1	4,560	1,189	P	300-1,000	207.0	4,353	—	—	
54	(D-3-1)17abb-1	4,577	515	P	205-498	190.8	4,386	—	—	
55	(D-3-1)18cba-1	4,414	1,150	U,O	252-1,150	79.6	4,334	84.0	4,330	
56	(D-3-1)18ddc-1	4,527	165	I	160-165	128.8	4,398	130.6	4,396	
57	(D-3-1)19ada-1	4,540	177	I	165-173	138.2	4,402	139.8	4,400	
58	(D-3-1)19bbd-2	4,450	326	I	226-306	107.2	4,343	111.3	4,339	
59	(D-3-1)21aaa-1	4,800	497	H	420-480	—	—	393.0	4,407	

ever, the magnitude of possible error should not introduce significant inaccuracies into a potentiometric-surface contour map with a contour interval of 25 feet.

After measuring water levels, the configuration of the potentiometric surface was obtained by plotting the altitudes of the water levels on a map and hand contouring the potentiometric surfaces for the two measurement periods (figs. 8 and 9). From the two potentiometric-surface contour maps, average hydraulic gradient along a given flow line can be derived by dividing the difference in hydraulic head by the distance between two points on that flow line. To illustrate the areal and temporal variability in hydraulic head gradient that is likely within the study area, values for slope of the potentiometric surface were generated by a computer program using bivariate quintic interpolation between line values represented by potentiometric contour lines and point values represented by water levels at wells. The program divides the potentiometric surface into triangular shapes and interpolates a slope and aspect for each triangle. Because these calculated average slopes may or may not represent the slope along a ground-water flow line, this and other computer-generated slope values should never be used for determining average linear ground-water velocity values. Figures 10 and 11 are presented only to show the approximate variations in slope of a potentiometric surface that are possible and how that slope can change as seasonal stress changes.

Two distributions (late winter and late summer 1990) of slope values for the potentiometric surface were derived on the assumption that no vertical hydraulic-head gradients exist between the control points or the potentiometric contours. Given the variability in the screened intervals of the wells used as control points, this assumption is probably correct only for the transition zone between the recharge and discharge areas. Ideally the potentiometric surface should be derived from water levels obtained from wells finished in the same part of the principal aquifer. The computer-generated slope values for late winter 1990 ranged from about 5 to 430 feet per mile, and for late summer 1990 from 5 to 520 feet per mile. The slope of the potentiometric surface for the late winter period was less than 50 feet per mile over a substantial area along the north boundary of the study area (fig. 10). The size of this area had decreased substantially by the time water levels were measured again in late summer 1990. From late winter to late summer 1990, pumping increased in the central and southwest parts of the study area. This resulted in steepening the slope of the potentiometric surface over a large area around the pumping wells (fig. 11) and broadening the zones of capture for the main pumping wells.

The distribution of values for the slope of a potentiometric surface along a given flow line can be easily derived with hand calculations. These hand calculations were made for the segments along a flow line on the potentiometric-surface map for late winter 1990, which extends from near the mountain front to well 42 (fig. 12). Assuming the aquifer is isotropic, flow lines were drawn perpendicular to the potentiometric contours. Slope of the potentiometric surface changes as water levels fall or rise in response to pumping or the cessation of pumping; thus, a slope value is valid only for the time it was actually calculated.

The distance traveled along a given flow line for any given time is determined by accumulating the individual travel times for smaller complete flow-line segments radiating from a well until the given travel time is exceeded. The length of the flow line is equal to the sum of the lengths of the complete flow-line segments that were required plus the length of the last partial flow line within which the given travel time was exceeded. The length of that partial flow-line segment to be included in the total length of the flow line is determined by multiplying the average linear velocity for that last segment by the travel time remaining after the travel time through the complete flow-line segments is subtracted from the given travel time. The limited number of wells from which water levels could be obtained prevented a detailed definition of the coalescing cones of depression created by pumping numerous wells with adjacent zones of capture.

CALCULATION OF AVERAGE LINEAR VELOCITY AND TRAVEL TIME TO A WELL

On the basis of the contour map of hydraulic-conductivity values (fig. 6), the porosity values (fig. 7), and the slope of the potentiometric surface along approximate flow lines derived from potentiometric-surface contours for late winter and late summer 1990, the average linear velocity of ground water moving from the area near the mountains to the Jordan River ranges from 0.06 to 144 feet per day. The median velocity is about 3 feet per day. The velocity is greatest in the areas with the steepest slope values near the Wasatch Range, near pumping wells, or in areas where hydraulic conductivity is large, such as the drainage of Little Cottonwood Creek and south of Dry Creek near the mountain front.

The increase in pumping from the winter to the summer of 1990 had a small but noticeable effect on average linear velocity of ground water. The size of the area where velocity was greater than 2.5 feet per day during the winter of 1990 increased by about 8 percent by the end of the late summer pumping period. The dis-

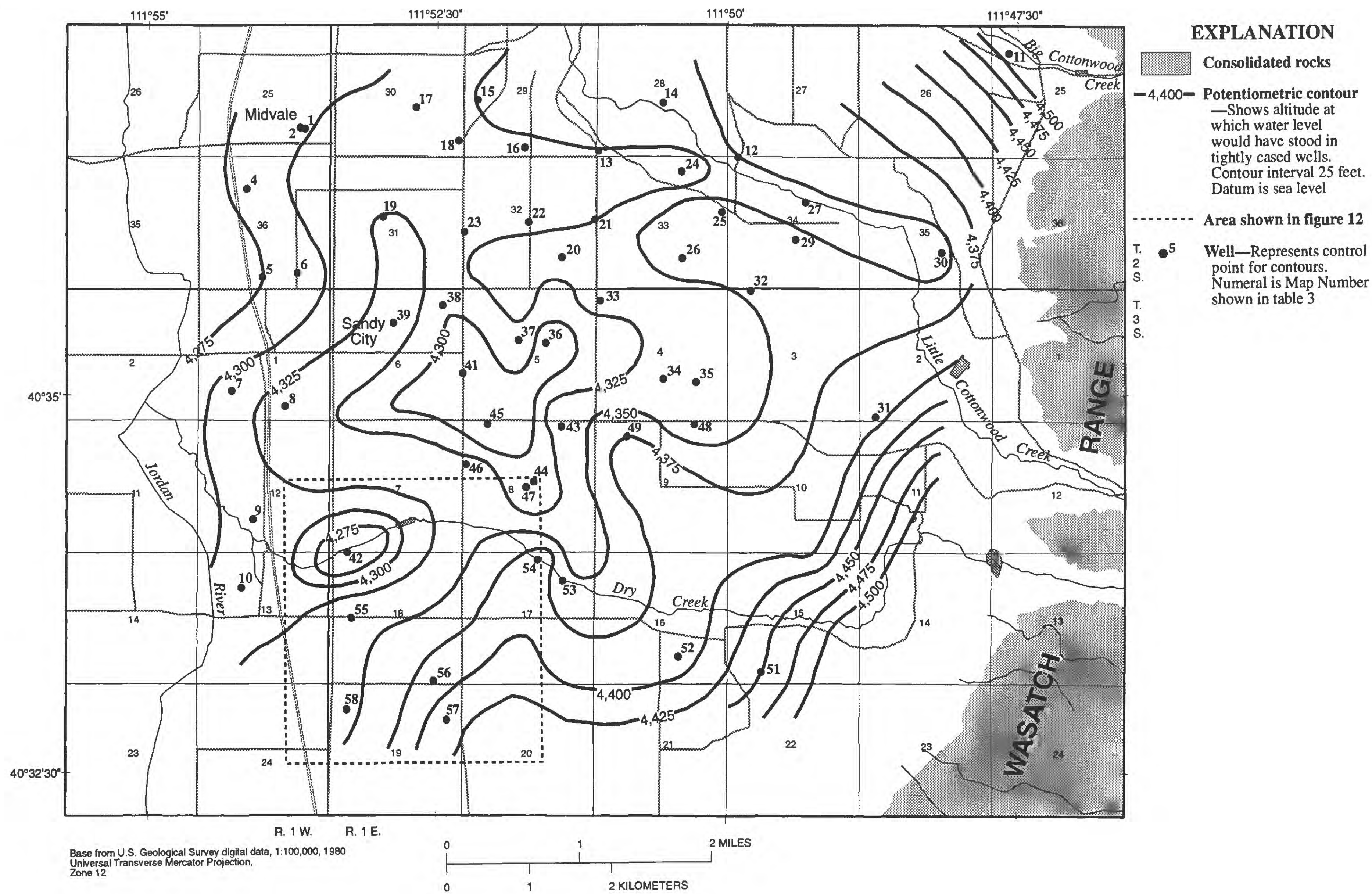


Figure 8. Potentiometric surface for late winter 1990.

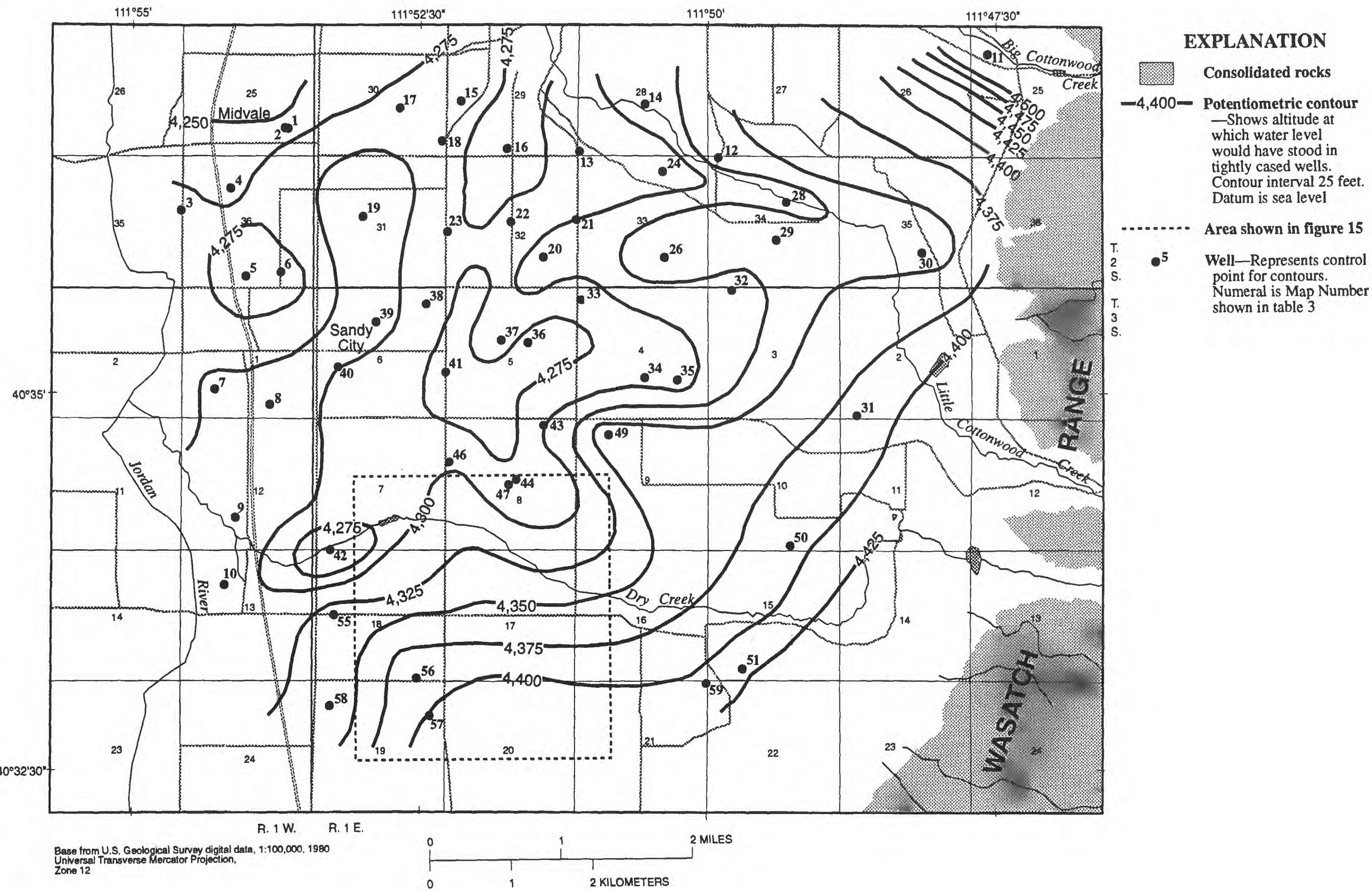


Figure 9. Potentiometric surface for late summer 1990.

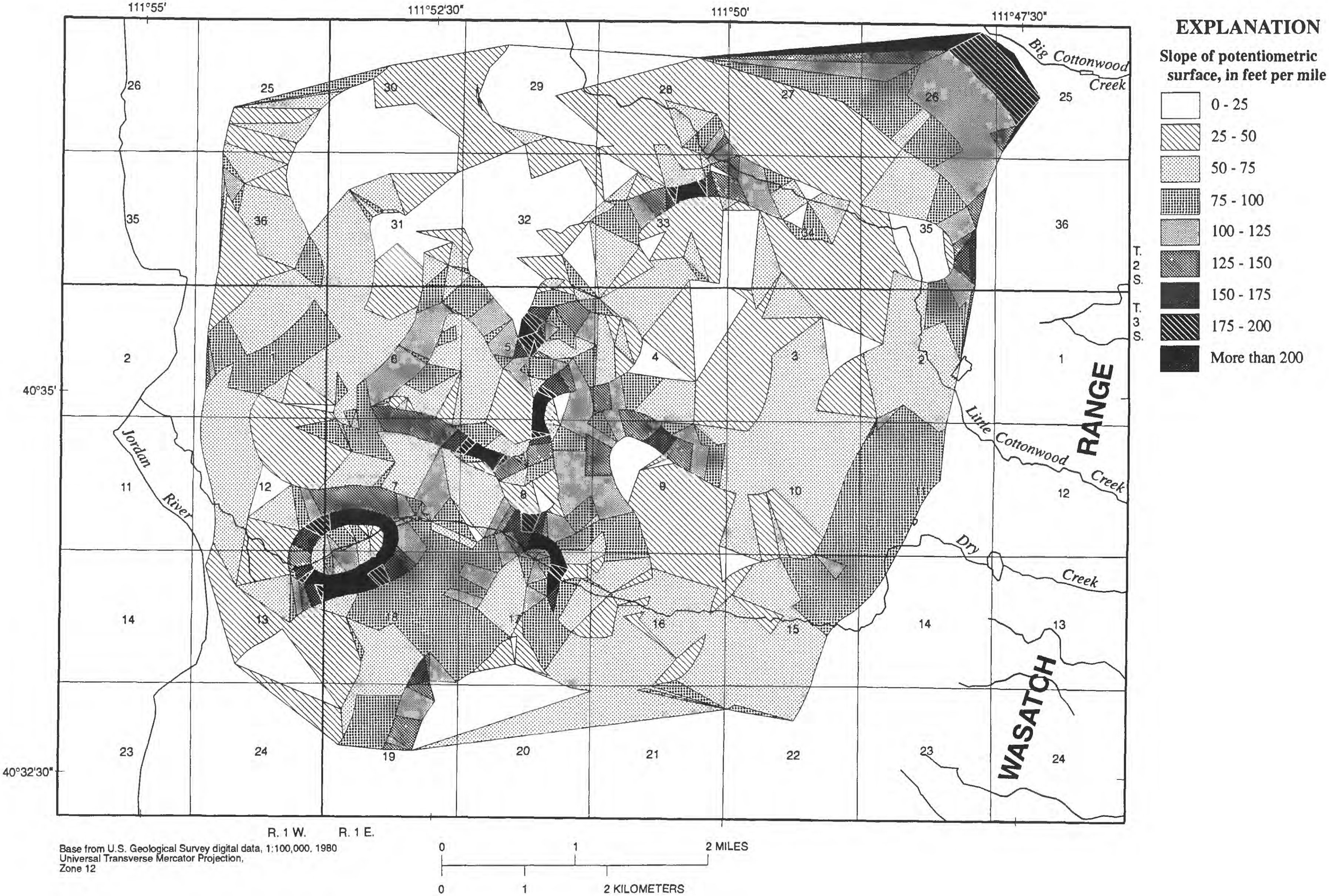


Figure 10. Computer-generated distribution of the slope of the potentiometric surface for late winter 1990 (direction of flow can be implied from fig. 8).

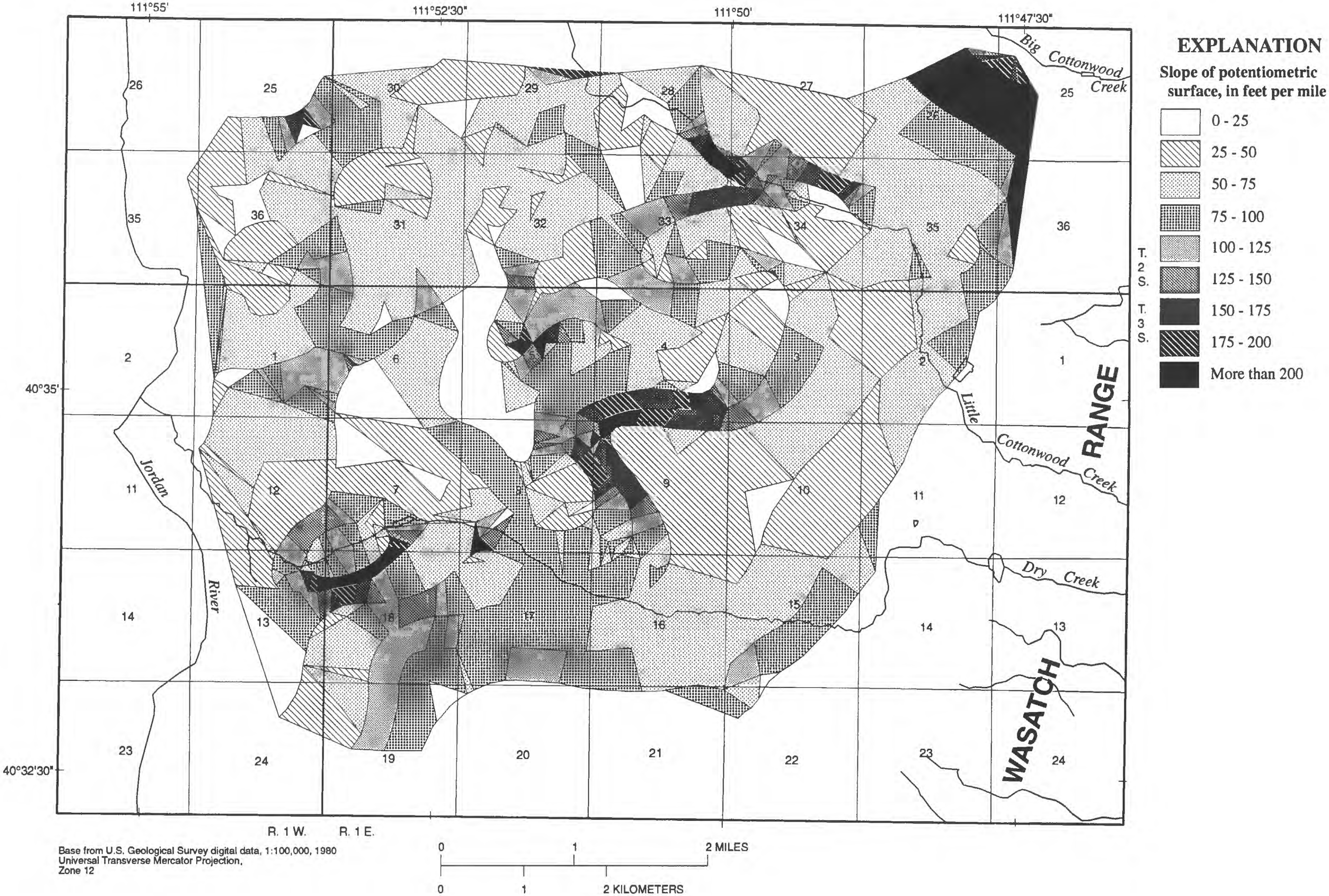
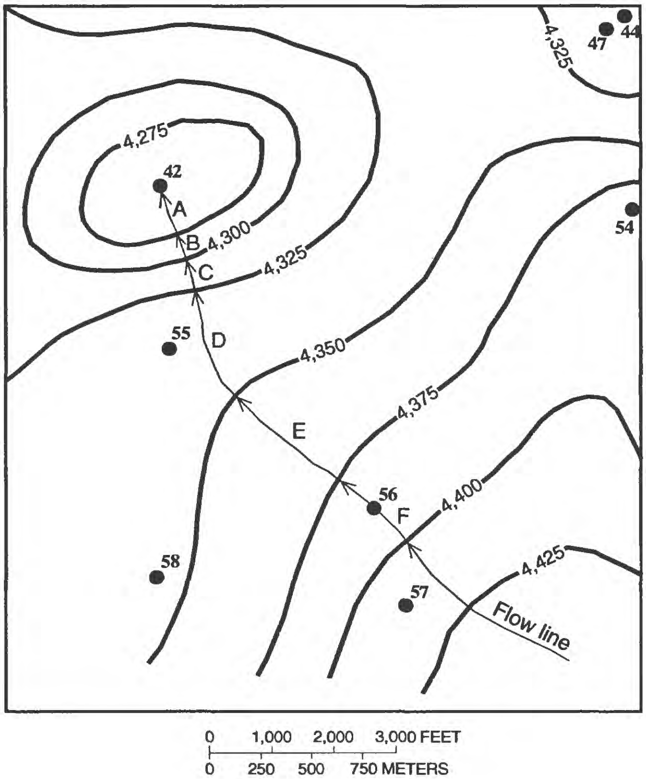


Figure 11. Computer-generated distribution of the slope of the potentiometric surface for late summer 1990 (direction of flow can be implied from fig. 9).



EXPLANATION

—4,300— **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells in late winter 1990. Contour interval 25 feet. Datum is sea level

←E— **Flow line**—Arrows show direction of flow. Letter is flow-line identification in calculation table (below)

● 58 **Well**—Number refers to map number in table 3

Calculation table

Segment identification	Difference in head (dh) (feet)	Length of segment (dl) (feet)	Hydraulic-head gradient (dh/dl)
A	17	825	.021
B	25	430	.058
C	25	530	.047
D	25	1,870	.013
E	25	2,130	.012
F	25	1,480	.017

Figure 12. Example of calculation of hydraulic-head gradient for segments along part of a ground-water flow line for winter 1990 (location of area shown in figure 8).

tribution of average linear velocity had no apparent relation to the distribution of porosity values.

The areal distribution of average linear velocity in the study area (figs. 13 and 14) was derived by integrating the components in equation 1, and is greatly generalized. As a result of this generalization and because the distributions are time dependent, the maps are not meant to be used for future determinations of site-specific wellhead protection areas. These maps are meant to be used only as a guide to indicate the variability that is likely in a typical basin-fill aquifer and to show the general variation between high and low water-level periods during a typical year.

Calculations of average linear velocity of ground water along a single flow line to a well were made from summer 1990 data (fig. 15).Total calculated travel time along the approximately 2-mile flow line was about 11 years. Each segment of the flow line will have a different set of values to be used in the equation, which will result in a slightly different average linear ground-water velocity. Average linear velocity will usually be fastest in the area nearest to a pumping well because of the steep slope of the potentiometric surface created by the cone of depression. The cone of depression near the pumped well is not usually well defined because of the lack of nearby observation wells. These fast average linear-velocity values near wells might be an important part of determining a protection area that is defined by a 250-day travel time to the well. Identification of these faster velocities near wells will require measuring water levels in adjacent observation wells.

Travel time along a ground-water flow line typically is determined from the well back toward the recharge area. Travel time is determined for each segment of the flow line that is characterized by a different average linear velocity. An example of the distances from a well that represent the 250-day and 15-year travel times in a confined aquifer if the well is not pumping is shown in figure 16A, and an example of the same distances if the well is surrounded by a cone of depression caused by pumping is shown in figure 16B. The assumptions and properties used to estimate these distances are (1) that pumping is from a nonleaky confined aquifer with a uniform hydraulic conductivity of 50 feet per day, a transmissivity of 13,000 feet squared per day, a porosity of 30 percent, and a storage coefficient of 0.0001; (2) that the initial hydraulic-head gradient with no pumping is a uniform 33 feet per mile; (3) that a constant rate of discharge of 1,500 gallons per minute took place for 6 hours; (4) that the aquifer recovers fully after pumping ceases; and (5) that there is no vertical ground-water movement. Values for hydrologic properties approximate conditions in the study area, but the assumptions of nonleaky conditions, fully penetrating wells, no vertical flow, and complete aquifer recovery after pumping stops are probably not representative of the study area. Figure 16 is presented only to demonstrate the possible variation in distance that ground water can travel in 250 days and in 15 years, as a result of a cone of depression around a pumped well.

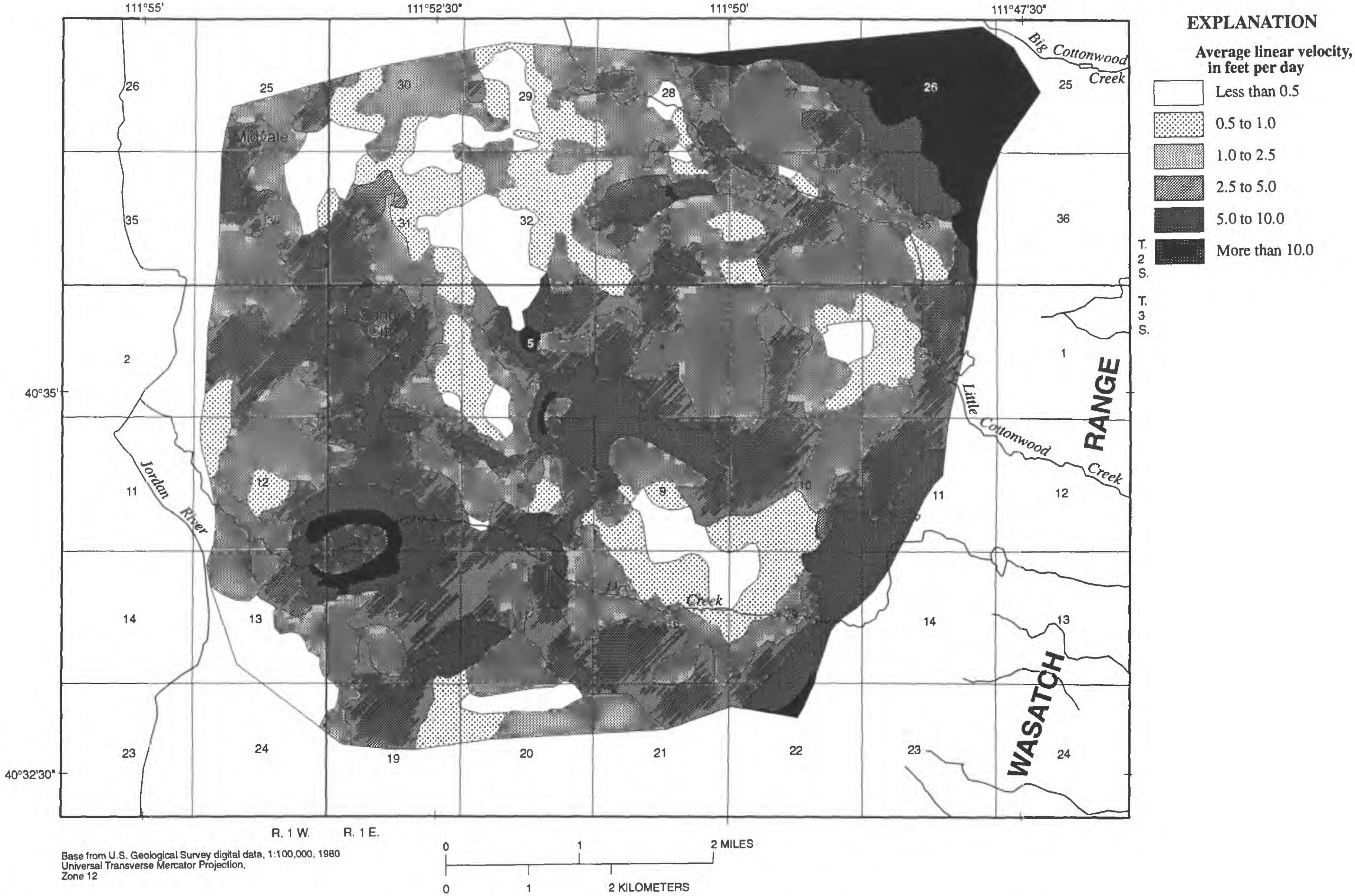


Figure 13. Generalized distribution of average linear velocity of ground water in the principal aquifer in Salt Lake Valley for late winter 1990.

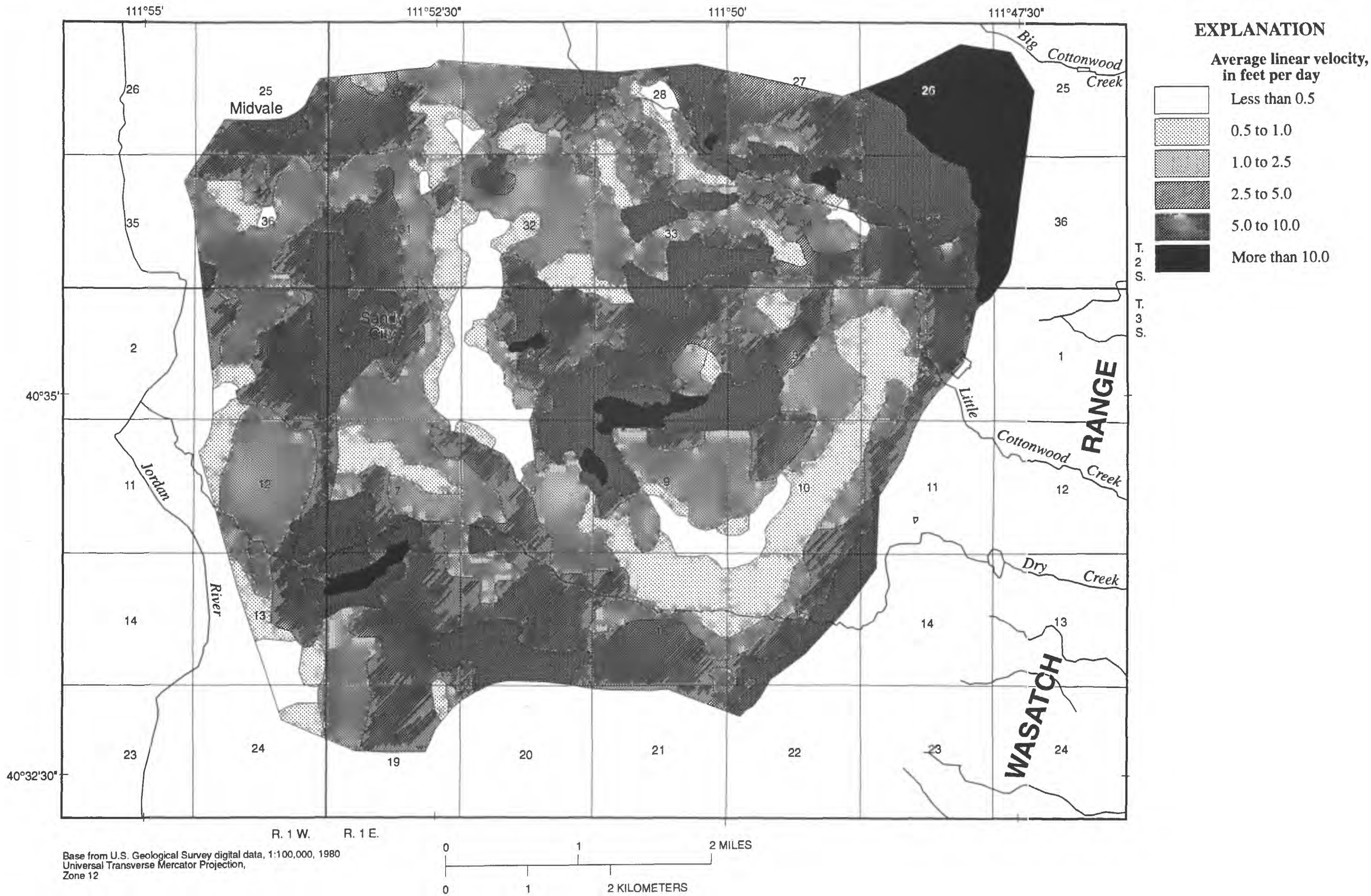
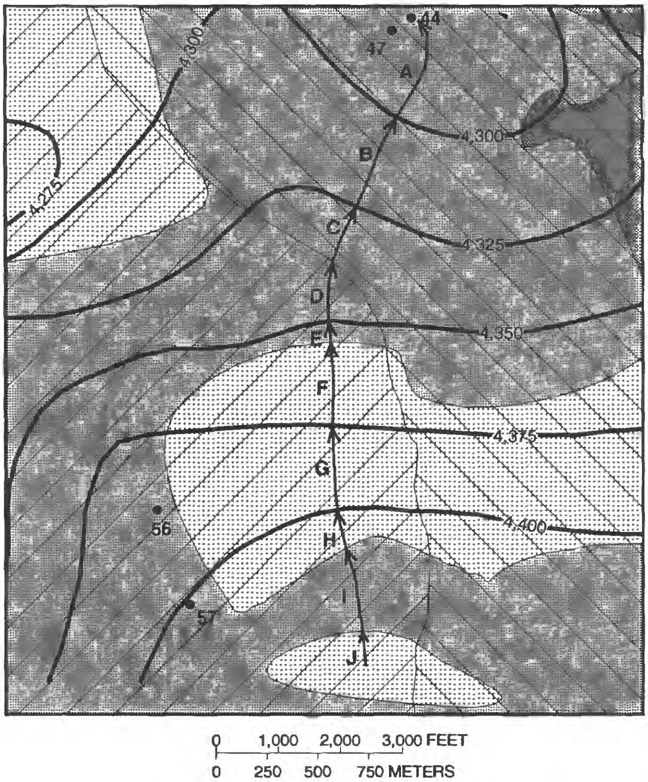


Figure 14. Generalized distribution of average linear velocity of ground water in the principal aquifer in Salt Lake Valley for late summer 1990.



EXPLANATION

Hydraulic conductivity, in feet per day

80 to 120

40 to 80

0 to 40

Porosity, in percent

35

30

—4,300— **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells in late summer 1990. Contour interval 25 feet. Datum is sea level

J **Flow line**—Arrows show direction of flow. Letter is flow-line segment identification in calculation table (below)

●47 **Well**—Number refers to Map Number in table 3

Calculation table

Segment identification	Hydraulic conductivity (feet per day)	Porosity (percent)	Difference in head (feet)	Length segment (feet)	Hydraulic-head gradient	Average linear velocity (feet per day)	Travel time per segment (days)	Travel time per segment (years)	Cumulative time (years)
A	60	35	17	1,666	.0102	1.75	952	2.61	2.61
B	60	35	25	1,588	.0157	2.70	588	1.61	4.22
C	60	35	25	973	.0129	2.20	442	1.21	5.43
D	60	30		960	.0129	2.60	369	1.01	6.44
E	60	30	25	335	.0148	2.95	114	0.31	6.75
F	100	30		1,352	.0148	4.90	276	0.76	7.51
G	100	30	25	1,366	.0183	6.10	224	0.61	8.12
H	100	30		631	.0097	3.25	194	0.53	8.65
I	60	30	25	1,359	.0097	1.95	697	1.91	10.56
J	100	30		574	.0097	3.25	177	0.48	11.04

Figure 15. Example of calculation of travel time for segments along part of a ground-water flow line for summer 1990 (location of area shown in figure 9).

LIMITATIONS IN APPLICATION OF METHODS AND USE OF RESULTS

The methods used rely on hydrologic and geologic data already available or data that could be easily obtained. Computer techniques were used to process the data for this report, but because of the associated costs and training required, hand-held calculators for calculations may likely be more cost effective for municipalities to estimate travel times to wells.

The methods described in this report are most applicable to an unconfined aquifer that is exposed to possible contamination from surface spills of chemicals. Determining travel time of ground water in a confined aquifer would be inappropriate if the overlying confining layer was thick and laterally extensive enough to impede the vertical movement of contaminants introduced at land surface to the aquifer. On the other hand, using the methods described in this report would be appropriate if the goal was to estimate travel time of ground water from the recharge area of a confined aquifer, if there was a possibility of contamination being introduced around the annulus of a well penetrating the confining layer, if contaminants could be introduced through forced injection through wells, or if the confining layer was leaky or laterally discontinuous and the direction of vertical ground-water movement was downward through that layer. The principal aquifer in the Salt Lake Valley could be in any of these categories, although if travel time during downward movement through a leaky or discontinuous confining layer were to be estimated, data on head gradients and hydraulic conductivity in the vertical dimension would be needed. The principal aquifer is unconfined near the mountain front, overlain by a leaky, discontinuous confining layer with a downward vertical direction of ground-water movement in areas between the mountain front and the center of the valley, and confined with an upward vertical direction of ground-water movement in the center part of the valley. For the purposes of this study, only horizontal ground-water movement in the principal aquifer is considered in calculating travel time.

Average linear velocity is not an actual velocity for a particle of water. The actual velocity of a water particle has a large range of values; thus, using an average velocity to determine the travel time of a contaminant in the ground water can be misleading, especially if the segment of the flow line for which the calculation is being made represents a large range in hydraulic conductivity and/or slope of the potentiometric surface. In general, the best predictions for travel times would result from using short segments of the ground-water flow lines in narrow zones of the aquifer along which

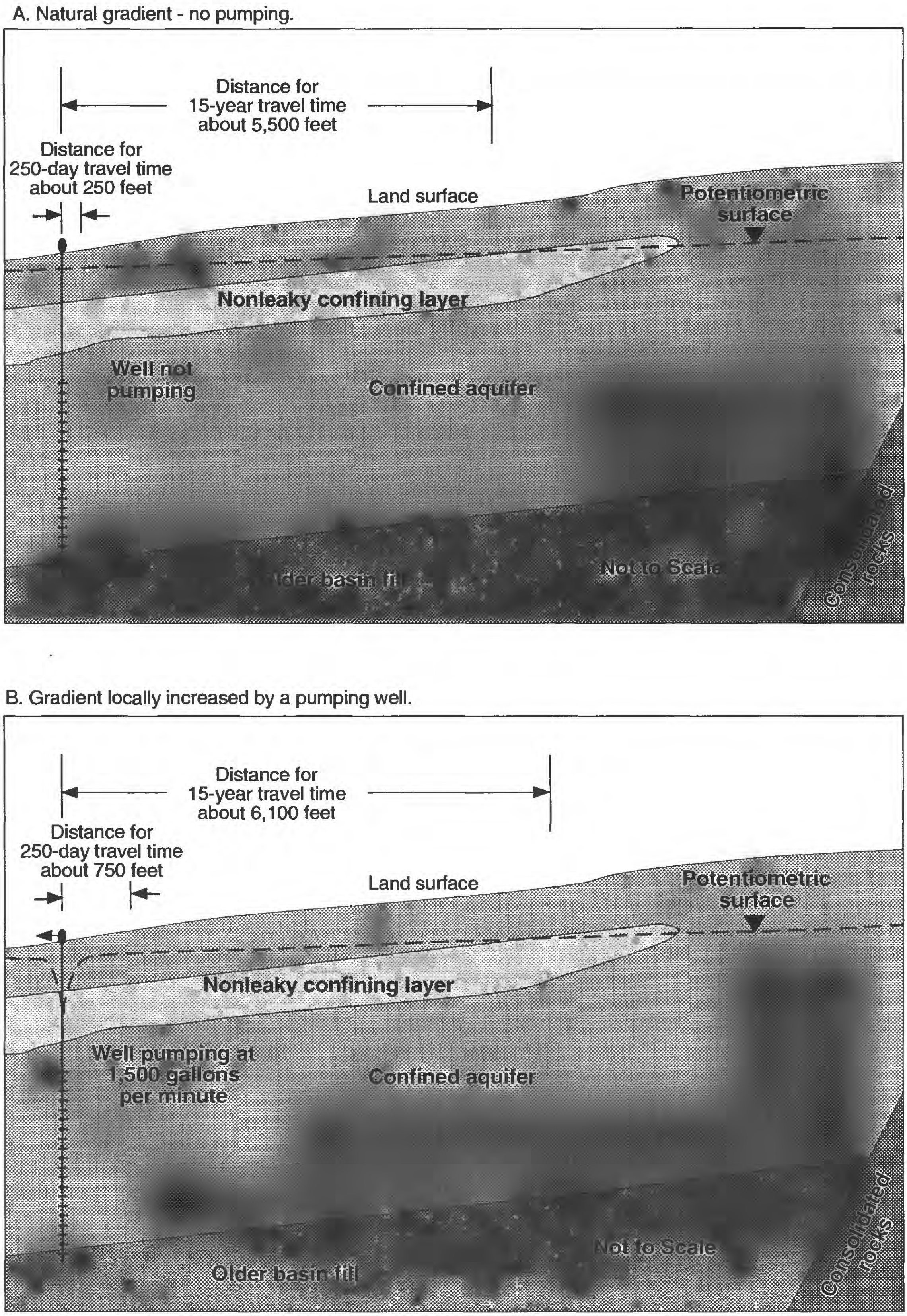


Figure 16. Schematic diagrams showing distances representing 250-day and 15-year ground-water travel times in a confined aquifer under (A) a natural gradient, and in the same aquifer where (B) the gradient is locally increased by a pumping well.

and within which the hydrologic properties and head gradients are explicitly defined.

The reliability of calculated average linear-velocity values depends largely on the ability of the investigator to correctly interpret the maps and conclusions of previous investigations and to transform these interpretations into valid concepts that result in the most reasonable estimates of hydrologic properties for a particular study area. Determining the magnitude and areal distribution of these hydrologic properties with great accuracy over an area the size of Salt Lake Valley, or even for a selected well field, is cost prohibitive at this time (1991).

Multiple-well aquifer testing and laboratory analysis of cores are usually considered the most reliable means of obtaining hydraulic conductivity and porosity of an aquifer. For this study, results from aquifer tests done on nine wells were available; however, only five of these could be used for adjusting values assigned to the various sediments described by the drillers. Laboratory analyses were available for similar basin-fill sediments, but not for the deposits in the study area; thus, the degree to which estimated values for these two hydrologic properties are representative of actual values in the study area is not known.

Wells for which water levels were measured were adequately distributed to obtain a good area-wide variation in the slope of the potentiometric surface, but were inadequate to define variations in slope around specific pumping wells within the study area. As a result, the average linear-velocity values near pumping wells are not adequate to delineate accurately the area around that well defined by a ground-water travel time of 250 days or less.

Conceptualization of the hydrologic system and the limiting assumptions associated with that conceptualization would probably be the source of greatest error when attempting to determine a protection area around a well. Vertical and horizontal variations in the lithologic character of an aquifer and a confining layer result in flow lines that are not simply horizontal and always perpendicular to the potentiometric contours drawn on a map. Many different preferential flow lines probably exist along which ground water could move faster than would be indicated by a hydraulic-conductivity value determined from an aquifer test. This would be true if the aquifer test yielded a value that was a result of averaging estimated hydraulic-conductivity values from many layers having slightly different lithologic character.

Because of the limited scope of this investigation, numerous aspects that affect the rate and direction of movement of a contaminant to and in an aquifer were not considered, such as anisotropy of the basin fill, vertical flow in the principal aquifer, the variable thickness of the unsaturated zone overlying the principal aquifer and the travel time through this zone, movement downward through an overlying shallow unconfined aquifer that exists locally in the area, and the variable rate of recharge that probably occurs with distance from the mountains. Other physical and chemical factors that influence movement of contaminants, such as mechanical and molecular dispersion, variable density of the fluid carrying the contaminant, temperature variability, mechanical diffusion, adsorption, or changes in chemical or biological composition with time, also were not considered. In a more comprehensive calculation of total travel time to a well, all of these aspects would have to be evaluated to determine if they would substantially affect the estimate of travel time computed for idealized horizontal fluid movement in a saturated isotropic aquifer.

CONCLUSIONS

The calculation of average linear velocity of ground water moving horizontally in the principal aquifer of the Salt Lake Valley, and the calculation of distances along flow lines that represent travel times of 250 days and 15 years requires that three properties be known: hydraulic conductivity, porosity, and slope of the potentiometric surface. Reasonable estimates for these properties can be obtained through semiquantitative hydrogeologic investigations, but reliability of the results is more questionable when data used to make these estimates are less numerous.

The information necessary to estimate average linear velocity was obtained from interpretive information sources such as geologic maps, water-level maps, and drillers' logs located in the files of private or government agencies and from comparative values measured for other similar aquifer systems. The estimates of values and their lateral distribution as interpreted for the southeastern part of Salt Lake Valley during this investigation are consistent with the current (1991) conceptualization of the hydrologic system by the USGS. Estimated hydraulic-conductivity values ranged from 20 to 250 feet per day. Porosity values ranged from 15 to 35 percent, potentiometric-slope values from 5 to 520 feet per mile, and calculated average linear-velocity values from 0.06 to 144 feet per day. Calculated travel time along one 2-mile part of a flow line was about 11 years.

The method demonstrated in this report can be used to develop travel time along flow lines to a well if supporting data for the site such as additional observation wells, drillers' logs, and aquifer-test results are available. If only limited supporting data are available, reliability of travel-time results needs to be carefully considered before using them to establish a WHPA.

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