Application of Digital Profile Modeling Techniques to Ground-Water Solute Transport at Barstow, California

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ILLUSTRATIONS

CONVERSIONS FACTORS

For readers who may prefer to use metric units rather than English units, the conversion factors for the terms used in this report are listed below:

Multiply English unit	*By—*	*To obtain metric unit*
acres	4.047×10^{-1}	ha (hectares)
ft (feet)	3.048×10^{-1}	m (meters)
ft/d (feet per day)	3.048×10^{-1}	m/d (meters per day)
ft/s (feet per second)	2.633×10^{3}	m/d (meters per day)
ft^3/s (cubic feet per second)	2.832×10	L/s (liters per second)
gal/min (gallons per minute)	6.309×10^{-2}	L/s (liters per second)
in. (inches)	2.54×10	mm (millimeters)
mi (miles)	1.609	km (kilometers)

APPLICATION OF DIGITAL PROFILE MODELING TECHNIQUES TO GROUND-WATER SOLUTE TRANSPORT AT BARSTOW, CALIFORNIA

By S. G. Robson

ABSTRACT

This study investigated the use of a two-dimensional profile-oriented water-quality model for the simulation of head and water-quality changes through the saturated thickness of an aquifer. The profile model is able to simulate confined or unconfined aquifers with nonhomogeneous anisotropic hydraulic conductivity, nonhomogeneous specific storage and porosity, and nonuniform saturated thickness. An aquifer may be simulated under either steady or nonsteady flow conditions provided that the ground-water flow path along which the longitudinal axis of the model is oriented does not move in the aquifer during the simulation time period. The profile model parameters are more difficult to quantify than are the corresponding parameters for an areal-oriented water-quality model. However, the sensitivity of the profile model to the parameters may be such that the normal error of parameter estimation will not preclude obtaining acceptable model results.

Although the profile model has the advantage of being able to simulate vertical flow and water-quality changes in a single- or multiple-aquifer system, the types of problems to which it can be applied is limited by the requirements that (1) the ground-water flow path remain oriented along the longitudinal axis of the model and (2) any subsequent hydrologic factors to be evaluated using the model must be located along the land-surface trace of the model. Simulation of hypothetical ground-water management practices indicates that the profile model is applicable to problem-oriented studies and can provide quantitative results applicable to a variety of management practices. In particular, simulations of the movement and dissolved-solids concentration of a zone of degraded ground-water quality near Barstow, Calif., indicate that halting subsurface disposal of treated sewage effluent in conjunction with pumping a line of fully penetrating wells would be an effective means of controlling the movement of degraded ground water.

INTRODUCTION

Miscible displacement problems are of great economic importance in a society increasingly confronted by problems of ground-water quality degradation. As a result, increasing emphasis in water-resource evaluations has been placed on use of mathematical models that describe the movement and dispersion of contaminants in aquifers. This study involves the use of one such solute-transport computer program as a profile-oriented (x-z

plane) model. The program was applied to a well-documented case of ground-water-quality degradation near Barstow, Calif., to evaluate the profile model as a potential tool in this geographic area and in studies of ground-water-quality degradation in general. The field application also provides a means of examining the data requirements, model-parameter sensitivity, and the advantages and disadvantages of the profile model with respect to an areal-oriented solute transport model of the Barstow area.

The program used for this work was developed by Bredehoeft and Pinder (1973) and employs an implicit alternating-direction iterative technique to solve the ground-water flow equation coupled with a method of characteristics solution to the mass-transport equation. Although the program was originally written as an areal-oriented (x–y plane) model, the program modifications made in changing the x–y orientation of the program to an x–z orientation were of a supportive nature and did not alter the numerical methods used in the original program.

The geohydrologic character of the shallow alluvial aquifer near Barstow, Calif., has been documented by previous water-quality modeling work in the area (Robson, 1974); basic data on the location, nature, and source of pollutants (Hughes, 1975; Hughes and Patridge, 1973); and other interpretive studies in the Barstow area (geologic and hydrologic characteristics of the aquifer) (Hughes and Robson, 1973; Hardt, 1971; Miller, 1969).

Barstow is 96 mi northeast of Los Angeles in the Mojave Desert region of southern California and is adjacent to the normally dry Mojave River (pl. 1). Precipitation averages about 5 in. per year and produces negligible ground-water recharge. Ground water in storage is the only reliable source of water for the main water purveyors (the city of Barstow and the U.S. Marine Corps Supply Center at Nebo). The quantity of ground water in storage is large in relation to the local demand and is of good chemical quality in areas not affected by serious degradation (Miller, 1969).

The main aquifer near Barstow consists of permeable younger alluvium of Holocene age, deposited by the Mojave River, and tributary alluvial fans. The younger alluvial aquifer is about 1 mi wide and about 100 ft thick. Periodic floodflow in the Mojave River is the main source of recharge. The younger alluvial aquifer is underlain in some areas by less permeable older alluvium of Pleistocene age and in other areas by consolidated rocks of Quaternary and Tertiary age that yield very little water to wells (pl. 1). Along the south side of the Mojave River the older alluvium contains water of poorer chemical quality than that in the younger alluvium, but the older unit is of relatively lower permeability and contributes only a fraction of the total recharge to the younger alluvial aquifer.

Water levels in the younger alluvial aquifer show the effects of the intermittent surface flow in the Mojave River. Steady ground-water-level declines in some areas exceed 40 ft during a period of dry years when no surface flow occurs and may be followed by as much as 50 ft of recovery during a year with ample surface flow (Hardt, 1969, p. 9).

The chemical quality of water in the main aquifer east of Barstow has been deteriorating since 1951 (Miller, 1969, p. 37). The city of Barstow and the Atchison, Topeka and Santa Fe Railway Co. sewage-treatment ponds were in the Mojave River north of Barstow prior to 1969 and were adjacent to and upgradient from an area of degraded ground-water quality. In 1969 a new treatment plant and ponds designed to meet the combined needs of the city of Barstow and the railway company went into operation about 3 mi east of Barstow (pl. 1). Treated effluent percolating from the new ponds is producing a second zone of degraded ground-water quality. This second zone was chosen as the primary topic of concern in the profile model, and the ground-water flow path downgradient from the lower Barstow sewage ponds dictated the location of the profile to be modeled.

In some areas ground-water contamination results from the deep percolation of irrigation water. Prior to 1973 the Marine Corps Supply Center irrigated a 30-acre golf course (pl. 1) with effluent from its sewage-treatment plant. This practice produced ground-water recharge of much poorer chemical quality than would have occurred if fresh water were used for irrigation. The resulting zone of degraded water was also considered in the profile model.

CONDITIONS AMENABLE TO PROFILE SOLUTE-TRANSPORT MODELING TECHNIQUES

The model may be used to simulate either confined or unconfined aquifers with nonhomogeneous anisotropic hydraulic conductivity, nonhomogeneous specific storage and porosity, and nonuniform saturated thickness. The movement of conservative (nonreactive) chemical constituents may be simulated, with sources of contamination varying both in location and chemical concentration, and in duration of occurrence.

To simulate ground-water flow in the x–z plane, the longitudinal axis of the profile model must be oriented along a ground-water flow path, because in general the part of the aquifer to be modeled must function as a two-dimensional flow system. As a result of this requirement the profile model can, in a strict sense, be applied only to steady-flow conditions or nonsteady flow conditions that do not appreciably alter the location of the initial ground-water flow path.

Near the Barstow sewage ponds some ground-water flow is orthogonal to the plane of the model because of the alinement of the ponds. The effects of the orthogonal flow are considered in the model head and mass-transport calculations by simulating a quasi three-dimensional ground-water flow system near the model nodes affected by orthogonal flow. A constant potential boundary is assigned the head values that occur in the aquifer beyond the effective radius of influence of the ponds. These heads, in conjunction with the horizontal hydraulic conductivity of the aquifer and the distance from the profile to the constant potential boundary, enable the model to simulate head and water-quality conditions in the part of the profile affected by orthogonal flow.

The land-surface orientation of a source of ground-water recharge or discharge may determine whether or not orthogonal flow will result. For example, had the Barstow sewage ponds been alined at right angles to the direction of ground-water movement, a nearly uniform flow field would have occurred near the ponds, and no orthogonal flow would have occurred in the model.

MODEL PARAMETER SENSITIVITY AND ESTIMATION TECHNIQUES

The parameter data required by a model have an important bearing on the potential usefulness of the model. If, for example, the model results are particularly sensitive to the value of a parameter which is difficult to quantify, the accuracy and usefulness of the model could be seriously compromised. Because the profile model uses the same numerical methods as the areal model, there are similarities in data requirements and sensitivity. Reports by Bredehoeft and Pinder (1973), Konikow and Bredehoeft (1973), Konikow (1976), Robertson (1974), and Robson (1974) describing other flow and transport models have discussed many of these parameter requirements, and emphasis in this report will be given only to those data requirements that are unusual because of the vertical orientation of the model.

NODE SIZE

The implicit alternating-direction technique used in the solution of a flow equation can be subject to head convergence problems when the node length greatly exceeds the node height in heavily stressed nonuniform aquifers (Trescott and others, 1976; Winters, 1976). Node sizes of 10x500 ft (vertical x horizontal) and 18x1,220 ft were tried in this study. In both cases head convergence was achieved with 5-10 iterations per time step, and the head and concentration distributions produced using the two node sizes were identical. The model used for subsequent simulations used a 10x21 grid of nodes 10x500 ft in order to simulate a zone 100 ft high and 10,500 ft long.

Work by Trescott, Pinder, and Larson (1976) and Winters (1976) indicates that the strongly implicit procedure for solving the ground-water flow equation can in some cases provide a much more efficient solution to the equations than the implicit-alternating direction technique. The alternating-direction procedure was used successfully, however, by Pinder and Cooper (1970) and was found to be satisfactory for the conditions simulated in this study.

The length of the simulation period in conjunction with grid size and ground-water velocities can combine to consume excessive computer time in the solution of the transport equation. Because this limitation is inherent in the mathematics of this model, use of small grid spacing should be avoided if lengthy simulation periods are required. Computer execution time was about 90 seconds on a Univac 1108,[1] about 120 seconds on an IBM 360/75,[1] and about 50 seconds on a CDC CYBER-74[1] with a 10x500-ft grid spacing, a simulation period of 5 years, and maximum ground-water velocities of about 450 ft per year.

It has been common practice in profile modeling to consider the model to have a unit width with hydraulic conductivity and specific storage as parameters and recharge and discharge scaled per unit width (Society of Petroleum Engineers, 1973; Grisak and Cherry, 1975; Pinder and Cooper, 1970; Winters, 1976). The model used in this investigation is configured so that a profile (or narrow section) of any width may be simulated and the hydraulic conductivity and specific storage data adjusted for the nonunity width. A profile width of 500 ft was used in this study and the rates of underflow recharge and discharge and of recharge from the lower Barstow sewage ponds and the U.S. Marine Corps Supply Center golf course were scaled to this width.

HYDRAULIC CONDUCTIVITY

The profile model allows variations in the hydraulic conductivity parameter in order to simulate anisotropic and nonhomogeneous conditions in the aquifer.

To paraphrase Weeks (1969, p. 197), most aquifers composed of

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[1]The use of brand names in this report is for identification purposes only and does not imply endorsement by the U.S. Geological Survey.

clastic sediments have a higher hydraulic conductivity (K) along the bedding plane than across the bedding plane and thus are anisotropic with respect to K. This anisotropy occurs in part because plate-shaped grains within the aquifer tend to be deposited with their flat surfaces parallel to the bedding plane. Such orientation increases the tortuosity of vertically interconnected pores, thus reducing the vertical K of the aquifer. Anisotropy caused by this occurs in even very small samples of sediment.

For the purposes of this study, anisotropic K is also assumed to occur because of small-scale interbedding of fine-grained and coarse-grained sediments within the aquifer. Such interbedding affects hydraulic conductivity of vertical flow much more than it does that of horizontal flow. Although the interbedding represents nonhomogeneity, rather than anisotropy, its effects on the hydraulic conductivity of a large sample of aquifer may be approximated by treating the sample as homogeneous but anisotropic.

Nonhomogeneous conditions due to large-scale interbedding of fine-grained and coarse-grained sediments were considered separate from anisotropy if the beds exceeded an arbitrary thickness of 2 ft (0.6 m). Subsequent work indicated that this definition of isotropy and homogeneity did not create significant problems in quantifying the K parameter for use in the model.

Because aquifer K data that quantify the effects of aquifer anisotropy and nonhomogeneity are not commonly available and field determinations can be complex and costly, it is important to know the sensitivity of the profile model to these parameters. A profile flow model was constructed to evaluate the effects of varying the ratio of horizontal hydraulic conductivity (K_h) to vertical hydraulic conductivity (K_v) produced by anisotropy in a homogeneous aquifer. The model simulated recharge at the upstream model boundary and Barstow sewage ponds (0.24 ft^3/s and 0.12 ft^3/s) and discharge (0.36 ft^3/s) at the downstream model boundary in a 40-ft interval near the base of the profile. A uniform K_h of $1.2x10^{-3}$ ft/s was used, and the K_v was varied for each of four runs in order to produce K_h/K_v ratios from 1 to 1000. The results (fig. 1) indicated that large vertical head differences can be produced if the ratio of K_h/K_v is greater than one or two orders of magnitude. The work of Johnson (1963a) indicates that alluvial aquifers can have K_h/K_v ratios within this range. Gillham and Farvolden (1974) conducted a sensitivity analysis of parameter data in a profile-oriented hypothetical flow model. They considered water table and vertical head gradients that were about 10 and 100 times greater than those found near Barstow. Their results indicated a greater sensitivity of head to hydraulic conductivity changes than is shown in figure 2 because of the more heavily stressed aquifer. These results emphasize that the sensitivity of a model is a function of the parameter data, and the transfer value of sensitivity analyses must be judged on the basis of similarities in model parameters.

Two field techniques for determining K_h/K_v that were investigated include (1) a direct estimate of the average K_h and K_v for the aquifer, based on the horizontal and vertical extent of movement of the zone of degraded water produced by percolation from the city of Barstow sewage-treatment plant and (2) an aquifer test on a partially penetrating well using the techniques of Weeks (1969) or Mansur and Dietrich (1965).

It is possible to calculate the average K_h/K_v ratio in a zone of degraded ground water if (1) the extent of longitudinal and vertical movement of the zone over a period of time is known and (2) the steady-flow longitudinal and vertical head configuration in the degraded zone is known. If steady-flow conditions are assumed, the distance the degraded zone has moved may be expressed as:

$$D_h = V_h\ t = \frac{K_h\ I_h}{\phi}\ t$$

$$\therefore K_h = \frac{D_h\ \phi}{I_h\ t}$$

$$D_v = V_v\ t = \frac{K_v\ I_v}{\phi}\ t$$

$$\therefore K_v = \frac{D_v\ \phi}{I_v\ t}$$

and

$$\frac{K_h}{K_v} = \frac{D_h\ I_v}{D_v\ I_h},$$

where the subscripts h and v denote horizontal and vertical components and

D = distance the degraded zone has moved,
V = velocity,
t = time,
K = hydraulic conductivity,
I = head gradient, and
ϕ = porosity.

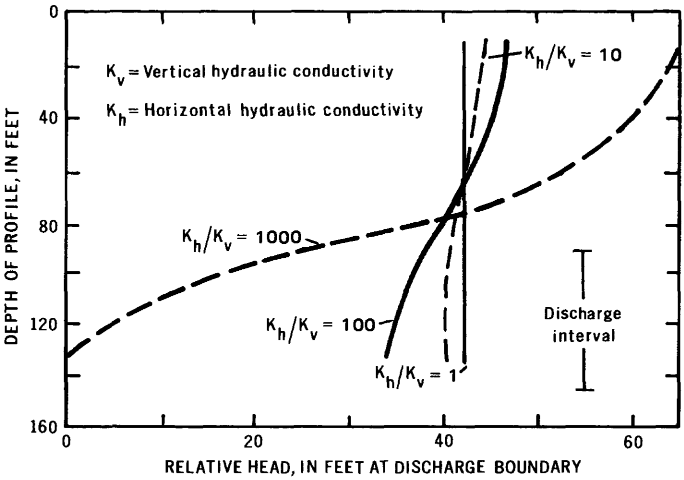


FIGURE 1.—Head variation produced by anisotropic hydraulic conductivity in profile model.

The distance the degraded zone near the Barstow sewage-treatment ponds moved during the first 5 years of operation was used to determine D_v and D_h. Heads in nearby observation wells were used to determine I_v and I_h. Using these data and a uniform ϕ of 40 percent (Robson, 1974, p. 26) in the above equations, a K_h/K_v ratio of about 200 is obtained, and K_h = 130 ft/d and K_v = 0.6 ft/d. This technique gave results that were very close to those indicated by previous studies; however, the technique provides only an estimate of K_h/K_v, as D_v and D_h are often difficult to determine accurately. As a result, the K_h/K_v ratio is probably estimated with an accuracy of about ±50 percent.

Subsequent model simulations made with a K_h/K_v ratio of 200 and 300 produced model results that were, for all practical purposes, identical. This indicates that the likely magnitude of any error that might be present in the estimate of K_h/K_v will not have a significant effect on the model results.

By pumping a well perforated in the lower part of an aquifer and monitoring head changes in shallow observation wells, Weeks (1969) and Mansur and Dietrich (1965) were able to calculate the ratio K_h/K_v in shallow glacial outwash and alluvial aquifers. A similar aquifer test was undertaken in this study to provide a basis for evaluating the results of the previous technique, but instead of pumping from the aquifer, a 2-in.-diameter well was used to inject water in an interval 95–100 ft below the water table. Head changes were monitored at 16 observation wells, 5 of which were located within 10 ft of the injection well. Water was injected at a rate of 100 gal/min under a pressure head of as much as 75 ft above the water table. No significant head changes were detected in any of the observation wells, which suggests a high ratio of K_h/K_v.

Both of the above techniques may provide a means of estimating the value of K_h/K_v resulting from anisotropy but are not well suited to estimating the value of K for the various beds in the aquifer. The profile model was shown to be sensitive to such nonhomogeneity by the comparison of two model runs with uniform anisotropy and differing values for K. In the first run, K stratification similar to that shown in figure 2 was assumed to range between plus and minus 50 percent of the mean K in the

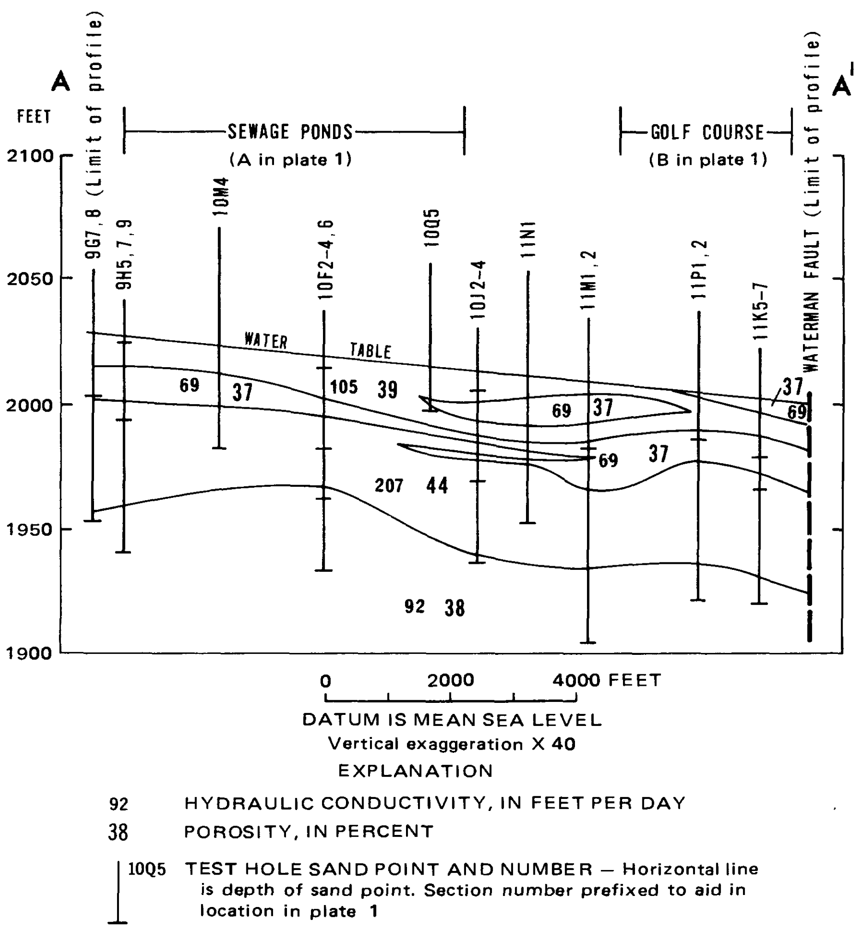


FIGURE 2.—Profile model showing distribution of hydraulic conductivity and porosity.

profile. The mean *K* in the profile (calculated as a weighted average) was 138 ft/d. In the second run a mean *K* of 138 ft/d was used uniformly throughout the profile. The nonhomogeneous *K* produced the following changes from the homogeneous *K* conditions: (1) A minimal (3 percent) increase in longitudinal head gradient, (2) a 20-percent increase in vertical head gradient, and (3) a maximum concentration reduction of about 10 percent at some points in the zone of degraded water. The quantitative interpretation of geophysical logs was next investigated as a possible means of quantifying the effects of *K* stratification within the profile.

Thirteen observation wells located near the profile penetrate most of the saturated thickness considered in the model. Natural gamma, gamma-gamma, and neutron geophysical logs were run on these wells. Seventeen drive-core samples were collected from two of the observation wells during drilling. Vertical and horizontal hydraulic conductivity, bulk density, porosity, and grain-size analyses were run on selected samples. The data were used in an attempt to calibrate the geophysical logs in a manner similar to the work of either Bredehoeft (1964) or Rabe (1957). In general, Bredehoeft developed a relation between porosity and hydraulic conductivity for an aquifer in which differences in these two characteristics were primarily due to grain size, sorting, and degree of cementation. Rabe, by contrast, worked with an aquifer in which clay content was the primary factor affecting hydraulic conductivity and developed a relation between *K* and natural gamma radiation.

The geophysical logs and core data strongly indicate that the *K* of the aquifer in the study area is controlled primarily by grain size and sorting rather than by clay content. Even though a correlation could be shown between core-sample porosity and neutron log response and between core-sample porosity and core-sample hydraulic conductivity, the use of these relations to calibrate the geophysical logs and thereby estimate the aquifer *K* was not successful. This was due to abnormally low values for the laboratory-determined core-sample hydraulic conductivity, which probably resulted from compaction of the samples during collection. As a result of the compaction, the core samples probably do not possess the hydraulic characteristics of the aquifer.

A comparison of the length of the cores in the barrels with the distance the core barrels were driven indicates a potential sample compaction of about 30 percent. This degree of compaction is further indicated by the porosity of the core samples, which averaged about 0.30 in an aquifer where other studies have indicated an average porosity of about 0.40 (Robson, 1974). These data suggest that the compaction of a sample by 30 percent or less may produce a 10- to 100-fold decrease in the hydraulic conductivity of the sample.

In highly porous unconsolidated coarse-grained aquifers, undisturbed core samples are difficult to obtain. Without these data it is not possible to adequately calibrate geophysical logs. Uncalibrated geophysical logs can, however, be used on a qualitative basis and in this study appear to be the best technique available for determining the nonhomogeneous variations in hydraulic conductivity in a vertical section of aquifer.

A modification of the above method of estimating K involves empirical correlations of the core sample particle grain size and sorting data (data that are little affected by compaction of the sample) with hydraulic conductivity. Curves presented by Johnson (1963b, fig. 23) and Masch and Denny (1966, fig. 8) show similar relations among K, sorting coefficients, and median grain diameter. These curves were extrapolated to include the range of uniformity coefficients found in the core samples for this study (fig. 3). The K for each core sample was then estimated, on the basis of the median grain diameter and uniformity coefficient of the sample.

If it is assumed that, because of the similarities of aquifer materials and sampling techniques, the core samples all underwent about the same degree of compaction (or no compaction), then the core porosity data may be treated as porosity indices. After using the porosity indices to calibrate the neutron log response, the method of least squares may be used to relate the core sample K estimated from the particle-size distribution to the porosity indices as measured from the neutron log response (fig. 4). This enabled the neutron log data to be related to the K of the aquifer and allowed estimation of K in wells without core-sample data.

The average K for the saturated thickness of the aquifer calculated by this technique was smaller than that found in other studies (Robson, 1974; Hardt, 1971) by a factor of about 3.5. By applying this factor uniformly to the log K data, a K distribution (fig. 2) was produced that agreed with previous data and generated the proper head and water-quality responses in the profile model.

Because of the assumptions made in applying this technique, the magnitude of the above factor is probably indicative of the error to be expected with use of this technique. In spite of the inaccuracy, the procedure provides an orderly means of estimating K with an accuracy that is probably better than could be obtained by an intuitive interpretation of the geophysical logs.

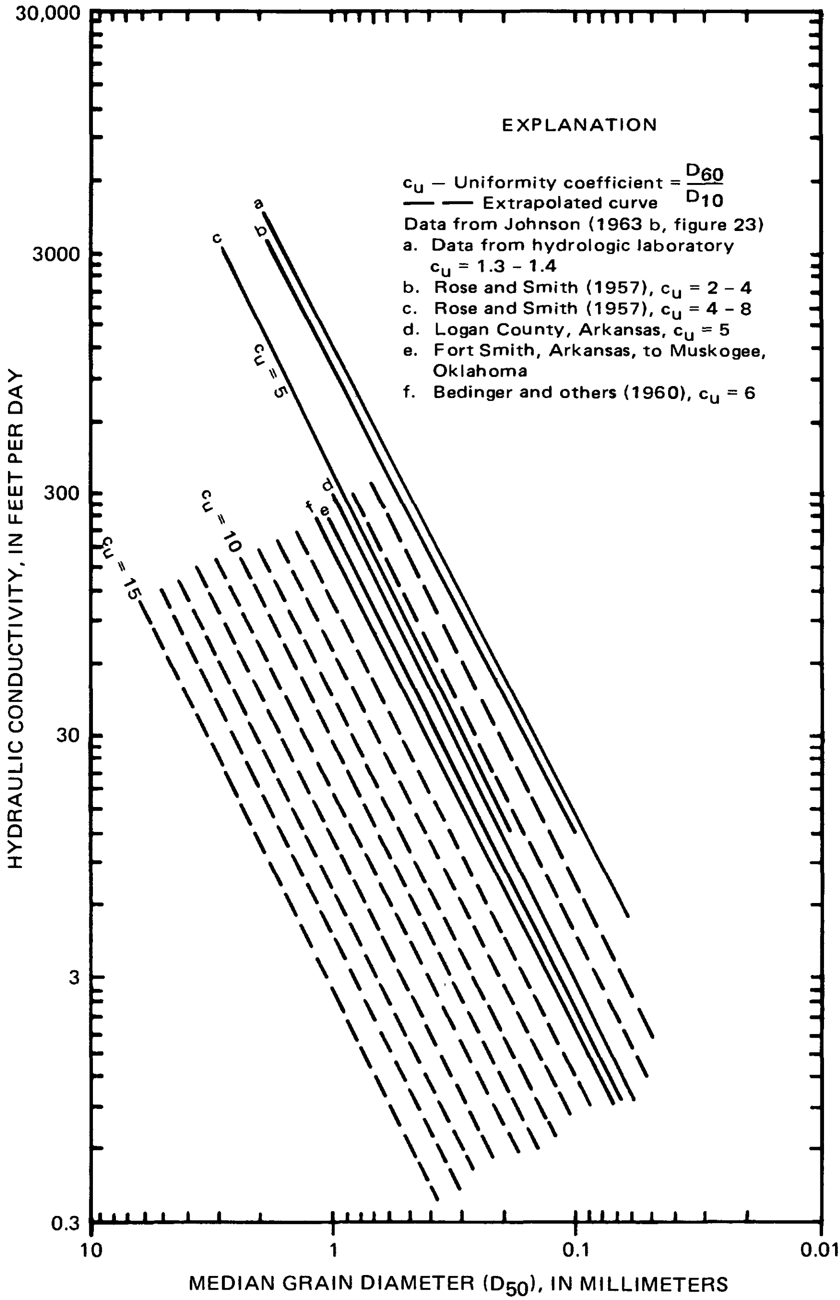


FIGURE 3.—Relations among hydraulic conductivity, uniformity coefficient, and median grain diameter.

The nonhomogeneity shown in figure 2 will have an effect on the ratio of horizontal to vertical hydraulic conductivity of the aquifer. The mean horizontal and vertical hydraulic conductivity of the aquifer may be calculated by:

$$K_h = (K_1 b_1 + K_2 b_2 + ... + K_n b_n)/(b_1 + b_2 + ... + b_n) \quad (1)$$

$$K_v = (b_1 + b_2 + ... + b_n)/\left(\frac{b_1}{K_1} + \frac{b_2}{K_2} + ... + \frac{b_n}{K_n}\right), \quad (2)$$

where K_h = mean horizontal hydraulic conductivity,
K_v = mean vertical hydraulic conductivity,
b = vertical thickness of each zone of differing hydraulic conductivity, and
K = hydraulic conductivity of each zone (K assumed to be isotropic).

The effects on the K_h/K_v ratio of various thicknesses and hydraulic conductivities of beds in an aquifer are shown in figure 5. The ranges in bed thickness and K shown in figure 3 are such that the K_h/K_v ratio resulting from large-scale nonhomogeneity is less than 2.0. Thus, the large-scale bedding in the aquifer does not significantly contribute to the ratio of $K_h/K_v \simeq 200$ attributed to

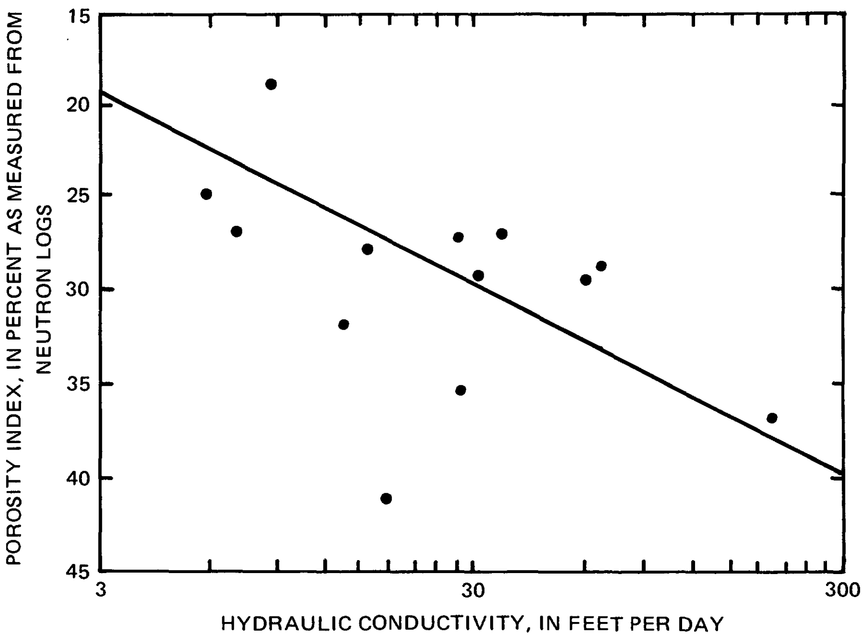
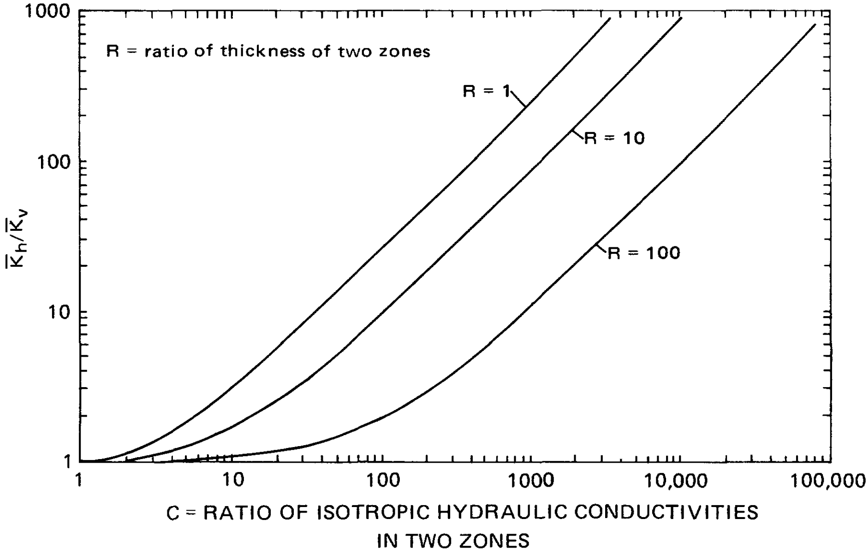


FIGURE 4.—Relation between porosity index and hydraulic conductivity.



EXPLANATION

Example 1	Example 2
$b_1 = 2$ $K_1 = 10$	$b_1 = 2$ $K_1 = 1000$
$b_2 = 20$ $K_2 = 1000$	$b_2 = 20$ $K_2 = 10$
$R = \frac{b_2}{b_1} = 10$	$R = \frac{b_2}{b_1} = 10$
$C = \frac{K_2}{K_1} = 100$	$C = \frac{K_1}{K_2} = 100$

From graph or equations (1) and (2)

$\frac{\overline{K}_h}{\overline{K}_v} = 9.1$ $\frac{\overline{K}_h}{\overline{K}_v} = 9.1$

FIGURE 5.—Relations among bed thickness, hydraulic conductivity, and K_h/K_v.

anisotropy. This is consistent with observed stratification in the aquifer, for although thick silt or clay layers were not encountered, thin stringers of silt or clay are common and could well exert a controlling influence on K_h/K_v attributed to anisotropy.

SPECIFIC STORAGE, POROSITY, AND DISPERSIVITY

Specific storage (S) data must be considered in the profile model if nonsteady flow conditions are to be simulated. Unlike areal models, a profile model of an unconfined aquifer must consider the effects of both unconfined and confined storage. The unconfined S applies to the water table, and the confined S applies to the aquifer below the water table. The profile-model computer program approximates this condition by assigning the unconfined S to nodes at the water table and reassigning the S to adjacent nodes if the water table declines into these nodes.

In unconfined aquifers, the quantity of water released from (or taken into) storage in the confined S parts of the aquifer will be negligible in comparison to the quantity released from (or taken into) storage at the water table. For this reason the magnitude of and variations in the confined S generally need not be considered when modeling an unconfined aquifer. An exception occurs if delayed gravity responses of the water table during pumping tests are to be considered (Neuman, 1972 and 1974). In confined nonsteady flow conditions the magnitude and nonhomogeneity of S will affect the computed head and chemical concentration distributions. Although an evaluation of techniques that could be used to quantify the S nonhomogeneity in confined aquifers was beyond the scope of this study, an approximation of the confined S can be made by multiplying the thickness of the aquifer by 1×10^{-6} after the technique discussed by Lohman (1972, p. 53). The unconfined S used in this study was based on specific-yield data used in previous hydrologic models of the area (Hardt, 1971; Robson, 1974).

The sensitivity of the water-quality model to varying degrees of nonhomogeneous porosity can be shown by a comparison of model runs having the same mean porosity ($\bar{N}$) but differing in the magnitude of nonhomogeneity within the profile. The porosity variations were patterned after those developed in the course of estimating K (fig. 2). In the first simulation the porosity of the Barstow profile model was homogeneous (40 percent). In the second simulation the zones of highest porosity in figure 2 were increased 10 percent, and the zones of lowest porosity were decreased 10 percent, with intermediate values adjusted to maintain an average porosity of 0.40 in the model. For the third simulation the range of adjustment was ±50 percent. Porosity changes affect ground-water velocity and consequently the rate of contaminant movement in the model aquifer. As shown in figure 6, the variations from the solute concentration produced by the homogeneous porosity were minimal when the porosity nonhomogeneity was $\bar{N}$ ±50 percent. The sensitivity of the model to porosity nonhomogeneity is such that errors in estimating the porosity (fig. 2) are probably not a serious source of error in the model results. Because porosity can be calculated from geophysical logs, aquifer porosity nonhomogeneity should not be difficult to estimate in other studies, provided the geophysical logs can be properly calibrated either by use of core samples or by other techniques.

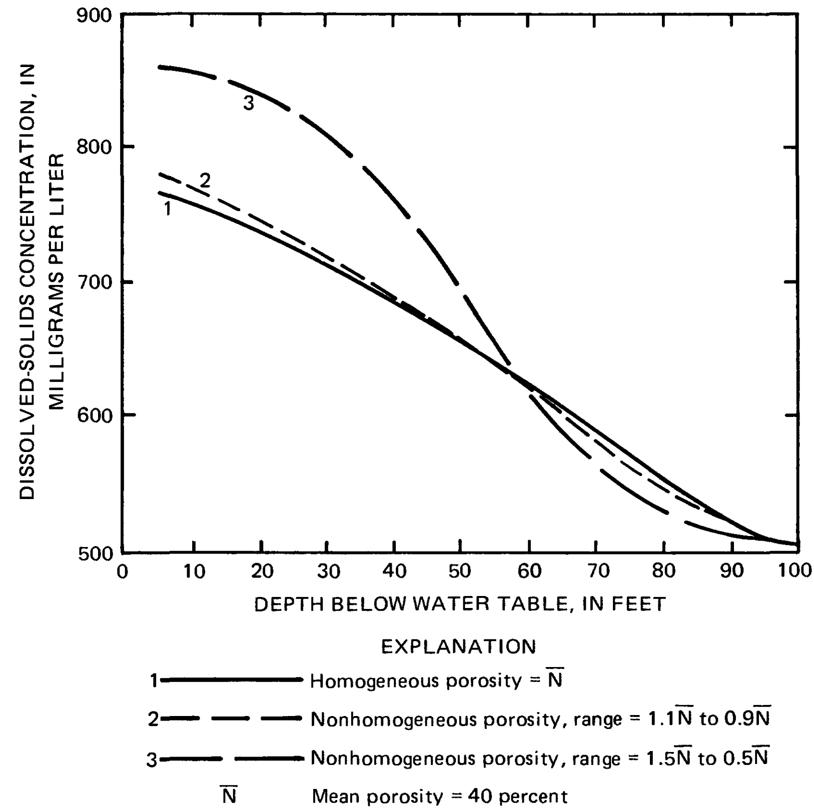


FIGURE 6.—Solute concentration in column near center of profile model, produced by varying degrees of nonhomogeneous porosity.

Errors in estimating the inverse of porosity ($1/n$), hydraulic conductivity, or total flux produce errors in model-calculated velocities in direct proportion to the magnitude of the parameter error. Because the ground-water velocity distribution controls the movement of contaminants in the aquifer, an error in one of these parameters will produce a similar error in the concentration distribution. Because a significant error in any one of the above parameters can adversely affect the predictive accuracy of the model, care must be exercised in calibrating the model to assure that the best-defined parameters are used to maximum advantage in modifying the values of the more poorly defined parameters. Such modification must make maximum use of available data. Gillham and Farvolden (1974) found that in a nonhomogeneous anisotropic aquifer identical head distributions could be calculated using different sets of parameter values.

The effects of variations in dispersivity on model-generated concentrations have been shown in previous studies (Robson, 1974) and have transfer value to the profile model. In the areal-oriented water-quality model, the ratio of lateral dispersivity (D_L) to transverse dispersivity (D_T) was 3.3. In the profile model this ratio was increased to about 330 in order to produce a reasonable model response. Varying this ratio changes the magnitude of the dispersivity applied to the vertical component of flow and produces responses in both the vertical and horizontal model-generated concentration gradients. The difference between the ratios of D_L to D_T used in the two models is probably reasonable. In the areal-oriented model D_L and D_T are essentially measures of mixing along aquifer bedding planes, as is D_L in the profile model, whereas D_T in the profile model is primarily a measure of mixing across bedding planes. Because much greater differences in aquifer characteristics occur across bedding planes than along bedding planes, it is reasonable to expect much steeper concentration gradients in the vertical than in the horizontal. In order to produce steep vertical concentration gradients in the profile model, small values of D_T are required, thus necessitating the large D_L/D_T ratio.

HISTORICAL DATA

Historical data, such as ground-water chemical quality, potentiometric head, and distribution, quantity, and chemical quality of ground-water recharge and discharge, are needed in both areal and profile water-quality models. Of the two models, the data needed for a profile model are more difficult to obtain, because the orientation of the model requires that these data be defined through the depth of the aquifer. Although most of these data may be obtained from a series of multiple-depth piezometers, most study areas would probably not have adequate wells available, necessitating what could be an extensive test-well drilling and sampling program.

In this study a series of 37 piezometers was monitored for head and water-quality changes over a period of about 3½ years (as of 1974). These and other data indicate that the ground-water flow system has been in near dynamic equilibrium since the present city sewage-treatment plant went into operation (subsequent to the flood of January-February 1969). The head and dissolved-solids configuration associated with percolation from the sewage-treatment ponds can thus be considered to have resulted from transient solute transport in a steady-flow system (fig. 7).

The quantity and distribution of underflow recharge (0.16 ft^3/s)

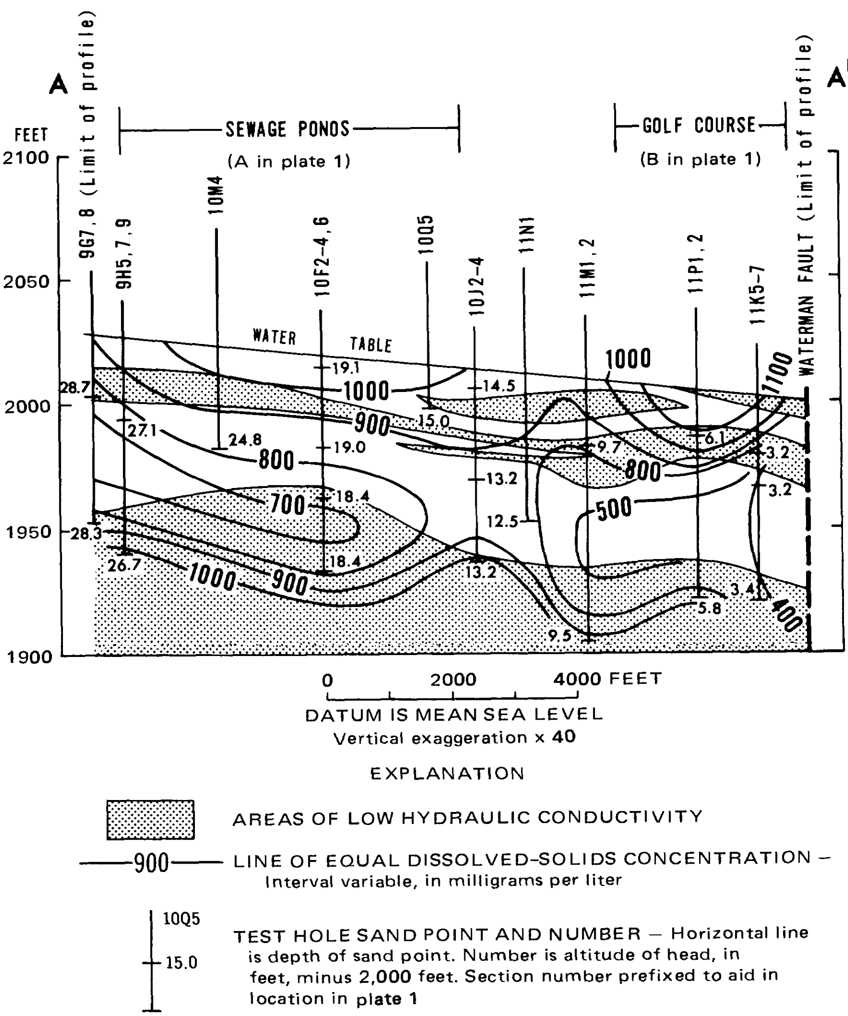


FIGURE 7.—Modeled profile showing observed data on dissolved-solids concentration and head, November 1973.

and discharge (0.19 ft³/s) were estimated by use of Darcy's law once the hydraulic conductivity and head gradients in the profile were determined. Recharge from the sewage ponds (0.72 ft³/s) and golf course (0.12 ft³/s) was estimated on the basis of the profile width and the recharge rates used in previous plan-oriented modeling. All recharge and discharge were modeled by use of constant-flow boundaries. The dissolved-solids concentration of the recharge from the golf course and sewage ponds was 1,600 and

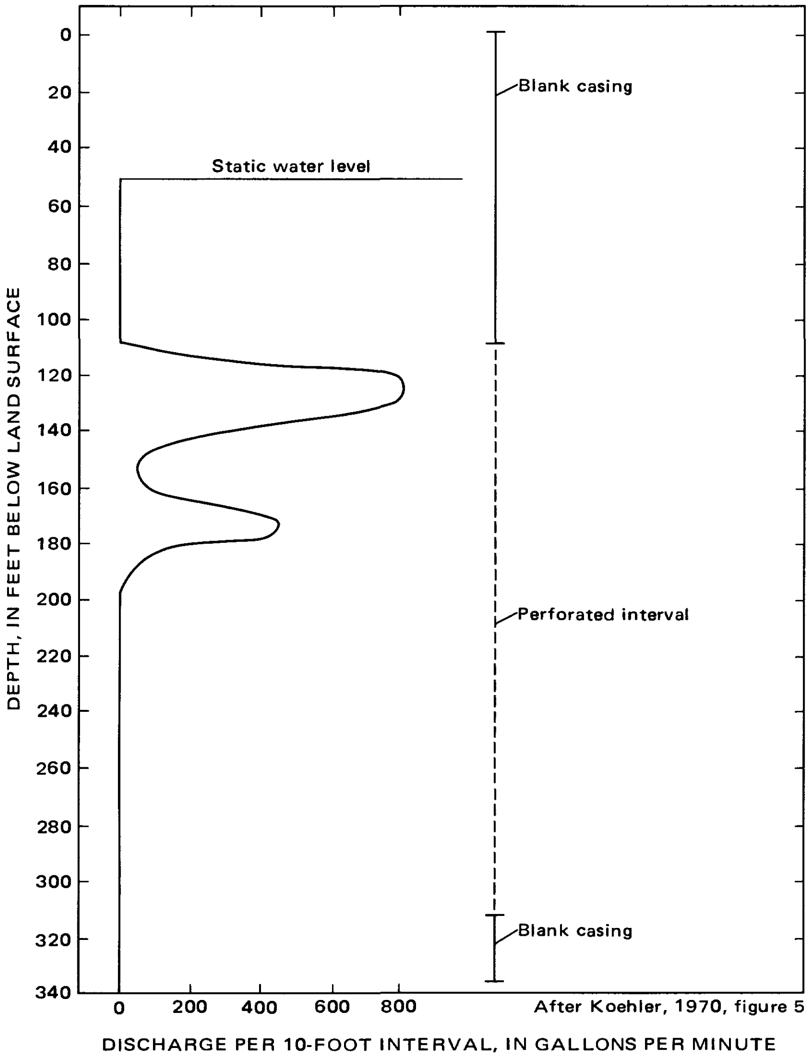


FIGURE 8.—Vertical distribution of discharge from U.S. Marine Corps supply well B3.

1,000 mg/l (milligrams per liter) respectively. The initial dissolved-solids concentration in the aquifer was assumed to be 500 mg/l.

Wells generally do not yield water uniformly throughout the perforated interval (fig. 8). As a result, the vertical distribution of flow from a recharging or discharging well should be estimated by use of a deep-well current meter or other techniques (Patten and

Bennett, 1962) if the well is to be considered in a profile-oriented model.

APPLICATION OF MODEL TO GROUND-WATER AND WASTE MANAGEMENT PROBLEMS

Before applying the profile model to management problems a calibration procedure is used to check the ability of the model to simulate field conditions. The calibration involved a comparison of 1973 potentiometric head and dissolved-solids concentrations in the aquifer with the 1973 values calculated by the model. Potentiometric head differences between the observed data (fig. 7) and model calculations were generally less than 1.0 ft. Calculated dissolved-solids concentrations were within about 100 mg/l of the observed concentrations shown in figure 7. The zone of degraded quality near the base of the profile was not considered in the calibration because of the lack of data on its duration and source concentration. The agreement between the observed and calculated values is adequate to enable the model to be used for illustrative purposes. If the model were intended for actual evaluation of management alternatives, a more stringent calibration might be required.

The vertical orientation of the profile model allows the simulation of changes in the vertical and longitudinal distribution of head and water quality in an aquifer. The extent to which this ability can be of use in evaluating the effects of various ground-water management practices is shown by examples on the following pages.

Ground-water management practices that can be simulated in the profile model may involve changes in the distribution of recharge or discharge in the aquifer, changes in the quantity of water recharging or discharging from the aquifer, or changes in the chemical concentration of the water recharging the aquifer. Several hypothetical ground-water management practices were simulated with the profile model in order to illustrate (fig. 9*A*–*F*) the applicability of the model to each of these categories. In each case the ground-water degradation near the base of the alluvial aquifer and near the golf course (fig. 7) was not considered in the model in order to simplify the cause-and-effect relations to be shown.

In the first management practice to be considered it was assumed that no remedial measures would be taken to limit the spread of degraded ground water near the city of Barstow sewage-treatment facilities. In particular, it was assumed that the historical quantities, distribution, and chemical quality of ground-water recharge and discharge would be continued during the next 5 years and that the ground-water system would continue to operate under steady-flow conditions. The distribution of dissolved-solids concentrations in 1979 shown in figure 9*A* was generated by the model as a result of this "do-nothing" management alternative.

If a series of fully penetrating pumping wells were installed along a line perpendicular to the axis of the profile model (fig. 9*B*), a nonradial flow field could be maintained in the model plane near the wells. Pumping from these wells could form a partial barrier to the movement of ground water that could be used to retard the expansion of the zone of degraded ground water below the city of Barstow sewage-treatment facilities. The pumped water would be transported out of the area of influence of the profile model. The model-generated distribution of 1979 dissolved-solids concentrations produced by pumping the barrier well in the profile at a rate of 0.15 ft^3/s (three-fourths of the total underflow in the profile) is shown in figure 9*B*. This management practice assumes that no other changes occur in the hydrologic system. The only significant difference in concentration between the barrier-well condition (fig. 9*B*) and the previous "do-nothing" condition (fig. 9*A*) would occur downgradient from the barrier well where the dissolved-solids concentration in the aquifer is reduced by as much as 100 mg/l after 5 years. Water-level declines in excess of 2.0 ft but not exceeding 3.0 ft occur in the shaded area shown in figure 9*B*.

Another management alternative investigated by use of the profile model involves pumping from barrier wells with a small perforated interval rather than the full saturated thickness previously considered. In this case it was assumed that the barrier wells were perforated only in the upper 10 ft of saturated thickness (fig. 9*C*) and that no other changes would occur in the hydrologic system. The model-generated distribution of dissolved-solids concentrations in 1979 produced by pumping a shallow perforated barrier well in the profile at a rate of 0.15 ft^3/s is shown in figure 9*C*.

This management practice would have less effect on longitudinal movement in the deeper parts of the degraded zone than would the fully penetrating barrier well. The larger vertical components of ground-water flow produced by this management practice would tend to retard downward movement of the zone of degraded water, however. As a result, the degraded zone of ground water would have a lesser vertical extent than would occur with the fully penetrating barrier wells. Water-level declines in excess of 2.0 ft but not exceeding 5.0 ft would occur in the shaded area shown in figure 9*C*.)

A fourth approach to managing the chemical quality of the basin

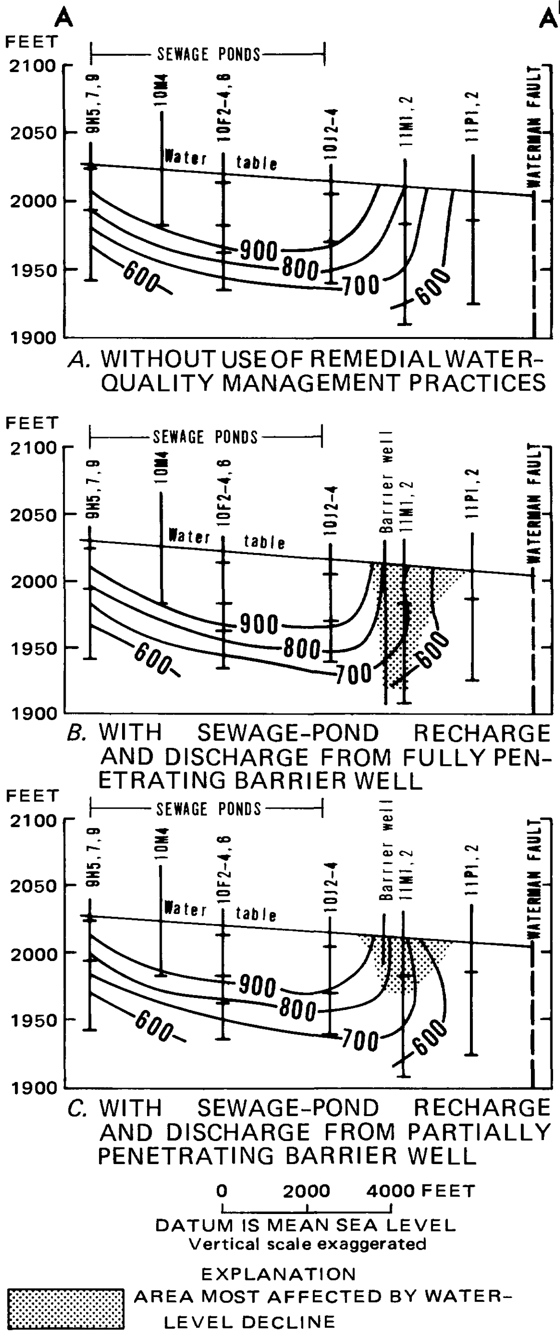
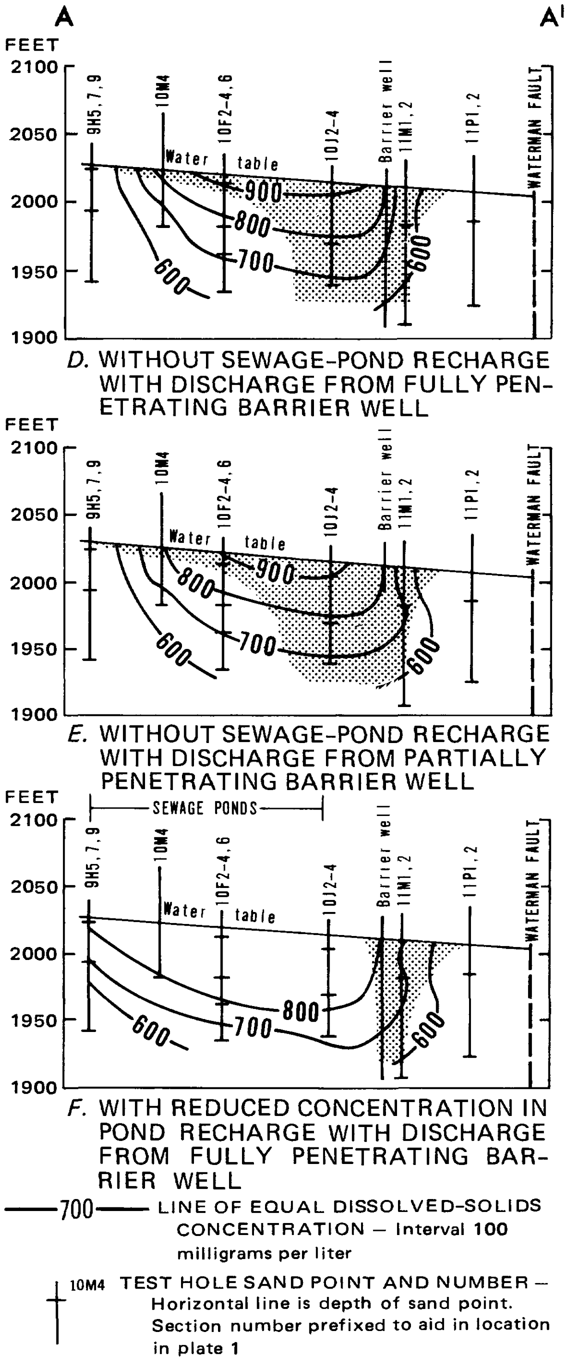


FIGURE 9.—Profiles showing model-generated vertical distribution



of dissolved solids and area of water-level decline, 1979.

that can be simulated in the profile model involves sealing the city of Barstow sewage-treatment ponds to prevent percolation of water of degraded quality. For modeling purposes, it was assumed that the treated effluent from the ponds would be disposed of in areas out of immediate hydraulic continuity with the modeled aquifer.

The distribution of dissolved-solids concentrations in 1979 shown in figure 9*D* would occur when percolation recharge from the sewage-treatment facility was eliminated when a fully penetrating barrier well in the profile was pumped at the previous rate. Similar conditions were considered in figure 9*E* except that the barrier well in the profile was considered to be perforated only in the upper 10 ft of saturated thickness. Both illustrations indicate that the loss of recharge from the sewage-treatment ponds would be a much more significant factor with regard to the ground-water quality of the area than the method of pumping water from the barrier wells. Water-level declines in excess of 4.0 ft but not exceeding 7.0 ft would occur in the shaded areas in figures 9*D* and 9*E*. The large water-level declines would result from the loss of recharge caused by sealing the sewage-treatment ponds.

The effects of changing the chemical concentration of recharge to the aquifer can be illustrated by assuming that the dissolved-solids concentration of the recharge from the city of Barstow sewage-treatment ponds could be reduced from 1,000 to 850 mg/l. A fully penetrating barrier well was simulated in the profile as in previous examples. The resulting distribution of dissolved-solids concentrations in 1979 (fig. 9*F*) indicates that this management practice would tend to reduce the concentration in the aquifer adjacent to the treatment facilities but would have little effect on the concentrations at greater distances from the recharge source. Modest water-level declines from 2.0 to 3.0 ft would occur near the barrier well.

Simulation of these management practices illustrates how the profile model could be used in evaluating the effects of various ground-water management practices. Many similarities exist between the simulation capabilities of the areal and profile water-quality models because of their origin as a common computer program. The orientation of the profile model places certain practical limitations on the capabilities of the model but in other respects permits greater flexibility by allowing simulation of head and water-quality changes with depth in the aquifer. A brief appraisal of the advantages and disadvantages of the profile and areal models sould be helpful in exploring the applicability of the two models to field problems.

ADVANTAGES AND DISADVANTAGES OF AREAL AND PROFILE MODELS

An areal water-quality model can be used to simulate areal changes in head and water quality in confined or unconfined aquifers with steady or nonsteady uniform or radial flow systems, but it is not suited to simulation of multiple aquifer systems and cannot simulate the vertical distribution of head and water quality in an aquifer. A profile water-quality model can readily simulate the vertical distribution of head and water quality in a single aquifer or in a multiple-aquifer system. The model can simulate these aquifers as confined or unconfined aquifers under conditions of steady or nonsteady flow but is not suited to conditions involving radial ground-water flow. The profile model is further limited in that the longitudinal axis of the model must follow a ground-water flow path. Thus, if two or more points in a study area are to be considered in a profile model they must be located on the same ground-water flow path. It is further required that the flow path not be subject to significant changes in location in the aquifer during the time period to be modeled.

The type of ground-water management practices to be investigated and the form of the model results necessary to make evaluations of the alternative practices play obvious roles in the choice of a model. Management practices that are concerned primarily with areal changes of head or water quality, such as changes in the locations of well fields or areas of natural recharge or discharge, are best considered in an areal model. By contrast, management practices that are concerned primarily with changes in the vertical distribution of head and water quality are best dealt with by use of a profile model. Examples of management practices of this type include changes in the location of perforated intervals in pumping wells, changes from surface liquid-waste disposal to injection-well disposal, or evaluating the change in head and water quality in a shallow aquifer owing to pumping-induced leakage of water of differing quality from deeper or shallower aquifers.

The availability of basic data necessary to define the model parameters also plays an essential role in the choice of a model. Although both the areal and profile models require similar types of parameter data, the parameters for the profile model are generally more difficult to define. Areal model parameters such as areal distributions of hydraulic conductivity, saturated thickness, storage coefficient, recharge and discharge, and water-level and water-quality data may often be estimated using the existing distribution of wells or other land-surface data in the model area.

In the profile model essentially the same parameters must be defined vertically through the saturated thickness to be modeled. The normal distribution of wells and perforated intervals would seldom be adequate to define the profile model parameters, thus necessitating the installation of additional wells or piezometers. The added time and expense that may be required to estimate the parameters for a profile model limit the applicability of the model to problems in which the changes in vertical distribution of head and water quality are of importance and cannot be adequately simulated by other techniques.

The possibilities of using the profile model at some later data to evaluate the effects of other ground-water management practices are probably more limited than they would be with an areal model. This stems from the requirement that the profile model be alined with, and only consider conditions along, a particular ground-water flow path. The chances of a subsequent problem in the study area being located on the flow path considered in the profile model are considerably less than the chances of the problem being located in the area considered by an areal model. One means of partially alleviating this problem would be to use the profile model in conjunction with an areal model. The combination of these models provides a better simulation of the real-life three-dimensional flow system than does either model separately.

The calibration techniques needed to assure the accuracy of either an areal or a profile model are virtually the same. Head and water-quality contour maps are well suited to making comparisons between model-generated data and observed data when a continuous trend of change in head and water quality is being simulated. Hydrographs of head and water-quality data are usually more suitable if the model is simulating a period of fluctuating head and water-quality conditions in the aquifer.

The computer costs for running an areal model are slightly less than for a profile model because it requires smaller core storage and fewer computations resulting from the more nearly square grid spacing of the model. No significant differences exist between an areal and profile model study in the type of computer required or in the needed expertise of the hydrologist.

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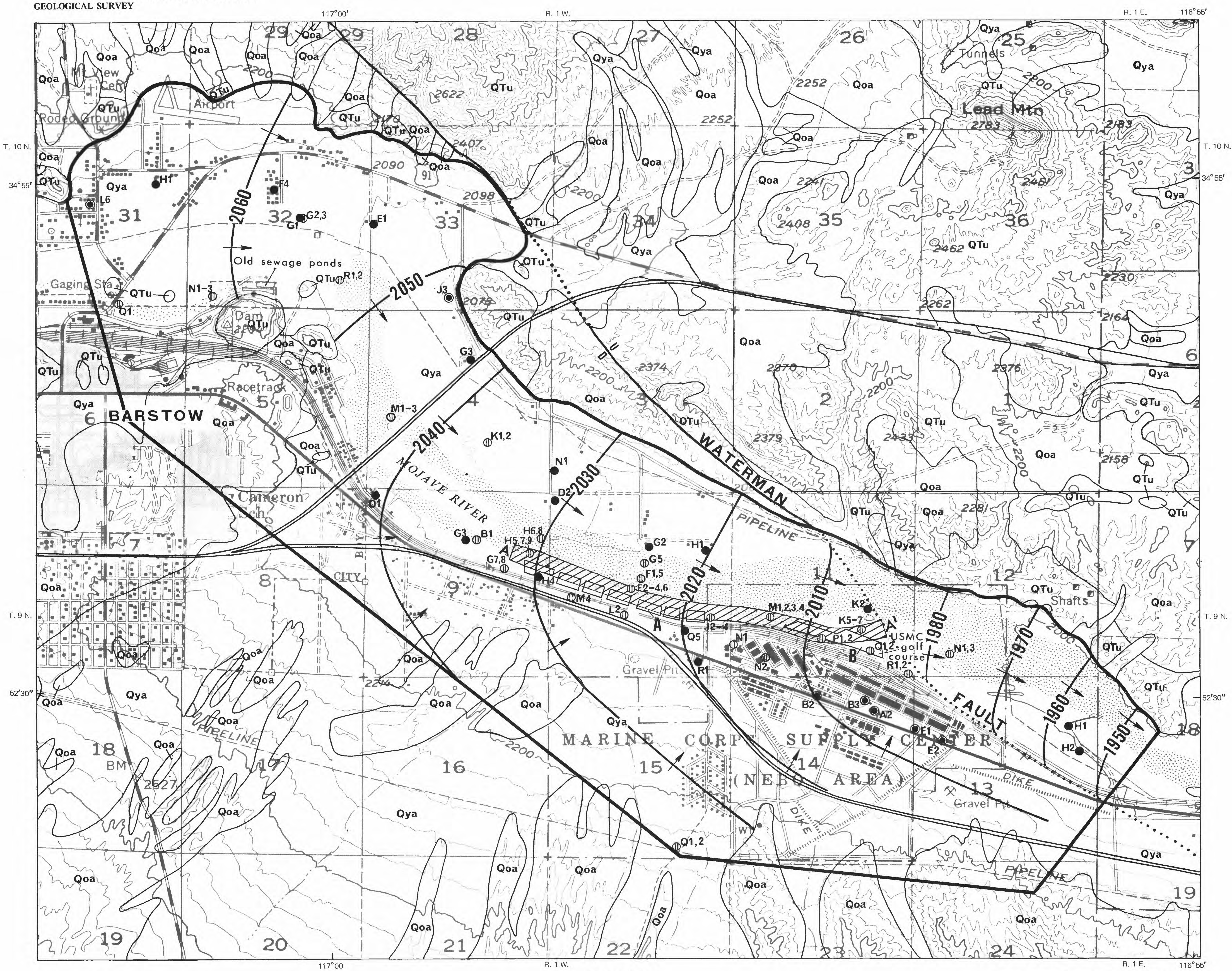


Base from U. S. Geological Survey Barstow, 1956, and Daggett, 1956 1:62,500. Highways as of 1972

TRUE NORTH
MAGNETIC NORTH
15°
APPROXIMATE MEAN DECLINATION, 1978

SCALE 1:24,000

1 ½ 0 1 MILE

1 .5 0 1 KILOMETER

CONTOUR INTERVAL 40 FEET
DATUM IS MEAN SEA LEVEL

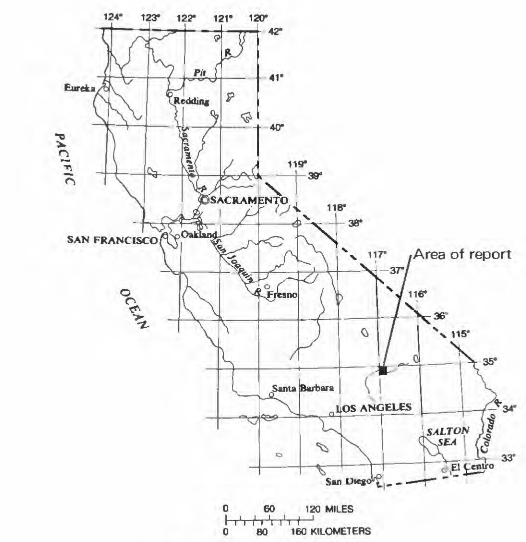
☆Interior – Geological Survey, Reston, Va., – 1978 – W77150
Geology modified after G. A. Miller (1969)

EXPLANATION

Qya	Younger alluvial deposits of Holocene	QUATERNARY
Qoa	Older alluvial deposits of Pleistocene age	QUATERNARY
QTu	Consolidated rocks, undifferentiated	QUATERNARY AND TERTIARY
	GEOLOGIC CONTACT	
U / D	FAULT – Dotted where concealed. U, upthrown side; D, downthrown side	
1980	WATER-LEVEL CONTOUR, 1972 – Shows altitude of water level. Contour interval 10 feet and 30 feet. Datum is mean sea level. Arrow shows direction of ground-water movement	
A A′	LOCATION OF PROFILE MODEL – Showing 500-foot width simulated; see text	
	BOUNDARY OF PLAN-ORIENTED WATER-QUALITY MODEL (Robson, 1974)	
A	CITY OF BARSTOW SEWAGE PONDS	
B	UNITED STATES MARINE CORPS SUPPLY CENTER GOLF COURSE	
B1	TEST HOLE AND NUMBER	
G3	IRRIGATION OR DOMESTIC WELL AND NUMBER	
B2	SUPPLY WELL AND NUMBER	

D	C	B	A
E	F	G	H
M	L	K	J
N	P	Q	R

Wells are assigned numbers according to their location in the rectangular system for the subdivision of public land. The letter designates the 40-acre tract of a section. Within each tract the wells are numbered serially as indicated by the last number.



INDEX MAP OF CALIFORNIA

GEOLOGY, WATER-LEVEL CONTOURS FOR SPRING 1972, AND LOCATION OF PROFILE MODEL, TEST HOLES, AND WELLS