THE CARBONATE AQUIFER OF THE CENTRAL ROSWELL BASIN:

RECHARGE ESTIMATION BY NUMERICAL MODELING

by

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ABSTRACT

The flow of groundwater in the Roswell, New Mexico, Artesian Basin has been studied since the early 1900's and varied ideas have been proposed to explain different aspects of the groundwater flow system. The purpose of the present study was to delineate the spatial distribution and source, or sources, of recharge to the carbonate aquifer of the central Roswell Basin. A computer model was used to simulate groundwater flow in the carbonate aquifer beneath and west of Roswell and in the Glorieta Sandstone and Yeso Formation west of the carbonate aquifer. The resulting spatial distribution of recharge strongly indicates that a major component of recharge to the carbonate aquifer is derived from the upward leakage of water from the underlying formations. The model results agree with tritium analyses which indicate that much of the water in the carbonate aquifer is relatively old.

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INTRODUCTION AND PURPOSE

The flow of groundwater in the Roswell Artesian Basin has been studied since the early 1900's with varied ideas proposed to explain different aspects of the groundwater flow system. The purpose of the present study is to help delineate the distribution and source, or sources, of recharge to the Roswell Basin by using a computer model to simulate groundwater flow in the carbonate aquifer beneath and west of Roswell and in the Glorieta Sandstone and Yeso Formation west of the carbonate aquifer. The use of the computer model offers the unique opportunity to evaluate different theories by simply changing various model parameters such as transmissivity, storage coefficient, interaquifer leakage, and recharge.

The results obtained are approximate, but represent the best estimate of the spatial distribution of recharge in the Central Roswell Basin to date. The compatibility of the model results with previously proposed ideas will lend credence to some and refute others. The model results should generate some new ideas, hopefully provoke further research, and serve as a stepping stone to future modeling attempts of the Roswell Basin.

DESCRIPTION OF THE STUDY AREA

The study area is an east-west strip in the central part of the Roswell Artesian Basin in Chaves and Lincoln Counties, New Mexico, and includes much of the Rio Hondo Drainage Basin (Figures 1,2). The study area was chosen because of the relative abundance of data in the Rio Hondo valley region as compared to other parts of the Roswell Basin.

Physiography

Following Motts and Cushman (1964), the area can be divided into a lowland and an upland. The lowland includes the three alluvial terraces, proposed by Fiedler and Nye (1933), known as the Lakewood, Orchard Park, and Blackdom. The upland area includes part of the Diamond-A Plain and the Vaughn-Macho Plain (called the gravel capped Mesas by Fiedler and Nye). The eastern part of the Diamond-A Plain includes the northern part of the Principal Intake Area of Fiedler and Nye.

Climate

The Roswell area has a semi-arid climate averaging about 12 inches of precipitation per year (Saleem and Jacob, 1971) most of which occurs during summer thunderstorms. The mean

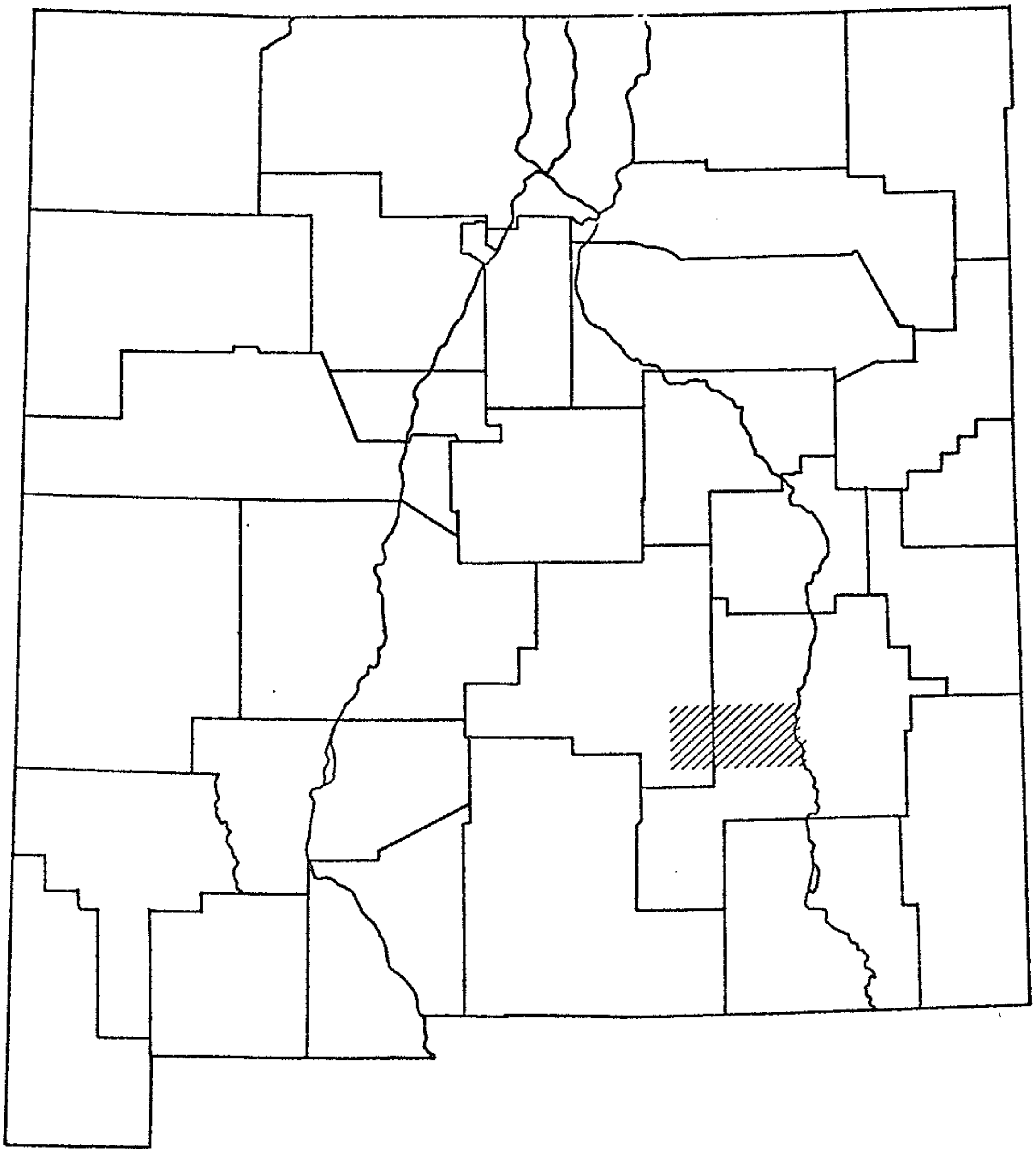

FIGURE 1: LOCATION OF THE CENTRAL ROSWELL BASIN STUDY AREA

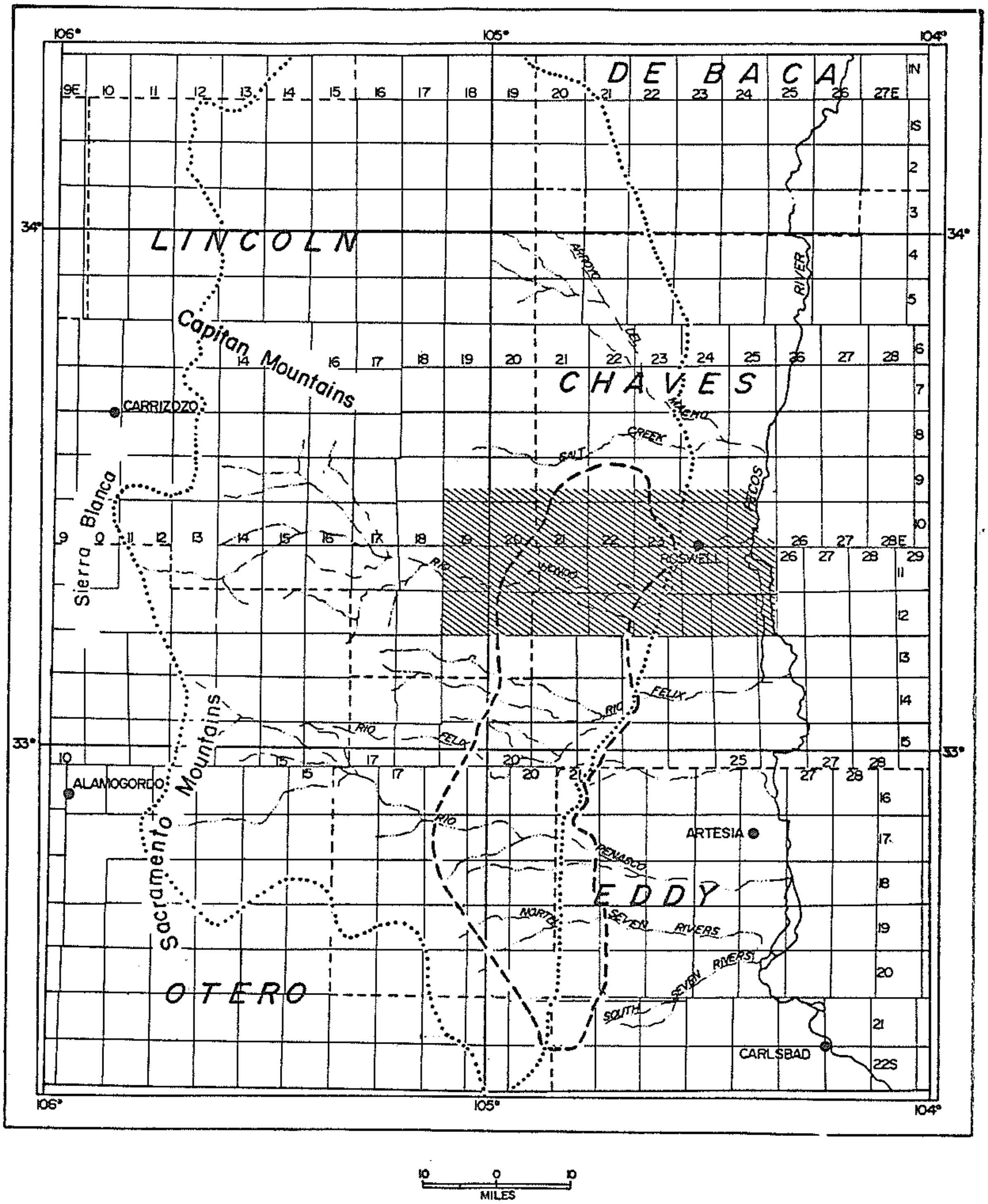

FIGURE 2: LOCATION OF THE STUDY AREA RELATIVE TO THE ROSWELL HYDROLOGIC BASIN

-------PRINCIPAL INTAKE AREA (PIA-FIEDLER & NYE, 1933)

........TOTAL INTAKE AREA (BEAN, 1949)

annual air temperature at Roswell is about 59 degrees Fahrenheit with mild winters and hot summers.

The Diamond-A Plain has a similar climate along its eastern edge, but changes near the mountains where air temperatures are cooler and yearly precipitation is greater.

Hydrogeology

This report is concerned mainly with the flow of groundwater in the carbonate aquifer which is composed of the San Andres Limestone and the Grayburg Formation. Groundwater in the carbonate aquifer is affected by groundwater in the underlying formations, the Yeso and Glorieta, and by the overlying formations, the Queen and the Alluvium. A general stratigraphic column is presented in Table 1.

Water is unconfined in the carbonate aquifer west of about Range 24 East and confined in and east of Range 24 East. The western boundary of the carbonate aquifer occurs where the water table intersects the Glorieta-San Andres contact and the eastern boundary is approximately the Pecos River. The northern and southern boundaries are estimated to be near Arroyo Del Macho and South Seven Rivers respectively. The hydrologic boundaries are shown in Figure 3.

Table 1

General Stratigraphic Column

(taken from Kelley, 1971)

Age	Series	Group	Formations & Members	Equivalents	Thick	Description
Holocene and Pleistocene			Assorted surficial deposits		0-300	Valley alluvium, terrace and pediment gravel, caliche soils, aeolian sand, travertine
Pleistocene-Pliocene			Gatuna Formation		0-200	Sandstone, sand gravel, siltstone, limestone, red, brown, tan, gray, yellowish
Oligocene			Sierra Blanca Volcanics		700-4,000	Andesite breccia and tuff; some flows
Paleocene			Cub Mountain Formation		500-2,000	Sandstone, mudstone, conglomerate, arkose; white, buff, lavender, purple, maroon
Cretaceous			Mesaverde Formation		500-1,500	Sandstone, shale, coal, conglomerate; buff, gray, black
Cretaceous			Mancos Shale		400-700	Shale, siltstone, with local thin sandstone and limestone; black, grayish-black
Cretaceous			Dakota Sandstone		100-150	Sandstone, conglomerate, black shale; gray to tan
Upper Triassic			Chinle Shale		0-300	Mudstone with some claystone and thin sandstone; reddish brown
Upper Triassic			Santa Rosa Sandstone		0-300	Sandstone, conglomerate, mudstone; brown, buff, lavender
PERMIAN	Ochoan Series		Dewey Lake Formation		200-250	Sandstone, siltstone; orange-brown; commonly laminated
PERMIAN	Ochoan Series		Rustler Formation: Upper Member		150-200	Dolomite, gypsum, mudstone, white, red-brown, green, gray, deep orange; Magenta dolomite at base
PERMIAN	Ochoan Series		Rustler Formation: Lower Member		100-250	Dolomite, gypsum, mudstone, sandstone; white, red-brown, gray, green; salt in subsurface; Culebra dolomite at base.
PERMIAN	Ochoan Series		Salado Formation		0-2,500	Gypsum, mudstone, thin local dolomite; white, red, brown, green, deep orange; breccia residue at surface, thick salt, potash in subsurface
PERMIAN	Ochoan Series		Castile Formation: Upper Member* (surface)		1,000±	Gypsum (anhydrite), salt; white, gray
PERMIAN	Ochoan Series		Castile Formation: Lower Member (surface)		1,000±	Laminated gypsum (anhydrite) and limestone, laminated limestone, laminated gypsum; gray, black, white
PERMIAN	Guadalupian Series	Artesia Group	Tansill Formation	Capitan Ls.† / Bell Canyon Fm.*	200-300	Dolomite and siltstone (south); dolomite, gypsum, and anhydrite (north); Ocotillo siltstone tongue near exposed top
PERMIAN	Guadalupian Series	Artesia Group	Yates Formation	Capitan Ls.† / Bell Canyon Fm.*	250-350	Siltstone, sandstone, dolomite, limestone and gypsum (south); gypsum, siltstone and thin dolomite (north)
PERMIAN	Guadalupian Series	Artesia Group	Seven Rivers Formation	Capitan Ls.† / Bell Canyon Fm.*	450-600	Dolomite, siltstone (south); gypsum and siltstone (north)
PERMIAN	Guadalupian Series	Artesia Group	Queen Formation	Goat Seep Do.† / Cherry Canyon Fm.*	200-400	Dolomite and sandstone (south); gypsum, red mudstone, dolomite (north); Shattuck member near top
PERMIAN	Guadalupian Series	Artesia Group	Grayburg Formation	Goat Seep Do.† / Cherry Canyon Fm.*	250-450	Dolomite and sandstone (south); gypsum, mudstone, dolomite (north)
PERMIAN	Guadalupian Series		San Andres Formation: Fourmile Draw Member		0-700	Dolomite, gypsum, reddish mudstone; sandstone locally at top; thin-bedded
PERMIAN	Guadalupian Series		San Andres Formation: Bonney Canyon Member		0-300	Dolomite, local limestone; gray, light-gray, local black; thin-bedded
PERMIAN	Leonardian Series		San Andres Formation: Rio Bonito Member		250-350	Dolomite, limestone, sandstone (Glorieta); gray, brownish gray; thick-bedded
PERMIAN	Leonardian Series		Yeso Formation		0-1,400	Sandstone, siltstone, dolomite, gypsum; tan, red-yellow, gray, white
Precambrian			Syenite, gneiss, and diabase			

* Delware basin facies only

† Reef facies only

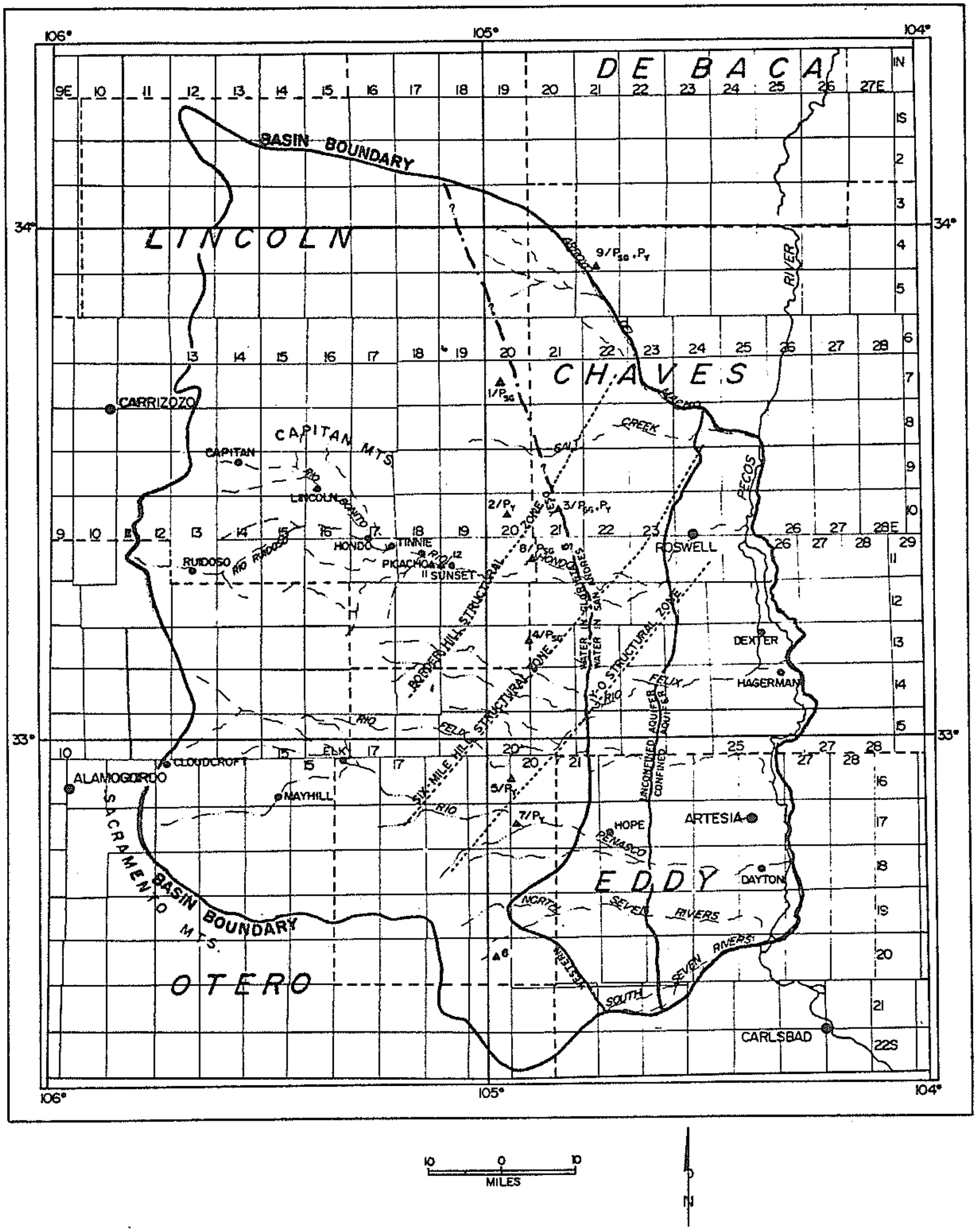

FIGURE 3: HYDROLOGIC BOUNDARIES OF THE ROSWELL BASIN

The carbonate aquifer is assumed to be recharged in the area of unconfined flow. Fiedler and Nye (1933) described what they called the Principal Intake Area as the area in which the majority of recharge occurs (Figure 2). Bean (1949) described a total intake area in which all the recharge occurs (Figure 2). One of the objectives of this report is to determine, from the model, the areal recharge distribution.

Moving from west to east in the Roswell Basin, progressively younger formations are encountered because the formations dip to the east-southeast at a greater angle than the topography. The slope of the water table in the western (unconfined) portion of the study area is less than the dip of the strata and consequently the water table intersects progressively younger formations from west to east.

Interaquifer leakage occurs between the carbonate aquifer and the alluvial aquifer, through the Queen aquitard, in a band about 20 miles wide adjacent to, and west of, the Pecos River. Leakage is generally greatest in the vicinity of Roswell and decreases to the south and southwest (Hantush, 1957). Prior to the development of irrigation wells, water leaked vertically upward from the carbonate aquifer to the alluvial aquifer. At present, the large drawdown of the potentiometric surface in the carbonate aquifer during the summer irrigation season reverses the direction of vertical leakage and the net yearly leakage may be nearly zero. A good summary of the hydrogeology is given by Gross and others (1976).

PREVIOUS WORK

Many excellent reports have dealt with various aspects of the Roswell Artesian Basin. In general, each report can be placed into one or more of the following three broad subject areas: (1) geology, (2) hydrogeology, and (3) hydrologic modeling.

Geology

Relatively few reports have dealt exclusively with the geology of the Roswell Basin. Kelley (1971) presents an excellent description of the stratigraphy and structural features of the Basin. Several reports deal with aspects of the Glorieta Sandstone such as the structure (Borton, 1972a), genesis, provenance, and petrography (Milner, 1978), and stratigraphic relationships (Harbour, 1970). Hawley and others (1976) describe the Quaternary stratigraphy of part of the Roswell Basin.

Hydrogeology

The first study of the relationship between geology and groundwater in the Roswell Basin was done by Fisher (1906). In 1933, Fiedler and Nye published the most comprehensive study of the Basin to date. Subsequent work in the Basin has, in general, substantiated the conclusions reached by Fiedler and Nye almost 50 years ago. Theis and Sayre (1942) summarized the geology and groundwater conditions for the Pecos River Joint Investigation. Maddox (1969a,1969b) and Kinney and others (1968) used oil exploration data as well as data from the Pecos Valley Artesian Conservancy District (PVACD) to detail the relationships between rock units and their effects on groundwater flow. Motts and Cushman (1964) present an appraisal of the possibilities for artificial recharge in the Roswell Basin. Gross and others (1976), Gross and Hoy (1979), and Hoy and Gross (1982) present the results of various isotope studies and their interpretation in terms of recharge to, and movement of, groundwater in the Basin.

Other authors have chosen to study selected areas of the Roswell Basin in greater detail. Bean (1949) and Theis (1951) looked at the area around the Hondo Reservoir. Mourant (1963) detailed the geology and hydrology of the Hondo Drainage Basin. Duffy and others (1978) examined stream leakage along the Rio Hondo and Rio Penasco with a stochastic analysis technique. Gross and others (1979) investigated the recharge characteristics of Paul Spring which is located on the Rio Penasco, near Elk, New Mexico. Havenor (1968) and Bunte (1960) examined the northern part of the Basin. Havenor presents some interesting ideas concerning the effects of the structural zones on the movement of groundwater in the Basin and Bunte examines the role of the Glorieta Sandstone in recharging the groundwater of the Basin.

Hood and others (1960), Hood (1963), and Henninghausen (1970) discuss the possible sources of the saline water encroaching into the Roswell area from the northeast. Renick (1926) and DeWilde (1961) studied the geology and groundwater conditions of the upper Rio Penasco and Rio Felix drainages respectively. Wasiolek (1981) did an excellent hydrogeologic study of the upper Rio Penasco. Her work supports the hypothesis of Gross and others (1976) that some recharge to the Roswell Basin is derived from the groundwater system along the east slope of the Sacramento Mountains.

Flow Models

The broad category of models includes studies that involved analytical solutions with a few parameters to studies involving many parameters.

On the basis of pumping test analysis Hantush (1957,1961) modeled the carbonate aquifer as a leaky confined system. He used a simple linear model to estimate recharge from precipitation and quantified many hydrologic parameters in the Basin. Maddox (1966,1969b) attempted to use an electrical analog model to simulate flow in the Basin as part of his work, but was unsuccessful. Rabinowitz and Gross (1972) and Rabinowitz and others (1977a, 1977b, 1977c) modeled the transport of tritium in the groundwater to determine the rate of groundwater movement. Hernandez (1971), and Saleem and Jacob (1971) used modeling to determine an optimal plan for the management of the water resources of the Basin.

Three bibliographies are available concerning the Pecos River Basin that contain the above mentioned studies and many others. They are Hernandez and Eaton (1968), Borton (1972b), and Wright (1979).

The above section on previous work was intended to acquaint the reader with some of the more important work done in the Basin. Many of the reports will be discussed in more detail in the Discussion of Results section where the results of the current modeling project will be compared to previously published ideas.

DESCRIPTION OF THE FINITE-DIFFERENCE MODEL

The computer model used is a two-dimensional finite-difference model written by Trescott and others (1976), herein called the Trescott model or the model. The model was chosen because it was easily obtained and extremely well documented. The application of a computer model to an aquifer is a three-step process of calibration, verification, and prediction.

Calibration is the trial and error process of adjusting the aquifer parameters in a model in order to match the computed head distribution to the observed head distribution for some historic period of time. If the computed head map does not match the observed head map, the parameters are adjusted and another computed head map is generated. This process is repeated until a satisfactory match is obtained.

Following the calibration, the calibrated parameters are verified against another historic period of time. The purpose of the verification is to provide a check on the calibration. For example, if the observed water levels used for the calibration were the result of some anomalous condition, the verification would produce a poor match between the computed and observed heads at the end of the verification period. On the other hand, if the verification produces a good match, we can be reasonably sure the calibrated parameters are correct.

The final step in the modeling process is prediction. The model is used to predict future water levels, given expected pumpage and recharge. Predictions using the model were not performed.

Two major assumptions were needed in order to apply the Trescott model to the Roswell Basin. First, that the application of a porous media flow model is valid for the fractured carbonate flow system in the Roswell Basin. In general, the carbonate aquifer will approximate a porous medium if the fractures are closely spaced with respect to the scale of the model and are approximately uniformly distributed. The model area is divided into equal areas of one square mile. Therefore, if an unspecified large number of approximately uniformly distributed fractures and solution channels occur per square mile, the assumption of porous media flow is valid.

The density and distribution of fractures and solution channels has not been determined, but some evidence points towards closely spaced, uniformly distributed fractures. Fiedler and Nye (1933) state that the Picacho (San Andres) limestone has remarkably uniform characteristics. They also say that in the area of confined flow most beds are less than

2 feet thick, "worm-eaten" limestone is not uncommon in the upper part of the Picacho (San Andres), and a limestone breccia is present at or near the top of the Picacho (San Andres) in parts of the Basin. However, in contrast, they state that the solution cavities vary greatly in number, size, shape, and distribution and that the "water rocks" as given by driller's logs are not stratigraphically continuous and represent erratic solution zones.

The fractures and solution zones must, however, be continuous hydrologically or the continuous potentiometric head distribution observed in the carbonate aquifer could not exist. Also, few wells, if any, fail to encounter water in the carbonate aquifer indicating that the fractures and solution cavities must be closely spaced. In terms of the area of the model, the use of a porous media flow model appears valid.

The second assumption was the result of a lack of appropriate data. As stated before, the area modeled includes both confined and unconfined flow. The Trescott model will simulate combined artesian-water table conditions, but the top and bottom elevation of the aquifer must be known. Unfortunately, the base of the unconfined aquifer in the San Andres, Glorieta, and Yeso is unknown. The assumption was made to model the unconfined flow area as a confined system. In that way the top and bottom of the aquifer are not required. This assumption is valid if the changes in the water table elevation are small compared to the aquifer thickness. For the period of time modeled, 1967-1975, the head in the unconfined area did not, for the most part, change significantly. Therefore the second assumption will not induce significant errors.

The model is capable of incorporating the effects of anisotropy, but no data are available concerning the degree of anisotropy or the principal directions of the transmissivity tensor for the Roswell Basin. The area was assumed isotropic.

Interaquifer leakage is calculated by the model, given the hydraulic head in the carbonate and alluvial aquifers plus the thickness and hydraulic conductivity of the aquitard.

Governing Equation

The following brief description of the governing flow equation is taken from Trescott and others (1976). The governing flow equation used to model the groundwater system is

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y} \right) = S\frac{\partial h}{\partial t} + W(x,y,t) \qquad (1)$$

where:

T_{xx}, T_{yy} - principal components of the transmissivity tensor;

h - hydraulic head;

S - storage coefficient;

$W(x,y,t)$ - volumetric flux of recharge or withdrawal per unit surface area of the aquifer.

The source term, $W(x,y,t)$, includes well discharge, transient leakage from the aquitard, and a recharge flux. The source term is

$$W_{i,j,k} = \frac{Q_{w[i,j,k]}}{\Delta x_j \; \Delta y_i} - q_{re[i,j,k]} - q'_{i,j,k} \tag{2}$$

where:

Q_w - well discharge;

q_{re} - recharge flux per unit area;

q' - flux per unit area from a confining bed;

i,j - y and x coordinate locations, respectively;

k - time step.

The well discharge and recharge flux terms are self explanatory, but the leakage term needs some elaboration.

The leakage term is the sum of a transient and a steady term. The steady term is the leakage due to the initial gradient across the aquitard. The transient term accounts for changes in leakage due to changes in the hydraulic head of the carbonate aquifer. The leakage term also considers the effect of storage of water in the aquitard.

The leakage flux is approximated by

$$q'_{i,j,k} \cong (h_{i,j,o}-h_{i,j,k}) \frac{K'_{i,j}}{\left(\frac{\pi K'_{i,j} t}{3m^2_{i,j} S_{s[i,j]}}\right)^{1/2} m_{i,j}} \cdot \left\{1+2\sum_{n=1}^{\infty} \exp\left[\frac{-n^2}{\left(\frac{K'_{i,j} t}{3m^2_{i,j} S_{s[i,j]}}\right)}\right]\right\} + \frac{K'_{i,j}}{m_{i,j}}(\hat{h}_{i,j,o}-h_{i,j,o}) \quad (3)$$

where:

$h_{i,j,o}$	- hydraulic head in the aquifer at the start of the pumping period;
$h_{i,j,k}$	- hydraulic head in the aquifer at time k;
$\hat{h}_{i,j,o}$	- hydraulic head on the other side of the aquitard;
$K'_{i,j}$	- vertical hydraulic conductivity of the aquitard;
$m_{i,j}$	- thickness of the aquitard;
$S_{s[i,j]}$	- specific storage of the aquitard;
$(K'_{i,j}\ t/m_{i,j}S_s)$	- dimensionless time;
t	- elapsed time of the pumping period.

In the leakage calculation, the hydraulic head on the other side of the aquitard (alluvial aquifer) does not change during a pumping period. Fortunately, the annual fluctuation in the hydraulic head of the alluvial aquifer is on the order of only 2 feet as determined from the U.S. Geological Survey continuous recording observation well 12.25.23.344a. The error introduced by assuming no change in the alluvial aquifer should be small. The maximum water level fluctuation in this well was only 5 feet in the period 1967-1975.

During the calibration, a contouring routine was employed. The routine, CONTUR, is available on the New Mexico Tech Computer Center system library and saved many hours of work.

The model uses a block-centered finite difference grid and a five point approximation scheme. The harmonic mean is used to calculate the transmissivity along nodal boundaries. The grid for the present study is presented in Figure 4. The blocks, or nodes, are all one square mile in area.

In summary, the study area was modeled as an isotropic, confined system with transient leakage occurring between the carbonate aquifer and the alluvial aquifer.

Boundaries

To model the area, the boundaries of the system must be defined. In the model, boundaries can be either constant head, constant flux, or no flux. At a constant head boundary the water level is held fixed, but the amount of water flowing across the boundary varies according to the changes in the water level of the node immediately interior. On a flux boundary, the amount of water entering or leaving the system at that boundary is specified. The no flux boundary

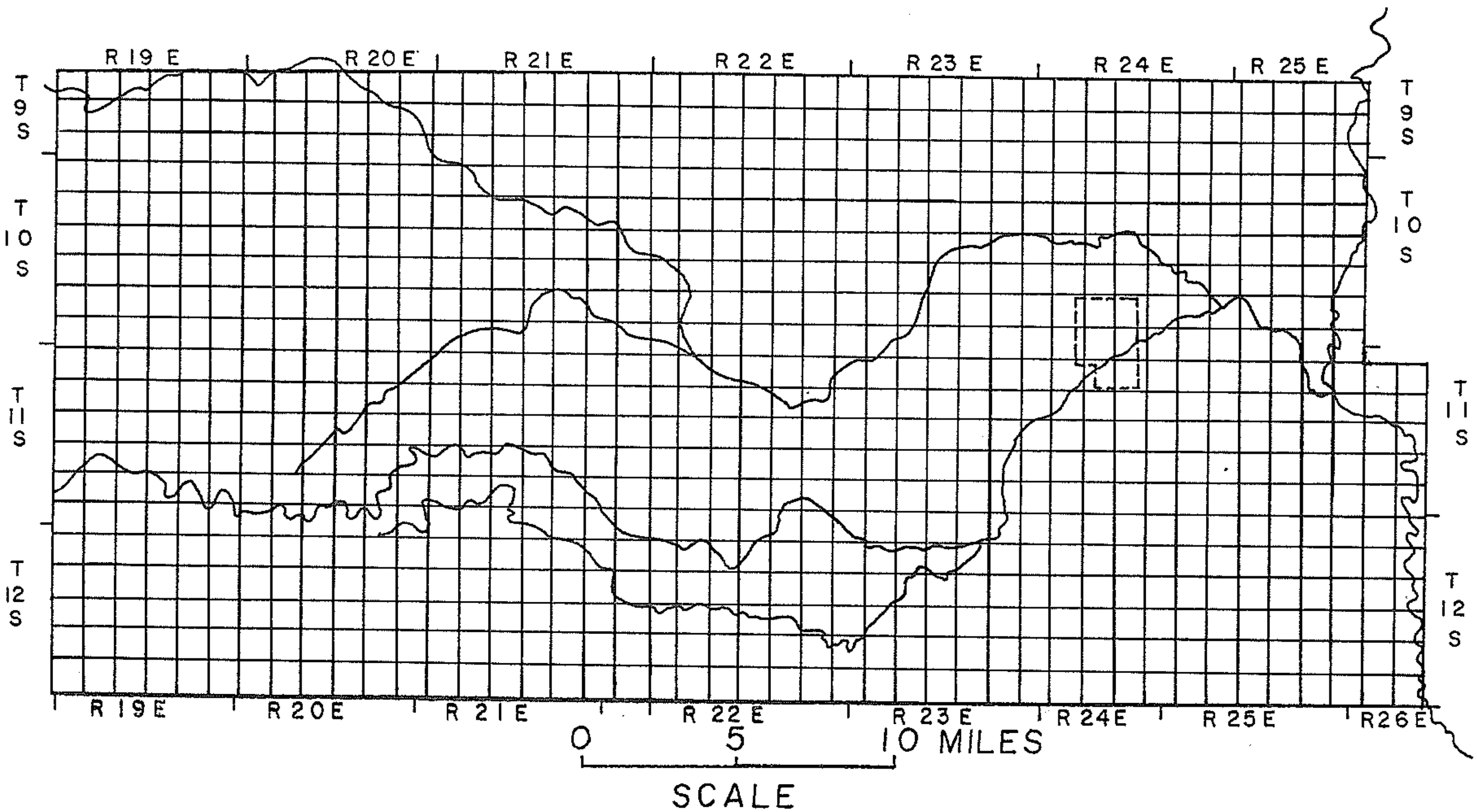

FIGURE 4: FINITE-DIFFERENCE GRID OF THE STUDY AREA

is the special case where the specified flux is zero.

In the model, the western and eastern boundaries were treated as of constant head. The water level in the unconfined portion of the Yeso is largely unaffected by pumping near Roswell as is demonstrated by PVACD observation wells located along the Border Hills. The water level in these wells does not fluctuate in a yearly pattern as do wells located in the irrigated area (Duffy and others, 1978). The unconfined Yeso also has maintained approximately the same water table configuration for many years. Therefore, treating the western boundary as a constant head is justified. On the other hand, treating the eastern boundary as constant head is not as easily justified.

The general vicinity of the Pecos River is assumed to be the eastern boundary of the Roswell Basin. Some groundwater leaves the Basin by flowing eastward, but the amount is small (Kinney and others, 1968; Hantush, 1957). Unable to determine the eastward flux of groundwater, the eastern boundary was assumed a constant head and the model calculated an eastward flux. Because the flux is small, the transmissivity of the eastern boundary was arbitrarily set many orders of magnitude smaller than values obtained for the rest of the basin. In that way, the eastern boundary acted largely as a barrier, but did allow small amounts of water to leave the basin.

The northern and southern boundaries are treated as no flux. For the most part the equipotential lines are perpendicular to the northern and southern boundaries of the model indicating flow along, but not across the boundary. The use of a no flux boundary worked well except in the southeast corner of the model area where the flow is in a southeasterly direction due to the large pumping center near Dexter (Maddox, 1969b; Saleem and Jacob, 1971). A hypothetical large well was placed in the southeast corner of the study area to simulate the pumpage of the Dexter area. The amount of water that actually flows out of the study area through the southeast corner is unknown and could only be estimated.

INITIAL DATA

The following parameters are required at each node in the model in order to simulate the groundwater flow system: (1) initial head, (2) storage coefficient, (3) transmissivity, (4) areal recharge flux, and (5) pumping rate. In the portion of the model with confined flow and leakage the following parameters are also required: (1) vertical hydraulic conductivity of the aquitard, (2) specific storage coefficient of the aquitard, (3) thickness of the aquitard, and (4) head in the alluvial aquifer.

The available data were reviewed in order to choose the time period offering the greatest amount of data to be used as the calibration period. A period of steady-state groundwater flow would be ideal because the storage coefficient is then eliminated from the governing equation. Unfortunately, steady-state conditions existed only prior to the development of artesian irrigation wells. Therefore, that period of time following the development of irrigation wells was selected which offered the greatest amount of data and the least number of unknowns. The only time dependent parameters are: (1) head distribution, (2) recharge, and (3) pumpage. Detailed head data are available from the 1950's to date. Recharge will be a function of stream flow and precipitation, both of which have extensive records. Note, however, that recharge is a complex function of stream flow and precipitation and is largely an unknown for the Roswell Basin. The first year that accurate pumping data were available was 1967. Therefore, the calibration period of January 1967 to January 1968 was chosen because the pumpage was a known parameter.

The data used in the model must be in consistent units with seconds as the unit of time. Feet was chosen as the length unit. Conversion of all the data to feet-second units would have been extremely tedious. Fortunately, the program allows data of any units to be input along with a conversion factor which changes the units to feet and seconds.

Potentiometric Head

Data on potentiometric head of wells in the basin were obtained from the U.S. Geological Survey. The data are also available from the New Mexico State Engineer in yearly technical reports entitled Ground-Water Levels in New Mexico. The head data are concentrated in the area of confined flow and are sparse to non-existent in portions of the unconfined area. Potentiometric head data are available for the unconfined Yeso, Glorieta, and western San Andres from Mourant (1963). Although Mourant's water table map was compiled in 1961, the water level in the Yeso and Glorieta has not been affected by the pumping along the Pecos River and has remained in a largely steady-state condition. To check for steady-state, water levels measured in wells by both Mourant (1963) and Fiedler and Nye (1933) were compared (Table 2). The data are limited and inconsistent, but no regional rise or decline in the water table is demonstrated. In the model, water levels west of about the center of Range 21 East were taken from Mourant (1963) and east of about Range 21 East from USGS records. As further evidence that the water levels west of Range 20 East do not change significantly, the water level in wells equipped with continuous recorders by the PVACD, located approximately in the north-south line between Ranges 20 and 21, fluctuated relatively little compared to wells near the Pecos River. The fluctuations are unrelated to pumpage and are probably related to long term climatic trends (Duffy and others, 1978).

Generally, the water level in the wells in the vicinity of Roswell is measured once a year in January, but occasionally some wells are sampled more often. Several wells in the confined zone are equipped with continuous water level recorders maintained by the USGS and the PVACD.

The potentiometric surface contour maps of the carbonate aquifer for January of 1967, 1968, and 1975 were drawn from USGS data and from Mourant (1963) and are presented in Figures 5, 6, and 7, respectively. The data for each of the three years were transferred to the finite-difference grid by

Table 2

Water Level Comparison of Wells in the Yeso and Glorieta, Western Region
1920 and 1960

Location	Altitude (feet)	Depth of well (feet)	Depth to water (feet)	Aquifer	Source
9.22.35.300	4100	540	535	Picacho	FN
9.22.35.330	4080	540	530	San Andres	M
10.20.22.110	4480	615	500	Nogal-sandstone	FN
10.20.22.110	4430	615	375	Glorieta	M
10.21.13.220	4120	510	475	Picacho	FN
10.21.13.224	4060	575	500	---	M
10.22.31.400	4010	450	440	Picacho	FN
10.22.31.433	4000	450	440	---	M
10.23.21.420	3755	160	130	Picacho	FN
10.23.21.244	3710	150	90	---	M
10.23.30.420	3795	300	150	Picacho	FN
10.23.30.422	3790	300	240	---	M
11.15.27.330	--	575	535	Nogal	FN
11.15.27.332	6720	565	510	Yeso	M
11.16.19.133	--	400	360	Nogal	FN
11.16.19.144	6300	410	365	Yeso	M
11.16.22.313	--	360	40	Nogal	FN
11.16.22.321	--	320	275	Yeso	M
11.22.02.130	3905	327	295	Picacho	FN
11.22.02.131	3880	326	313	San Andres	M
11.22.22.110	3965	410	360	Picacho	FN
11.22.22.111	3940	410	363	San Andres	M
11.23.27.420	3755	150	145	Picacho	FN
11.23.27.424	3720	-	149	San Andres	M
12.23.05.310	--	250	230	Picacho	FN
12.23.05.311	3810	250	237	San Andres	M

M - Mourant (1963)
FN - Fiedler and Nye (1933)

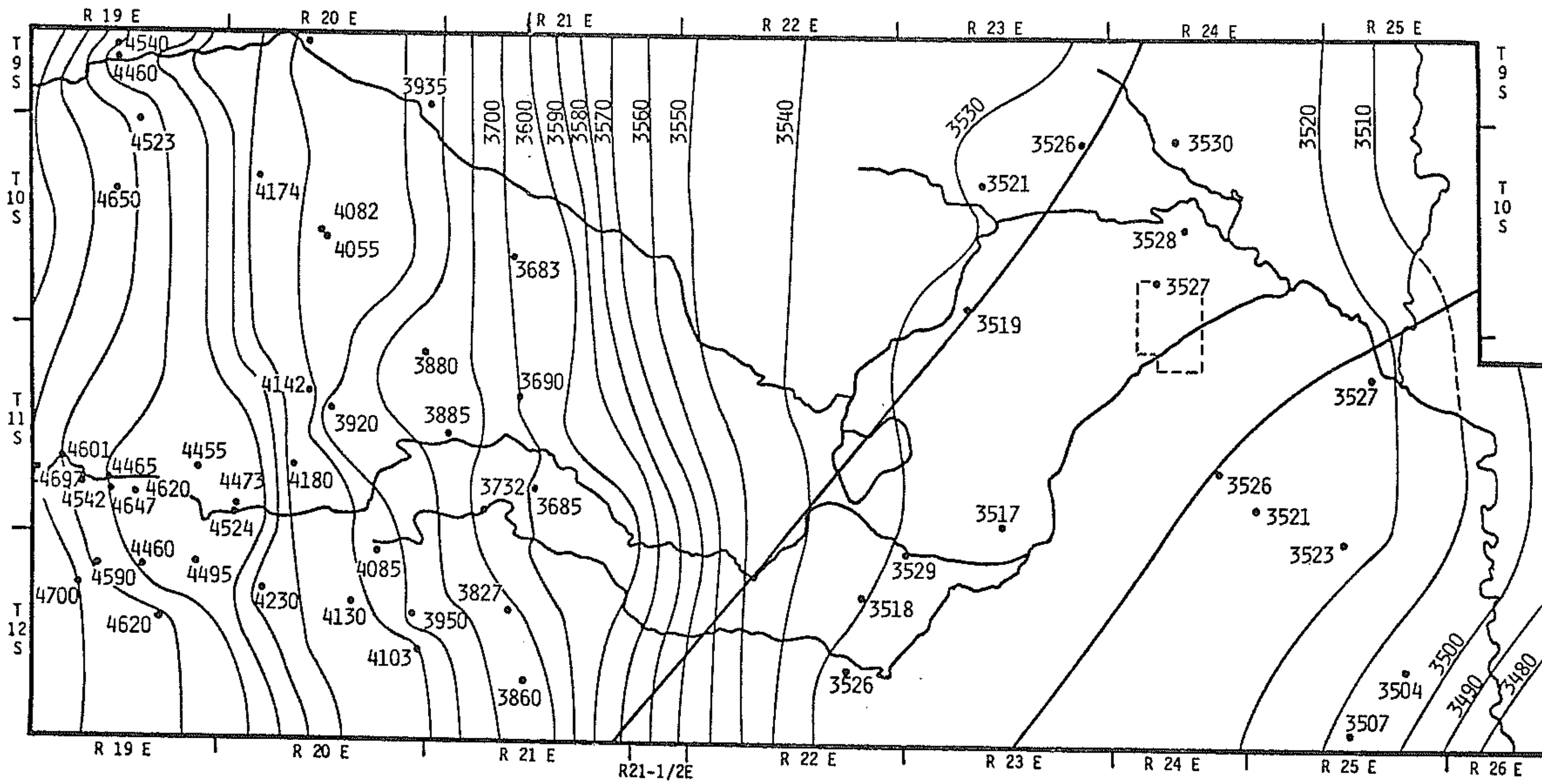

FIGURE 5: HYDRAULIC HEAD CONTOUR MAP FOR THE CARBONATE AQUIFER AND THE WESTERN REGION JANUARY, 1967

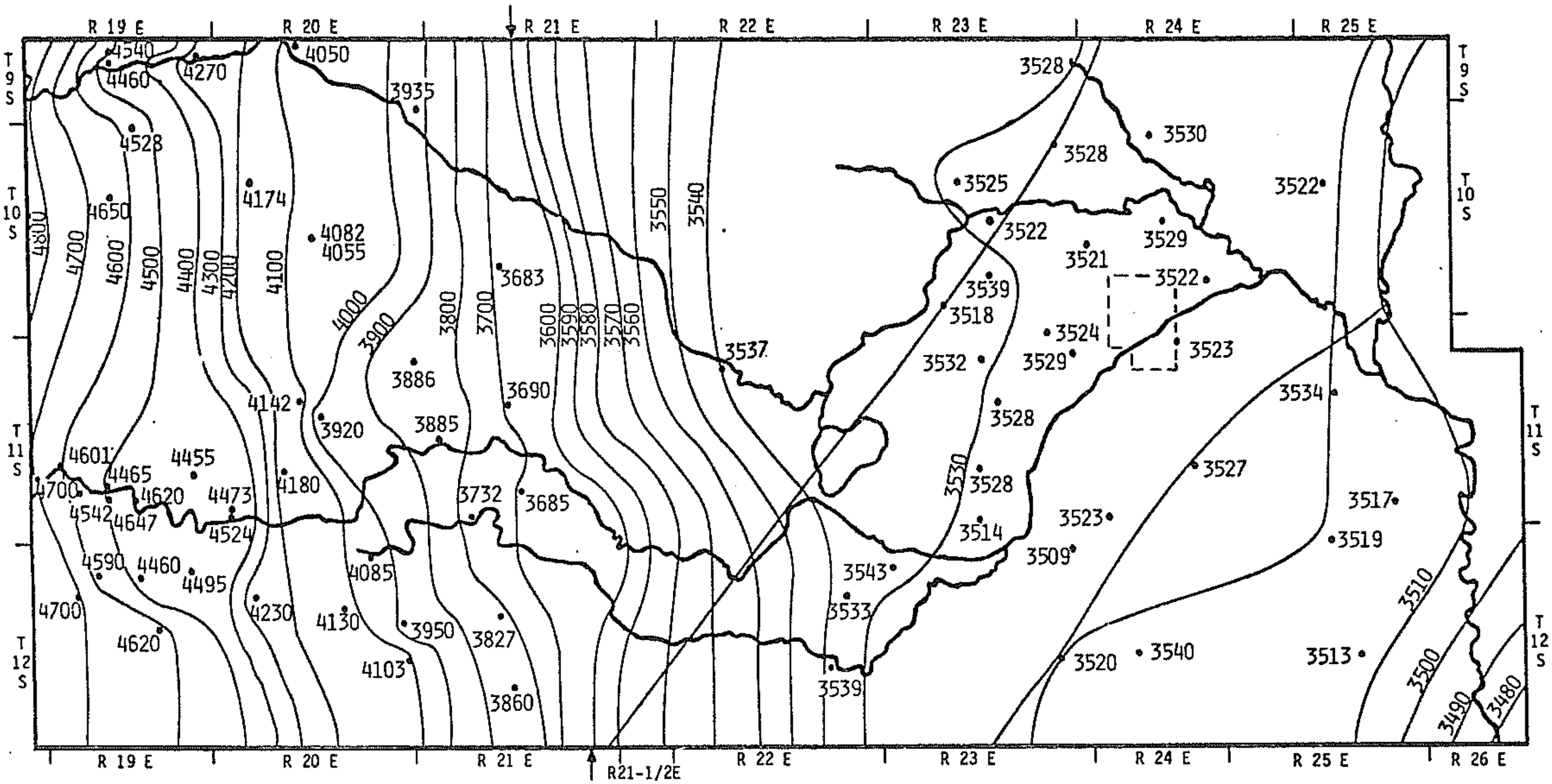

FIGURE 6: HYDRAULIC HEAD CONTOUR MAP FOR THE CARBONATE AQUIFER AND THE WESTERN REGION. JANUARY, 1968

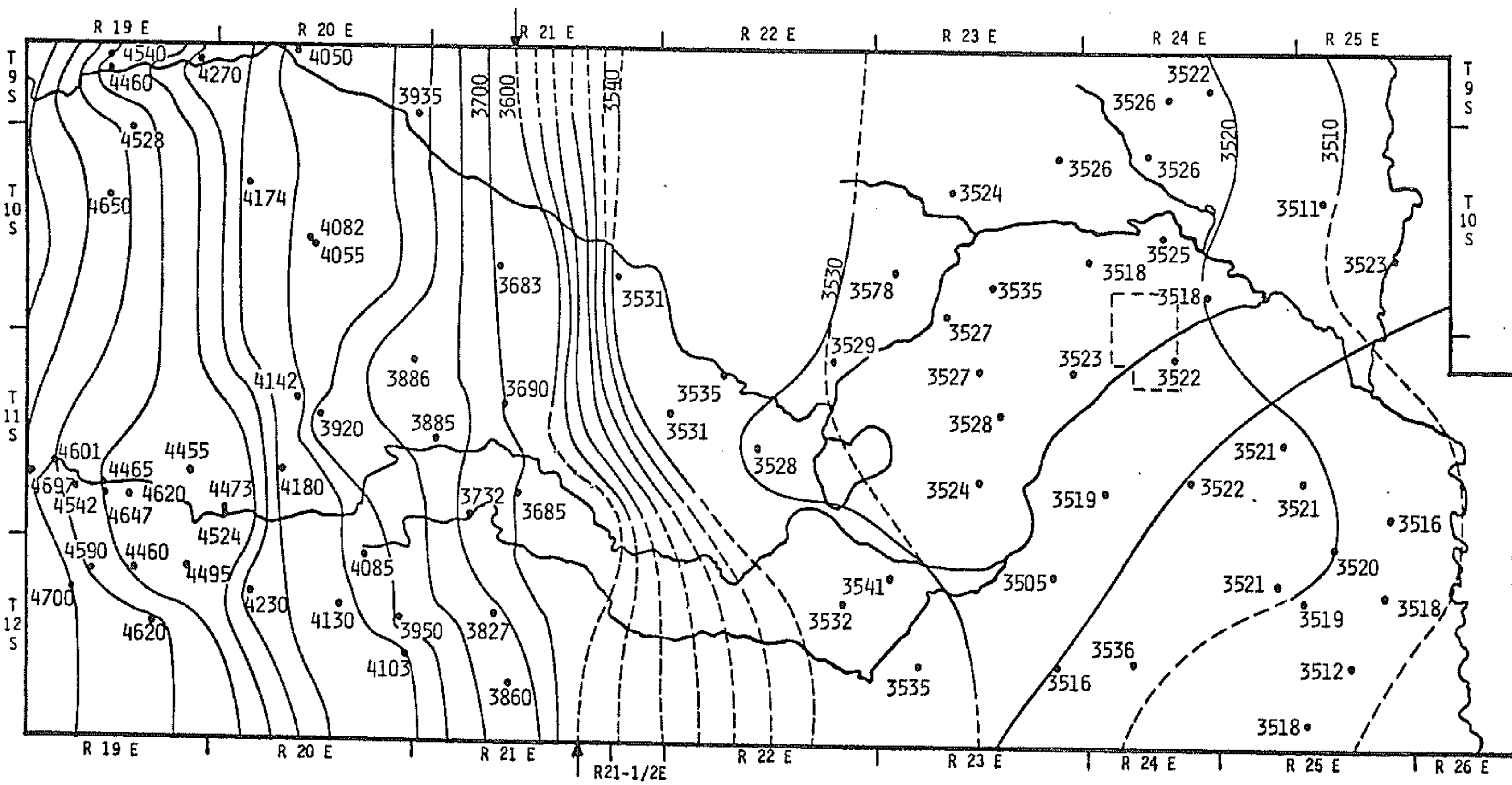

Figure 7: Hydraulic head contour map for the carbonate aquifer and the western region January, 1975

visually averaging the values over the nodal area. The average values were then plotted using CONTUR and the plot was compared to the hand drawn maps to check for errors in averaging and also to check that the computer generated plot was accurate. The program was modified to compute and plot the difference between the computed 1968 head values and the measured 1968 head values. The same was done for the 1975 data. This made the location of problem areas easier during calibration and verification.

USGS data were also used to draw the water table contour map of the alluvial aquifer. In the model, the head in the alluvial aquifer is assumed constant. A variable head is unnecessary because the head in January of 1967 and 1968 was almost the same, as illustrated by Table 3. The seasonal variation in head is small in the Roswell region as seen from the records of wells equipped with continuous recorders located near Roswell.

Again during the 8 year verification simulations, the head in the alluvial aquifer was assumed constant. Table 4 contains head values measured in 1975 and 1967 or 1968 for selected wells. Most wells do not differ by more than 2 or 3 feet. Figure 8 is a map of the water table in the alluvial aquifer drawn for 1968 and used for the entire simulation.

Table 3

Alluvial Aquifer Head Data for 1967 and 1968

(elevation above mean sea level)

Location	1967	1968	1968-1967
10.24.16.131	3538.57	3539.44	.87
10.25.15.323	3474.77	3475.07	.30
10.25.17.142	3496.91	3496.29	- .62
10.25.31.413	3485.42	3485.67	.25
10.25.33.331	3465.73	3465.52	- .21
11.24.14.331	3424.50	3422.11	-2.39
11.24.20.333	3424.36	3421.88	-2.48
11.25.06.421	3492.51	3494.08	1.57
12.25.12.233	3488.50	3487.77	- .73
12.25.23.344	3432.77	3435.18	2.41
12.26.17.143	3433.48	3432.50	- .98

Table 4

Alluvial Aquifer Head Data 1967 or 1968, and 1975

(elevation above mean sea level)

Location	1975	1967 or 1968
10.24.24.331	3511.88	3511.57
10.24.35.444	3527.12	3529.62
10.25.15.323	3475.55	3474.77
10.25.17.142a	3497.15	3496.91
10.25.31.413	3485.65	3485.42
10.25.33.331	3464.46	3465.73
11.23.13.232	3525.88	3529.35
11.24.14.331	3522.00	3424.50
11.24.20.333	3522.14	3424.36
11.24.24.231	3516.11	3515.37
11.25.06.421c	3495.20	3492.51
11.25.25.114	3439.15	3436.00
12.25.09.442	3486.56	3486.81
12.25.23.344	3431.89	3432.77
12.25.34.211	3451.47	3461.00
12.26.17.143	3434.58	3433.48

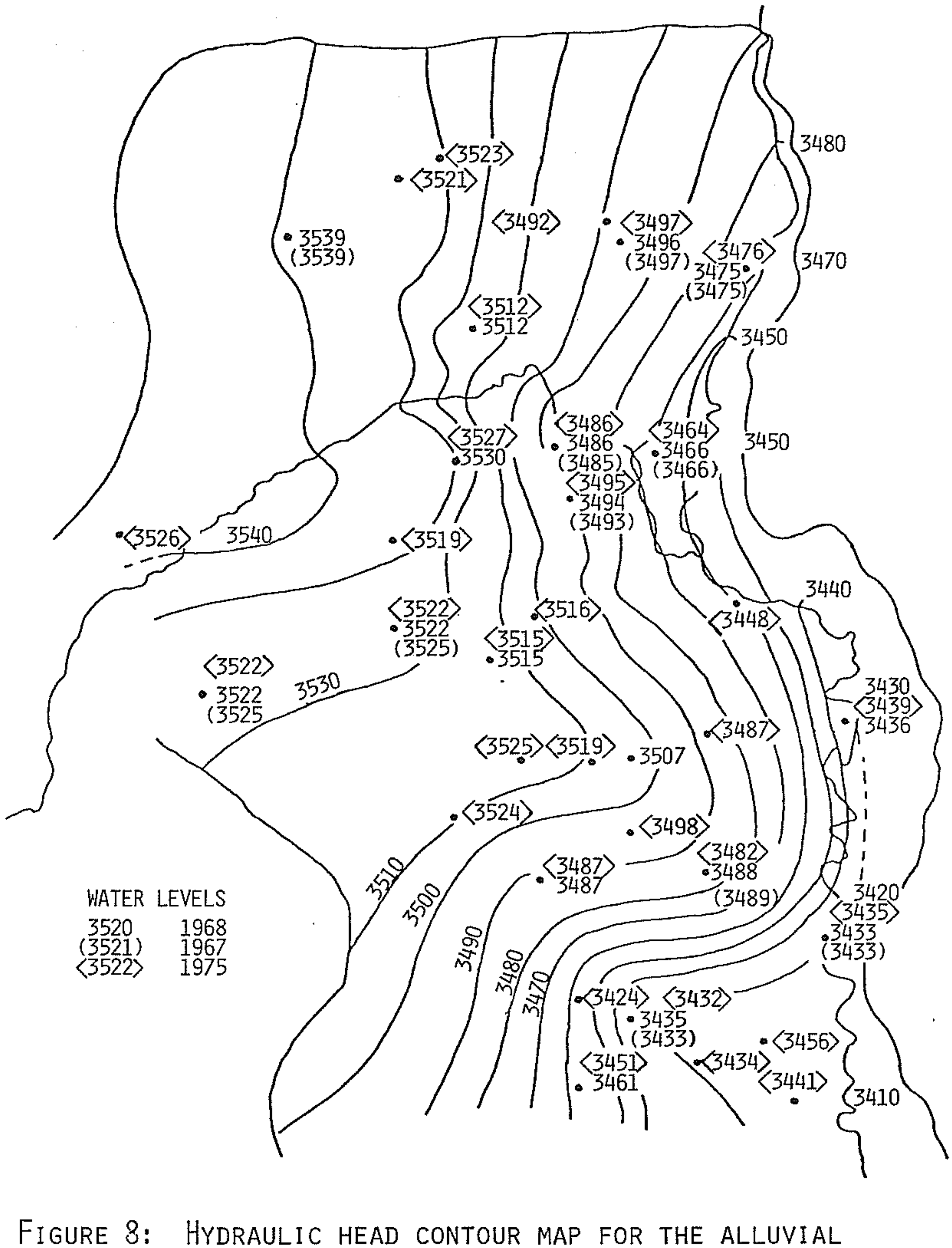

FIGURE 8: HYDRAULIC HEAD CONTOUR MAP FOR THE ALLUVIAL AQUIFER, JANUARY, 1968

The error in the predicted hydraulic head in the carbonate aquifer caused by a constant head alluvial aquifer is probably less than 2 feet.

The location of the western limit of the alluvial aquifer is taken from maps drawn by G. E. Welder of the USGS and Saleem and Jacob (1971). The western limit of the aquitard was assumed to be coincident with the western limit of the alluvial aquifer. The storage coefficient was modified to account for the transition zone from artesian to water table conditions.

Transmissivity and Storage Coefficient

In the area of confined flow, few numerical values are available for the transmissivity and storage coefficient of the carbonate aquifer. Values of transmissivity as derived from pumping tests are summarized in Maddox (1969b, p55). His values are compiled from Hantush (1957,1961) and from preliminary results of W.K. Summers. Saleem and Jacob (1971) list 76 values of transmissivity derived from routine step drawdown tests. The individual values vary greatly from well to well, but the average values for each Township are similar. The values obtained from pumping tests are more accurate and were used as initial values in the model. The pumping test values and the Township average step drawndown values are presented in Table 5 . As an initial first guess,

Table 5

Transmissivity and Step Drawdown Data

Location	gpd/ft	m^2/day	Source
10.24.09.333	1.44×10^6	1.79×10^4	Hantush (1957)
10.25.33.441	1.89×10^6	2.35×10^4	Hantush (1961)
10.25.32.423	1.16×10^4	1.44×10^2	Hantush (1961)
11.24.26.433	1.45×10^6	1.80×10^4	Hantush (1957)
11.25.14	7.70×10^5	9.57×10^3	Maddox (1969b)
12.24	1.50×10^6	1.86×10^4	Havenor (1968)
12.26.33	5.20×10^4	6.46×10^2	Maddox (1969b)
13.25.23.311	7.50×10^4	9.29×10^2	Hantush (1957)
13.25.36	5.00×10^5	6.21×10^3	Maddox (1969b)
13.26.03	4.00×10^5	4.97×10^3	Maddox (1969b)
8.24	2.30×10^5	2.85×10^3	Saleem and Jacob (1971)
9.24	1.17×10^5	1.46×10^3	Saleem and Jacob (1971)
10.23	2.53×10^5	3.14×10^3	Saleem and Jacob (1971)
10.24	3.73×10^5	4.63×10^3	Saleem and Jacob (1971)
11.23	4.19×10^5	5.20×10^3	Saleem and Jacob (1971)
11.24	3.40×10^5	4.23×10^3	Saleem and Jacob (1971)
11.25	2.51×10^5	3.12×10^3	Saleem and Jacob (1971)
12.23	2.89×10^5	3.60×10^3	Saleem and Jacob (1971)
12.24	2.80×10^5	3.47×10^3	Saleem and Jacob (1971)
12.25	2.26×10^6	2.81×10^4	Saleem and Jacob (1971)

the transmissivity of the entire confined zone was assumed to be 1.4×10^6 gpd/ft (Hantush, 1957).

The transmissivity in the unconfined area of flow has not been measured directly. The transmissivity west of Roswell is generally estimated to be the same as in the confined area (Maddox, 1969b; Theis, 1951; Kinney and others, 1968; Motts and Cushman, 1964). Hantush (1957) and Saleem and Jacob (1971) estimated values substantially lower than in the area of confined flow near Roswell. Their values, however, are an average estimated for the entire area of unconfined flow in the carbonate aquifer and are not directly comparable to the other estimates which are for only the area west of Roswell. Note that the area west of Roswell is much different from any other unconfined portion of the carbonate aquifer. This point will be brought out again when the recharge distribution to the carbonate aquifer is discussed.

Because no measurements for transmissivity are available for the unconfined zone in the study area, it was estimated as follows. First, the study area was divided into 3 regions: (1) confined carbonate, (2) unconfined carbonate, and (3) unconfined Glorieta and Yeso. Using Darcy's law and assuming that the flow rate in each region is the same and that no recharge occurs, we can write

$$Q = T_1 \frac{\Delta h_1}{\Delta L_1} D = T_2 \frac{\Delta h_2}{\Delta L_2} D = T_3 \frac{\Delta h_3}{\Delta L_3} D \qquad (4)$$

where

Q = flow rate

T = transmissivity

h = change in head

L = distance along the flow path corresponding to h

D = width (assumed equal to 1 unit)

and the subscripts are:

1 - confined zone

2 - unconfined limestone

3 - unconfined Glorieta and Yeso.

If the values of h and L are estimated from Mourant (1963, plate 1) as

$\Delta h_1 = 10$ feet,

$\Delta h_2 = 60$ feet,

$\Delta h_3 = 700$ feet,

$\Delta L_1 = 8$ miles,

$\Delta L_2 = 8$ miles,

$\Delta L_3 = 9$ miles,

and T is assumed to be 1.4×10^6 gpd/ft (Hantush, 1957), then

$T_2 = 233{,}333$ gpd/ft

$T_3 = 22{,}500$ gpd/ft.

This was meant to be only a rough estimate with which to begin calibration. If recharge were taken into account we would get

$$Q - QRE\ (DA) = T(\Delta h/\Delta L)\ D \tag{5}$$

where QRE(ΔA) is the depth of recharge multiplied by the area. Therefore the actual transmissivities should be smaller than the ones estimated using no recharge. Transmissivities in regions 1 and 2 were arbitrarily reduced by a factor of 10 to account for recharge. The map of the initial transmissivity distribution is shown in Figure 9.

Values for storage coefficient in the confined zone are available from Hantush (1957,1961) and are given in Table 6. As a starting point, a storage coefficient of $1\text{x}10^{-5}$ was used uniformly over the area of confined flow. In the area of unconfined flow only a few estimates of specific yield are available.

The specific yield of the unconfined carbonate aquifer was estimated by Hantush (1957), Theis (1951), Saleem and Jacob (1971) and Maddox (1969b) and is given in Table 7. The values are about the same although they were determined in different ways.

No estimates are available for transmisivity or specific yield of the Yeso or Glorieta. In general, the transmissivity is less than in the carbonate and the specific

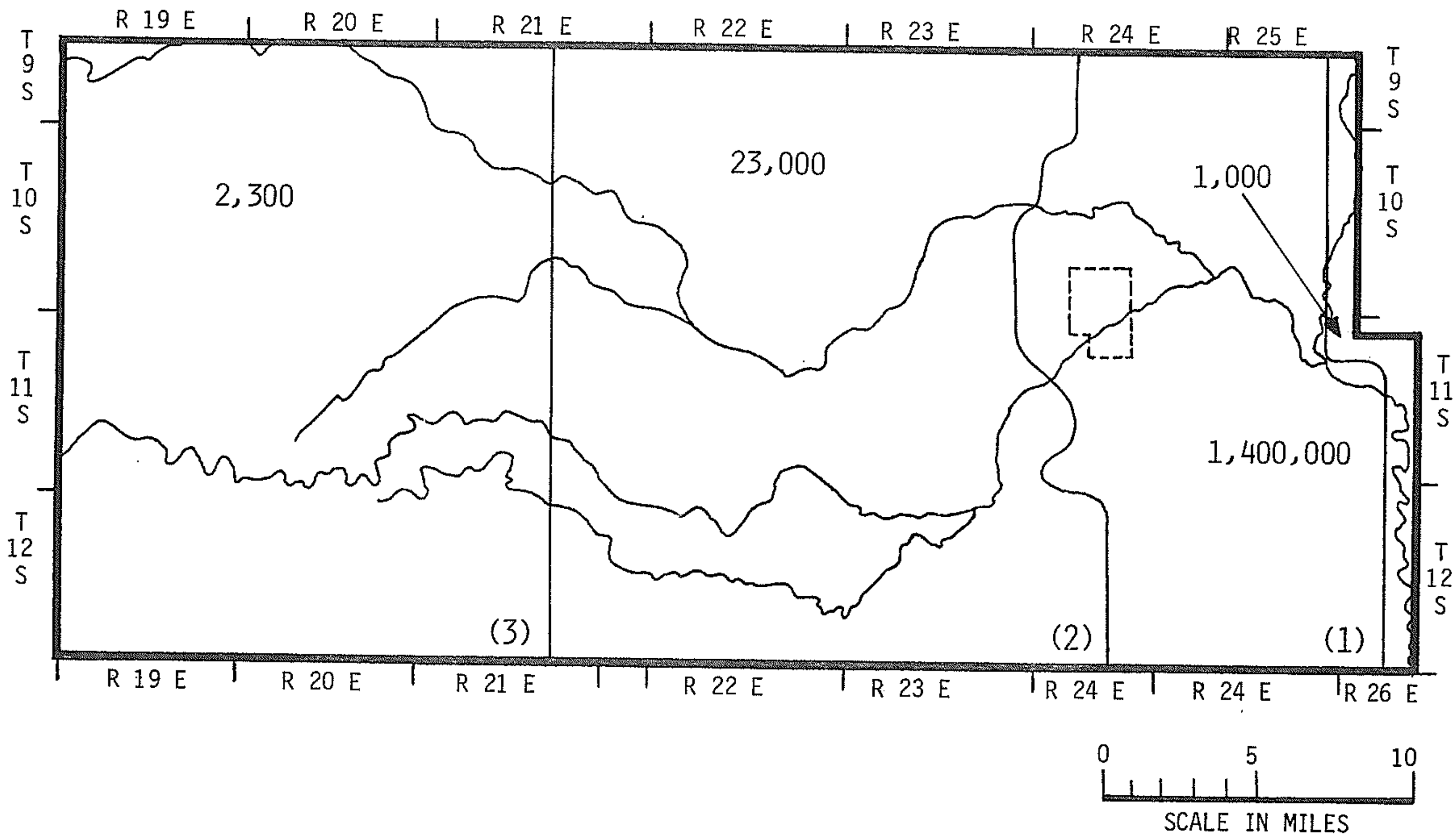

FIGURE 9: INITIAL TRANSMISSIVITY DISTRIBUTION (GPD/FT) CHARACTERIZING THE 3 GROUNDWATER FLOW REGIONS DEFINED IN THIS WORK; THEY ARE FROM EAST TO WEST: (1) CONFINED CARBONATE AQUIFER; (2) UNCONFINED CARBONATE AQUIFER; (3) UNCONFINED GLORIETA AND YESO

Table 6

Storage Coefficient of the

Confined Aquifer

Location	Storage Coefficient	Source
10.24.09.333	$1.5x10^{-5}$	Hantush (1957)
10.25.33.441	$6.7x10^{-5}$	Hantush (1961)
10.25.32.423	$5.7x10^{-5}$	Hantush (1961)
11.24.26.433	$8.4x10^{-6}$	Hantush (1957)
13.25.23.311	$1.3x10^{-5}$	Hantush (1957)
10.25-11.25	$5.5x10^{-5}$	Hantush (1961)

Table 7

Specific Yield of the Unconfined Carbonate Aquifer

Storage Coeficient	Source
<.05	Hantush (1957)
.025	Saleem and Jacob (1971)
.022	Maddox (1969b)
.040	Theis (1951)

yield is greater. A value of 0.01 was used as the specific yield in the unconfined carbonate and 0.02 in the unconfined Yeso and Glorieta. The map of initial storage coefficients is given in Figure 10.

Hydraulic Conductivity of the Aquitard

Hantush (1957,1961) calculated the parameter K′/b′ from his pumping test analyses where K′ is the vertical hydraulic conductivity of the aquitard and b′ is the thickness of the aquitard. The initial values of b′ were obtained from an isopach map of the Queen aquitard (Kinney and others, 1968). The initial distribution of K′ is given in Figure 11. Shortly after the initial simulations, more values of K′ were obtained from Maddox (1969b).

For the initial simulations, it was assumed that K′ was greatest along the river because Hantush's values increase in that direction. Later work indicated that this is only true in the area north of the Y-O structural zone (Fig. 3). For later simulations the data were obtained as follows.

Maddox (1969b,p68) gives values for the vertical hydraulic conductivity of the aquitard derived from Hantush's (1957,1961) pumping tests and from measurements made by R.W. Stallman (USGS), of the vertical temperature distribution across the aquitard. Stallman (1963) presented the governing

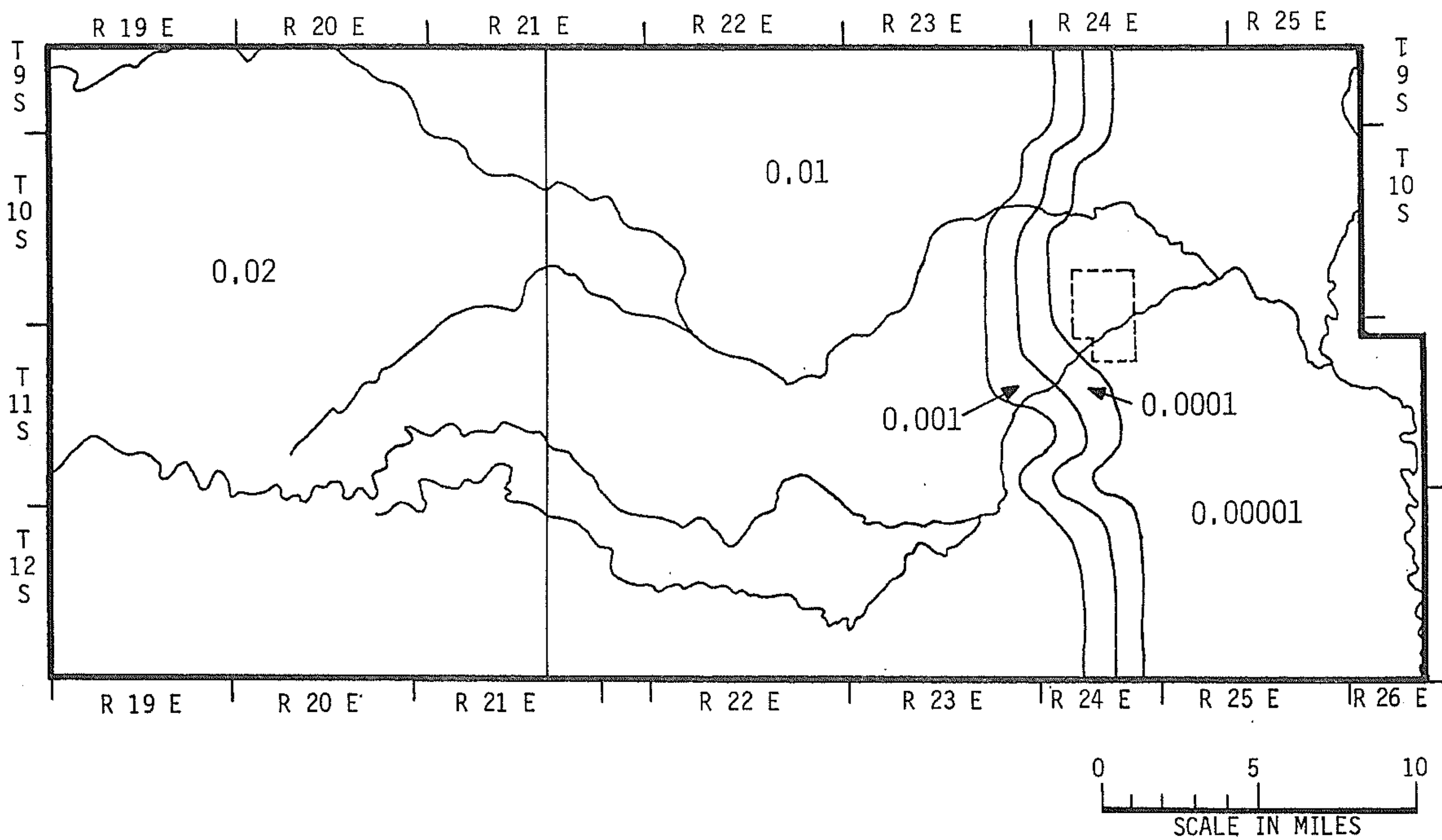

Figure 10: Initial distribution of the storage coefficient

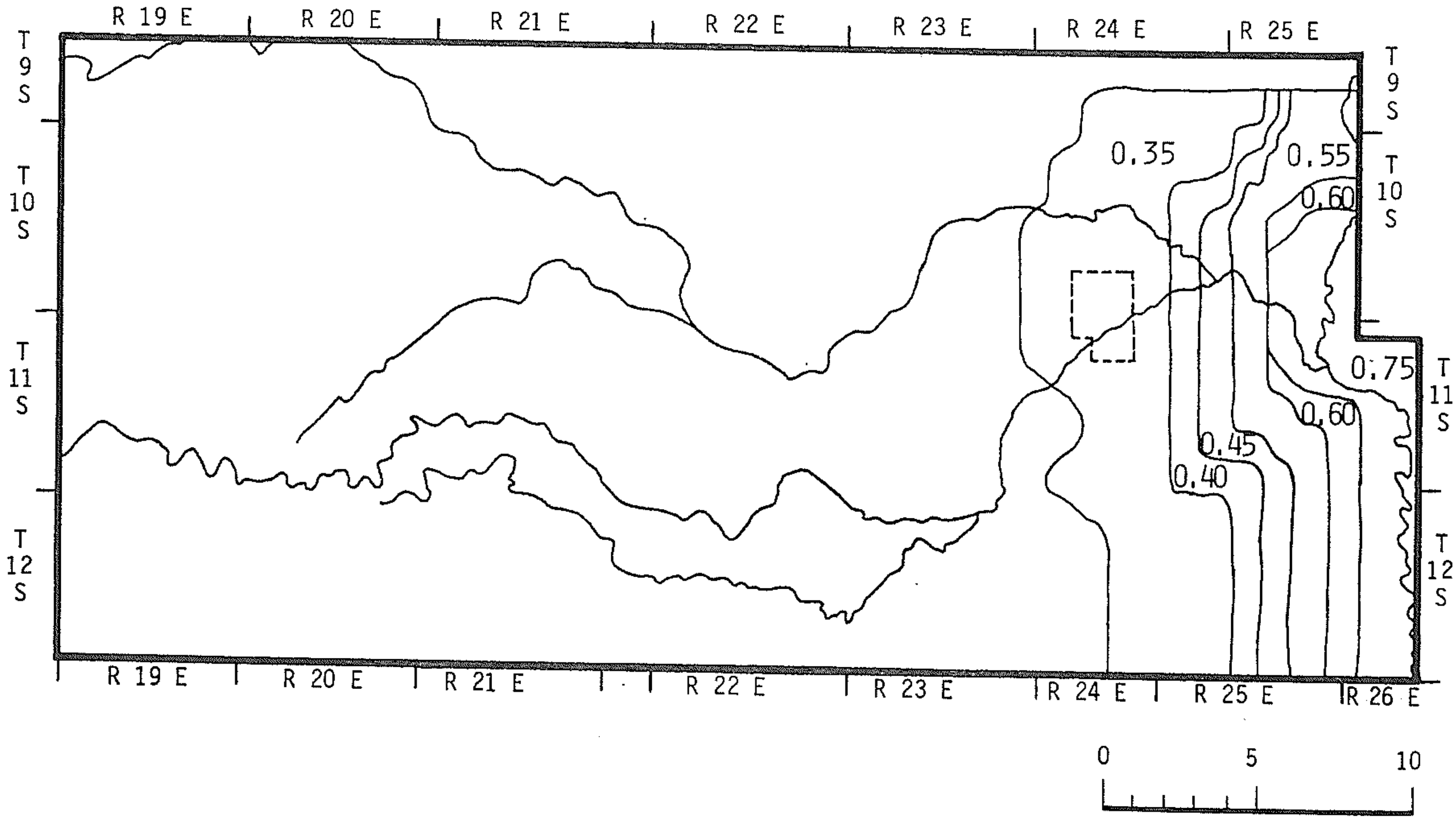

FIGURE 11: INITIAL DISTRIBUTION OF THE VERTICAL HYDRAULIC CONDUCTIVITY (GPD/FT^2) IN THE AQUITARD

equations and Bredehoeft and Papadopoulos (1965) gave an analytical solution for the case of vertical groundwater flow, as in an aquitard. Sorey (1971) also examined temperature profiles in the aquitard and obtained values of the upward flow rate similar to the values obtained by Hantush. In all cases, the thickness of the aquitard is needed in order to calculate the vertical hydraulic conductivity. Maddox (1969b), and Kinney and others (1968) both present maps of the thickness of the aquitard, but Maddox is more reliable because he used the top of the confined aquifer and the bottom of the alluvial aquifer to define the aquitard thickness whereas Kinney and others (1968) give the thickness of the rock units supposedly comprising the aquitard without regard to the varying stratigraphic position of the artesian and alluvial aquifers. Initial aquitard hydraulic conductivity estimates (Table 8) were obtained using Hantush's K'/b' values and the value of b' from Maddox as well as the values of K' obtained from thermal gradient studies.

As a matter of reference, Neuman and Witherspoon (1969a, 1969b, and 1972) developed an improved method of determining aquifer parameters in leaky artesian systems such as the Roswell Basin. In their 1969b paper, they compare their method to that of Hantush and state that in some cases Hantush's method of pumping test analysis may overestimate

Table 8

Vertical Hydraulic Conductivity of the Aquitard

(gallons per day per foot squared)

Location	K′	Source
10.24.09.333	.22	Hantush (1957)
10.25.32.423	.49	Hantush (1961)
10.25.33.441	.52	Hantush (1961)
11.24.26.433	.15	Hantush (1957)
12.25.23.344	.017	Maddox (1969b)
13.25.23.311	.03	Hantush (1957)
13.23.27.211	.016	Maddox (1969b)

the value of the aquifer transmissivity by up to 40% and underestimate the vertical hydraulic conductivity of the aquitard by a factor of 5. Hantush's values were not adjusted for the model because Neuman and Witherspoon only analyzed hypothetical situations and the amount of adjustment needed, if any, in the Roswell Basin could not be determined. Also, the values obtained by Stallman (as given in Maddox, 1969b) and by Sorey (1971) are similar to those of Hantush. The vertical hydraulic conductivity of the aquitard is an important variable in the model and more work needs to be done in order to determine if Hantush's values are in error.

Specific Storage Coefficient of the Aquitard

The specific storage coefficient of the aquitard has not been estimated in the Roswell Basin. Havenor (1968) described the Queen aquitard as being composed primarily of very fine-grained red sandstones and siltstones. The specific storage can be calculated from

$$S_s = \rho g \ (\alpha + n\beta) = \gamma_w \ (\alpha + n\beta) \tag{6}$$

where:

γ_w - the specific weight of water

α - the compressibility of the formation

β - the compressibility of water

n - porosity.

Freeze and Cherry (1979,p55) list values of α for different materials. A value of $1x10^{-8}$ is estimated for the material as described by Havenor. β is also given by Freeze and Cherry (1979). The porosity was estimated to be in the range of 0.06 to 0.27 (Summers, 1972). Fortunately the equation is not sensitive to porosity and the difference in calculated specific storage using 0.06 and 0.27 was insignificant. The specific storage calculated by equation 6 is $1x10^{-4}$ $feet^{-1}$.

Pumping

Pumping data, in one form or another, are available from 1938 to date. Prior to 1967, the volume of water pumped was only estimated (Fiedler and Nye, 1933; Mower, 1960). After 1967, pumpage from irrigation and municipal wells was measured with flow meters. Records of the metered pumpage are available from the State Engineer Watermaster reports and from individual well schedules. The spatial distribution of pumpage for 1967 was determined from the available records. The magnitude of total yearly pumping was varied according to the Watermaster records for the period 1968 to 1975.

According to Fiedler and Nye (1933) and Ray Wyche of the PVACD (personal communication) irrigation usually begins around March 1 and ends about September 30. The calibration period, January 1967 to January 1968, was divided into 3 pumping periods which are: (1) January 1 to February 28, (2) March 1 to September 30, and (3) October 1 to December 30. All the metered pumping is assumed to occur during the second time period. The assumption of no pumping in the winter causes less than a 2% error in the pumping data because 98% of the total yearly pumpage is for irrigation (State Engineer Watermaster Reports). The pumpage per node is given in Table 9. Most of the study area is what the State Engineer Watermaster calls the Roswell-East Grand Plains Area. The study area also includes a small portion of the Northern Extension and the Dexter-Hagerman Area. Therefore, the pumpage for 1967 in the study area (84,959 acre-feet) is greater than the Watermaster's reported pumpage (70,140 acre-feet) in the Roswell-East Grand Plains Area.

In order to simulate 8 years, 1967 to 1975, without having to break down individual yearly pumping data, the spatial distribution of pumpage in 1967 was assumed to be constant over that period. This is a reasonable assumption because the basin is considered closed in terms of the drilling of additional new wells unless old ones are taken out of service. Also, the model is not sensitive to small changes in the pattern of pumping. For example, no difference in computed head was seen when the pumpage at a node was distributed evenly to the surrounding 4 nodes and then set to zero at that node.

Table 9

Pumpage per node (cubic feet/sec)

Y	X	Pumpage	Y	X	Pumpage
2	31	.75	12	42	5.03
2	32	1.11	12	43	1.13
2	36	.60	13	33	1.06
3	31	2.72	13	34	.33
3	35	.09	13	36	3.73
3	36	.16	13	37	2.15
5	28	1.07	13	38	3.28
5	29	.18	13	39	5.29
5	32	.04	13	40	2.87
5	42	.45	13	41	6.89
6	29	1.44	13	42	.58
6	30	1.20	13	43	3.95
6	33	.91	14	29	3.38
6	34	1.84	14	30	.66
6	35	1.40	14	35	1.47
6	36	1.66	14	36	2.16
7	30	1.72	14	37	4.60
7	31	1.51	14	38	4.53
7	33	2.65	14	39	4.13
7	34	1.38	14	40	6.68
7	35	.49	14	41	.98
7	36	2.05	15	28	.86
7	37	1.08	15	29	2.32
8	29	1.68	15	30	.34
8	30	4.80	15	39	3.48
8	35	.37	15	40	3.47
8	36	2.73	15	41	3.06
8	37	3.49	16	27	2.45
8	38	.26	16	37	.31
8	39	.03	16	40	.99
9	28	.52	17	27	5.44
9	35	.43	17	28	2.46
9	36	2.64	17	31	1.51
9	37	1.74	17	32	1.53
9	38	.41	17	37	.43
10	31	1.01	17	38	.11
10	32	1.12	17	41	.42
10	33	.06	17	42	.65
10	36	2.31	18	38	.49
10	37	3.06	18	39	1.50
10	38	1.05	18	41	1.04
11	29	.82	18	43	.15
11	32	3.91	19	32	2.43
11	33	1.47	19	34	1.65

X	Y	Pumpage	X	Y	Pumpage
11	35	.25	19	43	1.12
11	36	3.87	20	39	.62
11	37	3.71	20	40	1.02
11	38	2.94	20	42	2.19
11	40	3.44	20	44	.35
12	33	.41	21	39	.52
12	34	.15	21	40	2.08
12	36	1.96	21	41	1.34
12	37	3.78	21	42	.33
12	39	5.21	21	43	1.46
12	40	1.92	21	44	35.00
12	41	2.29			

To generate 8 years of pumping data, the 1967 pumping distribution was multiplied by the ratio of the yearly metered pumpage in the Roswell-East Grand Plains Area for that year divided by that of 1967 (Table 10).

Recharge

Recharge is probably the least known parameter. Fiedler and Nye's (1933) assumption, that the major source of recharge is precipitation falling on the Principal Intake Area, has been adopted by most subsequent workers. Gross and others (1976,1979), however, have proposed that some of the recharge is derived from upward leakage from the Glorieta and Yeso. Most investigators acknowledge the recharge derived from the Rio Hondo and some of the other streams that cross the Principal Intake Area.

In determining the initial recharge, the assumption was made that all recharge comes from precipitation. The purpose was to test whether the groundwater system could be modeled without an upward leakage source. Precipitation is unevenly distributed during the year with most of it falling during the summer. The recharge was then distributed in approximately the same manner as the precipitation. Three recharge periods were used corresponding to the three pumping periods (see above). In Table 11 the average distribution of precipitation at Roswell, Bitter Lakes, and Picacho is given

Table 10

Yearly Pumpage, 1967 - 1975

(expressed as a fraction of 1967 pumpage)

(acre-feet)

YEAR	ROSWELL PUMPAGE	% OF 1967
1967	70,139.9	100
1968	63,442.1	90
1969	72,023.8	103
1970	72,224.9	103
1971	78,627.4	112
1972	72,512.4	103
1973	79,264.9	113
1974	77,794.3	111
1975	74,058.5	106

Table 11

Seasonal distribution of precipitation

Period	Percentage of annual		
	1967	1955-1974	model
Jan - Feb	2.3	9.0	3.0
Mar - Sep	87.0	75.0	87.0
Oct - Dec	10.7	16.0	10.0

Table 12

Stream Leakage Estimates - Rio Hondo

(acre-feet/year)

Water Year	Picacho	Diamond-A	Loss
1958	32,110	23,660	8,450
1959	14,350	8,880	5,470
1960	8,530	3,490	5,040
1961	6,660	1,620	5,040

for 1967 and for the period 1955-1974, along with the initial distribution of recharge used in the model which was based on the 1967 distribution of precipitation. Summer evaporation of rainfall is ignored in the estimation of recharge to give an upper estimate of precipitation recharge.

Bean (1949), and Gross and others (1976) believe that the percentage of precipitation falling on the Principal Intake Area that becomes recharge, estimated to be 25% by Fiedler and Nye (1933), is too high. As a first attempt to solve this problem, the initial precipitation percentage input to the model was estimated to be 5%, and was calculated as follows:

Average total precipitation = 10.62 inches = .885 feet (Roswell and Picacho for 1967)

Area of unconfined flow = 661 sq. miles = 423,040 acres

Total volume of precipitation = 374,214 acre-feet

5% = 18,711 acre-feet

28.31 acre-feet/square mile/year.

Then, using the percentages of yearly rainfall distribution, the amount of recharge per node per pumping period is

Pumping period	Recharge acre-ft/node	Months in period	Recharge acre-ft/month/node
1	.75	2	.379
2	24.72	7	3.532
3	2.83	3	.944

In addition to the recharge from precipitation, the Rio Hondo also looses appreciable amounts of water that may become recharge (Duffy and others, 1978; Bean, 1949). The Rio Hondo generally has a perennial flow at Picacho. The flow record at Picacho is incomplete, but the loss between the Picacho gaging station and the Diamond-A gaging station can be estimated from the stream flows at the two stations which is given in Table 12.

The flow at Diamond-A in 1967 was 6,600 acre-feet which occurred between June and October. The flow is within the range given in Table 12 for 1959-1961. The loss of flow between Picacho and Diamond-A was about 5,000 acre-feet per year for each of those years. The recharge for 1967 will be assumed to be about 5,000 acre-feet also. Water rarely flows to Roswell except during floods. Therefore, if evapotranspiration is neglected, the total flow at Diamond-A will be lost in the Principal Intake Area. This will overestimate the actual recharge from the Rio Hondo, but this may be offset, at least in part, by not considering recharge by other streams. The recharge for 1967 above and below the Diamond-A gaging station is 5000 and 6600 acre-feet, respectively. Above Diamond-A station the flow was assumed perennial and was distributed evenly over the 25 nodes the river occupies in the model to yield 16.67 acre-feet/month/mile. The recharge above the Diamond-A station was held constant for the verification period. This will cause, at most, a 3000 acre-feet error during periods of high stream flow.

Below the Diamond-A gaging station, the recharge was distributed above and below the Diamond-A dam based on the recorded flow at both stations. The flow is assumed ephemeral and to occur between March 1 and September 30. The initial recharge is given in Figure 12.

During the 8 year simulations, the amount of leakage below the Diamond-A gaging station was varied every year in accordance with the measured yearly flow in the Rio Hondo at the gaging station and the dam. Table 13 contains the yearly flow at the Diamond-A Ranch and at a point just below the Diamond-A dam along with the monthly recharge per node from the Rio Hondo above and below the Diamond-A dam.

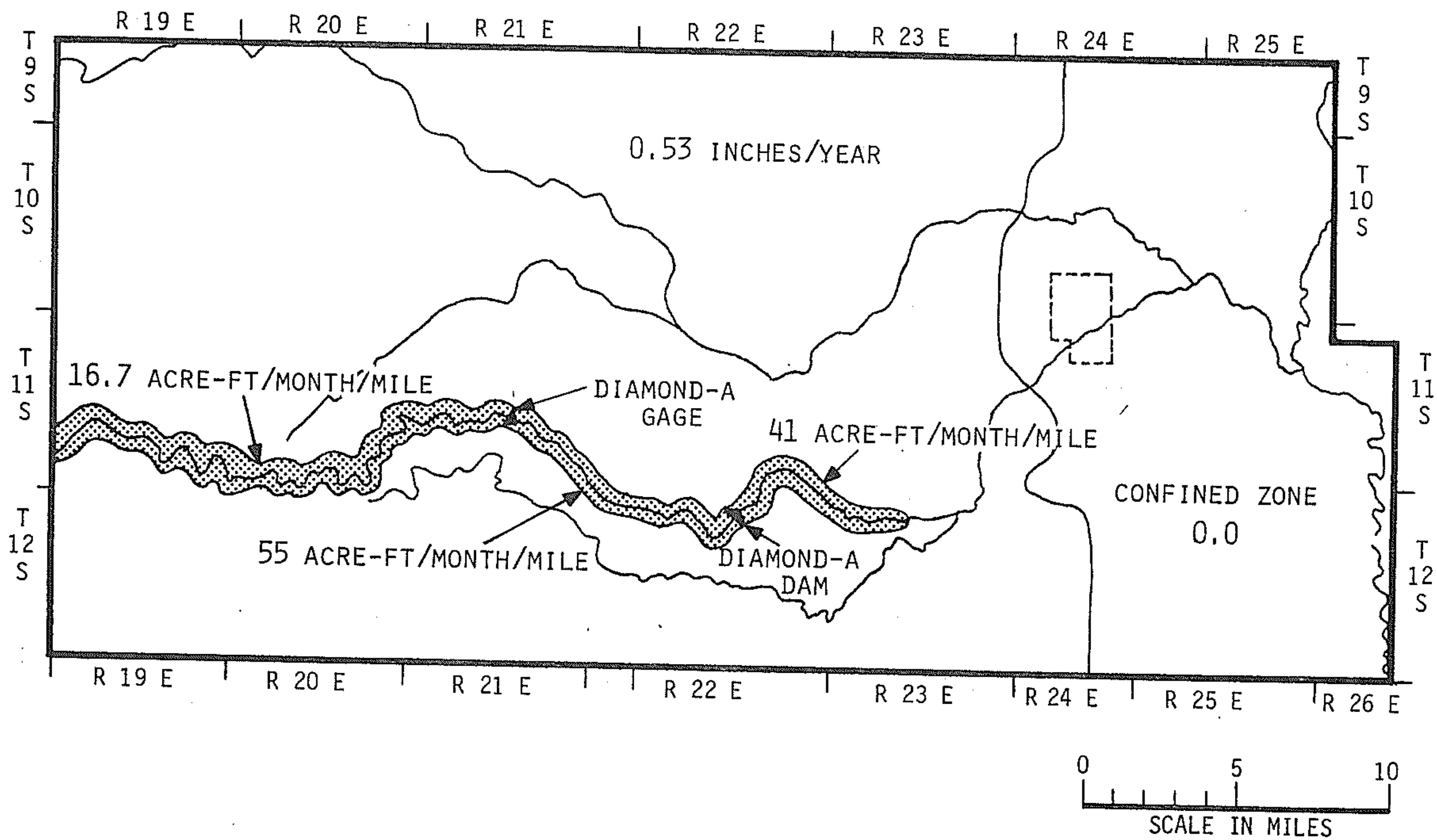

FIGURE 12: INITIAL RECHARGE DISTRIBUTION

Table 13

Rio Hondo; Yearly Discharge;

Monthly Recharge per Node

	Yearly Flow Diamond-A Gage (acre-ft)	Yearly Flow Diamond-A Dam (acre-ft)	Recharge between Diamond-A Ranch and Diamond-A dam (acre-ft/node)	Recharge Below Dia.-A dam (acre-ft/node)
1967	6,600	3,960	55	41
1968	16,250	6,830	195	71
1969	9,400	7,640	36	80
1970	790	422	7	4
1971	6,620	3,480	65	36
1972	19,170	12,480	138	130
1973	17,290	13,840	71	144
1974	10,770	8,150	54	85

CALIBRATION

During calibration some criterion must be established to determine if the computed heads match the observed heads. An exact match is unwarranted because the observed head map is approximated from discrete points and probably somewhat in error, and the added expense in terms of effort and computer time yields a diminishing return. For the present study, a node is considered in agreement when the computed head was within 3 feet (high or low) of the observed head. This criterion was relaxed for the western portion of the study area because the hydraulic gradients are much steeper than near Roswell.

The calibration process is not unique in that more than one set of parameters will produce the same head distribution. To help alleviate the nonuniqueness problem, constraints can be imposed on some of the parameters if field measurements are available. For example, a range of possible values for transmissivity and storage coefficient can be imposed in an area, based on pumping test results. The following is a brief description of the calibration procedure, especially the reasoning behind the various parameter changes that result in the final parameter distributions to be presented below.

The results of the initial simulation are presented in Figure 13. The match between the computed and observed head distribution is poor. The area of best fit, or least error, was the region of groundwater flow in the Yeso and Glorieta; therefore parameters in that area were adjusted first. Transmissivity was adjusted while storage coefficient and recharge were unchanged because transmissivity determines the amount of water that flows downslope and ultimately enters the carbonate aquifer from the west. In the areas away from the Rio Hondo the value of transmissivity generally decreased. The calibrated values ranged from 10 to 4,000 gpd/ft.

In addition to reducing transmissivity, the recharge was uniformly decreased, because the initial recharge was greater than the ability of the aquifer to transmit the water, and water levels were rising. The storage coefficient was increased from 0.02 to 0.05 to help alleviate groundwater mounding.

The recharge derived from the leakage of the Rio Hondo also required adjustment. As stated previously, the river leakage above Diamond-A gage was initially distributed evenly along the channel. It soon became apparent that leakage was not evenly distributed, but was concentrated largely between the 4400 and 4500 foot water level contours (Figure 13). The storage coefficient of nodes associated with the river was

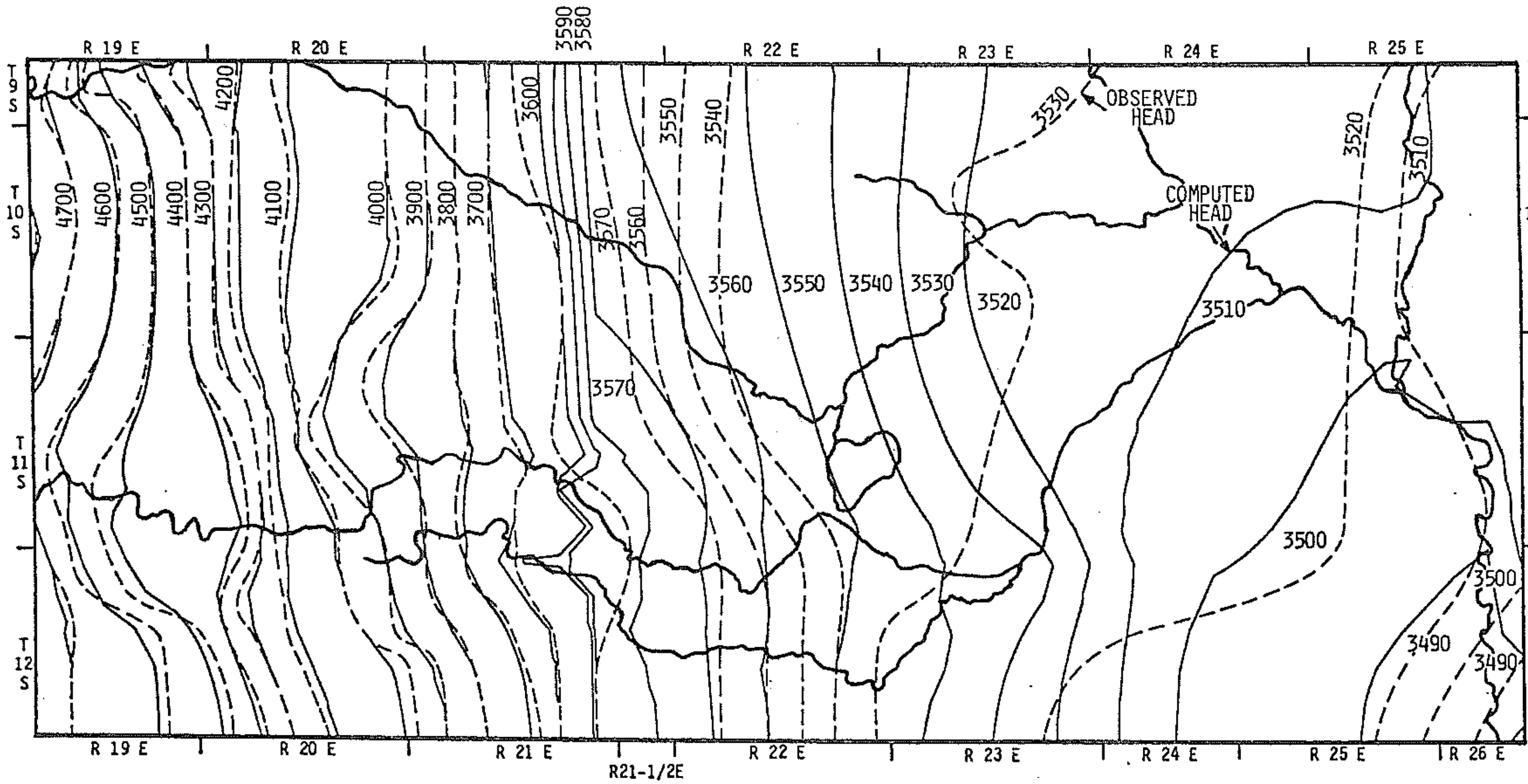

FIGURE 13: HYDRAULIC HEAD, JANUARY 1968, SIMULATED WITH THE INITIAL PARAMETERS (SOLID LINES), DASHED: OBSERVED HEAD DISTRIBUTION

increased from 0.02 to 0.10 because the aquifer along the river is largely alluvium. The river leakage was decreased to account for consumptive use and evapotranspiration which had been previously neglected. The maximum error between the computed and observed heads in the western region was 13 feet, with the vast majority less than 2 feet. With the western region calibrated, the next area of emphasis was the confined and unconfined carbonate aquifer.

As mentioned earlier, the hydraulic head in the alluvial aquifer, the thickness of the aquitard, and the pumpage were assumed known and therefore were not adjusted during calibration. The remaining parameters to be adjusted during calibration were transmissivity, storage coefficient, vertical hydraulic conductivity of the aquitard, and recharge.

After a few more calibration runs it became evident that 4 unknown parameters were too many and that one or more would need to be specified while the other parameters were adjusted.

The storage coefficient does not directly affect the amount of water entering or leaving a node, but it does affect the magnitude and rate of the water level response. The problem at hand was one of excessive drawdown and lack of recovery in the confined aquifer east of Roswell. The difficulty was due to an insufficient amount of water entering the area from the recharge area to the west. The storage coefficient exerts a lesser effect on the movement of groundwater than does transmissivity, therefore the storage coefficient distribution was reevaluated, replotted, and then left unchanged.

The vertical hydraulic conductivity (K′) of the aquitard has a strong effect on the artesian water level as do transmissivity (T) and recharge (QRE). Although more data exist for T in the confined zone than for K′, the need to calibrate T was deemed greater. Therefore, K′ was reevaluated and fixed.

The available data for K′ and S were plotted and extrapolated to generate Figures 14 and 15. The value of S, 8.4×10^{-6}, was further reduced to 4.4×10^{-6} in the southeast corner of the model area because the annual fluctation of water level is greater than in the area to the north and west and is due, in part, to a smaller value of S. Now only T and recharge remained to be determined from the calibration.

The problem remaining in the calibration was that the water level in the artesian zone was not recovering as it should. Recharge was added to the area of unconfined flow to increase groundwater flow to the pumping centers. As recharge was added, the transmissivity was also increased to

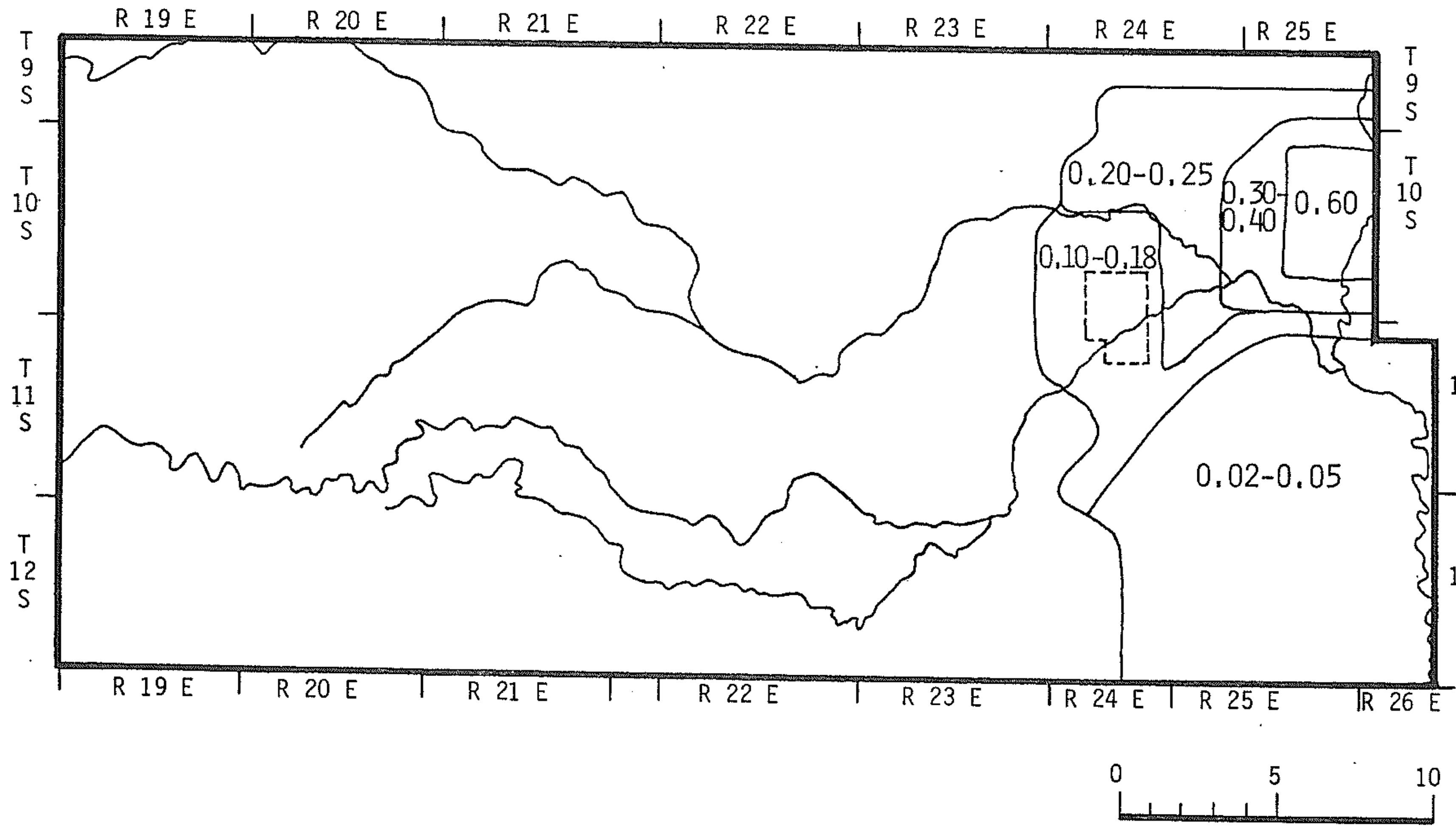

FIGURE 14: IMPROVED VERTICAL HYDRAULIC CONDUCTIVITY (GPD/FT^2) DISTRIBUTION IN THE AQUITARD

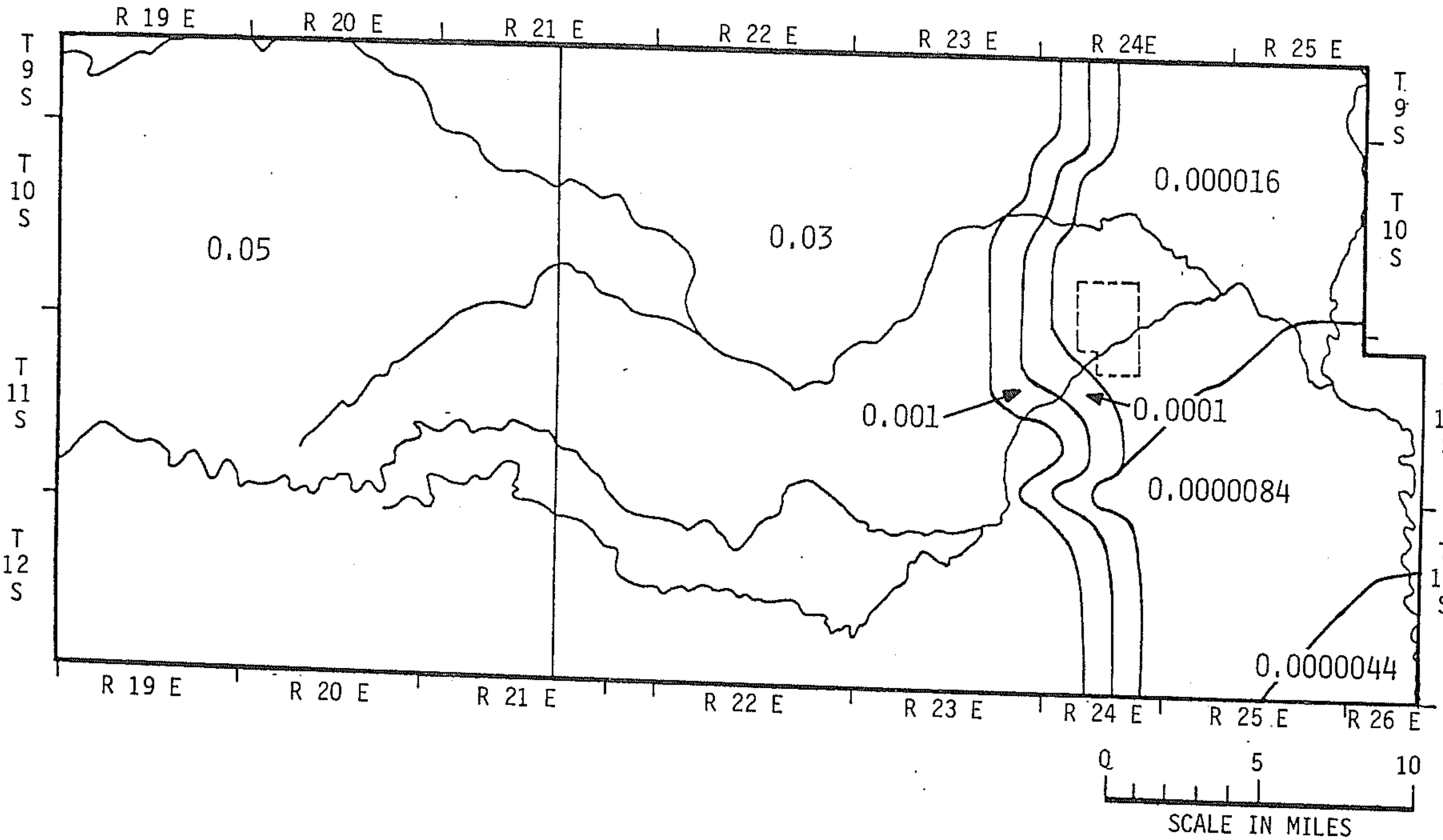

FIGURE 15: IMPROVED STORAGE COEFFICIENT

prevent groundwater mounding in the recharge area. The transmissivity of both the recharge and artesian area was finally increased to 1.8×10^6 gpd/ft. Even this value failed to solve the problem. As a final resort, recharge was added directly to the artesian area and finally the computed water levels matched the observed water levels. The source of the recharge added to the artesian zone is not direct precipitation; leakage derived from above must also be excluded because it is calculated separately in the model. The other possible source is leakage from below. As an indication of this, tritium data presented by Gross and Hoy (1979) from the Clardy well (11.25.15.343) show a decrease in tritium activity during the pumping season indicating that older, deeper water is being drawn into the well.

The southeast corner of the model area is difficult to analyze. In order to match the observed water levels, a zone of low transmissivity and recharge was needed. The anomalously low values were readjusted during the verification runs and were then generally consistent with the other values in the model. The reason for the discrepancy will be discussed later.

The final computed head map and the superimposed observed head map are given in Figures 16a and 16b. The

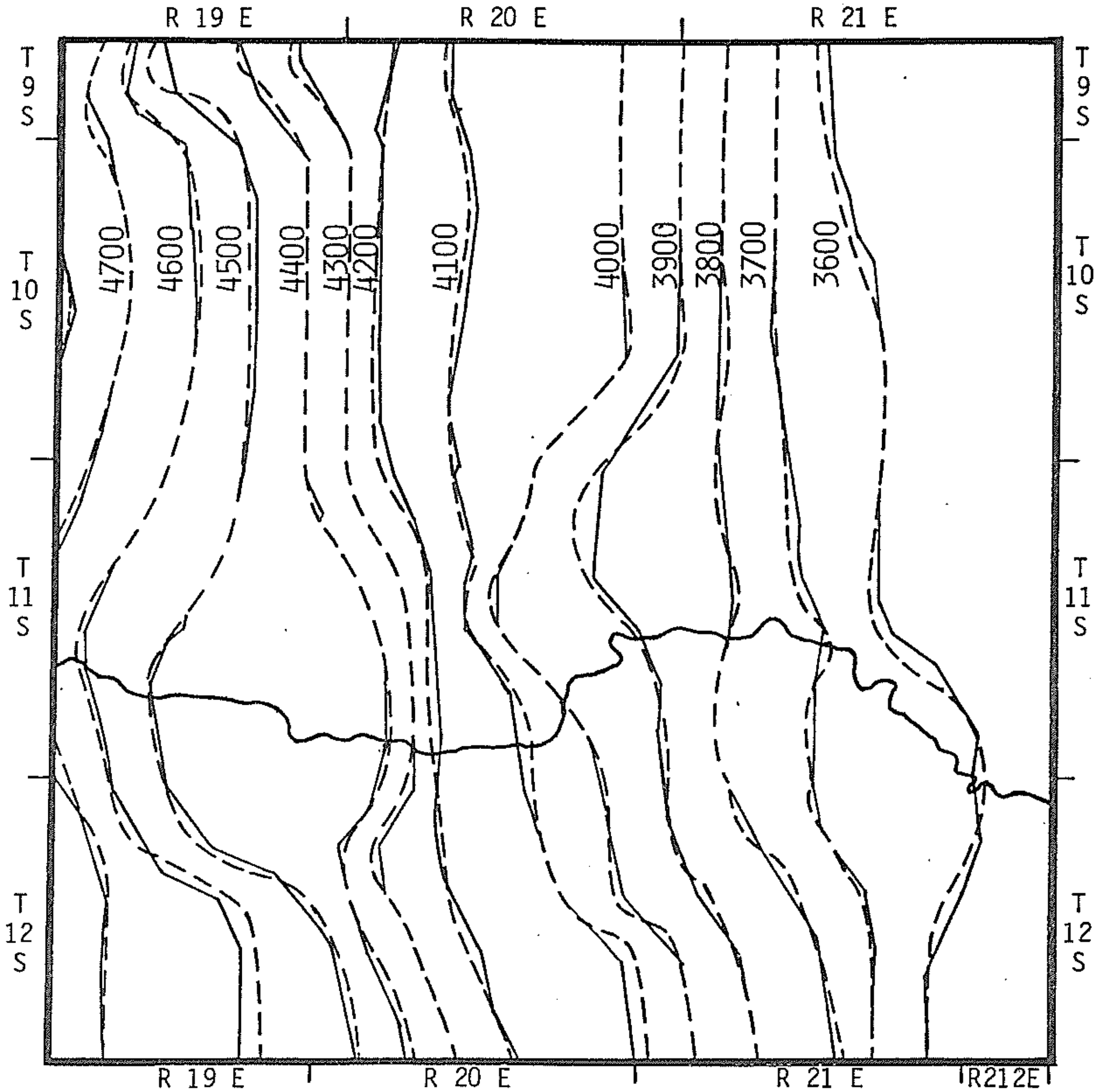

FIGURE 16A: HYDRAULIC HEAD DISTRIBUTION IN THE WESTERN REGION, JANUARY, 1968. SIMULATED WITH THE CALIBRATED PARAMETERS (SOLID LINES)
DASHED: OBSERVED HEAD DISTRIBUTION

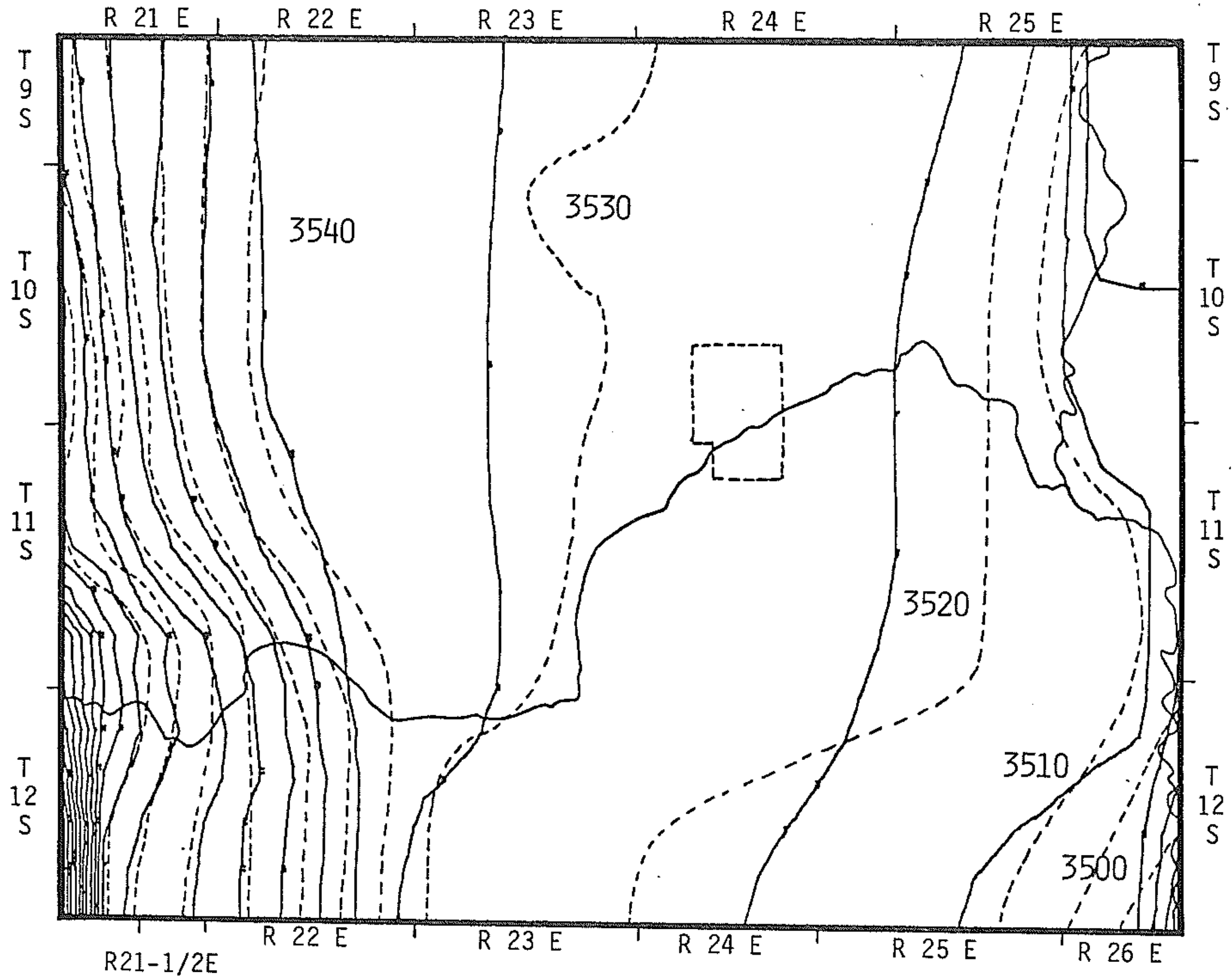

FIGURE 16B: HYDRAULIC HEAD DISTRIBUTION IN THE EASTERN REGION, JANUARY, 1968. SIMULATED WITH THE CALIBRATED PARAMETERS (SOLID LINES). DASHED: OBSERVED HEAD DISTRIBUTION final calibration parameters are presented in the Discussion of Results section. The results are not perfect, but the fit is reasonably good as indicated by Figure 17.

The next step in the calibration process is the verification of the calibrated parameters by simulating a different period of time and checking the results. The computed head and the observed head for the first verification simulation are presented in Figure 18 for the verification period of January 1967 to January 1975. In general, the match is not as good as for the calibration period. The areas of poor fit are the northern area just east of the 3600 foot contour, the southeast section of the model, and the western region. The error in the northern area is probably due to the lack of water level data in that region which means the observed water level map is probably in error. The error in the southeast corner is due to the anomalous parameters in that area (see above). In the western region, water levels continued to rise. In an effort to determine the changes needed to bring the computed and observed 1975 heads into agreement, the 1967 to 1975 period was also calibrated.

The following general changes in storage coefficient were made. The storage coefficient in the western region was increased from 0.05 to 0.08 away from the Rio Hondo and from 0.10 to 0.20 along the Hondo. Transmissivity was decreased

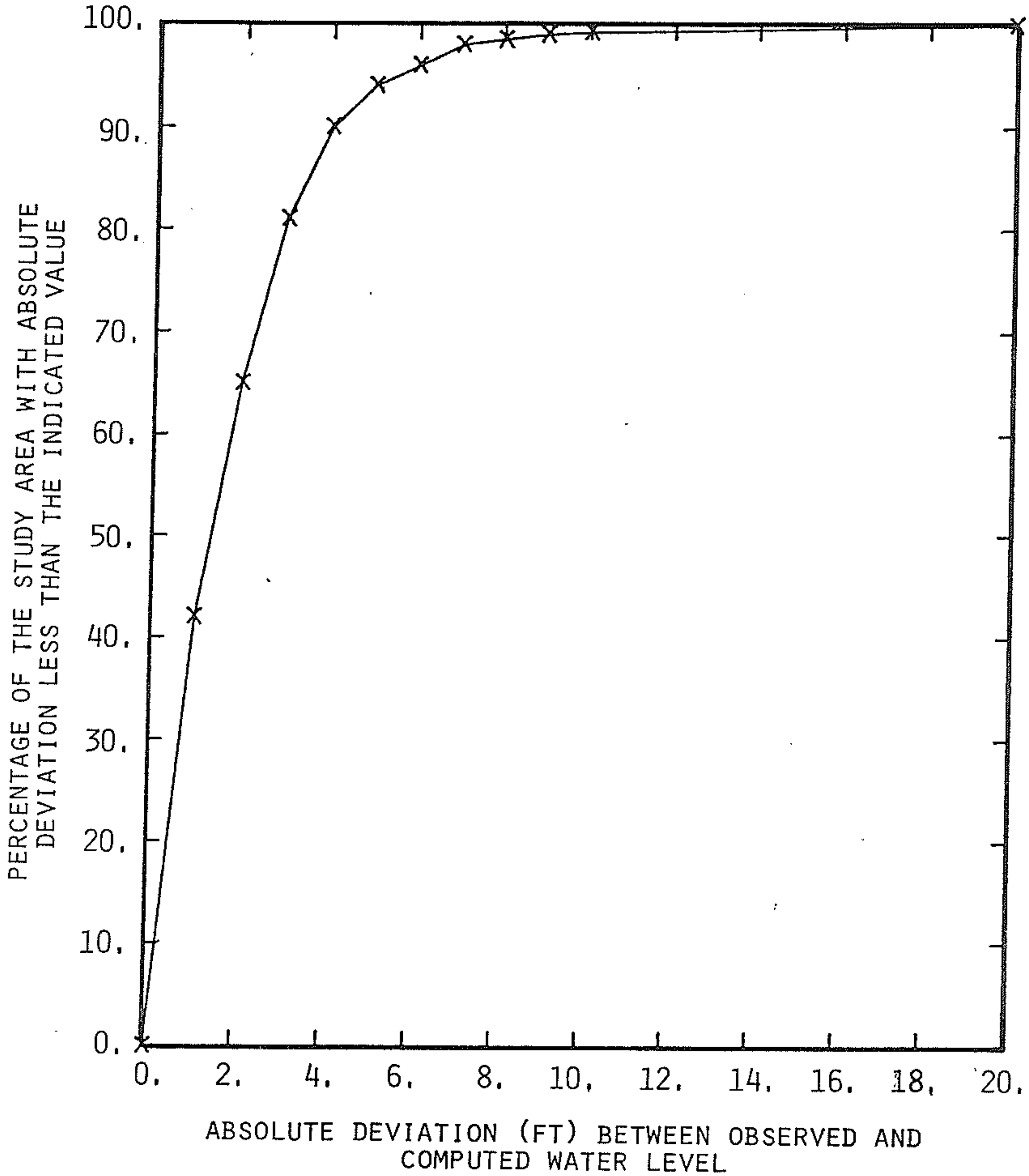

FIGURE 17: GOODNESS OF FIT - CALIBRATION
JANUARY, 1968

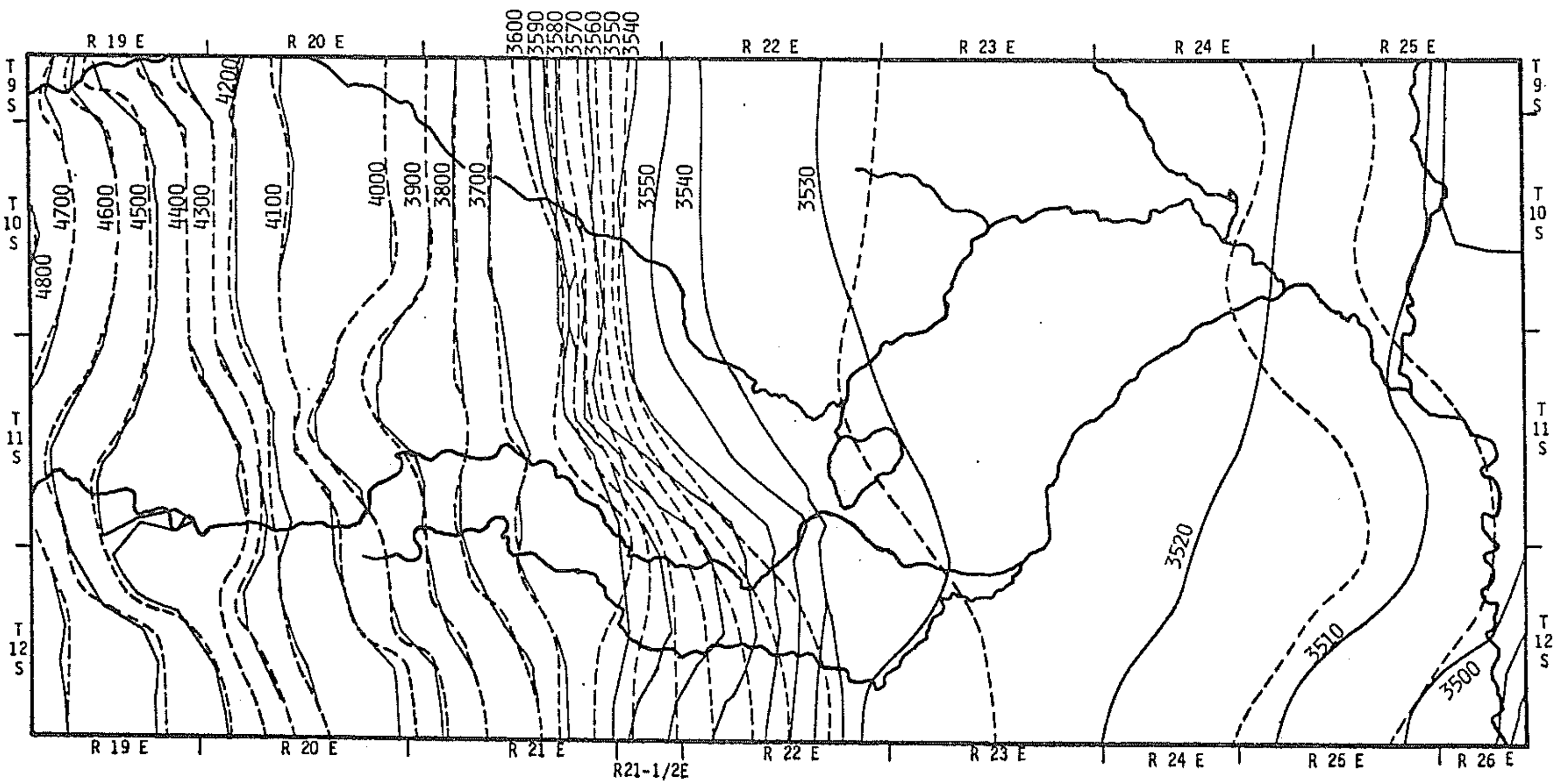

FIGURE 18: HYDRAULIC HEAD DISTRIBUTION, JANUARY, 1975.
INITIAL VERIFICATION.
SIMULATED: SOLID LINES
OBSERVED: DASHED LINES

from $1.8x10^6$ to $1.4x10^6$ north of the Rio Hondo area while recharge was also decreased slightly. In the southeast corner, transmissivity and recharge were both increased and the resulting values were more consistent with the other values in the model. Finally, the area around the Bitter Lakes was modeled as a groundwater discharge area by increasing the vertical leakage and increasing the outflow of water across the eastern boundary there. The final verification results are given in Figure 19. As a matter of convenience, only the eastern 2/3 of the study area is plotted because of the vast difference in water table gradients in the western and eastern regions. The western region remained largely as it did in Figure 16.

Generally, the verification simulation should not need to be calibrated. If the parameter changes are too drastic, the model should be recalibrated. As a check, the parameters derived from the verification simulation were used for the calibration period, January, 1967, to January, 1968. The computed results using the verification parameters is given in Figure 20. The fit is not as good as the calibrated results, but the same general trend remains. When the goodness of fit results are plotted, the curve deviates only slightly from the calibrated curve (Figure 21).

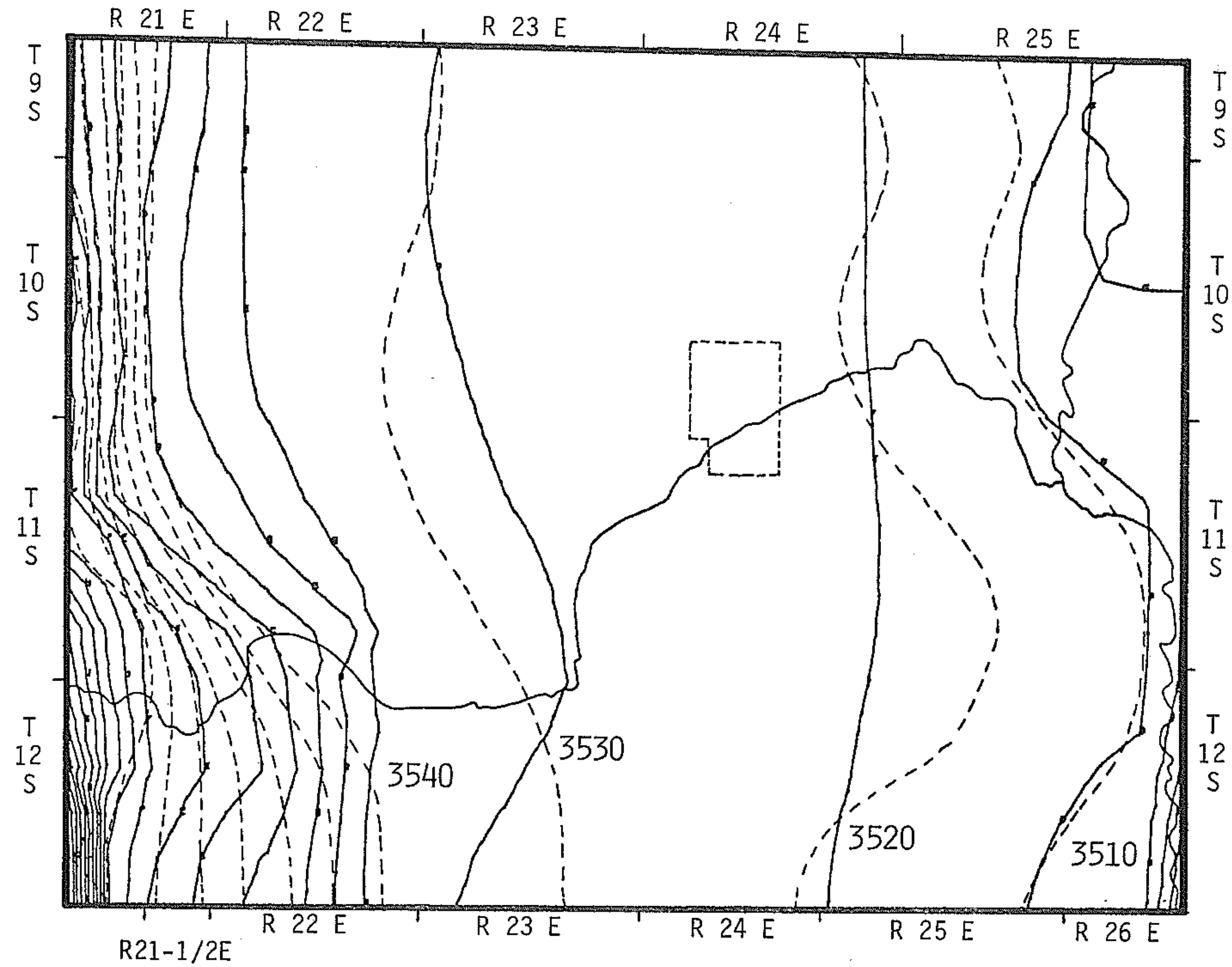

FIGURE 19: HYDRAULIC HEAD DISTRIBUTION, JANUARY, 1975.
FINAL VERIFICATION
SIMULATED: SOLID LINES
OBSERVED: DASHED LINES

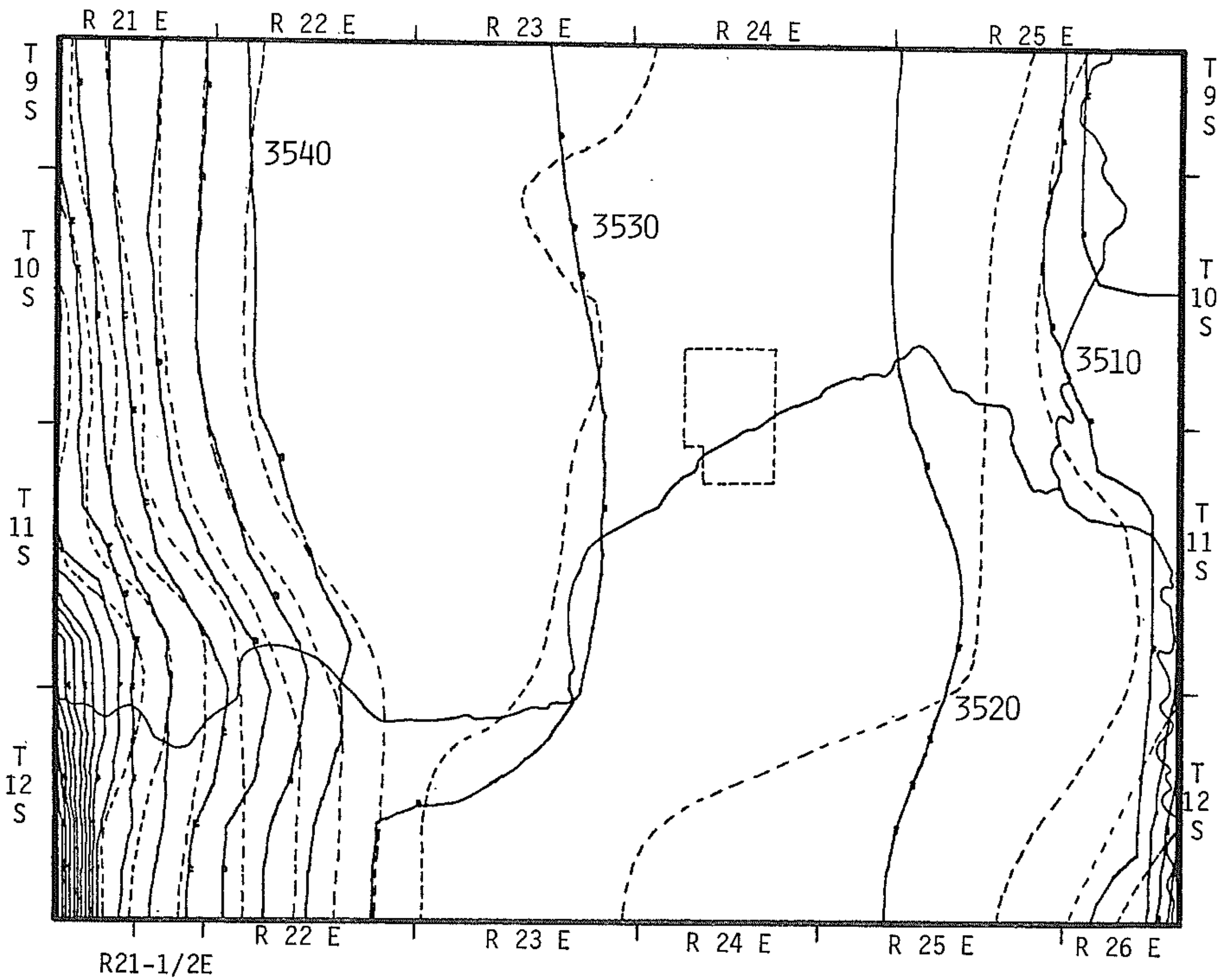

Figure 20: Hydraulic head distribution, January, 1968. Simulated with parameters from the final verification (solid lines). Dashed: Observed head distribution

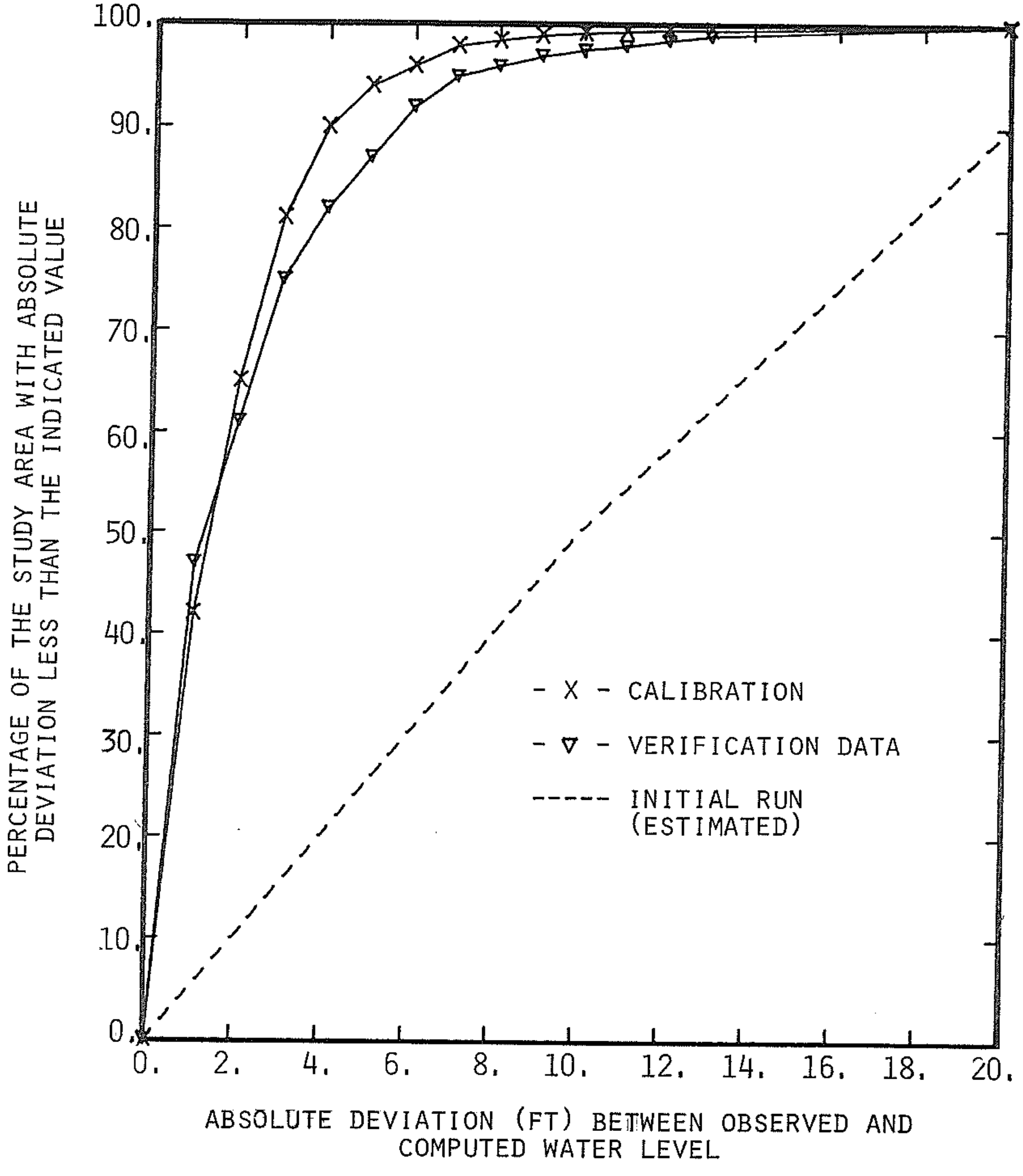

FIGURE 21: GOODNESS OF FIT - VERIFICATION DATA
JANUARY, 1968

The differences between the two sets of parameters are not significant except in the southern and southeastern parts of the model where the differences are due largely to poor head data and the inability to accurately model the southeast boundary.

PRESENTATION OF RESULTS

During the calibration, parameters were assumed known one by one, based on published values, until only two parameters, transmissivity and recharge, were actually determined by calibration.

Transmissivity and recharge were adjusted during the one year calibration, but also during the 8 year verification. Ideally, the calibrated parameters should not have needed to be changed, but then the real world is never ideal.

Transmissivity

Transmissivity maps for both the calibration and verification are shown in Figures 22 and 23, respectively. Only 2 areas were changed during the verification runs. The transmissivity was changed from $1.8x10^6$ to $1.4x10^6$ gpd/ft in the area of Townships 9 and 10, Ranges 23 and 24. The other area changed was Township 12, Ranges 23,24, and 25. The verification derived transmissivities are more consistent with previous studies and more consistent with other values in the basin than the calibrated values.

The change in the northern area is probably unnecessary because the resultant improvement in match between the computed and observed head in 1975 is small.

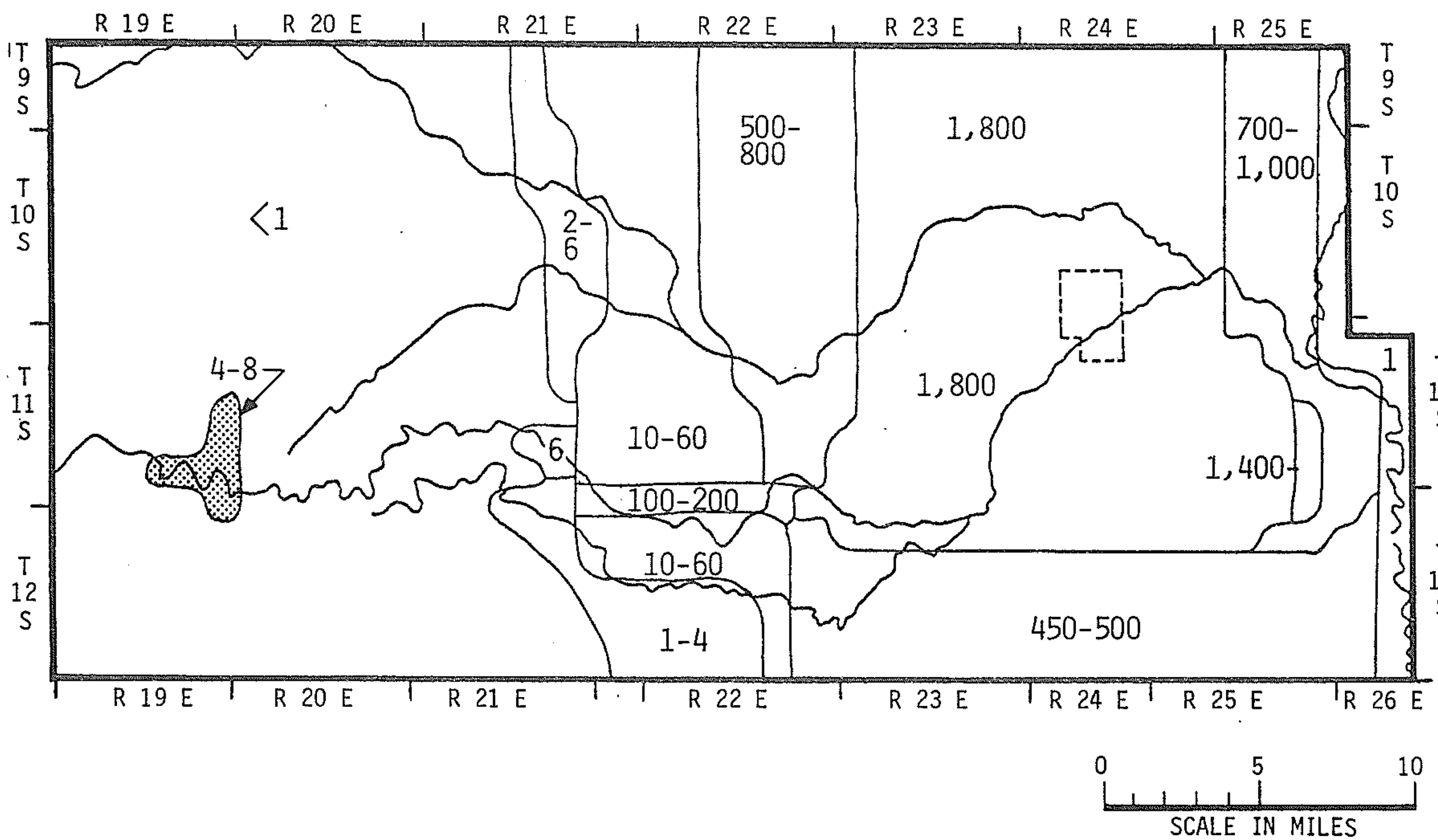

FIGURE 22: TRANSMISSIVITY DISTRIBUTION FROM CALIBRATION (1000 GPD/FT)

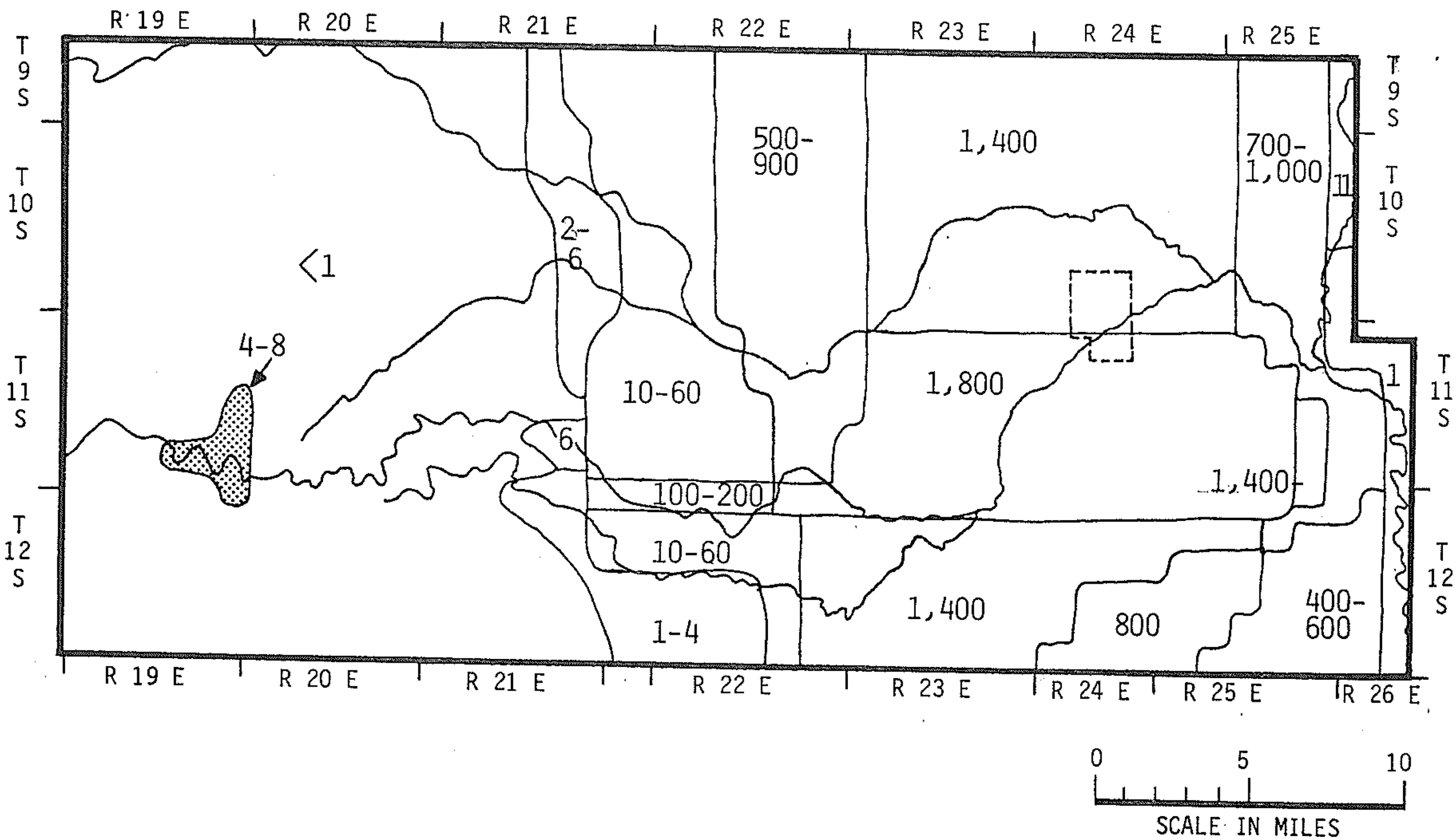

FIGURE 23: TRANSMISSIVITY DISTRIBUTION FROM VERIFICATION (1000 GPD/FT)

In the southern portion, the changes are larger and are likely the result of errors in the 1968 (Figure 6) head map. On that map, the 3520 foot contour swings to the west in Township 12 because of a single data point. Corresponding points in the area are either of questionable accuracy or simply missing. The data used for the 1975 head map are also poor in the same area, and it is difficult to say which map is more correct. The 1975 map, however, contains more data points and the general trend of contour lines in the area in question is consistent with other published maps of hydraulic head of the same area (Saleem and Jacob, 1971). Therefore, the transmissivity distribution based on the verification runs (Figure 23) is assumed for this report to be the correct one. The general trend of the transmissivity is an increase from west to east, with a decrease southeast of the Y-O structural zone.

Recharge

The two calculated distributions of recharge are given in Figures 24 and 25. The two maps differ in the same regions as the transmissivity and for the same reasons. Again, the recharge distribution obtained from the 8 year verification runs is assumed more nearly correct.

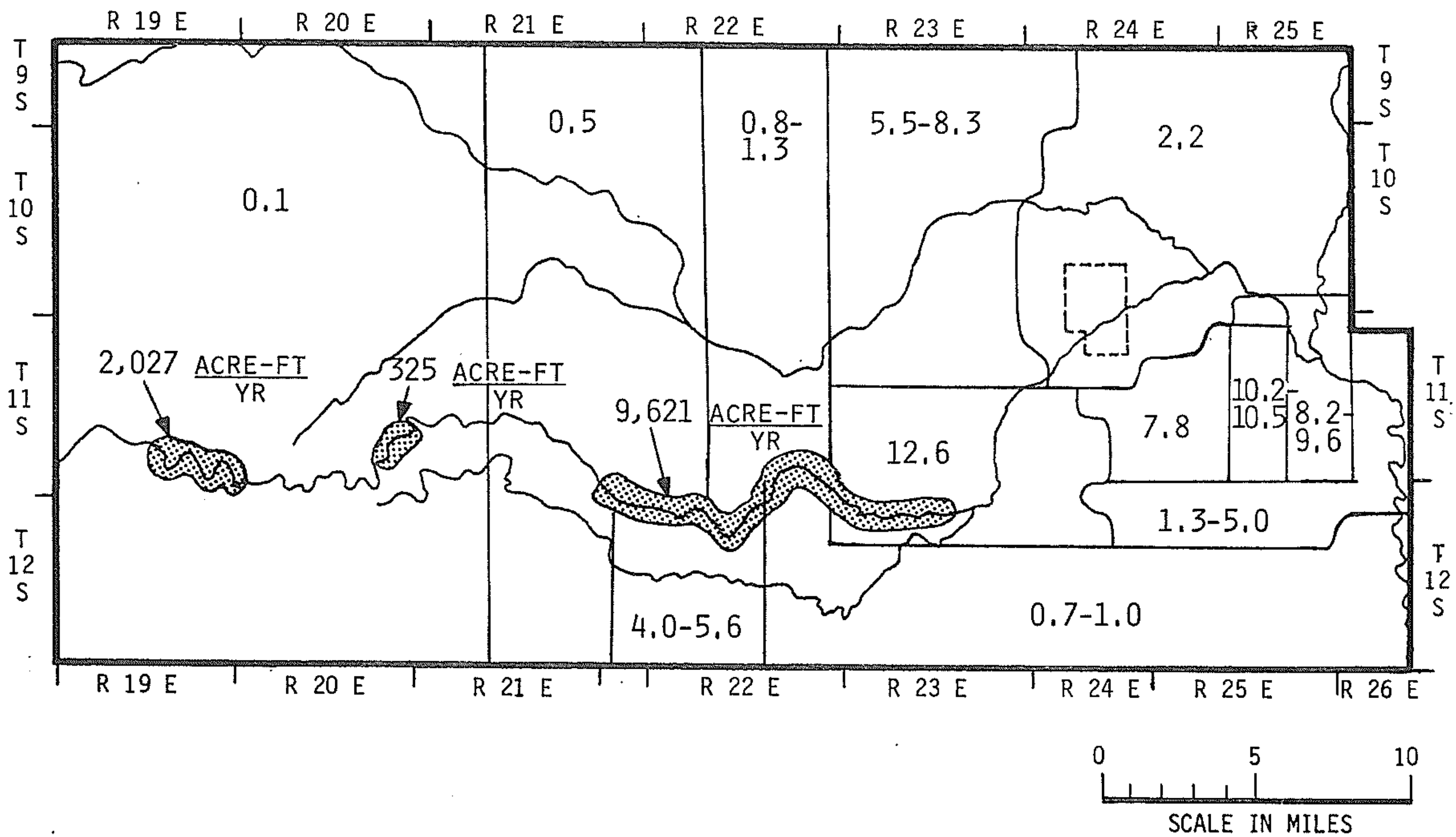

FIGURE 24: RECHARGE DISTRIBUTION FROM CALIBRATION (INCHES/YR)

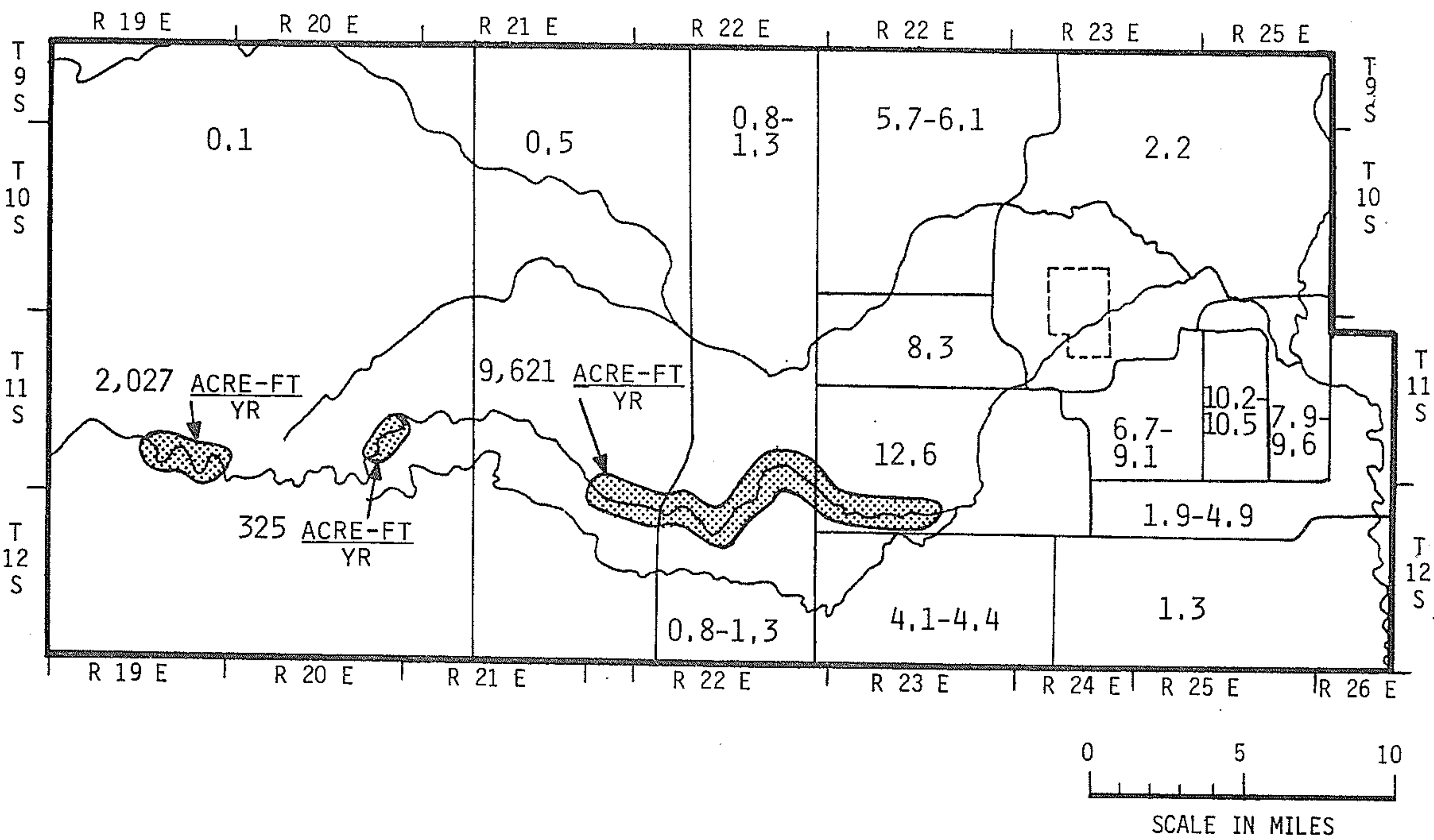

FIGURE 25: RECHARGE DISTRIBUTION FROM VERIFICATION (INCHES/YR)

The amount of recharge, as calculated by the model, varies over the basin. In the western region, where the flow occurs largely in the Yeso, the calculated recharge was about 0.1 inches per year. This is the amount that actually recharges the deep, or regional, water table as opposed to perched systems. Undoubtedly more water actually infiltrates into the overlying San Andres because of the presence of many springs which issue from perched systems along the Yeso-Glorieta (San Andres) contact (Davis and others, 1979). Also, some water may flow eastward along the contact until it reaches the saturated portion of the carbonate aquifer.

Progressing eastward, a zone of about 0.5 inches per year occurs at about the location where the water table lies in the Glorieta Sandstone (Figure 3). The Glorieta, being more permeable than the Yeso allows more water to infiltrate and may in fact absorb water that is flowing eastward along the Glorieta-San Andres contact.

East of the line where the water table intersects the San Andres (Figs. 3, 25), also called the western edge of the Principal Intake Area (PIA), the amount of recharge increases to between 0.8 and 1.3 inches per year. The increase may be due to a lessening of the land surface gradient which allows more water to infiltrate. East of the above region and west of the line where the San Andres becomes confined, something unusual occurs. The amount of recharge jumps from the range

5.7 to 8.3 inches per year above Township 11 to almost 13 inches per year in the region east of the Hondo Reservoir.

In the confined zone "recharge" was needed over and above leakage. Large amounts of this "recharge" was added to Township 11, Ranges 24 and 25, which is an area of heavy pumping. Possibly the value of transmissivity could have been increased to a point where recharge in the PIA would reach the pumping centers. However, that would have been inconsistent with pumping test values obtained by Hantush (1957,1961) which are an overestimate of the actual transmissivity (Neuman and Witherspoon, 1969b). Therefore, the "recharge" is probably a real phenomenon and not a failure to calibrate the model properly. The source of the "recharge" in the confined zone is obviously not precipitation and is probably not leakage from the overlying aquitard because that is calculated separately in the model. In fact, when the hydraulic conductivity of the aquitard was increased to allow more water in during the pumping season, the net upward leakage remained about the same while artesian water levels fell slightly.

Cumulative Mass Balance

The cumulative mass balances for the calibration and verification simulations are presented in Table 14. Some of the terms require an explanation. The recharge includes the

Table 14

Mass Balance - Calibration January 1967-January 1968

(units are acre-feet)

Sources		Discharges	
Recharge	123,644	Constant head	45
Constant head	1,123	Pumpage	101,166
Leakage	33,919	Leakage	48,280
Total	158,686	Total	149,491

Net gain in storage = 9,195 acre-feet
Net leakage (upward) = 14,361 acre-feet

Mass Balance - Verification January 1967 - January 1975

Sources	8 years	per year
Recharge	995,226	124,403
Constant head	8,937	1,117
Leakage	228,252	36,032
Total	1,292,415	161,552

Discharges	8 year	per year
Constant head	16,099	2,012
Pumpage	846,931	105,866
Leakage	385,800	48,225
Total	1,248,830	156,103

Net gain in storage = 43,585 acre-feet 5,448 per year
Net leakage (upward) = 97,548 acre-feet 12,194 per year

amount distributed over the area, which includes precipitation and upward leakage from the Yeso and Glorieta, and the amount contributed by the Rio Hondo. The constant head source is largely flow across the western boundary and represents the amount of water entering the carbonate aquifer by flow along the regional water table. The constant head discharge is the amount of water flowing east of the Pecos River. The leakage is the amount of water entering and leaving the confined portion of the carbonate aquifer through the overlying aquitard. The pumpage is self explanatory. The fluxes obtained appear to be reasonable when compared to similar estimates by Fiedler and Nye (1933), Hantush (1957), and Saleem and Jacob (1971).

The pumpage and recharge from the Rio Hondo were varied on a yearly basis in the verification simulation. The water levels, in the confined zone near Roswell, generally declined during the verification period, so the net gain in storage declined, but the leakage remained about the same. The constant head discharge increased dramatically because the eastward flow of water in the Bitter Lakes area increased. One will note that in Table 14 there is a net gain in storage over the model area. This appears inconsistent because the head in the confined zone is actually declining. Recall that the western region showed a general increase in water level. An excess recharge of only 1/6 inch over the 600 square miles of unconfined aquifer will produce the calculated gain in storage. In terms of the 8 to 13 inches of "recharge" in the vicinity of Roswell, the net gain in storage is not a significant problem. This also serves to indicate that the model results are not the last word, but merely a first attempt at a solution.

Hydrograph

Figure 26 is a hydrograph of the water level in the Berrendo-Smith Recorder Well (10.24.21.212-Hudson, 1978) and the hydrograph of the corresponding node in the model. The trend of the peaks appears to be similar, but the drawdown at the close of the irrigation season is too small. Several reasons for the lack of fit can be proposed: (1) a disproportionate amount of recharge is added during the pumping period resulting in less drawdown, (2) the transmissivity is much too high, or (3) the storage coefficient is too high. The transmissivity and storage coefficient are known reasonably well from pumping test results in the area (Hantush, 1957) and would not be in error enough to cause the observed mismatch in the hydrographs. As will be shown later, by changing the proportion of recharge added during the pumping period and during the winter, the hydrographs can be matched. In any event, the general trend of water levels is consistent with known hydrographs in that

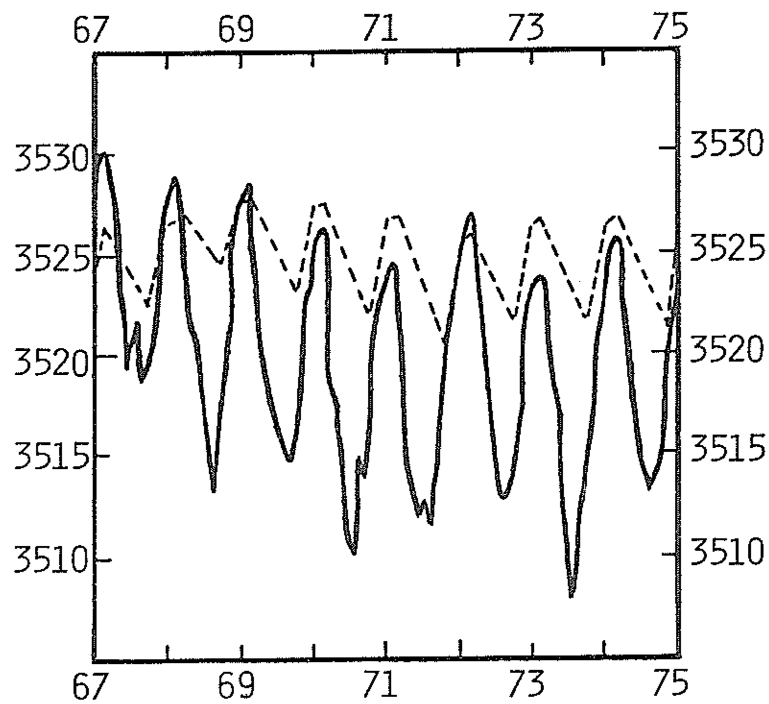

FIGURE 26: HYDROGRAPH OF THE BERRENDO-SMITH RECORDER WELL (SOLID LINE) AND THE CORRESPONDING NODE IN THE MODEL (DASHED LINE)

the water levels fall during the pumping season and rise during the non-pumping season.

The Orchard Park-A (12.25.23.344) well equipped with a chart recorder shows the same yearly fluctuations as does the Berrendo-Smith well (Hudson, 1978) except that the Orchard Well fluctuates over 100 feet per year. In the model, the same trend of yearly water level fluctuations increasing toward the southeast corner was simulated by decreasing transmissivity, storage coefficient, and leakage as was consistent with previous studies. The magnitude of fluctuation of the model results was about 30 feet per year. The reasons for the lack of fit are the same as for the Berrendo-Smith well plus an added reason resulting from the model itself. The boundary of the model, located very near to the Orchard Park-A node, has a strong influence on adjacent nodes. The boundary problem is the major reason the Orchard Park-A hydrograph is in error because, as before, the transmissivity and storage coefficient are not in error enough to generate the mismatch in hydrographs, and the reproportioning of the recharge has almost no effect in this area. The boundary problem will be discussed in the section on Errors.

DISCUSSION OF RESULTS IN TERMS OF RECHARGE TO THE CARBONATE AQUIFER

The recharge pattern derived from the calibration and verification of the model (Figure 25) is consistent with some theories, but raises some interesting questions. In general, 5 sources of recharge were identified or postulated in the model. They are: (1) reduction of net vertical leakage through the aquitard, (2) stream leakage, (3) flow from the Yeso and Glorieta, (4) precipitation, and (5) vertical leakage from the Yeso and Glorieta near Roswell.

In the study area the net annual loss of water from vertical leakage through the aquitard was found to be about 14,000 acre-feet for the period of January 1967 to January 1975. Fiedler and Nye (1933) and Hantush (1957) estimated the net loss of artesian water by leakage as 80,000 acre-feet per year for the entire basin, for 1927 and 1928, respectively. Hantush (1957) also obtained the same value for 1944. He determined the percentage of total leakage for various sections of the basin. The leakage in the area of this study comprises over 60% of the total leakage according to Hantush. In that case, the net leakage calculated from the model and extrapolated for the entire basin is 23,000 acre-feet per year or a saving of 57,000 acre-feet per year over the leakage estimated in 1944. The change in net leakage is due to the fact that for the period of 1944 to 1968 the average water level in the artesian aquifer declined 1.5 times as much as in the alluvial aquifer (Saleem and Jacob, 1971,p.56).

Stream leakage is an important source of recharge, as has been shown by Fiedler and Nye (1933), Bean (1949), and Duffy and others (1978). The stream leakage in the model was varied yearly in accordance with measured streamflow. Evaporation and irrigation losses from streamflow east of the Diamond-A gaging station were not accounted for in the model. On the other hand, leakage from intermittent streams was not modeled because of a lack of data, so ignoring the streamflow losses may, in part, balance the recharge from intermittent streams.

In the model, the amount of water flowing down the regional water table from the Yeso and Glorieta into the San Andres was found to be only 1,100 acre-feet per year. Duffy and others (1979) estimated that flow to be about 133,000 acre-feet per year for the entire recharge area or about 26,600 acre-feet for the 20 mile wide strip examined in the present study. The calibrated value is almost 25 times smaller than that of Duffy and others. Their value, however, can be altered using their own data as follows. In estimating transmissivity (T), Duffy and others started with the parameter S/T (S is storage coefficient). An average value of S/T was determined from 3 values; two of which were determined for Range 18 east and the third for Range 21 east. The water from the Yeso and Glorieta enters the San Andres approximately along Range 21 East. Therefore, why not use the S/T value for Range 21 East. Secondly, the value of storage coefficient must be estimated. Duffy and others used a value of 0.10 for S, but L. W. Gelhar, one of the co-authors, has calculated a value for S of 0.04 (personal communication) which is similar to the value of 0.03 used in the model. Using the S/T value from Range 21 East and S of 0.04, the flow becomes 5,500 acre-feet per year for the 20 mile strip. Therefore, the flow of water down the regional water table from the Yeso and Glorieta to the San Andres does not contribute more than about 5% of the total recharge to the carbonate aquifer. This is not to say that the Yeso and Glorieta contribute only 5% of the total carbonate aquifer recharge. In fact, there is evidence to suggest that most of the water in the carbonate aquifer may be derived from the Glorieta and Yeso as upward leakage occurring in the eastern half of the Principal Intake Area and in the confined zone.

The amount of yearly recharge in the Principal Intake Area jumps from a maximum of 1.3 inches west of Range 23 East to a minimum of 4 inches in Range 23 East (Figure 25). Much of Range 23 East has a recharge of greater than 6 inches and a maximum of 12.6 inches occurs in the lower 2/3 of Township

11 South. The area of greater than 8 inches of recharge will be named the "High Recharge Area". It should be noted that the High Recharge Area occurs only in the vicinity of the Rio Hondo and does not extend the full north-south length of the basin. In fact, the amount of recharge entering the system in the PIA in a 10-15 mile wide strip west of Roswell may be equal to the amount of recharge in the 60 miles of PIA south of the Rio Hondo. The question arises as to how 6 to 12 inches of recharge can occur in an area where the average annual precipitation is about 13 inches (Mourant, 1963, average of Roswell and Picacho).

Two possible explanations can be proposed: (1) precipitation infiltrates rapidly through solution features, cracks, and along stream channels, (2) water in the Glorieta and Yeso is leaking vertically upward into the carbonate aquifer. Evidence for either explanation is available and it appears that the answer is probably a combination of both with the second source predominant.

In support of the first explanation, Motts and Cushman's (1964) Northern Evaporite Area corresponds closely to the High Recharge Area and the area just to the north in the model. They describe the Northern Evaporite Area as having good to excellent recharge capacity, numerous sinkholes, and that the seepage loss per mile of stream channel is probably greater than in other parts of the intake area.

Precipitation will infiltrate rapidly and reach the water table sooner than in other parts of the intake area because the water table is closer to the land surface (Fiedler and Nye, 1933). Also, water from the west will enter the area in the stream channels and will be lost through leakage. Therefore, the potential for large amounts of infiltration exists along the eastern edge of the Principal Intake Area in the High Recharge Area.

Rabinowitz and Gross (1972) also described the above area as one of rapid recharge as opposed to slower recharge to the west. However, Gross and others (1976), have shown that Rabinowitz and Gross's interpretation of the data may be questionable and are the most recent proponents of the idea of upward leakage from the Glorieta and Yeso.

Fiedler and Nye (1933) were the first to propose a possible deep flow component from the Yeso and Glorieta. They said water in the Yeso and Glorieta had a greater artesian head than the carbonate aquifer and that water may be forced upward along joints and fractures, the amount was assumed small although it was never measured. Hantush (1957) and Saleem and Jacob (1971) also state that some recharge may be leakage from the Yeso and Glorieta. Bunte (1960) showed that the Glorieta is a major conduit of recharge north of the study area. Havenor (1968) presents some data which indicate the presence of upward vertical flow. The combined water level in the City of Roswell Test Well No. 2 (11.22.04) was 5 feet higher with both the San Andres and Glorieta producing than the water level of the San Andres alone. Many authors have said that the Glorieta and Yeso are not permeable enough to produce much water. However, over an area as large as the Roswell Basin, significant amounts of water can be produced from leakage through very tight formations. The overlying aquitard, for example, allows about 40,000 acre-feet per year to leak through in either direction. The strongest evidence for a deep flow comes from tritium data (Gross and others, 1976; Gross and Hoy, 1979).

Briefly, tritium is a naturally produced radioactive isotope of hydrogen that decays with a half-life of 12.3 years. Prior to 1953, only small amounts of tritium were present in precipitation and groundwater. With the onset of atmospheric testing of thermonuclear devices in 1953, the amount of tritium in precipitation increased nearly 3 orders of magnitude to a peak in 1963 (Rabinowitz and Gross, 1972). Since the Test Ban Treaty in 1963, atmospheric tritium activity has dropped, but is still above the pre-testing level. Consequently, the tritium activity in groundwater has increased as the high tritium precipitation recharges the groundwater.

If precipitation was the only source of recharge in the High Recharge Area, one would expect the tritium activity of the groundwater to be very nearly that of the precipitation. This, however, is not the case. Gross and others (1976) found the tritium activity of water from the Principal Intake Area to be well below expected values.

Based on the published tritium measurements of Gross and Hoy (1979), the average tritium activity in precipitation at Roswell for the period of 1972 to 1978 was 56.8 TU (1 TU = tritium unit = 1 tritium atom per 10^{18} hydrogen atoms) while the average tritium activity for 14 wells in the intake area for 1968 to 1978 was 13.2 TU. The average tritium activity in 6 wells in the High Recharge Area (Range 23) for 1972-1978 was 15.2 TU. Tritium activity of wells just east of the High Recharge Area was only 9.3 TU. The difference between tritium activity of groundwater and precipitation is significant when viewed in terms of the 12.3 year half-life of tritium because it would take about 24 years for a set volume of water of tritium activity 56 TU to reach 14 TU by natural decay of tritium. One might argue that the groundwater tritium activity is lower because of mixing of recharge with the ambient groundwater.

This is not the case in the carbonate aquifer. If we assume the aquifer is 200 feet thick with a porosity of 0.03 we have 6 feet of water per unit area in storage. With half a foot of recharge per year, the water in the aquifer should be replaced every 12 years. In the 12 years or so prior to the 1968 to 1978 period used above, the tritium activity of precipitation was much greater than at present (Gross and others, 1976, p. 54). Therefore, the ambient groundwater activity in 1968 should have been quite high, perhaps even higher than the precipitation in 1968. The low tritium activity of water in the Principal Intake Area, therefore, cannot be explained simply as the mixing of precipitation recharge and ambient groundwater.

A deep recharge component appears to be the best way to explain the low tritium levels in the Principal Intake Area, the High Recharge Area, and the needed "recharge" in the confined zone. One may ask how the Yeso and Glorieta can contribute only 1,100 acre-feet of water to the carbonate aquifer from the west and yet be considered the source of upward leakage in the eastern part of the study area.

Recent work by Wasiolek (1981) in the upper Penasco indicates that the Yeso transmits water through a series of highly permeable layers separated by low permeability layers. The high permeability layers appear to be quite extensive laterally. The 1,100 acre-feet is probably derived from only a few of these layers and much water is passing beneath the carbonate aquifer in other highly permeable layers. The water in the deeper layers is confined and under considerable pressure.

The deep recharge component theory is reasonable, but a mechanism is needed that allows the water in the deep, highly permeable layers of the Yeso to suddenly leak vertically upward in large volumes. The Six Mile Hill and YO structural zones could provide such a mechanism. Havenor (1968) has noted many possible effects of the structural zones on the flow of groundwater. The areas of high upward leakage correspond approximately to the inferred locations of the structural zones. These zones are areas of increased fracturing and would be ideal conduits for upward moving water. A rough calculation to determine the percentage of recharge from precipitation and from the deep flow can be performed using the tritium data.

If the measured average tritium of the groundwater in the PIA is assumed to be a mixture of deep water and precipitation we can use a simple mixing model.

The model (Figure 27) assumes that part of the groundwater is derived from precipitation of average tritium activity of 56.8 TU and the remaining groundwater is derived from a deep flow. The tritium activity of the deep flow can be estimated from the tritium activity of the PVACD

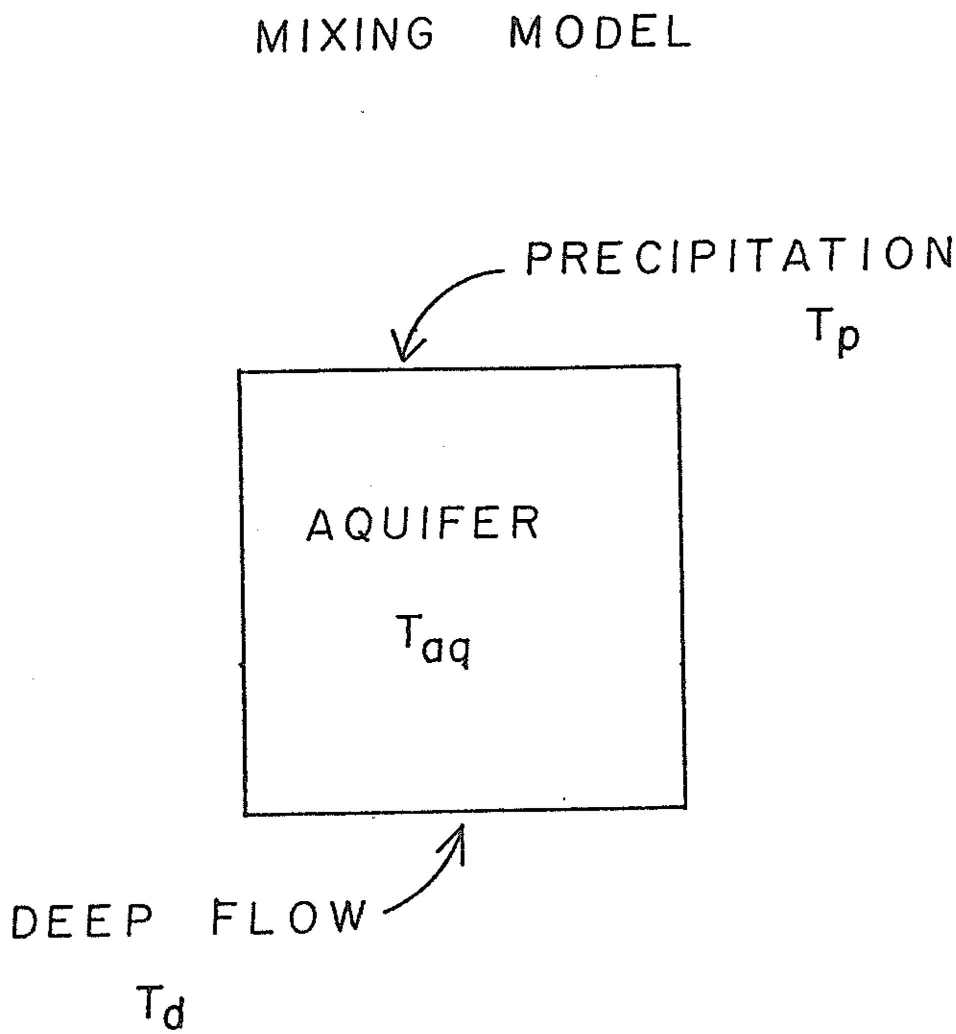

FIGURE 27: DIAGRAM OF THE TRITIUM MIXING MODEL

observation wells located at the western edge of the PIA. Four of the wells, numbers 1, 2, 3, and 4, are located in the study area and monitor water levels in the Glorieta and Yeso (Gross and others, 1976, p.71). For the period of 1974 to 1976, the average tritium activity of the four wells was 5.08 TU. The percentage of each recharge component is calculated as follows

$$T_{aq} = xT_p + (1-x)\ T_d \qquad (6)$$

where

T_{aq} = average tritium activity of the PIA

T_p = average tritium activity of precipitation

T_d = average tritium activity of deep flow

x = percentage of recharge from precipitation

$1-x$ = percentage of recharge from deep flow

T_{aq} = 13.2 TU, T_p = 56.8 TU, T_d = 5.1 TU.

From equation (6), the percentage of recharge from precipitation is only 16%. Deep flow accounts for over 80% of the recharge. If the calculation is done using the aquifer tritium level of 15.2 TU from the wells in the High Recharge Zone, the percentage of recharge from precipitation increases to only 20%.

Assuming that 20% is approximately correct, one would not expect normal fluctuations in precipitation to cause a significant fluctuation in water levels. Although a strong correlation appears to exist between precipitation events and water level rises, Hantush (1957) and Mourant (1963) have stated that the water level rise is caused mainly by the reduction in pumping associated with the precipitation. Mourant (1963,p.22) presents the hydrograph of a well located in the High Recharge Area (11.23.03.342). The water level in the well rose in June and July because the local rainfall caused a decrease in pumpage. In October, when pumping is small, a heavy rainfall produced no noticeable response in water level. Therefore, it appears that local fluctuations in precipitation have little effect on the water level in the carbonate aquifer. Duffy and others (1978,p20) present some evidence to suggest that long term trends in precipitation affect the water level in the PVACD observation wells located just west of the Principal Intake Area. That being the case, long term trends in precipitation will also affect the deep recharge component and will have a considerable effect on water levels in the carbonate aquifer. More work needs to be done, however, before any definite conclusions can be drawn.

The major argument against a substantial deep recharge component is that the Glorieta and especially the Yeso are said to contain saline water.

Little water quality data exist for the Yeso in the High Recharge and confined area. Some salt beds were logged in wells in the PIA (8.24.31.240, Bunte, 1960; 11.23.29.421, Hood and others, 1960). Water in the Yeso appears to be saline in the Roswell Block (Havenor, 1968) where the carbonate aquifer is also saline. No specific water quality data were found concerning the Yeso in the Orchard Park Block (Havenor 1968). West of the Principal Intake Area, wells in the Yeso encounter fresh water (Hood and others, 1960; Mourant, 1963; Wasiolek, 1981). The few Yeso wells in the western part of the PIA (Mourant, 1963) have good quality water according to local standards. In general, the water is suitable for domestic use west of the Principal Intake Area and for an unknown distance into the PIA. The water is almost always of poorer quality than in the carbonate aquifer. Again note that the Yeso is a stratified aquifer. Different layers may have different quality water. As no deep wells in the Yeso exist in the PIA, blanket statements concerning the quality of the Yeso water are highly questionable. The quality of water derived from the Glorieta is generally much better than that of the Yeso or San Andres in and west of the Recharge Area (Table 15). The water quality of the carbonate aquifer decreases from west to east with a marked change near the Six Mile Hill structural zone (Mourant, 1963) and the postulated High Recharge Area. One would expect such a change if poorer quality Yeso water were mixed with the ambient water in the carbonate aquifer.

Table 15

Water Quality--Intake Area

Range of Values (from Mourant, 1963, Plate 4)

(parts per million)

	Yeso	Glorieta	San Andres
Hardness	404-2520	287-830	340-1050
Sulfate	541-2130	128-617	119- 760
Chloride	22- 155	14- 98	14- 340

Table 16

Water Quality--Average Values

	Yeso	Glorieta	San Andres
Hardness	1030	459	576
Chloride	59	33	44
Sulfate	820	257	351

Hardness is given as $CaCO_3$

Wells tapping the Glorieta near the Pecos River, or in the area of saline water encroachment in the carbonate aquifer, encounter saline water (10.26.30.200, 11.24.04.114d, Hood and others, 1960; 9.25.17.110, Bunte, 1960; 10.23.19.440, 10.24.34.444, Havenor, 1968). Water of good quality in the Glorieta was encountered in and near the High Recharge Area (11.22.04, Havenor, 1968; 11.23.08.200, Fiedler and Nye, 1933). Havenor also found that the Glorieta yielded moderately high quantities of water of chloride content less than 100 ppm on the western edge of the Roswell Block. Until more is known about the groundwater quality of the Yeso and Glorieta in the Roswell region, the deep recharge source cannot be ignored.

As stated in the introduction, the computer model can be easily changed to incorporate new theories. In the calibration, the recharge was unevenly distributed throughout the year in the same manner as precipitation. As the model results were analyzed, the effect of precipitation was shown to be small, and deep flow became the major component of recharge. The deep flow is relatively unaffected by yearly variation in precipitation and should be essentially constant. Therefore, two more computer simulations were performed; one of 1 year and the other of 8 years corresponding to the calibration and verification simulations, respectively. In each, the recharge was held constant throughout the year. In addition, during the 8 year simulations, only the stream recharge and pumpage were changed from year to year while the areal recharge remained constant. The computed hydraulic head distribution for both simulations is presented in Figures 28 and 29. The goodness of fit for one year with a constant recharge is given in Figure 30.

The results are very similar to the results obtained with recharge variable throughout the year. In general, the water levels in the High Recharge Area increased slightly and the net upward leakage through the aquitard was decreased by almost 4,000 acre-feet per year. The decrease in net upward leakage is the result of increased drawdown in the confined zone during the irrigation season caused by the redistribution of recharge.

The hydrograph of the Berrendo-Smith recorder well (10.24.21.212- Hudson, 1978) is compared to the hydrograph of the corresponding node in the model in Figure 31. The predicted hydrograph using constant recharge is a much closer approximation to the observed hydrograph than the predicted one with variable recharge. This also affirms the theory that a substantial portion of the recharge to the carbonate aquifer is derived from the underlying Glorieta and Yeso.

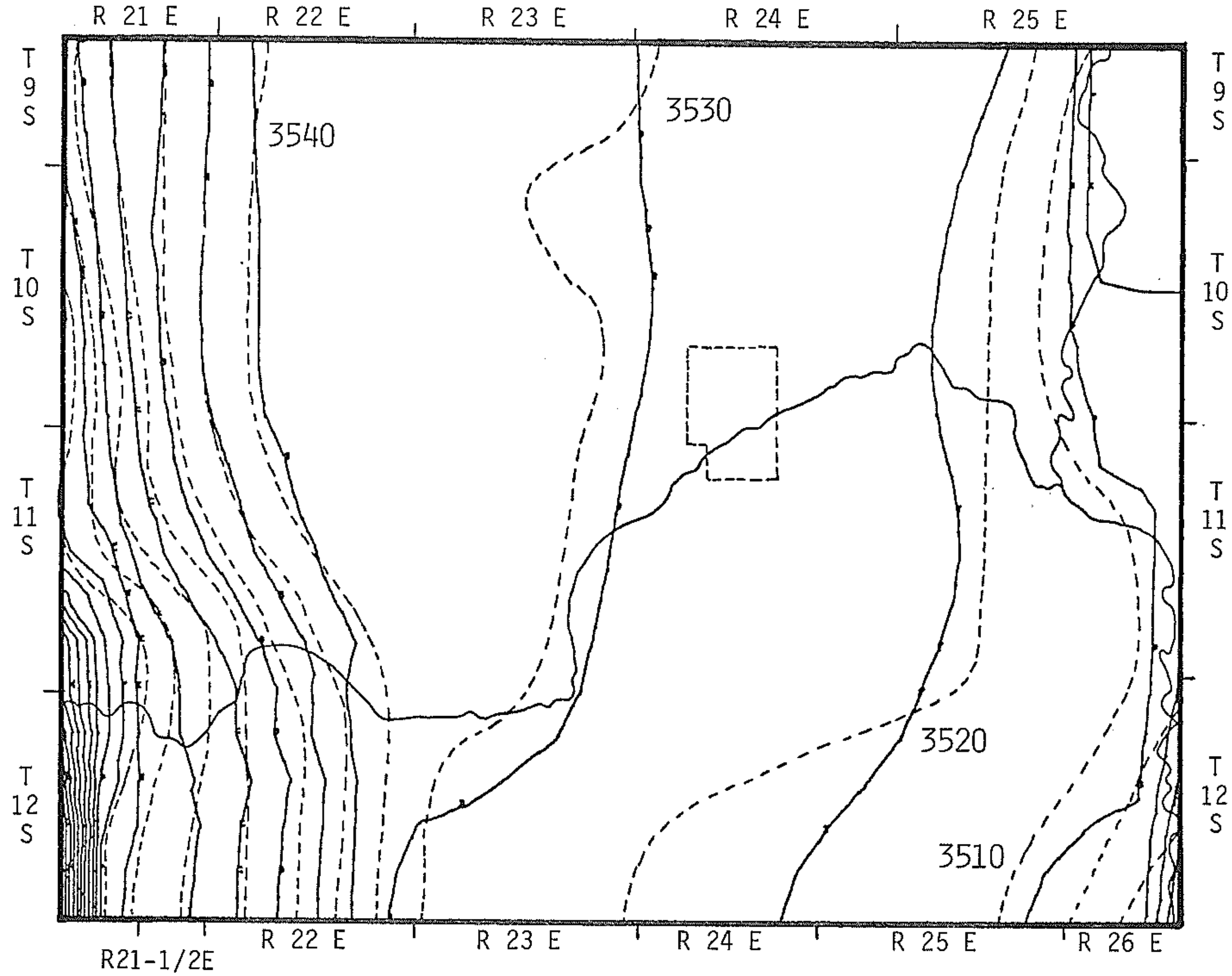

FIGURE 28: HYDRAULIC HEAD, JANUARY 1968
SIMULATED WITH CONSTANT RECHARGE (SOLID LINES)
DASHED: OBSERVED HEAD DISTRIBUTION

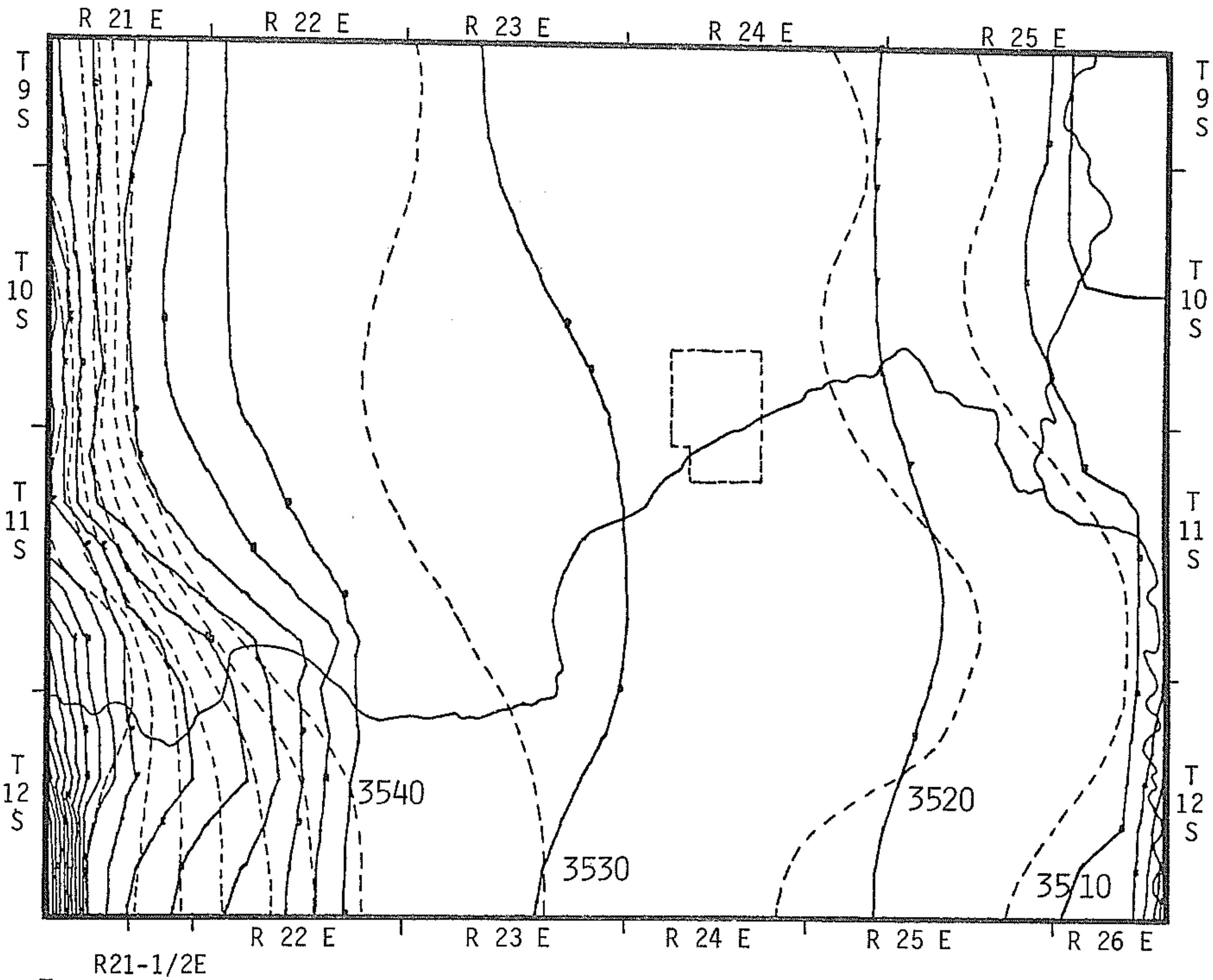

Figure 29: Hydraulic head, January 1975.
Simulated with Constant Recharge (Solid Lines)
Dashed: Observed head distribution

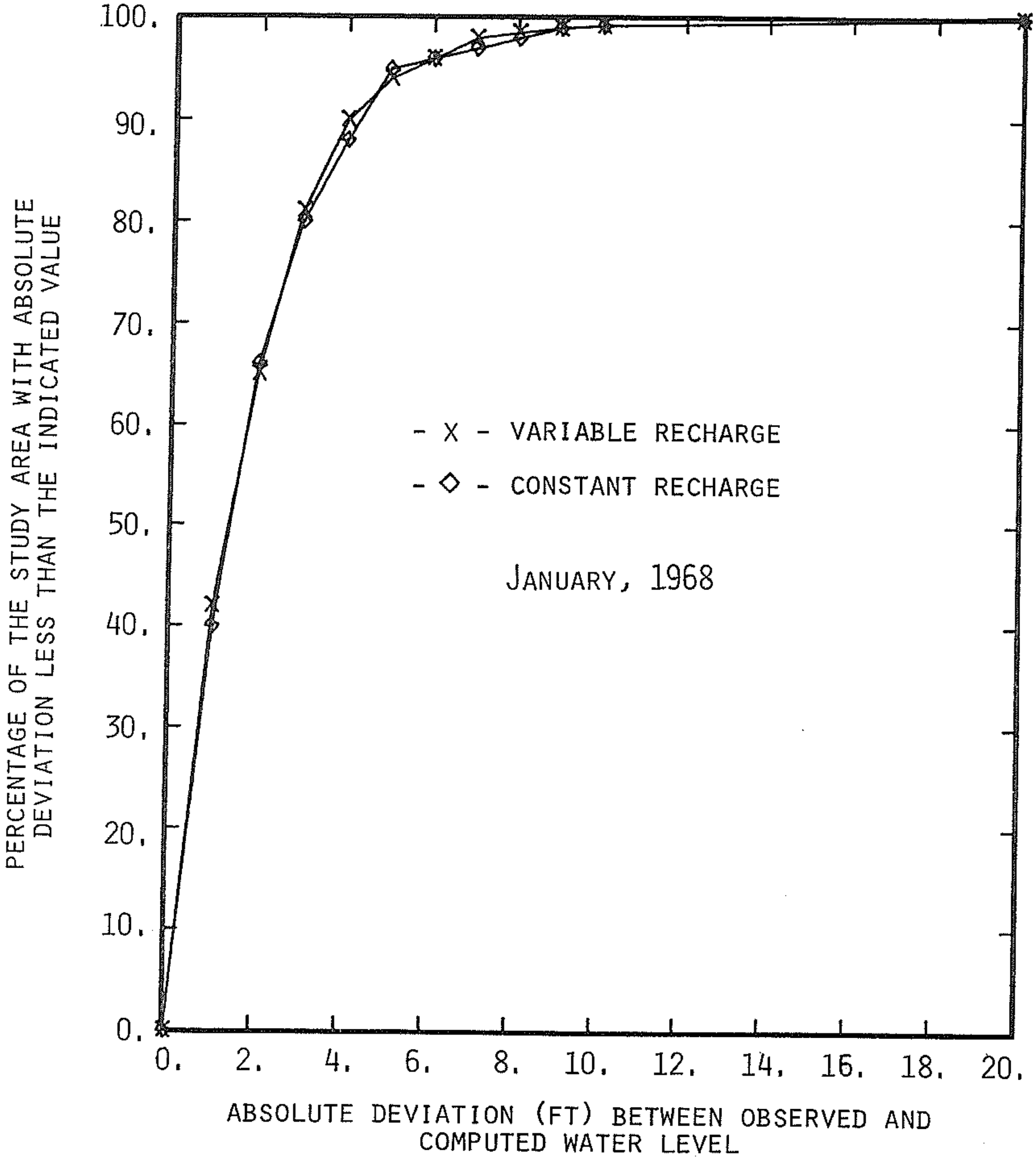

FIGURE 30: GOODNESS OF FIT - CALIBRATION FOR CONSTANT AND NON-CONSTANT RECHARGE

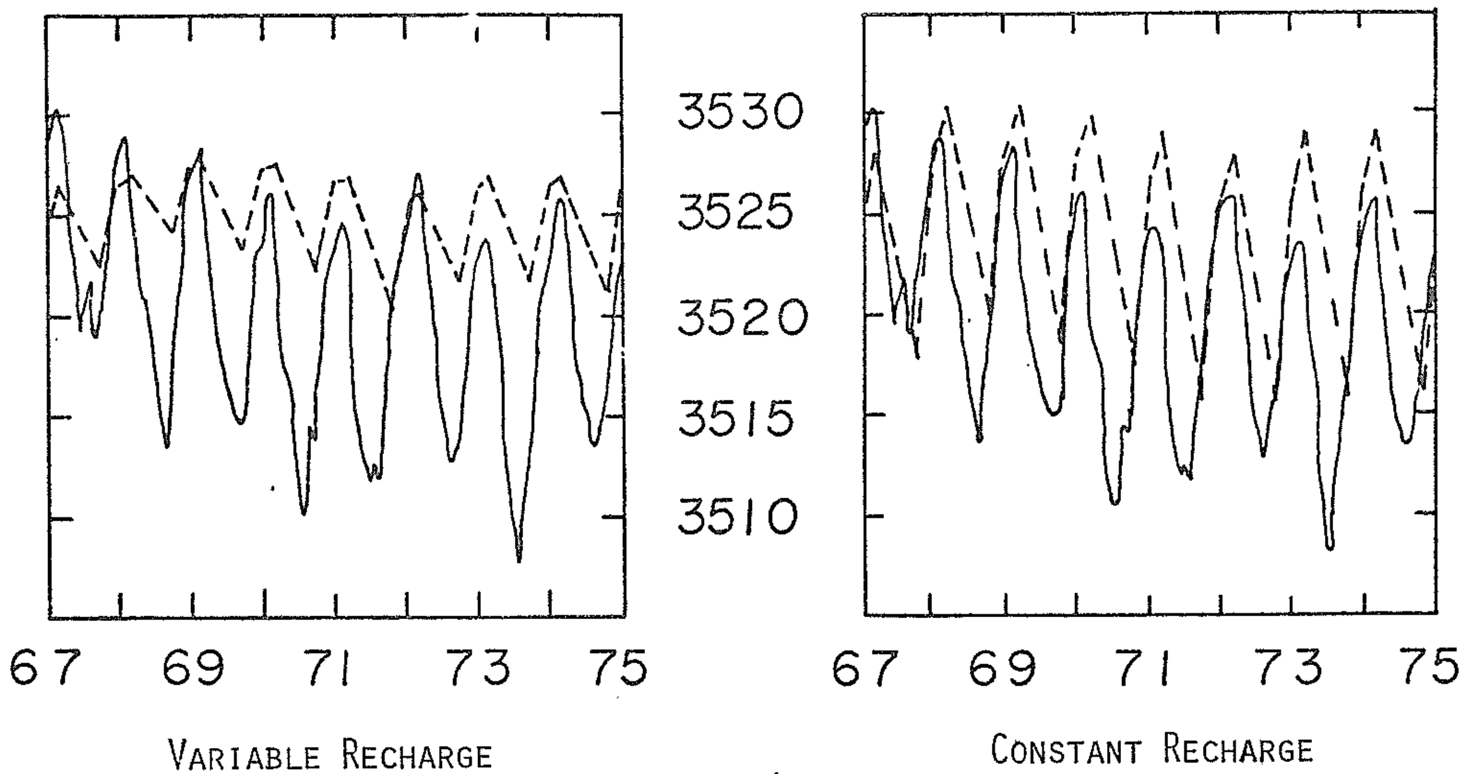

FIGURE 31: HYDROGRAPH OF THE BERRENDO-SMITH RECORDER WELL (SOLID LINES) AND THE CORRESPONDING MODEL NODE (DASHED LINES) FOR VARIABLE AND CONSTANT RECHARGE

SOURCES OF ERROR

The computed head distribution did not match the observed head distribution exactly because of errors. The source of the errors can be either the model or the data. Some of the errors introduced by the model have been discussed earlier. Those are the errors due to modeling the fractured carbonate system as a porous medium, modeling the entire area as a confined flow system, the assumption of isotropy, and the assumption of a constant water level in the alluvial aquifer. The error introduced from the first 3 sources cannot readily be determined, but the error is assumed to be small. Based on the experience gained from using the model, the assumption of a constant head alluvial aquifer probably does not amount to more than a 1,000 acre-feet per year error in total leakage. That translates to less than a one foot error in predicted head. The largest model-induced error is due to the boundaries.

The constant head west boundary does not induce a significant error. The no-flow north and south boundaries probably do not cause much error except in the area east of Range 23 East. Most authors believe that some component of flow enters the Roswell area from the northwest. The magnitude of that flow could not be determined, so a known flux boundary could not be established. The error introduced might, at most, be the 2 inches of "recharge" added to the area north of Roswell. If a northern flow existed, it would take the place of some of the "recharge". This error probably does not affect the Roswell or East Grand Plains area.

The southern no-flow boundary caused some error in the southeast corner of the model. The groundwater has a southeasterly flow component near the southeast corner as a result of the heavy pumping in the vicinity of Dexter. The inability of the model to predict the magnitude of yearly water level fluctuations in the vicinity of the Orchard Park A recorder well (Hudson, 1978) is due to the boundary problem in the corner. Of all the results, the ones obtained for the southeast corner are the least reliable. Again, the error probably did not affect the Roswell and East Grand Plains area to any great degree.

The eastern constant head boundary allowed a small amount of water to flow east out of the basin. The only significant errors were probably introduced in the southeast corner where large drawdowns induced flow from the east. The constant head also prevented the model from predicting the magnitude of water level fluctuation. Until the eastern boundary is known more precisely, the present results are assumed reasonable.

Finally, the data used will also induce errors. McWhorter and Sunada (1977), and Volker and Guvanasen (1975) found that errors in the initial head data produce the largest errors in calibration. Errors in the final observed head also produce errors in calibration because the computed head is matched to observed heads. If the observed head is in error, so is the computed head.

In parts of the model area, head measurements were scarce or non-existent. The western half of the Principal Intake Area has almost no head data, and the head contours drawn in that area are largely guesses. The other area of poor head data is located just east of the Pecos River. The head there was extrapolated from data west of the Pecos River.

In summary, the errors are least in the High Recharge Area and in the Roswell-East Grand Plains Area and are greatest along the southeast boundary.

SENSITIVITY ANALYSIS

A sensitivity analysis was performed to determine the response of the model to changes in the parameters. Two reasons for such an analysis are: (1) during calibration the most sensitive parameters must be determined to greater accuracy than less sensitive parameters, and (2) future data collection should concentrate on the more sensitive parameters.

Procedure

The general procedure was to alter a parameter at a selected node and then to calculate the difference in computed hydraulic head between the simulations using the altered and unaltered parameters, respectively. Each parameter was altered individually in order to examine its relative effects. The parameters were altered plus and minus 50%, except the length parameters such as hydraulic head, or the thickness of the confining bed, which were altered plus and minus 25 feet.

The total difference in computed head was obtained by calculating the sum of the absolute values of the difference in computed hydraulic head at each node in the model for each parameter. In this manner, the parameters could be ranked in terms of the total deviation in head over the model area.

The above procedure was performed at 3 nodes in the model representing 3 different flow regimes (Figure 32). Node A represents an area of confined flow where leakage between the limestone aquifer and the overlying alluvial aquifer is important. This is also an area of low deep flow. Deep flow refers to the "recharge" derived from upward leakage of water in the Glorieta and Yeso. Node B represents the unconfined flow area where substantial deep flow occurs. Node C is representative of the confined flow region south of the Y-O structural zone where confined flow, large deep flow, and small leakage occur. Node C also is a pumping node.

The sensitivity analysis was performed at each of the 28 time steps in the calibration period; January, 1967, to January, 1968. The time trend of the difference in computed head is plotted to determine whether the difference increases, decreases, or remains constant in time. The final time step yields the value of the difference at the end of the calibration period.

The following parameters were altered plus and minus 50%:

1) transmissivity (T)
2) storage coefficient (S)
3) areal recharge (QRE)

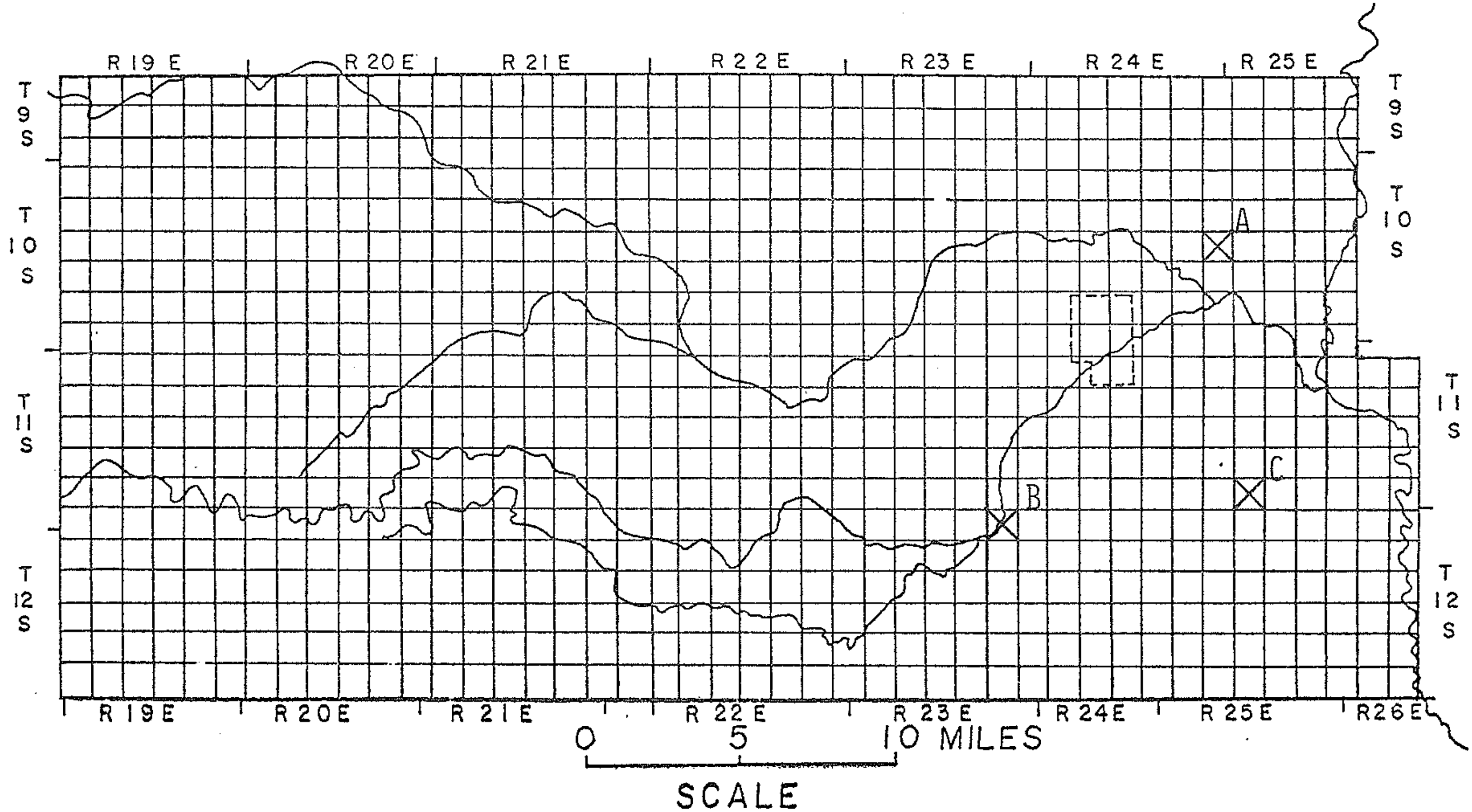

FIGURE 32: LOCATION OF THE NODES USED IN THE SENSITIVITY ANALYSIS

4) vertical hydraulic conductivity of the aquitard (KV)

5) pumping rate (PUMP).

The remaining parameters were altered plus and minus 25 feet:

1) initial hydraulic head (STRT)

2) head in the overlying alluvial aquifer (HALL)

3) thickness of the aquitard (M).

The parameters for each of the three nodes are given in Table 17.

A direct comparison between the first group of parameters, altered 50%, and the second group, altered 25 feet, is not possible. However, to compare the effects of their alteration on the head distribution is justified because the bounds of alteration were chosen to represent a likely error range.

Results

Table 18 contains the total difference in computed head due to the change in each parameter, at each of the 3 nodes, at time step 28, the end of the calibration period. For each

Table 17

Parameter Changes used in Sensitivity Analysis

Parameter	change	Node A (7,39)	Node B (16,32)	Node C (15,40)	units
	-25 ft	3495	3496.4	3495.7	
STRT	0	3520	3521.4	3520.4	feet
	+25 ft	3545	3546.4	3545.7	
	-50 %	.08	150	.042	
S	0	.16	300	.084	
(1e-04)	+50 %	.24	450	.126	
	-50 %	900	900	900	
T	0	1800	1800	1800	ft^2/sec
(1.55e-03)	+50 %	2700	2700	2700	
	-50 %	15	-	2.5	
KV	0	30	-	5	ft/sec
(1.55e-08)	+50 %	45	-	7.5	
	-25 ft	3477	-	3480	
HALL	0	3502	-	3505	ft
	+25 ft	3527	-	3530	
	-25 ft	295	-	475	
M	0	320	-	500	ft
	+25 ft	345	-	525	
	-50 %	49	279	232.5	
QRE	0	98	558	465	ft/sec
(6.03e-11)	+50 %	147	837	697.5	
	-50 %	-	-	-1.735	
PUMP	0	-	-	-3.47	ft^3/sec
	+50%	-	-	-5.205	

Numbers in parentheses beneath the parameter symbols are multiplication factors used as input to the model to convert the units to feet and seconds.
'e' represents the exponential

example 1e-04 = $1x10^{-4}$

Table 18

Results of the Sensitivity Analysis

Total difference in computed head (feet)
over the model area

Node	Parameter	Parameter Decreased	Parameter Increased
A	HALL *	37.37	37.36
	KV	11.53	10.66
	T	8.29	3.86
	STRT *	5.41	5.41
	QRE	3.03	3.03
	M *	1.67	1.48
	S	0.001	0.002
B	QRE	21.82	21.82
	T	10.70	5.19
	S	3.05	2.96
C	PUMP	37.57	37.36
	QRE	19.58	19.64
	T	6.75	3.27
	HALL *	5.42	5.41
	STRT *	5.33	5.33
	KV	2.01	1.60
	M *	0.09	0.08
	S	0.002	0.002

*-values were distance measures and were altered plus and minus 25 feet; non-starred values 50%

node, the parameters are ranked from the greatest to the least sensitive from top to bottom.

At all three nodes the following were observed:

1) storage coefficient was the least sensitive
2) transmissivity was more sensitive when decreased 50% than when increased 50%.

At nodes A and C in the confined zone, the thickness of the aquitard (M) was less sensitive than the other parameters.

The high sensitivity of the model at node A to the leakage parameters HALL and KV, and the low sensitivity to the areal recharge, which in this area is all deep flow, would indicate that the flow system northeast of Roswell is dominated by interaquifer leakage between the carbonate aquifer and the overlying alluvial aquifer.

An opposite effect is observed at node C. Areal recharge (deep flow) is more sensitive than the leakage parameters. Note that although pumpage was the most sensitive when altered 50%, it is a known value because most pumpage is metered. It is indeed fortunate that the irrigation and municipal wells are metered by the PVACD. At node B, areal recharge, which is about 80% deep flow, is the most sensitive parameter, followed by transmissivity.

Time trends in computed head were also observed. Head changes induced by changes in the storage coefficient fluctuated unpredictably about a mean value and appeared to remain approximately constant in time (Fig. 33). Changes of transmissivity, pumpage, or thickness of the aquitard increased the head difference through the initial nonpumping period and the following pumping period, but decreased it again during the final nonpumping period (Fig. 34). Changing the hydraulic head in the alluvial aquifer, a change of recharge, or of vertical hydraulic conductivity increased the head difference through all three periods (Fig. 35).

The effect of a 25 foot change in initial head decreased with time through all three periods. Although this would appear to imply that head values are not extremely critical, one must be cautious. If an error of 25 feet were present in the January, 1968, head data used for calibration, a change of greater than 50% in another parameter would be needed to compensate for the erroneous head value. Therefore, the head values are in reality the most critical data as reported by McWhorter and Sunada (1977), and Volker and Guvanasen (1975).

The more sensitive the model is to a given parameter, the more accurately we must determine its value in order to match observed and computed heads, and the more confidence we can place on that parameter value. In the model, every attempt was made to use published values for all parameters except recharge. If we accept the published values as

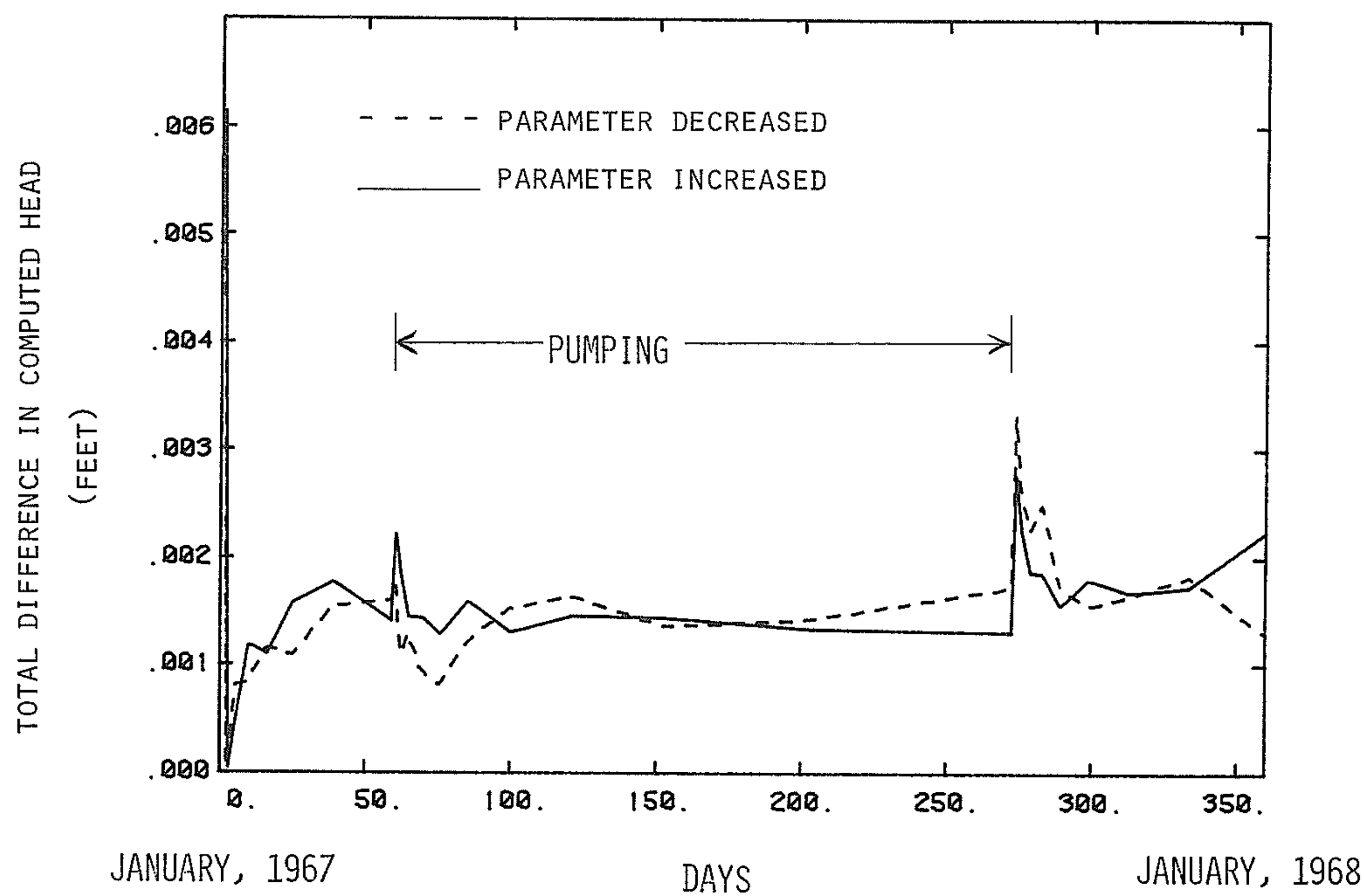

FIGURE 33: PLOT OF THE MODEL SENSITIVITY TO STORAGE COEFFICIENT AT NODE A

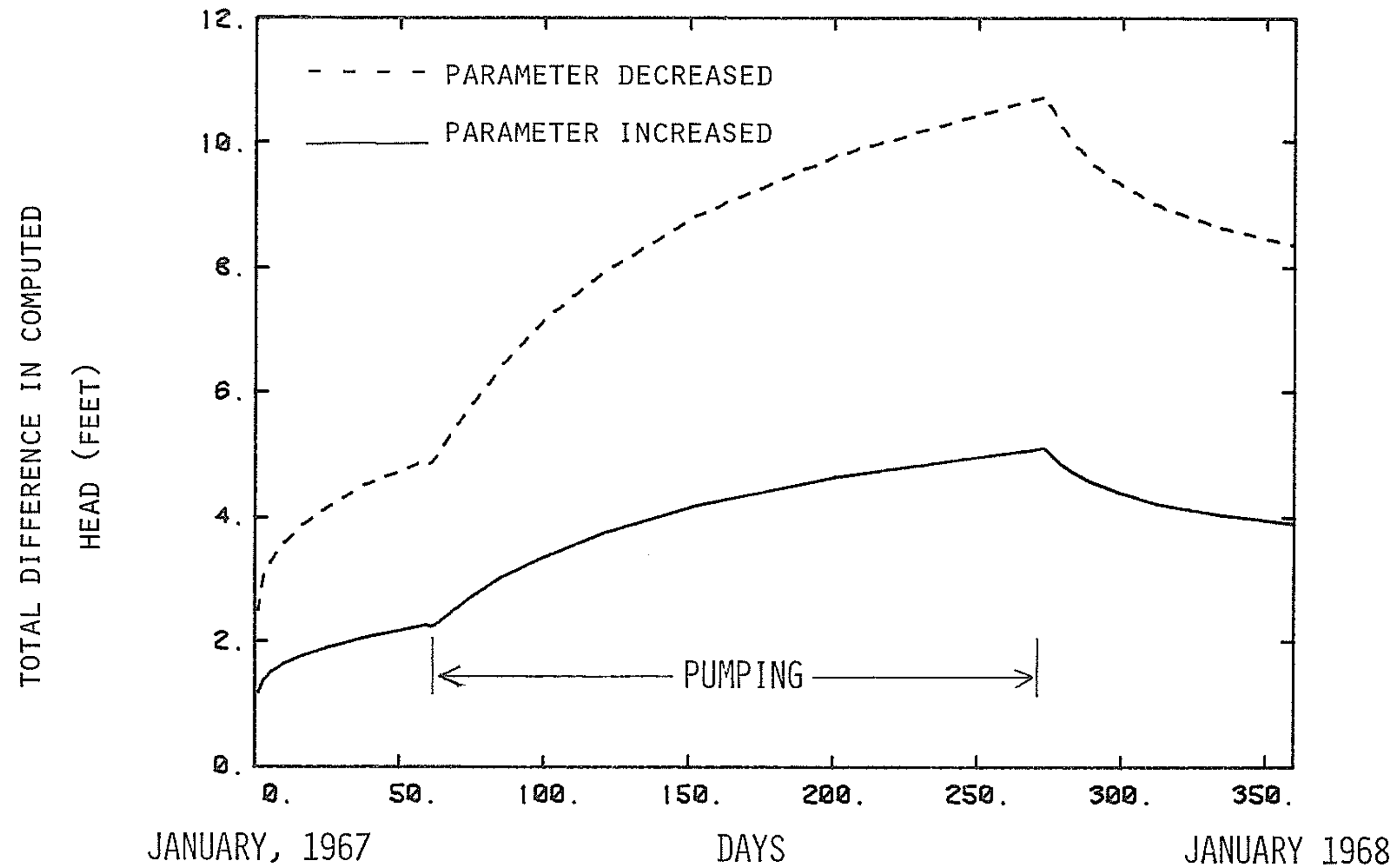

FIGURE 34: PLOT OF THE MODEL SENSITIVITY TO TRANSMISSIVITY AT NODE A. A SIMILAR TREND WAS OBSERVED FOR CHANGES IN PUMPAGE AND AQUITARD THICKNESS

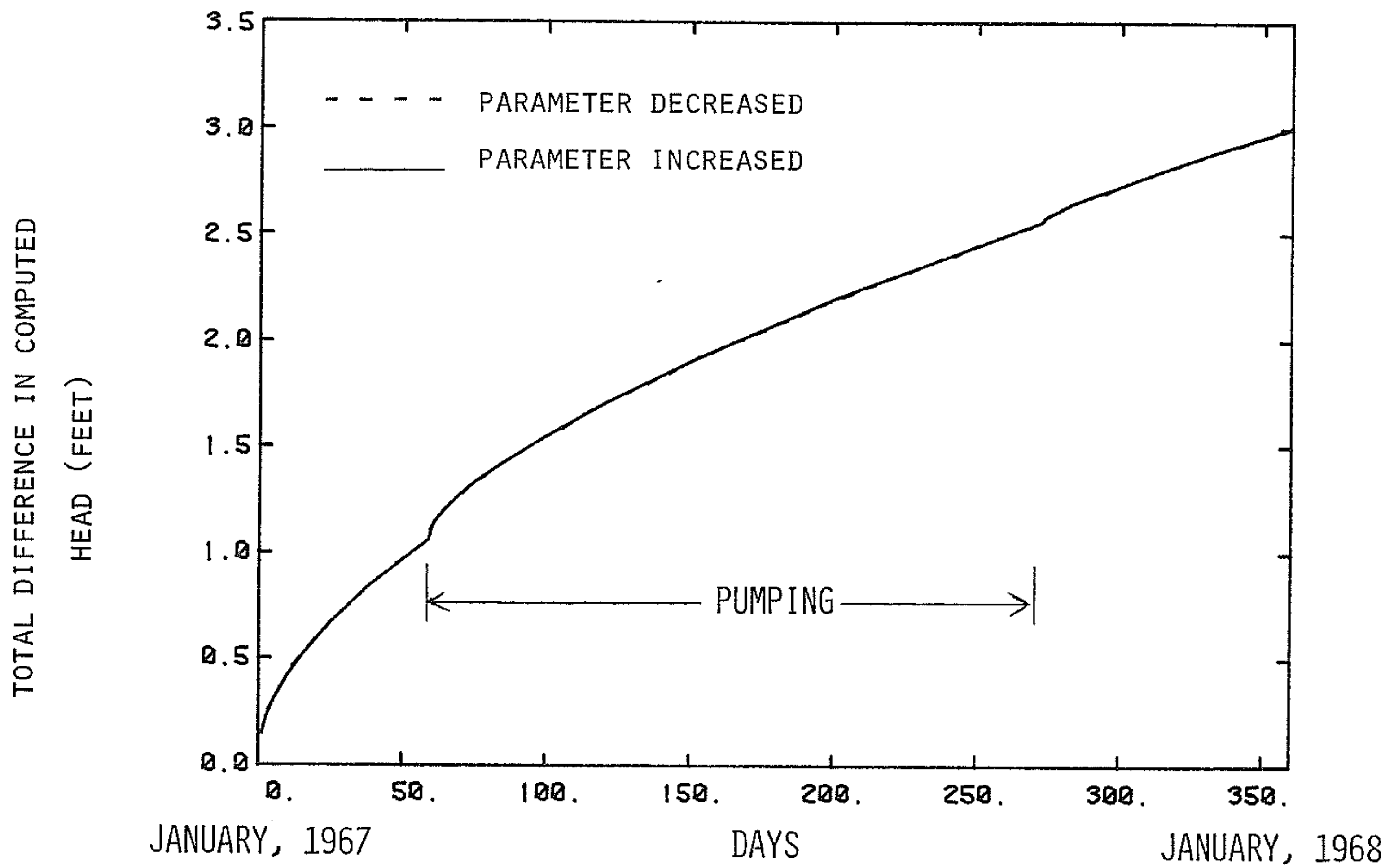

FIGURE 35: PLOT OF THE MODEL SENSITIVITY TO RECHARGE AT NODE A. A SIMILAR TREND WAS OBSERVED FOR CHANGES IN THE VERTICAL HYDRAULIC CONDUCTIVITY OF THE AQUITARD

reasonable, then the sensitivity analysis results lead one to conclude that the calibrated recharge estimates are reasonable. The only exception to this is the area northeast of Roswell (node A) where recharge was not as sensitive as other parameters. In that area, the calibrated recharge (deep flow) is much less certain than in other areas.

SUMMARY OF CONCLUSIONS

1) Most of the recharge to the carbonate aquifer occurs in Fiedler and Nye's Principal Intake Area.
2) Only 20% of the recharge to the carbonate aquifer occuring in the PIA comes from precipitation.
3) 80% of the recharge occurs as deep flow.
4) The percentage of precipitation that becomes recharge is about 7%.
5) The Yeso and Glorieta contribute much more water to the carbonate aquifer than previously reported.
6) Groundwater conditions along the east slope of the Sacramento Mountains have an impact on groundwater conditions in the Roswell Basin.

MODIFICATIONS TO THE COMPUTER PROGRAM

Some modifications to the program as listed in Trescott and others (1976) were needed in order to run the model on the DEC system - 2040 computer at New Mexico Tech. Other changes were made for one or more of the following reasons:

1) an extra data set was input containing the observed final head matrix;
2) a plot of the error between the computed and observed final head was generated on the line printer and a contouring routine was added;
3) recharge changed in every pumping period;
4) the necessary data for the head-contouring routine were generated in the program and stored in a separate file.

Details of these changes can be obtained by writing to Dr. G. W. Gross, Mew Mexico Institute of Mining and Technology, Socorro, NM 87801.

RECOMMENDATIONS FOR FUTURE STUDY

1) Determine the hydraulic conductivity of the aquitard. Ideally the method developed by Neuman and Witherspoon (1969a, 1969b, 1972) should be used. Temperature profiles could also be used (Stallman, 1963; Bredehoeft and Papadopoulos, 1965).
2) Model vertical sections of the basin to get a better understanding of the vertical flow components, especially the leakage.
3) Perform tracer tests in the basin to determine anisotropy.
4) Delineate the eastern groundwater boundary and its properties.
5) Model more of the basin to eliminate some of the boundary problems.
6) Determine the reasons why the Berrendo-Smith recorder fluctuates only 20 feet per year while the Orchard Park recorder fluctuates 120 feet per year.
7) Further investigate the High Recharge Area.
8) Determine the cause of the long term trends in the PVACD observation wells and the possible effect on the deep recharge component.
9) Improve the head data, especially west of Roswell.
10) Determine the bottom of the aquifer in the PIA using geophysical methods.

11) Correlate the Paul Spring record (Gross and others, 1979) to the PVACD observation well records.

12) Determine the aquifer properties in the PIA from pumping tests, slug tests, or the well network method (Stallman, 1956).

13) Use the method of kriging (Delhomme, 1978) to estimate the transmissivity distribution in the carbonate aquifer from the step drawdown estimates of T given by Saleem and Jacob (1971).

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APPENDIX A

Computer Input Data and Modified Computer Program

The data files used for the simulations are stored on tape at the New Mexico Institute of Mining and Technology. The files can be obtained by writing to Dr. G. W. Gross at the address below. The data sets and the associated file names are given below.

Data Set	File Name
1967 - 1968 time variable recharge	PINDR6.DAT
1967 - 1968 constant recharge	PINDR5.DAT
1967 - 1975 time variable recharge	PINDR4.DAT
1967 - 1975 constant recharge	PINDR3.DAT
Modified Computer Program	PINDR5.FOR

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