Case No. 84739

IN THE SUPREME COURT OF THE STATE OF NEVADA

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Nov 08 2022 04:38 p.m.
Elizabeth A. Brown
Clerk of Supreme Court

ADAM SULLIVAN, P.E., NEVADA STATE ENGINEER, et al.

Appellants,

vs.

LINCOLN COUNTY WATER DISTRICT, et al.

<u>JOINT APPENDIX</u>

VOLUME 36 OF 49

Geologic interpretations concerning the nature and geometry of faults and the location and properties of geologic formations (stratigraphic units) are used to delineate potential aquifers in which to site wells and to locate the thickest sections of regional and local flow. Distinction between carbonate and noncarbonate rocks, and between rocks types within the carbonate-rock layers, is important. Hess and Mifflin (1978, p. 32) concluded that carbonate-rock units of certain ages exhibit more indications of high permeability than units of other ages. The present study (in several series of detailed observations of outcrops) has yet to verify this result. In addition, analysis of available aquifer-test data indicate no correlation between permeability and carbonate-rock age.

Although the age of a carbonate-rock unit does not appear to define its permeability, the composition of the rock in a particular unit does. Generally, the shaly or silty carbonate-rock units are less permeable than the cherty or "clean" carbonate rocks in Nevada. Two wells in shaly and silty carbonate rocks in the study area that reflect the tendency for lower productivity from shaly carbonates are the MX well CE-VF-2 in Coyote Spring Valley and CSV-2 well in Upper Moapa Valley. Specific capacities at these wells in shaly carbonates—5.9 and 3.4 (gal/min)/ft, respectively (table 1)—are smaller than at wells drilled in "cleaner" units in the same area, MX wells CE-DT-4, -5, and -6, with specific capacities of 154, 309, and 11.4, respectively (table 1). Outside of Nevada, a similar relation has been found in other carbonate terranes by Sweeting and Sweeting (1969), Dreiss (1974), and Rauch and White (1970, 1977). On the other hand, three of the six most productive wells drilled in the carbonate-rock aquifers (well C at the Nevada Test Site and Tracer wells number 2 and 3 near Ash Meadows) were drilled into the carbonate-rock units nearest the bottom of the Paleozoic-rock sequences. This means that these wells tap thin carbonate-rock aquifers in comparison to most of the other wells that tap carbonates in southern Nevada. Thus, productivity probably is not closely tied to the overall thickness of carbonate rocks underlying a given well. It seems likely that the structures intercepted in a particular well (and the history of those structures) are determining the productivity of a well to a greater extent than either the stratigraphic unit or the thickness of the carbonate-rock section available for ground-water flow.

Water Quality

Water quality (chemistry) is an important consideration in obtaining a water supply. Chemical considerations differ according to the intended uses of the water, but, in general, domestic and municipal uses (especially drinking water) have more stringent chemical requirements than irrigation and other agricultural uses (Nevada Bureau of Consumer Health Protection Services, 1977, p. 8-9). Industrial and commercial uses have requirements that generally range between these two extremes. Because water quality varies from place to place in southern Nevada, ground water in some areas may not be suitable for most uses without expensive treatments to improve the chemistry of the water. Fortunately, regional patterns in water quality can be observed that allow delineation of areas of unacceptable quality.

Geochemical Controls on Quality

Chemistry of ground water in southern Nevada (carbonate rock, volcanic rock, and basin fill) is directly related to the type and amount of minerals present within the flow systems and environmental conditions upon recharge and discharge of the water. As a result, the quality of ground water tends to reflect the conditions along its flow path. Overall, however, the quality of ground water in regional systems in the carbonate-rock province of Nevada is notable for its large-scale homogeneity because of the dominance of carbonate minerals throughout the aquifers (Hess and Mifflin, 1978, p. 38). With a few exceptions, the water quality is also notably good; concentrations of dissolved solids (a general measure of the quality or salinity of water) in water sampled from the carbonate-rock aquifers commonly are less than 500 mg/L, and predominantly contain calcium, magnesium, and bicarbonate ions (Lyles and others, 1986, p. 13).

Within the large-scale homogeneity, local variations in water quality are caused by geologic complexities and localized interaquifer flows between rock types. In some areas, the quality of water discharging from or near carbonate rocks may have been changed by flowing through lake-bed deposits (mostly of late Tertiary age) that commonly contain some dissolvable evaporite minerals. In the area north and west of Las Vegas, some evaporite minerals are found in the basin-fill deposits. The presence of evaporite-bearing deposits will generally be recognized by geologic mapping or by the presence of water with dissolved-solids concentrations unsuitable for public supply and most other uses in basin-fill aquifers. An example of the relation between water quality in these younger basin-fill deposits and in a nearby carbonate-rock aquifer is described by Winograd and Thordarson (1975, p. C106-107). They state that the chemistries of water from the two aquifers are independent of each other and will remain so unless mixing of water from the two aquifers were induced or happened naturally.

Depth and Quality

In much of the carbonate-rock province, the depths from which water can be extracted might not be limited by water-quality constraints. Previous studies have noted that the quality of water in the carbonate-rock aquifers in much of the carbonate-rock province does not increase in dissolved-solids concentration with increasing depth (Winograd and Thordarson, 1975, p. C102-103; Hess and Mifflin, 1978, p. 42). The present study (together with results of the MX Missile Water Resources Program of the early-1980's) found no information to contradict that conclusion. The concentrations of major ions and dissolved solids (as well as some physical properties) of water samples obtained from deeper parts of the carbonate-rock aquifers (between 1,200 and 11,000 ft beneath land surface) are listed in table 6. The locations of these sample sites and the relation between depth below land surface and dissolved-solids concentration are shown in figure 23. As mentioned previously, these data do not indicate a significant increase in dissolved-solids concentration with depth. All the sites in table 6 are located north of the line that delineates the western limit of unsuitable quality water in figure 20, and because the samples selected are from deep in the aquifers, they probably are water that has not contacted young lakebed evaporites. Of the 21 samples, 15 contained less than 500 mg/L (milligrams per liter) dissolved solids; none contained more than 1,000 mg/L, and the median concentration was 366 mg/L. However, saline to briny water frequently has been found in localized pockets and traps penetrated by deep oil-test wells in eastern and southeastern Nevada. These pockets and traps are common in carbonate as well as volcanic and younger rocks (Chamberlain, 1988).

Mesozoic Evaporites and Water Quality

Southeast and east of Las Vegas (southeast of the line that shares the western limit of water with ionic compositions dominated by sulfate and chloride in fig. 20), depths from which water suitable for most uses can be extracted from the carbonate rocks may be severely limited. Water that is unsuitable for most uses is much more commonly found in that area, the dissolved-solids concentration in this water is derived most commonly from rocks that overlie or are in shallow structural juxtaposition with the carbonate-rock aquifers. The water (fig. 24) contains much more sulfate than does most water from farther north, and concentrations of dissolved solids commonly exceed the drinking water standard of 1,000 mg/L (Nevada Bureau of Consumer Health Protection Services, 1977, p. 8-9; Lyles and others, 1986, p. 13). Tertiary lake-bed sediments (such as the Muddy Creek and Horse Spring Formations) and other deposits contain abundant evaporite minerals in this southeastern area (Bohannon, 1984, p. 1-2). In addition to these basin-filling deposits, this southeastern area is underlain by extensive evaporite-rich sedimentary rocks of latest Paleozoic and Mesozoic age (fig. 24); they are older than the lakebed sediments but younger than most of the Paleozoic carbonate rocks. The latest-Paleozoic Kaibab Formation and Mesozoic Moenkoepi Formations are examples of such units, and contain hundreds of feet of soluble gypsum (Longwell and others, 1965, p. 37-39). Dissolution of evaporite minerals from these formations is a widespread source of sulfate in both basin-fill aquifers and carbonate-rock aquifers of southeast Nevada. Rogers Spring (pl. 1, fig. 24), which discharges from carbonate rocks above the Overton Arm of Lake Mead, contains about 1,700 mg/L of sulfate (as compared with a drinking-water standard of 500 mg/L, Nevada Bureau of Consumer Health Protection Services, 1977, p. 8-9) and 3,000 mg/L of dissolved solids (Lyles and others, 1986, p. 51). Even poorer quality water has been penetrated in shallow and deep wells in the carbonate rocks. For example, water in the Callville Limestone (a unit deposited during the middle of the Paleozoic Era) sampled from 17,470 ft in the Virgin River USA 1-A oil-test well (fig. 24) is reported to contain concentrations of dissolved solids of about 80,000 mg/L.

Both the lake-bed deposits and the sedimentary rocks of intermediate (latest Paleozoic and Mesozoic) age most commonly overlie the carbonate-rock aquifers, regardless of the structural deformation of the area, because they are much younger than the rocks

Table 6. Chemistry of water from wells more than 1,000 feet deep in carbonate-rock aquifers

[Dissolved solids.--r, residue on evaporation at 180 degrees Celsius; s, sum of silica and major cations and anions except bicarbonate, plus one-half of bicarbonate (to make result comparable with residual values); and ss, sum of major cations and anions except bicarbonate, plus one-half of bicarbonate (silica not determined); --, no data available]

Site No. (fig. 23)	Site name	Latitude (Degrees, minutes, seconds)	Longitude (Degrees, minutes, seconds)	Sample date	Total well depth (feet)	Temperature (Celsius)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO3)	Sulfate (SO4)	Chloride (Cl)	Silica (SiO2)	Dissolved solids	Reference
							Chemical constituent or property (milligrams per liter)									
1	S18/63-34C	36 20 28	114 55 36	9/30/86	1,205	31	128	48	134	12	226	380	190	23	1,000 s	Unpub. USGS analysis
2	TW-4	36 34 54	115 50 08	9/13/62	1,490	26	34	17	13	2.5	197	17	6.1	20	225 r	Schoff and Moore, 1964, p. 68
3	Army 1	36 35 30	116 02 14	3/18/71	1,946	31	44	22	37	5.2	262	51	15	19	301 r	Claassen, 1973, p. 18-19
4	TW-10	36 35 31	115 51 04	2/4/64	1,301	27	37	19	69	0	201	14	5	15	198 s	R.A. Young, USGS, written commun., 1965
5	TW-F	36 45 34	116 06 59	6/17/62	3,400	64	68	30	63	9.6	278	181	11	31	536 r	Schoff and Moore, 1964, p. 69
6	TW-3	36 48 30	115 51 26	5/10/62	1,853	38	51	21	83	7.6	328	84	23	24	444 r	Schoff and Moore, 1964, p. 70
7	UE25P#1	36 49 38	116 25 21	5/12/83	5,922	56	100	39	150	12	710	160	28	41	784 r	Benson and McKinley, 1985, p. 4-5
8	MX CE-VF-2	36 52 27	114 56 51	2/05/86	1,200	34	47	21	81	11	303	90	34	34	470 s	Berger and others, 1988, table 4
9	Well C	36 55 08	116 00 35	3/29/71	1,701	37	72	30	120	14	589	66	33	29	628 r	Claassen, 1973, p. 66-67
10	Well C-1	36 55 07	116 00 34	3/29/71	1,650	38	72	30	120	14	588	66	33	29	621 r	Claassen, 1973, p. 74-75
11	TW-D	37 04 28	116 04 30	1/9/60	1,950	26	17	10	107	14	274	71	20	18	343 r	Thordarson and others, 1962, p. 40
12	TW-1	37 09 29	116 13 23	8/11/62	4,206	31	20	11	35	3.2	150	35	8.7	19	220 r	Schoff and Moore, 1964, p. 72
13	WW-2	37 09 58	116 05 15	3/21/71	3,422	34	31	14	27	6.7	197	21	6	44	228 r	Claassen, 1973, p. 98-99
14	MX DL-DT-3	38 05 31	114 53 42	12/--/80	2,395	27	76	30	18	6.5	404	20	5	24	366 r	Bunch and Harrill, 1984, p. 36
15	MX CV-DT-1	38 07 58	115 20 41	12/--/80	1,837	23	38	18	18	4	221	20	5	35	253 r	Bunch and Harrill, 1984, p. 32
16	N7/55-28CA	38 26 13	115 47 44	10/06/55	1,711	60	12	5	[a]189	--	410	99	16	--	526 ss	Van Denburgh and Rush, 1974, p. 53
17	N7/56-2DAB	38 29 43	115 38 17	11/24/54	10,123	109	7	6	[a]192	--	[b]293	50	68	--	491 ss	do.
18	MX SV-DT-2	38 55 21	114 50 36	1/19/81	2,447	11	66	14	15	4.4	218	57	12	28	305 s	Unpub. USGS analysis
19	Fad shaft	39 30 20	115 59 05	1/21/53	2,465	13	52	26	83	1.4	238	38	10	11	178 s	Harrill, 1968, table 14
20	N23/56-22C	39 50 43	115 39 31	4/22/85	5,800	30	43	19	13	3.4	200	43	8.3	19	350 r	Unpub. USGS analysis

[a] Calculated sodium plus potassium, expressed as sodium.

[b] Also detected, 43 milligrams per liter of carbonate (CO_3); included in dissolved-solids sum.

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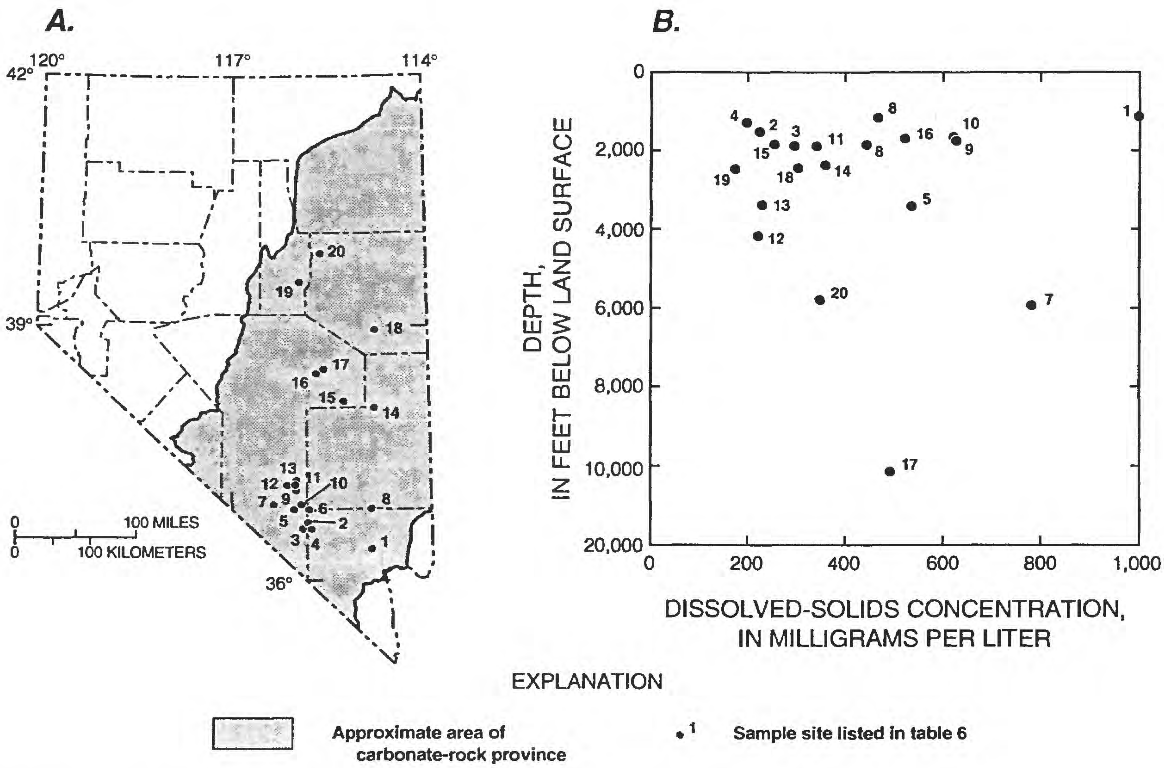


Figure 23. Relation between dissolved-solids concentration and depth below land surface.

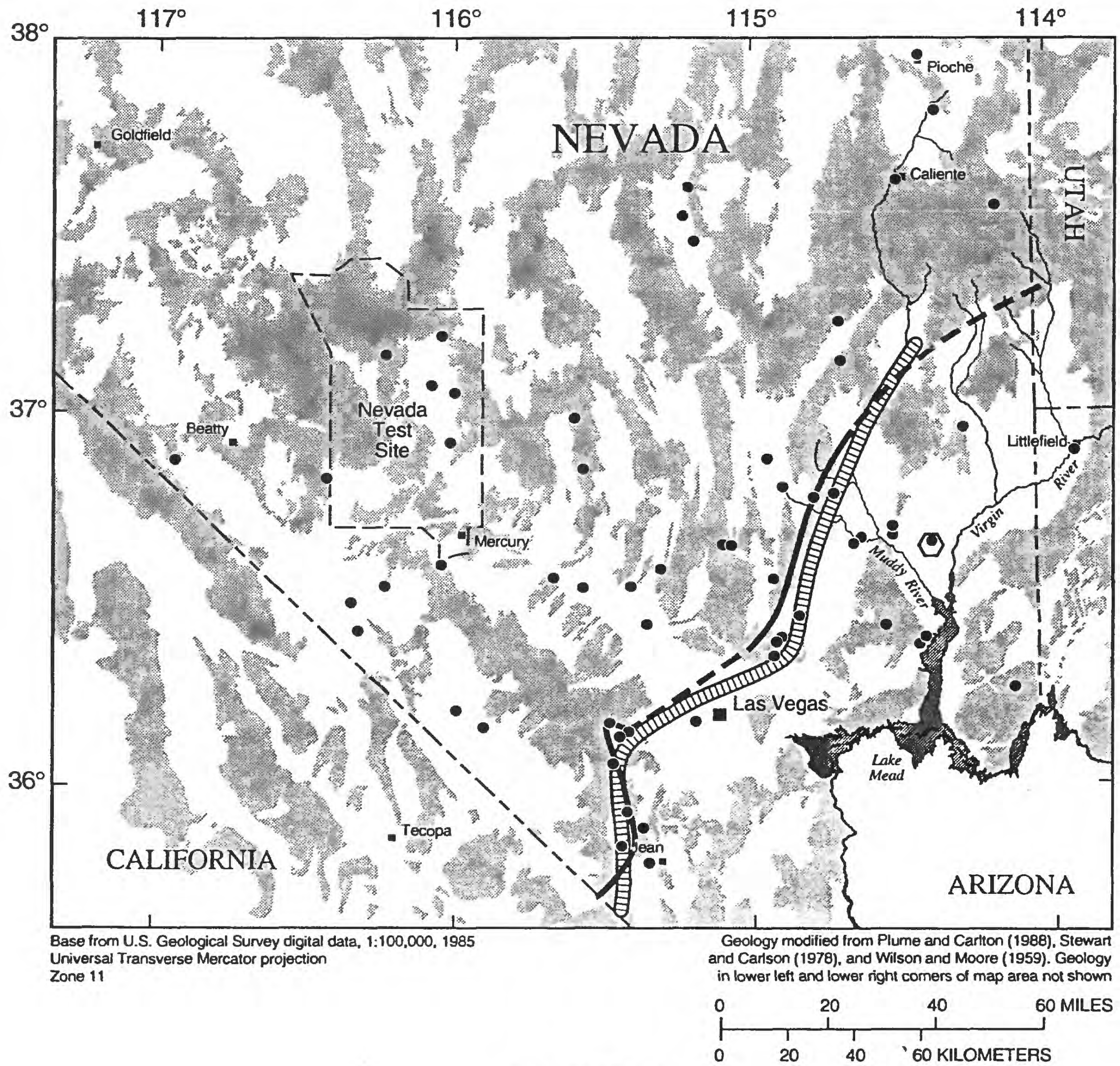
comprising the aquifers. Thus, at great enough depths in the carbonate rocks, freshwater might underlie the zones of water contaminated by the younger evaporite minerals. However, the complex geology of southern Nevada includes many places where the stratigraphic relations are reversed (just upgradient from Rogers Spring, for example). This complexity, together with flow paths that may traverse several rock types as water moves from recharge areas to discharge areas, can greatly extend the chemical influence of the contaminating deposits under even natural conditions. Once the carbonate-rock aquifers are developed, it is difficult to estimate how long pumping can proceed before capturing water of unacceptable quality.

At a regional scale, areas south and east of the lines shown on figure 24 have a relatively low potential for successful aquifer development because of the presence of water with dissolved-solids concentration commonly in excess of 1,000 mg/L. Therefore, assessments of the resources of this area were not emphasized in this study. This area is only a small fraction of the total area under consideration.

HOW MUCH WATER POTENTIALLY CAN BE WITHDRAWN FROM THE CARBONATE-ROCK AQUIFERS?

The water resource of carbonate-rock aquifers is the sum of sustained yield of the aquifers and a one-time reserve of water stored in the thick rocks. The magnitude of each of these components is estimated as described in the following paragraphs.

The sustained yield of the aquifers is the quantity of water that could be withdrawn annually for an indefinite period of time without depleting the ground-water reservoir (Scott and others, 1971, p. 13). In order to use the concept of sustained yield in this area of sparse data, the aquifers are assumed to be in a state of dynamic equilibrium—the long-term average recharge equals the average discharge and the quantity of water in storage remains constant except for minor seasonal and short-term climatic fluctuations. In arid areas such as southern Nevada, pumping from carbonate-rock aquifers is unlikely to induce additional recharge, except by inducing leakage from overlying basin-fill aquifers. Thus, the sustained yield of the



EXPLANATION

Basin fill

Consolidated rock

Northwestern limit of area known to contain thick or abundant evaporite beds — Evaporites are in rock of late Paleozoic and Mesozoic age

Boundary between areas where ionic composition of water in carbonate-rock aquifer is dominated by sulfate and chloride (to southeast) and by bicarbonate (to northwest) — Modified from Lyles and others (1986, fig. 13), using data from Winograd and Thordarson (1975, pl. 3), Glancy and Van Denburgh (1969, table 17), Trudeau and others (1983, table 1), and compiled for present study. Dashed where uncertain

Sample site — Used to delineate between bicarbonate-type and sulfate-chloride-type water

Petroleum-exploration well referenced in text

Figure 24. Location of transition between areas where ionic composition of water in carbonate-rock aquifer is dominated by sulfate and chloride (to southeast) and by bicarbonate (to northwest) in Nevada part of study area.

carbonate-rock aquifers is limited mostly to the fraction of natural discharge that can ultimately be captured by pumping. Because most of the discharge from the carbonate-rock aquifers is springflow, which has been appropriated for use, development of large parts of the sustained yield by pumping eventually will affect existing water rights. Protection of prior appropriations will place limits on pumping from the carbonate-rock aquifers. However, given conditions of high water demand, pumping water directly from the aquifers may use the resource more efficiently in the sense that pumpage can be turned off or varied to meet fluctuating demands, whereas spring discharge is relatively uniform and may not be readily used or stored during periods of low demand.

Ground water is stored in the water-saturated sections of carbonate rocks. Thicknesses of these sections range from less than 3,000 to about 20,000 ft. Because of the great thickness and vast areal extent of the rocks, volumes are measured in tens of thousands of cubic miles. Even though only a small percentage of the total volume of the aquifers is open for storage of water (fractures and interstitial spaces), the volume of water stored must greatly exceed the annual recharge or discharge rate. The water discharged in a given year has been in transit through the aquifers for thousands of years on average (for example, Thomas and Welch, in press). This large volume of water in slow transit through the carbonate rocks is a reservoir that potentially could be developed for use. Once this stored volume of water is tapped and a significant part is removed, however, the large disparity between the volume in storage and annual recharge means that very long periods without further pumping would be required for the storage reserve to refill.

The quantity of water that can be removed from storage without effects that would be unacceptable to the State is unknown. Minimum-effect strategies to develop stored water probably could be based on observations that recharge areas are separated from discharge areas (springs) by more than 100 mi in some flow systems, that relatively fresh water has been found at depths of 10,000 ft (implying that active ground-water flow occurs deep within the carbonate rocks), and that certain rock types and geologic structures form local barriers to prevent flow within regional ground-water flow systems. These observations imply that the aquifers to be developed, and the distances between developments, may be chosen to be large enough to minimize effects in key areas. Also, development might be strategically located behind geologic barriers that prevent or reduce the effect from extending in critical directions. The possibility of pumping stored water from some parts of the carbonate-rock aquifers appears to offer the most potential for additional development and is the component of the resource being most closely evaluated by this study.

Sustained Yield

The sustained yield of the carbonate-rock aquifers is equal to that fraction of the total discharge that can be captured by pumping without causing unacceptable effects. This will be less than or equal to the total discharge, and will be less than or at most equal to the total rate of natural recharge and inflow to an area.

Precipitation in the mountains is the ultimate source of virtually all recharge to the carbonate-rock aquifers. Of the precipitation that remains after sublimation, evaporation, and transpiration in the mountains, some or all percolates down into the rocks as recharge to the ground-water flow system and the rest supplies runoff and recharge to the adjacent basin-fill ground-water reservoirs. How much total recharge and runoff is generated by precipitation on the mountains is uncertain; how much of the recharge enters the carbonate-rock aquifers as opposed to the basin-fill aquifers is even less certain.

The quantity of water recharging the carbonate rocks is uncertain because once water enters the rocks it can follow one of several flow paths, and at each point will follow the one defined by least resistance. Flow paths may involve lateral movement through the carbonate-rock aquifers into adjacent basin-fill aquifers. The water that follows these paths leaves the carbonate-rock aquifers and becomes part of the resource of the basin-fill aquifers where it may be extracted as pumpage, discharged by evapotranspiration in areas of shallow ground water, or returned to an adjacent carbonate-rock aquifer. Some basin-fill aquifers receive recharge from adjacent carbonate-rock aquifers, some basin-fill reservoirs provide recharge to adjacent carbonate-rock aquifers, and others both receive and provide recharge. Because these interactions are deep within the subsurface, they are difficult to identify and quantify. Consequently, it also is difficult to estimate the quantity of water recharging the carbonate-rock aquifers. At present (1989), attempts at quantification are only adequate (1) to derive an estimate of the upper to the estimated recharge to all aquifers in the area, and (2) to estimate a lower limit based on discharge from the carbonate-rock aquifers at regional springs and by flow from the State through carbonate-rock aquifers. This latter estimate is a lower limit because it neglects the water that is moving from the carbonate-rock aquifers, through basin-fill aquifers, to discharge areas that have not been recognized as discharging water from the carbonate-rock aquifers. This unrecognized discharge from the carbonate-rock aquifers is not an additional water supply because it already has been included in the water budgets of the basin-fill aquifers.

Water Budget for All Southern Nevada Aquifers

The total of natural recharge to mountains and subsurface inflow for the study area is estimated to be about 160,000 acre-ft/yr. This estimate is based on the empirical method used to calculate recharge resulting from precipitation in mountains for water budgets in all basins of Nevada (Scott and others, 1971, p. 40) and on numerical models of ground-water flow in basin-fill aquifers around the Spring Mountains (Harrill, 1976, p. 50; Harrill, 1986, p. 46). The major mountain ranges of southern Nevada are shown in figure 25, and estimates of the recharge contributed by each are listed in table 7.

Some recharge enters the carbonate-rock aquifers and some enters the basin-fill reservoirs. Of the part that enters the carbonate-rock aquifers (directly or indirectly though basin fill), about 77,000 acre-ft/yr discharges from the carbonate rocks at major discharge areas for regional flow and identifiable regionally fed springs (fig. 25 and table 8). This discharge rate is believed, on the basis of the present studies, to include about 21,000 acre-ft/yr that enters the area from recharge sources in the high mountains of east-central Nevada along the northern part of the White River flow system (table 7; Kirk and Campana, 1988, scenario 1; Thomas, 1988; Thomas and Welch, in press). This table does not include springs believed to be supplied by local recharge (for example, Indian Springs). Local flow through the carbonate rocks represents a local resource that is little different from and probably intimately connected with the local basin-fill flow systems. Consequently, the local springs are included in estimates of discharge from the basin-fill aquifers that follow. Estimates of the quantity of water locally recharged to and discharged from carbonate rocks may have to be modified according to the extent that presumptions about the local recharge are modified by findings of future studies.

Assuming that the discharge information in table 8 accounts for most of the regionally derived discharge from the carbonate rocks in southern Nevada, all the recharge (listed in table 7) other than that supplying these discharge areas is believed to be discharged from the basin-fill aquifers. This is illustrated by a ground-water budget encompassing flows in both basin-fill and carbonate-rock aquifers (table 9). Discharge information in table 8 was supplemented by summing published estimates (see table 7 for references) of ground-water evapotranspiration—excluding areas where evapotranspiration was supported by discharge from regional springs—the Virgin River, and the Colorado River. Table 9 shows the total estimated inflow for all aquifers is about 161,000 acre-ft/yr and the total estimated discharge is about 165,000 acre-ft/yr. The difference, about 4,000 acre-ft/yr, may be attributed to estimation errors or unresolved hydrologic factors. This budget accounts for virtually all the replenishable water supply in southern Nevada.

Water Budget for the Central Corridor

The basin-fill and carbonate-rock water budgets are so intimately intertwined that an estimate is difficult to develop for the total flow through the carbonate-rock aquifers that does not also include the total flow through basin-fill aquifers. Either the total of flow in both aquifers can be estimated or that part of the flow that discharges directly from the carbonate-rock aquifers (neglecting that part of flow through carbonate-rock aquifers that discharges from the basin-fill aquifers) can be estimated, but not just the total flow through the carbonate rocks. As noted earlier (fig. 20), the central corridor contains the principal carbonate-rock aquifers of southern Nevada; it also contains 21 basins with basin-fill aquifers. Table 10 shows the estimated ground-water budget for the central corridor. The component of flow moving through the rocks of the central corridor that ultimately supplies regional discharge areas totals about 77,000 acre-ft/yr, 73 percent of which is derived from recharge areas in the corridor and 27 percent of which flows into the area from ground-water systems farther north. Water in a system having a dynamic equilibrium between recharge and discharge (which the authors assume is an adequate description of the carbonate-rock aquifers of southern Nevada) will not flow unless flow is ultimately toward

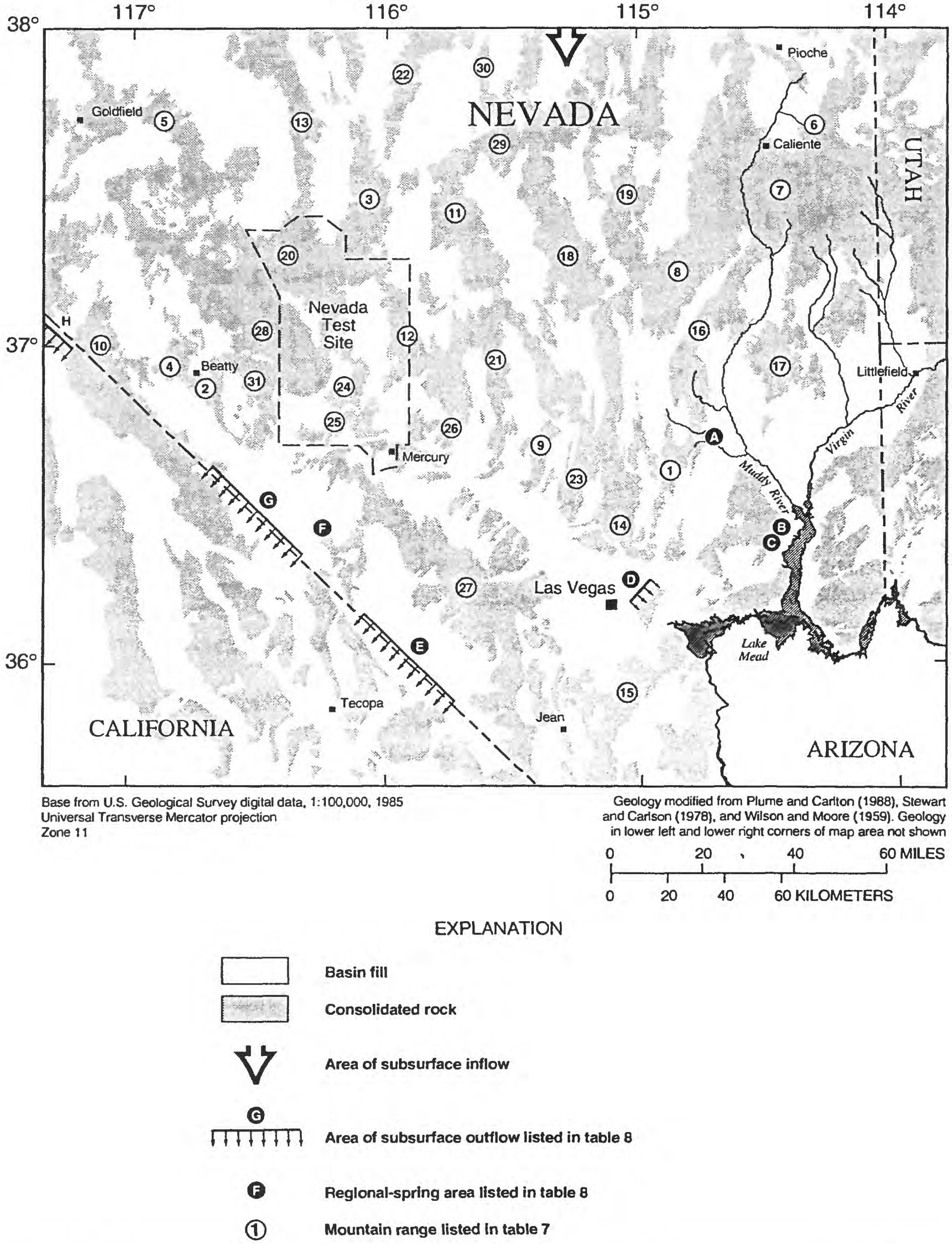


Figure 25. Mountain ranges where ground water is recharged and areas where regional ground-water flow is discharged in Nevada part of study area.

Table 7. Estimated recharge from the 31 most significant mountain ranges

[Abbreviations: N., North; S., South; V., Valley; W., West]

Map No. (fig. 25)	Mountain range	Total recharge[1] (acre-feet per year)	Adjacent hydrographic areas and referenced reports[2]
1	Arrow Canyon Range	400 *	Garnet V., Hidden V., California Wash (R50)
2	Bare Mountain	800	Crater Flat, Amargosa Desert, Oasis V. (R14, R54)
3	Belted Range	4,400	Buckboard Mesa, Groom Lake V., Kawich V., Yucca Flat, Penoyer V. (R54, R60)
4	Bullfrog Hills	700	W. Amargosa Desert, Oasis V. (R14, R54)
5	Cactus Range	700	Gold Flat, Ralston V., S. Stone Cabin V. (R12, R54)
6	Cedar Range	400	Meadow V. Wash (R27)
7	Clover Mountains	5,500	Meadow V. Wash, Tule Desert, Virgin River V. (R27, R51)
8	Delamar Mountains	1,500 *	Delamar V., Coyote Spring V., Meadow V. Wash (R16, R25)
9	Desert Range	5,200 *	Groom Lake V., Las Vegas V., N. Three Lakes V., S. Three Lakes V., N. Tikapoo V., S. Tikapoo V. (R54, B44)
10	Grapevine Mountains	1,100	W. Amargosa Desert, Sarcobatus Flat (R14, R10)
11	Groom Range	2,100 *	Groom Lake V., N. Tikapoo V., Penoyer V. (R54, R60)
12	Halfpint Range	200 *	Frenchman Flat, Yucca Flat (R54)
13	Kawich Range	3,400	Kawich V., Gold Flat, S. Stone Cabin V. (R12, R54)
14	Las Vegas Range	3,100 *	Coyote Spring V., Las Vegas V., Hidden V., Garnet V. (R25, R50, B44)
15	McCullough Range	1,000	Las Vegas V., El Dorado V., Ivanpah V. (B44, R46, B44)
16	Meadow Valley Mountains	1,400 *	Coyote Spring V., Meadow V. Wash (R25, R27)
17	Mormon Mountains	2,100	Meadow V. Wash, Tule Desert, Lower Virgin River V. (R27, R51)
18	Pahranagat Range	2,700 *	N. Tikapoo V., S. Tikapoo V., Pahranagat V. (R21, R54)
19	Pahroc Range	1,300 *	Delamar V., Pahranagat V. (R21, R16)
20	Pahute Mesa	4,000	Buckboard Mesa, Gold Flat, Kawich V., Oasis V. (R54, R10)
21	Pintwater Range	3,300 *	Groom Lake V., Indian Springs V., N. Three Lakes V., S. Three Lakes V. (R54)
22	Quinn Canyon Range	1,300 *	Penoyer V. (R60)
23	Sheep Range	11,000 *	Las Vegas V., Coyote Spring V., S. Tikapoo V. (B44, R54, R25)
24	Shoshone Mountains	800	Buckboard Mesa, Jackass Flat, Yucca Flat (R54)
25	Skull Mountains	200	Jackass Flat, Yucca Flat (R54)
26	Spotted Range	1,100 *	Frenchman Flat, Indian Springs V. (R54)
27	Spring Mountains	72,000 *	Amargosa Desert, Las Vegas V., Pahrump V., Mesquite V., Ivanpah V., S. Three Lakes V., Indian Springs V. (R14, R54, R46, B44, W2279)
28	Timber Mountain	400	Buckboard Mesa, Oasis V. (R54, R10)
29	Timpahute Range	1,200 *	N. Tikapoo V., Penoyer V. (R54, R60)
30	Worthington Mountains	900 *	Penoyer V., Garden V. (R18, R60)
31	Yucca Mountain	400	Crater Flat, Jackass Flat, Oasis V. (R10, R54)
	TOTAL (rounded)	[3]140,000	

[1] Asterisk indicates recharge estimate for mountain range within central corridor. Estimate based on published estimates of recharge in adjacent hydrographic areas, prorated for appropriate mountain ranges.

[2] Adjacent hydrographic area numbers in ( ) refer to references cited: B44, Harrill (1976); R10, Malmberg and Eakin (1962); R12, Eakin (1962); R14, Walker and Eakin (1963); R16, Eakin (1963a); R18, Eakin (1963b); R21, Eakin (1963c); R25, Eakin (1964); R27, Rush (1964); R36, Rush and Huxel (1966); R46, Glancy (1968); R50, Rush (1968); R51, Glancy and VanDenburgh (1969); R54, Rush (1971); R60, Van Denburgh and Rush (1974); W2279, Harrill (1986).

[3] Carbonate-rock aquifers of southern Nevada also receive about 21,000 acre-feet per year of subsurface inflow from northern parts of White River flow system, in east-central Nevada. Thus, total recharge plus subsurface inflow is about 160,000 acre-feet per year.

Table 8. Regional springs and areas of subsurface outflow through carbonate-rock aquifers

Map No. (fig. 25)	Name or area	Type of discharge	Total discharge (acre-feet per year)	Discharge derived from east-central Nevada (acre-feet per year)
A	Muddy River Springs	Springflow	36,000	14,000
B	Blue Point Spring	Springflow	240	0
C	Rogers Spring	Springflow	920	0
D	Frenchman Mountain	Underflow toward Colorado River	2,100	0
E	Pahrump Valley	Underflow to California	18,000	0
F	Ash Meadows	Springflow and evapotranspiration	17,000	6,800
G	Amargosa Desert	Underflow to Death Valley	[a]3,000	0
H	Grapevine Canyon	Underflow to Death Valley	400	0
TOTAL (rounded)			[b]77,000	21,000

[a] Does not include underflow (amount unknown).

[b] Total does not include flow from Hiko, Crystal, and Ash Springs in Pahranagat Valley, which discharge about 26,000 acre-feet per year (as measured during this study) near north edge of southern Nevada study area (fig. 1). This discharge is just inside study-area boundary but virtually all springflow derived from outside study area (Eakin, 1966, p. 265; Kirk and Campana, 1988, p. 32-34).

a discharge area (however indirect the path). Stated another way, the quantity of water in storage in a given area does not significantly change with time in a system in dynamic equilibrium. Therefore, unless the carbonate-rock aquifers of southern Nevada are not in dynamic equilibrium, the quantity of discharge from the carbonate rocks (observed at springs or inferred as underflow) must equal to annual net inflow and recharge to the aquifers. There can be no more water entering the study area unless some other mechanism of discharge has been overlooked, and unless the carbonate and basin-fill budgets are to be managed as a single system, the sustained yield of the carbonate-rock aquifers can be no greater than this discharge directly from them. Development of flow through the carbonate-rock aquifers on a sustained basis, however, ultimately would deprive some springs of some of their natural discharge. Thus, the sustained yield may be much smaller than the quantity of water directly discharging from the carbonate-rock aquifers, depending on how effects and development are managed and how "sustained yield" is ultimately defined.

Table 9. Estimated ground-water budget for entire study area

[All rates in acre-feet per year]

Inflow	
Locally generated recharge (table 7)	140,000
Subsurface inflow from the north (table 7)	21,000
Total	161,000
Outflow	
Discharge at regional springs and areas of subsurface outflow (table 8)	77,000
Evapotranspiration from basin fill, excluding that supplied by regional springs, Virgin River, or Colorado River (from published estimates for individual basins)[1]	88,000
Total	165,000

[1] Estimates were published in Nevada Division of Water Resources reports in the Reconnaissance and Bulletin series, and summarized by Scott and others (1971, table 3).

Development and Sustained Yield

If the basin-fill and carbonate-rock aquifers are managed together (locally or regionally), then the overall sustained yield of the area would be closer to the total recharge from mountains in the area plus regional inflow. This total is estimated to be about 160,000 acre-ft/yr. Joint management of the basin-fill and carbonate-rock aquifers would mean deciding where effects would be allowed to develop (in the carbonate rocks, in the basin fill, or both), where water would be extracted (again, in the carbonate rocks, in the basin fill, or both), and what degree of interaction between aquifers would be accepted. An understanding that precipitation in or near the

Table 10. Estimated ground-water budget for the central corridor

[All rates in acre-feet per year]

Recharge source	Rate	Discharge mechanism	Rate
Mountains of southern Nevada	110,000	Regional springs	54,000
Subsurface inflow from east-central Nevada	21,000	Subsurface outflow	about 23,000
		Discharge from basin fill under natural conditions	about 50,000
Sum of estimated rates	131,000		127,000
TOTAL RECHARGE (rounded)	130,000	TOTAL DISCHARGE (rounded)	130,000

mountains of southern and eastern Nevada (and vicinity) is the one source of all natural recharge to both the basin-fill and carbonate-rock aquifers of southern Nevada is needed to manage the resources for optimum effects. Confusion can arise when water from that common source is found in several different places as it moves through the different aquifers. For example, the water that recharged a given area may be present as infiltration of snowmelt, ground water flowing through a carbonate mountain block, ground water in a basin-fill aquifer, ground water in deeply buried carbonate-rock aquifers, and as springflow or evapotranspiration. Double-accounting or double-allocating of the water would be very easy given the complex and unseen interactions between aquifers. However, this would lead to eventual overallocation of the joint resource. Sustained development of the natural resource ultimately will be limited by the quantity of water supplied from their shared source of recharge—precipitation in the mountains of the carbonate-rock province.

Uncertainties in the Water Budgets

The ground-water balances developed in the preceding section contain several sources of uncertainty: empirical estimates of recharge, uncertain rates of ET, and relatively small errors in measuring springflow. The largest sources of uncertainty, however, are probably in the estimates of subsurface outflow to other States, the potential for subsurface inflows besides that from east-central Nevada, and the rough estimates of recharge to mountain ranges of the area. The potential for errors in estimates of recharge in mountains of the area and for subsurface inflow to southern Nevada near the Littlefield Springs in Arizona will be discussed in this section. Subsurface outflows are the most difficult budget components to identify and quantify. Four such outflows are listed in table 8: flow toward the Colorado River, flow toward Death Valley at Ash Meadows, flow toward the southern Death Valley area from Pahrump Valley, and flow towards northern Death Valley. This section also will summarize evidence for the occurrence and approximate magnitude of these outflows.

Potential for Additional Recharge in Mountains

Current estimates of recharge to the mountains of southern Nevada are based on relatively simple empirical methods developed during reconnaissance studies of basins throughout Nevada. These methods do not take geology and micro- or local meteorology into account, but rather are based on estimates of average annual precipitation or land-surface altitude. Despite these limitations, the estimates have proved useful and surprisingly robust in the face of regional studies subsequent to the original estimates (Harrill and others, 1988; Dettinger, 1989b).

The largest source of recharge in southern Nevada is the Spring Mountains (table 7), which receive about 72,000 acre-ft of recharge per year. The Sheep Range is a distant second, receiving about 11,000 acre-ft/yr. As a result, a 10-percent error in the estimate for the Spring Mountains would dwarf any other individual error in other ranges. Fortunately, because the range is near two highly developed basins, recharge has been studied in some detail. Models of ground-water flow in basin-fill aquifers of Las Vegas Valley (east of the Spring Mountains) and Pahrump Valley (west of the range) have been developed in previous studies, and provide additional estimates of recharge from the range (Harrill, 1976, 1986; Morgan and Dettinger, 1994).

Calibration of the model of flow in Las Vegas Valley required about 10,000 acre-ft/yr more recharge from the Kyle Canyon area of the Spring Mountains than estimated by the empirical methods (Harrill, 1976,

p. 50; Dettinger, 1989b). Recharge to the valley was much less than expected from the Sheep Range, north of Las Vegas. The higher than expected recharge to the valley from the Spring Mountains is attributed to the high altitude areas of the mountains where significant winter snowpacks accumulate and where underlying permeable carbonate rock facilitates copious and rapid percolation of ground water. Thus, the Spring Mountains may indeed receive much more recharge than would be expected from mountains of similar altitude in other parts of the State. The meager recharge from the Sheep Range is probably caused by local geologic barriers that direct flow north and east through carbonate-rock aquifers toward the Muddy River Springs rather than south toward Las Vegas Valley, and is in agreement with water budgets for the Muddy River Springs developed in this study (Kirk and Campana, 1988; Thomas and Welch, in press). A subsequent three-dimensional flow model of the valley (Morgan and Dettinger, 1994) required virtually the same conditions for adequate calibration.

The model of Pahrump Valley required recharge rates from the Spring Mountains that are about 10,000 acre-ft/yr higher than previously had been estimated (Harrill, 1986, p. 46).

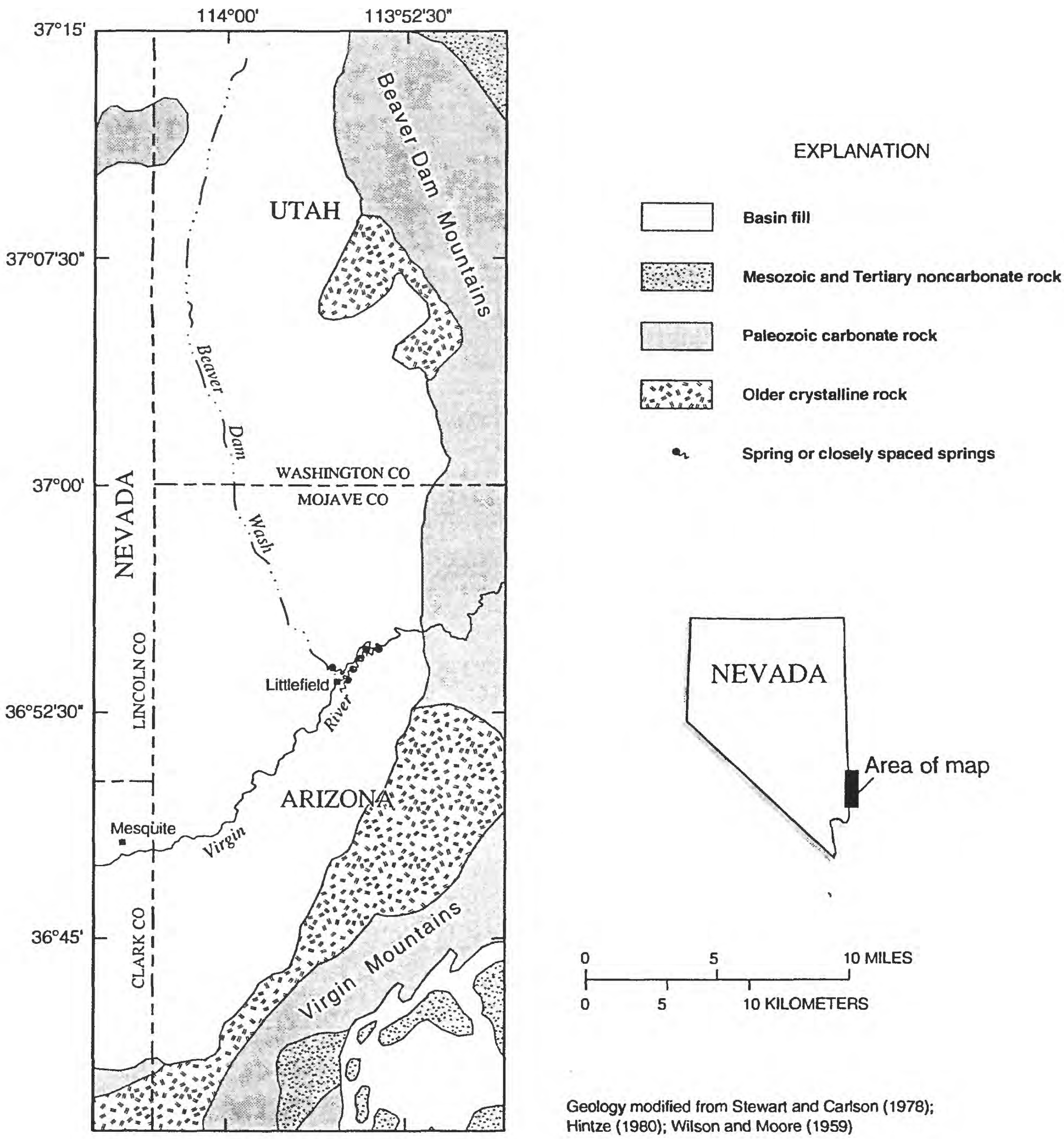
Together these models indicate about 40 percent more recharge to the Spring Mountains than would be estimated using empirical methods. This additional recharge is included in tables 7, 9, and 10, and is detected by calibrations that attempt to adjust the models to match observed discharge rates and observed water levels, simultaneously. A similar error in all other estimates is unlikely since the additional recharge is probably related to large snowpacks that develop on the Spring Mountains and nowhere else in southern Nevada. However, if the empirical estimates were generally 40 percent low in southern Nevada, then an additional 27,000 acre-ft/yr of recharge would be circulating through the different aquifers of southern Nevada. Perhaps 15,000 acre-ft/yr of that additional water would be recharged in the mountains of the central corridor where they might enter the large-scale carbonate-rock aquifers. Nearly all of that water, if it existed, would be discharging from the basin-fill aquifers or to the Death Valley area. The alternative—discharging from the carbonate-rock aquifers in areas other than the Death Valley area—is mostly precluded because other avenues of discharge from the carbonate-rock aquifers are not observed (no large quantity of springflow is being missed and ET is from basin-fill aquifers, not carbonate-rock aquifers, in southern Nevada) or are precluded by geologic barriers to outflow (which will be discussed below). Even if as much as 15,000 acre-ft/yr additional recharge is circulating to eventually discharge from the basin-fill aquifers (perhaps after flowing through carbonate-rock aquifers for a time), the water is probably already accounted for in water budgets of those basins.

Potential for Additional Subsurface Inflow

Regional inflow of ground water into southern Nevada through carbonate-rock aquifers from east-central Nevada has been studied and quantified repeatedly since the 1960's (Eakin, 1966; Winograd and Friedman, 1972; Welch and Thomas, 1984; Kirk and Campana, 1988; Thomas, 1988). Recent studies (Welch and Thomas, 1984; Kirk and Campana, 1988; Thomas, 1988) have tended to generally corroborate each others' budgets. The reader is referred to the previous studies for discussions of the uncertainties involved in the present estimates of inflow from east-central Nevada (especially, Kirk and Campana, 1988). This section will present evidence that other sources of inflow through the carbonate-rock aquifers from less intensely studied areas are unlikely.

Subsurface inflows can be ruled out along most of the boundaries of the present study area because water levels in the study area are higher than those of Death Valley and vicinity and the Colorado River (pl. 2). The one other area where large quantities of ground water might flow into the carbonate-rock aquifers of southern Nevada is along the eastern edge of the Great Basin region near the town of Littlefield, Ariz. The following review of hydrology and geology indicates that such inflow is unlikely.

At Littlefield, Ariz. (fig. 26), about 10 mi east of the Nevada border on Interstate 15, springs and seeps are estimated to discharge about 50,000 acre-ft/yr of water to the Virgin River (Glancy and Van Denburgh, 1969, p. 37). This flow is believed to be supplied from the Upper Virgin River in southeastern Utah, and from the Beaver Dam Mountains north of Littlefield (Glancy and Van Denburgh, 1969, p. 36; Trudeau and others, 1983), and reaches the springs by flowing through carbonate-rock aquifers of the Colorado Plateau. The Colorado Plateau is the geologic province east of the Basin and Range Province of the Great Basin. The Plateau is east of the Beaver Dam and Virgin Mountains,



Base from U.S. Geological Survey digital data, 1:100,000, 1985
Universal Transverse Mercator projection
Zone 11

Geology modified from Stewart and Carlson (1978); Hintze (1980); Wilson and Moore (1959)

Figure 26. Area of regional ground-water flow near Littlefield, Arizona. Modified from Glancy and Van Denburgh (1969, pl. 1).

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and in this area is underlain by sedimentary rocks similar to those in southern Nevada but that are much less deformed than those of the Great Basin.

Subsurface inflow beyond the springs, however, is unlikely unless it is supplied by downward leakage from the thick basin fill in the Lower Virgin River Valley, which in turn is supplied by downward leakage from the river and springs. The springs near Littlefield probably are located in that particular area because the carbonate rocks pinch out against crystalline basement rock that crop out to the north and south and against the very thick, probably low-permeability basin fill of the Lower Virgin River Valley (fig. 21, section *C-C'*). Both east and west of the Beaver Dam Mountains, the carbonate rocks thicken to 10,000 ft or more, but along much of the western edge of the range, carbonate rocks have been removed by erosion and structural events (Smith and others, 1987, p. 389). The area where the carbonate rocks were removed extends north of Littlefield for about 20 mi and to the south about 40 mi (along the Virgin Mountains; Wilson and Moore, 1959; Stewart and Carlson, 1978; Hintze, 1980). Directly east of Littlefield (fig. 26), no crystalline basement rocks are exposed; the exposed carbonate rocks might explain why the springs are located there. Even though carbonate rocks are exposed between the Virgin and Beaver Dam Mountains, the Littlefield area itself is underlain by basin fill and basement rock; therefore, no carbonate-rock aquifers are present through which ground water could flow.

Potential for Subsurface Outflow to the Colorado River

The principal evidence for ground-water discharge to the Colorado River through the subsurface are water-level gradients and numerical-model results for basin-fill aquifers in the northeast part of Las Vegas Valley. Water levels observed in basin-fill wells just east of Nellis Air Force Base (location shown in fig. 27) infer that flow is toward the base of Frenchman Mountain (Loeltz, 1963, p. Q5). A numerical modeling study of principal basin-fill aquifers of Las Vegas Valley by Harrill (1976, p. 50) inferred subsurface outflow from that same area of 1,200 acre-ft/yr. A subsequent modeling study that simulated more aquifers in the basin required a similar outflow (Morgan and Dettinger, 1994). This outflow is listed in table 8 based on the assumption that it is underflow toward the Colorado River. Another, more regional model (Prudic and others, 1993) showed as much as 10,000 acre-ft/yr flowing toward the Virgin River and Overton Arm of Lake Mead. This model result is as yet unverified in the field. Supporting the regional model results is the regional slope of water levels in carbonate-rock aquifers toward the Colorado River (Thomas and others, 1986).

Locally, however, several lines of evidence imply that little water reaches the river through carbonate-rock aquifers. First, geologic (Longwell, 1936) and topographic mapping of the floor of the present-day Lake Mead basin (including the Overton Arm of Lake Mead) located no large springs or spring deposits (except near Hualapai Wash, Ariz. [fig. 27], where subsurface flow to the lake is from the southeast [Longwell, 1936, p. 1431; Laney, 1977, p. 47; Remick, 1981]). Geologic mapping on the floor of the lake as well as of those parts of Nevada that border it (south and west of the Overton Arm of Lake Mead) shows that the carbonate-rock aquifers do not extend to or under Lake Mead except possibly in isolated areas (Longwell, 1936; Bohannon, 1983, pl. 1). Lake Mead is in that part of southern Nevada (and adjacent States) underlain by Tertiary deposits laid directly on crystalline basement rocks. Geologic and seismic studies have shown that the Overton Arm of Lake Mead is physically and hydraulically isolated from underlying rocks (including carbonate rocks) by a thick and impermeable salt body (fig. 27 in this report; Anderson and Laney, 1975). Finally, historical water balances of Lake Mead rule out large ground-water inflows—that is, inflows greater than about 50,000 acre-ft/yr (Harbeck and others, 1958, p. 11; Smith and others, 1960, p. 103). Thus, it is unlikely that the Colorado River is a sink for large flows through the carbonate-rock aquifers. Undiscovered flows to the Colorado River are likely to be few (if any) and to be of a magnitude comparable to the underflow from Las Vegas Valley at Frenchman Mountain: about 1,000 acre-ft annually.

Basin-fill aquifers along the Muddy River (which is supplied by the Muddy River Springs) and Virgin River conduct some flow toward and into Lake Mead, but outflow to the lake from bedrock aquifers is less likely. Nearly all the flow that is, at regional scales, directed toward the Colorado River is actually discharged from the carbonate-rock aquifers about 25 mi upgradient at Muddy River Springs before reaching the river. The Muddy River Springs, the authors believe, drain all the water flowing in the carbonate-rock aquifer because the carbonate rocks pinch out against geologic barriers to the east and south. One barrier to

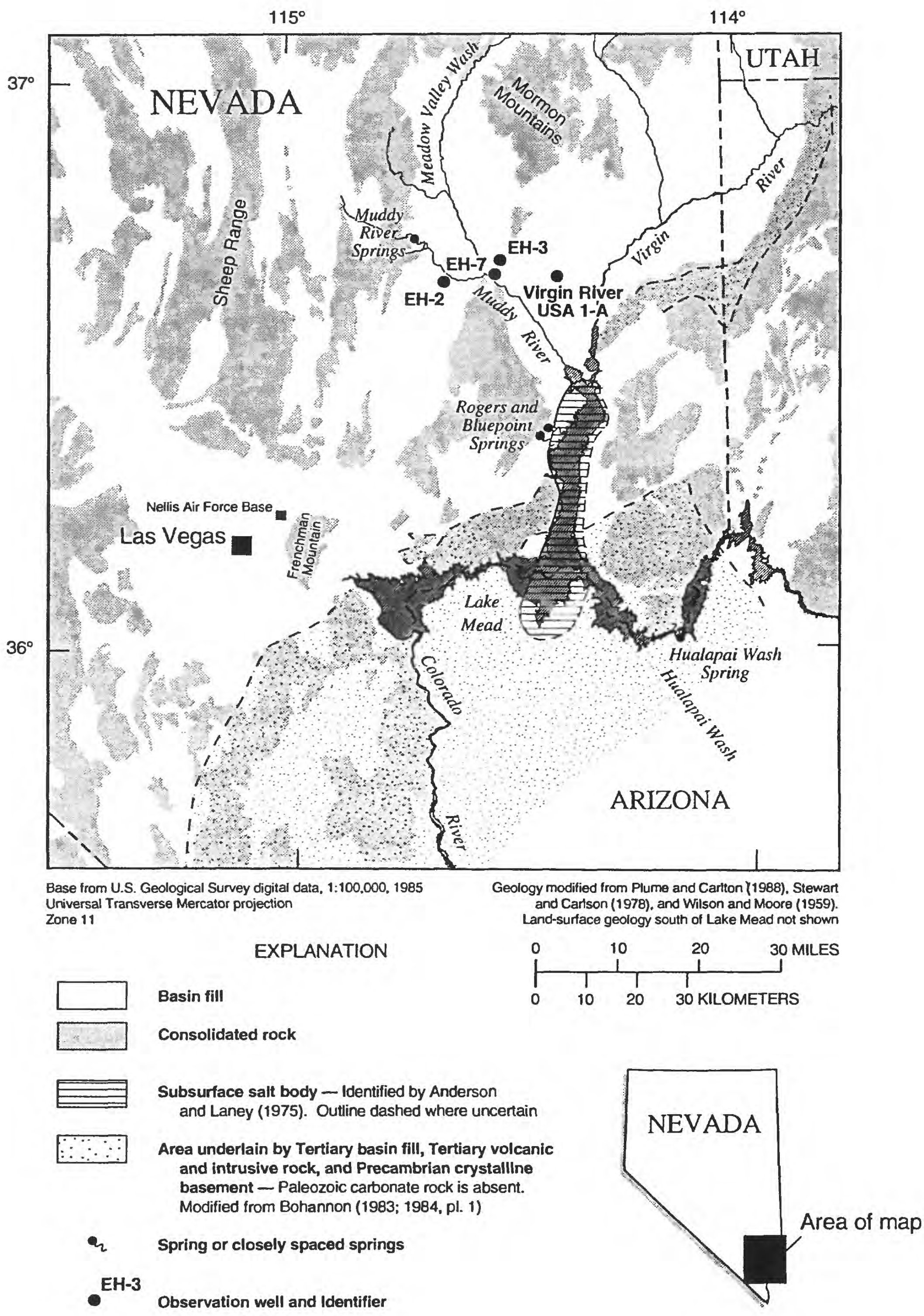

Figure 27. Region of outflow from carbonate-rock aquifers to Lake Mead and Colorado River in southeastern part of study area.

eastward flow is crystalline basement, which shallows to the east from great depth below the springs area and is exposed at the surface along the western base of the Mormon Mountains (fig. 21, section *C-C'*, this report; Wernicke and others, 1985, fig. 15). Another probable barrier is the deep basin that lies just east of the springs. A well was drilled in this basin to a depth of 4,000 ft without penetrating carbonate-rock aquifers (well EH-2 at the Nevada Power Company Reid-Gardner facility near Moapa, fig. 27). The basin-fill sediments and rocks penetrated consisted of lake-bed clay and silt all of which were of low permeability. This deep basin probably extends northward many miles along the Meadow Valley drainage as interpreted from the area's low gravity anomaly (Kane and others, 1979), and in some places probably directly overlies impermeable crystalline-basement rock. Farther east beneath Mormon Mesa, the basin-filling deposits are as much as 7,000 ft thick as measured in the Virgin River USA 1-A oil-test well (fig. 27). Finally, the isotopic composition (deuterium composition, -93 permil SMOW) and temperature (24°C) of water from the carbonate-rock wells EH-3 and EH-7 (fig. 27) are sufficiently different from those of water at Muddy River Springs (deuterium, -98 permil; temperature, 32°C) that the regional flow system that supplies the springs cannot be the source of water in the carbonate rocks east of Meadow Valley Wash (Schroth, 1987, p. 65, 82, 86). Thus, regional flow does not continue past the Muddy River Springs in an easterly direction.

The possibility that regional flow continues south beyond the Muddy River Springs is difficult to eliminate. As indicated earlier, the Overton Arm of Lake Mead appears isolated from underlying aquifers by a large salt body within the basin fill. The volcanic rocks of the Black Mountains area directly north of Lake Mead are underlain by Precambrian crystalline rocks. Finally, the water level in carbonate rocks beneath a 750-mi^2 area around, south, and west of the Muddy River Springs is extremely uniform between about 1,800 and 1,830 ft above sea level. The water levels surrounding this area, in contrast, have much steeper gradients (pl. 2), and thus, give the impression of a dead end in the White River system. One interpretation of this observation is that the area south of the springs offers no outlet for regional flow that trickles past the springs area. If the aquifers of the central corridor are isolated from the Colorado River (as discussed above) then flow that passes the Muddy River Springs area moving south might be virtually ponded at the same water level as the springs with little through flow except to supply leakage to basin-fill aquifers along the Muddy River. This interpretation may be supported by the general uniformity of chemistries and, in particular, isotopic compositions in most of the same area. The chemistry is uniform because the water is scarcely moving through the area and because it is at equilibrium with the aquifer materials. In contrast, upgradient from the Springs, water chemistry and isotopic compositions evolve in readily detectable patterns along flow distances no longer than those in the uniform area south of the Springs (Thomas and Welch, in press).

Another possibility that cannot be discounted, however, is that the flat water levels and uniform chemistries reflect high transmissivities in the area. If transmissivities are very high, water-level gradients would be flat regardless of flow rates, but then a vertical low-permeability barrier would be required to explain the steep gradients outside the uniform area. If the flow rates are large, this alternative requires significant discharge areas somewhere to the south that are at present (1989) unrecognized. The most likely discharge for water flowing southward past the Muddy River Springs are Rogers and Bluepoint Springs, which are warm-water springs that together discharge about 1,100 acre-ft annually from carbonate rocks above the Overton Arm of Lake Mead. Chemistry of water discharging from these springs indicates dissolution of large amounts of gypsum which tends to mask possible similarities to water in rocks further north. Isotopic composition of the ground water (-92 permil deuterium), however, implies that the water is not derived entirely from the White River flow system (as represented by water from the Muddy River Springs at -98 permil deuterium) but at most could contain about 20 percent of that water. To generate water of the isotopic composition of Rogers Spring, that small component of White River water would have to be mixed with 80 percent water derived from recharge in the relatively low mountains of southeastern Nevada. Chemical effects such as gypsum dissolution would not alter the isotopic composition of the water. Discharge at Rogers Spring is best attributed to recharge in either nearby Sheep Range or Mormon Mountains.

Thus, for practical purposes, springflow from the Muddy River Springs, along with upward leakage into overlying basin-fill aquifers in the area, constitutes virtually all the subsurface outflow from the White River flow system into southeastern Nevada. No other large discharge areas have been identified and geologic constraints prevent large-scale flow past the region. This implies that the White River flow system is contained within Nevada, and does not discharge directly into the Colorado River.

Potential for Subsurface Outflow to Death Valley

Ground water may flow toward Death Valley through the carbonate rocks of Nevada from three general areas (pl. 1 and fig. 28): (1) the Grapevine Mountains, (2) Ash Meadows, and (3) south from beneath Pahrump Valley. Geologic barriers limit how far these flows can extend because Death Valley (at least in the south and possibly as far north as the Furnace Creek area) is sparsely underlain by carbonate rocks. Large uncertainties in estimates of the rate of ground-water discharge from the basin fill in Death Valley make it difficult, however, to argue where regional flow in the Ash Meadows flow system ends or how much underflow passes beneath Ash Meadows.

Ground water from beneath Sarcobatus Flat, Lida Valley, and possibly Stonewall Flat flows toward northern Death Valley through consolidated rocks in the Grapevine Mountains (locations shown on pl. 1). Grapevine and Staininger Springs in central Death Valley discharge about 400 acre-ft/yr (Rush, 1968, p. 32), which probably represents the total outflow through consolidated rocks in that area. Few carbonate rocks crop out in the area that is well outside the central corridor (Stewart and Carlson, 1978). Thus, the outflow is probably through isolated blocks of carbonate rock, through some of the noncarbonate rocks, or through thin basin-fill connections between Sarcobatus Flat and Death Valley.

In the Ash Meadows-Furnace Creek area, regionally fed ground water discharges from two lines of springs—at Ash Meadows and from the Furnace Creek area on the eastern flank of Death Valley (fig. 28). Between the two lines of springs, water levels seem to vary smoothly rather than in discontinuous steps (Thomas and others, 1986). The chemistry and isotopic compositions of the water at both sets of springs are similar enough for the water to have originated from the same source (Winograd and Friedman, 1972, p. 3704). Finally, the subsurface geology and aquifer geometries between the two lines of springs are poorly understood and provides considerable latitude for postulating flow paths.

Several authors have postulated that discharge from springs in the Furnace Creek area is fed by water that flows past the springs at Ash Meadows (Winograd and Friedman, 1972, p. 3704; Czarnecki, 1987). This underflow then discharges at Death Valley at several springs and as ET. Czarnecki (1987) hypothesized that the Ash Meadows-Furnace Creek connection is in a carbonate-rock aquifer overlain by flow mostly in basin fill toward Franklin Lake (also known as Alkali Flat; Czarnecki, fig. 3-8). Others, however, have stated that the springs in the Furnace Creek area might be supplied by flow from the north in Amargosa Valley and Pahute Mesa (Waddell and others, 1984, p. 37). Considerable uncertainty with respect to the hydrology and geology controlling underflow precludes any strong conclusions about any of these hypotheses.

In addition to uncertainties about the general occurrence of underflow from Ash Meadows, the rate of underflow from beneath Ash Meadows is uncertain. Springflow is about 3,000 acre-ft/yr from the springs near Furnace Creek (Miller, 1977, p. 27), but many phreatophyte stands and an enormous area of salt pan may account for much more discharge by ET (Miller, 1977, p. 55). The origins of this water may include surface-water inflows (during floods), recharge from surrounding ranges, and regional inflow; all may be present in some unknown proportions. The discharge rates of the regional component could be significantly greater than those estimated in table 8.

Despite uncertainties about the existence and rate of underflow from Ash Meadows to Death Valley, flow and discharge of water from Nevada in areas beyond Death Valley are unlikely. Any underflow must discharge in Death Valley because the valley is a regional topographic and hydrologic low point. Also, the mountain ranges farther north, south, and west are composed of impermeable rocks.

Ground water flows toward the southern Death Valley area from beneath Pahrump Valley. Malmberg (1967, p. 32) estimated that between 5,000 and 15,000 acre-ft/yr of ground water flows southwest from beneath Pahrump Valley in carbonate-rock aquifers. A numerical model of ground-water flow beneath Pahrump Valley suggests that the total outflow is about 18,000 acre-ft (Harrill, 1986, p. 46). This flow is assumed to pass mostly through the carbonate rocks

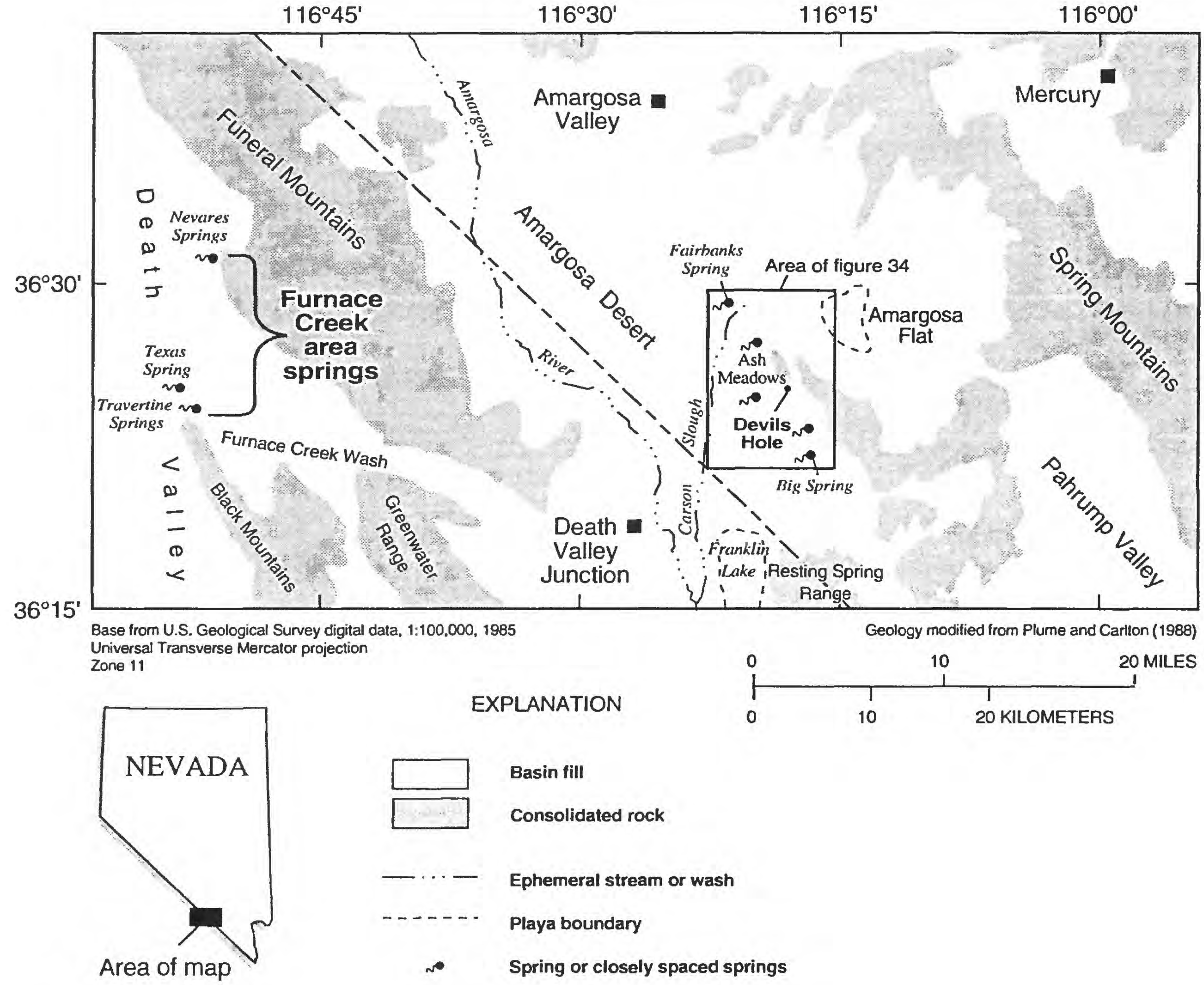

Figure 28. Region of possible ground-water throughflow from Ash Meadows area, Nevada, to Furnace Creek area, California. Modified from Winograd and Szabo (1986, fig. 1) and Plume and Carlton (1988).

of the Nopah and Resting Spring Ranges prior to discharging at warm water springs and phreatophyte stands near Tecopa and Shoshone, Calif., and into the lower reaches of the Amargosa River, also in California (Harrill, 1986, p. 15; locations shown in fig. 29). These estimates are based on interpretations and simulations of ground-water conditions in the basin-fill aquifers of Pahrump Valley.

Flow beyond the Amargosa River area almost certainly is impeded by the structural removal of the carbonate-rock aquifers. West of the Nopah Range, Wright and others (1981) and Wright and Troxel (1973, p. 404 and 406) show that in Resting Spring Range only small fragments of carbonate rock overlie predominantly older noncarbonate rocks and that mountain ranges even farther west consist of low-permeability granites and crystalline basement (fig. 21, section *E-E'*). Together these rocks constitute a formidable barrier to westward or southward flow beyond the Amargosa River.

Summary of Budget Uncertainties

Overall, this discussion of subsurface inflows and outflows to the carbonate-rock aquifers of southern Nevada implies that considerable uncertainty exists in the estimates in table 10. In each area, however, with the exception of the possible Ash Meadows-Furnace Creek connection, uncertainties are not large enough to allow for major regional flow paths to have been missed in previous analyses. Each of the other major discharge areas discussed (Muddy River Springs, the Tecopa-Shoshone area, and the Littlefield Springs) appear to comprise nearly all of the regional flow into

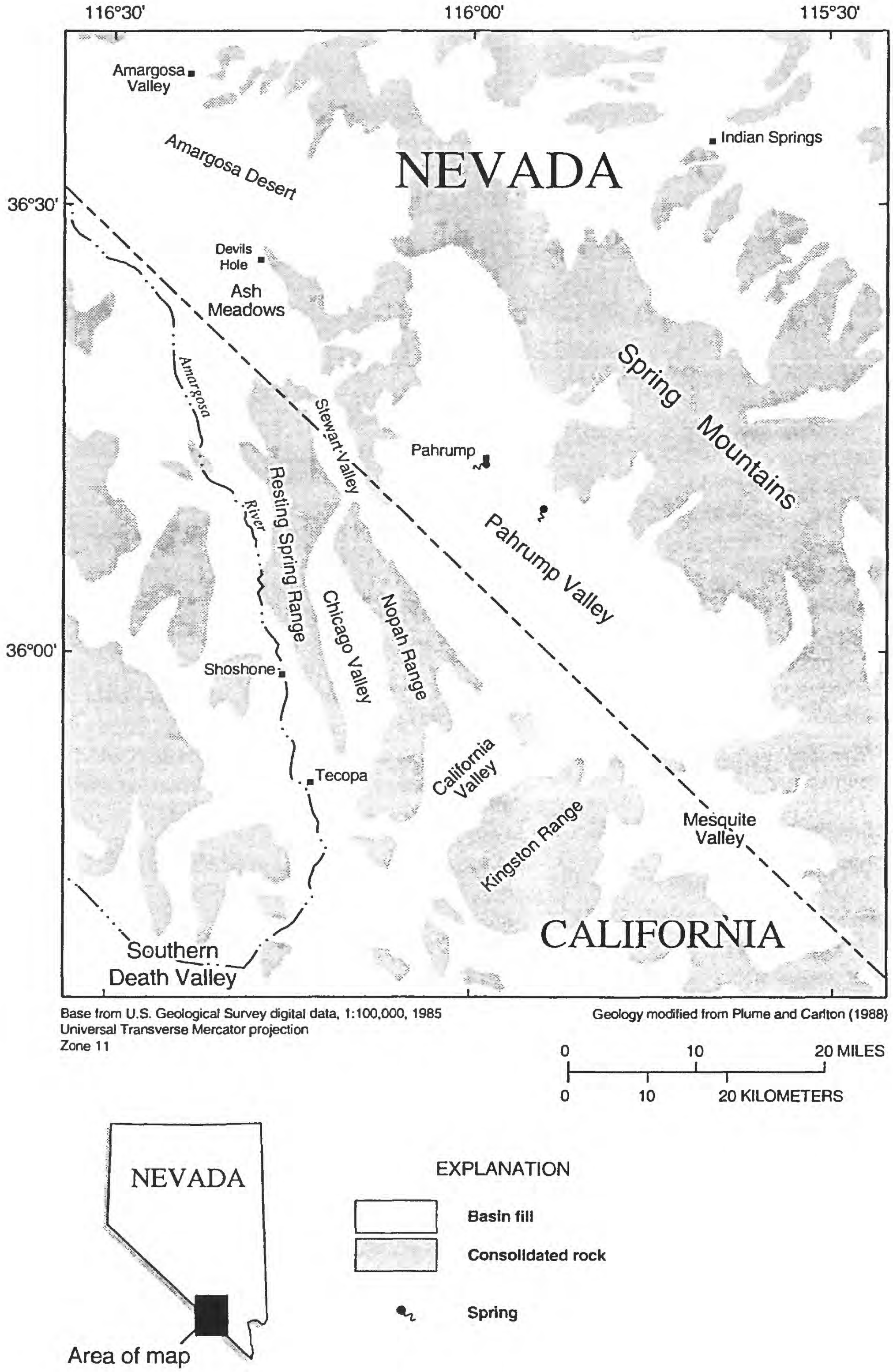

Figure 29. Area of underflow from Pahrump Valley, Nevada, to southern Amargosa River area, California. Modified from Harrill (1986, fig. 1).

each area. In each setting, geologic conditions have dictated that little water passes beneath the major springs and discharge areas to be consumed elsewhere. Thus, the same conditions that result in major spring discharges also account for virtually all flux through the regional flow systems.

Storage of Water in the Carbonate Rocks

Estimates of the total quantity of water stored as a one-time reserve in the carbonate-rock aquifers are based on current (1989) understanding of the present-day geometry of the rock sequences and of how much water is actually stored per cubic foot of rock.

The geometry of the layers and blocks of carbonate rock defines the maximum volume in which water may be stored. In order to calculate this volume, the thickness of carbonate-rock sequences must be estimated. Thickness is dependent on the geologic history of the area: the original depositional thicknesses, the amount of compressional thickening during Mesozoic time, the amount of extensional thinning during late-Tertiary time, and the amount of erosional thinning of the section during geologic time. Surficial geology, regional and local models of geologic structures at depth, and interpretations of geophysical data are all used to estimate present-day thicknesses in selected areas. Once the thickness is determined or estimated, it can be multiplied by an area for which that thickness is representative to estimate the total volume of rock.

The volume of water stored in the rocks is a small fraction of the total volume of rock. Water is stored in fractures, other secondary openings in the rock mass, and in tiny interstitial or intercrystalline (primary) openings in the unfractured rock. Secondary openings can comprise about 0.1 percent of total rock mass, whereas primary openings probably comprise about 1-10 percent. The volume associated with these two types of openings constitute the quantity of water available if all extractable water were drained from a given volume of carbonate-rock aquifer.

The ratio of the volume of water to the volume of saturated rock is commonly called the specific yield of the carbonate-rock medium. The total quantity of water stored in the aquifer is equal to the volume of the aquifer multiplied by its specific yield. If the spatial variability of the yield can be quantified, then stored water equals the integral of this product over the entire aquifer volume.

Pumping all the water stored in aquifers as thick as the carbonate-rock aquifers is neither economical or practical. The quantity of water that could be extracted from one-time storage in the carbonate-rock aquifers is limited by the depth of rock that can be dewatered before technical, economic, or environmental problems are experienced. Thus, the carbonate-rock province probably contains much more water than could ever be extracted. To estimate the fraction of the total stored water that could be realistically extracted for use, the fraction of the rock volume that could be dewatered has to be estimated. This fraction of rock volume would then be multiplied by the total volume of stored water to estimate practical limits of the stored-water resource. In this report, the fraction of rock volume that could be dewatered will be estimated in terms of the fraction of rock thickness that might under extreme conditions be dewatered.

Volume of Water Stored in a Unit Volume of Rock

In a confined aquifer, the measure of the volume of water stored in a unit volume (cubic foot) of water-saturated rock is the storage coefficient. In an unconfined aquifer, the measure is the specific yield of the aquifer. These numbers are generally quite different and are the result of differing storage processes. As a consequence, the volume of stored water that must be extracted to dewater a given thickness of an aquifer depends on, among other things, whether the aquifer is under unconfined ("water table") or confined ("artesian") conditions. Heath (1983, p. 6) describes unconfined aquifers as aquifers where water only partly fills the transmissive rocks so that the upper surface of water in the openings in the rocks can rise and decline in response to pumping. In contrast, a confined aquifer is one where water fills the entire thickness of transmissive rocks up to a layer or cap that restricts vertical flow. In confined aquifers, water pressure—not water level—rises and declines in response to pumping. Water added to storage is accommodated by compressing both the rock matrix and water into smaller volumes. Accomplishing this compression requires high pressures in the aquifer that force water to rise in a well to levels above the low-permeable cap and in some places to flow freely from a well that penetrates the aquifer. In contrast, water added to storage under unconfined conditions raises the water table by filling previously dry (or unsaturated) pores with water increasing the volume of the aquifer. The level of water in wells rises as high as the water table (top of the aquifer) under these conditions.

On the basis of a standard equation for estimating the storage coefficient of a confined aquifer from Bear (1979, p. 107) and assuming physically reasonable values for physical and mechanical properties of the carbonate-rock aquifers, the estimated storage coefficient of a confined part of the carbonate-rock aquifers is about 0.0001. In making this estimate, total porosity is assumed to be about 10 percent, the reciprocal of the modulus of elasticity of the rock itself is assumed to be about 10^{-10} ft^2/lb, the reciprocal of the modulus of elasticity of the water is assumed to be 2.3×10^{-8} ft^2/lb, and the aquifer thickness is assumed to be about 3,000 ft. A storage coefficient of 0.0001 indicates that if an acre-foot of stored water were extracted from a 1-acre block of carbonate rock that was responding as an isolated and confined aquifer, fluid pressure in the block would decline by the equivalent of the weight of a column of water 10,000 ft high. Consequently, in an area where the carbonate-rock aquifer is confined, relatively little water needs to be extracted to cause large reductions in the water pressure in the aquifer and the consequent water level in wells.

Under unconfined conditions, water to supply a pumped well is derived principally from drainage of water from previously saturated openings (matrix and secondary) in the aquifer (following a short initial period before drainage becomes fast enough to supply the demand). The specific yield is equal to that fraction of the porosity near the water table that is drained as the water table declines in the aquifer. Some pores are left undrained—pores that are poorly connected or unconnected to the permeable parts of the medium. As a result, the specific yield is always less than or equal to the total porosity of the aquifer and generally is equal to an "effective" porosity of the aquifer (Bear, 1979, fig. 5-4).

Estimating effective porosity of the carbonate-rock aquifers is difficult. Effective porosity is directly dependent on the fraction of total porosity that yields water to pumped wells. Field observations and laboratory measurements indicate that the large-scale average fracture porosity of carbonate rocks ranges from less than 0.01 percent to about 0.5 percent with a median that is probably on the order of 0.1 percent (Winograd and Thordarson, 1975, p. C17; Alan M. Preissler, U.S. Geological Survey, written commun., 1987). Laboratory measurements and geophysical logs, in contrast, indicate that the matrix porosity is commonly between 0 and about 20 percent with a median around 5 percent (Winograd and Thordarson, 1975, p. C17; Berger and others, 1988). Of this total matrix porosity, Winograd and Thordarson (1975, p. C17) determined that 20 percent is "effective" in rock samples from the Nevada Test Site area.

Water-level responses to constant pumping stress provides important clues as to the volume of storage that might be yielded by long-term pumping of carbonate-rock aquifers. Measured drawdowns in pumped wells completed in carbonate rocks in Nevada have two or more distinct breaks in slope when plotted against the logarithm of time since pumping began. As examples specific to this study, figure 30 shows drawdown curves for two of the wells tested as part of the present study. Distinct breaks in slope are evident in the upper curve (for MX well CE-DT-6) and a similar break is indicated in the lower curve (for well CSV-2). Similar responses also have been noted in aquifer tests in carbonate rocks of the Nevada Test Site area (several wells: Winograd and Thordarson, 1975, p. C20; well UE16d: Dinwiddie and Weir, 1979; well UE25p#1: Craig and Robison, 1984, p. 26). This response is characteristic of fractured porous media where flow to the well is through high-permeability, low-storage fractures but where most water is from the low-permeability, high-storage pores of the unfractured rock (Streltsova, 1988, p. 368). Media of this sort are called "double-porosity" media (Streltsova, 1988, p. 370), and would be expected to yield water from both the fractures and matrix over the long term—initially water is drawn from the fractures and later water is extracted from pores in the unfractured rock. Thus, the overall volume of stored water that might be yielded by long-term pumping of the carbonate-rock aquifers is most likely to approximate that contained in the effective part of the primary porosity (and is not limited to the fracture porosity).

Two alternative explanations for the breaks in the drawdown curves are: initial depletion of borehole storage or a transition from confined storage to gravity drainage as drawdowns develop. In large-diameter wells, the first storage depleted by the pumped well is the water in the borehole itself, and this can lead to breaks similar to those shown in figure 30. However, other characteristics of such initial depletion include a unit-slope when the drawdowns are plotted against time on a log-log graph. This slope is not found in the wells completed in the carbonate rocks being discussed

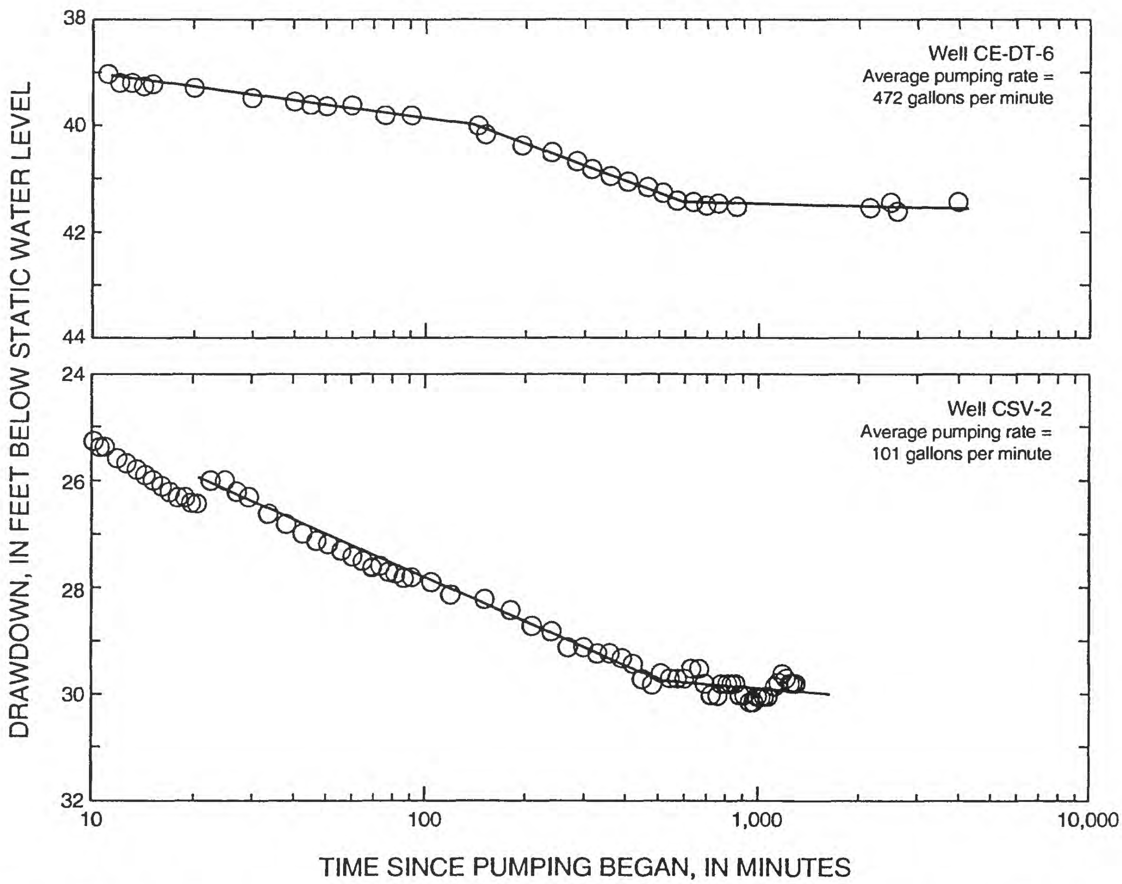

Figure 30. Water-level drawdown during aquifer tests at two carbonate-rock wells in Upper Muddy River Springs area. Solid lines indicate generalized straight-line slopes of drawdown curves.

and so is considered an unlikely explanation for the breaks. After moderate-to-long periods of pumping, water flowing to a pumped well in an unconfined aquifer is from gravity drainage of pores (and fractures) at the water table. Initially, however, water is derived from release of pressure (as in a confined aquifer) because gravity drainage is not instantaneous. This change in the storage mechanism that is being depleted by a well results in a set of changes in slope of drawdowns similar to that discussed above—although usually not as abrupt. This explanation for the breaks is inappropriate in several of the wells because they are clearly open to confined aquifers. Aquifer tests for wells 88-66 (Winograd and Thordarson, 1975, p. C30) and UE-25p#1 (Craig and Robison, 1984, p. 25-26) provide clear examples of the breaks in the drawdown curves and yet are confined by confining units that are hundreds or thousands of feet thick. Breaks in the slope of drawdown curves for other wells may result from a combination of dual-porosity response and delay of gravity drainage.

The preceding discussion is based on indirect inferences. Direct measurements of specific yields of the carbonate-rock aquifers in Nevada would be useful and would add reliability to storage calculations that follow. However, direct measurements of long-term storage properties of aquifers are difficult and expensive. Direct measurement of specific yield requires a aquifer test with observations of water levels in both a pumped well and an observation well. In general, the few production wells that draw water from Nevada's carbonate-rock aquifers are not accompanied by observation wells. Only three "direct" measurements of specific yield in the carbonate rocks of Nevada are available. A 30-day aquifer test at well CE-DT-5 (as observed from well CE-DT-4 that is 300 ft away) yielded an estimated storage coefficient of about 0.15

(based on data reported in Ertec Western, Inc., 1981; and Berger and others, 1988, table 9). A 30-day aquifer test made with several observation wells at the Fad Shaft near Eureka in east-central Nevada yielded storage coefficients between 0.00064 and 0.0014 (Stuart, 1955, table II). Finally, analyses of tidal and barometric fluctuations of water levels in well UE25-p#1 near Yucca Mountain indicated that the aquifer is confined (probably by the 5,000 ft of volcanic rocks overlying the carbonate rocks) and that the storage coefficient for the aquifer is 0.0003 (Galloway, 1986). Galloway also estimated the total porosity of the aquifer as 9 percent. In the future, improved estimates of the volume of water stored in the carbonate-rock aquifer will depend largely on collection of more data describing specific yields.

Despite the limited data available, and solely on the purposes of deriving estimates of the overall quantity of water stored in the carbonate-rock aquifers of southern Nevada, the specific yield or storage coefficient is assumed to be 0.01. This estimate actually is in the center of the probable range, and specific yield could be an order of magnitude smaller or larger in any given area. This assumed specific yield is about 100 times greater than the confined-aquifer storage coefficient values, and will be used in these storage calculations for two reasons:

1. In an isolated block, the quantity of water available from dewatering the top 100 ft of the aquifers under unconfined conditions would be equal to that required to lower water pressure under confined conditions by 10,000 ft (which would be enough to convert aquifers from confined to unconfined conditions almost everywhere). Thus, the quantity of water that would be available from dewatering a confined block of rock is so much smaller than the quantity available from an unconfined block that most water available by dewatering is likely to be from unconfined conditions.

2. At the regional scales to be considered here and when the issue being considered is the one-time storage of water, the carbonate-rock aquifers probably will respond as if unconfined. This hypothesis is based on the regional-scale geologic heterogeneity of the Basin and Range Province. In this block-faulted province, it is unlikely that any one confining unit is laterally continuous enough to maintain regional-scale confinement of the aquifers. The ubiquity of high-angle faults that function as vertical flow paths through confining units reduces the chance for large-scale confinement even where layers are present that otherwise could confine the aquifers.

Thickness of Aquifers

Most water stored in the carbonate rocks of southern Nevada is localized within the central corridor. The thickness of carbonate-rock aquifers within the central corridor of southern Nevada is estimated in this section. Blocks of carbonate rock outside the corridor are apt to be small and isolated and many are relatively thin. Some estimates of thickness of rocks in these blocks are presented in this section but the total volume of carbonate rocks (or water stored) in these isolated blocks and in areas assumed to contain poor-quality water is not estimated.

Thickness of the carbonate rocks within the central corridor ranges from several thousand feet to possibly greater than 19,000 ft. Variations of thickness are smooth in some areas and abrupt across structures in others. No one structural style or structural feature is predominate in the entire corridor, and at a regional scale, spatial variations of thickness depend on the particular structural features in a given area. The central corridor with some major structures used to delineate broad areas having distinct structural styles or boundaries is shown in figure 31; the structures are described in table 11. These areas are the framework used below for making estimates of representative thicknesses of carbonate rocks.

The few locations and estimates of the total thickness of Paleozoic rocks available in southern Nevada are shown in figure 32 and table 12. The numbered areas on figure 31 are each influenced by an overall structural style (for example, unextended contrasting to highly extended) or a controlling structure (for example, extension along a particular fault system). The thickness in one area is not structurally related to the thickness in any other area. Because of this relative independence among areas, and because thickness estimates are available only in a few locations, estimates of thicknesses of carbonate rock within the central corridor can usefully be combined into generalized estimates of thickness for each area. These generalized estimates (quantitative and qualitative) indicate that the carbonate rocks within the White River flow system (areas 3 and 4, fig. 31) are thickest, and those in areas

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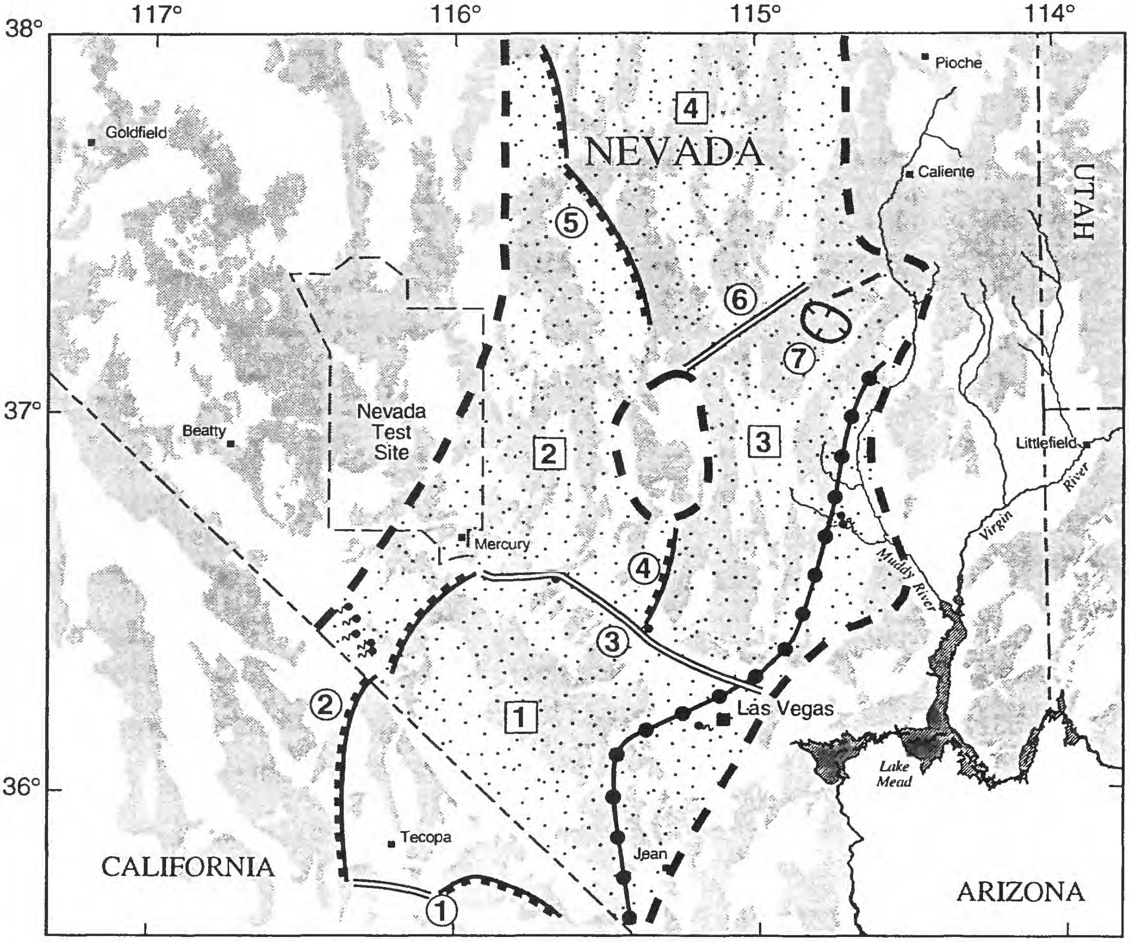

Base from U.S. Geological Survey digital data, 1:100,000, 1985
Universal Transverse Mercator projection
Zone 11

Geology modified from Plume and Carlton (1988), Stewart and Carlson (1978), and Wilson and Moore (1959). Geology in lower left and lower right corners of map area not shown

0 20 40 60 MILES

0 20 40 60 KILOMETERS

EXPLANATION

Basin fill

Consolidated rock

Central corridor of carbonate-rock aquifer in Nevada — Area is underlain by thick sequences of carbonate rock; outside corridor, carbonate rock is thin, or present as isolated bodies

Western limit of ground water generally too mineralized for most uses

Shear zone or high-angle fault

High- to low-angle normal fault

Outer margin of volcanic complex

1 **Geologic structure listed in table 11**

4 **Storage-estimate area listed in table 13**

Spring or closely spaced springs

Figure 31. Storage-estimate areas and geologic structural boundaries within central corridor of carbonate rock in study area.

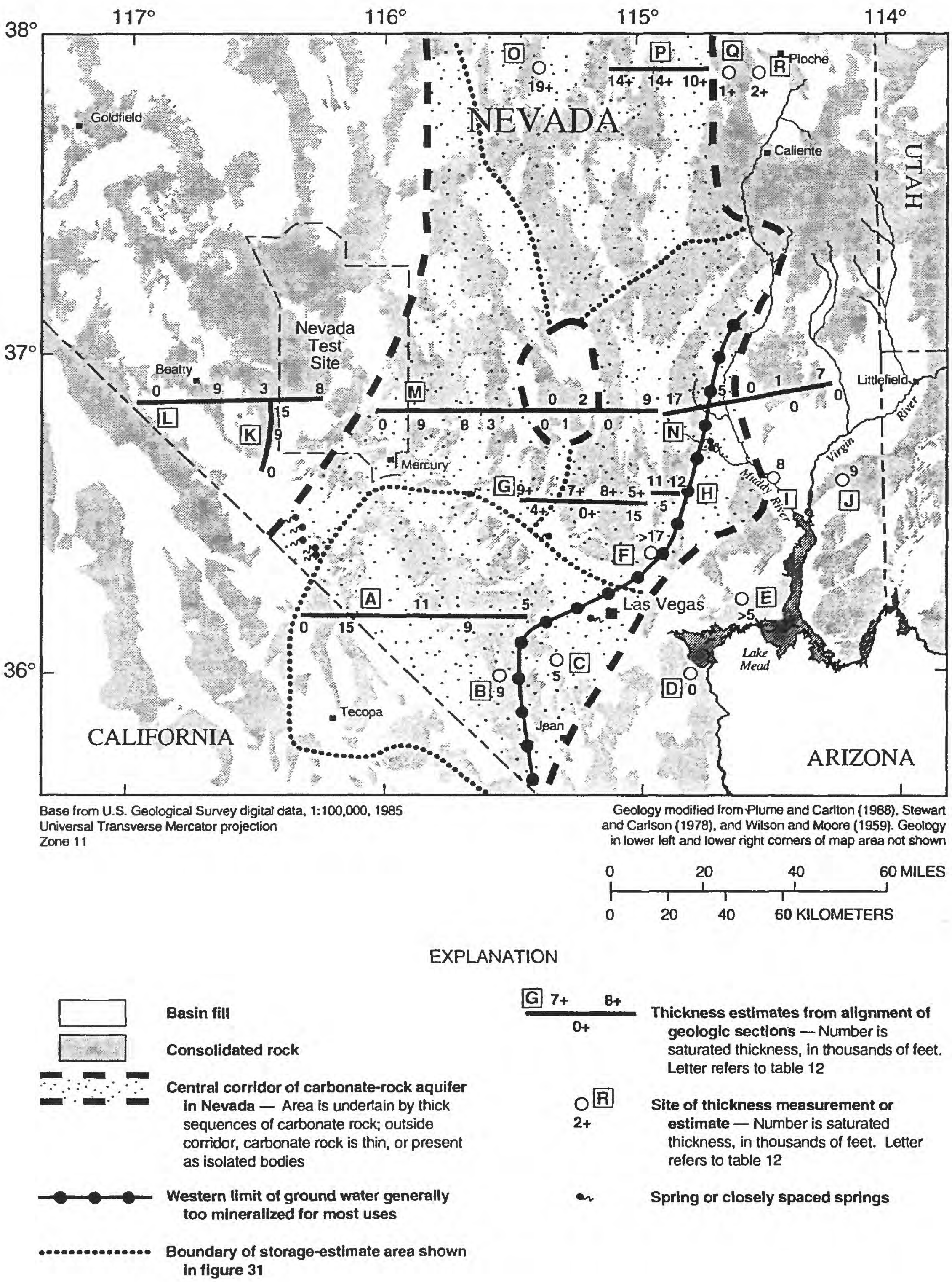

Figure 32. Estimated present-day saturated thickness of Paleozoic rock in study area.

Table 11. Description of structures used to define boundaries of storage-estimate areas

Structure number (fig. 31)	Name	Description	Reference
1	Sheephead fault zone and Kingston Range detachment	Shear zone separating denuded areas to south from thick Paleozoic sections in Nopah Range and Pahrump Valley	Stewart, 1983, figs. 1, 2; Jennings and others, 1962; Jennings, 1961
2	Detachment at Resting Spring Range	Zone of extreme surficial extension and thinning (separating thick Paleozoic sections to east from thinned and missing Paleozoic sections to west)	Stewart, 1983, p. 154; Wright and others, 1981
3	Las Vegas Valley shear zone	Shear zone separating areas of extreme extension to north from much-less extended Spring Mountains to south	Guth, 1981, p. 769
4	Sheep Range detachments	Major faults within and along west side of Sheep Range that juxtapose noncarbonate rocks within the range against carbonate rocks to west	Guth and others, 1988; Guth, Petmar Trilobite Breeding Ranch, written commun., 1988
5	Detachment(s) beneath Tikaboo and Penoyer Valleys	Boundary between extensional thinning beneath Tikaboo and Penoyer Valleys and less extended terrains farther east beneath Pahranagat and Garden Valleys	Jayko, 1988; Martin and Bartley, 1989
6	Pahranagat shear zone	Shear zone separating areas that experienced major extension at different times	Axen and others, 1987; Wernicke and others, 1984, p. 48
7	Kane Springs volcanic center	Volcanic center that interferes with present-day ground-water flow in much the same way as a small caldera complex	Ekren and others, 1977

Table 12. Estimated total and saturated thickness of Paleozoic rocks

[Letters A through R refer to locations indicated in figure 32. Symbol: ±, more approximate than most; <, less than; >, greater than. All numbers are estimates. Numbers followed by + are thicknesses or depths to bottom of uppermost thrust sheet; underlying thrust sheets may exist and contain additional Paleozoic carbonate rocks at much greater depths]

Area (and reference)	Thickness of Paleozoic section (feet)	Depth to top of carbonate rocks (feet)	Depth to water (feet)	Approximate saturated thickness of Paleozoic rocks (feet, rounded)
A. Resting Spring - Blue Diamond alignment (Wright and others, 1981; Harrill, 1986, p. 17)				
Resting Spring Range	0	--	--	0
Nopah Range	16,000	0	>1,000	15,000
Pahrump Valley	11,000	4,000	<1,000	11,000
Spring Mountains	11,000	0	500-3,000	9,000
Blue Diamond	5,000	0	200	5,000
B. Spring Mountains aeromagnetic anomaly (Blank, 1988)	10,000	0	500-2,000	9,000
C. Karmarden oil well (Garside and others, 1977, p. 6)	5,000	0	500	5,000
D. River Range (Smith and others, 1987, p. 2)	0	--	--	0
E. Bowl of Fire oil well (Schilling and Garside, 1968, p. 8)	>5,000	1,000	800	>5,000
F. Arrow Canyon oil well (unpublished records from Nevada Bureau of Mines and Geology)	>17,000	1,000-2,000 ±	700±	>17,000
G. Desert - Las Vegas alignment (Guth, 1980, pl. 2)				
Desert Range	10,000 +	0	1,100	9,000 +
Black Hills	5,000 +	0	1,500	4,000 +
Hoodoo Hills	11,000 +	0	4,000	7,000 +
East of Hoodoo Hills	3,000 +	0	4,000	0 +
Sheep Range	12,000 +	0	>4,000	8,000 +
North of Yucca Forest	8,500 +	0	3,500	5,000 +
Las Vegas Range	16,000	0	0-2,000	15,000

Table 12. Estimated total and saturated thickness of Paleozoic rocks—Continued

Area (and reference)	Thickness of Paleozoic section (feet)	Depth to top of carbonate rocks (feet)	Depth to water (feet)	Approximate saturated thickness of Paleozoic rocks (feet, rounded)
H. Las Vegas - Meadow Valley alignment (Dwight Schmidt, written commun., 1985)				
East Las Vegas Range	12,000	0	1,400	11,000
Arrow Canyon Range	7,000	0	2,300	5,000
South Meadow Valley Range	>13,000	0	800	12,000
I. Virgin River USA No. 1-A oil well (unpublished records from Nevada Bureau of Mines and Geology)	8,000	11,000	11,000	8,000
J. Basement outcrops in Virgin Mountains (Longwell and others, 1965, pl. 1)	0	--	--	0
K. Yucca - Lathrop Wells alignment (Robinson, 1985, fig. 9)				
Yucca Mountain	15,000.±	7,000	2,500	15,000 ±
Fortymile Wash	9,000.±	3,000	400	9,000 ±
Lathrop Wells	0	--	?	0
L. Bull Frog - Calico alignment (Scott and Whitney, 1987; R.B. Scott, U.S. Geological Survey, written commun., 1988)				
Bull Frog Hills	0	--	--	0
Bare Mountain	11,000	0	1,800	9,000
Crater Flat	3,000	11,000	1,000 ±	3,000
Calico Hills	10,000	0	1,600	8,000
M. Frenchman - Las Vegas alignment (Guth, 1989, fig. 4-2)				
Frenchman Flat	0	--	--	0
Spotted Range	12,000	0	2,600	9,000
Indian Springs Valley	8,000	2,000±	<2,000	8,000
Pintwater Range	5,000	0	2,000	3,000
Three Lakes Valley	0	--	--	0
North Desert Range	0	--	--	0
East Desert Range	4,00	0	2,500	1,500
Mule Deer Range	4,00	0	2,000	2,000
North Sheep Range	3,000	0	3,500	0
North Las Vegas Range	9,000	800	600	9,000
N. Meadow Valley - Mormon alignment (Axen and Wernicke, 1989, fig. 2-2)				
Meadow Valley Mountains	8,000 ±	0	1,000	17,000
Lower Meadow Valley	5,00	>2,000	200	5,000
Western Edge, Mormon Mountains	0	0	--	0
Western Mormon Mountains	4,000	0	3,000	1,000
Central Mormon Mountains	1,000	0	>1,000	0
Eastern Mormon Mountains	8,000	0	1,500 ±	7,000
East Mormon Mountains	0	0	--	0
O. Golden Gate Range (Tschanz and Pampeyan, 1970, p. 38-39, 95; Bartley and others, 1987)	20,000 +	0	1,250	19,000 +
P. Seaman - Drylake alignment (G.J. Axen, Harvard University, oral commun., 1988; Tschanz and Pampeyan, 1970, p. 38-39, 95)				
Seaman Range	15,000 +	0	1,400	14,000 +
North Pahroc Range	15,000 +	0	>1,000	14,000 +
Dry Lake Valley	10,000 +	5,000-10,000	<1,000	10,000 +
Q. Ely Springs Range (G.J. Axen, Harvard University, oral commun., 1988)	1,000 +	0-2,000	0-50	1,000 +
R. Meadow Valley (G.J. Axen, Harvard University, oral commun., 1988)	2,000 +	0-2,000	0-300	2,000 +

west of the Sheep and Pahranagat Ranges (area 2, fig. 31) are thinnest. The carbonate rocks beneath the Spring Mountains and Pahrump Valley (area 1, fig. 31) are of intermediate thickness.

Regardless of the structural differences between the areas designated in figure 31within the central corridor, the areas probably are not distinct hydrologic units; that is, flow and water levels in aquifers in one area cannot be assumed to be independent from conditions in adjoining areas. The major extensional structures in the corridor can only locally be shown to restrict flow (Dettinger, 1987, 1988). These barriers commonly can only be shown to restrict shallow flow (less than about 5,000 ft). Deeper flow, in principle, could pass beneath these barriers at greater depths. The Las Vegas Valley and Pahranagat Valley shear zones (fig. 31, table 11)—to the extent that they are barriers to flow—may extend deep enough to restrict both deep and shallow flows (Gary Axen, Harvard University, oral commun., 1987). Most of the structures that form the outer edges of the corridor (for example, the Sheephead fault and Kingston Range detachment) also are assumed to be barriers to both deep and shallow flow.

Outside the corridor, saturated thicknesses are generally less than within the corridor but are variable. For example, along the line through the Mormon Mountains water-saturated thickness estimates vary—5,000; 0; 1,000; 0; 7,000; and 0 ft—along a west-to-east distance of 20 mi.

The storage calculations that follow assume the entire thickness of carbonate rock beneath the central corridor constitutes carbonate-rock aquifers. Oil wells provide evidence that these rocks include permeable zones and function as aquifers to depths of at least 10,000 ft (Hess and Mifflin, 1978, p. 19). In southern Nevada, drill-stem tests in the Colorock Quarry No. 1 (T. 18 S., R. 65 E., section 22dcd) and Virgin River USA No. 1 (T. 15 S., R. 68 E., section 09cd) oil test wells are notable for having recovered thousands of feet of fluid during drill-stem tests at depths 10,000 ft and greater. These kinds of test results, together with freshwater recoveries in some deep oil-test wells, indicate that where carbonate rocks are 10,000 ft or more thick, they can constitute aquifers throughout any part or all of that thickness.

Total Volume of Aquifers and Stored Water

The total area of the central corridor of carbonate rocks south of 38°N. latitude is about 10,000 mi^2. The corridor is divided into four areas in figure 31, and estimates of their areas and representative thicknesses are listed in table 13. Together these estimates imply the volume of carbonate-rock aquifers in southern Nevada may be more than 70 billion acre-ft (as much as 23,000 mi^3 of rock).

If this estimate of the volume of the aquifers is correct, and if the quantity of storage in a given volume is equal to the specific yield or about 0.01, then about 800 million acre-ft of water are stored in the carbonate-rock sequences of southern Nevada. The largest volume of stored water is in the storage-estimate area centered on Pahranagat Valley (area 4, fig. 31; table 13).

Practical Limits on Stored-Water Resources

Not all of this large volume of water is extractable for use under any practical development strategy. Practical issues that might limit development of the stored water in the carbonate-rock aquifers include pumping lifts, water quality, economic constraints, and adverse effects.

Constraints resulting from well and pump performance are probably not the limiting aspect of developing the stored-water resource in the carbonate-rock aquifers. Pumps can be designed to lift water from depths of greater than 1,000 ft (Driscoll, 1986, p. 602). The principal constraints on achieving such pumping lifts and rates are well construction and aquifer properties. For instance, well diameter must be large enough to accept the pump. For example, a 16-in. casing diameter is required to contain a pump that can efficiently pump 2,000 gal/min from such depths (Driscoll, 1986, table 13.1). The aquifer also must be sufficiently permeable and thick to allow a given pump rate to be sustained (for both short and long times) without having water levels in the well drop below the intakes on the pump.

Other than in the zone of poor-quality water shown in figure 24, water quality also should not limit the resource severely. Most water recharging the aquifers rapidly becomes chemically saturated with respect to calcite and dolomite in the carbonate rocks of the recharge areas. Whole-rock chemical analyses of the carbonate rocks show that they contain only trace amounts of minerals other than calcite and dolomite (Thomas and Welch, in press). As a result, once equilibria with respect to those dominant minerals are attained, constituent concentrations do not significantly increase as water circulates through the aquifers.

Table 13. Estimated areas and representative thicknesses of saturated carbonate rocks in central corridor, and calculated total volume of stored ground water

[Symbols: +, thickness or depth is estimated to bottom of uppermost thrust sheet; underlying thrust sheets may exist and contain additional Paleozoic carbonate rocks at much greater depths; >, greater than; --, not applicable]

Area No. (fig. 31)	Area (thousands of acres)	Range of reported thickness (thousands of feet)	Assumed representative thickness (thousands of feet)	Estimated specific yield (dimensionless)	Calculated volume of water stored (millions of acre-feet)
1	1,700	5 - 15	10	0.01	170
2	1,800	3 - 9+	8	.01	140
3	1,200	5 - >17	15	.01	180
4	1,800	10+ - 19+	15	.01	270
AVERAGE	--	--	12	0.01	--
TOTAL	6,500	3 - 19+	--	--	760

Perhaps the principal water-quality constraint on how deep water can be taken from and used is the increase in temperature of water in rocks with increasing depth. Temperature of the water in the southern carbonate-rock province increases by about 10°F for every 1,000 ft of depth (Kron and Heiken, 1980). This means that water from rocks 10,000 ft below land surface would be about 150°F (assuming that water near land surface is about 50°F), which could limit some uses of the water.

Removing water from storage lowers water pressure or water levels within aquifers. The possible effects of developing the carbonate-rock aquifers will be discussed later in this report, but to estimate the practical limits of how much water might be withdrawn from storage in the carbonate-rock aquifers, it is instructive to consider water-level declines that have been induced by regional development in other large aquifers. Declines in selected aquifers are summarized in table 14. Average declines greater than 50 ft are not common in the larger aquifers. However, even 50-ft declines generally are viewed as excessive when prevalent over large areas. In order to develop such extensive declines, large drawdowns are induced in the vicinity of the pumping centers. In Nevada, some of the most severe declines are in the basin-fill aquifers of Las Vegas, Diamond Valley (in east-central Nevada), and Paradise Valley (in west-central Nevada) where local declines as large as 50 to 240 ft have been observed together with broader scale declines of as much as 30 ft over tens to hundreds of square miles.

A More Practical Estimate of Stored-Water Resources

From a regional and historical perspective, therefore, water-level declines greater than 100 ft over large areas are unlikely to be tolerated. Table 15 presents the estimated volumes of water stored in the upper 100 ft of carbonate-rock aquifer for the four areas shown in figure 31. As noted above, water-level declines in and near well fields developed to withdraw this volume of water from the aquifers would be many times greater than 100 ft. Dewatering on this scale would alter ground-water flow paths to deplete springflow and subsurface outflow from Nevada. Dewatering also would change flow directions and rates between the carbonate-rock aquifers and adjacent aquifers of other rock types.

The estimated total volume of water of quality acceptable for most uses present in the top 100 ft of carbonate-rock aquifers in the central corridor south of 38°N latitude is about 6 million acre-ft (table 15). This volume of water is about one-quarter of the average volume of water stored in Lake Mead. If this volume of water were withdrawn suddenly from the carbonate-rock aquifers, about 100 years would be required to refill the aquifers to present levels (assuming the recharge rate discussed in a previous section).

Comparison of Carbonate-Rock and Basin-Fill Storage

The estimated 6 million acre-ft of water stored in the top 100 ft of carbonate-rock aquifers in the central corridor south of 38°N latitude is about one-sixth of the volume of water stored in the upper 100 ft of the over-

Table 14. Water-level declines in areas of major municipal and agricultural development

Aquifer name and location	Aquifer type	Reported large-scale average decline (feet)	Estimated area of decline (square miles)	Maximum decline (feet)	Sources of data
Central Valley Alluvial Aquifer System (Central Valley, California)	Alluvial sediments	10	12,000-15,000	400	Williamson and others, 1985
Northern Midwest Regional Aquifer System (Chicago, Illinois area)	Sandstone	100	7,000	800	Young and others, 1986
Floridian Regional Aquifer System (Southeast Georgia and northeast Florida, Fort Walton Beach and Tampa, Florida areas)	Carbonate rock	10	15,000	30	Bush and Johnston, 1986
Ogallala Formation (Texas, Kansas, Nebraska, Colorado, Oklahoma, and New Mexico)	Alluvial sediments	10	50,000	200	Weeks, 1986
Southwest Basin and Range Alluvial Aquifers (Area between Phoenix and Tucson, Arizona)	Basin-fill sediments	50	8,000	400	Anderson, 1986
Chicot and Evageline Aquifers (Houston, Texas area)	Alluvial sediments	50	5,000	400	Meyer and Carr, 1979
Las Vegas Valley (Las Vegas, Nevada)	Basin-fill sediments	20	300	240	Harrill, 1976
Diamond Valley (Diamond Valley, Nevada)	Basin-fill sediments	30	50	50	James R. Harrill, oral commun., 1988
Paradise Valley (Paradise Valley, Nevada)	Basin-fill sediments	10	40	80	Prudic and Herman, in press

lying saturated basin-fill aquifers (about 39 million acre-ft [Scott and others, 1971, table 1]). This large quantity of water stored in basin-fill aquifers (relative to the carbonate-rock aquifers) relates directly to the large specific yield of unconsolidated sand and gravel (10 times that of fractured carbonate rocks). The disparity between the volumes of water stored in basin-fill aquifers and carbonate-rock aquifers can be viewed as a final practical limit on development of the stored water in the carbonate-rock aquifers. The basin-fill aquifers are much more accessible and familiar as water supplies and developing a one-time resource from an unfamiliar type of aquifer may seem impractical. However, withdrawing water stored in the carbonate-rock aquifers has some possible advantages over

Table 15. Estimated volume of water stored in upper 100 feet of saturated carbonate rocks in central corridor

Area No. (fig. 31)	Area (thousands of acres)	Estimated specific yield (dimensionless)	Calculated volume of water stored (millions of acre-feet)
1	1,700	0.01	1.7
2	1,800	.01	1.8
3	1,200	.01	1.2
4	1,800	.01	1.8
TOTAL	6,500		6.5

that stored in basin-fill aquifers, including: (1) Water is stored beneath large areas in which flow is well integrated. As a consequence, effects of the withdrawals would be dispersed over larger areas. Dispersal of effects is a disadvantage if the effects remain large everywhere but is an advantage if the effect at any given location is small. (2) Some areas in the carbonate-rock aquifers may prove to be sufficiently isolated from other aquifers so that effects of storage depletion could be contained far from other water wells, springs, or aquifers. Basin-fill aquifers have been more widely developed and storage depletions from basin-fill aquifers accessible for development are more likely to be near and to affect other water wells or springs. (3) The water stored in the carbonate-rock aquifers is water that generally has not been accounted for in previous budgets (unlike basin fill). (4) Extracting water from the carbonate rocks in certain areas might actually allow water from inaccessible basin-fill reservoirs (such as beneath lands withdrawn from public use) to be extracted. Under these conditions, in addition to producing water from storage, the carbonate-rock aquifers would function as a collector of basin-fill water and a conduit conducting flow from inaccessible areas to the carbonate-well fields.

WHAT EFFECTS MIGHT FOLLOW DEVELOPMENT OF THE CARBONATE-ROCK AQUIFERS?

Two broad areas of concern regarding the potential for effects from developing the carbonate-rock aquifers are adverse effects within the carbonate-rock aquifers and adverse effects on adjacent basin-fill aquifers presently used for water supply. Effects resulting from withdrawing water from the carbonate-rock aquifers might include reduction of springflow from carbonate-rock and basin-fill aquifers; water-level declines in wells in all types of aquifers; drying up of some streams, playas, and meadows; and changes in water chemistry. These effects are direct or indirect responses to water-level changes associated with aquifer development, and are related to disturbances of the natural equilibrium between aquifer recharge and discharge. These changes in turn can affect the water rights of current users, affect wildlife habitat, and cause increased pumping lifts. The extent and severity of effects are dependent on the local and regional hydrogeologic setting.

Many of the potential effects from development of the carbonate-rock aquifers can be simulated providing sufficient hydrogeologic data are available. These data include transmissivities, storage coefficients, vertical hydraulic conductivities and thicknesses of various aquifers, distances between points of interest, and times involved. On the basis of the above information one can calculate the spread of water-level declines induced in different aquifers by pumped wells. The data taken together can be used to calculate the extent and magnitude of declines surrounding an area of development, using simple mathematical equations, using complex numerical models of ground-water flow, or by analogy to historical experiences with other aquifer developments in the area. These calculations can be used to assess the acceptability of effects over the time period of interest.

Of particular concern when evaluating effects associated with development of the carbonate-rock aquifers (in contrast to basin-fill aquifers), is the probability that effects may be asymmetrical around pumped wells. Conditions within the carbonate-rock aquifers may favor flow in "preferred" directions along such localized pathways as fracture zones. For example, distributions of water-level declines may be elongated in a direction parallel to major fracture traces (Jenkins and Prentice, 1982, p. 15). The extent of declines depends on local hydrogeologic conditions but would be spread greatly along paths of high transmissivity.

In this section, the potential for simulating effects within the carbonate-rock aquifers will be addressed through hydraulic calculations and by a review of existing numerical models of flow in the study area. The potential for effects in adjacent basin-fill aquifers will be addressed through a review of the lessons derived from historical experiences of development. Considerable uncertainties are inherent in resulting calculations and conclusions; thus, a strategy for safeguarding against severe effects is described also.

Potential for Effects Within the Carbonate-Rock Aquifers

The primary effects anticipated from development of the carbonate-rock aquifers are water-level declines within the aquifers. These declines, in turn, may induce decreases in springflow and other natural discharge when they become significant in areas of natural discharge. Sustained effects within the carbonate-rock aquifers resulting from development of wells in the same aquifers have not been observed to date, but this reflects lack of development of the aquifers rather than a lack of potential for effects if development were undertaken. The hydraulic properties together with the geometry of a carbonate-rock aquifer will determine the distribution of water-level declines around pumped wells.

Influence and Range of Aquifer Properties

Under idealized conditions, water-level declines around a pumped well would be distributed as a "cone of depression," with a shape (depth and areal extent) determined by the capacity of the aquifer to transmit and store water (properties described by the transmissivity and specific yield, respectively). The cones of depression around production wells in parts of the aquifers with high transmissivities will be broader in extent but generally shallower than cones around production wells in aquifers with moderate transmissivities. Water-level declines or drawdowns around production wells in aquifers with low transmissivities will be large near the well but the large drawdowns will be restricted to areas close to the well. The extent to which an aquifer is confined or unconfined also will effect the drawdown around a production well. Confined parts of aquifers will be drawn down more and the drawdown will occur more quickly than parts of the aquifer that are unconfined.

Water-bearing properties of the carbonate-rock aquifers are uncertain and vary from place to place. In order to calculate the general range of drawdowns that might be induced around a well (or well field) that is developed in the carbonate-rock aquifers, a 'general' range of variability must be hypothesized or assumed. The best available estimate of this variability is from actual observations, and table 1 shows the available estimates of pertinent aquifer properties from tests in water wells in carbonate-rock aquifers (39 estimates of hydraulic conductivity and 4 estimates of storage coefficients). The estimates of hydraulic conductivity (fig. 5) are lognormally distributed (at a 99-percent confidence level using a Shapiro-Wilk test [Ryan and others, 1982, p. 48-49]), with a median value of 4.3 ft/d and a coefficient of variation (the standard deviation divided by the mean) of 2.6. (By definition [Benjamin and Cornell, 1970, p. 262-264], the logarithm of a lognormally distributed variable is statistically distributed according to a normal distribution—the familiar bell curve.) The relative frequency with which hydraulic conductivities of different values have been measured in the carbonate-rock aquifers (listed in table 1) are hypothesized to approximate the likelihood that a well in an untested location will penetrate aquifers with similar hydraulic conductivities. This hypothesis implies likelihoods of inducing a particular levels of drawdown by pumping at an untested location. For example, the odds of drilling a new well with a specific capacity of at least 5 (gal/min)/ft of drawdown are about 5 out of 10, whereas the odds of drilling a new well with a specific capacity of at least 25 (gal/min)/ft of drawdown are about 2 out of 10.

Only four hydraulic determinations of storage coefficient are available. As discussed in an earlier section, the storage coefficient or specific yield of the carbonate rock aquifers is estimated to range from about 0.01 to 0.10. Storage coefficient of the confine parts of the aquifers is probably on the order of 0.0001 or 0.01 (for example, at well UE25p#1 on the Nevada Test Site, storage coefficient is 0.03). The specific yield of unconfined parts of the aquifers is probably, at least, on the order of 0.1 (for example, at well CE-DT-5, storage coefficient is 0.14). Current (1989) understanding of the distribution of confined as opposed to unconfined parts of the aquifers or the change from confined to unconfined conditions as a result of pumping is limited. Thus, a statistical description of storage coefficient is not presented.

Predictions of Potential for Drawdowns on the Basis of Measured Aquifer Properties

Where aquifer properties are known, water-level drawdowns caused by pumping can be projected roughly using hydraulic analyses of Theis (1935). This method assumes temporal and spatial invariance of aquifer properties, strictly horizontal flow, and no lateral barriers (an infinitely extensive aquifer; Bear, 1979, p. 318-319). These assumptions at best are only approximately applicable to the carbonate-rock aquifers, and thus the assumptions represent limitations of varying severity on the projections presented herein.

The assumption of spatial invariance (homogeneity and, in simplest form, isotropy) is probably poorly met in the carbonate-rock aquifers. Transmissivity of these aquifers is predominantly associated with flow through fractures that are neither uniformly distributed nor uniformly oriented. Thus transmissivity (or hydraulic conductivity, to be more precise) varies from point to point and also varies depending on the direction of flow. Point-to-point variations in transmissivity are assumed small when averaged over relatively large areas. The validity of this assumption is difficult to assess with available information, but the assumption itself has precedent (for example, Harrill and others, 1983, p. 26). Possible effects of fracture orientations on transmissivity and drawdowns probably cannot be so easily neglected. Simple drawdown projections based on the Theis equation include the assumption that drawdowns will be identically distributed in all directions around the pumped well. If flow is preferentially conducted over long distances through a set of fractures of a single orientation, then the calculations will underestimate the drawdown at points along those fracture sets and will overestimate the drawdown at points along lines from the well not parallel to the fractures (Boehmer and Boonstra, 1987, fig. 3). This limitation is potentially severe and should be considered before applying results of simple calculations based on the Theis equation to a real-world situation.

The analyses by Theis were derived for confined-aquifer conditions so that time-invariant transmissivities and storage coefficients could be assumed. In some parts of southern Nevada, the carbonate-rock aquifers are confined under natural conditions (for example, at Well UE25p#1 on the Nevada Test Site). Other areas are not confined (for example, at MX well CE-DT-5), and some confined aquifers could become unconfined if development and drawdowns were extensive. The Theis analysis can however be applied under unconfined conditions provided the projections are (1) for times sufficiently long after pumping began so that water is all derived from drainage of open spaces, (2) for parts of the aquifer far enough from the pumped well so that flow is virtually horizontal, and (3) for aquifers with saturated thicknesses that are large compared to the amount of drawdown projected (Freeze and Cherry, 1979, p. 324-327).

Finally, Theis analyses assume that aquifer boundaries are far enough from the pumped well so that water-level declines at the boundaries never become significant during the period over which calculations are being made. Figure 33 compares these idealized aquifer geometries to those of block-faulted aquifers in the carbonate-rock province. The idealized aquifers of Theis analyses (fig. 33*A*) are assumed to be separated by an impermeable layer (if the separating layer is not completely impermeable, analytical methods developed by Hantush, 1956, can be substituted). The idealized aquifer also is infinite in lateral extent. Block-faulted aquifers, however, may be sufficiently deformed so that parts of the carbonate rocks and basin fill may be laterally isolated by juxtapositions of permeable and impermeable material along fault zones. These faults (which formed the Basin and Range topography) create partially isolated blocks within regional flow systems. These blocks commonly may be smaller than the calculated area of influence of the pumped well. Thus, the assumption of an infinite aquifer may not be satisfied in many areas of the carbonate-rock province. Because this condition is not satisfied, the Theis solution may over- or underestimate drawdowns. If the faults or offsets form barriers to flow, over long periods of time the Theis equation will underestimate drawdowns within the pumped block and overestimate drawdowns at larger distances. If the faults act as conduits for flow, water levels in other blocks somewhat removed from the developed block may be drawn down because of their connection to the conduits, and drawdowns will be less localized. Wells either within the pumped block or in adjacent and connected fault-conduits will experience concentrated drawdowns by pumping. Because the fault-conduits can extend for many miles and might conduct flow that supports springs, this potential for error is a major shortcoming of analyses made using the Theis equation. As with assumptions of spatial invariability of aquifer properties, site specific evaluations of this potential error are needed.

Regardless of the limitations outlined in the preceding paragraphs, the Theis equation can be used to develop generalized information about the magnitudes of drawdowns that might be expected around a production well in the carbonate-rock aquifers. Calculating drawdowns using Theis's equation requires that the time since pumping began and distance from the pumped well be specified and that a transmissivity and storage coefficient be estimated for the aquifer. For purposes of developing an example calculation of drawdowns resulting from small, long-term developments of the regional aquifers, drawdowns around a production well in a large, continuous and hydraulically uniform part of the carbonate-rock aquifers will be projected. The aquifer tapped is assumed to have a hydraulic conductivity of 0.35, 4.5, and 57 ft/d. These values represent the hydraulic conductivity that is less than all but one-sixth of the tests in carbonate-rock aquifers in table 1, the median hydraulic conductivity, and the conductivity greater than all but one-sixth of the tests. The perforated interval of the well and thickness of the aquifer penetrated both are assumed to be 2,000 ft. Storage coefficient is assumed to be 0.0001 for confined conditions and 0.01 for unconfined conditions. Table 16 shows the drawdowns projected at 1 and 10 mi from the pumped well after 1 and 10 years of pumping 1,000 gal/min (1,614 acre-ft/yr). All projections are based on the Theis equation for drawdown described in Bear (1979, p. 318-319).

These simple calculations show that water-level declines associated with a pumped well in the carbonate-rock aquifers may be moderate to small in most locations where the aquifer is unconfined (water table as opposed to confined or artesian) and would be much larger where the aquifer is confined. The projected water-level declines in extensive, unconfined carbonate-rock aquifers would commonly be between about 1 and 22 ft at a distance of 1 mi from a production well after 1 year of pumping. The drawdown at that distance in a part of the aquifer with the median hydraulic conductivity would be about 6 ft after 1 year. Those drawdowns would increase to range between 1 and 70 ft after 10 years. At a distance of 10 mi from the pumped well, drawdowns after 10 years are calculated to range between a fraction and several feet. If, instead, the

Table 16. Calculated drawdowns after 1 and 10 years of pumping at 1,000 gallons per minute from a fully penetrating well in a 2,000-foot-thick aquifer for selected hydraulic conductivities and storativities

Distance from pumped well (miles)	Drawdown after 1 year, in feet, for the following hydraulic conductivities, in feet per day			Drawdown after 10 years, in feet, for the following hydraulic conductivities, in feet per day		
	0.35	4.5	57	0.35	4.5	57
CONFINED (STORAGE COEFFICIENT=0.0001)						
1	120	13	1.4	170	17	1.7
5	48	7.9	1.0	97	12	1.3
10	21	5.6	.8	67	9.5	1.1
UNCONFINED (STORAGE COEFFICIENT=0.01)						
1	21	5.6	0.8	67	9.5	1.1
5	.0	.9	.4	8.5	4.1	.7
10	.0	.1	.2	.4	2	.5

developed aquifer is confined, then calculated water-level declines after 10 years range between 1 ft and about 170 ft at 1 mi from the pumped well and between 1 and 70 ft at 10 mi. The small area projected as experiencing 10 ft or more of drawdown in unconfined parts of the aquifers—less than a 1-mi radius for aquifers with the median conductivity and less than a 5-mi radius for aquifers for the less-permeable aquifers—implies that in many settings the effects of development of the carbonate-rock aquifers might be manageable. In confined parts of the aquifers, the projected drawdowns are substantially larger and might cause unacceptable effects.

Influence of Aquifer Geometry

Water-level declines or drawdowns resulting from development of a part of the carbonate-rock aquifers will be influenced by the geometry of the aquifer and by hydraulic properties. The present study has shown that the carbonate-rock aquifers are not a single massive, continuous layer beneath southern Nevada but rather have been broken into corridors and isolated blocks. The generalized hydrogeologic sections shown in figure 21 illustrate the irregular external boundaries of the carbonate aquifers. Development in a block or part of the corridor isolated from other carbonate-rock aquifers by impermeable barriers will not induce drawdowns in those other aquifers (or if the isolation is only partial, may induce much smaller effects). Within the developed block, however, drawdowns may be much greater than those expected if the barriers were not present. The influence of aquifer geometries is expected to be large, site-specific, and most important over large distances and long times. Bear (1979, section 8-10) describes techniques that can be used to simulate the effects of barriers when using simple analytical models such as the Theis equation. Use of these techniques where the geology is understood well enough to delineate barriers would significantly improve projections of drawdown.

Modeling Flow in Complex Geometries

Accurate prediction of drawdown in the carbonate-rock aquifers will require techniques that take full account of variations in hydraulic properties of the aquifers, aquifer geometry, and vertical components of flow. Numerical models are the tools most capable of taking of these factors into account. However, a great deal of detailed factual information is required to calibrate these models and, currently (1989), this type of information is lacking over most of eastern and southern Nevada. Until additional detailed hydrogeologic information is obtained, results of model simulations are probably only slightly better than results based on careful evaluation of the simple analytical calculations described in the previous paragraphs. However, some models developed for other uses are available and can provide insights into the complexities of projecting effects within the carbonate-rock aquifers. The following paragraphs briefly describe two such models.

Bedinger and others (1989) simulated flow paths and travel times along several profiles beneath the Death Valley region of Nevada and California. The profiles included flow through basin-fill, volcanic-rock, and carbonate-rock aquifers, and addressed the influence of probable geologic deformations on aquifer geometry. Flow paths were affected by changes in hydraulic conductivity between aquifer types and hydraulic barriers formed by impermeable rocks and geologic structures. Results indicate that ground water will preferentially flow in the upper parts (about the upper 3,000 to 4,000 ft) of the aquifers and that flow paths become convoluted in areas containing major structures or abrupt changes in rock type. No simulations of the effects of development in the carbonate-rock aquifers were undertaken with these profile models.

Prudic and others (1993) constructed a regional-scale model of the carbonate-rock aquifers within virtually the entire carbonate-rock province of the Great Basin shown in figure 1. The model includes both basin-fill and carbonate-rock aquifers, and simulates flow in two layers that are intended to represent shallow and deep ground-water flow systems, respectively. Simulated flow patterns in the upper layer can be divided into 17 largely independent flow systems that closely match flow systems delineated in previous basin studies. Flow was more integrated at depth. Flow in the bottom layer can be divided into five deep flow systems that drain toward major regional sinks. Flow boundaries in the deeper layer did not coincide with boundaries in the upper layer, which imply that flow-system boundaries are not always vertical. Thus, development in one flow system (as delineated in the shallow aquifers) could have influences on much larger flow systems that are only poorly understood at present (1989). In addition, the success of the regional model in reproducing observed regional conditions imply that its explicit assumption of hydraulic continuity among all aquifers throughout the region is a viable (and conceivably, a necessary) assumption. This, in turn, implies that effects of development are not precluded from spreading over very large distances and between the carbonate-rock aquifers and adjacent basin-fill ground-water reservoirs.

These two modeling studies imply that flow in the aquifers is complex at regional scales. The geologic complexity of southern Nevada is a constraint on development of more detailed models and on accurate predictions of effects of development. Numerous fault-zone orientations, changing aquifer thicknesses, and difficult-to-delineate masses of noncarbonate rock that function as barriers to flow make detailed aquifer descriptions a necessary input to site-specific (or even regional) predictions of development effects. An adequate incorporation of those detailed descriptions into predictions will probably involve use of new, specially constructed models of flow in the study area.

Potential for Effects in Basin-fill Aquifers

Effects in adjacent basin-fill aquifers resulting from development of the carbonate-rock aquifers are another concern. As water levels are drawn down in the carbonate-rock aquifers by withdrawals from those aquifers, leakage may be induced into those aquifers from adjacent aquifers or naturally occurring leakage into the adjacent aquifers from the carbonate-rock aquifers could be reduced. These diversions of water from the adjacent aquifers could result in water-level declines, depletions of springflow and evapotranspiration from those aquifers, and changes in distributions of water quality.

As with effects within the carbonate-rock aquifers, water-level declines in the basin-fill aquifers will be determined by aquifer properties of the two aquifer systems—carbonate-rock and basin-fill aquifers—and by the geometry and properties of any intervening units that restrict hydraulic contact between the aquifers. In the ideal aquifer system shown in figure 33, these factors are described by the aquifer properties of the upper and lower aquifers and the vertical hydraulic conductivity and thickness of the intervening layer of low-permeability material. Methods for hydraulic calculation of effects in an aquifer that overlies a second, stressed aquifer (Hantush, 1956)—similar in style and simplicity to the Theis (1935) equations and based on the ideal aquifer system shown in figure 33—are probably not applicable to the complex geologic setting in southern Nevada at the scales being considered here. However, historical experience with such conditions is available at two areas in southern Nevada that have experienced ground-water development from basin-fill aquifers adjacent to carbonate-rock aquifers: Ash Meadows and Muddy River Springs. Observations in these areas provide information concerning the potential for interaction of aquifers under developed conditions. However, the historical conditions are the reverse of the development being assessed herein (the basin-fill aquifer were developed instead of the carbonate-rock aquifers). One can presume that reversing the pumping conditions (withdrawing water from the carbonate rocks instead of the basin fill) might lead to effects that would be related to—although not a mirror image of—the general distribution and magnitude of the observed effects. Thus, a look at these two areas with different hydrogeologic settings gives some insight into what to expect under similar hydrogeologic conditions elsewhere.

Ash Meadows

The Ash Meadows discharge area (fig. 34) is located in the southeastern and east-central Amargosa Desert. It represents the terminus of the Ash Meadows regional ground-water flow system that encompasses approximately 4,500 mi^2. Water discharges at about

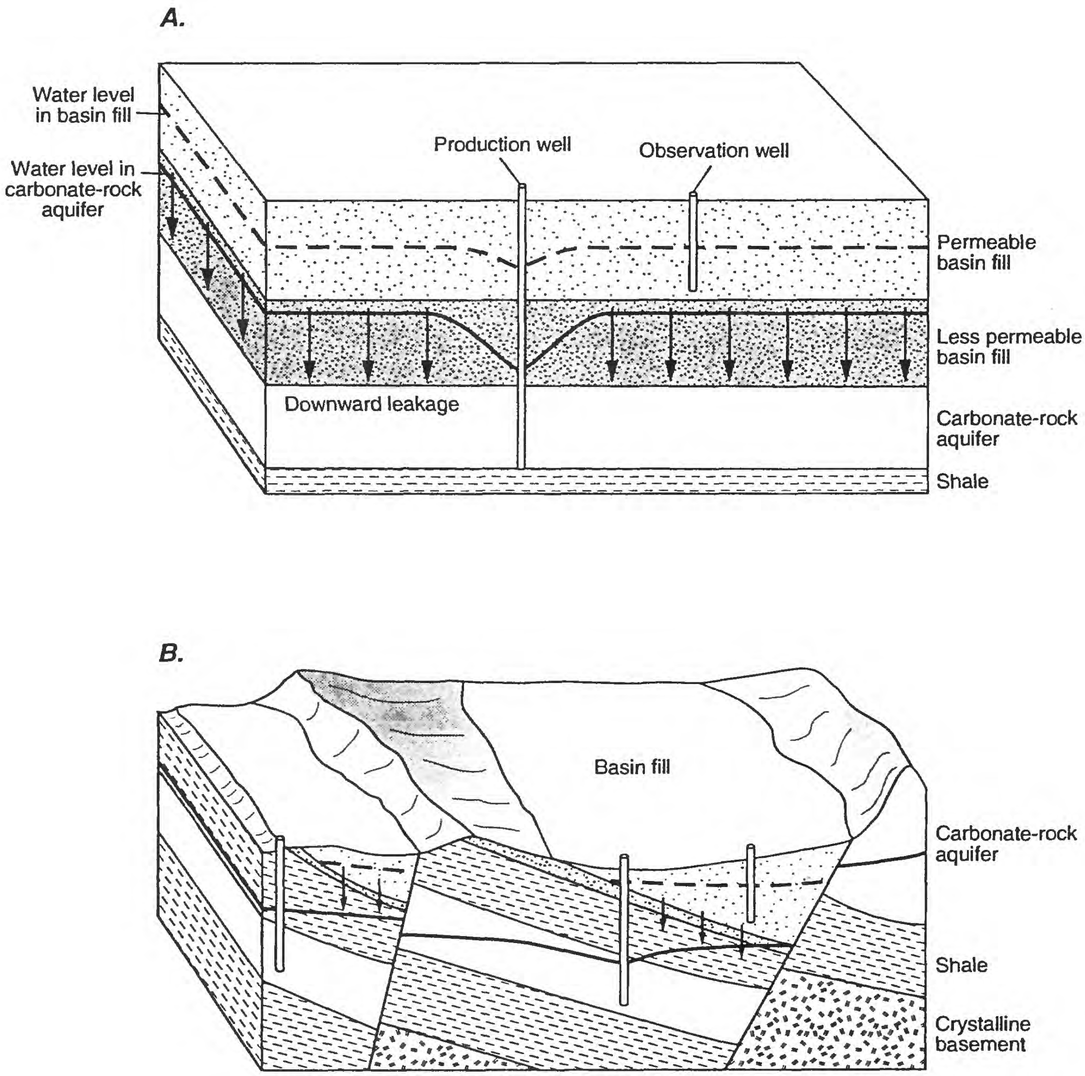

Figure 33. Examples of (A) ideal, laterally infinite aquifers (as assumed by Theis, 1935; Hantush, 1956) and confining units, and (B) block-faulted aquifers and confining units of carbonate rock province.

30 carbonate springs along a generally northwest-southeast line that is about 10 mi long. Discharge totals about 17,000 acre-ft/yr (Winograd and Pearson, 1976). All the major springs emerge from circular pools, are relatively warm, and discharged at a nearly constant rate from 1953 (and possibly from 1910) until agricultural development in the area beginning in 1969. Devils Hole, a collapse depression in the bordering carbonate rock (Carr, 1988b, p. 10), also contains a pool of water. The springs and Devils Hole support several species of *Cyprinodon*, commonly known as desert pupfish, which are unique to the area.

Irrigation using water from the basin-fill aquifers in the area around Devils Hole began in 1969. The subsequent progress of pumping is shown in figure 34. During 1969, approximately 2,100 acre-ft of ground-water were pumped. Pumping increased to in excess of 6,000 acre-ft during 1970. These high pumping levels were maintained during 1971 and declined starting in 1972 (Bateman and others, 1974). In August 1971, a Federal law suit stopped pumping in wells most directly affecting water levels in Devils Hole. A subsequent court order restricted pumping to maintain the water level in Devils Hole at a fixed level.

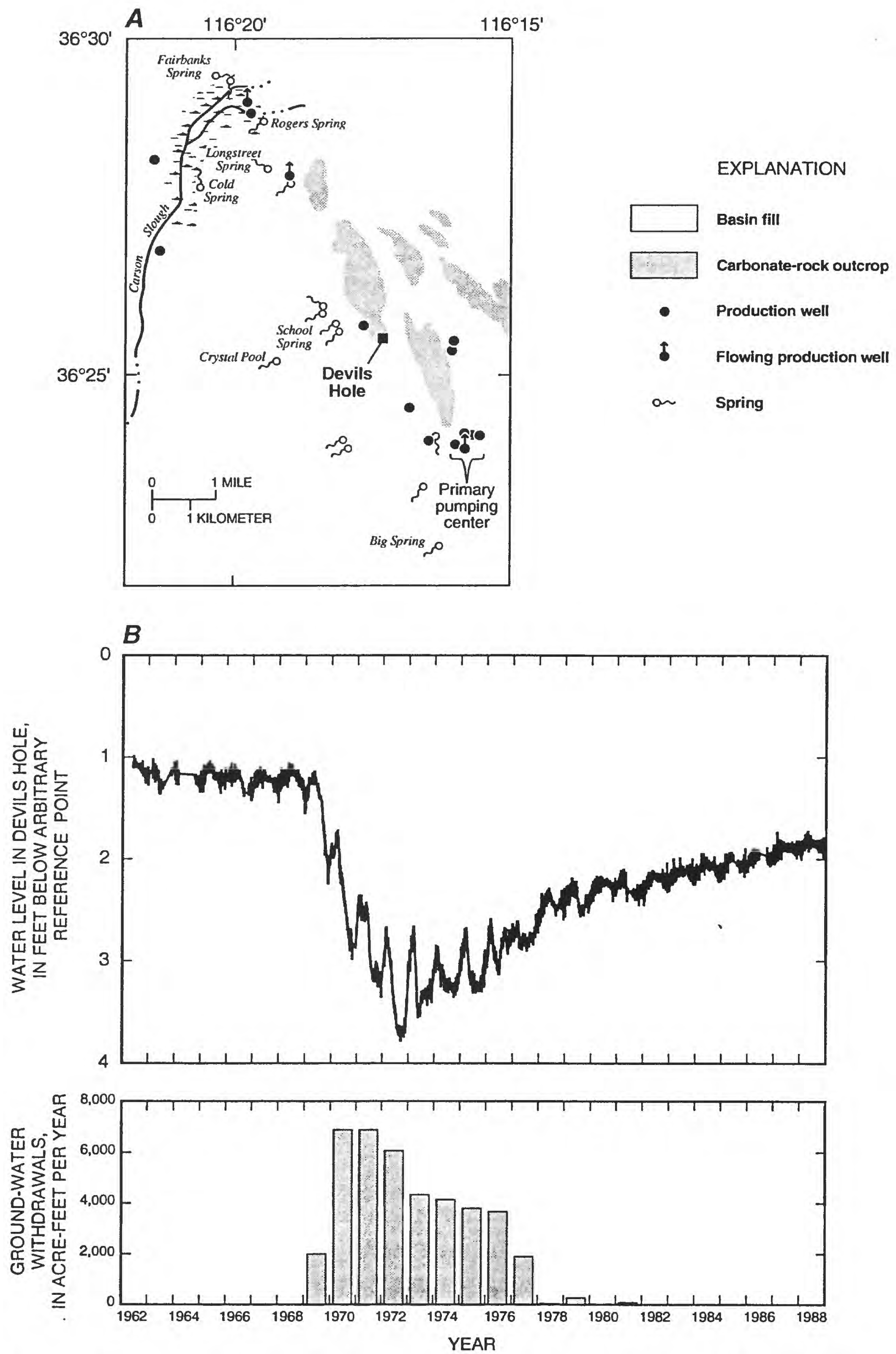

Figure 34. Interaction of basin-fill and carbonate-rock aquifers in Devils Hole (Ash Meadows) area. *(A)* Location of springs, production wells, and Devils Hole (from Larson, 1974a); *(B)* water-level history at Devils Hole, 1963-87, and available estimates of annual pumpage, 1969-75. Pumpage reported by Bateman and others (1974); Dudley and Larson (1976); Larson (1974a, 1974b, 1975); Hanes (1976); Carson (1980); and Westenburg (1993).

The effect of this pumping on hydrologic conditions in the area was extensive and was manifested primarily by water-level declines observed in Devils Hole (fig. 34), the decrease or temporary cessation of flow of several major carbonate springs in the area during pumping, and the gradual recovery of water levels and springflow after some pumping stopped.

In addition, the effect of pumping on individual springs differed indicating that a variable degree of hydraulic connection exists between the basin-fill ground-water reservoir and the carbonate-rock aquifer. The complex network of fractures that connected pumped wells to Devils Hole are distributed such that withdrawals from some wells caused greater declines than similar withdrawals from other wells closer to the Hole. This illustrates potential difficulties in attempting to simulate precise changes in localized areas of the carbonate-rock aquifers. Two modeling efforts have simulated the effects of nearby pumping on the water level in Devils Hole (Bateman and others, 1972; Bateman and others, 1974; and more recently, Rojstaczer, 1987). In general, however, special models of flow through specific faults or other corridors may be required before accurate predictions of small changes in localized areas are feasible. Application of such models requires much detailed and site-specific hydrogeologic data.

Muddy River Springs Area

The Muddy River Springs area is northwest of Glendale, Nev., and is an area of extensive surface- and ground-water development for irrigation, public water supply, and cooling water for power generation. Water has been developed directly from the Muddy River, from the regionally fed Muddy River Springs, and from wells completed in the basin-fill aquifer. Ground water has been withdrawn at varying rates during the last 20 years (Eakin, 1964, p. 24). Monitoring wells are completed in the basin fill and in the underlying and surrounding carbonate-rock aquifers to address concerns that increased ground-water development in the basin fill (as well as new development from nearby carbonate-rock aquifers) will reduce discharge rates from the Muddy River Springs. Locations of these features are shown later in figure 36.

Carbonate-rock aquifers are the main aquifers in the area. In the valleys, carbonate-rock aquifers are overlain by basin-fill deposits consisting of sands and gravels above thick sections of older less-permeable lake-bed deposits in the late Tertiary Muddy Creek and Horse Spring formations. The older basin-fill formations generally are not productive aquifers and probably act as semipermeable layers separating the aquifers of the uppermost basin fill from underlying carbonate-rock aquifers. The younger, productive basin-fill aquifers are localized along the valley floor of the White River drainage. In contrast, the carbonate-rock aquifers extend over large areas and have continuity with several upgradient areas. The basin-fill aquifers are recharged by springflow and seepage from the carbonate-rock aquifers and by secondary recharge from irrigated lands. The general direction of flow is down valley (to the southeast). The depth to water in the valley generally is less than 10 ft. In many parts of Upper Moapa Valley, wells drilled through the basin fill into the underlying carbonate-rock aquifers would flow naturally. Natural discharge occurs as surface outflow to the Overton Arm of Lake Mead and as evapotranspiration. During summer months, flow at the Muddy River gage (located on the river between Muddy River Springs and the town of Moapa) is reduced by evapotranspiration. Measured flow rates are below the winter flow rates by about 1,500 acre-ft/yr (Eakin and Moore, 1964). .

Short-term water-level fluctuations in two adjacent wells—Lewis North and EH-5—completed in the basin fill and carbonate rocks, respectively, were analyzed to help determine whether or not the aquifers are hydraulically connected (Pohlmann and others, 1988). Generally, water levels in the carbonate-rock well (EH-5B) rose (depth to water decrease) as atmospheric pressure or earth-tide dilation increase, with no discernable delay. This response is characteristic of confined or semiconfined conditions, which was indicated also by lithologies encountered during the drilling of the well. This interpretation implies the carbonate-rock aquifer in this locale functions as a confined aquifer in response to short-term pressure fluctuations, and, locally, may not be hydraulically connected to other aquifers. Meanwhile, the response in the basin-fill well (Lewis North) was similar in magnitude, but with a phase shift. This may imply unconfined conditions in the basin-fill aquifer, but alternatively could be related to other factors (Pohlmann and others, 1988). The uniformity and periodicity of these diurnal fluctuations and their correlation to changes in atmospheric pressure and earth tides imply that these short-term water-level changes do not reflect pumping in the area.

At longer time scales, however, fluctuations may reflect seasonal ground-water use in the Muddy River Springs area, which is characterized by pronounced seasonal fluctuations. Pumpage peaks in the summer due to irrigation demands and increased demands for water by the Nevada Power Company power plant nearby. A small (less than 0.5 ft) decline of water levels was measured during the summer of 1987 in two monitoring wells completed in the carbonate-rock aquifers on the edges of the valley floor (Pohlmann and others, 1988). This decline was most likely in response to increased pumping and drawdowns in the basin-fill aquifers during the summer months. Mifflin & Associates (M.D. Mifflin, written commun., 1987) mapped a drawdown distribution for the 1987 pumping-season in the basin fill around the Nevada Power Company Lewis Wells (fig. 36) with a maximum drawdown of greater than 15 ft. The seasonal water-level declines measured in carbonate-rock wells are small in comparison and difficult to attribute to water-level changes in the basin-fill aquifer (fig. 35). Other factors that could cause the seasonal water-level fluctuations in the carbonate-rock aquifers include pumping from elsewhere in the carbonate-rock aquifers (such as at MX well CE-DT-6 about 6 mi away, fig. 36) and natural seasonal variations in ground-water flux in the carbonate-rock aquifers. No corresponding decline in spring discharge has been observed during more than 3 years of continuous monitoring at these springs for the present study

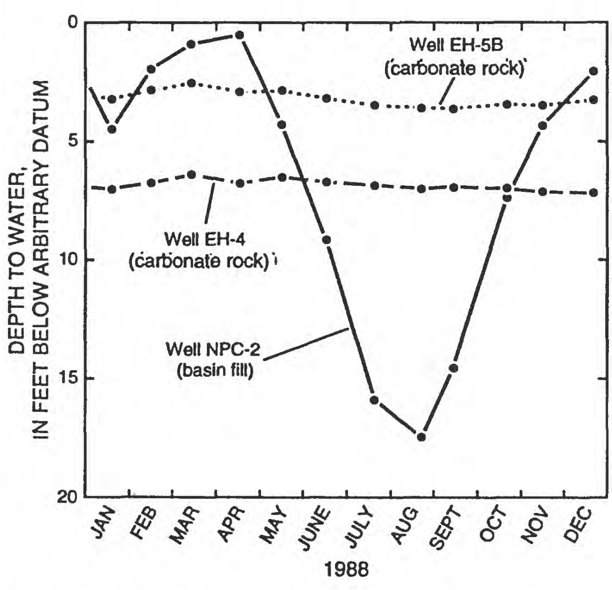

Figure 35. Water levels in wells completed in basin-fill and carbonate-rock aquifers in the Upper Muddy River Springs area, 1988.

(fig. 7). Continued monitoring in the Muddy River Springs area will help to better evaluate the relation between the carbonate-rock aquifers and local basin-fill aquifers.

Comparison of Historical Experiences

Experience from Ash Meadows and Upper Muddy River Springs area indicates the site specific nature of the magnitude and extent of ground-water development effects. In both cases, observations were of the effect of development of the basin-fill aquifers on the carbonate-rock aquifers. It is presumed that opposite types of effects (but not magnitudes of effects) might be observed if the situations were reversed, that is to have the pumping occur in the carbonate-rock aquifers. In Ash Meadows, an immediate and measurable water-level decline in the carbonate-rock aquifers and springflow decline resulted from development of ground water from the basin-fill aquifer. This indicates a direct hydraulic connection between the carbonate and basin-fill aquifer systems.

In the Upper Muddy River Springs area, only small changes in water levels in the carbonate-rock aquifers were measured. The carbonate-rock aquifers are confined in at least some areas, which implies that the carbonate-rock aquifer is isolated hydraulically (little exchange of pressure or water) from the overlying basin-fill aquifers. Causes of the small seasonal water-level changes remain uncertain and no decline in springflow has been measured. Thus, the hydraulic connection between the two aquifers is less than at Ash Meadows.

These two available case studies of development suggest that future development will need to be assessed on a site-by-site basis. In general, each hydrogeologic setting will respond differently to pumping stresses, ranging from immediate responses that directly reflect greater drawdowns in adjacent aquifers to slowly developing responses that only gradually reflect pumping in nearby aquifers.

Simulation of Interaquifer Effects in Pahrump Valley

Although based on a hypothetical development scenario rather than historical experience with aquifer development, numerical simulations of ground-water flow beneath Pahrump Valley by Harrill (1986) provide another insight into controls on interaquifer effects.

The simulations indicate that about half of the ground-water budget of Pahrump Valley discharges from the area as subsurface outflow through carbonate-rock aquifers (Harrill, 1986). Simulations of responses to hypothetical, long-term basin overdraft conditions showed almost no capture of this subsurface outflow despite widespread declines in water levels and spring-flows in the basin-fill aquifers (Harrill, 1986, p. 43). This response implies that developing basin-fill aquifers can result in severe effects within the basin-fill aquifer with little or no effect on the underlying carbonate-rock aquifers, in some places. In part, this response was simulated in Pahrump Valley because water levels in the basin-fill and carbonate-rock aquifers could not be drawn down (by any reasonable rate of pumping) to a level that would reverse or even flatten the steep gradients driving flow toward the southern end of Death Valley. Thus, the relation between the drawdowns associated with a pumped well and the natural gradients driving water toward regional discharge areas may determine what the mix of storage depletion, capture of discharge close to the well, and capture of regional discharge will be. If drawdowns are small compared to the head differences associated with the regional gradient then regional flow will continue unimpeded and the source of water to the pumped well will be depletion of storage or capture of discharge from nearby areas at altitudes close to that of the bottom of the drawdown cone. If the head differences associated with a regional gradient are smaller than or about equal to those induced by pumping, the drawdown cone will more readily capture the regional flow.

Monitoring and Prevention of Effects

Large uncertainties exist in prediction of the general effects of developing the carbonate-rock aquifers and even larger uncertainties where site-specific issues are considered. Given these uncertainties and the wide range of possible effects, observation and interpretation of the actual response to development remains the best approach to developing site-specific predictions and more general models capable of predicting a real response to pumping stress. In light of these uncertainties, how can development proceed with assurances that effects will not overshadow the benefits of the expanded water supplies? At present (1989), the best answer seems to be staging ground-water developments with adequate monitoring of related effects.

All ground-water development causes declines in water level and some depletion of stored water. If development is gradually undertaken (rather than undertaken in one sudden step), then water-level declines can be observed and judged in terms of the attendant benefits and costs at each step. Most effects are reversible, as the water-level changes at Devils Hole demonstrate (fig. 34). Once decisions to develop have been made, the prevention of adverse effects is best accomplished through active monitoring to provide early warning of undesirable changes.

Monitoring needs include more than just repetitive measurements at convenient locations. The information obtained needs to be indicative of changes in the aquifers, specific enough to indicate probable cause-and-effect relations, and reliable enough to support the start of corrective actions as needed.

Given these requirements, the initial monitoring need is to develop a basic understanding of the hydrologic conditions in the area of concern. This includes recognition of the major aquifers present, general directions of ground-water flow, general magnitudes of recharge and discharge, areas and magnitudes of pumpage, water levels in wells, and knowing if a trend in water-level changes exists.

This information can then be used to formulate a monitoring strategy. This generally involves determining the resources that need to be protected and the pumping stresses or other factors that might affect those resources. The monitoring network should provide specific information consistent with the general monitoring strategy. In addition to water-level measurements in wells, the network might have to include discharge measurements of springs and streams, determinations of pumpage, water-quality analyses, and other hydrologic characteristics. In most cases, the network is designed to use existing structures as much as possible for economic reasons. Measurement frequencies are based on long and short-term needs and historical data. Regular reporting and analysis of results is also a desirable part of operation of the monitoring network. Annual changes in monitored parameters should be supplemented by evaluations of long-term trends where data permit. The adequacy of the data needs to be periodically reviewed and the network modified as hydrologic and other needs change.

A Case History: Monitoring Around the Muddy River Springs

The Coyote Spring-Moapa area (fig. 36) is an example of how monitoring can help to identify (and provide information that could be used to prevent) undesirable changes in hydrologic conditions that could result from development. Water resources are supplied primarily by the discharge of about 36,000 acre-ft/yr from the Muddy River Springs. These springs are the regional discharge area for the lower White River flow system and are supplied by carbonate-rock aquifers that transmit water from mountain ranges far upgradient. Some water derived from the north travels more than 100 mi through carbonate-rock aquifers before discharging from the springs (Eakin, 1964, p. 1).

Water has been developed directly from the Muddy River, from the regionally fed Muddy River Springs, and from wells completed in the basin-fill aquifer. Discharge from the Muddy River Springs has been used for irrigation since the area was settled. Water is pumped from the basin-fill aquifer for irrigation, power plant cooling, and domestic water (pumpage in 1985 was about 7,000 acre-ft [Pohlman and others, 1988, p. 12; E. James Crompton, U.S. Geological Survey, written commun., 1989]). In 1986, the Moapa Valley Water District began pumping a well in the adjacent carbonate-rock aquifer (well CE-DT-6). This well is about 4 mi northwest of the springs and the pumped water is transported by pipe to the valley. Pumpage from the well in 1987 was about 200 acre-ft. Another well, CE-DT-5 drilled in the carbonate-rock aquifers of southern Coyote Springs Valley about 10 mi northwest of the springs, is slated for industrial development. This well could be pumped to provide as much as 5,000 acre-ft of water per year.

Despite these long-term developments of water, no significant long-term trend in water levels or spring discharge has been noted as of 1988. Short-term and seasonal fluctuations of water levels in both the basin-fill and carbonate-rock aquifers have been observed.

The water rights for the springs and for the river are longstanding and are protected under Nevada water law. In addition, desert fish live in the ponds and discharge channels of some of the springs and are protected by law. Several of the springs are included in the Moapa National Wildlife Refuge.

The general monitoring strategy discussed in this report includes continuous flow measurements at some springs and on the river as it flows from the area to determine the extent of any induced changes. Frequent measurements of other springs are desirable also. At the same time, the general monitoring strategy recognizes that causal relations between pumping and springflow declines will be difficult to verify without monitoring of ground-water levels. Monitoring of water levels also can provide early warning of impending springflow declines. Where possible and economically feasible, pumping rates at production wells and water-level changes in a network of observation wells would be monitored to determine the magnitude and areal extent on water-level declines associated with pumping. In the event that spring discharges decline, a critical question will be whether the changes were being caused by pumping in one specific area or whether combined effects from several areas were causing the changes. Detailed records of the actual times and rates that individual wells were pumped would be a useful addition to the monitoring data, and could be compared with the spring-discharge records to show correlations between specific pumping practices and springflow declines. Identifying cause-and-effect relations also would require that water levels in all aquifers be monitored, at pumped wells, near pumped wells, and between the wells and the springs. Finally, observations of short-term and naturally occurring seasonal fluctuations would be useful data for sorting out the relations under stressed conditions.

The foundation of an adequate monitoring network is already in place as a result of several separate efforts in the past 10 years. The success of nearby MX drilling programs in 1981 not only demonstrated the potential for high yield from some carbonate-rock aquifers but also provided valuable background information on hydrologic conditions prior to development in areas upgradient from the springs. Test wells drilled as part of an exploration program by Nevada Power Company provided similar information in areas generally downgradient from the springs (Desert Research Institute, 1986, p. 1). Two monitoring wells also were drilled into the carbonate-rock aquifers adjacent to major springs as part of this program. These wells can monitor small water-level changes in the immediate vicinity of the springs that might provide early indications of changes in spring discharge. Three wells were drilled as part of the present study to provide information in areas where data are lacking. Discharge from several springs also has been monitored with continuous recorders as part of the present study. In addition to continuous monitoring of several springs, nearly all other springs are measured semi-annually. This mix of occasional discharge measurements at most springs

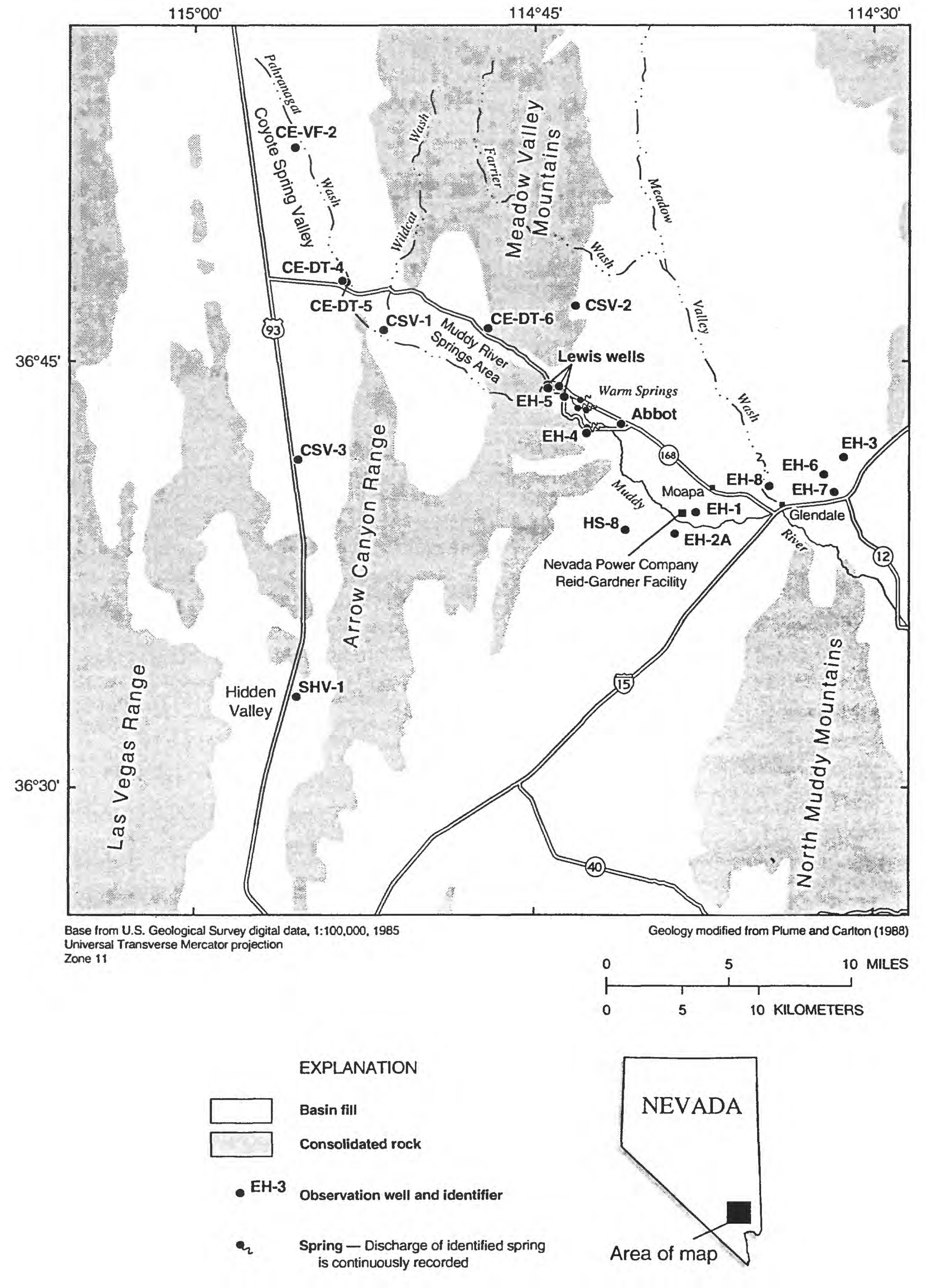

Figure 36. Observation wells and selected springs in and near Upper Muddy River Springs area.

and continuous discharge measurements at a few springs is designed to allow widespread identification of effects in an economical manner. The continuous measurements should detect changes in flow rates and the occasional measurements could be correlated to those continuous records to estimate the areal scope of the changes.

The existing network of well and spring discharge measurements coupled with a detailed inventory of pumpage will provide most of the information required for a comprehensive monitoring network. One or two additional monitoring wells might be needed upgradient from the springs to improve definition of possible effects of upgradient pumping from the carbonate-rock aquifers. Regular reporting of the results will provide managers and local officials with early warnings of detrimental changes and provide data to support decisions and actions that they might subsequently take. As such, the monitoring program that has developed will provide data to use in protecting against effects at this, likely the first area of significant development of the carbonate-rock aquifers in southern Nevada. It also can serve as a model for the kinds of considerations and detail required at other potential development sites.

A General Development and Monitoring Strategy

The monitoring network in the Muddy River Springs area has been developed during more than 10 years of investigations in the area. As a result, this real-world network has many features specific to data availability and history of development there. A broader range of monitoring wells penetrating the carbonate-rock aquifers is available than would be expected around some new developments. The proximity of MX well-siting investigations is propitious but not likely to be the norm near other developments.

Considering the network that evolved around Muddy River Springs and the prospects for development of carbonate-rock aquifers in other settings, however, leads to a more general strategy for managing effects around future aquifer developments. Many uncertainties that will plague predictions of effects have been discussed in this section, and a basic lesson of these discussions is that initially, assurances that effects from development will not overshadow benefits cannot be made with a high degree of confidence. However, if staged development was undertaken together with adequate monitoring, effects of continued or increased development could be estimated with progressively higher degrees of confidence. Staging development means not developing the resources in one large step but rather starting with small developments that are increased gradually as conditions and confidence warrant. This approach allows effects to be observed and analyzed as they develop so that the effects and benefits of development can be judged and, if necessary, reversed if the development proves too costly (in economic and environmental terms). Adequate monitoring means monitoring that provides data for a timely and sound judgement of effects and benefits. Monitoring of all aquifers that may be effected, and all hydrologic conditions that may reflect these effects, would be required for a sound basis. Regular monitoring, that recognizes early signs of effects and that allows natural fluctuations to be distinguished from development-induced variations, would be required for timely judgments.

SUMMARY

Rocks that were consolidated from layers of ancient marine sediments underlie the basins and mountains of a 50,000-mi^2 area of southern and eastern Nevada that is referred to as the carbonate-rock province. These rocks, which are predominantly limestone and dolomite, also extend beneath western Utah and into southeastern Idaho and eastern California. The large area underlain by the rocks, together with their capacity to transmit large volumes of water, indicate that the carbonate-rock province of Nevada contains aquifer systems of regional scale and significance.

In 1985, the State of Nevada entered into a cooperative effort with the U.S. Department of the Interior to study and test the carbonate-rock aquifers, in an effort to assess the potential for developing these ground-water resources. The studies were proposed in a report by the U.S. Department of the Interior (1985) and funded through Nevada State Senate bills S.B. 277 (in 1985) and S.B. 209 (in 1987). Technical work was funded by several agencies and done by the U.S. Geological Survey, the Desert Research Institute, and the Bureau of Reclamation.

During 1985-88, the technical studies were focused on southern Nevada, mostly north of Las Vegas and south of Pioche and Tonopah, although some activities—notably basic-data collection and geochemical analyses—extended farther to the north in Nevada and into adjacent states. The technical studies were intended to address the following basic concerns:

1. Where is water potentially available in the aquifers?
2. How much water potentially can be withdrawn from the aquifers?
3. What effects might result from development of the aquifers?

The carbonate rocks and subordinate noncarbonate rocks of southern Nevada were deposited as layers of ancient marine sediment that cumulatively were as much as 40,000 ft thick on the continental shelf off the ancestral west coast of North America between about 570 million and 250 million years ago (during that period, the west coast of the continent was in present-day Utah). In aggregate, the carbonate rocks are massive and widely distributed, and as a consequence, the aquifers they contain provide avenues for ground-water flow beneath much of the province. All the ancient sedimentary rocks, where deformed and fractured, can transmit some ground-water flow, but the carbonate-rock layers—because of their brittleness and tendency to dissolve into flowing water—are believed to contain the principal water-bearing (or aquifer) zones.

The carbonate-rock aquifers have complex shapes and are connected to aquifers in other rock types. Volcanic activity and erosion of the mountain ranges have resulted in the deposition of younger rocks atop the carbonate rocks. Among these younger rocks are basin-fill aquifers that are the sources of most of the ground water now used in the State. Although the carbonate rocks were deposited as widespread layers, geologic forces subsequently deformed the rocks into innumerable "blocks" that are bounded by faults, and folded rock masses of all sizes. Because of this deformation, rocks of widely differing geologic age are intermingled, and the distribution of rocks constituting aquifers is greatly complicated (as are the paths followed by ground-water flow through the rocks).

Accurate definition of the potential for developing carbonate-rock aquifers requires an understanding of the configuration and physical properties of the carbonate and intervening noncarbonate rocks, coupled with estimates of the volume of water contained in and moving through them. To develop the required understanding, the following activities were undertaken during 1985-88.

Several types of basic hydrologic data were collected in eastern and southern Nevada, including semiannual measurements of cumulative rain and snowfall at 15 high-altitude sites; continuous measurements of discharge at 8 representative springs; regularly scheduled measurements of discharge at 59 other springs; continuous measurements of water levels in 4 wells open to the carbonate-rock aquifers; regularly scheduled measurements of water levels at 11 other wells; and measurements of meteorological conditions used to estimate evaporation rates and water consumption by native plants at 2 sites. Continuous measurements made as part of previous studies indicate that the discharge from two of the larger of the Muddy River Springs remains nearly uniform at the rates (totaling about 8,000 acre-ft/yr) and that the discharge from Corn Creek Springs also is nearly uniform at about the rate measured during 1947-55—about 200 acre-ft/yr. Small, short-term fluctuations in spring discharge and water levels in carbonate wells, however, are observed; they represent responses to atmospheric pressure change, tides, local precipitation, and other natural stresses.

Geology was mapped in an area of about 1,800 mi^2 north of Las Vegas and centered on the Sheep Range. The mapping emphasized carbonate rocks and younger sediments, and was based on detailed field observations, interpretation of geophysical measurements, and correlation of observations with aerial photography and satellite imagery. Results of the mapping indicate that small openings (primary porosity) through which small volumes of water can flow probably are ubiquitous in the carbonate rocks throughout the region. Along many high-angle normal faults, fractured and fragmented carbonate rocks develop and maintain larger openings through which large volumes of water can flow. These fault zones may constitute the principal paths through which most water flows in the carbonate rocks.

Surface geophysical measurements were made to estimate thicknesses of basin-fill aquifers overlying the carbonate-rock aquifers and to locate faults deep beneath land surface. Measurements detected spatial variations in the density, strength, and electrical properties of rocks in the subsurface. Interpretation of these measurements indicate that thicknesses of basin fill overlying carbonate rocks range from less than 500 ft in Hidden Valley to more than 6,000 ft beneath South Tikaboo Valley. Estimates of thickness are critical in siting exploration wells in these valleys. The geoelectrical measurements also indicated that highly productive wells in Coyote Spring Valley penetrate a fault zone along the east edge of the Arrow Canyon Range, rather than the more visible fault zone on the steep west side of the range.

Water samples were collected and analyzed to characterize the quality of ground water and to delineate flow paths in terms of physical and chemical properties of the water. Samples were collected from 209 springs, streams, and wells during 1985-88. Hundreds of analyses from previous studies also were compiled. Geochemical balances and models were used to update water balances and to identify flow and mixing rates in six parts of southern Nevada, including the Spring Mountains, Las Vegas Valley, Ash Meadows area, and Muddy River Springs area. The results alter previous concepts of regional flow beneath southern Nevada by indicating that most recharge from the Sheep Range flows toward the Muddy River Springs rather than radially toward other adjacent valleys including Las Vegas Valley. Also, flow in the basin-fill aquifers of the Las Vegas Valley shear zone was shown to be impeded in some areas and enhanced in others, presumably by the influence of deeply buried geologic structures. Elsewhere, the studies generally verify the overall water budgets developed by previous investigators for regional flow beneath the White River drainage in east-central Nevada. A broad reconnaissance of water quality in southern Nevada demonstrated that aquifers beneath and east of Las Vegas Valley and elsewhere in southeasternmost Nevada are likely to contain water of quality not suitable for many uses.

Wells were drilled, logged, and tested in the basin-fill and carbonate-rock aquifers to provide direct observations of aquifer characteristics. Nine wells were drilled (a total of about 4,500 ft in basin fill and 5,500 ft in carbonate rocks) and two abandoned wells in key locations were rehabilitated. Borehole-geophysical logs were made from selected depth intervals in all these wells and in four more wells. The aquifers were tested at three of the new wells and at three of the wells during 1985-88 to determine transmissivity. Results from reports of specialized tests in 13 petroleum-exploration wells and aquifer tests at 33 other wells in the carbonate-rock aquifers of Nevada were compiled. The capacity of the carbonate-rock aquifers to transmit water ranges from low to very high, depending on location. Whereas the transmissivity of aquifers at the MX wells in Coyote Spring Valley is extremely high (greater than 200,000 ft^2/d) and productivity also is high, most other wells drilled in carbonate rocks are much less productive. In fact, the aquifer at the MX wells is somewhat more transmissive than at any other site for which test results are available. The average aquifer properties, as reported in compiled tests at water wells, are similar to those at Army Well 1 near Mercury (about 5,200 to 11,500 ft^2/d).

All these activities provided useful information about the movement of water in aquifers near and immediately north of Las Vegas. To address the regional distribution of the carbonate-rock aquifers throughout southern Nevada and their potential for development, however, synthesis of the results from all these activities, together with results from many other hydrogeologic and geologic investigations was made.

Understanding where water potentially could be developed from the carbonate-rock aquifers requires an understanding of where the carbonate rocks are present, and where they are continuous enough to form local and regional aquifers. Although the carbonate rocks and subordinate noncarbonate rocks are widespread and originally were thick (as much as 40,000 ft thick in some places), subsequent geologic forces disrupted and partly or completely removed these sedimentary rocks from large parts of southern Nevada. As a result of the action of those forces, much of present-day southern Nevada is underlain by areas where the carbonate rocks remain only as isolated blocks having dimensions that range up to miles on a side. In contrast, the central third of southern Nevada is underlain by a north-south corridor of thick, laterally continuous carbonate rocks. Even within the central corridor, the thickness of carbonate rocks was reduced to between about 3,000 and 19,000 ft. North of Las Vegas, within the central corridor, two areas are underlain by thick and relatively continuous carbonate rocks—the Pintwater Range-Spotted Range area and the Coyote Spring Valley area. These two areas are connected to a similarly thick carbonate-rock mass farther north. The thick carbonate rocks probably contain the principal conduits for regional flow from east-central Nevada into southern Nevada, where the flow is ultimately discharged at Ash Meadows and Death Valley, and at the Muddy River Springs. East and west of the central corridor are blocks of carbonate rock that are thick but largely isolated by noncarbonate materials of low transmissivity from the carbonate-rock aquifers of the central corridor. These carbonate blocks transmit little water.

Farther south, the central corridor underlies the Spring Mountains-Pahrump Valley area. Water flowing through the corridor at this latitude is derived mostly from recharging snowmelt in the Spring Mountains. This recharge moves radially away from the high-altitude areas of the Spring Mountains to discharge near Tecopa, in Pahrump Valley, at Indian Springs, and (in the past) at Las Vegas Springs. The carbonate rocks beneath Las Vegas Valley are believed to thin abruptly to the east toward Lake Mead.

Some zones within the central corridor are highly transmissive, as indicated by large spring discharges that are fed by parts of the aquifers having imperceptibly sloping water tables, and by geologic mapping of ancestral flow paths. The present study developed the hypothesis that highly transmissive zones may function as large-scale drains, collecting water from adjacent, less transmissive rock that underlies most of the study area. The drains would ultimately conduct much of the flow that discharges at large regional springs. The present study also hypothesizes that such zones might stay highly transmissive only if large volumes of water continue to flow through them. Otherwise, openings in the rock gradually fill with fault breccia and the rock reconsolidates. Small, filled fractures observed in outcrops of the carbonate rocks throughout the area appear to have been sealed while still below the water table, and they generally show no evidence of prior mineral dissolution from the fracture walls. In contrast, some flow tubes observed along major fault zones exhibit some characteristics of wall-rock dissolution by flowing ground water. This dissolution widened the openings and enhanced already high transmissivities. If this hypothesis proves true, then wells that tap conduits of concentrated regional flow probably will be most productive. Maps of recently active faults and delineations of regional flow confluences might be used to systematically locate the most productive structure beneath areas being considered for development.

A natural transition exists between (1) areas in southeasternmost Nevada where the carbonate rocks either are intermingled with—or depositionally juxtaposed against—other rocks containing abundant salts (evaporite minerals such as gypsum and halite), and (2) areas elsewhere in southern Nevada where these other rocks are nearly absent. The present investigation minimized study efforts in the former areas because development of the carbonate-rock aquifers risks the encounter of ground water that is unsuitable for most uses.

The water resource of the carbonate-rock aquifers is the sum of the perennial yield of the aquifers and reserve of water stored in them. The perennial yield can be no greater than the total rate of flow through the aquifers, and it probably is less. At present (1989), the total rate of flow through the carbonate-rock aquifers cannot be estimated directly, but rather is bracketed by other rates that can be estimated. The total flow is assumed equal to the recharge to the carbonate-rock aquifers alone, which is less than the total rate of recharge to all aquifers of southern Nevada, both carbonate and noncarbonate. In contrast, the total rate of flow through the carbonate-rock aquifers is greater than the rate of land-surface discharge directly from the carbonate-rock aquifers, because some discharge from the carbonate-rock aquifers is by unseen subsurface leakage of water into adjacent basin-fill aquifers. Thus, the total flow rate is bracketed between a regional total-recharge rate and a land-surface discharge rate.

Natural recharge in the mountains of southern Nevada has been estimated to total about 140,000 acre-ft annually, of which about 110,000 acre-ft is from within the central corridor of thick carbonate-rock aquifers. In addition to ground-water recharge in southern Nevada, geochemical balances computed in the present study indicate that another 21,000 acre-ft/yr is supplied to southern Nevada by inflow through carbonate-rock aquifers from east-central Nevada. Together, recharge plus inflow totals about 160,000 acre-ft/yr.

Part of the total ground-water inflow (recharge plus inflow) moves directly or indirectly into the carbonate-rock aquifers and discharges (1) at regional springs, (2) by flowing out of the study area through carbonate rocks that extend into California, or (3) by leaking into basin-fill aquifers. The first two mechanisms discharge about 77,000 acre-ft from carbonate-rock aquifers of the central corridor. The remaining water either leaks upward into the basin fill or directly recharges the basin fill, and ultimately discharges at local springs, playas, meadows, and streams. Previous studies have estimated natural discharge from basin-fill aquifers in the central corridor to be about 50,000 acre-ft/yr.

The different inflows to and outflows from all aquifers of the central corridor total about 130,000 acre-ft/yr. Under natural conditions, an equal quantity was discharged from springs, playas, meadows, and streams, in southern Nevada and nearby parts of California each year. As a result of this equilibrium between inflow and outflow, future wells that continuously extract any part of the annual recharge eventually can be expected to decrease the discharge from one or more of the aquifers. A review of other possible avenues of regional inflow and outflow along the boundaries of the central corridor of southern Nevada identified no sources of regional inflow, other than the White River system. The review also indicates that only along the California border are the carbonate-rock aquifers continuous enough to transmit large quantities of water to areas where its discharge at land surface could have been overlooked in this study. In particular, no large currently (1989) unidentified quantities of water are likely to be flowing out of southern Nevada to the Colorado River or Arizona. No sources of regional inflow, other than the White River system, were identified.

The perennial yield of the carbonate-rock aquifers cannot exceed the total flow through them. However, part of this flow discharges by leaking into adjacent basin-fill aquifers. This component of flow probably is accounted for already in the water budgets (and estimated perennial yields) of those basins and cannot be included properly in estimates of the perennial yield of the carbonate-rock aquifers unless it is first subtracted from the basin-fill budgets. The perennial yield of the carbonate-rock aquifers of southern Nevada, therefore, should be defined in terms of the remainder of the total flow. Thus, the perennial yield is no more than the combined rates of discharge at regional springs in southern Nevada and at discharge areas in the Death Valley region (total, about 77,000 acre-ft/yr).

The actual rate at which water can be withdrawn without continual depletion of the ground-water reservoir might depend on how the resource is developed. This is because practical strategies to capture springflow and outflow can entail inefficiencies that allow part of the flow to avoid capture. As a result, for practical purposes, if wells some distance from regional springs are pumped in an effort to capture the springflows at logistically convenient locations, the sustainable rates of withdrawal might be considerably less than the present flow rates from the regional springs and discharge areas. Alternatively, if the basin-fill and carbonate-rock aquifers were managed together, the overall perennial yield still could not be more than the total recharge to the area. Management of both aquifer types together would not lessen overall effects on natural discharge, but rather it would permit tradeoffs as to what effects would be allowed, and where.

The other component of the carbonate-rock water resources is the large volume of water stored in the rocks. Because of the areal extent (10,000 mi^2) and great thickness of the carbonate rocks in the central corridor (between 3,000 and 19,000 ft and averaging about 12,000 ft), the total volume of rock is enormous. Carbonate rocks that might store and transmit water south of Pioche and Tonopah are estimated to have a total volume of about 20,000 mi^3. Borehole geophysical logs made during 1985-88 indicate that the total amount of open space within these rocks might be on the order of 1 to 10 percent, and previous studies at the Nevada Test Site indicate that about one-fifth of that space is interconnected and will allow ground water to move through it. Therefore, if selected parts of the carbonate-rock aquifers were dewatered, they might yield a water volume on the order of 1 percent of the aquifer volume. Assuming the same percentage of recoverable water in each cubic foot of aquifer, the total quantity of water stored in the rocks south of Pioche and Tonopah would be about 800 million acre-ft. For practical purposes, not all this water can be extracted. However, if an average of 100 ft of the aquifer's thickness could be dewatered, the central corridor could yield a volume of stored water of about 6 million acre-ft. However, the water stored in the upper 100 ft, if depleted regionally, would be replenished only by the equivalent of decades or centuries of recharge.

Ultimately, long-term development of the carbonate-rock aquifers would result in depletion of stored water, or capture of water that otherwise would discharge from the aquifers of southern Nevada and vicinity, or both. In many places, development might extract water from both carbonate rock and basin-fill aquifers. Reasonable tradeoffs among these alternative sources might be possible given (1) improved local understanding of the aquifers and the effects of development, and (2) careful planning of the developments.

Possible effects of developing the carbonate-rock aquifers include declining water levels, decreasing springflow rates, drying up of some streams, playas, and meadows, and changing water quality. These effects are direct or indirect responses to water-level changes associated with aquifer development, and are related to disturbances of the natural equilibrium between aquifer recharge and discharge.

Sustained effects within the carbonate-rock aquifers resulting from development of those aquifers have not been observed to date (1989). The magnitude and extent of water-level changes and, eventually, changes in springflow rates caused by development would depend on the geometry of the aquifers and their capacity to transmit and store water. Such effects can be calculated roughly by using either complex hydraulic calculations or computer models of ground-water flow. Hydraulic calculations indicate that water-level declines in extensive, **unconfined** carbonate-rock aquifers commonly would be between about 1 and 70 ft at a distance of 1 mi of a production well after 10 years of pumping at 1,000 gal/min. Projected water-level declines as far as 10 mi from the pumped well in a **confined** part of the aquifer would be of similar magnitude.

Neither set of calculations (for unconfined and confined conditions) accounts for the large directional influence of narrow, highly transmissive zones in the aquifers. Calculated declines around wells drawing water from these zones could be underestimated along the zones and overestimated away from the zones. The calculated drawdowns are based also on an assumption that the aquifer extends far beyond the area affected by pumping. Currently (1989), data that allow a more realistic representation of the real-world influence of aquifer boundaries are not available; however, several previous studies have included computer models that provide insight into the limitations of estimates. A review of those models indicates primarily that flow in the aquifers varies complexly at every scale. Regardless of whether the effect of pumping near Devils Hole is being modeled at a small scale or whether the flow toward Death Valley is being simulated at a regional scale, the geologic complexity of southern Nevada is hydrologically important. The numerous fault-zone orientations, differences in aquifer thicknesses, and difficult-to-delineate masses of noncarbonate rock that are barriers to flow require (1) that detailed aquifer descriptions be available before accurate site-specific estimates can be made, and probably (2) that those estimates be based on effective use of sophisticated computer models.

The potential for adverse effects in adjacent aquifers resulting from development of the carbonate-rock aquifers is another concern. Historical experience with such conditions is available at two areas in southern Nevada that have undergone development of basin-fill aquifers adjacent to carbonate-rock aquifers: Ash Meadows and Muddy River Springs. At Ash Meadows, direct connections between pumping from basin-fill aquifers and water-level declines in the carbonate rocks were demonstrated. Around the Muddy River Springs, in contrast, differing levels of development of water from basin-fill aquifers have resulted in minimal changes in water levels of the carbonate-rock aquifers. The difference between historical responses at these areas probably could not have been estimated before the aquifers were pumped and early effects observed.

Simple projections of water-level declines in the carbonate-rock aquifers, together with a review of historical experiences with the effects of drawing water from basin-fill aquifers adjacent to carbonate-rock aquifers, indicate that the carbonate-rock aquifers might be developed in some areas without introducing unacceptable effects. In other areas—for example, where a stressed aquifer is fully confined or where adjacent aquifers are closely connected—effects could be more severe. The effects of depletion of stored water and capture of water that otherwise would discharge from the aquifers would depend on site-specific conditions around areas where water is withdrawn from the aquifers. Confidence in the projection of effects that might result from the carbonate-rock aquifers will remain limited until observations are available that document changes as the aquifers respond locally to long-term pumping stresses.

Initially, assurances that the adverse effects of development will not overshadow benefits cannot be made with a high degree of confidence. However, if staged development were undertaken together with adequate monitoring, effects of continued or increased development could be estimated with progressively higher degrees of confidence. Staging means not developing the resources in one large step but rather starting with small projects that are augmented gradually as conditions and confidence warrant. This approach allows the effects of development to be observed and analyzed continually, so that the adverse effects and the benefits of development can be judged and the effects reversed or mitigated if they prove to be too costly (in economic and environmental terms). Adequate monitoring means monitoring that provides data for a timely and sound judgment of adverse effects and benefits. Monitoring of all aquifers that may be affected, and all hydrologic conditions that may reflect these effects, should provide a basis for sound judgments. Regular or continuous monitoring that permits the recognition of early effects of development and that allows these effects to be distinguished from natural fluctuations, would provide information for timely judgments.

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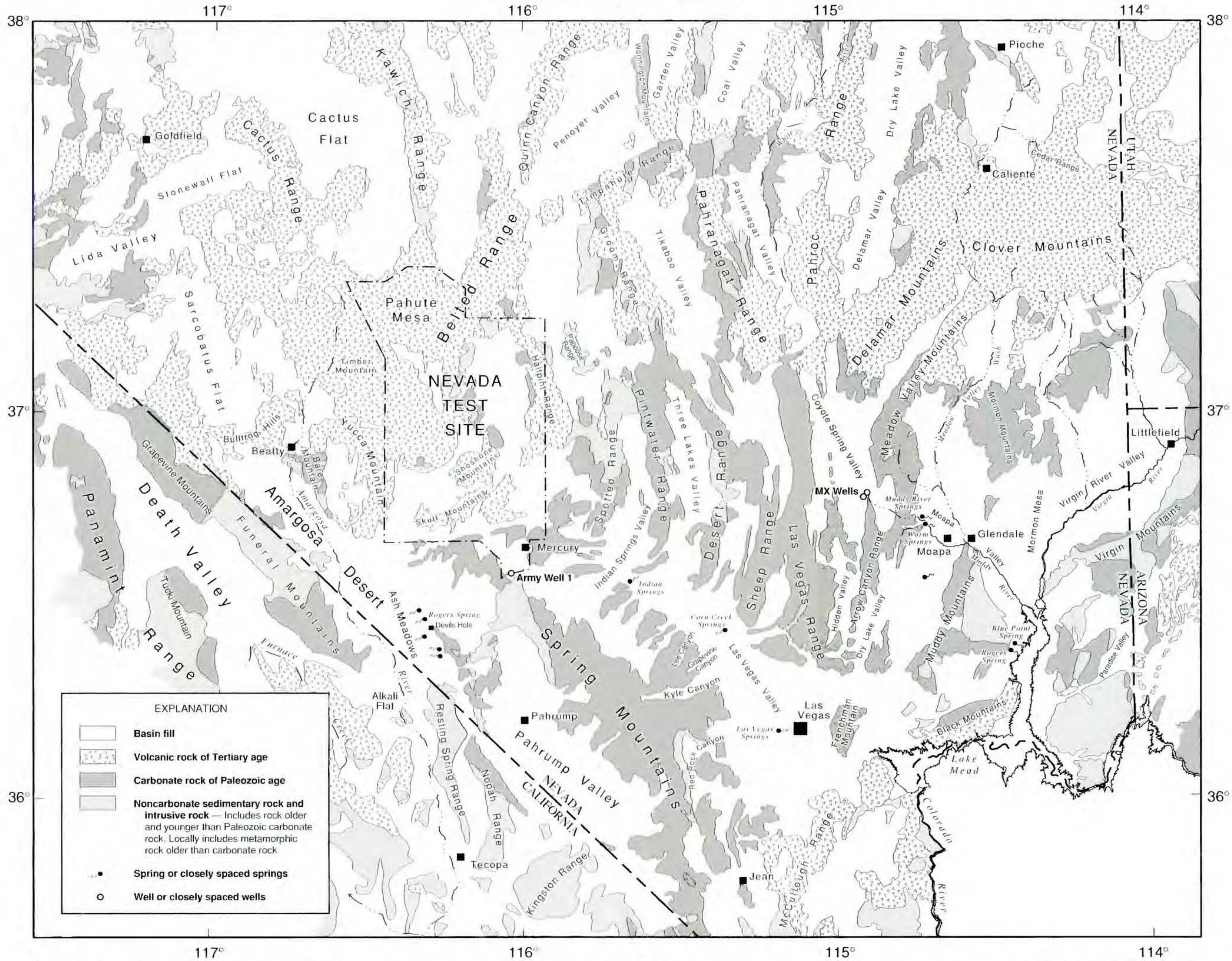
U.S DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY

PREPARED IN COOPERATION WITH THE
STATE OF NEVADA
AND
LAS VEGAS VALLEY WATER DISTRICT

WATER-RESOURCES INVESTIGATIONS REPORT 91-4146
Geology and Geographic Features—PLATE 1
Dettinger, M.D., and others, 1995, Distribution of Carbonate-Rock Aquifers and the Potential for Their Development, Southern Nevada and Adjacent Parts of California, Arizona, and Utah



Water bodies from U.S. Geological Survey digital data, 1:100,000 scale, 1978-87; Universal Transverse Mercator Projection, Zone 11. Hydrology from U.S. Geological Survey, 1:24,000 scale, 1985; Universal Transverse Mercator Projection, Zone 11

SCALE 1:1,000,000

MILES: 10 0 10 20 30 40 50

KILOMETERS: 10 0 10 20 30 40 50 60 70

Geology modified from Plume and Carlton (1988), Stewart and Carlson (1978), and Wilson and Moore (1959). Geology in lower left and lower right corners of map area not shown

MAP SHOWING GEOLOGY AND GEOGRAPHIC FEATURES OF SOUTHERN NEVADA AND ADJACENT PARTS OF ARIZONA, CALIFORNIA, AND UTAH

by
Michael D. Dettinger, James R. Harrill, and Dwight L. Schmidt, U.S. Geological Survey,
and
John W. Hess, Desert Research Institute
1995

U.S DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY

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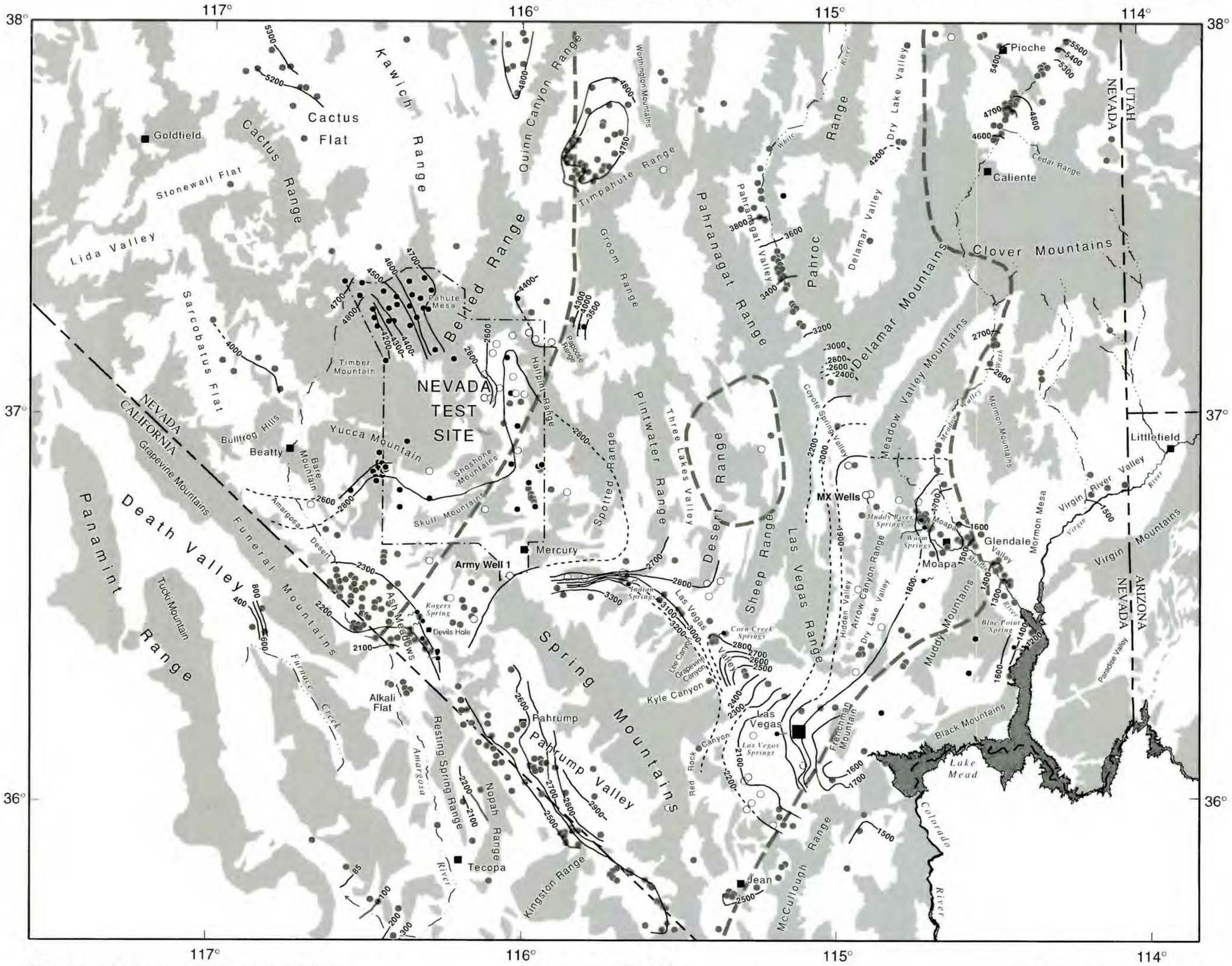
WATER-RESOURCES INVESTIGATIONS REPORT 91-4146
Ground-Water Levels—PLATE 2
Dettinger, M.D., and others, 1995, Distribution of Carbonate-Rock Aquifers and the Potential for Their Development, Southern Nevada and Adjacent Parts of California, Arizona, and Utah



Water bodies from U.S. Geological Survey digital data, 1:100,000 scale, 1978-87; Universal Transverse Mercator Projection, Zone 11. Hydrology from U.S. Geological Survey, 1:24,000 scale, 1985; Universal Transverse Mercator Projection, Zone 11

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Geology modified from Plume and Carlton (1988), Stewart and Carlson (1978), and Wilson and Moore (1959). Geology in lower left and lower right corners of map area not shown

EXPLANATION

- **Basin fill**
- **Consolidated rock**
- **Area of low hydraulic gradient**—Water-level altitudes are generally between 1,800 and 1,820 feet above sea level
- **Boundary of central corridor of carbonate-rock aquifer in Nevada**—Area is underlain by thick sequences of carbonate rock; outside corridor, carbonate rock is thin, or present as isolated bodies
- 2800 **Water-level contour** — Modified from Thomas and others (1986, sheet 2) by Thomas and Welch (in press) to account for additional water-level data. Contour interval, in feet, is variable. Datum is sea level
- **Spring or closely spaced springs**
- **Basin-fill well**
- **Carbonate-rock well**
- **Consolidated rock other than carbonate rock well**

MAP SHOWING GROUND-WATER LEVELS IN SOUTHERN NEVADA AND ADJACENT PARTS OF ARIZONA, CALIFORNIA, AND UTAH

by
Michael D. Dettinger, James R. Harrill, and Dwight L. Schmidt, U.S. Geological Survey,
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