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PRELIMINARY PROJECTIONS OF THE EFFECTS OF CHLORIDE-CONTROL STRUCTURES ON THE QUATERNARY AQUIFER AT GREAT SALT PLAINS, OKLAHOMA

U.S. GEOLOGICAL SURVEY
WATER-RESOURCES INVESTIGATIONS 80-120

Prepared in cooperation with the
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About 1,200 tons of chloride per day are added to the salt load of the Salt Fork Arkansas River at Great Salt Plains Lake from brine discharge from the rocks of Permian age in the vicinity of the lake. The U.S. Army Corps of Engineers has planned a chloride-control project. The Corps requested that the U.S. Geological Survey use a digital model to project the effects of the chloride-control structures on ground water. Ground-water flow and ground-water transport models were calibrated to represent the Quaternary aquifer (the near-surface part of the flow system). Although ground-water levels are projected to rise as much as 19 feet in areas near three reservoirs, changes in ground-water levels caused by the project will be small throughout most of the area. Chloride concentration of ground water is projected to increase by more than 90,000 milligrams per liter at one location. However, significant increases in chloride concentration during the 50-year period simulated are projected to be limited to areas where the ground water already contains excessive chloride concentrations.

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CONTENTS

ILLUSTRATIONS

TABLES

CONTENTS

CONVERSION FACTORS

The inch-pound units used in this report may be converted to metric units by the following conversion factors. Concentration of solute is expressed only in milligrams per liter (mg/L), a metric unit. The second (s) is common to both systems. Degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) by subtracting 32 and then dividing by 1.8.

Inch-pound units	Multiply by	To obtain metric units
foot (ft)	0.3048	meter
foot per second (ft/s)	0.3048	meter per second
foot per day (ft/d)	0.3048	meter per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
ton per day	0.9072	megagram, or metric ton, per day

PRELIMINARY PROJECTIONS OF THE EFFECTS OF CHLORIDE-CONTROL STRUCTURES ON THE QUATERNARY AQUIFER AT GREAT SALT PLAINS, OKLAHOMA

By J. E. Reed

ABSTRACT

About 1,200 tons of chloride per day are added to the salt load of the Salt Fork of the Arkansas River at Great Salt Plains Lake from natural sources. The source of this chloride is brine discharge from the rocks of Permian age in the vicinity of the lake. The U.S. Army Corps of Engineers has planned a chloride-control project. The Corps requested that the U.S. Geological Survey use a digital model to project the effects of the chloride-control plan on ground water. Ground-water flow and ground-water transport models were calibrated to represent the Quaternary aquifer that is the near-surface part of the flow system. The models were used to project the effects of planned chloride-control structures. Based on model results, ground-water levels are projected to rise as much as 19 feet. However, these water-level rises will occur only in areas near three reservoirs. Changes in ground-water level caused by the project will be small throughout most of the area. Chloride concentration of ground water is projected to increase by more than 90,000 milligrams per liter at one location. However, significant increases in chloride concentration during the 50-year period simulated are projected to be limited to areas where the ground water already contains excessive chloride concentrations.

INTRODUCTION

Great Salt Plains Lake is in Alfalfa County in north-central Oklahoma (fig. 1). The Great Salt Plains Lake, a flood-control reservoir, is formed by a dam, completed during 1941, on the Salt Fork of the Arkansas River. The Great Salt Plains adjacent to the lake consist of areas of barren ground termed "salt flats," which are seasonally encrusted with precipitated salt. Salt flats, or salt plains, are level to gently sloping areas that are part of stream flood plains. Hummocky areas of windblown sand anchored by vegetation may occur at elevations slightly above that of the surrounding salt flat. These mounds usually occur near the margins of the salt flats. A salt crust that is usually a fraction of an inch in thickness forms during dry weather at the surface of the salt flats.

One of the first discussions of the Great Salt Plains in hydrologic literature was by Gould (1905, p. 101). Theis (1934) made a geologic and hydrologic reconnaissance of the Salt Plains reservoir site. A study of the ground-water resources in the Cherokee area was made by Schoff (1950). Ward (1961) studied the geology and ground water in the vicinity of Great Salt Plains and in other brine discharge areas in Oklahoma, Kansas, and Texas.

The hydrologic problem is natural salt pollution in the Salt Fork Arkansas River. Chloride loads for three stations: one just downstream from Great Salt Plains Lake dam, one on the Salt Fork about 20 river miles upstream from the dam, and one on Medicine Lodge River about 40 river miles upstream from the dam (near Kiowa, Kans.), are listed in table 1. The locations of the first two sites are shown on figure 1. The average chloride load downstream from the reservoir for 1970-77 was about 1,300 tons per day. The average chloride load for 1974-76 increased by more than 1,400 tons per day between the stations near Ingersoll and Kiowa and the station near Jet. The average chloride load for 1974-76 was nearly 200 tons per day more than the load for 1970-77 at the station near Jet. The increase in the average chloride load at the station near Jet during 1970-77 was probably more than 1,200 tons per day.

In order to reduce the amount of salt being added to the Salt Fork, a chloride-control project was begun by the Corps of Engineers. The Corps' plan for controlling the salt discharge is the development of an evaporating brine-detention reservoir. (See page 37, this report.) Hydraulic-head changes resulting from construction of the proposed control works would have some effect on the ground water in the vicinity of the lake and Great Salt Plains. Accordingly, the Corps requested that the U.S. Geological Survey use a digital model to make projections of these effects.

Table 1.--*Chloride loads, in tons per day, and discharge, in cubic feet per second, for three stations in the Great Salt Plains area*

Station	Drainage area (square miles)		Water year 1970	1971	1972	1973	1974	1975	1976	1977
Salt Fork Arkansas River near Jet, Okla.	3,202	Chloride	1,040	784	1,230	1,960	1,790	1,890	792	839
		Discharge	235	81.4	132	887	673	588	183	153
Salt Fork Arkansas River near Ingersoll, Okla.	1,140	Chloride	-----	-----	-----	-----	71.4	71.9	32.9	27.9
		Discharge	-----	-----	-----	-----	180	157	79.1	65.1
Medicine Lodge River near Kiowa, Kans.	903	Chloride	-----	-----	-----	-----	42.7	26.9	22.6	-----
		Discharge	110	85.3	80.7	263	237	136	108	84.4

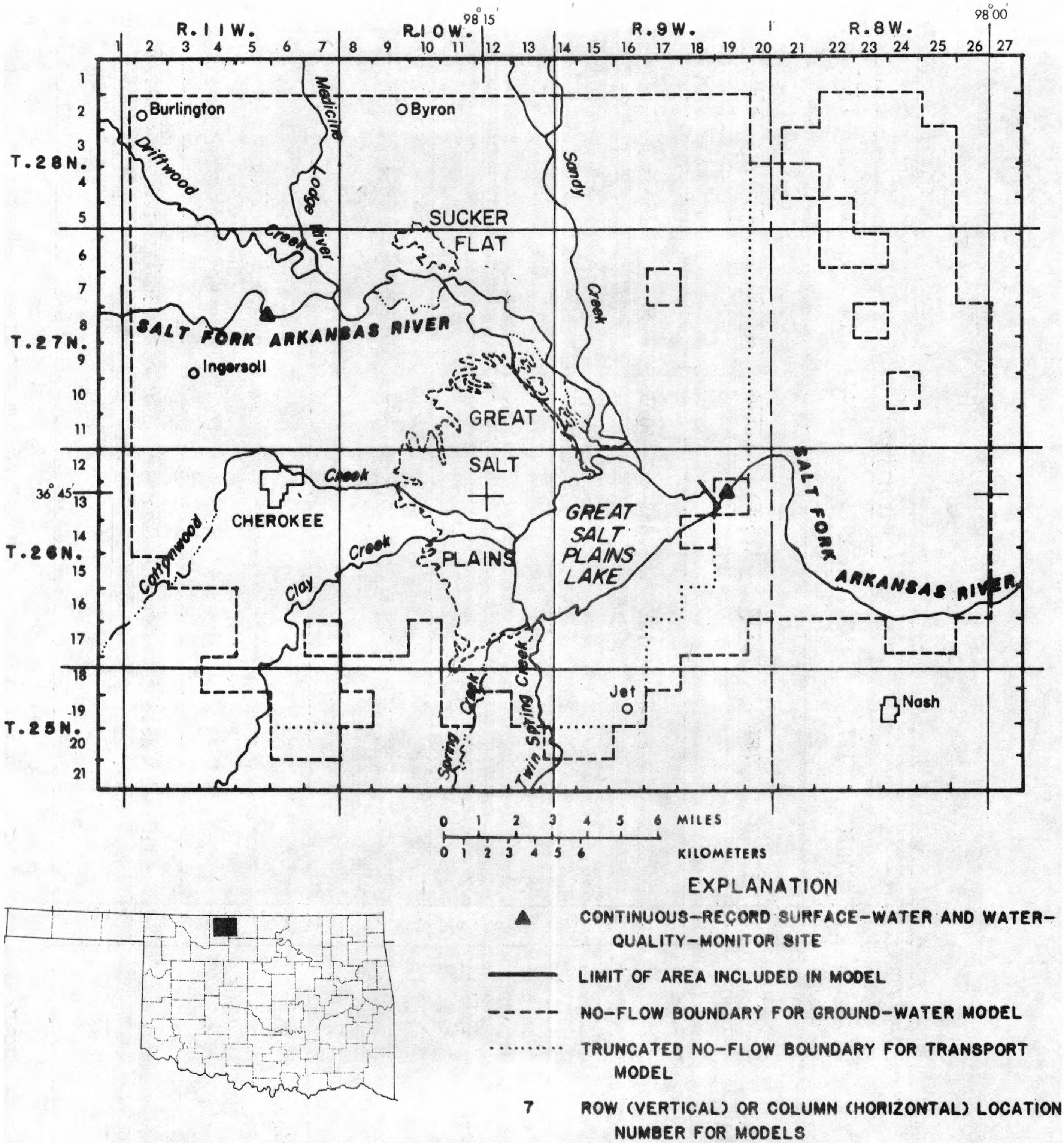


Figure 1.--Location of study area, model boundaries, and data-collection sites.

The purpose of this report is to describe the model used to define the ground-water flow system adjacent to and underlying the Great Salt Plains Lake, and to project the effects on ground water by stresses imposed by the structures designed to improve the quality of the outflow from Great Salt Plains Lake.

The study area includes two principal areas of salt flats; the Great Salt Plains, an area originally of about 40 mi^2, but now partly covered by the Great Salt Plains Lake; and Sucker Flat, a smaller area of about 1 mi^2 north of the Great Salt Plains.

Data on which the model was based were derived from reports, maps, and listings furnished by the Tulsa District, U.S. Army Corps of Engineers. Information on water use was taken from reports of the Oklahoma Water Resources Board (1968, 1969) and Oklahoma State University (Schwab, 1969; Duffin, 1967). Assistance and advice on modeling solute transport was given by Leonard F. Konikow, U.S. Geological Survey, Reston, Va. The SIP computer program was coded by Richard Healy, U.S. Geological Survey.

GEOHYDROLOGY

Sedimentary rocks of Quaternary and Permian age occur in the area; where the deposits of Quaternary age are at the land surface they are underlain by deposits of Permian age. Nomenclature for the Permian units as used by the U.S. Army Corps of Engineers is shown in table 2 and on figure 2. The surficial distribution of Quaternary and Permian units used in the model was taken from a geologic map compiled by Morton (1980). The Quaternary deposits have a maximum thickness of 131 ft, as indicated by test holes, but in most of the area the thickness is only about 30 ft. The Quaternary deposits are composed predominately of sand, with considerable amounts of silt, and some clay, and minor amounts of gravel.

The Permian deposits consist of shale, siltstone and some fine grained sandstone, usually red in color but with occasional gray beds. Beds of halite, and shale containing disseminated salt, are present in the subsurface.

Table 2.--*Subdivisions of the Permian*
[Modified from U.S. Army Corps of Engineers, 1978b]

System	Series	Formation	Member	Thickness (feet)	Lithology
Permian	Lower Permian	Cedar Hills Formation		170-180	Coarse-grained siltstone and very fine grained sandstone; light-gray siltstone and sandstone at base.
		Salt Plains Formation		250 ±	Fine to coarse-grained siltstone and very fine-grained sandstone; seams of red-brown blocky shale at top.
		Equivalent of Harper Formation of Kansas	Equivalent of Kingman Member	70-90	Light-gray siltstone at top; 3 to 5 feet of light-gray siltstone at base.
			Equivalent of Chikaskia Member	120-170	Shale with interbedded siltstone.
		Equivalent of Ninnescah Formation of Kansas	Lower "Cimarron" salt (economic term)	0-96	Equivalents of Stone Corral Dolomite and Runnymeade Sandstone at top of Ninnescah are absent locally; top of salt is used locally as top of Ninnescah Formation.
				500 ±	Shale
		Wellington Formation	Hutchinson Salt Member	450 ±	Shale, salt, and anhydrite.

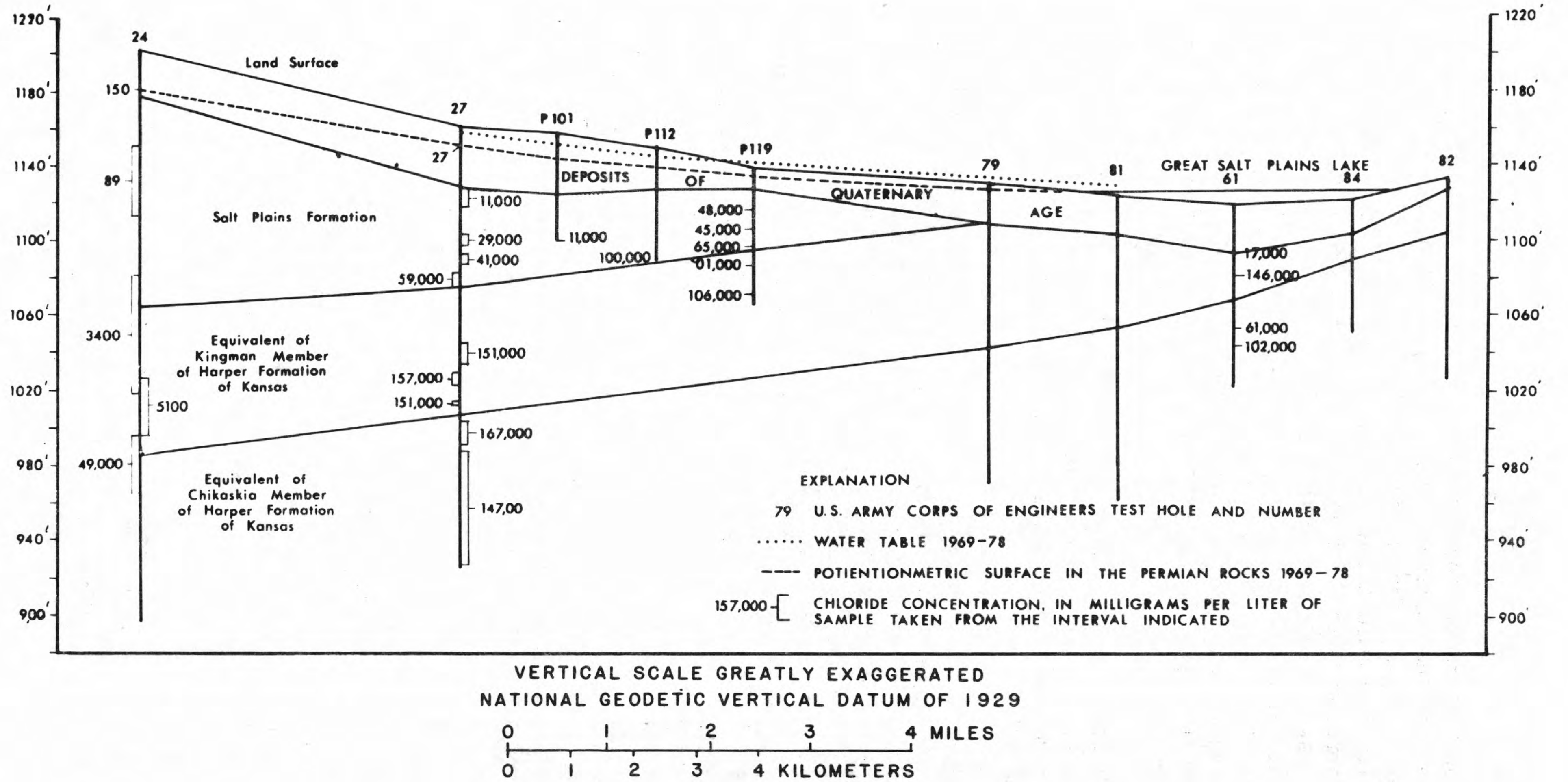
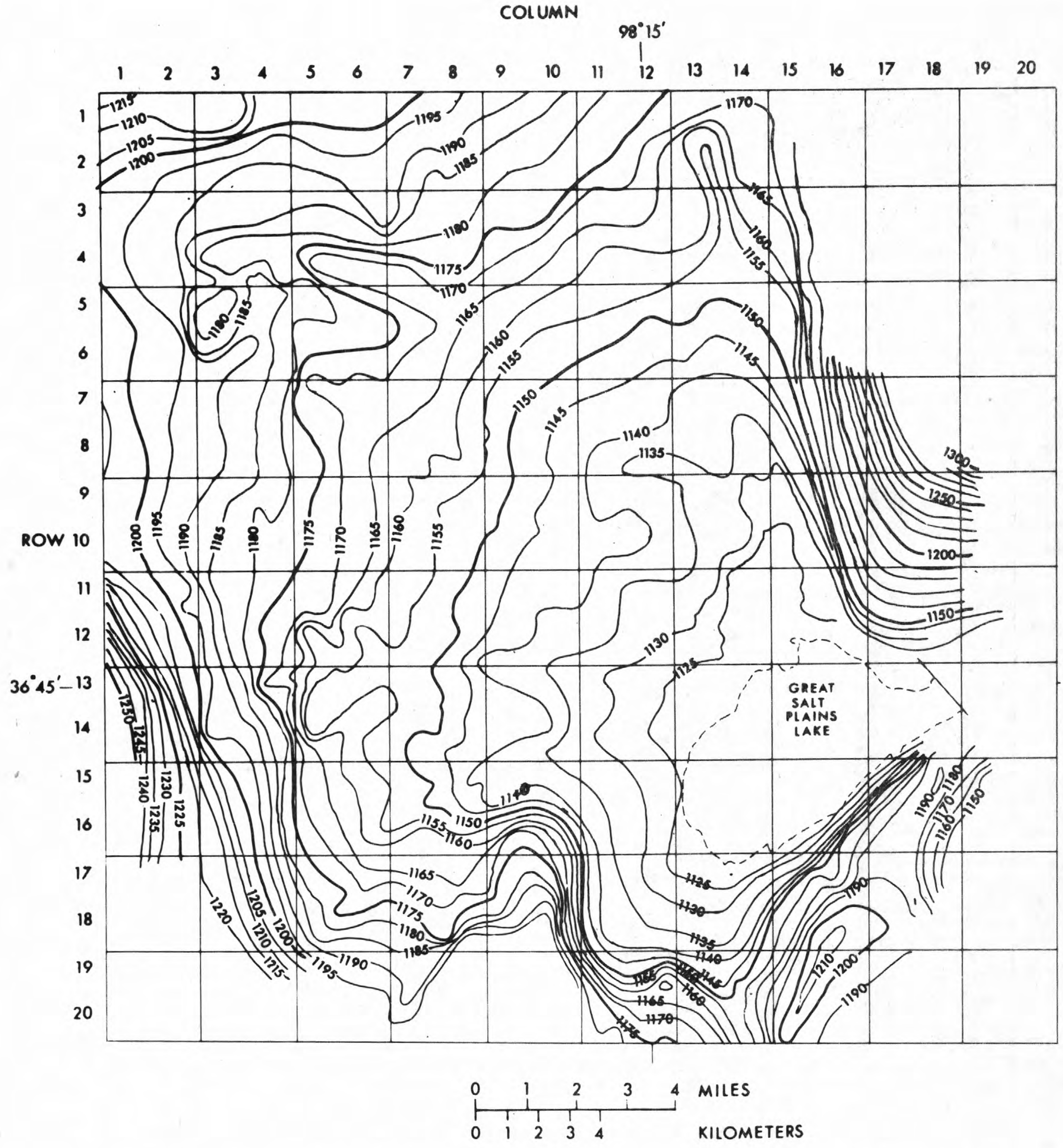


Figure 2.-- Geohydrologic section through Great Salt Plains. (From U.S. Army, Corps of Engineers, 1978b.)

The Quaternary deposits form a mantle over an erosional surface developed on the Permian bedrock. The direction of movement of water in the Quaternary aquifer, as indicated by the water-table map (fig. 3), generally is toward the Great Salt Plains Lake. A ground-water divide that corresponds to a topographic ridge is located in the area southeast of the lake. Fader and Morton (1975), indicate that another ground-water divide probably exists northeast of the lake, beyond the area shown in figure 3. The steep hydraulic gradients northeast and southeast of the lake occur where the altitude of the base of the aquifer rises steeply (fig. 4). The water-table surface generally follows the topographic contours.

Recharge to the Quaternary aquifer probably is greater northeast and southeast of the lake. Surficial materials in these areas consist largely of sand; large areas of sand dunes are found northeast of the lake.

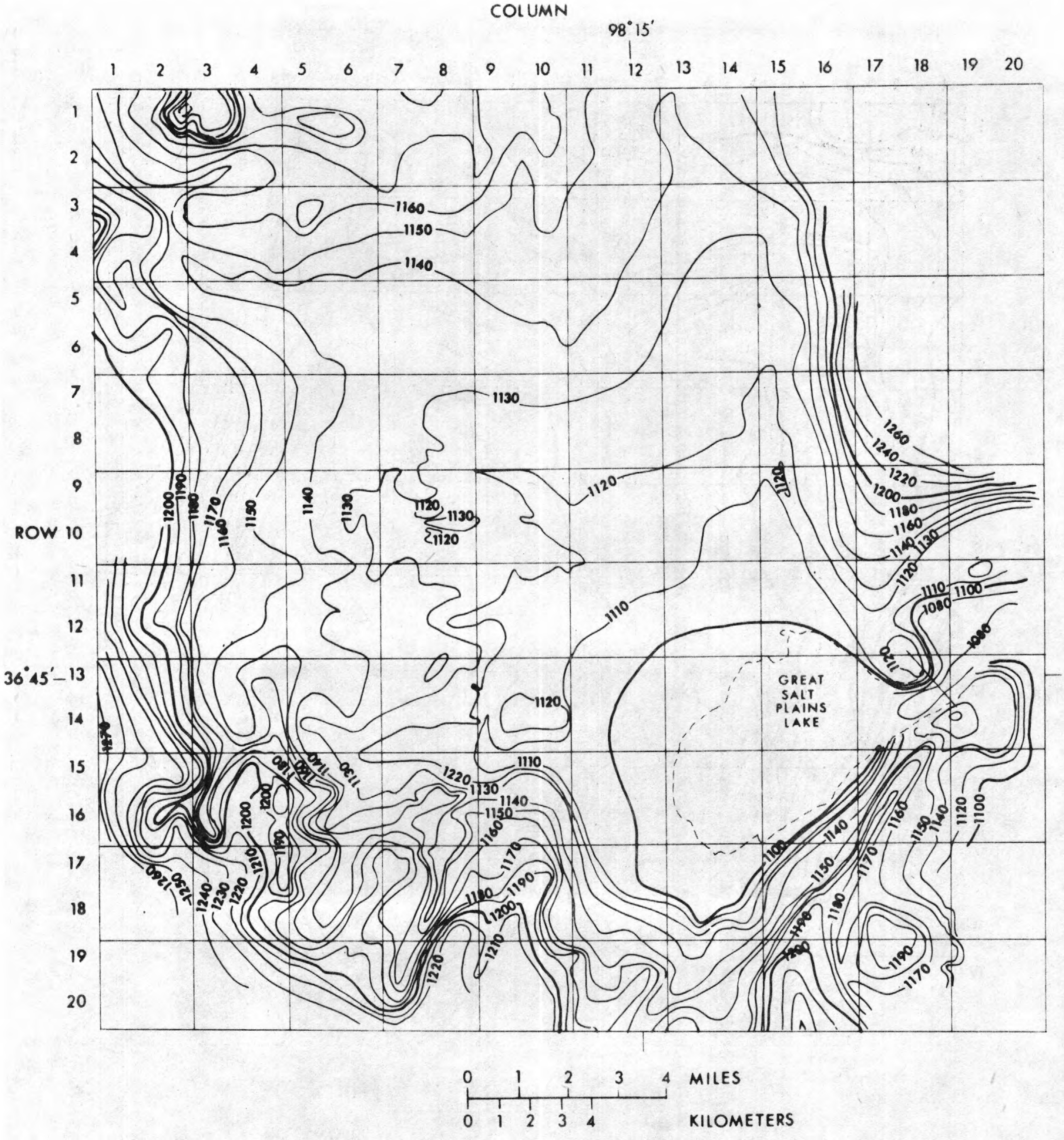
The chemical quality of the water in the Quaternary deposits is extremely variable in this area. The water level is within a few feet of the surface in the barren salt flats, and the water is brine (water having a dissolved-solids concentration greater than 35,000 mg/L). The quality of the water improves away from the salt flats, and the water is used for households, stock watering, irrigation, and public supply. Chloride concentration of ground-water samples collected at or near the water table is shown in figure 5, and chloride concentration of ground-water samples collected at the top of Permian bedrock is shown in figure 6. The main area of salt flats, the Great Salt Plains, corresponds approximately with the large area enclosed by the 10,000 mg/L line in figure 5. Sucker Flat, the smaller area of salt flats north of the Salt Fork, also is enclosed by a 10,000 mg/L line. Although a salt crust forms on the surface of the salt flats during dry weather, the salt crusts are a result of the brine contamination in the Quaternary aquifer and not the cause of it.



—1170— WATER-TABLE CONTOUR — Shows altitude of water table, 1968-9. Contour interval 5, 10, or 20 feet. National Geodetic Vertical Datum of 1929

Marginal numbers refer to model location

Figure 3.--Configuration of the water table. (From U.S. Army, Corps of Engineers, 1976.)



—1100— BEDROCK CONTOUR--Shows altitude of bedrock surface. Contour interval 10 feet. National Geodetic Vertical Datum of 1929

Marginal numbers refer to model location

Figure 4.--Configuration of the top of Permian bedrock.(From U.S. Army, Corps of Engineers, 1976.)

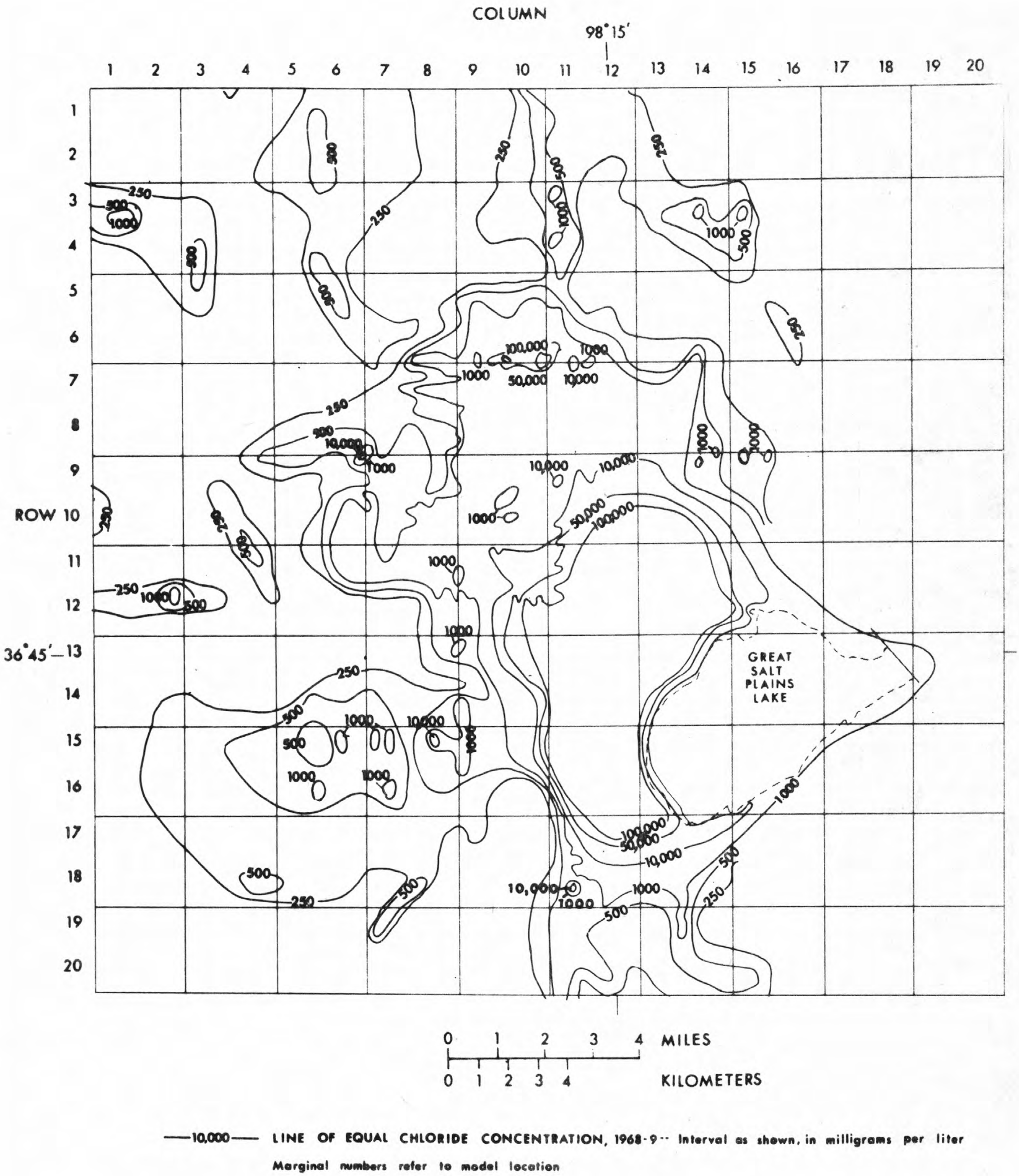
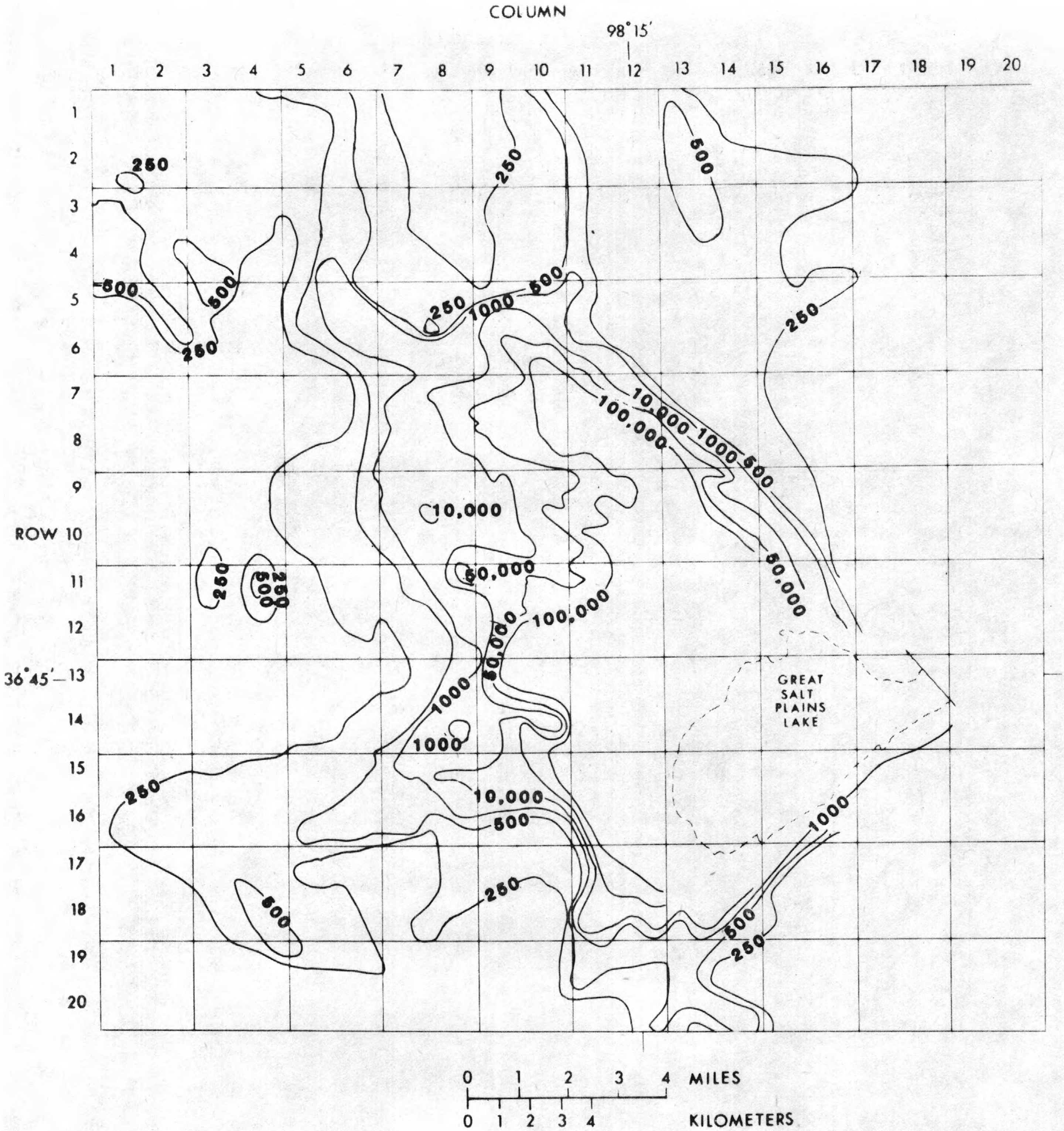


Figure 5.--Chloride concentration of samples collected at the water table. (From U.S. Army, Corps of Engineers, 1976.)



—10,000— LINE OF EQUAL CHLORIDE CONCENTRATION, 1968-9 -- Interval as shown, in milligrams per liter

Marginal numbers refer to model location

Figure 6.--Chloride concentration of samples collected at the top of Permian bed rock. (From U.S. Army, Corps of Engineers, 1976.)

SOURCE OF THE BRINE

The source of the excessive chloride concentration in the Quaternary aquifer is brine that is flowing upward from the underlying Permian rocks. The potentiometric surface of the Permian deposits is higher than the water table in the Quaternary deposits near the Great Salt Plains (fig. 2).

The brine discharge at Great Salt Plains is one of several areas in Kansas, western Oklahoma, West Texas, and southeastern New Mexico where brine is discharging from rocks of Permian age. Although the brines are probably related to salt deposits in the Permian deposits in these areas, differing hypotheses have been used to explain their occurrence. Frye and Schoff (1942) considered deep circulation and lateral movement of many tens of miles to be a plausible explanation for the salt-discharge areas in northwestern Oklahoma. Ward and Leonard (1961) viewed shallow circulation and short lateral travel distances to be a more probable explanation. Stevens and Hardt (1965) considered the brine discharge areas in the upper Brazos River of Texas to be controlled by local geologic features and related to a regional body of brine.

Two salt-bearing units are present in the subsurface of the study area. The upper unit is the lower "Cimarron" salt (economic term) (fig. 7), which is present only in part of the area and which lies 250 ft or more below the land surface. The lower unit is the Hutchinson Salt Member of the Wellington Formation, which includes less than 450 ft of salt, anhydrite, and shale, and lies at depths of 1,100 to 1,300 ft (U.S. Army Corps of Engineers, 1978a, p. ii-10). Although either or both of the salt-bearing units are possible sources of brine, the mechanism by which the brine might move into the Great Salt Plains is not known.

The Great Salt Plains is a shallow, semicircular basin. This indicated to Ward and Leonard (1961, p. D15) that the Great Salt Plains "...was formed, at least in part, by collapse resulting from the removal of underlying salt deposits."

Sinkholes and surface collapse features commonly are associated with local solution of salt beds in some of the other salt plains in Oklahoma. Such surface features seem to be lacking in the Great Salt Plains area, but because of the widespread mantle of unconsolidated Quaternary deposits they would be difficult to detect. Test holes, such as those shown on figures 2 and 7, indicate that the dip of Permian beds seems to be gentle and regular. However, local slumping of the Permian beds may have occurred as a result of salt solution.

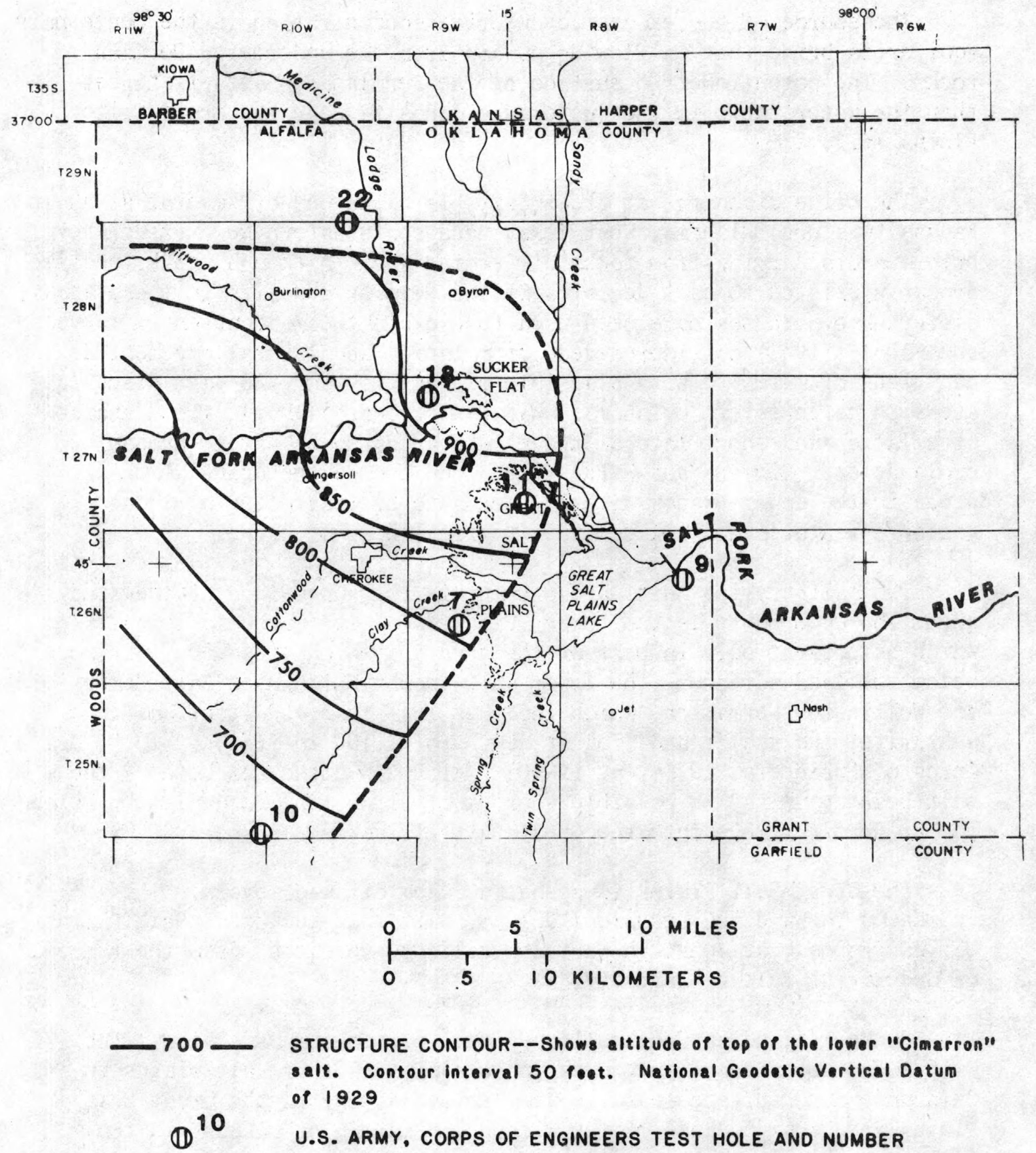


Figure 7.--Altitude of the top of the lower "Cimarron" salt. (From U.S. Army, Corps of Engineers, 1978a.)

MODELING OF THE FLOW SYSTEM

Two digital models are the basis for the construction of the model used to simulate, in a simplified manner, the complex ground-water hydrology of the alluvial aquifer in the area around and underlying the Great Salt Plains. Both of these models are well documented in existing reports, and little discussion of their approximating equations and solution algorithms is necessary in this report. One model is the two-dimensional ground-water flow model of Trescott, Pinder, and Larson (1976); the other is the two-dimensional ground-water solute-transport model of Konikow and Bredehoeft (1978).

In the discussion that follows, references to the "ground water model" will indicate a reference to the model of Trescott, Pinder, and Larson (1976), and references to the "transport model" will indicate the model of Konikow and Bredehoeft (1978).

The ground-water model encompasses the area shown in figure 1. This area is subdivided into a network of 21 by 27 square blocks, each having a width of 5,000 ft and extending the full thickness of the aquifer. The transport model covers a slightly smaller area as shown in figure 1, and consists of a network of 20 by 20 square blocks, each having a width of 5,000 ft and extending the full thickness of the aquifer.

Both the ground-water and transport models require that the outermost nodes have a transmissivity equal to zero. This creates a no-flow boundary located between the peripheral nodes and the interior of the model. The "useful" area of the model, therefore, is limited to an area within but slightly smaller than the modeled area.

The locations of nodes (centers of the square blocks) in the model are referred to by row number first and then by column number. For example, node (11, 13) refers to the node that is located at the intersection of row number 11 and column number 13. The starting point for numbering rows and columns is the same for both the ground-water and transport models; therefore, nodes representing the same geographic area have the same nodal location number for both the ground-water and transport models.

Ground-Water Flow Model

The partial differential equation for steady-state two-dimensional flow of a uniform incompressible fluid through a nonhomogeneous isotropic aquifer may be written as:

$$\frac{\partial}{\partial x}\left(T \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T \frac{\partial h}{\partial y}\right) = Q\ (x,y) - \frac{K'}{b'}\ (H_s - h) \qquad (1)$$

where

T	is the transmissivity, L^2/T;
h	is the hydraulic head, L;
x and y	are the Cartesian coordinates, L;
Q	is the rate of withdrawal (positive sign) or recharge (negative sign), L/T;
K'	is the vertical hydraulic conductivity of the confining layer, streambed, or lakebed, L/T;
b'	is the thickness of the confining layer, streambed, or lakebed, L; and
H_s	is the hydraulic head in the source bed, stream, or lake, L.

For a derivation of the unsteady-flow equation, the interested reader is referred to Pinder and Bredehoeft (1968). Equation 1 is simplified from Pinder and Bredehoeft's equation by the conditions that the aquifer is isotropic and flow is steady.

The procedure used subdivides the region of interest into a network of rectangular blocks of constant size. Each grid block is referred by its location indices (i,j) in the network. Equation 1 is transferred into a linear algebraic equation by replacing derivatives with finite differences. Equation 1 then becomes:

$$
\begin{aligned}
& T_{i-\frac{1}{2},j} \; \frac{h_{i-1,j} - h_{i,j}}{(\Delta x)^2} \\
& \quad + T_{i+\frac{1}{2},j} \; \frac{h_{i+1,j} - h_{i,j}}{(\Delta x)^2} \\
& \quad + T_{i,j-\frac{1}{2}} \; \frac{h_{i,j-1} - h_{i,j}}{(\Delta y)^2} \\
& \quad + T_{i,j+\frac{1}{2}} \; \frac{h_{i,j+1} - h_{i,j}}{(\Delta y)^2} \\
& \quad = \frac{q_{w(i,j)}}{\Delta x \, \Delta y} - \frac{K'}{b'} \quad H_{s(i,j)} - h_{i,j} \qquad (2)
\end{aligned}
$$

where

i, j are indices in the x and y dimensions, respectively;

Δx, Δy are increments in the x and y dimensions, respectively; and

q_w is the volumetric rate of withdrawal or recharge at the (i,j) node, L^3/T.

The interblock transmissivities in the above equation are replaced by the harmonic mean of the transmissivities of the adjoining blocks:

$$T_{i-\frac{1}{2},j} = \frac{2 \, T_{i-1,j} \, T_{i,j}}{T_{i-1,j} + T_{i,j}}$$

and similarly for:

$T_{i+\frac{1}{2},j}$, $T_{i,j-\frac{1}{2}}$, and $T_{i,j+\frac{1}{2}}$.

For modeling water-table aquifers, transmissivity is computed as the product of hydraulic conductivity and the saturated thickness for the previous iteration using the equation:

$$T^{n}_{i,j} = K_{i,j}\, b^{n-1}_{i,j}$$

where

K is the hydraulic conductivity, L/T;

b is the saturated thickness, L;

and the superscript indicates the iteration number.

This finite-difference approximation results in one linear algebraic equation, for each node, containing five unknown hydraulic head terms. For the entire network with M rows and N columns of active blocks, a system of MN equations in MN unknown hydraulic heads results. An efficient method of solving this system, for the values of the matrix coefficients that commonly are encountered in ground-water flow problems, is the strongly implicit procedure (SIP). Application of SIP to ground-water models is discussed in some detail in Trescott, Pinder, and Larson (1976).

Modeled Parameters

The Quaternary aquifer in the area was modeled as a water-table aquifer; that is, the transmissivity of the aquifer is the product of the average hydraulic conductivity at a given location and the saturated thickness of the aquifer at that location. The hydraulic conductivity undoubtedly changes from place to place because the type of material changes. However, aquifer-test results did not correlate well with lithologic logs at test sites. A uniform value for hydraulic conductivity equal to 100 ft/d was used in the model. Allison and others (1974, p. 6) report an average value for hydraulic conductivity of 25 ft/d for 12 sites with a range of 3 to 100 ft/d; Fader and Morton (1975, p. 33) estimate a range of 130 to 260 ft/d for the alluvium and 65 to 200 ft/d for the terrace deposits. A range of 0.6 to 4.5 ft/d for 10 permeability (hydraulic-conductivity) tests of near-surface sands and an aquifer-test value of 100 ft/d as the average for 25 feet of sand were reported by the U.S. Army Corps of Engineers (1974, p. 21).

Leakance, K'/b', is input to the model as two different leakage parameters, thickness of the confining bed, b', and hydraulic conductivity of the confining bed, K'. The model then computes leakance from these two input parameters. Leakance in the model represents vertical flow between overlying surface-water bodies and the aquifer. Flow between surface-water bodies such as the Salt Fork River, Medicine Lodge River, Great Salt Plains Lake, and the aquifer were modeled by coding values for these leakage parameters at nodes where the surface-water bodies occur. The hydraulic conductivity K', represents the vertical hydraulic conductivity, K_z, of the aquifer material, and b' represents the vertical flow path. The value of b' used in the model was 12.5 ft, which is approximately one-half of the typical thickness of the aquifer. The value of K' used (before adjustment for area of streambed) was 10 ft/d. This value of K' represents an anisotropic ratio, K_z/K_h, of 0.1, which would not be unusual for horizontally layered materials. The leakage parameters were applied throughout the entire area of a node even though a streambed may cover only a small part of the node. Input values of K' were reduced by the ratio of area of the streambed within a node to area of the node. This ratio ranged from 0.01 for nodes on the Medicine Lodge River and part of the Salt Fork River to 1.0 where nodes were totally overlain by Great Salt Plains Lake. The values of leakance ranged from 9.28×10^{-8} to $9.28 \times 10^{-6} s^{-1}$. These values are large enough that stream nodes are effectively constant hydraulic-head boundaries. The largest hydraulic-head difference between stream level and aquifer water level in the calibrated model is only 0.7 ft; it is usually 0.2 ft or less.

Modeled Stresses

Modeled stresses were recharge from infiltration of precipitation, discharge by evapotranspiration, pumpage, brine seepage from underlying formations, and leakage to and from streams and lakes. Two different values of recharge from infiltration of precipitation were used in the model. The larger value of 1.2×10^{-8} ft/s (4.5 in/yr) was applied to model nodes representing the sand dunes and other upland area east of the lake. The remainder of the model area had a recharge rate of 2×10^{-9} ft/s (0.8 in/yr). The average recharge rate was 1.3 in/yr.

Discharge of ground water by evapotranspiration is a function of depth to water in the model. The function is controlled by two input variables: QET, which is the maximum rate of evapotranspiration; and ETDIST, which is the depth to water level below land surface at which evapotranspiration ceases. If the water level is at or above land surface, the rate of discharge is at the maximum rate, QET. The rate of discharge is a linear function of depth to water, for depths to water below land surface and less than ETDIST. Input values used in the model were 2.6×10^{-8} ft/s (10 in/yr) for QET and 6.5 ft, for ETDIST.

Information on ground-water pumpage for irrigation was obtained from Duffin (1967, p. 2) and Schwab (1969, p. 2), and for irrigation and other uses from the Oklahoma Water Resources Board (1968, p. 2, 1969, p. 2). Withdrawals were modeled at the locations of the towns of Cherokee (1.2 ft^3/s), Burlington (0.1 ft^3/s), and Jet (0.1 ft^3/s). Withdrawals also were modeled at nodes where 22 water wells and a fish hatchery were indicated on 7½-minute topographic maps published during 1968 and 1969. The total ground-water pumpage modeled was 5.5 ft^3/s.

Brine seepage from underlying Permian rocks into the Quaternary aquifer was modeled as a specified flux into the model at nodes within the area of excessive chloride concentration in the Quaternary aquifer (figs. 5 and 6). The total seepage was modeled as 5.13 ft^3/s applied as a nonuniform rate ranging from 0.005 to 1.2 ft^3/s at 62 nodes.

Leakage occurs at those nodes for which non-zero values for leakage parameters K' and b' are specified. These nodes represent locations where the aquifer is overlain by surface-water bodies. Because the values for the vertical coefficients at leakage nodes are large relative to the horizontal coefficients, the nodes representing surface-water bodies are effectively constant hydraulic-head boundaries. The level specified in the model for Great Salt Plains Lake is 1,125 ft, the altitude of the conservation pool. Levels for the Medicine Lodge River and the Salt Fork Arkansas River upstream from Great Salt Plains Lake were taken from low-water profiles surveyed by the Corps of Engineers. Levels for the Salt Fork Arkansas River downstream from Great Salt Plains Dam were estimated from 7½-minute topographic maps.

Boundary Conditions

The Quaternary aquifer in the report area extends many miles upstream along the Salt Fork Arkansas and Medicine Lodge Rivers and downstream along the Salt Fork Arkansas River. Along the southwest, south, and northeast parts of the modeled area, the edge of the aquifer is approximately as modeled. On the north, northwest, and east sides of the modeled area, the boundary as modeled represents a truncation of a continuous aquifer. Along this truncated boundary, underflow was simulated by flux into the model on the north and northwest boundaries and flux out of the model on the eastern boundary, as indicated by the direction of slope of the water table at the respective model boundaries. Underflow varied from node to node along the model boundary, and is a function of the slope of the water table and transmissivity. An average value of underflow equal to 0.05 ft^3/s was used at nodes on a truncated boundary, either into or out of the model. Underflow was modeled as inward flux at 29 nodes on the northern and western edges of the modeled area and outward flux at 9 nodes on the eastern edge for a net inward flux of 1.00 ft^3/s. This average value of underflow is discussed further under the section on sensitivity tests.

Changes in water level computed by a model may be examined by comparing the effects of different model boundaries that can produce extreme values of hydraulic head at the boundaries. Specifying (1) a constant hydraulic-head boundary with no change in water level and (2) a no-flow or no-change-in-flow boundary would produce the extreme values for changes in water level for arbitrary boundary location. Boundary effects for the ground-water model was checked by examining changes in water level at the arbitrary boundaries. Changes in water level were less than 0.05 ft at the arbitrary boundaries, and no significant difference in water-level change would be produced by enlarging the model.

Sensitivity Tests

Sensitivity tests on the model were made by changing the value of a single input parameter, while maintaining the other parameters at their specified values, and observing the effects of this single parameter change on the computed hydraulic head for the model. When applied to model calibration, sensitivity tests indicate the relative response of the model to change in different input parameters. The most "sensitive" parameter is the one whose relative changes produce the most change in computed hydraulic head. The least sensitive parameter is one whose changes produce the least change in computed hydraulic head.

During model calibration, sensitivity tests may indicate that changes in a single parameter may produce a computed water-table map that is in general agreement with the historic water-table map, or calibration standard (fig. 3). However, changes in a parameter should be physically plausible, because physically absurd values may produce a good fit. Errors may be present in the historic data, or the model may not include all the physical processes that significantly affect the real system. Calibration of the model is a subjective process in which the modeler must decide the degree of correspondence between model results and the calibration standard. Factors involved in calibration are the modeler's concept of the physical system, quality of the historic data, and limitations of the model.

Results from the sensitivity tests on the calibrated model are listed in table 3. The results are listed in groups by parameter tested with the relative parameter values sorted by highest computed water table. Three measures of goodness of fit were applied to the difference between computed and historic water table and are shown in this table. These measures are the sum of the absolute values, mean values, and standard deviation.

The optimum values for these measures would be for all three to equal zero. Although this optimum can be approximately determined for the mean, the sum of the absolute values and the standard deviation do not approach zero, and the relative minimum values are the best values that can be determined by single parameter changes. However, the data in table 3 indicate that the best values for the three measures of goodness of fit between computed and historic values of hydraulic head usually do not indicate the same parameter value. Therefore, choosing a "best" parameter value becomes a subjective process.

The model was calibrated by choosing parameter values within a range indicated as reasonable by hydrologic information for the model area. A set of parameter values was determined by a subjective visual comparison between the computed and the historic water table. This set of parameter values was then tested by sensitivity analysis to determine if changes in parameter values would improve the match between the computed and the historic water table. Sensitivity tests indicated that the original set of parameter values were about as good as, or better than, any of the sets of changed values. The original set of parameter values was retained for further analysis of the system.

The mean value of computed minus historic water table generally showed the most response, and the standard deviation showed the least response to parameter changes. Most parameters were tested by making model analyses at values equal to one-half and twice the calibrated value. For the more sensitive parameters, this amount of variation prevented convergence of the algorithm to a solution. Therefore, the sensitive parameters were tested at a lesser variation, relative to the calibrated value.

The most sensitive parameters, as shown by changes in mean value of computed minus historic water table, are the hydraulic conductivity of the aquifer materials, recharge from infiltration of precipitation, and limiting depth for evapotranspiration from ground water (ETDIST). The least sensitive parameter is leakance at stream or lake nodes. Within the range of values in leakance tested in the model, the stream and lake nodes are effectively constant hydraulic-head boundaries.

Table 3.--*Results of sensitivity tests of the ground-water model*

X = (computed hydraulic head from the model) - (hydraulic head from water-table map)
N = (number of comparisons between computed hydraulic head and water table) = 281

Parameters	Calibrated values	Values (* - indicates calibrated value multiplied by number shown)	Sum of absolute values of *X* $\Sigma\lvert X\rvert$	Mean value of *X* $\bar{X}$	Standard deviation of *X* σ_x
QET (maximum rate of evapo-transpiration from ground water), in feet per second	2.6×10^{-8}	1.3×10^{-8}	1578.6	0.879	8.971
		2.6×10^{-8}	1363.9	-.308	8.340
		5.2×10^{-8}	1324.6	-.832	8.197
		1.0×10^{-7}	1318.1	-1.105	8.136
ETDIST (maximum depth below land surface for evapotrans-piration to extract ground water), in feet	6.5	5.5	1403.6	0.026	8.432
		6	1381.6	-.141	8.384
		6.5	1363.9	-.308	8.340
		9	1318.9	-1.150	8.144
		12	1367.6	-2.229	7.993
Underflow at truncated boundaries, in cubic feet per second	±0.05 per boundary node, at 38 nodes, sum = 1.00	*2	1439.7	0.227	8.471
		*1	1363.9	-.308	8.340
		*.5	1356.4	-.593	8.314
		*.25	1363.6	-.740	8.312
K (hydraulic conductivity of aquifer), in feet per second	1.16×10^{-3}	*0.75	1371.5	0.613	7.956
		*1	1363.9	-.308	8.340
		*1.25	1399.6	-.959	8.345
K'/b' (leakance at stream or lake nodes), in feet per second per foot	Ranges from 9.28×10^{-8} to 9.28×10^{-6}	*2	1364.1	-0.296	8.339
		*1	1363.9	-.308	8.340
		*.5	1363.3	-.332	8.345
Recharge from infiltration of precipitation, in feet per second	2×10^{-9} or 1.2×10^{-8} depending on location	*2	1705.0	3.500	8.482
		*1	1363.9	-.308	8.340
		*.75	1395.3	-1.517	8.765
Rate per node of brine seepage into the Quater-nary aquifer, in cubic feet per second	Ranges from 0.005 to 1.20, at 62 nodes, sum = 5.13	*2	1408.4	-0.055	8.482
		*1	1363.9	-.308	8.340
		*.5	1361.8	-.408	8.334
		*.25	1363.9	-.445	8.336
		*.1	1366.4	-.471	8.339

Transport Model

The transport model is described by Konikow and Bredehoeft (1978). Some minor modifications have been made to this model to represent the specific hydrologic conditions existing in the study area. The modeled process involves two-dimensional areal transport and dispersion for one dissolved chemical species in ground water. The chemical species is modeled as conservative (no loss of solute in the system through a physical or chemical process, such as radioactive decay) and nonreactive (no additional solution or precipitation of the species). Konikow and Bredehoeft (1978, equation 4) describe such a system in x_1, x_2, and t coordinates follows:

$$\frac{\partial (Cb)}{\partial t} = \frac{\partial}{x_i}\left(bD_{ij}\frac{\partial C}{\partial x_j}\right) - \frac{\partial}{\partial x_i}(bCV_i) - \frac{C'W}{\varepsilon}, \quad i,j = 1,2 \qquad (3)$$

where

C is the concentration of the dissolved chemical species, M/L^3;

D_{ij} is the coefficient of hydrodynamic dispersion (a second-order tensor), L^2/T;

b is the saturated thickness of the aquifer, L;

V_i is the seepage velocity in the direction of x_i, L/T;

C' is the concentration of the dissolved chemical in a source or sink fluid, M/L^3;

W is the volume flux per unit area for a source or sink (positive sign for outflow and negative for inflow), L/T; and

ε is the effective porosity of the aquifer, dimensionless.

The dispersion tensor, D_{ij} in equation 3, consists of four components for two-dimensional flow. For an isotropic aquifer, Konikow and Grove (1977, p. 26) showed that the dispersion tensor could be expressed in terms of the following: α_L, the longitudinal (in the direction of flow) dispersivity; α_T, transverse (normal to the flow direction) dispersivity; and the velocities, V_i and V_j, in the coordinate directions. Therefore:

$$D_{ii} = \alpha_L \frac{V_i^2}{|V|} + \alpha_T \frac{V_j^2}{|V|} \quad \text{and} \qquad (4)$$

$$D_{ij} = (\alpha_L - \alpha_T) \frac{V_i V_j}{|V|} \qquad i \neq j \tag{5}$$

where the magnitude of the velocity vector $|V| = (V_i^2 + V_j^2)^{1/2}$ and $i,j = 1,2$.

The ground-water interstitial (seepage) velocities are determined through a finite-difference solution to the ground-water flow equation. These velocities and a particle-tracking procedure (method of characteristics) are used to calculate the solute-transport equation. An explicit procedure is used to calculate the effects of dispersion and divergence of velocity that is due only to fluid sources and sinks for the Salt Plains model. A full description of the solution procedure is discussed in Konikow and Bredehoeft (1978, p. 5). The computed ground-water velocities in the model do not include the effects of variable fluid density on the ground-water flow system.

Modifications to the computer program were: (1) Replacement of the ADI (alternating direction implicit) numerical procedure with the SIP (strongly implicit procedure) algorithm for solving the ground-water-flow equations, (2) simulation of evapotranspiration (ET) from ground water as a linear function of depth to water, and (3) separate arrays containing source concentrations for recharge, leakage from streams, and brine seepage from the underlying Permian deposits. The SIP algorithm and the modeling of ET as a linear function of depth to water are discussed in Trescott, Pinder, and Larson (1976, p. 21 and p. 7).

The steady-state distribution of solute concentration in the aquifer was simulated with a single time step of 50 years. This time step was subdivided automatically into particle-tracing steps by the computer program according to input values for longitudinal dispersivity, maximum rate of all the fluid sinks and maximum distance of travel for a traceable particle. Each particle has a concentration and position associated with it and is moved through the flow field in proportion to the flow velocity at its location. The method of simulation may be visualized as tracing a number of fluid particles through a flow field and observing changes in chemical concentration in the fluid particles as they move. The maximum distance of travel for a particle was set equal to 2,500 ft or one-half of the node spacing; the simulation period then was divided into 85 particle-tracing steps. The adequacy of length of the simulation period was judged by comparing inflow mass-flux of solute (which is specified in the input data) to outflow mass-flux (which is determined by the solute distribution computed by the model). For calibration conditions, mass outflow in the model was 99.9 percent of model mass inflow; for projection conditions, mass outflow was 97.6 percent of mass inflow. An additional check was made by extending the steady-state simulation for an additional 50-year period. The change in chloride concentration during this extended period was small, averaging only 43 mg/L. This small change between 50 and 100 years also indicates that the model is approximately at steady-state conditions after the 50-year period.

Because the model represents steady-state conditions for ground-water flow, effective porosity (ε) does not affect the calibration of the model. Computed results for model runs with different effective porosities (0.15 and 0.30) and simulation times (50 and 100 years) were approximately the same.

The transport model represents a subarea within the area encompassed by the ground-water model (figure 8). Boundary conditions for the transport model are the same as for the ground-water model on the north and west sides, where the two model boundaries coincide. Part of the east boundary of the transport model north of the Salt Fork Arkansas River is a constant flux boundary, with the amount of flux indicated by flow between corresponding nodes of the ground-water model. The southeast boundary of the transport model is the ground-water divide shown on the water-table map, (fig. 3).

All of the input data used in the transport model to compute ground-water velocities, such as transmissivity, thickness, and initial hydraulic head, were the same as the computed steady-state results from the ground-water model. This reduced the number of iterations required for the transport model.

Input data for aquifer thickness are shown in table 4. These values represent saturated, and not total, thickness for the Quaternary aquifer. Input data for transmissivity were derived from the product of hydraulic conductivity (1.157×10^{-3} ft/s, or 100 ft/d) and the thickness shown in table 4. The thickness values were the computed saturated thicknesses from the calibrated ground-water model.

Input data for leakance are the same as used in the ground-water model and were discussed in the section on "Parameters Modeled." The leakance values reflect the part of each node that is covered by a surface-water body. Leakance values range from $9.26 \times 10^{-8} s^{-1}$ for the Medicine Lodge and Salt Fork Arkansas Rivers to $9.26 \times 10^{-6} s^{-1}$ for some of the lake nodes. The average concentration of chloride assigned to each stream node is different than that for each lake node. The average chloride concentrations of the Salt Fork Arkansas and Medicine Lodge Rivers are represented as 250 mg/L, and the average chloride concentration of Great Salt Plains Lake is represented as 3,000 mg/L.

Table 4.--*Aquifer thickness, in feet, for the transport model*

Row	Column 1	2	3	4	5	6	7	8	9	10
1	0	0	0	0	0	0	0	0	0	0
2	0	28.1	32.8	30.5	20.6	12.8	19.9	19.9	20.7	17.0
3	0	48.3	26.9	24.5	16.0	23.8	24.9	20.0	20.6	20.5
4	0	44.6	50.4	37.1	33.1	35.5	37.9	30.4	23.9	21.2
5	0	42.2	49.3	44.3	42.1	37.1	33.5	25.3	21.4	16.4
6	0	7.0	15.9	29.1	36.1	35.3	34.8	30.5	21.8	14.5
7	0	10.7	14.6	23.4	28.3	32.7	31.1	31.0	25.0	18.9
8	0	10.0	15.0	26.8	35.0	31.0	31.2	31.8	29.5	23.5
9	0	9.7	15.8	30.0	31.8	29.9	29.6	31.3	31.3	25.8
10	0	5.4	18.2	27.7	29.6	27.7	32.5	31.1	21.8	20.3
11	0	3.1	20.2	21.9	24.2	26.4	24.1	20.1	22.6	20.4
12	0	2.0	25.6	24.1	21.3	18.2	18.5	18.6	19.8	20.5
13	0	2.7	30.7	24.2	18.1	13.9	20.5	18.2	22.7	20.8
14	0	.7	12.2	19.7	27.5	32.2	34.8	30.8	23.6	14.0
15	0	0	14.8	1.5	19.6	40.0	32.6	31.8	33.8	24.6
16	0	0	0	0	23.8	26.5	8.8	21.1	1.1	.5
17	0	0	0	0	7.7	16.6	0	13.6	2.7	0
18	0	0	0	4.4	28.7	35.3	25.4	0	0	0
19	0	0	0	0	0	4.2	34.8	8.1	0	0
20	0	0	0	0	0	0	0	0	0	0

Row	Column 11	12	13	14	15	16	17	18	19	20
1	0	0	0	0	0	0	0	0	0	0
2	23.3	25.2	29.1	29.4	15.4	13.7	4.8	4.9	38.8	0
3	21.5	25.1	26.7	25.9	33.9	10.3	8.5	4.1	4.1	0
4	18.9	23.3	28.3	29.6	34.9	3.3	11.5	5.5	6.5	0
5	14.9	19.9	27.5	30.2	41.5	23.6	16.3	16.4	7.3	0
6	9.3	17.6	21.0	28.8	36.4	10.6	21.2	33.2	14.1	0
7	13.0	12.0	11.7	20.3	28.5	.4	0	16.9	10.1	0
8	15.1	13.1	13.9	15.9	22.5	3.9	.4	10.8	3.3	0
9	19.6	17.6	16.8	18.0	12.6	7.6	.6	9.8	5.9	0
10	18.4	18.8	20.7	21.8	18.0	9.9	1.9	49.4	52.2	0
11	20.0	20.3	19.0	19.0	19.0	11.0	15.5	45.0	54.0	0
12	21.5	25.6	26.3	24.9	25.0	24.0	20.0	48.1	53.0	0
13	21.9	26.5	25.4	25.0	25.0	25.0	19.0	14.0	0	0
14	22.4	25.5	25.0	25.0	25.0	25.0	25.0	0	0	0
15	22.5	25.2	25.0	25.0	25.0	25.0	6.2	2.3	0	0
16	20.0	23.9	25.0	25.0	25.0	12.9	2.0	0	0	0
17	10.8	25.0	26.4	29.1	8.0	12.7	0	0	0	0
18	4.5	5.0	21.8	18.3	6.9	9.9	0	0	0	0
19	3.5	0	5.1	3.6	.3	22.4	0	0	0	0
20	0	0	0	0	0	0	0	0	0	0

The only indicies of dispersion were at the site of a small freshwater impoundment on the Great Salt Plains located at node (11, 12). At this site a rectangular impoundment, approximately 250 by 750 ft, stored freshwater carried by a pipeline from the Salt Fork Arkansas River from mid-April to mid-December 1972. Data pertaining to this impoundment are described in a U.S. Army Corps of Engineers report (1974). A nonuniform-density model (INTERCOMP Resource Development and Engineering, Inc., 1976), of a cross section 2,400 ft long by 28 ft deep, was used to reproduce vertical chloride distribution under the impoundment, as shown in a U.S. Army Corps of Engineers memorandum (1974). Values of α_L of 10 ft and α_T of 3 ft produced model results for chloride at the base of the aquifer that were too small. Values of α_L of 1 ft and α_T of 0.3 ft produced model results that were similar to the historic data.

The values for dispersivity in the INTERCOMP model are somewhat dependent on, and biased by, the small part of the aquifer included in this model. Dispersivities are affected by the nonhomogeneity of the aquifer. A uniform aquifer would be expected to have a smaller dispersivity than a nonuniform aquifer. The dispersivity used in the small INTERCOMP model reflects the more uniform transmissive character in the area of the local impoundment than in the regional area of the Great Salt Plains. Therefore, it is reasonable to use a dispersivity that is many times larger for the regional model.

Konikow (1977, p. 24) studied the chloride movement in a thin alluvial aquifer near Denver, Colo., in an area that is somewhat similar to the Quaternary aquifer of the Great Salt Plains; the most appropriate values for longitudinal and transverse dispersivities were 100 ft. Cross-section and areal models used by Robson (1978, p. 17) to simulate solute transport at Barstow, Calif., indicated suitable ratios of α_T to α_L to be 0.3 for areal models and 0.003 for cross-section models. Values of α_L of 100 ft and α_T/α_L of 0.3 were chosen for use in the model of the Salt Plains. Suitability of the choice of α_L was tested by sensitivity analysis. The ratio of α_T/α_L was not tested. Three different values of longitudinal dispersivity and selected statistical measures of the corresponding model output of chloride concentrations are listed in table 5. Values of 10 and 100 ft for α_L were used to produce similar model output in each of the statistical measures. All values of α_L between 10 and 100 ft probably will reproduce approximately the historic chloride distribution. The use of 1,000 ft for α_L, however, produced model output that does not match the historic data very well; this value is not suitable to use in conjunction with the other parameter values.

Table 5.--*Results of sensitivity tests on longitudinal dispersivity*
[X = Chloride concentration computed by the model minus concentration, in milligrams per liter; N = 295, number of model nodes]

Longitudinal dispersivity α_L	Mean value of X $\bar{X}$	Standard deviation of X σ_x	Largest value of X	Smallest value of X	Average absolute deviation $\frac{\Sigma\lvert X\rvert}{N}$
1,000	890	6,211	73,514	-15,245	2,316
100	285	3,411	10,122	-12,480	1,721
10	160	3,506	10,502	-12,105	1,750

The direction of convective transport in the model is in the direction of the decrease in computed hydraulic head. Values of the altitude of the water table computed in the transport model are listed in table 6. Average interstitial velocities are computed as the product of hydraulic conductivity and hydraulic gradient divided by porosity. Because the porosity and hydraulic conductivity values are constant throughout the model, velocity in the Great Salt Plains' model is a function of hydraulic gradient only.

The rate of brine seepage from underlying Permian rocks into the Quaternary aquifer is not known. The total chloride load contributed by brine seepage is indicated by the records for the quality station on the Salt Fork Arkansas River near Jet. The rate of seepage can be estimated as a function of average chloride for the brine from the chloride load. The relation between seepage rate (cubic feet per second), load (tons per day), and chloride concentration (milligrams per liter) is:

$$\text{Brine seepage} = \frac{\text{Chloride load}}{0.002697 \text{ Chloride concentration of brine}}$$

If the chloride concentration of the brine seepage is not uniformly distributed in space, the relation above requires use of a discharge-weighted average chloride. The largest chloride concentration reported in a water sample from the Permian rocks was about 190,000 mg/L.

If the average chloride for the brine seepage is 170,000 mg/L, representing a range in chloride of about 150,000 to 190,000 mg/L, the seepage rate would be 2.62 ft^3/s to produce a chloride load of 1,200 ton/d. If the average chloride for the brine is only 150,000 mg/L,

Table 6.--*Altitude, in feet, of the water table as computed by the transport model*

[Blanks are nodes where transmissivity of the Quaternary aquifer is modeled as zero]

Row	Column 1	2	3	4	5	6	7	8	9	10
1										
2		1,194	1,194	1,192	1,187	1,184	1,191	1,186	1,182	1,178
3		1,194	1,193	1,191	1,187	1,185	1,186	1,181	1,177	1,174
4		1,194	1,191	1,188	1,184	1,182	1,181	1,174	1,169	1,167
5		1,193	1,190	1,185	1,181	1,176	1,173	1,164	1,162	1,159
6		1,196	1,190	1,184	1,179	1,174	1,171	1,165	1,158	1,151
7		1,197	1,191	1,184	1,177	1,172	1,166	1,162	1,156	1,151
8		1,196	1,191	1,186	1,176	1,168	1,164	1,161	1,154	1,148
9		1,196	1,188	1,181	1,173	1,161	1,161	1,157	1,152	1,147
10		1,196	1,185	1,179	1,172	1,164	1,159	1,154	1,148	1,142
11		1,199	1,183	1,177	1,169	1,163	1,157	1,151	1,144	1,139
12		1,203	1,182	1,175	1,164	1,157	1,156	1,150	1,141	1,135
13		1,204	1,182	1,175	1,164	1,157	1,156	1,149	1,141	1,134
14		1,224	1,183	1,176	1,169	1,163	1,158	1,152	1,145	1,136
15			1,186	1,183	1,171	1,166	1,159	1,153	1,145	1,137
16					1,175	1,173	1,165	1,160	1,164	1,158
17					1,184	1,183		1,165	1,172	
18				1,195	1,190	1,189	1,190			
19						1,197	1,196	1,199		
20										

Row	Column 11	12	13	14	15	16	17	18	19	20
1										
2	1,174	1,171	1,168	1,170	1,186	1,225	1,296	1,346	1,340	
3	1,171	1,168	1,166	1,164	1,175	1,221	1,290	1,345	1,365	
4	1,164	1,163	1,161	1,160	1,171	1,225	1,293	1,337	1,368	
5	1,157	1,158	1,158	1,158	1,173	1,215	1,278	1,318	1,349	
6	1,148	1,151	1,150	1,153	1,163	1,212	1,283	1,315	1,336	
7	1,144	1,141	1,139	1,142	1,152	1,192		1,319	1,332	
8	1,140	1,137	1,137	1,135	1,144	1,155	1,242	1,293	1,305	
9	1,140	1,136	1,135	1,131	1,134	1,144	1,201	1,232	1,229	
10	1,138	1,134	1,131	1,131	1,128	1,136	1,164	1,182	1,180	
11	1,136	1,132	1,129	1,125	1,126	1,127	1,147	1,163	1,162	
12	1,132	1,129	1,127	1,125	1,125	1,125	1,137	1,146	1,139	
13	1,130	1,127	1,125	1,125	1,125	1,125	1,125	1,125		
14	1,131	1,126	1,125	1,125	1,125	1,125	1,125			
15	1,131	1,126	1,125	1,125	1,125	1,125	1,140	1,179		
16	1,130	1,125	1,125	1,125	1,125	1,144	1,182			
17	1,132	1,127	1,127	1,129	1,144	1,167				
18	1,141	1,134	1,133	1,139	1,167	1,189				
19	1,152		1,139	1,155	1,199	1,202				
20										

representing a range of about 110,000 to 190,000 mg/L, the seepage rate would have to be 2.97 ft^3/s to produce the required load. Therefore, it seems reasonable to conclude that the rate of brine seepage is probably between 2.5 and 3 ft^3/s.

The areal distribution of the brine seepage is unknown and must be inferred from the pattern of increase in chloride concentration in water from the Quaternary aquifer (fig. 5 and 6). An average chloride concentration of 170,000 mg/L in the brine seepage was used in the transport model. Adjustments to the rate of brine seepage were made at each model node to improve the match between historic and model-computed chloride concentration.

Brine seepage was modeled as a specified flux into the model at certain model nodes. Initially brine seepage was modeled as a total seepage of 2.72 ft^3/s applied as a uniform seepage of 0.04 ft^3/s to 68 nodes representing the area of excessive chloride concentration in water from the Quaternary aquifer. Further model runs to improve the fit between historic and model-computed chloride concentrations from the transport model changed the total seepage to 5.13 ft^3/s applied as a nonuniform rate ranging from 0.005 to 1.2 ft^3/s at 62 nodes. Total seepage of 5.13 ft^3/s with a chloride concentration of 170,000 mg/L is a chloride load that is about twice as large as indicated by historic data. The final seepage distribution used in the transport model was a total of 2.81 ft^3/s applied to 70 nodes at rates ranging from 0.005 to 0.5 ft^3/s.

Concentrations of chloride computed in the transport model and the input data on rate of brine seepage are listed in table 7. The chloride concentration computed in the model represents the concentration averaged over a vertical section of the aquifer. This computed concentration approximates the average of the two values for chloride concentration at the water table (fig. 5) and at the top of bedrock (fig. 6).

In addition to the dispersion of solutes created by flow through the aquifer, the numerical technique used to solve the transport equations generally brings about other dispersive effects. This mathematical modeling effect is called "numerical dispersion." Numerical dispersion in the model is most noticeable at node (16, 10), which has a computed chloride concentration of 1,290 mg/L but no apparent brine source (fig. 7); in effect more dilute chloride concentrations exist up gradient from this node. Numerical dispersion also has increased the computed chloride concentrations at nodes (17, 15) and (16, 16) and, to a lesser amount, the computed chloride concentration for some other nodes on the south and east sides of the central area of excessive chloride concentrations. Numerical dispersion tends to reduce concentration gradients. The effects of numerical dispersion during calibration of the transport model probably has caused the model simulations to indicate less brine

Table 7.--*Chloride concentration, in milligrams per liter (upper number), computed by the transport model and input values for brine seepage, in cubic feet per second (lower number)*

[Chloride concentrations are rounded to nearest 10 milligrams per liter. Blanks are nodes where transmissivity of the Quaternary aquifer is modeled as zero. Brine seepage is zero where a lower number is not shown]

	Column																			
Row	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
1																				
2		250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	
3		250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	
4		250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	250	
5		250	250	250	250	250	250	360	430	270	250	250	250	250	250	250	250	250	250	
6		250	250	250	250	250	260	400	10,370 0.02	19,400 0.02	3,500	440	250	250	250	250	250	250	250	
7		250	250	250	250	250	290 0.01	4,840 0.005	13,380 0.01	45,540 0.05	40,380 0.01	17,130 0.01	320	260	250	250		250	250	
8		250	250	250	250	250	540	8,760 0.01	14,050 0.01	22,170 0.01	34,440 0.01	47,070	10,150 0.005	290	260	250	250	250	250	
9		250	250	250	250	250	590	13,800 0.005	13,920 0.005	17,450 0.01	27,230 0.01	60,600	51,050 0.03	4,660 0.005	350	260	250	250	250	
10		250	250	250	250	250	350	9,010 0.01	11,090	16,830 0.005	42,320 0.01	124,820 0.005	124,030 0.005	56,250 0.05	10,780 0.02	320	250	250	250	
11		250	250	250	250	250	310	5,490 0.005	19,820 0.02	24,150 0.005	82,580 0.02	137,900 0.01	159,660 0.02	57,220	8,160 0.005	320	260	250	250	
12		250	250	250	250	250	250	630	28,130 0.03	40,300 0.02	121,640 0.005	140,490 0.005	159,060 0.01	167,250 0.02	49,130	7,060 0.01	370	250	250	
13		250	250	250	250	250	270	640	23,400 0.03	56,380 0.05	154,450 0.02	166,310 0.02	168,970 0.02	155,700 0.01	110,000	21,310 0.01	360	250		
14		250	250	250	250	250	250	360	7,380 0.01	87,140 0.1	140,230 0.5	154,080 0.03	169,200 0.02	145,010	110,000	11,790 0.01	350			
15			250	250	250	250	260	1,970 0.005	6,930 0.02	39,440 0.1	115,480 0.5	132,950 0.05	169,090 0.02	140,000	52,320 0.05	8,280 0.03	340	250		
16					250	250	250	270	360	1,290	55,490 0.1	59,340 0.02	115,950 0.005	39,970 0.03	22,800 0.1	850	260			
17					250	250		250	250		32,700 0.02	69,880 0.2	50,180 0.1	24,980 0.1	850	280				
18				250	250	250	250				10,350 0.005	6,150	600	420	290	250				
19						250	250	250			430		260	250	250	250				
20																				

seepage on the margins of the excessive chloride area than actually exists. Numerical dispersion is a function of the spacing between model nodes. A reduction in node spacing probably would reduce the amount of numerical dispersion.

The total inflow of brine simulated in the model is 2.81 ft^3/s. The concentration of this brine is modeled as 170,000 mg/L. This is a chloride load of 1,290 ton/d, which is similar to that estimated from chloride records for the station near Jet.

The distribution of the rate of brine inflow or seepage into the model is shown in table 7. Increasing the brine seepage to values greater than those shown in table 7 produces a better match between the historic chloride and the computed chloride concentrations; however, this increases the total load above that indicated by records for the station near Jet. The best fit between historic and computed chloride concentrations would be an increase in brine seepage that produces a chloride load that is about twice the load indicated by the Jet station record. It is possible that the model parameters need to be changed and the model recalibrated so as not to produce the apparently excess total flux. The other possibility is that the simulation of the transport of solute at the salt flats between the water table and the land surface does not represent the actual transport processes.

Model simulations with the appropriate parameter changes were made to test the first possibility. Input values of transmissivity, leakance, recharge, maximum evapotranspiration, boundary flow, brine seepage, and pumpage were halved; this results in one-half the values of the previous seepage velocities, and the total simulation time for transport was doubled in order to produce the chloride load of record. This resulted in the same chloride distribution in the aquifer with a chloride load of 1,180 ton/d which is approximately what was estimated from records. Except for pumpage, the changes in the parameter values to one-half their estimated values appears permissible, in view of the information available on each parameter. Pumpage is a minor part of the total flux in the system, accounting for 13 percent of the total flux in the original model and 23 percent in the revised model. However, simulations with this revised model indicate excessive drawdown and dewatering of the aquifer at some of the pumpage nodes if hydraulic conductivity is reduced by one-half without also reducing pumpage. This reduction in hydraulic conductivity is too severe for the model as a whole.

Recalibration of the model by halving the model parameters is not possible unless pumpage estimates can be halved also. It seems unlikely that estimates of pumpage are twice the amount actually withdrawn.

For those areas where loss of water occurs through evapotranspiration (depth to water less than 6.5 ft), the net upward vertical flux is computed in the model as the algebraic sum of recharge and evapotranspiration. If the net flux indicates a gain of water, or recharge greater than evapotranspiration, solute is added to the model at a load rate based on the net flux and the concentration in the infiltration water. If the net sum indicates loss of water, solute is subtracted from the model at a load rate based on the net flux and the concentration of solute at the particular model node.

The process of solute removal, at nodes where evapotranspiration exceeds recharge, is by overland brine flow. On the salt flats during dry weather, brine is transported upward to the land surface where the water evaporates leaving a crust of salt behind. During rain, some or all of the salt crust is dissolved, forming a brine solution at land surface. Part of this brine solution infiltrates into the ground and becomes recharge to the aquifer, the remainder flows overland as brine runoff. During a wet season, the salt crust may disappear from the salt flats only to reappear during the next dry season.

The overland brine flow is not simulated by the transport model. Chloride and water balances for the transport model are tabulated below:

	Chloride concentration (tons per day)	Water flow (cubic feet per second)
Flows into the model:		
Recharge-----------------------	10.6	15.75
Boundary flow-----------------	1.2	1.82
Brine seepage-----------------	1,288.3	2.81
Streams-----------------------	1.5	2.26
Totals--------------------	1,301.6	22.64
Flows out of the model:		
Evaporation-------------------	958.7	11.59
Boundary flow----------------	1.6	2.33
Pumpage-----------------------	2.4	3.50
Great Salt Plains Lake--------	324.8	5.22
Totals--------------------	1,287.5	22.64

Because the model simulates only a single vertical layer representing the total thickness of the aquifer, vertical variations in transport cannot be represented. Brine seepage is mixed with lateral inflows containing much smaller concentrations of chloride. Actually, the more dense brine tends to remain near the base of the aquifer, and layers of water with progressively smaller concentrations are established from the base to the top of the aquifer. The simulation of water loss through evapotranspiration reflects the removal of more solute than in the actual condition, where the water removed by evapotranspiration is probably the less concentrated lateral inflow.

CHLORIDE-CONTROL PLAN

The Corps of Engineers' plan for controlling the brine discharge at the Great Salt Plains is shown in figure 8. Salient features of the plan include: Construction of a dike cutting through the thickness of the Quaternary aquifer and extending across the Great Salt Plains Lake; formation of a primary brine reservoir west of this dike; construction of a surface dike around three sides of Sucker Flat; routing of surface flow from Cottonwood and Clay Creeks through a diversion channel; impoundment of surface flows by reservoirs on Spring and Twin Spring Creeks, and routing of the outflow from these two reservoirs through a second diversion channel; and construction of a secondary brine reservoir (evaporation basin in fig. 8) in an area of the salt flats. Brine pipelines will carry surface flow from Sucker Flat, and from two small natural channels, under the northern diversion channel and onto the salt flats. These three pipelines will be gravity-flow systems. A pipeline will connect the primary brine reservoir with the secondary brine reservoir; this will be a pumped system designed to carry excess inflow from the primary brine reservoir to the secondary brine reservoir. Both areas of salt flats, Sucker Flat and Great Salt Plains, will receive surface inflow from a limited drainage area and have no downstream outlet into the drainage area of the Salt Fork Arkansas River, except in case of extreme flooding, when flows will pass over the dikes. Evaporation will concentrate the brines in the planned reservoirs.

Solute flux to the land surface is excessive in areas where evapotranspiration is simulated by the two-dimensional model. Solute movement in response to evapotranspiration can be simulated accurately only with a three-dimensional model.

The plan proposed by the Corps of Engineers will affect the flow system of the aquifer and the boundary conditions acting upon it. Construction of dikes cutting through the thickness of the aquifer will reduce the eastward underflow in the area. The new reservoirs, the secondary brine reservoir and the two reservoirs on Spring Creek and

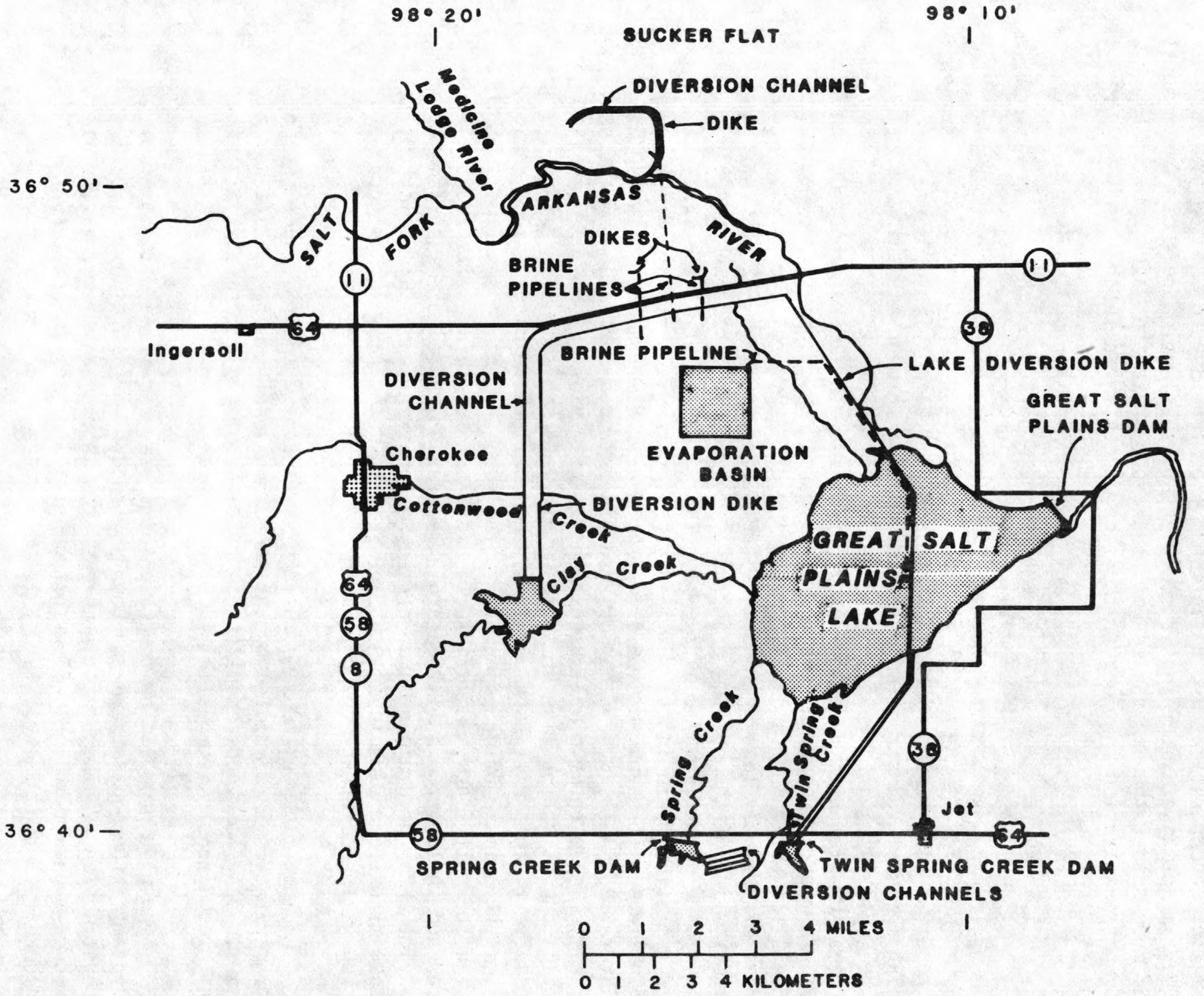


Figure 8.--Planned control structures. (From U.S. Army, Corps of Engineers, 1978b.)

Twin Spring Creek, will be new sources of inflow to the aquifer system. Any changes in hydraulic head in the primary reservoir and in the area downstream from the dike extending across the Great Salt Plains Lake will affect the flow system and the changes in chloride concentration in the two areas. The primary reservoir will be more mineralized, and the area downstream from the dike (freshwater part) will be less mineralized than at present. A resume of the conditions in the reservoirs is listed in table 8. Construction of the dikes will change existing drainage patterns on the Great Salt Plains. One node, (10, 13), represents a location where flow from the Salt Fork Arkansas River will be routed elsewhere under conditions of the Corps' chloride-control plan.

Table 8.--*Conditions in control-plan reservoirs*

Reservoir	Average water surface altitude (feet)	Average surface area (acres)	Average chloride concentration (milligrams per liter)
Freshwater partition of Great Salt Plains	1,125	3,000	Less than 250 (freshwater)
Primary brine (salt-water partition of Great Salt Plains)	1,125	5,400	Greater than 190,000 (saturated)
Secondary brine (evaporation basin)	1,150	864	Greater than 190,000 (saturated)
Spring Creek	1,171	166	Less than 250 (freshwater)
Twin Spring Creek	1,155	60	Less than 250 (freshwater)

The projected changes in water level, as a result of the stresses imposed on the ground-water model by the Corps' chloride-control plan, are listed in table 9. The boundary of the transport model on the southeast side of the lake is a ground-water divide. The ground-water model includes additional nodes beyond the limits of the transport model. Projections of hydraulic-head change using the ground-water model indicate a difference in water-level change, as compared to the transport model, of 0.5 ft at node (19, 15) and 0.1 ft or less at other nodes. The effect on computed velocities probably is not significant, if the other uncertainties in the transport model are considered.

Table 9.--*Changes in hydraulic head, in feet, resulting from the chloride-control plan as computed by the ground-water model*
[Minus indicates a decline in water level]

Row	Column 1	2	3	4	5	6	7	8	9	10
1										
2		0	0	0	0	0	0	0	0	0
3		0	0	0	0	0	0	0	0	0
4		0	0	0	0	0	0	0	0	0
5		0	0	0	0	0	0	0	0	0
6		0	0	0	0	0	0	0	0	0
7		0	0	0	0	0	0	0	0	0
8		0	0	0	0	0	0	0	0	0
9		0	0	0	0	0	0	0	0	0
10		0	0	0	0	0	0	0	0	0.1
11		0	0	0	0	0	0	0	0	.3
12		0	0	0	0	0	0	0	0	.1
13		0	0	0	0	0	0	0	0	0
14		0	0	0	0	0	0	0	0	0
15			0	0	0	0	0	0	0	0
16					0	0	0	0	0	0
17					0	0		0	0	
18				0	0	0	0			
19						0	0	0		
20						0	0			

Row	Column 11	12	13	14	15	16	17	18	19	20
1										
2	0	0	0	0	0	0	0	0	0	
3	0	0	0	0	0	0	0	0	0	
4	0	0	0	0	0	0	0	0	0	0
5	0	0	0	0	0	0	0	0	0	0
6	0	0	0	0	0	0	0	0	0	0
7	0	0	0	0	0	0		0	0	0
8	0	0	0	0	0	0	0	0	0	0
9	0.1	0.3	0	0	0	0	0	0	0	0
10	.5	2.2	-0.7	0	0	0	0	0	0	0
11	2.4	18.0	2.1	0	0	0	0	0	0	0
12	.6	2.4	.6	0	0	0	0	0	0	0
13	.1	.4	.2	0	0	0	0	0		0
14	0	.1	0	0	0	0	0		0	0
15	0	0	0	0	0	0	0	0	0	0
16	0	0	0	0	0	0	0	0	0	0
17	.2	.1	.1	.1	0	.1	0	0	0	
18	3.1	.2	1.0	.3	.2	.2	.1			
19	19.3		15.8	4.3	2.0	.3				
20				4.1	3.7					

The projected changes in chloride concentration, as a result of the stresses imposed on the transport model by the Corps' plan are listed in table 10. These changes are the result of simulating a 50-year period of solute transport using the computed results from the ground-water steady-state model after 50 years as initial conditions. The initial chloride distribution used in the simulation is the computed chloride listed in table 7. The projected chloride concentrations are listed in table 11. The largest chloride concentration (183,000 mg/L, an increase of more than 45,000 mg/L) was projected for node (11, 12), which is the location of the secondary brine reservoir. Larger increases in chloride concentration are projected for nodes adjacent to and in the vicinity of this node (table 10), but the total concentration (initial plus projected) is still largest at node (11, 12).

Although the projected chloride concentration at two nodes (10, 11) and (11, 11), will be doubled, the model projections indicate that significant increases in chloride concentrations will be present only where the existing water in the aquifer already contains excessive concentrations of chloride. Projections also indicate that the significant increases in chloride concentration will not extend into the part of the aquifer where the water is being used at present. Therefore, the chloride-control plan apparently will not significantly degrade the quality of water existing in the Quaternary aquifer.

The simulations include the effects of lateral inflow to a node with brine seepage, recharge from precipitation, and leakage from streams or the lake. Changes in chloride concentration reflect changes in rate and chloride concentration of lateral inflow. For nodes (11, 12), (19, 11), and (19, 13), which are new lake sources, the changes in chloride concentration also reflect changes in rate and concentration of leakage inflows. The concentration at a node is computed as the average for the particles within the block containing the node. Subtle changes in the seepage velocity can alter the flow pattern enough to produce significant changes in concentration.

Table 10.--*Changes in chloride concentration, in milligrams per liter, resulting from control plan as computed by transport model*

[Minus indicates a decrease in computed chloride. Values are rounded to nearest 10 milligrams per liter]

	Column									
Row	1	2	3	4	5	6	7	8	9	10
1										
2		0	0	0	0	0	0	0	0	0
3		0	0	0	0	0	0	0	0	0
4		0	0	0	0	0	0	0	0	0
5		0	0	0	0	0	0	0	0	0
6		0	0	0	0	0	0	0	0	0
7		0	0	0	0	0	0	0	0	30
8		0	0	0	0	0	0	110	90	60
9		0	0	0	0	0	0	0	10	250
10		0	0	0	0	0	0	10	100	480
11		0	0	0	0	0	0	-150	2,440	2,660
12		0	0	0	0	0	0	0	-10	-2,520
13		0	0	0	0	0	0	0	0	-60
14		0	0	0	0	0	0	0	0	0
15			0	0	0	0	0	0	0	0
16					0	0	0	0	0	0
17					0	0		0	0	
18				0	0	0	0			
19						0	0	0		
20										

	Column									
Row	11	12	13	14	15	16	17	18	19	20
1										
2	0	0	0	0	0	0	0	0	0	
3	0	0	0	0	0	0	0	0	0	
4	0	0	0	0	0	0	0	0	0	
5	0	0	0	0	0	0	0	0	0	
6	0	0	0	0	0	0	0	0	0	
7	20	0	0	0	0	0		0	0	
8	30	20	460	10	0	0	0	0	0	
9	390	230	51,620	-110	10	0	0	0	0	
10	54,650	55,960	44,270	8,290	-20	0	0	0	0	
11	93,100	45,550	22,520	64,900	-300	0	0	0	0	
12	50,160	42,340	22,690	90	420	-10	0	0	0	
13	90	350	300	180	-10	0	0	0		
14	0	10	130	-10	-10	0	0			
15	-10	-10	60	0	70	0	0	0		
16	-20	-40	220	-640	140	0	0			
17	-9,970	-3,330	1,110	-1,960	-10	0				
18	-1,630	1,400	-20	0	0	0				
19	-40		0	0	0	0				
20										

Table 11.--*Chloride concentration, in milligrams per liter, resulting from control plan as computed by transport model*
[Values are rounded to nearest 10 milligrams per liter]

Row	Column 1	2	3	4	5	6	7	8	9	10
1										
2		250	250	250	250	250	250	250	250	250
3		250	250	250	250	250	250	250	250	250
4		250	250	250	250	250	250	250	250	250
5		250	250	250	250	250	250	360	430	270
6		250	250	250	250	250	260	400	10,370	19,400
7		250	250	250	250	250	290	4,840	13,380	45,570
8		250	250	250	250	250	540	8,870	14,140	22,230
9		250	250	250	250	250	590	13,800	13,930	17,700
10		250	250	250	250	250	350	9,020	11,190	17,310
11		250	250	250	250	250	310	5,340	22,260	26,810
12		250	250	250	250	250	250	630	28,120	37,780
13		250	250	250	250	250	270	640	23,400	56,320
14		250	250	250	250	250	250	360	7,380	87,140
15			250	250	250	250	260	1,970	6,930	39,440
16					250	250	250	270	360	1,290
17					250	250		250	250	
18				250	250	250	250			
19						250	250	250		
20										

Row	Column 11	12	13	14	15	16	17	18	19	20
1										
2	250	250	250	250	250	250	250	250	250	
3	250	250	250	250	250	250	250	250	250	
4	250	250	250	250	250	250	250	250	250	
5	250	250	250	250	250	250	250	250	250	
6	3,500	440	250	250	250	250	250	250	250	
7	40,400	17,130	320	260	250	250		250	250	
8	34,470	47,090	10,610	300	260	250	250	250	250	
9	27,620	60,830	102,670	4,550	360	260	250	250	250	
10	96,970	180,780	168,300	64,540	10,760	320	250	250	250	
11	175,680	183,450	182,180	122,120	7,860	320	260	250	250	
12	171,800	182,830	181,750	167,340	49,550	7,050	370	250	250	
13	154,540	166,660	169,270	155,880	109,990	21,310	360	250		
14	140,230	154,090	169,330	145,000	109,990	11,790	350			
15	115,470	132,940	169,150	140,000	52,390	8,280	340	250		
16	55,470	59,300	116,170	39,330	22,940	850	260			
17	22,730	66,550	51,290	23,020	840	280				
18	8,720	7,550	580	420	290	250				
19	390		260	250	250	250				
20										

CONCLUSIONS

The models constructed to project the effects of the Corps' chloride-control plan on the Quaternary aquifer in the Great Salt Plains area are only applicable for preliminary projections because of several limiting factors. For example, the aquifer system is assumed to contain water of constant density, but the real system contains water with density differences of about 20 percent. Brine seepage from underlying formations into the aquifer is simulated at a constant rate, but the seepage rate and distribution probably are a function of the hydraulic head in the Quaternary aquifer. Increases in hydraulic head in the Quaternary aquifer at specific areas will tend to decrease the brine seepage at these locations. Dense brine water at the base of the Quaternary aquifer in some areas would tend to produce the same type of effect, and, in essence, the correct relationship between seepage rates and the vertical distribution of pressure need to be defined. These factors plus questions of correct transmissivity and model flux limit the model projections for use as only preliminary approximations.

A more comprehensive analysis of the effects of the chloride-control plan on the ground-water flow system, including both Quaternary and Permian deposits, may be effected with a three-dimensional, non-uniform density model. This type of model, plus the additional necessary data, need to be part of any further plans related to project.

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