DEVELOPMENT AND SENSITIVITY ANALYSIS OF STEADY-STATE AND TRANSIENT OGALLALA AQUIFER GROUND-WATER FLOW AND PARTICLE TRACKING MODELS

Final Report

by

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Prepared for the
U.S. Department of Energy
under
Grant No. DE-FG04-90AL65847

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1995

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ABSTRACT

Ground-water flow and transport models of the Ogallala aquifer were constructed and calibrated for the northeastern portion of the Southern High Plains to evaluate the spatial distribution and magnitude of recharge and ground-water flow and contaminant transport. The U.S. Department of Energy's Pantex Plant, near the center of the model area, was the primary focus of this study. Concerns about potential contamination of the Ogallala aquifer from activities at the Pantex Plant have necessitated this modeling effort to better understand the hydrogeologic processes controlling ground-water flow to and within the Ogallala aquifer.

An extensive literature review synthesized previous recharge hypotheses in an effort to develop an accurate conceptual model of recharge to the Ogallala aquifer. Most of these studies designated playas as either one of several possible areas of recharge or as the primary focal point of recharge to the Ogallala aquifer. The data base of geologic, geomorphic, geophysical, hydrologic, hydrochemical, and pedologic information supporting this hypothesis is extensive and expanding.

Historically, several published ground-water flow models, which were constructed primarily for water-resource evaluation, used spatially uniform recharge. However, contaminant transport velocities associated with focused recharge through playas were found to be significantly higher than those associated with spatially uniform recharge.

The purpose of this study was to develop an accurate ground-water flow model to evaluate the spatial distribution and velocity of contaminant transport to and in the Ogallala aquifer. Three basic methods were used to apply recharge and to evaluate the sensitivity of the system to various recharge scenarios. These included spatially uniform recharge, where recharge was applied at an equal rate throughout the model area; zonal recharge, where recharge was varied on the basis of regionally mapped geologic units; and modified zonal recharge, for which, in areas where playas are present at the surface, all recharge was focused through the playas and, in areas where playas are absent (interplaya areas), recharge was distributed uniformly. In the modified zonal recharge scenario, an equivalent regional rate of 6 mm yr^{-1} (0.24 inch yr^{-1}) was focused through the floors

of playas for an effective (focused or site specific) recharge rate of 219 mm yr^{-1} (8.6 inches yr^{-1}) and in zones where no playas are present, a uniform rate of 9 mm yr^{-1} (0.35 inch yr^{-1}) was used. Results for the modified zonal recharge scenario under steady-state conditions indicate that the statistical agreement between observed and simulated systems was as good or slightly better than that achieved with the other two methods.

The steady-state modified zonal recharge model was then used to develop a transient model incorporating ground-water withdrawals in the area since the late 1950's. Output files from the calibrated transient model were used to delineate potential directions and rates of flow from several points in the Ogallala aquifer at the Pantex Plant, the focus of this study, using a range of reasonable effective porosity values. Physical transport times from the various potential points of entry into the Ogallala aquifer to discharge points to the north and east of the plant ranged from several tens to hundreds of years.

INTRODUCTION

Steady-state, transient, and particle tracking, finite-difference models were constructed and calibrated to investigate the various hydrologic and geologic controls on the Ogallala aquifer system in the region of the U.S. Department of Energy's Pantex Plant. The model area includes all or parts of 11 counties in the Southern High Plains of Texas (figs. 1 and 2). In this area, the aquifer is heavily pumped to meet diverse water requirements, including agricultural, municipal, industrial, and domestic needs.

This section of the Ogallala aquifer is highly suitable for numerical simulation because most of it has natural hydrologic boundaries. The study area (fig. 1) is delimited by the Canadian River Valley to the north and the erosional limit of the Ogallala Formation at the Southern High Plains escarpment to the south-southeast. To the east, the Texas–Oklahoma border serves as the boundary, and the western edges of Randall and Potter Counties near the 102° meridian form the western boundary. The western boundary is an area where the structural base of the Ogallala

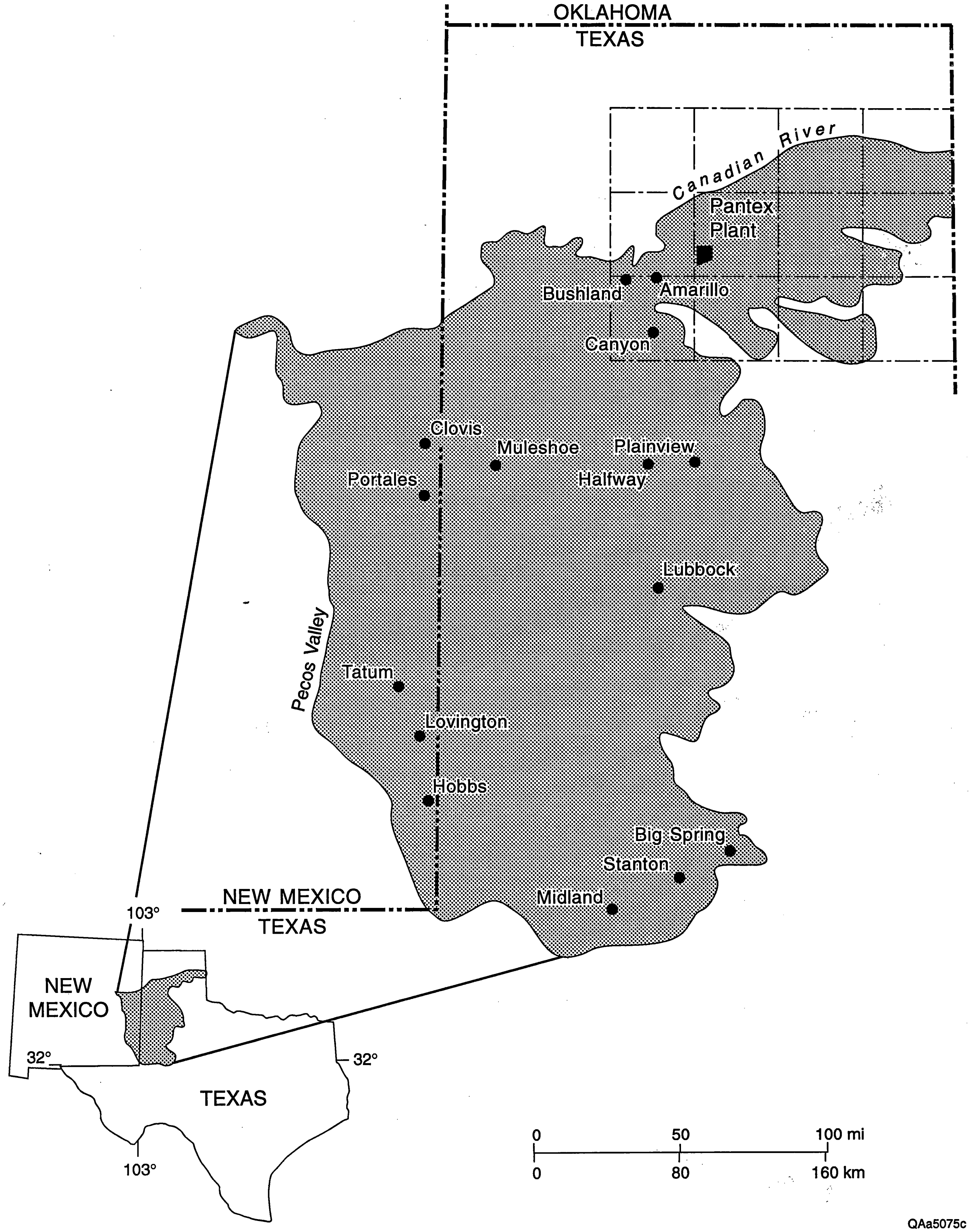

Figure 1. Location map of the study area showing 11 counties in the Southern High Plains of Texas included in model simulations.

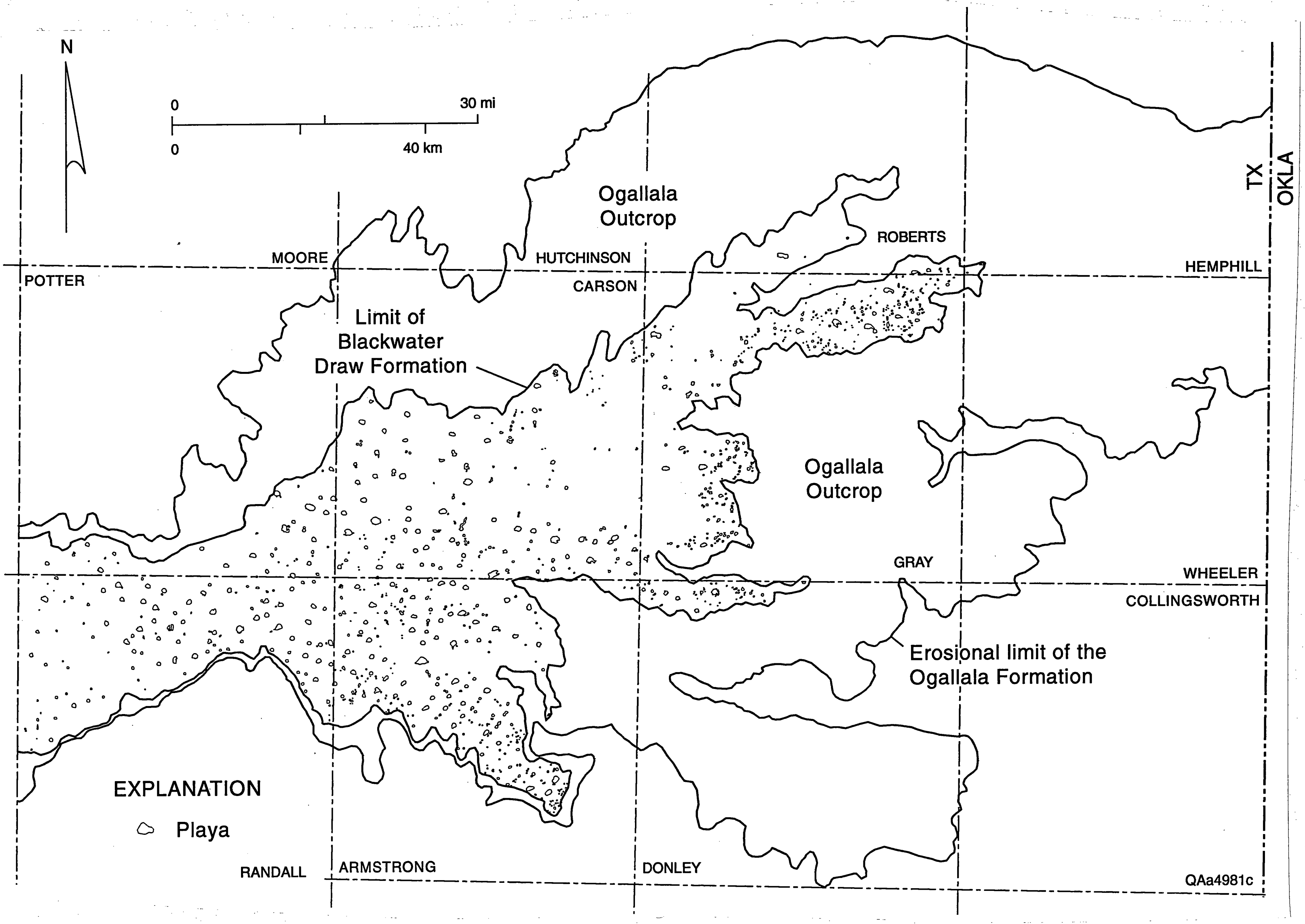

Figure 2. Geologic map of playas and the Quaternary Blackwater Draw and late Tertiary Ogallala Formation.

Formation is relatively high, the saturated thickness of the Ogallala aquifer is very thin, and the aquifer width is also at a minimum (at its narrowest point less than 22 km [13.7 mi] wide). The western boundary near the 102° meridian was also used in Ogallala aquifer models by Knowles and others (1984a), Luckey (1984), and Luckey and others (1986). Previously, Long (1961) described this area as a ground-water divide, thus implying that this area is a natural hydrologic boundary. Earlier studies using this model boundary have reported no problems with this boundary condition. Although the 102° meridian may not be a wholly natural hydrologic boundary, the narrow width and relatively thin saturated thickness of the aquifer in this region allows separation of the modeled area from the rest of the Ogallala aquifer to the west and south.

The three primary goals for the steady-state, transient, and particle tracking models of the Ogallala aquifer were to (1) delineate areas and quantify rates of recharge and discharge, (2) evaluate ground-water resources on the basis of current production trends, and (3) predict the directions, velocities, and fate of contaminants that may enter the aquifer.

In a previous report, Mullican and others (1994b) described initial results from steady-state model development and calibration. This steady-state ground-water flow model was used in developing a transient model to predict directions, rates, and controls on contaminant transport in the Ogallala aquifer in the vicinity of the Pantex Plant. The results presented here are an expansion of Mullican and others (1994b) and include steady-state, transient, and transport models of the Ogallala aquifer.

Nomenclature

Regional investigations that include this study area have used the term High Plains aquifer instead of Ogallala aquifer (for example, Gutentag and others, 1984). The term High Plains aquifer was used in those studies because stratigraphic units other than the Ogallala Formation, such as the Cretaceous Trinity, Fredericksburg, and Washita Groups and the Triassic Dockum Group, locally contain significant volumes of ground water in hydrologic continuity with the Ogallala aquifer (for example, Knowles and others, 1984a; Luckey and others, 1986; Nativ and Gutierrez, 1988).

Because all significant ground water produced in the study area is from the Ogallala Formation, the term Ogallala aquifer is used in this report. Furthermore, ground water in the study area is typically understood by landowners and members of the ground-water industry to be produced from the Ogallala aquifer. It should be understood, however, that this term can correctly be interchanged with the High Plains aquifer.

Geologic Setting

The principal ground-water unit and source of potable water in the study area is the Ogallala aquifer. This aquifer is within the late Tertiary Ogallala Formation, which underlies the Southern High Plains of Texas and eastern New Mexico. In the study area the Ogallala Formation consists of alluvial sediments (primarily sands and gravels) partly filling paleovalleys eroded into the pre-Ogallala surface and extensive, thick eolian sediments (silty, very fine sand or loamy clay with numerous buried calcic soils) capping paleouplands and most fluvial sections (Gustavson and Winkler, 1988). These sediments were probably deposited under arid- to subhumid-climatic conditions (Gustavson and Winkler, 1988). The Ogallala Formation is overlain by the Quaternary Blackwater Draw Formation; throughout much of the study area (see fig. 2), Ogallala strata are exposed along erosional escarpments. The Blackwater Draw Formation is the most widespread post-Ogallala unit throughout the Southern High Plains and consists of eolian sands and silts interbedded with numerous buried calcic soils (Holliday, 1989, 1990).

The Ogallala Formation unconformably overlies Permian and Triassic strata (McGookey and others, 1988). According to Seni (1980), the Ogallala Formation has a greater depositional thickness in western Carson County (immediately north of the Pantex Plant) than anywhere else in the High Plains of Texas. This depocenter, which Gustavson and Winkler (1988) referred to as the Panhandle Paleovalley, contains more than 274 m (900 ft) of Ogallala and Blackwater Draw Formation sediments (Long, 1961; McAdoo and others, 1964; Seni, 1980). This is also an area where the saturated section of the Ogallala aquifer on the Southern High Plains is at its maximum thickness. McAdoo and others (1964) described one well with 131.7 m (432 ft) of saturated sands and gravels, in addition to minor amounts of silts, clays, and caliches.

The most obvious topographic features of the Southern High Plains are the abundant, small, ephemeral lakes, commonly known as playas. Wood and Osterkamp (1987) described the playas as generally circular, and typically less than 1 km (0.62 mi) in diameter, and less than 4 m (13 ft) in depth. In one study of playas using digital mapping techniques and LANDSAT imagery, the Texas Department of Water Resources (1980) estimated the total number of playas on the High Plains of Texas at approximately 19,250.

The Southern High Plains, also referred to as the Llano Estacado, or "Staked Plains," is bounded on the east and west by the Caprock Escarpment, on the north by the Canadian River Valley, and grades into the Edwards Plateau to the south. The study area includes the northeasternmost part of the Southern High Plains and surrounding escarpments where the Ogallala and Blackwater Draw Formations are exposed. Drainage on the High Plains is internal into numerous playa basins, while drainage of the eroded margins of the High Plains is into the Canadian River, the Red River, and their tributaries.

The climate of this region is classified as continental steppe in the western part and subtropical subhumid in the eastern part (Larkin and Bomar, 1983). Annual precipitation ranges from approximately 45 to 56 cm yr^{-1} (18 to 22 inches yr^{-1}) and increases from west to east (fig. 3). The average annual gross lake surface evaporation rate varies from 183 to 190 cm yr^{-1} (72 to 75 inches yr^{-1}) (Larkin and Bomar, 1983), and the average evaporation pan rate (measured at the Amarillo–Bushland Station) is 196.26 cm yr^{-1} (77.27 inches yr^{-1}) (Dougherty, 1975).

PREVIOUS INVESTIGATIONS

Numerical models have been used as predictive tools to simulate ground-water flow in the Ogallala aquifer by several investigators (Claborn and others, 1970; Rayner, 1970; Bell and Morrison, 1979; Simpkins and Fogg, 1982; Knowles, 1984; Knowles and others, 1984a; McAda, 1984; Kier and others, 1984; Luckey, 1984; Luckey and others, 1986). Most previous modeling

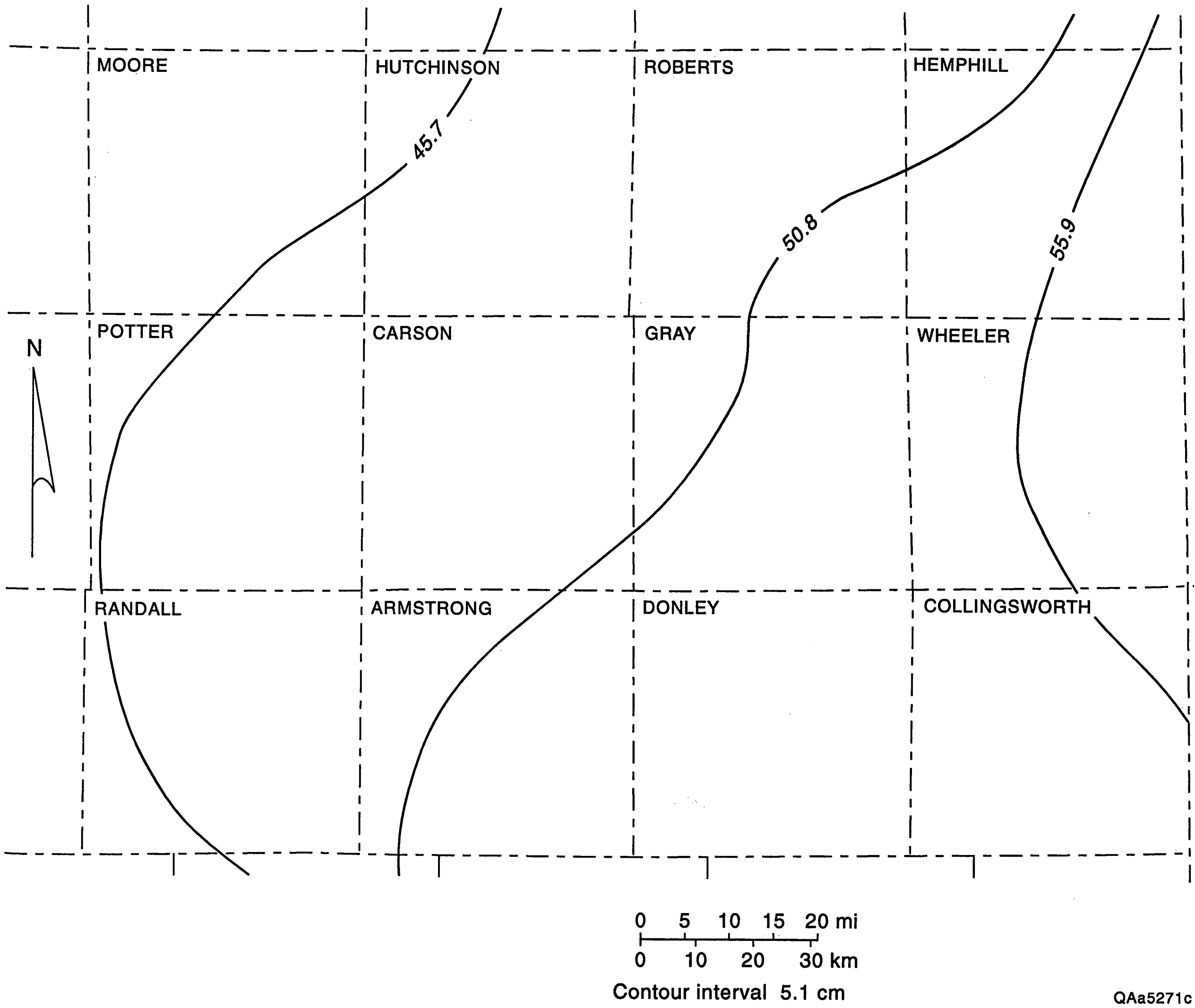

Figure 3. Precipitation map of the study area.

efforts attempted to quantify recharge, principally from infiltration of precipitation, to the Ogallala aquifer. The amount of recharge relative to natural discharge and artificial withdrawals indicates how efficiently the aquifer is being replenished or to what extent ground water is being "mined" from the aquifer. More importantly for this study, accurate prediction or measurement of recharge is required to quantify rates of contaminant transport to the aquifer.

Constructing ground-water flow models requires assigning recharge rates to areas within the model where recharge was measured or where it is believed to be active. Previous numerical models of ground-water flow in the Ogallala aquifer have relied on regionally applied, uniform rates of recharge throughout the model area. The use of regionally based recharge rates was largely constrained by the coarseness of previous model grids, in which recharge could only be varied over broad areas. The estimated rate(s) of recharge were typically derived through calibration of the model and not as the result of direct physical measurements. Although recharge values based on direct hydrologic and chemical measurements were available in the literature (for example, Johnson, 1901; Theis, 1937; White and others, 1946), the recharge rates used in previous models were often adjusted until a match between observed water levels, discharges, and withdrawals and those simulated by the model was obtained.

In developing the models described in this report, we have attempted to move away from the derivation of uniform or zonal recharge rates based upon model calibration. This model relies on field measurements and observations of the spatial location and magnitude of recharge to the Ogallala aquifer. An extensive literature review, presented in Appendix A of this report, provides the basis of our understanding of each of these two elements.

METHODS

The numerical models developed of the Ogallala aquifer were carried out in three stages. The first stage was used to calibrate the model and to investigate the plausibility of three recharge scenarios. In this stage, the aquifer was simulated in a "predevelopment" state before significant agricultural, municipal, or industrial withdrawals occurred. This approach was also used by

Gutentag and others (1984) and Knowles and others (1984a) in larger regional studies of the Ogallala (High Plains) aquifer. Under these predevelopment conditions, the aquifer is assumed to be in a long-term state of equilibrium, a "steady-state," where inflows and outflows of water are balanced and water-table elevations are stable through time.

In the second stage of modeling, the calibrated model was utilized to simulate the Ogallala aquifer in a transient manner for the period 1960–90 when large changes in water-table elevations occurred, both locally and regionally. This stage of modeling permitted the evaluation of whether the rates of recharge proposed in the predevelopment stage of modeling could be coupled with the substantial known withdrawals of water to predict water-level declines over the 1960–90 period. This stage of modeling also provided for the investigation of the hydrologic relationship of the Amarillo Carson County Well Field (ACCWF) to the changes in the water table in the vicinity of the DOE Pantex Plant.

The third stage of modeling involved a telescoping procedure to focus in on the Pantex Plant for the evaluation of rates and directions of potential contaminant transport within the Ogallala aquifer. In this stage, only recharge waters originating from playas on the Pantex Plant were evaluated along their flow paths to probable points of discharge from the system.

The finite-difference code MODFLOW (McDonald and Harbaugh, 1988) was used to model ground-water flow in the Ogallala aquifer. MODFLOW offers two iterative solution methods, each of which requires the specification of one or more parameters controlling this iterative process. A testing procedure was used to arrive at a set of parameters yielding a balance between computational speed and level of accuracy in the final set of hydraulic heads (McDonald and Harbaugh, 1988). Mass-balance calculations provided at the end of computations were used to evaluate the performance of the solution.

Steady-State Model Development

Four different types of data were required to establish and calibrate the predevelopment or steady-state portion of the model. These were aquifer geometry, hydraulic conductivity, boundary conditions, and inputs to or withdrawals from the aquifer (stresses).

Each cell of the model was assigned a specified size in three dimensions and a hydraulic conductivity. Additionally, cells on the boundary of the model must have boundary conditions describing their connection to unmodeled areas. All known stresses to the system, such as recharge and withdrawal, were entered into the model in the appropriate cells. For predevelopment conditions, recharge was the only stress applied to the model aquifer, with natural discharge through springs numerically simulated with the boundary conditions.

A one-layer finite-difference model was constructed, and all ground-water flow in the system was restricted to the Ogallala aquifer. This restriction ignores cross-formational flow to or from underlying Triassic and Permian strata. Simpkins and Fogg (1982) concluded that one of the three possible mechanisms for ongoing salt dissolution, especially the dissolution observed along the eastern escarpment of the Southern High Plains, was cross-formational flow from the Ogallala aquifer down to the salt-bearing horizons of the Permian. Wirojanagud and others (1986) used a ground-water flow model to simulate the magnitude of this cross-formational flow. On the basis of their published results, we have estimated the flux to be 0.019 to 0.026 mm yr^{-1} (6.23 to 8.53×10^{-5} ft yr^{-1}) for Carson County. For comparison, if a representative recharge rate to the Ogallala aquifer of 6 mm yr^{-1} (0.236 inch yr^{-1}) was applied, cross-formational flow out of the aquifer would account for approximately 0.3 percent of recharge to the aquifer. Fisher and Kreitler (1987), however, reported no evidence of cross-formational flow based on water chemistry of the deep-basin brines. Senger and others (1987), in a study of the hydrodynamics of the Palo Duro Basin, estimated vertical leakage rates from the Ogallala and Dockum aquifers down to the evaporite aquifers to be from 1.25 to 1.55×10^{-2} m^3 d^{-1} (4.41 to 5.47 ft^3 d^{-1}). The loss through the base of the Ogallala Formation therefore appears to be negligible, especially compared with the recharge and internal flow components.

Our goal during model construction was to use only geologic and hydrologic data based predominantly upon field measurements, at least until extensive calibration efforts had been made. The desire was to produce a deterministic, more accurate ground-water flow model of the Ogallala aquifer for the study area than previous larger scale, coarser grid models had presented. As is discussed in the following sections, obtaining input of hydrologic parameters was commonly straightforward and was aided by significant data sets, but in some cases hydrologic data were sparse. A description of each of the input data sets, its source, and data quality is presented in the following sections.

Aquifer Geometry

Aquifer geometry defines the spatial extent of the modeled part of the aquifer, both in areal and vertical dimensions. To approximate the areal extent and shape of the Ogallala aquifer, the model grid contains 171 rows and 210 columns for a total of 35,910 cells. Of these, 26,167 cells were used as active cells during the simulations. Nonactive cells represent areas where the Ogallala Formation is nonexistent or exterior to the area of interest (such as to the north of the Canadian River).

The model was designed so that increasing detail could be applied toward the central part of the model in the primary area of interest around the Pantex Plant and the adjacent ACCWF (fig. 4). Three basic cell sizes were used in this model (1.609 km by 1.609 km [1 mi by 1 mi], 1.609 km by 0.402 km [1 mi by 0.25 mi] and 0.402 km by 0.402 km [0.25 mi by 0.25 mi]) (fig. 4). Single intermediate rows and columns were used with cell dimensions of 0.402 km by 0.402 km (0.5 mi by 0.5 mi) to make the transition from large to small cells.

The base of the model was derived from maps and data of the base of the Ogallala aquifer (Knowles and others [1984a, their fig. 20]; supplemented by the maps of Knowles and others [1982, 1984b], Bowers and McReynolds [undated], and Seni [1980]) (fig. 5). These maps and

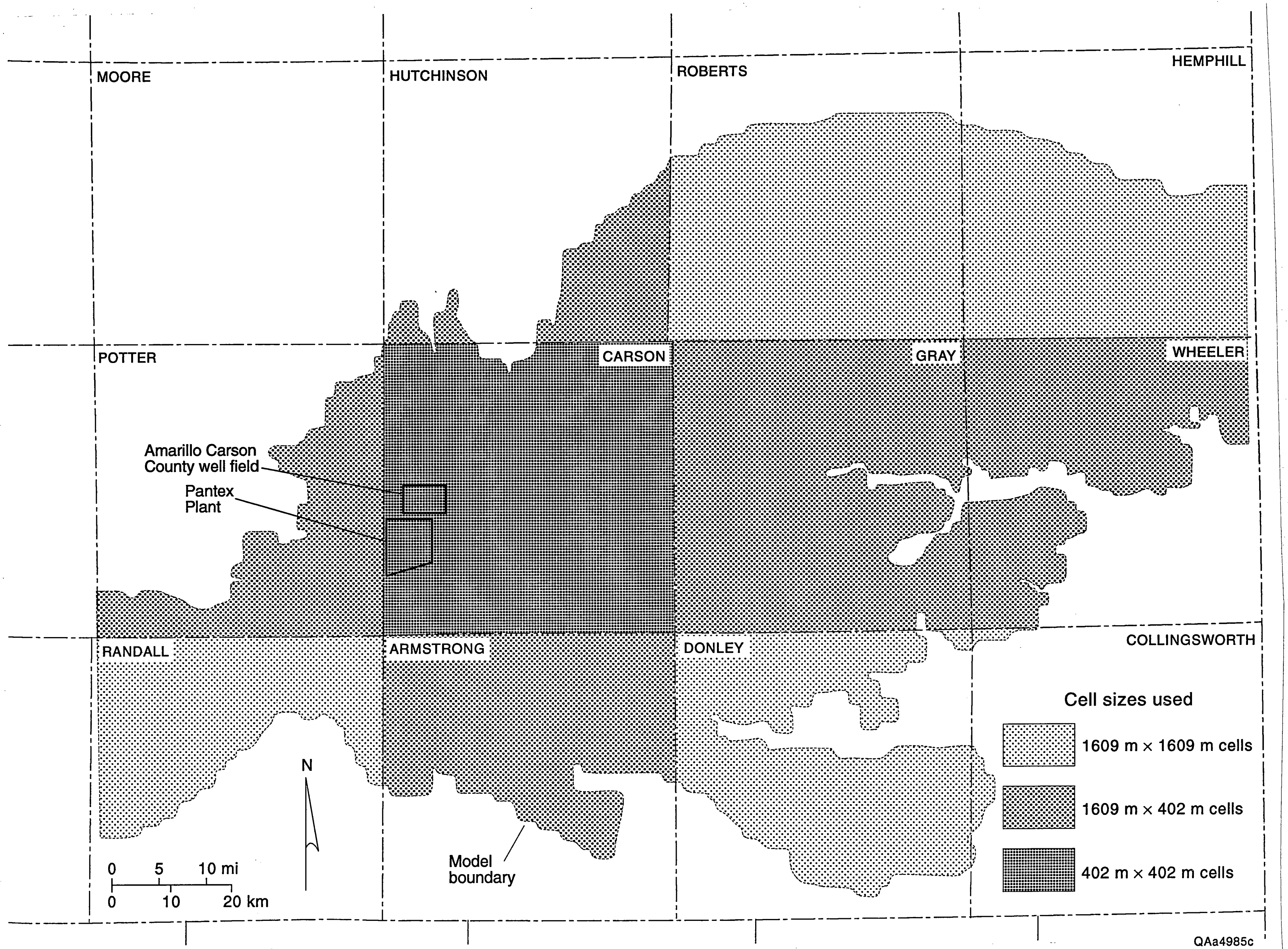

Figure 4. Finite-difference grid of the ground-water flow model constructed for this study.

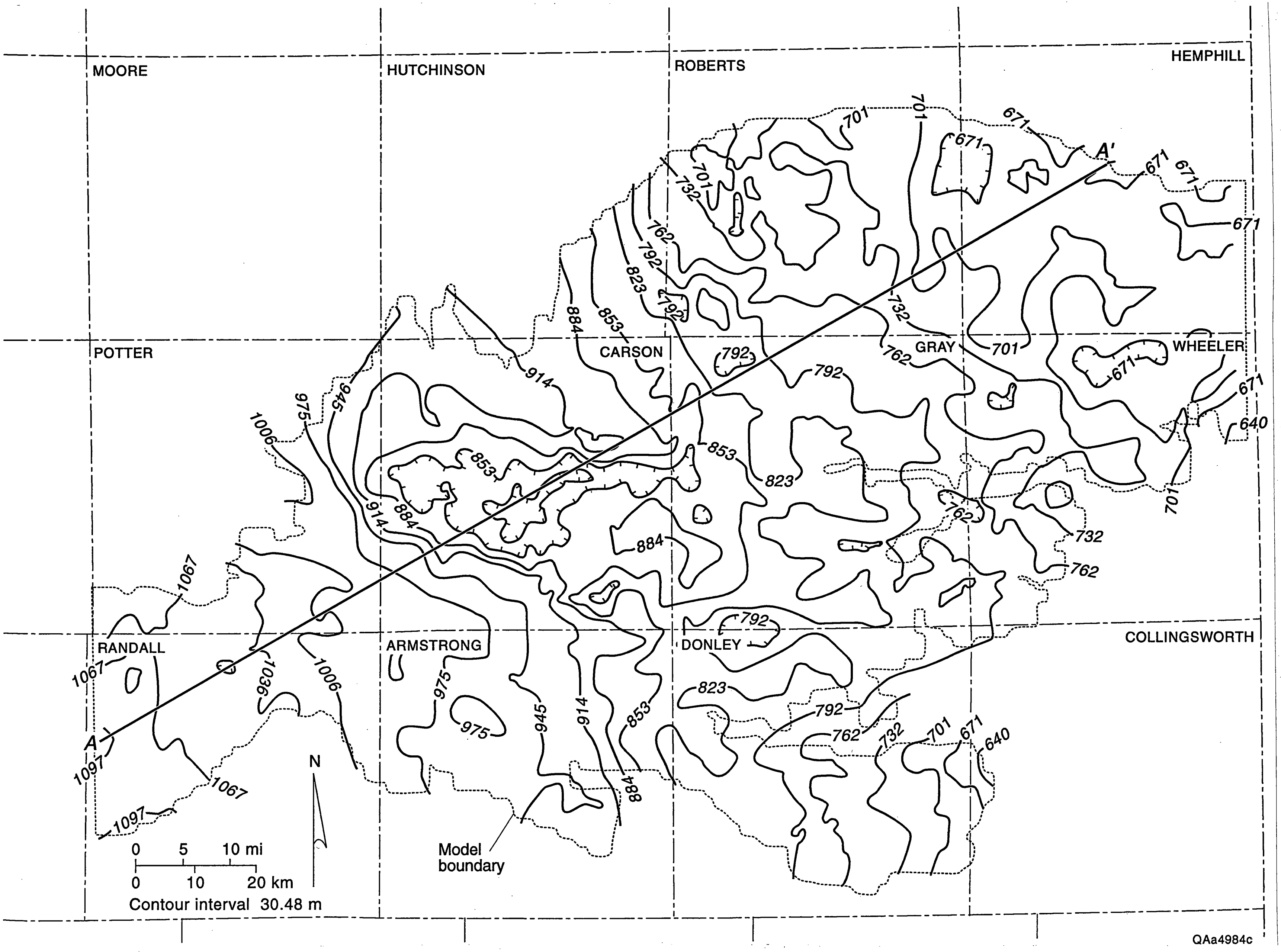

Figure 5. Structure map on the base of Ogallala aquifer showing elevation in meters above sea level and cross-section line.

other data were digitized to generate a composite digital map. Using this digital map of the base of the Ogallala aquifer, the model finite grid was overlain to interpolate a bottom elevation for each cell of the model.

As a means of evaluating the model calibration procedure for the predevelopment stage, the estimated water-table surface as of 1959–60 was used as a baseline for comparison with model simulation results. The 1959–60 water table was used because well withdrawals for agricultural, municipal, and industrial uses in the study area were rather limited at that time. The principal information source was the estimated 1959–60 water table presented by Knowles and others (1984a, their fig. 42), although they acknowledge that this map was based on limited data in much of the study area.

By subtracting the elevation of the base of the Ogallala aquifer from the 1959–60 water-table elevation, a map of 1959–60 saturated thickness results (fig. 6). Each cell of the finite-difference grid was assigned a "known" water-table elevation (top of model) and saturated thickness. As described below, the 1959–60 water-table surface was also used in constructing the set of boundary conditions for the model. Figure 7 illustrates the 1990 Ogallala aquifer water table.

Hydraulic Conductivity

The calibration of most ground-water flow models is limited by the amount of hydraulic conductivity data. This was also the case in this study area. Myers (1969), in a major compilation of approximately 968 aquifer test results in Texas, listed results from only 1 well in the study area (located in Randall County).

An initial goal during the construction of this ground-water flow model was to use published maps of hydraulic conductivity (permeability) (Knowles and others, 1984a; Luckey and others, 1986). However, these maps were based primarily on the calibration of numerical model. For the present study, hydraulic conductivity was derived by using water-well test data.

A common measure of a well's performance is the specific capacity test, which describes a well's ability to produce water. The specific capacity test is performed by pumping the well at a

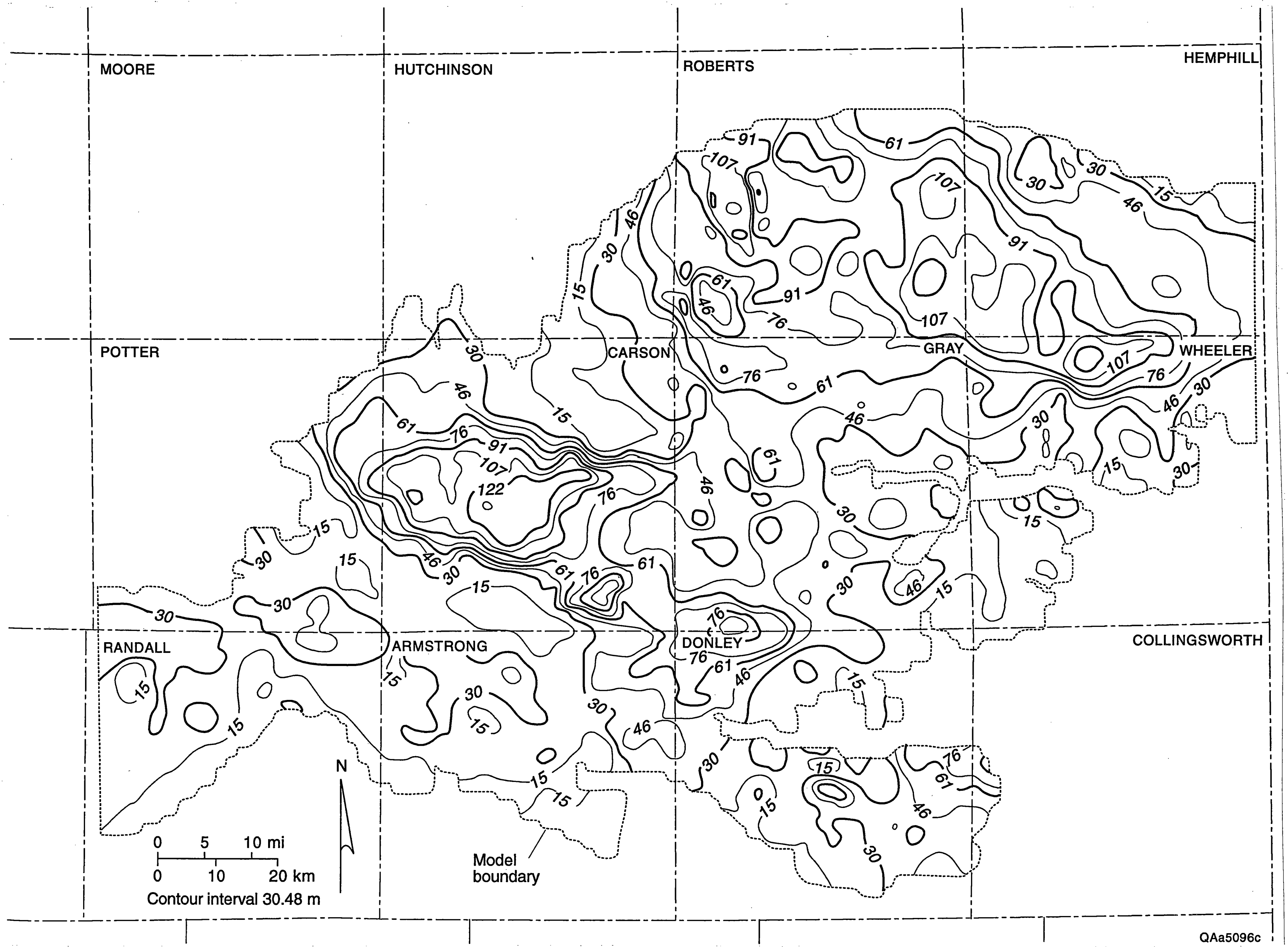

Figure 6. Estimated saturated thickness of the Ogallala aquifer during 1959 and 1960, which in this area is considered to be prior to significant ground-water withdrawals.

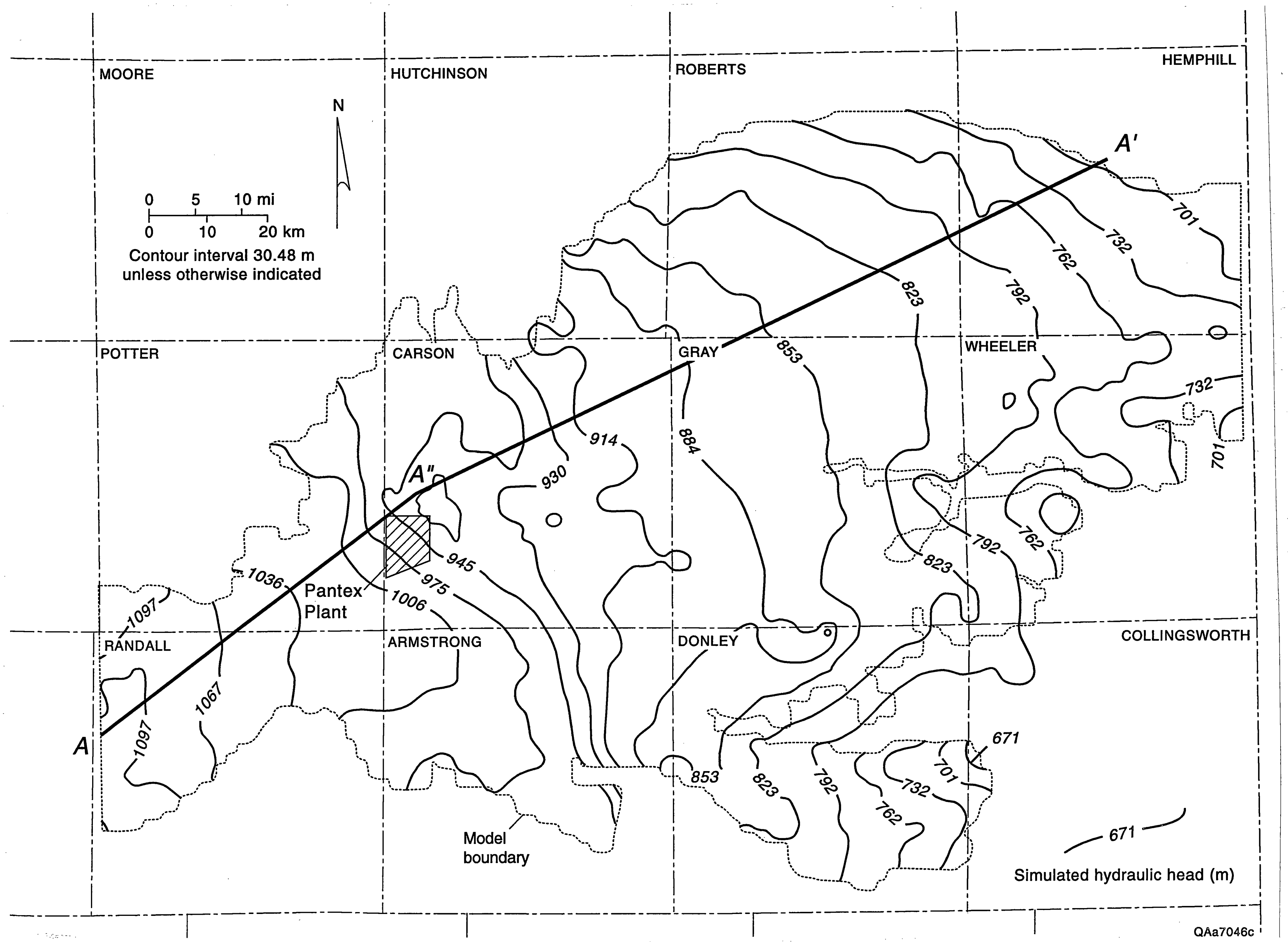

Figure 7. Ogallala aquifer water-table surface map for 1990.

constant rate (discharge) and recording the decline in water level (drawdown) at the end of the test. The specific capacity is the ratio of the discharge to the drawdown. If duration of pumping for the test, radius of the well, and storativity are known, hydraulic conductivity can be derived for the aquifer in the vicinity of the well. Complete sets of aquifer test data (specific capacity, duration of the test, and well radius) were found for 63 wells in 7 counties of the 11-county study area (Coker and others, 1992). Data were found for an additional 7 wells that are less than 2.25 km (1.4 mi) to the west of the model area, just west of the Deaf Smith–Randall county line. Figure 8 illustrates the spatial distribution of wells for which usable specific capacity test data were available. Storativity was set at 0.15 for all of the tests, based on several published literature values. The sensitivity of the model to changes in storativity is evaluated later in this report.

The method for converting the aquifer test data into an estimate of hydraulic conductivity is based on the Theis (1935) solution for the dynamic response of water level of areas in and near a pumped well, and for pumping times of at least several hours. Theis (1963) simplified his original formula to the following (presented in the original units):

$$T = \frac{114.6Q}{D}\left[-0.577 - \ln\left(\frac{1.87r^2S}{Tt}\right)\right] \qquad (1)$$

where

T = aquifer transmissivity (gallons / ft - day)
Q = well discharge (gallons / minute)
D = drawdown in well (ft)
r = radius of well (ft)
S = aquifer storage coefficient (dimensionless)
t = elapsed time of pumping (days)

Aquifer transmissivity, as expressed above, is related to the hydraulic conductivity by the following formula:

$$K = 0.0124\frac{T}{h} \qquad (2)$$

where

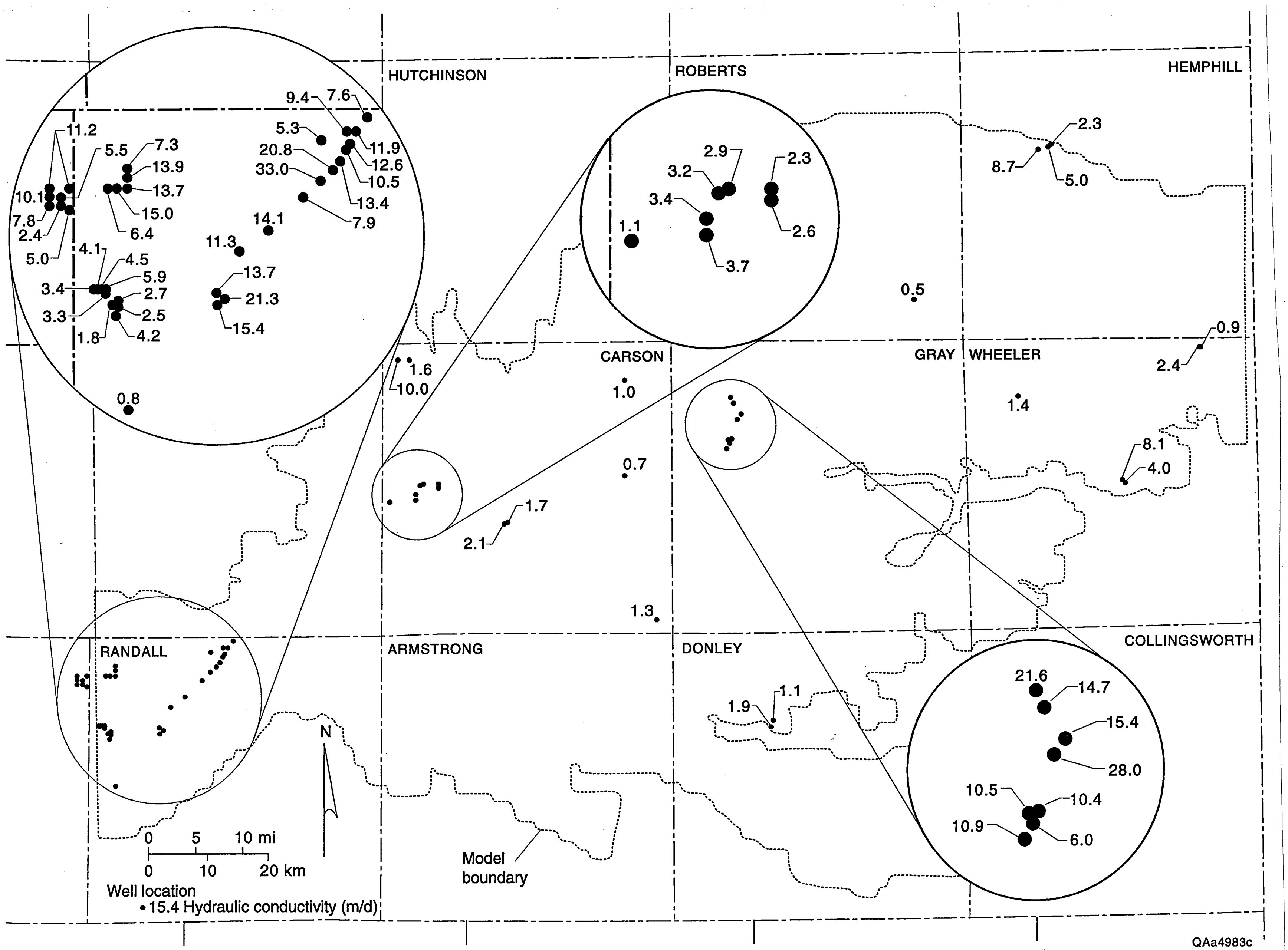

Figure 8. Map illustrating distribution and values of hydraulic conductivity derived from specific capacity tests.

K = hydraulic conductivity $(\mathrm{m\,d^{-1}})$
h = saturated thickness of aquifer (m)

Equation 1 reveals that aquifer transmissivity appears on both sides of the equality. This type of equation cannot be solved directly for T but can be solved through an iterative procedure. First, equation 1 is transformed into the equation

$$f(T) = \frac{114.6Q}{D}\left[-0.577 - \ln\left(\frac{1.87r^2S}{Tt}\right)\right] \tag{3}$$

The secant method can be employed to find the zero solution of equation 3, that is, the value T such that $f(T) = 0$. This is equivalent to finding the value of T in equation 1 such that the equality holds (Moursund and Duris, 1967). As illustrated in figure 9, the function $f(T)$ is a continuous curve passing through a point T where $f(T) = 0$. The secant method is initiated by making two beginning approximations for the solution of $f(T) = 0$, T_1 and T_2. These two initial approximations are then used with equation 3 to find $f(T_1)$ and $f(T_2)$. These coordinate pairs $[T_1, f(T_1)]$ and $[T_2, f(T_2)]$, lying on the $f(T)$ curve, are then used to calculate a secant approximating the slope of the curve. The intersection of this secant line and the horizontal axis yields a third approximate solution for equation 3, T_3. This new approximation is substituted into equation 3 to find another point on the curve, $f(T_3)$. With this latest coordinate pair $[T_3, f(T_3)]$ and the previous pair $[T_2, f(T_2)]$ another secant line can be constructed. These steps are repeated iteratively until the process converges to the value of T, which satisfies equation 3 and therefore also satisfies equation 1.

To carry out the above procedure it was necessary to specify an approximate value for the storage coefficient. This was done with figure 36 in Knowles and others (1984a), which is a map of aquifer specific yield (storage for an unconfined aquifer) in the model area. From this figure, a representative value of 0.15 was selected.

After T was found for each well, it was then necessary to compute the hydraulic conductivity using equation 2. The saturated thickness for each well was determined as the depth to the base of the aquifer minus the reported initial depth to the water table before the test. Table 1 presents a

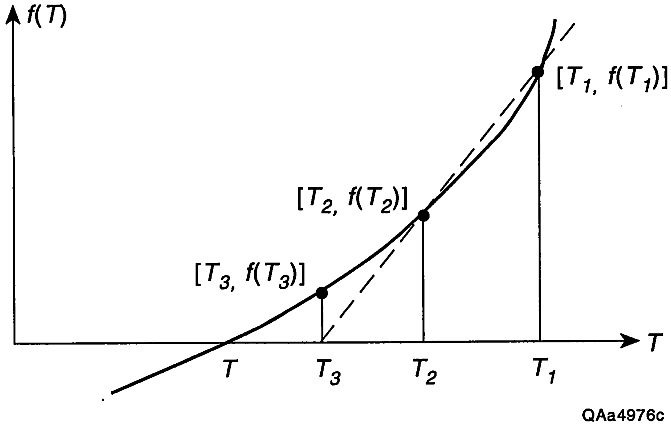

Figure 9. Illustration of the secant method (modified from Moursund and Duris, 1967). T_1, T_2, and T_3 represent iterative estimates of transmissivity ($m^2\ d^{-1}$) used to obtain a final solution for T such that $f(T) = 0$ and the Theis equation is satisfied.

Table 1. Summary of well test data and derived hydraulic conductivity values used in model.

County	ID no	Location of well Latitude	Location of well Longitude	Well radius (ft)	Test pump rate (gpm)	Drawdown (ft)	Depth to base of aquifer (ft)	Depth to initial water table (ft)	Duration of test (day)	Calculated aquifer transmissivity (gal/ft-day)	Calculated hydraulic conductivity (m/day)
Carson	06-36-707	35-22-59	101-36-25	0.67	825.0	105.0	800.0	439.0	1.25	9834.	1.1
Carson	06-36-812	35-23-04	101-33-33	0.67	1500.0	59.0	760.0	380.0	1.00	34858.	3.7
Carson	06-36-813	35-23-33	101-33-34	0.67	1500.0	64.0	762.0	377.0	1.00	31896.	3.4
Carson	06-36-819	35-24-26	101-32-41	0.67	1500.0	61.0	845.0	376.0	1.00	33612.	2.9
Carson	06-36-820	35-24-18	101-33-03	0.67	1594.0	113.0	708.0	460.0	2.08	19596.	3.2
Carson	06-36-920	35-24-10	101-31-10	0.67	1705.0	102.0	814.0	460.0	1.25	22508.	2.6
Carson	06-36-921	35-24-30	101-31-10	0.67	1744.0	114.0	812.0	455.0	1.17	20297.	2.3
Carson	06-56-409	35-12-09	101-06-49	0.67	500.0	70.0	570.0	302.0	0.92	8574.	1.3
Carson	06-28-104	35-35-53	101-35-30	0.50	800.0	127.0	350.0	173.0	1.00	43315.	10.0
Carson	06-28-208	35-35-44	101-34-22	0.67	600.0	52.0	534.0	160.0	0.79	14336.	1.6
Carson	06-45-311	35-21-02	101-23-22	0.67	970.0	93.0	695.0	372.0	1.50	13666.	1.7
Carson	06-45-315	35-20-58	101-23-40	0.67	550.0	47.0	692.0	408.0	1.00	14905.	2.1
Carson	06-31-508	35-34-02	101-10-24	0.50	200.0	68.0	398.0	286.5	0.17	2769.	1.0
Carson	06-39-507	35-25-16	101-10-13	0.50	430.0	63.0	814.0	318.0	0.50	8127.	0.7
Donley	05-57-613	35-02-54	100-53-54	0.50	500.0	147.0	180.0	18.0	3.00	4520.	1.1
Donley	05-57-910	35-02-16	100-54-08	0.33	140.0	110.0	126.0	94.0	1.00	1487.	1.9
Gray	05-25-707	35-30-22	100-57-39	0.67	1000.0	20.0	468.0	362.0	1.00	72770.	28.0
Gray	05-25-709	35-32-15	100-58-24	0.75	460.0	22.0	412.0	358.0	1.50	28601.	21.6
Gray	05-25-710	35-31-47	100-58-06	0.67	600.0	23.0	450.0	347.0	1.50	37171.	14.7
Gray	05-25-803	35-30-54	100-57-19	0.67	1056.0	36.0	471.0	363.0	1.00	40742.	15.4
Gray	05-33-103	35-28-29	100-58-42	0.67	730.0	45.0	402.0	319.0	1.00	21328.	10.5
Gray	05-33-104	35-28-16	100-58-35	0.67	560.0	50.0	406.0	310.0	1.00	14204.	6.0
Gray	05-33-106	35-27-36	100-58-40	0.67	1080.0	44.0	447.0	322.0	1.00	33545.	10.9
Gray	05-33-109	35-28-34	100-58-16	0.67	900.0	49.0	413.0	317.0	1.00	24435.	10.4
Hemphill	05-05-603	35-55-01	100-22-31	0.50	150.0	49.0	121.0	57.0	1.00	3603.	2.3
Hemphill	05-05-909	35-54-35	100-23-52	0.50	750.0	40.0	128.0	5.0	1.00	26342.	8.7
Hemphill	05-06-704	35-54-47	100-22-49	0.50	450.0	45.0	186.0	77.0	1.00	13263.	5.0
Randall	06-49-601	35-10-36	101-53-47	0.67	700.0	50.0	267.0	170.0	1.00	18147.	7.6
Randall	06-49-803	35-08-15	101-55-12	0.67	880.0	36.0	258.0	150.0	2.00	35508.	13.4
Randall	06-49-805	35-07-56	101-55-33	0.67	1100.0	33.0	233.0	141.0	1.00	46830.	20.8
Randall	06-49-808	35-09-15	101-56-14	0.67	620.0	50.0	286.0	165.0	1.00	15884.	5.3
Randall	06-49-909	35-09-40	101-54-20	0.75	750.0	42.0	267.0	162.0	30.0	30702.	11.9
Randall	06-49-911	35-09-06	101-54-39	0.75	750.0	37.0	270.0	156.0	30.0	35166.	12.6

Table 1. Cont.

County	ID no	Location of well Latitude	Longitude	Well radius (ft)	Test pump rate (gpm)	Drawdown (ft)	Depth to base of aquifer (ft)	Depth to initial water table (ft)	Duration of test (day)	Calculated aquifer transmissivity (gal/ft-day)	Calculated hydraulic conductivity (m/day)
Randall	06-49-912	35-08-50	101-54-52	0.75	750.0	46.0	270.0	162.0	30.0	27850.	10.5
Randall	06-49-913	35-09-40	101-54-39	0.75	750.0	44.0	289.0	163.0	30.0	29209.	9.4
Randall	06-57-102	35-05-14	101-59-13	0.67	1340.0	44.0	256.0	126.0	2.00	45071.	14.1
Randall	06-57-206	35-07-28	101-56-17	0.67	1090.0	18.0	254.0	137.0	2.00	94777.	2.5
Randall	06-57-212	35-06-43	101-57-21	0.67	720.0	49.0	236.0	137.0	1.08	19284.	7.9
Randall	07-56-703	35-07-51	102-06-52	0.67	1025.0	50.0	292.0	138.0	1.00	27554.	7.3
Deaf Smith	07-63-203	35-05-59	102-10-05	0.67	433.0	49.0	289.0	105.0	1.00	10942.	2.4
Deaf Smith	07-63-204	35-06-08	102-10-34	0.67	760.0	49.0	253.0	86.0	1.00	20305.	5.0
Deaf Smith	07-63-205	35-06-34	102-10-34	0.67	961.0	50.0	304.0	115.0	1.00	25679.	5.5
Deaf Smith	07-63-207	35-07-00	102-10-05	0.67	1150.0	32.0	320.0	136.0	1.00	50825.	11.2
Deaf Smith	07-63-208	35-06-10	102-11-05	0.67	1268.0	50.0	283.0	102.0	1.00	34762.	7.8
Deaf Smith	07-63-209	35-06-34	102-11-06	0.67	1218.0	41.0	298.0	131.0	1.00	41309.	10.1
Deaf Smith	07-63-210	35-06-57	102-11-06	0.67	1130.0	37.0	295.0	140.0	1.00	42573.	11.2
Randall	07-63-302	35-06-59	102-07-56	0.67	930.0	52.0	300.0	148.0	1.00	23733.	6.4
Randall	07-64-103	35-07-28	102-06-52	0.67	1075.0	33.0	282.0	148.0	1.00	45672.	13.9
Randall	07-64-104	35-07-01	102-06-52	0.67	975.0	34.0	267.0	149.0	1.00	39749.	13.7
Randall	07-64-136	35-07-02	102-07-26	0.67	1200.0	32.0	294.0	149.0	1.00	53233.	15.0
Randall	07-64-605	35-04-22	102-00-45	0.67	1115.0	35.0	264.0	103.0	1.00	44576.	11.3
Randall	07-64-912	35-02-12	102-01-31	0.42	1280.0	50.0	200.0	127.0	1.00	38119.	21.3
Randall	07-64-915	35-02-24	102-02-01	0.42	940.0	26.0	200.0	35.0	1.00	55382.	13.7
Randall	07-64-916	35-01-54	102-01-53	0.42	910.0	31.0	205.0	88.0	1.00	44209.	15.4
Randall	07-63-623	35-02-34	102-08-08	0.67	508.0	38.0	242.0	122.0	1.00	17251.	5.9
Randall	07-63-624	35-02-33	102-08-21	0.67	508.0	51.0	221.0	100.5	1.67	13125.	4.5
Randall	07-63-625	35-02-33	102-08-33	0.67	400.0	50.0	237.0	140.0	1.00	9806.	4.1
Randall	07-63-626	35-02-33	102-08-48	0.67	400.0	58.0	256.0	157.0	1.00	8324.	3.4
Randall	07-63-925	35-01-51	102-07-52	0.50	240.0	65.0	243.0	142.0	1.00	4433.	1.8
Randall	07-63-926	35-02-24	102-08-10	0.67	450.0	75.0	238.0	145.0	1.50	7444.	3.3
Randall	07-63-927	35-01-22	102-07-41	0.50	400.0	55.0	162.0	65.0	2.00	9987.	4.2
Randall	07-63-928	35-02-03	102-07-36	0.50	200.0	41.0	205.0	107.0	2.54	6601.	2.7
Randall	07-63-929	35-01-45	102-07-37	0.50	280.0	59.0	161.0	66.0	1.00	5849.	2.5
Randall	10-08-418	34-57-09	102-06-52	0.50	100.0	52.0	224.0	114.0	1.00	2149.	0.8
Robertson	05-19-608	35-41-08	100-38-02	0.50	350.0	149.0	505.0	270.0	1.50	2806.	0.5
Wheeler	05-32-108	35-36-23	100-06-05	0.25	60.0	10.0	231.0	125.0	0.04	6203.	2.4

Table 1. Cont.

County	ID no	Location of well Latitude	Location of well Longitude	Well radius (ft)	Test pump rate (gpm)	Drawdown (ft)	Depth to base of aquifer (ft)	Depth to initial water table (ft)	Duration of test (day)	Calculated aquifer transmissivity (gal/ft-day)	Calculated hydraulic conductivity (m/day)
Wheeler	05-32-109	35-36-23	100-06-05	0.25	60.0	20.0	246.0	110.0	0.04	2832.	0.9
Wheeler	05-29-820	35-32-14	100-26-28	0.50	250.0	52.0	250.0	64.0	1.79	6286.	1.4
Wheeler	05-39-706	35-24-14	100-14-55	0.75	300.0	36.0	136.0	85.7	1.00	10009.	8.1
Wheeler	05-39-711	35-24-01	100-14-41	0.33	75.0	24.0	75.0	34.0	1.00	4006.	4.0

summary of the well data and the derived hydraulic conductivities for the 70 wells used in the study. As to whether or not each well is fully penetrating the aquifer is unknown.

Although 70 sets of data were located for the study area, all available data were from municipal wells and were highly clustered geographically (fig. 8). This resulted in large sections of the study area having little or no coverage. Additionally, for the areas having a high concentration of well data, there was a very high degree of variability in the derived hydraulic conductivity over relatively small distances, as shown in figure 8. Because of these problems, a statistical averaging approach was used before the hydraulic conductivities were interpolated to non-covered areas.

The high variation in horizontal hydraulic conductivity, as derived from well data, may be due partially to variability in the quality of the well completion, well loss at high pumping rates, or to data reporting errors. However, other studies of the Ogallala Formation have found it to be composed of layers with highly variable hydrologic properties. The stratigraphy ranges from layers of low-permeability cemented calcretes and caliches to highly permeable sands and gravels (Seni, 1980; Gustavson and Winkler, 1988; Gustavson, 1995). Ashworth (1980, 1984) used a core analysis procedure from 41 test wells to evaluate vertical relationships in horizontal hydraulic conductivity and reported that the vertical variability in hydrologic properties was randomly distributed.

Other studies of heterogeneous aquifers have found that horizontal hydraulic conductivity values are distributed in a log-normal fashion (for example, Prudic, 1991). For such distributions, a geometric mean is the best representation of an average value for the variability in hydraulic conductivity. Therefore, the wells used to derive the hydraulic conductivity were first clustered into spatially distinct subsets and then the distribution of hydraulic conductivity within each subset was examined. Data from 37 clustered wells in western Randall and extreme eastern Deaf Smith Counties fall in a log-normal distribution (fig. 10), with a geometric mean hydraulic conductivity of 7.4 m d^{-1} (24.2 ft d^{-1}). This procedure of replacing tightly clustered values with their geometric mean was repeated to arrive at 14 representative data points, which were used to extrapolate to non-covered areas using a computer-based gridding procedure and to create a map of hydraulic

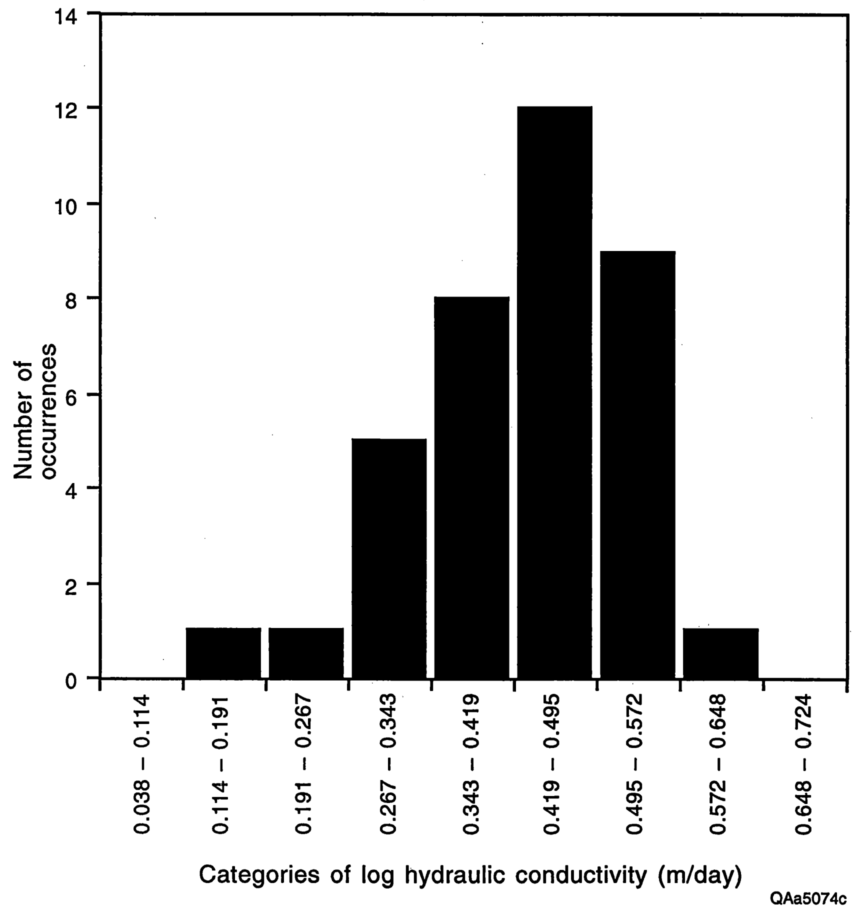

Figure 10. Distribution in measured hydraulic conductivity for wells producing from the Ogallala aquifer in Randall County.

conductivity (fig. 11). Then, in a procedure analogous to the steps used for the base of the aquifer and the 1959–60 water-table elevation, each cell of the model grid was assigned a hydraulic conductivity based on this map.

Subsequent to the completion of modeling efforts, a data set consisting of results from 11 pump tests from wells in the ACCWF were obtained. These pump tests were analyzed using Neuman's method for unconfined aquifers (Neuman, 1975). The mean hydraulic conductivity for these 11 pump tests was 3.65 m d^{-1} (11.99 ft d^{-1}) with a standard deviation of 1.19 m d^{-1} (3.9 ft d^{-1}) (Xiang, in prep.). Because these values compare very well with the values derived from specific capacity tests used during model calibration, no attempt was made to rerun the model using the additional data.

Lateral Boundary Conditions

Ground-water models require the specification of boundary conditions, which includes the conceptualization of, and mathematical expressions for, the connections between the modeled area and the exterior. Anderson and Woessner (1992) presented three types of boundary conditions:

(1) *specified head*—a fixed value of hydraulic head is applied to a given cell,

(2) *specified flow*—the volumetric flow rate for a given cell is fixed, and

(3) *head-dependent flow* or *mixed flow*—flow across the boundary is calculated within the model on the basis of a fixed boundary hydraulic head value.

Head-dependent boundaries were used extensively for this model. For such boundary cells, two parameters must be specified: the head on the boundary, h_b, and the hydraulic conductance, C_b, between the boundary and the node at the center of the cell. During the model simulation, MODFLOW calculates the flow of water across each boundary cell as

$$Q_{b=C_b}(h_b - h_n) \qquad (4)$$

where

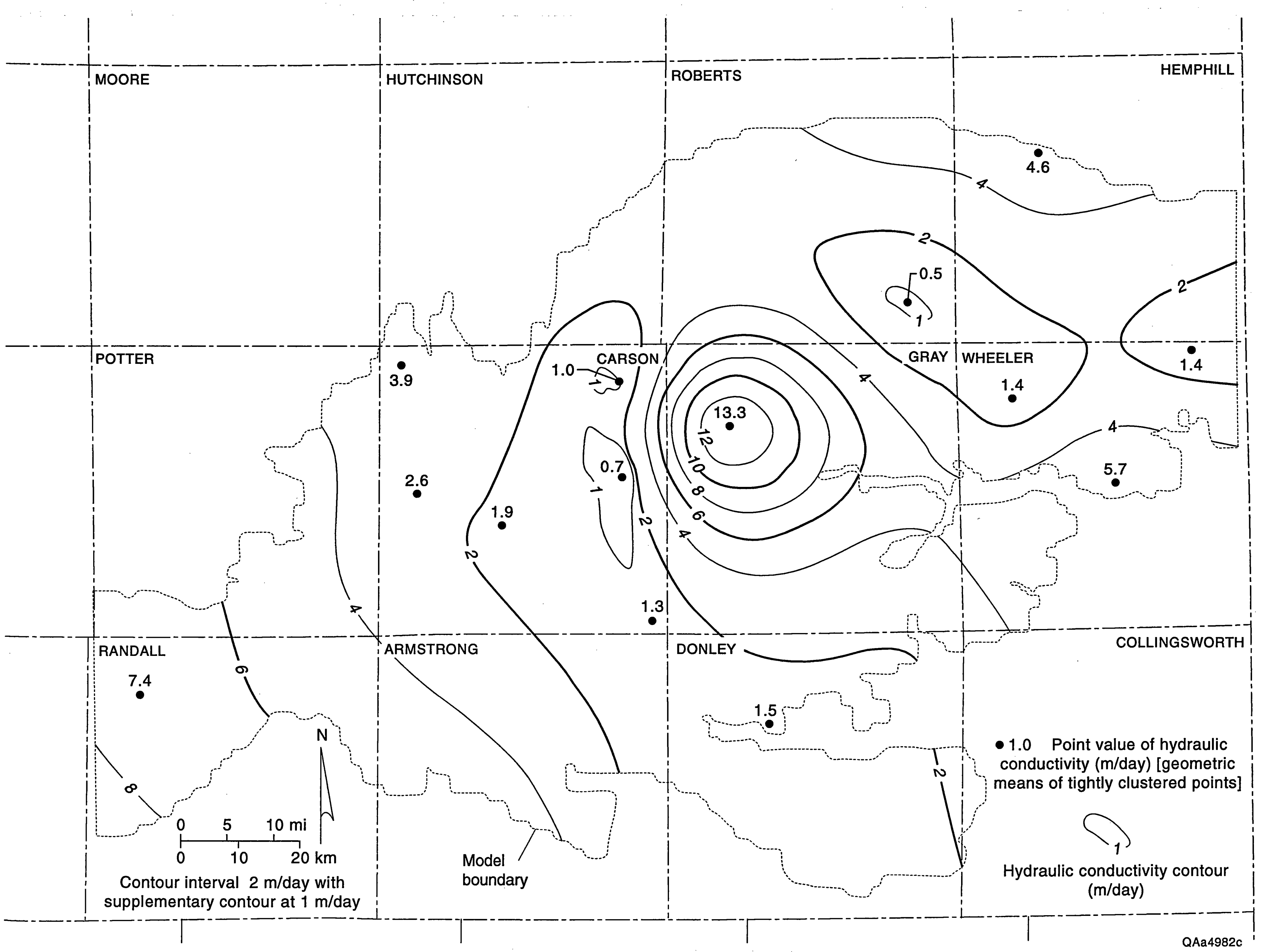

Figure 11. Hydraulic conductivity input data set used during model simulations.

Q_b = water flow across boundary of cell $(m^3\, d^{-1})$

C_b = hydraulic conductance between boundary edge and cell node $(m^2\, d^{-1})$

h_b = pressure head on boundary edge of cell (m)

h_n = pressure head at node in center of the cell (m)

Three variations or subtypes of head-dependent boundary conditions were applied to a total of 1,099 cells along the lateral boundary of the model grid.

(a) **Drain-type** boundary cells were used to simulate boundary conditions for 933 cells, which compose most of the perimeter of the model. These cells were used to represent the boundary conditions for all edges except the extreme northern edge, where the Canadian River contacts the Ogallala Formation, and along the extreme eastern and western boundaries, where the modeled area is joined to other parts of the Ogallala aquifer. Drain-type cells allow water to flow only out of the aquifer.

The drain-type cells were used to simulate the natural hydrologic conditions where water within the Ogallala aquifer moves toward the edges of the High Plains escarpment and outcropping edges of the Ogallala Formation and discharges. Prior to extensive development of ground-water resources in the study area, numerous springs and seeps of varying discharge were located throughout the escarpment area, which separates the High Plains and the Rolling Plains in the Texas Panhandle. As part of this study, an inspection of the 7.5-minute topographic maps covering boundary areas of this model, including the Canadian River Valley and the escarpment separating the High Plains from the Rolling Plains, and a review of Brune (1981), documented approximately 210 springs and seeps along the boundary. Gould (1906) reported the presence of "thousands" of springs in the 11 counties covered in our study. He provided the following description of the typical Ogallala aquifer spring (text in brackets added as explanation):

> Not infrequently springs occur at the line of contact between the Tertiary deposits [the Ogallala Formation] and the clay strata of the upper part of the red beds [strata underlying the Ogallala Formation]. This condition is due to the ready seepage of water through the Tertiary sands to the top of the impervious red beds, where it flows laterally until it reaches the surface. Many of these contact springs do not issue from a single opening but the water finds its escape along a zone of seepage extending sometimes for hundreds of yards along the side of a cliff.

As a result of the significant lowering of water levels in the Ogallala aquifer because of production, almost all of these springs dried up before any discharge measurements were collected. Our study approximated the springs and seeps as a long linear drain having a comparatively short vertical extent. In order to implement this, the parameters C_b and h_b of equation 4, the hydraulic conductance and boundary head, respectively, had to be assigned for each of the 933 boundary nodes around the perimeter of the model. On the basis of the description of Gould (1906), the value of h_b was approximated for each cell situated along the boundary of the model as a constant water level equal to 3 cm (0.1 ft) above the base of the aquifer.

During model calibration, the effective saturated thickness between the boundary and the cell center was adjusted for two reasons. Within MODFLOW the cell-to-cell conductance terms for interior parts of the model change with saturated thickness. However, no such adjustment is made for the boundary conductance term, C_b, with change in hydraulic head and saturated thickness. Therefore, a separate calibration is needed to adjust the initial estimate for C_b to match final water-table positions. Second, estimated saturated thickness is highly variable near the model edge. Knowles and others (1984a, b) indicated that saturated thicknesses generally range from 6 to 12 m (20 to 40 ft) at distances of 1,000 to 3,000 m from the outcrop edge, although in some extreme cases values range from 30 to 90 m (100 to 300 ft).

(b) **General head-dependent flow** boundaries were used for the 94 cells at the eastern and western edges of the model area that are hydrologically connected to other parts of the Ogallala aquifer. The saturated thickness at the north-east boundary was based on the elevation difference between the 1959–60 water table and the base of the aquifer. For the eastern and western edges, the head on the boundary was set to the 1959–60 water-table elevation.

(c) **River** boundary conditions were assigned to 72 cells where the Canadian River is in contact with the Ogallala Formation. This occurs along an approximately 108-km (67.1-mi) long reach of river that begins about 11 km (6.8 mi) upriver from the Hutchinson–Roberts County boundary, stretches through Roberts County, and ends approximately 16 km (9.9 mi) upstream of the Texas–Oklahoma border in Hemphill County, Texas. Along this part of the river, the Ogallala

Formation is particularly thick and extends under the river to the north. Upstream and downstream of this reach, in areas where the base of the Ogallala is above the river and exposed in banks of the river valley, drain-type cells were used.

The river boundary cells simulate head-dependent effluent and influent river reaches. When the hydraulic head in the aquifer exceeds that in the river, water moves from the aquifer to the river; water moves to the aquifer under reverse conditions. The outflow to the stream is proportional to the pressure head differential. The inflow process to the aquifer, however, has two stages; it is proportional to the head difference until the aquifer head falls below the river bottom, where it then achieves a constant value. To implement this approach, each cell requires essentially the same parameters as above: the hydraulic conductance term, modified to be an effective conductance of the stream-aquifer interconnection; the river water elevation, plus the specification of a river bottom elevation.

The hydraulic conductance term for such an aquifer-river connection is poorly known. The approach used for our study was to calculate a representative conductance based on known differences between water levels and inflow or outflow measurements as described by McDonald and Harbaugh (1988).

Luckey and others (1986) examined winter river flow records from 1938 to 1950 and found that the Ogallala aquifer contributes approximately 1.09×10^5 m^3 d^{-1} (3.86×10^6 ft^3 d^{-1} or 45 ft^3 s^{-1}) to the Canadian River between Amarillo and Canadian, Texas; that is, on average the Canadian is an influent stream in this reach. All of this flow was initially assumed to originate from the southern side of the Ogallala aquifer uniformly over the approximately 108 km (67.1 mi) long reach. For example, a typical grid cell 1,609 m (1.0 mi) in length was assumed to contribute approximately 1,628 m^3 d^{-1} (57,492 ft^3 d^{-1}) of flow. Inspection of the 1959–60 water-table map of the Ogallala aquifer reveals an approximate head gradient of from 8 to 15 m (25 to 50 ft) in hydraulic head per each 1.6 km (1.0 mi) of lateral distance in the vicinity of the Canadian River. The effective conductance can be calculated by using a median head drop of 12 m (39.4 ft) and assuming the effective distance from the cell node at the center of the block to the center of the river channel is about 1.6 km (1.0 mi). A flow rate of 1,628 m^3 d^{-1} (57,492 ft^3 d^{-1}) and a head drop of 12 m (39.4 ft) thus yields an average conductance of 135.7 m^2 d^{-1} (1460 ft^2 d^{-1}) for a cell of length 1,609 m (1.0 mi). The average conductivity in the aquifer for this region is approximately 4.5 m d^{-1} (15 ft d^{-1}), with some variation. Therefore, in addition to the average conductance value of 135.7 m^2 d^{-1} (1460.7 ft^2 d^{-1}) being adjusted for smaller grid cells, it was also adjusted on a cell-by-cell basis by the ratio of hydraulic conductivity of that cell to this average value of 4.5 m d^{-1} (14.8 ft d^{-1}). The quantification of Ogallala aquifer discharge to the Canadian River from the north was not determined since the model did not extend north of the river.

Recharge

All discharge from the model was incorporated into the various types of boundary conditions and wells, depending on the stage of modeling. The primary inflow was recharge applied as a flux to the upper face of specified grid cells.

In the predevelopment stage of modeling, three recharge distributions were evaluated. First, recharge was applied *uniformly* over the entire model area, with fluxes ranging from 6 to 9 mm yr^{-1} (0.236 to 0.354 inch yr^{-1}). This range was based on the recharge rates calculated to maintain local perched aquifers (Mullican and others, 1993a). The second type of recharge was based on the regionally mapped geologic units present at the surface. Under this *zonal* recharge scenario, fluxes were in the same range as before, but the recharge on the fine-grained Blackwater Draw Formation surface was reduced relative to that on the Ogallala Formation outcrop area (fig. 2). Third, a *playa-focused* recharge scenario was used. Under this scenario the *zonal* approach was modified to spatially restrict and focus the equivalent recharge of the entire Blackwater Draw Formation surface through the playas and to eliminate recharge to the nonplaya or interplaya areas. In this part of the Southern High Plains, playas are restricted to areas where the Blackwater Draw Formation is present at the surface. Therefore, recharge to the Ogallala Formation was held at the same range in fluxes and method as that used for the *zonal* scenario, and cells in the area of the model coinciding with playas within the Blackwater Draw Formation received an "equivalent recharge," based on equation 5:

$$ER = \frac{(IP + P)}{P} R \tag{5}$$

where

$ER =$ Equivalent (effective) recharge (mm yr^{-1})

$IP =$ Surface area of interplaya area (for this model area, $IP = 4.843 \; x \; 10^9 m^2$)

$P =$ Surface area of playas (for this model area, $P = 1.37x10^8 m^2$)

$R =$ Regionally averaged recharge rate (mm yr^{-1})

These three distributions (*uniform*, *zonal*, and *playa-focused* [modified zonal]) were evaluated in light of recent evidence that most of the recharge on the Blackwater Draw Formation is concentrated through the playas and little exists in interplaya areas (Gustavson and others, 1993; Scanlon and others, 1993, 1994; Scanlon, 1995). This recharge evaluation was performed under the predevelopment stage of modeling.

Transient Model Development

The second stage of modeling used only the *playa-focused* scenario. The simulations of this second stage, covering the period 1960–90, served to verify the ground-water model's soundness under the assumptions of the *playa-focused* scenario. Also, in the second stage of modeling, the focus was shifted from evaluating recharge scenarios to studying the hydrologic relationship between the ACCWF and the changes in the water table in the vicinity of the DOE Pantex Plant. For transient modeling it was necessary to specify the specific storage coefficient, initial conditions, and estimated effective porosities for the area.

For the transient stage of modeling, the performance of the simulations was assessed by a comparison to the "known" water-table surface at the end of the 1960–90 period. The 1990 water-table surface was derived by utilizing the averages of the available water-level measurements in many wells in the area over the period 1980–91 as found in the TNRIS database and data provided by the City of Amarillo. This 1990 water table is depicted in figure 7. One notable feature of this surface is the approximate 61 m (200 ft) "mound" visible in the extreme north of Donley County. This mound appears to be an anomaly. Wells in the area of the mound may not have been completed in the main Ogallala aquifer but instead, in locally perched aquifers above the Ogallala aquifer. In Carson County for example, multiple areas of localized perched aquifers have been identified above the Ogallala aquifer (Mullican and others, 1993a,b, 1994c).

Time periods

For the transient stage of modeling, the 31-year period of interest (1960–90) was subdivided into 6 multiyear subperiods. This was necessary principally because of the rapid changes over the 31-year period in the number of producing wells and the withdrawal rate per well. To facilitate the use of available county-by-county data for ground-water production data from the Ogallala aquifer as presented in Knowles and others (1984a), the same subperiods were used for 1960–79, namely subperiods 1960–64, 1965–69, 1970–74, and 1975–79. For the period 1980–90 two additional subperiods were used: 1980–85 and 1986–90. In addition to being a necessity for the well data input, using subperiods provides intermediate output to track the performance of the model results over the transient simulation.

Initial Condition

For the transient simulations the initial condition for the model area was based on output from the steady-state simulation using the playa-focused recharge package.

Withdrawals through Wells

Due to the lack of data, the simulation of withdrawal of water from the Ogallala aquifer through wells is not straightforward. For many parts of the study area the number and location of active wells is poorly known. C. E. Williams (pers. comm., 1994), estimates that only 10–15 percent of irrigation wells are recorded in databases such as the TNRIS data base. Furthermore, the volume of water withdrawn on a per well, per county, or regional basis is also unknown. Fortunately, good records of well locations and total well field withdrawals for much of the 1960–90 period are available for the ACCWF.

Because of the uncertainty in well production data outside the ACCWF, several combinations of locations and withdrawal rates per well were examined during transient simulations. All other aquifer properties were maintained as in the final calibrated predevelopment model.

The amount of water withdrawn from the Ogallala aquifer in the study area for the 1960–90 period was an important parameter during transient simulations. The basic source of ground-water withdrawal data is Knowles and others (1984a, their table 10) covering the time period 1960–79. For the period 1980–90, unpublished county-by-county withdrawal estimates were obtained from the Texas Water Development Board. The withdrawals from both sources were reported on a total county basis. Since several counties of the study area are not wholly within the modeled area, it was necessary to proportion their withdrawal estimates on the basis of area ratios.

Another type of information required was yearly pumpage totals for the ACCWF as shown in table 2. Although these figures were reported on a fiscal-year basis, no attempt was made to add and subtract appropriate partial year amounts because they were utilized in the model on a calendar-year basis. Since the rates represent a more or less constantly increasing trend, this is an acceptable simplification. As evident in table 2, the data on withdrawals are only available for the period 1972–92. For periods of modeling covering 1960–71, the withdrawal rate for the ACCWF was estimated with the average rate for 1973–74. With these considerations, a set of withdrawals on a county and period basis was derived for the 11 counties of the study area as presented in table 3. Also given in table 3 are the reported withdrawals from the ACCWF and an estimated withdrawal rate for the production wells on the Pantex Plant of 0.7×10^6 m^3 yr^{-1} (500,000 gal d^{-1}).

Three primary data sources were used for this information: the Texas Natural Resource Information System (TNRIS), which maintains a statewide data base on water wells, the City of Amarillo's records regarding their ACCWF, and the U.S. Army Corp of Engineers (1992)

Table 2. Summary of Withdrawals from Amarillo's Carson County Well Field.

Fiscal year	Total yearly withdrawals 10^6 gall/yr	10^6 m^3/yr	Averages 10^6 m^3/yr
1991-92	5088	19.26	
1990-91	5764	21.82	
1989-90	5782	21.89	
1988-89	4400	16.66	avg 86-90
1987-88	4901	18.55	18.5
1986-87	4818	18.24	
1985-86	4599	17.41	
1984-85	3003	11.37	
1983-84	2853	10.80	
1982-83	2886	10.93	avg 80-85
1981-82	2556	9.68	11.2
1980-81	3007	11.38	
1979-80	3426	12.97	
1978-79	2941	11.13	
1977-78	3008	11.39	avg 75-79
1976-77	2864	10.84	10.9
1975-76	2920	11.05	
1974-75	2603	9.85	
1973-74	2808	10.63	avg 73-74
1972-73	1199	4.54	7.6

Table 3. Summary of withdrawals from the modeled portion of the Ogallala aquifer for the 11-county study area over the period 1960-1990.

County	Annual withdrawals by period (10^6 m^3/year) 1960-64	1965-69	1970-74	1975-79	1980-85	1986-90	Totals† 1960-90 (10^6 m^3)
Armstrong	25.8	26.5	23.2	12.8	13.6	12.5	585.3
Carson - total (Amarillo Wellfield) (Pantex Plant)	189.6 (7.6) (0.7)	221.8 (7.6) (0.7)	230.9 (7.6) (0.7)	188.5 (10.9) (0.7)	164.7 (11.2) (0.7)	121.7 (18.5) (0.7)	5750.5 (328.8) (21.4)
Collingsworth	1.6	1.6	1.6	1.6	1.6	1.6	49.7
Donley	28.2	24.3	47.5	20.0	11.2	11.7	726.0
Gray	40.8	57.5	64.0	43.2	41.4	38.2	1467.3
Hemphill	2.2	2.7	5.4	9.0	7.4	6.0	171.5
Hutchinson	21.1	25.9	26.5	23.6	24.9	20.5	737.3
Potter	10.4	10.7	10.7	6.3	15.9	13.3	352.7
Randall	67.6	43.7	40.5	42.4	48.2	38.5	1452.5
Roberts	8.8	10.0	15.3	14.7	9.3	4.9	323.8
Wheeler	5.4	4.1	9.0	9.7	4.9	3.3	187.5
period total†† (10^6 m^3)	2007.5	2143.8	2373.2	1858.9	2058.9	1361.8	11804.0

† County totals are the sum over the 6 periods of the annual withdrawals for the county × period length. Respective period lengths are 5, 5, 5, 5, 6, and 5 years.

††Period totals are sum over the 11 counties of the annual withdrawals for the county × period length.

showing the location and establishment dates for the production wells on the Pantex Plant. Table 4 shows the total number of wells in the model area extracted from these sources, while fig. 12 illustrates the location of these wells.

Combining the withdrawal data of table 3 with the enumeration of known wells presented in table 4 was done to derive a set of withdrawals per well as presented in table 5. These values should be considered the "default" well withdrawals because evidence suggests that the known wells represent only a fraction of the actual wells in production.

The magnitude of many of these default withdrawals are very high, ranging from 0.04 to 0.9 m^3 min^{-1} (19 to 467 gallon min^{-1}). This can be seen by comparing these rates to the ACCWF withdrawals for which very good data exist, especially in the 1975–1990 time frame. The ACCWF wells are pumped on a year-round basis to meet the demands of a major metropolitan area, whereas most of the wells of the TNRIS data base are irrigation wells, generally used for only a portion of the year. In 1992, the production rate for individual ACCWF wells averaged approximately 1.0 m^3 min^{-1} (262 gallon min^{-1}) annually, while a typical irrigation well produces at a rate of approximately 2.65 m^3 min^{-1} (700 gallon min^{-1}) for roughly 1,500 to 2,000 hours per year (Williams, 1994, pers. comm.), which would be an equivalent yearly rate of from 0.45 to 0.60 m^3 min^{-1} (120–160 gallon min^{-1}).

As will be detailed below in the Results section, these default withdrawals proved to be too high to obtain adequate model performance in the transient simulations. For many wells, these default withdrawals resulted in an unrealistically rapid decline of water levels in the vicinity of the well, which often caused finite-difference cells to go "dry" and become inactive for the remainder of the simulation. A well cell going dry is not implausible, in that it may be the simulated equivalent of a well being taken out of production after precipitous water-level declines. In MODFLOW, however, the remainder of the simulation can be unrealistic if too many of these occur because these cells then remain inactive and the "known" well withdrawal targets will not be met.

Table 4. Summary of known wells in the modeled portion of the Ogallala aquifer for the 11-county study area over the period 1960 - 1990.

County	Number of known wells by period 1960-64	1965-69	1970-74	1975-79	1980-85	1986-90
Armstrong	20	30	43	80	81	81
Carson - total	82	94	105	118	130	131
(Amarillo Wellfield)	30	33	33	33	37	37
(Pantex Plant)	1	2	4	4	5	5
Collingsworth	2	2	2	2	2	2
Donley	16	36	39	74	74	74
Gray	34	76	96	109	113	113
Hemphill	7	11	36	57	67	67
Hutchinson	1	2	4	13	13	13
Potter	5	8	8	11	21	21
Randall	44	51	56	70	86	87
Roberts	11	21	35	61	64	64
Wheeler	35	45	64	83	88	88
period totals	288	411	525	715	781	783

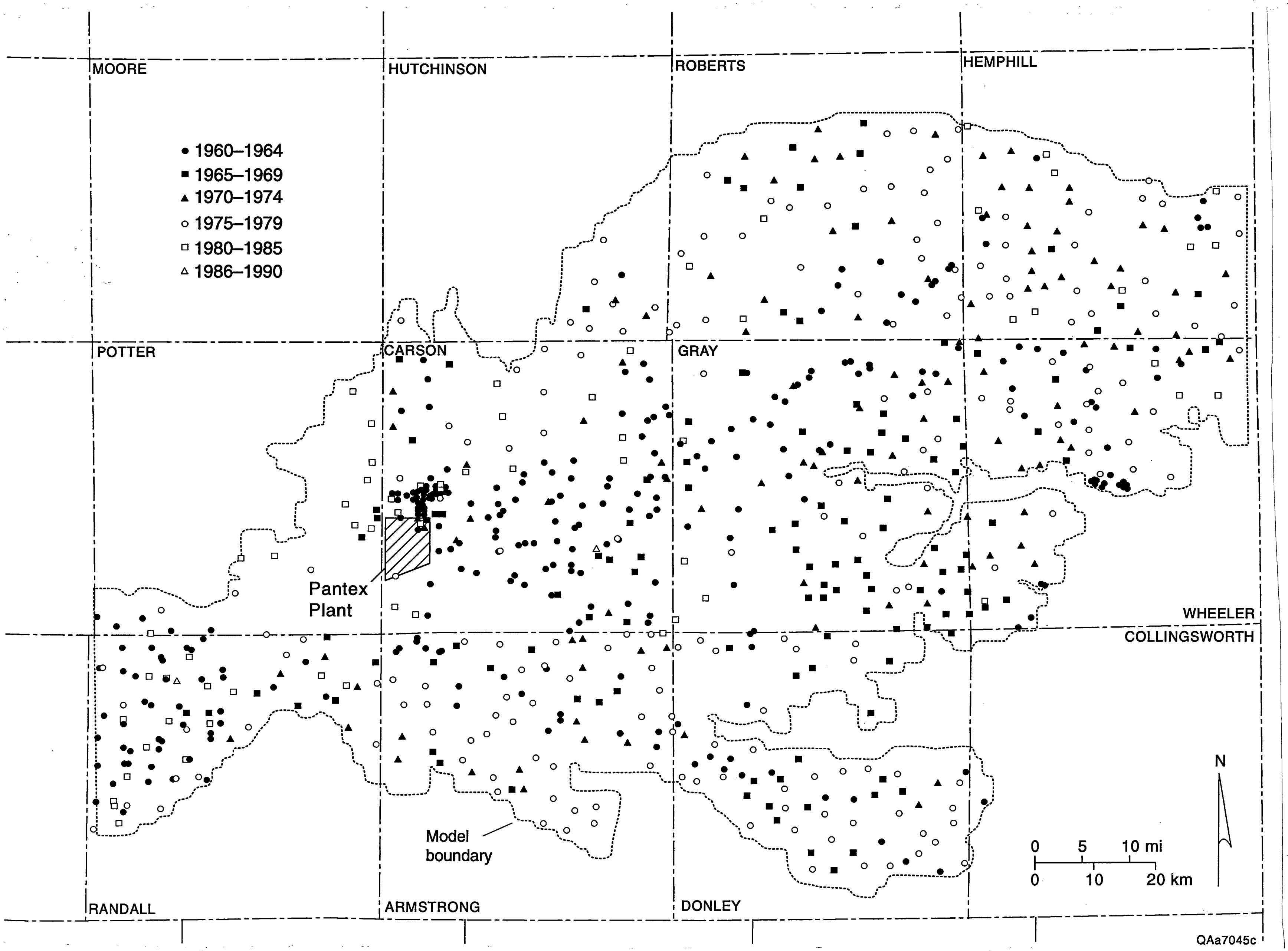

Figure 12. Map showing locations of all known wells for the study area.

Table 5. Summary of default withdrawal rates for the known wells of the modeled portion of the Ogallala aquifer for the period 1960-90.

County	Default withdrawal rates per well by period (106 m3/year & [gallon/min]) 1960-64	1965-69	1970-74	1975-79	1980-85	1986-90
Armstrong	1.289 [647.82]	0.884 [444.28]	0.539 [271.04]	0.160 [80.59]	0.168 [84.19]	0.154 [77.30]
Carson (total)	2.312 [1161.98]	2.359 [1185.77]	2.199 [1105.23]	1.597 [802.75]	1.267 [636.61]	0.929 [467.07]
-Amarillo wellfield	0.255 [128.12]	0.232 [116.47]	0.232 [116.47]	0.329 [165.31]	0.303 [152.47]	0.500 [251.32]
-Pantex wells	0.691 [347.16]	0.345 [173.58]	0.173 [86.79]	0.173 [86.79]	0.138 [69.43]	0.138 [69.43]
Collingsworth	0.802 [402.95]	0.802 [402.95]	0.802 [402.95]	0.802 [402.95]	0.802 [402.95]	0.802 [402.95]
Donley	1.765 [887.26]	0.675 [339.24]	1.218 [611.98]	0.270 [135.71]	0.152 [76.23]	0.158 [79.58]
Gray	1.201 [603.51]	0.756 [80.11]	0.667 [335.15]	0.396 [199.06]	0.367 [184.33]	0.338 [170.07]
Hemphill	0.317 [159.41]	0.247 [123.98]	0.151 [75.77]	0.158 [79.39]	0.110 [55.52]	0.090 [45.34]
Hutchinson	21.093 [10600.68]	12.952 [6509.19]	6.630 [3332.08]	1.812 [910.81]	1.917 [963.26]	1.575 [791.59]
Potter	2.072 [1041.47]	1.341 [674.17]	1.341 [674.17]	0.572 [287.42]	0.758 [380.81]	0.634 [318.82]
Randall	1.536 [772.09]	0.856 [430.30]	0.722 [363.10]	0.606 [304.65]	0.561 [281.85]	0.442 [222.32]
Roberts	0.796 [400.13]	0.476 [239.11]	0.437 [219.63]	0.241 [120.94]	0.145 [72.65]	0.077 [38.75]
Wheeler	0.155 [77.93]	0.090 [45.46]	0.141 [70.71]	0.117 [59.00]	0.056 [28.18]	0.038 [19.02]

To address the problem of the fact that only a small portion (10 to 15 percent) of the actual producing wells are included in available data bases (and thus an unrealistic volume of water must be withdrawn from individual wells to achieve the required total production), in subsequent transient simulations it was necessary to add make-up or phantom wells to distribute the total county withdrawals over more wells. Several combinations of the number of phantom wells and their withdrawal rates were used. Additionally, for areas of the model with low initial (1959–60) saturated thickness and/or low hydraulic conductivity, it was necessary to limit the rate of withdrawal on both known and phantom wells in order to prevent too many cells from going dry during simulation. This was done in a fashion similar to that used for the Ogallala model by Knowles and others (1984a). The discharge from wells in low saturated thickness or low hydraulic conductivity areas was limited by using an equation describing the long-term drawdown in the vicinity of a well (Todd, 1959, as presented in Knowles and others, 1984a). Although not rigorously correct for a transient simulation and for wells which may be interferring with each other, this equation proved to be an adequate means of limiting well discharges to prevent cells from drying. The equation is:

$$Q^* = \frac{\pi}{24} K H_0^2 \frac{(1-F^2)}{\ln(R_o / R_w)} \tag{5}$$

where

Q^* = permissible well discharge
K = hydraulic conductivity
H_o = initial saturated thickness
F = ratio of permissible saturated thickness to initial saturated thickness
R_o = distance from well where change in water table is negligible
R_w = radius of well

The values for R_w and R_o were set to constants of 20.3 cm (8 inches) and 229 m (750 ft), respectively. For the several transient simulations in which equation 5 was used to restrict pumping, each well's withdrawal was limited based on the value of H_0 and K for that cell. Several simulations were done with different values of F, with higher values representing a stronger reduction of well withdrawals and requiring more phantom wells to meet withdrawal targets. The details of these simulations with phantom wells and the level of withdrawal curtailments are presented in the Results section below.

Aquifer Storage Coefficient

Transient ground-water simulations require use of the aquifer storage coefficient, which describes changes in storage in an aquifer through time. Sensitivity analysis using a range in literature values was conducted to evaluate the response of the system to variations in storage coefficients. Figure 36 in Knowles and others (1984a) is a map of specific yield that covers the model area. From this figure, a representative value of 0.15 was selected.

In transient simulations, the value for aquifer storage can have a significant effect upon modeling results; it is a parameter requiring calibration, much like hydraulic conductivity. For the transient simulations the value of 0.15 was used initially, with two simulations evaluating the sensitivity of the model results to changes in this parameter.

Particle Tracking

MODPATH (Pollock, 1989) was used to evaluate flow paths and flow rates from several points in the immediate area of the Pantex Plant. MODPATH evaluates the path of a particle that travels at the mean ground-water or Darcy velocity. If a nonsorbing, nonreactive, and nondecaying contaminant is introduced as a pulse into the aquifer, the traveltime of the particle represents the arrival of the center of mass of the contaminant. MODPATH uses the head files generated in MODFLOW to define the paths or particle tracks. MODPATH required that the effective porosity and the initial placement of particles be assigned.

Effective Porosity

The accurate determination of effective porosity within an aquifer is problematic, especially in aquifers with significant heterogeneities. For this effort, a two-step approach was used. First, porosity logs from a geophysical survey in a monitor well drilled and completed in the Ogallala aquifer at the Pantex Plant, OM-105 (Bureau of Economic Geology, 1992), were evaluated to determine the range in total porosity within the saturated section of the Ogallala aquifer. The range in total porosity determined in OM-105 was then applied to the empirical relationships between total and effective porosity presented by Castany (1967, *in* Marsily, 1981), to determine the corresponding range in effective porosity. Sensitivity analysis was then performed using several values of effective porosity that were within the selected range.

Particle Placement

MODPATH requires the user to specify the initial placement of particles to be tracked during simulations. Both the individual cells and the position on the cell must be designated. The selected points for placement of particles for tracking in this model of the Ogallala aquifer at the Pantex Plant were the five playas located within the reservation. Although the five playas are clearly not the only potential points for entry of contaminants into the Ogallala aquifer in the area, the even spatial distribution of the playas at Pantex allows for a good evaluation of particle paths and traveltimes. Four particles were placed in a square pattern on the top of each cell in the model that corresponded to the location of each of the five playas. Twelve cells in the model were selected for the placement of particles; thus, 48 particles were tracked in the simulations. Ten- to fifty-year time intervals were used to track the particles as they moved through the system.

RESULTS

Predevelopment Model

The three primary hydrologic variables of interest during the predevelopment stage of modeling were (1) rate of recharge, (2) distribution of recharge, and (3) hydraulic conductivity of the aquifer. The level of confidence in other input parameters ranged from high (for example, base of the aquifer) to adequate (boundary conditions). A progression of model simulations were performed to calibrate the model hydraulic conductivity and boundary conductance terms and to evaluate the three recharge distributions.

Initial simulations were used to calibrate the edge conductance terms and the numerical solution parameters. For the drain-type boundary cells, the effective saturated thickness was initially tested by using the 1959–60 water-table elevations extrapolated to the cell edge. This method led to high values of saturated thickness around much of the model perimeter and very high values of edge conductance. Because edge conductance terms remain constant during the entire simulation, these high values caused extreme drops in predicted hydraulic heads around the edge of the model and some convergence problems. To alleviate these problems, the effective saturated thickness was set to a value of 6.1 m (20.0 ft) for all drain-type boundary cells, which substantially improved the model performance. This saturated thickness was used for the rest of the model simulations.

During initial simulations, the Strongly Implicit Procedure (SIP) iterative solution technique was found to perform well, although the large number of nodes (26,167) required a long computation time and a large number of iterations (from 200 to 1,000) for each model run. The maximum allowable head change criterion for the convergence of the iterative process was tested, and a value of 0.3 cm (0.01 ft) was found to yield a mass balance that had usually less than 1-percent error. This convergence criterion was used for all subsequent model runs.

Simulation A used the uniform spatial recharge scenario with a flux of 9 mm yr^{-1} (0.354 inch yr^{-1}) and the derived hydraulic conductivity (fig. 11), assigned to each cell of the model. Figure 13

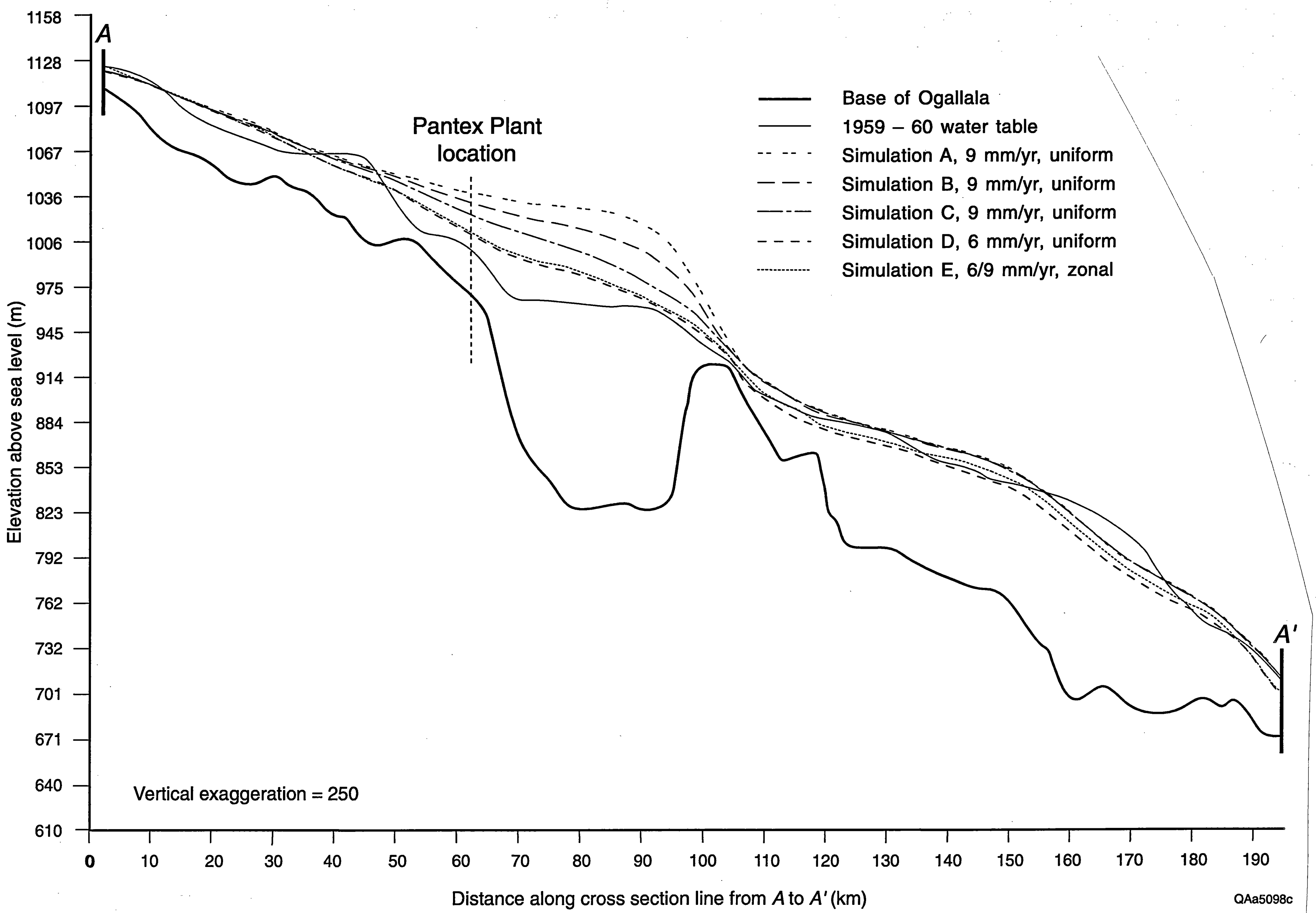

Figure 13. Cross section showing results of simulations A, B, C, D, and E.

shows a cross-sectional view along line A–A′ (as originally presented in fig. 5). This figure compares the calculated water-table surface of simulation A (and the other key simulations discussed below) with the 1959–60 predevelopment surface. The resulting water-table surface from this model crudely approximated the estimated 1959–60 surface in most of the model area. The most notable discrepancy was in the central and western parts of Carson County and in adjacent parts of Randall and Potter Counties, where the model greatly overpredicted the water-table elevation. In simulation A the maximum discrepancy in water-table elevation (between the model and the 1959–60 surface) was in the Carson County area, 75.8 m (249 ft). Because the 1959–60 water-table surface is an estimate, as discussed earlier, the discrepancies presented here should be considered as comparative rather than absolute measures of model results.

To assess the overall match of simulation A (and other simulations highlighted) with the predevelopment water levels, several other comparative measures were used. The first was the absolute average discrepancy over the entire model area's 26,167 finite-difference nodes. To prevent the high number of small cells in the central area of the model, 53 percent by number (13,924 of 26,167) but only 15.5 percent by area (2.25×10^9 of 1.45×10^{10} m^2 [2.42×10^{10} of 1.56×10^{11} ft^2]), from unduly influencing the calculation, each cell was weighted according to its area. Also, in order that positive and negative discrepancies did not cancel one another out, the calculation was done with the absolute value of the difference in head at each node and is referred to as the *absolute average discrepancy* (also referred to as mean absolute error). Comparisons between simulated and known water-table surfaces were made at the scale of the cell node, not at individual wells. For simulation A, the area-weighted absolute average discrepancy between the calculated water-table surface and the 1959–60 predevelopment surface was 16.9 m (55.4 ft). The maximum discrepancy in the center of Carson County and the absolute average discrepancy for the entire model area for simulation A are depicted in table 6.

A second measure of model accuracy was a comparison of the estimated amount of water in storage for the run versus the estimated 1959–60 conditions. This measure was especially important because the average head discrepancy measurement was not able to show the overall

Table 6. Summary of steady-state ground-water modeling results.

Run I.D.	Recharge scenario and flux (mm/yr)	Measures of discrepancy† from 1959-60 water table: mean error (m)	standard deviation (m)	mean absolute (m)	extremes - (m)	extremes + (m)	water in storage* volume (10^{10} m^3)	water in storage* percent 1959-60	mass balance flow terms** (× 10^5 m^3/day): inflow from recharge	inflow from west	discharge on east	spring & seepage discharge	discharge to Canadian River
A	*uniform*: 9 mm/yr	10.7	26.4	16.9	-71.2	75.8	12.85	122.1	3.56	0.19	0.10	2.68	0.96
B	*uniform*: 9 mm/yr	8.5	21.1	14.5	-70.0	61.0	12.37	117.5	3.56	0.19	0.10	2.67	0.99
C	*uniform*: 9 mm/yr	5.6	15.7	11.7	-68.1	48.1	11.75	111.6	3.56	0.19	0.10	2.65	1.02
D	*uniform*: 6 mm/yr	-4.0	12.9	9.9	-72.7	31.0	9.65	91.7	2.37	0.21	0.07	1.80	0.73
E	*zonal*: Blackwater Draw @ 6, Ogallala @ 9	0.3	12.6	9.2	-68.1	33.2	10.59	100.7	3.10	0.21	0.10	2.24	0.95
F	*playa*: Blackwater Draw @ 6, Ogallala @ 9	-0.3	13.6	9.9	-68.1	36.7	10.45	99.3	2.99	0.23	0.10	2.22	0.90
G	*playa*: Blackwater Draw @ 5, Ogallala @ 9	-1.4	13.2	9.7	-68.1	33.6	10.22	97.1	2.85	0.23	0.10	2.12	0.89
H	*playa*: Blackwater Draw @ 3, Ogallala @ 9	-4.4	12.1	10.1	-68.1	30.4	9.56	90.9	2.58	0.24	0.10	1.91	0.86
I	*playa*: Blackwater Draw @ 2, Ogallala @ 9	-6.8	11.4	10.8	-68.1	30.1	9.05	86.0	2.44	0.25	0.10	1.76	0.84

† Means and standard deviation based on the difference between head predicted by MODFLOW and the 1959-60 head for the 26,167 active cells, each weighted by proportional area. "Mean error" includes positive and negative values while the "mean absolute" uses only absolute values.

* All storage calculations based on storage coefficient of 0.15. The 1959-60 volume in storage = 10.52×10^{10} m^3.

** The sums of inflow and discharge terms may be slightly unequal due to the effects of rounding and slight mass balance errors in the original MODFLOW runs. The original mass balance errors for the nine runs above were, in percent, 0.1, 0.2, 0.2, 0.5, 0.6, 0.5, 0.5, 1.2, and 0.1, respectively.

underprediction or overprediction in the water-table surface. For simulation A, the calculated volume of water in storage, using a storage coefficient of 0.15, was 12.85×10^{10} m^3 (4.54×10^{12} ft^3) or 122.1 percent of the estimated predevelopment conditions. This represented an overall overprediction of water in storage.

Third, volume balance terms indicate the relative influence boundary conditions have on the results of any given model simulation. For the uniform recharge scenario in simulation A, the flux of 9 mm yr^{-1} (0.354 inch yr^{-1}) results in a total recharge of 3.56×10^5 m^3 d^{-1} (1.26×10^7 ft^3 d^{-1}) (table 6). In comparison, the hydraulic connections to other parts of the Ogallala aquifer at the western and eastern edges represent only a small part of the water budget with inflow of 1.9×10^4 m^3 d^{-1} (6.71×10^5 ft^3 d^{-1}) and an outflow of 1.0×10^4 m^3 d^{-1} (3.53×10^5 ft^3 d^{-1}), respectively. Most of the inflow from recharge and western inflow is discharged through spring and seepage flow around the nonriver perimeter of the model (2.68×10^5 m^3 d^{-1} [9.46×10^6 ft^3 d^{-1}] or 71 percent), represented by the 933 drain-type boundary cells. A smaller amount, 9.6×10^4 m^3 d^{-1} (3.39×10^6 ft^3 d^{-1}), or 29 percent of inflow, is discharged through the 72 Canadian River cells. The relative magnitude of these terms suggests that the dominant features affecting the overall predevelopment model results were recharge and the amount of discharge along the perimeter of the model area. On a per-cell basis, however, the amount of discharge through the river cells is almost an order of magnitude greater than the discharge through the drain-type boundary cells. To reduce the excess of water in storage resulting from simulation A, either reducing the recharge or increasing the conductivity of the drain- and river-type boundaries were the next options considered to improve overall model performance.

Because model stresses, boundary conditions, and aquifer properties collectively determine a ground-water model's results, it was also necessary to examine the aquifer properties for their influence on the model results. Figure 13 shows that the large overprediction in heads of simulation A in central to western Carson County lies upgradient from a region where the base of the Ogallala Formation is relatively high (fig. 5). Figure 11 shows that this is also an area of very low hydraulic conductivity, in the range of 0.7 to 1.3 m d^{-1} (2.3 to 4.3 ft d^{-1}). The combined effect is to back up the water-table heads calculated by MODFLOW to the southwest until sufficient gradient forms to drive the ground water across the flow restriction. Because this was the principal area of discrepancy between the water table of model simulation A and the 1959–60 surface, several subsequent runs examined the sensitivity of hydraulic conductivity in this area on the model results. The base of the aquifer was not altered because there was a relatively high level of confidence in these data.

The extremely low values in this critical area were based on three isolated data points in eastern Carson County (fig. 8). Much of the rest of the area had a greater data density. Because of the questionable model performance in this flow-restriction area, the confidence in these three isolated points was considered suspect. Furthermore, the earlier modeling efforts by Knowles and others (1984a) and Luckey and others (1986) used hydraulic conductivities in the range of 10.3 to 20.4 m d^{-1} (34 to 67 ft d^{-1}) and 7.6 to 15.2 m d^{-1} (25 to 50 ft d^{-1}), respectively. The hydraulic conductivities of these earlier studies were approximately 8 to 20 times higher than our study's 0.7 to 1.3 m d^{-1} (2.3 to 4.3 ft d^{-1}).

To evaluate the effects of the hydraulic conductivity, hydraulic conductivity was varied in two ways. First, *all* hydraulic conductivities over the entire model area were increased by a factor between 1.5 and 3.0. Second, *only* the hydraulic conductivity of the three isolated points in eastern Carson County were increased by factors of 2.0 to 4.0. Of these two approaches, the best results were obtained with the second approach. With the highest adjustment factor of 4.0, the average hydraulic conductivity in the flow-restriction area was 4.0 m d^{-1} (13.1 ft d^{-1}), well within the range of nearby values derived for this study and still below those used in previous modeling.

Results of simulations B and C are depicted in figure 13. As evident from this figure, both of these simulations resulted in improvements in the match to the 1959–60 water-table surface in the central model area and essentially left the downgradient area unchanged. In simulation B, doubling the hydraulic conductivity resulted in improvements of average and maximum discrepancy in water-table heads of 14 percent and 20 percent, respectively, over simulation A. The match to the estimated water in storage was improved by less than 5 percent. Quadrupling the hydraulic conductivity in the flow restriction area in simulation C resulted in improved average and maximum discrepancy in water-table head by 31 and 37 percent, respectively, over simulation A, whereas the estimated water in storage was improved by 11 percent. Figure 14 provides a map view for comparison of the model-calculated water table with that of the 1959–60 predevelopment surface for simulation C. Substantial improvements in the resulting water-table surfaces for simulations B and C were accompanied by only minor changes in the relative magnitudes of the overall water balance terms, as evident in table 6. Slightly less water was conducted through the spring and seepage drain-type boundaries, with the balance shifted to the Canadian River cells.

Because of improvements in model results from adjusting the hydraulic conductivity with the above procedure, the hydraulic conductivity arrangement of simulation C was preserved for all subsequent model simulations. Except for the influence of the altered hydraulic conductivity on the boundary conductance terms, no explicit reformulation of the boundary conditions was used.

Because simulation C still overpredicted the water-table surface elevation in the center of the model and the overall storage of water was still at 111.6 percent of the estimated 1959–60 level, the next step selected was to reduce the recharge. Key simulation D used a reduced recharge flux rate of 6 mm yr^{-1} (0.236 inch yr^{-1}), applied uniformly to the entire model area. The resulting water-table surface from simulation D is depicted in the cross section in figure 13 and in the map view in figure 15. The water-table surface in the center of the model in Carson County still was overpredicted, although the maximum discrepancy declined to 31.0 m (102 ft). The overall absolute average discrepancy for the whole model area declined to 9.9 m (32.5 ft). However, in more northeastern areas the resulting surface of simulation D was often substantially below that of 1959–60. This is also reflected by the overall storage term falling to 91.7 percent of the predicted 1959–60 level. The mass balance terms of table 6 also show the effects of the reduction in recharge compared with those of earlier runs. Slightly more water flows across the western boundary because the water-table surface is lowered, thereby increasing the hydraulic head gradient across these boundary cells. All other volume balance terms, representing discharge across boundaries, were reduced because of lowered gradients associated with a lower water-table surface.

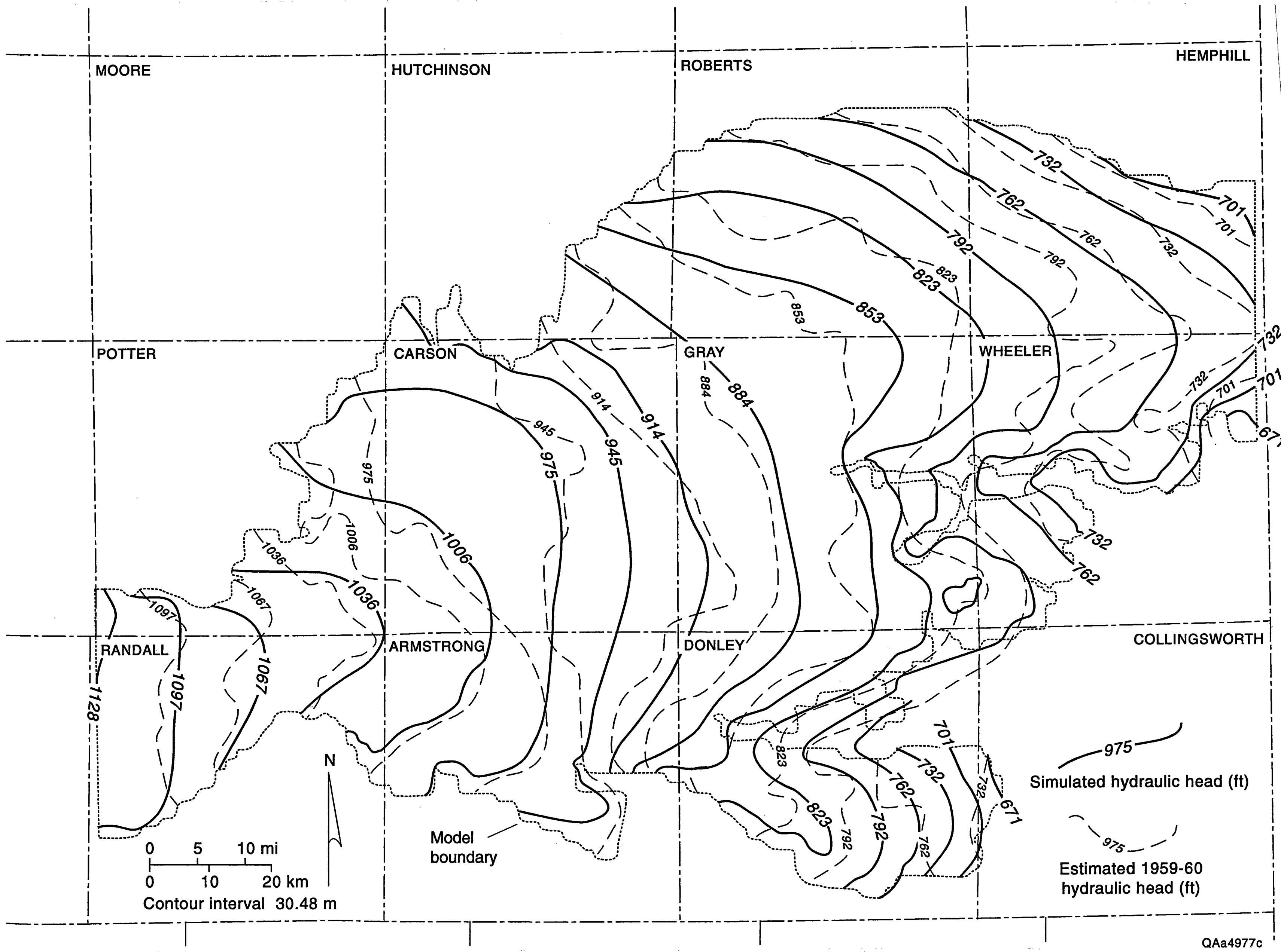

Figure 14. Simulation C water-table surface and estimated 1959 to 1960 measured water-table surface.

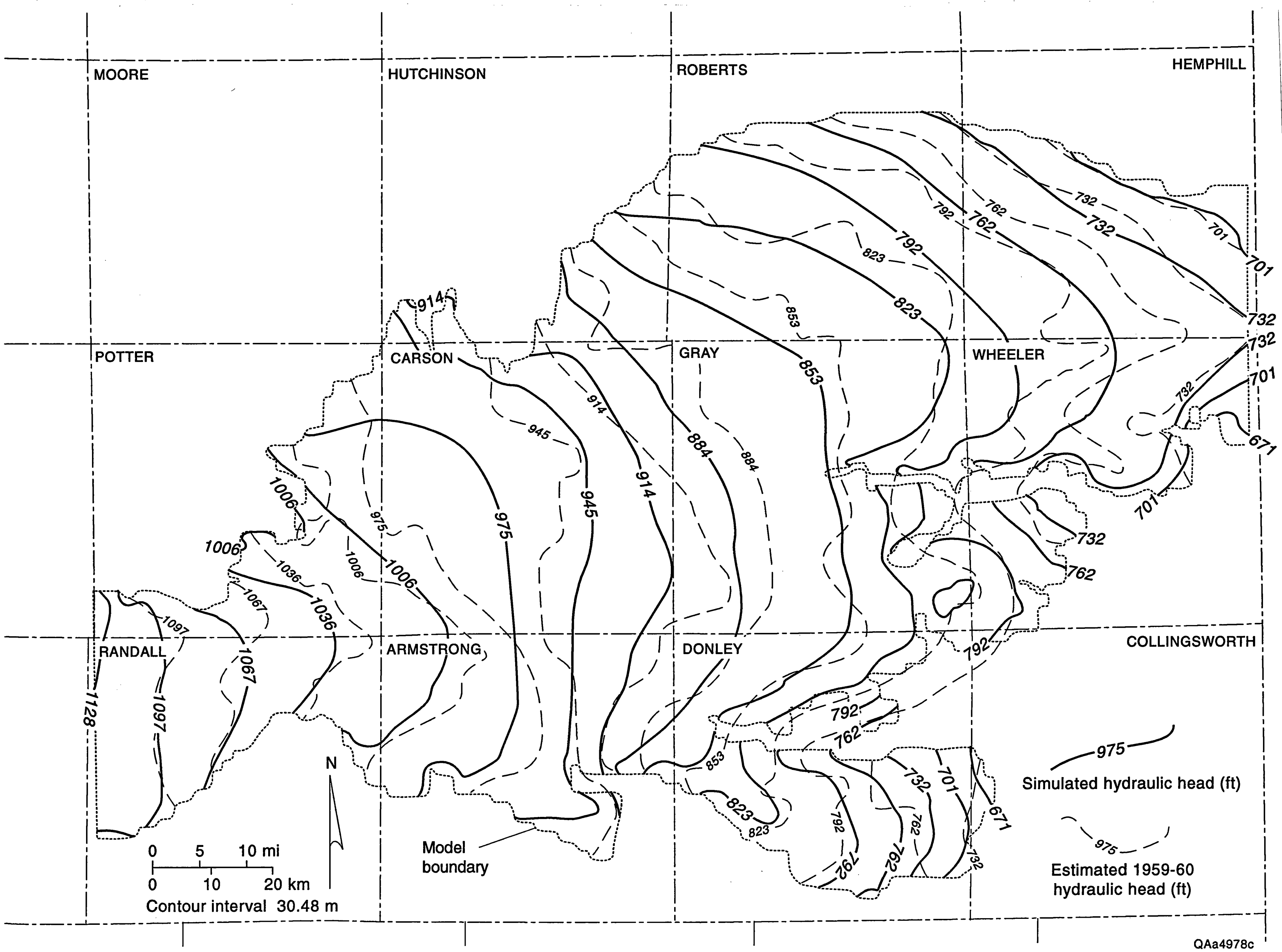

Figure 15. Simulation D water-table surface and estimated 1959 to 1960 measured water-table surface.

The transition from the 9 mm yr^{-1} (0.354 inch yr^{-1}) uniform recharge case of simulation C to the 6 mm yr^{-1} (0.236 inch yr^{-1}) uniform recharge case of simulation D was accompanied by an improved water-table surface match under visual inspection in the central to western parts of the model area, but a general underprediction of the water-table surface in more northeastern areas. As a result, a different recharge scenario was next evaluated. Simulation E used a zonal recharge scenario with the recharge flux set according to the regionally mapped geologic unit exposed at the surface. A comparison of these two (C and D) simulations illustrated that a better agreement was achieved in areas where the Blackwater Draw Formation was present at the surface using the uniform recharge rate of 6 mm yr^{-1} (0.236 inch yr^{-1}), whereas 9 mm yr^{-1} (0.354 inch yr^{-1}) achieved a better agreement in areas where the Ogallala Formation was present at the surface. The resulting water-table surface for simulation E is shown in the cross section in figure 13 and the map view in figure 16. For simulation E an improvement in overall match to the 1959–60 surface is shown by the absolute average discrepancy and water in storage entries in table 6. The maximum water-table discrepancy in central Carson County, however, increased slightly to 33.1 m (108.6 ft).

The remaining model runs of the predevelopment stage of modeling were designed to evaluate the recharge role of the numerous playas of the Blackwater Draw Formation of the Southern High Plains (fig. 2). Recent evidence has suggested that in this area of the Southern High Plains, recharge on this surface is restricted to the playas and negligible recharge occurs in the interplaya areas (Gustavson and others, 1993; Scanlon and others, 1993, 1994), which serves to confirm the conceptual models proposed by numerous investigators dating back to Johnson (1901). This scenario was simulated by focusing the total recharge volume of the Blackwater Draw Formation surface through the area of the playas.

To represent the playa recharge scenario, it was necessary to determine which cells of the finite-difference grid corresponded to a playa location and to determine the appropriate recharge flux of those cells. The areal extent of a playa was defined as the lowest closed contour on 7.5-minute USGS topographic maps (filled with blue pattern), which corresponds to an area

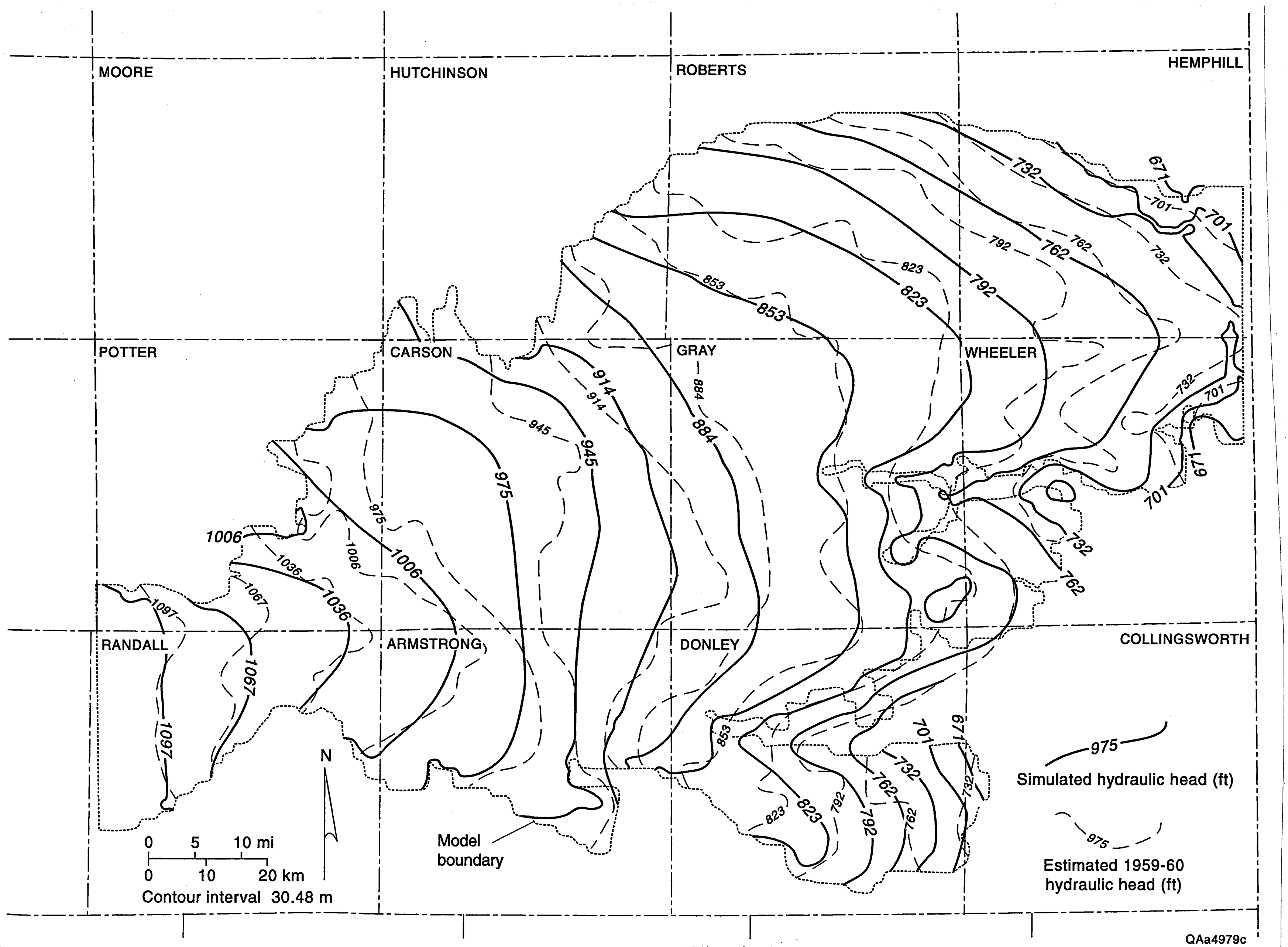

Figure 16. Simulation E water-table surface and estimated 1959 to 1960 measured water-table surface.

periodically flooded. One thousand forty-one playas were located on sixty-one 7.5-minute USGS topographic maps of the model area. The ARC/INFO® Geographic Information System (GIS) software was used to assign each playa a unique identification number and to convert the lowest contour to digital coordinates from which the playa area and center coordinates were computed. The ARC/INFO® analysis found that the Blackwater Draw Formation covers 4.98×10^9 m^2 (5.37 $\times 10^{10}$ ft^2) in the study area and playas cover 1.37×10^8 m^2 (1.47×10^9 ft^2), or 2.7 percent of this surface. Of the 13,511 cells of the Blackwater Draw part of the grid, 1,015 were found to correspond to playa coverage cells, with some cells holding more than 1 playa and some large playas covering up to 8 cells.

Simulation F was the first simulation in which the *playa-focused* recharge scenario was used. The regional Blackwater Draw Formation recharge rate of 6 mm yr^{-1} (0.24 inch yr^{-1}). was applied through playas for an effective recharge rate in the playas of 219 mm yr^{-1} (8.6 inch yr^{-1}), with no recharge occurring in interplaya areas. The Ogallala Formation outcrop area was set at 9 mm yr^{-1} (0.35 inch yr^{-1}). As can be seen in the volume balance terms of table 6, there is a slight difference (3.5 percent) in the total recharge flow of simulation E from that of F, although they are each based on the same recharge rates. This small difference is due to slight differences in the area of the Blackwater Draw Formation, as determined by the ARC/INFO® procedure compared with the area of the section of the finite-difference grid of MODFLOW used to represent the Blackwater Draw.

Simulation F results are shown on the cross-sectional view in figure 17 and the map view of figure 18. The results were very close to those of the zonal recharge scenario of simulation E. In the Blackwater Draw part of the model, the estimated water-table surface of the playa recharge scenario was above the zonal case in areas of high playa density and below the zonal case in areas of low playa density. For example, in the area near the Pantex Plant the playa recharge scenario results in a slightly higher surface than the zonal scenario, a trend which reverses to the northeast, an area of diminished playa density. This playa recharge simulation, as indicated by the water in storage, was as close to the estimated 1959–60 value as was simulation E, with respective +0.7

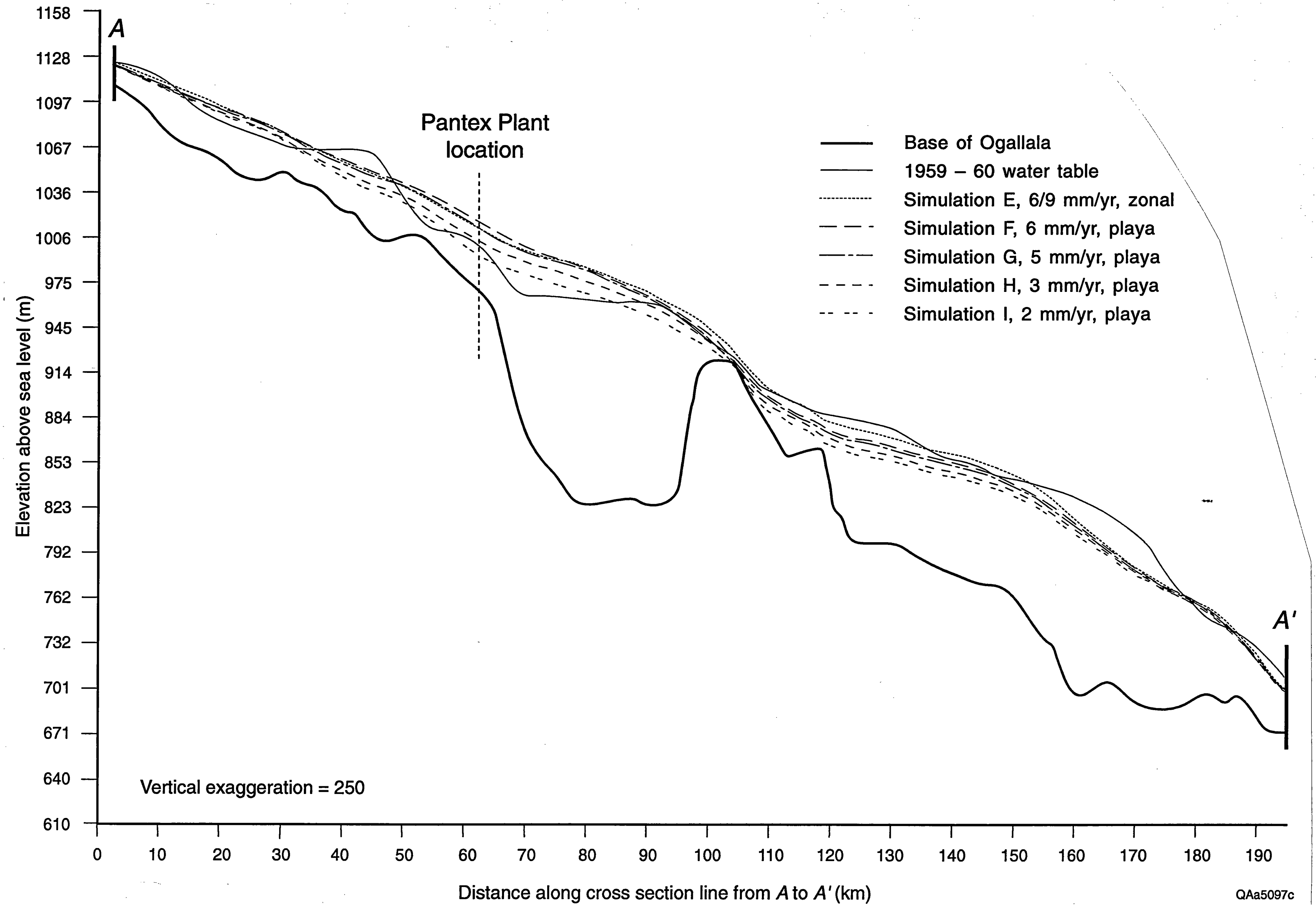

Figure 17. Cross section showing results of simulations E through I.

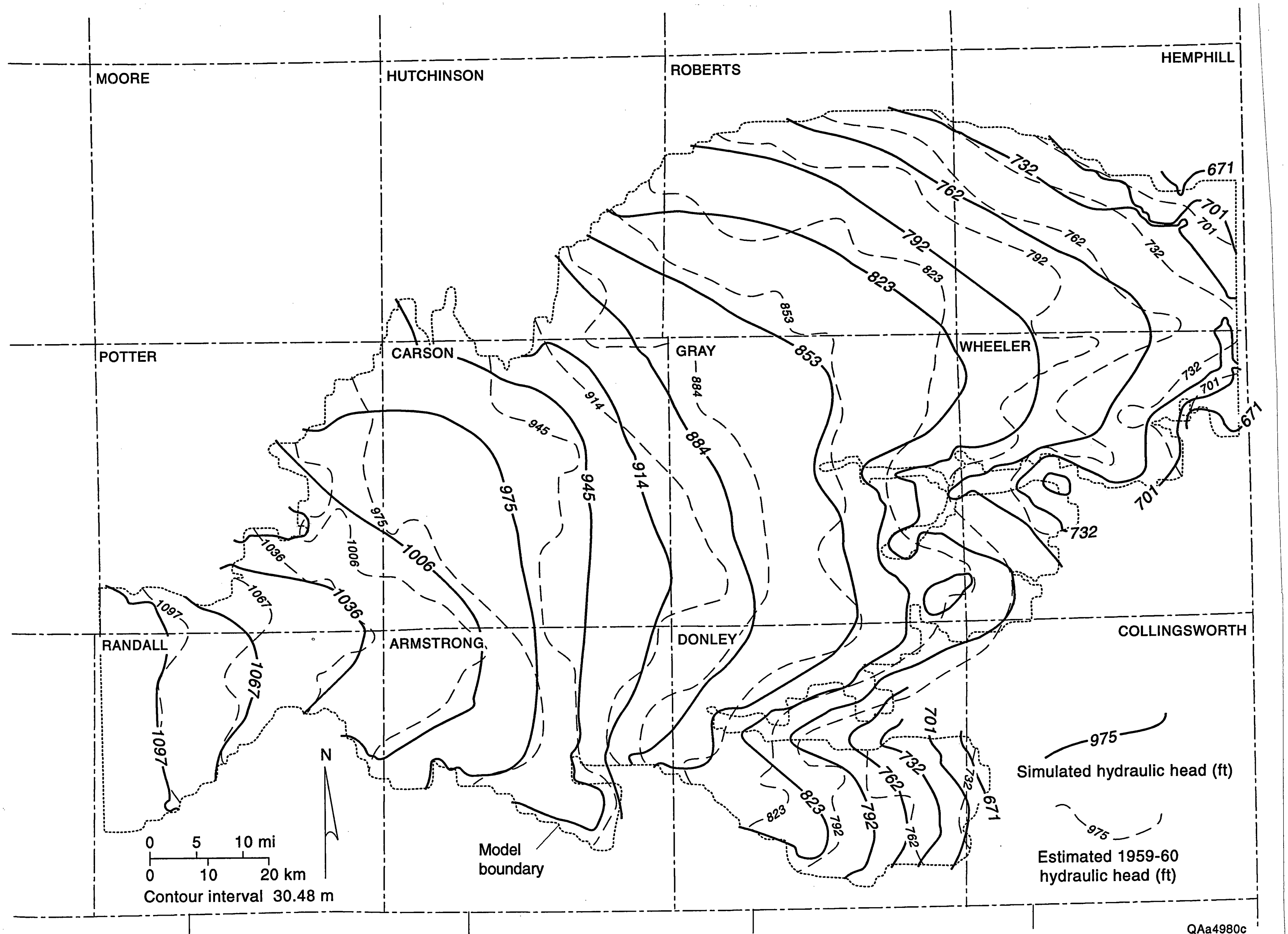

Figure 18. Simulation F water-table surface and estimated 1959 to 1960 measured water-table surface.

percent and –0.7 percent differences. The absolute average and maximum Carson County head discrepancy values of 9.2 m and 36.7 m were slightly worse than the zonal case.

Because the results of the playa recharge scenario with a 6 mm yr^{-1} (0.236 inch yr^{-1}) flux rate overpredicted the estimated water-table surface in central Carson County, a series of simulations were used to examine the effects of lowering the recharge rate of the playas. Simulations G, H, and I used the playa recharge scenario with flux rates on the Blackwater Draw surface of 5, 3, and 2 mm yr^{-1} (0.197, 0.118, and 0.079 inch yr^{-1}), respectively (figs. 19–22). This progression of lower recharge fluxes led to a closer match of the estimated water-table surface in the area of the Pantex Plant. Table 6 shows that the maximum discrepancies in Carson County declined from 33.6 m (110.2 ft) in simulation G to only 19.1 m (62.7 ft) in simulation I. Table 6, however, also indicates a worsening trend in absolute average match to the 1959–60 water-table surface and the estimated water in storage over the entire model area. Figure 19 also shows that near the area of low saturated thickness in northeastern Carson County, and in some areas to the northeast, the water-table surface falls far below the estimated 1959–60 level.

Transient Model

In the transient stage of modeling a series of simulations were used to simulate the time period 1960–90. Transient modeling permitted the evaluation of whether the rates of recharge proposed in the predevelopment stage of modeling could be coupled with the substantial known withdrawals of water to predict water-level declines over the 1960-90 period. Transient modeling also provided for the investigation of the hydrologic relationship of the ACCWF to the changes in the water table in the vicinity of the DOE Pantex Plant.

Since the three recharge scenarios evaluated in the predevelopment stage of modeling (*uniform, zonal,* and *playa-focused*) had relatively comparable results, only the *playa-focused* scenario was used here. The respective Blackwater Draw Formation and Ogallala Formation outcrop recharge rates of simulation F were used: 6 mm yr^{-1} (0.24 inches yr^{-1}) and 9 mm yr^{-1}

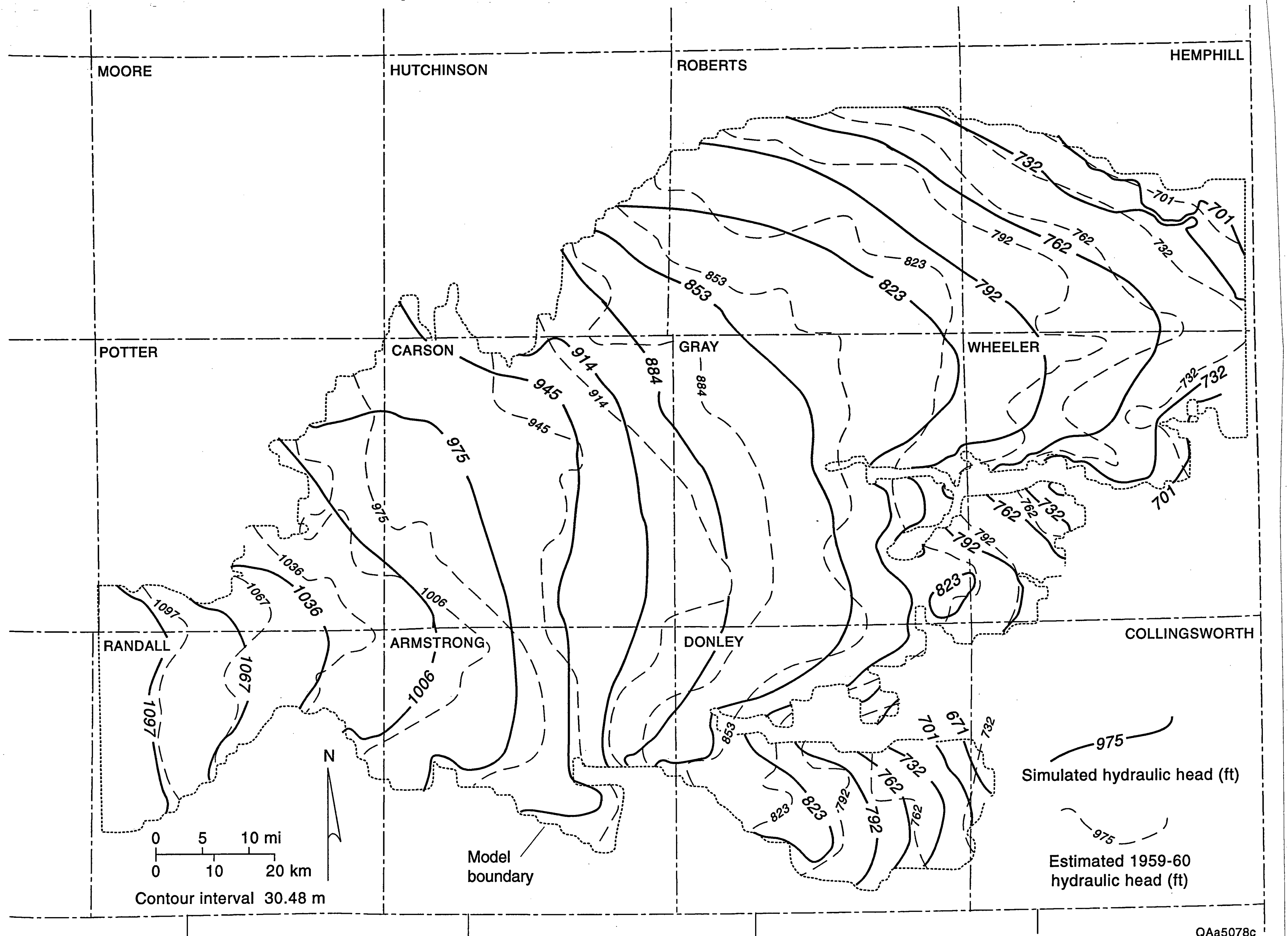

Figure 19. Simulation H water-table surface and estimated 1959 to 1960 measured water-table surface.

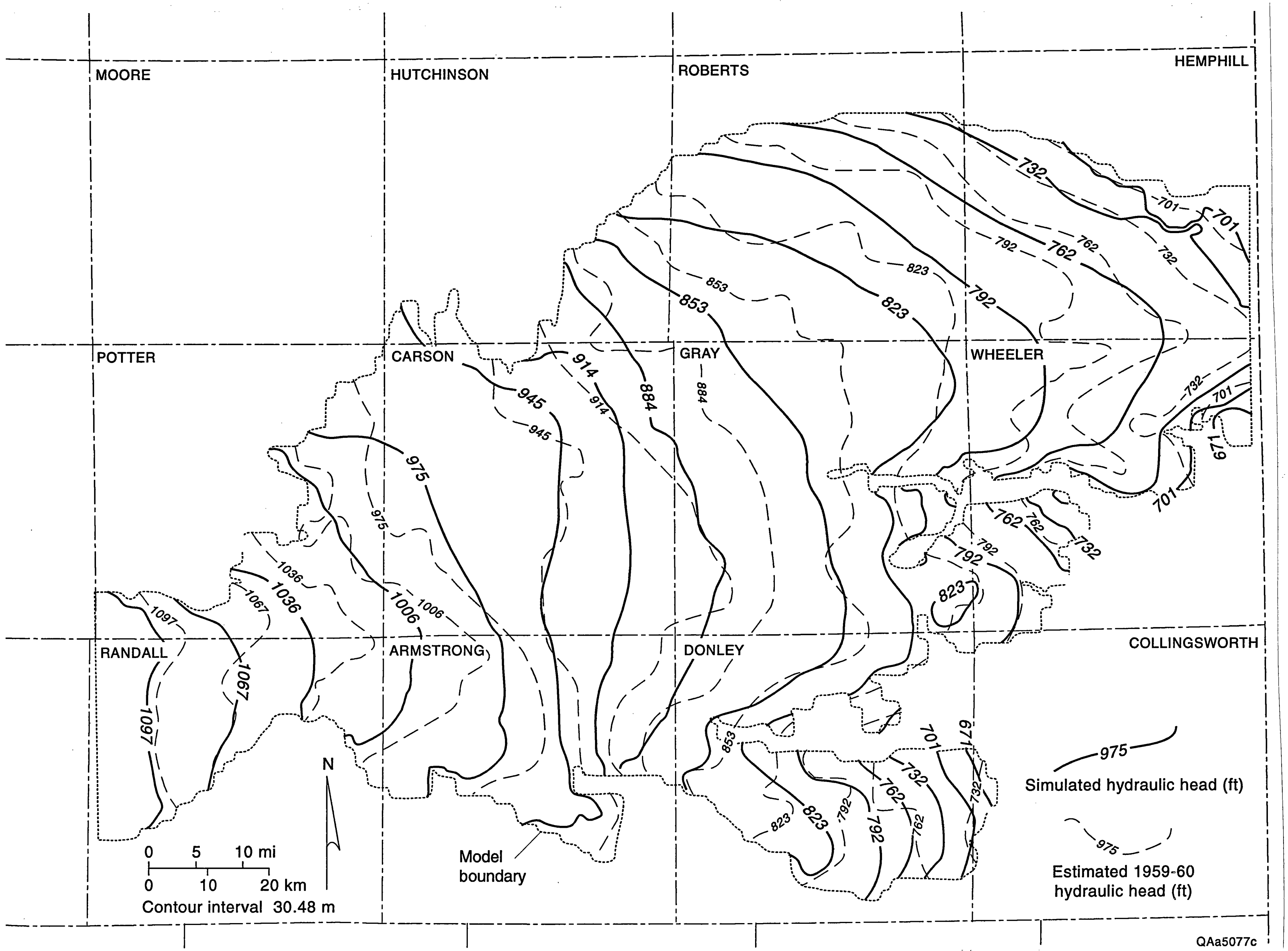

Figure 20. Simulation I water-table surface and estimated 1959 to 1960 measured water-table surface.

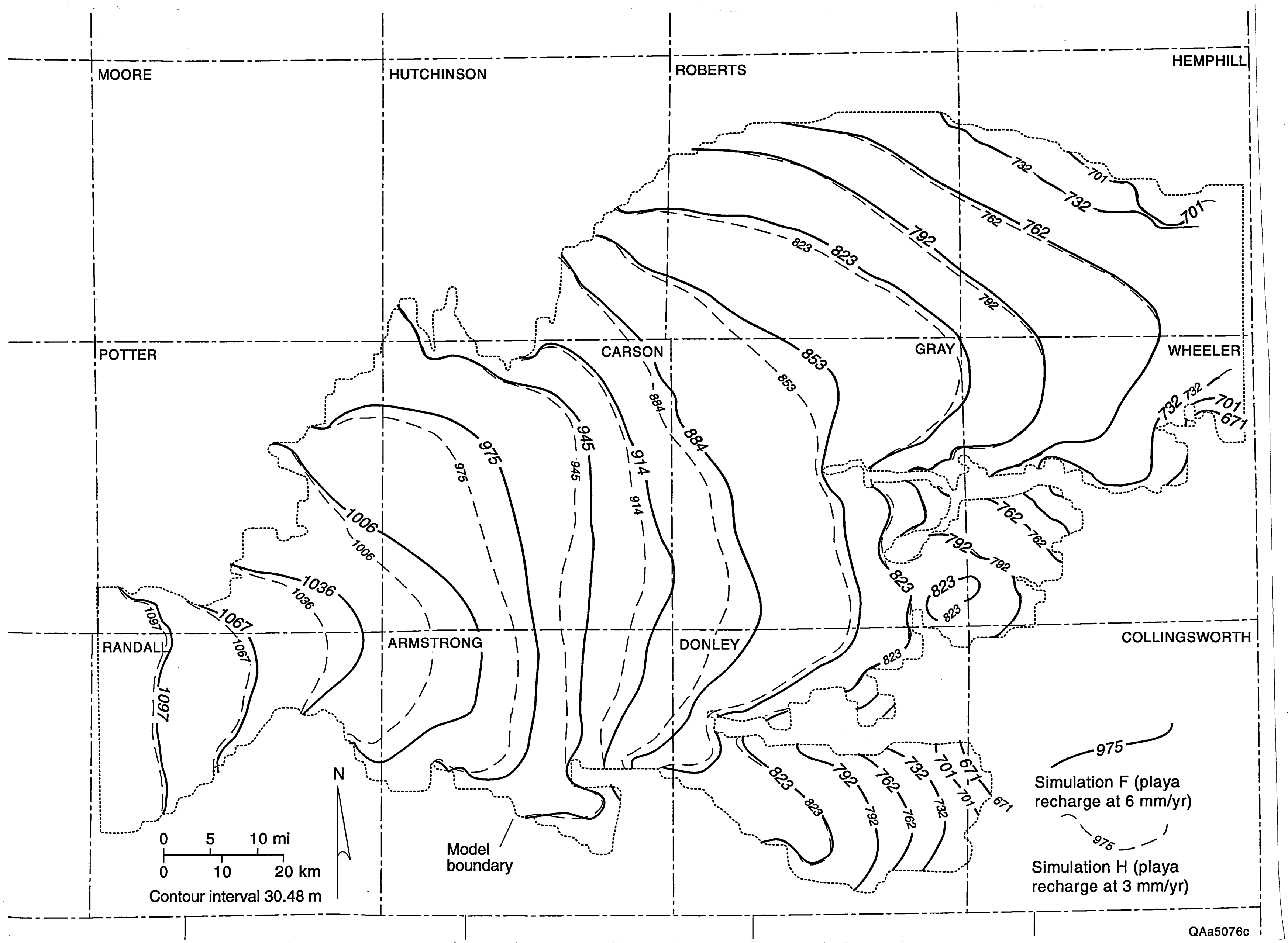

Figure 21. Comparison of water-table surfaces of simulations F and H.

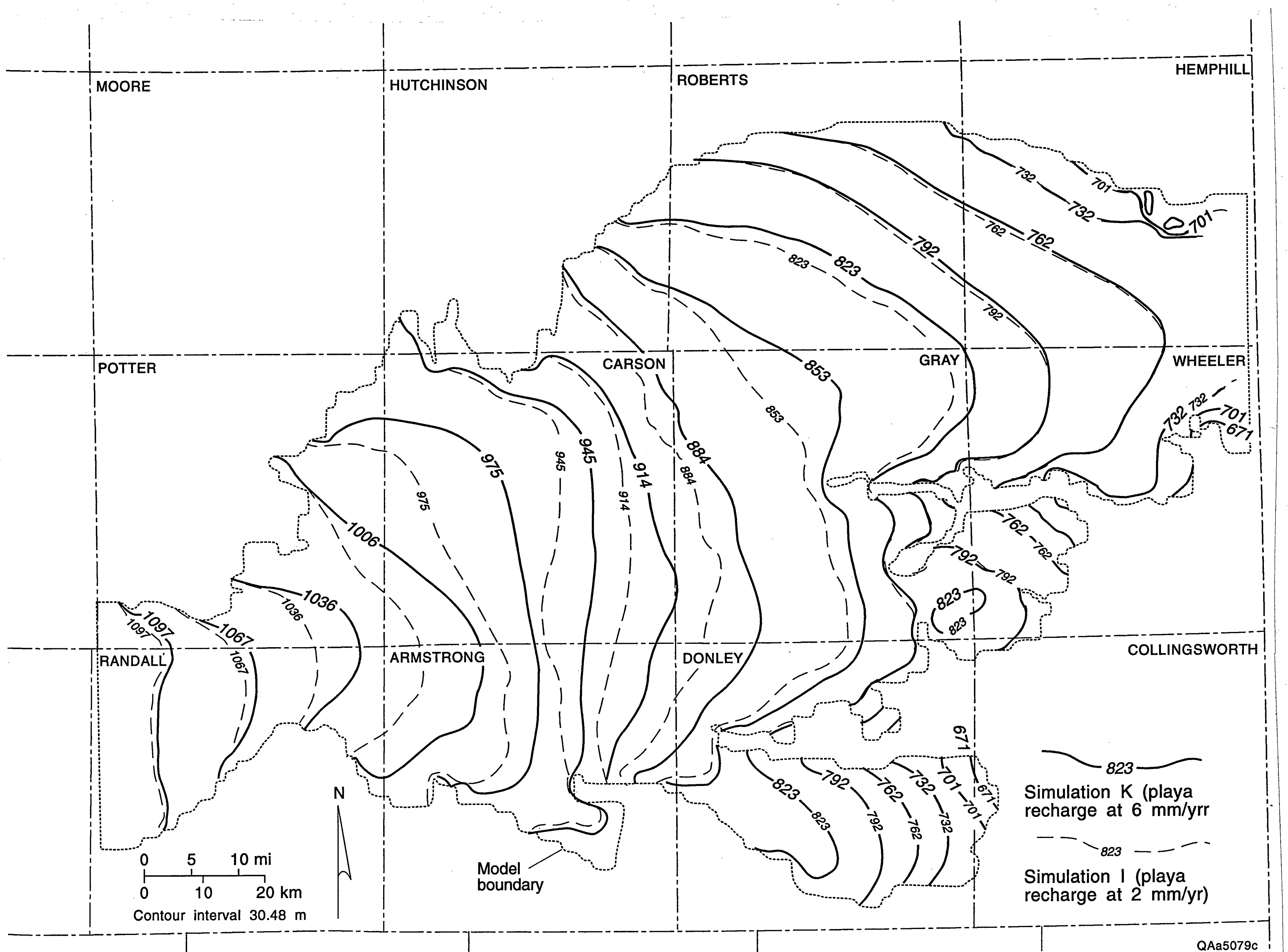

Figure 22. Comparison of water-table surfaces of simulations F and I.

(0.35 inches yr^{-1}). All of the transient simulations used the same boundary conditions as the predevelopment stage, and initial conditions were set to the 1959–1960 water table.

Simulation J utilized the set of known wells as presented in table 4 and figure 12, with default withdrawal rates as in table 5. The aquifer storage coefficient was set to 0.15 for the entire model area, and 6 time steps, equal to the subperiods presented earlier, were used. Figure 23 is a map view with the results of this simulation. The simulated 1990 water-table surface is compared to the "known" 1990 water-table surface.

In general, simulation J produced a very reasonable result with fairly close agreement between simulated and "known" 1990 water-table surfaces over most of the model area. A notable disagreement was in the extreme northern portion of Donley County where a substantial mounding of the "known" surface was not predicted. This is an area of high playa density (see fig. 2) and hence high recharge under the *playa-focused* scenario, which is also indicated by the shape of the predicted water-table surface contour at 853 m (2800 ft). Nonetheless, the simulated water-table surface was as much as 52.5 m (172 ft) too low in this area. As discussed above, the known water-table surface in this area appears to be an anomaly, possibly indicating the presence of a localized perched aquifer above the main Ogallala aquifer.

The water-table surface of simulation J shows the effects of the high rates of "default" withdrawals with the many closed and jagged contours representing precipitous changes in the water-table surface near wells. Table 7 offers a comparison of the simulated and known surfaces for this and the other simulations with measures of discrepancy between the surfaces on the left. The measures are similar to those in table 6 for the predevelopment modeling.

The right six columns of table 7 present mass balance terms for the entire 31-year period. For example, recharge under the *playa-focused* scenario represented an addition of 3.39×10^9 m^3 (1.20×10^{11} ft^3) of water, whereas the loss of water around the model boundary to springs and seeps totaled 4.27×10^9 m^3 (1.51×10^{11} ft^3). Obviously the dominant terms among these are the well withdrawals at 8.18×10^9 m^3 (2.89×10^{11} ft^3) and the decline of water in storage of 10.11×10^9 m^3 (3.57×10^{11} ft^3). These represent 69.3% and 86.4%, respectively, of the known values.

Table 7. Comparison of key transient model simulations for entire model area.

Run I.D.	Maximum number of wells used† for whole model/ discharge summary	Measures of discrepancy†† from 1990 water table						Mass balance volumes for period 1960-90** ($\times 10^9$ m^3)					
		mean error (m)	standard deviation (m)	mean absolute (m)	extremes - (m)	extremes + (m)	1990 storage* (10^9 m^3)/ [% known]	6/9 playa recharge inflow	east discharge/ west inflow	spring & seepage discharge	flow to Canadian River	withdrawn with wells/ [% known]	change in storage
J	741 T; 37 A/ discharges at default levels	0.7	13.9	6.2	63.9	36.7	94.9/[101.5%]	+3.39	-0.03/+0.39	-4.27	-1.42	-8.18/[69.3%]	-10.11
K	741 T; 37 A; 1165 M/ A at default; T & M at 36.35 m^3hr^{-1}	-0.1	14.0	6.3	52.5	36.4	93.3/[99.8%]	+3.39	-0.03/+0.34	-3.98	-1.35	-10.15/[86.0%]	-11.78
L	741 T; 37 A; 1538 M; 5 P/ A & P at default; T & M at max. of 2/3 drawdown limit or 36.35 m^3hr^{-1}	-0.6	13.8	6.4	52.5	35.4	92.2/[98.6%]	+3.39	-0.03/+0.37	-3.96	-1.34	-11.45/[97.0%]	-13.01
M	741 T; 37 A; 1453 M; 5 P/ A & P at default; T & M at max. of 9/10 drawdown limit or 36.35 m^3hr^{-1}	-0.6	13.9	6.3	52.5	35.4	92.3/[98.7%]	+3.39	-0.03/+0.38	-3.95	-1.35	-11.31/[95.8%]	-12.87

† T = wells from the Texas Natural Resources Information System database; A = City of Amarillo wells in the Carson County Wellfield; M = randomly placed "make-up" wells; P = Pantex Plant wells.

†† Means and standard deviation based on the difference between heads predicted by MODFLOW and the 1990 known heads for the 26,167 active cells, each weighted by proportional area. "Mean error" includes positive and negative values while the "mean absolute" uses only absolute values.

*All storage calculations based on storage coefficient of 0.15. Based on the 1990 known water-table surface, the 1990 volume in storage = 93.5×10^9 m^3. This compares to the 1959-60 "predevelopment" volume in storage = 105.2×10^9 m^3 for a known 1960-1990 change in storage of -11.7×10^9 m^3.

**The sums of inflow and discharge terms are slightly unequal due to the effects of rounding.

The large discrepancy in well withdrawal volume was due to cell drying caused by excessively high withdrawal rates per well as discussed above.

The overall results of simulation J are reasonable with regard to the water-table surface agreement and tend to verify that the *playa-focused* recharge scenario and the recharge rates used in simulation F are credible. However, to evaluate the hydrologic relationship of the ACCWF to the changes in the water-table surface in the vicinity of the Pantex Plant, it was also necessary to examine a “close-up” of this area as shown in figure 23. The measures of discrepancy were repeated for this area as presented in table 8.

Whereas the mean error and standard deviation of simulation J for the entire model were 0.7 m (2.3 ft) and 13.9 m (45.6 ft), respectively, they were 6.0 m (19.7 ft) and 2.4 m (7.9 ft) for the close-up area. These data indicate that the overall model had a mixture of positive and negative discrepancies which tended to cancel each other, whereas the close-up area has mostly a positive discrepancy. This means that in the vicinity of the Pantex Plant and ACCWF, the model water-table surface was higher than the known 1990 surface. Because of this discrepancy, coupled with the model’s wells only achieving 69.3 percent of known withdrawals, it was deemed necessary to modify the wells to spread the withdrawals among more wells.

For simulation K, all parameters were as in simulation J, except that the set of wells withdrawing water was expanded from the 778 known TNRIS and ACCWF wells to include a large number of phantom wells. For this and all transient simulations, the ACCWF wells were maintained at the withdrawal rates indicated in table 5. In simulation K each of the TNRIS and phantom wells was set to a maximum withdrawal rate of 36.35 $m^3\ hr^{-1}$ (160.0 gallon min^{-1}). This maximum rate was based on estimates of the typical withdrawals of irrigation wells in the area (Williams, 1994, pers. comm.) normalized over a whole year. For each county, the period was determined in which the maximum required number of phantom wells, if any, was needed to meet the known withdrawals. For other periods in which the maximum number of phantom wells were not needed in that county, the maximum number of such wells was maintained as actively pumping wells, but their withdrawals were proportionally decreased to meet the known withdrawal target.

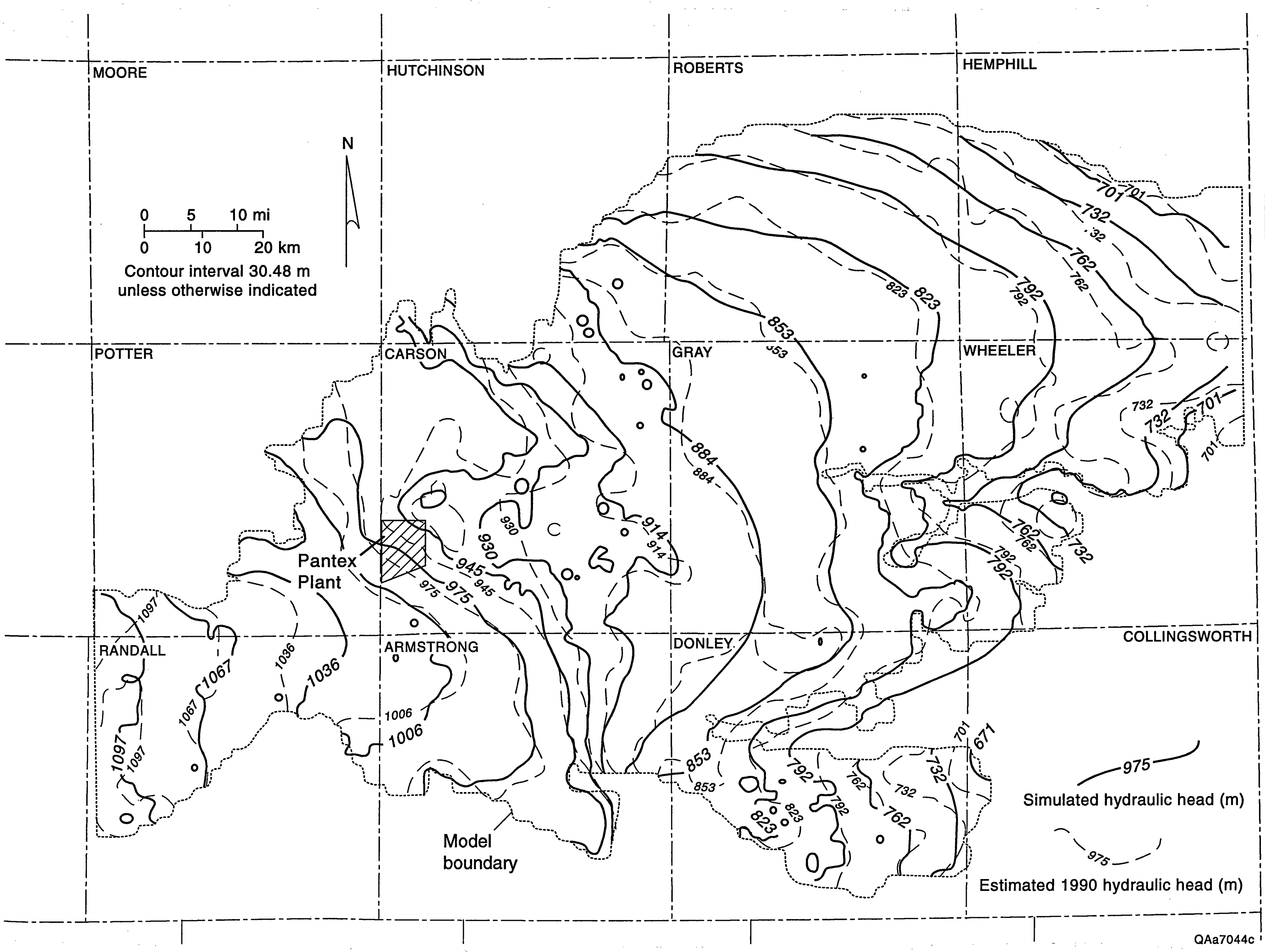

Figure 23. Simulation J water-table surface and estimated 1990 water-table surface.

Table 8. Comparison of key transient model simulations results in the vicinity of the Pantex Plant and ACCWF.

Run #	Maximum number of wells used† for whole model/ discharge summary	Measures of discrepancy†† from 1990 water table: mean error (m)	standard deviation (m)	mean absolute (m)	extremes - (m)	extremes + (m)	1990 storage* (10^9 m^3)/ [% known]
J	741 T; 37 A/ discharges at default levels	6.0	2.4	2.4	30.8	19.4	6.45/[109.0%]
K	741 T; 37 A; 1165 M/ A at default; T & M at 36.35 m^3hr^{-1}	4.9	2.3	1.9	22.2	19.2	6.36/[107.4%]
L	741 T; 37 A; 1538 M; 5 P/ A & P at default; T & M at max. of 2/3 drawdown limit or 36.35 m^3hr^{-1}	5.5	3.0	2.1	12.9	20.9	6.41/[108.3%]
M	741 T; 37 A; 1453 M; 5 P/ A & P at default; T & M at max. of 9/10 drawdown limit or 36.35 m^3hr^{-1}	5.7	3.0	2.1	14.7	21.2	6.43/[108.5%]

† T = wells from the Texas Natural Resources Information System database; A = City of Amarillo wells in the Carson County Wellfield; M= randomly placed "make-up" wells; P = Pantex Plant wells.

†† Means and standard deviation based on the difference between heads predicted by MODFLOW and the 1990 known heads for the 3,600 cells composing the Pantex Plant-Amarillo Carson County Wellfield vicinity, each weighted by proportional area. "Mean error" includes positive and negative values while the "mean absolute" uses only absolute values.

*All storage calculations based on storage coefficient of 0.15. Based on the 1990 known water-table surface, the 1990 volume in storage for the 3,600 cell area = 5.92×10^9 m^3. This compares to the 1959-60 "predevelopment" volume in storage = 7.85×10^9 m^3 for a known 1960-1990 change in storage of -1.93×10^9 m^3.

These conditions required 1,165 phantom wells to be added to meet the maximum known withdrawal rate for the total period of record. The phantom wells were placed randomly on a county-by-county basis (fig. 24).

The results of simulation K are depicted in figure 25. The overall results were a close match between the known and simulated water-table surfaces. Compared to simulation J, the use of phantom wells led to fewer closed and jagged contours because the withdrawals were spread among more wells. The measures of discrepancy are shown in tables 7 and 8. The most marked improvements were in the mean error terms for the entire model and the close-up area and in the percentage of known withdrawals through the wells, which rose to 86.0 percent. The simulation K results for water in storage at the end of 1990, and therefore the change in storage from 1960–90, was also much improved at 93.3×10^9 m^3 (3.29×10^{12} ft^3), or 99.8 percent of the known value.

Several other simulations were made as attempts to improve upon the performance of the wells and to minimize cell drying. Simulation L was used to evaluate the effects of different length time steps within each subperiod. Each subperiod (e.g., the 5-year period 1960–64) was further divided into 1-year length time steps. However, this achieved essentially identical results as simulation K.

In order to improve upon the model simulation with regard to the percentage of known withdrawals through wells, the well package was modified in one other fashion. In simulations J through L, it was observed that most of the wells which were going dry, whether known or phantom wells, were in areas of either low initial 1959–60 saturated thickness or low hydraulic conductivity, or both. This was the case for many wells in southwest and northeast Carson County, as can be seen by examining figures 7 and 11. Because actual wells pumping under such conditions are limited with respect to the maximum pumping rate that they can maintain without going dry (Theis, 1935), further simulations utilized a withdrawal curtailment algorithm as introduced in equation 5.

Simulation M used the withdrawal curtailment with Q^* for wells calculated with the factor F in equation 5 set to 0.1. This means that the withdrawals were curtailed such that the long-term

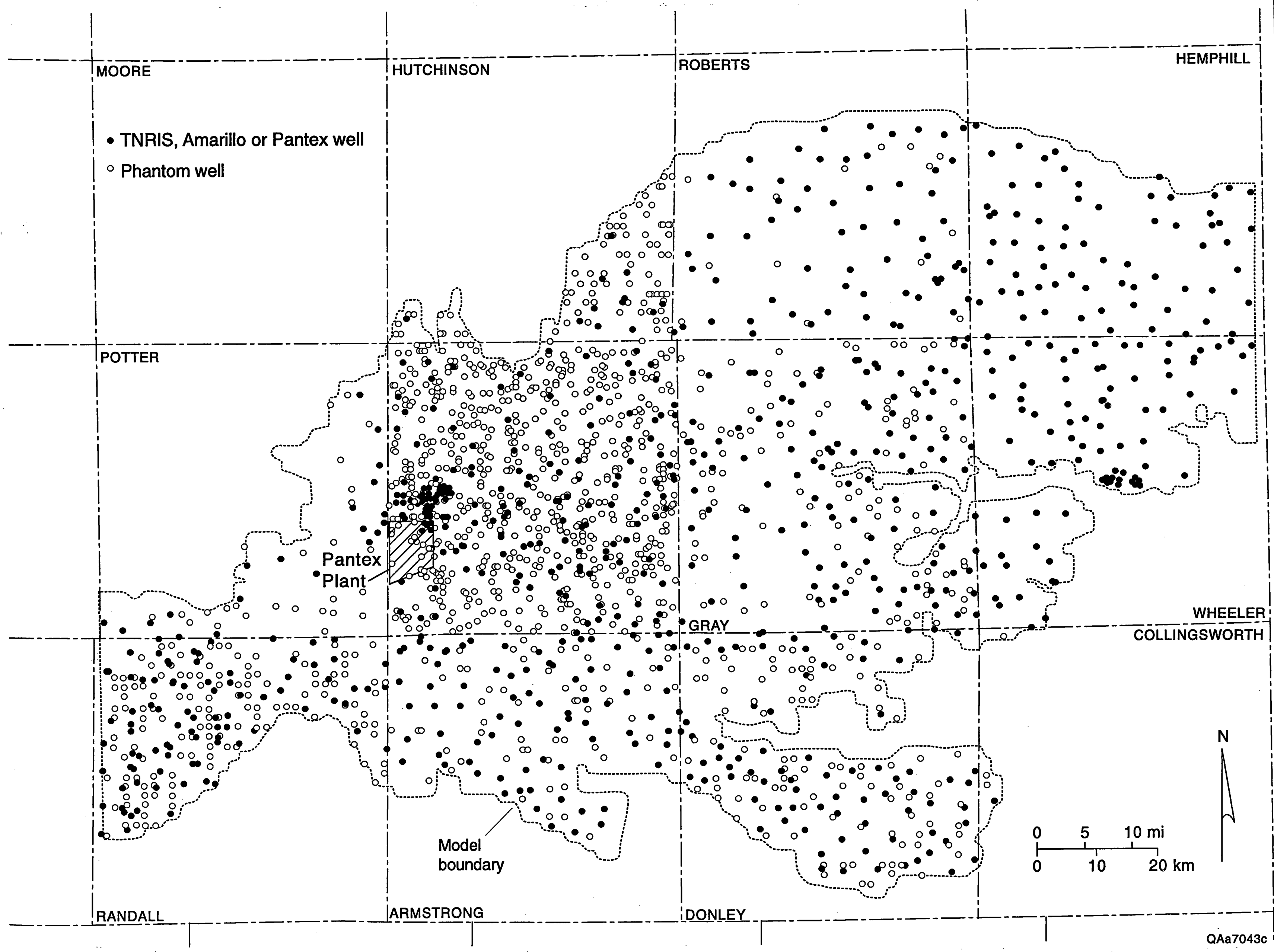

Figure 24. Map showing locations of the 1,165 phantom wells used in simulation K.

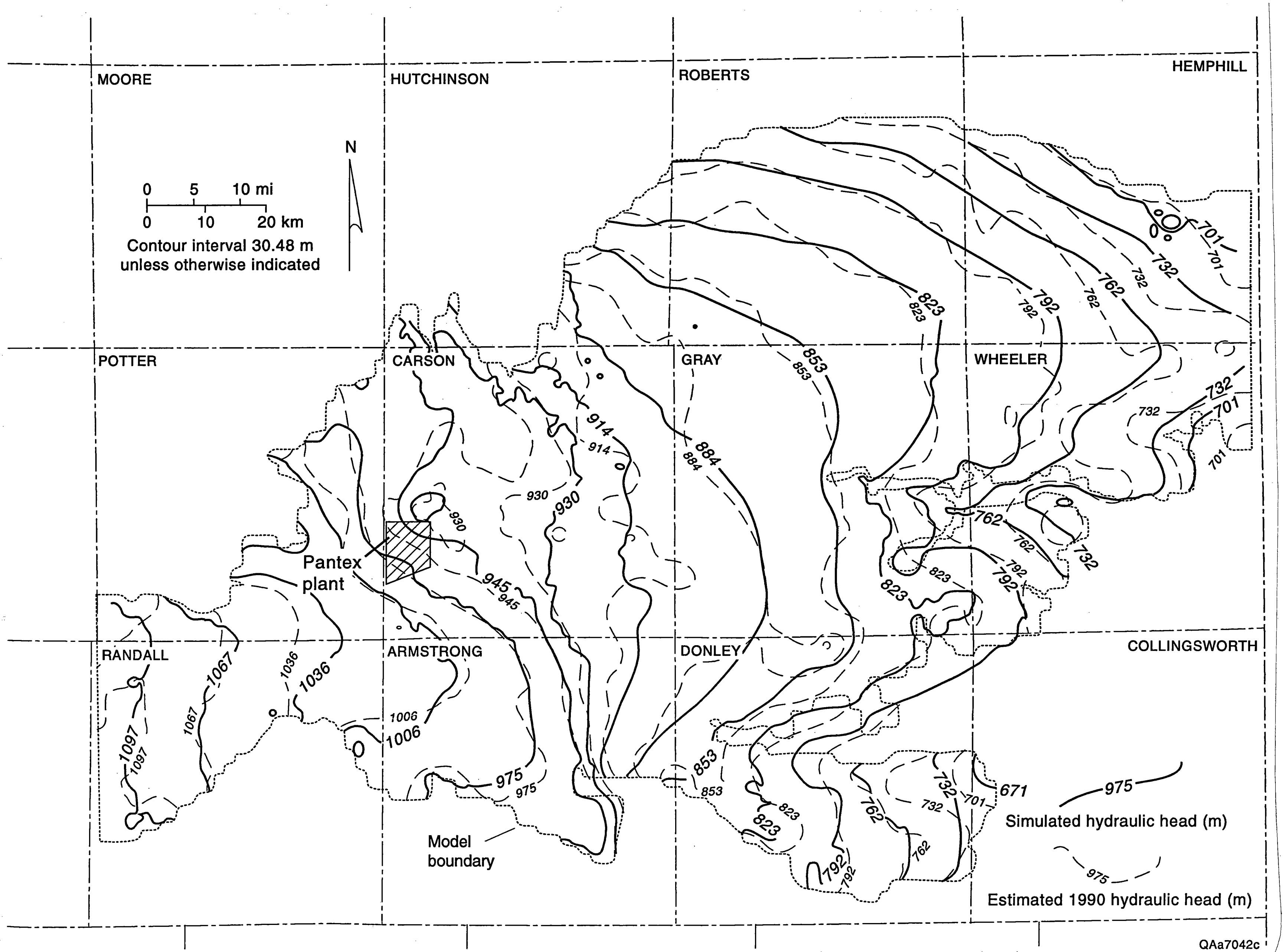

Figure 25. Simulation K water-table surface and estimated 1990 water-table surface.

decline in the water table for each well should not exceed 90 percent of the initial 1959–60 saturated thickness. The withdrawal for each TNRIS and phantom well was then limited to the minimum of the calculated Q* value or 36.35 m^3 hr^{-1} (160.0 gallon min^{-1}). The procedure described above for simulation K to determine the maximum number of phantom wells required in a county for each period was utilized with one modification: no phantom wells were placed within the boundaries of the Pantex Plant. With the withdrawal curtailment procedure it was necessary to place a total of 1,453 phantom wells based on the known withdrawal targets.

With the withdrawal curtailment in place, simulation M achieved substantial improvement with the percentage of known withdrawals achieved through the wells rising to 95.8 percent as shown in table 7. Figure 26 shows the water-table surface resulting from this simulation compared to the known surface at the end of 1990. A close-up of the water-table surface in the immediate vicinity of the Pantex Plant and the ACCWF is shown in figure 27.

The measures of discrepancy in tables 7 and 8 for simulation M illustrate a very similar performance to simulation J and K. The most notable change is that although simulation M achieves a better agreement for the well withdrawals, this leads to a slightly worse performance in regard to change in storage for the 1960–90 period. The value of -12.87×10^9 m^3 (4.54×10^{11} ft^3) is an overprediction of the known change in storage of -11.7×10^9 m^3 (4.13×10^{11} ft^3). Figure 28 illustrates in cross section the predicted water table from simulation M at the end of each time period.

The sensitivity of the modeled system to variations in storativity is illustrated in figure 29. At the scale of this map, most of the system does not appear to be very sensitive to moderate changes in storativity. The two notable exceptions occur in northwest and southeast Carson County. Because model results, for the most part, did not differ significantly with changes in storativity, the typical published value of 0.15 was used during the remainder of the simulations.

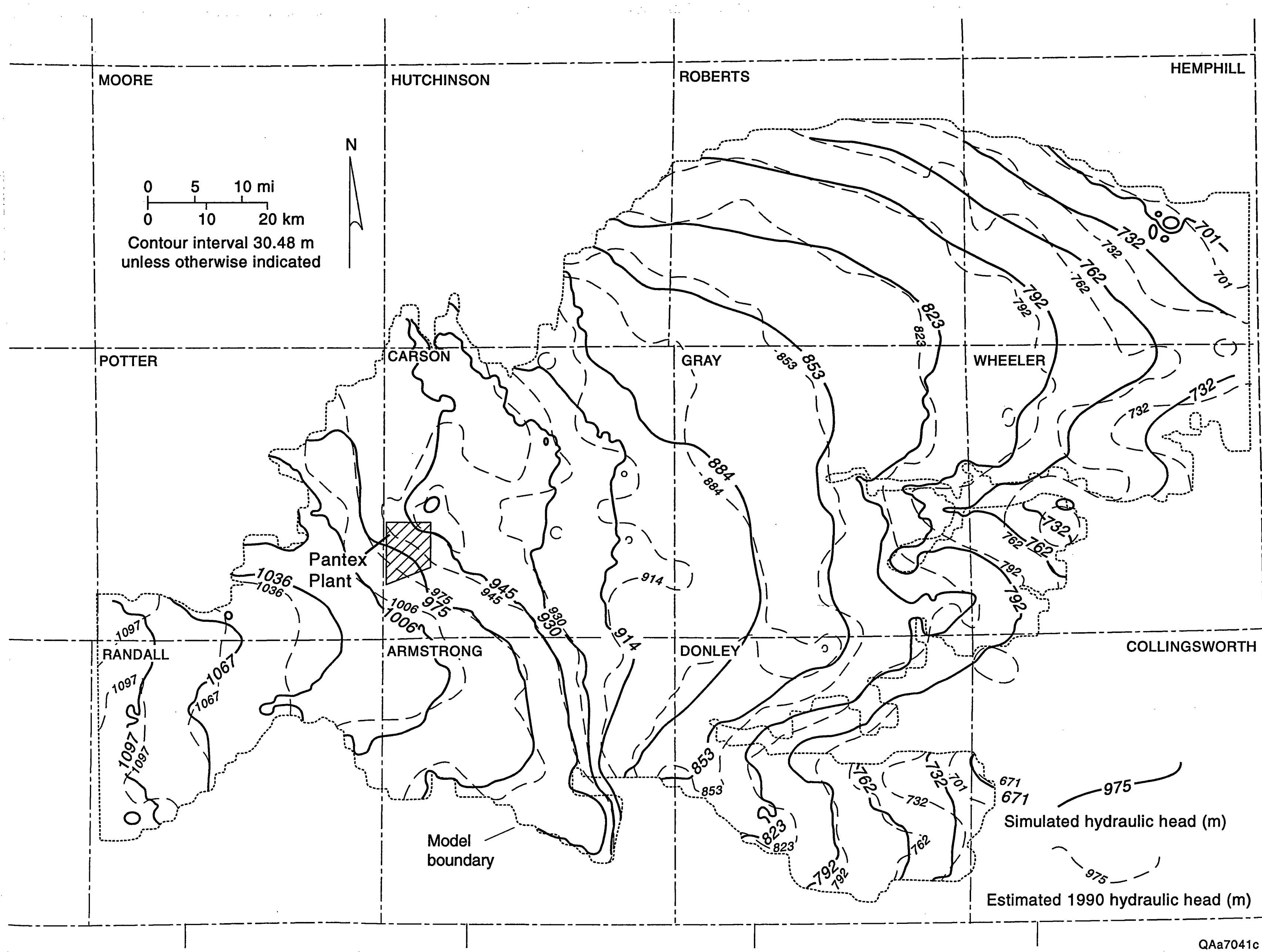

Figure 26. Simulation M water-table surface and estimated 1990 water-table surface.

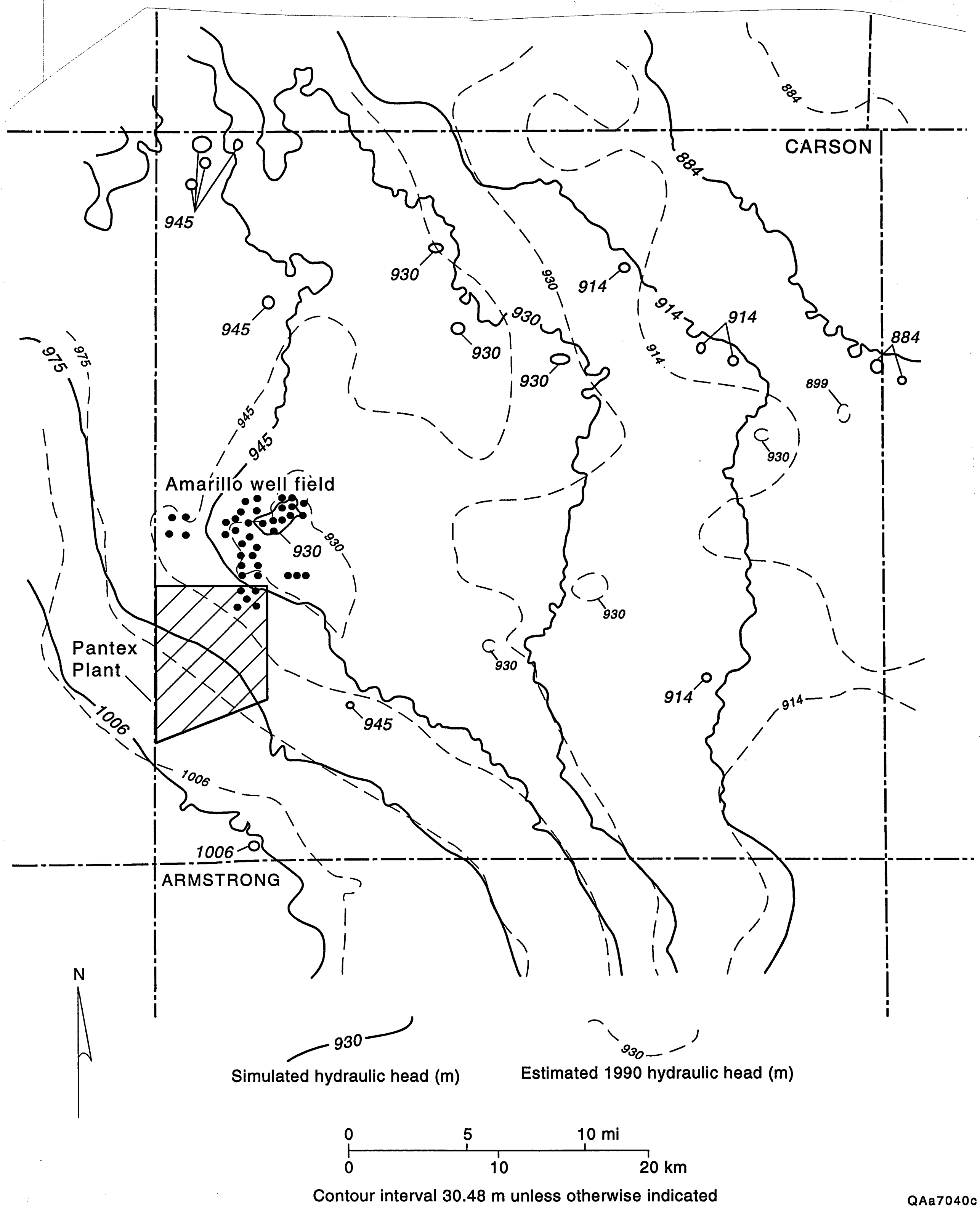

Figure 27. Simulation M water-table surface and estimated 1990 water-table surface for the immediate vicinity of the Pantex Plant and the City of Amarillo's Carson County Well Field.

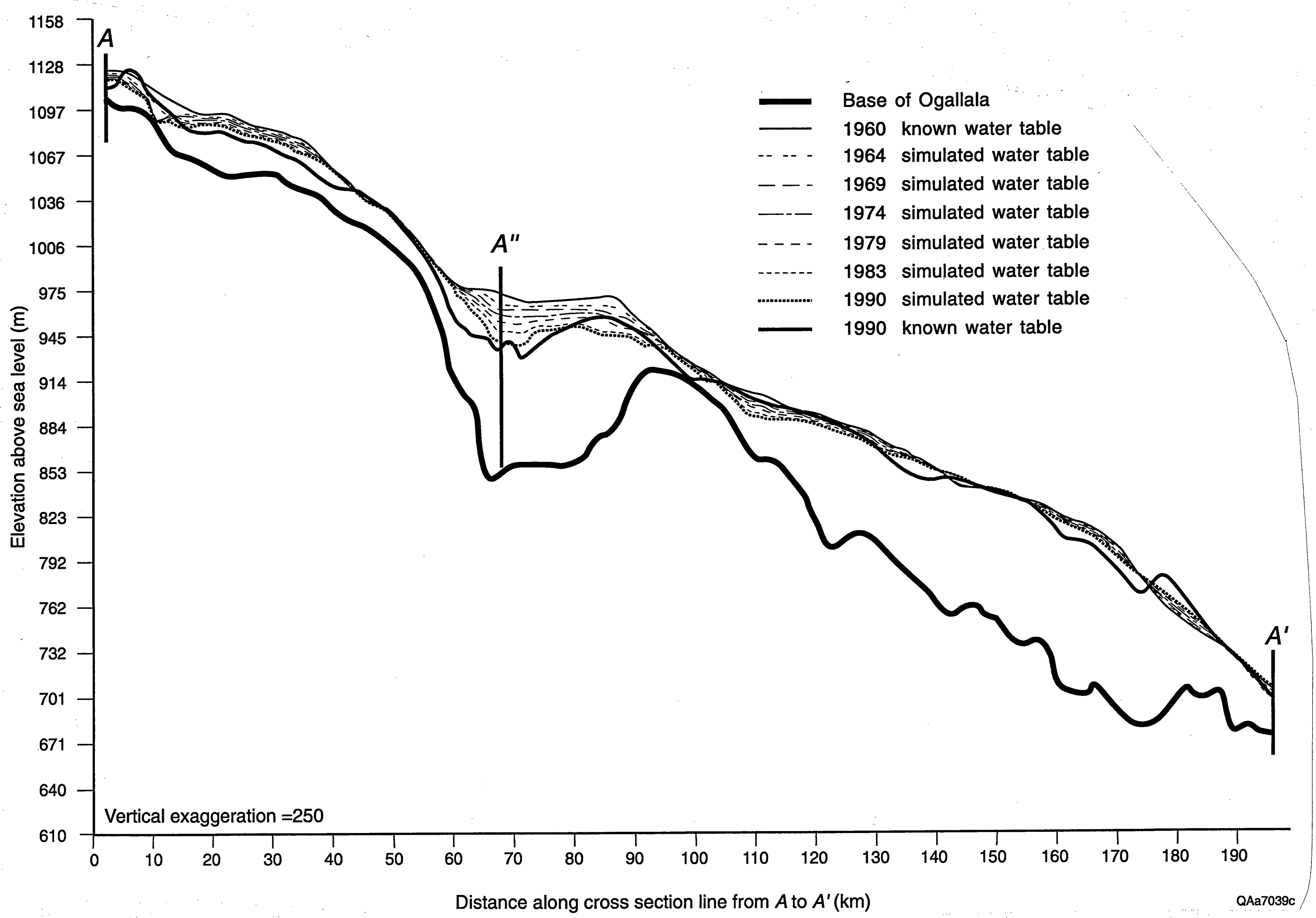

Figure 28. Cross-sectional view of simulation M illustrating predicted water-table surface from 1960 through 1990 and known 1990 water-table surface.

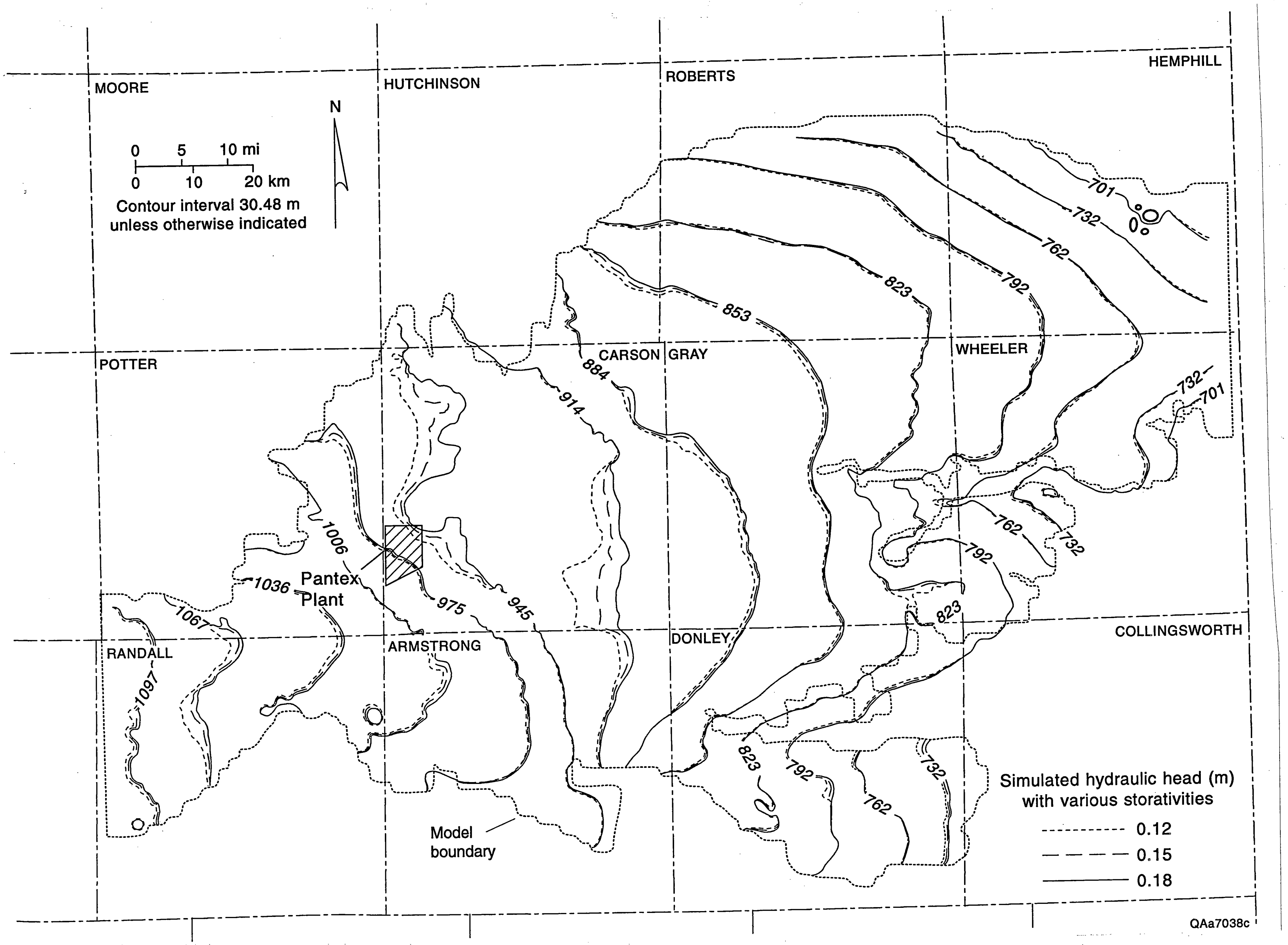

Figure 29. Comparison of simulations M, N, and O incorporating storativity values of 0.15, 0.12, and 0.18, respectively.

Particle Tracking

Flowpaths generated using hydraulic heads from transient simulation M were used to evaluate particle velocities in the area of the Pantex Plant. At least four particles were placed under each of the five playas located on the reservation and then tracked along the flowpath to the ultimate point of discharge. In each simulation, under the current hydrologic regime the points of discharge were production wells located in the northeast corner of the plant reservation or in the ACCWF to the north-northeast.

The range in effective porosity used during sensitivity analysis was determined using the empirical relationship between total and effective porosity (fig. 30) presented by Castany (1967, *in* Marsily, 1981). The dominant grain size distribution for Ogallala aquifer sediments in this area is from silt to coarse sand. Based on a range in total porosity from 20 to 37 percent as measured in the Ogallala aquifer on geophysical porosity logs from monitor well OM-105, the range in effective porosity used during sensitivity analysis was from 18 to 32 percent.

Figure 31 illustrates the flowpaths used during particle tracking simulations and traveltimes from origination points to discharge points based on an effective porosity of 25 percent. This value for effective porosity was selected because most of the saturated interval had an effective porosity equal to or greater than 25 percent. The flowpaths illustrated in figure 31 result from the significant withdrawal of ground water from the ACCWF since its construction in the late 1950's. As illustrated in the predevelopment model discussed earlier, prior to the late 1950's, the dominant direction in this area for ground-water flow was from east to west, whereas now it has shifted markedly to the north-northeast. This model predicts that all flow within the Ogallala aquifer from the area of the Pantex Plant is now being captured by production wells in the northeast corner of the plant or by wells in the ACCWF (fig. 31).

The shortest and longest distances from origination points to discharge points are from playa 3 to wells in the ACCWF and from playa 5 to Pantex production wells, respectively. Using an effective porosity of 25 percent, the total traveltime from playa 3 and playa 5 to point of discharge

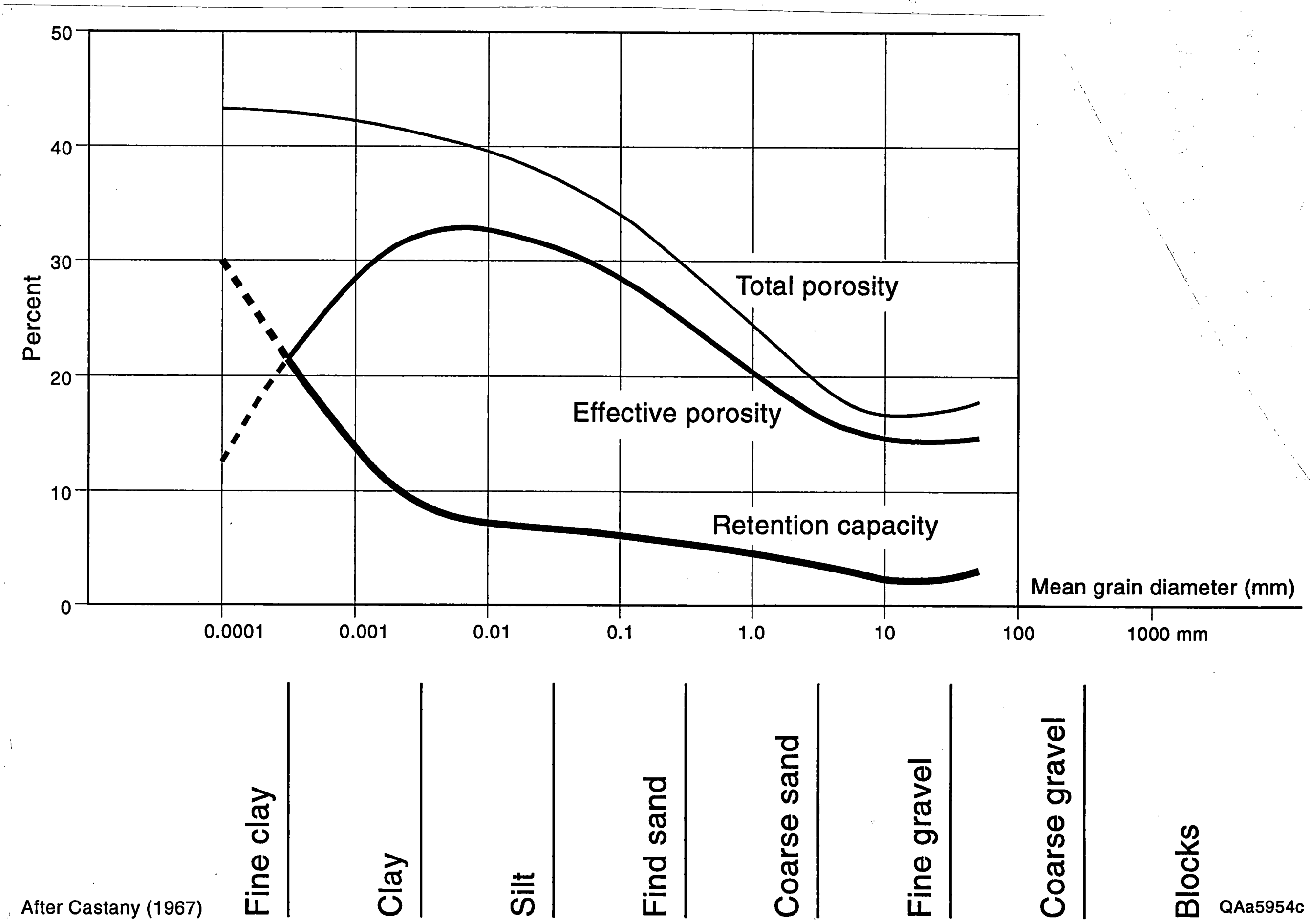

Figure 30. Empirical relationship between total and effective porosity and retention capacity based on textural distribution (after Castany, 1967, *in* Marsily, 1981).

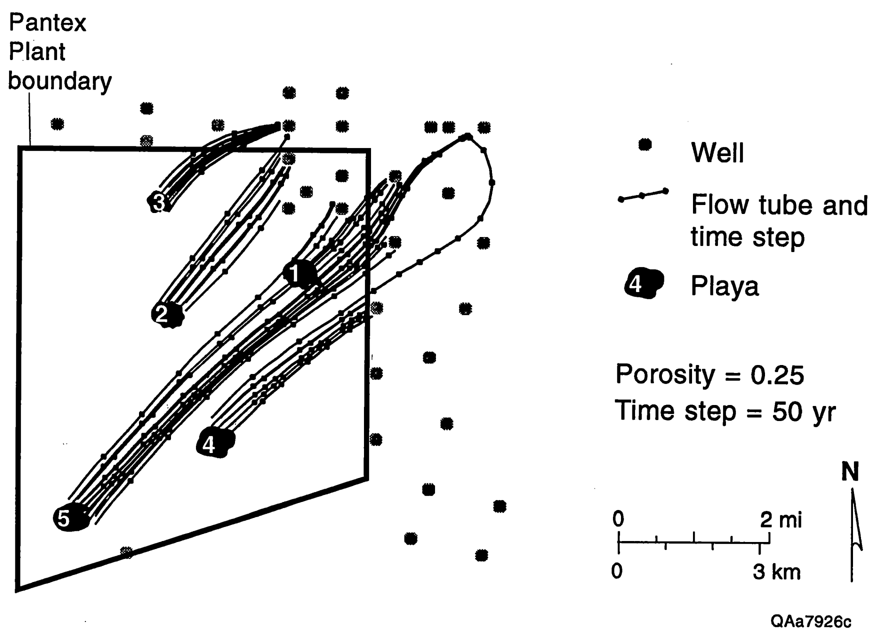

Figure 31. Results from transport simulation based on an effective porosity of 25 percent.

was 137 and 533 years, respectively. Figures 32 and 33 illustrate the total traveltimes predicted by the model for particles introduced into the Ogallala aquifer under the five playas at the Pantex Plant, based on effective porosities of 25 and 30 percent, respectively.

DISCUSSION

The historical evolution in the conceptual understanding of natural recharge to the Ogallala aquifer, from Johnson (1901) to Scanlon and others (1994) is extensive, cyclic, and possibly controversial. Wide differences with respect to rates and the spatial distribution of recharge to the Ogallala aquifer have been reported. Most investigators have attributed at least some recharge (sometimes described as minimal) to the Ogallala aquifer through playa lakes. It might be argued that various hydrologists began to describe the rate of recharge through playas and interplaya areas as insignificant or negligible only after withdrawal rates from the Ogallala aquifer for irrigation significantly exceeded natural recharge to the aquifer. In a relative sense, this was clearly the case because in many highly irrigated areas of the Southern High Plains, water-level declines were averaging in excess of 0.3 m yr^{-1} (1 ft yr^{-1}), in comparison to regionally averaged recharge rates ranging from 6 to 9 mm yr^{-1} (0.24 to 0.35 inch yr^{-1}). Weeks and Gutentag (1984) stated that the saturated thickness of the Ogallala aquifer in Texas has decreased more than 25 percent, and in 8 percent of the area the decrease has been more than 50 percent. Perhaps the prevailing attitude toward the future of the Ogallala aquifer can be summarized by the following quotes. Inman (1982, *in* U.S. Bureau of Reclamation, 1982) stated that "Water resource technologists expect underground water in Lubbock County, Texas, area to be depleted by 1994." Langford (1981, *in* U.S. Bureau of Reclamation, 1982) stated, "It is estimated that the Ogallala in Texas will be completely depleted in 40 to 50 years."

The desire to promote conservation and artificial recharge to the Ogallala aquifer apparently influenced the thinking of some authors from the 1960's to the 1980's. For example, Dvoracek and Peterson (1970) stated that there is "little opportunity for natural recharge . . . and . . . In order to preserve the usefulness of the aquifer, artificial recharge must be utilized." Another example can

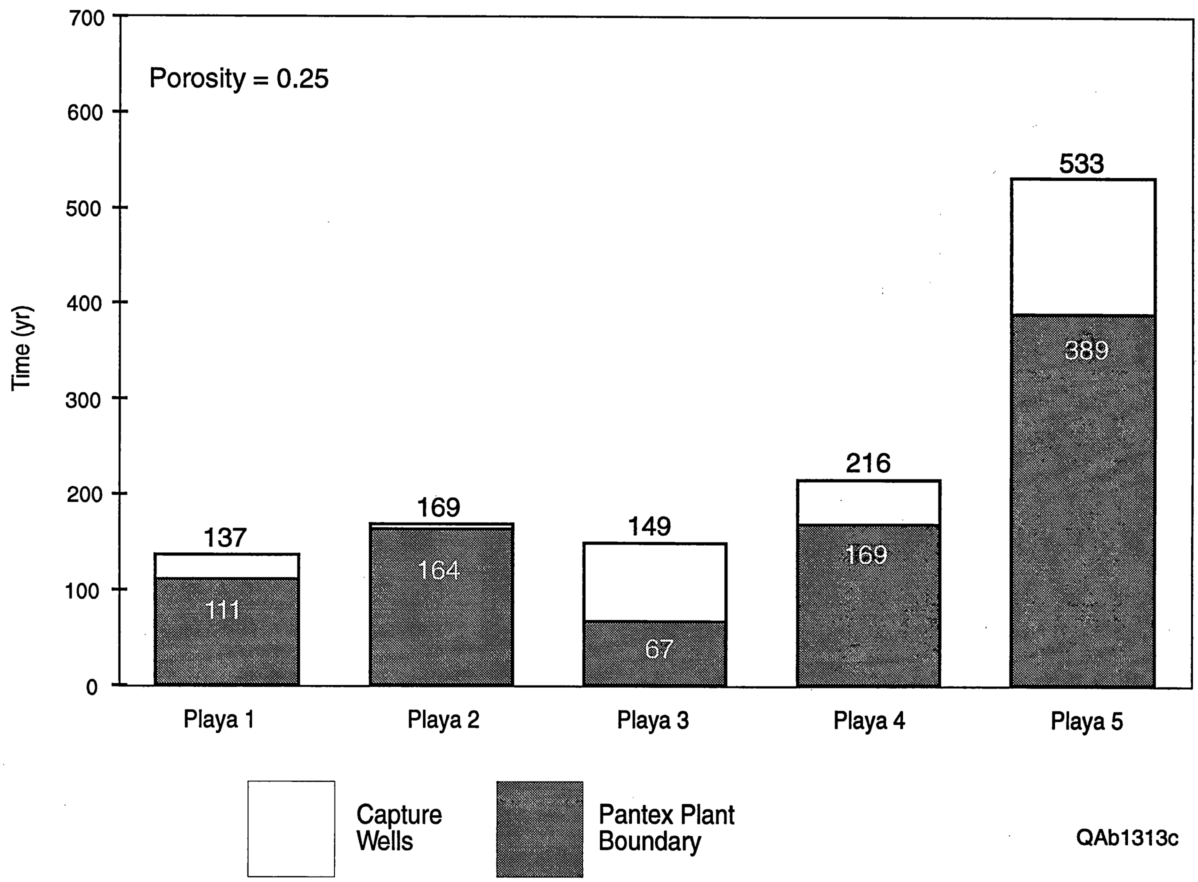

Figure 32. Graph illustrating traveltimes from origination points to plant boundary or points of discharge, whichever was encountered first, based on effective porosity of 25 percent.

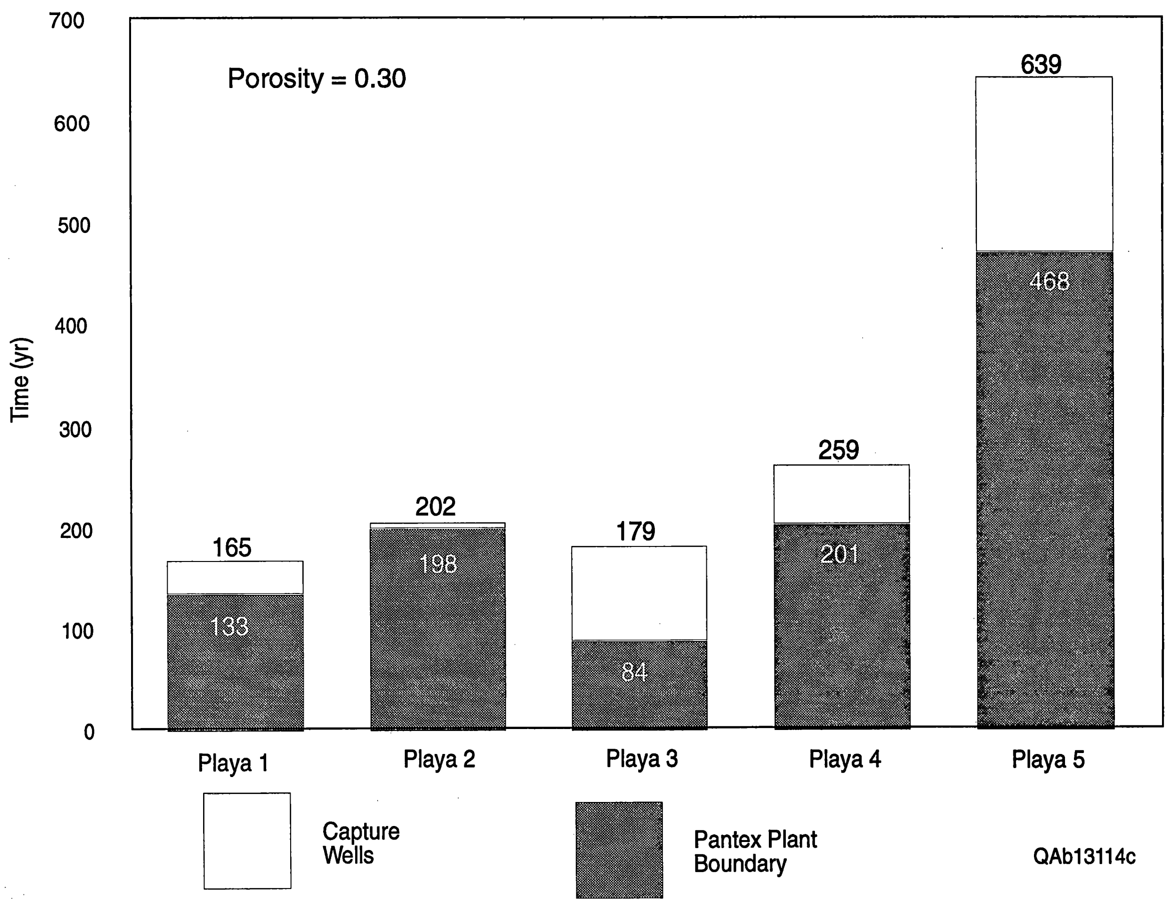

Figure 33. Graph illustrating traveltimes from origination points to plant boundary or points of discharge, whichever was encountered first, based on effective porosity of 30 percent.

be found in Knowles and others' (1984a) conclusion of negligible recharge and rapid depletion of Ogallala aquifer resources.

The use of regionally averaged recharge rates to the Ogallala aquifer in ground-water resource evaluation is and will continue to be a valid methodology. The use of regionally averaged recharge rates, however, is unsuited for understanding rates, directions, and controls on contaminant transport in areas where recharge may be locally focused or controlled by preferential pathways. Therefore, two important elements of this study should be clearly documented.

First, the numerical success (volumetric budget and history-matching) of ground-water flow simulation F, using playa lakes as sole points of recharge to the Ogallala aquifer in our study area, suggests that it is a viable model, as good as zonal recharge results from simulation E. Therefore, this model may be used to support field and laboratory observations of focused recharge through playa lake basins reported by Gustavson and others (1993, 1995) and Scanlon and others (1993, 1994).

Second, the overall results of simulation F demonstrate that there is very little difference in the resulting water-table surface when the playa recharge scenario is employed compared to zonal recharge. Although playas cover only 2.7 percent of the Blackwater Draw Formation surface in this study area, modeling results indicate that playas are a plausible route for all recharge to the Ogallala aquifer where the Blackwater Draw is present at the surface. In figure 19, with a contour interval appropriate for this regional-scale model, there are no visible mounds or peaks in the water-table surface. The hydraulic conductivity of the Ogallala aquifer, as determined in this study, appears to be high enough that the focused flow from the playas is rapidly conducted away from the recharge points.

To calibrate a predevelopment or steady-state ground-water flow model such as the one constructed for this study, data defining the predevelopment potentiometric surface are needed for comparing model results with the known surface. The data set used to construct the potentiometric surface of the Ogallala aquifer for this model was based on all water-level data available by 1960. In most areas of the model, this data set was sufficient to describe the aquifer in its predevelopment or equilibrium condition. Figure 13, however, illustrates an area where the potentiometric surface in 1960 does not appear to represent predevelopment conditions. Immediately downgradient from the Pantex Plant to the northeast, the water table appears to be anomalously flat (very low hydrologic gradient) in an area where the Ogallala Formation and the saturated section are very thick. This is also the approximate location of the ACCWF, which went into production in the late 1950's. Ground-water withdrawal for irrigation was also initiated at approximately the same time. Although this area had been in production for only 2 to 3 years, withdrawals may have been sufficient to reduce the hydrologic gradient on the water table in this area to the water levels measured in 1959 and 1960. Another possible contributing factor not incorporated into the model is ground-water production at the Pantex Plant. Clearly, some unknown volume of ground water was produced from 1942 to 1945 and from 1951 to 1960 during initial operations at the plant. No records exist, however, quantifying this production period and its potential impact on local hydrologic gradients are unknown.

In future modeling efforts, it may be beneficial in this area to use initial water levels measured in 1957 and 1958 instead of those measured in 1959 and 1960 while mapping the predevelopment water table. This alternative might provide better agreement between known and simulated water tables under predevelopment conditions.

For the predevelopment stage of modeling, the main focus of this study was to construct an accurate, deterministic ground-water flow model to examine the plausibility of various recharge scenarios. Clearly, input data sets such as the base of the aquifer contain ample data. The data set for hydraulic conductivity, although based mostly on published aquifer test results, is still deficient in the spatial distribution of hydraulic data.

For the transient modeling stage, the focus was to examine one of the recharge scenarios versus the substantial withdrawals of water through wells in the model area for the period 1960–90. Also important was the hydrologic relationship of the ACCWF to the changes in the water table in the vicinity of the DOE Pantex Plant. Clearly for both of these goals, the number, location, and withdrawals of wells through time is a critical set of data. Obviously, more accurate data would be beneficial, but the use of randomly placed phantom wells is a satisfactory approach.

Particle tracking modeling focused on the predicted direction and rates of ground-water flow in the immediate area of the Pantex Plant. Two elements of this modeling effort warrant discussion. First, the traveltimes predicted here do not take into consideration the time required for movement from land surface to the actual top of the water table. This time may be considerable if a perched aquifer, for example, is present beneath the point where the water enters the subsurface. The "clock" for particle tracking begins only after the particle has actually entered the aquifer. Therefore, the traveltimes presented in figure 32 could probably be considered to be minimum values, because the higher the effective porosity the slower the velocity, and because a majority of the aquifer sediments in the area appear to have effective porosities greater than or equal to 25 percent. Second, it is important to note that these results are strictly for advective decay and do not include the effects of dispersion, dilution, and sorption on potential contaminant transport.

It is noted that the results are based on total porosity values from only one monitor well. It is recognized that the Ogallala aquifer, even at a local scale, can be a very heterogeneous unit and the impact of these heterogeneities has not been evaluated. Only the sensitivity of the system to changes in overall effective porosities was evaluated. In addition, once dispersion, dilution, reactions, and sorption decay factors become known for the area, contaminant transport modeling will be warranted.

CONCLUSIONS

Planar, finite-difference, numerical ground-water flow models were constructed to evaluate three different conceptual models of recharge to the Ogallala aquifer and to evaluate the hydrologic relationship between the ACCWF and changes in water-table elevations in the region of the Pantex Plant. These conceptual models included spatially uniform recharge, in which recharge was applied at an equal rate throughout the model area; zonal recharge, in which recharge was varied on the basis of regionally mapped geologic units; and modified zonal or *playa-focused* recharge, in which all recharge was focused through the playas in areas where playas are present at the surface and recharge was distributed uniformly in areas with no playas. Previous ground-water flow models applied recharge uniformly throughout the aquifer or zonally, depending on soil types or precipitation rates. From a regional volumetric balance perspective, this approach (uniform or zonal recharge) is valid; however, in a system where focused or preferential flow dominates, the spatial distribution and velocity of a recharge pulse may be significantly different from that which is predicted using uniform or zonal recharge models.

Results using modified zonal recharge in which a regional rate of 6 mm yr^{-1} (0.24 inch yr^{-1}) was focused through the playas for an effective recharge rate of 219 mm yr^{-1} (8.6 inches yr^{-1}) and in which a uniform 9 mm yr^{-1} (0.35 inch yr^{-1}) was used where no playas are present indicate that agreement achieved using modified zonal recharge was as good as that using zonal recharge. The degree to which this recharge model results in the best fit between known and simulated water tables is not the critical issue of this investigation, however. The critical point is that a modified zonal recharge using playas as the focal points of recharge is a viable numerical model that has results compatible with those from ongoing investigations of spatial distribution and controls on recharge described by Scanlon and others (1993) and Gustavson and others (1993).

Although playas cover only 2.74 percent of the Blackwater Draw Formation surface in this study area, the simulation results indicate that they are a plausible route for all recharge to the Ogallala aquifer in areas where the Blackwater Draw Formation is present at the surface. The hydraulic conductivity of the Ogallala aquifer is high enough that the focused flow from the playas is rapidly conducted away from the recharge points, based on the absence of mounds on the potentiometric surface. This absence of mounds, however, may simply be the result of the course contour interval selected. In some areas, as is the case at the Pantex Plant, focused recharge may also be diffused in the unsaturated zone by the presence of low-permeability horizons that result in perched aquifers. In these areas, the focused recharge spreads out radially before moving vertically through the low-permeability horizon.

A second goal of this modeling effort was to investigate the hydrologic relationship of the ACCWF to the changes in the water table in the vicinity of the DOE Pantex Plant. The transient simulations add further support to the plausibility of the *playa-focused* recharge scenario by adequately matching the changes in water-table elevations throughout much of the study area. These simulations also suggest that the changes in water-table surface in the vicinity of the Pantex Plant are dominated by changes due to the withdrawals of the ACCWF. A significant impact of this withdrawal has been the change in direction of ground-water flow at the Pantex Plant from predominantly east to west to a north-northeast flow direction toward the ACCWF.

Particle tracking results predict that any contaminants that may eventually enter the Ogallala aquifer under the Pantex Plant will require a time period from approximately 60 to 400 years to travel from probable points of origin to discharge or capture points, based on an effective porosity of 25 percent.

ACKNOWLEDGMENTS

This project was funded by the U.S. Department of Energy under Grant No. DE-FG04-90AL65847. Thomas C. Gustavson, Robert E. Mace, Jay A. Raney, and Bridget R. Scanlon provided helpful reviews of this report. Others contributing to this report were Susan Lloyd, who provided assistance in word processing, and William C. Bergquist, Michele Bailey, Susan Krepps, Kerza Prewitt, and Randy Hitt, who drafted figures. Daryl Peckham and John Ashworth of the Texas Water Development Board provided data regarding county-by-county well withdrawals for the study area. Gary Denton, of the City of Amarillo, provided valuable information regarding withdrawals from the ACCWF. C. E. Williams of the Panhandle Underground Water Conservation District No. 3 provided valuable insight into the number of and typical withdrawal rates of wells.

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APPENDIX A

RECHARGE TO THE OGALLALA AQUIFER—A REVIEW

Johnson (1901) recognized two primary areas of recharge to the Ogallala aquifer: (1) the numerous basins that collect runoff water from precipitation (thus reducing evaporation and increasing absorption) and (2) the numerous sand and gravel surface exposures. One of the earliest proponents of the vast economic potential of the natural resources of the High Plains, Johnson (1901) discussed at length the possibility of using playa lakes as water storage basins. He recommended several solutions to prevent the "rapid loss by ground absorption" in the floors of the playa lakes. One proposed route was basin modification to enhance runoff into the larger playa lakes, wherein a large volume of silt that washed onto the playa floor would act to restrict absorption.

Gould (1906), in a discussion of the topography of the Llano Estacado, stated that since the land is so flat, drainages are wholly underdeveloped, and thus any significant precipitation runs into the depressions (playas), where it escapes either through evaporation or through seepage to the ground water. Gould (1906) also noted the occurrence of perched aquifers (local aquifers that lie above the main Ogallala aquifer), which he described as "first sheet and second sheet" or "first and second water." He stated that they were common occurrences in the area.

Baker (1915) was perhaps the first to recognize that desiccation cracks in the playa floors act as preferential flow paths for recharging waters to the Ogallala aquifer. In his discussion of recharge to the Ogallala aquifer, Baker (1915) listed both the large and small playa lakes and streams as some of the primary points of recharge.

Theis (1937) was one of the earliest to suggest that recharge was not occurring uniformly across the Southern High Plains. Instead, he suggested that most of the recharge occurred in areas where the Ogallala Formation was exposed at the surface or in areas where it was overlain by sandy permeable sediments. Later, however, Theis (1971) stated that although the playa lakes are "one of the chief sources of recharge of the High Plains under natural conditions," the "rate of infiltration under natural conditions is very slow."

Broadhurst (1942) described four areas of enhanced or focused recharge to the Ogallala aquifer: playas, sand stream beds and their adjacent flood plains, and sand-hill areas. Field data collected as part of this investigation included an extensive drilling program to determine the extent and nature of caliche in playas, the installation and monitoring of stake gages in playas, and the monitoring of water levels in numerous wells in upland (interplaya) areas and areas adjacent to playas, stream beds, and sand hills. Perhaps the most dramatic results from this effort were the clear differences between water-level responses in wells located in playas and those in uplands. After what Broadhurst (1942) described as "phenomenally heavy rains of 1941," water levels in some wells located adjacent to playas rose more than 3.0 m (10 ft), water levels in upland or interplaya areas, however, showed relatively little change.

White and others (1946) argued that the source of all modern recharge to the Ogallala aquifer is precipitation followed by infiltration. This discussion was counter to the prevalent belief that the Ogallala aquifer was a vast underground river running west to east and being continuously recharged in the Rocky Mountains, essentially an infinite source of ground water. It is interesting to note that some 40 years later, Knowles (1985) reported this fallacy was still a common belief in the area. White and others (1946) concluded that the water resource was not infinite and that as early as 1940, significant declines in water levels were being observed because of the production of large volumes of ground water from storage, mainly for irrigation.

White and others (1946) also delineated the probable areas of most of the recharge to the Ogallala aquifer, partly because of direct observations of water-level responses in playa lakes and in monitor wells adjacent to playa lakes. The primary areas of recharge listed are depressions or sinks (playa basins), sandy stream beds, sandy flood plains, and sand dunes. Within the playa lake basins, White and others (1946) thought that most recharge occurred through desiccation cracks in the lake bottom for the first few days after significant precipitation, through smaller sinks present as a result of solution channeling, and through a sandy belt surrounding the floor. White and others (1946) provided extensive precipitation and water-level data from five Ogallala aquifer water wells located close to playa lakes in Deaf Smith, Floyd, Hale, and Lubbock Counties. These data illustrate the direct response of aquifer water levels to the recharge from playa lakes after local precipitation events (fig. A-1). Additional water-level data illustrating the response of the Ogallala aquifer to recharge through the playas was also provided by Rettman and Leggat (1966) for a well and playa in Gaines County and by the U.S. Bureau of Reclamation (1982) for a study on playa lakes.

Barnes and others (1949) stated, "Water level measurements in wells have proved conclusively that a large proportion of the recharge to the underground reservoir is derived from surface water that collects in the thousands of depression ponds on the High Plains." This conclusion was based on the rise in water levels in wells reported by White and others (1946). Variable rates of recharge within individual depression ponds (playa lake basins) were attributed to the extent of caliche development and variations in permeability, the presence of silt, shrinkage or desiccation cracks, solution channels in the caliche, and the presence of "a belt of sandy material that absorbs water readily surrounds these deposits" (fine-grained silts and clays lining the bottom of the playa floors).

Cronin (1961, 1964) suggested that natural recharge to the Ogallala aquifer in Texas was from two sources: underflow from New Mexico (which should remain fairly constant) and direct infiltration of precipitation. Factors affecting the rate of direct infiltration include the amount, spatial distribution, and intensity of the precipitation event and the type of soil and vegetative cover. In a study of 50 playas, Cronin (1961) estimated that 35 percent of the surface water collected in playas eventually infiltrated into the subsurface and recharged the aquifer. Cronin (1961), like White and others (1946), also recognized recharge through the annular rings of playas (his term was "sandy-soil belt of the playa lake"). Water-level responses in wells adjacent to playas after precipitation events reported by White and others (1946) were also explained by Cronin (1964) in a two-part model. First, water percolates through the desiccation cracks in the playa

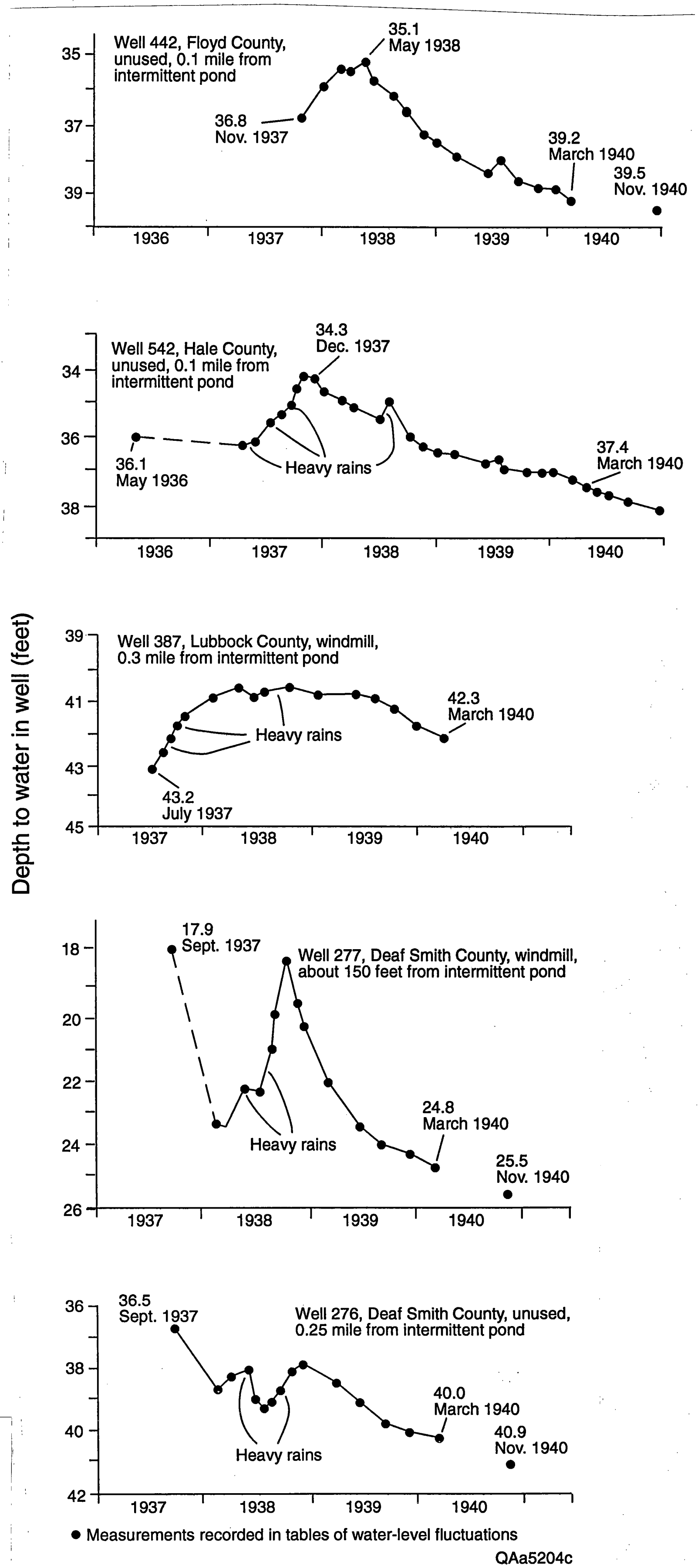

Figure A-1. Response of selected wells in the Southern High Plains to recharge through nearby playas (modified from White and others, 1946).

floors as runoff flows into playas. If surface runoff is sufficient to raise water levels in the playa to the elevation of the sandy, higher permeability annular belt, then rapid infiltration into this zone would occur.

Schwiesow (1965) estimated that between 1,233 and 1,845 hm^3 (1 and 1.5 million acre-feet) of water entered the approximately 37,000 playas annually as runoff. He also estimated that less than 10 percent of this runoff reached the aquifer by percolation through the playa sediments and soils. This statement, however, seems to contradict his description of playa floors as being "practically impermeable" (Schwiesow, 1965, p. 2).

Havens (1966), working primarily in the same area as Theis (1937) (eastern New Mexico), was another on the list of investigators to propose that most of the natural recharge was focused in the floors of the numerous playas in the area. After one precipitation event at a playa that was instrumented to monitor water levels both in the playa and at the water table, only 15 percent of the water loss from the playa lake was attributed to evaporation, while 85 percent was attributed to infiltration into the aquifer. This focused recharge resulted in temporary mounding of the water table beneath the playa after significant precipitation events. During the monitoring period, water levels in two water wells adjacent to the playa lake rose 0.3 and 0.55 m (1.0 and 1.8 ft). In addition to natural recharge, seepage from irrigation (or artificial recharge) was also recognized as contributing to the overall recharge process. Havens (1966) estimated that 20 to 80 percent of the water that collected on the floor of the playas during normal years moved into the subsurface to recharge the Ogallala aquifer. Havens (1966) also noted that recharge rates through playas were higher when water levels in the playas extended up on the slopes of the playas or on the annular region of the playa lake, where the soils were observed to be more permeable.

Throughout the 1960's to the early 1980's, an extensive effort was made to educate the public about water conservation and to develop a cost-effective method of artificially recharging the Ogallala aquifer. A representative sample of the results of investigations conducted at the Agricultural Research Service Center of the U.S. Department of Agriculture and the Texas

Agricultural Experimental Station at Bushland, Texas, a division of Texas A&M University, is reviewed here.

A common theme throughout the research effort at Bushland was that current rates of ground-water production were so much greater than natural ground-water recharge that essentially all of the ground water being produced was being mined or removed from storage (Hauser and Lotspeich, 1967). For example, Aronovici and others (1970) stated, "Natural recharge through the slowly permeable hardland soils of the High Plains is negligible, and the Ogallala is cut off from any other appreciable water source." Aronovici and Schneider (1972) demonstrated that no natural (nonirrigation) percolation was observed in the spatially dominant Pullman soils that occupy upland areas that surround playa lake basins on the Southern High Plains. Even areas where significant surface saturation from irrigation practices had occurred did not show deep percolation of water through the unsaturated zone. It is interesting to note that the Pullman soils are basically restricted to what are defined in this report as interplaya areas. Thus, Aronovici and Schneider (1972) concluded that recharge is not occurring in interplaya areas. Two reasons given for this absence of recharge were low infiltration rates of the fine-textured soils and the high ratio of evaporation to infiltration. On the basis of measurements made over time in several different natural and altered settings, they found no deep percolation except in very flat, irrigated cropland.

Clyma and Lotspeich (1966) reported that the principal source of water to recharge the Ogallala aquifer is surface water that collects in the playa lakes. They went on to report, however, that only a small volume of the surface water recharges the aquifer under natural conditions. The amount of surface water collected annually in the playas has been estimated to range from 2,219 to 4,562 hm^3 (1.8 to 3.7 million acre-ft) (Clyma and Lotspeich, 1966). Hauser (1966), in an experiment to quantify the amount of seepage that occurs from surface water in a playa, estimated that only 15 percent of the surface water infiltrates to the ground water. This estimate is based on a comparison of pan evaporation rates versus volumetric changes measured in surface water at a playa at Bushland. Reddell and Rayner (1962), *in* Hauser (1966), reported that in a study of five playa lakes in the Lubbock area, 54 to 84 percent of the surface water infiltrated to recharge the aquifer. Hauser and Lotspeich (1967) appeared to contradict this theme by stating that most of the surface water in playas evaporates.

This conceptual model of playas as evaporation pans was addressed by Lehman and others (1970) and Lehman (1972) in an investigation of water quality in playas used by cattle feedlots for waste-water impoundment. Lehman and others (1970) illustrated the absolute nature of the acceptance of the "evaporation pan" model by their statement: "Consequently, some cattle feeding operations are using playas as natural lagoons to impound all runoff from their operations. . . . at least historically, all water has been allowed to evaporate."

Several investigations have focused on possible techniques to artificially recharge the Ogallala aquifer, typically using surface runoff that collects in playa lakes. Although enhanced recharge to the Ogallala aquifer is not included in the models presented in this report, results from some of these artificial recharge experiments are worth noting because of their applicability to enhanced recharge that has occurred at the Pantex Plant through the numerous ditches and playas. Schneider and Jones (1984) reported results of long-term artificial recharge experiments in excavated basins constructed on the slopes and floor of a playa at Bushland. These small basins were excavated to a depth of 1.2 to 1.3 m (3.9 to 4.3 ft) and then flooded with both clear-well and turbid-playa water to evaluate their potential as artificial recharge basins. The excavation depth was selected in order to remove the relatively low permeability Pullman soils. During these tests, soil moisture measurements indicated that the wetting front was advancing at a rate of 0.13 to 0.14 m hr^{-1} (5.1 to 5.5 inches hr^{-1}) with long-term recharge rates ranging from 0.37 to 0.43 m d^{-1} (14.6 to 16.9 inches d^{-1}). Desiccation cracks up to 10 mm (0.39 inch) in width and 100 mm (3.9 inches) in depth were observed to significantly increase the rate of recharge through the floor of the excavated basins. Significant perched aquifers developed as the result of these recharge experiments on top of low-permeability carbonate zones.

Investigations have also been conducted to determine the potential impact that agrochemicals (pesticides and fertilizers) might have on the Ogallala aquifer as a result of artificial recharge activities (Felty and others, 1972). Their conclusion was that, although some agrochemicals were present in the playa lake waters, the overall quality of playa lake water was superior to Ogallala aquifer ground water and therefore posed no negative impact on the aquifer.

It has been estimated that 90 percent of all precipitation runoff collected in playa lakes is lost to evaporation and transpiration (Ward and Huddleston, 1972). In a 4-year study of 11 playa lake basins in Lubbock County, Ward and Huddleston (1972) reported that infiltration rates on the Southern High Plains directly relate to the amount of clay in the upper foot of soil. They also observed enhanced infiltration in the annular area of the playa lakes and noted that in all of the playa lake basins studied, initial infiltration rates always decreased rapidly to a point and then increased slightly before leveling off at a fairly constant rate.

In a study to evaluate the possibility of artificial recharge to the Ogallala aquifer of Texas and New Mexico, Brown and Signor (1972, 1973) and Brown and others (1978) stated that natural recharge may be higher in sandy soil and in sand dune areas. Regionally, however, they stated that recharge to the Ogallala aquifer is insignificant relative to the rate of withdrawal for agricultural purposes.

The "evaporation pan" conceptual model was also noted by Bell and Sechrist (1972) in such descriptions as "there is little infiltration into underlying strata, and thus most of the water is lost through evaporation" and "water in the shallow lakes could only be expected to remain for a few days." A later study in Carson County by Bell and Morrison (1979), however, concluded that it is probable that naturally occurring recharge to the Ogallala aquifer as a result of direct infiltration of precipitation is currently higher than it has been in the recent past. They attributed this accelerated rate of recharge to changes in the soil and land surface as part of agricultural practices related to irrigation. These include such practices as removing certain deep-rooted plants, deep plowing of fields, plowing of playa lake floors, contour farming, irrigating before precipitation events (resulting in initially higher moisture content), and increasing the humus level (Bell and Morrison, 1979). Enhanced or accelerated recharge was also attributed to the recirculation of irrigation waters in the area.

The U.S. Bureau of Reclamation (1982) reported that the playa lakes are the major source of natural recharge to the Ogallala aquifer but that most of the surface water that collects in the playa lakes is lost through evaporation because of the relative impermeability of the lake bottoms. In this study of playa lakes on the Llano Estacado, 36 playa lakes were selected for instrumentation with various types of monitoring equipment, such as rain gauges, staff gauges, and water-level recorders, for the measurement of surface water levels through time. In addition, LANDSAT data were used to determine the presence or absence of surface water in all playas in the study area. For the monitor period from 1972 to 1981, the wettest and driest periods were determined on the basis of precipitation records. The percentage of playa lakes that held water in the respective periods (wet or dry) was then determined. For example, during the dry period, 7 of 535 and 10 of 752 playa lakes held water (1.3 percent) in Carson and Gray Counties, respectively. During the wet period, however, 50 of 535 and 43 of 752 playa lakes held water (9.3 and 5.7 percent) in Carson and Gray Counties, respectively. Throughout the entire study area of this report, 15 percent of the playas held water during the wet period and 2 percent of the playas held water during the dry period. Because even in the wettest period of this study, 85 percent of the playa lakes did not hold water after significant precipitation events, infiltration rates in these playas may be higher than the rates suggested in their conclusions.

The U.S. Fish and Wildlife Service (1980), *in* the U.S. Bureau of Reclamation (1982), classified playa lakes into three categories: (1) perennial, which contain water more than 9 months a year, (2) intermittent, which hold water 3 to 9 months a year, and (3) ephemeral, which contain water less than 3 months a year. On the basis of this classification, 33 percent of the playa lakes were reported to be perennial, 26 percent to be intermittent, and 41 percent to be ephemeral (U.S. Fish and Wildlife Service [1980] *in* the U.S. Bureau of Reclamation [1982]).

In contrast, Knowles and others (1984a) concluded that recharge to the Ogallala aquifer occurs principally by infiltration of precipitation on outcrops. Similar conclusions were reported by Weeks and Gutentag (1984), who observed, "Recharge . . . is entirely from precipitation and seepage from streams." No mention is made of playa lakes being a significant or focal point of recharge. In addition, Knowles and others (1984a) also suggested that the caprock calcrete at the top of the Ogallala Formation would act as a barrier to recharge. Knowles (1985), however, reported that recharge is not uniform, is probably controlled by soil type, and that measured water-level declines have been much less than expected from estimates of pumpage from the aquifer. He attributed this discrepancy to one of two possibilities: either the recharge rates used in the calculation are too low or the pumpage rates are too high.

Wood and Osterkamp (1984a, b, 1987) and Osterkamp and Wood (1987) expanded on Broadhurst (1942), White and others (1946), and Barnes and others (1949) to conclude that much of the recharge to the Ogallala aquifer of the Llano Estacado was through the playa lakes (they estimated to be approximately 30,000). Furthermore, they specified the small annulus surrounding the playa lake floor as the principal route of recharge, as had been previously proposed by White and others (1946), Barnes and others (1949), Cronin (1961, 1964), Havens (1966), and Conselman (1970). The downward movement of recharging ground water through the playas is the critical element of the model proposed for the formation of playa lake basins on the Southern High Plains by Osterkamp and Wood (1987) and Wood and Osterkamp (1987). This model essentially proposes that playa lake basins are formed by focused recharging ground water, which results in dissolution of carbonate materials in the subsurface, piping, and eluviation. This model was based on several lines of evidence, including water-level records adjacent to playas reported by White and others (1946), Rettman and Leggat (1966), and the U.S. Bureau of Reclamation (1982). They also used unpublished U.S. Geological Survey tritium data, unsaturated zone chemistry data from playa and interplaya settings, playa lake water chemistry reported by Lotspeich and others (1969), playa lake vegetation studies by Reed (1930), chloride mass-balance profiles by Stone (1984, 1990) and Stone and McGurk (1985), and water budget studies by Myers (*in* Cronin, 1964) and Havens (1966). No significant recharge is attributed to the central floors of playas because of the predominance of fine-grained sediments. The presence of deep desiccation cracks in the floors of playa lakes when they are dry is acknowledged, but the effectiveness of the floors for recharge is not presented.

Wood and Petraitis (1984) also concluded that most of the Ogallala recharge probably occurs on and around playa lakes. This conclusion was based on observations of water levels in playa lakes after precipitation events, theoretical considerations of geomorphology, and solute concentrations of water in the unsaturated zone. They also observed that very little recharge is occurring in the upland areas between playas because of the rapid runoff of any significant precipitation into the playa lakes.

Kier and others (1984) provided direct evidence of focused recharge through playas in their study of the potentiometric surface of the Ogallala aquifer in the city of Lubbock by constructing a potentiometric surface map for the year 1937. This was significantly before any modification or cultural development occurred that would have enhanced recharge to the playas and before ground-water withdrawals for irrigation initiated a steep decline curve in regional water levels. On this map (their figure 2), two significant ground-water mounds are present directly under two of the playas present in the city. This indicates that the rate of recharge is higher at these locations than the flux of recharge water away from the focal point of recharge. The same process is attributed to the formation of mounds in the potentiometric surface of perched aquifers at the Pantex Plant (Mullican and others, 1993).

Stone (1984, 1985, 1990) and Stone and McGurk (1985) used the soil-water chloride mass-balance approach of Allison and Hughes (1978) in several different geomorphic environments on the Southern High Plains of east-central New Mexico to evaluate rates of recharge assuming all recharge is by piston flow (no preferential flow). Geomorphic settings evaluated included playas, nonirrigated and irrigated cover sands, and sand hills. On the basis of results from each setting, Stone (1984, 1990) and Stone and McGurk (1985) reported that the highest local rate of recharge was in the playa and the lowest rate was in the nonirrigated cover sand area (interplaya). Stone (1990), however, stated that "transmission loss along dry channels is probably the major source of recharge to the Ogallala."

Claborn and others (1985) conducted a study to determine the frequency of significant recharge to the Ogallala aquifer from precipitation runoff into playa lakes. This investigation was based on 22 playa lake basins on the Southern High Plains for which historical records of precipitation were available. One of the significant conclusions from this study was that 27 to 43 percent of the precipitation runoff that collected in the playas was recharged to the Ogallala aquifer; the rest was lost to evapotranspiration. Claborn and others (1985) stated that the playa lakes "are characterized by a naturally occurring liner of almost impermeable clay or clay/silt." Natural recharge, according to Claborn and others (1985), takes place when sufficient runoff occurs to raise water levels above the feather edge of the clay liner, where it is exposed to a much more permeable lithologic unit, such as silty sand or silty loam.

Nativ (1988) and Nativ and Riggio (1990) also concluded that the most likely method of ground-water recharge to the Ogallala aquifer is focused percolation of waters collected in the numerous playa lakes. This conclusion was based on rapid recharge rates calculated using bomb-tritium data and on recognition of only a slight enrichment in stable isotopic data (^{2}H and ^{18}O) from ground water with respect to precipitation. Nativ (1988) suggested that recharge might be controlled by factors such as shrinkage or desiccation cracks, solution channels in the caliche, and the upward discharge of ground water from underlying formations. Additional methods of recharge to the Ogallala aquifer were thought to be infiltration through riverbeds (because of the absence of an integrated drainage system on the Southern High Plains, this process is thought to be insignificant) and diffusive infiltration of precipitation directly onto Ogallala outcrops and through Quaternary deposits.

One of the more spatially extensive data sets documenting the process of active recharge through playas was presented by Mollhagen and others (1993). In this study, 99 playa lakes were sampled for chemical and pesticide analysis of surface waters in an effort to understand nonpoint source pollution in the Brazos River Basin. Using the principle of chloride mass balance (chloride concentrations in soil water are inversely proportional to the recharge rate when the chloride input is uniform and the only source of chloride in the system is from precipitation), the chemistry, especially chloride levels, with respect to local levels of chloride in precipitation, can be used to illustrate that the playa lakes are continuously flushing chlorides down through the unsaturated section with recharging waters. Mollhagen and others (1993) reported that for the 99 playa lakes sampled for surface-water chemistry analysis, total dissolved solids (TDS) ranged from 52.6 to 33,686 mg/L, with a mean of 365 mg/L. Of the 99 playa lakes sampled, only 2 samples had TDS levels greater than 3,000 mg/L. Therefore, considering that chloride in precipitation in the area is approximately 0.59 mg/L (based on chloride concentrations in precipitation measured at Amarillo, Texas, and reported in Lodge and others [1968]) and that climatic conditions have been stable for several thousand years and geologic conditions have been stable for at least tens of thousands of years (thus no anticipated change in the natural rate of chloride input from precipitation), chloride input into playa lakes has been quite significant, conservatively in the range of hundreds of thousands of mg/L. Therefore, if playa lakes were "evaporation pans," or non-leaky, then each playa lake should have extensive buildups of evaporite minerals, and when surface water was present, the chemistry of that surface water should be that of brine and not fresh water, as reported in Mollhagen and others (1993). Previous similar but smaller scale studies reporting comparable results for other playa lakes on the Southern High Plains include those by Wells and others (1970), Felty and others (1972), Lehman (1972), and the U.S. Bureau of Reclamation (1982). Mollhagen and others (1993) stated that, on the basis of theirs and previous studies, "Since the major portion of the water in playas infiltrates the lake basin soils, and recharges the underlying groundwater, the quality of playa lake water will affect the groundwater quality."

Scanlon and others (1993, 1994) and Scanlon (1995), using coupled physical and chemical studies in the unsaturated zone, documented that recharge is focused through playa lake bottoms and that negligible recharge is occurring in interplaya or upland areas. Their results were based on an extensive drilling program where vertical profiles illustrating water potentials, water content, and chloride concentrations were constructed to delineate areas of recharge. Chloride profiles in playa lake sediments documented a well-flushed system and water potentials illustrated a downward gradient. Interplaya sediments, however, recorded very high chloride concentrations near the surface and upward water potential gradients, both indicating negligible recharge.

In a discussion of the origin and development of playa basins, Gustavson and others (1995) compare calcic soil development in playas and upland areas and argue that the lack of well-developed calcic soils in playa sediments is physical evidence of a well-flushed system.

Rates of Recharge

The earliest report of a recharge rate to the Ogallala aquifer was made by Johnson (1901), who estimated that recharge ranged from 76 to 102 mm yr^{-1} (3 to 4 inches yr^{-1}). This rate was based on the dynamic equilibrium between recharge and discharge in the aquifer, where estimates for discharge were based on spring discharge. Gould (1906), expanding on Johnson's efforts, estimated a slightly higher recharge rate for the area of approximately 152 mm yr^{-1} (6 inches yr^{-1}). All of the various recharge rates reported are illustrated in figure A-2.

Baker (1915) followed with an estimate of how much precipitation falling on the Llano Estacado annually was actually infiltrating to the aquifer. A recharge rate of 76 to 102 mm yr^{-1} (3 to 4 inches yr^{-1}) is suggested, although no justification of the estimate was given.

Theis (1937) was the first to use physical measurements to calculate mass-balance values of recharge and discharge for the Ogallala aquifer on the Southern High Plains. Nine recharge rates are listed in his report, with a mean recharge rate of 8.9 mm yr^{-1} (0.35 inch yr^{-1}) and a standard deviation of 4.1 mm yr^{-1} (0.16 inch yr^{-1}) (fig. 5). In summarizing these various recharge rates, Theis (1937) concluded that the different rates converge at a value of "slightly less than 13 mm yr^{-1} (0.5 inch yr^{-1})". Brown and Signor (1972) reported that Theis (based on his oral communication to them in 1969) had adjusted his rate of recharge to the Ogallala aquifer downward but did not describe by how much. Long (1961) and Cronin (1961) suggested that recharge is less than 13 mm yr^{-1} (<0.5 inch yr^{-1}) on the basis of reviews of previously published recharge rates. Havens (1966) estimated a regionally averaged recharge rate of 20.6 mm yr^{-1} (0.81 inch yr^{-1}) (which was based on mass-balance calculations of recharge and discharge). In a similar exercise, Brown and Signor (1973) estimated regional recharge to range from 0.006 to 0.02 mm yr^{-1} (0.02 to 0.08 inch yr^{-1}). Brutsaert and others (1975) presented a summary of previously unpublished

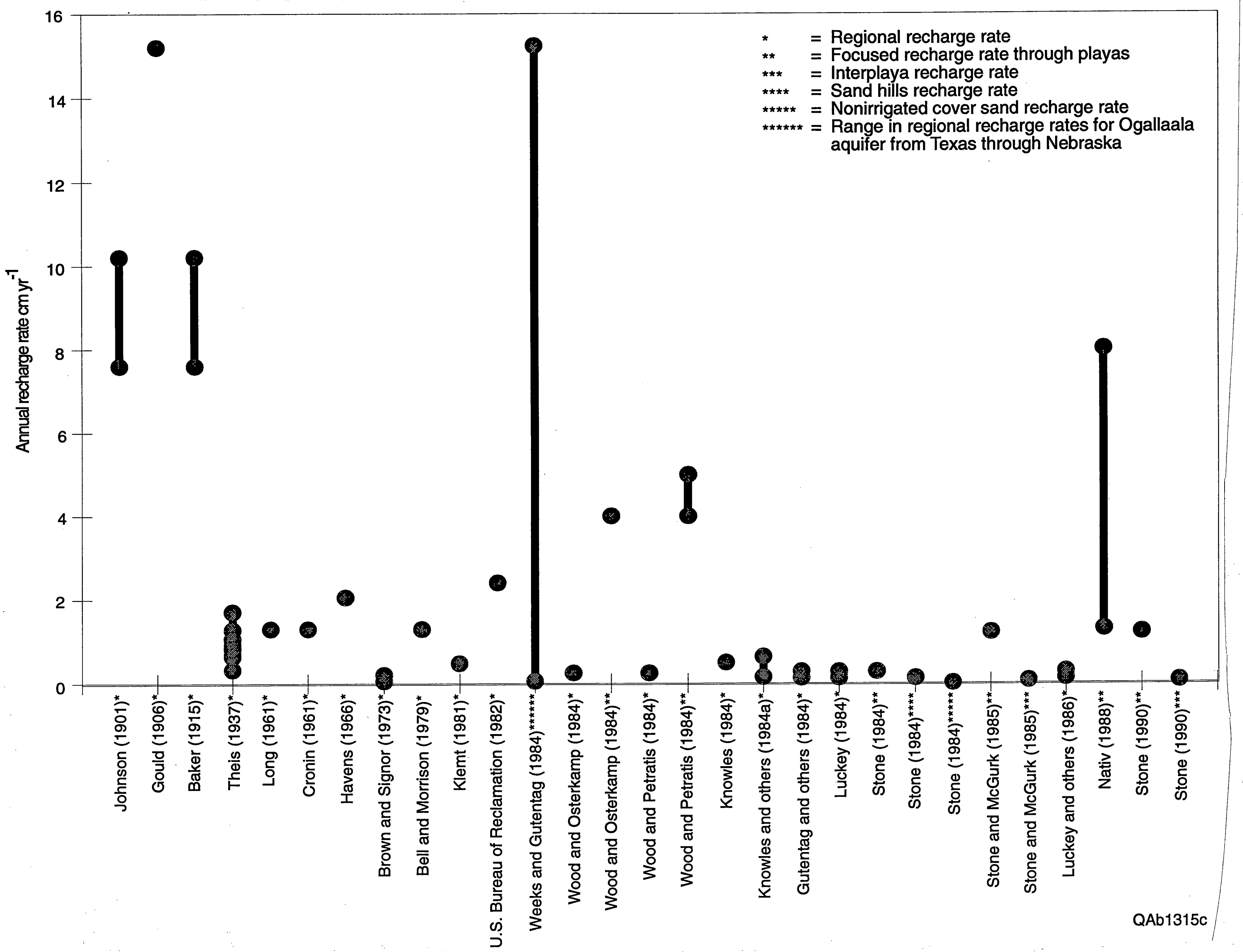

Figure A-2. Histogram of published recharge rates to the Ogallala aquifer on the Southern High Plains.

research performed by C. E. Jacob on the Ogallala aquifer of Texas and New Mexico. This summary noted that Jacob used a recharge value of 3.8 mm yr^{-1} (0.15 inch yr^{-1}) that was based on mass-balance calculations along several cross sections of the Southern High Plains.

Although Broadhurst (1942) did not specifically state a rate of recharge, he did document an average rise in water levels in the Plainview and Hereford districts after the heavy rains of 1941 to average 0.3 m (1.0 ft) and 2.0 m (6.5 ft) in the Muleshoe district. Data needed to make a direct calculation of this recharge rate were not provided. Other indirect predictions of recharge rates included several estimates of the percentage of surface water collected in the playa lake basins that eventually infiltrates to the Ogallala aquifer. Reported ranges of infiltration include 35 to 40 percent (Myers, *in* Cronin, 1964), 20 to 80 percent (Havens, 1966), 27 to 43 percent (Claborn and others, 1985) and 10 to 15 percent (U.S. Bureau of Reclamation, 1982).

Methods that have been used to measure recharge to the Ogallala aquifer, either directly or indirectly, include neutron-probe measurements of deep soil moisture, stable isotopes, chloride mass balance (in the unsaturated zone and in playa lake water), ponding tests, crop-water demands, comparison of simultaneous playa and evaporation pan water-level measurements, radioactive isotopes, and calculations incorporating various hydrologic conditions.

On the basis of irrigation practices and application rates, total crop-water demand, amount of precipitation, and observed water-level declines, Bell and Morrison (1979) observed that the volume of water removed from the aquifer was considerably less than that required by local crops. Their recharge estimate for Carson County was 13 mm yr^{-1} (0.5 inch yr^{-1}) with an additional 10-percent recharge due to the recirculation of irrigation water.

Klemt (1981) used measurements of deep soil moisture collected with a neutron probe over 1 year to estimate that 4.8 mm yr^{-1} (0.1875 inch yr^{-1}) of precipitation reaches the Ogallala aquifer. In an extensive study of 29 playa lake basins on the Southern High Plains, the U.S. Bureau of Reclamation (1982) estimated recharge to the Ogallala aquifer through playas to average 24 mm yr^{-1} (0.94 inch yr^{-1}).

Wood and Osterkamp (1984), using physical and chemical lines of evidence, concluded that much of the recharge to the Ogallala aquifer was occurring through the annular rings of the playa lakes (when playa lake water extends beyond the lateral extent of the Randall Clay). They reported a recharge rate through these annular rings of approximately 40 mm yr^{-1} (1.6 inches yr^{-1}). It is important to note that this value is reported as a flux rather than a recharge rate because of the author's desire to differentiate between an areally averaged and a point-specific recharge rate. In fact, this 40 mm yr^{-1} (1.6 inches yr^{-1}) recharge flux is based on an areally averaged recharge rate of 2.5 mm yr^{-1} (0.1 inch yr^{-1}) for the entire Ogallala aquifer of the Llano Estacado.

Wood and Petratis (1984), in their study on the origin and distribution of carbon dioxide in the unsaturated zone above the Ogallala aquifer, used a regional recharge rate of 2.5 mm yr^{-1} (0.1 inch yr^{-1}). The method used to derive this value is unknown, however, because it is referenced as a personal communication with J. Weeks (1984). On the basis of what were considered to be average surface areas for a playa lake, playa lake basin, and playa annulus, and using their hypothesis that recharge to the Ogallala aquifer is restricted to the playa annulas, they estimated that 6 percent of the surface area of the typical playa basin was covered by the playa annulus. This resulted in a focused recharge rate through the playa annulas of 40 to 50 mm yr^{-1} (1.6 to 2.0 inches yr^{-1}).

Stone (1984, 1990) and Stone and McGurk (1985) used a soil-water chloride mass-balance approach to estimate recharge rates to the Ogallala aquifer in a number of geomorphic settings in eastern New Mexico. They reported the highest rate of recharge in a playa lake to be 12.2 mm yr^{-1} (0.48 inch yr^{-1}). The lowest recharge rate (from an interplaya setting) was more than an order of magnitude less than that reported for playas. The recharge rate reported for upland (interplaya) geomorphic settings was 0.75 mm yr^{-1} (0.029 inch yr^{-1}). Stone and McGurk (1985), in a follow-up to Stone (1984), reported recharge in playas to be 12.2 mm yr^{-1} (0.48 inch yr^{-1}) and 0.8 mm yr^{-1} (0.03 inch yr^{-1}) for cover sand (their interplaya setting).

Knowles (1984, 1985) and Knowles and others (1984a) used an average recharge rate in their ground-water flow model of the High Plains aquifer in the Southern High Plains of 5.1 mm yr^{-1}

(0.2 inch yr^{-1}). This value for recharge appears to have been a simplification of the recharge rate reported by Klemt (1981) of 4.8 mm yr^{-1} (0.19 inch yr^{-1}). The portion of the Knowles and others (1984a) model that corresponded to our report's study area used three zonal rates of recharge. Recharge rates for these zones were 1.5, 3.8, and 6.3 mm yr^{-1} (0.058, 0.150, and 0.250 inch yr^{-1}).

The Regional Aquifer-System Analysis (RASA) of Gutentag and others (1984) used seven recharge rates for the Texas part of the High Plains aquifer. The ground-water flow model constructed as part of the RASA effort used two subregional recharge rates of 1.4 and 2.8 mm yr^{-1} (0.056 and 0.11 inch yr^{-1}) for the 11-county study area described in our report (Luckey, 1984; Luckey and others, 1986). One of the simulations from this effort used a "no recharge" scenario for the Ogallala aquifer, as had been proposed by several investigators through the 1960's to the early 1980's. This scenario was based on the hypothesis that the Ogallala aquifer was recharged during a wetter, cooler climate on the Southern High Plains, which is called the Lubbock subpluvial (Wendorf, 1970). It was proposed that during this time period, recharge was much more efficient and that water levels were much higher, in some areas intersecting the floors of the playa lake basins. With the change to current climatic conditions, no recharge has occurred. The primary result from the "no recharge" simulation was that the Ogallala aquifer would be dewatered rapidly, often in less than a few thousand years. In some areas, it was estimated that dewatering would take less than a thousand years. This RASA simulation result strongly argues against the "no recharge" scenario previously proposed. In a similar effort, Jacob (*in* Brutsaert and others, 1975) estimated that if the Ogallala aquifer was at its original, pre-production, steady-state levels and that all recharge was to be abruptly terminated, then the aquifer would totally dewater after a period of not more than 8,000 years.

Weeks and Gutentag (1984) estimated a range in recharge rates to the Ogallala (High Plains) aquifer from 0.0061 mm yr^{-1} (0.024 inch yr^{-1}) in parts of Texas to 152.4 mm yr^{-1} (6 inches yr^{-1}) in parts of Kansas and Nebraska.

Luckey and others (1986) also discussed that current recharge rates are clearly higher than predevelopment rates, primarily because of the return flow from irrigation and changes in land-use practices. In the RASA model area (Luckey and others, 1986) that corresponds to the model area of our report, significant adjustments in hydraulic conductivity from initial estimates were required to achieve model calibration. The greatest adjustments had to be made in areas such as western Carson County, which were described as areas of salt dissolution and collapse. Luckey and others (1986) also noted that the best simulation results occur when the rates of recharge are designated by soil type.

Nativ (1988), on the basis of elevated bomb-tritium values in Ogallala ground water, estimated a range in recharge rates of 13 to 80 mm yr^{-1} (0.5 to 3.24 inches yr^{-1}).

Dugan and others (1994) quantified recharge to the Ogallala aquifer using the relationship

$$R = P + S - O - E - C$$

where

R = recharge
P = precipitation
S = antecedent soil water (stored)
O = surface runoff
E = actual evapotranspiration
C = water storage capacity of the soil zone

They estimated recharge for this study area, based on the application of the above relationship, to be from 12.7 mm yr^{-1} (0.5 inches yr^{-1}) along the southwestern edge to almost 50.8 mm yr^{-1} (2.0 inches yr^{-1}) throughout most of the area.