GROUND-WATER DISCHARGE AND RECHARGE IN THE SODA LAKES AND UPSAL HOGBACK GEOTHERMAL AREAS, CHURCHILL COUNTY, NEVADA

By F. H. Olmsted

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CONVERSION OF UNITS OF MEASUREMENT

INCH-POUND TO METRIC

Multiply inch-pound units	by	To obtain SI units
	Length	
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.589	square kilometer (km^2)
	Volume	
cubic foot (ft^3)	0.02832	cubic meter (m^3)
	Flow rate	
gallon per minute (gal/min)	0.06308	liter per second (L/s)
	Pressure	
pound per square inch (lb/in^2)	6.895	kilopascal (kPa)
	Temperature	
degree Fahreheit (°F)	°C = 5/9 (°F - 32)	degree Celsius (°C)
	Viscosity (absolute)	
centipoise	0.001	Pascal second (Pa/s)
	Mass	
pound, avoirdupois (lb)	453.6	gram (g)

CONVERSION OF UNITS OF MEASUREMENT (Continued)

INCH-POUND TO METRIC

Heat flow

One heat-flow unit (HFU) =	4.184×10^{-2}	Watts per square meter (W/m^2)

METRIC TO INCH-POUND

Multiply SI units	by	To obtain inch-pound units
	Length	
millimeter (mm)	0.0394	inch (in.)
meter (m)	3.281	foot (ft)
	Mass	
gram (g)	0.002205	pound, avoirdupois (lb)
kilogram	2.205	pound, avoirdupois (lb)
	Thermal energy	
kiloJoule (kJ)	239	calory (cal)
	Temperature	
degree Celsius (°C)	°F = 9/5 °C + 32	degree Fahrenheit (°F)
	Permeability	
darcy (k)	2.43	foot per day (ft/d) at 60°F

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Mean Sea Level." NGVD of 1929 is refered to as "sea level" in this report.

GROUND-WATER DISCHARGE AND RECHARGE IN THE SODA LAKES AND UPSAL HOGBACK GEOTHERMAL AREAS, CHURCHILL COUNTY, NEVADA

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ABSTRACT

The Soda Lakes and Upsal Hogback areas, in the west-central Carson Desert about 80 kilometers east of Reno, Nevada, comprise the upflow or discharge parts of two hydrothermal systems in which thermal fluid rises from depth along steeply inclined conduits believed to be generally fault-controlled. Specific discharge and recharge were investigated as part of an earlier, more comprehensive, study of the Soda Lakes and Upsal Hogback systems in order to assist in deriving an estimate of upflow of thermal fluid into the deposits above a depth of 45 meters in the Soda Lakes system and to estimate the convective component of near-surface heat flow for both systems. In areas of major ground-water discharge, the logarithm of the specific discharge estimated on the basis of measured vertical hydraulic gradient and estimated vertical hydraulic conductivity was found to have a significant correlation with water-table depth. Scatter of the data is caused by errors in the estimates--chiefly in assumed values of vertical intrinsic permeability--and by factors other than depth to water such as variations in specific discharge with depth, and the type of soil, density and types of vegetation, presence or absence of a salt crust, and other surface conditions.

Local recharge in low-lying areas like the west-central Carson Desert, generally assumed to be almost negligible in earlier studies, probably is significant where (1) intense storms occur in areas of shallow water table, (2) irrigation and canal leakage maintain high soil-moisture levels, and (3) runoff is concentrated in surface-water bodies such as lakes, ponds, or stream channels. Clear evidence of the first and third types of recharge was obtained in the study, but the magnitude of such recharge in terms of a water budget for a large area such as the Carson Desert cannot be assessed with data presently at hand.

INTRODUCTION

This report presents the results of an investigation of vertical ground-water flow rates--specific discharge and recharge--that was part of a broader study of the geohydrology, aqueous geochemistry, and thermal regime of the Soda Lakes and Upsal Hogback geothermal systems in west-central Nevada (Olmsted and others, 1984). The Soda Lakes and Upsal Hogback geothermal areas are the upflow or characterized as the discharge parts of hydrologic systems in which thermal fluid rises from depth along steeply inclined conduits believed to be generally fault-controlled. Instead of emerging at the land surface as thermal springflow; however, the thermal fluid flows laterally toward the northeast and north through aquifers at varying depths in unconsolidated deposits and volcanic rocks of late Tertiary and Quaternary age. In the earlier, broader study, the specific-discharge and specific-recharge data were used to (1) assist in deriving an estimate of upflow of thermal fluid into the deposits above a depth of 45 m in the Soda Lakes system, and (2) estimate the convective component of near-surface heat flow for both systems (Olmsted and others, 1984, p. 66). In addition, data acquired after a series of storms in the winter of 1982-83 demonstrated the significance of local ground-water discharge resulting from infrequent filling of intermittent ponds. It is hoped that the data and interpretations presented herein will assist in designing studies of ground-water discharge and recharge, both in the Carson Desert and in similar desert basins in the northern Basin and Range province.

GEOHYDROLOGIC SETTING

The Soda Lakes and Upsal Hogback geothermal areas are in the west-central Carson Desert, about 80 kilometers (km) east of Reno, Nevada (fig. 1). The Carson Desert is a large intermontane basin in the northwestern Basin and Range province, a region characterized by thin crust, major east-west crustal extension, block faulting, and high regional heat flow (Thompson and Burke, 1974; Eaton and others, 1978). The youngest deposits in the Carson Desert are of fluvial and eolian origin; these deposits overlie and are interbedded with lacustrine and deltaic deposits of Lake Lahontan, a large lake that occupied much of northwestern Nevada during pluvial stages of the late Pleistocene and early Holocene. Underlying the Lahontan and post-Lahontan deposits are unconsolidated to semiconsolidated sediments and intercalated basalts of late Tertiary and Quaternary age. This basin fill, which locally exceeds 1 km in thickness, is underlain and surrounded by a wide variety of consolidated sedimentary, igneous, and metamorphic rocks of Mesozoic and Tertiary age; these rocks form the mountainous areas shown in figure 1.

The Carson Desert, like many other intermontane basins formerly occupied by Lake Lahontan, is undrained. Under pre-development conditions, surface inflow was largely from the Carson River, which rises in the eastern Sierra Nevada. During wet cycles, the Humboldt River also provided (and still provides) inflow by way of overflow from Humboldt Sink. (See fig. 1.) Since the advent of irrigation in the early 1900's, inflow from Lahontan Reservoir on the Carson River at the western margin of the Carson Desert has been augmented by water diverted from the Truckee River through the Truckee Canal (fig. 1).

Before irrigation development, recharge to the ground-water system in the Carson Desert was largely by seepage from the Carson River through channel bottoms in its natural distributory system and by overbank flooding. Today, ground-water recharge is largely from infiltration of applied irrigation water and canal leakage. Since extensive irrigation began about 1906, ground-water levels have risen throughout most of the west-central Carson Desert; the rise near Soda Lakes has been as much as 18 m (Rush, 1972). Recharge from surface inflow to the ground-water system of the Carson Desert since 1906 has been estimated to exceed one half of the approximately 480 cubic hectometers per year (hm^3/a) release from Lahontan Reservoir (Olmsted and others, 1975, p. 80).

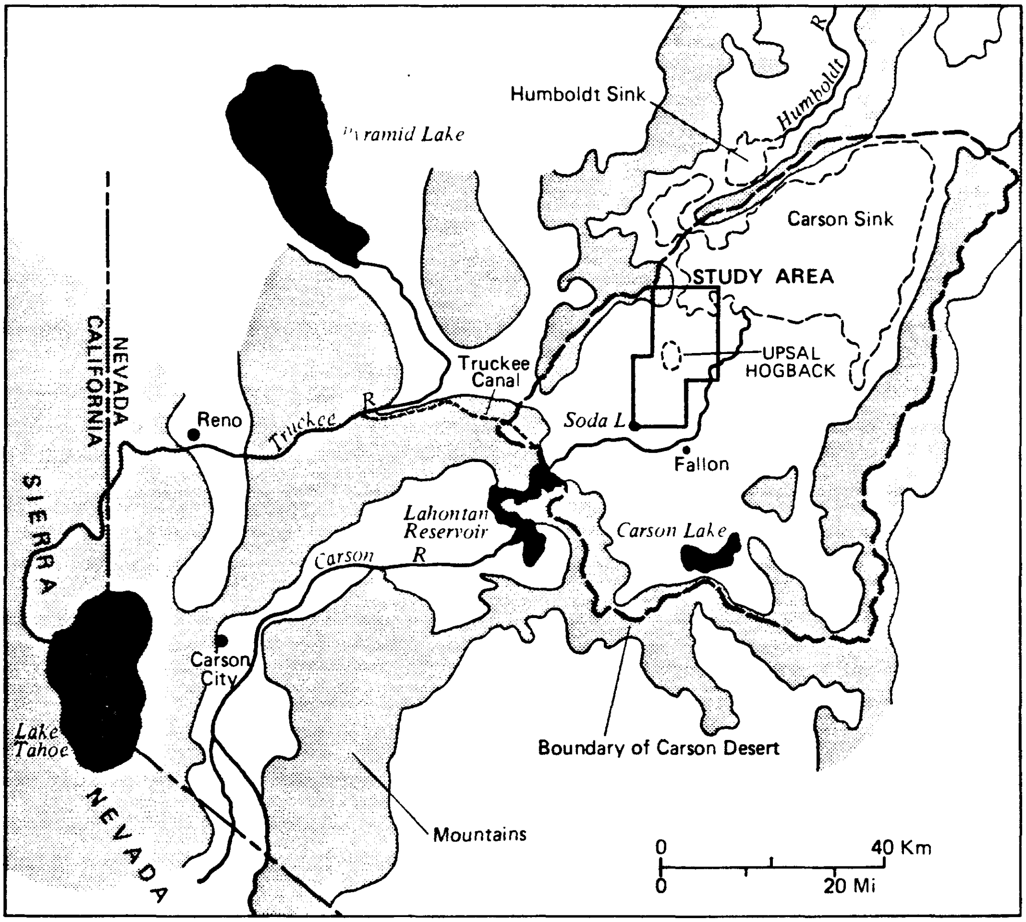

Figure 1. — Index map of the Soda Lakes and Upsal Hogback geothermal areas, Nevada.

Local precipitation on the Carson Desert commonly has been regarded as an almost negligible source of ground-water recharge. Average annual precipitation within the drainage basin ranges from less than 100 millimeters (mm) in the lowest parts of the basin to more than 400 mm on the highest ranges along the margin (Hardman, 1936; Hardman and Mason, 1949). Using an empirical method developed by Eakin and Maxey (1951), Glancy and Katzer (1975) estimated potential recharge annually from local precipitation on the Carson Desert to be 1.6 hm^3/a--this amounts to only 0.2 percent of the average annual precipitation and to only 0.7 percent of the estimated total recharge from surface inflow. However, results of the present study suggest that the method of Eakin and Maxey (1951) may yield too low an estimate because it does not include potential recharge in areas of shallow water table or irrigated areas, nor does it include recharge produced by infrequent meteorological events that cause local flooding and the filling of ponds and lakes.

Confined ground-water conditions prevail in most of the saturated deposits and rocks beneath the west-central Carson Desert. Ground water is locally unconfined where coarse, permeable deposits extend to depths as much as several meters below the water table. Large-scale lateral movement of ground water is toward the Carson Sink, which is largely northeast of the study area, (fig. 1). Sandy aquifers within a few tens of meters of the land surface transmit this water most rapidly. Confined ground water in and near areas of discharge moves upward through confining beds of clay and silt. Ground-water movement is mainly downward in recharge areas, especially in and near irrigated lands.

METHODS OF STUDY

Data for this study were obtained largely by test drilling. These data were supplemented by mapping of vegetation and surface conditions by P. A. Glancy (written commun., 1979). Density and vigor of phreatophytes were evaluated qualitatively in order to estimate ground-water evaporation in the Soda Lakes and Upsal Hogback geothermal areas.

The test drilling included the collection of drill cuttings and core samples for geologic and hydrologic analysis, borehole geophysical logging, and the installation of small-diameter (38- or 51-mm) wells, most of which are 45 m or less in depth. The wells were used for measurements of temperature and water levels (hydraulic heads), and were sampled for hydrochemical analysis. The preparation and analysis of detailed lithologic logs based on interpretation of the borehole geophysical logs, drill cuttings, and core samples, and water-level measurements were of particular importance for the present study.

Two to four wells were installed at most sites in order to determine the altitude of the water table and confined water levels and to measure the vertical component of the hydraulic gradient. Locations of the well sites are shown in figure 2. Many of the wells, drilled in 1972 and 1973, were bored with a power auger; most of the wells drilled in 1974 and later were drilled by the mud-rotary method. The deeper wells of the latter group, particularly those in the low-lying parts of the Upsal Hogback area, were completed by grouting the annulus between the casing and drill-hole walls with neat cement. The cement grout minimizes the possibility for upward or downward water flow outside the well casing and thereby assures more reliable water-level measurements than those obtained in other wells, in which the annulus was backfilled with drill cuttings or surface materials. Most of the cemented wells were completed originally with steel caps at the bottom and filled with water for temperature-gradient measurements. Subsequently, all these wells were perforated near the bottom with explosive charges. This method of casing perforation allowed the collection of water samples and water-level data. All the other wells were completed either with a screen or well point or with saw-cut perforations at or near the bottom of the well.

Water levels in all the test wells were measured at intervals of about 1 to 6 months from December 1973 to September 1983. A pressure gage was used to measure hydraulic head in flowing wells.

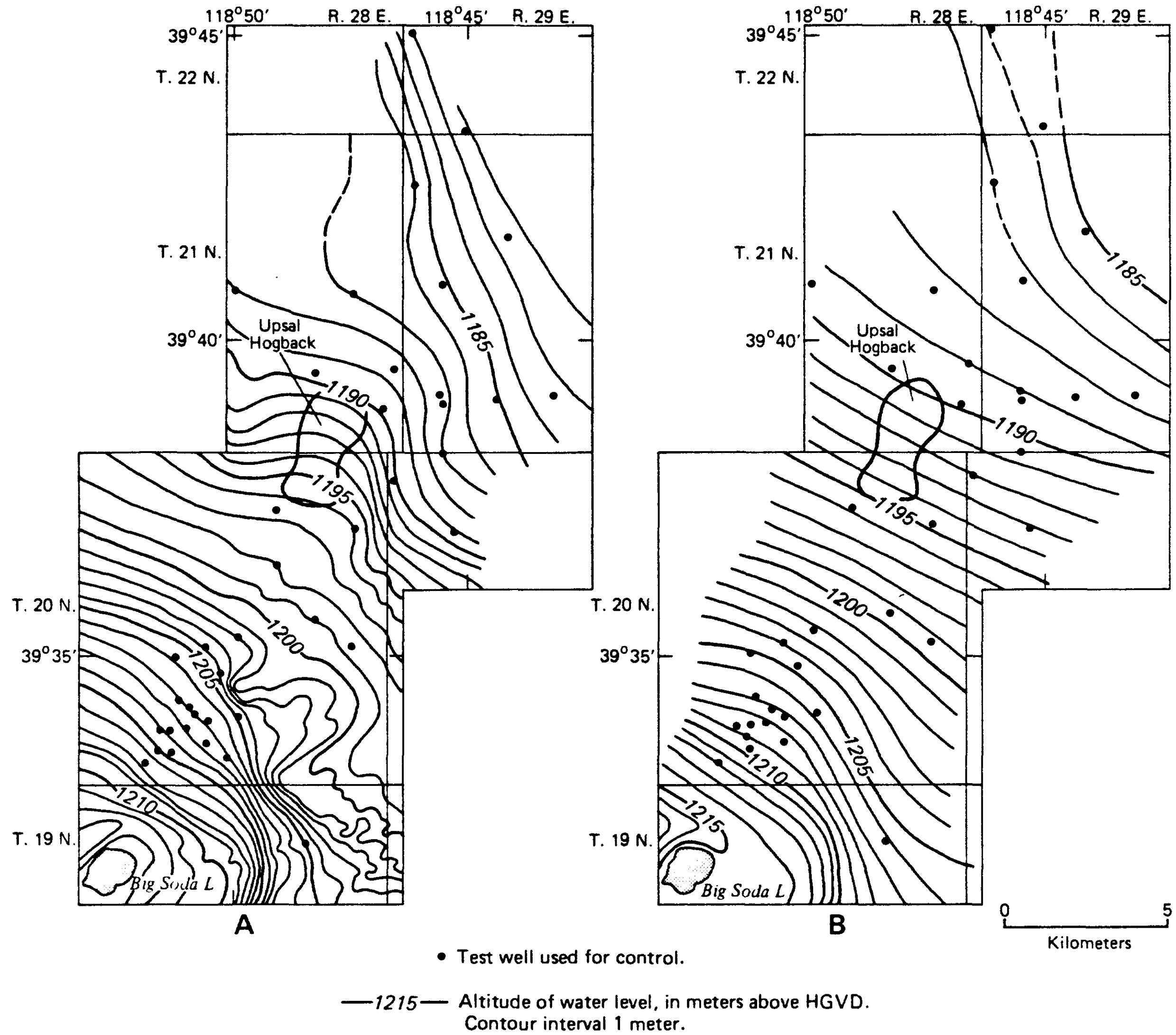

Figure 2. — Soda Lakes and Upsal Hogback geothermal areas.

A. Altitude of water table, December 1979.

B. Altitude of confined water level representing a depth of 30 m, December 1979.

SPECIFIC DISCHARGE OR RECHARGE ESTIMATED FROM HYDRAULIC DATA

The average specific discharge recharge for a given depth interval at a site may be estimated as the product of the vertical hydraulic gradient (or, more correctly, the vertical component of the hydraulic gradient) and the harmonic-mean vertical hydraulic conductivity of the deposits at the temperatures prevailing for the interval. In the present study, the estimates were based on gradients calculated from water levels measured in the test wells and hydraulic conductivities estimated from lithologic and temperature logs of the wells.

The vertical hydraulic gradient was calculated as the difference in depth of water levels in a pair of wells at a site, divided by the difference in depth between the bottom of the screen or perforations in the shallower well and the top of the screen or perforations in the deeper well. Where the depth to the water level in the deeper well was less than that in the shallower well, the gradient was assigned a positive value (upward component); where the reverse was true, the gradient was assigned a negative value (downward component).

At many sites, the shallower or shallowest well is screened or perforated at a substantial depth below the water table, and some degree of confinement is suggested by the presence of beds of clay or silt above the well screen or perforations. At these sites, the altitude of the water table (fig. 2A) was calculated by adjusting the altitude of the water level in the shallowest well, using the vertical hydraulic gradient indicated by measurements in that well and the next deeper well at the site. In a similar fashion, the altitude of the confined water level representing a depth of 30 m, shown in figure 2B, was calculated by adjustment of the measured altitude of water level in the well closest to 30 m in depth at the site, using the vertical hydraulic gradient for the appropriate depth level interval. The configuration of the confined potentiometric surface representing a depth of 30 m shown in figure 2B was used by Olmsted and others (1984, p. 66-75) as a basis for estimating lateral ground-water flow through the deposits above a depth of 45 m.

At some well sites, measured vertical hydraulic gradients were adjusted for the "short-circuit" effect of water flow through the annulus between the well casing and the walls of the hole. The magnitude of this effect is indicated by the ratio of the hydraulic gradient interpolated from the maps of unconfined and confined water levels (figs. 2A and 2B) to the gradient actually measured in the wells at each site; this ratio is termed the "gradient adjustment factor". (See table 2.)

The vertical hydraulic conductivity of a depth interval was computed using the harmonic-mean vertical intrinsic permeability and the weighted-average temperature for the interval. Most depth intervals used in the study comprise several layers of highly variable vertical intrinsic permeability. The harmonic-mean vertical intrinsic permeability, k_n, of a series of n layers is

$$k_n = \frac{z_t}{\sum z_i/k_i} , \qquad (1)$$

where z_i is the thickness of a given layer, k_i is the vertical intrinsic permeability of the layer, and z_t is the total thickness of the layers--the depth interval of interest. It is apparent from the equation 1 that the layers of low permeability have a dominant effect on the harmonic mean. The accuracy of the estimates of specific discharge and recharge is most strongly dependent upon the validity of the values of vertical intrinsic permeability assigned to the least permeable layers in the interval. Also important, however, are the reliability of the lithologic log, particularly with respect to the thickness and character of the confining layers, and the accuracy of the water-level data upon which the calculated vertical gradient is based.

The values of intrinsic permeability assigned to materials classified in the lithologic logs of the test wells are presented in table 1. the values are based on averages obtained for samples of unconsolidated sediments from a variety of locations analyzed by the Hydrologic Laboratory of the U.S. Geological Survey and grouped according to median particle size (Morris and Johnson, 1967). The lowest value in table 1, 0.3×10^{-15} m^2[1/] for clay or silty clay, is the geometric mean of average values of 0.1×10^{-15} m^2 for clay given by Morris and Johnson (1967, table 12) and 1×10^{-15} m^2 for silty clay interpolated from Morris and Johnson (tables 5 and 12). Fine sand and related materials in table 1 were assigned a value of 30×10^{-15} m^2, which actually

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1/ The value 1×10^{-15} m^2 is nearly equal to 1 millidarcy.

is based on the value of 32 X 10^{-15} m^2 given for silt by Morris and Johnson (1967, table 5). This adjustment was made because cores from several test wells in the Soda Lakes and Upsal Hogback areas indicated the presence of thin layers of silt and finer material within thicker sequences of fine sand or related deposits of similar hydrologic character; the fine-grained layers tend to control the vertical permeability of the zones in which they occur. Similarly, the values listed in table 1 for medium sand and coarser materials reflect the presence of thin layers of finer material.

Table 1. -- Values of vertical intrinsic permeability assigned to materials classified in lithologic logs of test wells

Materials	Vertical intrinsic permeability (X10^{-15} m^2)
Gravel and sand; sand and gravel; pebbly sand; coarse sand; medium to coarse sand.	300
Medium sand; fine to medium sand; sand; coarse sand with silt.	100
Fine sand; silty sand; sand and silt; clay and gravel; clay and coarse sand.	30
Silt and fine sand; clay and sand.	10
Silt; sandy clay; clay and fine sand; pebbly clay	3
Clayey silt; silty clay; and fine sand.	1
Clay; silty clay.	.3

Table 2 presents the estimates of specific discharge or recharge (vertical Darcian velocity) based on hydraulic data. The average vertical hydraulic conductivity was derived from average values of vertical intrinsic permeability by adjusting for weighted-average (by depth) temperature, which affects the kinematic viscosity of the water, and converting units from meters squared to millimeters per day. Measurements for a period of several years (1974-82 at most sites) were averaged to obtain estimates of long-term average vertical hydraulic gradients. However, measurements made in 1983, after the significant recharge event described later, were not included in the estimates. The plus-or-minus values in table 2 indicate two standard deviations from the long-term mean values and represent the approximate amplitude of variation in depth to water table, vertical hydraulic gradient, and specific discharge or recharge.

Table 2. — Specific discharge or recharge at test-well sites estimated from hydraulic data

[Positive values indicate discharge (upward flow); negative values, recharge (downward flow); plus-or-minus values indicate two standard deviations from mean values (approximately equal to the amplitude of variation); gradient adjustment factor is ratio of interpolated to measured vertical hydraulic gradient as explained in the text]

Well pair	Depth interval (m)	Period of record	Depth to water table (m)	Gradient adjustment factor	Vertical hydraulic gradient	Vertical hydraulic conductivity (mm/d)	Specific discharge or recharge (mm/a)	Comments
U.S. Geologic Survey								
2A,B	10.39-25.91	Oct 74-May 82	9.61 ± .07	1.5	+0.023 ± .007	2.7	+23 ± 7	1
3A,B	7.19-43.56	Oct 74-May 82	2.34 ± .17	1.0	+0.0036 ± .0042	2.4	+ 3.2 ± 3.6	2, 3
8A,B	3.63-37.95	Oct 74-May 82	3.29 ± .25	1.0	−0.063 ± .006	1.4	−32 ± 3	4
9A,B	10.18-39.32	Dec 74-Jul 79	9.13 ± .19	1.0	+0.018 ± .006	1.7	+11 ± 2	5
10A,C	3.86-32.31	Dec 75-Nov 82	2.83 ± .04	4.3	+0.046 ± .033	2.8	+44 ± 31	6
12A,B	14.60-21.46	Oct 74-Nov 82	10.03 ± .09	1.0	−0.068 ± .009	1.4	−35 ± 5	
13A,C	4.83-21.18	Dec 75-Nov 82	3.07 ± .15	1.2	+0.038 ± .008	2.1	+29 ± 6	7
17A,B	6.56-8.99	Oct 74-Nov 82	2.93 ± .39	1.0	−0.016 ± .034	5.7	−33 ± 63	8
18A,B	5.33-41.15	Oct 74-May 82	4.60 ± .76	1.0	−00087 ± .019	3.5	−11 ± 24	9, 10
27A,B	4.97-44.07	Dec 74-May 82	4.52 ± .35	3.3	+0.044 ± .021	2.4	+39 ± 18	11, 12
29A,B	6.68-43.95	Dec 74-May 82	6.19 ± .32	1.5	+0.035 ± .010	2.4	+31 ± 9	13
30A,C	3.17-39.93	Oct 74-Nov 82	1.78 ± .42	1.0	+0.021 ± .006	4.0	+31 ± 9	14
32A,B	7.19-44.33	Dec 74-May 82	6.88 ± .26	1.0	−0.0077 ± .0050	1.7	− 4.8 ± 3.1	15
35A,B	2.59-20.06	Oct 74-May 82	1.67 ± .18	1.3	+0.021 ± .006	2.6	+20 ± 6	17
41A,B	11.81-41.85	Dec 75-Aug 81	9.34 ± .08	1.3	+0.017 ± .005	2.1	+13 ± 4	18
46B,C	1.55- 3.93	Dec 75-Nov 82	1.14 ± .13	1.0	+0.0064 ± .0336	23	+54 ±280	19
46B,D	4.54-16.08	Dec 75-Nov 82	1.14 ± .13	1.9	+0.120 ± .049	1.2	+53 ± 21	11
46A,D	16.54-25.66	Nov 81-Nov 82	1.17 ± .11	1.9	−0.017 ± .033	7.5	−47 ± 90	11, 20
46A,C	1.55-25.66	Nov 81-Nov 82	1.17 ± .11	1.9	+0.025 ± .003	1.8	+16 ± 2	11
48A,B	4.57-30.94	Jan 75-Nov 82	2.19 ± .25	1.2	+0.058 ± .006	2.3	+49 ± 5	21
49B,C	1.85-4.38	Jan 76-Mar 81	1.00 ± .17	1.0	+0.046 ± .041	4.4	+74 ± 66	
49A,B	4.84-31.55	May 82-Dec 82	.96 ± .18	1.1	+0.148 ± .014	1.6	+86 ± 8	22
50A,B	2.53-19.29	Dec 75-Jan 77	1.59 ± .66	1.1	+0.142 ± .048	1.1	+57 ± 19	23
51A,B	12.41-43.91	Dec 75-May 82	11.79 ± .35	1.9	+0.106 ± .019	.88	+34 ± 6	24
52A,B	20.32-44.81	Dec 75-Nov 82	17.45 ± .12	1.0	+0.0031 ± .0045	5.6	+ 6.3 ± 9.2	20
53A,B	10.97-41.92	Dec 75-May 83	9.67 ± .15	2.1	+0.017 ± .011	4.5	+28 ± 18	25
55B,C	1.40-2.32	Jul 78-Nov 82	1.32 ± .22	1.0	+0.226 ± .454	.76	+63 ±126	26
55A,B	2.55-42.22	Mar 82-Dec 82	1.27 ± .20	1.0	+0.075 ± .007	3.5	+96 ± 9	27
56B,C	4.51-22.d81	Dec 75-Nov 82	3.38 ± .11	1.0	+0.080 ± .006	1.3	+38 ± 3	28
56A,C	23.27-42.46	Nov 81-Dec 82	3.40 ± .12	1.0	+0.168 ± .033	2.1	+129 ± 25	29
56A,B	4.51-42.46	Nov 81-Dec 82	3.40 ± .12	1.0	+0.121 ± .015	1.6	+71 ± 9	27
57B,C	1.73-5.71	Dec 75-Nov 82	1.07 ± .33	1.0	+0.22 ± .048	.90	+72 ± 16	
57A,B	6.16-41.35	Mar 82-May 82	1.00 ± .21	1.7	+0.101 ± .012	2.0	+74 ± 9	30
57A,C	1.73-41.35	Mar 82-May 82	1.00 ± .21	1.7	+0.126 ± .019	1.8	+83 ± 12	30
58B,C	1.02-2.63	Dec 75-Nov 82	.75 ± .22	1.0	+0.033 ± .033	7.5	+90 ± 90	31

Table 2. — Specific discharge or recharge at test-well sites estimated from hydraulic data (Continued)

[Positive values indicate discharge (upward flow); negative values, recharge (downward flow); plus-or-minus values indicate two standard deviations from mean values (approximately equal to the amplitude of variation); gradient adjustment factor is ratio of interpolated to measured vertical hydraulic gradient as explained in the text]

Well pair	Depth interval (m)	Period of record	Depth to water table (m)	Gradient adjustment factor	Vertical hydraulic gradient	Vertical hydraulic conductivity (mm/d)	Specific discharge or recharge (mm/a)	Comments
U.S. Geologic Survey (Continued)								
58A,B	3.08–42.08	Nov 81–Dec 82	.78 ± .14	1.0	+0.098 ± .009	2.7	+97 ± 9	27
59A,C	6.85–45.10	Dec 75–Nov 82	4.47 ± .19	1.7	+0.018 ± .006	3.7	+24 ± 8	32
60A,C	30.68–41.51	Mar 82–Dec 82	1.94 ± .03	1.0	+0.046 ± .024	5.3	+89 ± 46	
60DA,B	3.11–41.51	Mar 82–Dec 82	1.94 ± .03	1.0	+0.081 ± .006	3.1	+92 ± 7	27
63A,B	9.81–29.09	Jul 76–Oct 80	9.10 ± .20	1.0	+0.054 ± .009	2.6	+51 ± 9	33
U.S. Bureau of Reclamation								
13A,C	15.00–66.14	Dec 74–May 82	5.15 ± .45	1.0	−0.037 ± .004	3.2	−43 ± 5	4
13B,C	67.06–152.40	Aug 78–May 82	5.32 ± .30	1.0	+0.0037 ± .0008	.75	+ 1.0 ± .22	4
14A,B	12.65–158.96	Dec 74–May 82	6.18 ± .90	1.0	+0.024 ± .007	4.2	+37 ± 11	

Comments

1. Slow upflow in annulus above well screen.
2. Slow upflow in annulus, but well screen appears to be isolated.
3. Negative gradient March 1980.
4. No significant flow in annulus.
5. Lateral flow of cool water at about 30 m; possibly also at 20 m.
6. Strong upflow in annulus; possible lateral flow of cool water at 28 m.
7. Some upflow in annulus; possible lateral flow of warm water at 13 m.
8. positive gradient in March and November 1982.
9. Lateral flows of cool and warm water at several depths.
10. Positive gradient in July 1975 and July 1976.
11. Upflow in annulus.
12. Probable lateral flow of cool water at 36 m.
13. Upflow in annulus, mostly above about 28 m.
14. Lateral flow of warm water at 18 m.
15. Lateral flow of cool water at 37 m.
16. Lateral flow of cool water at 37 m; upflow below 37 m.
17. Possible lateral flow of warm water at 12 m.
18. Slow upflow in annulus; lateral flow of cool water at about 34 m.
19. Negative gradient October 1979, March 1980, and March 1981.
20. Gradient changes substantially with time but has not actually reversed.
21. Slow upflow in annulus.
22. Upflow in annulus, but well screen appears to be almost isolated.
23. Slow upflow in annulus, chiefly above 17 m.
24. Position near edge of terrace results in a larger vertical hydraulic gradient than normal for this depth to water table.
25. Upflow in annulus, especially above 20 m.
26. Negative gradient November 1982.
27. Annulus cemented; good data.
28. Annulus of deeper well not cemented; may be some upflow.
29. Annulus of shallow well not cemented.
30. Annulus cemented, but seal apparently incomplete.
31. Negative gradient July 1978.
32. Upflow in annulus, especially between 16 and 32 m.
33. Vertical hydraulic conductivity poorly known.

SPECIFIC DISCHARGE ESTIMATED FROM VEGETATION AND SURFACE CONDITIONS

Specific discharge near the top of the saturated zone may be estimated on the basis of empirical values of annual ground-water evapotranspiration assigned to various types of phreatophytes and surface conditions. The empirical values in table 3 are those used in numerous ground-water reconnaissance studies in Nevada by the Geological Survey. (See Nevada Department of Conservation and Natural Resources, 1960-74.) The evapotranspiration data, along with a phreatophyte and surface-conditions map of the Soda Lakes and Upsal Hogback areas, were provided by P. A. Glancy (U.S. Geological Survey, written commun., 1979).

Table 3. -- Estimated rates of evapotranspiration from ground water for various types of phreatophytes or surface conditions (Data from P. A. Glancy, U.S. Geological Survey, written commun., 1979)

Type of phreatophytes or surface conditions	Evapotranspiration rate (mm/a)
Perennially free water surface	1,200
Irrigated pasture or cropland; assumes 300 mm use per cutting of alfalfa	900-1,200
Seasonally free water surface	600
March grasses growing in a dominantly nonsaline environment and (or) dense assemblage of saltceder, greasewood, and saltgrass	300
Saltgrass dominant; locally includes minor amounts of greasewood and rabbitbrush	150
Greasewood dense or dominant, with or without rabbitbrush and (or) saltbrush with a thin understory of saltgrass	100
Greasewood dominant but of moderate density; locally includes rabbitbrush and (or) big sage and hairy horsebrush in sandy areas	60
Playa deposits containing scattered stands of pickleweed	45
Greasewood of low density and vigor; locally includes scattered rabbitbrush; playa deposits lacking vegetation	30
Greasewood of very low density and vigor	20
Area of deep water table lacking phreatophytes	0

EVIDENCE FOR LOCAL RECHARGE

For purpose of this study, ground-water recharge is defined as water that percolates to the saturated zone. Water in the unsaturated zone that discharges by evaporation or transpiration before reaching the saturated zone is not considered as recharge. As thus defined, ground-water from local precipitation in the Basin and Range province has been assumed to be negligible where average annual precipitation is less than about 200 mm (see Eakin and Maxey, 1951; Nevada Department of Conservation and Natural Resources, 1960-74). This assumption is based chiefly on the fact that, in the driest parts of the province, such as the lowest part of the Carson Desert, potential evapotranspiration may exceed the average annual precipitation by a factor of 10 or more. Seemingly, in such an environment, virtually all the meager influx from precipitation to the unsaturated zone would return to the atmosphere by evaporation or transpiration before it could percolate to the underlying saturated zone.

In principle, however, significant local recharge could occur where (1) the water table is sufficiently shallow so that, at times, the capillary fringe extends to the land surface, (2) the water content of the unsaturated zone is increased by irrigation or canal leakage so that some of the influx from local precipitation reaches the water table, or (3) runoff from local precipitation is concentrated in lakes, ponds, or stream channels and a part percolates to the water table before being discharged as evapotranspiration.

The first type of local recharge is suggested by hydraulic data obtained from well pairs 46B and C, 49B and C, 55B and C, 57B and C, and 58B and C, which are in areas of major ground-water discharge where the water table ranges in depth from about 0.5 to about 1.5 m. The deeper well of each pair is screened or perforated at fairly shallow depths ranging from about 1.0 to about 4.5 m below the water table (see table 2). All these well pairs might be expected to record reversals in the usual upward (positive) hydraulic gradient after precipitation events intense enough to produce local ground-water recharge. Such reversals have been recorded in four of the five well pairs during the period 1975-82; only at well pairs 49B and C has no reversal been observed.

At all the other sites in areas of major ground-water discharge, the depth interval represented by the shallowest well pair probably is too large to record a net downward vertical hydraulic gradient after a recharge event.

Hydraulic data for the second type of recharge from local precipitation --that in irrigated areas--would be nearly impossible to obtain from the wells used in the present study. Such recharge probably is small in comparison with that from the irrigation itself and from canal leakage. The resultant increase in downward hydraulic gradient recorded by a well pair probably could not be differentiated from the fluctuations in gradient resulting from the varied rates of infiltration from irrigation and canal leakage.

Hydraulic data for the third type of local recharge, from runoff concentrated in the intermittent ponds and streams, was obtained during the final stages of the present study. Two intermittent ponds between test-well sites 27 and 29 in the Soda Lakes area (fig. 4), which had been dry or nearly dry since the beginning of the study in the fall of 1972, filled with water during a series of storms in the winter of 1982-83. As a result, water levels near the ponds rose materially, especially in the shallow wells nearest the ponds. Unfortunately, the exact period and nature of the rise were not established. The last antecedent set of synoptic water-level measurements was made May 18-19, 1982; except for measurements made in wells 30A and C on November 17, 1982, the first measurements after the rise were made on April 20, 1983. The November 1982 data for wells 30A and C clearly indicate that the rise began before that time, however. (See fig. 4.)

Maximum recorded rise in the water table was 3.27 m in well 29B, a short distance northeast of the the northeastern pond (see figs. 3 and 4). The corresponding rise in the companion well 29A, in which the screen at a depth of 44 m is separated from the screen in 29B (at 6.7 m) by several confining beds of low vertical hydraulic conductivity, was only 0.60 m. As a result, the vertical component of the hydraulic gradient, which had been consistently positive (upward) since measurements began in October 1974, reversed and became negative (downward) sometime after May 1982, thus indicating a change from specific discharge to specific recharge. (See fig. 4.) A similar reversal occurred at well site 27, southwest of the southwestern pond, and probably also at site 39, about 0.32 km south of the southwestern pond. Water-level data at well-site 39

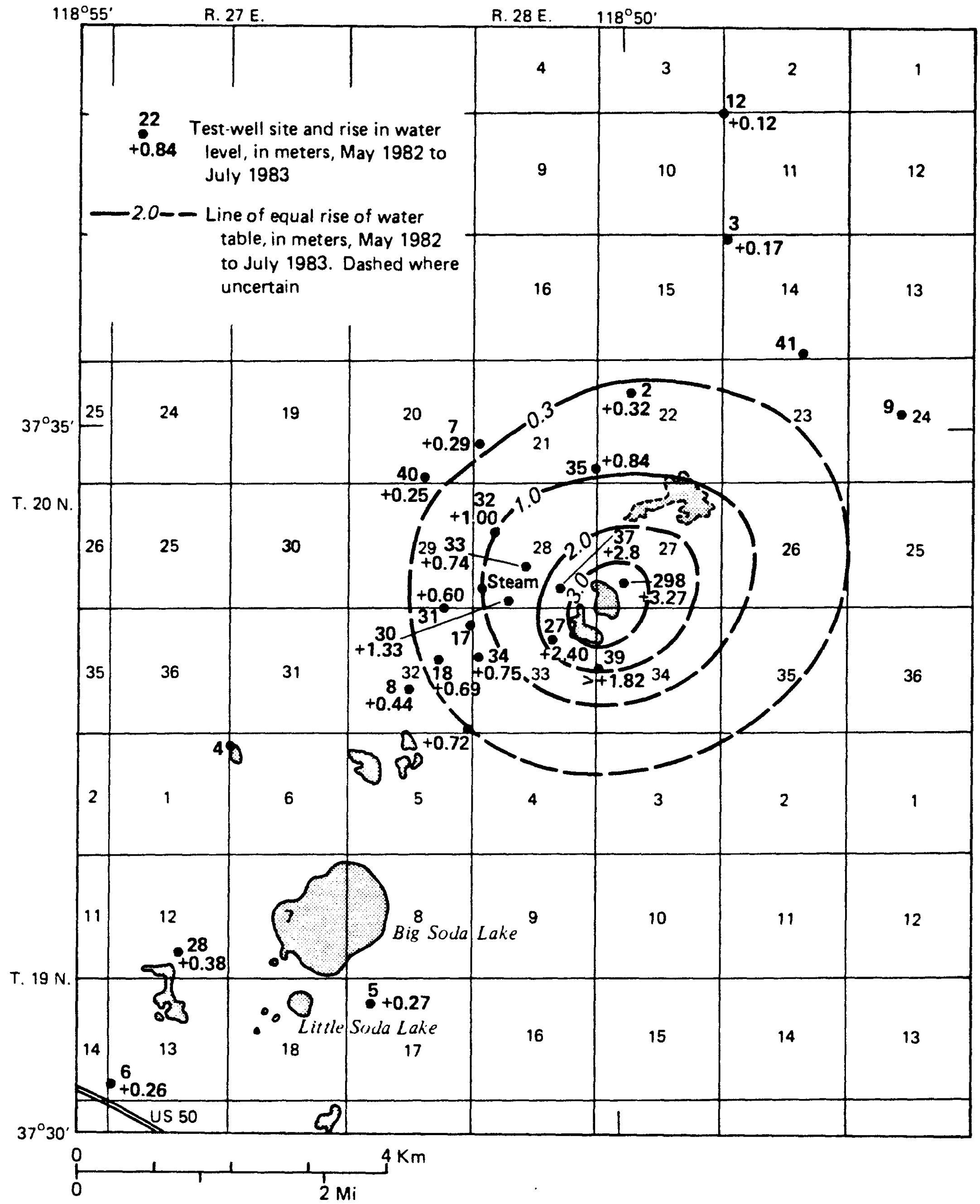

Figure 3. -- Soda Lakes geothermal area. Water-table rise attributable to rise in stage of ponds between test-well sites 27 and 29, May 1982 to July 1983.

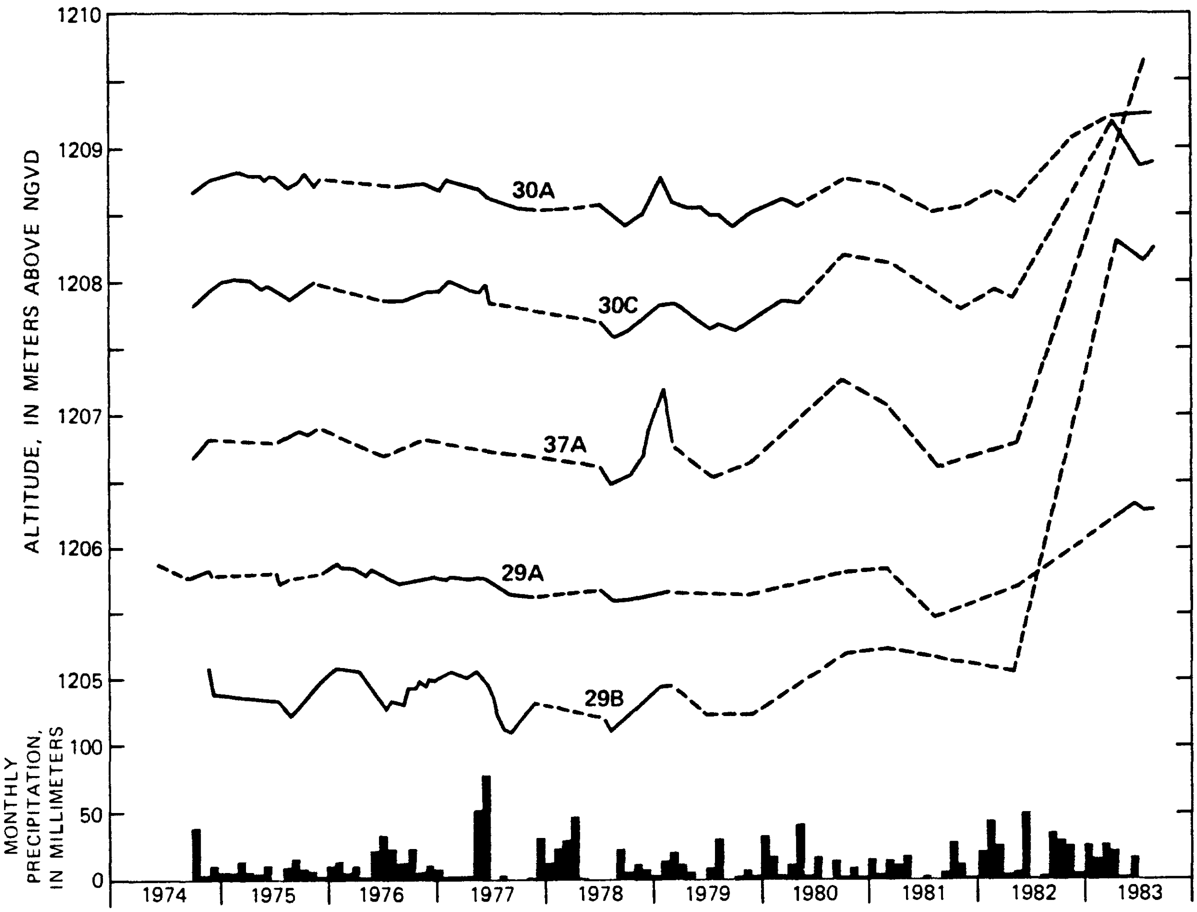

Figure 4. -- Water-level fluctuations in test wells 29A and B, 30A and C and 37A, and monthly precipitation at Fallon Experiment Station, October 1974 to August 1983.

do not conclusively indicate a gradient reversal, because the shallower well (39B) was dry before the rise. Other well pairs, more distant from the ponds, did not record reversals in vertical hydraulic gradient, and, in fact, the water-level rise was greater in the deeper well than in the shallower well at several of these sites. The greater rise in the distant deeper wells is not unexpected; the hydraulic diffusivity of the confined zone or zones in which those wells are screened presumably is greater than that of the generally unconfined zone tapped by the shallower wells because of the much smaller storage coefficient of the confined zone or zones.

The water-table rise caused by the filling of the ponds between well sites 27 and 29 is fairly well indicated by data west of the ponds, but data are lacking to the east; the interpreted pattern (fig. 3) is based on an assumption of bilateral symmetry. A further assumption is that part of the rise in the southwest (sites 8, 18 and BR 13) is attributable to a rise in stage in the ponds in the northwestern part of sec. 5, T. 19 N., R. 28 E. rather than to the rise in stage in the two ponds between sites 27 and 29.

The increase in ground-water storage represented by the recharge mound of water-table rise shown in figure 3 may be estimated roughly on the basis of the following assumptions (all of them probably conservative): (1) The outer limit of the mound is defined by water-level rise of 0.3 m; (2) maximum rise, near the ponds, is 3 m; and (3) average effective porosity of the deposits within the mound is 15 percent. On this basis, the total volume of the mound is 18 hm^3 and the volume of effective pore space is 2.7 hm^3.

The amount of recharge represented by the increase in ground-water storage in the mound adjacent to the ponds cannot be estimated from present data. Much of the rise in water levels, especially that in the outer parts of the mound (fig. 3), is attibutable to pressure-head increase and clearly does not represent water that has migrated from the ponds. A detailed study using a dense array of piezometers and hydrochemical analyses of pond water and adjacent ground-water would be required to relate ground-water-storage increase to recharge.

RESULTS AND THEIR SIGNIFICANCE

Several significant conclusions may be drawn from the data obtained in this study. First, in areas of major ground-water discharge, specific discharge as estimated from hydraulic data (table 2). varies with depth to the water table. The best least-squares fit to the data for thirty pairs of wells in discharge areas is given by the exponential function

$$\text{in } Q_z = 4.487 - 0.163\, z_s \qquad (2)$$

where Q_z is specific discharge in millimeters per year, and z_s is depth to the water table in meters. (See fig. 5.) The coefficient of determination (r^2) of 0.77 indicates a significant correlation of the logarithm of specific discharge and water-table depth.[1/] In addition to errors in the estimates of specific discharge, scatter of the data is caused by factors other than depth to water such as variation in specific discharge with depth, and type of soil, density and types of vegetation, presence or absence of a salt crust, and other surface conditions.

Estimates of specific discharge based on hydraulic data ("hydraulic" estimates) are in fair agreement with those based on vegetation and surface conditions ("vegetation" estimates) at some sites, whereas at other sites, the differences represent errors in one or both estimates. At many sites, however, the differences are real because the vertical Darcian velocity (specific discharge or recharge) varies with depth, and different depth ranges are represented by the two estimates. The "vegetation" estimates represent the specific discharge in only the uppermost part of the saturated zone, whereas many of the

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1/ Using more stringent selection criteria, Olmsted and others (1984, p. 52) obtained the relation $Q_z = 95.6\, e^{-0.155\, z}$ ($\ln Q_z = 4.56 - 0.155\, z_s$) with an r^2 of 0.94 for 11 test-well sites; their results are similar to those of the present study in spite of their exclusion of the less reliable data.

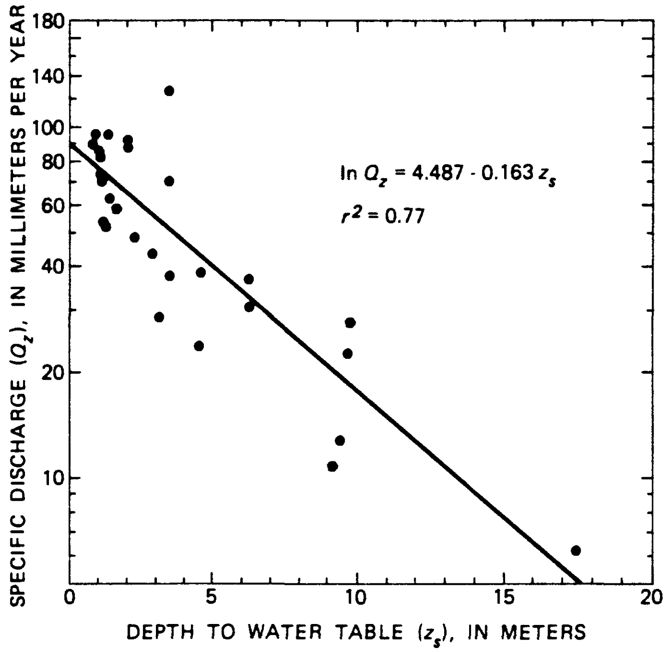

Figure 5. -- Relation of specific discharge to water-table depth at selected test-well sites.

"hydraulic" estimates in table 4 represent a much greater depth interval. Thus, consumptive use of water by phreatophytic vegetation (representing specific discharge) occurs at some sites where "hydraulic" data from well pairs indicate net specific recharge for the interval between the screens or perforations of the two wells. The data for USGS sites 8, 18, and 34, and USBR site 13 (table 2) all illustrate this case. At several other sites, both estimates indicate specific discharge, but the specific discharge near the water table is substantially different from that at greater depth.

Table 4. -- Summary of estimates of specific discharge or recharge at test-well sites based on hydraulic data, vegetation and surface conditions, and depth-to-water-table relation

[Values of specific discharge (+) or recharge(-) are in millimeters per year]

Test well site	Long term average depth to water table (m)	Depth interval for hydraulic data (m)	Specific discharge or recharge		
			Hydraulic data	Vegetation or surface conditions	Depth-to-water-table relation (fig. 5)
U.S. Geological Survey					
2	9.61	10.39-25.91	+23	+30	+18
3	2.34	7.91-43.56	+ 3.2	+60	+61
8	3.29	3.63-37.95	-32	+60	+52
9	9.13	10.18-39.32	+11	+30	+20
10	2.83	3.86-32.31	+44	+30	+56
12	10.03	14.60-21.46	-35	0	+17
13	3.07	4.83-21.18	+29	+30	+54
17	2.93	6.56- 8.99	-33	0	+55
18	4.60	5.33-41.15	-11	+60	+42
27	4.62	4.97-44.07	+39	+60	+42
29	6.19	6.68-43.95	+31	+30	+32
30	1.78	3.17-39.93	+31	+60	+66

Table 4. — Summary of estimates of specific discharge or recharge at test-well sites based on hydraulic data, vegetation and surface conditions, and depth-to-water-table relation (Continued)

[Values of specific discharge (+) or recharge(-) are in millimeters per year]

Test well site	Long term average depth to water table (m)	Depth interval for hydraulic data (m)	Specific discharge or recharge: Hydraulic data	Specific discharge or recharge: Vegetation or surface conditions	Specific discharge or recharge: Depth-to-water-table relation (fig. 5)
U.S. Geological Survey					
32	6.88	7.19-44.33	- 4.8	+30	+29
34	7.21	7.50-44.44	+13	0	+27
35	1.67	2.59-20.06	+20	+30	+68
41	9.34	11.81-41.85	+13	+30	+19
46	1.14	1.55-25.66	+54; +53	+30	+74
48	2.19	4.57-30.94	+49	+60	+62
49	.96	4.84-31.55	+86; +74	+30	+76
50	1.59	2.53-19.29	+57	+30	+69
51	11.79	12.42-43.91	+34	+20	+13
52	17.45	20.32-44.81	+ 6.3	+20	+ 5.1
53	9.67	10.97-41.92	+28	+20	+18
55	1.27	2.55-42.22	+96; +63	+30	+72
56	3.38	4.51-42.46	+71; +38; +129	+45	+51
57	1.07	6.16-41.35	+72; +74; +83	+30	+75
58	.78	3.08-42.08	+97; +90	+30	+78
59	4.37	6.85-45.10	+24	+30	+43
60	1.97	3.11-41.51	+92; +92; +89	+30	+65
63	9.10	9.81-29.09	+51	+30	+20
U.S. Bureau of Reclamation					
13	5.15	10.00-66.14	-43	+60	+38
14	6.18	12.65-158.96	+37	0	+32

In addition to the real differences described above, both the "hydraulic" and the "vegetation" estimates are subject to potentially serious sources of error. In the "hydraulic" method, the gradient-adjustment factor used to correct for the "short-circuit" effect of annular flow of water in the deeper well at the site may be incorrect. Even more serious are errors related to the large range of uncertainty in the estimates of vertical hydraulic conductivity and inaccuracies in the lithologic logs of the wells. In the "vegetation" estimates, the rates of specific discharge assigned to the categories listed in table 3 may not be appropriate for the study area, and the areal distribution of the various categories may have been mapped incorrectly. In addition, factors such as density and vigor of phreatophytic vegetation were not quantified; instead, discrete values of specific discharge were assigned to a few broad categories within which specific-discharge rates undoubtedly have a wide range.

In spite of the lack of direct comparability of the estimates of specific discharge or recharge by the two methods described above, and the unknown but possibly large errors in the values obtained, a few general conclusions may be drawn from the data. First, the "hydraulic" estimates appear to be reasonable in comparison with the "vegetation" estimates in areas of major ground-water discharge, in spite of the large range of uncertainty in the former estimates, especially with regard to values of vertical permeability assigned to the materials described in the lithologic logs of the test wells. That is, the two estimates do not differ by a factor of more than 3 or 4. (See table 4.) Second, unless the errors in the "hydraulic" estimates have an areal bias, comparison of the two estimates suggests that the specific discharge of 30 mm/a assigned to playa deposits lacking vegetation (table 3) may be too low, especially in areas of shallow water table, but that the same value may be reasonable in areas of greasewood of low-density and vigor where depths to water table are greater. Third, as discussed earlier, the logarithm of specific discharge is strongly correlated with water-table depth within areas of major ground-water discharge in spite of the perturbing effects of soil type, vegetation, presence or absence of a salt crust, and other extraneous factors.

On both theoretical and empirical grounds, local recharge in low-lying areas like the west-central Carson Desert probably is more significant than generally recognized heretofore (see Eakin and Maxey, 1951; Glancy and Katzer, 1975). Data obtained during the present study clearly indicate recharge from occasional intense storms in areas of shallow water table, but the data are not sufficient to indicate the magnitude of such recharge. Detailed site studies in areas of shallow water table are needed to define the magnitude of recharge in relation to intensity of precipitation, temperature, soil type, antecedent soil-moisture conditions, depth to water, ground cover, and other factors. Similar studies are needed in areas of deeper water table where high soil-moisture levels are produced by irrigation and canal leakage.

The major rise in the water table following the filling of the two ponds between well sites 27 and 29 after at least 8 years of dry or nearly dry conditions provided unequivocal evidence of recharge from surface-water bodies that collect and concentrate runoff from local precipitation. Again, however, the magnitude of such recharge in terms of a long-term water budget for a large area such as the Carson Desert cannot be assessed with the data presently at hand. The frequency of such events is not known, nor is the size and spacing of the areas so affected.

The reason why the ponds filled during the winter of 1982-83 but not during the previous 9 winters is not obvious from casual inspection of the precipitation and water-level data in figure 4. The winter of 1977-78 also was much wetter than average, but it followed several years of below-average precipitation (in spite of the large amounts in May and June of 1977), and ground-water levels in the wells shown in figure 4 actually declined to somewhat below-average values in late summer of 1978. The winter of 1982-83, on the other hand, followed the wet winter of 1981-82; soil moisture levels probably were higher than normal during the summer of 1982, so that the series of storms in the winter of 1982-83 produced unusually heavy runoff in the area tributary to the ponds. Another possible explanation for the heavy runoff is that the local precipitation was much greater than that recorded at Fallon Experiment Station --the record shown in figure 4. Whatever the reason, the data shown in figure 4 indicate clearly that precipitation, in itself, had little or no effect on local ground-water recharge at the sites near the ponds; only the filling of the ponds produced significant rise in ground-water levels during the 9-year period of record.

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