ANALYSIS OF DATA FROM TEST-WELL SITES ALONG THE DOWNDIP LIMIT OF FRESHWATER IN THE EDWARDS AQUIFER, SAN ANTONIO, TEXAS, 1985-87

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CONTENTS

FIGURES

TABLES

CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATIONS

Multiply	By	To obtain
acre-foot (acre-ft)	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per second (ft/s)	0.3048	meter per second
foot squared per day (ft^2/d)	0.09290	meter squared per day
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer
pound per cubic foot	0.01602	gram per cubic centimeter
square mile (mi^2)	2.590	square kilometer
Temperature		
degree Celsius (°C)	°F = 1.8×(°C) + 32	degree Fahrenheit (°F)

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Abbreviations:

g/cm^3, gram per cubic centimeter

µg/L, microgram per liter

µm, micrometer

µS/cm, microsiemens per centimeter at 25 degrees Celsius

mg/L, milligram per liter

Analysis of Data from Test-Well Sites Along the Downdip Limit of Freshwater in the Edwards Aquifer, San Antonio, Texas, 1985-87

By George E. Groschen

Abstract

Many researchers have studied the downdip limit of freshwater in the Edwards aquifer or various aspects of the saline-water zone and its relation to the freshwater zone. These studies were summarized and used to synthesize a consistent hydrologic and geochemical framework from which to interpret data from field studies. The concept derived from the previous work on the downdip limit of the freshwater zone is that fresh recharge water entered the aquifer and developed a vast flow system controlled by barrier faults. Some recharge water flows into the saline-water zone rather than toward major freshwater discharge points. The water that enters the saline-water zone continues to dissolve gypsum and dolomite, and calcite precipitates out of the water. This process of dedolomitization has helped to develop the large secondary porosity of the freshwater zone as the downdip limit of the freshwater zone progressively moved downdip in recent geologic time.

The drilling of test holes and installation of monitoring wells began in 1985 and helped to define the downdip limit of the freshwater zone at one location. Dolomite was found in greater amounts in rocks from the saline-water-zone test hole than in rocks from the freshwater-zone test hole. Other mineralogic and lithologic contrasts between the saline-water-zone test hole and the freshwater-zone test hole support the conceptual model of dedolomitization. Geophysical logs and test-hole survey logs helped to define the stratification of freshwater and associated altered rock textures, secondary porosity development, and water chemistry in the freshwater-, saline-water-, and transition-zone test holes.

The differences in actual measured head among the seven completed monitor wells varied over time, especially during periods of substantial water use. The water levels in two monitor wells completed at the freshwater zone (site D; wells D1 and D2) responded quickly and strongly to withdrawals. In the transition zone (site C; wells C1 and C2), the water level in the shallow monitor well (C1) completed in a cavern responded quickly, but because it was farther from the water-supply wells near site D, it did not respond as strongly to changes in withdrawals at the nearby well field. The water levels in the three saline-water wells at site A (A1, A2, and A3) and the deep site C well (C2) were less responsive to stresses relative to the water-level changes in the freshwater wells (D1, D2, and C1).

Large amounts of freshwater were produced from the upper 300 to 350 feet of the aquifer in the freshwater zone (site D). Water produced from below this interval was as saline as that from the saline-water zone (site A). The cavern near the top of the aquifer in the transition zone (site C) produced large amounts of freshwater. The freshwater produced from the cavern was of a different geochemical type than the water from the other wells. The saline-water-zone test hole (site A) produced small amounts of water having specific conductance generally about 5,800 to 6,200 microsiemens per centimeter at 25 degrees Celsius.

A consistent trend in the water quality was not detected in the monitor-well data for July 1986-April 1987. This was caused, in part, by the average to above-average rainfall and by the lack

of large withdrawals during the period. The water quality of samples from several of the wells was similar to the water quality determined by a previous study of the area. Geochemistry of the oil- or gas-well brines from downdip in the saline-water zone had slight resemblance to the geochemistry of the water at the downdip limit of freshwater; updip flow of saline water toward the freshwater zone was not indicated.

INTRODUCTION

The Edwards aquifer in south-central Texas supplies water for more than 1 million people in the San Antonio metropolitan area, for industry and agriculture, and for human and environmental needs downstream of major springs flowing from the aquifer (fig. 1). The updip part of the aquifer contains freshwater (freshwater zone) and the downdip part contains saline water (saline-water zone). Perez (1986) concluded that, under certain conditions, movement of saline water into the freshwater zone is possible. It is therefore critical to know: (1) if and under what conditions saline water will move into the freshwater zone; and (2) if it does, how rapidly and how much of the saline water will move. Should saline-water intrusion occur, the extensively leached freshwater part of the aquifer could allow rapid mixing of the saline water and freshwater, possibly causing contamination of the most productive part of the aquifer near San Antonio and northeastward toward the major springs at New Braunfels and San Marcos.

In 1985, the U.S. Geological Survey began a study to improve the definition of the downdip limit of freshwater and to assess the potential for saline-water movement into the freshwater zone. This study was conducted in cooperation with the San Antonio Water System (formerly the City Water Board), the Edwards Underground Water District, and the Texas Water Development Board.

The freshwater and saline-water zones of the aquifer are three-dimensionally complex and only roughly defined. The interface (surface) between the freshwater zone (dissolved-solids concentration less than 1,000 mg/L) and the saline-water zone (dissolved-solids concentration of 1,000 mg/L or greater) is defined for this report as the downdip limit of freshwater. The downdip limit of freshwater is an irregularly shaped surface, roughly concave upward, but tilted

toward the freshwater zone, similar to freshwater/seawater interfaces in coastal aquifers. The updip horizontal extent of the surface where it underlies freshwater in the Edwards aquifer is unknown, but its approximate intersection with the top of the aquifer is mappable on the basis of water-sample data from wells because most Edwards aquifer wells tap the uppermost section of the aquifer. The line on land surface directly overlying the intersection of the downdip limit of freshwater with the top of the aquifer locally is termed the "bad-water" line. With some exceptions, this imaginary "bad-water" line is the farthest horizontal extent of freshwater down the southeastward dip of the aquifer.

In addition to the freshwater and saline-water zones, a third zone of water quality in the Edwards aquifer is defined qualitatively for this report because the downdip limit of freshwater is not well characterized anywhere within the aquifer. The "transition zone," as used in this report, comprises subzones of the freshwater zone and the saline-water zone approximately within 1,000 ft on either side of the downdip limit of freshwater (fig. 2).

A series of test holes was drilled and tests were run in the open holes during drilling. Following those tests, permanent monitor wells were installed at three sites in or near the transition zone in San Antonio (fig. 3). Different tests were run in the open test holes than were run in the completed monitor wells. In this report, the terms "test hole," "open hole," or "primary test hole" refer to the first hole drilled at each of the three sites. Thus, three "test holes" or "open holes" are defined: the site A (saline-water-zone) test hole, in which monitor well A1 was completed; the site C (transition-zone) test hole, in which monitor-well C1 was completed; and the site D (freshwater-zone) test hole, in which monitor-well D1 was completed. The other test holes and the tests conducted in them during drilling generally are not discussed in this report.

The terms "monitor well" or "well" refer to the wells completed in either the primary or other test hole at each site. Three wells were completed at site A (A1, A2, A3), two wells at site C (C1, C2), and two wells at site D (D1, D2). In addition, water-level and water-quality data were obtained from several wells that were not part of the test-drilling and monitor-well program. These wells are referred to as "observation wells."

EXPLANATION

OUTCROP OF CONTIGUOUS HYDRAULICALLY CONNECTED UNITS OF THE EDWARDS AQUIFER

•••••• APPROXIMATE DOWNDIP LIMIT OF FRESHWATER ZONE OF THE EDWARDS AQUIFER

——— APPROXIMATE DOWNDIP LIMIT OF SALINE-WATER ZONE OF THE EDWARDS AQUIFER

Base from U.S. Geological Survey
1:250,000 quadrangles

Geology from University of Texas,
Bureau of Economic Geology, Geologic
Atlas of Texas, 1:250,000 quadrangles;
and Slagle and others (1986).

Figure 1. Location of study area.

INTRODUCTION 3
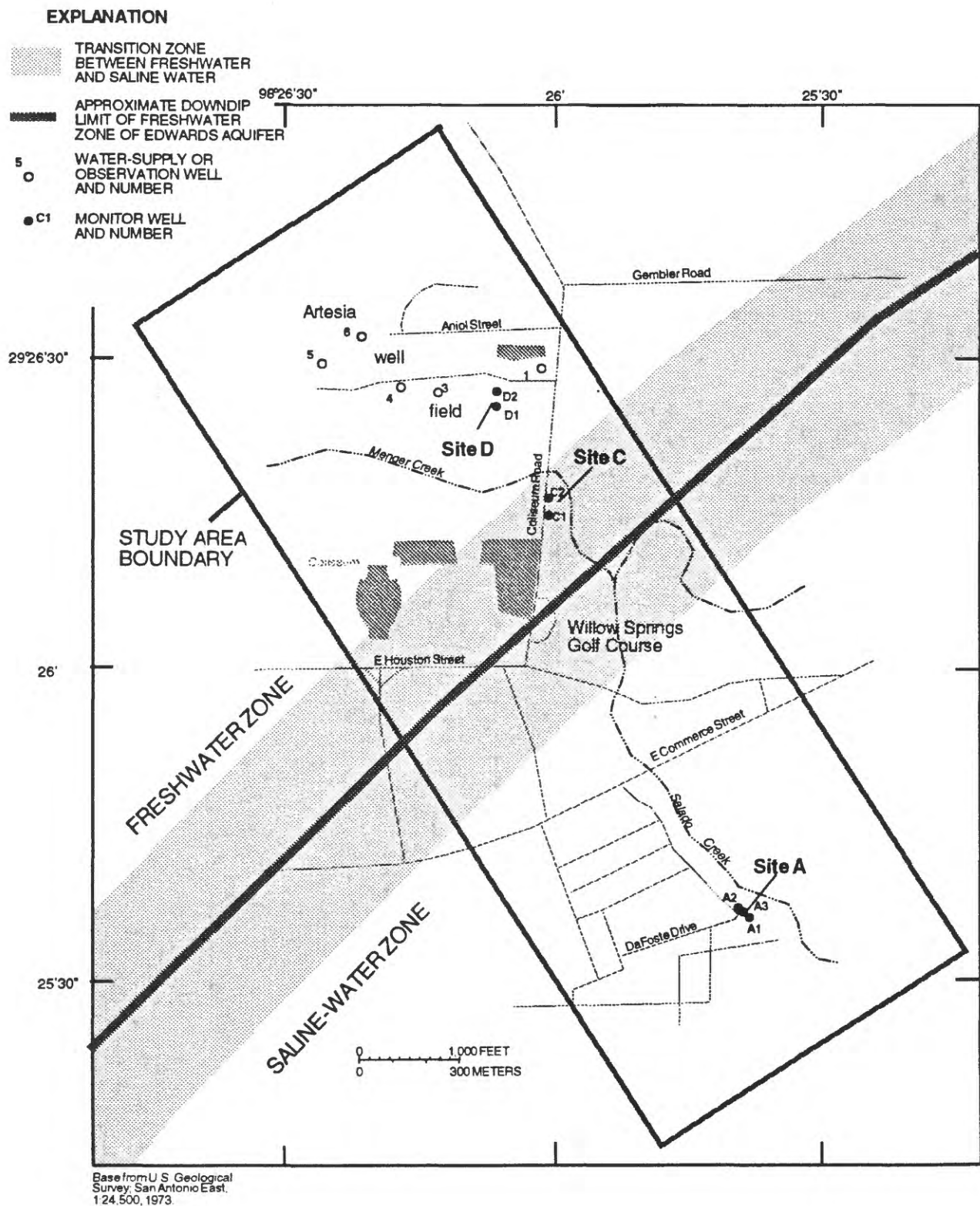
Figure 2. Location of wells and salinity zones in the study area, San Antonio, Texas.

Analysis of Data from Test-Well Sites Along the Downdip Limit of Freshwater in the Edwards Aquifer, San Antonio, Texas, 1985-87
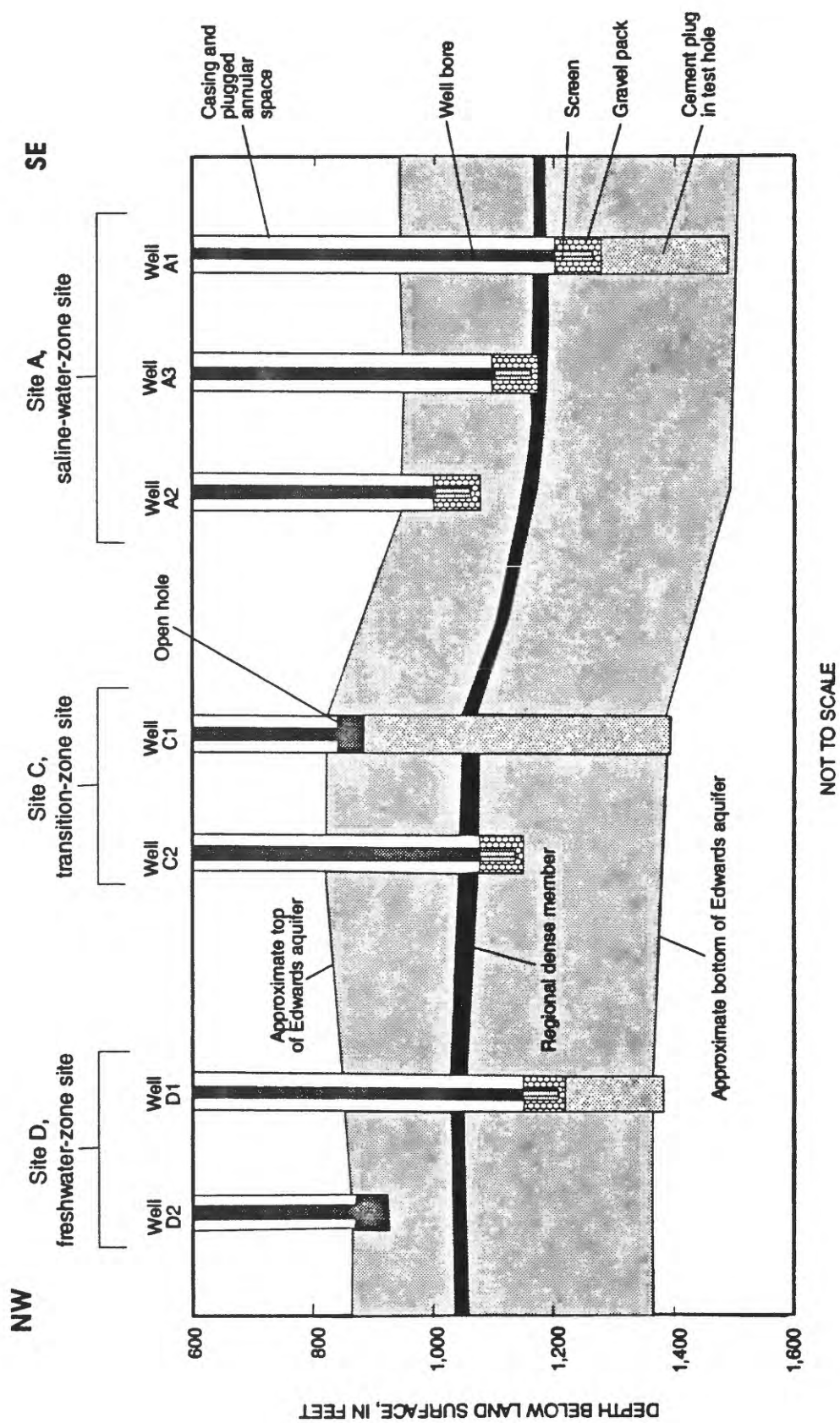
Figure 3. Diagrammatic section of test sites and design of monitor wells.

Purpose and Scope

This report describes the geology, hydrology, and water quality of the Edwards aquifer in and near the transition zone to improve the definition of the conditions that control the salinity and hydraulics of the saline water near the downdip limit of freshwater. Data from Pavlicek and others (1987) used in this report were collected between January 1985 and September 1986 from test holes and monitor wells in the study area (fig. 2). Additional data included in this report were collected from the monitor wells between October 1986 and April 1987. The focus of the study was limited to an area of about 1.2 mi^2 containing the seven monitor wells.

The specific objectives of the report are as follows:

1. To review the previous studies on the hydrologic system and summarize concepts about the downdip limit of freshwater and saline-water movement proposed by the authors of those studies as a basis for interpreting the data collected;

2. To describe the geology of the rocks that compose the aquifer near the downdip limit of freshwater on the basis of data collected during drilling and interpretation of the drill cuttings;

3. To analyze the geophysical logs of the test holes for correlation of stratigraphic units, aquifer characteristics, and aquifer conditions;

4. To describe the hydrology of the aquifer near the downdip limit of freshwater where the wells were drilled by analyzing the drawdown and water-level data; and

5. To analyze water-quality data to help interpret hydrologic concepts of evolution of water chemistry near the downdip limit of freshwater.

Approach

Previous studies of the Edwards aquifer that have interpretations of the relation of the aquifer to the downdip limit of freshwater, or of the origin of the saline-water zone, were reviewed before interpreting the data collected from the seven monitor wells. Many authors made separate and significant interpretations about various sections or aspects of the downdip limit of freshwater, or about the saline-water zone. These

were summarized to describe the saline-water zone and indicate the probability of saline-water intrusion from this zone. A more thorough interpretation of the data obtained from the test holes and monitor wells was possible because the data could be fit into a conceptual framework developed from the previous studies.

Several methods were used to analyze and interpret the data from Pavlicek and others (1987) and the data collected in 1986 and 1987. William F. Guyton Associates, Inc. (1986) made calculations to quantify the hydraulic properties of the aquifer near the monitor wells. The interpretations from that report are summarized in subsequent sections.

A test hole was drilled and logged at each of the three sites. The suites of geophysical logs, including compensated gamma-gamma (density), neutron, and acoustic velocity logs, were examined to determine the various aquifer properties that can be interpreted from them. These logs and the spontaneous potential and resistivity logs were used for geological correlation of the lithologic units that compose the Edwards aquifer. The fluid resistivity and temperature logs were used to interpret the water quality in the formations and the possibility of flow between members within the test hole drilled at each site.

Water-quality data from the test holes and monitor wells were examined to determine whether one or more models of aquifer or downdip limit of freshwater development were consistent with the data. Graphical analyses were used to further interpret the data.

Acknowledgments

William F. Guyton and Mervin Klug of LBG-Guyton Associates helped develop the study, provided technical assistance, and assisted in interpreting some of the data. Royce McDonald of the San Antonio Water System, Robert Bader of the Edwards Underground Water District, and Robert Bluntzer of the Texas Water Development Board provided technical assistance.

HYDROLOGIC SYSTEM

The freshwater zone of the aquifer is approximately 230 mi long from the Rio Grande to the Colorado River, and is about 5 to 40 mi wide (fig. 1). The freshwater zone of the aquifer is bounded to the north by the edge of the recharge zone (outcrop of contiguous and hydraulically connected rocks of the Edwards aquifer) and to the south and southeast by a zone of

chemical transition from freshwater to saline water. The top of the Edwards aquifer is about 1,000 ft below land surface at the downdip limit of freshwater near San Antonio, and the aquifer is about 500 ft thick.

In this report, the freshwater zone of the Edwards aquifer is defined as that part of the aquifer updip from the downdip limit of freshwater. The saline-water zone is that part of the Edwards aquifer downdip of the freshwater zone. Freshwater is defined as having dissolved solids of less than 1,000 mg/L and saline water is defined as having dissolved solids of 1,000 mg/L or greater. Slightly saline and brine refer to the opposite extremes of the range of saline-water concentrations. Brine is defined as having a dissolved-solids concentration greater than 35,000 mg/L.

Regional stratigraphic studies of the rocks that compose the Edwards aquifer by Rose (1972), and of equivalent rocks in south Texas by Fisher and Rodda (1969), Lozo and Smith (1964), and Tucker (1962), have resulted in subdivisions within the major depositional basins and correlations between the basins. The stratigraphically equivalent geologic units of the Edwards aquifer in this report are the Kainer and Person Formations and the overlying Georgetown Formation in the San Marcos Platform depositional province; the Devils River Limestone of the Devils River Trend depositional province; and the West Nueces, McKnight, and Salmon Peak Formations of Lozo and Smith (1964) in the Maverick Basin depositional province. Depositional provinces are shown in figure 4. The geologic unit stratigraphically below the Edwards aquifer is the Glen Rose Formation. This formation comprises a thick sequence of shale, calcareous shale, and dolomite in the upper part, and massive bedded limestone and dolostone in the lower part. The upper part of the Glen Rose Formation is the lower confining unit of the Edwards aquifer. The stratigraphic unit above the Edwards aquifer, the Del Rio Clay, is nearly impermeable and confines the ground water in units below.

Recharge to the aquifer results from precipitation on the outcrop area and from streams that drain the Edwards Plateau and lose water to the aquifer where the stream channels cross the outcrop area. During low-flow conditions, the streams lose virtually all their flows in the outcrop. The hydrologically significant faults, water-level contours for the winter of 1973, regional flow direction of ground water, and the major natural discharge points (springs) are shown in figure 5.

Recharge to and discharge from the aquifer have been reported by Nalley (1989, p. 14-21). Total recharge to and discharge from the aquifer averages about 636,000 acre-ft/yr. Recharge varies considerably depending on climatic conditions. Minimum recharge was about 44,000 acre-ft in 1956; maximum recharge was about 2,500,000 acre-ft in 1992 (Bader and others, 1993, p. 34, table 4.1). Discharge from the aquifer is primarily by pumping and springflow. Annual pumpage from the aquifer during 1980-88 ranged from about 364,000 acre-ft in 1987 to 540,000 acre-ft in 1988. The major pumping center is in the San Antonio area, where water levels from the index well "J-17" (AY-68-37-203) have ranged from 612 to 703 ft above sea level since 1932. Springflow has ranged from 69,800 acre-ft in 1956 to 803,000 acre-ft in 1992.

Recharge water generally moves parallel to fault barriers. As the water approaches the downdip limit of freshwater, it is deflected eastward or northeastward toward the major springs in the northeastern part of the aquifer.

Transmissivity in the freshwater zone near the downdip limit of freshwater is as great as 2 million ft^2/d (Maclay and Small, 1984, p. 50). Transmissivity on the saline-water side of the downdip limit of freshwater is not known but has been estimated to be about 11,600 ft^2/d or less (William F. Guyton Associates, Inc., 1986, p. 25). Average effective porosity or specific yield in the aquifer is about 6 percent in the unconfined zone. A substantial secondary porosity has been derived from the leaching of the once dolomitic formation. Conversely, carbonate-mineral precipitation has decreased the amount of primary porosity in the freshwater zone relative to the saline-water zone.

Previous Investigations

Published reports concerning the Edwards aquifer and the water resources in the San Antonio area contain references to, and some brief descriptions of, the downdip limit of freshwater. Only a few reports address the downdip limit of freshwater in detail.

Livingston and others (1936) concluded that the freshwater and saline water are found in the same geologic unit and are in contact. With respect to the aquifer in Bexar County, the report (p. 104) stated,

Figure 4. Location of oil or gas fields and depositional provinces of geologic units in the Edwards aquifer.
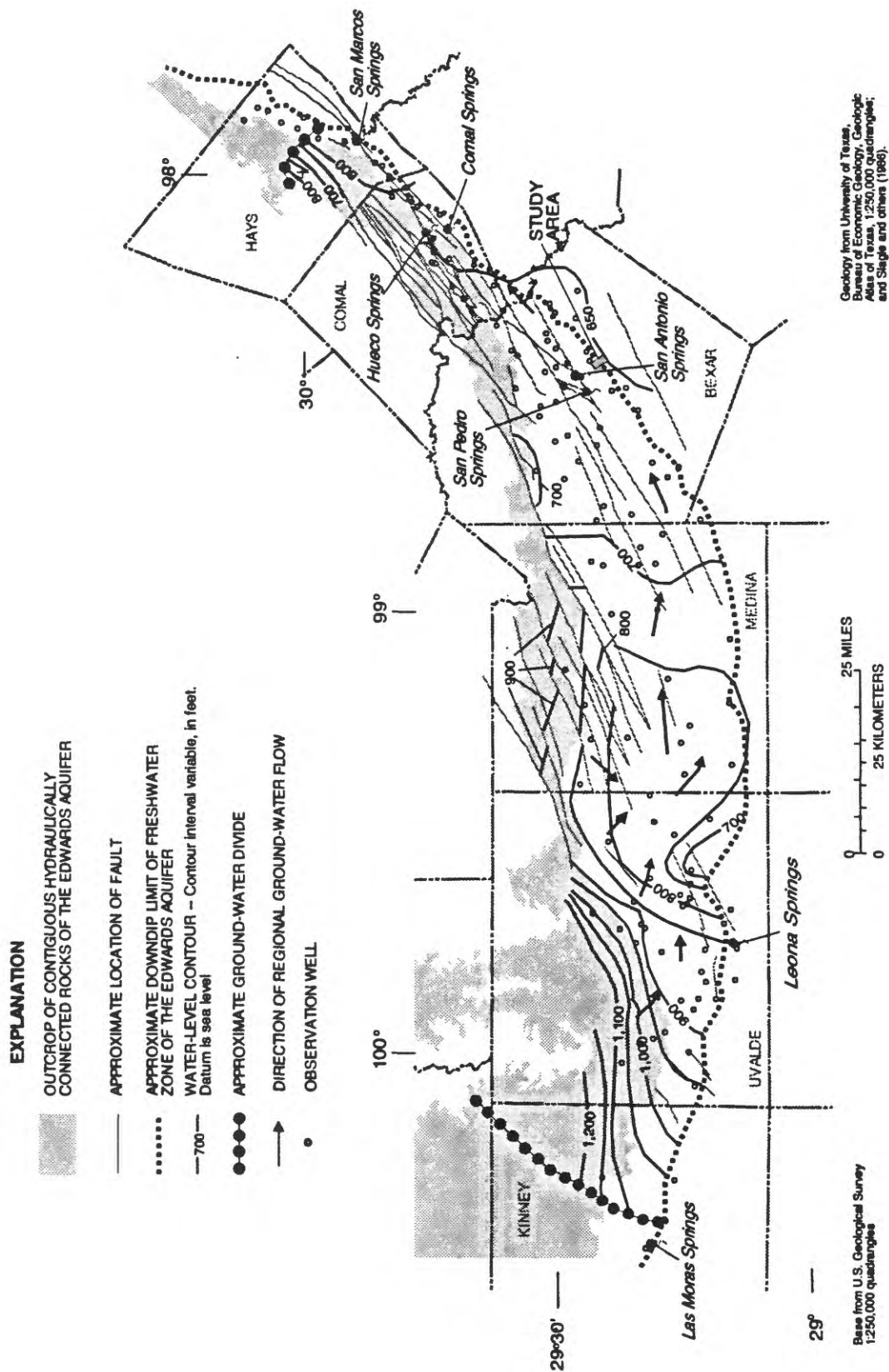
Figure 5. Location of selected observation wells and faults, water-level contours for winter 1973, regional flow direction in the freshwater zone, and major springs of the Edwards aquifer, San Antonio region, Texas. (Modified from Maclay and Land, 1988.)

"The line of demarcation between the normal Edwards water north of the fault and the highly mineralized water south of the fault is so sharp as to suggest that the fault retards free circulation of the water from the north to the south side and that this lack of circulation has been the chief cause of the high mineralization of the water south of the fault."

Petitt and George (1956, p. 73) indicated that the temperature gradient in the Edwards ground water generally is nonexistent between land surface and a depth of 600 ft; 0.3 °C per 100 ft between depths of 600 and 1,200 ft; and 1.7 °C per 100 ft between depths of 1,200 and 2,500 ft.

Holt (1959, p. 72) described the saline-water zone as a low-permeability zone with little circulation. The few wells that had been drilled in the saline-water zone had small yields and most were abandoned. Holt (1959, p. 72) also noted that many oil fields were associated with known faults and that the downdip limit of freshwater in parts of Medina and Bexar Counties seemed to be associated with the Pearson and Dunlay faults in southern Medina County. DeCook (1963, p. 1) concluded that the boundaries of the aquifer (freshwater zone) in Hays County were formed by faults.

Garza (1962a, p. 58a) completed the first detailed analysis of data specifically from the transition zone. Garza stated that hydrogen sulfide was not restricted to areas of the aquifer where the dissolved-solids concentration was 1,000 mg/L or greater as had been previously assumed, but also was found in many freshwater wells located in southern Medina and Uvalde Counties. Garza (p. 58a) also stated that:

"The mineralization of water from some wells in the transition zone between the water of good quality and the water of poor quality decreases as the artesian pressure increases."

Garza (1962a) compiled the first detailed map of the downdip limit of freshwater and mapped the sulfate and chloride data from water analyses to determine spatial trends in the data. Garza (p. 58a) concluded that the temperature of water with less than 500 mg/L dissolved-solids concentration increased about 0.3 °C per 100 ft of depth, and the temperature of water with dissolved solids in excess of 1,000 mg/L increased about 1.4 °C per 100 ft of depth. Garza's data were published in a separate report (Garza, 1962b).

On the basis of field work done between 1938-40, Bennett and Sayre (1962, p. 94) noted that hydrogen sulfide was detected in freshwater wells and spec-

ulated that hydrogen sulfide was derived in the aquifer. Hydrogen sulfide was generated from anhydrite by hydrocarbons in petroliferous layers found in some rock units of the Edwards aquifer.

Moredock and Van Siclen (1964, p. 269) concluded that the source of the hydrocarbons was the fine-grained sediments downdip (southeast) of the Stuart City reef trend (fig. 4), and that reverse-fractionation of the various hydrocarbons occurred as the hydrocarbon phase was expelled from the compacting sediments. The report concluded that: (1) differential migration was the best explanation for the distribution of the hydrocarbons found in the rocks of the oil or gas fields in the Edwards aquifer saline-water zone (fig. 4), although it did not completely explain the observations; (2) meteoric water from either Mexico to the west or southwestern Texas, where the Edwards aquifer or equivalent rocks crop out, imposed a southwest-to-northeast flow system on the fluid movement in the saline-water zone; and (3) the McKnight Formation also could be a source for some of the hydrocarbons found in the oil or gas fields in the Edwards aquifer saline-water zone.

The principal conclusions and recommendations of William F. Guyton Associates, Inc. (written commun., 1968) are as follows:

1. Most of the changes in the quality of water from individual wells to date were small. A large, lateral shift in the position of the downdip limit of freshwater was not apparent from the water-quality data available, and none was believed to have occurred in historical times.

2. The data indicate that considerably more wells with water-quality variations existed than were recognized, or were recognizable, in the past.

3. Water from wells located close to the downdip limit of freshwater tended to show larger and more easily recognizable changes in water quality than did other wells in the aquifer, but some changes in water quality were noted for water from wells located some distance north and south of the downdip limit of freshwater.

4. A more exhaustive study of the downdip limit of freshwater was not warranted in 1968, but more wells should be sampled, particularly in Bexar County.

William F. Guyton Associates, Inc. (written commun., 1968) concluded that some responses of water quality to water-level changes could be caused by time-lag effects; therefore, it is possible for the water level in the freshwater observation well to rise while the dissolved-solids concentration at a transition-zone well is increasing. William F. Guyton Associates, Inc., noted that the sulfate concentration generally increases along the direction from Brackettville in Kinney County to Barton Springs in Travis County. Water from Comal Springs has the smallest mean and least variation in sulfate concentration. The spring having the next lower land-surface altitude, San Marcos Springs, discharges water with the next largest variation and mean in dissolved sulfate. Barton Springs has the largest mean and greatest variation in sulfate and is the lowest freshwater spring from the Edwards aquifer south of the Colorado River.

Maclay and Rettman (1972) plotted the chemical facies of the water, but did not find any new insight into the hydrochemistry of the aquifer based on chemical facies. Pearson, as cited in Maclay and Rettman (1972, p. 17), stated that the ratios of carbon-13 to carbon-12 indicate that the carbonate dissolved in the freshwater has a mixed source of plant-derived carbonate and rock-derived carbonate. The carbonate from the saline-water zone could have a component derived from petroleum hydrocarbons.

R.W. Maclay (U.S. Geological Survey retired, written commun., 1973) presented data from a series of cored test holes drilled in the aquifer at various locations. The major findings of this study are the differences in the rock in core samples taken from the freshwater and saline-water zones:

Maclay concluded that there is a wide range in the amount to which the freshwater is saturated with respect to the mineral calcite, and the chemistry of the ground-water changes across the downdip limit of freshwater from strongly oxidizing freshwater to strongly reducing in the saline-water zone.

On the basis of radiocarbon and tritium data, Pearson (1973, p. 52-56) estimated the residence time of the main freshwater body was between 20 years and several thousand years, and that the water in the saline-water zone had a residence time of tens of thousands of years or more. On the basis of the stable isotopes of sulfur in sulfate and sulfide, Pearson (1973, p. 37) concluded that substantial sulfate reduction and associated generation of bicarbonate did not occur in the saline-water zone. Pearson also concluded that the hydrogen sulfide might be from deep in the saline-water zone.

Rightmire and others (1974) studied the sulfur isotope content of water from the Edwards aquifer. The authors concluded (p. 201) that the sulfate in the freshwater probably is derived from the Permian rocks in west Texas by atmospheric dust transport. The authors also suggest (p. 205) that bacterial reduction probably is not the source of the hydrogen sulfide and that the hydrogen sulfide and the dissolved sulfate in the saline-water zone are partly derived from a downdip brine source.

R.W. Maclay (U.S. Geological Survey retired, written commun., 1975) defined the lithologic and hydrologic differences between the freshwater-zone rocks and the saline-water-zone rocks. Maclay concluded that:

1. Dolomitic zones are more porous than limestone zones;

2. The pore spaces in dolomitic zones tend to be more interconnected, better sorted in size class, and more coarsely skewed in size than in the limestone zones;

Freshwater zone	Saline-water zone
a. Predominantly dense limestone	a. Dolostone or dolomitic
b. Light colored, gray	b. Dark gray to brown
c. Large secondary porosity	c. Mostly interparticle and intragranular primary porosity
d. Recrystallized, no highly soluble minerals or pyrite	d. Pyrite, gypsum and celestite
e. Sedimentary structures poorly preserved	e. Sedimentary structures retained
f. By Archie's (1952) matrix classification: type I and II	f. Mostly type III

3. The pore spaces in limestone zones are poorly connected, and the size distribution is variable; and

4. A direct correlation between matrix porosity and matrix permeability is evident for all samples of the core.

Maclay also concluded that:

1. Rocks south of the downdip limit of freshwater are more dolomitic with respect to the pure limestone of the freshwater zone;

2. Heat-flow experiments indicate that water moves in the saline-water zone;

3. Saline water is found near the bottom of the Edwards aquifer in the transition zone and in some areas where large amounts of freshwater are produced from wells tapping the upper Edwards aquifer; and

4. The regional dense member (RDM), an informal member proposed by Rose (1972), lies near the middle of the aquifer and apparently separates the freshwater overlying it from the saline water underneath it in some sections of the aquifer updip from the downdip limit of freshwater.

Pearson and others (1975) reported on the tritium data collected from wells, streams, and springs during 1963-71. The largest concentrations of tritium (produced by nuclear-weapon testing in the atmosphere) were from samples in and near recharge streams. The smallest measurable concentrations were from samples close to the downdip limit of freshwater and in Comal Springs in Comal County. Increased concentrations of tritium were not detectable in the saline-water zone.

Abbott (1975) described the hydrology of the Edwards aquifer with emphasis on the geomorphic development of the aquifer. Abbott (p. 260) concluded that the downdip limit of freshwater developed as:

" *** a by-pass boundary that meteoric groundwater moving under structural or hydrologic controls does not transgress. *** Originally the bad-water line was probably a random hydrologic boundary but it has become deeply ingrained with time."

On the basis of geochemical data, Pearson and Rettman (1976, p. 1-2), concluded that:

1. The water chemistry can be divided into five distinct groups on the basis of chemistry, isotopes,

and calculated mineral saturation indices—recharge group, main freshwater group, western varied group, western saline and transitional group, and eastern saline and transitional group (fig. 6); and

2. All the water groups are saturated with respect to calcite, and the western varied group and the saline and transitional groups also are saturated with respect to dolomite. Some saline-water samples are saturated with respect to gypsum ($CaSO_4 \cdot 2H_2O$), celestite ($SrSO_4$), strontianite ($SrCO_3$), and fluorite (CaF_2) (Pearson and Rettman, 1976, p. 22).

After further study of the cores from nine test holes in the aquifer, Maclay and Small (1976, p. 30-32) determined that the rocks of the saline-water zone, stratigraphically equivalent to the most permeable layers of the freshwater zone, are generally sucrosic dolostones with high porosity. Maclay and Small (1976, p. 22-25) developed a classification (subdivisions I to VIII) of the hydrogeology on the basis of the informal members described by Rose (1972) and the relative permeabilities of each member from laboratory tests and geophysical logs. The aquifer subdivisions that Maclay and Small (1984) developed for the San Marcos platform depositional province are shown in figure 7.

In terms of the position of the downdip limit of freshwater, Maclay and Small (1976, p. 32) stated that faults determine the location of most major springs and, thus, the zone of freshwater circulation—indirectly determining the location of the downdip limit of freshwater near major springs. In areas such as southern Medina County, a relation between faults and the downdip limit of freshwater is not apparent. On the basis of work by Jones (1975), Maclay and Small concluded that the thickness of the overburden could result in a condition where the net overburden pressure is sufficient to close vertical fractures, thus restricting circulation by preventing discharge of saline water. Maclay and Small (1976, p. 32) also concluded that because organic matter is found in the rocks of the saline-water zone, substantial amounts of freshwater have never flowed farther south than the location of the present-day downdip limit of freshwater.

On the basis of the constancy of the dissolved uranium at extremely low levels in freshwater in the Edwards aquifer, Cowart (1980, p. 281) stated that only small amounts of the rock matrix of the aquifer were

EXPLANATION

- OUTCROP OF CONTIGUOUS HYDRAULICALLY CONNECTED ROCKS OF THE EDWARDS AQUIFER

WATER-QUALITY GROUPS—defined by Pearson and Rettman, 1976

MF Main freshwater group

WV Western varied group

RG Recharge group

ES Eastern saline and transitional group

—WS— Western saline and transitional group

••••••• APPROXIMATE DOWNDIP LIMIT OF FRESHWATER ZONE OF THE EDWARDS AQUIFER

——— APPROXIMATE DOWNDIP LIMIT OF SALINE-WATER ZONE OF THE EDWARDS AQUIFER

Base from U.S. Geological Survey 1:250,000 quadrangles

Geology from University of Texas, Bureau of Economic Geology, Geologic Atlas of Texas, 1:250,000 quadrangles; and Slagle and others (1986).
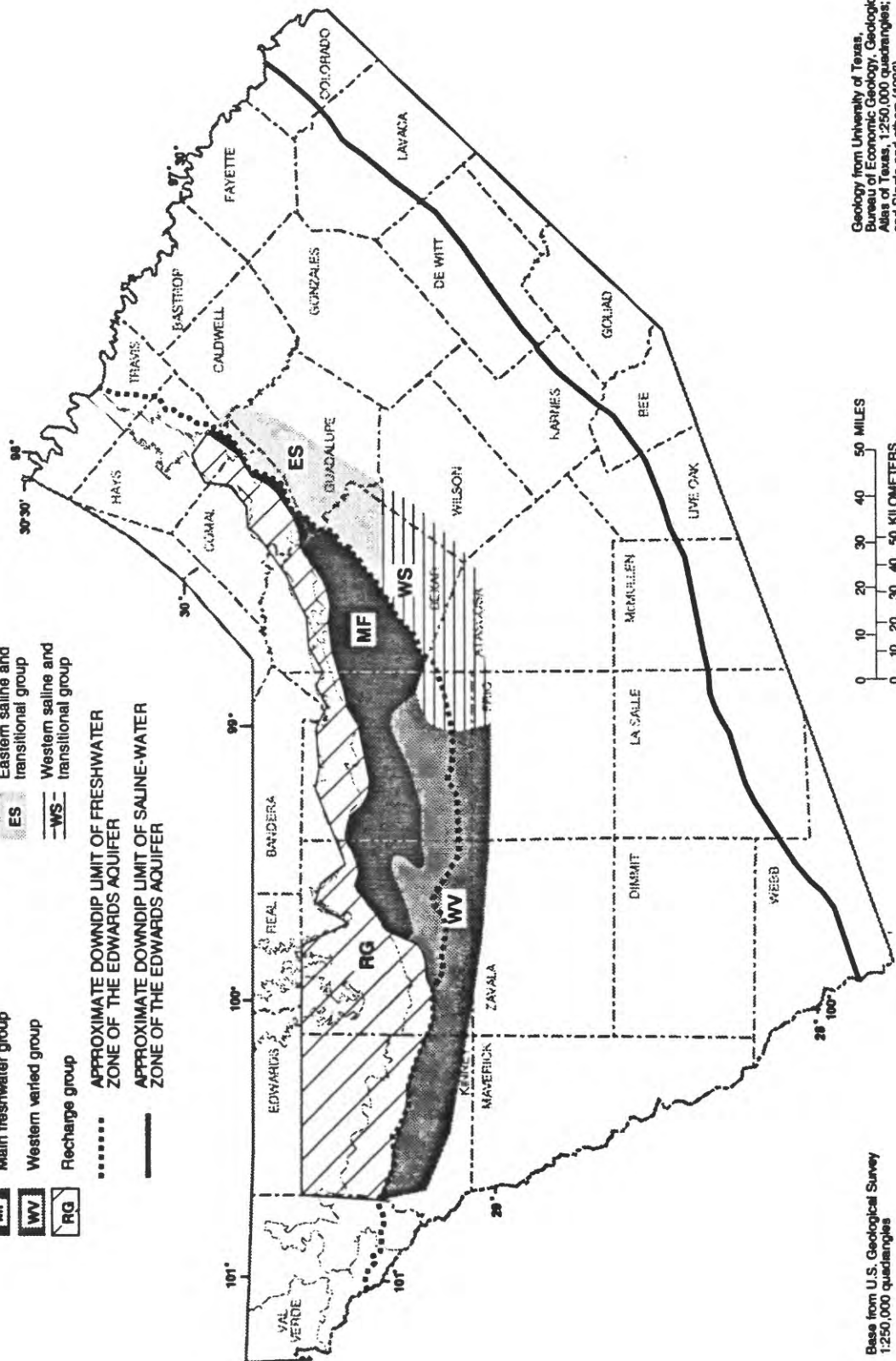
Figure 6. Approximate areas of water-quality groups in the Edwards aquifer, south-central Texas.
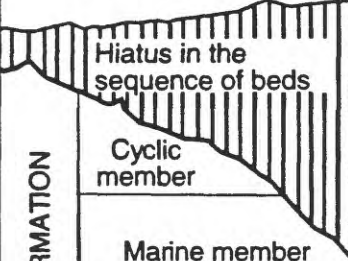
	RELATIVE PERMEABILITY	AQUIFER SUBDIVISIONS
BUDA LIMESTONE		
DEL RIO CLAY		
GEORGETOWN FORMATION	Low	I
Hiatus in the sequence of beds		
Cyclic member	High	II
Marine member		
Leached member	Very high	III
Collapsed member		
Regional dense member	Low	IV
Grainstone member	Moderate	V
Kirschberg evaporite	Very high	VI
Dolomitic member	Moderate	VII
Basal nodular member	Low	VIII
GLEN ROSE FORMATION		

PERSON FORMATION

KAINER FORMATION

Figure 7. Stratigraphic column showing subdivisions of the Edwards aquifer. (Stratigraphy from Rose, 1972; aquifer subdivisions from Maclay and Small, 1984.)

being dissolved. Uranium concentrations were less than 0.8 µg/L at both recharge and discharge areas of the aquifer. Several of the samples taken from the transition zone near the northwestern point of Atascosa County cannot be explained as a simple mixture of freshwater with saline water. Two samples indicate that the freshwater zone has intruded on the saline-water zone and moved the downdip limit of freshwater downdip, probably in recent geologic time (p. 282). One sample indicates that the opposite might have occurred.

Rye and others (1981) reinterpreted the data first presented by Pearson (1973) and Rightmire and others (1974). The isotopic composition of the dissolved sulfide and the dissolved sulfate from the saline-water zone indicated that a substantial amount of sulfate reduction occurs through microbial processes (Rye and others, 1981, p. 1949). The analytical data from a water sample taken from a well in the Charlotte oil field in Atascosa County (fig. 4) indicated that the isotopic composition of the sulfate and sulfide dissolved in the oil-field water was different from that of the saline water near the downdip limit of freshwater.

Prezbindowski (1981) and Land and Prezbindowski (1981) studied the aquifer near the downdip limit of the Edwards aquifer (fig. 1). Prezbindowski (1981, p. 174-185) concluded that the majority of the diagenetic changes in these rocks were caused by near-surface weathering processes that occurred shortly after the rocks were deposited. Deep burial of the rocks beneath the upper Cretaceous, Paleogene and Neogene sediments has had little effect on these rocks. Isotope compositions of the water and rocks indicate that the two phases are not in isotopic equilibrium throughout the saline-water zone, possibly indicating recent infusion of brines from deep in the Gulf of Mexico basin either updip or upfaults. Land and Prezbindowski (1981) concluded that these "basinal brines" developed through re-solution of evaporites in underlying Jurassic rocks to increase the salinity and through albitization to increase the calcium to sodium ratio. In the Edwards aquifer, the fluid causes dedolomitization as the fluid moves updip toward the freshwater zone.

Stoessell and Moore (1983) made a different interpretation of the sample data from oil or gas wells collected by Prezbindowski (1981). Stoessell and Moore (1983, p. 897-898) concluded that the chemical composition of the Edwards brines could be explained by the mixing of a fluid similar to modern seawater and a brine derived from the Jurassic Louann salt deposits

that had originally underlain the Cretaceous rocks. The authors further stated that the ratio of ions in the water samples indicate equilibration with progressively more highly ordered dolomite with increasing temperature.

Land and Prezbindowski (1983, p. 119-120) supported the concept of dissolution of dolomite on the basis of estimated volumes of brine from the Jurassic sediments and strontium stable isotopes.

Maclay and Small (1984, p. 44-66) concluded that:

1. The permeability of the freshwater zone is directly related to particular strata (lithofacies) and to leaching. Dissolution of pore walls resulted in permeable strata by forming cavernous or honeycomb porosity at stratigraphically-controlled intervals (aquifer subdivisions III, V, and VI in fig. 7).

2. The diagenetic changes that resulted in these stratigraphically controlled porous zones have not affected the saline-water-zone rocks. These rocks are typically dolomitic, medium to dark gray or brown, and contain unoxidized organic material and minerals such as pyrite (FeS_2), gypsum ($CaSO_4$), and celestite ($SrSO_4$). The matrix of the saline-water zone is more porous, but the pores are less connected than the matrix of the freshwater zone. Freshwater zone rocks are calcitic, light-colored, mostly recrystallized, and dense. They contain little to no pyrite or gypsum.

3. Faults with a substantial amount of offset tend to divert flow of water perpendicular to the dip of the Edwards aquifer.

4. In Bexar County, the water levels or pressures in the saline-water-zone wells change 1 hour to several days after the water-level changes in the freshwater zone.

5. Some of the permeability of the aquifer in the confined zone might be attributed to the remnants of paleokarst.

Mench-Ellis (1985, p. 221-225) reconstructed the diagenetic history of the Edwards aquifer saline-water zone after deposition. Before faulting, changes occurred early after the deposition of the sediments, primarily by meteoric water weathering and by mixing between freshwater and saline water or brine. The changes resulted in minerals with texture and isotopic composition distinct from minerals formed later. After

fault movement began in the Miocene Epoch, a through-flowing freshwater system developed in the faulted area. Further changes occurred that altered the rocks in the present freshwater zone. The fluids in the saline-water zone were not flushed.

Perez (1986) used available data and a one-dimensional solute-transport model to study the potential for saline-water intrusion into the freshwater zone. Because data were not available for all aquifer properties and conditions in the saline part of the aquifer, various sensitivity tests of a range of reasonable values for these undefined properties were conducted. Primary conclusions of the study were that the conditions of hydraulic head, transmissivity, porosity, and dispersivity in the saline-water zone need to be better defined with field data to adequately assess the problem of potential saline-water intrusion.

William F. Guyton Associates, Inc. (1986) presented a report describing the drilling and testing of the holes for the study of the downdip limit of freshwater, and made some preliminary interpretations of the hydraulic characteristics of the Edwards aquifer observed in the test holes.

A report on the distribution of chemical facies in the saline-water zone near the downdip limit of freshwater was presented by Clement and Sharp (1988). The westernmost (fig. 8, A) chemical facies discussed is a dilute (in terms of saline water) calcium sulfate water in the far western part of the saline-water zone. The authors concluded that the westernmost (A) facies was developed by freshwater recharge moving into the saline-water zone and dissolving gypsum. The facies to the east (fig. 8, C and D) (Clement and Sharp, 1988, p. 139-140) become progressively more enriched by sodium and chloride in an easterly direction. The proportion of magnesium increases to the east primarily because of an increasing proportion of dolomite in the aquifer. In the northeasternmost facies (D), sodium and chloride are major constituents in the water (Clement and Sharp, 1988, p. 140-141). The largest concentration of sodium and chloride and the most saline water in the saline-water zone near the downdip limit of freshwater are in Guadalupe and Comal Counties where the displacement of faults is most pronounced.

Clement and Sharp (1988, p. 139-140) concluded that upfault movement of sodium calcium chloride brine is hydrologically significant because:

1. The geothermal flux alone cannot explain the higher temperatures of water from wells in the Edwards aquifer near the downdip limit of freshwater;

2. A well about 10 mi south of the downdip limit of freshwater contains more sulfate and less sodium and chloride than a well only 5 mi south of the downdip limit of freshwater. This type of discontinuity can happen only if faults between the two wells yield brine from below the Edwards aquifer;

3. The ratio of the concentration of calcium plus magnesium to sulfate plus bicarbonate is about 1.0 or less in water from facies A where gypsum and carbonates are dissolved, and greater than 1.0 in water from facies C, indicating a source for the excess calcium or magnesium, or both, outside the aquifer;

4. The ratio of sodium to chloride of the facies C water is near 1.0 and similar to the ratios in water from the nearest oil fields in Caldwell and Guadalupe Counties (fig. 4); and

5. The concentrations in the water of facies C were adequately simulated by mixing water of facies A with a representative sodium calcium chloride brine.

Summary of Concepts about Saline-Water Movement

On the basis of the previous investigations, the following process probably has occurred and might still be occurring in the aquifer. Freshwater recharges the aquifer, dissolves some calcite (if it is not saturated with calcite already) in the recharge zone, and reaches saturation with respect to calcite. As the water moves downgradient (not necessarily downdip), it approaches the downdip limit of freshwater. Most water is diverted by faults to the discharge points of the major springs, while some water flows into the saline-water zone and dissolves dolomite and gypsum. As the gypsum dissolves and adds calcium to the water, calcite becomes supersaturated and precipitates into dead-end pores and tight fractures. As the precipitation of calcite depletes the dissolved bicarbonate, dolomite, which is abundant in the saline-water zone and less abundant in the freshwater zone, remains undersaturated and dissolves. Dolomite dissolves and calcite continues to precipitate until the water approaches saturation with respect to gypsum. Saturation is about 2,000 mg/L of sulfate,

Figure 8. Location of saline-water zone hydrochemical facies near the downdip limit of freshwater in the Edwards aquifer, south-central Texas.

typically the greatest sulfate concentration observed in the saline-water zone near the downdip limit of freshwater. Sulfate reduction can occur, depleting dissolved sulfate, resulting in more gypsum dissolving and, thus causing the dedolomitization process to continue. Because the permeability of rocks in the saline-water zone is small relative to the permeability of the freshwater zone, the movement of water in the saline-water zone is much slower than in the freshwater zone.

The flow system in the saline-water zone probably is controlled to some extent by barrier faults. The water movement in the saline-water zone probably is parallel to the flow in the freshwater zone. Updip movement of saline water from the deeply buried Edwards aquifer also can substantially affect the type of reactions and the composition of the water in the saline-water zone.

The factors that partly control the existence and position of the saline-water zone are divided into five categories. The categories are: (1) the relation between the openness of fractures and the hydrostatic and lithostatic pressure in the aquifer; (2) the contrasts in mineralogy between the freshwater zone and the saline-water zone; (3) the chemistry of the fluids in the rocks—also related to 2; (4) the effects of faults that control flow direction and the location of major discharge points; and (5) the dynamics of water flow in the freshwater zone as opposed to the slow-moving saline water, and the possibility of the mixing of different water types in the transition zone.

At present (1993), data are not available for evaluating the extent to which the lithostatic pressure could be restricting vertical movement of fluids through fractures, faults, or both. The contrast in mineralogy between the two zones of the aquifer is well established from cores and drill cuttings. The difference in water chemistry of the two aquifer zones is likewise well established. A consistent hydrologic conceptual system might not indicate whether the mineralogy and chemical environment in the saline-water zone constitutes the necessary and sufficient conditions to develop the saline water from meteoric freshwater.

The effects of faults as barriers to flow within the freshwater zone are discussed by Maclay and Small (1984) and Maclay and Land (1988). The effectiveness of faults as vertical pathways for fluid flow can only be positively identified at the major springs of the freshwater zone. The major springs, San Antonio, San Pedro, Comal, and San Marcos (fig. 5), all rise from the deep confined Edwards aquifer. These springs are

at the intersection of major faults or subsurface fault-bounded blocks (Maclay and Land, 1988; R.W. Maclay, U.S. Geological Survey retired, oral commun., 1988). Faults that are barriers to horizontal flow must be nearly impermeable to flow across them and probably are nearly impermeable to vertical flow. The potential for certain fault intersections to open vertical pathways for fluid movement from aquifers underlying the Edwards aquifer has not been evaluated in detail. The possibility of faults of large displacement juxtaposing deeper saline-water aquifers against the Edwards aquifer, thereby creating a hydraulic connection, also has not been evaluated. Allen (1989, p. 805) has demonstrated circumstances where a fault cuts a series of permeable arched strata and results in conditions for a spiral or zigzag pattern of upward flow to occur at a fault even if the fault surface itself is not permeable. This condition is referred to as a cross-fault vertical migration. Allen (1989) primarily discussed hydrocarbon migration, but the process is feasible for any subsurface fluid.

The effect of kinetically controlled reactions on water chemistry in the saline-water zone has not been addressed specifically. Relative to the fast-moving water in the freshwater zone, the slow-moving water in the saline-water zone should be more affected by slow reactions. Bacterially-mediated sulfate reduction could be one of these slow reactions. Dissolution of evaporite minerals usually is not considered a slow reaction, but the accompanying pore-filling calcite precipitation reaction could be considered a slow reaction.

A final factor that was not discussed by most researchers is the relation of density-related flow to the dip of the rocks that constitute the aquifer.

ANALYSIS OF DATA FROM THE TEST-WELL SITES

The data collected during the drilling of the wells at the test sites can be more fully interpreted using the conceptual framework described in the previous section. The data are divided into four categories: Geologic, geophysical, hydrologic, and water quality.

Geologic Data

Drill cuttings collected for this study were examined and the results are listed in Pavlicek and others (1987). Summary illustrations of the stratigraphy, lithology, textures, and porosity data, and the

identification of the formations from preliminary examination, are modified from Pavlicek and others (1987) in figures 9-11. The informal members of the Person and Kainer Formations (Rose, 1972) are shown in these figures. Also shown in these figures are the textural terms used in table 1, which is the carbonate-rock classification system of Dunham (1962, table 1, reprinted by permission). The porosity-classification system of Choquette and Pray (1970, fig. 2, reprinted by permission), shown in figure 12, also is used in figures 9-11. Comparison of figures 9-11 indicates that there are many distinct differences in geologic characteristics across the downdip limit of freshwater.

In the drill cuttings taken from the site A test hole in the saline-water zone (fig. 2), dolomite was identified in 7 of the 20 cuttings from above the RDM and in 14 of the 29 cuttings from below the RDM (fig. 9). In the site C test hole, located in the transition zone, dolomite was identified in 10 of the 21 cuttings from above the RDM and in 17 of the 32 cuttings from below the RDM (fig. 10). In the site D test hole in the freshwater zone, dolomite was not identified in any of the 18 cuttings from above the RDM, but was identified in 19 of the 33 cuttings from below the RDM (fig. 11). If these cuttings are representative of the rocks in the Edwards aquifer at the sites, less dolomite above the RDM is related to greater water yield and less salinity. This relation does not exist at site C, but a cavern, located in the top of the Person Formation at this test hole, dominated the results. The large amount of dolomite in the cuttings from all the test holes below the RDM, and the presence of saline water in these sections, indicate that freshwater has not circulated in the section of aquifer below the RDM as much as it has in the rocks above the RDM at all the sites.

Pyrite was identified in the Georgetown Formation at all sites and in the top 5 to 15 ft of the Edwards aquifer at sites A and C. The presence of pyrite indicates that oxygen rich freshwater probably has not circulated through these rocks. The Georgetown Formation generally is one of the least permeable hydrostratigraphic units in the Edwards aquifer (Maclay and Small, 1984).

If calcite precipitates in the aquifer matrix, as proposed by Mench-Ellis (1985), the presence of megascopic crystals, or aggregates of crystals, of clear or drusy calcite would be an indication of the progression of the dedolomitization reaction. In the cuttings from site A, single-crystal calcite or aggregate or calcite crystal druse were identified in eight of the cuttings

from above the RDM and in three of the cuttings from below the RDM (fig. 9). In the site C test hole, single-crystal calcite or aggregate was identified in five of the cuttings from above the RDM and in two samples from below it (fig. 10). In the site D test-hole samples, single-crystal calcite or aggregate calcite crystal druse was identified in nine cuttings from above the RDM and in eight cuttings from below the RDM (fig. 11). Substantially more of the site D cuttings contained single-crystal calcite or aggregate calcite crystal druse than did the site A cuttings. According to the dedolomitization reaction, calcite precipitation should precede the preferential dissolution of dolomite; if the cuttings are representative of the sites, this relation is borne out.

Celestite was identified in one sample from above the RDM in the site A test hole and in eight samples from below the RDM. In cuttings from site C, celestite was not identified in any of the samples from above the RDM, but was identified in two samples from below it. In cuttings from the site D test hole, celestite was identified in two samples from above the RDM, but was not identified in any samples from below it. Celestite identifications were tentative because celestite is difficult to identify accurately by optical microscopy alone. The cause of the distribution of celestite in these test holes is not well understood.

Another possible indication of the amount of water circulating in the rocks is the number of calcite filled microfractures identified in the drill cuttings. Microfractures filled with mineral precipitates (calcite) were identified in five cuttings from the site A test hole above the RDM, but not in any cuttings from below the RDM. In the site C test hole, filled microfractures were identified in five cuttings from above the RDM and in one cutting from below it. Filled microfractures were not identified in cuttings from site D. Filled microfractures could be an index of the relative state of saturation of the pore water residing in the rocks. Lack of filled microfractures indicates that the pore water in the rocks of the freshwater zone is not supersaturated with respect to any mineral; however, the presence of filled microfractures at site C and site A test holes indicated that the saline water is supersaturated with respect to calcite.

The most important geologic and hydrologic differences among the three lithologic columns (figs. 9-11) are the amount and type of porosity in the aquifer matrix. In the site A test hole, eight drill cutting samples from above or in the RDM and six cutting samples

DEPTH, IN FEET BELOW LAND SURFACE

EDWARDS AQUIFER

STRATIGRAPHY	LITHOLOGY	TEXTURE	MATRIX POROSITY

STRATIGRAPHY

Georgetown Formation

Person Formation — Cyclic, marine, leached, and collapsed members

Regional dense member

Kainer Formation — Grainstone member

Kirschberg evaporite and dolomitic member

Basal nodular member

1,000 — 1,050 — 1,100 — 1,150 — 1,200 — 1,250 — 1,300 — 1,350 — 1,400 — 1,450 — Total depth 1,489

EXPLANATION

STRATIGRAPHY
Stratigraphy from Rose, 1972 (fig. 5)

LITHOLOGY

Fossil allochem

  ∂   milliolid foraminifera

  ♫   caprinid rudistid

  C   *Toucasia* rudistid

  ♀   gastropod

  ∟   other mollusc fragments

Mineral constituent

  //// dolomite (otherwise calcite)

  ▲ chert

  ■ pyrite

  ▭ single crystal calcite or aggregate

  ⋈ calcite crystal druse

  ⊠? celestite?

  • pyrite replaced allochems, or
  •• black rotund bodies (BRB)

Sedimentary structure

  ⁓⁓ pressure solution boundaries or clay seams

Tectonic structure

  ⌗ calcite-filled microfracture

TEXTURE (from Dunham, 1962)

  m = lime mudstone (micrite)

  w = wackestone

  p = packstone

  g = grainstone

MATRIX POROSITY

Type (from Choquette and Pray, 1970)

  BP = interparticle

  WP = intraparticle

  BC = intercrystal

  MO = moldic

Degree

  n = negligible

  poor, fair, and good are qualitative modifiers

NOTE: Cuttings collected at approximately 10-foot intervals. All descriptions based only on hand-sample examination of cuttings.

Figure 9. Stratigraphy, lithology, textures, and matrix porosity of the drill cuttings, site A test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987, p. 21.)
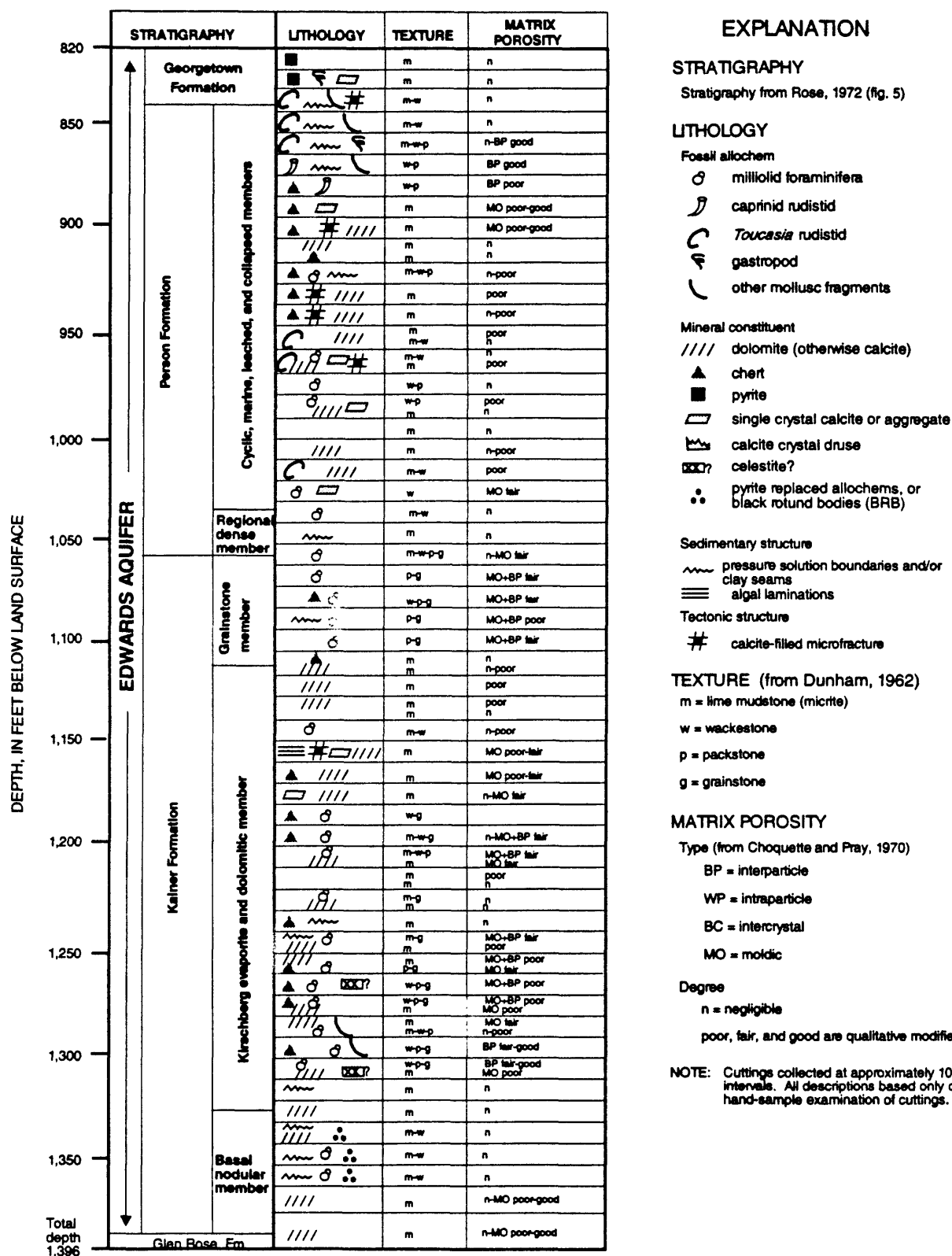
STRATIGRAPHY	LITHOLOGY	TEXTURE	MATRIX POROSITY

EDWARDS AQUIFER

Georgetown Formation

Person Formation — Cyclic, marine, leached, and collapsed members

Regional dense member

Grainstone member

Kainer Formation — Kirschberg evaporite and dolomitic member

Basal nodular member

Glen Rose Fm

820, 850, 900, 950, 1,000, 1,050, 1,100, 1,150, 1,200, 1,250, 1,300, 1,350

Total depth 1,396

EXPLANATION

STRATIGRAPHY
Stratigraphy from Rose, 1972 (fig. 5)

LITHOLOGY

Fossil allochem

- ♂ miliolid foraminifera
- ♬ caprinid rudistid
- C *Toucasia* rudistid
- ⸮ gastropod
- ⸜ other mollusc fragments

Mineral constituent

- //// dolomite (otherwise calcite)
- ▲ chert
- ■ pyrite
- ▭ single crystal calcite or aggregate
- ⸽ calcite crystal druse
- ⸽⸽? celestite?
- •• pyrite replaced allochems, or black rotund bodies (BRB)

Sedimentary structure

- ∿∿∿ pressure solution boundaries and/or clay seams
- ≡ algal laminations

Tectonic structure

- ⚹ calcite-filled microfracture

TEXTURE (from Dunham, 1962)
m = lime mudstone (micrite)

w = wackestone

p = packstone

g = grainstone

MATRIX POROSITY

Type (from Choquette and Pray, 1970)

BP = interparticle

WP = intraparticle

BC = intercrystal

MO = moldic

Degree

n = negligible

poor, fair, and good are qualitative modifiers

NOTE: Cuttings collected at approximately 10-foot intervals. All descriptions based only on hand-sample examination of cuttings.

Figure 10. Stratigraphy, lithology, textures, and matrix porosity of the drill cuttings, site C test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987, p. 47.)

Figure 11. Stratigraphy, lithology, textures, Neogene diagenetic features, and matrix porosity of the drill cuttings, site D test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987, p. 69.)
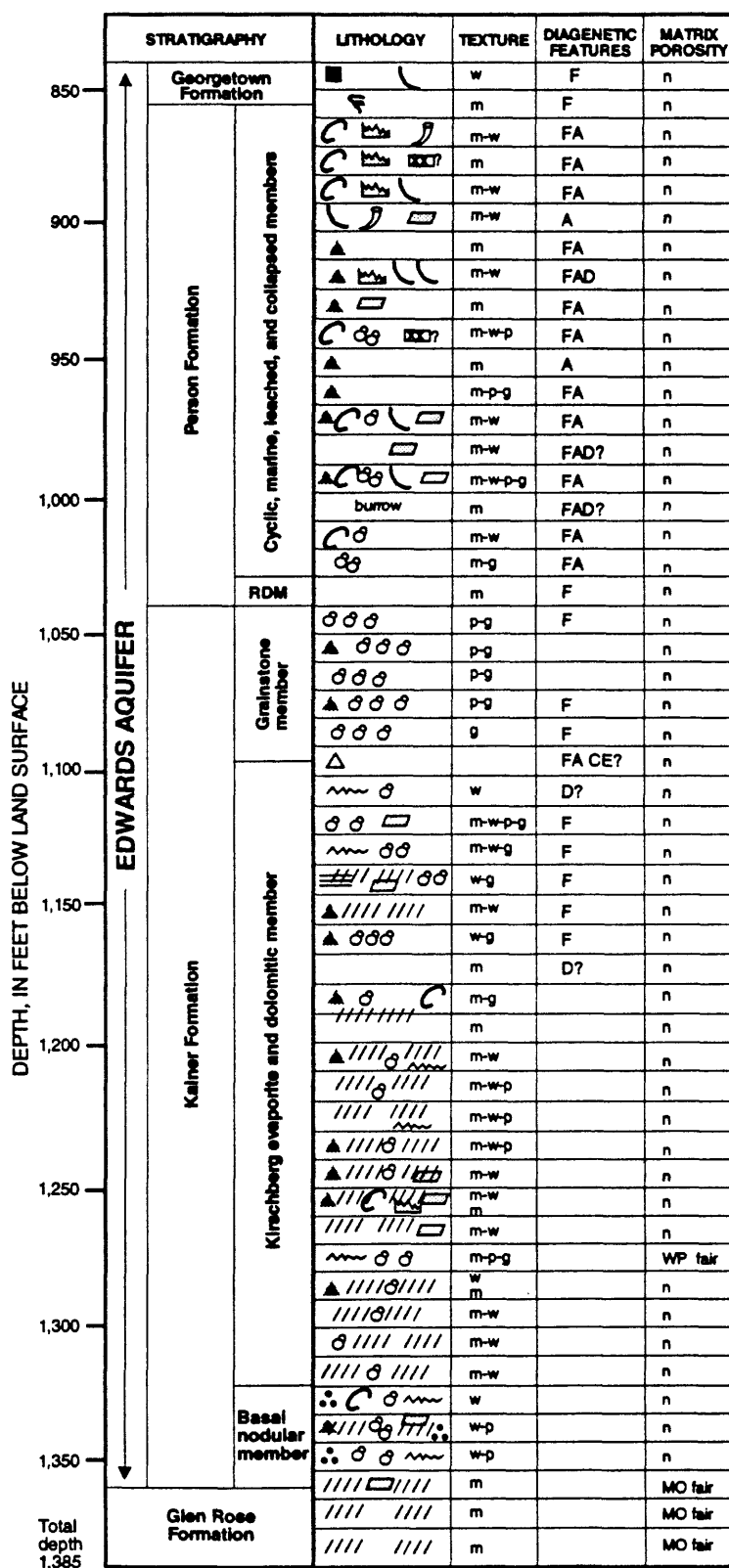
EXPLANATION

STRATIGRAPHY
   Stratigraphy from Rose, 1972 (fig. 5)
   RDM Regional dense member

LITHOLOGY
   Fossil allochem
   ♂   milliolid foraminifera
   ♫   caprinid rudistid
   C   Toucasia rudistid
   ⚲   gastropod
   ⌣   other mollusc fragments

   Mineral constituent
   ////   dolomite (otherwise calcite)
   ▲   chert
   ■   pyrite
   ▭   single crystal calcite or aggregate
   ⋈   calcite crystal druse
   ▨?   celestite?
   • •   pyrite replaced allochems, or black rotund bodies (BRB)

   Sedimentary structure
   ∿∿   pressure solution boundaries or clay seams
   ≡   algal laminations

TEXTURE   (from Dunham, 1962)
   m = lime mudstone (micrite)
   w = wackestone
   p = packstone
   g = grainstone

DIAGENETIC FEATURES
   F = ferric iron stain
   A = altered (associated with Neogene diagenesis)
   D? = dedolomite?
   CE? = calcitized evaporites?

MATRIX POROSITY
   Type (from Choquette and Pray, 1970)
   BP = interparticle
   WP = intraparticle
   BC = intercrystal
   MO = moldic

   Degree
   n = negligible
   poor, fair, and good are qualitative modifiers

NOTE: Cuttings collected at approximately 10-foot intervals. All descriptions based only on hand-sample examination of cuttings.

BASIC POROSITY TYPES

FABRIC SELECTIVE			NOT FABRIC SELECTIVE		
	INTERPARTICLE	BP			
	INTRAPARTICLE	WP		FRACTURE	FR
	INTERCRYSTAL	BC		CHANNEL*	CH
	MOLDIC	MO		VUG*	VUG
	FENESTRAL	FE		CAVERN*	CV
	SHELTER	SH			
	GROWTH-FRAMEWORK	GF	*Cavern applies to man-sized or larger pores of channel or vug shapes.		

FABRIC SELECTIVE OR NOT

BRECCIA BR	BORING BO	BURROW BU	SHRINKAGE SK

Figure 12. Porosity-classification system. (From Choquette and Pray, 1970, fig. 2, reprinted by permission.)

Table 1. Carbonate-rock classification system

[from Dunham, 1962, table 1, reprinted by permission]

Depositional texture recognizable					Depositional texture not recognizable
Original components not bound together during deposition				Original components were bound together during deposition... as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or questionably organic matter and are too large to be interstices.	Crystalline carbonate
Contains mud (particles of clay and fine silt size)		Grain-supported	Lacks mud and is grain-supported		(Subdivide according to classifications designed to bear on physical texture or diagenesis.)
Mud-supported					
Less than 10 percent grains	More than 10 percent grains				
Mudstone	Wackestone	Packstone	Grainstone	Boundstone	

from below it had negligible matrix porosity. In the samples from the site C test hole, negligible matrix porosity was identified in nine samples from above or in the RDM and in seven samples below it. In the site D cuttings, negligible matrix porosity was identified in all 18 samples from above the RDM; matrix porosity was identified in 4 of the 33 cuttings from below the RDM (including the Glen Rose Formation beneath the Edwards aquifer). This distribution follows the pattern identified by Maclay and Small (1984) in cores from the freshwater and saline-water zones.

Only the cuttings from the site D test hole had Neogene diagenetic features, on the basis of the description of diagenesis in the Edwards aquifer by Mench-Ellis (1985). These include ferric (reddish) iron staining, "altered" texture (primarily referring to extensive dissolution and recrystallization of the rocks), and tentatively identified dedolomite (Pavlicek and others, 1987). These are the major textures and petrographic characteristics of matrix rocks that were identified by Maclay and Small (1984) and Mench-Ellis (1985) as indicative of the final stages of the dedolomitization reaction and formation of the large secondary porosity responsible for the greater permeability of the freshwater zone. These diagenetic features were identified in all the cuttings from above the RDM, in the RDM cutting, and in 11 cuttings from below the RDM of the site D test hole. The diagenetic features were not observed in the deepest 20 cuttings from the site D test hole, but dolomite was identified in most of these cuttings. All of the petrographic observations about the drill cuttings were taken from Pavlicek and others (1987). Petrographic examination of rock thin sections is necessary to document, in detail, the Neogene diagenetic features.

Video logs made of the test holes revealed several interesting and important aspects of the geology of the rocks. The site A test hole had more vertical fractures that appeared to be open to water flow below the RDM than above it. Relatively good development of vugs and openings for water flow was apparent in the site A and the site C test holes. Most vugs and openings were restricted to distinct strata several feet thick. The site D test hole, however, had few indications of vugs or other strictly solution-mediated openings. Broken rubble and boulders were common from the top of the Person Formation to about one third of the way down the Kainer Formation. Below this zone, the Kainer Formation in the site D test hole appeared similar to that in the other two test holes.

Geophysical Data

Site A

Geophysical logs for the test hole in the saline-water zone at site A are shown in figure 13. The hole-diameter log indicates that the test hole was not competent through most of the section. A number of washouts and greatly fractured or cavernous zones coincided with places where the diameter of the test hole was substantially greater than 8 in.

The tool used to measure gamma radiation responds to gamma rays emitted by the naturally occurring radioactive minerals usually present in clays. The amount of natural gamma radiation is directly related to the amount of clays and clay minerals in the rocks, especially in the pure carbonate rocks that make up the Edwards aquifer. The spike at point η in the natural gamma log (fig. 13) was caused by the reduction in gamma rays from the larger diameter test hole because the tool does not compensate for hole diameter. The enlargement of the test hole can be seen on the hole diameter log at the same depth. The spike at point θ (fig. 13, RDM) was caused by the substantially greater clay content of the RDM. This is one of the reasons that the RDM usually can be positively identified on gamma logs (Small, 1985). Combined with the hole diameter, which shows a narrow "in-gage" hole (the hole diameter is close to the diameter of the drill bit), the natural gamma and hole diameter indicate a competent and clayey section that is distinct from the rest of the test hole. The spike at point ι (fig. 13) indicates the increase in clay content of the basal nodular member, which usually is a distinguishing trait of the base of the Kainer Formation (Small, 1985).

The spontaneous potential logs and resistivity logs generally are not useful in carbonate rocks with freshwater fluid in the hole. The test holes were drilled using a reverse air rotary process without drilling fluid, which explains why distinguishing characteristics were not apparent in the spontaneous potential log (fig. 13). The resistivity logs show distinct spikes of increased resistivity above the RDM, but it is not known whether these indicate water with less salinity than water below the RDM. The fluid resistivity and temperature logs (fig. 14) indicate that the fluid above the RDM has greater resistivity than the fluid below the RDM. According to these data, the responses of the resistivity logs probably were caused by salinity variations of the formation water or the rugosity of the holes.

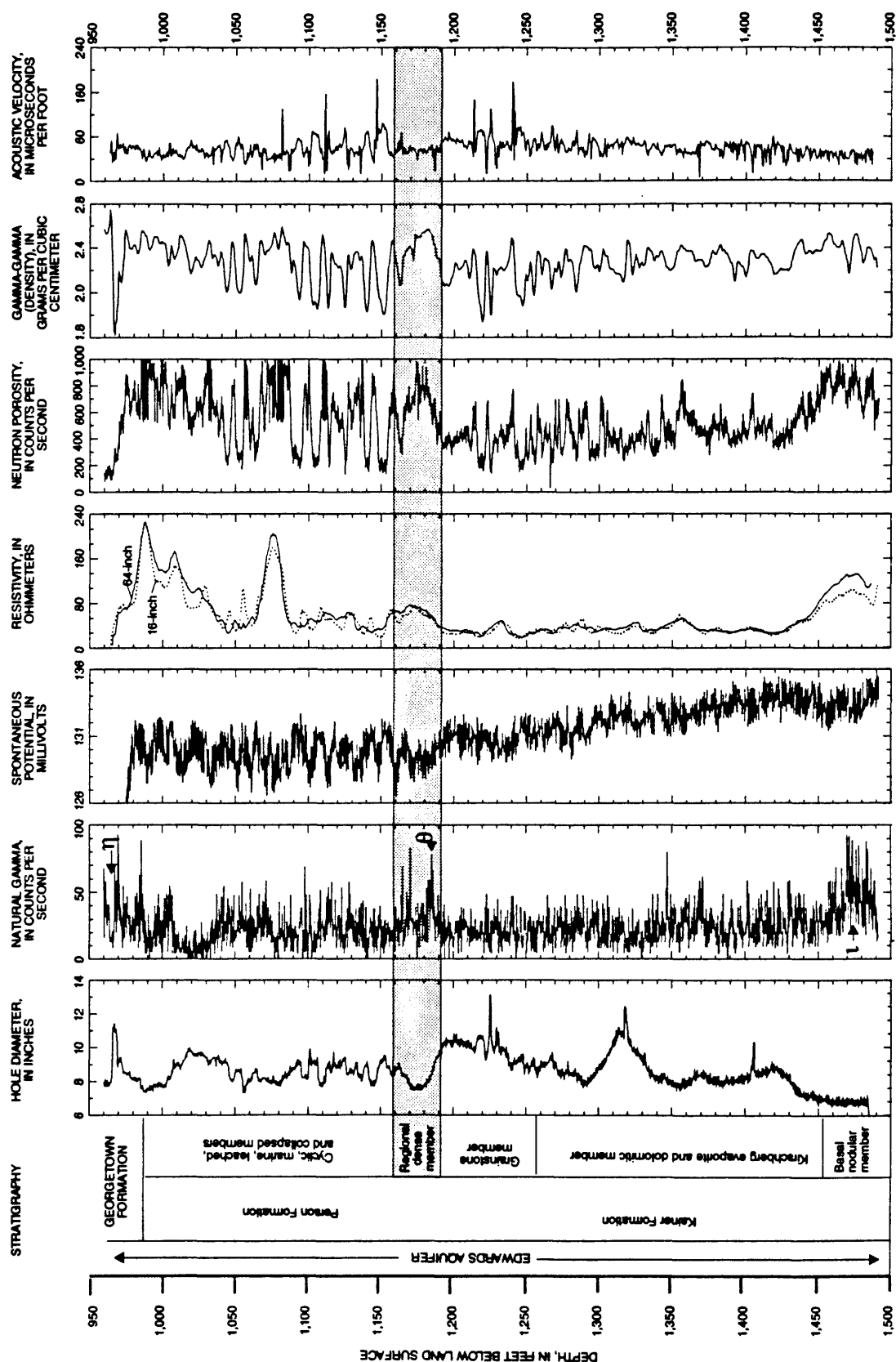**Figure 13.** Geophysical logs of site A test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987, 1987. η, θ, and ι denote deflections discussed in text.)
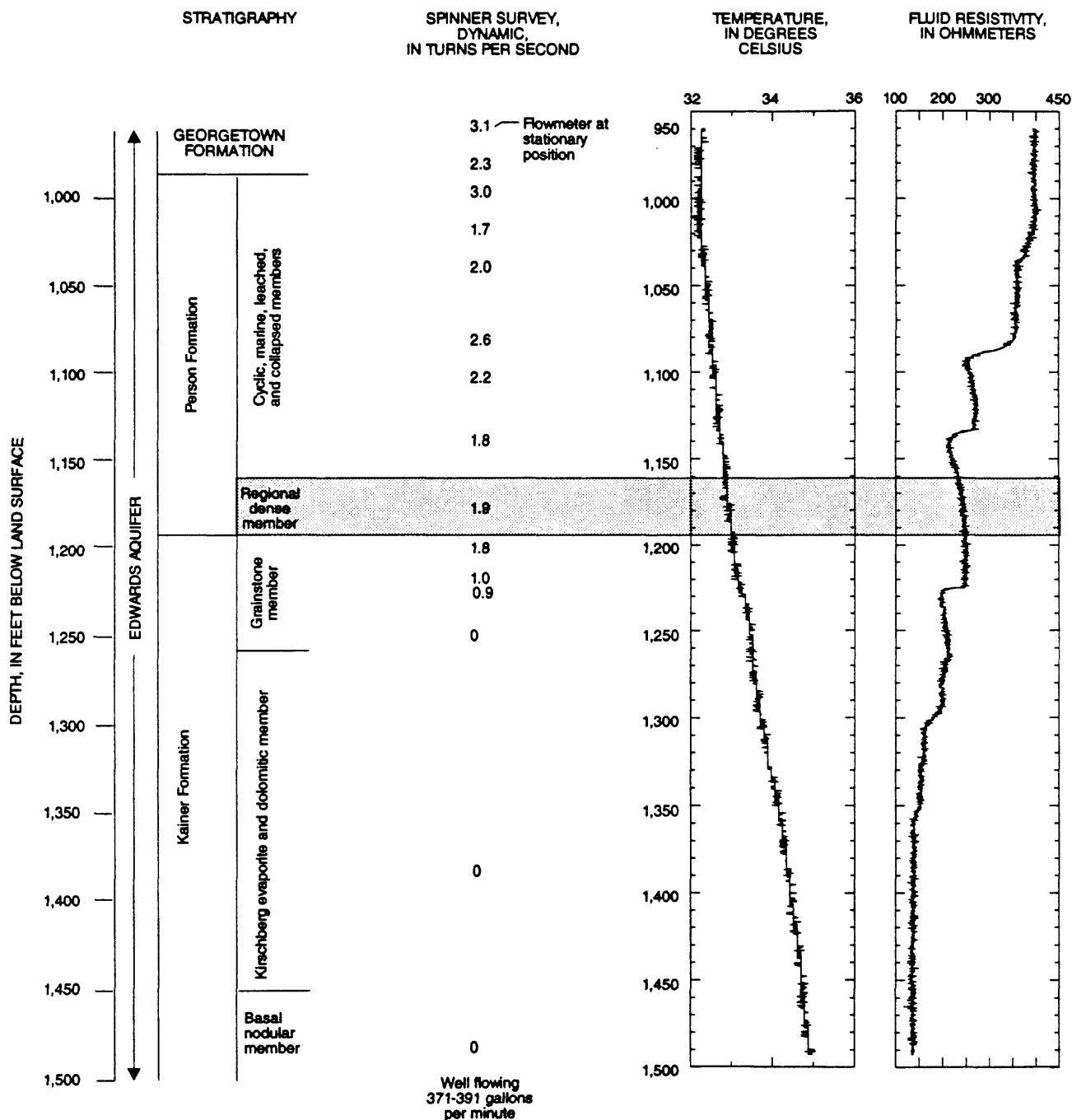
Figure 14. Spinner survey, temperature, and fluid resistivity logs of site A test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987.)

The neutron porosity log (fig. 13) is compensated for hole diameter. Therefore, the response of this tool can be interpreted directly as porosity where the test hole is not rugose based on a relation established between the tool response and a standard limestone with known porosity. Other minerals that affect the neutron response are dolomite and silica. The apparent limestone porosity represents the actual porosity only if the rock is pure limestone with no substantial amount of silica (either chert or sandstone) or dolomite. Substantial dolomite was present in the drill cuttings. For instance, a limestone porosity of 10 percent corresponds to a dolostone porosity of 7.5 percent. The log indicates a distinct difference in the nature of the porosity from above the RDM to below the RDM where the hole is rugose. A number of large spikes are apparent in the neutron porosity log of material above the RDM. Spikes are more numerous, but smaller in the porosity log of material below the RDM. The log shows slight porosity of material in the RDM.

The gamma-gamma (density) log was compensated for the diameter of the test hole by a single-arm caliper, but this cannot adequately compensate for strong hole rugosity or small caverns (MacCary, 1978). The compensated density log responds to the bulk density of a volume of rock and water around the tool. Less density indicates substantial secondary porosity filled with water—density about 1.0 g/cm^3. Some extensive secondary porosity filled with water is probable where the sections of test-hole diameter greater than 8 in. corresponded to spikes of lesser density on the density log. Pure limestone has a density of about 2.71 g/cm^3 and dolomite has a density of about 2.88 g/cm^3. The RDM has a bulk density of less than 2.71 g/cm^3 and only a few small layers have a density greater than 2.71 g/cm^3. Many of the spikes indicating lesser density in the log correspond to spikes in the neutron porosity log that indicate greater porosity, which further indicate water filled fractures and pores. The spikes in the gamma-gamma (density) log are subdued below the RDM as are the spikes in the neutron porosity log.

The acoustic velocity log (fig. 13) measures the speed of a sound wave transmitted through the rocks surrounding the tool. The velocity in rock is greater than the velocity in water, so relative velocity is indicative of the porosity and interconnections of the water filled pores and fractures in the rocks. The spikes showing long transit times just above and below the RDM indicate zones with substantial secondary porosity.

The spinner survey and the temperature and fluid resistivity logs (fig. 14) measure attributes of the fluid in the test hole. The spinner survey indicates that most of the flow from the open test hole was from above the RDM. The geophysical logs (fig. 13) indicate that this is caused by greater secondary porosity above the RDM. The uniform temperature with depth is indicative of the slow movement of the water in the saline-water zone. The slow movement of fluid in the saline-water zone allows the geothermal heat to flow up through the fluid and rocks uniformly. The fluid resistivity log indicates that the fluid usually is more saline with depth.

Site C

The logs for site C (the transition zone) are shown in figures 15 and 16. The hole diameter (fig. 15) indicates that this test hole was a more uniform in-gage hole than the site A test hole in the saline-water zone. Several spikes in the log indicate a hole wider than 8 in. (out of gage). At the top of the Person Formation, just below the Georgetown Formation, is a cavern about 8 ft deep (vertical measurement). The caliper did not give an indication of the true size of the cavern because the cavern extended more than several feet from the center of the test hole. Another relatively smooth spike is evident 20 to 30 ft below this cavern. This probably was caused by the widening of the test hole resulting from incompetent rocks or fractured rocks that fell into the hole. Again, the RDM appears to be a competent dense rock. The hole diameter shows slight openings of unknown origin in the test hole just below the RDM and midway between the RDM and the bottom of the hole.

The natural gamma log starts below the top of the Edwards aquifer. Major deflections in this log include one at the RDM and one near the bottom where an increase in gamma radiation was caused by the increased clay content of the basal nodular member.

The spontaneous potential log is ineffective in a test hole drilled without drilling mud. The only spike is at the RDM. The 16- and 64-in. resistivity logs have a major deflection at 10 to 20 ft below the cavern at the top of the Person Formation. The deflection might be related to the smooth spike in the hole diameter near this level, indicating a wider hole.

The neutron porosity log shows substantial porosity at the same level as the cavern and an apparent zero-porosity layer just below it. Effects of the
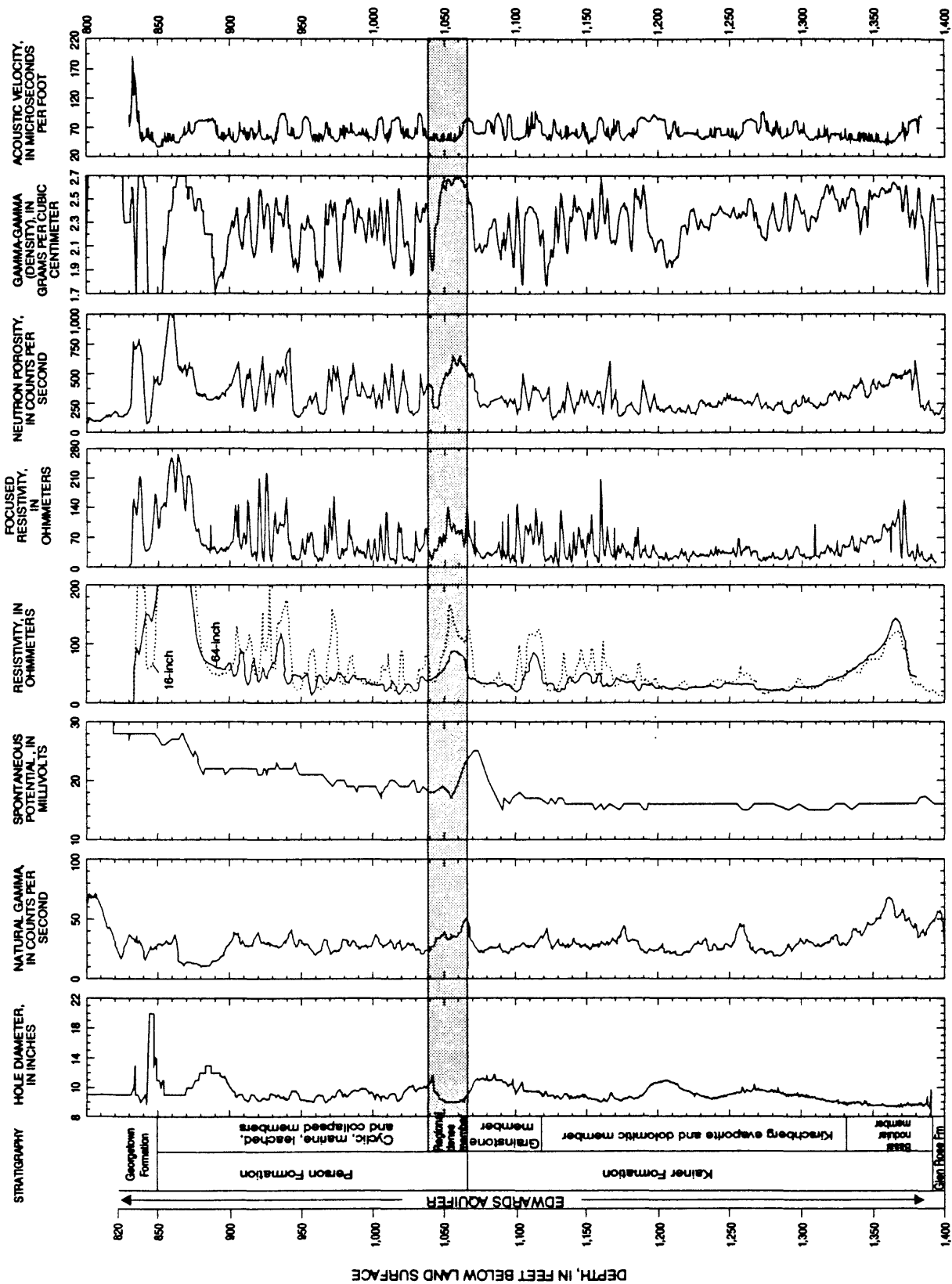
Figure 15. Geophysical logs of site C test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987.)
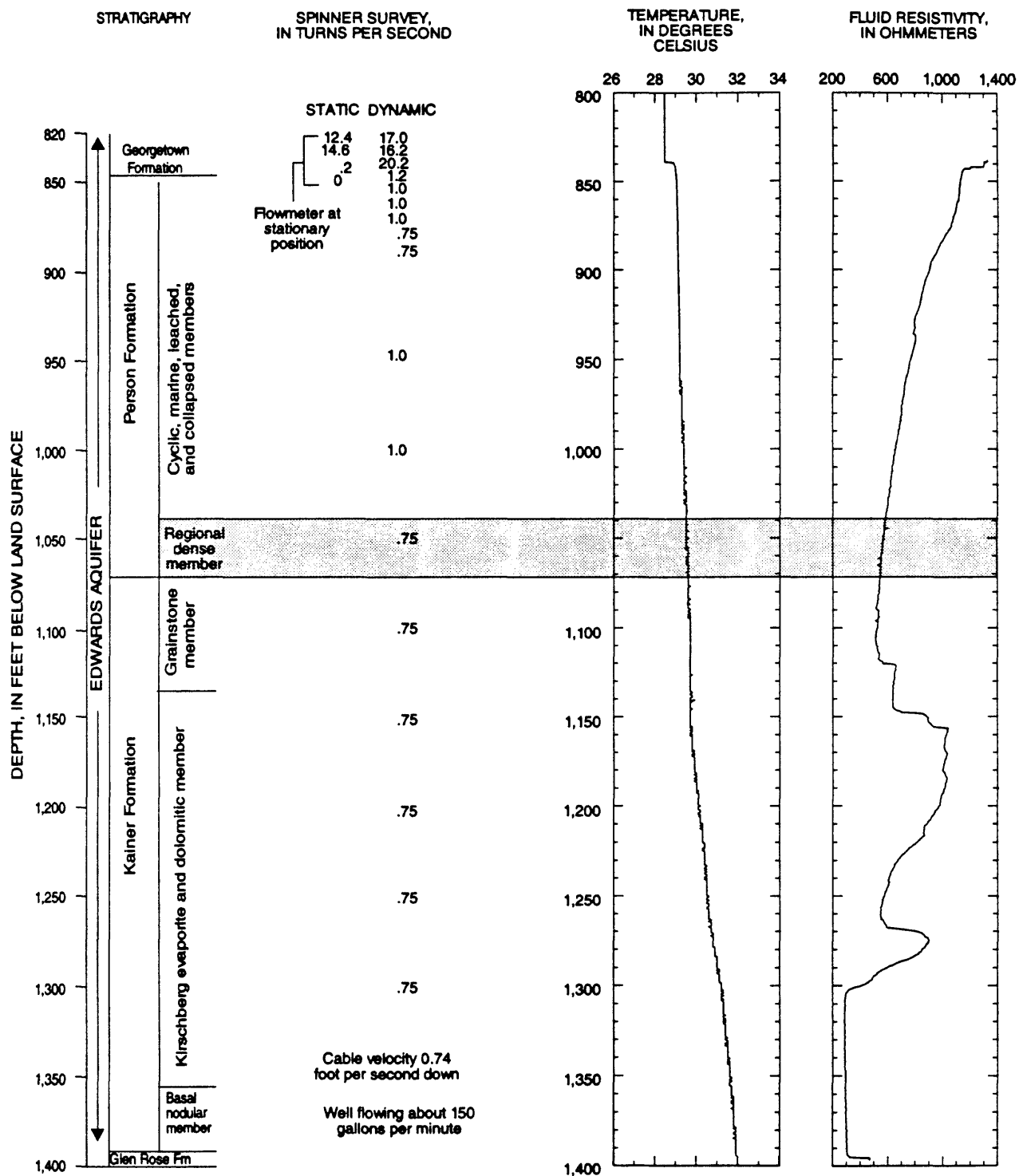
Figure 16. Spinner survey, temperature, and fluid resistivity logs of site C test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987.)

freshwater in the cavern at the top of the Person Formation are not known. Many large spikes in the neutron porosity log of the Person Formation correspond to porous layers. Examination of the drill cuttings revealed that a considerable amount of dolomite was present in this interval; therefore, the limestone porosity from the log appears to be greater than actual porosity. The porosity of the RDM is slight, as indicated by this log. The spikes in the neutron log are subdued in the Kainer Formation below the RDM. This could be caused by the less developed secondary porosity, such as that noted in the saline-water-zone rocks by Maclay and Small (1984). About midway down the Kainer Formation, the spikes in the neutron porosity log become noticeably smaller than those in the log of the upper half of the Kainer Formation. This indicates the level below which little to no alteration by freshwater has occurred.

The gamma-gamma (density) log starts above the Georgetown Formation and indicates that the formation is a dense carbonate rock with slight porosity. Maclay and Small (1984) stated that the Georgetown Formation is composed of limestone containing little or no dolomite. A large spike below the Georgetown Formation in the top of the Person Formation indicates the cavern at this level. The spike in the density log of the rock just below the cavern could be an indication of the same dense layer of slight porosity shown by the neutron log. This dense layer might prevent the freshwater in the cavern from moving down through the rocks below. The characteristics of the neutron and density logs from the site A test hole and from the site C test hole indicate that thin, porous, less dense layers alternate with thin, slightly porous, dense, layers through the Person Formation above the RDM.

The acoustic velocity log also indicates stratification of porous zones. The major spikes on this log are at the cavern near the top of the Person Formation and 20 to 40 ft above the RDM. Large spikes are not apparent below the RDM.

A static spinner survey (fig. 16) was conducted while the test hole was shut-in (no flow was allowed to come out of the casing at the surface), and a dynamic spinner survey was conducted while water was allowed to flow at the surface. On the basis of static spinner results, flow could have continued in the test hole even when flow was not allowed at the surface. The limited data were not corrected for hole diameter variations and do not indicate where the water was flowing. Water could have been flowing out of the cavern into the test

hole and then down the test hole into the saline-water zone below the cavern. The dynamic spinner survey was run as the tool was lowered down into the test hole at 0.74 ft/s. Because the tool was moving down the test hole, it was difficult to determine if there was flow downward in the test hole while the water was flowing at the surface. The constant 0.74 turns per second of the spinner tool as it was lowered below the RDM could indicate no-flow conditions in the test hole below the RDM.

The curve of the temperature log (fig. 16) with depth indicates that the saline water at depth was substantially warmer than the freshwater near the top. Unlike the smooth shape and slight slope of the temperature log for the site A test hole, the temperature log of the site C test hole is marked by breaks in slope and is steeper from top to bottom. This is caused by the greater velocity of the freshwater in the cavern near the top relative to the velocity of the saline water. The fast flowing water is able to absorb and rapidly transfer geothermal heat; therefore, the temperature of the freshwater was less than would be expected if the water was stagnant.

The stratification of salinity shown by the fluid resistivity log (fig. 16) could be related to this stratification of porosity. The flow tests to determine relative permeabilities also indicate that rock stratification affects flow in these rocks. Because of the drilling methods used and the structure of these deep test holes, detailed vertical profiles of water chemistry and rock stratification were not obtained. Maclay and Small (1984) indicated the presence of considerable permeability stratification in the cores of the Edwards aquifer rocks they studied from the saline-water and freshwater zones. The unexpected increase in resistivity of the fluid shown in the fluid resistivity log in the Kainer Formation could have been caused by fresher water flowing down into the aquifer in the lower part of the open hole.

Site D

The geophysical logs of the freshwater-zone test hole (site D) are shown in figure 17. The hole diameter has a distinctly different shape from the diameters of the other two test holes. The scale on the site A and site C test holes is 6 to 14 in.; the scale on the site D test hole is 6 to 42 in. to accommodate the larger diameter of this test hole. At least one of the spikes near the top of the Person Formation exceeded 42 in. The hole
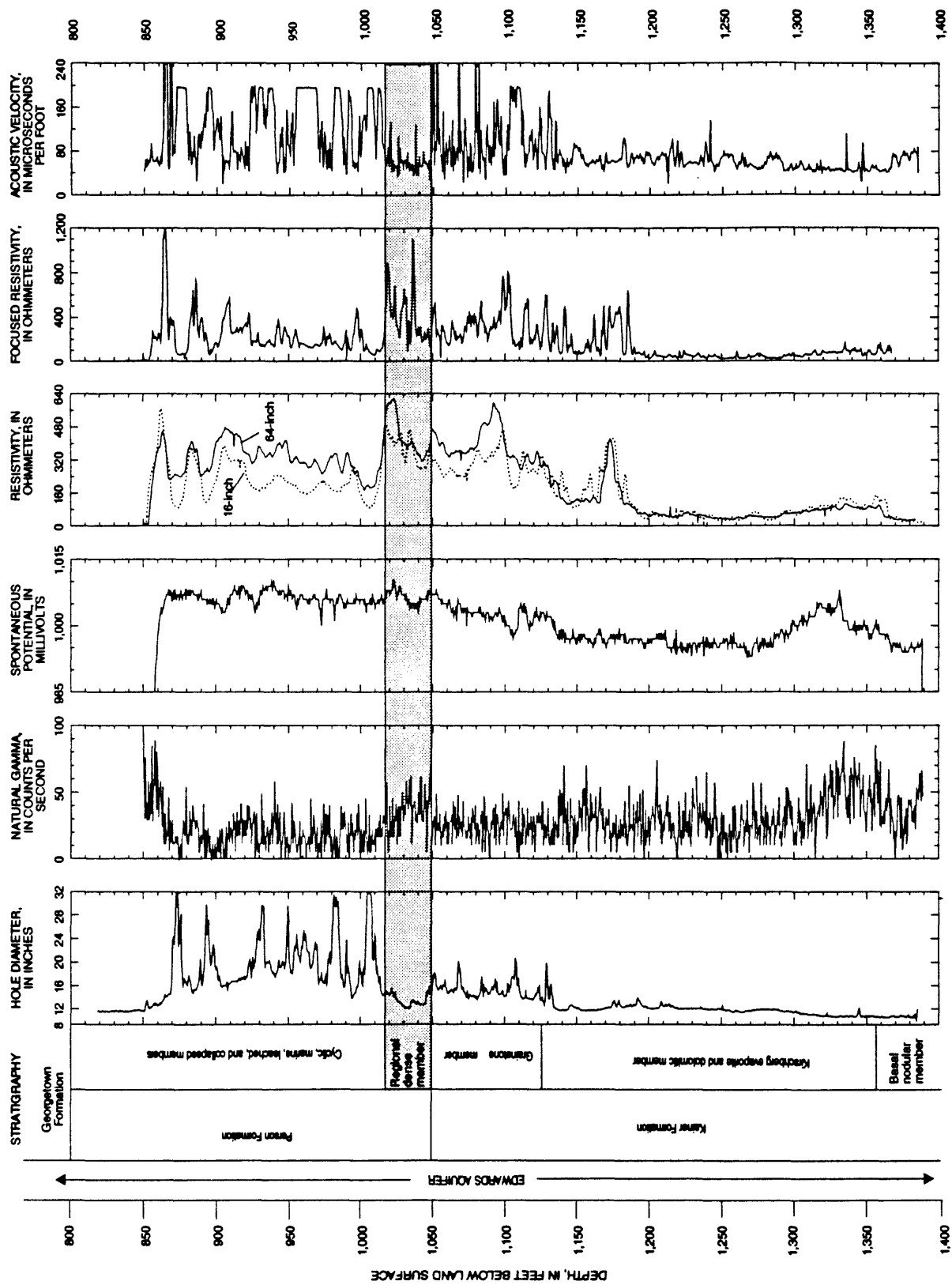
Figure 17. Geophysical logs of site D test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987.)

diameter indicates that the entire section of the Person Formation, excluding the RDM, is incompetent and broken. Much of the matrix rock in this section probably fell into the test hole and was pulverized by the drill bit. The section of the test hole that had large broken or fractured sections extends below the RDM about the same distance as the porous and less dense zones extend below the RDM in the site C test hole. Below this point, the site D hole diameter shows a much more competent rock that is closer to the size of the drill bit (in gage), although still much wider than the other test holes. This widening at the bottom could be the result of the instability of the drill stem caused by the caverns created as the drill bored through the incompetent and broken sections in the upper Kainer and Person Formations.

The fluctuations in the natural gamma log indicate the changes in test-hole diameter, except in the Georgetown Formation, the RDM of the Person Formation, and near the bottom of the Kainer Formation in the basal nodular member. In the RDM, small spikes appear in the natural gamma log, which could be the result of wide sections of test hole. This is confirmed by the large spike in the acoustic velocity log in the RDM.

The spontaneous potential log shows nothing substantial. The 16- and 64-in. resistivity logs exhibit a number of spikes, but these do not appear to be related to any noticeable feature of the rocks. Neutron porosity and density logs were not run in this test hole. The fractures and broken nature of the rocks of the Person Formation and the top of the Kainer Formation indicated that the rocks were unstable and could collapse on top of the tool, causing possible loss of a radioactive source in the freshwater zone of the aquifer.

The acoustic velocity log indicates that so little competent rock surrounded the tool that only water transmitted the sound wave from the transmitter to the receiver. This resulted in a condition whereby the signal received by the transmitter was received later than is normally expected for competent rock surrounding the tool; thus, the cycle of transmitted signal being picked up by the receiver was offset by one or more cycles. This phenomenon is called cycle skipping (MacCary, 1978). A cycle skip was evident in the RDM; even this otherwise competent layer has been extensively fractured or dissolved. The cycle skipping stopped at the same level in the Kainer Formation where the hole diameter corresponds to more in-gage conditions of the test-hole diameter, and where the neutron and gamma-gamma logs of the transition-zone

(site C) test hole show distinct changes. The logs indicate that the rock below this level was more intact than the rock above it.

Static and dynamic spinner surveys were made in the site D test hole (fig. 18). Three static measurements were made at depths of 847.5 ft (17.5 turns per second), 860 ft (5 turns per second), and 870 ft (5 turns per second), when no flow was allowed at land surface. These data indicate that more flow occurred at a depth of 847.5 ft than at the other two depths. Some of the increase in flow rate might have been caused by a slightly smaller diameter of the test hole at 847.5 ft. Flow occurred in this test hole even when flow was stopped at the surface, as it did in the test hole at site C. The dynamic spinner survey was run while water was allowed to flow at the surface. The tool was lowered down the test hole at about 1.8 ft/s. The data were not corrected for test-hole diameter. Because the log was not run in the opposite direction, data from this dynamic spinner survey were difficult to interpret. The smallest number of turns per second could indicate either that the tool was moving through still water, or water was moving slowly past the moving tool. The increases in the readings correspond to an increase in the velocity of water moving up past the downward moving tool. Increases in the upward movement of water are indicated near the top of the Person Formation, 40 to 50 ft above and below the RDM. Most of the flow from the test hole came from the top 100 ft of the Person Formation.

A slight increase in the fluid temperature (fig. 18) is evident in the Kainer Formation where the cycle skipping of the acoustic velocity log ceased. The fluid resistivity log recorded an increase in salinity in this part of the Kainer Formation.

A number of similarities are apparent between the logs of the sites C and D test holes. Both suites of logs indicated that the most porous and permeable sections of these test holes were above the middle of the Kainer Formation. The Person Formation appeared to be leached and porous in both test holes. The greatest difference was that the freshwater and most of the flow of the site C test hole was confined to the cavern near the top of the Person Formation. Flow in the freshwater section was more evenly distributed over the leached and porous zone of the site D test hole than in the site C test hole. Another difference between the sites C and D test holes is that, on the basis of texture and fracturing of the rocks at the freshwater zone, the site D test hole was drilled within a fault zone. A fault zone where
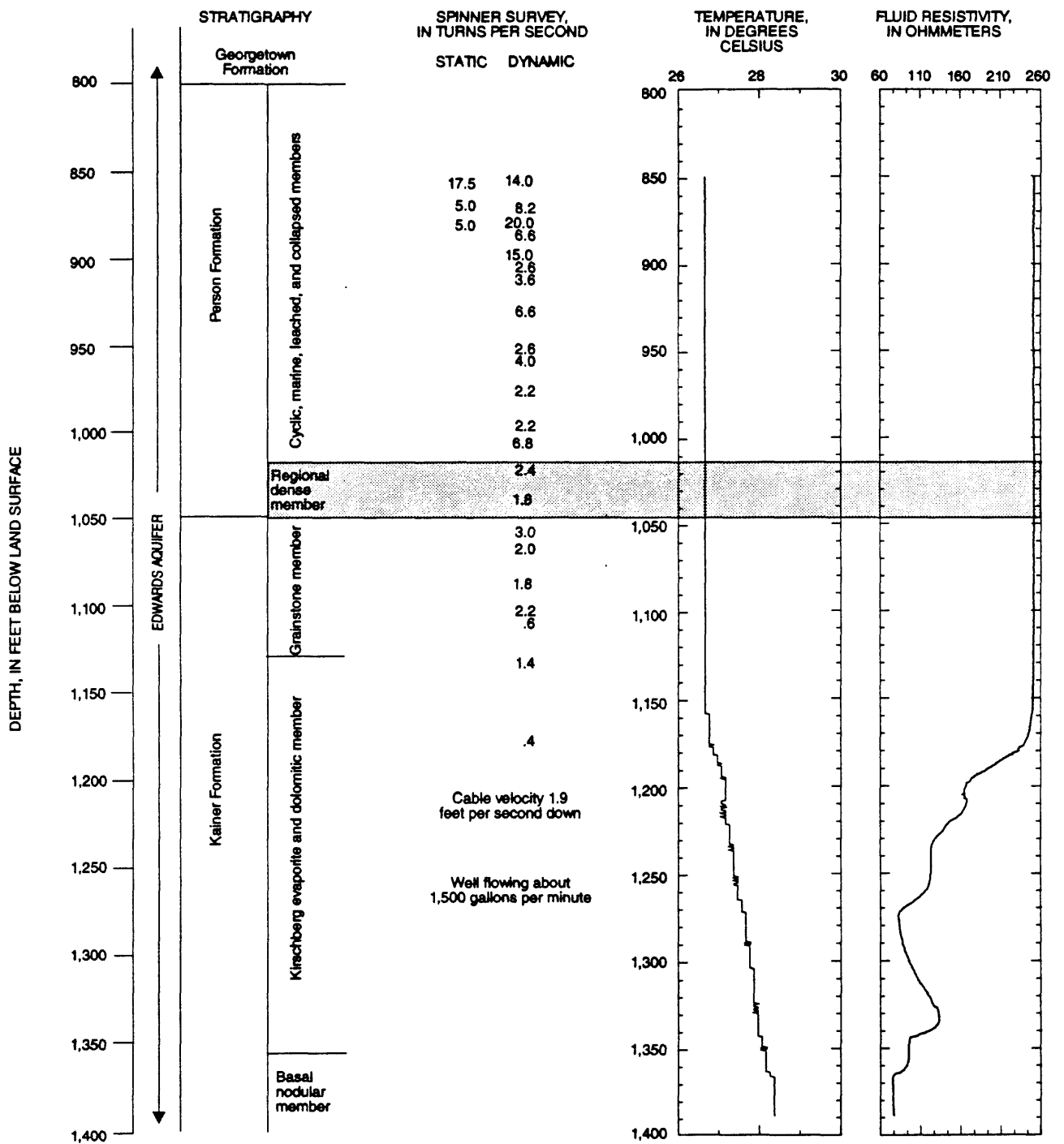
Figure 18. Spinner survey, temperature, and fluid resistivity logs of site D test hole, San Antonio, Texas. (Modified from Pavlicek and others, 1987.)

rocks are extensively broken up would be more permeable to freshwater infiltration and the associated alteration of the rocks as described by Mench-Ellis (1985).

Hydrologic Data

Hydrologic analysis of ground water commonly is based on observed fluctuations in water levels. These fluctuations could be caused naturally by changes in recharge and discharge or caused artificially by pumping water from wells. In this section, the following topics are discussed: (1) water-level fluctuations caused by allowing certain sections of the open drill hole to flow before installing well casing; (2) water-level fluctuations caused by changes in natural recharge and discharge and pumping (drawdown tests); and (3) water-level fluctuations caused by artificially imposed water withdrawals in completed wells.

Flow Tests Before Installing Well Casing

During drilling, and before installing well casing, tests of the flow of water from certain sections of each test hole were possible because the static hydraulic head was above land surface. In the following discussion of these tests, the terms cumulative-depth flow tests or cumulative-depth tests refer to the process of stopping the drilling after approximately each 50 ft, allowing the water to flow around the drill stem and other equipment in the test hole for 1 hour, and then measuring the rate of discharge from the test hole. The water level was measured before flow began and after an apparent equilibrium of flow was established. Open-hole tests were made after all the drilling equipment was withdrawn from the completed test hole and the water was again allowed to flow. Single-packer tests were made after the open-hole test. Equipment was lowered into the test hole to hydraulically divide the test hole into vertical sections by an inflatable packer. The packer was set at predetermined depths and inflated to press against what was assumed to be unbroken and solid low-porosity rock. Flow from above and below the packer was measured along with the water levels before flowing and after equilibrium flow was established. Flow was allowed from only one isolated section at a time. The data from the cumulative-depth flow tests at test holes A1, C1, and D1, plotted in figure 19, are modified from Pavlicek and others (1987).

William F. Guyton Associates, Inc. (1986) made a number of assumptions about the nonoptimal condi-

tions of the cumulative-depth flow tests to estimate transmissivities of each section of each hole tested. All transmissivities discussed below are from William F. Guyton Associates, Inc. (1986, p. 25-27). In the site A test hole, the transmissivity began at 5,030 ft^2/d in the top 100 ft and, with few exceptions, increased to 12,400 ft^2/d. The transmissivity from the open-hole test at A1 was 11,600 ft^2/d. Transmissivity in the secondary test hole at site A in which well A3 was completed began at 8.0 ft^2/d for the top 50 ft and increased to 2,400 ft^2/d for the open 200 ft.

At the site C test hole, the transmissivity was 42,400 ft^2/d for the top 50 ft and generally decreased with increasing cumulative depth of the hole. Slight increases were observed at certain levels in the wells; the decrease probably resulted from the poor conditions for measuring transmissivity, primarily for measuring discharge. The transmissivity from the open-hole test was 39,600 ft^2/d, which is similar to the largest value of the cumulative-depth flow tests. Almost all flow from the open test hole at site C was from the top 50-ft section.

The flow tests using a single packer to divide the site C test hole into two sections resulted in transmissivity for the top section of the test hole that was greater than that from cumulative-depth flow tests for the top section and much greater than the open test-hole transmissivity. These discrepancies resulted primarily from nonoptimal conditions for measuring aquifer-test data, but indicate that downward flow occurred in the test hole during the cumulative-depth and open-hole flow tests.

The single-packer or interval flow tests indicated that the transmissivity for most of the lower 500 ft of the site C test hole was 50 to 180 times less than the transmissivity for the top 39 to 100 ft (William F. Guyton Associates, Inc., 1986, p. 25-27). The greater observed transmissivity was caused primarily by the cavern located in the top 8 ft of the Edwards aquifer below the Georgetown Formation. It is possible that water from the cavern at the top of the aquifer in the transition zone was flowing down the open test hole into the lower section of the aquifer through the open hole. This flow could explain the discrepancy between the test results, assuming that the differences in the flow rates were not caused by measurement error.

A cumulative-depth flow test in the site C secondary test hole (in which well C2 was completed) yielded a small transmissivity. The transmissivity was 52 ft^2/d in the top 50-ft section (the same depth interval
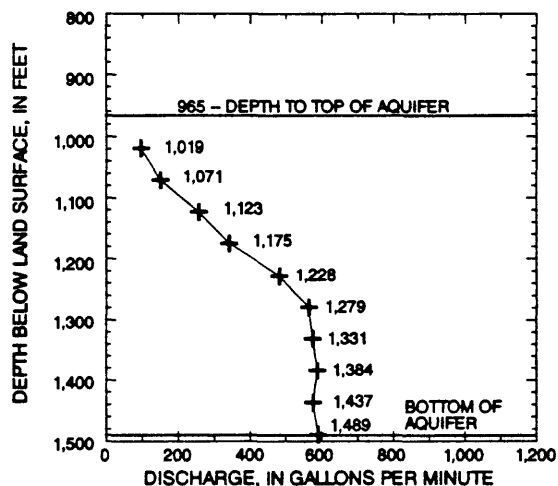
EXPLANATION

+ 941 DISCHARGE AND BOTTOM DEPTH
OF INTERVAL--Interval is from
top of aquifer to indicated depth.
Data are plotted at bottom of
interval

a. Site A test hole
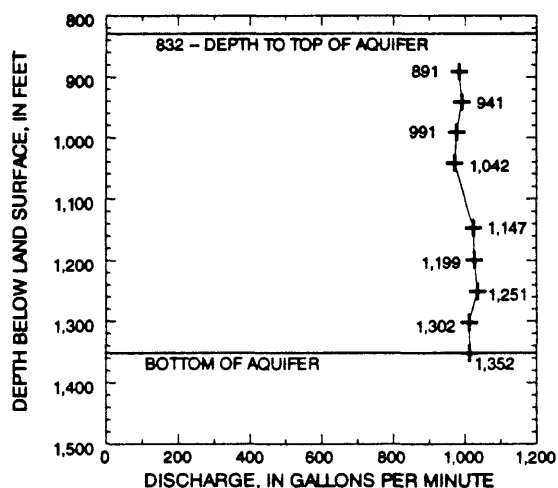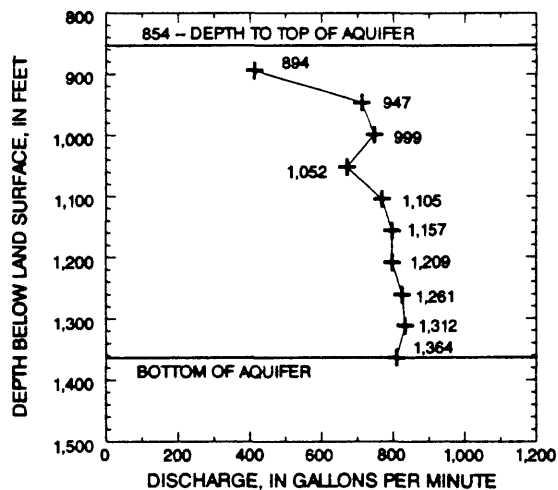
b. Site C test hole

c. Site D test hole

Figure 19. Discharge from test holes at sites A, C, and D, San Antonio, Texas, during cumulative-depth flow tests. (Modified from Pavlicek and others, 1987.)

where the cavern was encountered in primary site C test hole) and 48 ft²/d in the top 100-ft section. The transmissivity for the open-hole flow test at the secondary site C test hole was 694 ft²/d. The secondary test hole at site C apparently was in a zone that has a small transmissivity relative to the zone where the primary site C test hole was.

Because of difficulties in obtaining the necessary data, only one of the cumulative-depth flow tests at the site D test hole produced an estimate for transmissivity. The transmissivity was estimated as 72,300 ft²/d in the freshwater zone from a depth of 854 to 1,052 ft. The transmissivity from the open-hole test was estimated at 79,000 ft²/d. Single-packer tests indicated that most of the transmissivity (almost 95 percent) was in the section of the test hole above a depth of 1,040 ft. Depth to the RDM at this site is between 1,030 and 1,045 ft below land surface.

Drawdown Tests

Several drawdown tests were conducted before and after monitor wells were completed in the secondary test holes at sites C and D and in wells Artesia 1 and Artesia 3. Artesia 1 was an observation well that was open above the RDM at the depth interval 863 to 977 ft below land surface. Artesia 3 is a water-supply well that is open to most of the Edwards aquifer (860 to 1,108 ft depth). The drawdown tests were conducted to compare the response of the monitor-well water levels to drawdown resulting from allowing the primary test hole or completed well to flow. The data were published in Pavlicek and others (1987). The effect of drawdown on water levels in monitor-wells C1 and C2 and in observation well Artesia 1 while water flowed from the site C open hole is shown in figure 20a. The response of well C2 to drawdown in the C1 open test hole was slight. Over the period tested, the drawdown in the test hole was about 30 ft and the response of the well C2 water level was between 2 and 3 ft. The water level of the Artesia 1 well showed no discernible response to the flow from the site C test hole.

After the monitor well was completed in the site C test hole (well C1), a similar test was conducted. The water-level drawdown (fig. 20b) in the completed monitor-well C1 was greater by about 4 to 5 ft than was drawdown in the open test hole (fig. 20a). The response of the water level in well C2 was negligible. The water level in the Artesia 1 observation well did not respond to the flow of the completed well at C1.

The slight drawdown in well C2 caused by flow at site C open test hole, and the lack of response of C2 when the completed monitor-well C1 flowed, indicated vertical anisotropy in that part of the aquifer. The interval of test hole C1 that was completed as a monitor well (fig. 3) was above the RDM. When only this interval was allowed to flow at land surface, well C2, with its monitor interval below the RDM, did not produce a measurable response.

The drawdown test conducted at the site D test hole (fig. 20c) was equally inconclusive. The drawdown in the open test hole (depth interval 854 to 1,384 ft) was about 14 ft. The water levels in Artesia 1 and Artesia 3 did not respond to the flowing of the site D test hole. The lack of response of the two Artesia wells indicates horizontal anisotropy at the site D/Artesia well-field area. During the test, the site D test hole was open to the entire thickness of the Edwards aquifer. Related studies indicated that the site D test hole was in or near a fault zone, and that Artesia 1 probably was located outside this fault zone (P.L. Rettman, U.S. Geological Survey retired, oral commun., 1987). This could account for the difference in response to the water levels, but the results are inconclusive.

Water-Level Fluctuations in Completed Wells

Water levels were observed in the completed monitor wells from July 1986-May 1987 to interpret the effects of variations in precipitation and artificially induced discharge changes on the hydrology of the aquifer at the downdip limit of freshwater. The water levels of the monitor wells and corresponding precipitation amounts are shown in figures 21a-i. Water levels are expressed as absolute hydraulic head (altitude above sea level) in these figures, rather than as equivalent freshwater head, because equivalent freshwater head can lead to inappropriate interpretations of flow directions (Davies, 1987).

The relation among the wells at the various depths was generally consistent between site A and site C. The shallowest wells usually had the highest hydraulic heads and the deepest wells had the lowest heads. At site A, the differences in head were caused primarily by increased salinity (density) between the various monitor intervals and by the difference in the length of the water column in each well. The difference between the head in the shallow freshwater well (D2) and the deep saline-water well in the transition zone (C2) was frequently greater than could be accounted
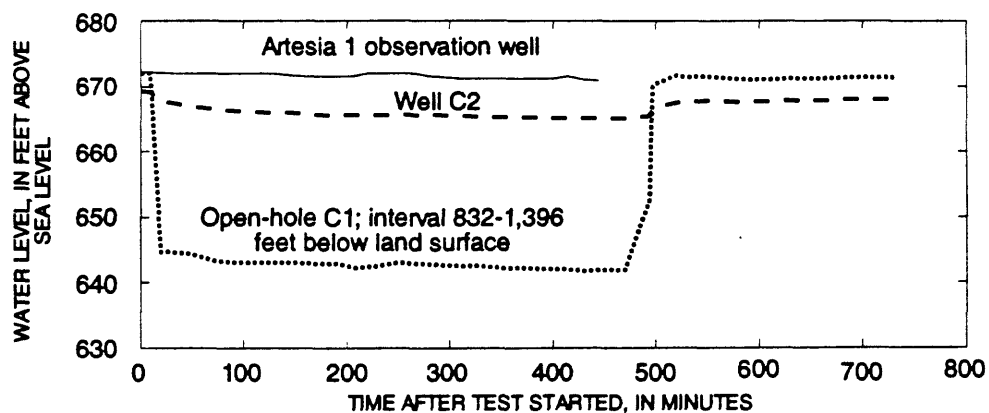
a. Transition-zone open-hole test
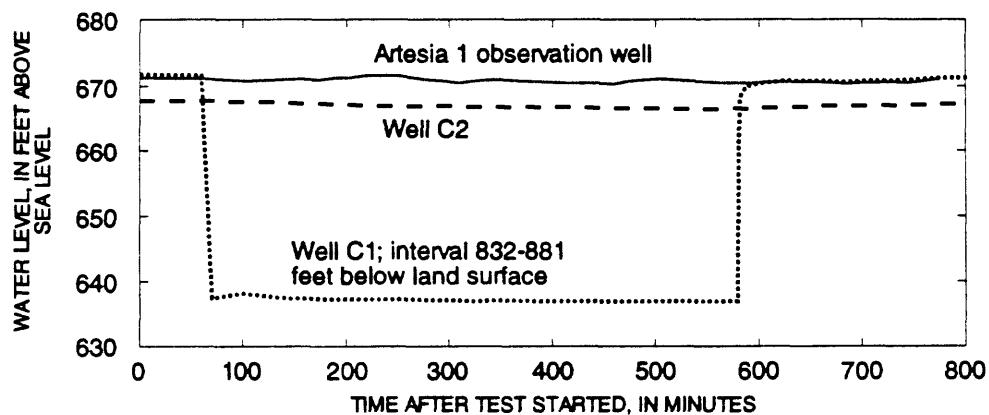
b. Transition-zone monitor-well (C1) test
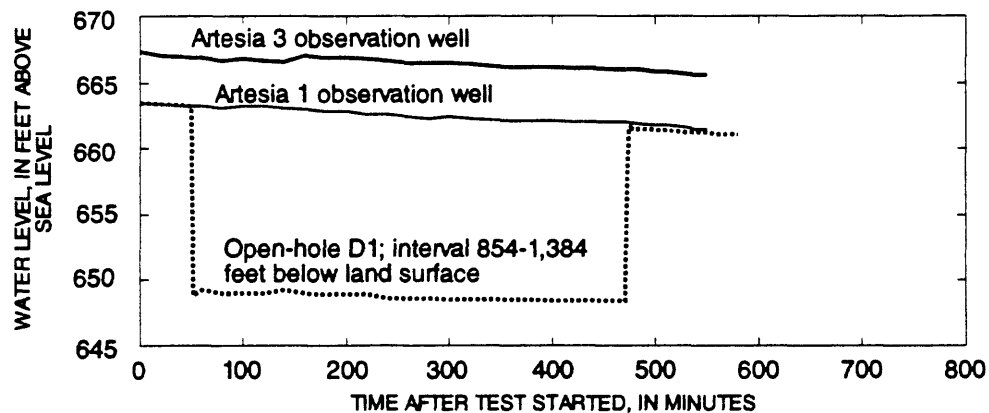
c. Freshwater-zone open-hole test

Figure 20. Water levels in test holes and monitor wells at sites C and D and observation wells Artesia 1 and 3, San Antonio, Texas, during drawdown tests.
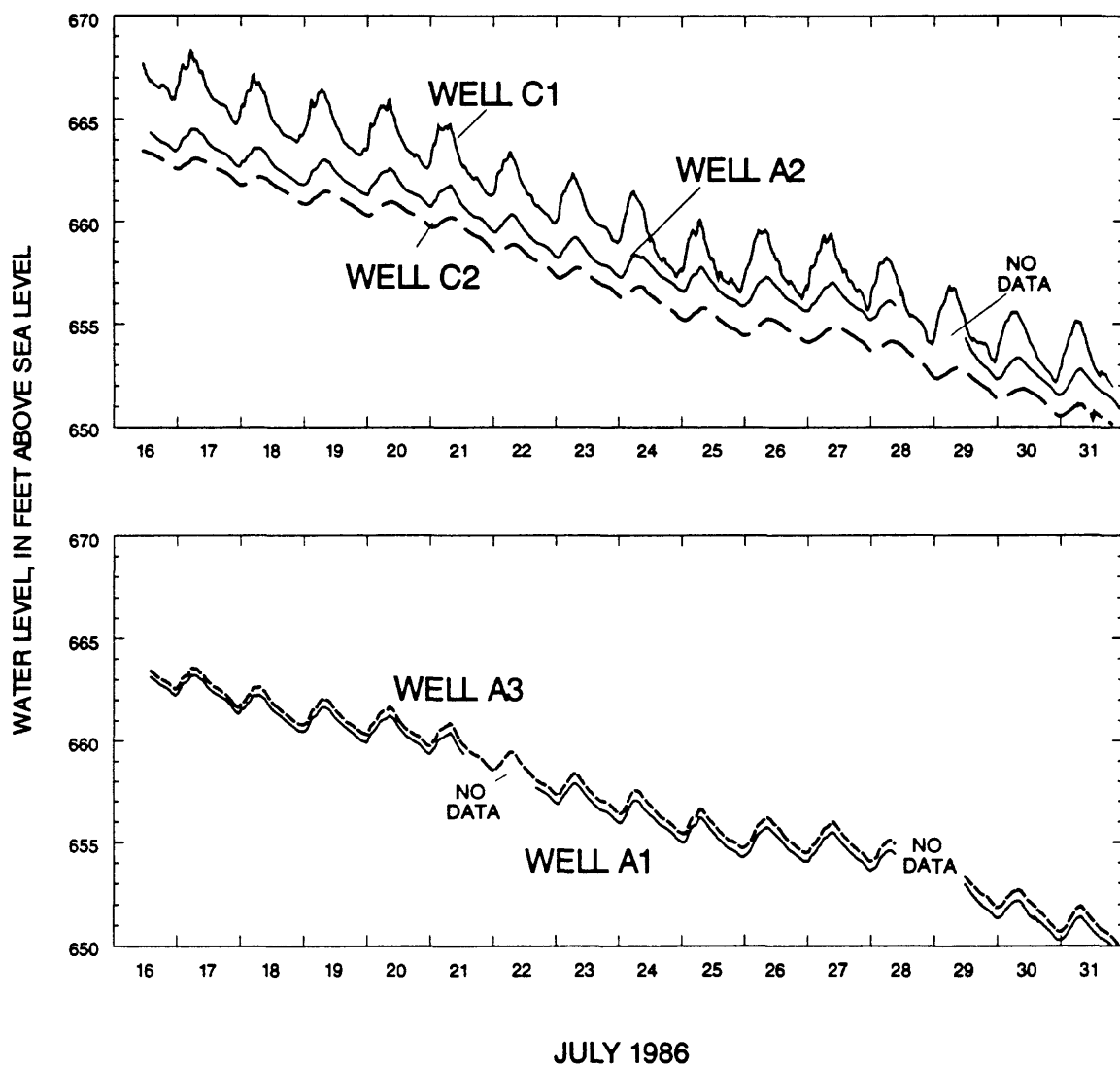
Figure 21a. Water levels in monitor wells A1, A2, A3, C1, and C2 and precipitation, San Antonio, Texas, July 16-31, 1986.
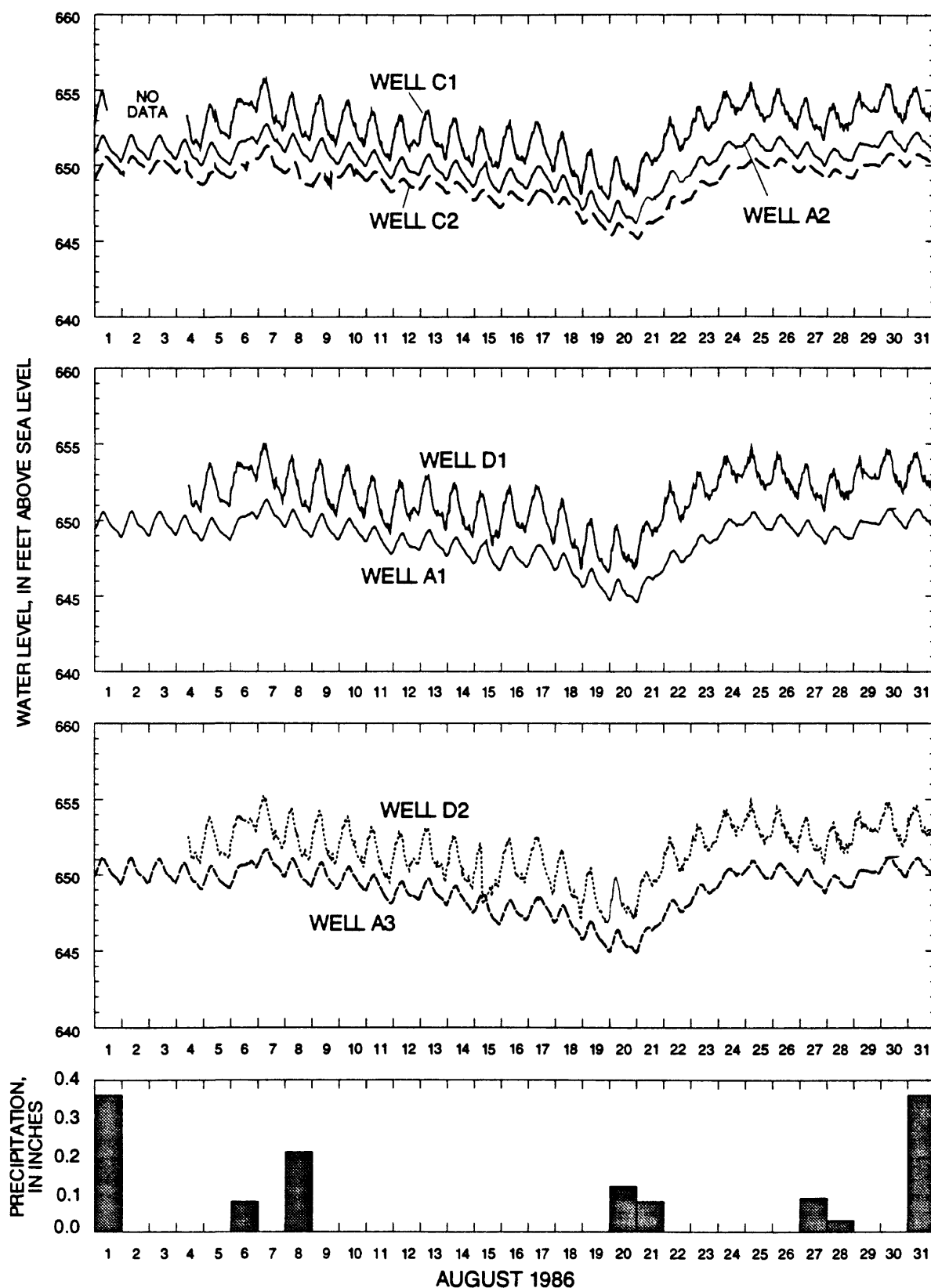
Figure 21b. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2 and precipitation, San Antonio, Texas, August 1986.
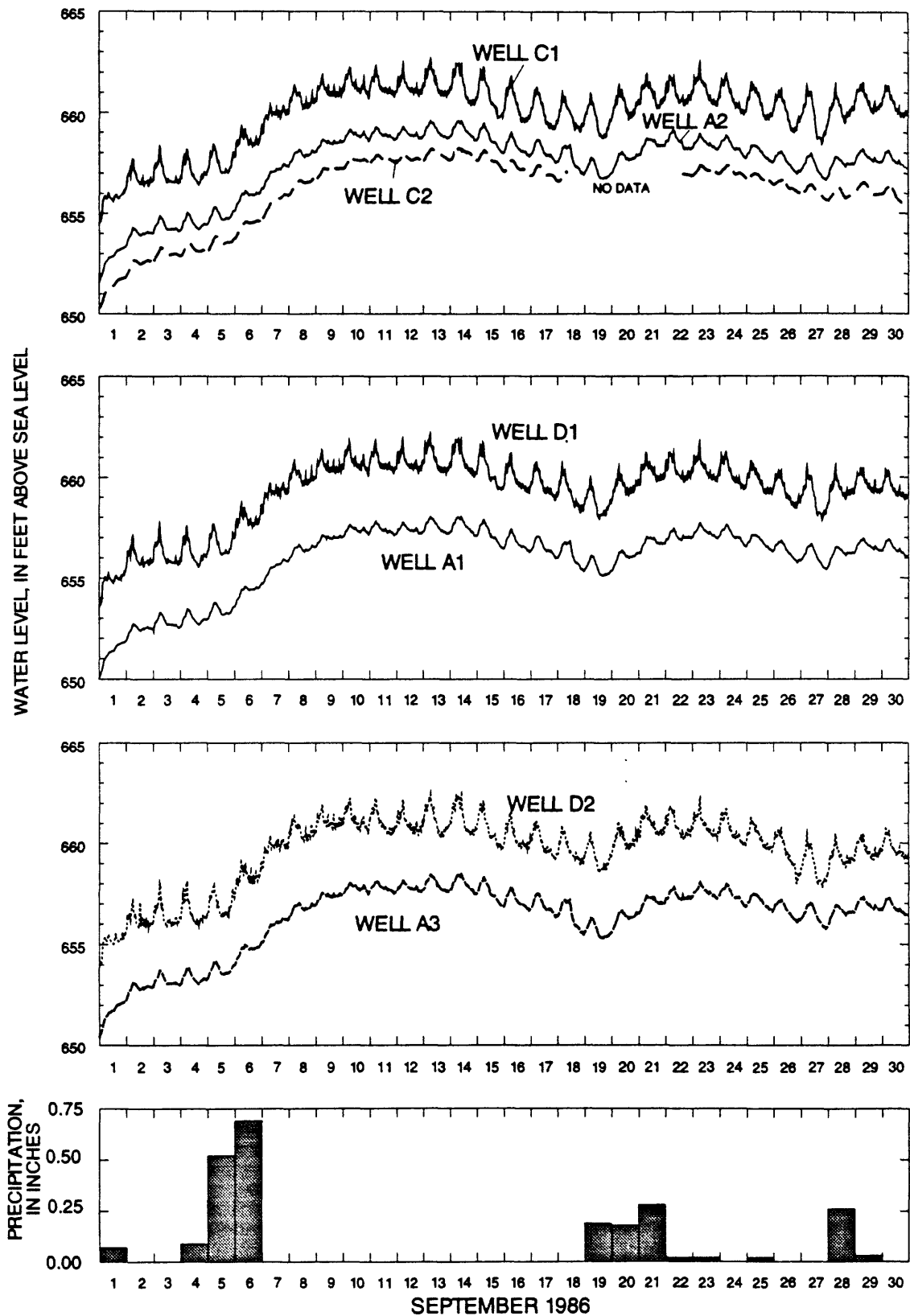
Figure 21c. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2 and precipitation, San Antonio, Texas, September 1986.

Analysis of Data from Test-Well Sites Along the Downdip Limit of Freshwater in the Edwards Aquifer, San Antonio, Texas, 1985-87

Figure 21d. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2 and precipitation, San Antonio, Texas, October 1986.
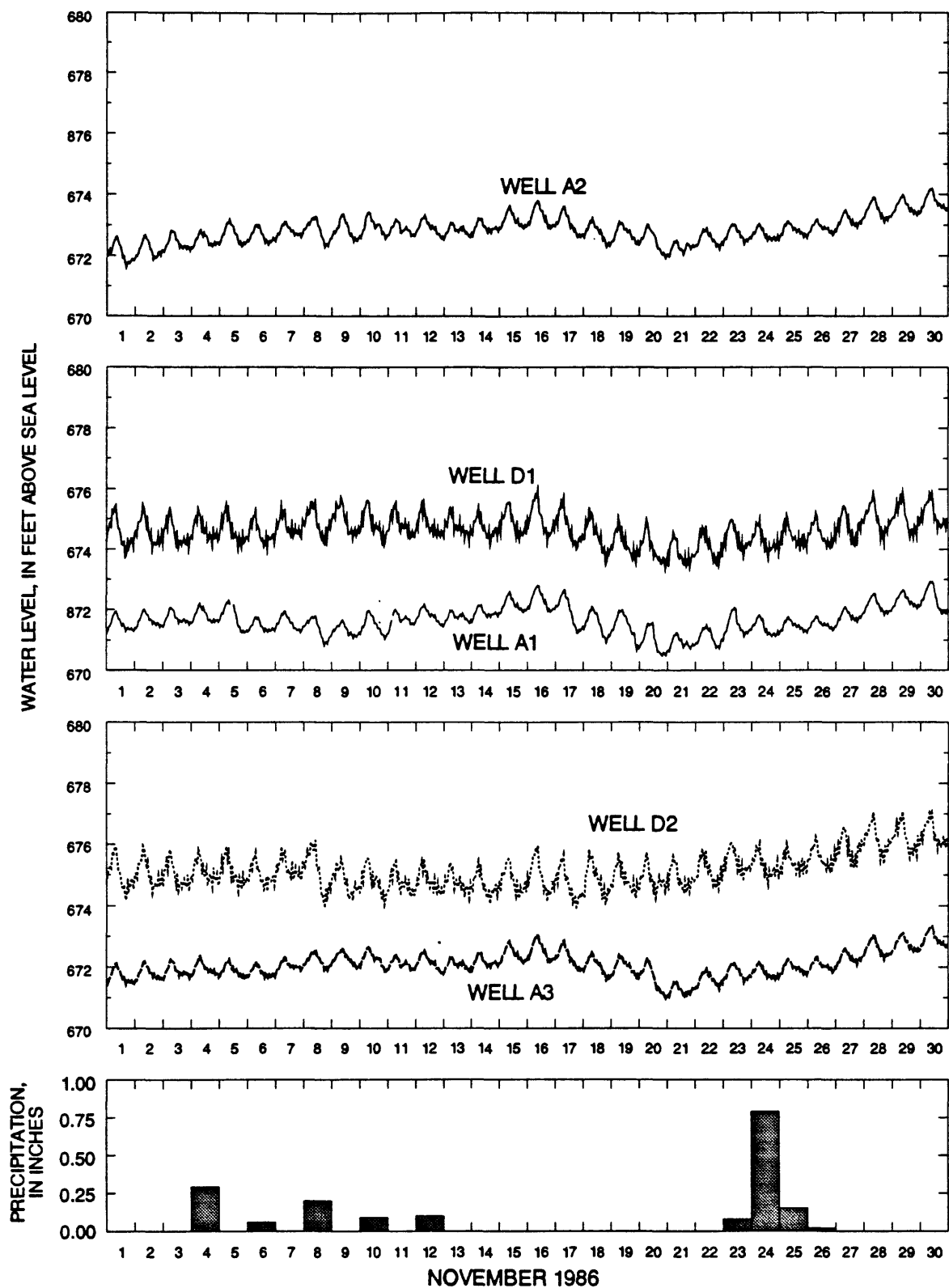
Figure 21e. Water levels in monitor wells A1, A2, A3, D1, and D2 and precipitation, San Antonio, Texas, November 1986.
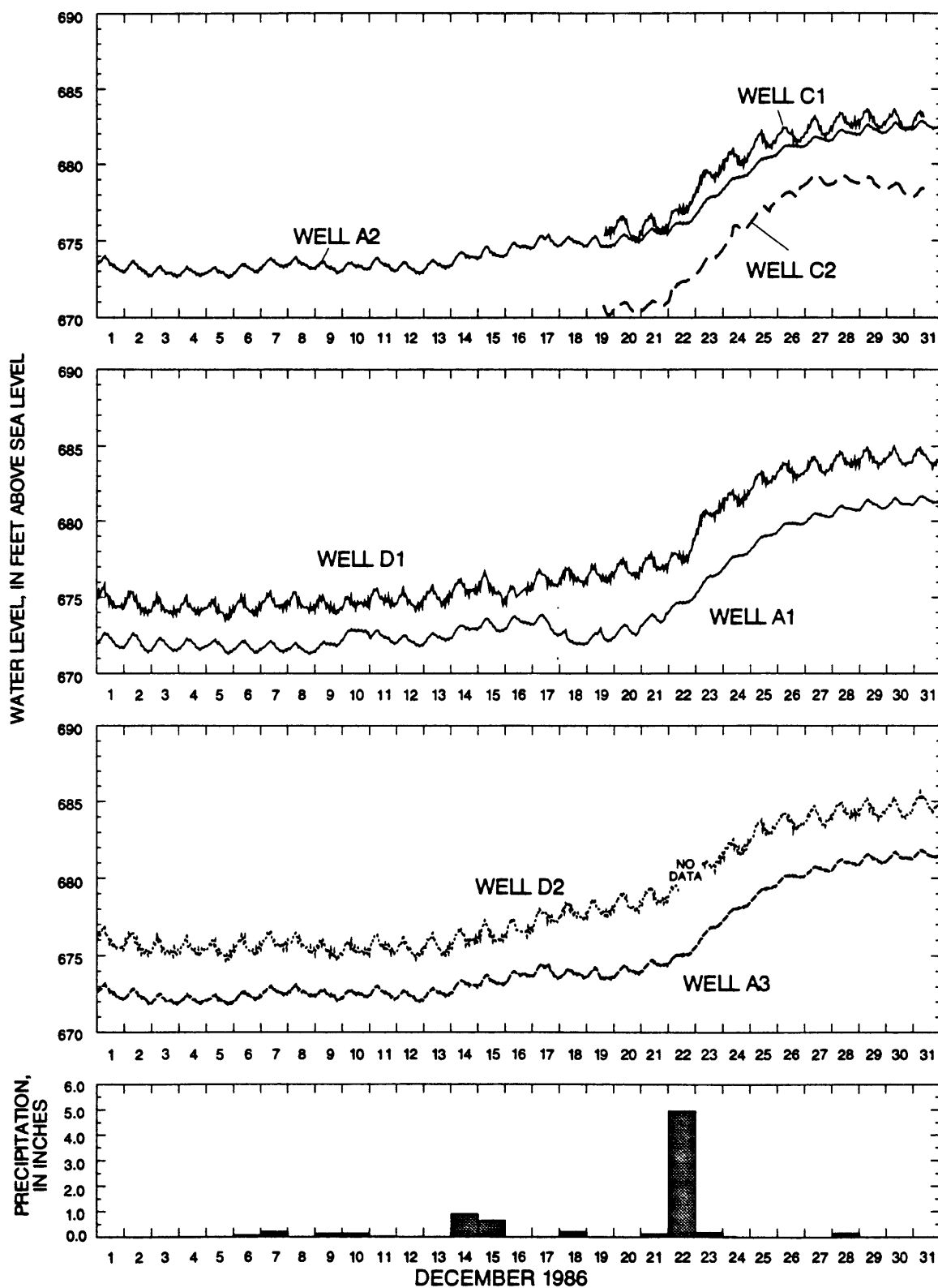
Figure 21f. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2 and precipitation, San Antonio, Texas, December 1986.
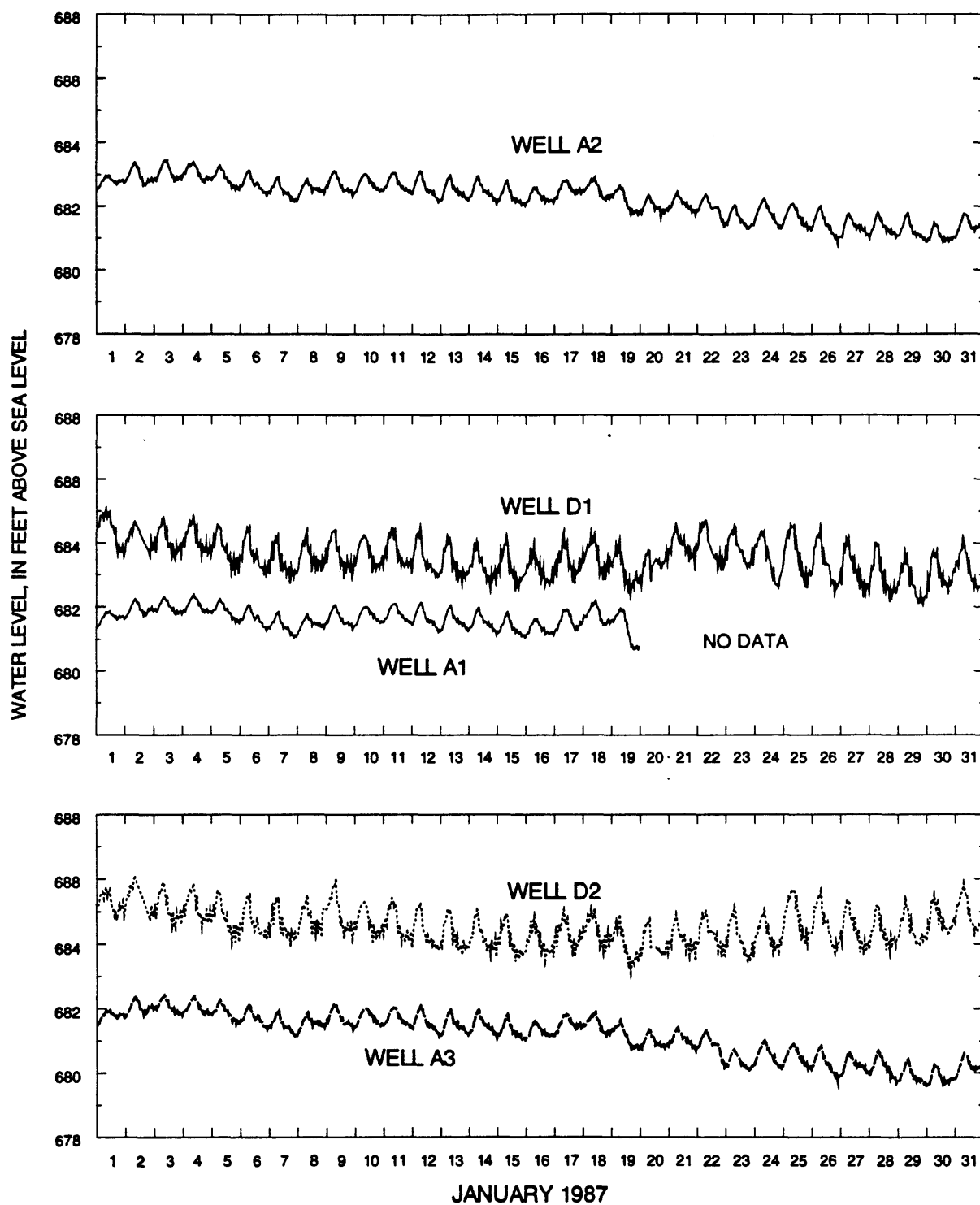
Figure 21g. Water levels in monitor wells A1, A2, A3, D1, and D2, San Antonio, Texas, January 1987.
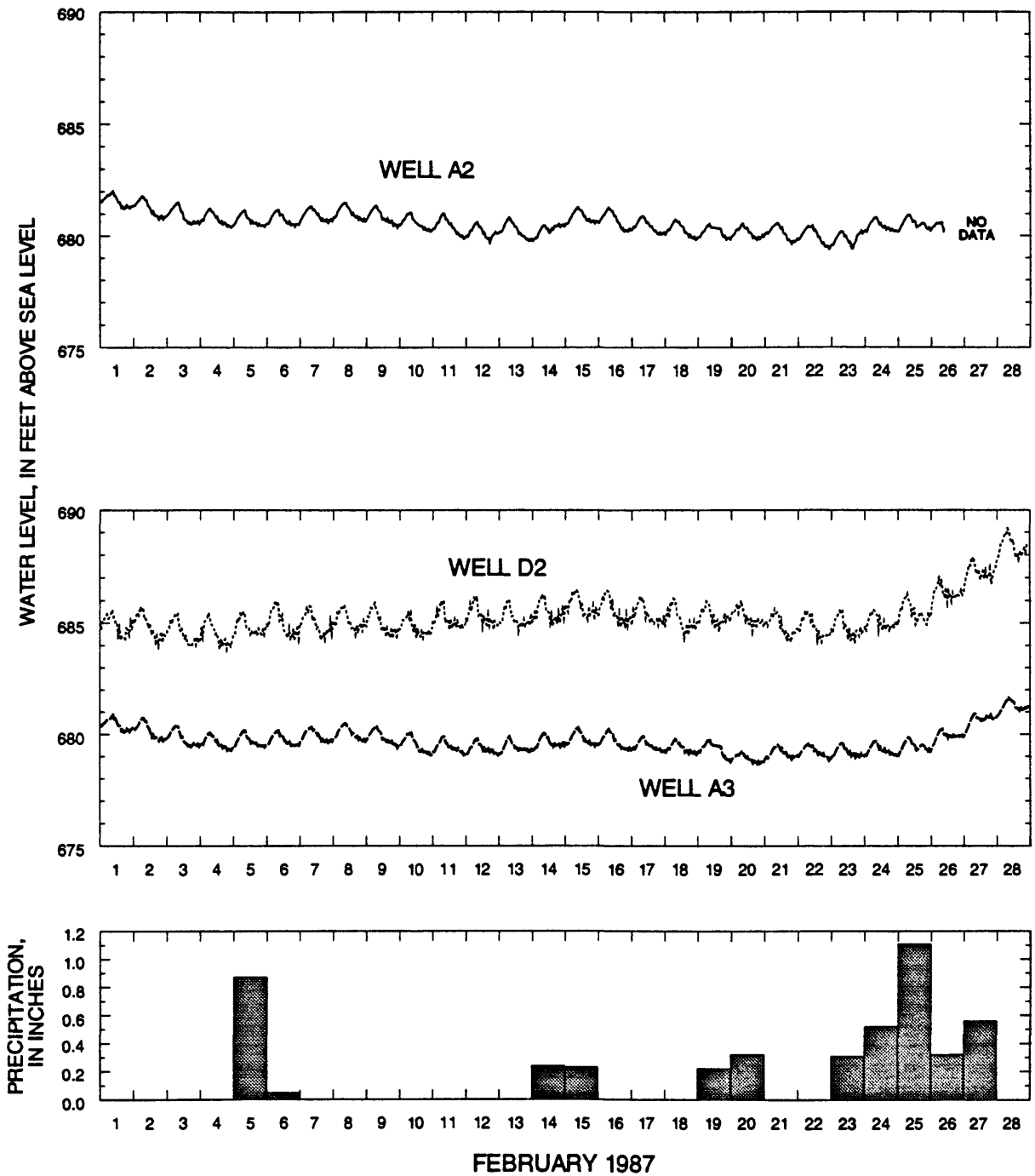
Figure 21h. Water levels in monitor wells A2, A3, and D2 and precipitation, San Antonio, Texas, February 1987.
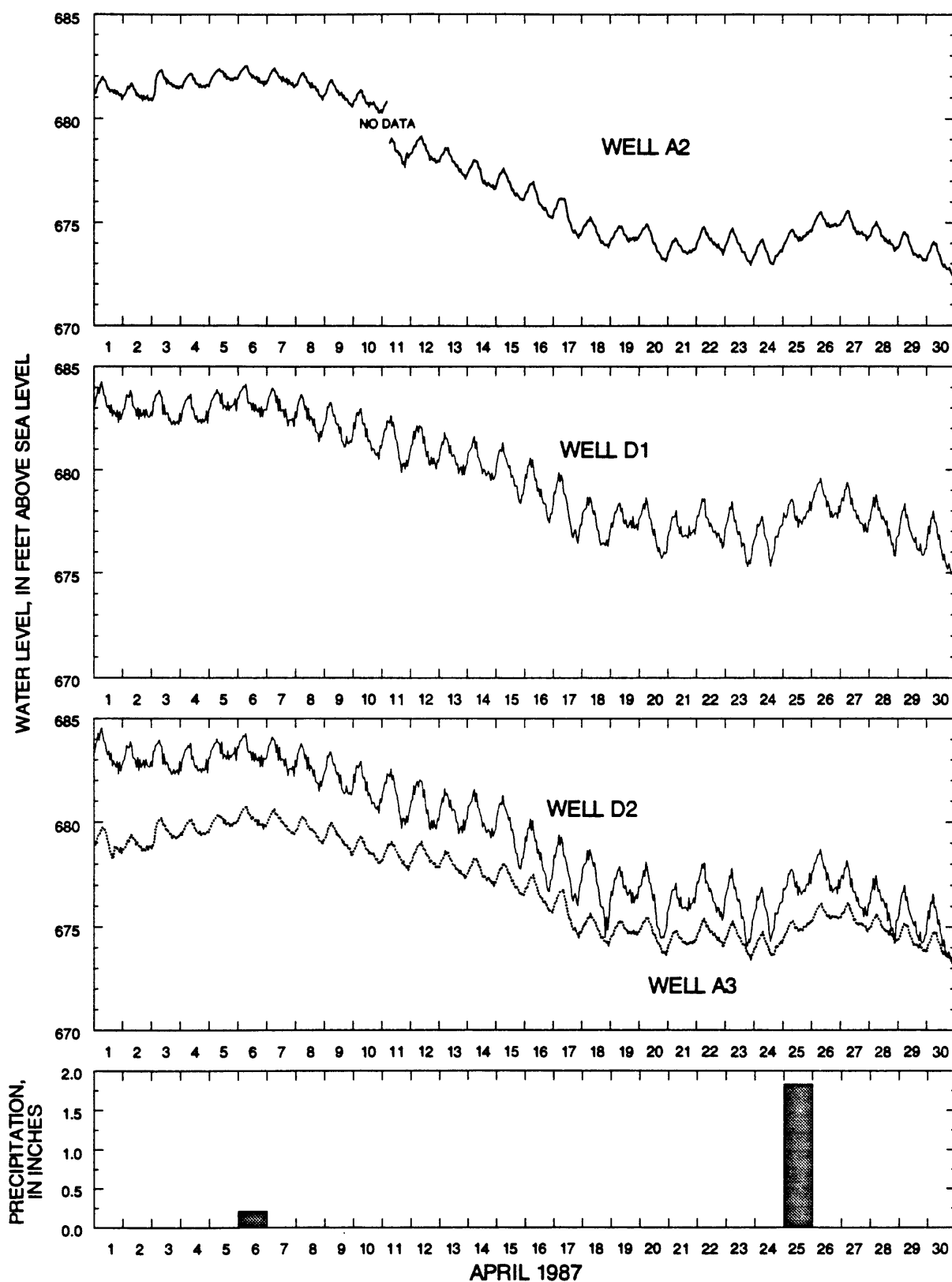
Figure 21I. Water levels in monitor wells A2, A3, D1, and D2 and precipitation, San Antonio, Texas, April 1987.

for by salinity differences and by different lengths of the water columns. The cause of this vertical head gradient is not well understood, but could be related to the difference in transmissivity between the cavernous zone and the saline-water zone below the RDM, and to the varying response of water levels in each zone to changes in stresses or recharge.

The relation between the water levels in wells D1 and D2 was variable during the summer months. Because of the shorter water column, the shallow well (D2) usually had a slightly higher water level. Changes in pumping, either regionally or locally, at the nearby Artesia well field periodically caused the water level in well D2 to decline below the water level in the deep well (D1). At these times there was a slight upward head gradient from the zone below the RDM to the zone above it. This could have been caused by the Artesia wells withdrawing more water from above the RDM than from below it. The zone above the RDM probably was more transmissive at all the Artesia wells as indicated at the site D test hole. The transient vertical gradients probably were short term because most of the Artesia wells (and most of the water-supply wells in San Antonio) are open holes for most or all of the thickness of the Edwards aquifer. At times when the water-supply wells are not being pumped, water flow between zones probably occurs in the open holes to equalize any head differences. When these open-hole wells were drilled, any naturally existing vertical gradient could have been reduced or eliminated. Only transient vertical gradients resulting from well withdrawals can be established in the aquifer near a water-supply well or well field.

Water levels in the freshwater wells at site D (D1 and D2) and at site C (C1) responded quickly and strongly to withdrawals and other aquifer stresses. Water levels in the saline-water wells at site A (A1, A2, and A3) and at site C (C2) were less responsive and responses were delayed from several minutes to several hours compared to those of the freshwater wells (figs. 21a-i).

On March 25, 1987, a test of the response of the monitor wells to discharge from all the Artesia water-supply wells was conducted. Because of equipment problems, the data from wells C1 and C2 are not reliable. The water levels in wells A2, A3, D1, D2, and Artesia 1 observation well are shown in figure 22. At 6:00 a.m., all the Artesia wells, except for Artesia 1 (fig. 2) were turned on and pumped for 12 hours. The

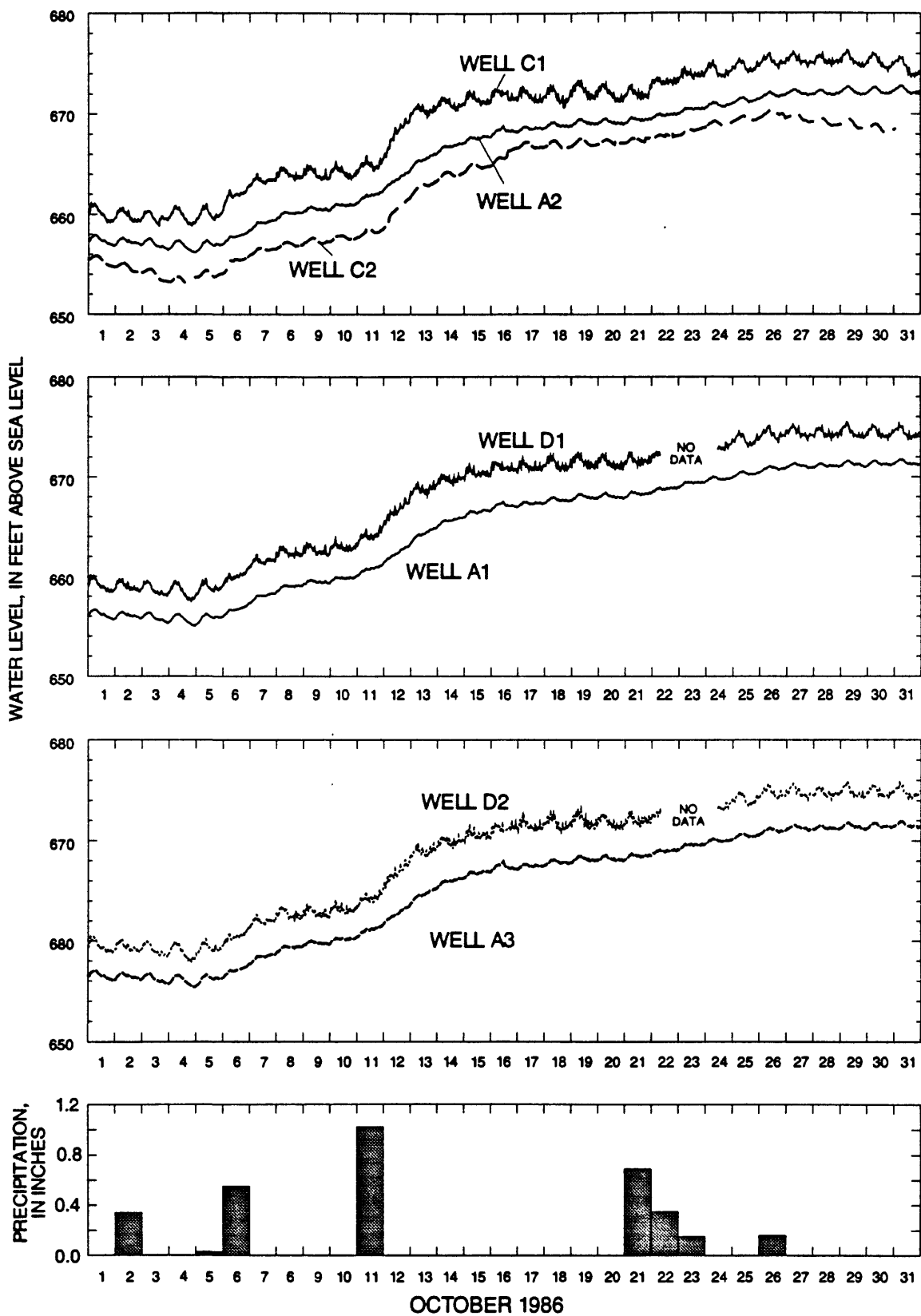discharge was estimated by comparing pressure readings to pressure/discharge curves. The estimated average discharge was 56 ft^3/s (25,000 gal/min) and the computed transmissivity was 790,000 to 950,000 ft^2/d (William F. Guyton Associates, Inc., 1988, table 1). The water level in Artesia 1 observation well did not respond as quickly or as strongly as the water levels in wells D1 and D2 even though Artesia 1 observation well was located closer to the pumped wells. Differences in the construction of wells D1 and D2 and the Artesia 1 observation well could account for the different responses. As previously mentioned, wells D1 and D2 might be completed in a fault zone, and the Artesia 1 observation well was outside the fault zone.

The responses of the water levels in wells A2 and A3 were much attenuated with respect to the instantaneous response shown by water levels in wells D1, D2, and Artesia 1 observation well. For wells A2 and A3, it is difficult to separate the water-level response caused by the pumping of the Artesia wells from the response caused by regional increase in water use. Water levels in most Edwards aquifer observation wells respond to changes in regional water use through the day. Water levels that decline because of municipal water use usually reach a minimum at about 6:00 p.m., and the water levels rise to maximum altitude in the early morning between 12:00 a.m. and 5:00 a.m. Between 6:00 a.m. and 7:00 a.m., water use begins to increase again and water levels in observation wells decline from the early morning peak. This regional effect was observed in wells A2 and A3. The test conducted at the Artesia well field, about 1 mi from the site A wells, did not cause a decline in the water levels in wells A2 or A3.

Hydrographs detailing some of the short-term relations of water levels among the monitor wells and comparing the water-level responses to abrupt changes in aquifer stresses are shown in figures 23a-d. These are extracted from the month-long hydrographs shown in figures 21a-i. The first short-term water-level hydrograph (fig. 23a) shows 3 days in August when withdrawals were large, a substantial portion of which went for lawn watering. Because of the large withdrawals, water levels in the wells that responded quickly to pumping—wells C1, D1, and D2—were drawn down near or below water levels in the slowly responding saline-water wells. Changes in the water levels in the four saline-water wells (A1, A2, A,3 and C2) lagged behind the changes in the freshwater well levels by several minutes to less than 2 hours. The effect of density
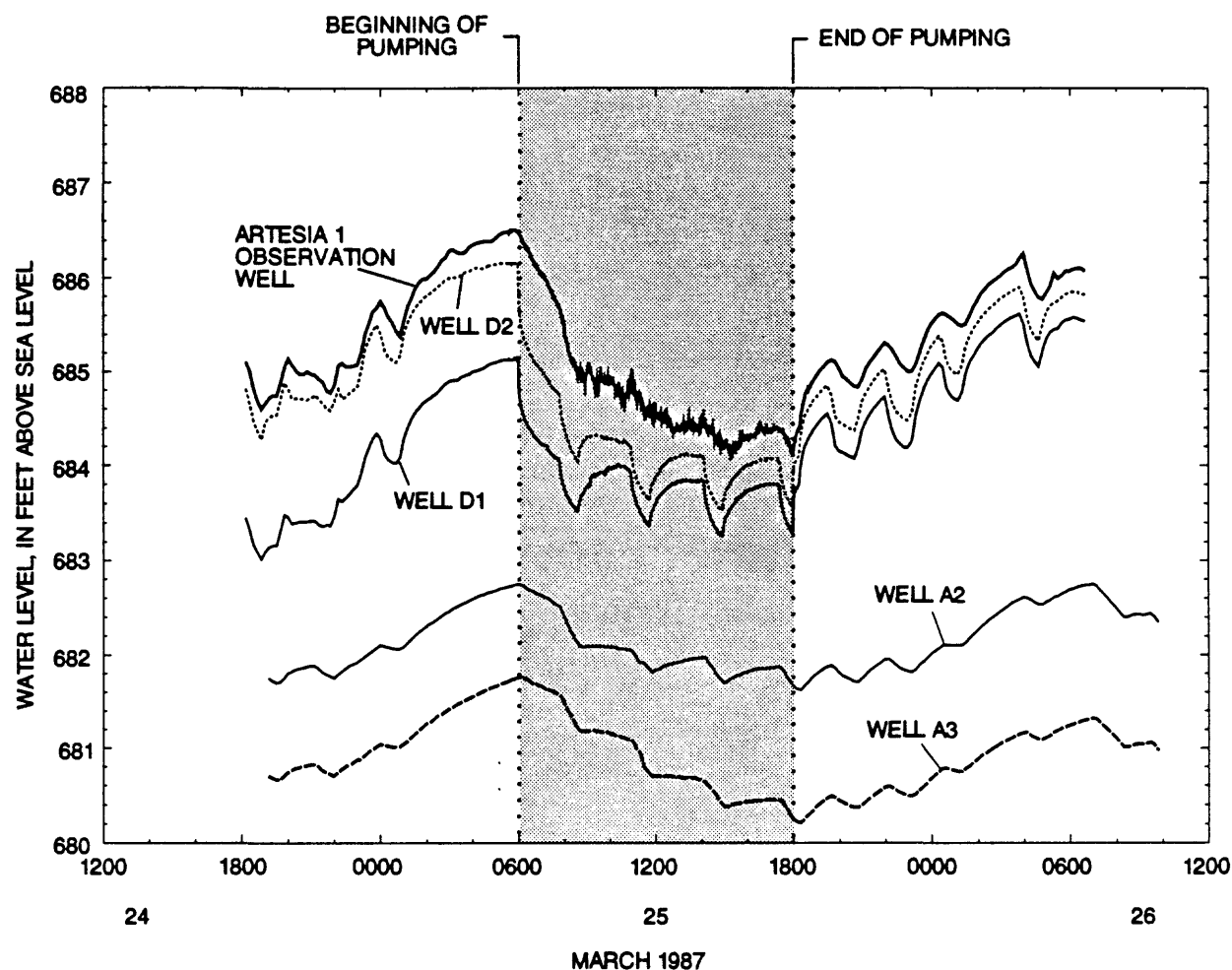
Figure 22. Water levels in monitor wells A2, A3, D1, and D2 and Artesia 1 observation well, San Antonio, Texas, before, during, and after an aquifer test, March 24-26, 1987.

contrasts on potential flow is not indicated because actual measured altitudes are plotted in these figures.

The short-term water-level hydrograph (fig. 23b) for September 5-7, 1986, includes a period of rain showers when lawn watering was still intense. The hydrographs show that the smaller water withdrawals caused small fluctuations in water levels in wells C1, D1, and D2, but caused few changes in the four slowly responding wells (A1, A2, A3, and C2). Water levels increased during the 3 days.

The relation between wells C1 and C2 for September 5-7, 1986, also is shown in figure 23b. The water level in the shallower freshwater well (C1) was consistently higher than the water level in the deep saline-water well (C2). The two hydrographs were usually parallel, although the deep well was less respon-

sive than the shallow well. The average difference in water levels was about 4 ft throughout the period shown. The difference expected from the contrast in salinity and lengths of the water columns was about 3 ft. Because the fluctuations of the shallow well were greater (daily fluctuations of about 2 ft) than the deep well fluctuations (less than 1 ft), the apparent vertical gradient varied throughout the day during periods of intense pumping. The vertical gradient between wells C1 and C2 could reverse during a day of intense withdrawals.

A steeper upward trend is shown for October 10-13, 1986 (fig. 23c). During this period, all the water levels rose substantially in response to a large rainstorm on October 11, 1986. Part of the rise was caused by recharge from rainfall, but most of the rise in
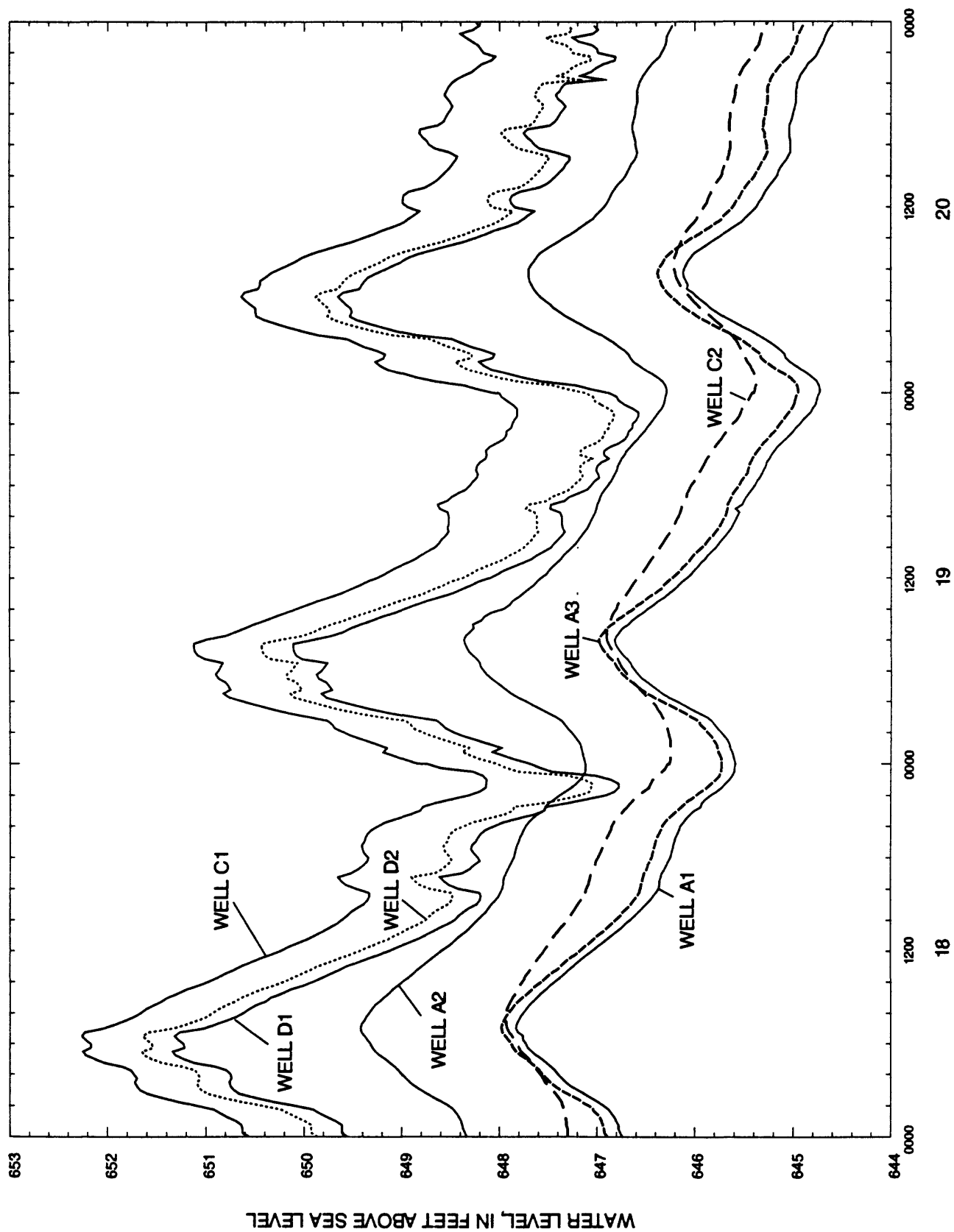
Figure 23a. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, August 18-20, 1986.
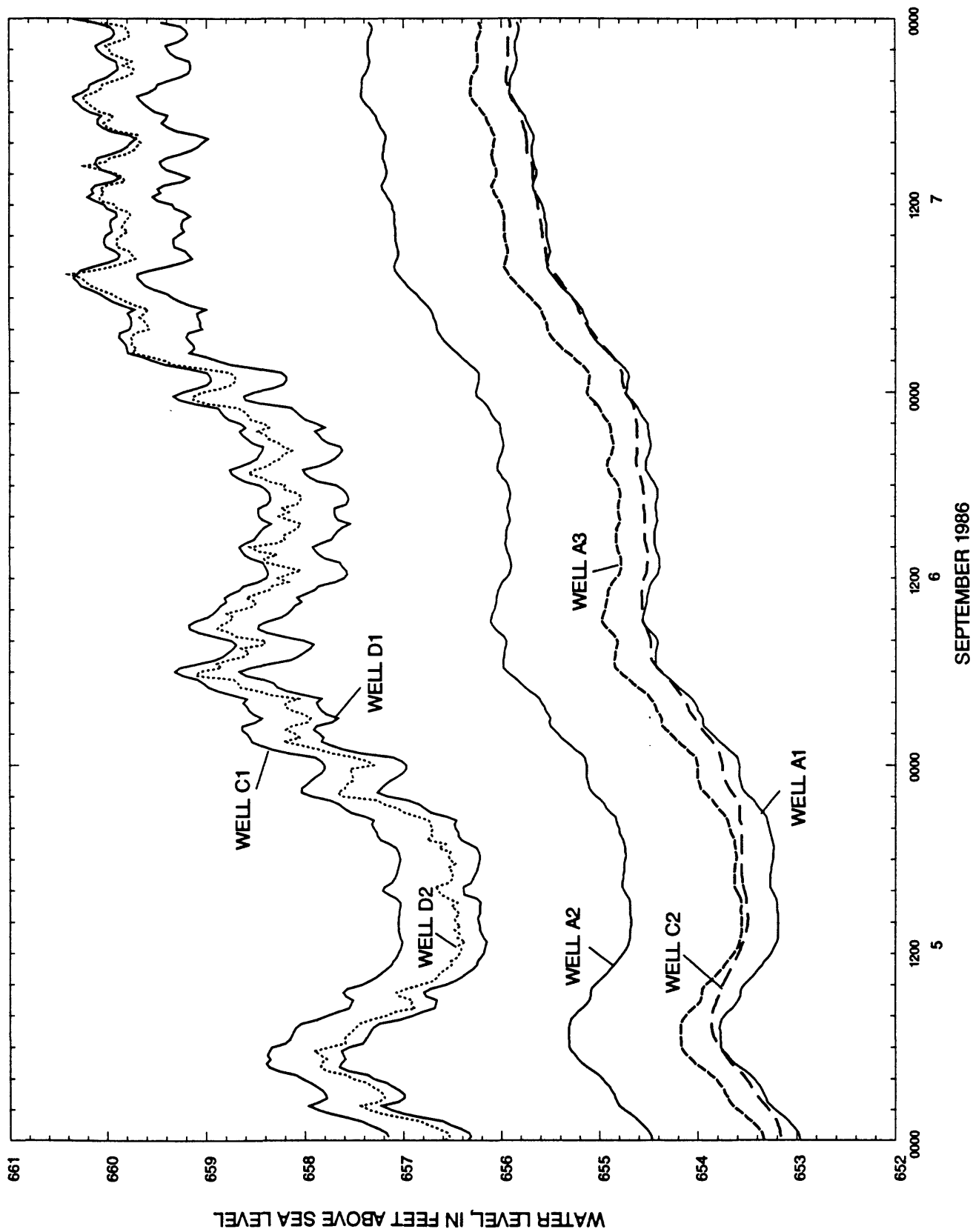
Figure 23b. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, September 5-7, 1986.
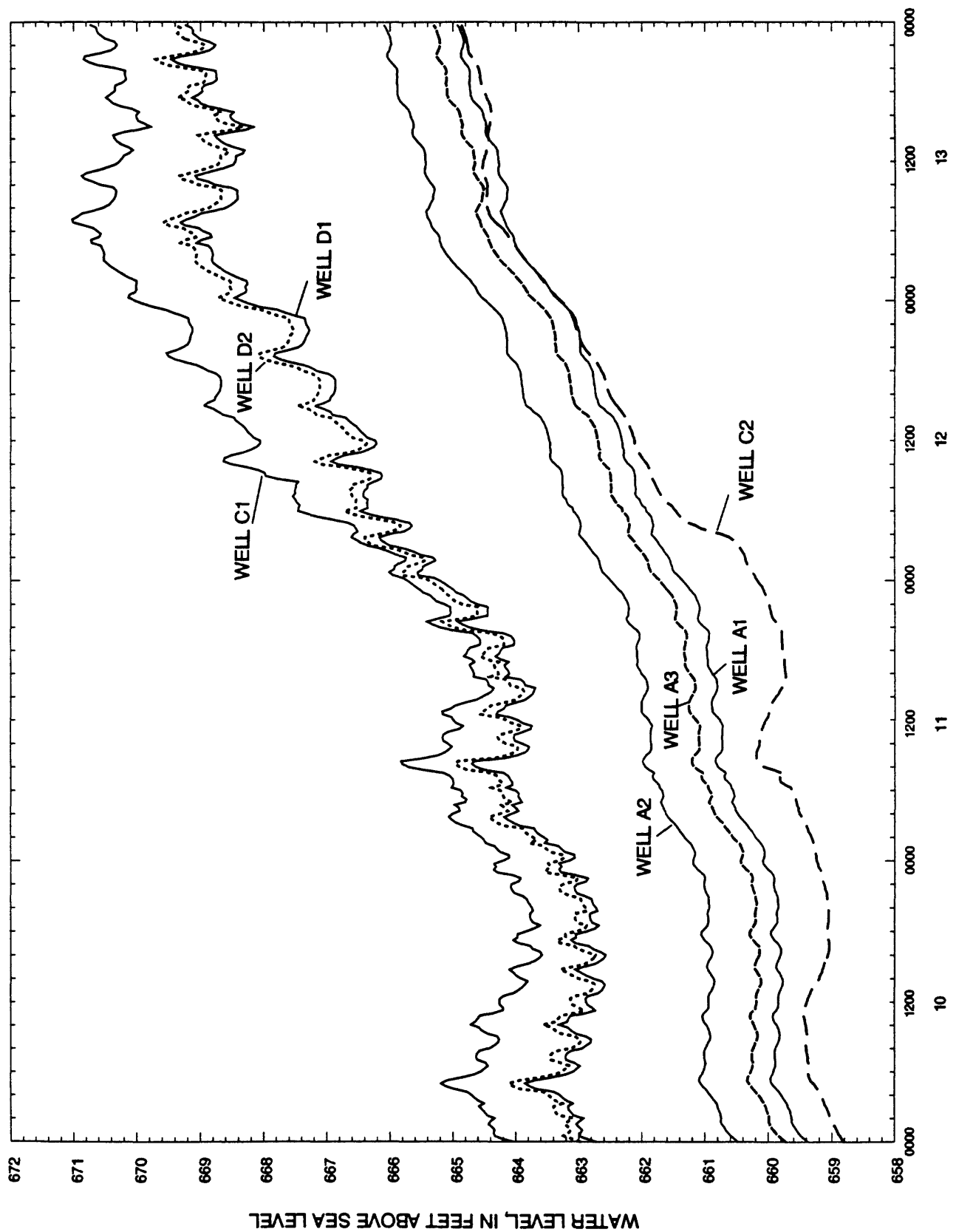
Figure 23c. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, October 10-13, 1986.
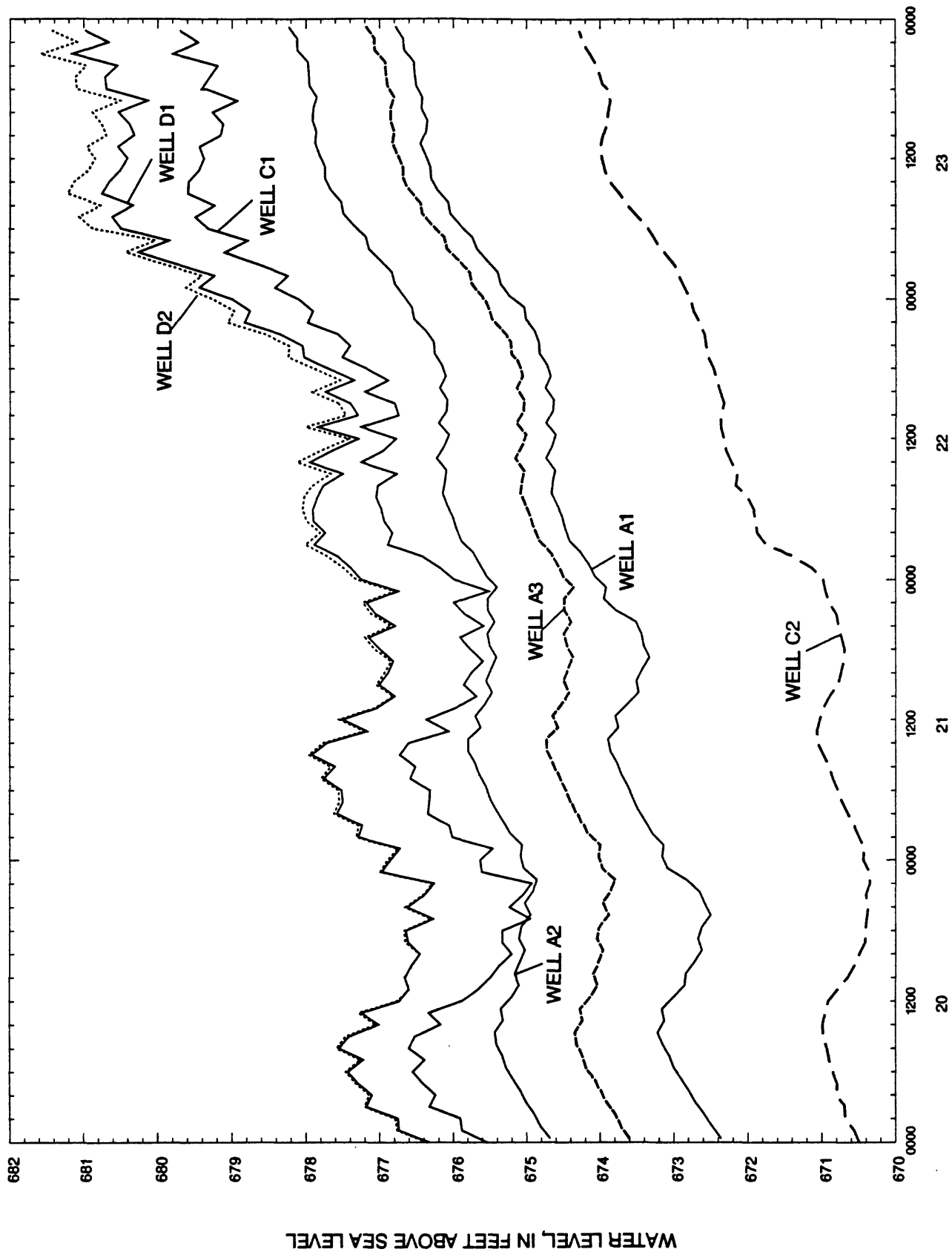
Figure 23d. Water levels in monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, December 20-23, 1986.

the water levels was caused by a decrease in lawn watering.

Another large rainstorm occurred on December 22, 1986; its effect on water levels in the monitor wells are shown in figure 23d. The rise in the water levels shown in figure 23d resulted mainly from recharge to the aquifer because lawn watering was minimal during December, and because more rain fell during December than during October.

Well D2 had the highest water levels during December 20-23, 1986. Wells D1 and D2 had similar high water levels, the water level in well C1 was below the water level in wells D1 and D2, and wells A1, A2, and A3 had water levels intermediate between the freshwater-zone wells (D1 and D2) and well C2. The salinities in water from wells C1, D1, and D2 were similar enough to negate the density differences between them. Therefore, the differences in water levels among these wells reflect the gradient. The gradient that was almost always from the shallow transition-zone well (C1) to the freshwater-zone wells (D1 and D2) was reversed in December. The salinities of water from the other four wells at the saline-water zone and at the transition-zone sites (A1, A2, A3, and C2) were similar. A comparison of the water levels indicates that the gradient was usually from A1, A2, and A3 toward well C2.

The dip of the aquifer and the gradient in equivalent freshwater head can be used to determine the direction of ground-water flow (Davies, 1987). In the study area, the dip of the aquifer is large enough and the relative difference in equivalent freshwater heads is small enough that dense saline water will tend to flow downdip regardless of the existence of a small updip gradient in equivalent freshwater head. The denser saline water usually is downdip from or beneath the freshwater, and tends to remain there because of its greater density and the dip of the aquifer.

Water-Quality Data

A consistent and unresolved problem exists with most of the analyses of the saline-water samples from wells A1, A2, A3, and C2, and for historical analyses of samples from the saline-water zone. The electrical balance check on the ratio of cations to anions indicates that major anion equivalents from the analytical data usually were greater than the equivalents of cations. An approximate check on the accuracy of a complete water analysis is a balance between cations and anions. A water solution is electrically neutral. The sum of the equivalents of cations is equal to the sum of the equivalents of anions. The difference between the two must be less than 1 or 2 percent of the total equivalents for a good complete analysis (Hem, 1985, p. 164). Hem (1985) also states that the cation-anion balance might not be a good indication of the accuracy of the smaller ion concentrations when the total dissolved-solids concentration exceeds 1,000 mg/L. The cation-anion balance error for saline-water samples ranged from +0.73 to -7.03 percent. The apparent excess of anions is restricted to those samples with individual anion concentrations that are greater than 1,000 mg/L and major individual cation concentrations greater than about 100 to 200 mg/L. Lack of precision (two significant digits) of the reporting format for concentrations that exceed 1,000 mg/L might contribute to the cation-anion imbalance. Also, the analytical methods used to determine the major ions in these saline-water samples might be inappropriate for samples where individual ion concentrations exceed 1,000 mg/L.

Undetermined sulfide anion concentrations are an unlikely factor of the charge imbalance because the anion concentrations in the saline-water samples were greater than the cation concentrations. The possibility that the imbalance was caused by undetermined strontium also is not likely because the available strontium concentration data indicate that strontium could not be present in concentrations as great as 7 percent of the total cation equivalents (Pearson and Rettman, 1976).

The data from the saline-water samples are questionable because many of the analyses of the saline water did not meet the requirement of analytical balance between equivalents of cations and anions. A comparison of individual ion concentration uncertainties would be useful because the source of the apparent problem in the saline-water samples is unknown. Unfortunately, the apparent error could be the result of a single ion determination; several ion determinations; all ion determinations; or none. Therefore, the error distribution is difficult or impossible to quantify. Nevertheless, the possibility of a significant error exists in the saline-water-zone analyses. The effects of data uncertainty on geochemical modeling of the type described in the section on water chemistry evolution are discussed in detail by Plummer (1984).

Changes in Specific Conductance with Flow and Depth

Site A

Plots of the specific conductance of water from the cumulative-depth flow tests with depth of the interval tested are shown in figure 24a. The data are modified from Pavlicek and others (1987). A comparison of specific conductance for the site A test hole (A1) (fig. 24a) with the discharge (fig. 19a) indicates that the specific conductance of the water and the flow rate changed appreciably with the length of the hole tested. These data were not from discrete samples; water from the more permeable top of the aquifer almost always was mixed with the more saline water from the less permeable lower part of the aquifer in the open hole. These data represent a qualitative indication of salinity and water productivity from the various rock strata. Discrete samples (samples from sections less than 10 ft long) and flow from multiple-packer tests could have delineated these vertical differences more accurately. Multiple-packer tests were assumed to be infeasible because of the nature of the rocks that make up the aquifer.

The increase in flow and specific conductance from the interval between the bottom of the George-town Formation (965 ft) and 1,279 ft indicate that this interval is uniform in permeability, but yields water of increasing salinity with depth. The water with the smallest dissolved-solids concentration was produced from the top half of the Person Formation. The section of the test hole below 1,279 ft produced less than one-tenth of the open-hole flow, but is slightly more saline than water from other intervals.

The specific conductance of the open-hole flow was 5,230 μS/cm. The depth of interval tested, average discharge, and measured specific conductance for the discrete-interval flow tests of the site A test hole are listed in table 2. The specific conductance of the interval between 1,276 and 1,489 ft (bottom of the test hole) was 6,650 μS/cm. The interval from 965 to 1,075 ft depth had a specific conductance of 2,680 μS/cm, smaller than the specific conductance of 3,198 μS/cm (fig. 24a) obtained from only the upper 50 ft of the Edwards aquifer during the drilling. A zone of less saline water could be present below the upper 50 ft of the Person Formation and above 1,075 ft depth. Data from the other shallow monitor wells drilled at this site were insufficient for further understanding of the stratification of the salinity in the saline-water zone.

Table 2. Selected data from the discrete-interval flow tests at the site A test hole, San Antonio, Texas

[ft, feet; gal/min, gallons per minute; μS/cm, microsiemens per centimeter at 25 degrees Celsius; --, no data]

Interval (ft)	Average discharge (gal/min)[1]	Specific conductance (μS/cm)
965-1,489	[2]378	5,230
1,276-1,489	[3]33	6,650
1,180-1,489	[4]44	6,060
965-1,180	[5]238	4,360
965-1,075	[6]128	2,680
[6]1,200-1,275	[3,7]23	--

[1] Duration of flow, 4 hours; duration of recovery, 2 hours.
[2] Discharge determined by manometer with 10-inch pipe × 8-inch orifice.
[3] Discharge determined volumetrically.
[4] Discharge determined by manometer with 4-inch pipe × 2.5-inch orifice.
[5] Discharge determined by 8-inch pipe × 5-inch orifice.
[6] Completed monitor well.
[7] End of 1-hour flow.
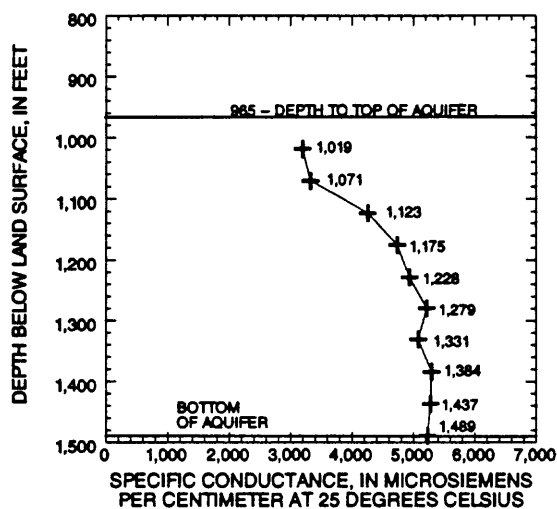
EXPLANATION

✛ 941 SPECIFIC CONDUCTANCE AND
BOTTOM DEPTH OF INTERVAL
—Interval is from top of aquifer
to indicated depth. Data are
plotted at bottom of interval

a. Site A test hole
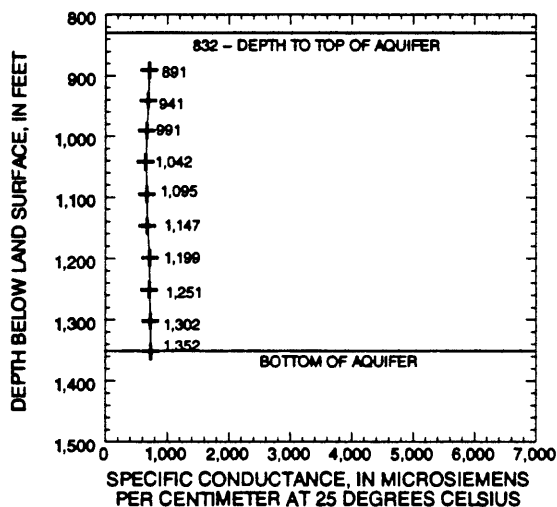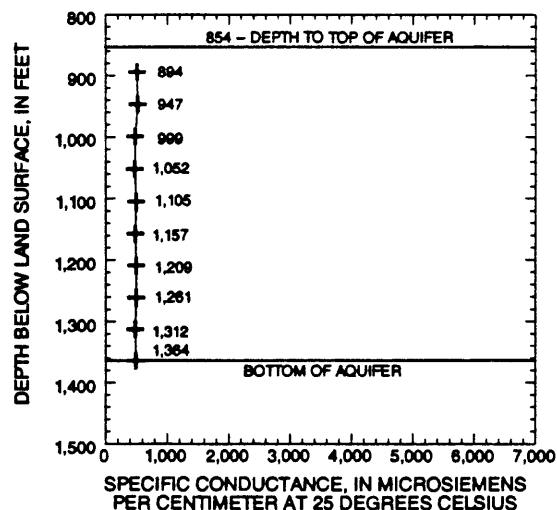
b. Site C test hole

c. Site D test hole

Figure 24. Specific conductance of test-hole discharge at sites A, C, and D, San Antonio, Texas, during cumulative-depth flow tests. (Modified from Pavlicek and others, 1987.)

Site C

The plots of specific conductance (fig. 24b) and discharge (fig. 19b) of cumulative-depth tests for the site C test hole are not representative of the conditions in the aquifer for this zone. The large amount of flow from the cavern near the top of the Person Formation diluted the water flowing from the section of the hole below it. Change was not evident in amount of water flow or in salinity for each progressively deeper interval because of the effect of the water from the cavern.

The data from the single-packer interval flow tests added some definition to the vertical stratification of flow and water quality in the transition zone. The depth of interval tested, average discharge, and measured specific conductance for the discrete-interval flow tests of the site C test hole are listed in table 3. The specific conductance of the water produced from the interval 832 to 859 ft depth was 772 μS/cm. The specific conductance from the section below 859 to 1,396 ft (bottom of the hole) was 3,860 μS/cm. The flow produced from the top 27-ft section (832 to 859 ft) was 80 percent of the entire flow from the open hole and had the smallest salinity measured in the test hole. Other

intervals that included the top 27 ft also produced about the same amount of flow and water of the same salinity, indicating that the flow from the top 27 ft (mostly from the cavern) diluted the flow of saline water from below it.

The hydraulic head of the water below the top 27 ft was lower than that of the hydraulic head in the cavern, and water could have moved downward in the open hole. The greatest salinities of water produced from the test hole were from the intervals between 1,056 and 1,396 ft depth and between 1,240 and 1,396 ft depth. The specific conductance of water from these two intervals was 5,860 and 5,870 μS/cm, respectively. The longer interval produced about 3 times the flow of the shorter, deeper interval (table 3). The final interval test was done on the section between depths of 840 to 881 ft and a small amount of water was produced relative to the other sections. The specific conductance of the water from this interval was not measured. This last test indicates that almost all of the freshwater produced from the test hole comes from the cavern in the interval between depths of 832 and 840 ft, although some freshwater could also be from 840 to 881 ft.

Table 3. Selected data from the discrete-interval flow tests at the site C test hole, San Antonio, Texas

[ft, feet; gal/min, gallons per minute; μS/cm, microsiemens per centimeter at 25 degrees Celsius; --, no data]

Interval (ft)	Average discharge (gal/min)[1]	Specific conductance (μS/cm)
832-1,396	[2]1,413	842
859-1,396	[3,4]95	3,860
832-859	[3]1,128	772
1,056-1,396	[3,4]76	5,860
832-1,056	[5]1,147	784
1,240-1,396	[3,4]27	5,870
832-1,240	[6]1,167	826
[7]840-881	[4,8]42	--

[1] Discharge determined by manometer with 10-inch pipe × 8-inch orifice.
[2] Duration of flow, 6 hours; duration of recovery, 2 hours.
[3] Duration of flow, 4 hours; duration of recovery, 2 hours.
[4] Discharge determined by manometer with 4-inch pipe × 2-inch orifice.
[5] Duration of flow, 3 hours; duration of recovery, 1 hour.
[6] Duration of flow, 2 hours; duration of recovery, 1 hour.
[7] Completed monitor well.
[8] End of 1-hour flow.

Results of cumulative-depth tests for the C2 monitor well (before casing), drilled in the transition zone, were consistent with those observed in the saline-water-zone wells at site A. Although the production is constant over the longer depth intervals, the salinity increases with increasing depth and length of the sampled interval.

Site D

The specific-conductance (fig. 24c) and discharge (fig. 19c) data from the cumulative-depth flow tests from the site D test hole indicate that most water probably was produced from near the top of the Person Formation. The flow-rate curve for these cumulative-depth flow tests shows that the largest volume of water was produced from the top 50 ft of the test hole and increased slightly as the test hole was drilled deeper.

Data from the single-packer interval flow tests provided an incomplete definition of the vertical stratification of flow and salinity. The depth of interval tested, average discharge, and measured specific conductance for the discrete-interval flow tests of the site D test hole are listed in table 4. The water with the smallest specific conductance (474 μS/cm) was produced by the interval from 1,040 to 1,384 ft depth. The most saline water was produced by the interval from 1,225 to 1,384 ft depth. The specific conductance of the water produced by this interval, 6,380 μS/cm, is about

as saline as the most saline water produced from the site A and site C test holes.

A test of the interval between a depth of 1,158 and 1,384 ft produced less water than the shorter interval between 1,225 and 1,384 ft; the specific conductance of the water produced from the first section was 1,862 μS/cm, which is less saline than the water from the shorter, deeper interval (6,380 μS/cm). Interstratal flow within the test hole is the most likely explanation for the discrepancy. Some slightly saline water from the interval not in common (between 1,158 and 1,384 ft depth) could have had greater hydraulic head than the saline water in the deeper interval. The difference in hydraulic head could have resulted in downward movement in the test hole. Water ranging from slightly saline to fresh from the interval between 1,158 and 1,384 ft depth could have flowed into the test hole and down to the interval between 1,225 and 1,384 ft depth and into the aquifer from the test hole.

The differences in hydraulic head required to cause the inferred reversals of flow in the test holes discussed in this report are unknown and probably are not great—on the order of 1 ft or less difference in equivalent freshwater head. The deep monitor well was not completed in the most saline interval of this test hole. Therefore, no information on the vertical potential gradient between the freshwater near the top of the aquifer and the saline water at the bottom of the aquifer could be collected from the two wells installed at this site.

Table 4. Selected data from the discrete-interval flow tests at the site D test hole, San Antonio, Texas

[ft, feet; gal/min, gallons per minute; μS/cm, microsiemens per centimeter at 25 degrees Celsius; --, no data]

Interval (ft)	Average discharge (gal/min)[1]	Specific conductance (μS/cm)
1,158-1,384	[2]32.4	1,862
1,225-1,384	[3]114.0	6,380
1,040-1,384	[4]64.0	474
[5]1,148-1,223	[6,7]7.56	--

[1] Discharge determined by manometer with 4-inch pipe × 2-inch orifice.
[2] Duration of flow, 7 hours; duration of recovery, 1 hour.
[3] Duration of flow, 8 hours, 40 minutes; duration of recovery, 1 hour.
[4] Duration of flow, 8 hours, 10 minutes; duration of recovery, 1 hour.
[5] Completed monitor well.
[6] Discharge determined volumetrically.
[7] End of 1-hour flow.

Changes with Time

The mean concentrations of chemical constituents for samples collected from the completed monitor wells, July 1986-April 1987, are shown in table 5 (at end of report). The mean concentrations of samples from an observation well at the Mission well field and the Artesia 1 observation well during this study are listed in table 6. In most of the figures that follow in this section, data for the saline-water wells are shown in the upper part of the figure and data for the slightly saline-water and freshwater wells are shown in the lower part. This was done to have comparable scales on the vertical axes of the plots.

A hydrograph of the water level at midnight in well D2 for August 5, 1986-April 30, 1987, is shown in figure 25. The hydrograph is shown to compare the water level with the data from the water samples collected during this same period (table 5). Comparing the water level and the salinity or dissolved constituents can help determine whether a relation exists between salinity in the monitor wells and the head in the freshwater zone of the aquifer. The observed water level in well D2 is the most complete water-level data set of the seven transect wells and parallels the water level in the index well in San Antonio, known as "J-17" (AY-68-

37-203). The water levels in both wells are representative of the combined regional withdrawals and recharge status. Except for brief periods at the beginning and ending of the hydrograph, the water level in well D2 remains steady or increases indicating a general recovery of the water level resulting from substantial recharge and decreased withdrawals.

Specific-conductance measurements in water samples from the monitor wells are illustrated in figure 26. The plots of specific conductance for all seven monitor wells indicate no discernible trend in the data through time. The fluctuations in the data from month to month can be explained partly by analytical and sampling uncertainty and partly by random fluctuations in the quality of water sampled from the wells.

The deep monitor well (D1) at the freshwater site had the largest coefficient of variation for specific conductance (0.099) of the wells monitored during the study (table 5). The coefficient of variation is defined as the ratio of the standard deviation to the mean. Water in this well could be the most sensitive indicator of changes in water quality through time. The wells that had the smallest coefficients of variation of specific conductance were A1 (0.035), A2 (0.024), A3 (0.035), and C2 (0.032); the four most saline wells. This is
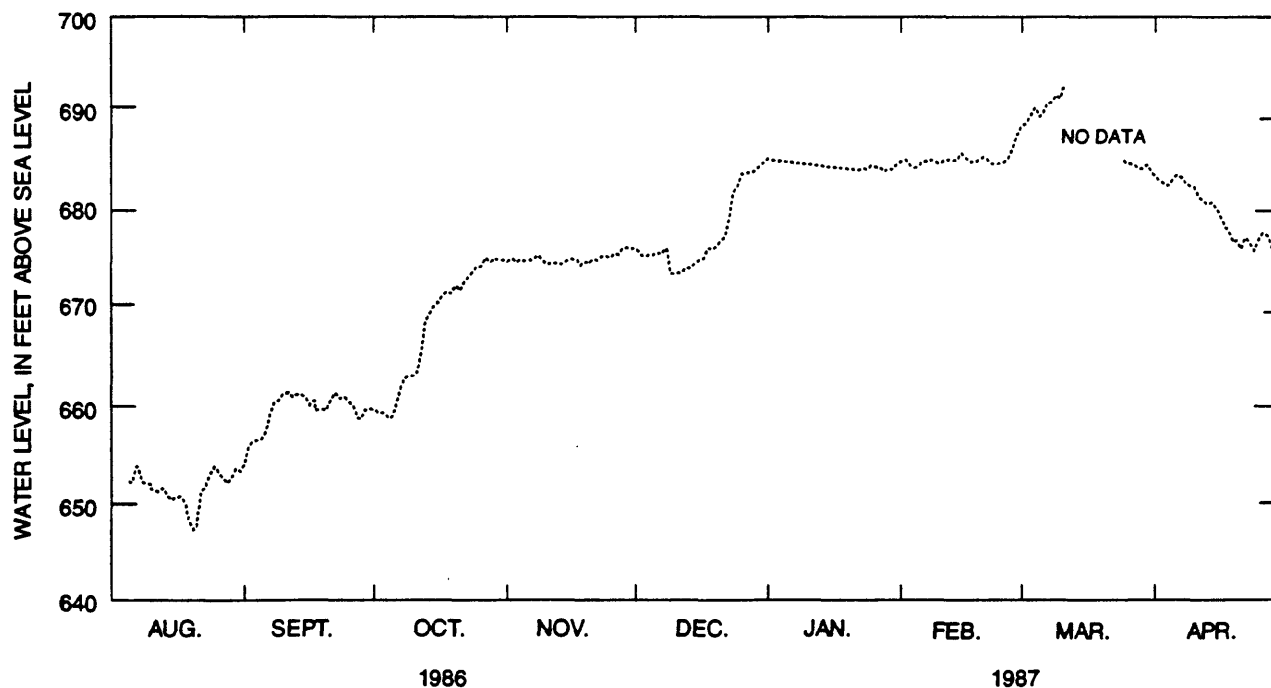
Figure 25. Water level in well D2, San Antonio, Texas, at midnight, August 1986-April 1987.

caused partly by the small relative changes in the specific conductance measured at these wells and partly by the attenuated response they have to changes in aquifer stresses. The coefficient of variation for specific conductance for well C1 is 0.067 and for well D2 is 0.052. Water in these wells probably is of intermediate sensitivity to changes in water quality. Although the water level in well D2 appeared to be sensitive to aquifer stresses, the water comes from a rather large body of freshwater (virtually the entire freshwater zone of the aquifer upgradient from it) and shows little change in quality through time.

Changes in specific conductance in water from two freshwater observation wells are shown in figure 27. The Artesia 1 observation well is adjacent to site D (fig. 2), and the Mission observation well is more than 5 mi southwest from the monitor wells (fig. 1).

The alkalinity of the water in the well samples can be assumed to result from the bicarbonate ion, which is the major anion in the samples from wells C1

and D2. The concentrations of alkalinity over time (fig. 28) show alkalinity as $CaCO_3$. A substantial increase in alkalinity was measured in water from all saline-water wells in the September or October 1986 samples. These changes are of unknown origin.

Calcium (fig. 29) is the main cation in the water from wells C1, D1, and D2. Calcium concentrations decreased consistently in the November and December 1986 samples from these three wells. The similarity in the three well-sample concentrations indicates that some event caused the change, but on the basis of these data, the nature of the event is difficult to determine. A slight, but consistent decrease in calcium concentrations also is apparent for the October, November, and December 1986 samples from wells A1, A2, and C2. Calcium concentration in the August 1986 sample from well A2 had a large increase paralleling the increase in magnesium concentration (fig. 30) in water from this well. Water from wells A1 and A3 also had slight increases in calcium concentration during August
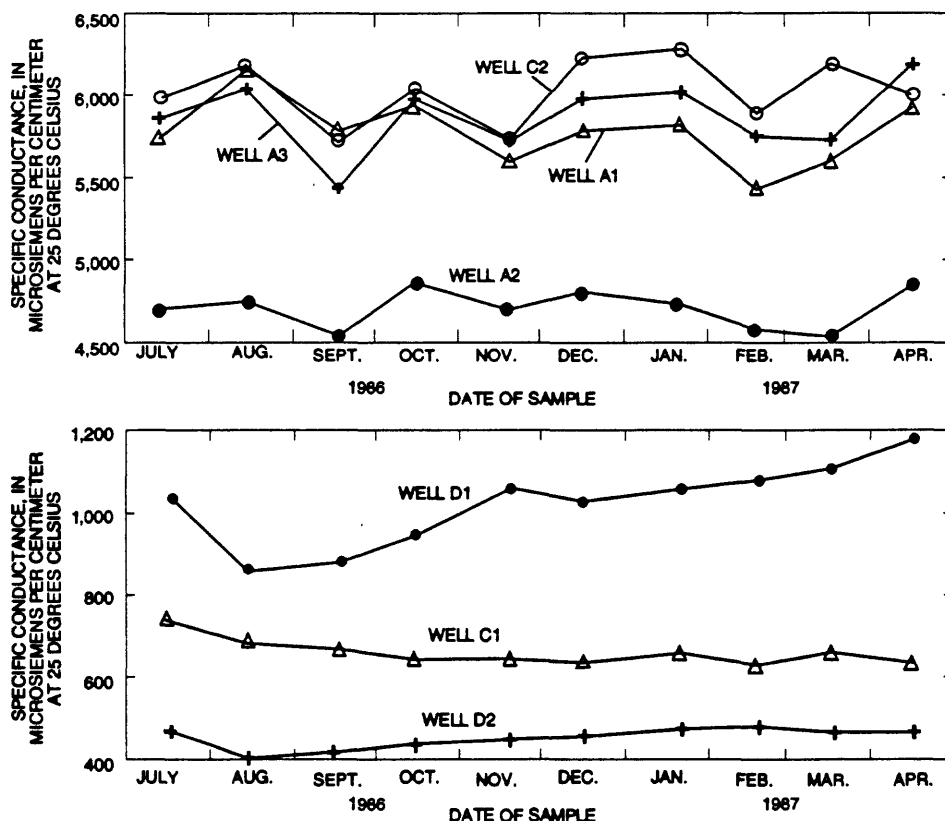
Figure 26. Specific conductance in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

Table 6. Summary statistics of the water quality from Mission and Artesia 1 observation wells, San Antonio, Texas

[μS/cm, microsiemens per centimeter at 25 degrees Celsius; °C, degrees Celsius; mg/L, milligrams per liter; --, no data]

Property or constituent	Number of samples	Mean	Median	Standard deviation	Standard error	Mini- mum	Maxi- mum	Coefficient of variation[1]
			Mission observation well					
Specific conductance (μS/cm)	8	480	480	22	8	460	520	0.046
pH (units)	8	--	7.10	--	--	6.60	7.40	--
Temperature (°C)	8	26.7	26.8	.4	.1	26.0	26.5	.015
Alkalinity (mg/L as CaCO₃)	8	196	198	6	2	184	201	.031
Calcium (mg/L)	8	70	70	4	2	60	70	.061
Magnesium (mg/L)	8	17	17	1	<1	16	18	.059
Sodium (mg/L)	5	10	10	.2	<.1	10	10	--
Potassium (mg/L)	5	1.1	1.1	.1	<.1	.9	1.3	--
Sulfate (mg/L)	8	25	25	1	<1	23	28	.040
Chloride (mg/L)	8	20	20	2	<1	17	24	.10
Fluoride (mg/L)	5	.3	.3	<.1	<.1	.3	.4	--
Silica (mg/L)	5	12	12	<1.0	<1.0	12	13	--
Dissolved solids (mg/L)	5	270	270	4	2	270	280	--
			Artesia 1 observation well					
Specific conductance (μS/cm)	7	467	468	27	10	421	528	0.058
pH (units)	7	--	7.00	--	--	6.60	7.40	--
Temperature (°C)	7	26.0	26.0	.7	.3	25.0	27.0	.027
Alkalinity (mg/L as CaCO₃)	7	200	200	4	1	200	200	.020
Calcium (mg/L)	7	65	65.	3	1	61	68	.046
Magnesium (mg/L)	7	17	17	1	<1	15	17	.059
Sodium (mg/L)	4	10	10	.2	<.1	10	10	--
Potassium (mg/L)	4	1.0	1.1	.1	<.1	.9	1.1	--
Sulfate (mg/L)	7	22	22	1	<1	21	24	.045
Chloride (mg/L)	7	20	19	3	1	17	25	.15
Fluoride (mg/L)	4	.3	.3	.0	.0	.3	.3	--
Silica (mg/L)	4	12	12	<1.0	<1.0	12	13	--
Dissolved solids (mg/L)	4	270	270	5	2	260	270	--

[1] Coefficient of variation is defined as the ratio of the standard deviation to the mean. Only values for maximum number of samples are shown.
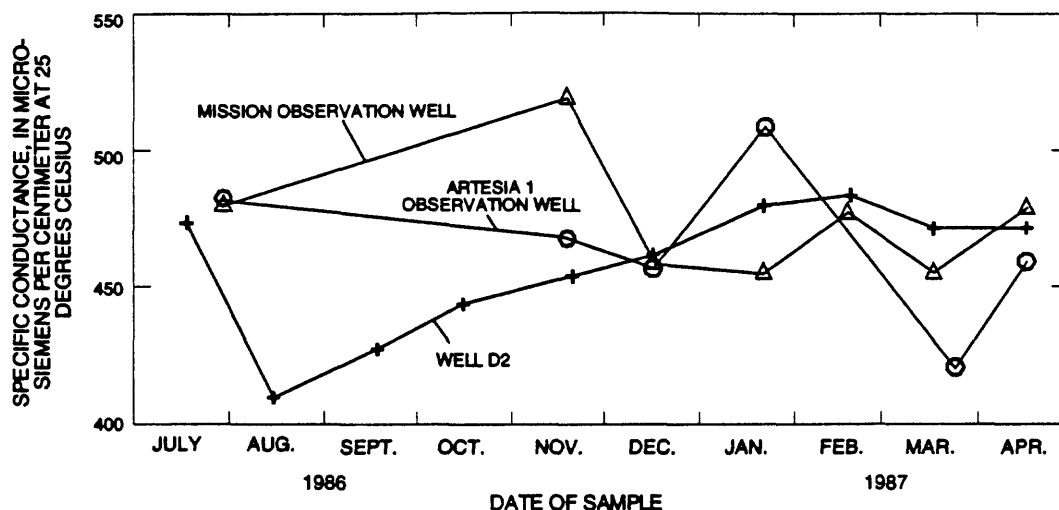
Figure 27. Specific conductance in water samples from Mission and Artesia 1 observation wells and monitor well D2, San Antonio, Texas, July 1986-April 1987.

Figure 28. Alkalinity concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

Figure 29. Calcium concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

1986. Insufficient data were collected to determine the cause of these changes in concentrations. Calcium concentration changes were more consistent in samples from all seven monitor wells than were concentrations of any other major ion studied.

Magnesium (fig. 30) had the second largest concentration of the cations in water samples from the freshwater wells (C1, D1, and D2). Random fluctuations of magnesium concentrations were the only trends apparent for water from these three wells. A substantial increase in the magnesium concentration was measured in the water from well A2 in August 1986. The cause of this large fluctuation is unknown. The concentration of magnesium in the water from well A2 remained constant through the rest of the study. Appreciable changes in the magnesium concentration in water from wells A1, A3, and C2 are not indicated.

Sodium (fig. 31) was a major cation in the saline-water samples, but was determined only in water samples taken every 3 months. The data from wells A3 and D1 show a steady increase in sodium concentration

over time, but the data for well C1 show a decrease. The data are too few to draw any conclusions.

Sulfate concentrations were small in water from well D2 and showed little change over time (fig. 32). Sulfate made up an increasing proportion of the anions in the water from wells C1 and D1. Except for a reduction in sulfate concentration in the August 1986 sample from well D1, these data show only slight trends. Sulfate concentration increases slightly over the period in well D1 and decreases slightly over the period in well C1 samples. The causes for these small changes are unknown. Sulfate was the major anion in the water samples from the other four wells. An appreciable change did not occur over the period monitored, on the basis of limited amount of data and the analytical uncertainty.

Chloride was a minor constituent in water from the two freshwater wells (C1 and D2) and was a major constituent in water from well D1. These data are plotted in figure 33 with the chloride concentrations of water from the four saline-water wells. Substantial

Figure 30. Magnesium concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

changes in chloride concentrations were not observed in the water from the three wells with the smallest salinity (C1, D1, and D2). Water from well A1, the deepest well in the saline-water zone, had several large changes in chloride concentration through time. These were assumed to be normal inter-sample variability in the chloride concentration of the deep part of the aquifer in this part of the saline-water zone. The chloride concentration of water from well C2, the deep transition-zone well, also changed with time, but the analytical uncertainty prevents better definition of chloride changes. Water from well A2, the shallowest well in the saline-water zone, had little change in chloride concentration over time. Water from well A3 had changes in chloride concentration, but the data are too limited to infer what caused the changes.

The major ions determined in water samples from the Mission well field and Artesia 1 observation wells are plotted in figures 34a-b. Concentrations of magnesium, sulfate, and chloride are shown in figure

34a, and the alkalinity and calcium concentrations are shown in figure 34b. Calcium data have similar concentrations and magnitudes of change to those in water from well D2 (fig. 28) and also indicate no trend with time.

The data do not indicate a definite trend in time for any of the dissolved constituents. Water samples taken from the freshwater wells show greater relative changes in dissolved constituents than the water samples taken from the saline-water wells. The salinity variations in water samples from most monitor wells appear to be random and do not correlate with the water level of well D2. The specific conductance of samples from monitor well D2 and the Mission and Artesia 1 observation wells (fig. 27) indicates that the range of variability of well D2 is similar to the range of variability of the other two observation wells.

Data from wells C1 and D1 appear to have slight trends with time. The increase in specific conductance, sulfate, and chloride in samples from well D1 does not
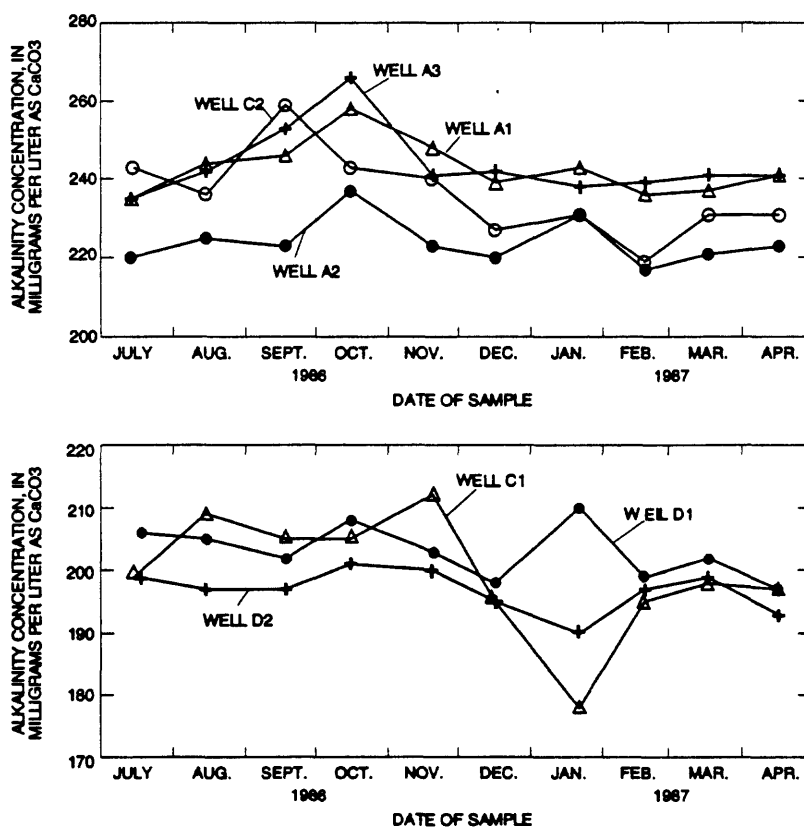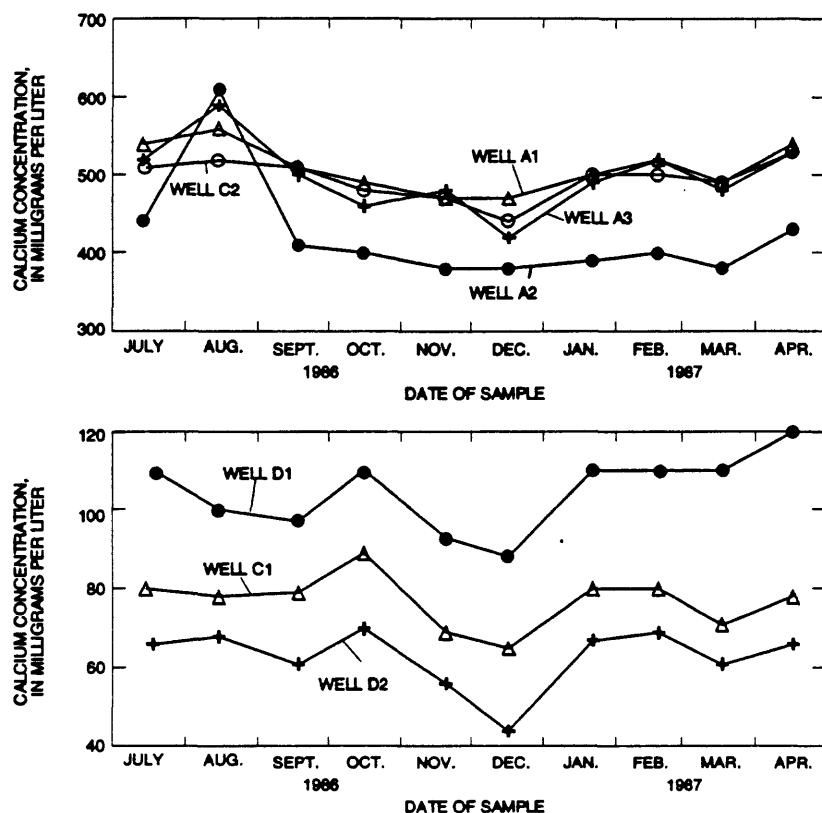
Figure 31. Sodium concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

correlate with the rising water levels in well D2. If the water levels were correlated with the change in salinity, the specific conductance of water in well D1 should decrease as the water level rises and increase as the water level declines. The specific-conductance data from well D1 tend to increase as the water level in D2 rises, exactly the opposite that would be expected from saline-water intrusion. The change from small to large values of specific conductance in the well D1 samples is about 29 percent of the median, 1,050 μS/cm (table 5).

The specific conductance, chloride, and sulfate of samples from well C1 decrease slightly during the period indicating a possible correlation between decreased salinity in water from well C1 and higher water level at well D2. The change in specific conductance in well C1 samples is much smaller relative to the opposite trend in well D1. The change is about 8 percent of the median (660 μS/cm).

Water samples were collected from all seven monitor wells on January 22, 1987. These water sam-

ples (table 7) were analyzed for a suite of trace elements, mostly metals. Before analysis, all water samples were filtered through a 0.10-μm filter to remove colloidal-size particles.

The January water samples collected had small concentrations of most of the elements tested. Arsenic concentration was less than the minimum reporting level (1 μg/L as As) in samples from wells A1, A2, C2, and D2. Arsenic was detected in samples from wells A3, C1, and D1, but was equal to or slightly above the minimum reporting level. Arsenic concentrations in samples from the wells had no apparent relation to each other.

Barium concentration was less than the minimum reporting level (100 μg/L as Ba) in samples from wells A1, A2, and A3. The minimum reporting level is greater in the saline-water samples (A1, A2, A3, and C2) than in the freshwater samples (C1, D1, and D2) because of the dissolved-solids concentrations. Barium was detected in samples from wells C1, C2, D1, and
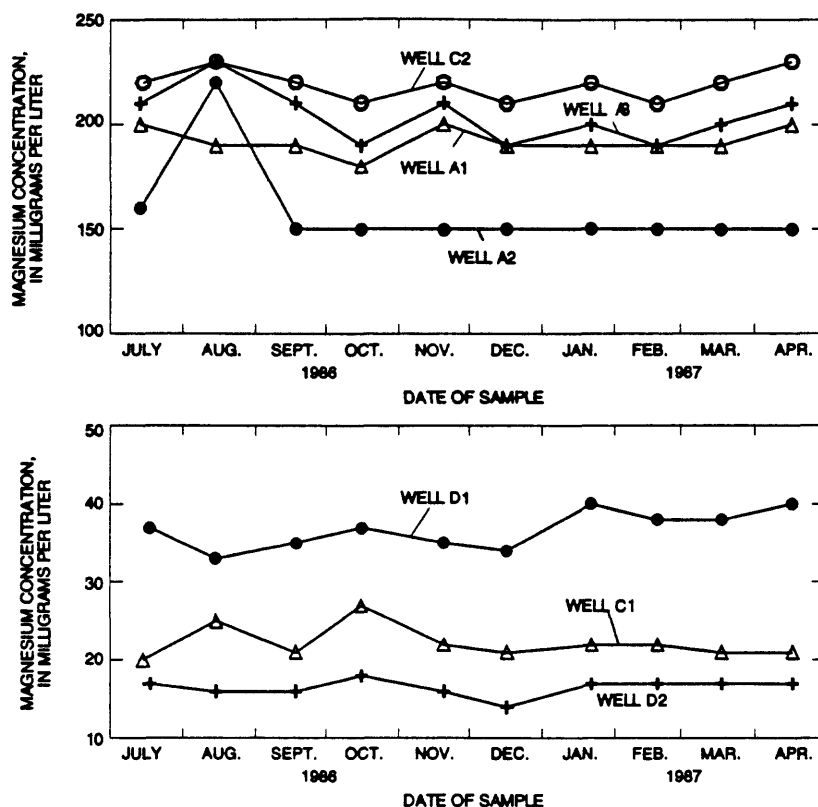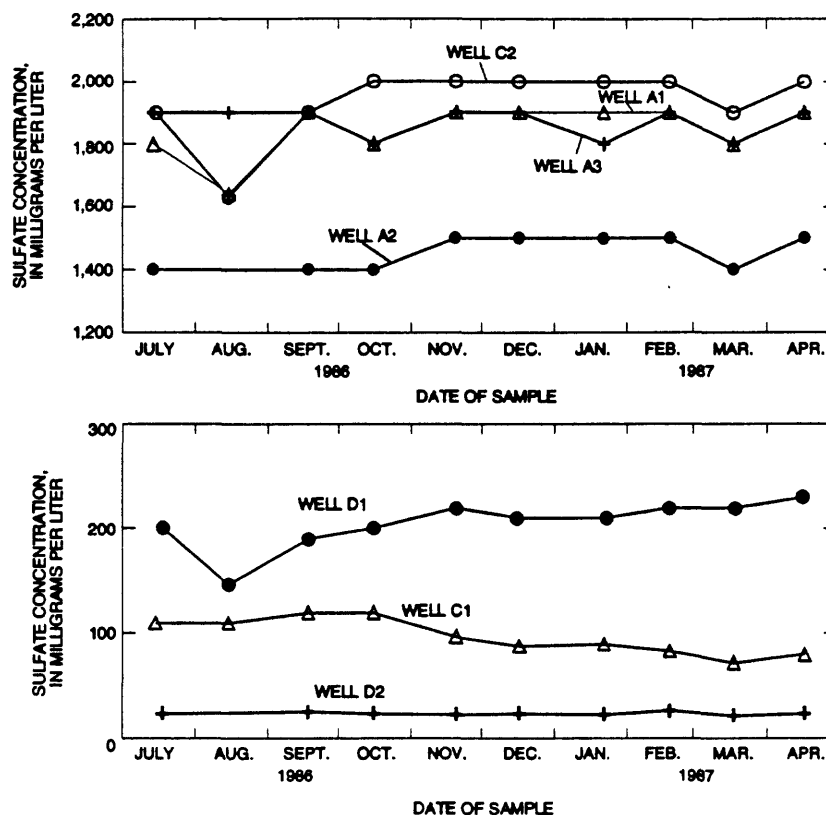
Figure 32. Sulfate concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

D2. The largest concentration was detected in the sample from well C2.

Differences in concentrations of dissolved iron and manganese are evident among samples from the wells. Water samples from wells A1, A2, and A3 had the greatest concentrations of manganese. Pearson and Rettman (1976) collected data on the oxidation/reduction conditions in the freshwater and saline-water zones. Pearson and Rettman (1976) indicated that the saline water was substantially reduced and contained hydrogen sulfide gas. Iron and manganese are more soluble in the reduced state, except in the presence of dissolved sulfide. However, the variations in the iron and manganese concentrations in water from the monitor wells apparently are unrelated to the presence of hydrogen sulfide.

Ferric iron staining is common in the freshwater part of the aquifer and pyrite (FeS_2) is present in the saline-water zone. The water sample from well D2 had the smallest concentrations of iron. The smallest concentrations of manganese were in water samples from

freshwater wells C1 and D2. The largest concentrations of manganese were in water samples from wells A1, A2, and A3, which had the most reducing conditions. The manganese concentrations in water samples from wells C2 and D1 were intermediate, indicating a transition from greater manganese solubility in the saline-water zone to lesser solubility in the freshwater zone.

The iron concentrations indicate that other controls in the saline-water zone also could be affecting the solubility of iron in the reducing water. Iron sulfides are relatively insoluble. Hydrogen sulfide and sulfide concentrations were not measured in samples from the monitor wells, but probably were present because a strong odor of hydrogen sulfide was emitted from the water. Dissolved sulfide, associated with hydrogen sulfide, probably controls the level of dissolved ferrous iron in the saline water. This can be seen by comparing the levels of dissolved iron in the samples from saline-water wells A1, A2, and A3, the shallow transition-zone well C1, and the deep freshwater-zone well D1.
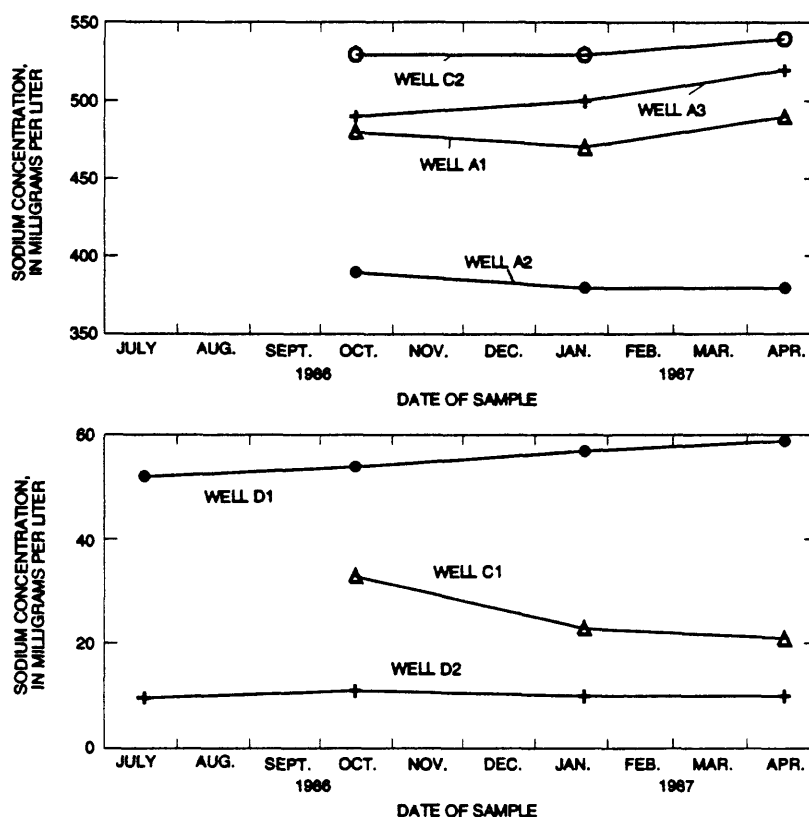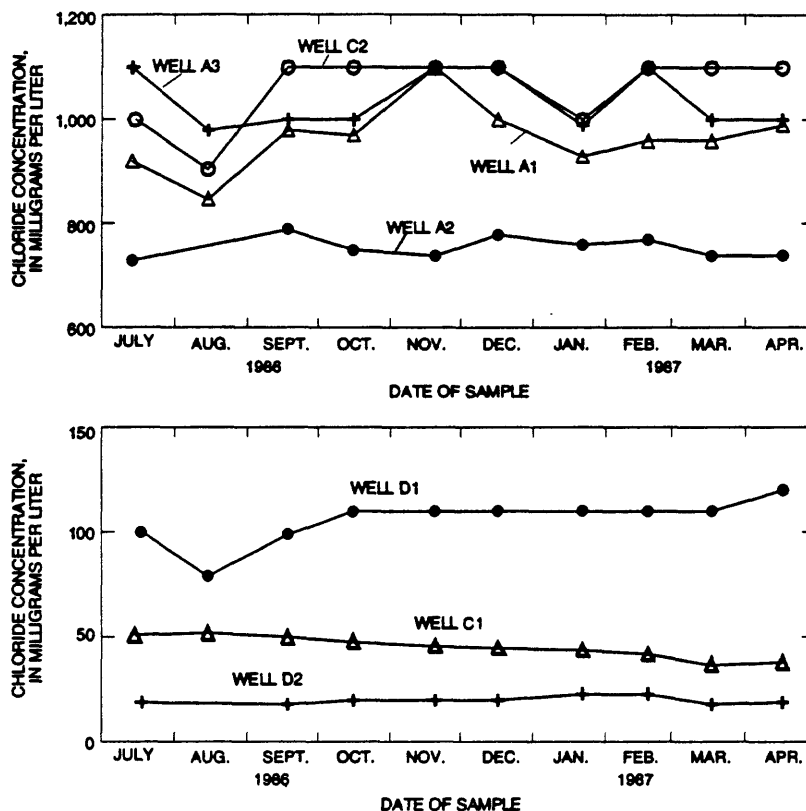
Figure 33. Chloride concentration in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

The largest iron concentrations were in samples from well D1. Buszka (1987) and Pearson and Rettman (1976) reported that all the freshwater sampled in the Edwards aquifer contains 2 mg/L or more of dissolved oxygen. Although well D1 yields freshwater, the monitor interval spans the downdip limit of freshwater · between oxygen rich freshwater and sulfide rich saline water. The large concentration of iron could be caused by the mixing of these two chemical environments and subsequent reactions. The water from well C1 contained some hydrogen sulfide, but apparently it had little effect on the iron concentration in the water from that well.

Water from wells A2 and D2 was analyzed on April 17, 1987, for a slightly different suite of major cations and trace elements (table 8). These analyses were conducted to determine the differences or similarities between the saline water (well A2) and the freshwater (well D2) with respect to these trace metals at lower minimum reporting levels. Barium concentra-

tions were greater in the freshwater than in the saline water. The concentration of barium in the April 17 sample from well D2 was the same as that in the January 22 sample.

Boron, cadmium, lead, and molybdenum were present at similar small concentrations in saline water from well A2 and freshwater from well D2. Concentrations of chromium, copper, lithium, manganese, strontium, and zinc were detected in the saline-water and freshwater samples, but the concentrations were substantially greater in the saline water—particularly lithium, manganese, and zinc. The sources of chromium, copper, lithium, and zinc are unknown. The source of the large concentrations of manganese could be the dissolution of manganese sulfides associated with pyrite. Carbonate rocks are the source of the large concentrations of strontium. The minerals that compose limestone and dolostone contain varying amounts of strontium (Kulp and others, 1952). The larger concentration of strontium in the saline water is caused by
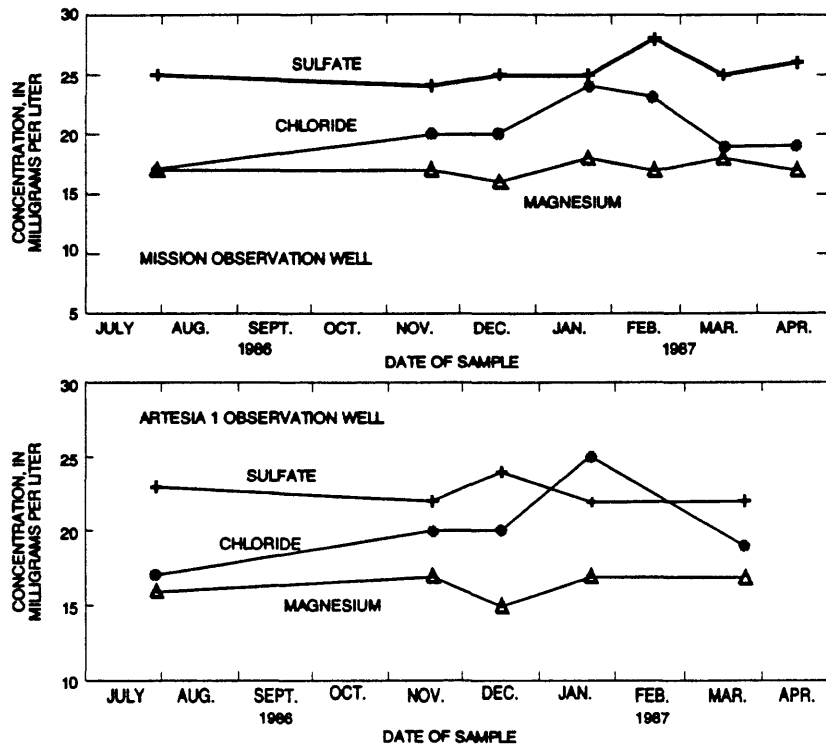
a. Magnesium, sulfate, and chloride
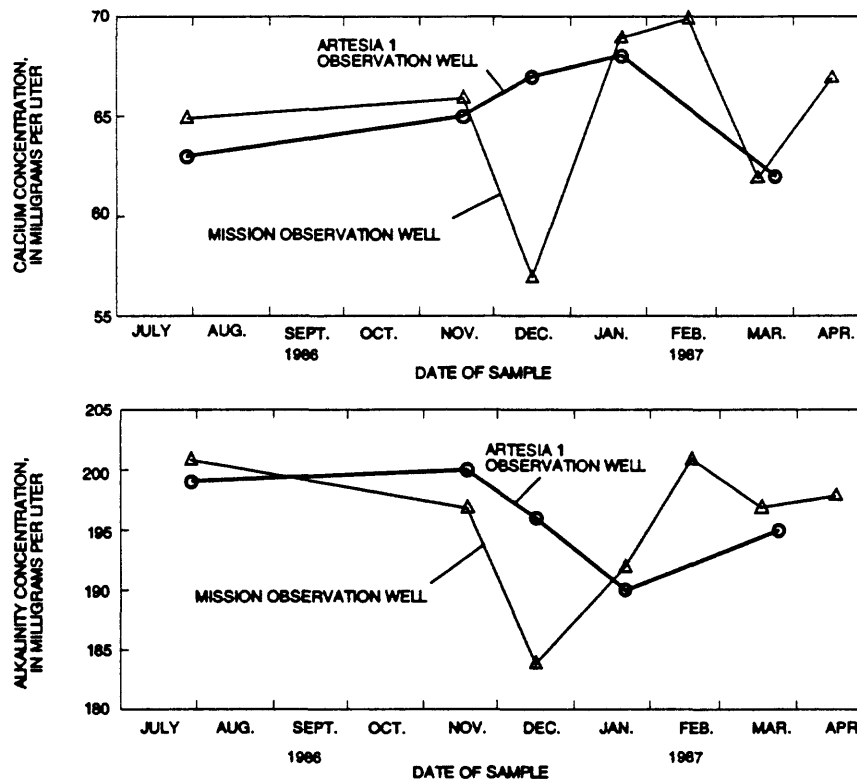
b. Calcium and alkalinity

Figure 34. Concentrations of dissolved ions in water samples from Mission and Artesia 1 observation wells, San Antonio, Texas, July 1986-April 1987

Table 7. Concentrations of selected trace elements in samples from monitor wells, San Antonio, Texas, January 22, 1987

[μg/L, micrograms per liter; <, less than]

Well number	Arsenic, dissolved (μg/L as As)	Barium, dissolved (μg/L as Ba)	Cadmium, dissolved (μg/L as Cd)	Chromium, dissolved (μg/L as Cr)	Copper, dissolved (μg/L as Cu)	Iron, dissolved (μg/L as Fe)
A1	<1	<100	1	<10	<1	120
A2	<1	<100	1	<10	<1	120
A3	1	<100	1	<10	<1	90
C1	2	84	<1	<10	2	450
C2	<1	200	1	10	<1	100
D1	1	110	<1	<10	<1	1,400
D2	<1	110	<1	<10	<1	38

Well number	Lead, dissolved (μg/L as Pb)	Manganese, dissolved (μg/L as Mn)	Selenium, dissolved (μg/L as Se)	Silver, dissolved (μg/L as Ag)	Zinc, dissolved (μg/L as Zn)
A1	<5	70	<1	<1	<10
A2	<5	50	<1	<1	30
A3	<5	60	<1	<1	30
C1	<5	4	7	<1	33
C2	<5	40	<1	<1	30
D1	<5	17	<1	<1	15
D2	<5	7	1	<1	40

the lack of freshwater diagenesis or longer residence time. The iron concentration was less in the saline-water sample (well A2) than in the freshwater sample (well D2), probably because of the presence of sulfide in the saline water.

Evolution of Water Chemistry

The water chemistry in the freshwater zone of the Edwards aquifer does not change substantially from the recharge to the discharge areas. The chemical evolution of water that enters the saline-water zone has been interpreted by previous studies through the use of ion ratios and graphical comparisons.

Molar Ratios of Major Ions

Pearson (1973) noted that the ratio of dissolved sulfate to dissolved chloride was variable in the freshwater zone primarily because of the variability in the amount of dissolved chloride in the recharge water. The molar ratios of sulfate to chloride were compared with molar ratios of magnesium to calcium in the water samples taken from the test holes at sites A, C, and D. The

Table 8. Concentrations of selected dissolved major and trace ions and silica in samples from wells A2 and D2, San Antonio, Texas, April 17, 1987[1]

[mg/L, milligrams per liter; µg/L, micrograms per liter]

Constituent	Units	Well A2 (saline water)	Well D2 (freshwater)
Calcium	mg/L	490	68
Magnesium	mg/L	160	17
Sodium	mg/L	380	9.8
Silica	mg/L	19	11
Barium	µg/L	12	110
Boron	mg/L	1.3	.78
Cadmium	µg/L	.7	.4
Chromium	µg/L	4.1	1.0
Copper	µg/L	9.2	.6
Iron	µg/L	62	170
Lead	µg/L	1.1	1.8
Lithium	µg/L	800	22
Manganese	µg/L	37	5.7
Molybdenum	µg/L	3.6	3.9
Strontium	mg/L	11	1.5
Zinc	µg/L	16	5.4

[1] Analyses made by J. Garbarino, U.S. Geological Survey.

ratio of dissolved magnesium to calcium could be indicative of the relative amount of dolomite in contact with the water in the rocks, and the ratio of sulfate to chloride could be an indicator of the relative importance of gypsum dissolution on the water chemistry, particularly in the saline-water zone.

The plot of the ratios from the cumulative-depth flow tests at site A is shown in figure 35. Water samples from the tests showed little change in the two ion ratios as the sampled depth interval increased. The data points are numbered for the respective cumulative-depth test sample. The number increases with increasing depth and the letters refer to the samples from the single-packer or interval flow tests. Samples 1 and 3, taken from the intervals between depths of 965 and 1,123 ft, are isolated from the main group of samples

indicating that the water from the top 50 to 150 ft of the aquifer at the saline-water zone is of a different type. The specific conductance of these samples shown in figure 24a indicates that the water from near the top of the aquifer at site A is less saline than the water from the rest of the aquifer at this zone. Samples 2 and δ from 965 to 1,075 ft, that includes the intervals of samples 1 and 3, are not grouped with these two samples. The reason for the larger molar ratio of magnesium to calcium of samples 2 and δ with respect to samples 1 and 3 is not known.

The relation between the molar ratios of sulfate to chloride and the molar ratios of magnesium to calcium in water samples from the cumulative-depth and interval flow tests from the site C test hole is shown in figure 36. Data from three interval flow test samples (α,
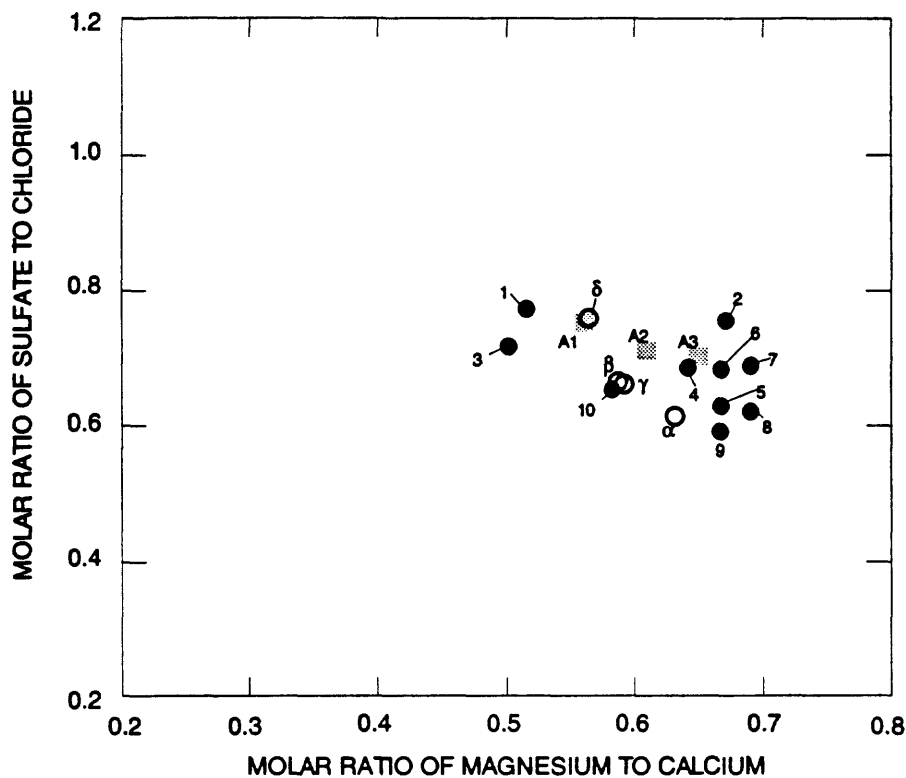
EXPLANATION

Sample number	Depth interval, in feet below land surface

CUMULATIVE-DEPTH FLOW TESTS

1	965-1,019
2	965-1,071
3	965-1,123
4	965-1.175
5	965-1,228
6	965-1,279
7	965-1,331
8	965-1,384
9	965-1,437
10	965-1,489

INTERVAL FLOW TESTS

α	1,276-1,489
β	1,180-1,489
γ	965-1,180
δ	965-1,075

MONITOR-WELL MEANS

A1	1,200-1,275
A2	1,001-1,075
A3	1,099-1,175

Figure 35. Relation between the molar ratio of sulfate to chloride and the molar ratio of magnesium to calcium in water samples from the cumulative-depth and interval flow tests and monitor-well sample means, site A, San Antonio, Texas.

γ, and ε in the right half of fig. 36) plot in the same area as most of the points for the site A samples in figure 35. Sample intervals α, γ, and ε are all below 859 ft depth. The specific conductance of these samples (table 3) indicates that all the water produced below 859 ft (below the freshwater producing cavern) is saline and of a different type than the water produced from the cavern. Samples β, δ, and ζ are from depth intervals that included the freshwater cavern (832 to about 840 ft depth) and are representative only of this interval because flow from the cavern diluted the flow from below it. Therefore, all the cumulative-depth samples also plot in the same area on figure 34 as do samples β, δ, and ζ.

The relation between the molar ratios of sulfate to chloride and the molar ratios of magnesium to calcium of the site D samples from the cumulative-depth and interval flow tests is shown in figure 37. All samples, except for 3, α, and β, are in the lower center of the graph. Data points α and β are from interval flow test samples from the two saline-water intervals that did not include the section of test hole above a depth of 1,158 ft. Samples α and β plot in the same area as do most of the saline-water-zone samples (fig. 35). At site D, the freshest water is near the top of the aquifer, and saline water, with specific conductance between 5,500 and 6,500 μS/cm, is near the bottom. The large ratio of sulfate to chloride of sample 3 could be the result of sampling or analytical error. The major difference between the site C water quality and the site D water quality is the water chemistry type and the section of the aquifer that produces freshwater. The fluid resistivity curve of the site D test hole (fig. 18) indicates that most of the section of aquifer from the bottom of the Georgetown Formation to about the middle of the Kainer Formation contains freshwater.

The pattern of the data plotted in figure 37 indicates that this type of plot is an important means to
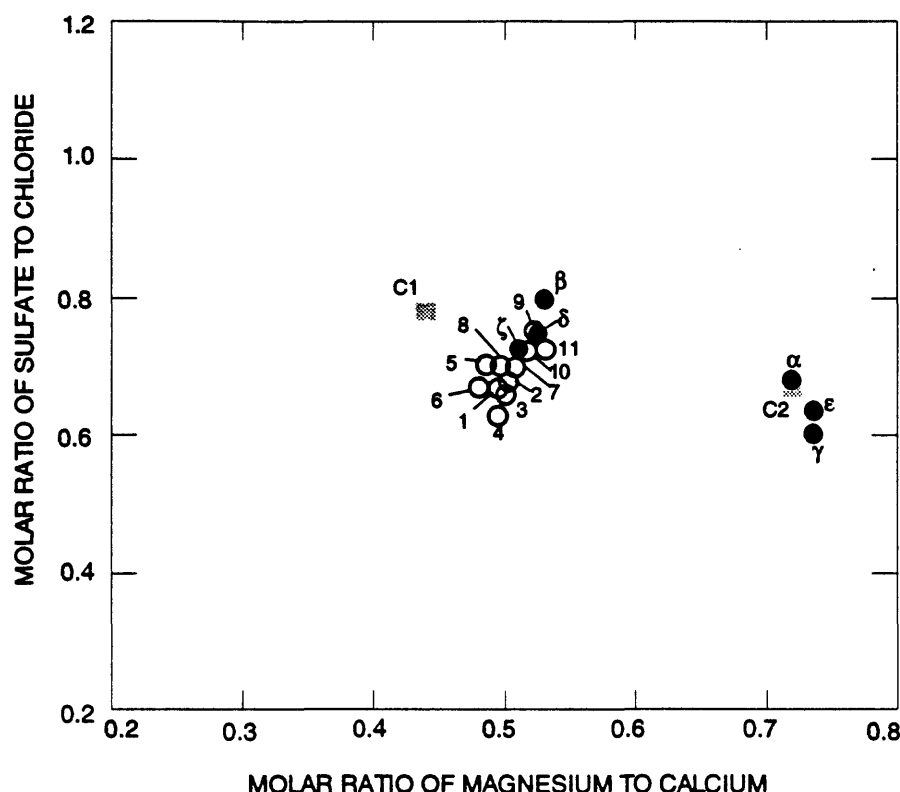
EXPLANATION

Sample number	Depth interval, in feet below land surface
CUMULATIVE-DEPTH FLOW TESTS	
1	832- 891
2	832- 941
3	832- 991
4	832-1,042
5	832-1,095
6	832-1,147
7	832-1,199
8	832-1,251
9	832-1,302
10	832-1,352
11	832-1,396
INTERVAL FLOW TESTS	
α	859-1,396
β	832- 859
γ	1,056-1,396
δ	832-1,056
ε	1,240-1,396
ζ	832-1,240
MONITOR-WELL MEANS	
C1	840- 881
C2	1,072-1,150

Y-axis: MOLAR RATIO OF SULFATE TO CHLORIDE (0.2 to 1.2)
X-axis: MOLAR RATIO OF MAGNESIUM TO CALCIUM (0.2 to 0.8)

Figure 36. Relation between the molar ratio of sulfate to chloride and the molar ratio of magnesium to calcium in water samples from the cumulative-depth and interval flow tests and monitor-well sample means, site C, San Antonio, Texas.

determine if saline-water intrusion occurs in the aquifer near these monitor wells. Movement of saline water toward water-supply wells in the freshwater zone might not show large increases in dissolved solids, but the relative proportions of the various ions, as used in the plot, could show distinct changes.

The molar ratios of the mean concentrations of sulfate to chloride and magnesium to calcium from the five water-quality groups (fig. 6) defined by Pearson and Rettman (1976) are shown in figure 38 with the ratios of the mean concentrations from the monitor-well samples. Samples from site A are similar to the western saline and transitional group, as is the sample from the freshwater well D1 that includes the downdip limit of freshwater. The water sample from freshwater well D2 yielded ratios that are close to those of the main freshwater group. The water sample taken from transition-zone well C1 yielded ratios between those of the main freshwater and western saline and transitional groups.

A plot of the ratios of ions from the water samples collected in the completed monitor wells from July 1986-April 1987 is shown in figure 39. Samples from the three saline-water-zone monitor wells (A1, A2, and A3) plot in the same general area as discrete-interval samples from the saline-water-zone test hole. Samples from well C2 are similar to the other saline-water well samples, but have a slightly larger ratio of magnesium to calcium. Samples from well D1 (slightly saline water) are similar to other saline-water well samples (A1, A2, A3, and C2), but have a slightly smaller ratio of magnesium to calcium. The difference between C2
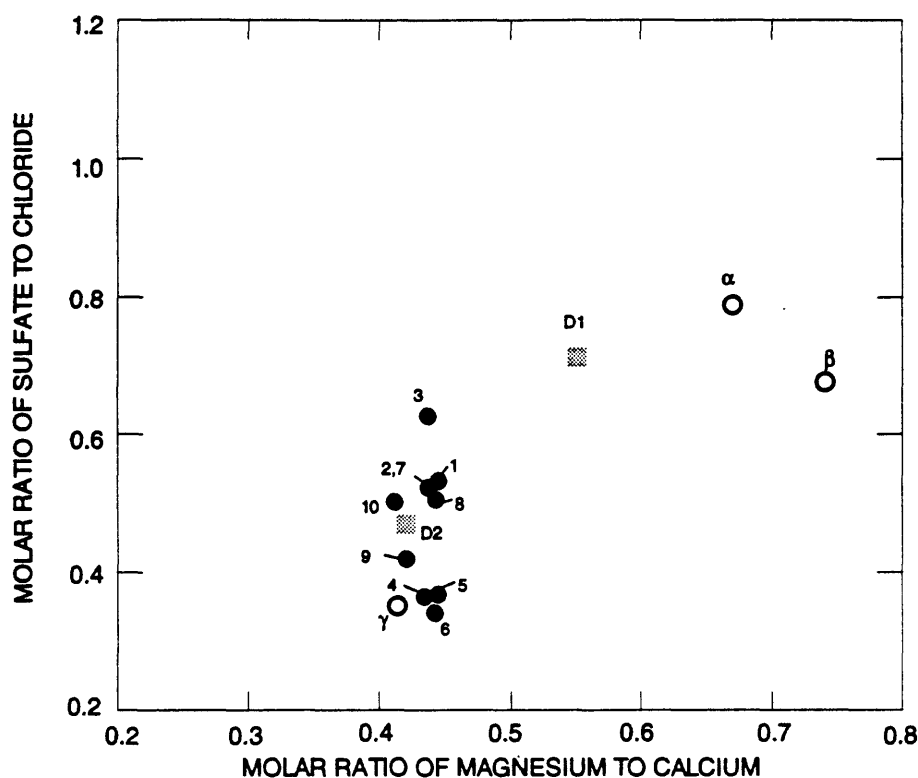
Figure 37. Relation between the molar ratio of sulfate to chloride and the molar ratio of magnesium to calcium in water samples from the cumulative-depth and interval flow tests and monitor-well sample means, site D, San Antonio, Texas.

and D1 could be the result of more dolomite in the rocks along the flowpath to well C2 relative to the amount of dolomite in the rocks along the flowpath to well D1, which would increase the magnesium concentration relative to calcium concentration. The ratio of sulfate to chloride is nearly constant for these five · saline-water wells, which agrees with a similar observation made by Pearson (1973) for most of the saline-water zone near the downdip limit of freshwater. Three discrete clusters of samples are indicated in figure 39. Two of the clusters represent water samples from two freshwater wells (C1 and D2). The third, an elongate cluster of samples in figure 39, represents water samples from D1 and the four saline-water wells (A1, A2, A3, and C2).

Graphical Comparison

Graphical comparison of concentrations of dissolved constituents in water samples is useful to interpret the evolution of water chemistry and to identify and define the types of chemical reactions that take place in an aquifer.

Major ions in water samples taken from monitor wells April 17, 1987, are shown on a Piper diagram in figure 40. The cation-anion balance errors for these analyses are listed in table 9. The data indicate increasing dissolved sodium and potassium relative to calcium and magnesium (fig. 40, left to right in the lower left triangular graph) and increasing dissolved chloride and sulfate relative to bicarbonate ion (fig. 40, left to right in the lower right triangular graph). The overall pattern in the water-sample data, from calcium magnesium bicarbonate dominated water to calcium sodium sulfate dominated water with some chloride, is indicated by the linear pattern from sample D2 to the four saline-water samples (A1, A2, A3, and C1) shown in the middle quadrilateral graph of figure 40. The local geochemical evolution just described for these seven monitor-well water samples is similar to the regional

Figure 38. Relation between the molar ratio of sulfate to chloride and the molar ratio of magnesium to calcium in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, and for the mean concentrations of five water-quality groups.

geochemical evolution described by Pearson (1973) and Rightmire and others (1974) for the development of Edwards aquifer saline-water chemistry from Edwards aquifer freshwater chemistry. Pearson (1973) and Rightmire and others (1974) postulated that the evolution occurred at or near the downdip limit of freshwater by mixing of Edwards aquifer freshwater (similar to the main freshwater group of Pearson and Rettman, 1976) with one of two saline-water "brines" originating tens of miles downdip in the saline-water zone.

The water type from well D1 could have resulted from the mixture of freshwater and saline water similar to the mixing reaction postulated by Pearson (1973) and Rightmire and others (1974). However, because the screened interval of D1 intersects the downdip limit of freshwater and likely draws from both the freshwater zone and the saline-water zone, water drawn from well D1 is an artificially induced mixture of the saline water and freshwater and might not represent natural conditions in the aquifer.

The composition of the water sample from well C1 is not quite intermediate between the composition of D2 and D1. The water composition of C1 cannot be explained as readily by mixing because it is drawn from a cave that is, at least locally, hydraulically isolated from both the freshwater zone tapped by well D2 and the saline-water zone tapped by C2.

Salt-Normative Assemblages

Assessing the probability of saline-water intrusion can be more accurate by increasing the understanding of the origin and nature of the salinity in the Edwards aquifer. One means that can be used to characterize the salinity and determine to some extent the origins of the salinity is to compute a salt-normative assemblage for each of the various samples.

A technique that computes the salt-normative (salt-norm) assemblage of minerals for a given water chemistry, SNORM (Bodine and Jones, 1986), was used to help define the evolution of saline-water-zone water chemistry. The salt norm is the quantitative ideal
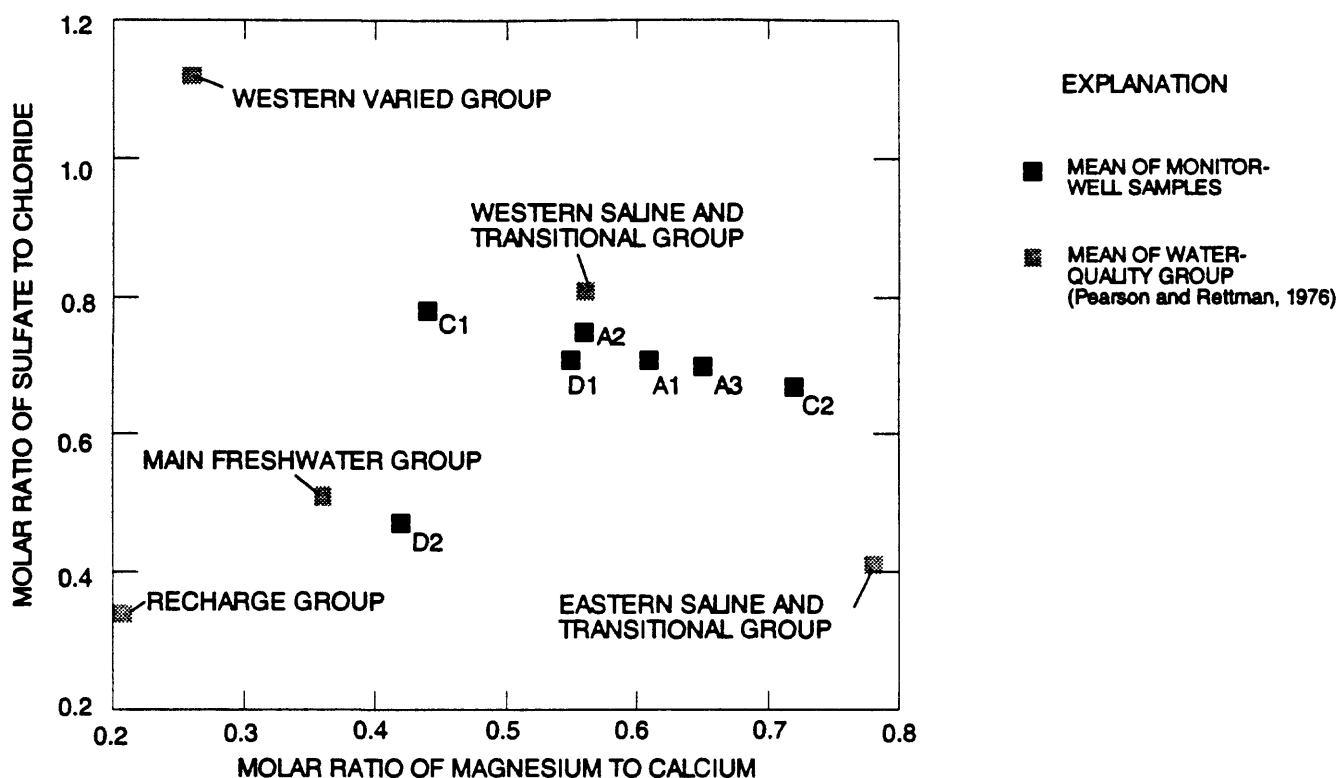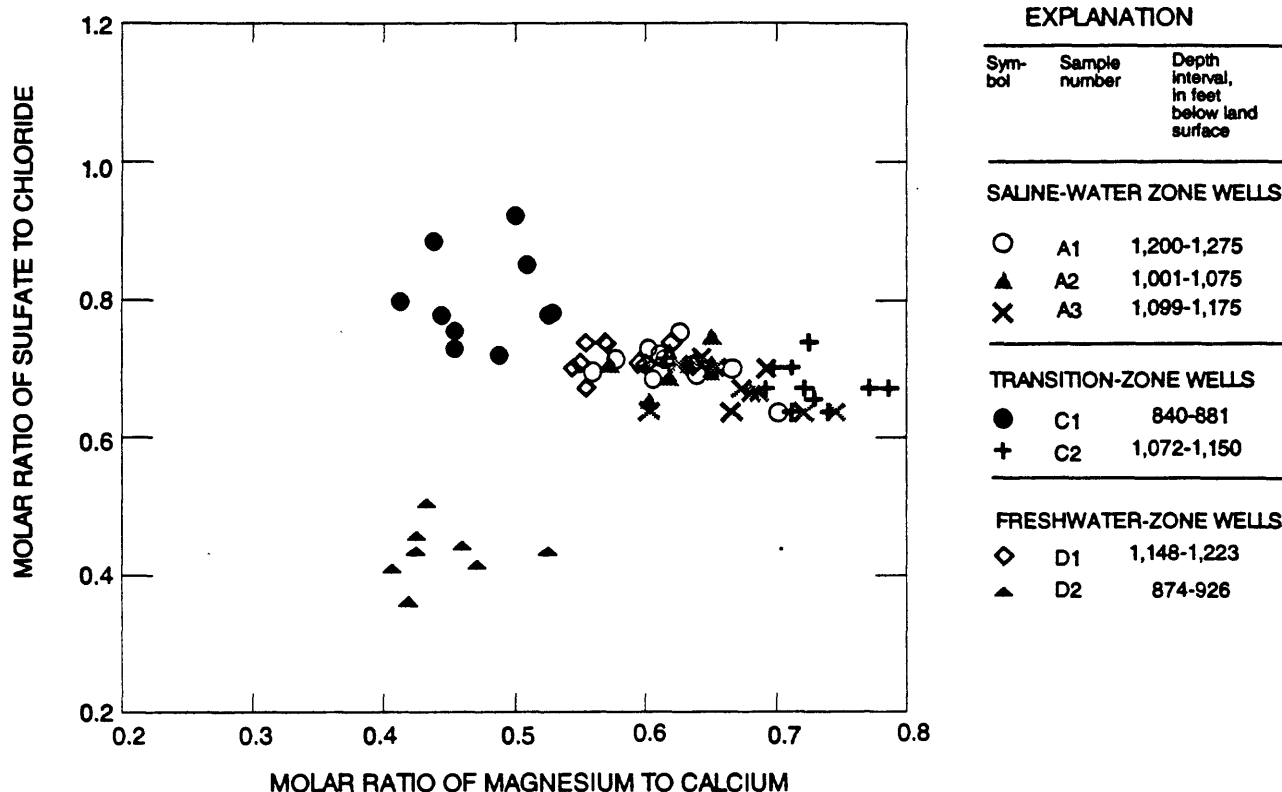
Figure 39. Relation between the molar ratio of sulfate to chloride and the molar ratio of magnesium to calcium in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, July 1986-April 1987.

equilibrium assemblage that would crystallize if the water evaporated to dryness at 25 °C and 1 bar total pressure and atmospheric partial pressure of CO_2. The results from SNORM help to identify three water types: (1) meteoric or weathering water types that are characterized by normative alkali bearing sulfate and carbonate salts; (2) connate marine-like water types that are chloride rich with a halite anhydrite kieserite bischofite carnallite association; and (3) diagenetic water types that are frequently of marine origin but yield normative salts, such as calcium bearing chlorides and sylvite, which indicate solute alteration by secondary mineral reactions.

Salt norms might help to: (1) identify lithologic sources of dissolved solids; (2) identify the origin of connate water types from normal marine, hypersaline, or evaporite mineral resolution processes; and (3) distinguish between dolomitization and silicate hydrolysis or exchange for the origin of diagenetic water types. Bodine and Jones (1986) examined hundreds of water analyses using this technique and developed several

relations that are used here to classify the water chemistry of the monitor wells and the regional chemistry of the saline-water zone, and to help determine the origin of these water types.

Percentage of anhydrous weight of the salt-norm assemblages for the water samples taken from the site A test hole is shown in figure 41. The assemblages for these samples do not differ appreciably over the intervals sampled. The assemblage, in order of decreasing percentage of anhydrous weight, is generally about 45 percent anhydrite ($CaSO_4$), 28 to 30 percent halite (NaCl), 10 to 15 percent kieserite ($MgSO_4•H_2O$) plus bischofite ($MgCl_2•6H_2O$), 5 to 10 percent magnesite ($MgCO_3$), and 3 to 4 percent carnallite ($KMgCl_3•6H_2O$) plus tachyhydrite ($CaMg_2Cl_6•12H_2O$).

Percentage of anhydrous weight of the salt-norm assemblages of minerals computed for the water samples taken from the site C test hole is shown in figure 42. The 11 cumulative-depth samples show little variation because most of the water produced by these intervals was from the cavernous zone at the top of the
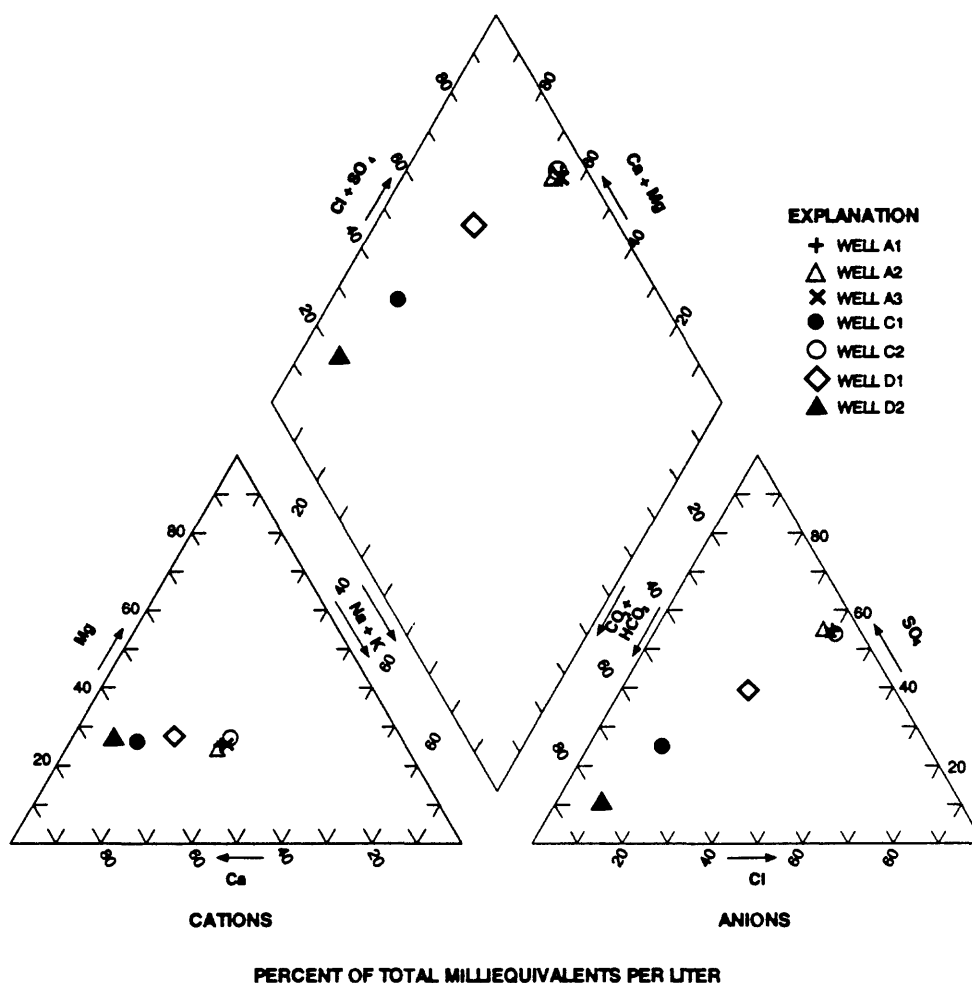
Figure 40. Piper diagram showing major ions in water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, April 17, 1987.

Table 9. Cation-anion balance errors for the analyses of samples from monitor wells, San Antonio, Texas, April 17, 1987

Well number	Error (percent)
A1	-5.1
A2	-5.3
A3	-4.0
C1	-.84
C2	-5.3
D1	-.87
D2	2.5

Figure 41. Percentage of anhydrous weight of the salt-normative assemblage of minerals computed for water samples from site A test hole, San Antonio, Texas.

Figure 42. Percentage of anhydrous weight of the salt-normative assemblage of minerals computed for water samples from site C test hole, San Antonio, Texas.

aquifer. In order of decreasing percentage of anhydrous weight, the assemblage is generally about 40 to 45 percent dolomite, 30 to 36 percent anhydrite, 17 to 20 percent halite, 1 to 5 percent calcite, 1 to 3 percent carnallite plus tachyhydrite (all carnallite in these samples), and negligible amounts of other minerals. Weights of less than 1 percent are not plotted or discussed. This general assemblage is quite different from the assemblages of the site A test-hole samples. The site C assemblage has less anhydrite, halite, and carnallite plus tachyhydrite, and no kieserite plus bischofite. Tachyhydrite is not present in any of the assemblages of the cumulative-depth samples from the transition zone (site C). Considerable dolomite and some calcite are present; neither are present in the saline-water-zone assemblage site A.

Three of the six assemblages computed from the interval flow samples in the site C test hole (α, γ, and ϵ) are distinctly different from the computed salt-norm assemblages from the cumulative-depth (1-11) sample. They show no dolomite but contain 15 to 20 percent kieserite plus bischofite, which are absent in cumulative-depth samples from site C flow-test samples. The interval flow samples, from below the cavern, have assemblages similar to those of site A. Assemblages computed from water samples from intervals that included the cavern (β, δ, and ζ) are similar to those computed from the cumulative-depth samples. This is caused by the flow of water from the cavern.

The salt-norm assemblages computed for the site D cumulative-depth and interval flow samples are shown in figure 43. The general assemblage for the cumulative-depth flow test samples, in order of decreasing percentage of anhydrous weight, is about 47 to 50 percent dolomite, 25 to 30 percent calcite, 10 to 13 percent anhydrite, 10 percent halite, about 1 percent carnallite plus tachyhydrite (all carnallite in these samples), and no kieserite plus bischofite. This general assemblage also is different from that of the site C cumulative-depth assemblage. The assemblages computed for the interval flow test samples that included only the lower section of the aquifer (α and β) were similar to site A assemblages.

The salt-norm assemblages computed for the April 17, 1987, samples from the monitor wells and the assemblages computed for the mean concentrations of the five water-quality groups (Pearson and Rettman, 1976) are shown in figure 44. Water samples from the three saline-water-zone wells (A1, A2, and A3) and well C2 are similar to the general assemblage com-

puted for the interval samples from the site A test hole shown in figure 41, with the exception of kieserite plus bischofite. The eastern saline and transitional and western saline and transitional groups (fig. 6) from Pearson and Rettman (1976) also are similar to the assemblage of the site A test hole. This similarity implies that all of these water types had similar origins. The western saline and transitional group assemblage is closest to the site A well assemblage. The well D2 assemblage is similar to the assemblage for the main freshwater group, with the exception of kieserite plus bischofite. The well C1 and western varied group assemblages apparently are intermediates between the assemblages of saline water and freshwater.

On the basis of work by Bodine and Jones (1986), the minerals that compose these salt-norm assemblages can help identify the origins of, and the relations among, the several water types of these water samples. The presence of the halite anhydrite kieserite bischofite carnallite association in the site A assemblages and in the assemblages computed for the saline and transitional and varied groups of Pearson and Rettman (1976) indicates that there might be a component of marine-like water in these assemblages (fig. 44). The excess of anhydrite over halite, relative to the mean seawater assemblage (plotted in fig. 46) from Bodine and Jones (1986), indicates that a substantial amount of anhydrite or gypsum dissolution occurs in and near the saline-water zone of the aquifer relative to halite dissolution.

On the basis of water samples from oil or gas wells (fig. 45), Land and Prezbindowski (1981) concluded that there is upfault or updip movement of diagenetic water types into the Edwards aquifer in the saline-water zone. Land and Prezbindowski (1981) did not analyze for dissolved sulfate, but stated that sulfate concentration generally is 300 mg/L or less in samples from the Edwards aquifer saline-water-zone oil or gas wells. Rye and others (1981) reported dissolved sulfate concentration of 2,870 mg/L obtained from an oil-field well that was sampled by Land and Prezbindowski (1981). The concentration of dissolved sulfate in this well sample is much larger than 300 mg/L and approximately at the saturation level (with respect to dissolved calcium) for anhydrite or gypsum. The analysis (Rye and others, 1981) indicates that appreciable concentrations of dissolved sulfate might be present in other samples from oil or gas wells. Therefore, dissolved sulfate concentrations were estimated for the analytical

Figure 43. Percentage of anhydrous weight of the salt-normative assemblage of minerals computed for water samples from site D test hole, San Antonio, Texas.

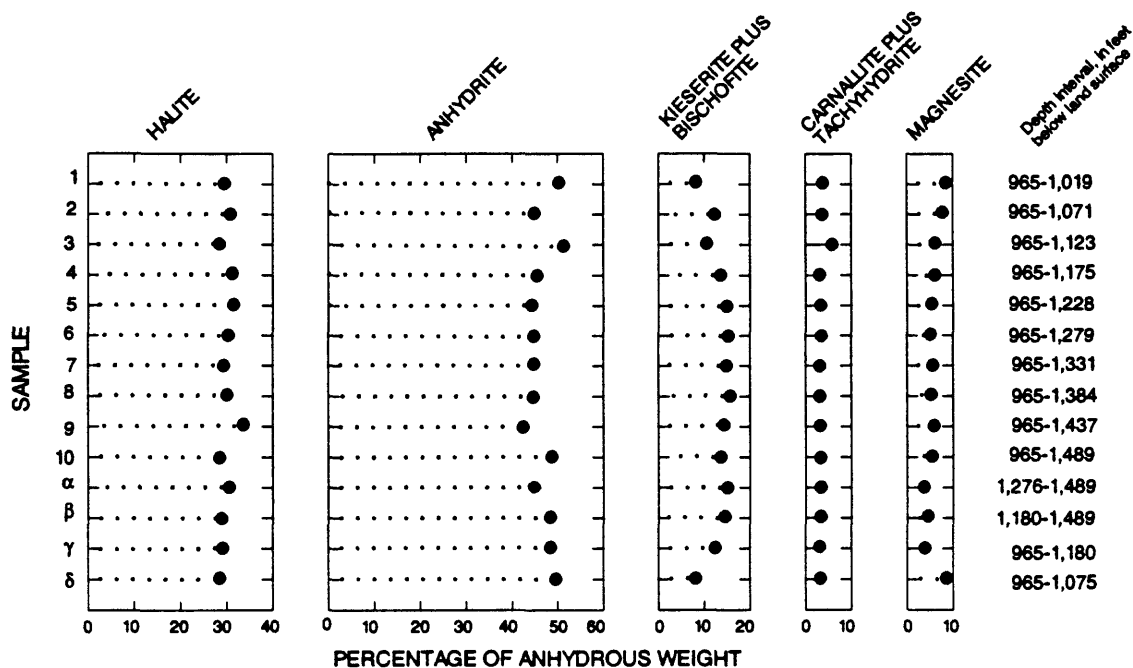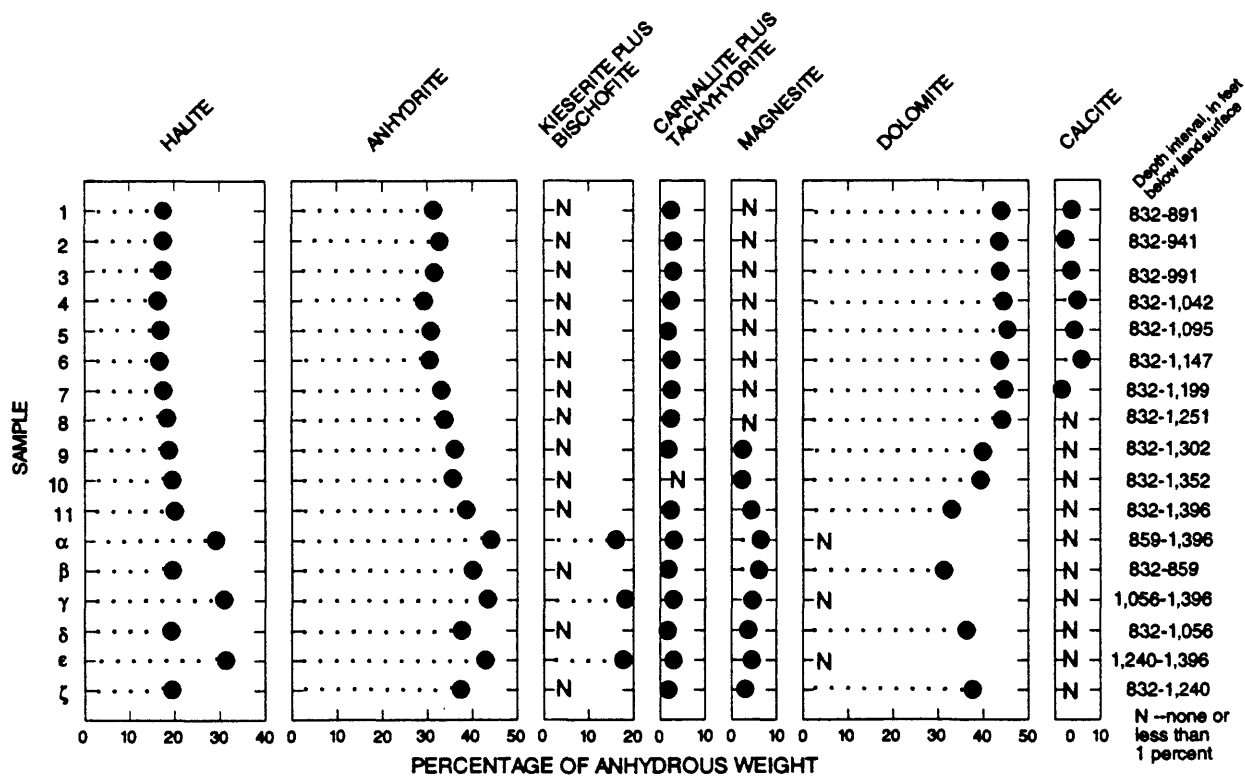**Figure 44.** Percentage of anhydrous weight of the salt-normative assemblage of minerals computed for water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, April 17, 1987, and the mean concentrations of the five water-quality groups of Pearson and Rettman (1976).

data sets of Land and Prezbindowski (1981) used in this study.

Because the level of dissolved sulfate obtained by Rye and others (1981) was approximately at the saturation level for anhydrite or gypsum, the upper limit for estimated concentrations of dissolved sulfate was the gypsum saturation level as determined by the computer program WATEQ4F (Ball and others, 1991) for the temperature and ionic strength of the water sample. Estimated concentrations of dissolved sulfate were constrained further so that the estimates would not result in a cation-anion charge imbalance nor exacerbate any imbalance that exists in the unadjusted data.

Initial charge imbalances were 5 percent or less in all the water-quality analyses used from Land and Prezbindowski (1981). Charge imbalances were still 5 percent or less after accounting for estimated concentrations of dissolved sulfate. Estimated dissolved sulfate concentrations for the well samples are: 200 mg/L

for Lodge 2 sample; 1,000 mg/L for the Knobloch 1 sample; 1,300 mg/L for the Strawn 1 and E.J. Pruitt 36 samples. The imposition of dissolved sulfate concentrations on these sample analyses did not substantially affect (less than 1-percent change) the cation-anion balance of the Lodge 2 and Strawn 1 analyses. The charge imbalances of the Knobloch 1 and E.J. Pruitt 36 sample analyses improved to 0.0 with the imposed dissolved sulfate estimates. The Land and Prezbindowski (1981) data sets, including estimated concentrations for dissolved sulfate, were used in the following discussion and in figure 46 presenting salt-norm assemblages, unless otherwise noted.

The mean seawater assemblage, the assemblages of four representative samples from Land and Prezbindowski (1981), and the salt-norm assemblages of the April 17, 1987, samples of the seven monitor wells are shown in figure 46. The salt-norm assemblages for A1, A2, A3, and C2 contain much greater

Figure 45. Approximate location of oil and gas fields and selected wells sampled by Land and Prezbindowski (1981) in the Edwards aquifer saline-water zone.
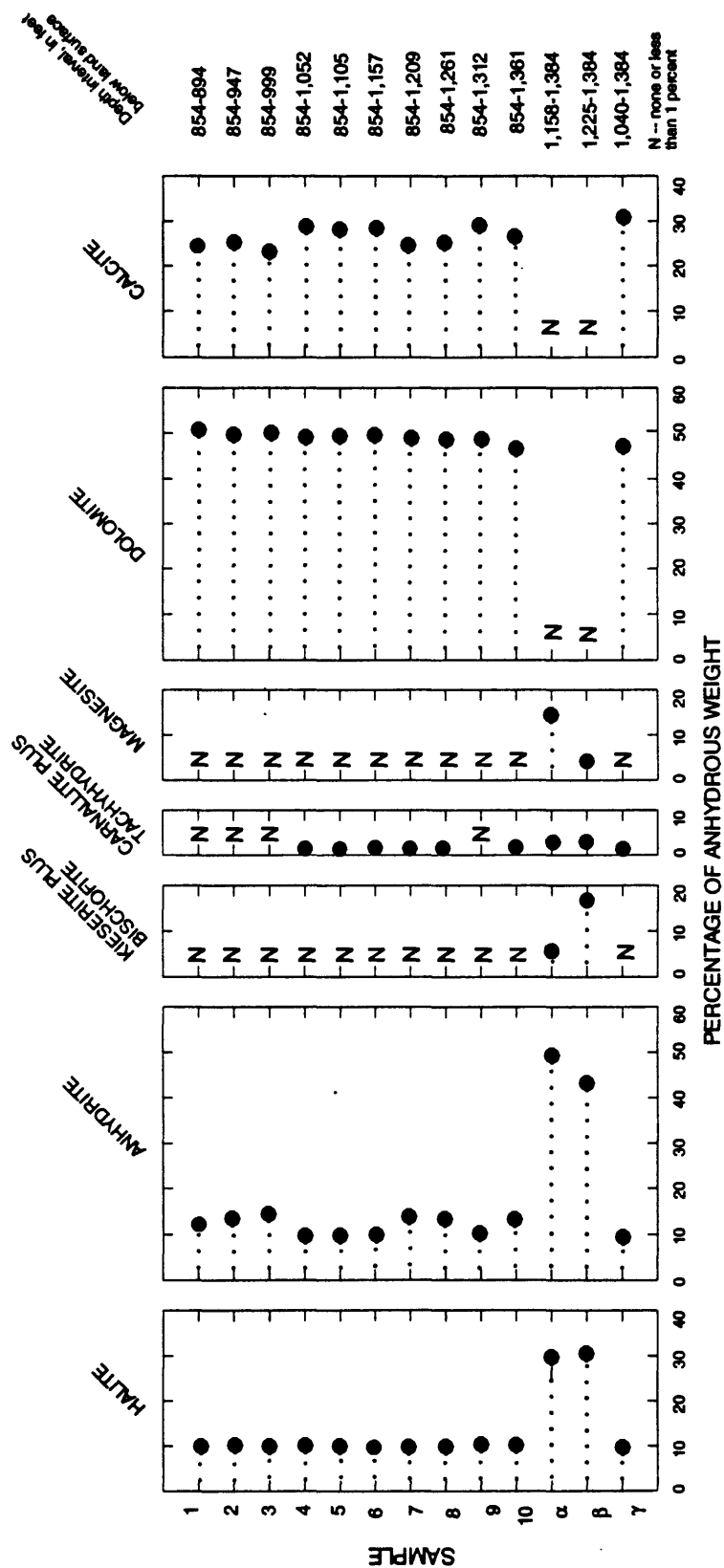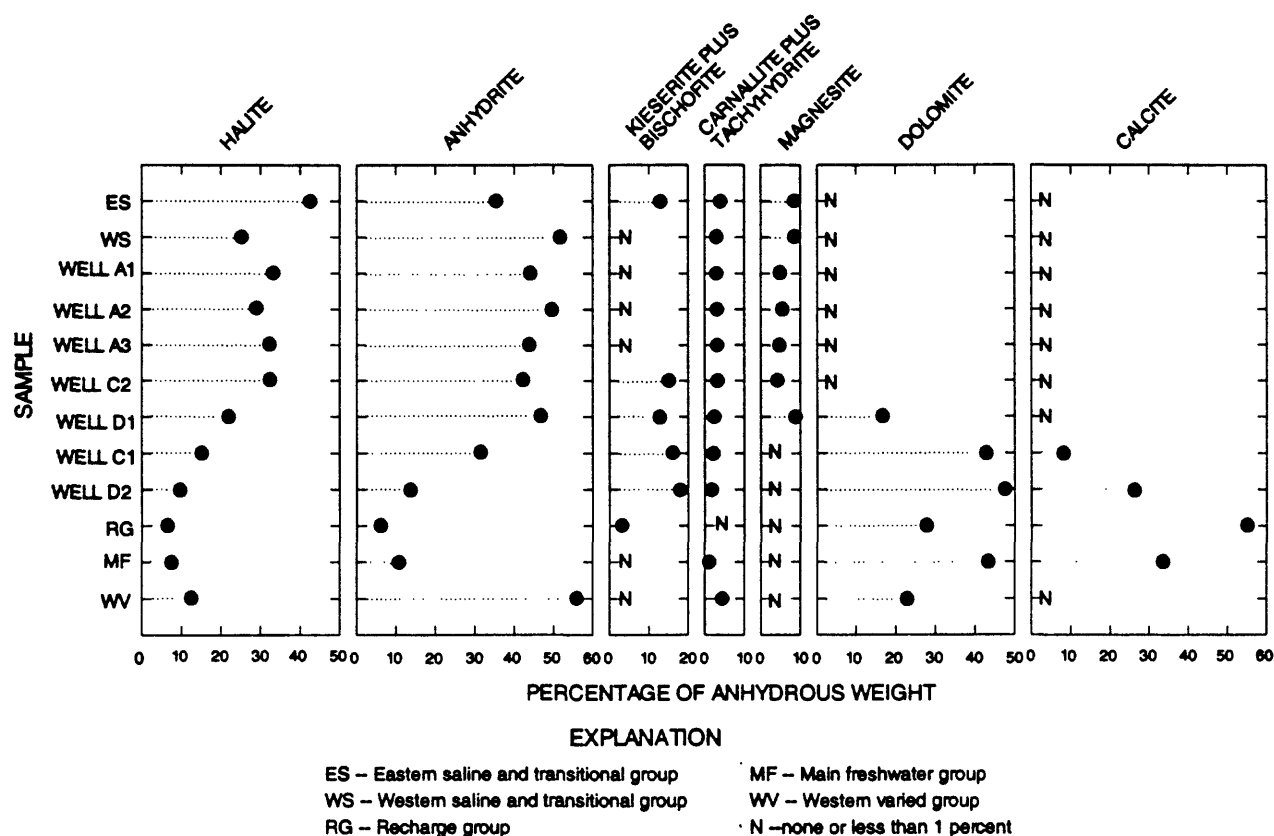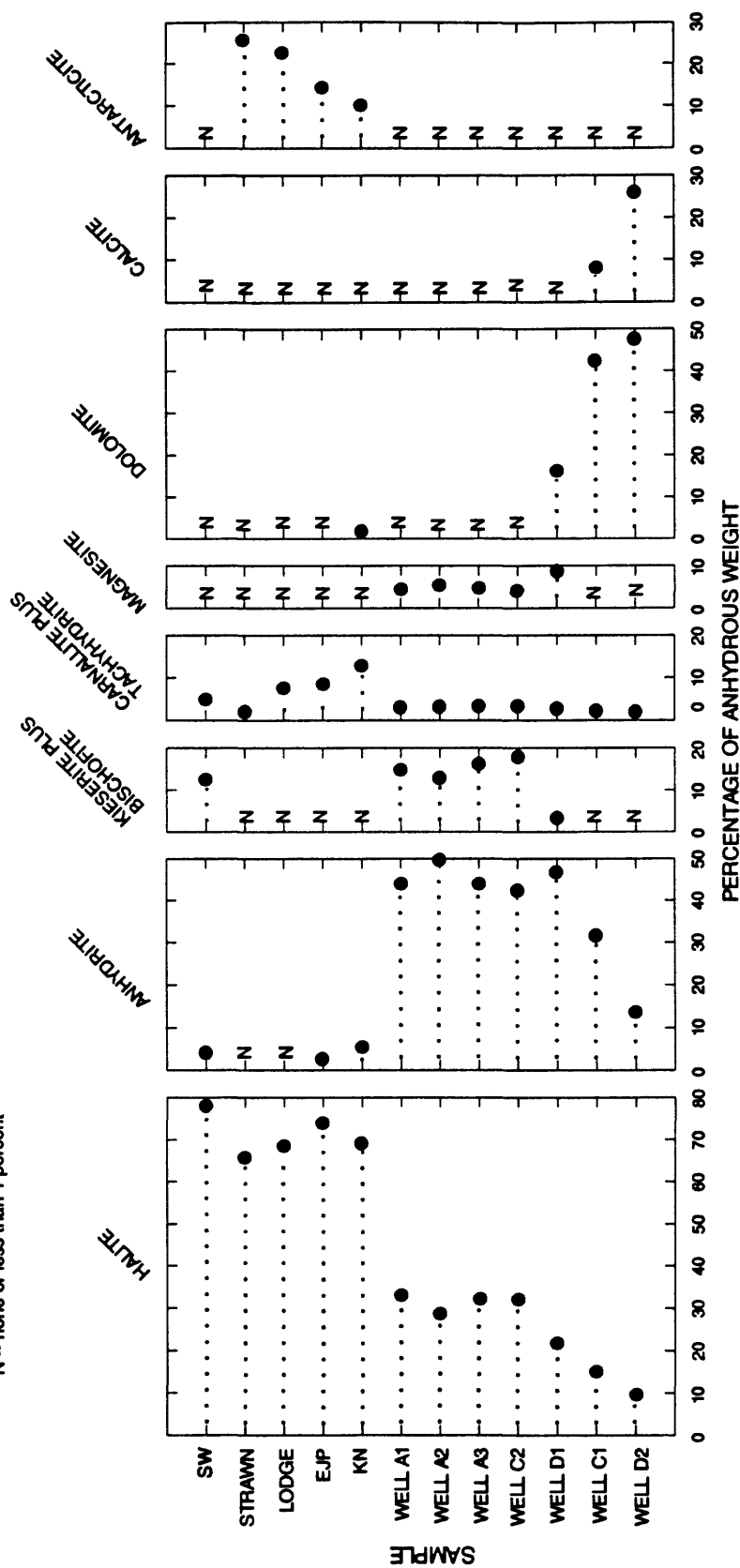
Figure 46. Percentage of anhydrous weight of the salt-normative assemblage of minerals computed for water samples from monitor wells A1, A2, A3, C1, C2, D1, and D2, San Antonio, Texas, April 17, 1987, mean seawater concentrations, and water samples from selected Edwards aquifer oil or gas wells.

amounts of anhydrite and much smaller amounts of halite than the seawater assemblage. The salt-norm assemblages for four oil- or gas-well samples (fig. 45) of Land and Prezbindowski (1981) contain no kieserite plus bischofite, unlike the seawater assemblage, and contain greater amounts of carnallite plus tachyhydrite and antarcticite than the seawater assemblage.

The analyses were used to compute a salt-normative mineral assemblage for comparing the chemistry of the samples. The sample from the deepest Edwards aquifer saline-water-zone well and the sample with the greatest dissolved-solids concentration is Strawn 1. In this assemblage, anhydrite is not present and the halite component is similar to that of the seawater norm. The lack of anhydrite might be caused by a component of hypersaline brine, which is depleted in sulfate. The addition of sulfate had little to no effect on the salt-norm assemblage. Kieserite and bischofite are absent from all the oil- or gas-well sample assemblages, which could be caused by the presence of redissolved halite or calcium sulfate in the brine. The lack of kieserite in the Strawn 1 assemblage probably is not caused by solution of calcium sulfate because of the near absence of anhydrite. Some solution of halite probably occurred to form this brine.

The salt-norm assemblage computed for the adjusted Lodge 2 analysis was not different from the unadjusted assemblage. For the samples with slightly smaller concentrations, such as Knobloch 1 and E.J. Pruitt 36, the addition of sulfate added sufficient anhydrite to make the computed salt-norm assemblage similar to the seawater assemblage. The added sulfate also caused a redistribution of small proportions of the calcium and chloride bearing minerals in the computed salt-norm assemblages. The two most diagnostic minerals or pairs of minerals in the assemblages for the oil- or gas-well samples are antarcticite $(CaCl_2 \cdot 6H_2O)$ and carnallite plus tachyhydrite (fig. 46). The addition of sulfate to the E.J. Pruitt 36 analysis substantially changed the resulting salt-norm assemblage. Without sulfate, the salt-norm assemblage contained more antarcticite than carnallite plus tachyhydrite. The added sulfate reversed the ranking of antarcticite to carnallite plus tachyhydrite in the computed salt-norm assemblage.

The large antarcticite norm relative to the seawater norm indicates that a component of the oil- or gas-well sample is made up of diagenetic brine, as hypothesized by Land and Prezbindowski (1981). The proportions of antarcticite to carnallite plus tachyhy-

drite can help distinguish diagenetic brine resulting from interaction with silicate minerals or carbonate minerals. Bodine and Jones (1986) concluded that when large amounts of antarcticite and carnallite plus tachyhydrite are present in an assemblage, and the antarcticite exceeds the carnallite plus tachyhydrite weight percent, then silicate dominated diagenesis has been involved in producing the water chemistry of the brine. The Strawn 1 and Lodge 2 assemblages and, to a lesser extent, the E.J. Pruitt 36 assemblage, contain more antarcticite than carnallite plus tachyhydrite. Therefore, there could have been silicate dominated diagenesis involved in the origin of the water chemistry at these wells.

If carnallite plus tachyhydrite exceeded antarcticite, such as in the Knobloch 1 assemblage, diagenesis dominated by carbonate minerals was involved in producing the water chemistry. If no sulfate had been added to this analysis, the amount of antarcticite would be about 5 percent greater, thus making it similar to the salt assemblages from water samples taken farther downdip. The effect that sulfate has on distributing the proportions of salts for this well sample strongly supports the assumed sulfate concentration.

A pattern is not apparent in these assemblages. The deepest water, represented by the Strawn 1 sample, might result from the mixing of marine-like connate brine with brine from another section of the Gulf Coast sediments where diagenesis of silicate minerals was present. The effect of silicate dominated diagenesis is to decrease the ratio of magnesium to calcium in the pore fluid brine by exchanging magnesium and sodium in solution for calcium in solids (Bodine and Jones, 1986). The ratio of magnesium to calcium in the brine is indicated by the relative amounts of antarcticite and tachyhydrite plus carnallite in the salt-norm assemblage.

The importance of the silicate dominated diagenesis becomes less in the water 6 mi or more updip, as represented by the Lodge 2 sample, and is even less important farther updip, as represented by the E.J. Pruitt 36 sample. At the oil or gas field closest to the downdip limit of freshwater of the Edwards aquifer, represented by the Knobloch 1 well sample, the silicate diagenetic water type apparently is insignificant, but water that is dominated by carbonate mineral diagenesis could have mixed with marine-like water in this part of the aquifer. This oil field is about 12 mi downdip from the downdip limit of freshwater.

The proportion of a possible marine-like water type in the water samples becomes less with proximity to the freshwater part of the aquifer. The carbonate dominated diagenetic water type becomes less at the zone represented by four saline-water-zone monitor-well assemblages: A1, A2, A3, and C2. This progression of water types does not indicate substantial movement of water updip toward the freshwater zone. The progression does indicate that separate masses of water, possibly compartmentalized by fault barriers, are affected by mixing with brine introduced into the Edwards aquifer. Originally, the rocks that today compose the Edwards aquifer contained marine water of fairly uniform composition. Over geologic time, part of the area underlain by the Edwards aquifer has been variably elevated above sea level and then subsided below sea level in part, and several distinct flow regimes were established.

The freshwater zone probably developed as suggested by Mench-Ellis (1985) and discussed in the section "Previous Investigations." Because the water flow in the freshwater zone is partly or completely controlled by barrier faults, the flow in the saline-water zone also could be controlled by barrier faults. Most oil fields are located where substantial offset of up-to-the-coast faults results in stagnation zones that prevent oil and other hydrocarbons from moving farther updip (Rose, 1972). Therefore, control of fluid flow in the saline-water zone by faults is likely.

The oil or gas fields in the Edwards aquifer saline-water zone are located primarily in the nearly stagnant zones near faults at the top of the Edwards aquifer, and samples taken from these wells could be biased toward the type of water in these stagnant zones. If bias toward one type of water is present, the actual nature of fluid in the zones between oil and gas fields is unknown. Land and Prezbindowski (1981) concluded that upfault movement of water is from older formations into the Edwards aquifer. It is possible that the withdrawal of fluids (hydrocarbons and saline water) in the oil and gas fields causes more upfault water flow and, thus, these oil and gas zones are potentially biased toward water chemistry influenced by upfault movement.

Land and Prezbindowski (1981) proposed that a mixture of Edwards saline ground water and diagenetic water from outside the aquifer moves updip toward the freshwater zone. If this water does not enter the freshwater zone, it must discharge elsewhere. Information on amount of saline water that moved into the

freshwater zone or on saline-water springs or seeps of the Edwards aquifer is limited, and Land and Prezbindowski (1981) did not present any data on locations or volumes of natural saline-water discharge.

This preliminary comparison of the water chemistry differentiation of the saline-water zone indicates the possibility that upfault movement of water could occur in the saline-water zone and by inference, the freshwater zone, unless prevented by mineral precipitation or other factors. The best examples of upfault water movement in the freshwater zone are the major springs—San Antonio, San Pedro, Comal, and San Marcos—all of which discharge water diverted along major faults from the deeply buried Edwards aquifer. It is possible for saline water from beneath the Edwards to flow up along faults and into the Edwards aquifer.

Assuming that upfault water flow from formations beneath the Edwards aquifer in the freshwater zone does occur, it might not have been identified because: (1) the volume of flow into the freshwater zone is too small to substantially affect the water chemistry; (2) the differences in the water chemistry in the freshwater zone, for example, the five water-quality groups of Pearson and Rettman (1976), are too broadly defined; or (3) the water from formations below the Edwards aquifer in the freshwater zone is similar to the freshwater in the Edwards aquifer. The water from formations underlying at least part of the freshwater zone of the Edwards aquifer is slightly saline to saline (G.L. Duffin, Texas Water Development Board, written commun., 1975). If upfault water movement from the underlying formations into the freshwater zone of the Edwards aquifer does occur, the volumes probably are small.

SUMMARY AND CONCLUSIONS

The Edwards aquifer can be separated laterally and vertically into a freshwater zone and a saline-water zone. The freshwater zone is the principal source of water for San Antonio and a number of small communities in the San Antonio area. The saline-water zone consists of slightly saline water to brine. The downdip limit of freshwater between these two zones is sometimes called the "bad-water" line. One of the questions raised when considering the maximum use of the freshwater part of the aquifer is whether saline water will intrude into the freshwater zone as a result of large water-level decline. This question is of greatest concern in those areas where water-supply wells are

located near the downdip limit of freshwater between the two aquifer zones.

Four area agencies have cooperatively undertaken a study to define the downdip limit of freshwater more accurately and to monitor water quality and water levels in the aquifer at the downdip limit of freshwater. The U.S. Geological Survey is responsible for collecting and interpreting the data from the test-drilling and monitoring programs. The first of the transects across the downdip limit of freshwater was drilled in 1985 and 1986. Geologic, geophysical, hydrologic, and water-quality data were collected during the drilling phase. Data were collected for 8 months from the 7 monitor wells completed along this transect, which extends southward from the Artesia well field.

Results of studies about the various aspects of the water quality at the downdip limit of freshwater, the hydrology in the aquifer affecting the downdip limit of freshwater or saline-water zone, and the water chemistry within the saline-water zone have been published by the U.S. Geological Survey and other agencies. These reports were reviewed to summarize the findings of the authors and synthesize a conceptual framework to interpret the new data. Results of these studies indicate that the nature and origin of the saline-water zone of the Edwards aquifer must be understood in order to understand the extent to which and how saline-water intrusion might occur.

The primary concept of the development of a large secondary porosity in the freshwater zone is that the porosity was caused by dedolomitization. Dedolomitization occurred when recharge water dissolved calcite near the recharge area and moved down into the aquifer through fractures or bedding planes that allowed for movement toward a distant discharge point. Along this flowpath, the water contacted dolomite and anhydrite. As anhydrite dissolved in the freshwater, calcite became supersaturated and precipitated into dead-end pores and fractures. The precipitation of calcite caused dolomite to become undersaturated in the water and it dissolved from the rocks.

The diagenetic processes that occurred early in the geologic history of the Edwards aquifer left a large amount of dolomite in the aquifer matrix rocks. As the reactions proceeded, much of the dolomitic rock was converted to pure calcium carbonate with a large secondary porosity and to dense matrix rocks with less porosity. The secondary porosity then led to the recharge of greater quantities of freshwater to dissolve more rock in the aquifer. Because dolomite in the rocks

of the Edwards aquifer is restricted primarily to several strata within certain members of the Edwards aquifer formations, the flow that developed was restricted vertically.

Neogene fault movement developed barrier faults where the most permeable layers of rock were juxtaposed against nearly impermeable rocks of the Edwards aquifer or formations above or below the Edwards aquifer. The flow of water was diverted parallel to these faults where freshwater had extended some distance into the rocks and where saline water was still extant. The faulting also opened a limited number of vertical pathways for ground water to discharge. The most permeable of these then developed into major discharge pathways for freshwater and possibly for saline water. The freshwater springs are well known locally, but little is known about the location and discharge of saline water. The saline water moves very slowly and probably discharges in small amounts.

The data collected from the test holes during drilling and from the monitor wells July 1986–April 1987 fit into the conceptual framework of the aquifer. Water flow is parallel to faults, which are generally parallel or subparallel to the downdip limit of freshwater. The water flow near the downdip limit of freshwater is also parallel or subparallel to the downdip limit of freshwater. This is indicated by the difference in water-quality types and salinity in the test holes and in the completed monitor wells.

Specific conclusions of the study are as follows:

1. The mineralogy of the aquifer at the test sites varies across the downdip limit of freshwater as described by previous investigators. There is a discontinuous progression of dedolomitization from unaltered, but water-producing, dolomitic rocks of the saline-water-zone test hole (site A) for the full aquifer thickness, to the highly solution eroded and fractured calcitic rocks that are present at the freshwater-zone test hole (site D) from the bottom of the Georgetown Formation to the midpoint of the Kainer Formation.

2. The shape of the downdip limit of freshwater is complex. Water salinity is stratified in the aquifer at the three test sites. At each test-well site, the most saline water occurs near the bottom of the aquifer; freshwater overlies saline water in the transition-zone site (C) and the freshwater-zone site (D).

3. Water-quality type also varies laterally in the Edwards aquifer in the study area. Variations in water-quality type were indicated by differences in dissolved-solids concentrations and relative proportions of dissolved ions in samples taken from test holes during drilling and interval tests and in the completed monitor wells. The water-quality type from the shallow freshwater site monitor well (D2) is a calcium magnesium bicarbonate water with small dissolved-solids concentration. This water type is similar to the main freshwater water-quality group of well samples from the Edwards aquifer freshwater zone identified by previous investigators. The water-quality type from the four monitor wells that contain the most saline water (A1, A2, A3, and C2) is a calcium sodium sulfate chloride type similar to a saline water-quality group also defined by previous investigators. The water-quality type of the deep freshwater-zone well (D1) is similar to samples from wells A1, A2, A3, and C2, but has a much smaller dissolved-solids concentration. Well C1, the shallow transition-zone monitor well, produces freshwater from a cave near the top of the Edwards aquifer; however, the water-quality type is dissimilar to all other monitor wells and water-quality groups as defined by previous investigators.

4. The saline water in Edwards aquifer oil or gas fields downdip of the freshwater zone may comprise distinct masses of marine-like connate water mixed with varying amounts of brine introduced into the Edwards aquifer from poorly characterized sources.

5. The chemistry of saline water near the downdip limit of freshwater is different in salinity and type from that much farther downdip. The water chemistry of saline water near the downdip limit of freshwater does not appear to have been affected by mixing with brine from other sources.

6. During July 1986-April 1987, the quality of water in samples from the seven monitor wells did not vary substantially. The water quality in all wells remained constant, even during a March 1987 aquifer test conducted at the Artesia water-supply wells adjacent to the freshwater (D) site.

7. During July 1986-April 1987, the water levels in the seven monitor wells fluctuated several feet over the period of a day in response to hourly variations in withdrawal rates from the freshwater zone. Water levels in all wells decreased slowly by several feet during July-August 1986 because of increased withdrawals and no rain during the period. Several times after August 1986, the water levels recovered a few feet to more than 10 ft in response to recharge from rain storms and to decreases in withdrawals related to rain or the expectation of rain. In general, water levels in the seven wells were similar, usually less than 3 ft difference among the wells, and the wells responded almost in unison, except that the water levels in the saline-water-zone wells—A1, A2, A3, and C2—responded several minutes to less than 2 hours later than water levels in wells C1, D1, and D2. The water-level fluctuations in wells A1, A2, A3, and C2 were smaller than the water-level fluctuations in wells C1, D1, and D2. The head in the saline-water zone near the downdip limit of freshwater varies along with the water levels in the freshwater zone.

8. Periodic samples from monitor-wells C1 and D1 are the most likely to indicate significant changes in water quality caused by saline-water intrusion if it occurs. Changes in the molar ratios of dissolved sulfate to chloride and dissolved magnesium to calcium and the relation between the two ratios might be more accurate and reliable indicators of saline-water intrusion than changes in dissolved-solids concentration or individual dissolved-ion concentrations.

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Table 5. Summary statistics of the water quality from completed monitor wells, San Antonio, Texas, July 1986-April 1987

[μS/cm, microsiemens per centimeter at 25 degrees Celsius; °C, degrees Celsius; mg/L, milligrams per liter; --, no data]

Property or constituent	Number of samples	Mean	Median	Standard deviation	Standard error	Minimum	Maximum	Coefficient of variation[1]
				Well A1				
Specific conductance (μS/cm)	10	5,780	5,780	205	65	5,430	6,160	0.035
pH (units)	10	--	6.60	--	--	6.30	6.90	--
Temperature (°C)	10	32.0	32.0	.4	.1	32.0	33.0	.012
Alkalinity (mg/L as CaCO₃)	10	242	242	7	2	235	258	.029
Calcium (mg/L)	10	510	510	31	10	470	560	.061
Magnesium (mg/L)	10	190	190	6	2	180	200	.031
Sodium (mg/L)	3	480	480	10	6	470	490	--
Potassium (mg/L)	3	28	29	1.2	1	27	29	--
Sulfate (mg/L)	10	1,800	1,900	97	31	1,600	1,900	.053
Chloride (mg/L)	10	970	960	64	20	850	1,100	.066
Fluoride (mg/L)	3	2.9	2.9	<.1	<.1	2.8	2.9	--
Silica (mg/L)	3	20	20	.0	.0	20	20	--
Dissolved solids (mg/L)	3	4,200	4,200	100	58	4,100	4,300	--
				Well A2				
Specific conductance (μS/cm)	11	4,710	4,710	112	34	4,540	4,860	0.024
pH (units)	11	--	7.00	--	--	6.00	7.00	--
Temperature (°C)	11	31.2	31.5	.3	.1	30.5	31.5	.0096
Alkalinity (mg/L as CaCO₃)	11	224	223	6	2	217	237	.027
Calcium (mg/L)	11	430	400	69	21	380	610	.16
Magnesium (mg/L)	11	160	150	21	6	150	220	.13
Sodium (mg/L)	4	380	380	8	4	370	390	--
Potassium (mg/L)	5	19	23	9	4	2.2	24	--
Sulfate (mg/L)	10	1,400	1,400	53	17	1,400	1,500	.036
Chloride (mg/L)	10	750	740	21	7	730	790	.028
Fluoride (mg/L)	4	2.8	2.8	.1	<.1	2.7	2.9	--
Silica (mg/L)	4	15	19	8.0	4.0	2.7	19	--
Dissolved solids (mg/L)	4	3,400	3,400	58	29	3,300	3,400	--

Footnote at end of table.

Table 5   89

Table 5. Summary statistics of the water quality from completed monitor wells, San Antonio, Texas, July 1986-April 1987--Continued

Property or constituent	Number of samples	Mean	Median	Standard deviation	Standard error	Minimum	Maximum	Coefficient of variation[1]
				Well A3				
Specific conductance (µS/cm)	11	5,870	5,900	205	62	5,440	6,190	0.035
pH (units)	11	--	7.00	--	--	6.00	7.00	--
Temperature (°C)	11	30.0	31.0	1.1	.3	28.0	32.0	.036
Alkalinity (mg/L as CaCO₃)	11	244	241	9	3	235	266	.037
Calcium (mg/L)	11	510	500	50	15	420	590	.098
Magnesium (mg/L)	11	200	210	13	4	190	230	.065
Sodium (mg/L)	4	500	500	13	6	490	520	--
Potassium (mg/L)	4	30	30	1	<1	30	30	--
Sulfate (mg/L)	11	1,900	1,900	50	15	1,800	1,900	.026
Chloride (mg/L)	11	1,000	1,000	60	18	940	1,100	.060
Fluoride (mg/L)	4	2.9	2.8	.1	<.1	2.8	3.1	--
Silica (mg/L)	4	15	19	8.7	4.4	2.3	21	--
Dissolved solids (mg/L)	4	4,200	4,200	130	63	4,100	4,400	--
				Well C1				
Specific conductance (µS/cm)	11	670	660	45	14	630	770	0.067
pH (units)	11	--	6.90	--	--	6.30	7.40	--
Temperature (°C)	10	28.0	28.0	.4	.1	28.0	29.0	.014
Alkalinity (mg/L as CaCO₃)	11	199	198	9	3	178	212	.045
Calcium (mg/L)	11	77	79	7	2	65	89	.091
Magnesium (mg/L)	11	22	22	2	<1	20	27	.091
Sodium (mg/L)	4	27	27	6	3	21	33	--
Potassium (mg/L)	4	2.7	2.7	.5	.3	2.2	3.2	--
Sulfate (mg/L)	11	99	97	18	5	72	120	.18
Chloride (mg/L)	11	46	46	5	2	37	52	.11
Fluoride (mg/L)	4	.8	.8	.1	<.1	.6	.9	--
Silica (mg/L)	4	13	13	.5	.2	13	14	--
Dissolved solids (mg/L)	4	410	410	44	22	370	460	--

Footnote at end of table.

Table 5. Summary statistics of the water quality from completed monitor wells, San Antonio, Texas, July 1986-April 1987--Continued

Property or constituent	Number of samples	Mean	Median	Standard deviation	Standard error	Minimum	Maximum	Coefficient of variation[1]
				Well C2				
Specific conductance (μS/cm)	11	6,000	6,000	192	58	5,700	6,300	0.032
pH (units)	11	--	6.60	--	--	6.00	6.90	--
Temperature (°C)	11	29.0	30.0	.9	.3	30.0	30.0	.031
Alkalinity (mg/L as CaCO₃)	11	237	236	11	3	219	259	.046
Calcium (mg/L)	11	500	500	38	12	440	590	.076
Magnesium (mg/L)	11	220	220	12	4	210	250	.054
Sodium (mg/L)	4	600	500	34	17	500	600	--
Potassium (mg/L)	4	30	30	1	<1	29	31	--
Sulfate (mg/L)	11	2,000	2,000	120	36	2,000	2,000	.063
Chloride (mg/L)	11	1,100	1,100	69	21	900	1,100	.063
Fluoride (mg/L)	4	2.8	2.8	.2	<.1	2.6	2.9	--
Silica (mg/L)	4	19	19	.0	.0	19	19	--
Dissolved solids (mg/L)	4	4,500	4,500	82	41	4,400	4,600	--
				Well D1				
Specific conductance (μS/cm)	10	1,020	1,050	101	32	860	1,180	0.099
pH (units)	10	--	7.10	--	--	6.80	7.30	--
Temperature (°C)	10	27.0	26.0	.4	.1	26.0	27.0	.0038
Alkalinity (mg/L as CaCO₃)	10	200	200	4	1	200	210	.020
Calcium (mg/L)	10	100	110	10	3	88	120	.10
Magnesium (mg/L)	10	40	40	2	<1	30	40	.054
Sodium (mg/L)	4	56	56	3	2	52	59	--
Potassium (mg/L)	4	4.0	4.0	.2	<.1	3.9	4.3	--
Sulfate (mg/L)	10	200	210	23	7	150	230	.115
Chloride (mg/L)	10	110	110	11	4	79	120	.10
Fluoride (mg/L)	4	.9	.9	.2	<.1	.8	1.1	--
Silica (mg/L)	4	13	13	.0	.0	13	13	--
Dissolved solids (mg/L)	4	670	660	31	16	640	710	--

Footnote at end of table.

Table 5   91

Table 5. Summary statistics of the water quality from completed monitor wells, San Antonio, Texas, July 1986-April 1987--Continued

Property or constituent	Number of samples	Mean	Median	Standard deviation	Standard error	Minimum	Maximum	Coefficient of variation[1]
				Well D2				
Specific conductance (μS/cm)	10	458	467	24	8	410	484	0.052
pH (units)	10	--	6.90	--	--	6.50	7.20	--
Temperature (°C)	10	27.0	26.0	.4	.1	26.0	27.0	.015
Alkalinity (mg/L as CaCO₃)	10	197	197	3	1	190	201	.015
Calcium (mg/L)	10	60	70	8	2	40	70	.13
Magnesium (mg/L)	10	16	17	1	<1	14	18	.062
Sodium (mg/L)	4	10	10	.5	.3	9.7	11	--
Potassium (mg/L)	5	1.7	1.3	1.0	.4	1.1	3.4	--
Sulfate (mg/L)	9	24	24	1	<1	22	26	.042
Chloride (mg/L)	9	20	20	2	<1	18	23	.10
Fluoride (mg/L)	4	.3	.3	<.1	<.1	.3	.4	--
Silica (mg/L)	4	12	12	<1.0	<1.0	12	13	--
Dissolved solids (mg/L)	4	270	270	5	3	270	280	--

[1] Coefficient of variation is defined as the ratio of the standard deviation to the mean. Only values for maximum number of samples are shown.